FUNDAMENTALS OF SEDIMENTOLOGY

Sreepat Jain

CRC Press
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Fundamentals of Sedimentology

This new textbook is a modern look at key concepts of sedimentology. With lavish, colorful, and abundant illustrations and easy-to-understand explanations, the book focuses on the concepts required to understand physical, chemical, and biological characteristics of sedimentary rocks and the processes involved in their formation. This includes transportation, deposition, and transformation of sediments. It also emphasizes how the understanding of sedimentary rocks can be used to interpret all continental, marginal marine, and deep-water oceanic environments. Written with undergraduate-level students in mind, it serves as a primary textbook for the new generation of students.

Features

- Fully up-to-date coverage, using the latest studies in the field of sedimentology.
- Many colorful illustrations to facilitate the understanding of key concepts.
- Explanations that are jargon-free and easy to understand for the undergraduate-level reader.
- Examples to interpret ancient environmental conditions in sediment source areas and depositional sites.

Written by an experienced researcher and academic who has taught the course at different universities and countries for over 20 years, *Fundamentals of Sedimentology* is an excellent resource for upper-level undergraduate and graduate students studying Geology, Geomorphology, Physical Geology, and Geography, and it serves as a great reference for entry-level researchers who work in the same fields.

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Designed cover image: The author with the backdrop of sedimentary Quartzite (consisting mainly of quartz; >90%) and formed by recrystallization of sandstone (that had undergone low-grade metamorphism showing various deformational features; prominently joints, vertical and closely spaced) belonging to the Aravali Range (Alwar Group, Delhi Supergroup; 900-1600 Ma) at Galtaji (Jaipur; Rajasthan, western India; 26.918139, 75.858240).

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Dedication

I dedicate this book to my son Parth Jain and my wife Archana Mamgain.

Contents

SECTION I Basic Concepts

SECTION IVa Depositional Systems: Continental Environments

Preface

Sedimentology is a vast interdisciplinary subject; hence, explaining all the topics in a book of manageable page numbers is almost impossible. This book is thus designed as a brief summary of major topics in sedimentology, explained through well-labeled line diagrams so that students can better understand and appreciate the distinctive characteristics of sedimentary rocks produced via different processes. This book is specifically designed for undergraduate students and as a tool for those instructors who are involved in teaching them.

The deposition of sediments produces characteristic sedimentary structures, textures, and bedding features that are used as robust proxies for interpreting and inferring sedimentary depositional environments. It is the characteristics of these environments that form the heart of the book. Additionally, through a series of well-labeled line diagrams, the characteristics of each environment and subenvironment are elaborated, including their sedimentary structures, associated facies, faunal assemblages, and depositional settings.

The book does NOT provide any case studies of individual ancient sedimentary environments as these are well-documented as the concluding section in almost all sedimentary books. Moreover, no two such examples are similar and opinions vary on their mode of origin as well, depending on the type of tools used for inferring the depositional environment! Hence, the emphasis of the book is very traditional, to provide the tools that could be used by a student and/or a budding sedimentologist for paleoenvironmental reconstructions.

Acknowledgments

I am grateful to my wife Archana Mamgain and my son Parth Jain for their constant support and encouragement, to Professor Mariusz A. Salamon (University of Silesia, Sosnowiec, Poland) for providing computer software that enabled the drawing of figures used in this book, and above all to Ethiopia, the country that provided a fertile ground to pen my thoughts into a book. I would also like to thank Irma Shagla Britton (Senior Editor) and Chelsea Reeves (Senior Editorial Assistant), CRC Press (Florida, USA) for their patience and continued support.

About the Author

I started collecting “stones” when I was 5 years old, and I am still enamored by wondrous nature and its bounty, like a boy smitten by colorful candies in a candy store!

I am a Professor working in the Department of Applied Geology, Adama Science and Technology University, Adama, Ethiopia, with 22 years of teaching and over two and a half decades of research experience. I have two doctorates, one from India and the other from the United States. I was awarded the TA Excellence in Teaching in the United States. In addition, I have also received the prestigious Prof. S. K. Singh Memorial Gold Medal from the Paleontological Society of India for best research paper. I am also a recipient of the Indo-German DST-DAAD Fellowship. I have traveled extensively in the United States and Europe in the pursuit of academic scholarship.

My research work is multidisciplinary, covering various aspects of paleontology (ammonites, benthic and planktic foraminifera, calcareous nannofossils, palynomorphs, echinoids, crinoids, trace fossils, plant fossils, and plant-animal interactions), sedimentology, sequence stratigraphy, and stable isotopes, stretching from the Triassic period to recent times, with an emphasis on the Jurassic-Cretaceous.

I have published several research papers in national and international peer-reviewed journals and authored three books with Springer: *Fundamentals of Physical Geology* (2014), *Fundamentals of Invertebrate Paleontology – Macrofossils* (2017), and *Fundamentals of Palaeontology – Microfossils* (2020). All three books received excellent international reviews and have been recommended for undergraduate Earth Science semester courses in several international universities.

My motto has been “I love what I do and do what I love”!

Sreepat Jain

Structure of the Book

The book is divided into four broad sections: Section I deals with basic concepts (Chapters 1–3), Section II details sedimentary structures (Chapter 4), Section III provides the composition and classification of sedimentary rocks (Chapters 5–7), and Section IV outlines and describes various depositional systems (Chapters 8–21).

Section I on basic concepts gives an overview of weathering and soils (Chapter 1), properties of sediments (Chapter 2), and the mechanisms of sediment transport (Chapter 3). Weathering (physical, chemical, and biological), along with erosion, disintegrates older rocks to produce rock fragments, resistant minerals, and dissolved chemical substances as a final byproduct. Some of these products accumulate in situ to form soils that are preserved in the geologic record, but most are removed by erosion and transported to depositional sites through diverse processes. Thus, the origin of sedimentary rock involves (a) weathering of pre-existing rocks to produce materials that form them, (b) the weathered debris and soluble constituents (the end products of weathering), through erosion and transport are moved to depositional basins, to be deposited in continental (terrigenous), marginal marine or marine environments, and (c) the deposited sediments are diagenetically altered during their burial to form lithified sedimentary rocks.

Section II deals with sedimentary structures and their environment of deposition (Chapter 4). Sedimentary structures, profusely abundant in siliciclastic rocks (and relatively less in non-siliciclastic sedimentary rocks), faithfully reflect environmental conditions during the time of their deposition, enabling paleoenvironmental reconstructions at local (section where studied), regional (basinal), and global (at large) level. They also provide valuable information on the mechanisms of sediment transport, flow directions (paleocurrent), relative water depth, and current velocity, among others. They have also been used successfully to infer whether the sedimentary sequence was in the correct depositional stratigraphic order or was overturned by tectonic forces, using the concept of identifying the tops and bottoms of beds. In this section, the emphasis is on illustrating the main types of sedimentary structures, showing their major characteristics and how to use them for interpreting lithofacies and their features vis-à-vis their depositional environment.

Section III deals with the composition and classification of sedimentary rocks (Chapters 5–7). This section details major compositional types, depositional environments, origin, and diagenesis of sedimentary rocks, such as siliciclastic sedimentary rocks (Chapter 5: sandstones, conglomerates, and shales), carbonate rocks (Chapter 6: limestones and dolomites), and chemical, biochemical, and carbonaceous sedimentary rocks (Chapter 7: evaporites, cherts, iron-rich sedimentary rocks, phosphorites, oil shales and coals). The emphasis of this section is on detailing the characteristics and classification of each rock type and elaborating on their environments of deposition.

Section IV on depositional systems (Chapters 8–21) deals with depositional environments: continental (Chapters 8–12), marginal marine (Chapters 13–17), siliciclastic marine and pelagic (Chapters 18–19), and carbonate (chapters 20–21). The continental depositional environment includes fluvial, river, lacustrine, eolian desert, and glacial systems (Chapters 8–12, respectively). The marginal marine depositional environment includes deltaic, beach and barrier-island, estuarine, lagoonal, and tidal-flat systems (Chapters 13–17, respectively). The siliciclastic marine and pelagic depositional environment includes shelf and oceanic (deep-water) environments (Chapters 18–19, respectively). Special mention is given to carbonate and carbonate shelf environments due to their immense economic and environmental importance (Chapters 20–21, respectively). Herein, each depositional environment details a note on its current classification/subdivisions, major characteristic features, associated sedimentary structures, and its environment of deposition.

Section I

Basic Concepts

1 Weathering and Soils

1.1 INTRODUCTION

Weathering, through decomposition (chemical breakdown) and disintegration (physical breakdown), breaks rocks into smaller fractions that are in equilibrium with the prevailing environment (Bland and Rolls, 1998). Erosion, on the other hand, is the movement (either by moving water, wind, glaciers, and gravity) of particles away from their place of formation. Thus, the rocks are made of sediments that were weathered, eroded, transported, and deposited in sedimentary basins. Weathering also releases ion salts from rocks and minerals into the oceans and makes the ocean saline. Often weathering and erosion are used interchangeably but the fundamental processes behind them are quite different. Weathering is without movement and involves breaking down of the substrate by chemical or physical means, but in erosion, movement is the main driving force. Mass Wasting is another name for erosion which is the down slope movement of sediments affected by gravity (such as rock falls, slumps, and debris flows). However, it is erosion, if the rock particles are moved by a flowing agent, such as air, water, or ice.

Broadly, weathering is of three types: physical (mechanical), chemical, and biological (although this forms only a small fraction). Physical weathering causes reduction in size of the rock by breaking it into smaller fragments, but with no change in its chemical composition, whereas chemical weathering changes the composition of the parent rock through chemical reactions between minerals and chemical reagents, such as H_2O, CO_2, O_2, or organic acids, resulting in either the formation of new minerals and/or the dissolution of elements from minerals. Both physical and chemical weathering processes operate in tandem and often assists each other.

1.2 ROCK CYCLE

Rock cycle expresses the interrelationship between the three major groups of rocks, igneous, sedimentary, and metamorphic (Figure 1.1). Igneous rocks cool, crystallize and solidify from the hot molten lava and magma. Thereafter, they undergo weathering and erosion to form sediments that are then transported/deposited and lithified by compaction and cementation forming sedimentary rocks. These are then buried deep within the earth, and subjected to increased pressure and temperature, causing them to undergo metamorphism (metamorphic rocks). Increased burial and heating of metamorphic rocks results in their melting, forming magma. The magma eventually cools and crystallizes to form plutonic igneous rocks, or is erupted onto the earth's surface as lava, which cools and crystallizes to form volcanic igneous rocks. These Igneous rocks undergo weathering and erosion to form sediments, and thus, this cycle goes on and on (Figure 1.1).

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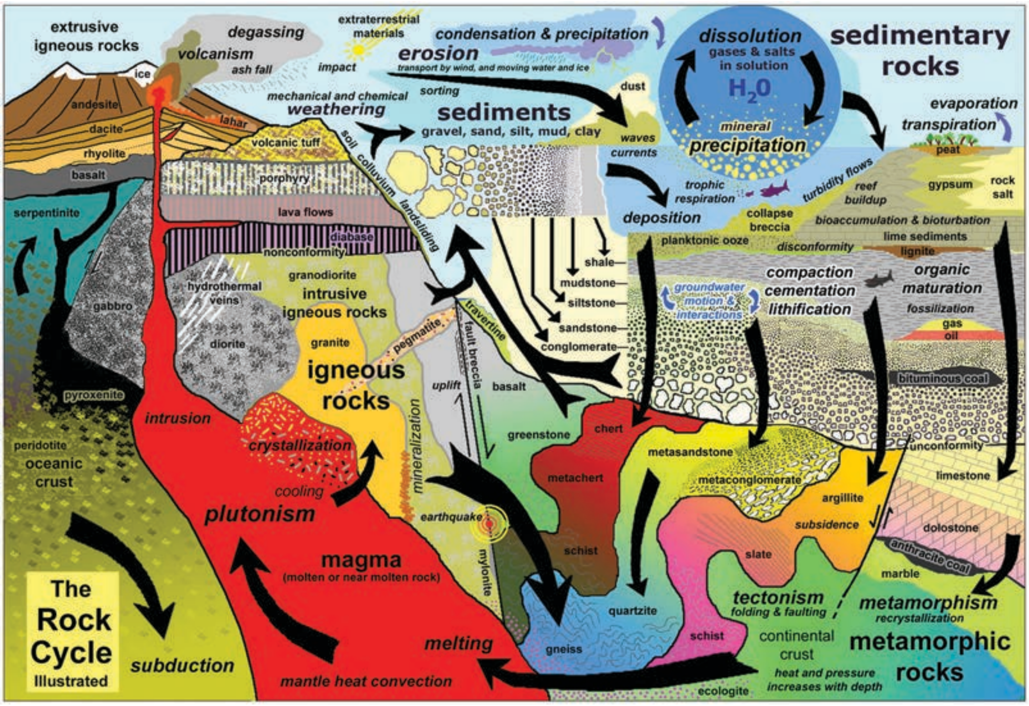

FIGURE 1.1 The rock cycle. (Reproduced with permission from Dr. Phil Stoffer, www.geologycafe.com.).

1.3 TYPES OF WEATHERING

Chemical, physical, and biological weathering work together to break down rocks and minerals into smaller fragments and more stable minerals that are in equilibrium with the prevailing environment.

1.3.1 Physical Weathering

Thornbury (1969) defined physical weathering as the breakdown of minerals or rock materials by entirely mechanical means with no change in the chemical composition of the weathered rock (or mineral). The increased strain (internal and external) ruptures the rock either due to abrasion, growth of salt crystals (crystallization), unloading (pressure-release), thermal insolation, or by several cycles of wetting and drying. Additionally, due to the breakage of the rock, more surface area is exposed for weathering. On a microscopic scale, the mineral grain boundaries provide potential areas of weakness within the rock, and thus future locations of weathering. Sedimentary rocks are often layered and those that are more massive have joints that open as the rock is exposed to erosion; physical weathering, with time, widens these ruptures and fractures.

1.3.1.1 Plants (Roots)

Root wedging occurs when the root of a plant begins to grow in a crack or pore within a rock (Figure 1.2). As the roots grow larger, they break the rock apart (Figure 1.2); plant roots can split even the hardest rocks and is commonly noted in city sidewalks and foundations, where roots push from underneath, thus, raising the concrete and subsequently breaking it. Additionally, some acid-producing microorganisms (such as fungi and lichens) that live on rocks dissolve nutrients (such as phosphorus or calcium) within rocks. Thus, these microorganisms also assist in the breakdown and weathering of rocks.

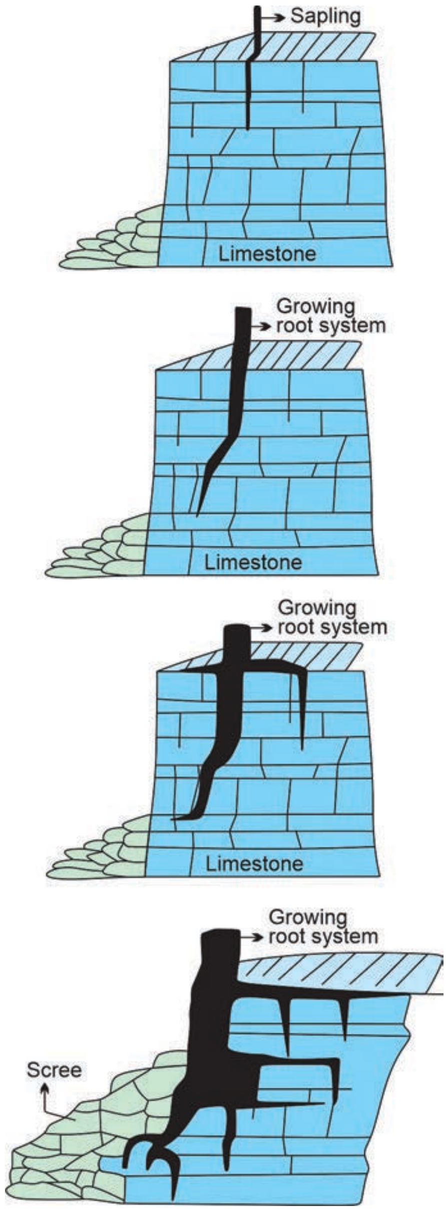

FIGURE 1.2 Root wedging.

1.3.1.2 Animals

Burrowing by animals (such as worms, termites, reptiles, rodents, etc.) into earth's substrate moves rock fragments and sediments enabling the disintegration of rocks. Additionally, digging by animals also results in the slow breaking of rocks.

1.3.1.3 Salt Wedging (Crystallization or Growth of Salt Crystals)

As water evaporates moisture from rocks, commonly noted in arid climates, mineral salts develop from mineral crystals; the mineral grains are moved apart by these crystal growths, thus breaking the rock (Figure 1.3). Crystallization causes stress that promotes the mechanical rupturing of both

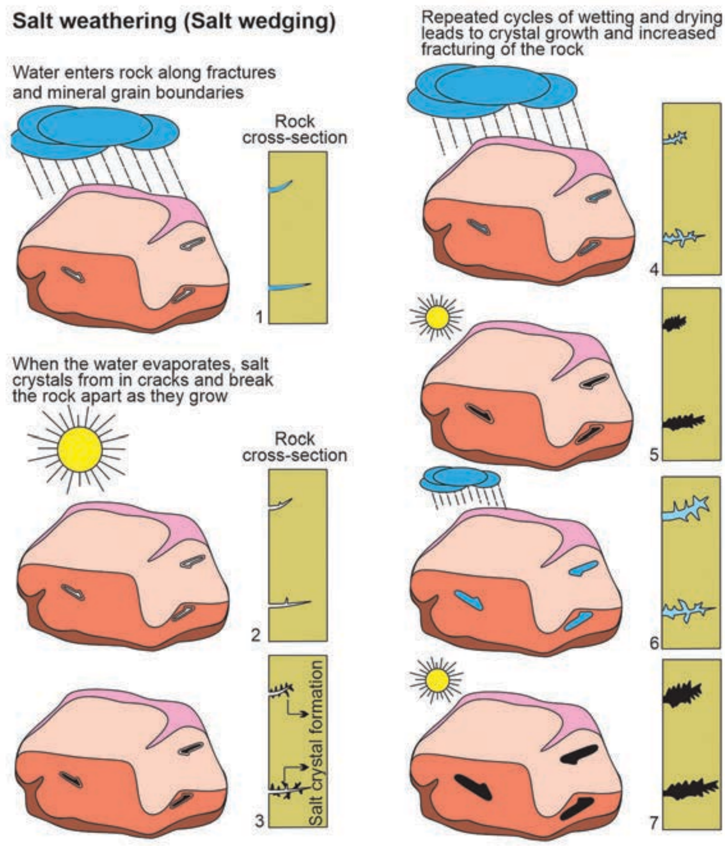

FIGURE 1.3 Salt weathering (or salt wedging). Crystal growth causes stress due to the compound's or an element's change in both physical state and volume change which in turn results in mechanical action and rock rupture. As the saltwater seeps into cracks in the rock, and then evaporates on a hot sunny day, the salt crystals grow within these cracks and pores. This crystal growth, through several cycles of wetting and drying, pushes the grains apart, causing the rock to weaken and break.

rocks and minerals (Figure 1.3). Crystal growth causes stress due to the compound's or an element's change in physical state and volume change which in turn results in the mechanical action and eventual rock rupture. Ice and salt are the two primary types of crystal growth that occurs. Salt crystallization is an example of volumetric change (from 1–5%) depending on the temperature of the rock or mineral surface. Most salt weathering occurs in hot arid regions, but instances are also noted to occur in cold climates. Salt wedging is the process by which the rocks weaken and eventually break (Figure 1.3). As the saltwater seeps into cracks in the rock, and then evaporate on a hot sunny day, the salt crystals grow within these cracks and pores. This crystal growth, through several cycles of wetting and drying, pushes the grains apart, causing the rock to weaken and break (Figure 1.3).

1.3.1.4 Grinding or Rubbing

This is the disintegration of rocks by grinding or by the rubbing of moving rocks against each other.

1.3.1.5 Abrasion

Abrasion, aided by erosional transport of materials by wind, water, and ice, occurs when two rock surfaces come together causing mechanical wearing or grinding. Pure water, although not abrasive, results in weathering due to the collisions among rock, sand, and silt. The wind throws sand against rocks, resulting in sandblasting, and eventual weathering. Glaciers also cause abrasion as they drag particles (clay-sized to boulders) across the bedrock; both the rock fragments (embedded within the ice) and the bedrock beneath, are abraded.

1.3.1.6 Unloading or Pressure-Release (Exfoliation)

Unloading or pressure-release is the removal of sediments (often thick layers) overlying deeply buried rocks either by erosion or uplift (Figure 1.4). This removal decreases pressure on the buried rocks (Figures 1.4A–B). Rocks are slightly elastic; hence, they expand in response to this pressure reduction by sediment removal forming fractures (cracks or fissures) parallel to the surface (Figure 1.4C). With continued erosion, these rocks are exposed on the surface as slabs of rock which eventually break off along the newly formed pressure-release fractures resulting in the formation of bare rock surfaces that are more resistant than the surrounding rocks (Figure 1.4C). These newly formed structures are called exfoliation domes (large, rounded masses of rock) and the slabs of rock that are broken off are called exfoliation sheets (Figure 1.4C). This is best seen in granites that split away like the layers of an onion (Figure 1.4D). Unloading plutonic igneous rocks form depth also creates zones of weakness in them. Hence, when these rocks are exposed, they expand, and the zones of weakness open up as joints. Some workers attribute exfoliation to a process called hydrolysis where feldspars and other silicate minerals react to form clay, and not to pressure-release fracturing. This water addition (i.e., the change from orthoclase feldspar to kaolinite) leads to a greater volume in clay as compared to the original mineral. Hence, the formation of clay (kaolinite) and the subsequent mechanical expansion of the clay, leads to exfoliation fractures (onion-shells; Figure 1.4D).

1.3.1.7 Wetting and Drying

The alternate wetting and drying of rocks is an important contributor to weathering. It is noted that about 20 cycles of alternating wetting and drying can disintegrate a rock sample. In soils, this wetting and drying results in swelling and contracting of soil particles. The soil shrinks when dry, and cracks develop, creating an irregular boundary between horizons. As rocks are alternately heated and cooled, they expand and contract; minerals also expand and contract by different amounts, and this differential expansion and contraction stresses the rocks and cracks them open, thereby, facilitating increased weathering.

1.3.1.8 Variation in Temperature

Insolation weathering, expansion and contraction due to diurnal (daily) temperature changes (which can be as large as 30°C or more), results in the physical breakdown of rocks. Insolation weathering is the inability of the rock to conduct heat properly, resulting in differential rates of expansion and contraction. Heat causes expansion, and cooling causes contraction. Hence, the surface of the rock expands more than its interior, resulting in stress that eventually causes the rock to break. Additionally, dark colored minerals expand more due to their increased absorptive properties. Hence, in a rock peppered with colored grains, rupturing occurs at differential rates at various mineral boundaries. Different minerals expand and contract at different rates causing stresses along mineral boundaries. Repeated heating and cooling of such rock causes the breakdown of the rock, resulting in enhanced weathering.

1.3.1.9 Freezing and Thawing

As water converts to ice, it expands as much as 9% of its volume. Freezing water exerts a pressure of 150 tons per square foot; repeated freezing and thawing that occurs on a daily cycle can split any mineral or rock. Thus, cracks that are filled with water are forced further apart due to freezing; this is called frost wedging – filling of a crack that freezes and expands (Figure 1.5). The expanding ice presses against the rock and wedges open the crack thereby facilitating enhanced weathering.

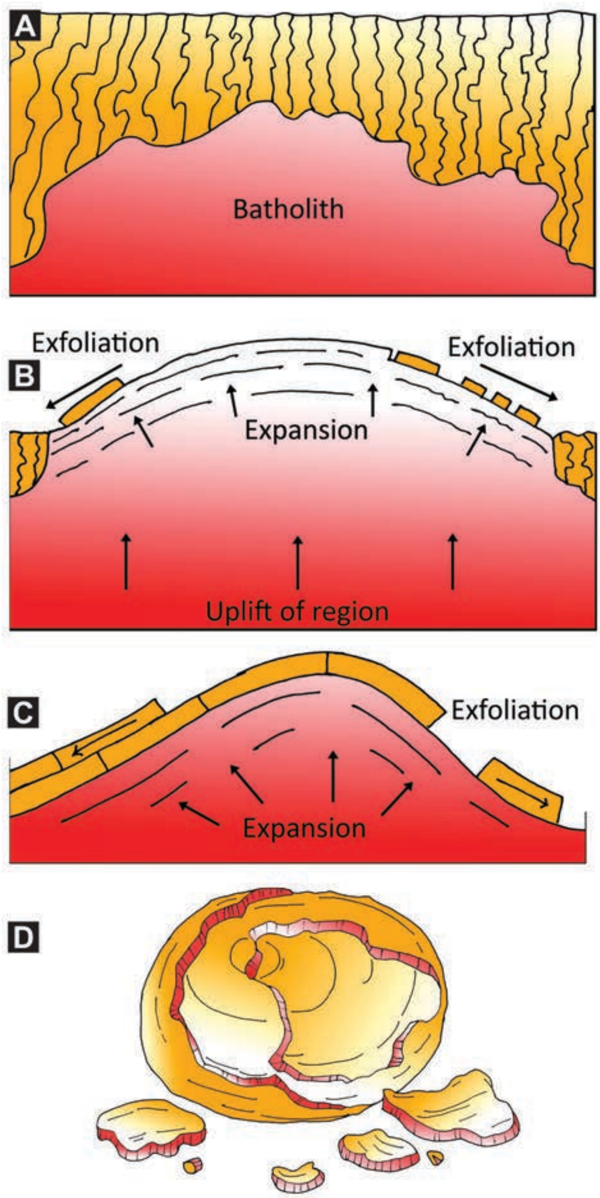

FIGURE 1.4 Unloading and exfoliation. A: These rocks were emplaced as molten bodies, or plutons, deep underground, raising the Sierra Nevada range. B: Erosion unroofed the plutons and took away the pressure of the overlying rock. C: As a result, the solid rock acquired fine cracks through pressure-release jointing. Mechanical weathering opened up the joints further and loosened these as slabs. D: Granite commonly fractures by exfoliation, a process in which large plates or shells split away like the layers of an onion. (Modified after Jain, 2014.)

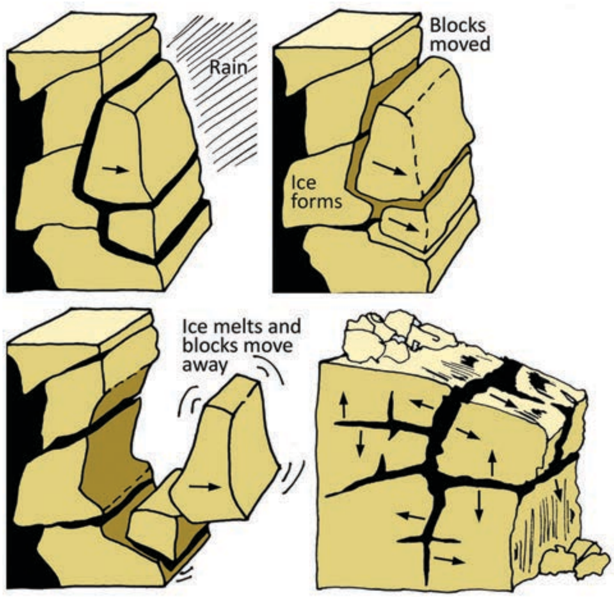

FIGURE 1.5 Frost wedging takes place when water seeps into cracks (A), expands as much as 9% of its volume as it freezes (B), and pries loose angular pieces of rock (C–D). (D: Modified from Christiansen Hamblin, 2014; with permission of Eric H. Christiansen.)

1.3.1.10 Heating and Cooling (Thermal Expansion and Contraction)

Differential expansion and contraction and the resulting stresses within a rock that occur due to differences in temperature cause minerals to fracture and exfoliate, as often noted in granitic terrains. Thermal expansion and contraction are more significant in large outcrops.

1.3.2 Chemical Weathering

Chemical weathering results in the decomposition and decay of minerals in rocks through chemical alteration (Table 1.1). Both increasing precipitation (rain) and temperature accelerate chemical weathering causing minerals to degrade; water is an essential agent of chemical weathering (Figure 1.6). Climate is another factor that considerably affects chemical weathering. It controls the rate of weathering by regulating the amount of precipitation and temperature and physical weathering in cooler and drier conditions (Figure 1.7A). Thus, chemical weathering is strongest in tropical wet to monsoon climates and physical weathering in cooler ones such as the subarctic and tundra (Figures 1.7A–B).

In general, weathering is a combined product of the effects of precipitation, temperature, and vegetation with minor inputs based on the nature of topography, rock type, and time (Figures 1.6 and 1.7). Thus, where temperature, moisture and vegetation are at their maximum, such as in the tropics, the weathering rates are three and a half times higher than the rates noted in temperate environments (Figure 1.7A). Additionally, the solubility of minerals depends upon solution (often in water), hydrolysis (the reaction of elements with water) and carbonation (the reaction with HCO_3^{-1}). Chemical weathering is faster where temperatures are higher and water is present (Figure 1.6).

TABLE 1.1

Weathering products of common rock-forming minerals (under the influence of CO_2 and H_2O)

Original material	Main solid products	Additional products (Ions in solution)
Amphibole (Ca, Mg, Fe Silicate)	Limonite, clay	Ions (K^+, Mg^{2+}, SiO^{4-}_{4})
Biotite (Fe, Mg, K, Al Silicate)	Clay mineral	Ions (K^+, Mg^{2+}, SiO^{4-}_{4})
Calcite	-	Ions (Ca^{2+}, HCO^{-}_{3})
Ferromagnesian minerals (including biotite)	Clay mineral	Ions (Na^+, Ca^{2+}, K^+, Mg^{2+}), SiO_2, Fe oxides
Gypsum ($CaSO_4$ plus water)		Ions (Ca^{2+}, SO^{2-}_{4})
Halite (NaCl)		Ions (Ca^{2+}, Cl^-)
Muscovite (K, Al Silicate)	Clay mineral	Ions (K^+, SiO^{4-}_{4})
Olivine (Mg, Fe Silicate)	Limonite, clay	Ions (Mg^{2+}, SiO^{4-}_{4})
Plagioclase feldspar (Ca, Na, Al Silicate)	Clay	Ions (Na^+, Ca^{2+}, SiO^{4-}_{4})
Potassium feldspar (K, Al Silicate)	Clay	Ions (K^+, SiO^{4-}_{4})
Pyroxene (Ca, Mg, Fe Silicate)	Limonite, clay	Ions (Ca^{2+}, Mg^{2+}, SiO^{4-}_{4})
Quartz (SiO_2)	Quartz grains (sand)	Ion (SiO^{4-}_{4})

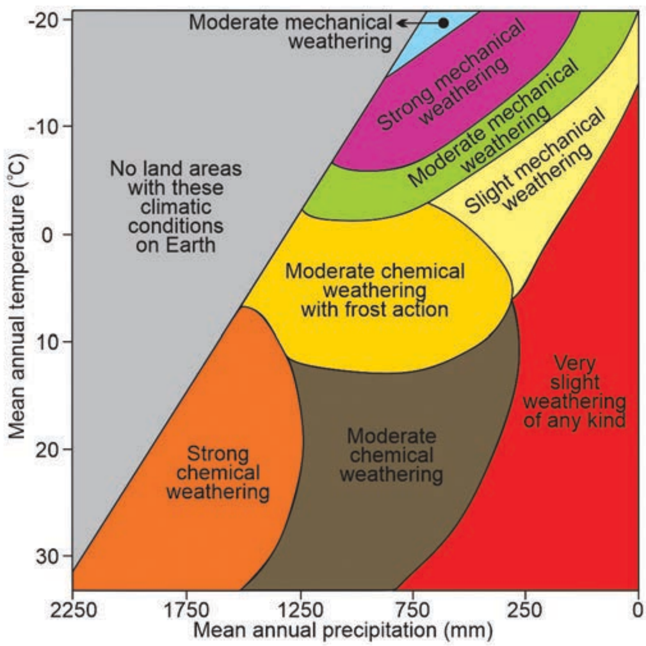

FIGURE 1.6 Role of precipitation and temperature (climatic conditions) in controlling the rate of weathering. (Modified from Peltier, 1950, Fig. 3, p. 219, and with permission from Taylor and Francis.) At high temperature and precipitation (bottom left of the diagram), chemical weathering is the strongest, and physical weathering in cooler and drier conditions (top right of the diagram). Hence, chemical weathering is strongest in tropical wet to monsoon climates and physical weathering in subarctic and tundra (see Figure 1.7A).

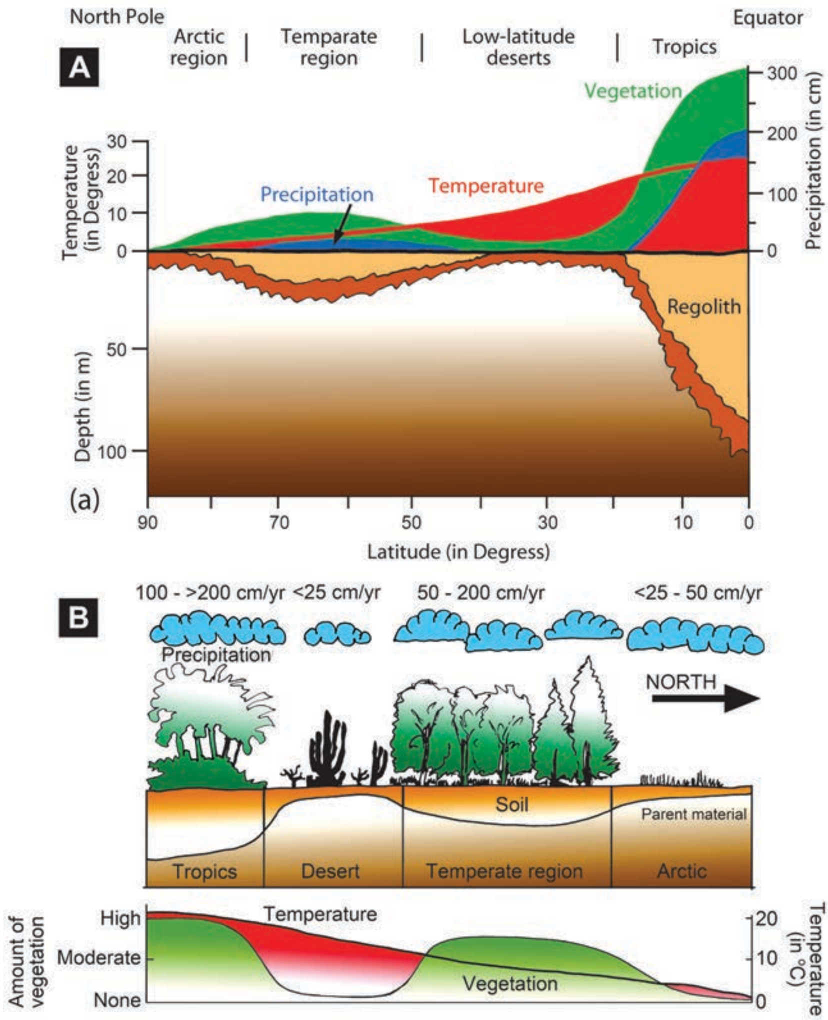

FIGURE 1.7 Relationship between precipitation, temperature, and vegetation and tropical weathering rates. A: The diagram shows the combined effects of precipitation, temperature, and vegetation. Weathering is most pronounced in the tropics, where precipitation, temperature, and vegetation reach a maximum. (Modified from Christiansen Hamblin, 2014; with permission of Eric H. Christiansen.) B: Tropical weathering rates are three and a half times higher, where temperature, moisture, and vegetation are at their maximum, than the rates noted in temperate environments (note the higher amounts of eroded soil).

1.3.3 Biological Weathering

Biological weathering is the process where living organisms, ranging from bacteria to plants (tree roots) to animals, break down rocks (see Figure 1.2). Lichen that live on bare rocks, demonstrate a symbiotic relationship between fungus and algae. The fungal part of the association secretes the acids, which reacts to dissolved minerals that are then used by the algae. Later, water seeps into crevices etched by the acid, and assists in the final break down through freezing (frost wedging) and chemical weathering. Tree roots grow into cracks and widen them, thus, facilitating physical weathering (Figure 1.2).

In summary, it must be kept in mind that, in a given area, a single weathering process may dominate, but all three weathering processes (physical, chemical, and biological) occur simultaneously. Physical weathering helps chemical weathering by breaking rocks into smaller pieces, thus, allowing more surface area to be exposed. With more surface area, chemical reactions occur faster. Chemical weathering helps physical weathering by weakening the mineral grains that bind the rock together. Biological weathering helps both physical and chemical weathering. Trees fracture rocks with their roots, which makes the rocks easier to break up physically, thereby, exposing more surface area for chemical weathering. Bacteria secretes acid solutions which speed up the chemical weathering process. Thus, working in tandem, these three weathering processes reduce a once-resistant rock into nothing or erode into weaker materials (such as clays).

1.3.4 Weathering Products

Weathering byproducts include (a) the complete loss of particular atoms or compounds from the weathered surface, (b) the addition of specific atoms or compounds to the weathered surface and (c) a breakdown of one mass into two or more masses, with no chemical change in the mineral or rock. Thus, the residue of weathering consists of chemically altered and unaltered materials. The most common unaltered residue is quartz which is resistant to chemical decay, and therefore, only undergoes size reduction during transportation. Other minerals, such as feldspar, olivine, augite, and hornblende, react with chemical reagents to produce various products. Generally, the Si, Al, and Fe ions from these minerals are used up in generating secondary solid products like quartz, muscovite, kaolinite, hematite, and goethite.

1.3.5 Weathering and Associated Rocks

When rock weathers, they usually do so by working inward from a surface that is exposed to the weathering process. This results in spheroidal weathering, exfoliation and weathering rinds.

1.3.5.1 Exfoliation

The onion-skin weathering (also called exfoliation) are shells of weathering that form on the outside of a rock. The shells (slabs) become separated from the rock due to the release of stresses that result from volumetric changes in minerals (Figure 1.4).

1.3.5.2 Spheroidal Weathering

The rock is broken into cube-like pieces separated by fractures, if joints and fractures form a three-dimensional network in the rock beneath the surface (Figure 1.8). Water enters through cracks and dissolves the cement that binds particles together and also erodes sharp edges and corners of the rock, making it appear spheroidal in shape; the central part remains unaltered, whereas the outside edges are weathered. This progression of weathering is called spheroidal weathering.

1.3.5.3 Weathering Rinds

A rock shows an outer weathered zone (called a weathering rind) and an inner unweathered zone (Figure 1.9). The thickness of the weathering rind increases as weathering progresses. In this process, the ferromagnesian silicates (minerals such as olivine, pyroxene, hornblende, and biotite) rust to produce hematite and other oxide minerals.

1.4 FACTORS AFFECTING THE RATES OF WEATHERING

Time, rock type, slope (topography), mineral composition (the type of rock), rock structure, and the prevailing climate control the rates of weathering (Figures 1.6 and 1.7). Rates of weathering vary

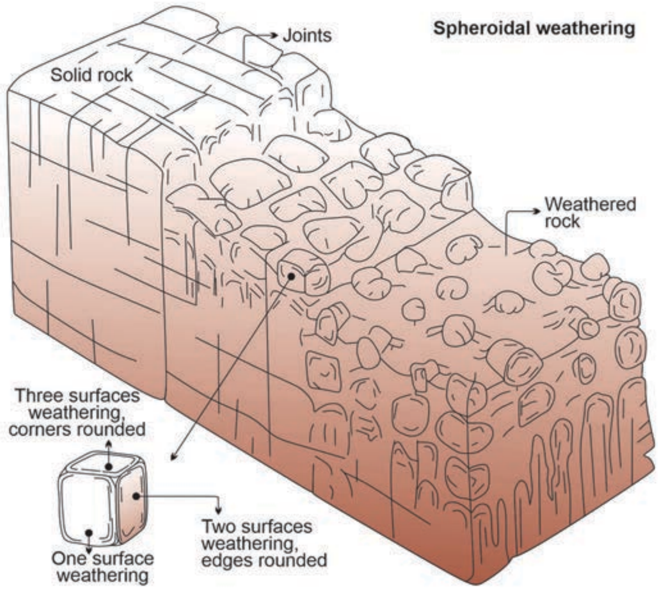

FIGURE 1.8 Spheroidal weathering. (Adapted from Fig. 43, *The Geologic Story of Yosemite National Park* [1987] by N. King Huber [www.yosemite.ca.us].)

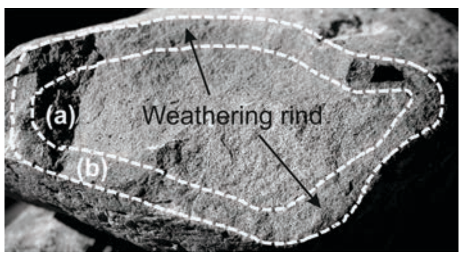

FIGURE 1.9 Weathering rind. Note the presence of an outer weathered zone (weathering rind) and an inner unweathered zone. As weathering progresses, the thickness of the weathering rind increases. It is, thus, sometimes also used as an indicator of the amount of time the rock has been exposed to the weathering process.

between rapid to extremely slow and are largely a function of slope, climate, and the type of rock. If all things are equal, the longer a rock is exposed at the surface, the more weathered it will be. When a slope is gentle, water stays in contact with rock for longer periods of time, and thus, results in higher weathering rates. Hence, on gentle slopes, the weathering products accumulate. The constituent minerals of a rock type and their susceptibility to breakdown define the type of weathering (Table 1.1). Thus, sandstones that are largely or exclusively made of quartz, a mineral that is very stable on the earth's surface, will not weather at all in comparison to a limestone which is composed of calcite, a mineral that easily dissolves in a wet climate (Table 1.1). Calcite, halite, and olivine are highly soluble, while quartz dissolves very slowly (Table 1.1). Additionally, the earlier a mineral crystallizes from the melt, the less stable it is at surface conditions.

Joints, fractures, fissures, and bedding planes provide easy pathways for the entry of water enabling enhanced physical and chemical weathering. Additionally, as soils hold moisture and organisms; rocks weather more quickly when they are in contact with a soil. High amounts of water and higher temperatures accelerate chemical reactions. Hence, in warm humid climates, there are more weathered rocks, and subsequently, the rates of weathering are much higher in comparison to cold dry climates; climatic conditions control the type and rates of weathering (Figures 1.6 and 1.7). Thus, areas with little water or low temperatures have relatively slow rates of chemical weathering (Figure 1.6). Physical weathering is most common in cold climates where frost action occurs. Frost action is most prevalent when temperatures are low and moisture is abundant (Figures 1.6 and 1.7). Hence, weathering rates are always differential and often, both physical and chemical weathering occurs together. For example, chemical weathering is fastest at the edges of a rock as they provide a larger surface area, resulting in faster chemical reactions, and thus converting the sharp edges into rounded ones. Physical weathering, on the other hand, breaks rocks into smaller pieces. Hence, more surfaces are exposed to chemical weathering. Thus, both physical and chemical weathering work in tandem.

1.5 STABILITY OF MINERALS

Uplift and/or erosion brings the rocks near the surface, thereby providing conditions that are very different from those under which they were originally formed, i.e., deep within the earth. This new environment facilitates minerals in rocks to react and thus produce new minerals that are stable under newer conditions (i.e., near the surface). Table 1.2 lists minerals in order of most stable to least stable ones.

1.6 WEATHERING OF COMMON ROCKS

Most rocks form clay minerals as the weathering end product; quartz and muscovite remain as residual minerals as they are resistant to weathering (Figure 1.10). The common rocks and their weathering products are provided in Table 1.3. In general, the ferromagnesian minerals present in igneous rocks, calcite and dolomite of sedimentary rocks (limestone), and iron oxides and sulfides, salts and gypsum, volcanic debris, and organic material are weathered to clay, highly soluble ions and suspended matter.

TABLE 1.2
Distribution of minerals in order of their stability with respect to weathering

	Slow weathering Rapid weathering		
	Stable in atmosphere	**Unstable in atmosphere**	**Dissolve and re-precipitate**
Least stable			Halite
			Gypsum
			Calcite
		Pyrite	Dolomite
		Olivine	
		Ca-plagioclase	
		Pyroxene	
		Amphibole	
		Biotite	
		Na-plagioclase	
		K-feldspar	
		Muscovite	
	Quartz		
	Clay		
	Aluminum oxide		
Most stable	Iron oxide		

TABLE 1.3
Weathering of common rocks

Rock	Primary minerals	Residual minerals	Leached ions
Granite	Feldspars	Clay minerals	Na^{+}, K^{+}
	Micas	Clay minerals	K^{+}
	Quartz	Quartz	-
	Fe-Mg minerals	Clay minerals + hematite + goethite	Mg^{+2}
Basalt	Feldspars	Clay minerals	Na^{+}, Ca^{+2}
	Fe-Mg minerals	Clay inerals	Mg^{+2}
	Magnetite	Hematite, goethite	-
Limestone	Calcite	None	Ca^{+2}, CO_3^{-2}

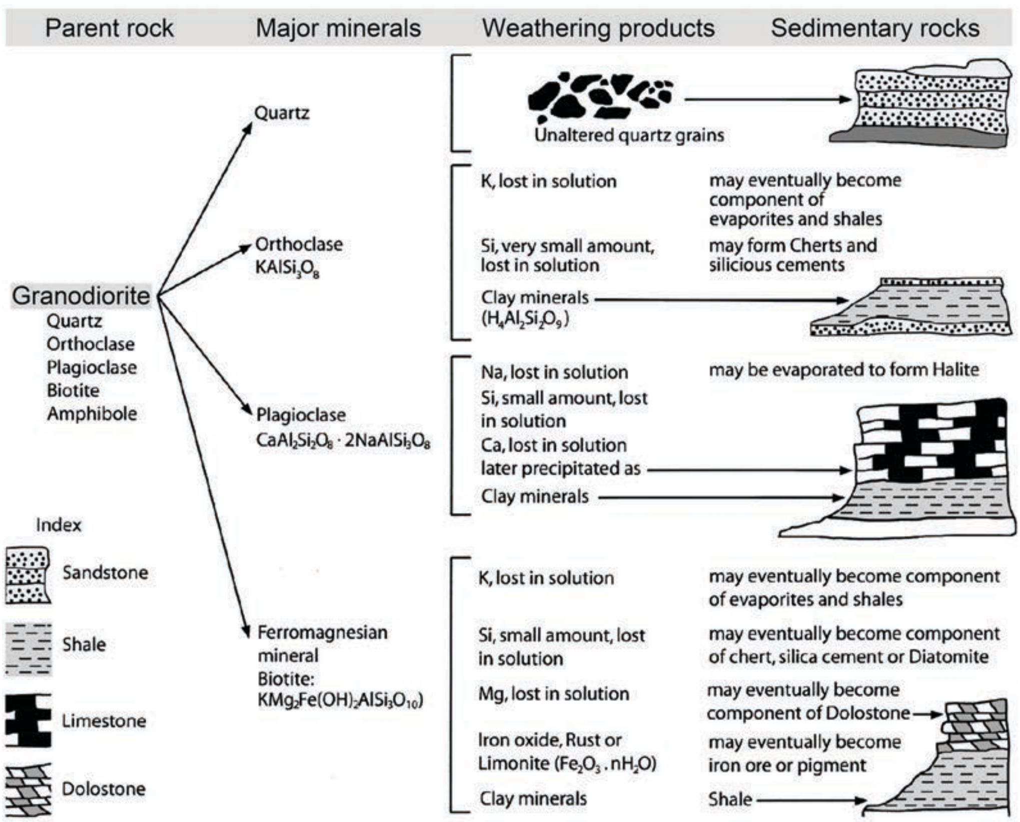

FIGURE 1.10 The weathering products of a granodiorite. The resultant products are used for the formation of various sedimentary rocks. (Modified after Jain, 2014.)

Granodiorite, an igneous rock, provides a good example of weathering byproducts of its major minerals. The unweathered granodiorite contains minerals such as plagioclase feldspar, potassium feldspar, quartz, and small amounts of biotite, amphibole, orthoclase, and sometimes even muscovite (Figure 1.10). Weathering under warm, humid conditions forms a rock called saprolite (also referred to as rotten rock). The feldspars undergo hydrolysis and form clay (kaolinite) and sodium and potassium ions. The Na and K ions are removed through leaching and carried in solution along with the running water. Biotite and/or amphiboles undergo hydrolysis to form clay, and after oxidation, to form iron oxides. Quartz (and muscovite, if present) are resistant to weathering, hence remain as residual minerals (Figure 1.10). The weathered byproducts are subjected to various physical processes, such as transportation and deposition. Quartz grains are eroded, and incorporated into sedimentary rocks or just become quartz sand which is eventually transported to the ocean (as bed load), where it accumulates to form beaches. Clays are ultimately eroded and washed out to the ocean. The fine-grained clays remain suspended in the water column (as suspended load) and are eventually deposited in quiet waters. Dissolved ions are transported by rivers to the ocean as dissolved load, and eventually become part of the salts in the ocean.

1.7 CONSTITUENTS OF SOIL

Soil is the byproduct of weathering of a parent rock that may be just below the soil or at some distance from it. All soils are formed through a series of changes brought about by weathering (physical, chemical, and biological), its intensity and duration, the physico-chemical properties of the parent rock, climatic and geologic factors, among others. Soil contains five major components: (a) mineral constituents such as clay, silt, sand, and gravel; (b) organic matter made of decaying plant and animal parts; (c) water; (d) air; and (e) organisms (soil biota ranging from bacteria, fungi, and earthworms).

In general, soil contains about 50% solid space and 50% pore space (Figure 1.11). The total solid space of the soil is made up of 45% mineral matter and 5% organic matter, whereas the total pore space (the other 50%) is made by 25% air and 25% water (Figure 1.11); the latter proportion varies depending upon weather and environmental factors. Soils with more than 20% organic constituents are called organic soils, and those in which inorganic constituents dominate are called mineral soils. The inorganic constituents include silicates (primary or secondary; primary such as pyroxene, amphibole, olivine, mica, and secondary minerals such as clays), and a proportion of carbonates, soluble salts, and free oxides of iron, aluminum, and silicon, with some amorphous silicates. In general, soil formation is a very long process and it takes several million years to form a thin layer of soil. Broadly, four stages of soil formation are noted.

1.7.1 Stage of Soil Formation

In Stage 1 (Figure 1.12A), the bedrock is exposed and is subsequently fractured by continued weathering (physical, chemical, and biological). Lichen and moss grow on the rock surface as do small plants in cracks and depressions, where small amounts of sediments start accumulating.

In Stage 2 (Figure 1.12A), weathering is at an advanced stage, and the parent material is much softer and weaker due to continued weathering; weathering products now accumulate in between the remaining rock fragments. The C horizon has developed from the products of weathering (consisting of small rock fragments, sand, and clay); organic matter accumulates near the surface.

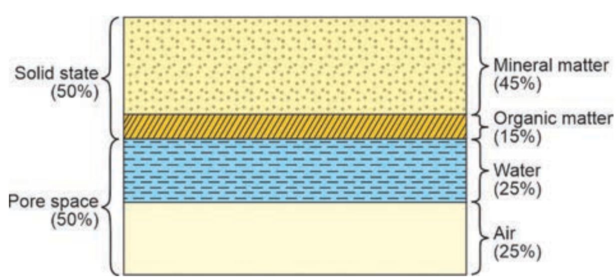

FIGURE 1.11 Soil components.

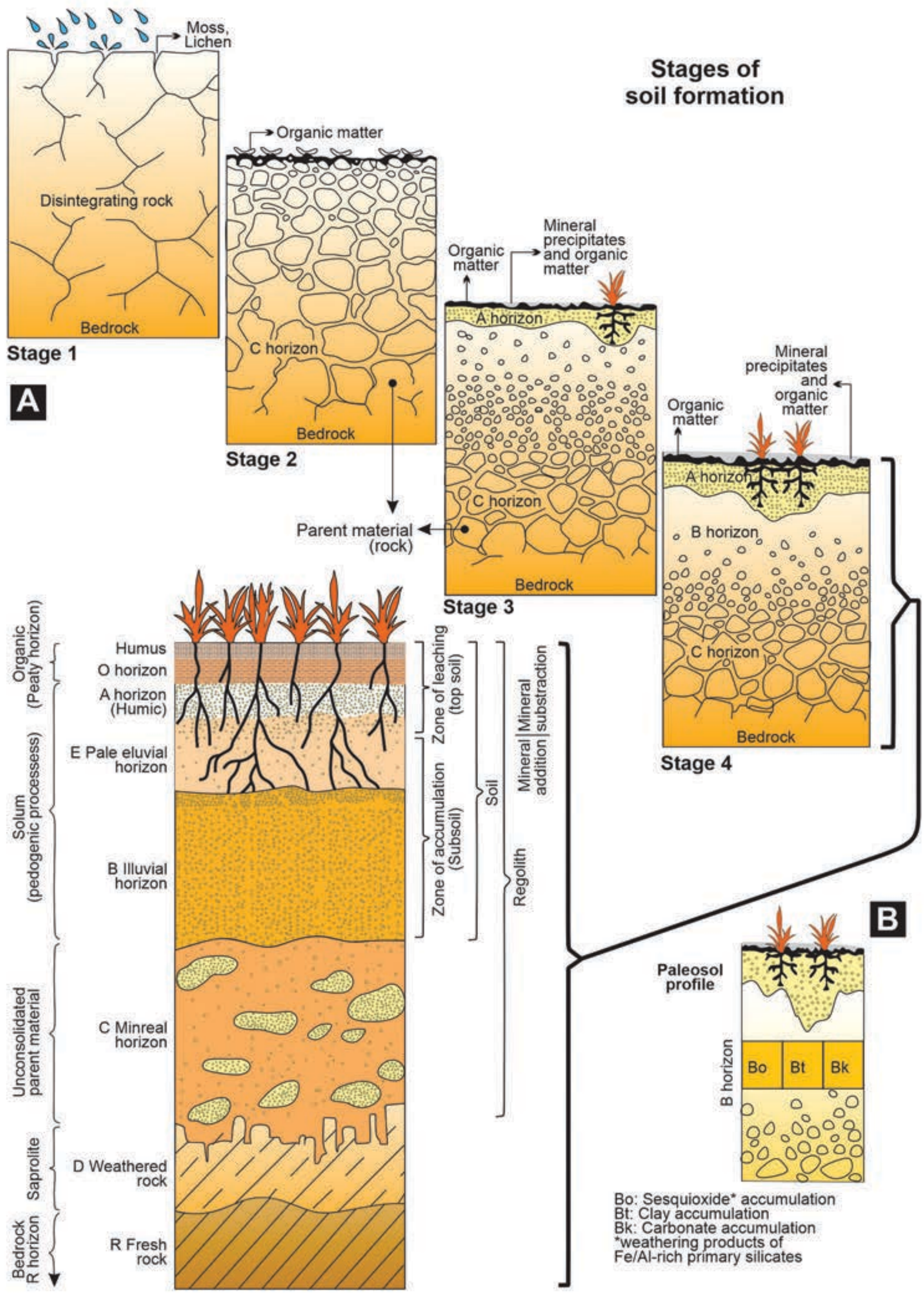

FIGURE 1.12 Stages of soil formation.

In Stage 3 (Figure 1.12A), the soil profile has started to evolve, largely due to chemical changes brought about by the downward and upward motion of ions in water, and the transfer of water and chemicals by the roots of plants and the mycorrhizal network (i.e., the network of plant roots that connects individual plants together to transfer water, nitrogen, carbon, and other minerals).

In Stage 4 (Figure 1.12A), the soil is now well-developed with significant weathering of minerals like feldspar and amphibole within the soil to produce clay minerals associated with the upward and downward movement of chemicals (such as iron, manganese, potassium, sodium, calcium, magnesium, and aluminum). At this stage, relatively large amounts of carbon is stored in the soil (as organic matter, charcoal, carbon dioxide and methane) due to the abundance of living organisms (roots, mycorrhizae, worms, and insects). Additionally, based on the type of the parent material, and the rate of soil formation (influenced by local climate), the soil is now several centimeters to over a meter thick. All horizons are now well-developed.

1.7.2 Mechanisms of Soil Formation

The soil formation process begins with a parent material that determines the mineral composition and contributes to the chemical and physical properties of the soil. There are several mechanisms involved in soil formation, such as weathering, accumulation of materials, leaching, transformation, and calcification. These are briefly enumerated below.

1.7.2.1 Weathering

Weathering (physical, chemical, and biological) is the breakdown of rocks and minerals at or near the earth's surface into products that result in equilibrium with the conditions found in this environment. Physical weathering is the breakdown of mineral or rock material by mechanical methods. Chemical weathering, on the other hand, results in the change of chemical and mineralogical composition of the weathered material through myriad processes, such as hydrolysis, oxidation, reduction, hydration, carbonation, and solution. Biological weathering, always a minor contributor, disintegrates rocks and minerals due to the chemical or physical agents of an organism such as bacteria, plants, and animals.

1.7.2.2 Accumulation of Materials

Over time, organic matter, and decomposing or new mineral materials are added to the soil. In poorly drained soils, organic matter accumulates more as waterlogging prevents it from being oxidized. But, in well-drained soils, accumulation only occurs when the root systems hold them up. Additionally, particle depositions by wind, water, or ice aid in the accumulation of new materials. Some plants with the help of symbiotic bacteria fix atmospheric nitrogen and ammonia compounds into the soil as nitrates, thus adding new mineral materials.

1.7.2.3 Leaching

Leaching removes the soil's soluble components by water by carrying bases like calcium, that are held as exchangeable ions in the clay-humus complex, or by causing acidification by the substitution of hydrogen ions. Through the movement of water or by the uptake of the accumulated materials by plants, new materials such as organic matter, clay, silt, or other chemical compounds get leached or are taken up by plants. Hence, through leaching, the parent material is altered by the accumulated new materials. In simple terms, leaching is the loss of soluble substances from the top layer by percolating water. The lost materials are carried downward (eluviated) into the E horizon and are redeposited (illuviated) in the lower layer, B horizon (see Figure 1.13). A good example of leaching is the formation of the laterite soil. It is formed by the leaching or weathering of laterite rocks when high temperature and heavy rainfall is present, and with alternating dry and wet spells. The laterite soil is red in color with a high iron oxide content and poor in nitrogen (see Figure 1.13).

1.7.2.4 Transformation

Chemical weathering, and to an extent physical breakdown, of soil materials (such as silt, sand, and clay minerals), and the conversion of organic materials into its degradation-resistant form is called transformation. These are then washed from the upper layer and deposited in the lower horizons.

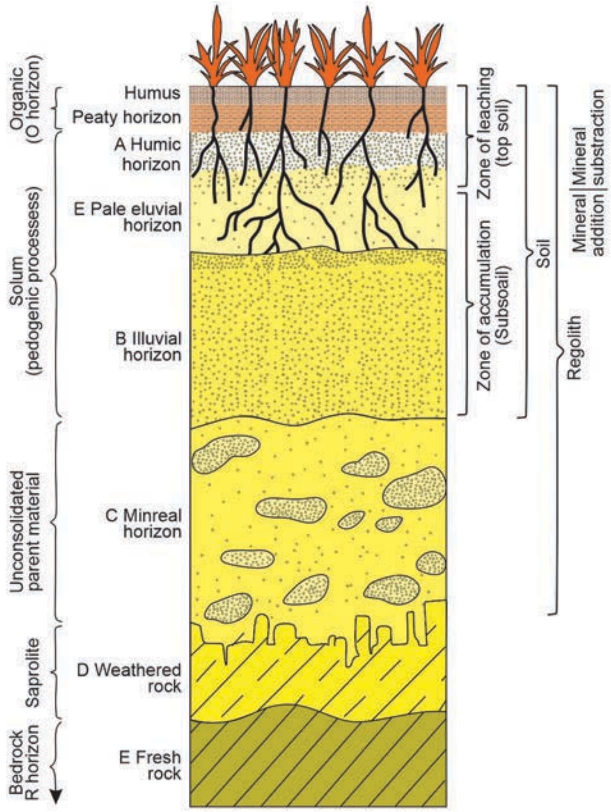

FIGURE 1.13 Soil profile. The figure shows the transition from bedrock to regolith through a sequence of layers. The most obvious product of weathering is a blanket of loose, decayed rock debris known as regolith, which forms a discontinuous cover over the solid, unaltered bedrock. Eluviation (= E horizon) is the mobilization and translocation of certain constituents, viz. clay, F_e2O_3, Al_2O_3, SiO_2, humus, $CaCO_3$, other salts, etc. from one point of soil body to another. Illuviation (= B horizon) is the immobilization and accumulation of the eluviated constituents at a depth beneath the soil surface.

1.7.2.5 Calcification

It occurs when the removal of water through evapotranspiration is more than precipitation, thus causing the upward movement of dissolved alkaline salts from the groundwater. On the other hand, the movement of rainwater causes a downward movement of the salts.

1.7.3 Factors Influencing Soil Formation

In particular, five factors influence the process of soil formation. These include parent material, topography, climate, organisms, and time. These are briefly enumerated below.

1.7.3.1 Parent Material

This is the initial solid mass that forms the soil, such as consolidated rocks or unconsolidated particles (bed rock; Figure 1.13). The type of soil formed is based on the composition of the parent material; for example, iron-containing rocks will yield iron-rich soil characterized by darker color and higher pH. Coarse-grained parent material composed of minerals resistant to weathering will yield a soil characterized by coarse-grained texture, whereas parent material composed of unstable minerals that readily weather will yield fine-grained soils. Similarly, parent materials that are rich in soluble ions, such as calcium, magnesium, potassium, and sodium, will get dissolved easily in water and are made available to plants. Contextually, parent materials that have a high content of soluble bases, such as limestones and basaltic lavas, form fertile soils in humid climates. However, those that are low in soluble ions, the soil becomes acidic and unsuitable for agriculture, as water moving through the soil removes the bases and substitutes them with hydrogen ions. Soils developed over sandstone are low in soluble bases and coarse in texture; these conditions also facilitate leaching.

1.7.3.2 Topography/Slope

This includes slope, and geological structures with elevation above sea level; these influence hydrologic cycle, transpiration, and other similar processes. Hence, soil profiles on the convex side of the slopes are shallower with less defined horizons than those noted on the top of the concave slopes. However, organic matter, an important constituent, is higher at lower slopes due to reduced runoffs. Similarly, in steep, long slopes, soil development is poor as the mean water run down is faster and thus erodes the surfaces of slopes more; conversely, the foots of the slopes have better soil development.

1.7.3.3 Climate

Weather (including rainfall, temperature, and storm patterns) over long time periods (i.e., climate), influences soil formation. Climate acts via water and solar energy; water gives life (and various soil organisms), whereas sunlight affects the concentration of water in the soil. Hence, desert soils, usually present around the equatorial region, have high solar and water energy. However, in the temperate regions, climate is humid, resulting in tropical soil with enough moisture content.

1.7.3.4 Organisms

Soil formation is greatly influenced by organisms (macro- and microorganisms), and vegetation. Hence, soils present under trees tend to be more acidic and contain much less humus than those under grass. Soil animal inhabitants also affect soil formation as they influence the organic content of soil and the texture due to their metabolic and physical activity. Microorganisms facilitate chemical reactions or excrete organic substances that improve water infiltration in the soil, hence better soil development. But some macro-organisms, such as gophers, inhibit soil development as they dig and mix soil materials, thus destroying soil horizons.

1.7.3.5 Time

The effects of time on the soil profile are reflected in its composition, where increased accumulation of clay and lime in the horizons occurs due to downward translocation. The humus content in the soil horizons differs with increasing time (aging). It must be noted that it can take over 500 years to form just one centimeter of soil from some of the harder rocks.

1.7.4 Soil Horizons

The soil layers are called horizons designated by capital letters O, A, E, B, and C; O is the organic horizon, whereas A, E, B, and C are mineral horizons, overlaying the bedrock (R horizon) (see Figure 1.13). These horizons are very briefly enumerated below.

O horizon is the layer of organic matter, formed from organic litter derived from plants and animals. It sits on the top of the soil (Figure 1.13). Due to rich organic content, this horizon is black or dark brown in color.

A horizon, in the absence of O horizon, is known as the topsoil or sometimes referred to as the root zone (see Figure 1.13). It is made of sand, silt, and clay with high amounts of organic matter. The horizon is characterized as the zones of “washing out” or of maximum leaching (Figure 1.13).

E horizon is sometimes used for eluvial horizon or parts of the B horizon (see Figure 1.13). It is lighter in color, often rich in nutrients leached from the above A and O horizons. Hence, this is the mineral horizon in which the loss of silicate clay, iron, aluminum, or some combination of them, is noted resulting in the concentration of sand and silt particles; almost all of the original rock structure is destroyed. This horizon is lighter in color than the underlying B horizon, as it contains less organic matter (see Figure 1.13).

B horizon (subsoil) is also called the illuviation zone due to the accumulation of minerals; it has more clay than topsoil (see Figure 1.13). This horizon is also characterized by the accumulation of illuvial, amorphous, dispersible organic matter (i.e., the illuvial accumulation of organic matter). This horizon is also referred to as the zone of "washing in" or the accumulation of materials such as iron and aluminum oxides and silicate clays; in arid conditions, calcium carbonate, calcium sulfate, and other salts accumulate during evaporation.

C horizon (weathering rock) is also known as saprolite. This layer is often very deep, and may not be present everywhere. It is predominantly made up of broken bedrock with no organic material (Figure 1.13). It has cemented sediments with little or no activity; additions and losses of soluble materials may occur. In general, some soils that are formed from material that is already highly weathered, and that do not meet the requirements of A, E, or B horizons, are designated as the C horizon.

R horizon is the hard bedrock, such as granite, basalt, quartzite, and indurated limestone or sandstone (Figure 1.13). It must be noted that, in some cases, the B horizon is absent due to mixing of the A and B horizons; such profiles have an AC horizon.

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2 Sediment Properties

2.1 INTRODUCTION

Sedimentary rocks are characterized by their distinctive grain parameters such as size, shape (form, roundness, surface texture), and grain fabric (grain orientation and grain-to-grain relationship). All textural parameters, primarily produced by physical processes of sedimentation, are fair proxies for inferring the nature of sediment transport and depositional processes. The nature of these primary textural properties (of siliciclastic sedimentary rocks as discussed in this chapter) also has a bearing on the derived, textural properties such as rock bulk density, porosity, and permeability. The textures of some non-siliciclastic sedimentary rocks such as limestones and evaporites (see Chapters 6–7 for details, respectively) are also formed (in part or whole) by similar physical transport processes and part by chemical or biochemical sedimentation processes. Recrystallization and/ or diagenetic changes may remove the original textures and produce secondary crystalline textural fabrics reflecting a strong diagenetic imprint. These are also genetically quite different from siliciclastic sedimentary rocks.

2.2 GRAIN PARAMETERS

Grain parameters refer to various characteristics and properties of mineral grains within a rock or sediment. These parameters are used to describe and analyze size, shape, composition, and texture of individual grains, thus, providing valuable information about the origin, transport, and depositional processes of sediments, as well as their metamorphic or diagenetic history.

2.2.1 Grain Size and Grain Scales

Grain size describes the average size of individual mineral grains within a rock. It is typically reported in terms of an average value, such as the mean grain diameter or the median grain size. Grain size ranges from being fine-grained (with small, closely packed grains) to coarse-grained (with larger, more widely spaced grains). Grain size provides information about the conditions under which they were formed. Grain size can be measured using several techniques, such as optical microscopy, electron microscopy, or laser diffraction. For sediments that are composed of sand-sized particles, hand-lens is used to determine the dominant grain-size class; it is usually possible to distinguish between very coarse/coarse, medium and fine/very fine sand classes.

The sediment particles show large size variations ranging from several meters to less than a micrometer, hence any scale using uniform divisions by size places more emphasis on the coarse range and less on the finer fractions. To mitigate this, a geometric scale, the Udden-Wentworth scale (see Figure 2.1), was introduced which places equal emphasis on small differences in the fine

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fraction and larger differences for the coarse ones. The two commonly used grain-size scales are Udden-Wentworth and Phi scales (Figure 2.1). These describe and categorize the range of grain sizes found in a rock and thus provide a standardized way to quantify and display grain sizes.

2.2.1.1 Udden-Wentworth Scale (Millimeter Scale)

This is a widely used geometric scale that describes and classifies sediments (specifically detrital ones) based on their grain diameter (see Udden 1914; Wentworth 1922, 1935; Tanner 1969; Folk et al. 1970; Blair and Mcpherson, 1999). It is a ratio scale where the grade boundaries

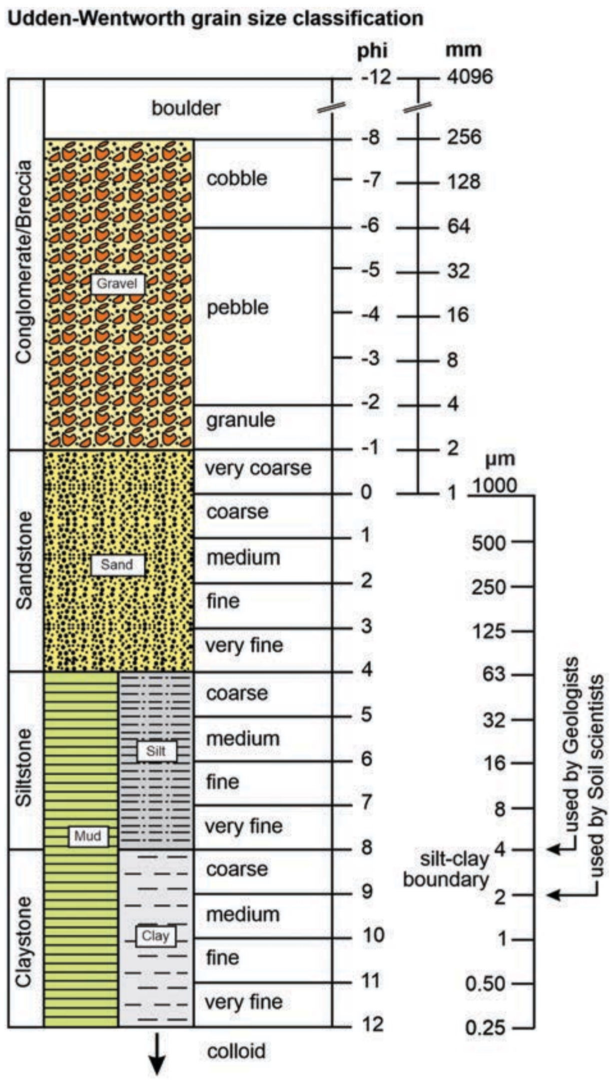

FIGURE 2.1 Grain-size scale. Note that the arithmetic scale in phi units is used for grain-size data, as it has the advantage of making mathematical calculations easier. The scale gives equal significance to size ratios, whether they relate to gravel, sand, silt, or clay. The Udden-Wentworth scale is the most widely used grain-size scale. (Modified after Wentworth, 1922; Krumbein, 1934, 1968; Krumbein and Pettijohn, 1938; Krumbein and Sloss, 1963.)

differ by a factor of 2. One grade coarser is twice the size of its predecessor and one grade finer is half the size (Figure 2.1). This scale recognizes three fractions, gravel (2 to 4096 mm), sand (1/16 to 2 mm), and mud (<1/16 mm). The mud fraction has been divided into silt and clay classes, and the gravel fraction into granule, pebble, cobble, and boulder classes. The silt-clay boundary is variously placed either at 2 µm (Briggs, 1977; Friedman and Sanders, 1978) by soil scientists, or at 4 µm by sedimentologists, as proposed in the original Udden-Wentworth system (see Tanner, 1969; Pettijohn, 1975) (Figure 2.1). The scale gives equal significance to size ratios, whether they relate to gravel, sand, silt, or clay. The Udden-Wentworth scale is the most widely used grain-size scale.

2.2.1.2 Phi Scale (Krumbein's phi or Logarithmic Scale)

For the better representation of grain-size analysis, it is imperative to use logarithmic scale graph to give visual equality to scale divisions. Contextually, Krumbein (1934) introduced the phi transformation, where the grain size is expressed as $\Phi = \log_2 d$; d is the grain diameter in millimeters (see Krumbein, 1934, 1938; Krumbein and Pettijohn, 1938; Krumbein and Sloss, 1963). Later, this scale was modified as $\Phi = - \log_2 (d/d_o)$, where d_o is the diameter of a 1 mm grain, thus enabling the use of this dimensionless numbers for statistical analyses and to derive factors such as standard deviation, skewness, and kurtosis of grain-size distributions (McManus, 1963; Krumbein, 1964). The larger the particle phi number the smaller (finer) the particle is. The phi scale allows for easier mathematical calculations and comparisons between grain sizes. There are three major advantages of using the phi scale: (a) grain-size distributions can be plotted easily on the arithmetic paper instead of the log paper; (b) statistical parameters can be conveniently interpolated; and (c) the boundaries between different size units are in whole numbers (Figure 2.1).

Grain-size analysis is an essential tool for classifying sedimentary environments. Broadly, the grain size of siliciclastic sediments reflects the hydraulic energy of the environment: coarser sediments are transported and deposited by faster-flowing currents than finer sediments; fine-grained mudrocks tend to accumulate in quieter waters. The sorting of a sandstone reflects the depositional process, and this improves with increasing agitation and reworking. In contrast, the grain size of carbonate sediments generally reflects the size of the organism skeletons and calcified hard parts which make up the sediment; these are affected by currents.

2.2.2.1.3 Sieve Numbers

Sieve numbers, also known as mesh size or mesh numbers, are used to classify and measure the size of grains. Sieve numbers are typically assigned to specific mesh sizes based on the number of openings per linear inch or per centimeter in a sieve. The higher the sieve number, the smaller the particle size that can pass through the sieve (see Table 2.1).

The American Society for Testing and Materials (ASTM) E11 standard (a scale in millimeter, mm or micrometer, µm) is commonly used to define sieve numbers (see Table 2.1). The standard provides a table that correlates sieve numbers to specific mesh sizes. For example, a sieve number of 4 corresponds to a mesh size of 4 openings per linear inch, while a sieve number of 200 corresponds to a mesh size of 200 openings per linear inch. In addition to sieve numbers, the corresponding opening size or particle size range is often provided. For example, a sieve with a mesh size of 4 may have openings of approximately 4.75 mm (0.187 inches), while a sieve with a mesh size of 200 may have openings of approximately 0.075 mm (0.0029 inches) (see Table 2.1).

2.2.2.2 Presentation of Grain-Size Data

When presenting grain-size data, there are several common ways to effectively communicate the information such as by using histograms, frequency curves, cumulative distribution curves, and classification charts. A few approaches using the data are briefly explained (see also Table 2.2).

TABLE 2.1
Sieve numbers (mesh size or mesh numbers) based on American Society for Testing and Materials (ASTM) E11 standard

ASTM E 11		Nominal sieve opening mm	Nominal sieve opening inch
	Coarse sieves		
Standard	Alternate		
75 mm	3 inches	75	3
63 mm	2-1/2 inch	63	2.5
50 mm	2 inches	50	2
37.5 mm	1-1/2 inch	37.5	1.5
25 mm	1 inch	25	1
19 mm	3/4 inch	19	0.75
12.5 mm	1/2 inch	12.5	0.5
9.5 mm	3/8 inch	9.5	0.375
	Fine sieves		
4.75 mm	No. 4	4.75	0.1870
2.36 mm	No. 8	2.36	0.0937
1.18 mm	No. 16	1.18	0.0469
600 μm	No. 30	0.6	0.0234
300 μm	No. 50	0.3	0.0117
150 μm	No. 100	0.15	0.0059
	Very fine sieve		
75 μm	No. 200	0.075	0.0029

TABLE 2.2
Sieves commonly used for sieve analysis

Φ size class	raw weight (gm)	individual weight percent	cumulative weight percent
-1	0.39	0.5	0.5
-0.5	1.96	2.5	3
0	3.92	5	8
0.5	6.26	8	16
1	8.61	11	27
1.5	11.75	15	42
2	12.53	16	58
2.5	11.75	15	73
3	8.61	11	84
3.5	6.26	8	92
4	3.92	5	97
4.5	1.96	2.5	99.5
5	0.39	0.5	100
Total	78.31		

Histogram is the simplest method that displays the number or percentage of grains falling within different size ranges (Figure 2.2A). The x-axis represents the grain-size range (in phi scales), and the y-axis represents the frequency or percentage (Figure 2.2A). Histograms are useful for visualizing the distribution and range of grain sizes within a sample (Figure 2.2A).

Frequency curve is where a table is used to present grain-size data in a tabular form. The table includes columns for the grain-size range, the frequency or percentage of grains falling within each range, and any additional statistical parameters such as mean, median, or standard deviation. This format allows for a clear and concise presentation of the grain-size distribution.

The cumulative distribution curve is also known as an empirical cumulative distribution function (ECDF), which shows the cumulative percentage or frequency of grains below a given size (Figure2. 2C). It plots the grain size on the x-axis and the cumulative percentage or frequency on the y-axis (Figure 2.2C). This curve provides a comprehensive view of the cumulative distribution of grain sizes.

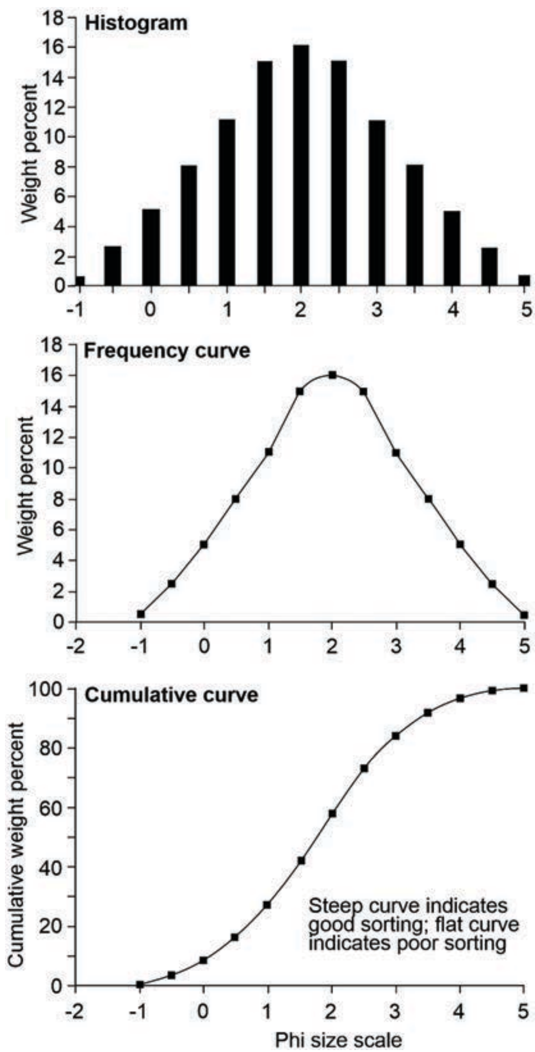

FIGURE 2.2 Presentation of grain-size methods, illustrating the character of grain-size data using histogram, frequency curve, and cumulative distribution curve methods.

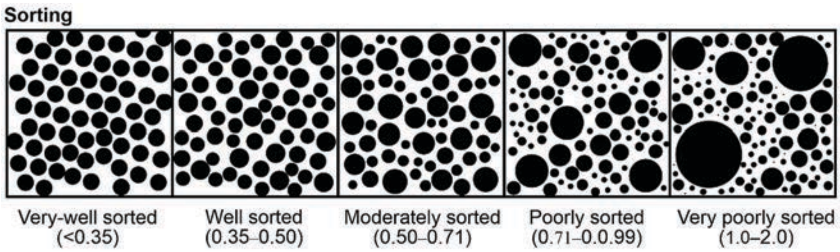

FIGURE 2.3 Sorting. This is the degree of uniformity in the size, shape, and composition of sediment within a sedimentary deposit. It displays the degree of homogeneity or heterogeneity of particle sizes from very well sorted to very poorly sorted.

The classification chart classifies the grain sizes into different categories or classes. This is done using a grain-size classification chart that assigns descriptive terms (e.g., fine, medium, coarse) to specific grain-size ranges. This approach simplifies the interpretation of the grain-size data and facilitates comparisons between samples.

In addition to numerical or graphical presentations, visual representations such as photomicrographs or scanning electron microscope (SEM) images are also used to illustrate grain shapes and textures. These methods provide a more detailed view of the grains and help in better understanding their characteristics and relationships.

2.2.2.2.2 Sorting

Sorting refers to the degree of uniformity in the size, shape, and composition of sediment within a sedimentary deposit. It is a measure of the degree of homogeneity or heterogeneity of particle sizes (Figure 2.3). Sorting of sediments provides information about the processes that formed it and the energy conditions under which it was transported and deposited. Well-sorted sediments have particles that are similar in size and shape (Figure 2.3), indicating that they were transported and deposited under relatively consistent and stable energy conditions. Poorly sorted sediments that have a wide range of particle sizes and shapes (Figure 2.3), suggest a mixture of different source materials and/or variable energy conditions during transport and deposition. Well-sorted sediments or rocks are often associated with high-energy environments, such as rivers, beaches, or wind-blown dunes, where the particles are transported over long distances and sorted by size. Poorly sorted sediments are typically associated with low-energy environments, such as lakes, swamps, or glacial deposits, where the particles are deposited close to their source and not significantly sorted. Sorting is assessed visually or quantitatively. Visual assessment involves examining the sediment and assigning a descriptive term to describe the degree of sorting (Figure 2.3). Quantitative assessment involves measuring the particle sizes within the sample and calculating the standard deviation (SD) to quantify the degree of sorting, such as very well-sorted (where the SD value is <0.35), well-sorted (0.35–0.50), moderately well-sorted (0. 50–0.71), moderately sorted (0. 71–1.00), poorly sorted (1.00–2.00), very poorly sorted (2.00–4.00), and extremely poorly sorted (>4.00); some values are shown in Figure 3 (see Folk and Ward, 1957).

2.2.2.2.3 Skewness and Kurtosis

Skewness and kurtosis are statistical measures used to describe the shape and distribution of grain-size data (Figure 2.4). Both methods provide information beyond the average grain size and thus help in the detailed characterization of grain size distribution. But, to better understand skewness and kurtosis, it is imperative to understand mean (MZ), median (Md), and mode (MO) (Figure 2.4A). Mean (MZ) is a graphic measure for determining overall size and is given by the formula MZ = (ΦI6 + ΦSO + Φ 84)/3. Median (Md) is when half of the particles by weight are coarser than the median,

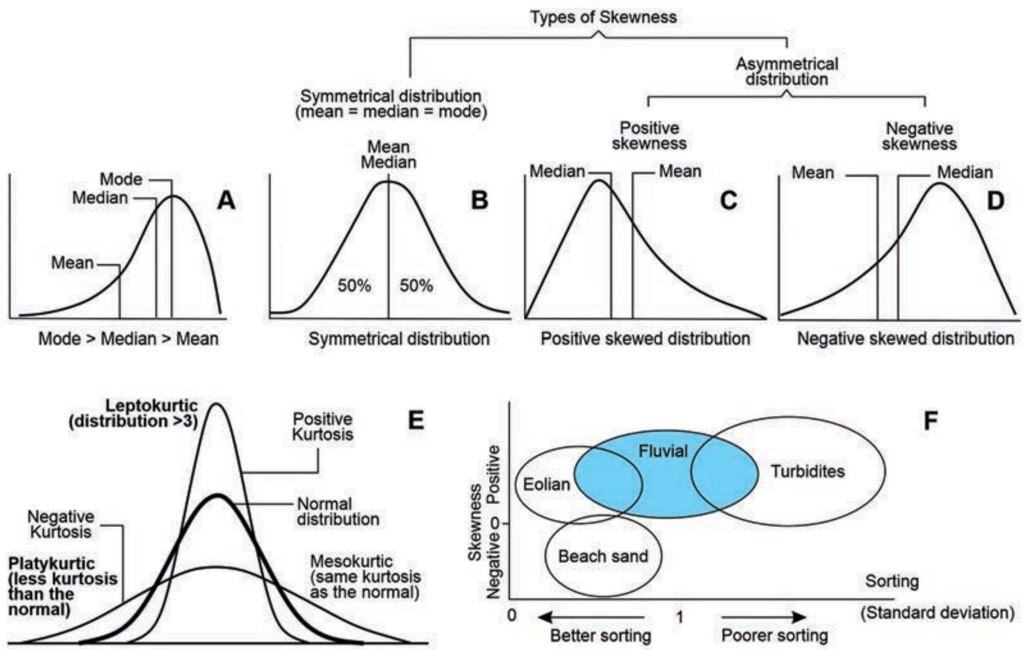

FIGURE 2.4 Statistical measures. A: Mean, median and mode. B–D: Skewness. E: Kurtosis. F: Spread of sedimentary deposits in context of skewness and kurtosis.

and other half are finer. It is the diameter corresponding to the 50% mark on the cumulative curve and is expressed either in Φ or mm (Md Φ or Md mm) (see Figure 2.4A). Mode (MO) is the most frequently occurring particle diameter and is the diameter corresponding to the steepest point (point of inflection) on the cumulative curve; it corresponds to the highest point on the frequency curve (Figure 2.4A).

Skewness measures the asymmetry of a distribution and indicates whether the distribution is skewed to the left (negative skewness) or to the right (positive skewness), or if it is approximately symmetrical (zero skewness) (see Figures 2.4B–D). Zero skewness indicates a symmetrical distribution, where the number of smaller and larger grain sizes are approximately equal (Figure 2.4B). Positive skewness means that the tail of the distribution is longer on the right side, indicating an excess of larger grain sizes (see Figure 2.4C). This is observed in sediment deposits influenced by the deposition or accumulation processes that preferentially retain larger grains. Negative skewness means that the tail of the distribution is longer on the left side, indicating an excess of smaller grain sizes (see Figure 2.4D). This is observed in sediment deposits influenced by erosion or transport processes that preferentially remove larger grains.

Fluvial sediments normally show poor sorting and positive skewness, i.e., have a wide spread towards the finer grain sizes (higher phi values) and a sharp delimitation at large grain-sized end (Figure 2.4F). Major variations in river flow velocity may occur during floods and cause poor sorting in river sediments. Eolian (wind) sediments also have positive skewness as there is an upper limit of coarse grain size that wind can transport. However, there may be a tail of finer grain size in an eolian dune sediment as fine-grain sediments are protected from erosion and transport under the armor of relatively larger sand-sized grains. Eolian sediments are well to very well-sorted (Figure 2.4). Beach deposits, on the other hand, are negatively skewed, i.e., the distribution curve shows a definite lower limit of grain size where there is often a tail of larger grain size, i.e., granules and pebbles. On beaches, repeated breaking of waves takes sediments in suspension and finer sediments (fine sand, silt, clay) are swept away. Coarse grains, particularly medium to coarse sand, rapidly settle down from suspension and get deposited on the beach again. Hence, beach sediments are well-sorted.

Kurtosis measures the degree of peakedness or flatness of a distribution as compared to a normal one (Figure 2.4E). Kurtosis provides information about the presence of outliers or extreme values within the data. Normal distribution has a kurtosis value of zero, indicating a bell-shaped curve with no outliers (Figure 2.4E). Negative kurtosis means that the distribution is flatter and has lighter tails than a normal distribution (Figure 2.4E). This indicates a lack of extreme values or outliers in the grain-size data. Positive kurtosis means that the distribution has a sharper peak and heavier tails than a normal distribution (Figure 2.4E). This indicates the presence of more extreme values or outliers in the grain-size data.

Sediments deposited from suspension are poorly sorted and positively skewed such as the turbidite sediments. The sediments remain in suspension in turbidite currents due to fluid turbulence that results in suspension settlement of sediments, hence, the deposits are characterized by poor sorting and positive skewness (Figure 2.4F). Although this character is nearly similar to fluvial deposits, the distinction between these two products can be made based on average grain size; the average grain size of fluvial sediments is much coarse as compared to those of turbidite sediments (Figure 2.4F).

2.2.2.2.4 Moments of Particle Size Distribution

The moments of a particle size distribution refer to statistical measures that describe the distribution of grain sizes within a sample. These moments provide information about average grain size, the spread or variability of grain sizes, and the presence of extreme values or outliers. Some commonly used moments include:

Mean (first moment): The arithmetic mean is the average grain size and is calculated by summing the individual grain sizes and dividing by the total number of grains. It provides a measure of the central tendency of the distribution (Figure 2.4A).

Median (second moment): The median is the middle value in the sorted list of grain sizes. It represents the 50th percentile of the distribution and provides a measure of the central tendency that is less influenced by extreme values (Figure 2.4A).

Mode (third moment): The mode is the most frequently occurring grain size in the distribution. It represents the peak or the highest point of the distribution and can provide insights into the dominant grain-size class (Figure 2.4A).

Standard Deviation (second central moment): The standard deviation measures the spread or variability of grain sizes around the mean. It provides information about the degree of dispersion within the distribution.

Skewness (third central moment): Skewness measures the asymmetry of the distribution. Positive skewness indicates a tail to the right, suggesting the presence of larger grain sizes, while negative skewness indicates a tail to the left, suggesting the presence of smaller grain sizes (Figures 2.4B–D).

Kurtosis (fourth central moment): This measures the peakedness or flatness of the distribution compared to a normal distribution. Positive kurtosis indicates a sharper peak and heavier tails, suggesting the presence of extreme values or outliers, while negative kurtosis indicates a flatter distribution with lighter tails (Figure 2.4E).

These moments provide a comprehensive description of the grain-size distribution and are commonly used to characterize sedimentary deposits, assess sediment transport processes, and interpret depositional environments.

2.2.2 Grain Morphology

The morphology of grains has three aspects: shape (or form) determined by various ratios of the long, intermediate and short axes; sphericity – a measure of how closely the grain shape approaches

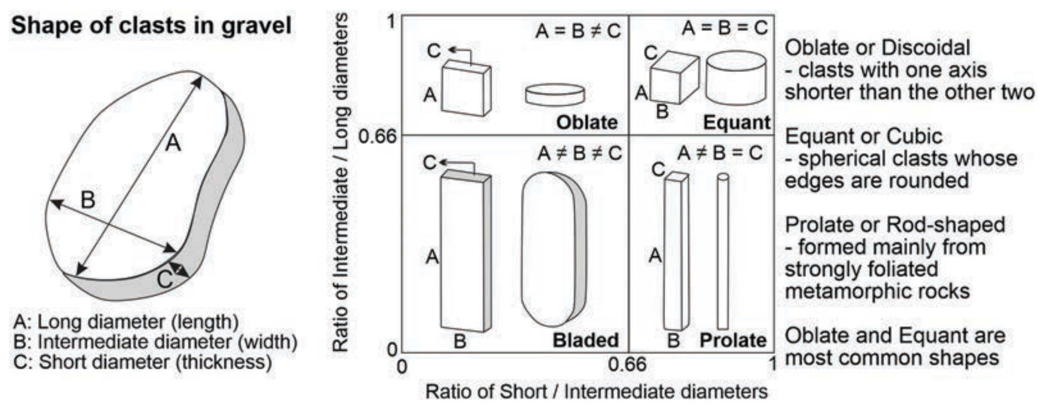

FIGURE 2.5 Grain morphology showing the axis ratios (A: long, B: intermediate, and C: short diameters or axes) and the four major classes of shapes: spheres, discs, blades, and rods.

that of a sphere; and roundness that concerns the curvature of the corners of the grain (see Figures 2.5 and 2.6).

2.2.2.1 Particle Shape

Based on ratios involving the long (A axis), intermediate (B axis), and short diameters or axes (C axis), four major classes of shapes are noted: spheres, discs, blades, and rods (see Figure 2.5). These terms are useful for describing clast shape in gravels, conglomerates, and breccias. The shape of pebbles is largely a reflection of the composition and any planes of weakness (such as bedding/lamination, cleavage, or jointing in the rock). Rocks of uniform composition and structure, such as granites, dolerites, and thick sandstones, will give rise to equant/spherical pebbles; thin-bedded rocks will generally form tabular and disk-shaped clasts, and highly cleaved or schistose rocks, such as slates, schists, and some gneisses, will form bladed or rod-shaped pebbles (see Figure 2.5). Thus, the shape of particles is influenced by several factors, such as the parent material, weathering processes, transport mechanisms, and depositional environments. For example, in high-energy environments, particles are often more angular and less rounded due to limited transport and minimal abrasion, whereas the particles in low-energy environments tend to be more rounded and well-sorted due to extensive transport and abrasion.

2.2.2.2 Sphericity and Roundness

It is important to first understand the difference between grain sphericity and roundness (see Figure 2.6). Sphericity is the degree of roundness or circularity of sediment grain or clast; it is a measure of how closely a particle resembles a perfect sphere, whereas roundness refers to the shape of sediment grain or clast and describes the degree to which a grain is rounded or angular. Roundness is a measure of how smooth or rough the surface of a grain is (see Figure 2.6A). In simple terms, sphericity is a measure of the degree to which a particle approximates the shape of a sphere, and is independent of its size, whereas roundness is the measure of the sharpness of a particle's edges and corners (see Figure 2.6A).

The Krumbein roundness and sphericity chart has been used since the 1960s (Krumbein and Sloss, 1963) (see Figure 2.6B). By manually sieving the sample into its various size fractions, 20 particles from each fraction are examined under a microscope. Then the analyst visually compares the shape of the particles on each slide with shapes given in the Krumbein roundness and sphericity chart (Figure 2.6B) and assigns values of the silhouette on the chart which

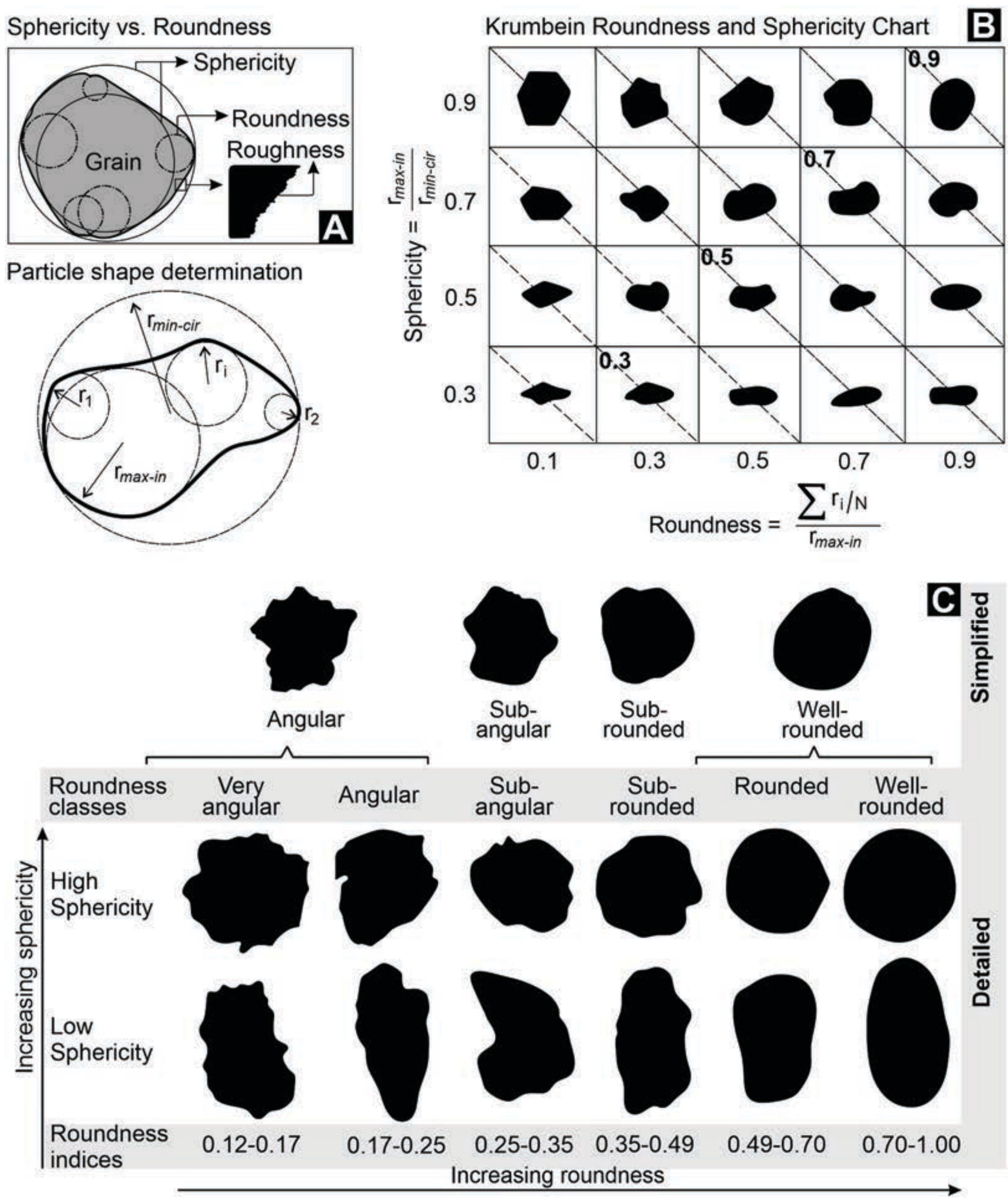

FIGURE 2.6 Sphericity and roundness. A: Sphericity is a measure of the degree to which a particle approximates the shape of a sphere. B: Krumbein roundness and sphericity chart. C: Sedimentary particles range from very angular to well-rounded and from high to low sphericity.

most closely matches the grains on each of the slide. This method has several drawbacks: it is very time-consuming; the improbability of such a small sample size being representative of an entire batch; different analysts have different subjective opinions of which grains pass the specifications; and the manual sieve size estimation and shape estimation are prone to errors. However, this is now being replaced by automated digital image analysis which uses software to extract data from digital images.

In terms of sphericity, the sedimentary particles range from very angular to well-rounded (Figure 2.6C). Angular particles have sharp edges and corners, indicating limited transport and minimal rounding (Figure 2.6C). Well-rounded particles, on the other hand, have smooth surfaces and lack sharp edges, suggesting extensive transport and significant rounding (Figure 2.6C). Sphericity is an important property in describing and classifying sedimentary rocks. It can also indicate the distance and energy of transport, as well as the source rock from which the particles originated. Thus, sphericity provides valuable information about the depositional environment, such as whether the sediments were deposited by fast-moving currents or slow-moving water bodies. For example, well-rounded and highly spherical particles (Figure 2.6C) are often associated with long-distance transport and high-energy environments like rivers or beaches, while angular particles are more commonly found in low-energy environments like lakes or stagnant water bodies.

Roundness is another important property that has been used to analyze and classify sedimentary rocks (Figure 2.6C). It also indicates the distance and energy of transport, as well as the source rock from which the particles originated. Sedimentary particles can range in roundness from very angular to well-rounded. Angular particles have sharp edges and corners, indicating limited transport and minimal rounding. Well-rounded particles, on the other hand, have smooth surfaces and lack sharp edges, suggesting extensive transport and significant rounding. The roundness of sedimentary particles provides valuable information about the depositional environment. For example, well-rounded and highly rounded particles are often associated with long-distance transport and high-energy environments like rivers or beaches, where they undergo abrasion and rounding due to the constant movement and collision with other particles. Angular particles, on the other hand, are more commonly found in low-energy environments like lakes or stagnant water bodies, where they experience limited transport and minimal rounding.

Thus, both sphericity and roundness are important properties in describing and classifying sedimentary rocks, as they provide valuable information about the processes of erosion, transportation, and deposition that have shaped the particles.

2.2.2.3 Grain Surface Texture

Particle surface texture refers to the characteristics of the outer surface of sediment grains or clasts. It describes the texture, roughness, and features present on the surface of individual particles. Surface texture is an important property that enables in the analysis and classification of sedimentary rocks, as it provides insights into the processes of weathering, erosion, transportation, and deposition, influencing the grain. For example, smooth and polished surfaces are often associated with long-distance transport and high-energy environments like rivers or beaches, where particles undergo abrasion and smoothing due to constant movement and collision with other particles. Rough and irregular surfaces indicate limited transport and minimal weathering, suggesting a shorter distance of transport or a depositional environment with lower energy. Surface texture can also provide clues about the type of weathering that has affected the grains. For example, grains with pitted surfaces may have been subjected to chemical weathering, while those with scratch marks or striations may have experienced mechanical weathering. Grain surface texture is also an important control on the economically significant relationship between porosity and permeability. Geologists use various methods to examine and describe particle surface texture, including visual inspection, microscopic analysis, and scanning electron microscopy. These techniques allow for the identification and characterization of surface features such as cracks, pits, grooves, ridges, and mineral coatings.

2.2.3 Grain Fabric and Orientation

Grain fabric, also known as sediment fabric, particle fabric or preferred orientation, is the arrangement and orientation or alignment of individual sediment grains within a sedimentary deposit. It is a key feature used to describe the texture and structure of rocks and provide insights into their formation and subsequent deformation processes. Grain fabric has also been used to interpret the depositional environment and subsequent geological history of an area. It also provides information about the direction and intensity of currents, the types of sedimentary structures (such as cross-bedding or ripple marks) formed, and the potential for post-depositional deformation or modification.

Grain fabric can be observed at various scales, ranging from microscopic to macroscopic. At the microscopic scale, it is noted using thin sections under a polarizing microscope, where the orientation of individual mineral grains is observed and analyzed. At the macroscopic scale, grain fabric is observed in hand specimens or in field outcrops, where the overall arrangement and orientation of mineral grains or of sedimentary layers or bedding planes is observed, respectively.

Grain fabric is influenced by various factors, including sediment transport, depositional processes, and subsequent digenetic or tectonic deformation. For example, in environments with strong currents, sediment grains become aligned parallel to the flow direction, resulting in a preferred orientation or a fabric known as imbrication (see Figure 2.7; detailed below). In other cases, the presence of sedimentary structures like cross-bedding or graded bedding indicates the direction and intensity of sediment transport. Deformation events, such as tectonic forces or metamorphic processes, cause significant changes in grain fabric. For example, during tectonic deformation, rocks undergo folding or faulting, leading to the rotation or reorientation of mineral grains. Metamorphic processes, such as recrystallization or pressure solution, also cause changes in grain fabric by promoting the growth of certain mineral grains or the realignment of existing ones. Thus, the analysis of grain fabric is important in understanding the mechanical behavior and deformation history of rocks. It also provides information about the stress and strain conditions during rock formation and subsequent deformation events. Additionally, grain fabric influences the physical properties of rocks, such as their strength, permeability, and anisotropy.

2.2.3.1 Imbrication

Imbrication refers to the arrangement and orientation of sediment grains within a sedimentary rock layer or deposit. It describes the overlapping and inclined orientation of individual grains, typically in a preferred direction (Figure 2.7). The degree of imbrication varies, ranging from loosely stacked grains with minimal overlap to tightly packed ones with significant overlap. The angle of inclination of the grains also varies, depending on the intensity and direction of the current. Imbrication commonly occurs in environments with strong currents, such as rivers, beaches, or shallow marine settings. The movement of water or wind causes sediment grains to become aligned parallel to the flow direction (Figure 2.7). As a result, the grains stack up and overlap each other, forming the characteristic imbricated fabric (Figure 2.7). Imbrication is commonly observed in sedimentary rocks, such as conglomerates or sandstones, where the imbricated fabric is preserved. It is observed at various scales, ranging from individual grains within a thin section to larger-scale features in outcrops or in the field. Thus, imbrication is an important sedimentary structure that provides valuable information about the depositional environment and the direction of sediment transport. It indicates the presence of energetic currents capable of moving and reorienting sediment grains. By studying imbrication, geologists can reconstruct the paleocurrent direction, which is the direction of the ancient currents that transported and deposited the sediment.

2.2.3.2 Grain Packing

Grain packing refers to the arrangement and distribution of individual grains within a sedimentary rock or deposit. It describes how the grains are packed together and the spaces or voids between them. In well-sorted sedimentary deposits, where grains are similar in size and shape, the packing tends to be more efficient, with minimal void spaces between grains. This leads to higher porosity and permeability, as there is more space for fluids to flow through the rock. On the other hand, poorly sorted deposits, with a wide range of grain sizes, tend to have less efficient packing, resulting in lower porosity and permeability. Grain shape also plays a role in grain packing. Angular or irregularly shaped

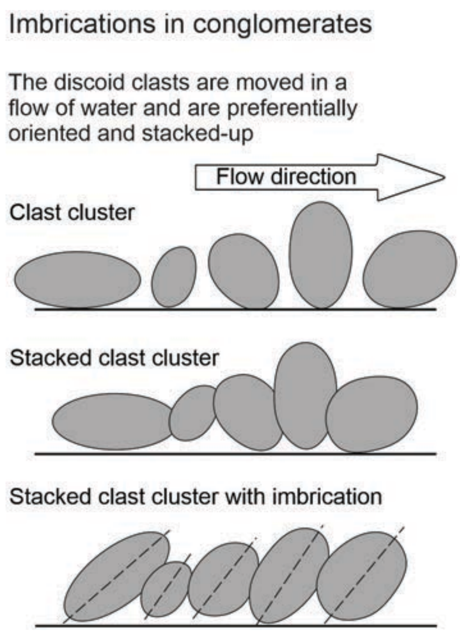

FIGURE 2.7 Imbrication. This shows the arrangement and orientation of sediment grains within a sedimentary rock layer or deposit, where the angle of inclination of the grains varies, depending on the intensity and direction of the current (flow direction).

grains do not fit together as tightly as rounded grains, leading to increased porosity and decreased packing efficiency. The presence of cement or binding materials, such as calcite or silica, also affects grain packing. These materials fill in the void spaces between grains, thus, reducing porosity and increasing the strength and cohesion of the rock. The packing of grains also has a significant impact on the physical properties and behavior of sedimentary rocks, affecting porosity, permeability, and compaction. The packing of grains is influenced by various factors, including grain size, shape, sorting, and the presence of cement or other binding materials. Thus, understanding grain packing is important in the field of reservoir geology, as it affects the storage and flow of fluids within sedimentary rocks, such as in oil and gas reservoirs. Grain packing also has implications for engineering geology, as the packing of grains influences the stability and strength of rock masses. The packing of grains is studied and quantified by using various techniques, including grain-size analysis, porosity measurements, and imaging techniques such as computed tomography scans.

2.2.3.2.1 Grain Contact

Grain contact refers to the points or areas where individual grains within a sedimentary deposit or rock come into contact with each other; grain contact describes the physical interaction between adjacent grains (Figure 2.8). Four types of grain contacts are noted: point, long, concavo-convex, and sutured contacts (Figure 2.8). In point contact, the grains touch only at a point (Figure 2.8). In long contacts, the long sides of grains touch each other (Figure 2.8). In concavo-convex contacts, the concave sides of grains penetrate into the convex sides of other grains (Figure 2.8). In sutured contacts, the wriggly boundaries with grains interpenetrating each other (Figure 2.8).

The nature of grain contact varies depending on factors such as grain size, shape, sorting, and the presence of cement or other binding materials. In well-sorted deposits, where grains are similar in size and shape, the contact between grains tends to be more point-to-point, with minimal intergranular space. This leads to a more efficient packing of grains and higher grain-to-grain

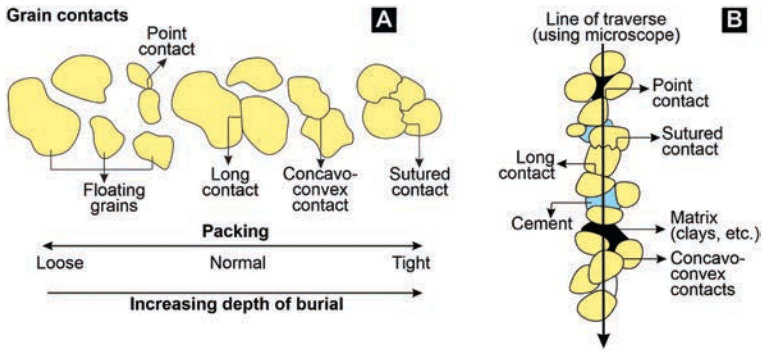

FIGURE 2.8 Grain contact. A–B: The diagrams show the points or areas where individual grains come into contact with each other.

contact. The type of contacts between grains determines the porosity, and hence the percentage of matrix, if pores are filled. Point and long contacts form at deposition, whereas concavo-convex and sutured contacts indicate progressive compaction (reduction of pore spaces) during burial diagenesis (Figure 2.8). A rock with prevailing long, concavo-convex, and sutured contacts has pore spaces, and matrix, less than 15%. This can be useful to recognize mature, low-matrix rocks such as arenites and orthoconglomerates. Efficient grain contact leads to lower porosity and permeability, as there is less space for fluids to flow through the rock. Grain contact also contributes to the mechanical strength and stability of the sediment or rock. In poorly sorted deposits, where there is a wide range of grain sizes, the contact between grains is more irregular and less efficient. Larger grains have limited contact with smaller grains, resulting in more intergranular space and lower grain-to-grain contact. The presence of cement or binding materials also influences grain contact. These materials fill in the intergranular spaces, thus enhancing the contact between grains and increasing the cohesion and strength of the sediment or rock. Grain contacts are studied by using various techniques, such as microscopic analysis, imaging techniques, and laboratory experiments. These methods allow for the quantification and characterization of grain contact, providing insights into the packing and physical behavior of sedimentary deposits or rocks.

2.2.3.2.1.1 Grain-Supported Fabric

The grain-support fabric is the arrangement and orientation of grains within a sedimentary deposit or rock (Figure 2.9A). It describes how the grains are supported or held in place by neighboring grains or the presence of cement or other binding materials (Figure 2.9B). The grain-support fabric can have important implications for the physical properties and behavior of sedimentary rocks. It affects parameters such as porosity, permeability, and mechanical strength (see Figure 2.10). In well-supported sedimentary deposits, grains are tightly packed and supported by neighboring grains, resulting in higher grain-to-grain contact and lower porosity resulting in lower permeability, as there is less space for fluids to flow through the rock (see Figure 2.10). Well-supported deposits are typically more mechanically stable and resistant to deformation. In poorly supported deposits, grains are loosely packed, with limited grain-to-grain contact and higher porosity resulting in higher permeability, as there are more interconnected void spaces for fluid flow (Figure 2.10). Poorly supported deposits are typically more mechanically weak and prone to deformation. The presence of cement or binding materials enhances grain support by filling in the intergranular spaces and binding the grains together (see Figure 2.9B). This increases the mechanical strength and cohesion of the rock, reducing porosity and permeability. Thus, understanding the grain-support fabric is important in various fields of geology, such as reservoir geology, where it influences fluid flow and storage in sedimentary rocks, and engineering geology, where it affects the stability and strength of rock masses.

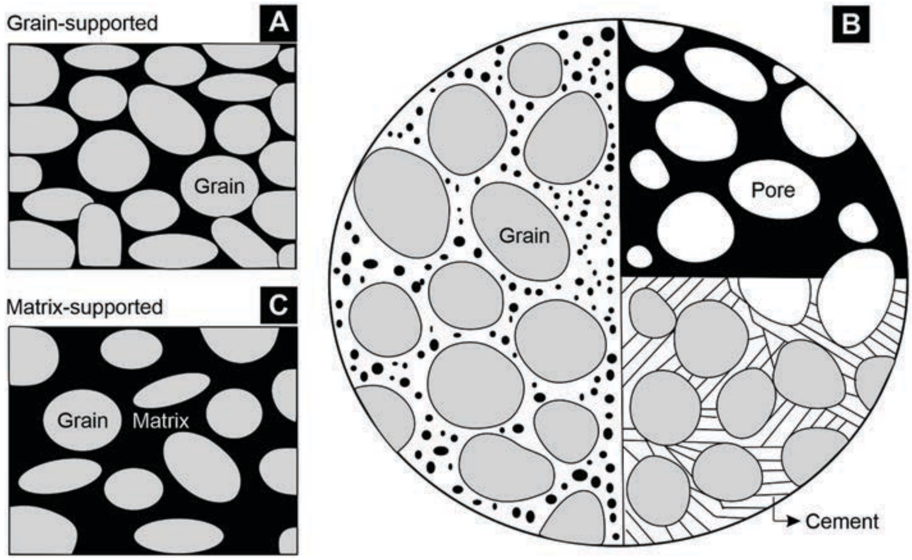

FIGURE 2.9 Fabric type. A: Grain-support. B: Matrix-supported fabric. C: Relationship of pores, cement, and matrix within a sedimentary deposit or rock.

By studying the grain-support fabric, geologists can gain insights into the depositional processes, diagenesis, and subsequent behavior of sedimentary deposits or rocks.

2.2.3.2.1.2 Matrix-Supported Fabric

The matrix-supported fabric is the arrangement and composition of grains and matrix within a sedimentary deposit or rock (Figure 2.9C). It describes the dominance of the matrix material in holding the grains together, rather than the grains being supported by grain-to-grain contact (Figure 2.9C). The grains are typically dispersed within the matrix, and their contact with each other is limited (Figure 2.9C). The matrix material consists of various clay minerals, silt, sand, or even cement. It can be fine-grained or coarse-grained, depending on the composition and size of the matrix particles. Matrix-supported fabric is commonly found in sedimentary deposits that have undergone significant diagenesis, where the original grains have been partially or completely dissolved or replaced by the matrix material. This occurs due to cementation, compaction, or chemical alteration. The presence of a matrix significantly influences the physical properties of the sediment or rock. As the matrix material fills the void spaces between the grains, restricting the ability of fluids to flow through the rock, leading to reduced porosity and permeability. Matrix-supported fabric also affects the mechanical strength and stability of the sediment or rock. The matrix material provides cohesion and support, enhancing the overall strength of the deposit. However, if the matrix material is weak or poorly cemented, the sediment or rock is more prone to deformation or failure. Thus, by studying matrix-supported fabric, geologists gain insights into the diagenetic processes, depositional environments, and subsequent behavior of sedimentary deposits or rocks.

2.2.4 Textural Maturity

Textural maturity is the degree of rounding and sorting of sediment grains within a sedimentary rock. It is an important characteristic in determining the distance of sediment transport and the energy of the environment in which the sediment was deposited. Texturally mature rocks have well-rounded

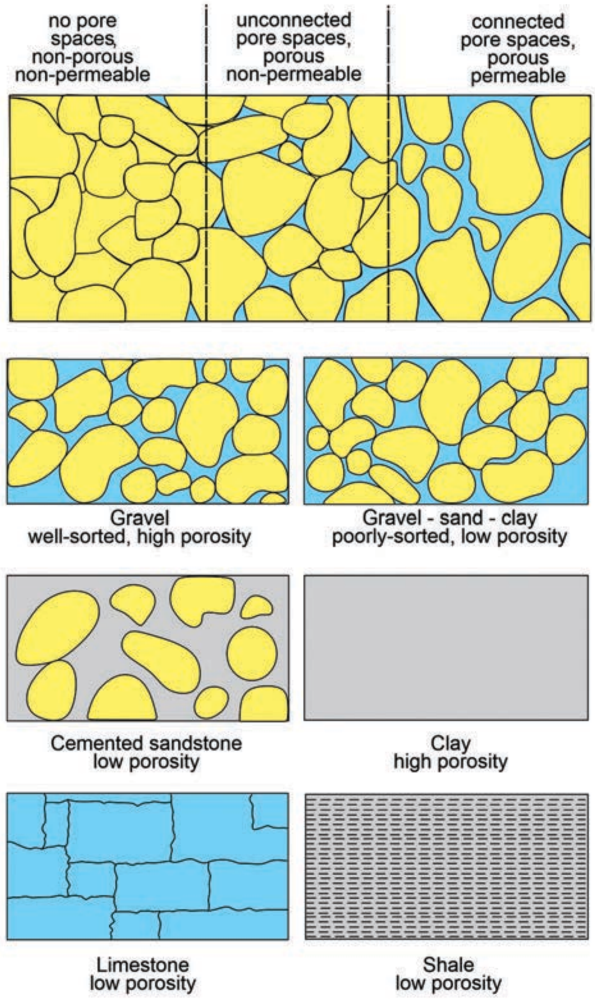

FIGURE 2.10 Porosity and permeability. Rocks with high porosity and high permeability, such as sandstones or fractured limestone, are considered good reservoir rocks, as they can store and transmit large volumes of fluids. In contrast, rocks with low porosity and low permeability, such as shales, are typically poor reservoirs, as they have limited storage and flow capacity.

and well-sorted grains, indicating that they have been transported over long distances and deposited in high-energy environments such as rivers, beaches, or wind-blown dunes. In contrast, texturally immature rocks have angular and poorly sorted grains, suggesting that they have been transported over shorter distances and deposited in low-energy environments such as lakes or deep ocean basins. In general, as sediment is transported by water or wind, it undergoes erosion, abrasion, and attrition, resulting in the rounding of grains. Sediment grains that have been transported for longer distances are more rounded; than those that have undergone less transport, and are more angular. On the other hand, well-sorted rocks have grains that are similar in size, whereas poorly sorted rocks have a wider range of grain sizes. Sediments that have been transported for longer distances tend to be better sorted, as the finer grains are carried further away, leaving behind coarser grains. Thus, the degree of sorting, the roundness and the matrix content contribute towards the textural maturity of the sediment.

2.2.4.1 Porosity and Permeability

Porosity and permeability (see Figure 2.10) are important properties of rocks and sediments that determine their ability to store and transmit fluids, such as water, oil or gas. They play a crucial role in various geological processes, including groundwater flow, and hydrocarbon reservoir characterization. Porosity refers to the amount of empty space, pores or voids within a rock or sediment and is expressed as a percentage and represents the ratio of the volume of pores or voids to the total volume of the rock or sediment (see Figure 2.9B). Porosity can be primary, formed during the deposition or formation of the rock, or secondary, created by subsequent processes such as fracturing or dissolution. Permeability, on the other hand, refers to the ability of a rock or sediment to transmit fluids through interconnected pore spaces (see Figure 2.10). It is a measure of the ease with which fluids can flow through a material. Permeability is influenced by the size, shape, and connectivity of the pores, and the viscosity of the fluid. It is expressed in units of darcy or millidarcy. High porosity indicates a greater volume of voids, which means there is more space available for fluid storage (see Figure 2.10). But a rock with high porosity may not necessarily be highly permeable if the pores are not well connected or are too small for the fluid to flow (see Figure 2.10). Conversely, a rock with low porosity may have high permeability if the pores are well connected and large enough to allow the movement of the fluid (see Figure 2.10). Thus, the combination of porosity and permeability is critical for understanding fluid flow in subsurface environments. Rocks with high porosity and high permeability, such as sandstones or fractured limestone, are considered good reservoir rocks, as they can store and transmit large volumes of fluids. In contrast, rocks with low porosity and low permeability, such as shales or unfractured igneous rocks, are typically poor reservoirs, as they have limited storage and flow capacity.

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3 Sediment Transport Mechanisms

3.1 INTRODUCTION

Sediment transport is the movement of sediment particles by various natural processes such as gravity, water (fluvial transport), wind (aeolian transport), ice (glacial transport), and waves and currents (marine transport). These sediment transport mechanisms often work in combination, with multiple processes contributing to the overall movement and deposition of sediments. The specific mechanisms involved depend on factors such as the energy of the environment, the size and composition of the sediment particles, and the availability of the medium (such as water, wind, etc.). The source materials (largely siliciclastic rocks) include, among others, conglomerates, sandstones, shales, silicate minerals, rock fragments, and pyroclastic materials that are weathered from older rocks. These source materials are eroded from highlands, transported to lower elevations, and deposited. Mass-wasting processes like slides and slumps facilitate the initial short-distance downslope sediment movement to sites where other transport processes take over such as fluid flows (moving water) or sediment-gravity flows, such as mudflows. In general, sediment transport takes place under a variety of conditions, both subaerially by wind and certain kinds of sediment-gravity flows, and subaqueously in rivers, lakes, and the ocean by currents, waves, tides, among others. Hence to better understand sediment transport, some basic understanding of fluid flow and sediment-gravity flow dynamics is needed. The involved mechanisms by which sediment is transported are briefly enumerated below.

3.2 BASICS OF FLUID FLOW

Fluid density and fluid viscosity are the two basic properties of water, and water containing suspended sediments; differences in them affect the ability of fluids to erode and transport sediments. Fluid density, commonly referred to as p (rho), is defined as mass per unit fluid volume. Density affects the magnitude of forces that act within a fluid, on the bed, and the rate at which particles fall or settle through a fluid (they settle slower in denser fluids). Density, under the influence of gravity, impacts the movement of fluids, downslope. Different fluids have different densities and density increases with decreasing temperature of a fluid. Hence, this density difference influences the abilities of water and air to transport sediment; water can transport particles of much larger size than those by wind. Fluid viscosity is a measure of the ability of fluids to flow. Fluids with low viscosity (such as water) flow readily whereas those with high viscosity flow sluggishly (such as honey). Like density, viscosity increases with decreasing temperature of the fluid. Increasing viscosity reduces turbulence, thus slowing the rate at which particles settle through water, thereby greatly influencing the transport of suspended sediments. Decreased turbulence also reduces the ability of running water to erode and entrain sediments. Two types of fluid flow patterns occur in geological processes (such as

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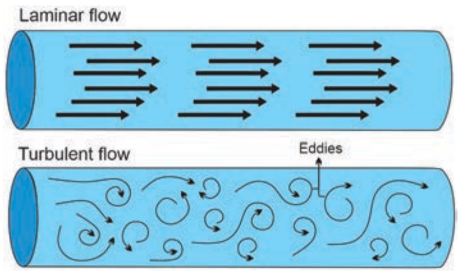

FIGURE 3.1 Laminar and turbulent flow. Laminar flow is a smooth and stable flow pattern in which the fluid moves in parallel layers or streamlines with minimal mixing or turbulence. Turbulent flow, on the other hand, is a chaotic and irregular flow pattern characterized by the mixing and swirling (eddies) of fluid particles.

sediment transport), subject to fluid viscosity, flow velocity, and the roughness of the bed over which the flow takes place – laminar and turbulent flow (see Figure 3.1). These are briefly described below.

3.2.1 Laminar versus Turbulent Flow

Laminar flow is a smooth and stable flow pattern in which the fluid moves in parallel layers or streamlines with minimal mixing or turbulence, and is characterized by low velocities (see Figure 3.2A). It occurs where fluid viscosity is high, flow is slow or is confined to narrow channels, over smooth beds (see Figure 3.2A).

Turbulent flow, on the other hand, is a chaotic and irregular flow pattern characterized by the mixing and swirling of fluid particles (see Figures 3.2B–C). In turbulent flow, the fluid moves in a random and disordered manner, with eddies forming throughout the flow (Figures 3.2B–C). Turbulent flow is associated with higher velocities and is noted where the fluid viscosity is low, the flow is fast, or there are obstacles or irregularities in the flow path (see Figure 3.2B). If flow velocity increases or viscosity of the fluid decreases, the stream is no longer maintained as a coherent stream but breaks up and becomes highly distorted, i.e., it moves as a series of constantly changing and deforming masses and the streamlines are intertwined in a very complicated manner (Figure 3.2C). This type of flow is called turbulent flow due to the transverse movement of fluid masses. These highly turbulent water masses are called eddies (see Figure 3.2C). Most flow of water and air under natural conditions is turbulent, although flow of ice and mudflows is largely laminar.

In terms of sediment transport, laminar flow is typically associated with the suspension and transport of smaller, fine-grained particles, such as silt or clay, where the flow velocities are relatively low. On the other hand, turbulent flow is more likely to occur during the transport of coarser, larger particles, such as sand or gravel, where the flow velocities are higher and the particles are more prone to being lifted and transported by turbulence. Thus, understanding the flow regime, whether laminar or turbulent, is important as it can influence the transport and deposition of sediments, and the behavior of fluids in geological formations.

The boundary conditions separating laminar and turbulent flow is given by two dimensionless parameters, Reynolds and Froude numbers. Both are important in understanding fluid flow and sediment transport in various geological processes. They help determine the flow regime, stability of sediment deposits, and potential for erosion and sediment transport in rivers, coastal areas, and other geological environments. They are briefly enumerated below.

3.2.2 Reynolds Number

Reynolds number (R_e) is a dimensionless parameter that characterizes fluid flow for better understanding the behavior of sediment transport. It is used to determine the type of flow in a fluid,

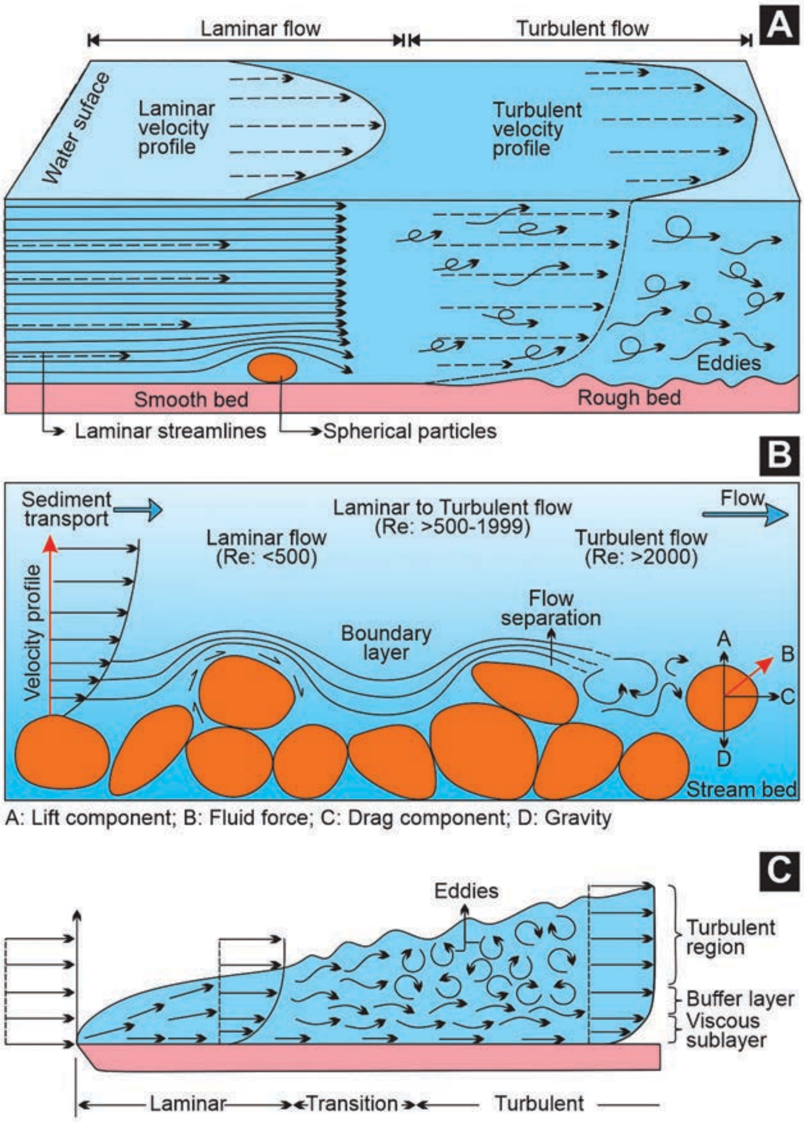

FIGURE 3.2 Laminar and turbulent flow. A–B: Laminar flow occurs when the flow is slow or is confined to narrow channels over a smooth bed and fluid viscosity is high. In turbulent flow, the fluid moves in a random and disordered manner, with eddies forming throughout the flow. Turbulent flow is associated with higher velocities and low fluid viscosity; the flow is fast, or there are obstacles or irregularities in the flow path. C: Turbulent flow and the formation of eddies. The dashed lines with arrows show different velocity profiles between laminar and turbulent flows.

whether it is laminar or turbulent. The fundamental differences in laminar and turbulent flows arise from the ratio of inertial forces to viscous forces. Inertial forces are related to the scale and velocity of fluids in motion, and tend to cause fluid turbulence. Viscous forces, which increase with increasing viscosity of a fluid, resist deformation of a fluid and thus tend to suppress turbulence. When viscous forces dominate, as in highly concentrated mudflows, Reynolds numbers are small and the flow is laminar (see Figures 3.2A–B). When the inertial forces dominate and flow velocity increases, as in most flow in rivers, the Reynolds numbers are large and the flow is turbulent (see Figures 3.2A–B). Thus, most flows under natural conditions are turbulent.

The Reynolds number (R_e) is calculated by dividing the product of the fluid velocity, characteristic length (channel width), and fluid density by fluid viscosity. Mathematically, it is expressed as: [$R_e = (\rho * v * L)/\mu$], where ρ is fluid density, v is fluid velocity, L is the characteristic length, and μ is fluid viscosity. At low Reynolds numbers (typically <2000), the flow is laminar, and is characterized by smooth and orderly flow patterns (see Figures 3.2A–B). At higher Reynolds numbers (>2000), the flow becomes turbulent, and is characterized by chaotic and irregular flow patterns with mixing and eddies (see Figures 3.2A–B). Thus, the Reynolds number is used to derive some idea of the magnitude of turbulence and predict whether the flow is laminar or turbulent. As the Reynolds number is a dimensionless parameter, it is also very useful in modeling natural systems.

3.2.3 Froude Number

The Froude number (Fr) is another dimensionless parameter that is used to better understand the behavior of fluid flow, particularly in open channels or rivers. It is calculated by dividing the fluid velocity by the square root of the product of the gravitational acceleration and a characteristic length (water depth or channel width). Mathematically, it is expressed as: $Fr = v/\sqrt{(g*L)}$, where: v is the fluid velocity, g is the gravitational acceleration and L is a characteristic length (e.g., water depth or channel width). The Froude number helps determine whether the flow is subcritical, critical, or supercritical. In subcritical flow (Fr <1), the flow velocity is less than the wave velocity, resulting in smooth and tranquil flow. In critical flow (Fr = 1), the flow velocity is equal to the wave velocity, leading to a transition between subcritical and supercritical flow. In supercritical flow (Fr >1), the flow velocity is greater than the wave velocity, resulting in rapid and turbulent flow. Thus, the Froude number is used to define the critical velocity of moving water.

3.2.4 Boundary Layers and Velocity Profiles

When a fluid flows over a streambed (boundary), the flow, close to the boundary is retarded by the frictional resistance of the boundary; this zone of resistance is called a boundary layer (Figure 3.2B). The flow within boundary layers may be laminar or turbulent or may grade from laminar to turbulent (Figure 3.2B). As greater shear stress is required to maintain a particular velocity gradient in a turbulent flow, turbulent flow-velocity profiles have different shapes than do laminar flow-velocity profiles; the latter depends upon the nature of the bed over which the flow takes place (Figure 3.2B). For smooth beds, there is a thin layer close to the bed boundary where molecular viscous forces dominate. Due to molecular adhesion, the fluid next to the boundary remains stationary, but successive overlying layers slide relative to those beneath, at a rate dependent upon the fluid viscosity (see Figure 3.2B). Flow within this thin layer tends toward laminar, but it has slower and faster moving fluid, and hence is not truly laminar. This layer is called the viscous sublayer, or laminar sublayer (Figure 3.2C). Over a rough or irregular bed such as coarse sand or gravel, the viscous sublayer is destroyed by irregularities (eddies) that extend through the layer into the turbulent flow (Figure 3.2B). Hence, the flow is affected by the roughness of the boundary (Figures 3.2B–C). Obstacles on the bed generate eddies at the boundary; the larger and more abundant the obstacles are, the more turbulence is generated (Figure 3.2B). Most sediment transport takes place within boundary layers. Hence, the turbulent boundary flow is more effective in eroding and transporting sediments than the laminar flow. The presence or absence of a viscous sublayer is an important factor in initiating grain movement, i.e., very small grains that lie entirely within the viscous sublayer are very difficult to move.

3.3 FLUID TRANSPORTATION

The erosion and entrainment of sediments from the bed, and the sustained downcurrent movement of sediments along or above the bed are two steps in the transport of sediments by fluid flow. Entrainment is lifting the resting grains from the bed, or putting them into motion. More energy is required to initiate particle movement than to keep them in motion after entrainment. Once particles are lifted from the sediment bed into the overlying water, the rate at which they fall back to the bed (the settling velocity) is an important factor in determining how far the particles would travel downcurrent before they again come to rest on the sediment bed.

3.3.1 Particle Entrainment by Currents

The entrainment by currents is the process of picking up and carrying sediments or rocks within the fluid flow. As the current moves, it entrains particles of various sizes and transports them downcurrent. Contextually, as the velocity and shear stress of a fluid moving over a sediment bed increases, a critical point is reached and the grains begin to move downcurrent; smaller and lighter grains move first. This threshold for grain movement depends on factors such as boundary shear stress, fluid viscosity, particle size, shape, and density, and indirectly on the velocity of the flow. In simple terms, forces caused by gravity act downward to resist grain motion and hold them against the bed (Figure 3.2B). Fine, clay-sized particles have increased resistance to movement due to their cohesive nature caused by the electrochemical bonds between them. Both the drag (drag component) and lift forces (lift component) enable the resistance to movement imposed by these retarding factors, over the grains, to be overcome. For the grain to move, both drag and lift forces combine to produce the total fluid force and this force must be large enough to overcome the gravity and frictional forces, thereby enabling grain movement (Figure 3.2B).

The simplest plot that shows the initiation of grain movement is the Hjulström diagram (see Figure 3.3). It shows the critical velocity needed for the movement of quartz grains on a plane bed at a water depth of 1 m. The critical entrainment velocity for grains larger than about 0.5 mm increases gradually with increasing mean grain size, whereas the entrainment velocity for consolidated clay and silt grains smaller than 0.05 mm increases with decreasing grain size (Figure 3.3). This is due to the increasing cohesion of finer sized grains, making them more difficult to erode than larger, noncohesive one (Figure 3.3). Also, very small grains lie within the viscous sublayer, where little grain movement takes place.

3.3.2 Particle Settling Velocity

The particle settling velocity plays a crucial role in the transport of sediments and the formation of various geological features. Some key aspects of its role are: (a) the particle settling velocity determines the rate at which sediments settle out of a fluid medium; this process is called sedimentation. It is responsible for the formation of sedimentary rocks, as the settling particles accumulate and eventually become lithified over time, (b) the settling velocity of sediment particles influences their ability to be transported by erosional agents such as water, wind, or ice. Sediments with higher settling velocities are more resistant to transport and tend to settle out in low-energy environments, such as in lakes or deep ocean basins. Whereas sediments with lower settling velocities are transported over long distances by high-energy agents like rivers or ocean currents, (c) sediment particles with different settling velocities segregate during transport, leading to the process of sorting. This results in the separation of sediment particles based on their size and density. For example, in a river, the coarser and denser particles settle first, while the finer and lighter ones remain suspended for longer distances. This sorting process leads to the deposition of sediments in distinct layers, creating sedimentary structures like graded bedding, (d) sediment settling velocity

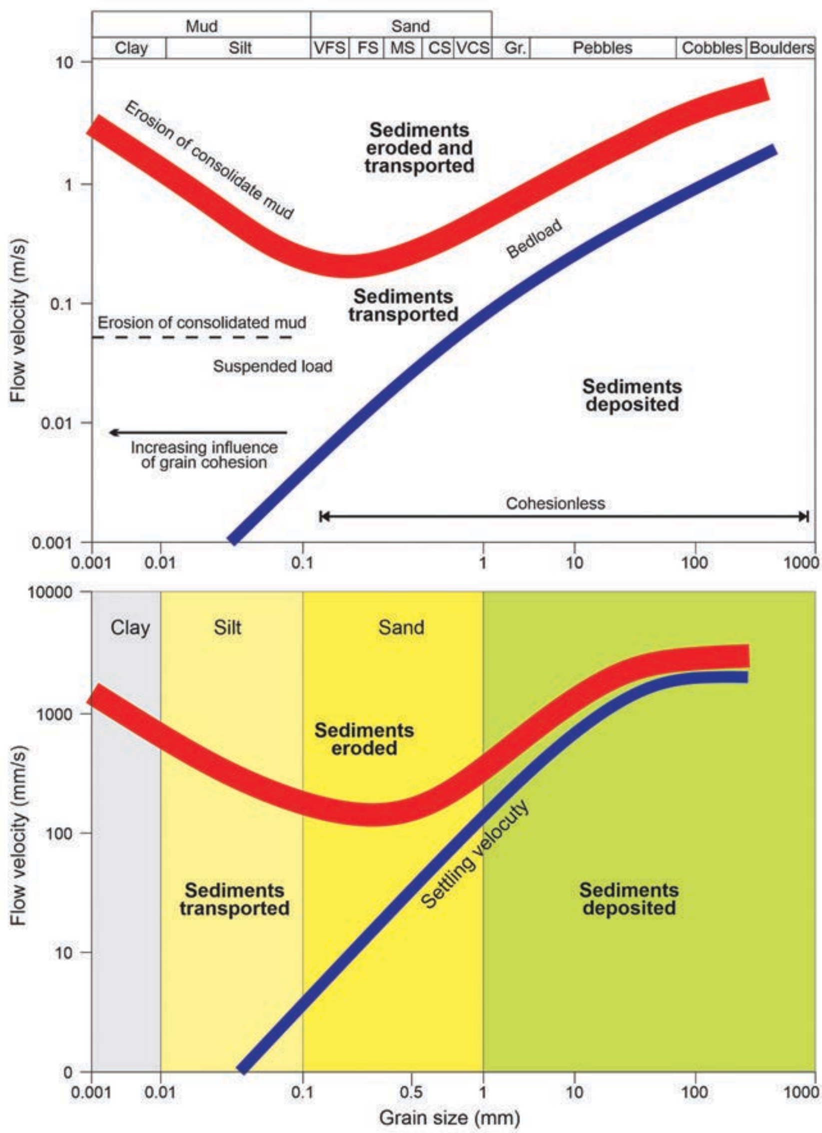

FIGURE 3.3 Hjulström curve diagram. A–B: The curve shows the relationship between sediment size and the velocity needed to erode, transport, or deposit sediments. The upper thick line shows the critical erosion velocity needed to kick start sediment erosion, whereas the lower thin line shows the fall or settling velocity, and between the two is the transportation of sediments. A big gap between the critical erosion and deposition line indicates that sediments will be transported further; the opposite happens for a small gap where a relative drop in velocity (critical fall velocity) causes sediments to be deposited. The curve indicates that sediments such as sand need low velocities to be dislodged from the bed as the sand particles are non-cohesive, i.e., they do not stick together. But clay and silt need higher velocities (as do cobbles and boulders) to be displaced or entrained, due to the cohesive nature of clay particles (i.e., they stick together). But once the clay particles are displaced from the bed, then they are carried further distances in suspension (= the big transport gap on the diagram). The diagram also illustrates that erosion (the picking up of sediments) requires higher velocities than transportation. It must also be kept in mind that the Hjulström curve also has limitations. It neglects some factors such as vegetation and gradient that influence particle movement. On gentler gradients, particles tend to resist movement more than on steeper terrains.

influences the formation of various sedimentary structures. For example, when sediment-laden water enters a standing body of water, such as a lake or an ocean, the sudden decrease in velocity causes sediments to settle rapidly resulting in the formation of deltas. Similarly, in coastal settings, the settling velocity of sediment particles affects the formation of beaches, sandbars, and spits, and (e) particle settling velocity can also serve as an environmental proxy. For example, the settling velocity of particles reflects the energy level of the flow, which in turn provides information about the depth, current speed, and turbulence of the water. By studying the settling velocity of sediment particles, depositional environmental conditions can be inferred and reconstructed.

3.3.3 Sediment Loads

Sediment load is the amount of sediment carried by a fluid, such as a river, glacier, or wind, and is measured as the volume or weight of sediment transported per unit time. Sediment loads can vary greatly depending on the erosional processes, the size and type of sediment grains, and the energy of the transporting medium. Once sediment has been eroded and put into motion, its transport downcurrent is a function of the settling velocity of the particle and the magnitude of the current velocity and turbulence. Coarse sediments, such as sand and gravel, move on or very close to the bed during transport as bed load, whereas finer ones are carried higher up in the main flow above the bed forming the suspended load (Figure 3.4A). These are briefly enumerated below.

3.3.3.1 Bed Load Transport

Bed load transport (also called traction transport) is the movement of sediment particles along the bed of a river or stream. It is one of the three main modes of sediment transport, along with suspended load (see Figure 3.4A). Bed load transport occurs when sediment particles are too large (larger than sand size) or are too heavy to be carried in suspension by the flowing water and instead roll and slide (Creep), or bounce (saltation) along the riverbed (see Figure 3.4A). Thus, the size and shape of sediment particles play a significant role in bed load transport (see Figure 3.4B). Larger, more angular particles are less likely to be entrained in the water flow and are more likely to move as bed load (Figure 3.4A). Smaller, rounded particles are more easily transported in suspension as suspended load (Figure 3.4A). The threshold size for bed load transport depends on water velocity, sediment density, and bed roughness (see Figure 3.4B).

Creep results from the grains being pushed to short distances along the bed in the downcurrent direction due to the impact of other moving grains (see Figure 3.4A). In saltation, the grains, particularly sand-sized ones, move in intermittent contact with the bed. Thus, the saltating grains move by a series of jumps or hops, rising off the bed and then falling back to the riverbed (see Figure 3.4A). Saltation transport is an intermediate between traction and suspension transport, but since most saltating grains remain relatively close to the bed, it is considered under bed load transport (see Figure 3.4B).

The rate of bed load transport is influenced by factors such as water velocity, sediment supply, and the characteristics of the riverbed. Higher velocities increase the potential for bed load transport, as they generate greater shear stress on the bed (see Figure 3.4B). Sediment supply from upstream sources affects the amount of bed load available for transport. The roughness of the riverbed, influenced by the presence of large rocks or other obstructions, affects the rates of bed load transport. The bed load sediments are deposited when the water velocity decreases, such as in areas of reduced flow or in eddies. As the water slows down, it loses its ability to transport sediments, causing them to settle and accumulate on the riverbed. This process leads to the formation of sedimentary structures like bars or ripple marks. Additionally, bed load transport has a significant impact on the morphology of river channels. The movement of sediment particles along the riverbed erodes the channel, leading to the formation of pools, riffles, and other features. The deposition of bed load also contributes to the formation of bars and islands within the channel.

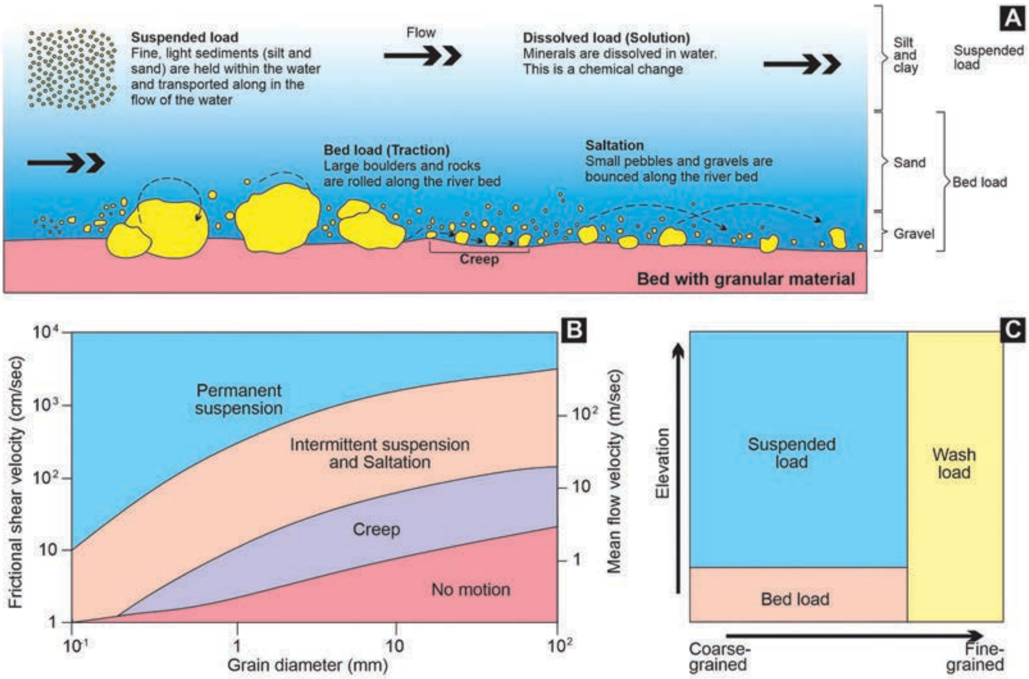

FIGURE 3.4 Sediment load. A: Coarse sediment such as sand and gravel move on or very close to the bed during transport as bed load, whereas finer ones are carried higher up in the main flow above the bed, forming the suspended load. B: Domains of transport modes for quartz sand. The frictional shear velocity is plotted against grain diameter. C: Wash load is that portion of sediment load that is carried in suspension. It consists of fine particles, typically smaller than sand-sized grains, that remain suspended in the fluid due to the velocity of the transporting medium.

3.3.3.2 Suspended Load Transport

In water, sediment particles are transported through traction (rolling or sliding along the bed), saltation (bouncing), or suspension (carried within the water column as suspended load) (Figure 3.4A). As the flow strength of a current increases, the particle trajectory becomes longer, more irregular, and higher up from the bed than that of the saltating particles (see Figure 3.4C). The upward components of fluid motion due to increased turbulence balance the downward gravitational forces on the particles, thus, allowing the particles to stay suspended above the bed; this behavior is called intermittent suspension (Figure 3.4B). It differs from saltation as the suspended particles are carried higher and remain off the riverbed for longer periods of time. Smaller particles have settling velocities that may be so low that they remain in nearly continuous suspension and are carried along at almost the same velocity as the fluid flow. Additionally, the size, shape, and density of sediment particles influence their transportability. Thus, smaller particles are more easily transported in suspension, while larger particles tend to move as bed load. Round and well-sorted particles are more easily transported than angular and poorly sorted ones. The density of sediment particles affects their ability to be entrained by the transporting medium, with denser particles requiring higher velocities to be transported. More so, the rate at which sediment is transported depends on various factors, including the velocity of the transporting medium, the availability of sediment supply, the characteristics of the sediment particles and the transporting environment. Higher velocities generally increase the potential for sediment transport, while reduced velocities lead to deposition. The sediment supply from upstream sources, such as erosion of hill-slopes or sediment input from tributaries, also influences the rates of transport.

3.3.3.3 Wash Load

Wash load is that portion of sediment load that is carried in suspension. It consists of fine particles, typically smaller than sand-sized grains that remain suspended in the fluid due to the velocity of the transporting medium (see Figure 3.4C). As water flows, it entrains and suspends small sediment particles, creating a turbid appearance; this suspended sediment is known as the wash load (see Figure 3.4C). The suspended sediments are derived either from upstream source areas or by the erosion of the bank, rather than coming from the streambed. The size and composition of the wash load vary depending on the characteristics of the sediment source and the erosional processes occurring in the catchment area. It typically consists of silt, clay, and organic matter, but can also include finer sand-sized particles that have very low settling velocities. As the wash load travels in continuous suspension at about the same velocity as the water, it is transported rapidly through river systems. The transport of wash load is influenced by the velocity and turbulence of the water. Higher water velocities increase the potential for wash load transport, while decreased velocities lead to deposition, as the suspended sediment settles out of the fluid. The transport of wash load plays a significant role in shaping river channels, as it erodes the channel bed and banks, and contributes to sediment deposition in areas of reduced flow or in floodplains.

3.3.4 Transport Paths

Sediment transport paths are the routes taken by sediments as they move from their source to their ultimate destination. These paths are influenced by the topography, geology, and hydrological conditions of the area. For example, in river systems, sediments are transported downstream along the river channel, with variations in velocity and direction, influenced by factors such as channel morphology, slope, and flow rate. The commonly noted transport mechanisms, wind-aided, glacial ice-aided, and gravity-aided, are briefly enumerated below.

3.3.4.1 Wind-Aided Transport

The processes involved in the entrainment and transport of particles by wind (eolian transport) are similar to those for water; but with different threshold values (see Figure 3.5). Smaller or fine-sand-sized particles are transported by wind and moved by traction (surface creep) and saltation; the dust-sized ones are done by suspension (see Figure 3.5). Sediment transport takes place only at relatively high wind velocities; the flow is commonly turbulent, and characterized by eddies moving with different velocities and directions. The suspended load carried by the wind is called the dustload, which is carried to heights of hundreds or even thousands of meters during a volcanic eruption (see Figure 3.5). These remain in suspension for relatively longer periods of time and are then spread over a very wide area, including the ocean basins. The very fine-grained components of deep-sea pelagic sediments are largely of windblown origin. Like water transport, eolian transport also results in the deposition of bedforms ranging in size from ripples (a few centimeters high) to dunes (hundreds of meters high).

3.3.4.2 Glacial Ice-Aided Transport

Glacial ice-aided transport is the movement of sediment by glaciers (Figure 3.6). The glaciers are powerful agents of erosion and transportation, and are capable of moving vast amounts of sediment over long distances, although this movement is very slow due to the high viscosity of glacial ice. Glacial erosion occurs by plucking and abrasion (Figure 3.6). Plucking occurs when the glacier freezes onto rock fragments and pulls them out as it moves, whereas abrasion is the grinding and

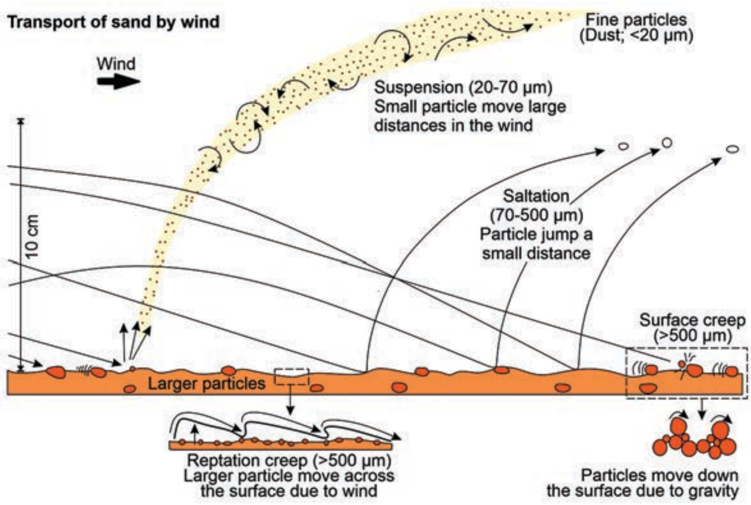

FIGURE 3.5 Sediment transport processes in deserts. In saltation, wind transports sand-sized particles in a bouncing or hopping motion. Surface creep (traction) is the movement of coarse sand and pebbles (up to 6x larger than saltating grains) as they slide and roll, impacting one another and transferring momentum. In suspension, the smaller particles, such as silt and dust, are carried in the air by wind. Grains less than 0.2 mm in diameter are suspended in air as turbulent eddies, and are carried as dust for thousands of meters upward, and 1,000 km downwind, forming dust storms.

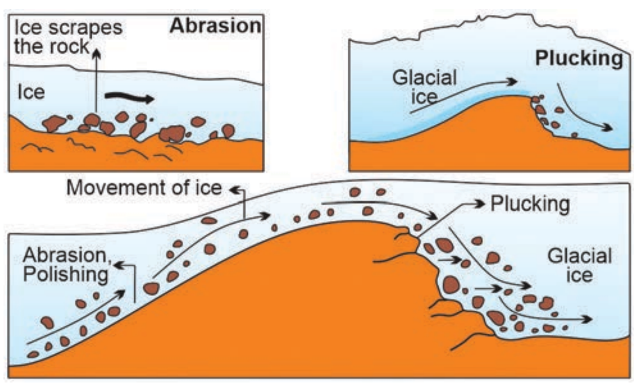

FIGURE 3.6 Abrasion, plucking, and the movement of ice.

scraping of the sediment-laden ice against the bedrock, resulting in the smoothing and polishing of the underlying surfaces. As the glacial ice moves, it entrains sediments. This sediment is either derived from the glacier's bed or from the surrounding landscape. Thus, glaciers pick up a wide range of sediment sizes, ranging from clay and silt to large boulders.

The glacial ice transports sediment in several ways such as by basal slip (basal sliding) and internal deformation (internal plastic flow) (see Figure 3.7). Basal slip occurs when the glacier slides over a thin layer of meltwater at its base, carrying the sediments along with it (see Figure 3.7A). Internal deformation, on the other hand, is the movement of sediment-laden ice within the glacier, causing the sediment to be transported along with the mass of ice (see Figure 3.7B). The sediments can also be transported at the surface of the glacier, either by being carried within crevasses or by being deposited on the ice surface and then transported through supraglacial streams or meltwater

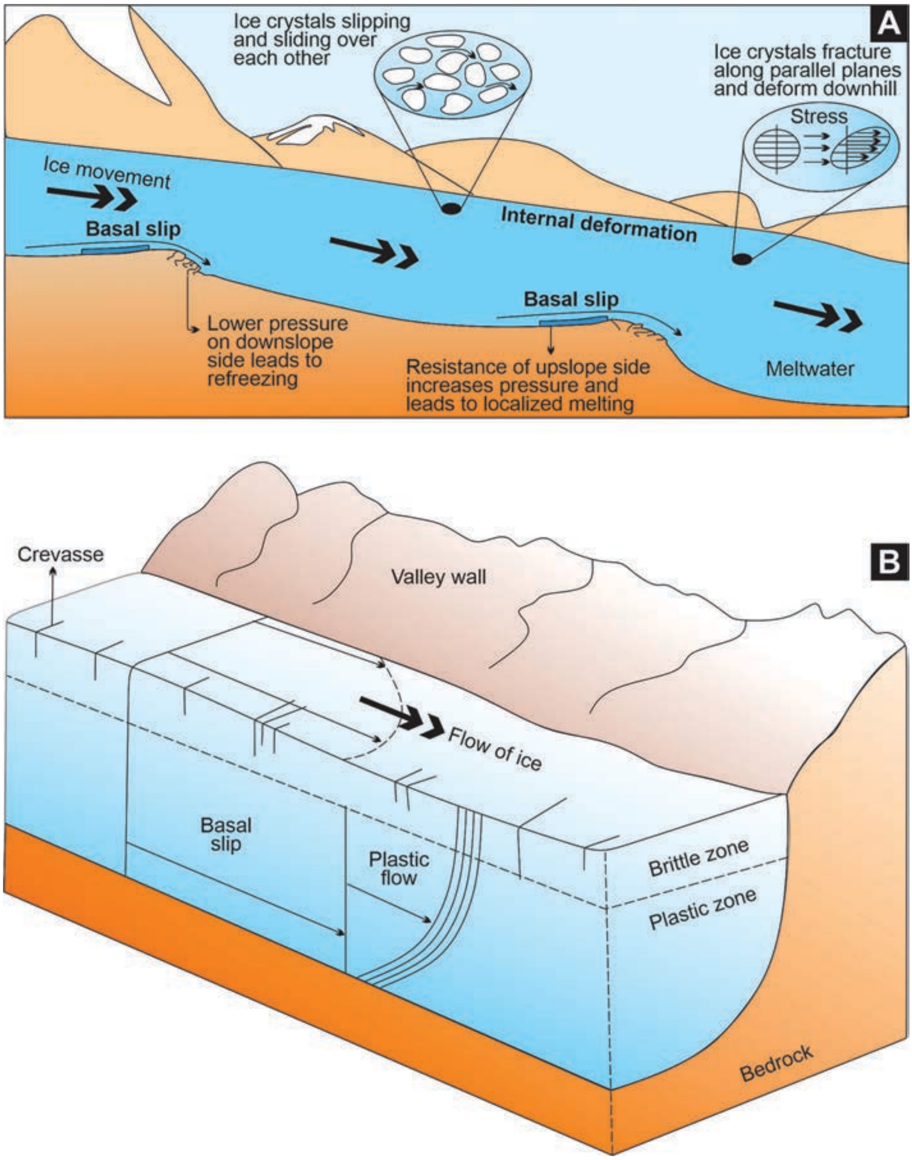

FIGURE 3.7 Basal slip (sliding) and plastic flow: the two mechanisms of glacier movement. A: Basal slip is sliding over the underlying surface. B: Plastic flow involves internal deformation within the ice. If a glacier is frozen to its bed, it moves only by plastic flow. Plastic flow is the dominant form of movement where all parts of the glacier are below the freezing point, including the base. The flow of glacial ice produces vertical to nearly vertical, wedge-shaped cracks called crevasses; these are formed as the upper portion of a glacier is brittle.

channels. As the ice melts, it releases these entrained sediments, leading to the formation of various landforms, such as moraines, drumlins, eskers, and outwash plains. These glacially transported sediments are typically poorly sorted and with a wide range of particle sizes; the sediments are angular and unsorted, reflecting the abrasive nature of the glacial ice. The presence of striations, polish, and other glacially induced features on the sediment grains provides another evidence of glacial transport.

3.3.4.3 Gravity-Aided Transport (Gravity Flows)

Gravity flows are the movement of glacial sediments under the influence of gravity; the force of gravity acts as the primary driving force for the movement of particles downslope. Gravity flows can occur in various environments, including terrestrial, subaqueous, and submarine, and are classified into different types based on the nature of the material being transported and the mechanisms of flow. The sediment movement under the influence of gravity creates a flow, and as and when the flow stops, the sediment load is deposited. Sediment-gravity flows are of interest as they are capable of rapidly transporting large quantities of sediments, including very coarse ones, and even to very deep ocean settings. In submarine environments, gravity movements range from the en-masse movement of sediments where fluids act largely to reduce internal friction by lubricating the grains, to those in which fluids play a role in supporting the sediment during transport, i.e., on a grain-by-grain basis. Broadly, the gravity movements are grouped into rock falls, slides, and sediment-gravity flows. Rock falls are free falls of blocks or clasts from cliffs or steep slopes, whereas slides are en-masse movements of rock or sediments caused by shear failure and occur with little accompanying internal deformation of the mass. Several factors influence the occurrence and behavior of gravity flows. These include slope gradient, amount and type of sediment or material being transported, presence of water or fluidization agents, and triggers such as rainfall, seismic activity, or volcanic eruptions. The properties of the sediment, such as grain size, cohesion, and internal friction, also play a role in determining flow characteristics. Gravity flows also result in the deposition of sediments and the formation of distinct landforms such as alluvial fans, deltaic deposits, submarine fans, turbidites, and slump deposits.

3.3.4.3.1 Turbidity Currents

Turbidity currents are gravity-driven flows that occur in subaqueous or submarine environments (Figure 3.8). They involve the downslope movement of sediment-laden water due to density differences when the sediment-laden water becomes denser than the surrounding water. This happens due to various triggers, such as earthquakes, underwater landslides, or the sudden release of sediment-laden waters from rivers or glaciers. When triggered, turbidity currents flow downslope under the influence of gravity (Figures 3.8A–B). These currents are generated by a variety of mechanisms, including sediment failure, storm-triggered flow of sand and mud into canyon heads, bed load inflow from rivers and glacial melt water, and flows during eruption of airfall ash. Turbidity currents are efficient agents of sediment transport and deposition in submarine environments. As they flow, turbidity currents pick up and suspend sediment particles, ranging from clay to sand and even larger fractions. The suspended sediment gives turbidity currents their characteristic cloudy appearance. As the flow slows down, the sediment settles out and is deposited on the seafloor, forming distinctive sedimentary deposits known as turbidites (see Figures 3.8C–E). Turbidity currents are also responsible for the formation of submarine canyons and deep-sea sedimentary deposits.

Turbidites are often characterized by graded bedding, which reflects the progressive settling of sediment particles of different sizes (Figures 3.8D–E). Bouma (1962) proposed an ideal turbidite sequence, called the Bouma sequence (see Figure 3.8E). This ideal sequence consists of five structural units; however, most turbidites do not contain all of these units (Figure 3.8E). Thick, coarse-grained turbidites tend to show well-developed A and B units, but C through E units are commonly poorly developed or absent (see Figure 3.8E). Thin, finer-grained turbidites may display well-developed C through E units, but A and B units are poorly developed or absent (see Figure 3.8E). Turbidity currents can shape the seafloor by eroding, transporting, and depositing sediment. Turbidity currents can also deposit sediment in the form of submarine fans (deep-sea fans), which are fan-shaped accumulations of sediment at the base of the continental slope (see Figure 3.8A). These features are important for understanding the geological history of an area and provide valuable information about past environmental conditions. Turbidity currents also contribute

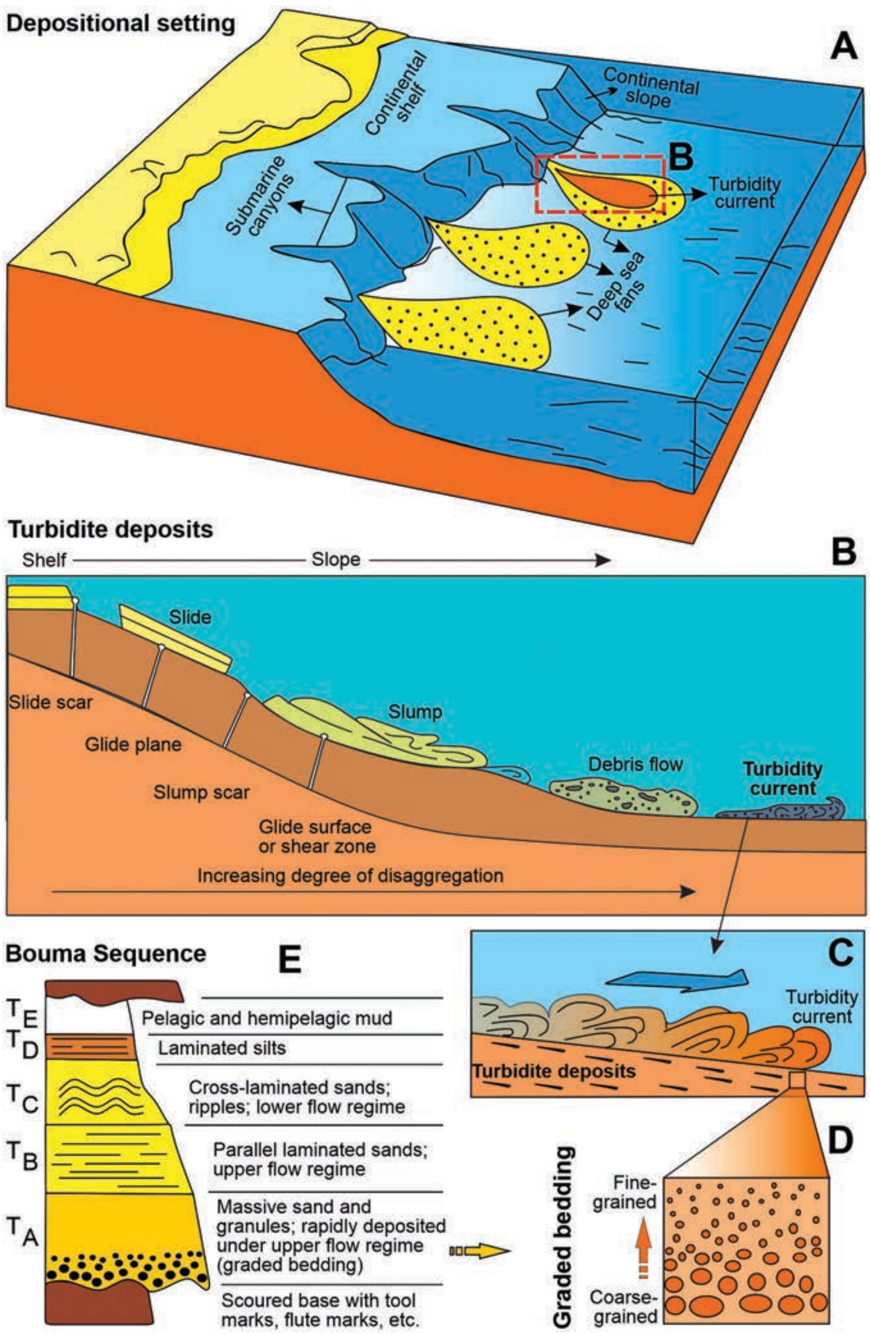

FIGURE 3.8 Bouma sequence and turbidites. A: Depositional setting. The turbidites form when a mass of sediments on the continental shelf tumbles down the steep continental slope and into deep water in a density-driven underwater landslide. Turbidites typically display an upward-fining sequence known as graded bedding. B: General process of turbidite formation through sediment-gravity flow triggered by slope failure. C–D: Close-up of turbidite current, leading to the formation of turbidite deposits. E: In a complete Bouma sequence, five divisions are noted: A) massive or graded bedding interval; B) sandy parallel laminations; C) rippled and/or convoluted bedding; D) fine and parallel interlaminations of silt and mud; and E) mud introduced by turbidity currents and hemipelagic background mud of the basin.

to the formation of hydrocarbon reservoirs in sedimentary basins, as they can transport and deposit organic-rich sediments.

3.3.4.3.2 Liquefied Flows

Liquefied flows, also known as liquefaction flows, are a type of gravity-driven flow that occurs when saturated or partially saturated sediment loses its strength and behaves as a fluid-like mass. These flows typically occur during earthquakes or other seismic events. When an earthquake occurs, the shaking of the ground causes the pore pressure within saturated or partially saturated sediments to increase. This increase in pore pressure reduces the effective stress and weakens the sediment, causing it to lose its strength and thus behave as a fluid. Once liquefaction occurs, the sediment can flow downslope under the influence of gravity. The flow can be rapid and can travel considerable distances, depending on the slope gradient and the volume of liquefied material. Liquefied flows exhibit characteristics of both debris flows and mudflows, with a mixture of sediment, water, and air. The deposits of liquefied flows are typically thick, poorly sorted sandy units. They are characterized by fluid escape structures, such as dish structures, pipes, and sand volcanoes (see Chapter 4 for sedimentary structures). Some liquefied flows may become turbulent as the flowing sediment mass is accelerated downslope and thus changes into turbidity currents. Liquefaction flows create depositional fans, lobes, or levees, depending on the prevailing topography and flow dynamics.

3.3.4.3.3 Grain Flows

Grain-flow deposits are formed by the movement of grains in a fluidized state. The fluidization of the sediment occurs when the interstitial water between the grains is pressurized, causing the grains to become buoyant and flow like a fluid. Grain flows are commonly found in submarine environments, but can also occur in terrestrial settings. Grain flows are characterized by the rapid movement of sediment grains in a cohesive mass. The grains are typically sand-sized particles, but can range from silt to gravel. The formation of grain flows is often associated with high-energy events such as underwater landslides, turbidity currents, or storm surges. These events mobilize large volumes of sediments and transport them downslope or along the seafloor. As the sediment moves, it picks up additional grains and incorporates them into the flow, leading to an increase in their size and volume. Grain flows often exhibit massive or laminated structures, with little to no visible bedding or layering. The lack of internal structure is due to the chaotic nature of the sediment movement and the depositional process. In some cases, grain flows display internal deformation features such as folding or faulting. Grain-flow deposits are commonly massively bedded (generally <5 cm thick) with little or no internal laminations and reverse grading in the base (as noted in some sandstone units). Reverse grading occurs when the fine-sized grains grade upward to coarse-sized ones. This is formed as the smaller particles filter down through the larger ones while they are in the dispersed state, through a process called kinetic sieving. Grain flow is similar to liquefied flow in many respects and may, in fact, grade into these flows. In contrast to liquefied flows, grain flow can occur under subaerial conditions as well as subaqueous conditions.

3.3.4.3.4 Debris Flows and Mudflows

Debris flows involve the rapid downslope movement of a mixture of sediments that typically occur in mountainous areas, triggered by heavy rainfall or the melting of the snow. Mudflows are similar to debris flows but consist primarily of fine-grained sediments, such as silt and clay. They are often associated with volcanic eruptions, heavy rainfall, or the failure of sediment-rich slopes. Both debris flows and mudflows are types of mass-wasting events that involve the rapid movement of sediments and water, downslope.

Debris flows, also known as debris avalanches or subaerial debris flows or mudslides, are fast-moving slurry like flows composed of highly concentrated, poorly sorted mixtures of sediment and water that behave in a different manner from fluid flows. They typically occur in mountainous or hilly regions with steep slopes and are triggered by heavy rainfall, snowmelt, or earthquakes. They are common in arid and semiarid regions where they are usually initiated after heavy rainfalls. They are also common in volcanic regions where volcanic debris becomes water-saturated during heavy rains that accompany eruptions or from melting ice and snow that accumulate on volcanic cones between eruptions. Debris flows travel at high speeds and have a viscous, flowing consistency. They occur in both subaerial and subaqueous environments; the latter possibly as a result of mixing at the downslope ends of subaqueous slumps. As subaqueous debris flows move rapidly downslope, they are diluted by mixing with more water, their strength is reduced, and they pass into turbidity currents. Debris-flow deposits are thick, poorly sorted, and lack internal layering. Hence, they typically consist of chaotic mixtures of particles that may range in size from clay to boulders. The large particles commonly show no preferred orientation. Flows that are composed predominantly of mud-sized grains are mudflows and those with a substantial mud fraction (>5 % by volume) are muddy debris flows. The grains in these mud-bearing debris flows are supported in a matrix of mud and interstitial water that has enough cohesive strength to prevent larger particles from settling but not enough strength to prevent flow. Debris flows that have a matrix composed predominantly of cohesionless sand and gravel, are called mud-free debris flows.

Mudflows are composed of fine-grained sediments, such as silt and clay, mixed with water. They are often triggered by intense rainfall on slopes with loose, unconsolidated sediments. Mudflows have a more fluid-like consistency as compared to debris flows, resembling a thick slurry. Mudflows can travel rapidly and can carry a significant amount of sediment and debris downstream. Lahars are mudflows that specifically occur in volcanic settings. They are triggered by the mixing of volcanic ash and water, often during rainfall or the melting of snow and ice on volcanic slopes. Both debris flows and mudflows cause significant erosion and deposition of sediments and can transport large boulders, trees, and other debris downslope, causing widespread damage and altering river channels. These deposits have a chaotic, jumbled appearance.

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Section II

Sedimentary Structures and their Environment of Deposition

This section is dedicated to sedimentary structures as they play a fundamental role in the interpretation of sedimentary processes, the first step in inferring depositional environments. They are also robust proxies for inferring sediment transport mechanisms, paleocurrent flow directions, bathymetry, and current velocity. They can also be used as an effective tool to identify the tops and bottoms of beds and infer whether the sedimentary sequences are in depositional stratigraphic order (i.e., “stratigraphic way-up” or “younging upwards”) or have been overturned by tectonic forces. Hence, the study of sedimentary structures uniquely brings together various disciplines, geology, physics, chemistry, and biology (biogenic sedimentary structures).

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4 Sedimentary Structures

4.1 INTRODUCTION

The sedimentary structures that are largely formed contemporaneously with sediment deposition, the focus of this chapter, are called primary sedimentary structures. Those that are formed sometime after deposition, i.e., during burial diagenesis, are secondary sedimentary structures; they are briefly discussed, herein. Sedimentary structures are more abundant in siliciclastic rocks, although they do occur in other sedimentary rock types such as limestones and evaporites. Sedimentary structures are hard to classify due to their varied origin and geometries. Two broad classifications are noted that are based on (a) the kind of mechanism that produces them (physical, chemical, and biogenic structures), and (b) the time of development relative to the time of deposition (primary and secondary sedimentary structures). The latter approach is followed here. The primary sedimentary structures are broadly classified into four categories: (a) depositional structures (generated mainly by deposition); (b) erosional structures (generated by processes that involve an episode of erosion followed by deposition); (c) deformation structures (generated by deposition followed by physical soft-sediment deformation); and (d) biogenic structures (biogenically mediated deposition or non-biogenic deposition followed by biogenic modification).

4.2 DEPOSITIONAL STRUCTURES

4.2.1 Stratification

To better understand depositional sedimentary structures, it is imperative to first understand stratification. It is the most useful aspect of sedimentary rocks in terms of interpreting depositional environment. Most, but not all, sedimentary rocks are stratified in one way or another (Figure 4.1A). Stratification is the layering that is brought about by deposition, and layering is the arrangement of rocks in approximately planar-tabular form (Figure 4.1A). Stratification manifests itself in six ways (Figure 4.1B), as: (a) zones of larger or smaller concentration of pebbles or fossils in an otherwise homogeneous sediment; (b) difference in composition; (c) shape/color difference due to difference in composition; (d) difference in grain size; (e) preferred orientation of non-spherical components (technically, this is not stratification, but it can reveal stratification as in unstratified conglomerates); and (f) differential weathering caused by differences in composition/texture (Figure 4.1A). Stratification is often used synonymously with bedding.

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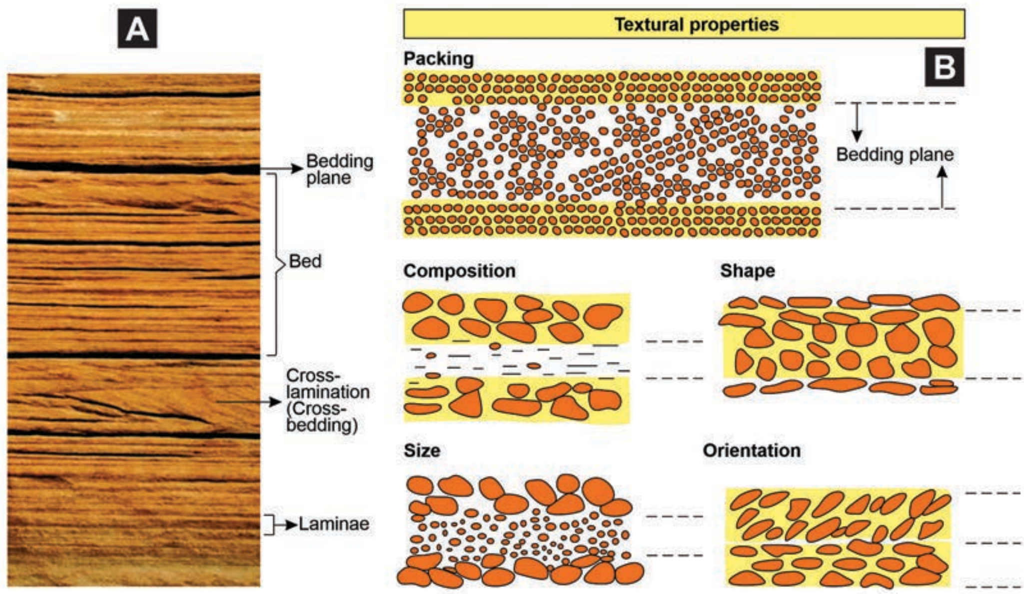

FIGURE 4.1 Stratification (layering) in sedimentary rocks. A: Stratification is layering of rocks in approximately planar-tabular form. B: Stratification manifests itself in several ways in the form of variations in textural properties, including packing arrangement, composition, shape, size, and orientation of its constituent grains.

4.2.2 Bedding and Lamination

Stratification, depending upon the thickness of the strata (bed), is divided into bedding and lamination which in turn are further subdivided (Figures 4.2A–B). Bedding is a fundamental characteristic of sedimentary rocks and refers to horizontal layering (Figure 4.2A). Beds are tabular or lenticular layers of sedimentary rock (also called "strata") that have lithologic, textural, or structural unity, distinguishing them from the strata above and below. Thus, bedding is formed by the sequential deposition of sediments over time, wherein each layer or bed layer represents a separate episode of sedimentation, and is distinguishable from each other by differences in grain size, composition, color, mineral composition, or fossil content. Laminated bedding provides valuable information about the depositional environment and the processes that might have occurred during sedimentation. For example, alternating layers of fine-grained and coarse-grained sediments indicate changes in energy levels or fluctuations in water depth. Fossilized remains found within specific layers also provide clues about the organisms that lived in the past (paleoenvironment).

Beds are separated by bedding planes or bedding surfaces (see Figure 4.2A), most of which represent planes of non-deposition, or an abrupt change in composition (due to a change in depositional conditions), or an erosion surface (see Campbell, 1967). The upper and lower surfaces of beds are known as bedding planes or bounding planes (Figure 4.2A). Beds are defined as strata where they are thicker than 1 cm (McKee and Weir, 1953); layers less than 1 cm thick are called laminae (Figure 4.2B). The bedding surfaces themselves may be planar, wavy, or curved (Figure 4.2C). Depending upon the combination of these characteristics, beds can have a variety of geometric forms such as uniform-tabular, tabular-lenticular, curved-tabular, wedge-shaped, and irregular (Figure 4.2C) (see Ingram, 1954; Collinson et al., 2006).

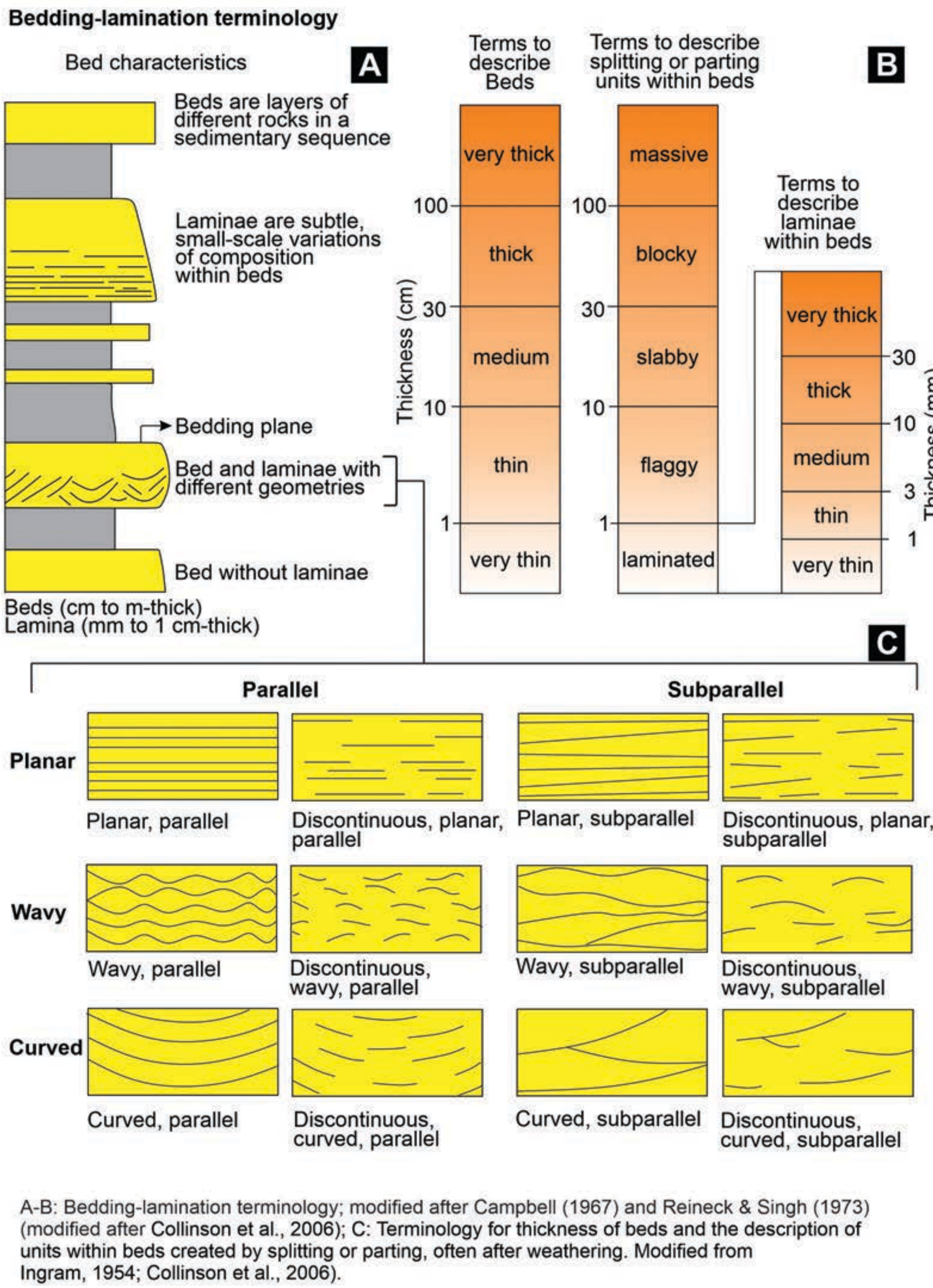

FIGURE 4.2 Bedding-lamination terminology. A: Terminology used in the present study. (Modified after Campbell, 1967; Reineck and Singh, 1973; Collinson et al., 2006.) B: Terminology used for bed thickness. (Modified after Ingram, 1954; Collinson et al., 2006.) C: Bedding terminology based on their geometric forms. (Modified after Collinson et al., 2006.)

4.2.2.1 Planar Bedding and Lamination

Planar bedding refers to the horizontal or near-horizontal layering of sedimentary rocks (Figure 4.2A). The layers, or beds, are typically parallel to each other and can vary in thickness and composition. Planar bedding is commonly observed in sedimentary rocks such as sandstones, siltstones, and shales. It is formed by the sequential deposition of sediment in a horizontal or nearly horizontal fashion (Figure 4.2A). Planar bedding provides information about the original depositional environment, such as whether marine, fluvial, or eolian (wind). The interior of beds (the interval between two bedding planes) contains layers and laminae that are essentially parallel to the bedding planes; i.e., beds display internal planar stratification (laminated bedding or planar bedding) (Figure 4.2A).

Laminations are fine-scale layering within individual beds characterized by thin, parallel layers that are often less than a centimeter thick (see Figures 4.1A and 4.2A–B). Laminations are caused by a variety of processes, such as variations in sediment supply, changes in sedimentation rate, or the presence of currents or waves during deposition. The layers in lamination are often distinguishable by differences in grain size, color, or texture (see Figure 4.1). Both bedding and lamination provide important information about the depositional environment and processes that might have occurred during the formation of sedimentary rocks. They indicate the type of sedimentary environment (such as marine, fluvial, or aeolian) and help interpret energy conditions (high or low) at the time of deposition. Additionally, bedding and lamination are used to determine the direction of current flow or to identify changes in sediment sources or depositional conditions over time.

Groups of similar beds or cross-beds are called bedsets (Figure 4.3A). A simple bedset consists of two or more superimposed beds characterized by similar composition, texture, and internal structures (Figure 4.3A). A bedset is bounded above and below by bedset surfaces (bedding planes) (Figure 4.3B). A composite bedset refers to a group of beds differing in composition, texture, and internal structures but associated genetically, and thus, representing a common type of deposited succession, i.e., beds are produced under similar physical, chemical, or biological conditions (Figure 4.3B) (see also Reineck and Singh, 1980). When layers and laminae are deposited at an angle to the bounding surfaces of the bed (i.e., curved bounding plane), they are called cross-strata or cross-laminae (see Figures 4.3B and 4.4A). Beds composed of such cross-stratified or cross-laminated units are called cross-beds (Figure 4.3B). Thus, cross-bedding is characterized by distinct layers that are inclined at an angle to the overall bedding plane (Figure 4.3B). The angle of inclination varies depending on the depositional environment and the type of sediment involved. Cross-bedding is typically formed by the migration of ripples or sand dunes within a depositional environment such as a riverbed or a desert, respectively. Cross-bedding is used to interpret past depositional environments, to infer the history of a particular sedimentary rock, the processes that formed it, and to understand the direction and intensity of ancient currents or wind patterns (Figure 4.3B).

The bounding surfaces of cross-beds may be either parallel (tabular or planar cross-bedding) or non-parallel (trough cross-bedding) (see Figure 4.4). Both planar and trough cross-beddings are common in eolian, shoreface (above fair-weather wave base), tidal and fluvial (in point bars) environments. Trough cross-bedding (Figures 4.4 and 4.5A) is often confused with hummocky and swaley cross-bedding (see Figures 4.5B–C). In trough cross-bedding, the trough cross-beds cut across troughs of other beds (Figure 4.5A), whereas in hummocky and swaley cross-bedding, beds do not cut across each other sharply, and the curves are at a much shallower angle (Figures 4.5B–C). Hummocky and swaley cross-beddings are common in shallower lower and middle shoreface settings as opposed to upper shoreface settings for trough cross-bedding (see Figure 4.6).

4.2.2.2 Graded Bedding

Graded bedding is characterized by a systematic change in grain size within a single sedimentary layer (see Figure 4.7). It occurs when sediment is deposited in a specific environment, such as a river or a submarine fan, and is sorted by the energy of the transporting medium. Thus, graded bedding is a gradual change in grain size from coarse-grained clasts at the bottom to fine-grained ones at the

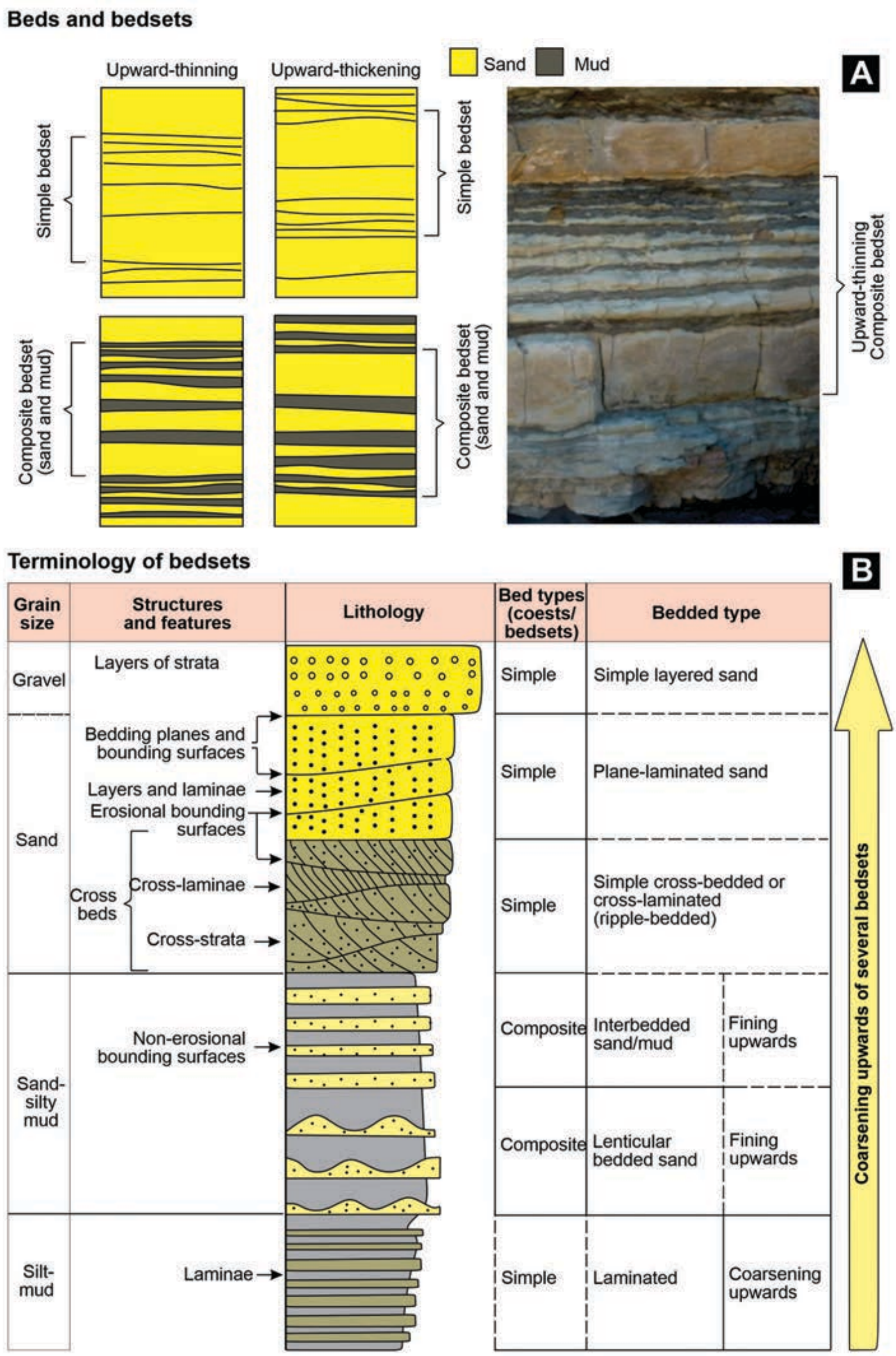

FIGURE 4.3 Beds and bedsets. A: Groups of similar beds are called bedsets, wherein a simple bedset consists of two or more superimposed beds characterized by similar composition, texture, and internal structures. B: Use of bedding in interpreting environments of deposition. Beds composed of cross-stratified or cross-laminated units are called cross-beds, wherein the angle of inclination varies depending on the depositional environment and the type of sediment involved.

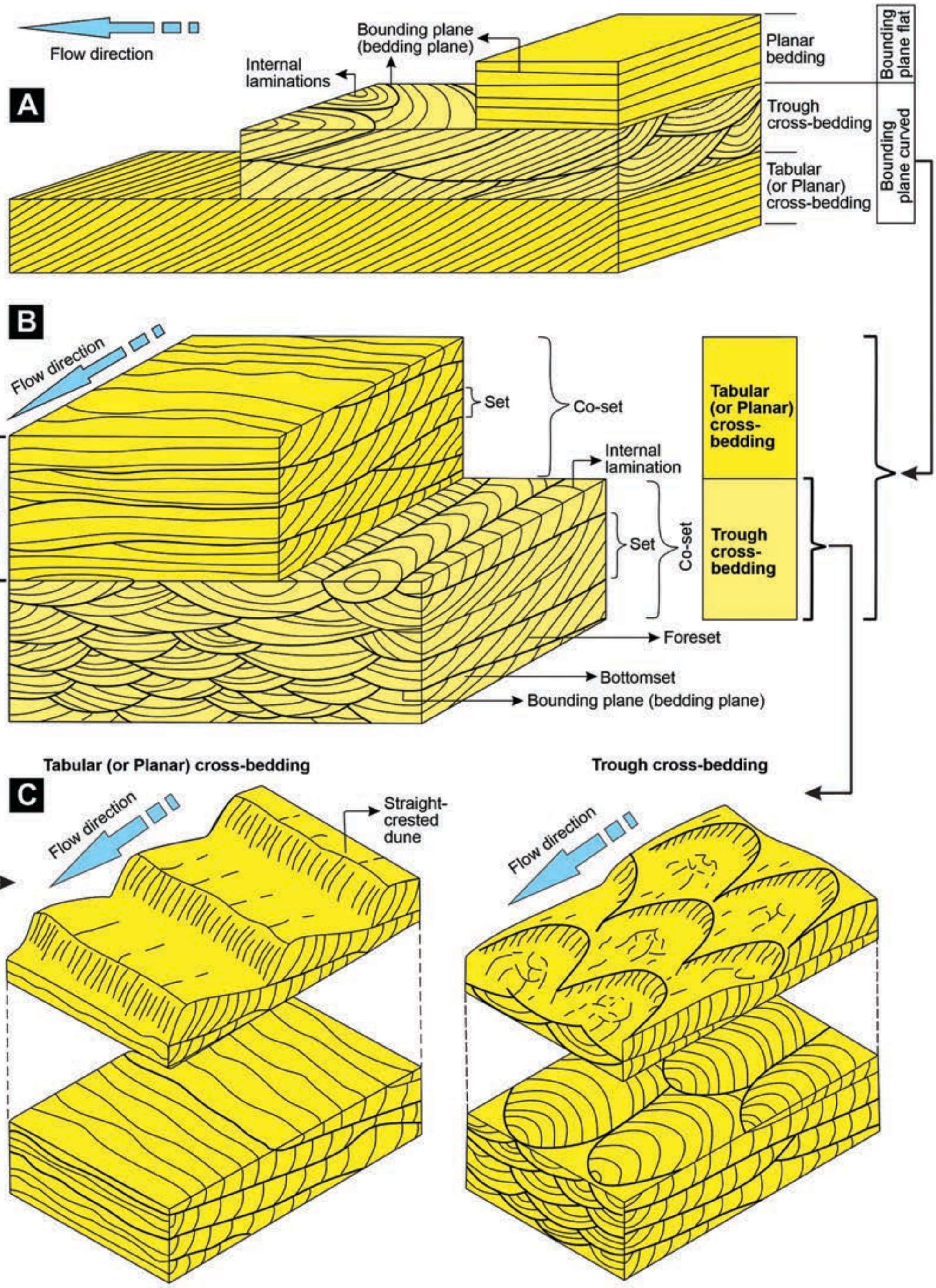

FIGURE 4.4 Cross-stratification. A–B: The bounding surfaces of cross-beds may be either parallel (tabular or planar cross-bedding) or non-parallel (trough cross-bedding). C: Cross section of tabular or planar and trough cross-beddings.

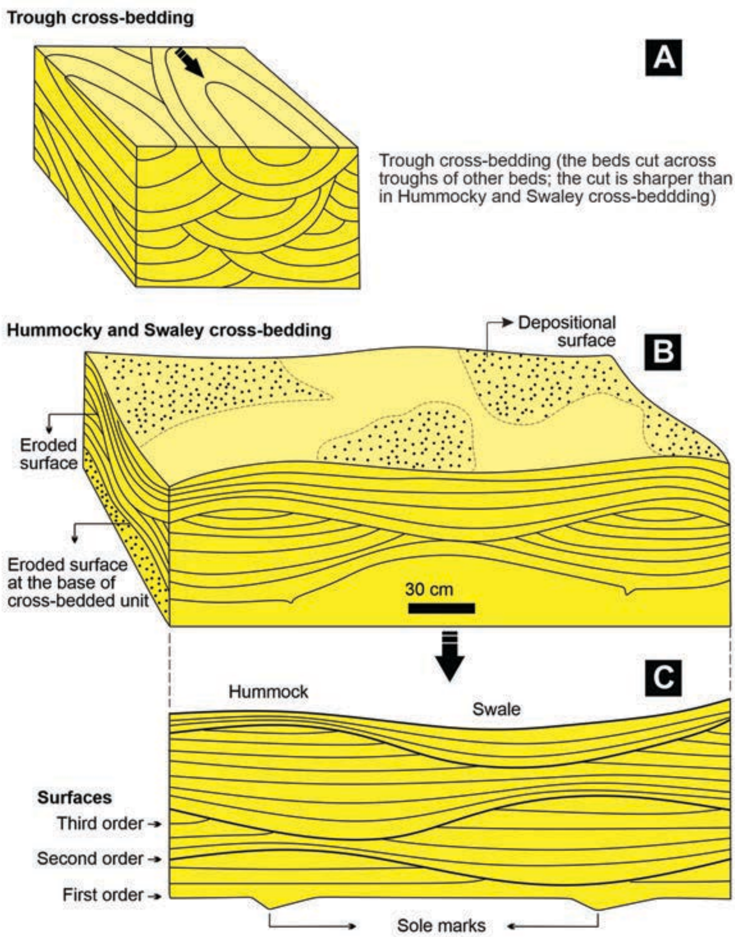

FIGURE 4.5 Types of cross-stratification. A: In trough cross-bedding, the trough cross-beds cut across troughs of other beds at a much steeper angle. B–C: Hummocky and swaley cross-bedding. The beds do not cut across each other sharply, and the curves are at a much shallower angle as compared to trough cross-bedding.

top of a siliciclastic bed such as commonly noted in sandstones or siltstones (referred to as normal grading) (see Figure 4.7A). The coarser grains settle out first due to their higher settling velocity in higher-energy conditions, while the finer grains settle out later in lower-energy conditions (see Figure 4.7). Graded beds range in thickness from a few centimeters to a few meters and are commonly characterized by a sharp basal contact.

Four types of graded bedding are noted (Figure 4.7): (a) normal grading where there is a gradual upward decrease in grain size within a bed (Figure 4.7A); (b) coarse-tail grading is where there is a gradual upward decrease in the size of coarse grains within a bed (i.e., decrease in the percentage of the coarsest grains, and maximum grain size) (Figure 4.7B); (c) reverse (or inverse) grading is when the grain size increases upward within a bed (rare) (Figure 4.7C); and (d) compound (multiple) grading where several graded subunits occur within a bed (Figure 4.7D). A distinction can be made between two kinds of grading: distribution grading, whereby the entire frequency distribution of the sediment shifts toward a finer or a coarser mean size; and coarse-tail grading, where the frequency distribution of the main mass of the sediment stays about the same but the percentage of sediment in the coarse tail of the distribution changes significantly (as mentioned above).

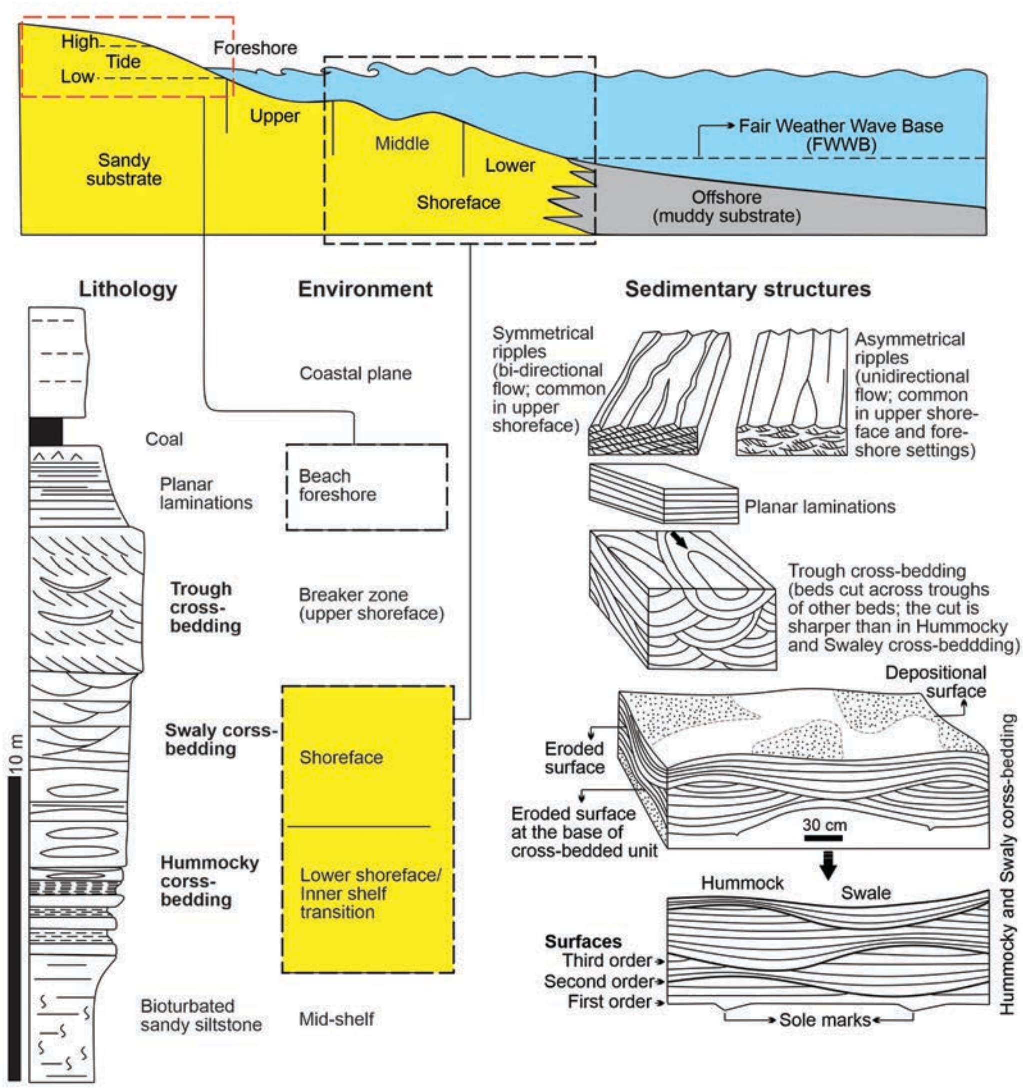

FIGURE 4.6 Major depositional environment of trough and hummocky and swaley cross-beddings. The hummocky and swaley cross-beddings are common in shallower lower and middle shoreface settings as opposed to upper shoreface settings for trough cross-bedding.

Graded bedding is produced by turbidity currents as the sediments settle out of suspension, normally during the waning phase of a turbidity flow (Figures 4.8A–D). The T_A bed of the Bouma sequence characterizes a graded bedded sequence (Figure 4.8E), particularly coarse-tail graded bedding (see Figure 4.7B). The Bouma sequence is a characteristic set of sedimentary structures typically preserved within a graded sand or silt-mud couplets. From base to top, Bouma (1962) differentiated the following intervals above an erosional surface or sharp boundary (Figure 4.8E): (T_A) massive to graded sand, (T_B) plane-parallel-laminated sand, (T_C) cross-laminated sand and silt, (T_D) parallel-laminated sand to silt, and (T_E) laminated to homogeneous mud. However, due to the non-uniform grain size distribution and flow transformations, the complete sequence is rarely preserved (see Fisher, 1983; Mulder and Hüneke, 2014).

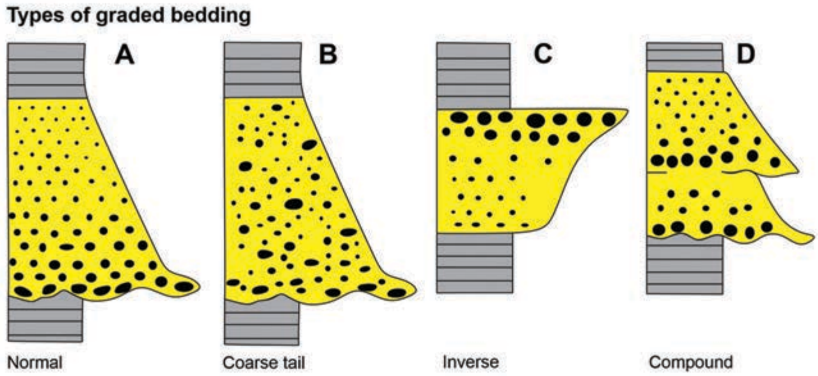

FIGURE 4.7 Graded bedding. This is the systematic change in grain size within a single sedimentary layer. A: Normal graded bedding is characterized by a gradual change in grain size from coarse-grained clasts at the bottom to fine-grained ones at the top. B: Coarse-tail-graded bedding. A gradual upward decrease in the size of coarse grains within a bed is noted. This decrease is in the percentage of the coarsest grains, maximum grain size. C: Reverse (or inverse) grading occurs when the grain size increases upward within a bed. D: Compound (multiple) grading is when several graded subunits occur within a bed.

Graded bedding commonly occurs at the basal part of a turbidite deposit (see Figure 4.8E) and is often associated with flute (Figures 4.9A–B), groove (Figure 4.9C) and tool marks (Figure 4.9D) (these erosional features are dealt later in the chapter). Graded bedding helps to determine the base and top of a bed. Graded bedding is also used to interpret depositional environments, processes, and the energy conditions prevailing during the time of sedimentation (i.e., the dynamics of sediment transport). For example, a well-developed graded bedding sequence in a sandstone layer indicates a rapid decrease in water flow velocity, suggesting deposition in a deep water setting or reflecting a sudden decrease in sediment supply. By studying the grain size variations within a graded bedded sequence, geologists gain insights into changes in sediment sources, transport mechanisms, and energy conditions over time.

4.2.2.3 Massive (Structureless) Bedding

Massive bedding, often noted in sandstones, siltstones, mudstones, and conglomerates is characterized by the absence of distinct layers or bedding planes within a sedimentary unit. Instead of having visible layers or stratification, the rock appears as a homogeneous mass without any discernible internal structure. Massive bedding is typically associated with rapid deposition or sedimentation processes that do not allow for the development of distinct layers. This can happen in environments with high sediment supply, strong currents, or turbulent conditions. There are several factors that contribute to the formation of massive bedding. For example, in high-energy environments such as river channels or nearshore areas, sediment is rapidly deposited and compacted, preventing the development of well-defined layers. In addition, certain types of sediment, such as cohesive mud or volcanic ash, may settle and compact without forming distinct layers. Massive bedding can also be a result of post-depositional processes. For instance, the compaction and diagenesis of sediment over time can obliterate any original layering, leading to a massive appearance. While massive bedding lacks the detailed information that can be obtained from well-defined layers, it can still provide valuable insights into the depositional environment. For example, the presence of massive bedding in a sedimentary rock may indicate a high-energy depositional environment, such as a river channel or a beach, where the sediment was rapidly transported and deposited.

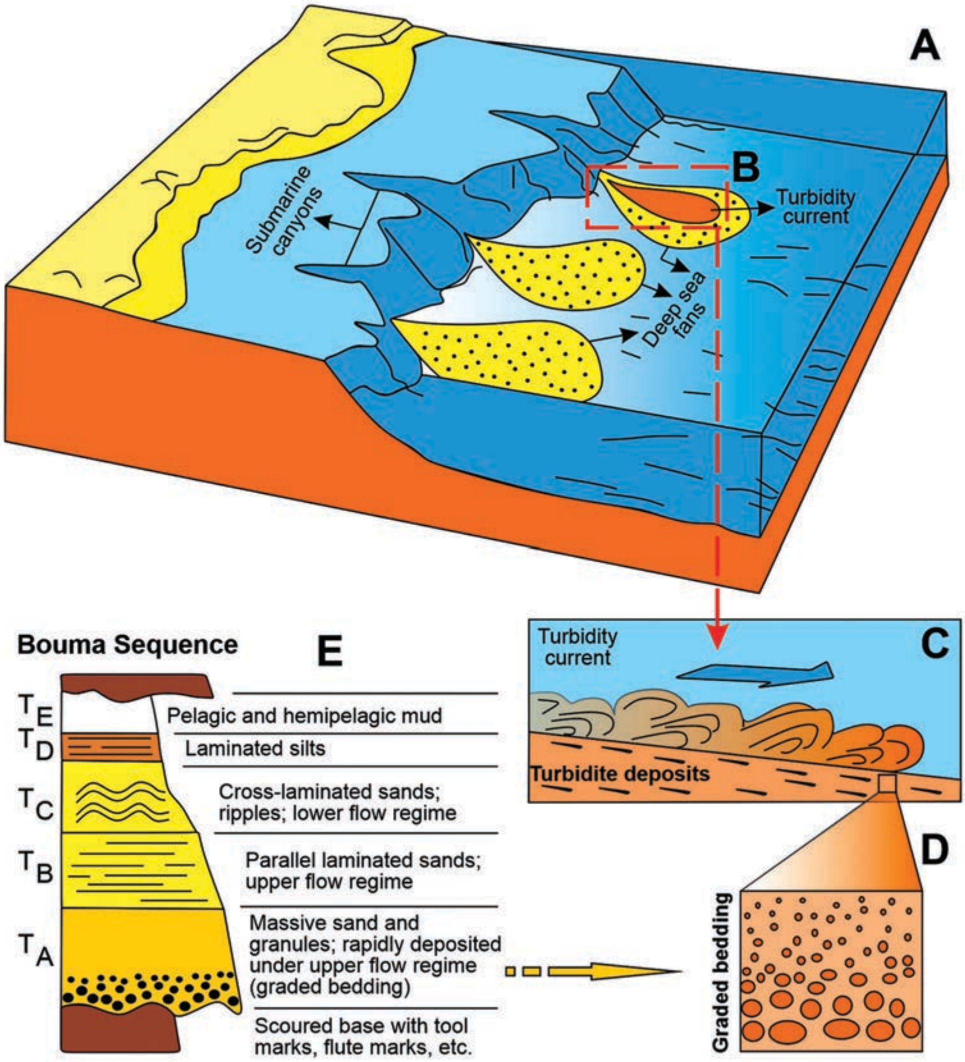

FIGURE 4.8 Graded bedding, turbidites, and Bouma sequence. A–D: The basal part of a turbidite deposit commonly displays graded bedding which is produced by turbidity currents as the sediments settle out of suspension, normally during the waning phase of a turbidity flow. E: Bouma sequence displaying graded bedding at the base and its constituent units, T_A to T_E. However, due to the non-uniform grain-size distribution and flow characteristics, a complete sequence is rarely preserved.

4.2.2.4 Ripples

Ripples are small, wave-like patterns or ridges that form on the surface of sediments, such as sand or silt, due to the movement of water or wind (Figure 4.10). They are a common feature in both terrestrial and aquatic environments and provide information about past depositional processes. In an aqueous environment, there are two main types of ripples: current and wave (Figure 4.10). Ripples generally have a crest-to-crest distance/wavelength of <50 cm and a wave height of <4 cm (see Figures 4.10C–D); bedforms with larger dimensions are called dunes or sandwaves (as in deserts, and discussed later in the chapter).

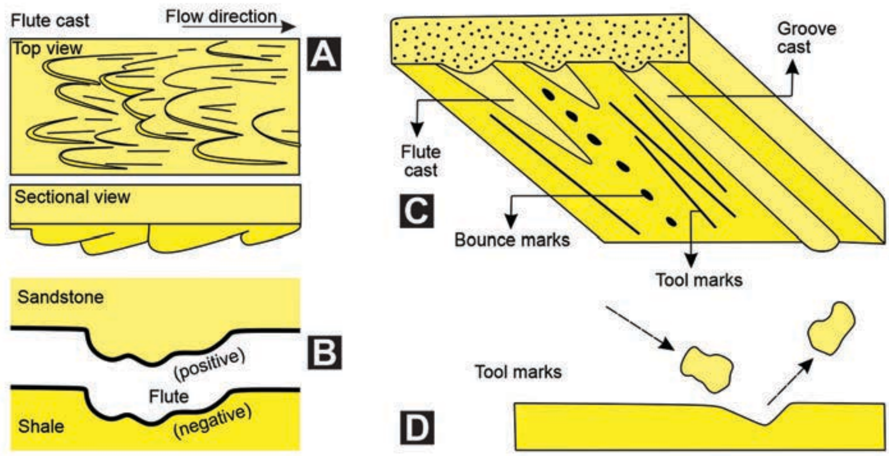

FIGURE 4.9 Sedimentary structures associated with Bouma sequence. A–B: Flute cast. C: Plan view of groove cast, bounce, and tool marks. D: Sectional view of tool marks.

4.2.2.4.1 Current Ripples

Current ripples or unidirectional ripples, form in unidirectional currents such as in streams or rivers (Figures 4.10A–C). They have an asymmetrical profile, i.e., an asymmetric cross section, hence called asymmetric ripple marks (Figures 4.10B–D). They have a steep slope on the downstream side, and a gentle slope on the upstream side with rounded crests, peaks and troughs (Figure 4.10C). They are used to determine the paleocurrent direction. The current ripples occur in a variety of forms depending largely upon the prevailing water depth and flow velocity (see also Allen, 1968).

It must also be kept in mind that the flow velocity (i.e., the speed of a current) and the depth of the water column have a profound impact on the geometries of the crests of ripples (Figure 4.10D). Those ripples with straight crests are stable at low flow velocity or at relatively large depth; they are commonly produced by wind, and linked to the formation of tabular/planar cross-bedding (Figure 4.10E). At higher flow velocity or lower water depth, ripple crests become progressively more sinuous, then catenary (the shape of a flexible hanging chain fixed to poles, like the Greek letter ω), and finally highly irregular (linguoid, tongue-shaped, or lunate, crescent-shaped) (Figure 4.10D), resulting in the formation of trough and herringbone cross-laminations (Figure 4.10E). At higher flow velocity or at lower depths, ripples cannot form and the dominant bedform is a plane bed that produces plane laminations (Figure 4.10D). It must be noted that the above-mentioned bedforms have been formed by unidirectional currents, i.e., wind or water flowing in the same direction. However, there are also bedforms that are formed by oscillatory or bidirectional currents, where the waves move alternatively in two opposite directions forming oscillatory ripples, thus producing symmetric ripples or wave ripples (see Figures 4.10F–J).

4.2.2.4.2 Oscillatory Ripples (Wave Ripples)

Oscillatory ripples, also called wave ripples or bidirectional ripples, are produced by waves or oscillating waters (i.e., due to the back-and-forth motion of water) (see Figures 4.10F–J) (Komar, 1974). They have a smooth symmetric cross section with relatively long straight and sharp crests and rounded troughs, i.e., with gentle slopes on both sides – hence their name, symmetric ripple marks (Figures 4.10F–H). They are generally formed in bodies of standing water such as a lake or a pond. The wave ripples tend to be smaller and more closely spaced as compared to current ripples (compare Figures 4.10 A and 4.10F). The oscillatory ripples indicate an environment with weak currents

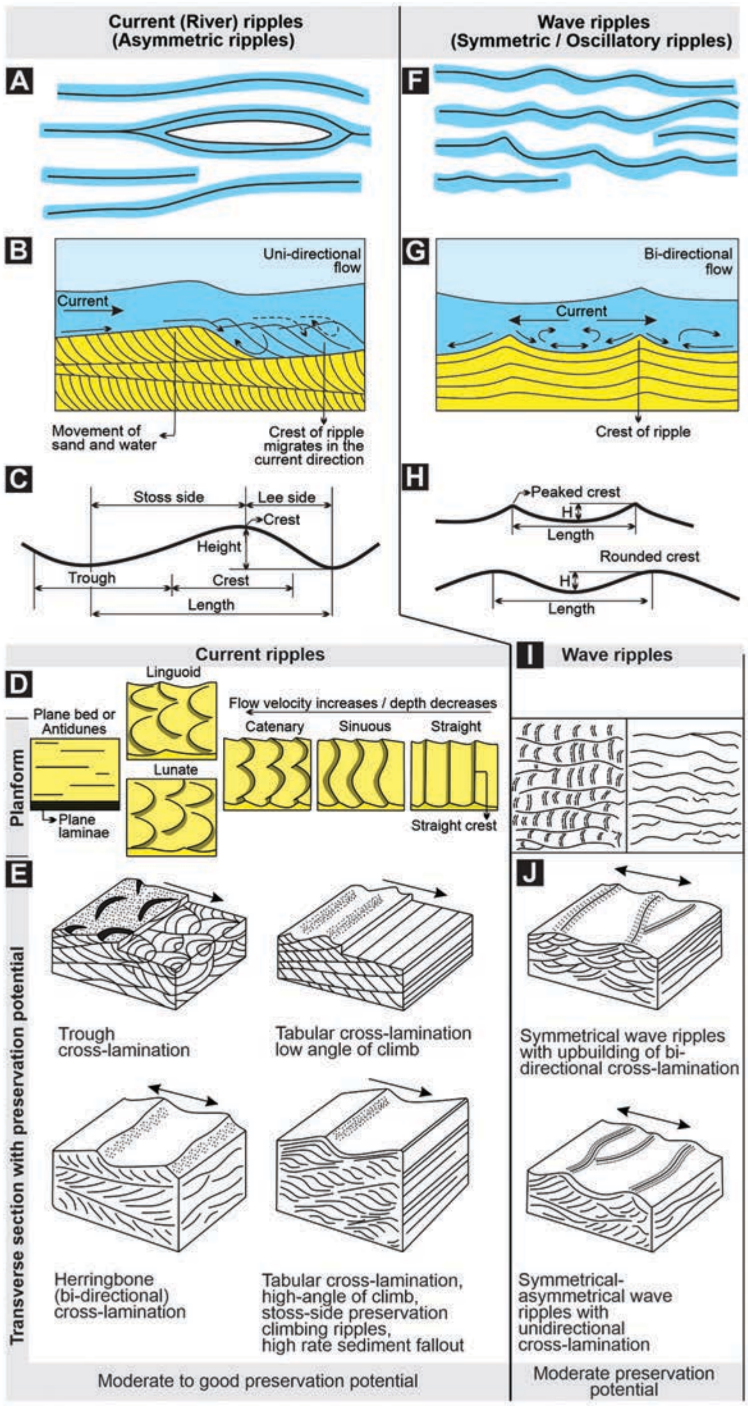

FIGURE 4.10 Current and wave ripples. A–E: Current ripples. A–B: Current ripple morphology. C: Wave length terminology. D: Types of current ripples. Flow velocity (i.e., the speed of a current) and the depth of the water column profoundly impact the geometries of the crests of ripples, ranging from straight to highly irregular (linguoid) crests. E: Preservational potential and cross section of the types of current ripples. F–J: Wave ripples. F–G: Wave ripple morphology. H: Wave length terminology. I: Types of wave ripples. J: Preservational potential and cross section of the types of wave ripples.

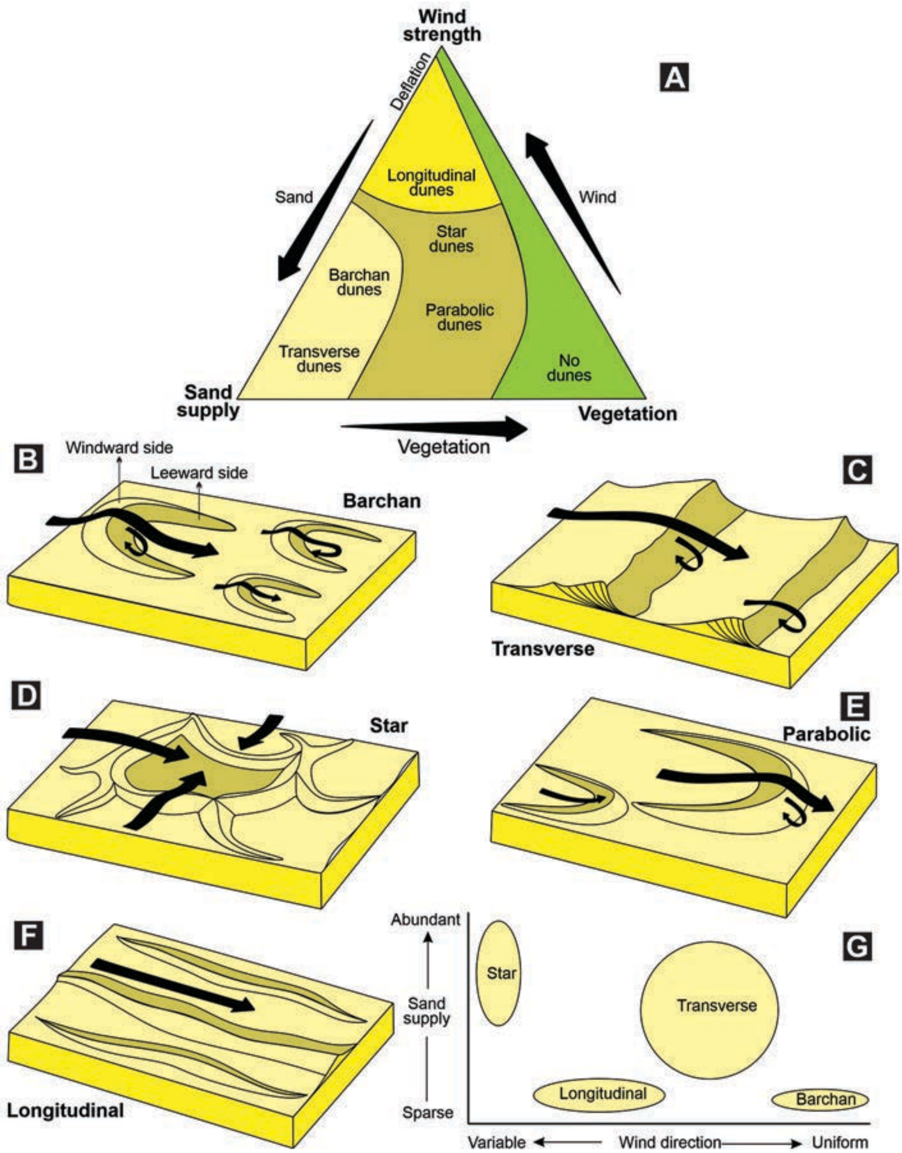

FIGURE 4.11 Dunes. A: Ternary diagram of dunes based on three parameters, sand supply, wind strength, and vegetation cover. A constant unidirectional wind is assumed. (Modified after Hack, 1941.) B–F: Types of dunes. G: Dune development with respect to available sand supply and wind direction.

where water motion is dominated by wave oscillations (the sand grains are moved in both directions, back and forth).

In general, the ripples vary in size, ranging from a few millimeters to several centimeters in height and wavelength. The size of the ripples is influenced by factors such as the velocity of the water or wind, the size of the sediment particles, and the duration of the sediment transport. In addition to their shape and size, ripples also provide information about the energy conditions and sedimentary processes that occurred during their formation (i.e., about the dynamics of sediment transport in both ancient and modern settings). For example, ripples with well-defined crest lines and symmetrical shapes suggest a consistent and unidirectional flow, while ripples with irregular or disrupted patterns may indicate turbulent or fluctuating flow conditions.

As large-scale irregularities begin to develop, and as ripples flatten out, dunes develop (see Figure 4.11). Hence, the fundamentals of dune and ripple formation are the same except for their size; the area of flow separation (spacing) is much larger in dunes. The ripples have spacings less than 0.6 m (Allen, 1984), whereas dunes have spacings larger than 0.6 m (Costello and Southard, 1981; Allen, 1984; Nichols, 2009).

4.2.2.5 Dunes

Dunes are large-scale (60 cm-100's m wavelength and 10's of cm to meters in height), asymmetrical landforms that are formed by the movement of wind or water over loose sediments, such as sand or silt. They are dynamic landforms that can change shape and migrate over time due to the continuous movement of wind or water. They are common features in deserts, coastal areas, and riverbeds, and can vary in size and shape depending on the prevailing environmental conditions. In desert environments, wind-blown dunes, also known as eolian dunes, are commonly noted (Figure 4.11). These are formed by the transport and deposition of sand grains by the wind. The dune size and shape depend on factors such as wind speed, sand supply, and vegetation cover (Figure 4.11A). Common types of eolian dunes include barchan, transverse, star, parabolic, and longitudinal (Figures 4.11B–F). The barchan dunes are crescent-shaped with a gentle slope on the windward side and a steep one on the leeward side (Figure 4.11B). They form in areas with limited sand supply and unidirectional winds (Figure 4.11A). The transverse dunes, on the other hand, are long, linear dunes that form perpendicular to the wind direction (Figure 4.11C). They occur in areas with abundant sand supply and strong, consistent winds (Figure 4.11A). The star dunes are complex dunes with multiple arms radiating from a central point, and form in areas with variable wind directions (Figures 4.11A and D). The parabolic dunes are U-shaped dunes with convex noses trailed by elongated arms (Figures 4.11A and E). They are formed from blowout dunes, where the erosion of vegetated sand leads to a U-shaped depression. The longitudinal dunes are large and elongated, lying parallel to the wind direction (Figures 4.11A and F). They form in areas that are located behind an obstacle where there is strong and constant winds, and sparse sand (Figures 4.11A). Their formational setup with respect to sand supply and wind direction is provided in Figure 4.11G.

4.2.2.6 Antidunes

Antidunes are a type of bedform that form in rivers or in other flowing water bodies (Figure 4.12). They are the opposite of regular dunes, as they form in response to the flow of water rather than wind. They are characterized by their asymmetrical shape and their movement against the direction of the flow. Antidunes typically occur in rivers with fast and turbulent flows, such as during periods of high-water discharge or when there is a sudden increase in flow velocity (see also Figure 4.10D). They form when the velocity of the flowing water exceeds the settling velocity of the sediment particles on the riverbed. As a result, the sediment is temporarily suspended in the water column and transported downstream. As the suspended sediment moves downstream, it accumulates in the troughs of the water surface waves, creating a mound or ridge that is higher than the surrounding riverbed. This mound is the antidune (Figure 4.12) (see Harms and Fahnestock, 1965).

It must be kept in mind that the flow-regime model considers three states of flow starting with no bed movement where there is too little energy in the system to initiate and maintain sand grain movement (Figure 4.12). At the lower flow regime, the plane bed (i.e., parallel, planar laminae with no ripples) represents the lowest velocity, or energy conditions where sediment movement is initiated (Figure 4.12). As the flow energy increases, the size of bedforms increases from ripples to large subaqueous dunes (Figure 4.12). The dune types also change from straight crests and planar cross-bed bounding surfaces (i.e., from two-dimensional structures), to sinuous, arcuate, and lunate outlines and spoon or scour-shaped bounding surfaces (i.e., trough cross-beds) (i.e., to three-dimensional structures). At the upper flow regime where the power of stream flow increases and washes out ripples and dunes, these are replaced with plane beds (with parting lineations), antidunes, and

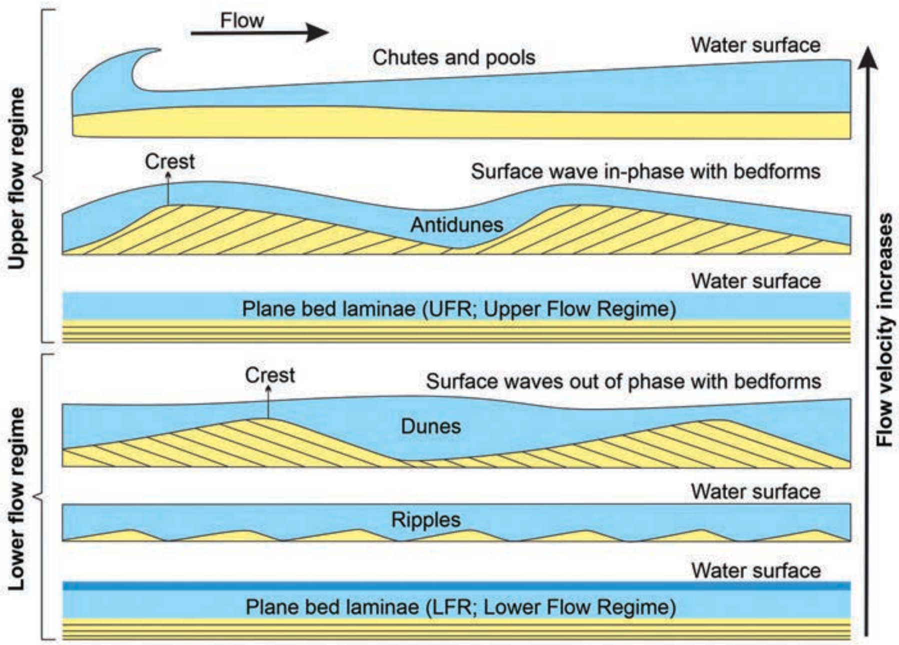

FIGURE 4.12 Antidunes. (Modified from Harms and Fahnestock, 1965.)

erosional chutes and pools (Figure 4.12). As the stream flow increases, the transition from lower to upper flow regime produces antidunes (Figure 4.12). They are commonly noted in shallow channels (e.g., fluvial and tidal). If high flow is maintained, the antidunes will migrate upstream. But, once the flow weakens, they tend to wash out; hence, the preservation potential of antidunes is low.

The flow of water over the antidune causes it to migrate upstream, opposite to the direction of the overall flow. The migration of antidunes is driven by a feedback mechanism. As the water flows over the crest of the antidune, it accelerates and erodes the sediment, causing the crest to collapse. This collapse leads to a decrease in water velocity, allowing sediment to settle and rebuild the mound downstream. The process repeats, causing the antidune to migrate upstream. Antidunes can reach heights of several meters and have wavelengths of tens to hundreds of meters. They are often found in rivers with sandy or gravelly beds and are more common in steep channels with high flow velocities. Antidunes enhance sediment mixing and promote the exchange of sediments between the riverbed and the water column. They can also affect the flow patterns and create turbulence, influencing the distribution of sediments and the formation of other bedforms. Antidunes are rarely preserved in the rock record as they are often reworked into other sedimentary structures as the flow speed decreases.

Numerous flume experiments have established that under unidirectional fluid flow, small ripples begin to develop in sandy sediment as soon as the critical entrainment velocity for the sediment is reached. The exact sequence of other kinds of bedforms that develop with increasing velocity depends upon the grain size of the material. If flow is over a bed of sediment ranging in size from about 0.25 mm to 0.7 mm (medium to coarse sand), for example, the succession of bedforms illustrated in Figure 4.12 is generated, beginning with ripples; ripples are the smallest bedforms. They form in sediments ranging in size from silt (0.06 mm) to sand as coarse as 0.7 mm. Larger bedforms with spacing, or wave length, ranging from under 1 m to over 1000 m are called dunes (see Figure 4.12). Dunes are similar in general appearance to ripples except for their larger size. They form at higher flow velocities in sediment ranging in grain size from fine sand to gravel. The

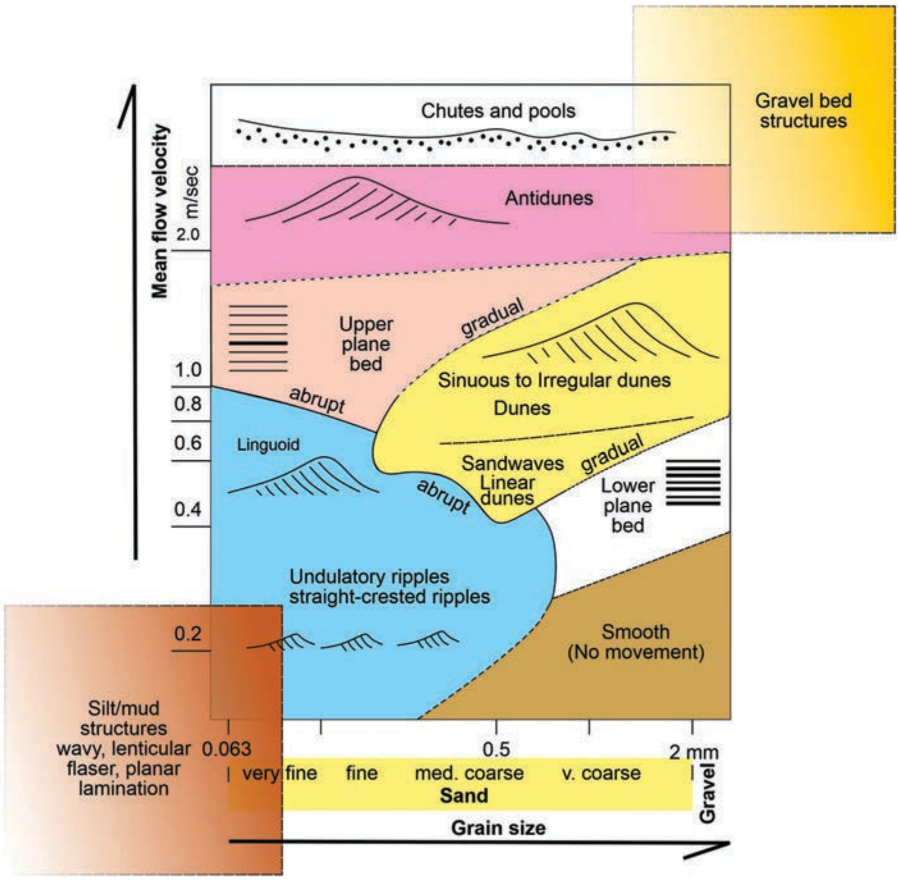

FIGURE 4.13 The formation of ripples, dunes, and antidunes. Their formation is a function of changes in flow velocity and grain size. (Modified from Southard, 1991.)

formation of ripples, dunes, and antidunes is thus a function of changes in flow velocity and grain size, *sensu* the phase diagram of Southard (1991) (see Figure 4.13).

Hence, bedforms (such as ripples, dunes, and antidunes, among others) of different characteristics are formed at varied ranges of flow conditions and sediment grain size, i.e., bedform phases or bed phases (in the sense of Southard, 1991; see also Ohata et al, 2017). Such unidirectional flow-based bedform phase diagrams (such as Figure 4.12) are based both on laboratory and field observations. These phase diagrams enable to infer paleohydraulic conditions but such applications of bedform phase diagrams are not limited only to the analysis of sedimentary structures in fluvial deposits but also to turbidites, megaflooding from glacial lakes, tidal deposits (Mitchell et al., 2010), and tsunami deposits (Fujiwara and Tanigawa, 2014).

4.2.3 Cross-Stratification Structures

Cross-stratification is the layering or bedding of sedimentary rocks that is inclined relative to the horizontal plane (Figure 4.14) commonly noted in deposits from various environments, such as in rivers, deserts, beaches, and (ancient) marine deposits. Cross-stratification is formed when sediments are deposited in a dynamic environment where the direction and intensity of the flow changes over time. As the sediment is transported by water or wind, it accumulates in layers that are inclined to the horizontal plane; these inclined layers are called cross-beds (Figure 4.14). The cross-beds can have different shapes and orientations depending on the specific depositional environment. In rivers, for example, cross-stratification often occurs as a result of migrating ripples or dunes. As the sediment is transported downstream, it is deposited on the inclined foresets of the dunes or ripples, creating cross-beds that dip in the direction of the flow (see Figures 4.14A–B). In eolian (wind) environments, cross-stratification is commonly observed in sand dunes (Figure 4.14C). As wind blows the sand across the dune, it accumulates on the windward side and forms inclined layers or cross-beds (Figure 4.14C). The angle of the cross-beds varies depending on the wind direction and the shape of the dune. In beach and coastal environments, cross-stratification is formed by the action of waves and currents. As waves approach the shore, they transport and deposit sediment in inclined layers that dip landward. Thus, these cross-beds provide clues about the direction of ancient shorelines and the movement of sediment along the coast. In general, cross-stratification is an important sedimentary structure that provides information about the depositional environment and the processes that shape sedimentary rocks. By analyzing the orientation and characteristics of cross-beds, geologists can reconstruct ancient flow patterns, determine the direction of ancient currents, and interpret the dynamics of ancient environments.

4.2.3.1 Ripple Cross-Lamination

Ripple cross-lamination is characterized by the presence of small-scale ripples within the larger sedimentary deposit (Figure 4.15). These ripples are formed when sediment is deposited and subsequently reworked by currents or waves. Ripple cross-lamination is commonly observed in environments such as rivers, beaches, and shallow marine settings. It is indicative of dynamic water flow or wave action that continuously moves and redistributes sediments. In rivers, ripple cross-lamination is often found on the bed of the channel. As water flows over the sediment, it creates small-scale ripples with crests and troughs. These ripples migrate downstream or change shape as the flow conditions change. On beaches, ripple cross-lamination is formed by wave action. As waves reach the shore, they transport and deposit sediments, forming small-scale ripples that are preserved in the sedimentary record, forming cross-laminated layers that dip landward. In shallow marine environments, ripple cross-lamination is formed by tidal currents or storm events. The movement of water and sediments create small-scale ripples that are subsequently preserved as cross-laminated layers. Ripple cross-lamination provides information about the depositional environment and the processes that shaped the sedimentary rock record. By analyzing the size, shape, and orientation of the ripples, geologists gain insight into the direction and intensity of ancient currents or waves (see Figure 4.15). Ripple cross-lamination also helps to determine energy conditions and the type of sedimentary environment in which the rocks were deposited.

Depending on the proportion between sand and mud availability, different types of ripple and beddings [flaser (sand > mud), wavy (sand ≈ mud) and lenticular/nodular (sand < mud)] are formed (see Reineck and Wunderlich, 1968; Reineck and Singh, 1980; Maciaszek et al., 2019) (see Figure 4.15). The ripple-related sedimentary structures are largely grouped into ripple to climbing-ripple cross-laminations and heterolithic bedding (i.e., flaser, wavy, lenticular/nodular beddings) (Figure 4.15). All these small-scale structures are a result of deposition from traction and suspension. The first group (ripple to climbing-ripple cross-lamination) is mostly built by pure sand grains with minor (and only occasionally) mud (finer-grained particles) (Figure 4.15). Climbing-ripple cross-lamination is formed during periods of waning flow in water that is rich in suspended sediment load (Figure 4.15). The second group (heterolithic bedding), is composed of lithologically different laminae, i.e., of sand and mud (>50% volume of clay and silt) (Figure 4.15). Heterolithic bedding does not indicate a specific depositional environment, but provides evidence of very low energy conditions (i.e., slow current to almost stagnant waters) and the presence of grains of various sizes. Fine sand is transported during times of a relatively faster-moving flow (< 0.5 m/s), but still very slow, and creates ripples. On the other hand, during intervals of almost stagnant water (~0 m/s), mud

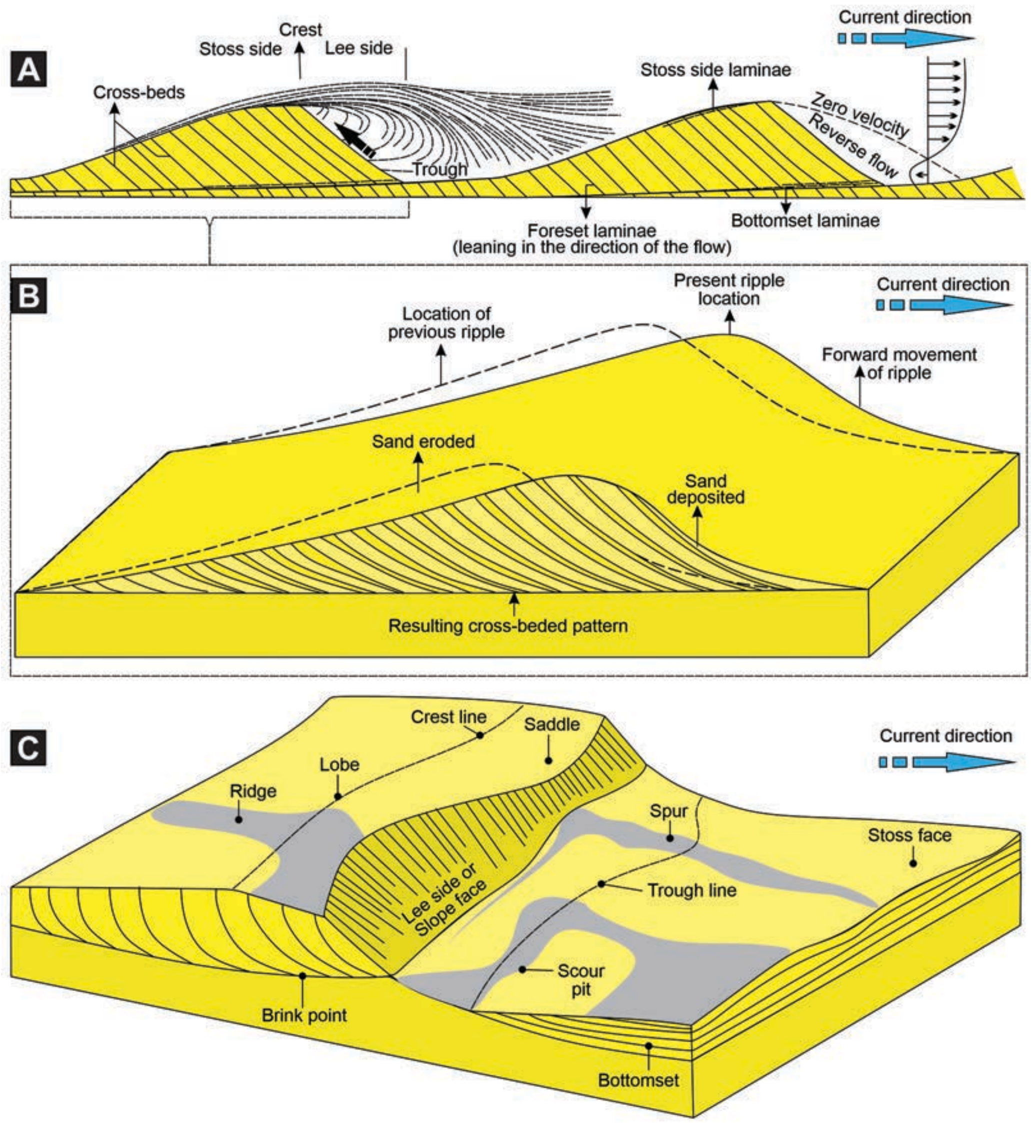

FIGURE 4.14 Cross-stratification (cross-bedding). A–B: Formation and movement of ripple and dunes. Cross-bedding (or cross-stratification) is characterized by layers that intersect at an angle with each other through planar erosional surfaces that truncate inclined beds and laminae. This results in the migration of dunes and ripples produced either by wind or water currents in sand-rich sediments. The sand grains are pushed by currents and grains accumulate as small mounds or ridges, called ripples. The current pushes the ripples ahead, thus eroding sand from their stoss side and depositing them on the lee side, the steeper side of the bedform, and the direction of the current. The accretion of sand on the lee side forms foreset laminae in small ripples and inclined beds in larger bedforms such as dunes. These foreset laminae are inclined at about 34° (angle of repose of sand) with respect to the horizontal. These foreset beds, i.e., inclined beds and laminae in cross-bedded sequences, are thus a very useful paleocurrent indicator, as the laminae are inclined in the direction of the current that produced them. Sand ripples are a few centimeters high, whereas sand dunes are meters to tens of meters high. Thus, migrating sand dunes produce cross-bedded sequences alternating at the meter scale, but ripples produce cross-laminations, intersecting at the scale of only a few centimeters. C: Morphological characteristics of a sand dune in cross section (of B).

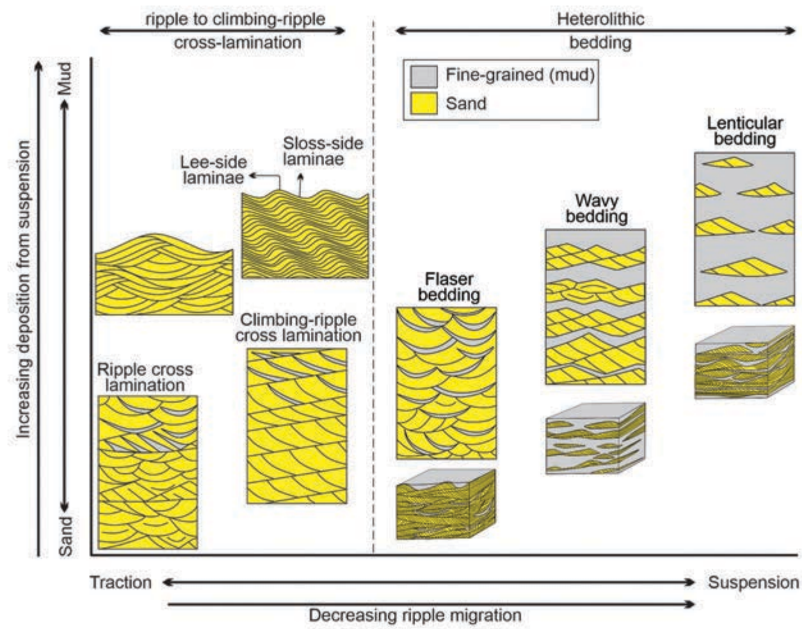

FIGURE 4.15 Diagram showing the relationship between traction and suspension processes, and between sand and mud content during formation of ripple-derived sedimentary structures. (Modified from Maciaszek et al., 2019.)

(i.e., clay and silt), is deposited between ripples or drapes these. Thus, depending on the proportion between sand and mud, flaser (sand > mud), wavy (sand ≈ mud) and lenticular/nodular (sand < mud) bedding is formed (see Reineck and Wunderlich, 1968; Reineck and Singh, 1980).

4.2.3.2 Flaser Bedding

The name "flaser" comes from the French word for "flaky" and describes the appearance of the bedding, which resembles flakes or layers of different sediment types stacked on top of each other (see Figure 4.15). As such, flaser bedding is characterized by alternating thin layers of different sediment types within a larger sedimentary deposit (Figure 4.15). These thin layers (typically ranging from a few millimeters to a few centimeters in thickness) are composed of different grain sizes, mineral compositions, or sedimentary structures. These layers can be parallel or slightly inclined to the bedding plane. The contact between the different sediment layers is usually sharp or erosional, indicating a change in depositional conditions. Flaser bedding is formed when sediment is deposited in an environment where the current strength or direction varies over time. During periods of stronger currents, coarser sediments are transported and deposited, forming sand layers. During periods of weaker currents, finer sediments settle out and form mud layers; this alternating deposition of sand and mud results in the formation of flaser bedding. Thus, flaser bedding is commonly noted in environments where sedimentation is influenced by intermittent or changing currents such as tidal environments, where the ebb and flow of tides results in the deposition of alternating layers of sand and mud. Hence, flaser bedding provides useful information about the dynamics of ancient environments and the processes that influenced sedimentation. By analyzing the characteristics of flaser bedding, geologists can infer the presence of tidal currents, changes in water depth, and variations in sediment supply. This information can help reconstruct the paleoenvironment and understand the depositional history of sedimentary rocks.

4.2.3.3 Lenticular Bedding

Lenticular bedding, also known as lensoidal bedding, is characterized by lens-shaped or lenticular bodies of sediments within a larger sedimentary deposit (Figure 4.15). These lens-shaped bodies are composed of different sediment types, grain sizes, or mineral compositions as compared to the surrounding sediments. The formation of lenticular bedding is often associated with the deposition of sediments in areas of varying energy or sediment supply. For example, in a river environment, lenticular bedding forms when the flow velocity decreases, causing sediment to settle out and accumulate in lens-shaped bodies. These bodies are composed of coarser sediments, such as sand or gravel, within a finer-grained sediment matrix such as mud (Figure 4.15). In general, lenticular bedding is commonly observed in environments where sedimentation is influenced by fluctuating or changing conditions. Lenticular bedding can occur in various depositional settings, including fluvial (river), deltaic, lacustrine (lake), and marine environments. In a deltaic environment, lenticular bedding results from changes in the distribution and deposition of sediments. As the river enters a delta and encounters different flow patterns and sedimentation rates, lens-shaped bodies of sediment form within the overall deltaic deposit. In a lacustrine or marine setting, lenticular bedding is formed by variations in water depth, currents, or sediment supply. Changes in these factors lead to the deposition of lens-shaped bodies of sediment within the overall sedimentary record. Thus, lenticular bedding provides information about the dynamic nature of ancient environments and the processes that influenced sedimentation patterns. By analyzing the characteristics of lenticular bedding, geologists can infer changes in energy conditions, sediment supply, and depositional processes enabling the reconstruction of the paleoenvironment and the depositional history of sedimentary rocks.

4.2.3.4 Hummocky Cross-Stratification

Hummocky cross-stratification (HCS) is characterized by a series of undulating, irregularly shaped layers that are inclined to the overall bedding plane (i.e., cross-stratified) with varying thicknesses (see Figures 4.5B–C). This inclined layering is a result of the sediment being deposited at an angle by turbulent water currents or waves. The undulating nature of layers and their varying thickness are a result of the interaction between the sediment transport processes and the irregularities on the seafloor or lakebed (Figure 4.5B). As the sediment is transported and deposited, it accumulates in the troughs or depressions of the bedforms, creating thicker layers (Figure 4.5B). In contrast, the crests or higher points of the bedforms experience less sediment accumulation, resulting in thinner layers (Figures 4.5B–C). The formation of HCS is associated with the deposition of sediments during periods of high-energy hydrodynamic conditions, such as storms or strong wave action. During these events, the water currents and waves rework and redistribute sediments on the seafloor or lakebed, creating irregular bedforms. These bedforms are then preserved as HCS layers within the sedimentary deposit. Thus, HCS is particularly common in shallow marine settings that are influenced by storm events or strong wave action (see Figure 4.6). HCS can also occur in lacustrine (lake) or fluvial (river) environments under similar energetic conditions. Thus, HCS provides information about the paleoenvironment and the processes that influenced sedimentation. By analyzing the characteristics of HCS, geologists can infer the presence of high-energy events, such as storms or strong wave action, in ancient marine or lacustrine environments. This information enables reconstruction of the paleoenvironment and helps toward a better understanding of the prevailing depositional history of sedimentary rocks.

4.2.3.5 Irregular Stratification

Irregular stratification is characterized by the absence of consistent layering within a sedimentary deposit or a chaotic arrangement of sedimentary units, with varying thicknesses, orientations, and compositions. It is noted at various scales, from small-scale features within a single bed to larger-scale patterns across an entire sedimentary deposit. It is often associated with other sedimentary structures, such as cross-bedding, ripple marks, and erosional surfaces that attest to the dynamic nature of the depositional environment. Irregular stratification occurs in a variety of sedimentary environments, including fluvial (river), deltaic, coastal, and marine settings. It is commonly associated with high-energy conditions, such as strong currents, waves, or turbulent flow, that often disrupt the normal pattern of sediment deposition and create a mixture of different sediment types. The formation of irregular stratification is largely influenced by three factors: (a) the high-energy conditions prevent the settling and sorting of sediment particles, resulting in a mixture of grain sizes and compositions, (b) the irregular topography of the underlying substrate creates areas of local scour and erosion, resulting in uneven deposition, and (c) the sediment supply and transport dynamics play a role in shaping irregular stratification, as fluctuations in sediment input leads to variations in thickness and composition within the deposit. Thus, irregular stratification provides information about the paleoenvironment and the processes that might have influenced sedimentation. By analyzing the characteristics of irregular stratification, the presence of high-energy events, the nature of sediment transport and deposition, and the overall dynamics of ancient environments is inferred.

4.3 EROSIONAL STRUCTURES

Erosion structures provide clues about the depositional history and environmental conditions that existed prior to the formation of the current rock unit. They are also used to reconstruct the paleoenvironment and understand the processes that have shaped the landscape over time. On a broader scale, erosion structures provide geologists with an insight into the tectonic activity, climate, and other factors that have influenced the evolution of the earth's surface.

4.3.1 Bedding-Plane Marking Structures

Bedding-plane markings are sedimentary structures that are preserved on the surfaces of bedding planes, which are the horizontal layers or strata in sedimentary rocks (Figures 4.16 and 4.17) (see Stow, 2005; Collinson et al., 2006 and references therein). There are numerous types of bedding-plane markings observed in sedimentary rocks (Figure 4.16). Some of the major ones are: groove casts and striations, bounce marks, brush marks, prod marks, roll marks, and skip marks, flute casts, load casts, and biogenic structures. Bedding-plane markings are important for geologists as they can be used to interpret the depositional environment, reconstruct ancient landscapes, and understand the processes that have shaped the sedimentary rock. By studying these markings, geologists can gain insights into the paleoenvironmental conditions, such as the presence of ancient currents or waves, the behavior of organisms, or the changes in sediment supply and energy.

4.3.1.1 Sole Marks

Sole marks are geometrical features produced on a sediment bed by erosion by a strong current or by mechanical disruption of the bed by large objects carried by a strong current (see Stow, 2005; Collinson et al., 2006 and references therein) (see Figure 4.16). For the formation of sole marks, a short-lived current acting upon a semi-cohesive bed, usually of mud, is the culprit; although strong currents in other conditions can also form sole marks. In general, scour marks are erosion structures formed by the erosion of pre-existing sedimentary rocks. They are typically found in areas where the surface has been exposed to erosion by wind, water, or ice. Sole marks are broadly grouped into two broad categories, Scour and tool marks (see Figure 4. 16). These are briefly enumerated below.

4.3.1.1.1 Obstacle Scours (Scour-and-Fill Structures)

Obstacle scours (also referred to as scour-and-fill structures, erosional scours, or erosional remnants) are formed from the erosion and subsequent infilling of pre-existing sediments (see Figure 4.16).

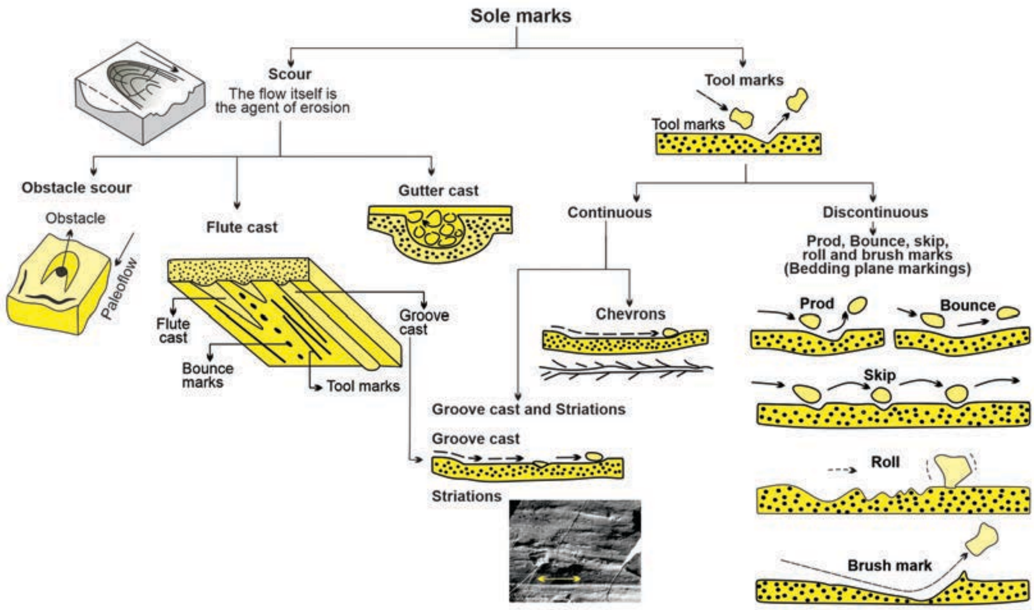

FIGURE 4.16 Sole marks (bedding-plane marking structures).

Their formation typically involves several stages. Initially, a high-energy current or flow erodes the underlying sediments, creating a depression or scour (see Figure 4.16). This erosion is caused by factors such as increased flow velocity, changes in sediment supply, or the presence of obstacles in the flow path (hence also referred to as obstacle scours) (Figure 4.16). Once the scour is formed, sedimentation processes take place, and the depression is filled with new sediments. This infilling occurs through the deposition of sediments from the same current or flow that caused the erosion, or through the accumulation of sediment from other sources. The infilling process results in the formation of distinct layers or beds within the scour. Thus, the resulting structure is characterized by a concave-upward shape, with the in-filled sediment conforming to the shape of the original scour (Figure 4.16). Hence, these structures are used to infer stratigraphic way-up, the up-side of a bed. Such scour-and-fill structures range in shape and size; from small pits to large channels or troughs, depending on the strength of the erosive forces. They are commonly noted in fluvial, coastal, and marine environments where the erosive power of water or currents is strong enough to remove sediments and create depressions or channels. Scour-and-fill structures are significant for geologists as they provide valuable information about the depositional and erosional processes that occurred in a particular environment. They can help reconstruct paleoenvironmental conditions, such as the presence of high-energy currents or flows, changes in sediment supply, or variations in water depth. Additionally, they can provide insights into the sedimentary history and the evolution of a sedimentary basin or depositional system.

4.3.1.1.2 Groove Casts and Striations

Groove casts and striations are observed on bedding-plane surfaces of sedimentary rocks (Figure 4.16). They are typically formed by the movement of sediment or other materials across a surface, leaving behind distinct marks and features (Figure 4.16). Groove casts are elongated depressions or channels that are preserved on a bedding plane (Figure 4.16). They are often formed by the movement of larger particles or objects, such as boulders or ice, across the sediment surface. As these particles move, they carve out grooves in the underlying sediment that are then filled

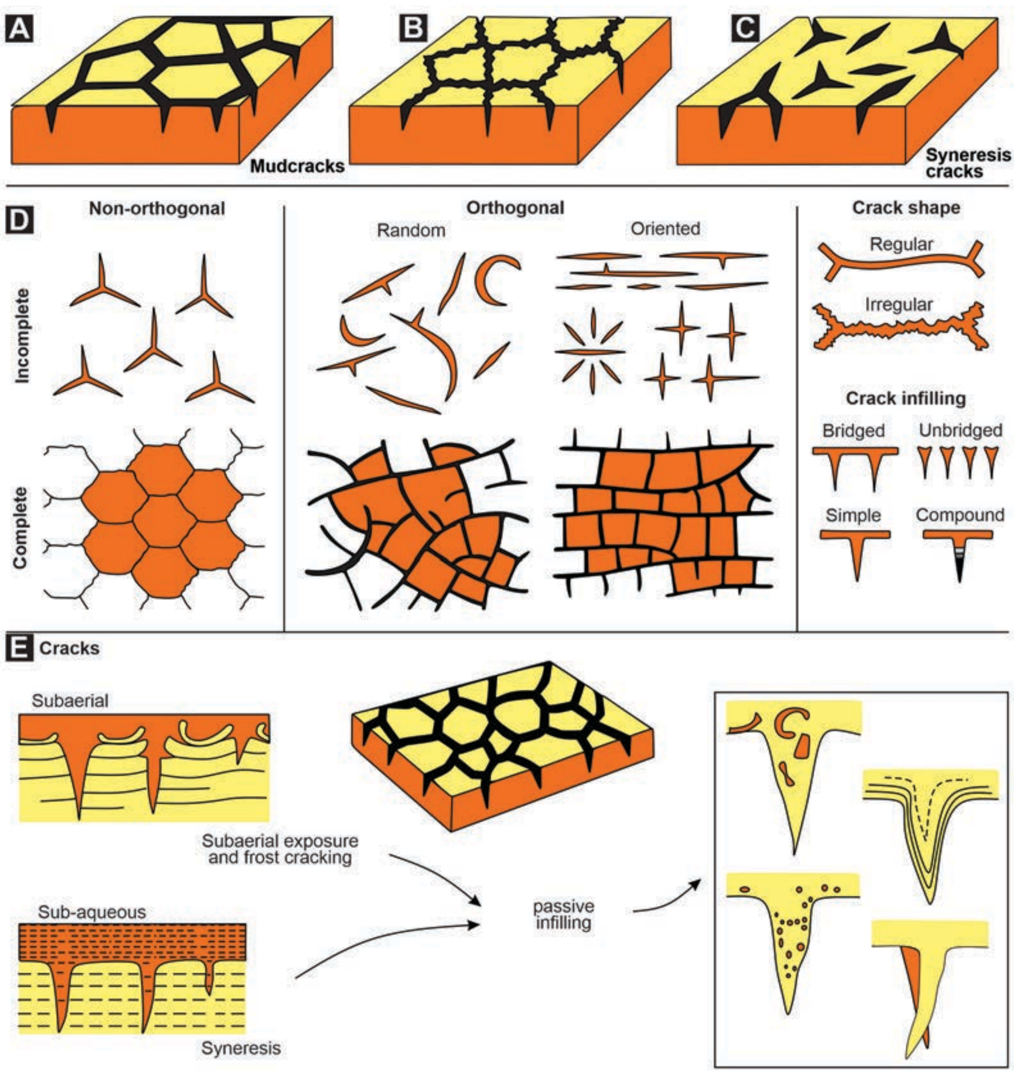

FIGURE 4.17 Mud cracks and syneresis cracks (shrinkage cracks) (bedding-plane marking structures). (Modified after Collinson et al., 2006.) A–B: Mud cracks (shrinkage cracks) formed by subaerial desiccation. C: Syneresis cracks formed by shallow subaqueous water escape. D: Types of shrinkage cracks. (Modified after Allen, 1982.) E: Formation of syneresis cracks. (Modified after Tanner, 1998.)

with finer sediment or other material. When the sediment hardens into rock, the groove casts are preserved as distinct features.

Striations, on the other hand, are linear or curvilinear marks that are left behind by the movement of smaller particles or abrasive materials across the bedding surface (Figure 4.16). They are either parallel or crisscross, and are typically formed by the scraping or scratching action of these smaller materials. Striations are commonly observed on bedding planes, glacially eroded surfaces, or within sedimentary structures, such as cross-bedding.

The formation of groove casts and striations provides important information about the processes and conditions that were present during the deposition and the subsequent formation of the rock. For example, grooves formed by glacial movement indicate the direction and strength of ancient ice flows. Striations provide evidence of the direction and intensity of currents, waves, or wind that were present during sediment deposition. Thus, by studying these structures, a geologist can gain insights into the prevailing paleoenvironmental conditions, such as the presence of ancient glaciers, the movement of water or wind, or the dynamics of sediment transport. These structures help reconstruct past landscapes and understand the processes that have shaped sedimentary rocks over time.

4.3.1.1.3 Flute Casts

Flute casts are observed on bedding planes of sedimentary rocks (Figure 4.16). They are elongated, streamlined depressions that are formed by the erosion of sediment by a fluid flow, such as water or wind. Flute casts are typically asymmetrical, with a steeper side (the upstream side) and a gentler side (the downstream side). The formation of flute casts begins with the erosion of sediment by a fluid flow. As the flow moves over the sediment surface, it carves out a depression in the sediment, with the steeper side facing upstream and the gentler side facing downstream (see Figure 4.16). This depression is then filled with finer sediment or other material, which hardens into rock over time. The resulting structure is a streamlined depression that is preserved on the bedding plane or other surface (Figure 4.16). Flute casts provide important information about the direction and strength of the fluid flow that was present during sediment deposition. They also indicate the nature of the sediment that was being eroded and transported by the flow. By studying these structures, geologists gain insights into the paleoenvironmental conditions, such as the presence of ancient rivers or streams, the direction and intensity of ocean currents, or the behavior of wind in ancient landscapes.

4.3.1.1.4 Gutter Casts

Gutter casts (cast is an infilling) are products of eroding currents that form cup-shaped to cylindrical pillars of sandstone by the infilling of potholes, or rounded nonlinear erosional depressions (Whitaker, 1973) (Figure 4.16). Thus, these are downward bulging sole structures produced by the process of (scour) erosion and followed by deposition. Gutter casts range in shape from discs to rounded loaf-like forms, and in size from 1 cm to 20 cm in diameter or more. They are generally associated with tempestites deposited on a muddy ramp, where the beds also contain hummocky cross-stratified sediments (Perez-Lopez, 2001).

4.3.1.1.5 Prod, Bounce, Skip, Roll, and Brush Marks

These are types of bedding-plane markings observed on the surfaces of sedimentary rocks (Figure 4.16) and provide information about the processes and conditions that were present during the deposition and subsequent formation of the rock. These are briefly enumerated below.

Prod marks are elongated depressions or grooves that are formed when a relatively sharp object is pushed or prodded into the sediment surface (Figure 4.16). These marks are caused by the movement of animals, such as burrowing organisms or grazing animals. They record the impact at a steep angle, thereby producing an asymmetric mark with the deep end downstream. Prod marks provide evidence of the presence and behavior of organisms within the depositional environment.

Bounce marks are small, circular depressions or pits that are formed when sediment particles are dropped or bounced onto a surface (Figure 4.16). These marks are typically small and can be caused by the impact of raindrops, falling debris, or even the movement of animals. They are caused by shallower angled impact, thus, producing symmetrical features. They do not record directional information. Bounce marks can provide evidence of the energy of the environment and the size of the particles that were being transported.

Skip marks are irregularly spaced depressions or pits that are formed when sediment particles are skipped or bounced across a surface (Figure 4.16). These marks are typically larger and deeper than bounce marks and can be caused by the impact of larger particles or objects. Skip marks provide evidence of the energy of the environment and the size of the particles that were being transported.

Roll marks are elongated depressions or ridges that are formed when sediment particles are rolled or dragged across a surface (Figure 4.16). They can be caused by the movement of water, ice, or wind, as well as the movement of animals or other objects. Roll marks provide information about the direction and intensity of the sediment transport and the nature of the particles involved.

Brush marks are linear or curvilinear grooves or scratches that are formed by the movement of sediment particles across a surface (Figure 4.16). They can be caused by the action of currents, waves, or wind, as well as the movement of ice or other abrasive materials. Brush marks can provide information about the direction and intensity of the sediment transport and the nature of the particles involved.

By studying these bedding-plane markings, geologists gain insights into the processes and conditions that were present during the deposition and subsequent formation of the rock. These markings help to reconstruct past environmental conditions, better understand the dynamics of sediment transport, and interpret the behavior of organisms prevailing within the depositional environment.

4.3.2 Bedding-Plane Markings of Miscellaneous Origin

Bedding-plane markings of miscellaneous origin refer to various features and structures that are found on the surfaces of bedding planes in sedimentary rocks. These markings are not directly related to the process of sediment deposition or the formation of the rock itself, but rather represent secondary features that have formed after the rock was deposited and lithified. Some examples of bedding-plane markings of miscellaneous origin are: load casts, raindrop imprints, mud cracks and syneresis cracks, and biogenic structures.

4.3.2.1 Mud Cracks and Syneresis Cracks (Shrinkage Cracks)

Mud cracks are tension cracks or fractures that extend downward from the top of the bed into the sediment below (Figures 4.17A–B). They are arranged in a network of regular hexagons or rectangles but more commonly with irregular geometry and commonly taper downward to a sharp lower end, at depths of centimeters to a few decimeters. The lateral shrinkage of sediments causes tensile stresses that result in cracking. Sometimes also referred to as desiccation cracks, they provide evidence of subaerial exposure of the sediment surface and are also a top-and-bottom indicator. The polygonal patterns of mud cracks are formed when wet mud or clay dries and contracts (Figures 4.17A–B). As the sediment dries, it shrinks and develops cracks that are preserved in the rock record. The cracks intersect at angles of 120 degrees (see 4.17A) and vary in size, ranging from a few millimeters to several centimeters in width. Mud cracks are commonly found in environments that experience periodic drying and wetting, such as tidal flats, lake beds, and river floodplains.

Syneresis cracks, also known as shrinkage cracks, are similar to mud cracks but form in different types of sediment, such as gel-like materials or colloidal suspensions (Figure 4.17C). Syneresis cracks occur when the sediment undergoes volume reduction due to the loss of water or other volatile components. These cracks also have a polygonal shape and are typically smaller in size compared to mud cracks. They also develop by the shrinkage of sediments but without desiccation. Syneresis cracks are commonly observed in gelatinous materials, such as certain types of clays or gels used in laboratory experiments.

Both mud cracks and syneresis cracks provide important information about past environmental conditions. The presence of mud cracks in sedimentary rocks indicates periods of drying and wetting, which can be used to infer past climatic conditions and hydrological processes. Syneresis cracks can also provide insights into the physical properties of the sediment and the processes that led to its formation.

4.3.2.2 Pits and Small Impressions

These are common bedding-plane markings of miscellaneous origin found on the surfaces of sedimentary rocks. These markings have various causes and provide information about the processes that occurred during or after the deposition of the sediment. Examples of pits and small impressions include borings, gas escape structures, tool marks, nodules or concretions, and crystals or crystal molds.

Borings are small holes or tunnels formed by organisms (these are further described under biogenic structures). These organisms create openings in the sediment or rock by burrowing or drilling through it. Borings provide evidence of past biological activity and the presence of organisms in ancient environments.

Gas escape structures, also known as gas escape pipes or gas escape structures, are small cylindrical or tubular features that form when gas bubbles are trapped within the sediment or rock during deposition. As the sediment compacts and lithifies, the gas bubbles escape, leaving behind cylindrical or tubular voids. These structures indicate the presence of gas-rich sediments or the release of gases during sediment compaction. Tool marks are small impressions or scratches that are left on the surface of sedimentary rocks during geological processes. These marks are caused by natural processes such as abrasion or erosion.

Pits and small impressions provide information about the processes and conditions that occurred during or after the deposition of sedimentary rocks. They help in reconstructing ancient environments, identify biological activity, and better understand the diagenetic processes that shaped the rocks.

Crystal molds create small impressions or pits on the bedding plane of sedimentary rocks. These markings are the result of crystal growth or dissolution processes that occurred after the deposition of the sediment.

Nodules or concretions are compact masses or aggregates of minerals that form within sedimentary rocks. Nodules create small impressions or pits on the bedding plane when they are eroded or weathered out. Nodules or concretions vary in composition and provide information about the diagenetic processes that occurred during the formation of the rock.

4.3.2.3 Rill and Swash Marks

Rill marks are small channels or grooves formed on the ground by the flow of water (see Figure 4.18). They are typically seen on slopes or in areas with loose soil or sediment. Rill marks are commonly found after heavy rainfall or during the melting of snow or ice. They indicate the path and intensity of water runoff, as well as the erosion and deposition of sediment. Swash marks, on the other hand, are marks left on the beach or shoreline by the movement of waves. When a wave reaches the shore, it carries water and sediment up onto the beach in a swash motion. As the water recedes back into the ocean, it creates swash marks in the form of small ridges or lines in the sand. These marks provide

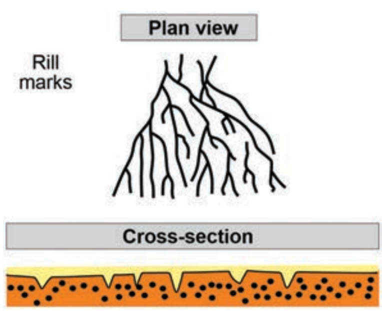

FIGURE 4.18 Rill marks.

information about the direction and strength of the waves, as well as the movement of sediment along the shoreline.

4.3.2.4 Sedimentary Sills and Dikes

A sedimentary sill is a horizontal sheet-like intrusion of magma that is injected between layers of sedimentary rock. It occurs when magma is forced into a horizontal crack or bedding plane within the sedimentary rock and solidifies. Over time, erosion may expose the sill at the surface. Sills are typically parallel to the layering of the surrounding sedimentary rock and can vary in thickness. They are commonly found in areas with extensive volcanic activity or in regions with thick sequences of sedimentary rock.

A dike, on the other hand, is a vertical or near-vertical (inclined) sheet-like intrusion of magma that cuts across pre-existing rock layers. Dikes are formed when magma is injected into fractures or cracks in the surrounding rock and solidifies. Dikes vary in thickness and length and are often seen as narrow, wall-like structures that cut across the layering of the surrounding rock. They are commonly found in volcanic regions and can be exposed through erosion or uplift. Both sedimentary sills and dikes provide important insights into the geological history and processes of an area. They help understand the timing and sequence of magma intrusions, as well as the deformation and alteration of the surrounding rock. Additionally, the study of sills and dikes provides information about the tectonic activity and volcanic processes that have shaped the earth's crust.

4.3.2.5 Raindrop Imprints

Raindrop imprints are small depressions or pockmarks that are formed on the surface of sedimentary rocks when raindrops impact wet or loose sediment. Thus, when a soft, moist surface of freshly deposited sediment is exposed to a brief, light shower of large raindrops, tiny craters, circular in outline and with a slightly raised rim, are imprinted upon the sediment surface (Figure 4.19). If these are buried by later deposition, then they are preserved within the sedimentary record. They provide evidence of subaerial emergence and enable the identification of top and bottom of beds. These imprints also provide evidence of past climatic conditions, such as periods of rainfall or wet environments.

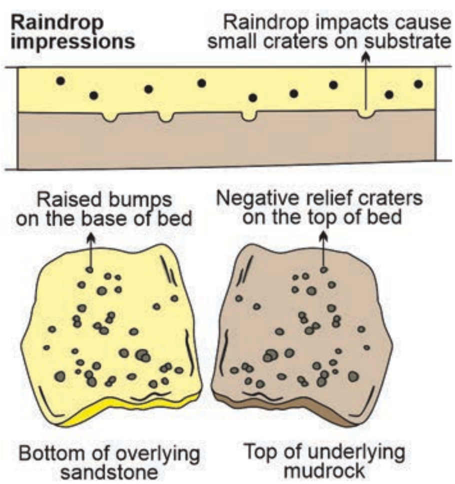

FIGURE 4.19 Raindrop marks.

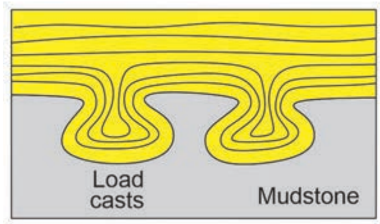

FIGURE 4.20 Load cast.

4.3.2.6 Load Casts

Load casts are structures that form when denser sediment is deposited on top of less dense sediment, causing the underlying material to be squeezed and displaced (Figure 4.20) (see also Allen, 1982). This displacement creates elongated, lens-shaped structures that are often filled with different sediment from the surrounding rock. Load casts typically appear as elongated or bulbous features within sedimentary layers, resembling fingers or blobs (Figure 4.20) and are formed when the denser sediment sinks into the less dense one, displacing and deforming it. The shape and size of load casts can vary depending on factors such as the viscosity and composition of the sediment or rock involved. Load casts provide valuable insights into past geological processes and the conditions under which they have occurred. They indicate the direction and magnitude of the forces applied, as well as the nature of the sediments or rocks involved. By studying load casts, geologists can better understand the history and dynamics of the earth's crust and the forces that have shaped it over time.

4.4 SOFT-SEDIMENT DEFORMATION STRUCTURES

These structures are formed due to the deformation or alteration of pre-existing sedimentary layers (Figure 4.21) (see also Seilacher, 1969; Owen, 1987; Moretti and Sabato, 2007). These are also called penecontemporaneous deformation structures, i.e., deformation involving folding and contortions developed before the sediment was lithified, i.e., shortly after burial, when the sediment was still non-cohesive and buried only less than a few meters (see Shanmugam, 2016). Some deformations occur even earlier, i.e., during, not after, deposition (Figure 4.21B). Sediments ranging from coarse silt up to gravel can be viewed as a packed framework of grains in mutual contact. So, any kind of disturbance to the sediment bed, like an earthquake, can cause a sudden repacking of grains; the grains fall into a new, closer packing in a kind of wave that sweeps through a more or less large volume of the sediment (Figure 4.21C). As this happens, the repacking sediment finds itself suffused with excess pore water, and the sediment is then in a liquefied state, hence, not locked into packing by being in contact with surrounding grains. This process is called liquefaction. In finer sediments, like silts and very fine sands, due to their low permeability (because of the smallness of pore passageways), the sediment remains in a liquefied state for some time, and long enough to get deformed under the influence of stress field present within the sediment. The soft-sediment deformation structures are thus robust proxies providing valuable information about tectonic forces, stress conditions, and geological processes that might have affected the sedimentary rocks.

4.4.1 Types of Soft-Sediment Deformation

4.4.1.1 Loading

When a bed has a greater bulk density than the bed underlying it, stratification becomes gravitationally unstable (Figure 4.22). Then, if the sediment becomes mobilized, there is a tendency for the material of the overlying bed to sink down into the underlying bed and for the material of the

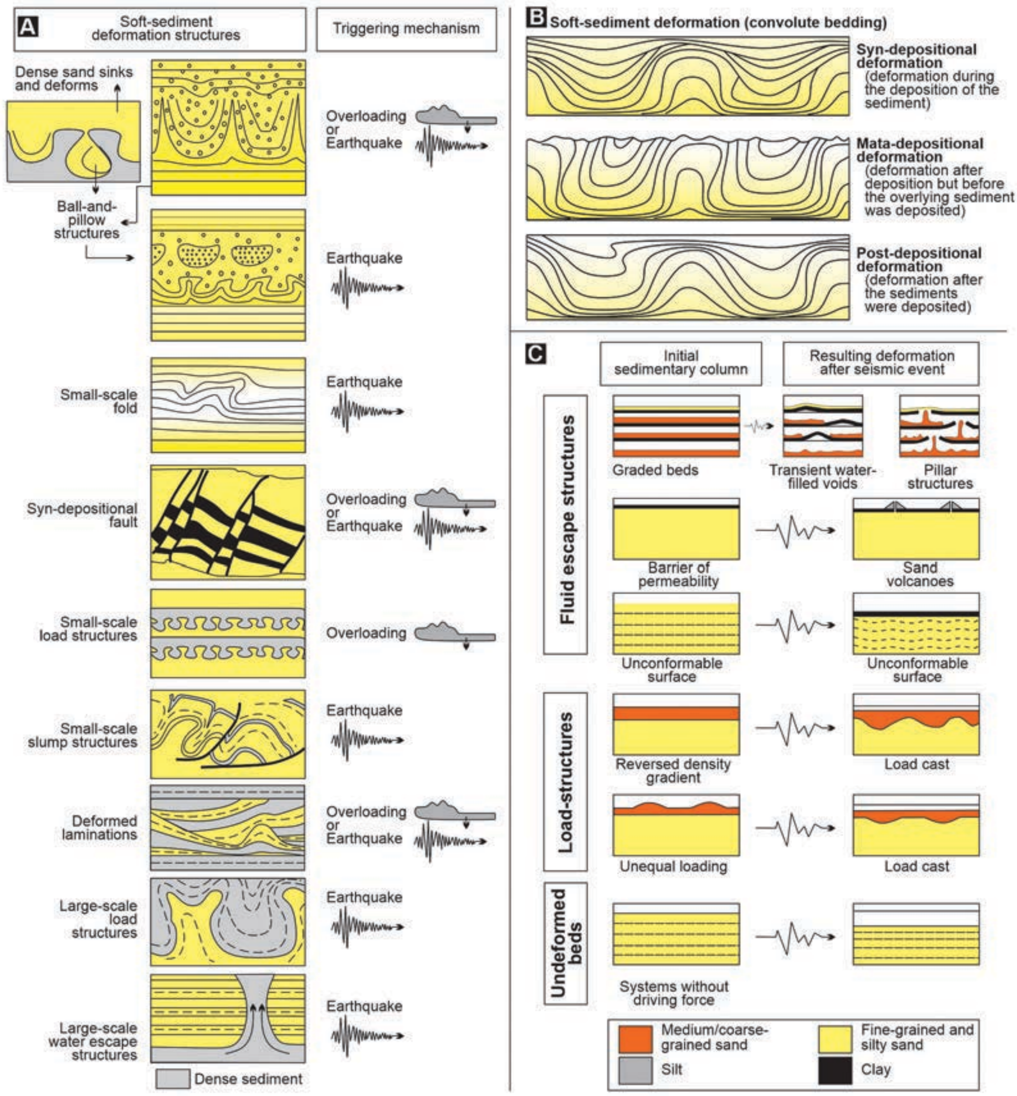

FIGURE 4.21 Soft-sediment deformation structures. A: Soft-sediment deformation structures and their triggering mechanisms. (Modified after Morettie and Sabato, 2007.) B: Types of convolute laminations. (Modified after Allen, 1982.) C: Sedimentary structures and their resulting deformation caused by seismic events. (Modified after Moretti et al., 1999.)

underlying bed to rise up into the overlying material. This is called loading. Loading is often noted where a water-rich sand bed is deposited over a semi-cohesive mud bed. These results in the formation of various sedimentary structures; the major ones, among them are load cast, flame, ball-and-pillow structures, and pseudo-nodules. Extreme development of load casts are pseudo-nodules and ball-and-pillow structures (Figure 4.22).

4.4.1.2 Convolute Bedding and Laminations

Convolute bedding (convolute lamination or convolute stratification) consists of folded or contorted layers that have been deformed during sedimentation or the early diagenetic process, and is typically

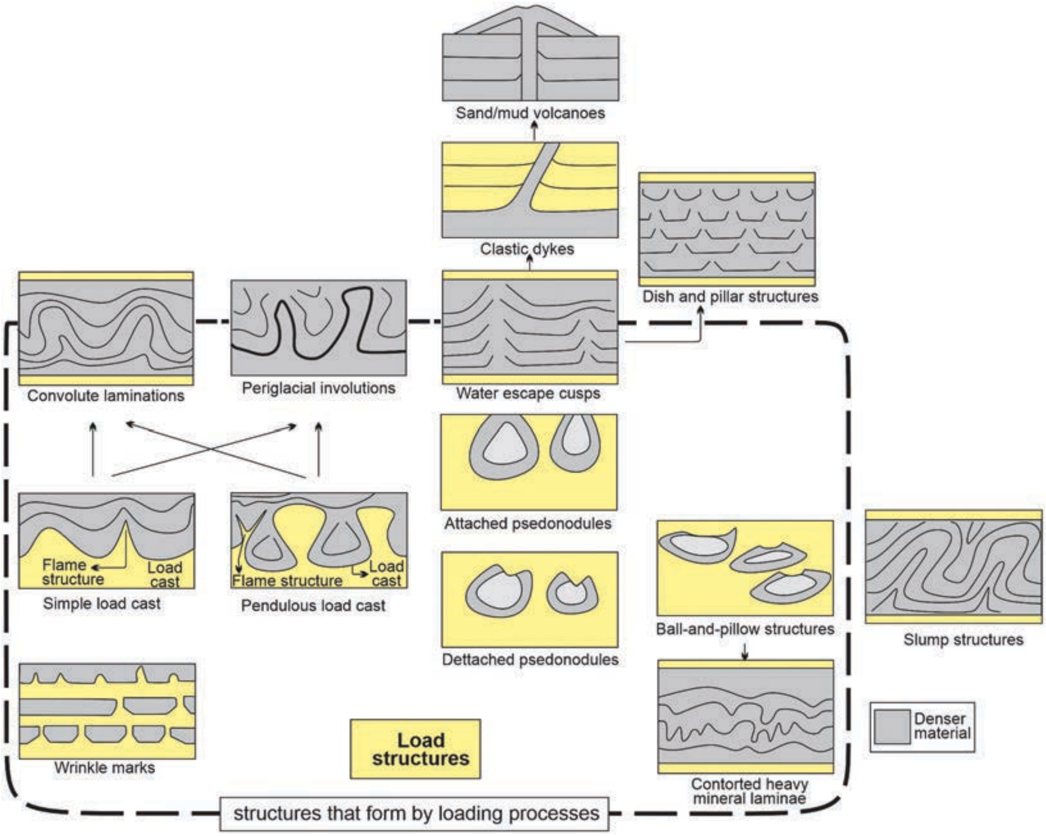

FIGURE 4.22 Load structures. Relationships between load structures and other structures suggestive of sediment mobilization. The thick dashed line encloses structures that form by loading processes. Shading (gray color) represents denser sediment. Arrows indicate links between different structures. (Modified after Owen, 2003.)

characterized by tight, irregular folds that may intersect or overlap with each other, resulting in a convoluted appearance of the rock unit (Figure 4.23). Convolute bedding forms when sediment layers are subjected to compressional forces, typically caused by the weight of overlying sediments or tectonic activity causing the originally horizontal or gently inclined sediment layers to buckle and fold, resulting in the contorted appearance of a convolute bedding (Figure 4.23). The beds have planar lower and also upper contacts, but the bed is internally folded into broad synclines and sharp to dome-shaped or even mushroom-shaped anticlines, that usually die out upward. Convolute bedding formation is influenced by several factors such as (a) the presence of water in the sediments that act as a lubricant, facilitating the deformation and folding process, (b) the sediment composition and grain size affect the susceptibility of the layers to deformation, where fine-grained sediments (such as siltstone or mudstone) are more likely to deform and fold compared to coarser-grained sediments, and (c) the intensity and direction of the compressional forces play a role in shaping the convolute bedding structures. Convolute bedding can occur in a variety of sedimentary environments, such as in the deep-sea turbidite deposits, deltaic environments, and submarine fan systems. Convolute bedding is often associated with the deposition of fine-grained sediments, such as silt or mud that have a high water content and are more prone to deformation. Thus, convolute bedding provides evidence of the deformation and compaction processes that occurred during sedimentation. It also indicates the presence of tectonic activity or other geologic events that caused the deformation.

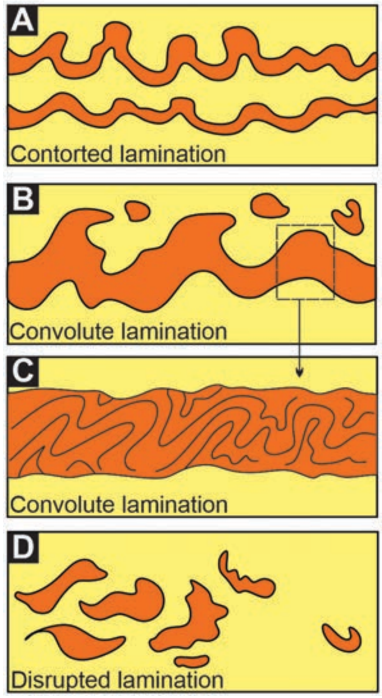

FIGURE 4.23 Small-scale sedimentary structures formed by the disruption of beddings/laminations. A: Contorted laminations. B–C: Convolute laminations. D: Disrupted laminations. (Modified after Collinson et al., 2006.)

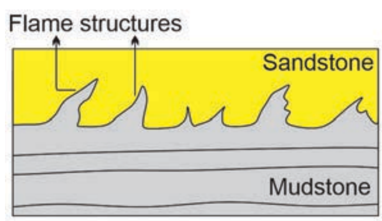

FIGURE 4.24 Flame structures.

Thus, by studying convolute bedding, geologists can gain insights into the stress conditions, paleoenvironment, and tectonic history of an area.

4.4.1.3 Flame Structures

Flame structures are characterized by the presence of thin, flame-like projections or fingers that extend into adjacent layers within a rock unit, commonly noted in fine-grained sedimentary rocks, such as mudstone or siltstone (see Figure 4.24). They are formed by the expulsion of fluid or gas from a more permeable layer into a less permeable one. They can be straight or curved, branching or non-branching, and can extend over short or long distances within a rock unit (see Figure 4.22). The spacing between flame structures can also vary, ranging from closely to widely spaced. The formation of flame structures is influenced by several factors such as (a) the presence of a fluid or gas source within the more permeable layer is necessary for the expulsion process. This can be due to the presence of water, hydrocarbons, or other fluids trapped within the sediment, and (b) the contrast in permeability between the layers is crucial for the migration of the fluid or gas. The more permeable layer allows for the upward movement, while the less permeable layer restricts lateral or vertical migration.

Flame structures typically form in environments where there is a contrast in permeability between sediment layers. The more permeable layer, often a sand or coarser-grained layer, acts as a conduit for the expulsion of fluid or gas. As the fluid or gas is forced upward, it creates finger-like projections that penetrate the overlying or surrounding less permeable layer, which is typically composed of finer-grained sediment. Flame structures vary in size, shape, and orientation. The presence of flame structures provides information about the fluid dynamics and depositional processes that might have occurred during sedimentation. They also indicate the presence of fluid flow pathways, the migration of fluids or gases, and the permeability variations within a sedimentary sequence. The flame structures, also throw light into the prevailing paleoenvironment, fluid migration patterns, and diagenetic processes that might have affected the rock unit.

4.4.1.4 Ball-and-Pillow Structures

Ball-and-pillow structures, also known as ball-and-socket structures, are formed when a denser sediment or fluid is injected into a less dense sediment or fluid (see Figures 4.21A and 4.22). They are commonly observed in fine-grained sedimentary rocks, such as mudstone or shale, and are indicative of fluid or sediment injection processes. The ball-and-pillow structures typically form in environments where there is a contrast in density or viscosity between the injected material and the surrounding sediment or fluid. The denser material, often mud or sand, is injected into the less dense material, causing it to deform and form rounded or bulbous structures resembling balls or pillows (Figures 4.21A and 4.22). These structures vary in size, shape, and orientation. The size of the structures can range from small centimeter-scale to larger meter-scale features. They can be spherical or elongated in shape, depending on the injection dynamics and the properties of the injected material and surrounding sediment (Figures 4.21A and 4.22). The orientation of the structures varies depending on the direction of the injection and the deformation of the surrounding sediment. In general, their formation is influenced by several factors such as (a) the presence of a fluid or sediment source that is capable of injection is necessary for the process to occur. This can be due to the presence of overpressured fluids, such as water or hydrocarbons, or the injection of sediment from a nearby source, and (b) the contrast in density or viscosity between the injected material and the surrounding sediment or fluid dictates the behavior and shape of the structures. The denser material tends to form rounded or bulbous shapes, while the less dense material deforms and accommodates the injected material (Figures 4.21A and 4.22). The presence of ball-and-pillow structures in sedimentary rocks provides valuable information about fluid or sediment injection processes and the dynamics of sedimentary systems. They indicate the presence of fluid flow pathways, the migration of sediment or fluids, and the permeability variations within a sedimentary sequence. Studying ball-and-pillow structures enables a better understanding of the prevailing paleoenvironment, fluid migration patterns, and depositional processes affecting the rock unit.

4.4.1.5 Synsedimentary Folds

These are folds that form during or shortly after sedimentation, while the sediment is still soft and unconsolidated (see Figure 4.21A). Synsedimentary folds form due to a variety of factors, including tectonic activity, sedimentation processes, and changes in sea level. Tectonic activity can cause the sediment to be deformed and folded as it is being deposited, while changes in sea level can cause the sediment to be compressed and folded as it is buried. Sedimentation processes, such as the accumulation of sediment in a deltaic environment, can also lead to the formation of synsedimentary folds. In general, synsedimentary folds are typically found in fine-grained sedimentary rocks, such as shales or mudstones. They range in size from small centimeter-scale folds to large kilometer-scale structures. Their shape and orientation vary depending on the type and intensity of the forces that caused the deformation. They can be symmetrical or asymmetrical, and can have a variety of fold geometries, including anticlines, synclines, and domes. The orientation of the folds also varies, depending on the direction of compressional forces. Synsedimentary folds provide evidence of the deformation and tectonic activity that occurred during the early stages of sedimentation. They can also provide information on the paleoenvironment, sea level changes, and sedimentation processes that occurred during the formation of the rock unit, and about the tectonic history and evolution of an area.

4.4.1.6 Synsedimentary Faults

These form contemporaneously with sedimentation, i.e., they develop while the sediment is still being deposited or are formed shortly after the sediment has been deposited (see Figure 4.21A). Synsedimentary faults form due to a variety of factors such as tectonic activity, sediment loading, and changes in sea level. Tectonic activity, such as extension or compression, can cause the development of faults as sediment is being deposited. Sediment loading can also lead to the development of faults as the weight of the sediment causes the underlying rocks to deform and fracture. Changes in sea level can cause the sediment to be compressed or stretched, leading to the formation of faults. The orientation and geometry of synsedimentary faults vary depending on the type and direction of the tectonic forces or sediment loading. They can be normal faults, where the hanging wall moves down relative to the footwall, or reverse faults, where the hanging wall moves up relative to the footwall. The displacement along synsedimentary faults can range from small millimeter-scale movements to larger kilometer-scale displacements. Synsedimentary faults provide evidence of deformation and tectonic activity that might have occurred during sedimentation. They also help determine the paleostress conditions and the tectonic history of an area. Additionally, they can influence sedimentation patterns, basin architecture, and the distribution of sedimentary facies within a rock unit.

4.4.1.7 Slump Folding

As the sediment on a slope is liquefied, it flows or slides down the slope forming another type of soft-sediment deformation called slump folding (Figure 4.25A). It is often intra-formational, i.e., bounded above and below by undeformed/undisturbed layers (Figure 4.25B). The folds are characteristically tight (such as isoclinal and recumbent) and range up to a few meters. These folds are sometimes truncated by erosion and are overlain by very similar sedimentary material with only a slight depth of burial, thus, providing evidence of penecontemporaneous deposition (i.e., formed almost at the same time as the original deposition of the layers) (Figure 4.25B).

4.4.1.8 Dish and Pillar Structures

Dish and pillar structures are characterized by alternating concave-upward depressions (dishes) and cylindrical or pillar-like structures (pillars) (see Figure 4.22). The dish structures are shallow depressions that form due to the accumulation of sediment or the erosion of underlying layers; they typically have a concave-upward shape and vary in size from a few centimeters to several meters in diameter. Dish structures are often formed by the action of currents or waves that erode or scour the sediments in certain areas, thus, creating the depressions.

Pillar structures, on the other hand, are cylindrical or pillar-like structures that rise above the surrounding sediments. They form as a result of the deposition of sediment or the accumulation of organic material in specific areas. They are often found in environments with low sedimentation rates, where the sediment is more resistant to erosion or compaction as compared to the surrounding sediments. The dish and pillar structures are typically arranged in a regular pattern, with the dishes and pillars alternating along the bedding plane. This pattern results from the interaction of different sedimentation processes, such as the migration of tidal currents, the accumulation of organic material, or the deposition of sediment in specific areas. They are noted in certain depositional

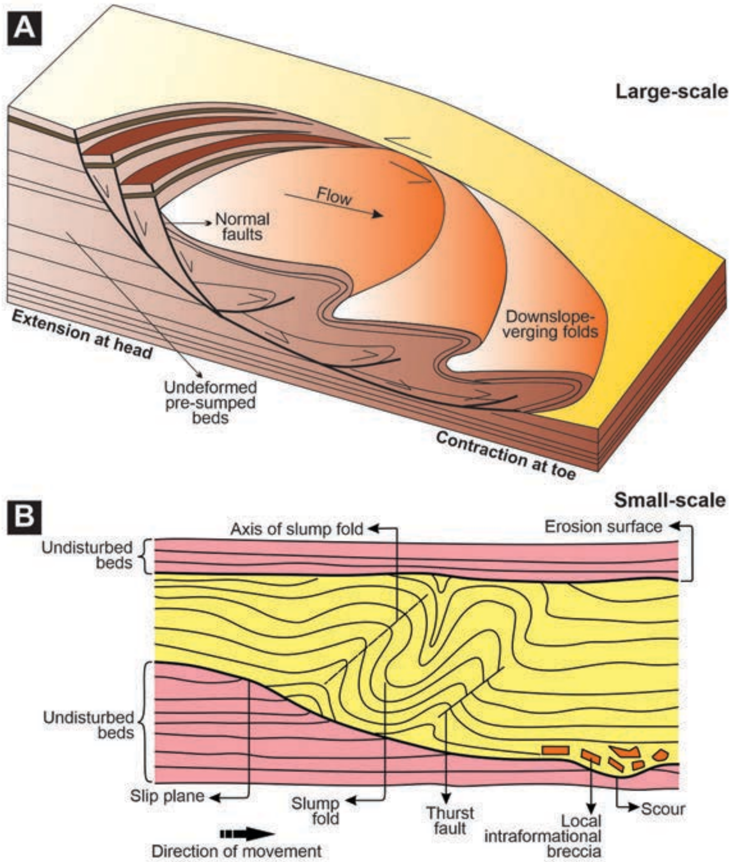

FIGURE 4.25 Major characteristics of a slide–slump structure. A: Large scale. B: Small scale. (Modified after Tucker, 1996.). Lateral translation of sediment mass along a basal dislocation zone–slip plane and internal disruption of bedding occurs in slumping. Both can occur at all scales from large-scale (km-scale) to small-scale (cm-scale). (Modified from Tucker, 1996.)

environments, such as tidal flats, lagoons, or shallow marine settings. Dish and pillar structures provide information about depositional processes and environmental conditions that existed during the formation of the sedimentary rock in which they were found. They also indicate the presence of tidal or wave action, the distribution of sedimentary facies, and the dynamics of sediment transport and deposition. Thus, they enable the paleoenvironment to be reconstructed and provide information about the sedimentary processes that shaped the rock unit in which they are found.

4.5 BIOGENIC STRUCTURES

Biogenic structures are formed by the activities of organisms. These structures are noted in various types of sedimentary rocks and provide evidence of past life and ecological conditions. Some common examples of biogenic structures include: tracks, trails, and burrows, stromatolites, trace fossils, shell beds and shell hash, and coral reefs, among others. These biogenic structures are valuable in paleontological and paleoecological studies as they provide evidence of past life and ecological interactions. They can also help in reconstructing ancient environments and understanding the evolution of organisms and ecosystems over time.

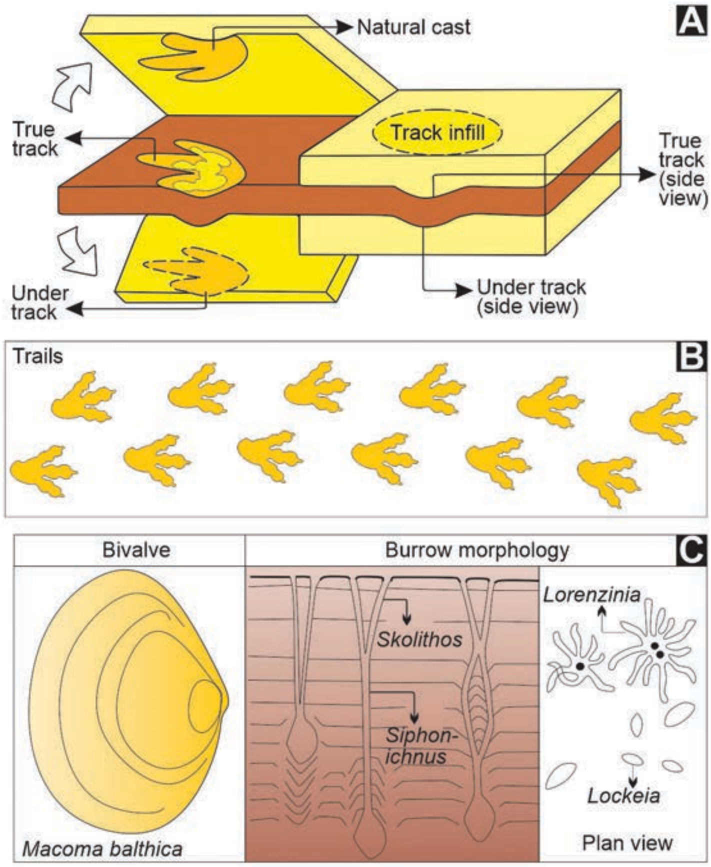

FIGURE 4.26 Tracks, trails, and burrows. A: Tracks. Tracks are imprints or marks left by an animal's feet or paws on the ground. B: Trails. Trails are pathways or routes created by animals as they move through their environment. C: Burrows. These are tunnels or holes dug by animals for shelter, protection, or for nesting purposes.

4.5.1 Tracks, Trails, Burrows

Tracks, trails, and burrows are forms of evidence left behind by animals (Figure 4.26). Tracks are imprints or marks left by an animal's feet or paws on the ground (Figure 4.26A). They provide valuable information about the animal's size, shape, and movement patterns. Tracks are used to identify the species of animal that made them, as different animals have distinct footprints. Trails are pathways or routes created by animals as they move through their environment (Figure 4.26B). These can be formed by repeated use over time, and they may indicate common travel routes or foraging areas for the animal. Trails can be seen in various habitats, such as forests, deserts, or grasslands. Burrows are tunnels or holes dug by animals for shelter, protection, or for nesting purposes (Figure 4.26C). Burrows vary in complexity and size, depending on the species. They provide a safe space for animals to rest, raise their young, or escape from predators. All three of these forms of evidence are useful for studying and understanding animal behavior, habitat use, and ecological interactions. Scientists often use tracks, trails, and burrows to identify animal populations in different ecosystems.

4.5.2 Stromatolites

Stromatolites are layered structures formed by the growth and trapping of sediment by microbial communities, primarily cyanobacteria (also known as blue-green algae) (Figure 4.27A). These microbial structures, largely domed or columnar in form (Figure 4.27A), develop at the sediment–water interface in shallow marine environments, such as tidal flats, lagoons, or carbonate platforms; they are also known from freshwater and evaporitic environments (see Riding, 2010). These structures are some of the oldest known forms of life on earth, with fossil evidence dating back over 3.7 billion years (Figure 4.27B), reported from the Isua Greenstone Belt of southwestern Greenland (see Nutman et al., 2016). However, recently, Zawaski et al. (2020) noted that these structures, deformed conical stromatolites, are not of biogenic but tectonic origin.

Stromatolites are not as common today as they were in the past (Figure 4.27B), but they can still be found in a few locations worldwide. Modern stromatolites are noted in hypersaline lakes or lagoons, such as Shark Bay, Australia, or other hypersaline lakes in Western Australia, such as Lake Thetis. Some freshwater stromatolites have also been noted. The only modern stromatolite that grows in an open marine environment is known from the Exuma Cays, Bahamas in tidal channels, sandy embayments, and intertidal settings. These modern stromatolites provide insights into the processes and conditions that existed during the early stages of life on earth.

In general, stromatolites are built up by the activity of cyanobacteria, which trap and bind sediment particles together with their sticky extracellular substances. Over time, layer upon layer of sediment is deposited and cemented, creating a distinctive laminated structure (Figure 4.27C). The growth of stromatolites is influenced by a variety of factors, including the availability of light, nutrients, and the presence of grazing organisms. The cyanobacteria in stromatolites are photosynthetic, using sunlight to convert carbon dioxide and water into organic matter, and releasing oxygen as a byproduct. This photosynthetic activity helps to create an environment favorable for other organisms to colonize and contributes to the development of the layered structure. Although cyanobacteria – photosynthetic microbes – are thought to be the dominant organisms that formed ancient stromatolites, some stromatolites were definitely formed in the deep ocean or in caves where sunlight cannot penetrate. This suggests that at least some stromatolites are formed by non-photosynthetic microbes, such as chemotrophic bacteria (sulfate-reducing bacteria).

4.5.3 Trace Fossils

Trace fossils are preserved evidence of an organism's activity resulting in the formation of tracks, trails, burrows and feeding marks (see Figure 4.26). Tracks and trails are impressions left by an organism as it moved across the surface of the sediment, while burrows are simple tubes or complex networks created by an organism that lived within the sediment (see Figure 4.26). Trace fossils, thus, provide insights into the behavior, locomotion, and feeding habits of ancient organisms. The associations of trace fossils (i.e., trace fossil assemblages) reflect environmental conditions such as water depth (bathymetry), salinity, and the nature of substrate in or on which they were formed (see Seilacher, 1964, 1967). This concept of using trace fossil assemblages for inferring environmental conditions is called ichnofacies (see Seilacher, 1964, 1967).

The tracks, trails, burrows, borings, and other structures made by organisms on bedding surfaces or within beds are collectively known as ichnofossils and its study (of trace fossils) forms a discipline called ichnology. Trace fossils, based on their particular characteristics that relate to major behavioral traits, are grouped into ichnogenera constituting several individuals, called ichnospecies. Sedimentary facies defined on the basis of trace fossils are called ichnofacies, wherein each ichnofacies includes several ichnogenera (and ichnospecies) (Seilacher, 1967). Seilacher (1967) established six ichnofacies, based on characteristic ichnogenera such as *Skolithos*, *Cruziana*, *Zoophycos*, and *Nereites* (based on bathymetric criteria) (Figure 4.28A). The *Glossifungites* ichnofacies was erected to represent traces noted in firm to hardgrounds marine surfaces, and *Scoyenia* ichnofacies represented

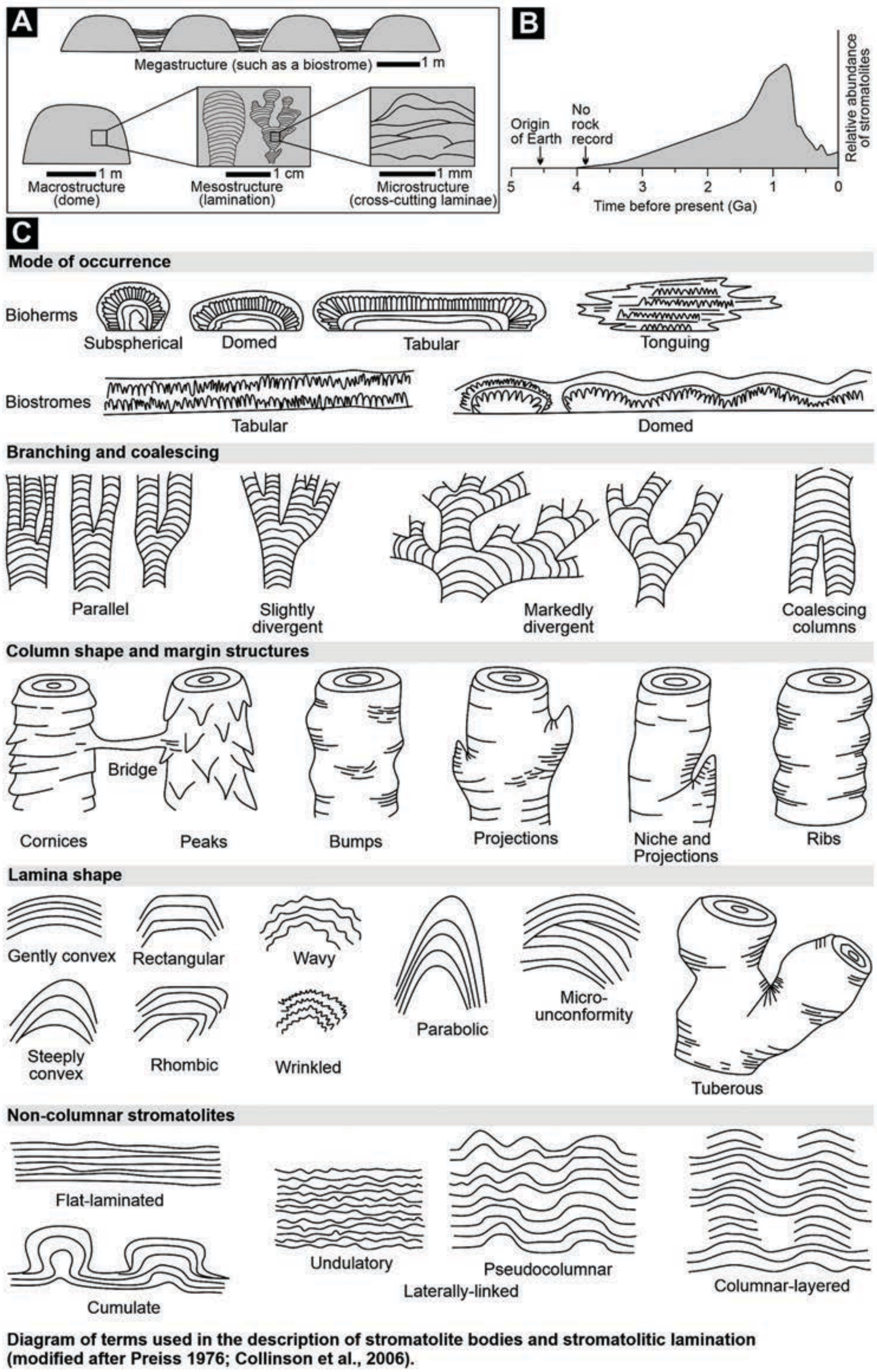

FIGURE 4.27 Stromatolites. A: Descriptive terms used for size differentiation into mega-, macro-, meso-, and microscales. B: Relative abundance of stromatolites through time. C: Stromatolite terminology. (Modified after Preiss 1976.)

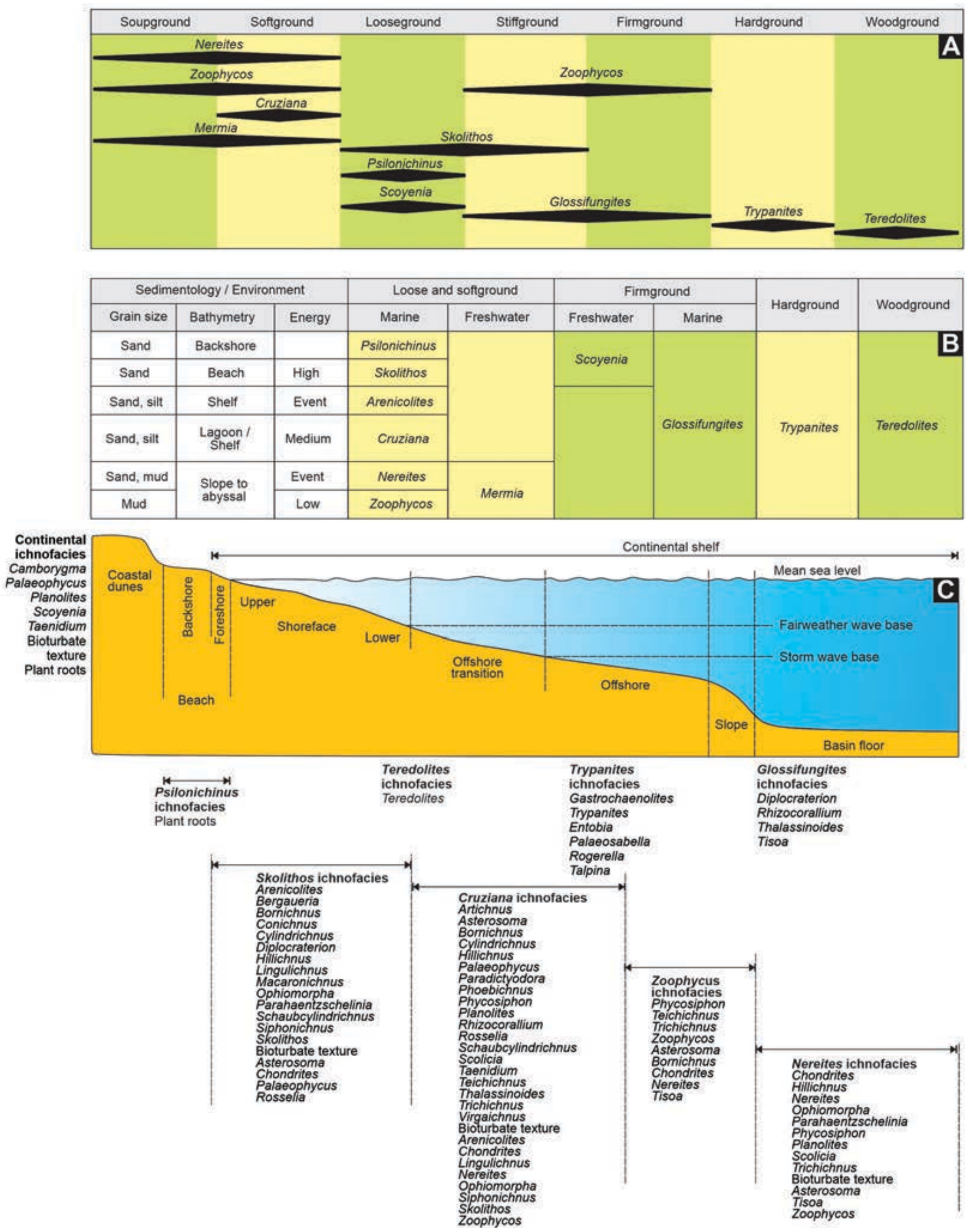

FIGURE 4.28 Ichnofacies. A: Position of major ichnofacies in marine and continental environments. (Modified after Benton and Harper, 1997; Miller, 2007.) B: Relationship of major ichnofacies with the depositional environment (Modified after Bromley, 1996.) C: Distribution of ichnofacies with major ichnospecies through marine and continental environments. (Modified after Buatois and Mángano, 2011.)

non-marine environments (freshwater) (Figure 4.28A). Later, Frey and Seilacher (1980) established the *Trypanites* ichnofacies for hardgrounds, *Teredolites* ichnofacies (by Pemberton and Rahmani, 1984) for borings in wood (woodgrounds), and *Psilonichus* ichnofacies by Frey and Pemberton (1987) for softgrounds in the marine to non-marine environment (Figure 4.28A). Later Bromley (1996) proposed some additional ichnofacies (Figure 4.28B). However, for sedimentologists, the *Skolithos*, *Cruziana*, *Nereites*, and *Zoophycos* ichnofacies are particularly useful as they provide the maximum information for inferring paleoenvironmental conditions and bathymetric control (Figure 4.28C) (see also Buatois and Mángano, 2011; Knaust, 2017 and references therein).

4.6 STRATIGRAPHIC WAY-UP

A brief note on the usefulness of sedimentary structures vis-à-vis their use for identifying the right way up of a bed as they were originally deposited (i.e., "stratigraphic up" or "younging upwards") is warranted (see Figure 4.29). Several sedimentary structures enable us to determine the relation of "top" to "bottom" of a bed at the time it was deposited or may have been overturned by subsequent deformation (see Tucker, 2001; Collinson et al., 2006).

Cross-stratification (cross-bedding or cross-lamination) can be used as a "stratigraphic up" indicator in three ways: (a) the troughs of trough cross-beds, in which the concave part of the trough always points up; (b) the tangential (end) part of the cross-bed is always on the bottom of the cross-bed; and (c) many cross-beds are stacked, and truncation always happens up-section (Figure 4.29A). Graded bedding displays changes in grain or clast sizes from the base of the bed to its top (either increase or decrease in size) (Figure 4.29B). In a normally graded beds, the grain or clast sizes are largest at the base and increasingly smaller at the top of the bed, i.e., fining upwards (often noted in turbidite deposits) (Figure 4.29B). Reverse grading or coarsening upwards is noted in some alluvial fan deposits (Figure 4. 29B). Sole markings are noted in beds deposited by density underflows (such as turbidites) that are characterized by erosional bases (i.e., at the base of a bed) (Figure 4.29C). These structures include flute casts, grooves, and tool marks. Load casts are formed when a higher-density layer (i.e., composed of sand) is deposited on a lower-density one (such as mud) (Figure 4.29D). Characteristic structures include flame structures that are forced up from the underlying low-density layer (Figure 4.29D). Mud cracks are formed in a desiccated mud layer, on a dried-out lake bed or tidal flat, and are often filled by an overlying sand layer (Figure 4.29E). In cross section, they typically show a profile wherein the filled sand layers are narrow at the base and wider at the top resembling a V-shape. Ripple marks are typically rounded at the base (trough) and sharp (or truncated) at the top (crest) (Figure 4.29F). Rain drops or rain prints are small, concave imprints made by rain when it falls on soft sediment or in fine-grained rocks like siltstones and shales (Figure 4.29G). They occur as circular or ovate depressions. These impressions or small craters are produced by the force of the falling raindrops onto the sediment, hence, these structures are good way-up indicators. Crenulation lineations (Figure 4.29H) are very small folds that deform the cleavage, hence can be used for identifying the way-up of a bed. Water escape structures (Dishes; Figure 4.29I) are structures noted in thick sand (or sandstone) and are characterized by concave-up, bowl-like shapes. Hence, these structures are good way-up indicators. Erosional bed contacts are another example of identifying stratigraphic way-up (Figure 4.29J). Erosion across a sediment surface produces scours where coarser sediments may either fill the entire scour, or may occur as a veneer or lag on the erosion surface. These are commonly noted in channels but can also form over any depositional surface. Fossils in life position (such as stromatolites, coral, and articulated crinoids) provide a paleo up direction (Figure 4.29K). Similarly, trace fossils such as the U-shaped burrows of suspension feeders like *Diplocraterion* (and of *Cruziana* ichnofacies in general) along with traces of root systems provide a good evidence of stratigraphic way-up (Figure 4.29L). Grazing traces (pascichnia) such as the inchnogenera *Nereites* (Figure 4.29M) provide good evidence of stratigraphic way-up. Pascichnia are feed intake traces (horizontal feeding traces) left by the grazing

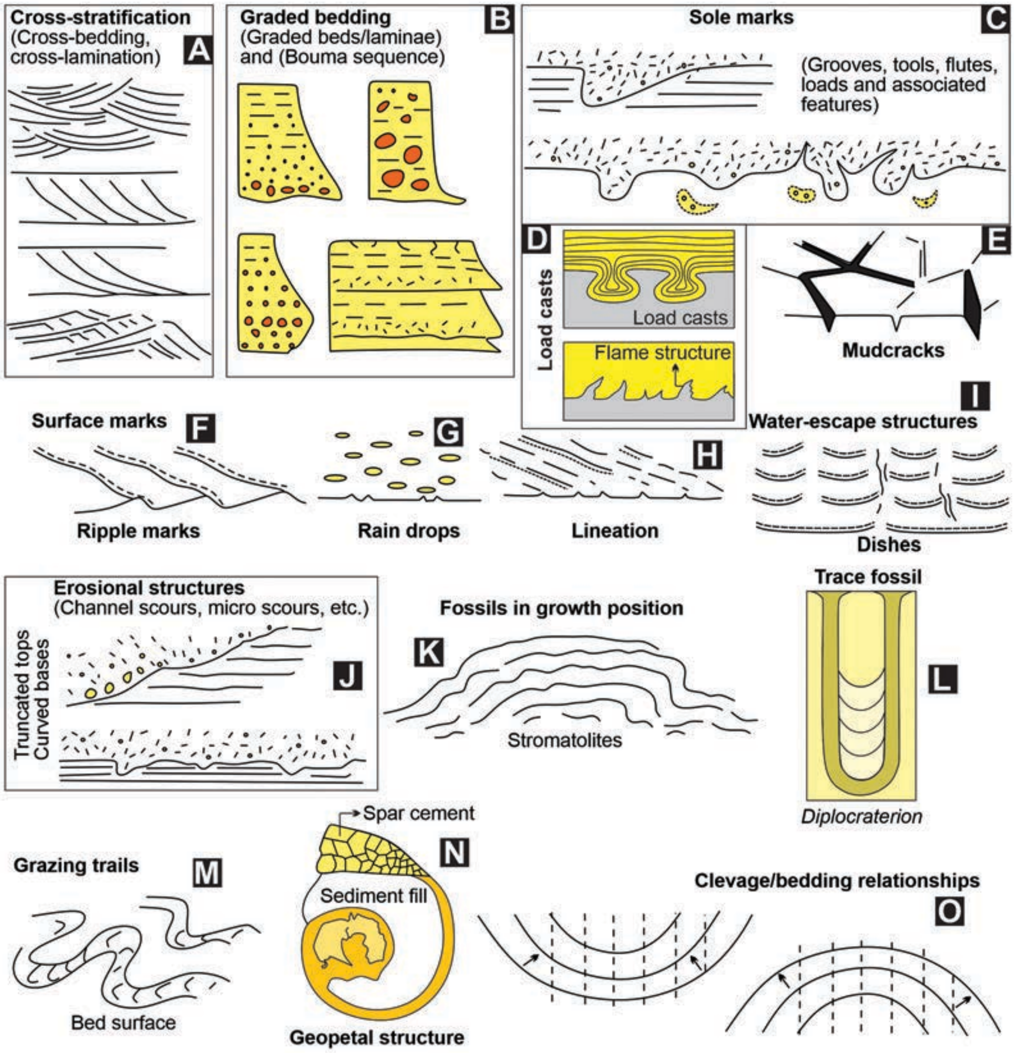

FIGURE 4.29 Stratigraphic way-up structures. (Modified after Collinson et al., 2006.)

animal on soft soil or the substrate's interface as noted in *Nereites* (Figure 4.29M). Additionally, geopetal structures also called Cavity fills (where the initial fill is often of the original sediments but the remaining part is filled by calcite) provide another evidence of way-up direction (Figure 4.29N). Neptunian dikes are formed when the top of a bed is temporarily exposed and a fissure develops either by an earthquake or through the solution of carbonates. These fissures are then filled either during the period of exposure by contemporaneous, possibly wind-blown material, or as and when deposition resumes, by the same material as the overlying bed.

4.7 PALEOCURRENT ANALYSIS FROM SEDIMENTARY STRUCTURES

Paleocurrent analysis, is identifying the orientation and shape of a sedimentary structure to enable the direction of currents and sediment transport. This information helps in better understanding depositional environments, sediment transport pathways, and on a much broader scale, paleogeographic reconstructions. The commonly used sedimentary structures for inferring paleocurrent direction are cross-stratification (cross-bedding), ripple marks, graded bedding, and mud cracks. The direction of the inclined layers (as in cross-bedding) reflects the direction of currents, i.e., if the cross-beds are inclined towards the north, then the sediment was transported from the south. Similarly, the shape and orientation of ripple marks also provide clues about the direction of the current; in asymmetrical ripple marks, the steeper side of the ripple faces upstream. In graded bedding, coarser particles are found at the bottom of the layer, and finer particles accumulate on the top. The direction of the graded bedding thus indicates the direction of the sediment transport. The orientation of mud cracks indicates the direction in which the sediment was exposed to air and dried out, and is related to the direction of water flow. Clast imbrication (in conglomerates) is also used as pleoflow direction indicator wherein the discoid gravel clasts become oriented and where one of the two longer axes dips upstream, when viewed side-on. Hence, the direction of dip of the clasts will be 180 degrees from the direction of palaeoflow. Flute casts are local scours in the substrata generated by vortices within a flow leaving an asymmetric mark on the floor of the flow, with the steep edge in the direction of the upstream side. Channel and scour margins can also be used as a pleoflow direction indicator as the cut bank of a channel lies parallel to the direction of flow.

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Section III

Composition and Classification of Sedimentary Rocks

This section includes Chapters 5–7, on the composition, classification, depositional environments, origin, and diagenesis of sedimentary rocks such as siliciclastic sedimentary rocks (Chapter 5: sandstones, conglomerates, and shales), carbonate rocks (Chapter 6: limestones and dolomites), and chemical, biochemical, and carbonaceous sedimentary rocks (Chapter 7: evaporites, cherts, iron-rich sedimentary rocks, phosphorites, oil shales and coals). The emphasis of this section is on detailing the characteristics and classification of each rock type and elaborating on their respective depositional environments.

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5 Siliciclastic Sedimentary Rocks

5.1 SHALES AND MUDROCKS

A fine-grained, siliciclastic sedimentary rock containing more than 50% of siliciclastic grains that are less than 0.062 (1/ 256) mm is referred to as shale, i.e., a rock type that is dominantly made up of silt-sized (1/16-1/256 mm) and clay-sized (< 1/256 mm) grains (Table 5.1). However, for all such fine-grained rocks, some prefer to use the term mudrock rather than shale, and divide mudrocks into non-laminated (mudstones) or laminated (shales) types (see Blatt et al., 1980) (Figure 5.1).

Mudstones are then categorized as fine-grained, non-laminated, non-fissile rocks with a blocky or massive texture (Figure 5.1A). In mudrocks, the clay minerals are poorly or partially oriented (Figure 5.1A). Hence, fissility is either poorly developed or absent in mudstones (Figure 5.1A). This is primarily due to bioturbation, the presence of increased amounts of quartz silt or calcite, and the flocculation of clays during sedimentation, resulting in a random fabric that is retained on compaction (Figure 5.1A). Hence, mudrocks have a tendency to break as elongated, irregular blocks along conchoidal fractures. Mudrocks are also referred to as pelites (Greek name) or lutites (Latin root name). Prothero et al. (1996), using size fraction, subdivided a mudrock into siltstone (containing >2/3 silt over clay), mudstone (containing between 1/3 and 2/3 clay with the rest being silt) and claystone (clay constitutes more than 2/3 of the total fraction).

The term shale, on the other hand, is restricted to fine-grained rocks that display lamination or fissility (Figure 5.1B). Fissility is the ability of the rock to split easily into thin layers or possess a strong tendency to break in one direction, parallel to the bedding plane; the thin layers are generally <10 mm. Compaction, diagenesis, and deformation reorganize clay minerals in strong parallelism with each other, thus producing strong fissility (Figure 5.1B). Thus, in contrast to mudrocks, shales have a strong tendency to break along nearly parallel surfaces, forming thin flat chips (fissility) (Figure 5.1B).

Both shales and mudrocks contain roughly 95% of the organic matter in all sedimentary rocks. However, this amounts to less than 1% by mass in an average shale. Additionally, shale is the most common sedimentary rock that accounts for about 70% of the earth's crust; ~55% of all sedimentary rocks are shale. However, shales have often been overlooked and understudied as a group primarily for four reasons: (a) they are fine-grained, hence difficult to study with an ordinary petrographic microscope; (b) advanced techniques such as scanning electron microscopes and x-ray diffraction analysis are needed to know their individual constituents and internal structure; (c) they generally do not show clear sedimentary and biogenic structures as noted in coarser clastic rocks and limestones; and (d) they do not generally form visible steep cliffs, and soil (supported by vegetation) covers most of their outcrops.

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TABLE 5.1
Simplified classification of clastic sediments and sedimentary rocks based on mechanical origin

				Rock type			
Clast size	**Size class[a]**	**Sediment / Rock name**		**If rounded to subangular**	**If angular**	**If rounded to subangular**	**If angular**
<0.32 mm	clay	Mudstone, siltstone, shale	Lutaceous or argillaceous	Mud	Clay	Mudrock	Claystone
0.32–0.63 mm	silt			Mud	Silt		Siltstone
0.63–0.125 mm	very fine sand	Sandstone	Arenaceous	Sand (Latin: *arenite*)	Sand	Sandstone	Sandstone
0.125–0.25 mm	fine sand						
0.25–0.5 mm	medium sand						
0.5–1 mm	coarse sand						
1–2 mm	very coarse sand						
2–4 mm	granules	Conglomerate	Rudaceous	Gravel (Latin: *rudite*)	Rubble	Conglomerate	Breccia
4–64 mm	pebbles						
64–256 mm	cobbles						
256 mm	boulders						

Note:
[a] Based on Udden-Wentworth sediment size classification (after Wentworth, 1922).

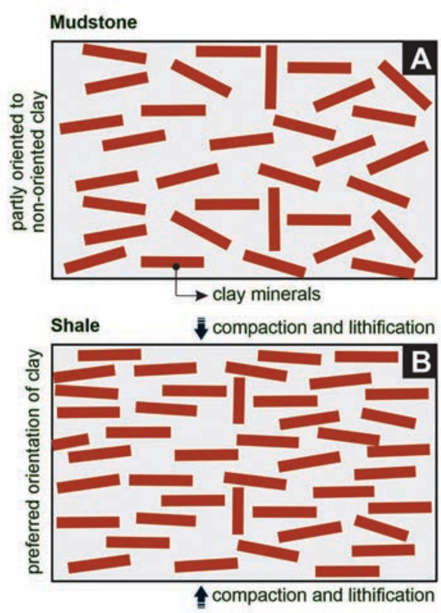


FIGURE 5.1 Differences in clay mineral alignment between mudrock and shale. A: In mudrocks, the clay minerals are poorly or partially oriented. B: In shales, the clay minerals are in strong parallelism with each other, thus producing strong fissility.

5.1.1 Mineralogical and Chemical Composition of Shales

Clay minerals, such as kaolinite, illite, smectite, and chlorite, form major components. The other associated components are fine-grained quartz and feldspars, carbonate minerals (mainly calcite and dolomite, with little siderite), sulfides (pyrite and marcasite), iron oxides (goethite), heavy minerals, and a small amount of organic carbon. During burial diagenesis, the carbonate minerals and sulfides form cements or replacement minerals; the terrigenous (detrital) minerals such as quartz, feldspars, and clays, may also form the same. In general, the clay minerals are strongly affected by diagenesis and their relative proportion changes with time such that the Mesozoic rocks have higher proportions of illite and chlorite as compared to kaolinite and smectite; kaolinite and smectite form illite and chlorite via diagenetic alteration.

The mineralogical composition of shales drives its chemical composition. Hence, SiO_2 is the most abundant constituent (57–68 %), followed by Al_2O_3 (16–19 %) which mainly comes from clay minerals and feldspars. Iron (Fe) comes largely from hematite and goethite (iron oxide minerals), biotite, and in minor amounts from siderite, and ankerite. The clay minerals supply K_2O and MgO, although some Mg may be supplied by dolomite and K from feldspars. Additionally, clay minerals (such as smectites) and sodium plagioclase provided Na, whereas carbonate minerals (calcite, dolomite) and calcium-rich plagioclase provide Ca.

5.1.2 Classification

The classifications (see also Spears, 1980; Potter et al., 1980) subdivide shales based on: (a) grain size, lamination, and degree of induration (this is the most widely used classification scheme); (b) texture (i.e., the amount of fine-grained silt and clay particles); (c) mineral composition (i.e., based on x-ray diffraction analysis); (d) cementing materials or the type of cementation; (e) depositional environment; and (f) the amount of organic matter content.

The classification based on grain size, lamination, and degree of induration uses the relative amount of silt and clay, degree of induration (the hardness of a rock), and the presence or absence of fissile laminations (see Table 5.2). The classification emphasizes the importance of clay-sized constituents and bedding thickness, i.e., whether shale is bedded or laminated. Based on this classification, shales are subdivided into mudstone (33–65% clay-size constituents and bedded) or mud shale (33–65% clay-size constituents and laminated), and claystone (66–100% clay-size constituents and bedded) or clay shale (66–100% clay-size constituents and laminated); fine-grained siliciclastic rocks with <33% clay-sized constituents are grouped as siltstones (see Table 5.2).

The classification based on texture (see Spears, 1980; Potter et al., 2005) uses the amount of fine-grained silt and clay particles. When the clay fraction (less than 4 microns) forms more than 67% of the rock, the terms claystone or clay shale are used (see Table 5.3). When the clay fraction forms 33–66%, the terms mudstone or mud shale are used, but when silt (more than 4 microns) forms 67% or more, the term siltstone is used (Table 5.3). Silt-shale and clay shale are called argillaceous shales, whereas those that contain appreciable amounts of sands are called sandy shales or arenaceous shales.

The classification based on mineral composition uses x-ray diffraction analysis to determine shale mineralogy within the rock (see Pettijohn, 1957; Picard, 1971; Lewan, 1978). Based on this, shales are classified as quartzose (predominance of quartz), feldspathic (predominance of feldspar), or micaceous (predominance of mice).

The classification based on cementing materials (or the type of cementation) uses the presence of common cementing materials such as silica, iron oxide and calcite or lime. Accordingly, shales are classified as siliceous, ferruginous, or calcareous (sometimes also called limey), respectively.

Shales are also classified based on their depositional environment, ranging from lacustrine (continental) to deltaic (transitional) and marine, and are correspondingly referred to as lacustrine, deltaic,

TABLE 5.2
Classification of shales and siltstones based on grain size, lamination, and degree of induration (more than 50% grains are less than 0.063 mm size)

Type	% Clay-sized constituents	0–32	33–65	66–100
Non-indurated	Field description	Gritty	Loamy	Fat or slick
	Bed >10 mm	Bedded silt	Bedded mud	Bedded claymud
	Laminae <10 mm	Laminated silt	Laminated mud	Laminated claymud
Indurated	Bed >10 mm	Bedded siltstone	Mudstone	Claystone
	Laminae <10 mm	Laminated siltstone	Mudstone	Claystone
	Low	Quartz argillite	Argillite	
Metamorphosed	Degree of metamorphism	Quartz shale	Shale	
	High	Phyllite and/or mica schist		

Source: Modified after Spears (1980); see also Potter et al. (1980).

TABLE 5.3
Classification of mudrock

Particle size and proportions	Non-fissile	Fissile
4–63µm, >67%	Siltstone	Silt-shale
<4µm, >67%	Claystone	Clay shale
<63µm, no proportions	Mudstone	Mud shale

Source: Modified after Stow (1980); see also Potter (2005).

and marine shales (see Boggs, 2009). The lacustrine deposits are characterized by a mixture of clay, silt, and sands, and inorganic carbonate precipitates, with invertebrates and plant remains. Deltaic deposits are generally paralic (i.e., with sequences of shales and sandstones formed as a result of alternating marine transgressions and regressions). Marine environment shales have a homogeneous rock sequence (i.e., non-paralic), formed at greater depth with oxygen deficiency and a higher concentration of illite/montmorillonite clay minerals. These are generally darker in color and richer in marine planktic fossils as compared to shales deposited in lacustrine and deltaic environments.

Shales are also classified based on their organic matter content (Krumbein and Sloss, 1963). When the dominant organic matter is from a plant source, such as pollen grains, stems, and leaves, the shale is classified as carbonaceous, and the depositional environment is usually continental (lacustrine) or transitional (deltaic or lagoon). When the dominant organic matter content is from animal fragments such as fossils, the shale is classified as bituminous and its depositional environment is usually deltaic or marine. Both carbonaceous and bituminous shales are important source rocks for the generation of petroleum oil and gas depending on their amount, and type of kerogen content.

5.1.3 Sedimentary Structures in Mudrocks

By studying the sedimentary structures is mudrocks, more so when analyzed in association with the occurrences of associated body and trace fossils, the depositional history of a basin can be inferred and reconstructed. Briefly, these are enumerated below (see Chapter 4 for details and the types of sedimentary structures). Lamination is one of the most important and commonly noted sedimentary structures in mudrocks. It is mostly noted as variations in the proportions of silt and clay, but can occur as thin layers of carbonates (pelagic rain or algal mats), minor textural variations within muddy layers, and very rarely by variations of organic matter (kerogen). Laminations can be continuous or discontinuous, wavy or planar, massive or graded, inclined (cross-laminated) or convoluted, random or rhythmic. Present on the bottoms of some thin silty laminations are small flutes and load-cast structures with borings, traces, and impressions made by bottom dwellers. The orientation of flutes and cross-laminations provide useful information on the direction of paleocurrents. Additionally, these sedimentary structures also help to identify the types of bottom currents such as distal turbidity currents, contour currents, or hemipelagic rains of fine terrigenous debris.

5.1.4 Concretions in Mudrocks

Carbonate concretions occur in sedimentary rocks of all geological ages, mostly in sandstones and mudstones as ovoid rock bodies that bulge out from natural outcrops; harder or better cemented than their host rocks (see Marshall and Pirrie, 2013). These concretions are, in simple terms, patches of cemented sediment where the mineral cement between grains is different from that in the host (which may not be cemented at all).

Calcite (calcium carbonate) is the most common cement but siderite (iron carbonate) is commonly noted in non-marine sediments. Magnesium-rich and mixed phases (such as dolomite or ankerite) also occur. Pyrite or fools' gold is often associated with many carbonate concretions in marine mudstones. Many mudstones also contain networks of fractures, "septarian" cracks, and are called septarian concretions (Figure 5.2) (see also Pratt, 2001; Paxton et al., 2021). These carbonate concretions in marine mudrocks contain calcite cements that and are robust proxies of ambient marine

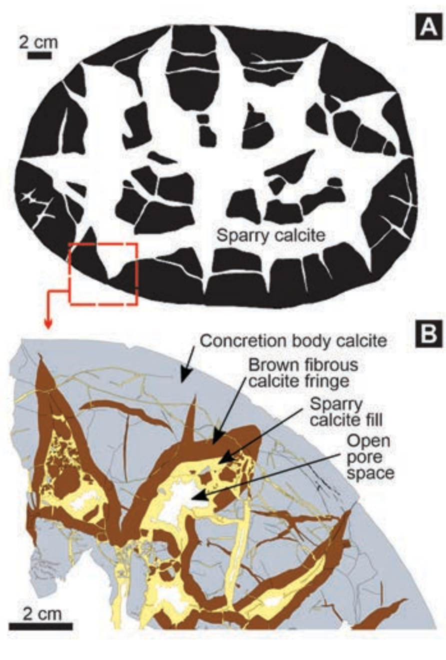


FIGURE 5.2 Septarian carbonate concretions. A: Network of spar-filled cracks within the mudstone-hosted concretion. (Modified from Pratt, 2001.). B: Calcite phases (cement) within the septarian concretion. (Modified after Paxton et al., 2021.)

benthic palaeotemperatures, shallow subsurface environments, and burial diagenetic conditions (see Paxton et al., 2021). Most concretions range in size from a few centimeters to several meters in diameter: the smaller (cm-scale ones) are commonly called nodules. Carbonate concretions occur in both marine and non-marine mudstones where they tend to be associated with dark gray or black sediments. They are generally concentrated in particular layers within the succession rather than randomly dispersed through the rocks.

Thus, mudrocks (mudstones) play an important role in understanding the history of sedimentary basins. In marine and lacustrine basins, they are (a) the most widespread and most continuously occurring lilthologies; (b) extensively used to infer paleo-oxygen levels; (c) used to infer provenance, where their silt content provides a rough estimate for shoreline distance (silty mudrock is noted in more proximal setting, whereas more clay-rich rocks are deposited more distally; (d) excellent tools for inferring paleoenvironment as mudrocks almost always contain rich assemblages of pelagic microfossils (such as foraminifera and ostracods, among others) and spores; and (e) good indicators of thermal history after deposition based on the study of their vitrinite reflectance properties.

5.2 SANDSTONES

Of all the sedimentary rocks, sandstones are pervasive, and make up 20–25% of the sedimentary rock record. The silicate grains, ranging from 0.0625 to 2 mm in size, make up the framework fraction. The 2 mm size upper boundary separates sand from gravel and their lithified counterparts, conglomerate and breccia, respectively. The 0.0625 mm size lower boundary is the limit of the human eye to discern grains; we can only see grains that are larger than 0.0625 mm in diameter. Grains smaller than that (i.e., mud (silt + clay)) are not discernible by the human eye. Hence, rocks consisting of them, like mudstones, appear homogeneous to us. To qualify itself as a sandstone, a clastic sedimentary rock must contain more than 25% sand over mud. Moreover, if enough particles with grain size >2 mm (rudite) are present, in general more than 5%, the rock is a conglomerate (or a breccia) (discussed later in the chapter).

To better understand sandstones, it is imperative to understand the idea of cement, matrix, framework grains and interstitial pore spaces (pores) (see Figure 5.3). The interstitial pore spaces among the framework grains (of detrital origin) contain various amounts of cement (i.e., the mineral precipitate in pore spaces after deposition) and very fine-sized (<0.03 mm) material called matrix (the fine-grained fraction consisting of mud) (Figure 5.3). The most common sandstone cements are of

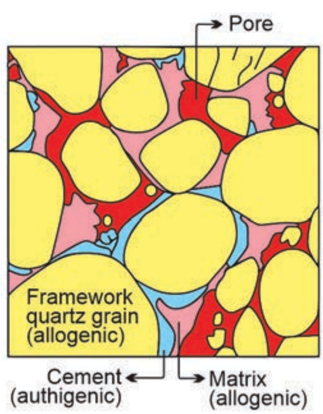


FIGURE 5.3 Relationship of framework grains, cement, matrix, and interstitial pore spaces. Matrix is fine-grained (silt and clay size) and occurs in between coarse-grained particles, whereas cement is the binding material that is present between grains.

siliceous material (such as quartz, chalcedony and opal) or of carbonates (such as calcite), but other minerals like oxides, feldspars, zeolites, and authigenic clays can also form cement. The pore spaces (i.e., porosity) between framework grains can be empty or partly or entirely filled. The pore filling is often a combination of fine-grained clastic matrix, cement (such as calcite, quartz, chert, or hematite), and fluids (such as gas, air, oil, and groundwater) (see Figure 5.3).

5.2.1 Framework Mineralogy

The framework fraction of the sandstones is made up of sand- and coarse silt-sized particles such as silicate minerals and rock fragments. In general, only a few kinds of minerals make up the bulk of all sandstones (see Table 5.4); these are briefly discussed below.

5.2.1.1 Quartz

Quartz (SiO_2) is the dominant mineral making up ~50% of the framework fraction in most sandstones. Most quartz grains (mostly single-crystals or monocrystalline quartz) display some

TABLE 5.4
Common minerals and rock fragments in siliciclastic sedimentary rocks

Major minerals (abundance > ~ 1–2%)

Stable minerals (resistant to chemical decomposition)

Quartz – makes up ~65% of sandstone, 30% of shale; 5% of carbonate rock

Less stable minerals

- **Feldspars** – includes K-feldspars (orthoclase, microcline, sanidine, anorthoclase) and plagioclase feldspars (albite, oligoclase, andesine, labradorite, bytownite, anorthite); make up ~10–15% of sandstone, 5% of shale, < 1% of carbonate rock
- **Clay minerals and fine micas** – clay minerals include the kaolinite, illite, smectite (montmorillonite the major component), and chlorite; fine micas are mainly muscovite (sericite) and biotite; they make up 25–35% of total siliciclastic minerals, but may be > 60% in shales

Accessory minerals (abundances < ~ 1–2%)

- **Coarse micas** – mainly muscovite and biotite
- **Heavy minerals** (specific gravity > 2.9)
- **Stable nonopaque minerals** – zircon, tourmaline, rutile, anatase
- **Metastable nonopaque minerals** – amphiboles, pyroxenes, chlorite, garnet, apatite, staurolite, epidote, olivine, sphene, zoisite, clinozoisite, topaz, monazite, among others
- **Stable opaque minerals** – hematite, limonite
- **Metastable opaque minerals** – magnetite, ilmenite, leucoxene

Rock fragments (~10–15% of the siliciclastic grains in sandstone and most of the gravel-sized fractions in conglomerates; shales contain few rock fragments)

- **Igneous rock fragments** –includes clasts of any igneous rock; fine-crystalline volcanic rock and volcanic glass are most common fragments in sandstones
- **Metamorphic rock fragments** – includes metaquartzite, schist, phyllite, slate, argillite, and less commonly gneiss clasts
- **Sedimentary rock fragments** – any type of sedimentary rock possible in conglomerates (such as clasts of fine sandstone, siltstone, shale, and chert; limestone clasts are comparatively rare in sandstones)

Chemical cements (abundance variable)

- **Silicate minerals** – predominantly quartz; associated with chalcedony, opal, feldspars, and zeolites.
- **Carbonate minerals** – mainly calcite; rarely aragonite, dolomite, siderite
- **Iron oxide minerals** – hematite, limonite, goethite, and sulfate minerals such as anhydrite, gypsum, barite

Source: Modified after Boggs (2009).
Note: All numbers are average values.

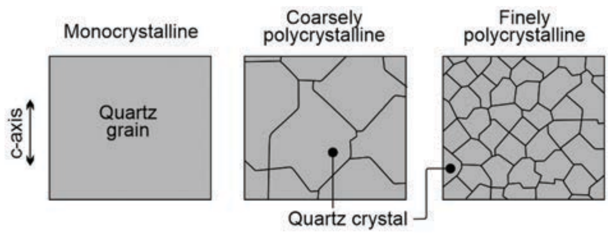


FIGURE 5.4 The types of quartz framework grain.

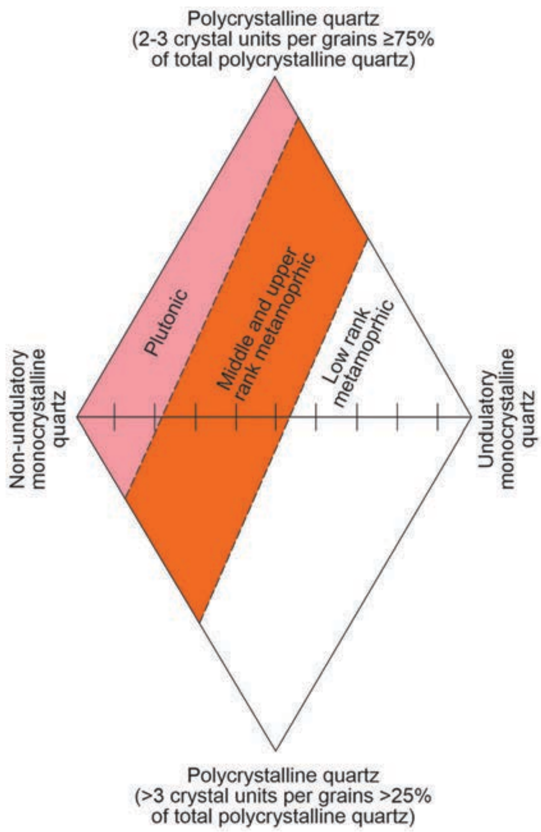


FIGURE 5.5 Diamond diagram. (Modified after Basu et al., 1975; Tortosa et al., 1991.) Quartz grains that show undulatory extinction and polycrystallinity are used as proxies for provenance.

degree of rounding due to abrasion during transport. Quartz occurs as single (monocrystalline) grains or as composite (i.e., with multiple interlocking quartz crystals or polycrystalline) grains (see Figure 5.4). Many quartz grains display undulatory extinction wherein, as the petrographic microscope stage is rotated and the thin section examined under crossed polarizing light, a sweeping pattern of extinction (alternating light and dark fields) is noted. Both polycrystallinity and undulatory extinction have been used as proxies for inferring provenance. The polycrystalline quartz grains are largely derived from metamorphic and plutonic rocks (particularly felsic plutonic ones such as granites), as also from hydrothermal vein deposits and fractures (Figure 5.5) (see Folk, 1974; Basu et al., 1975).

5.2.1.2 Feldspars

On average, in most sandstones, feldspars make up 10–20% of the framework grains and hence are the second most abundant mineral. Feldspars are divided into two broad groups: alkali and plagioclase feldspars. Alkali feldspars constitute a group of minerals in which chemical composition ranges through a complete solid-solution series from $KAlSi_3O_8$ through (K, Na)$AlSi_3O_8$ to $NaAlSi_3O_8$ (Figure 5.6). As potassium-rich feldspars are very common, the alkali feldspars are also called potassium feldspars (or K-spars or K-feldspars). Common members include orthoclase, microcline, and sanidine (Figure 5.6). The plagioclase feldspars form a solid-solution series ranging in composition from $NaAlSi_3O_8$ (albite) through $CaA1_2Si_2O_8$ (anorthite) with a general chemical formula for the series as (Na, Ca)(Al, Si)Si_2O_8 (see Figure 5.6). The plagioclase feldspars are commonly distinguished from K-spars on the basis of twinning, an optical property examined under a petrographic microscope (Figure 5.7). The plagioclase feldspars commonly display albite (polysynthetic; striped appearance) and Carlsbad twinning (with two distinct domains; divided into half by a twin plane) (Figure 5.7A), whereas K-feldspars display two sets of polysynthetic twins that are at right angles to each other, thus forming a distinctive crosshatched pattern called tartan twinning (Figure 5.7B). However, it must be mentioned that some K-feldspars (such as sanidine) and some plagioclase feldspars are untwinned; thus, differentiation between them becomes very difficult and also from quartz. The orthoclase (K-feldspars) may show no twinning or Carlsbad twinning.

Compared to quartz, the feldspars are chemically less stable and thus more susceptible to chemical destruction during weathering and diagenesis (see Figure 5.8). Due to their softness, the feldspars

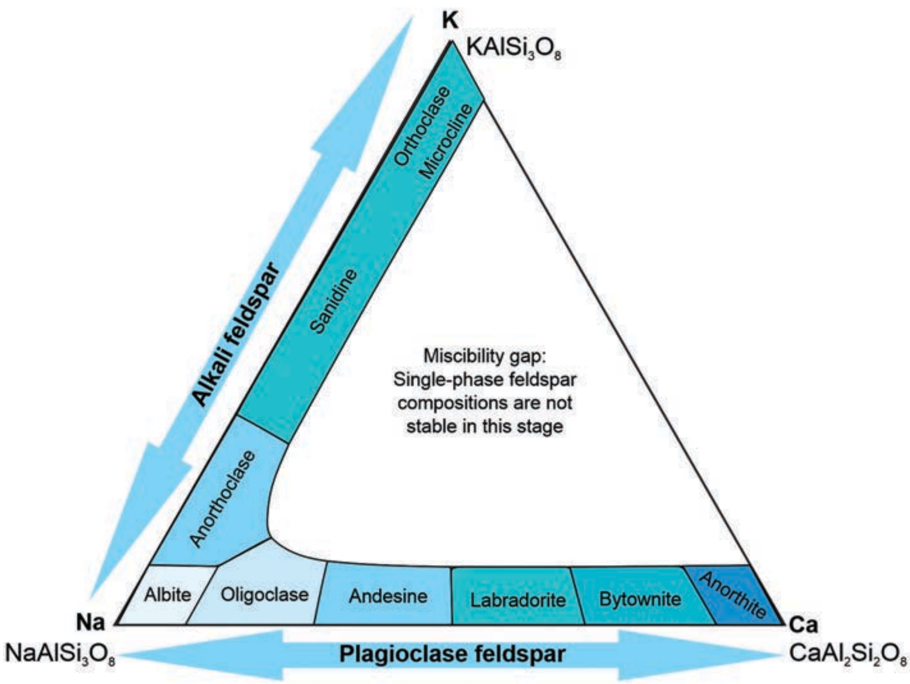


FIGURE 5.6 Feldspar classification. Feldspars are solid-solution minerals with a common formula as (Ca,Na,K)(Si,Al)$_4$O$_8$ expressed as three end members: albite ($NaAlSi_3O_8$), anorthite ($CaAl_2Si_2O_8$), and orthoclase ($KAlSi_3O_8$). The alkali feldspar is mainly solutions of orthoclase and albite (may contain up to 15 wt % anorthite), whereas the plagioclase feldspar is between albite and anorthite (may contain up to 10 wt % orthoclase).

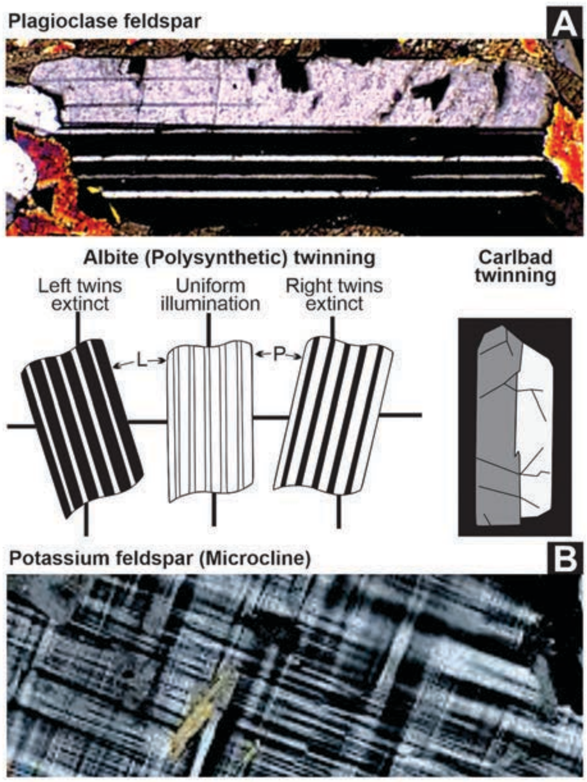


FIGURE 5.7 Twinning of feldspars. A: Polysynthetic and Carlsbad twinning are the two most common twinnings in plagioclase feldspar. Carlsbad twinning is also commonly noted in orthoclase feldspar. B: Crosshatched or tartan twinning of microcline (K-feldspar).

Sandstone minerals	Relative chemical stability
Quartz	Most stable
Clay	↓
Muscovite mica	
Orthoclase feldspar (microcline)	
Albite feldspar	
Anorthite feldspar	Least stable

FIGURE 5.8 Stability of minerals. The chemically less stable minerals are more susceptible to chemical destruction during weathering and diagenesis.

are more readily rounded during transport (in comparison to quartz) and are also more prone to mechanical shattering and breakup due to their cleavage. Feldspars, in comparison to quartz, rarely survive multiple episodes of recycling; hence, the presence of a few feldspar grains in a sedimentary rock does not always suggest that the rock is composed of first-cycle sediments derived directly from crystalline igneous or metamorphic rocks. On the other hand, a high content of feldspars (>25%) is often suggestive of a derivation directly from crystalline source rocks.

To better understand sandstone maturity and mineral composition, a brief note on the stability of minerals is warranted. The femic (i.e., magnesium-, or calcium-rich) minerals like olivine, pyroxene, and amphibole, and Ca-rich sialic (silica- and aluminum-rich) minerals like plagioclase break down to clay minerals very fast, hence, rarely survive multiple cycles of erosion and sediment transport. On the other hand, alkali feldspar, Na-rich plagioclase, and the micas (biotite and muscovite) also alter to clay minerals, but at a much slower rate, thus are present in sands and sandstones. Quartz is the most stable rock-forming mineral, as it does not dissolve in water and is also resistant to physical erosion due to its high hardness (Figure 5.8). Several cycles of erosion and transport concentrate quartz in sediments, as feldspars and other less stable minerals are progressively removed. Hence, beach and aeolian sands, that undergo several cycles of erosion and transport, are quartz-rich (Figure 5.8). Quartz-rich sediments also contain erosion-resistant minerals, such as zircon, tourmaline, and rutile. Deep sea turbidites that are rapidly deposited often contain abundant feldspars and micas.

5.2.1.3 Accessory Framework Minerals

Accessory minerals are those that have an average abundance of ~1–2% within a sedimentary rock such as mica (muscovite: white mica, and biotite: dark mica; these are detrital components), and heavy minerals. The micas, on average, make up < 0.5% in a siliciclastic sedimentary rock; they rarely make up 2–3% in some sandstones. Micas have a characteristic platy or flaky habit and are largely derived from metamorphic and some plutonic igneous rocks. Biotite, a type of mica, is more abundant in sandstones than the other type, muscovite; the latter is also chemically more stable than biotite.

Heavy minerals are those that have a specific gravity >2.9 (or minerals with a density > 2.9 g/cm^3) (see Table 5.4). Generally, heavy minerals constitute minerals that have specific gravities greater than the two main framework components of sands and sandstones, quartz (2.65) and feldspar (2.54–2.76). Heavy minerals make up <1% of the siliciclastic rocks. They are both chemically stable and unstable (labile) (see Table 5.4). Zircon and rutile are stable heavy minerals that can survive multiple recycling episodes and are generally rounded, suggesting sedimentary sources. Magnetite, pyroxenes, and amphiboles are less stable fractions and are commonly first-cycle sediments that reflect the composition of proximate source rocks. Thus, heavy minerals are robust proxies of sediment source rocks as different source types yield a different set of heavy minerals. But, in general, heavy minerals can be derived from a variety of igneous, metamorphic, and sedimentary rocks.

5.2.1.4 Rock (Lithic) Fragments or Clasts

These are fragments of pre-existing rocks (igneous, metamorphic, or sedimentary) that have not yet been broken down to yield individual mineral grains. They range in size from less than a micron to as big as an apartment block. However, rock fragments of fine-grained source rocks are sand-sized fragments, whereas very coarse-grained source rocks such as granites have clasts of coarse sand size or larger. In an average sandstone, the clasts generally make up ~15–20% of the framework grains, but this range may vary, from 0 to >95%. In sandstones, volcanic rocks are the most common rock fragments, and slate, phyllite, schists in fine-grained metamorphic rocks. In younger rocks and quartzite, volcanic glass is quite common. Chert (microcrystalline quartz), largely derived by the weathering of bedded chert or chert nodules in limestones, is a common rock fragment in sandstones. Rock fragments are very useful proxies in provenance studies (i.e., the study of sediment source rocks) (discussed later in the chapter).

5.2.2 Mineral Cements

In most siliciclastic sedimentary rocks, the framework grains are bound together by some type of mineral cement (see Figure 5.9). The cementing materials may be silicate minerals (quartz and opal) or non-silicate ones (calcite and dolomite); often quartz is the most common cement. It must be noted that in sandstones all cements are secondary minerals that form after deposition and during

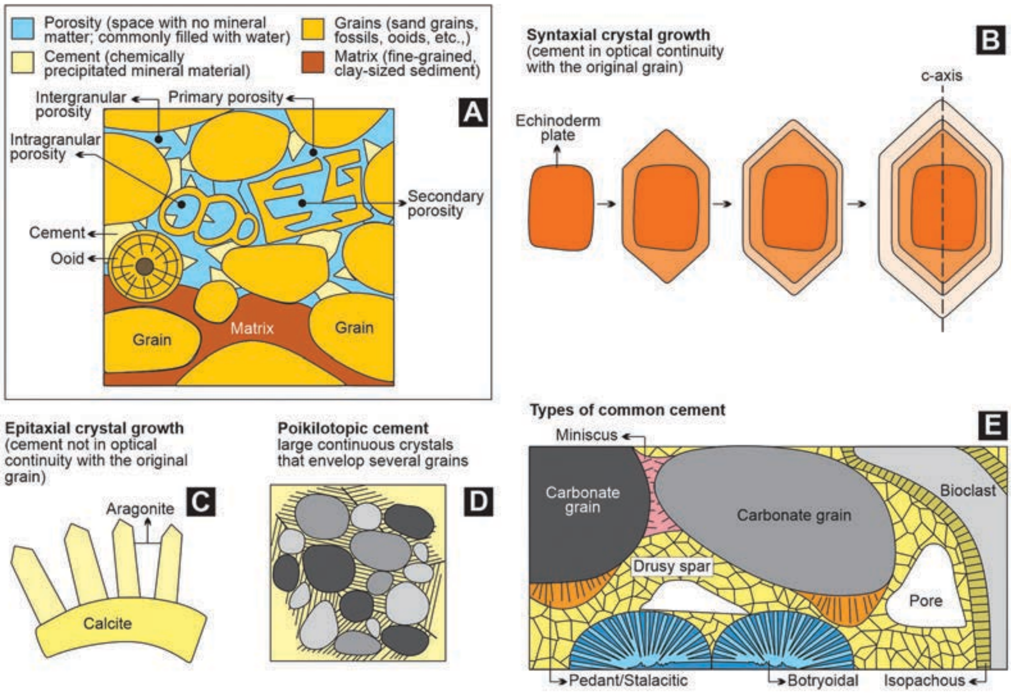


FIGURE 5.9 Relationship of porosity, cement, and crystal growth. A: Types of cement. The cements precipitate in available pore space: intergranular (between grains), intra-granular (within the whorls of a gastropod), or in larger voids. B: Syntaxial overgrowth. The quartz cement is chemically attached to the crystal lattice of an existing quartz grain and the overgrowth retains the crystallographic continuity of the grain. C: Epitaxial growth. Optical continuity is not maintained and the new crystals stick out from the underlying one. D: Poikilotopic cement. In this, large optically continuous crystals envelop several framework grains. E: Types of non-silicate mineral cements.

burial. Cements precipitate in available pore space: intergranular (between grains), intragranular (like the whorls of a gastropod), or in larger voids (see Figure 5.9A).

Overgrowth in sandstones occurs when the quartz cement is chemically attached to the crystal lattice of an existing quartz grain, forming a rim (Figure 5.9B). Such overgrowths that retain the crystallographic continuity of a grain are referred to as syntaxial (Figure 5.9B). In syntaxial overgrowth, optical continuity with the original grain is maintained, such that both the overgrowth and the grain go to extinction in the same position when rotated on the stage of a polarizing microscope (Figure 5.9B). However, in epitaxial growth, this optical continuity is not maintained and the new crystals (with faces) stick out from the underlying one (see Figure 5.9C). In poikilotopic cement, large optically continuous crystals envelop several framework grains (see Figure 5.9D).

The carbonate minerals are the most common non-silicate mineral cements and the types include drusy, isopachous, pendant/meniscus, and botryoidal, among others (see Figure 5.9E). Drusy cements consist of calcite rhomb mosaics that line and fill pores and intra-skeletal chambers (see Figure 5.9E). Isopachous cements are made of fibrous aragonite, that commonly rim grains or bioclasts; the cement fringe is of equal thickness throughout, hence called isopachous (see Figure 5.9E). Pendant (stalactitic or dripstone) cements accumulate on the low point of grains during gravity drainage of interstitial fluid, whereas the meniscus cement forms at grain contacts where water is retained by surface tension forces (see Figure 5.9E). Less common carbonate cements are dolomite and siderite (iron carbonate). In summary, the most common cements are carbonates (calcite, aragonite, dolomite, and siderite), silicates (quartz, opal, clay minerals, and zeolites), sulfates (gypsum and anhydrite) and chlorides (halite).

5.2.3 Matrix and Clay Minerals

Grains in sandstones smaller than about 0.03 mm that fill interstitial spaces among framework grains are matrix minerals (see Figure 5.4). These include fine-sized micas, quartz, and feldspars; clay minerals are the most abundant ones. The clay minerals belong to the phyllosilicate mineral group; the most common ones are illite [$K_2(Si_6Al_2)Al_4O_{20}(0H)_4$], smectite (montmorillonite) [$(Al, Mg)_8(Si_4O_{10})_3(OH)_{10} \cdot 12H_2O$], kaolinite [$Al_2Si_2O_5(0H)_4$], and chlorite [$(Mg, Fe)_5(Al, Fe^{3+})_2Si_3O_{10}(OH)_8$]. The clay minerals principally form as secondary minerals during subaerial weathering and hydrolysis, although they can also form by subaqueous weathering within the marine environment and during burial diagenesis.

5.2.4 Classification of Sandstones

The descriptive classification of sandstones is based fundamentally on framework mineralogy, although the relative abundance of matrix (clastic material finer than 30 mm, i.e., coarse silt) also plays a major role in some classifications. Although mineralogy is the principal basis for classifying sandstones, finding a classification that is suitable for all types of sandstones and acceptable to most geologists has proven to be an elusive goal. In fact, more than 50 different classifications for sandstones have been proposed (see Friedman and Sanders, 1978; Garzanti, 2017), but none has received widespread acceptance. The range of classifications includes those that are either all-inclusive and thus tend to be too complicated and unwieldy for general use, and those that are oversimplified and thus may convey too little useful information. The classification that is widely used is the one proposed by Dott (1964; this classification was later modified after several workers; for details see Pettijohn et al. 1987).

5.2.4.1 Textural Classification of Sandstone

The textural classification includes particles ranging in size from mud (clay and fine silt) to gravel (see Figure 5.10A). Note that the textural boundaries in this classification scheme are not entirely symmetrical. Ideally, we might expect the boundary between gravel and mud-sand to be set at 50%; however, this is not always the case (see Figure 5.10A). As the particles of gravel size are less abundant than sand and mud, geologists consider a sediment with 30% gravel-size fragments as gravel. If the sediments contain only particles of sand size or smaller, then such a textural classification scheme uses sand, silt, and clay as end members (see Figure 5.10B). Once the textural nomenclature has been established, rocks within each textural group can be further classified on the basis of composition.

5.2.4.2 Mineralogical Classification of Sandstone

A combination of small numbers of framework components such as quartz, feldspars, and lithic (rock) fragments (such as chert and volcanic clasts), make up most sandstones, and hence these three components are used for its classification (Figure 5.11). However, despite these low numbers, no single, acceptable sandstone classification has emerged, over time. The most commonly used sandstone classification involves the QFL plot, a triangular diagram on which quartz (Q), feldspars (F), and lithic (rock) fragments (L) are plotted as end members (Figure 5.11). Recognizing lithic fragment group needs attention as the clasts (of igneous, sedimentary, and metamorphic rocks) consist of more than one crystal. In general, besides containing more than one crystal, the lithic fragments also show an internal texture inherited from the parent material (e.g., igneous, metamorphic, and sedimentary textures). However, this parameter becomes tricky when dealing with

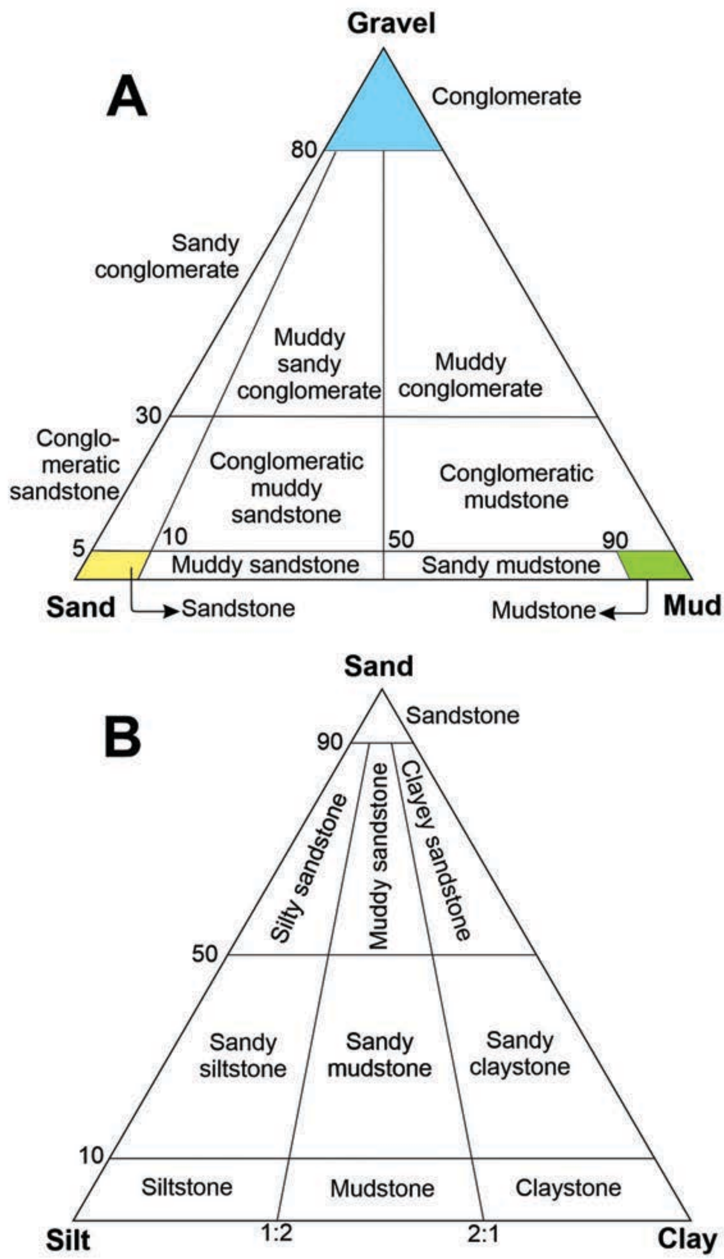


FIGURE 5.10 Grain-size classification (grain size is based on the Wentworth scale). A: Three grades are used: sand (between 1/16 mm and 2 mm), mud (anything smaller than sand and includes silt and clay-sized grains), and gravel (anything larger than sand and includes granules, pebbles, cobbles, and boulders). Notice that the lines that fan downward from the gravel vertex are based on values, expressed as percentages, of the expression mud/sand and mud, meaning that each point on the line, regardless of the gravel content, has the same proportion of sand to mud. B: Sandstone and mudstones (grain-size classification based on Folk, 1951). Based on the fact that $<5\%$ of the rock is larger than sand (gravel), three grades are used: sand (grains between 1/16 mm and 2 mm), silt (grains between 1/16 mm and 1/256 mm), and clay (grains smaller than 1/256 mm).

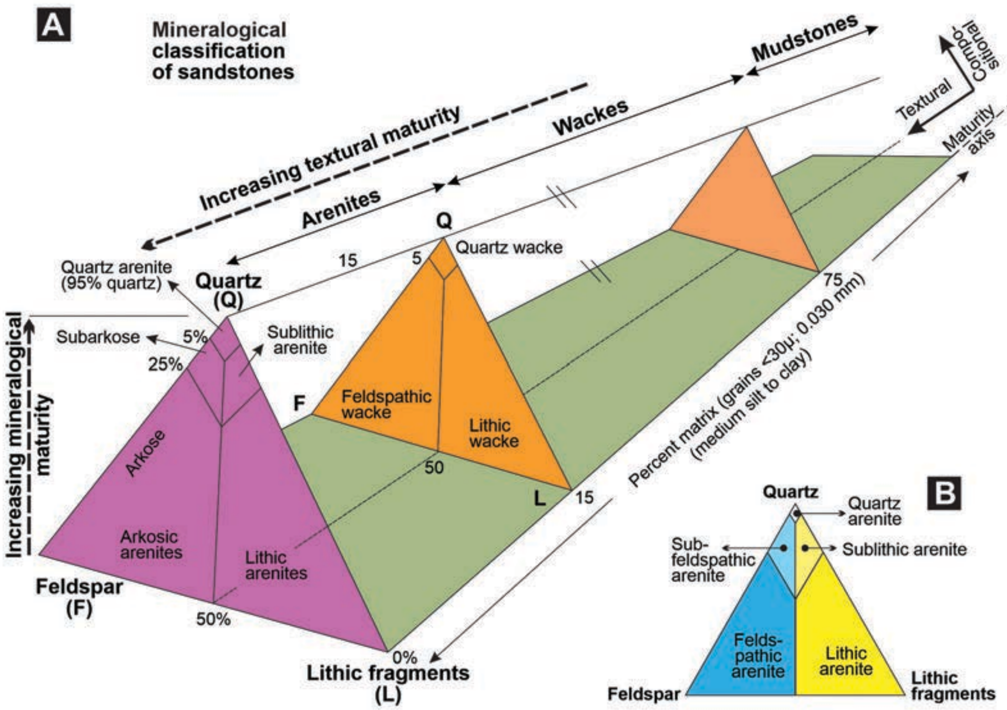


FIGURE 5.11 Mineralogical classification of sandstones. A–B: Classification by Dott (1964). The classification is based on the combination of small numbers of framework components such as quartz (Q), feldspars (F), and lithic (rock) fragments (L). The classification is based on (a) the percentage of matrix (grain size <30 μm) and (b) the proportion of quartz, feldspar, and lithic fragments (>30 μm framework grains). Based on this diagram, a mudstone is if the percentage of matrix is >75%. Sandstones are subdivided into arenites (matrix is <15%) and wackes or graywackes (matrix is >15%). Arenites and wackes are further subdivided based on quartz (Q), feldspar (F), and rock or lithic fragments (L) composition. If quartz is >95% (F + L < 5%), they are classified as quartzarenite and quartzwacke, respectively. Arkosic arenite and arkosic wacke occur when feldspars dominate, but if lithic fragments dominate, the sandstones are classified as lithic arenite and lithic wacke. Feldspathic can be used instead of arkosic. The arenites are further subdivided as subarkose (where quartz is between 75 and 95%, and feldspar > lithic fragments) and sublitarenite (quartz is between 75 and 95%, and lithic fragments > feldspar).

very fine-grained rocks, as the clasts can be misidentified as monomineralic grains – for example, the clasts of cherts of fine-grained limestones; a thin section is the best solution in such cases.

The classification by Gilbert (1982) (modified from Dott, 1964) is the simplest one where sandstones that are mostly free of matrix (<5%) are classified as quartz arenites, feldspathic arenites, or lithic arenites depending upon the relative abundance of the QFL constituents (Figure 5.11). If matrix is there (at least 5%), the terms quartz wacke, feldspathic wacke, and lithic wacke are used (Figure 5.11). Thus, the classification by Gilbert (1982) is based on (a) the percentage of matrix (grain size <30 μm) and (b) the proportion of quartz, feldspar, and lithic fragments in the > 30 μm framework grains (Figure 5.11). Based on this, mudstones are characterized as rocks with the percentage of matrix > 75% (see Figure 5.11). Sandstones are divided into arenites (<15% matrix) and wackes or graywackes (> 15% matrix) (Figure 5.11). Based on their composition in terms of quartz (Q), feldspar (F), and rock or lithic fragments (L), the arenites and wackes are further subdivided. If quartz is >95% (F + L <5%), they are classified as quartzarenite and quartzwacke respectively

(Figure 5.11). When feldspars dominate, they are classified as arkosic arenite and arkosic wacke, but if lithic fragments are more abundant, the sandstones are classified as lithic arenite and lithic wacke (Figure 5.11). Feldspathic can be used instead of arkosic. Further subdivisions of arenites are subarkose (quartz between 75 and 95%, feldspar > lithic fragments) and sublitarenite (quartz between 75 and 95%, lithic fragments > feldspar) (Figure 5.11). Other, later, classifications do not include matrix as part of the classification scheme (see McBride, 1963; Folk et al., 1970).

5.2.5 Sandstone Maturity

Maturity in sandstones is either textural or compositional. Textural maturity is determined by the relative abundance of matrix (i.e., clay content; clay is defined as any grain with a diameter <30 mm), and the degree of sorting and rounding of framework grains (Figure 5.12) (see also Folk, 1951).

5.2.5.1 Textural Maturity

Textural maturity ranges from immature (much clay, framework grains are poorly sorted and poorly rounded) to super-mature (little or no clay, framework grains are well sorted and well

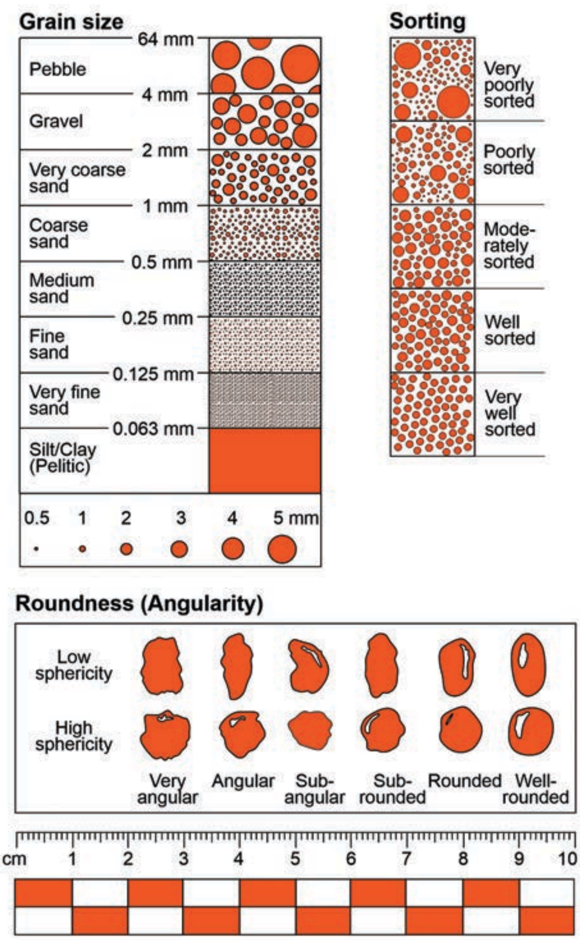


FIGURE 5.12 Textural maturity is a function of grain size, and the degree of sorting and rounding of framework grains. (Modified after Folk, 1951.)

rounded) (see Figure 5.13). Textural maturity largely reflects sediment transport and reworking but it may also be due to diagenetic processes where the clay minerals are formed in pore spaces during burial diagenesis. Nevertheless, textural maturity, reflecting the degree of sediment transport, is based on the premise that higher the energy, (a) the lower the matrix content (clay particles and silt <30 μm; Folk, 1951), and (b) the higher the sorting and degree of rounding (see Figures 5.12 and 5.13).

Folk (1951) linked the degree of maturity of sandstones to sedimentary environments where they are deposited (see Figure 5.13). Thus, as sands are transported, they lose clay content and become more sorted and rounded (Figure 5.13A). Using these three criteria (i.e., loss of clay, sorting, and roundedness; see also Figure 5.12), a sand can be characterized as immature, sub-mature, mature, or super-mature (Figure 5.13B). In a texturally immature sandstone (Figure 5.13A), irrespective of grain sorting and rounding, the proportion of clay-sized material exceeds 5%. Such sandstones also contain high proportions of feldspars and rock fragments and characteristic of alluvial fans, turbidite sands, and fluvial overbank sands (Figure 5.13B) (see Folk, 1951). In texturally sub-mature sandstones (Figure 5.13A), the clay fraction is <5% but contains moderately sorted sand grains; the grains are slightly more rounded than those noted in immature sandstones. The feldspars and lithic fragments are somewhat less than noted in immature sandstones. These texturally sub-mature sandstones are commonly noted in fluvial channels and in turbidites (see Figure 5.13B). In both mature and super-mature sandstones, the clay content is <5% with negligible amounts of rock fragments (possibly chert). In mature sandstones, the grains are subangular to subrounded, whereas in super-mature sandstones, they are well to very well sorted, and rounded to subrounded (Figure 5.13A). Mature sandstones are commonly noted in fluvial channels but also occur on beaches and in eolian dunes; the super-mature sandstones occur in the latter two settings (see Figure 5.13B). A summary of textural and compositional maturity of sandstones is provided in Figure 5.14 (see also Folk, 1951, 1974; Dott, 1964; Garzanti, 2017).

5.2.5.2 Compositional Maturity

Compositional maturity is the relative abundance of stable and unstable framework grains in a sandstone. Quartz-dominated sandstones are compositionally mature, whereas those with unstable minerals (such as feldspars) or unstable rock fragments are compositionally immature. The unstable rock-forming minerals (olivine, pyroxene, and amphibole, and Ca-rich sialic minerals like plagioclase) are more susceptible to alteration due to chemical weathering and erosion than the stable ones or those minerals that undergo weathering at a much slower rate (such as the alkali feldspar, Na-rich plagioclase, and the micas: biotite and muscovite) (see also Figure 5.8). Quartz is the most stable mineral, as it does not dissolve in water, and its hardness makes it resistant to physical erosion (see Figure 5.8). Hence, continued transport concentrates quartz in the sediment, whereas feldspars and other less stable minerals are progressively altered or destroyed by erosion; quartz-rich sediments contain many erosion-resistant minerals like zircon, tourmaline, and rutile.

5.2.6 Some Major Types of Sandstones

Framework mineralogy enables three groupings: quartz arenites, feldspathic arenites, and lithic arenites (these groups include also wackes) (see Figure 5.11). In general, based on 15% matrix content, two groups of sandstones are noted: arenites and wackes (or graywackes) (see Figure 5.11). Arenites are texturally "clean," matrix-free (or matrix-poor) sandstones, whereas wackes are argillaceous, matrix-rich, texturally immature, or "dirty" sandstones (see Figure 5.11). Further, the percentage of sand-sized grains (quartz, feldspar, and rock fragments) allows separation between arenites and wackes (see Figure 5.11). Pettijohn (1957) used the percentage of matrix finer than 0.03 mm as the criterion to separate wacke (matrix-rich sandstones) and arenite (matrix-poor sandstones; typically, with 1% to 2% matrix).

5.2.6.1 Quartz Arenites

Quartz arenites (or quartz sandstone) consists of >95% siliceous grains such as quartz (mainly; with sand-sized monocrystalline grains), chert, and quartzose rock fragments (Figure 5.11); most quartz arenites are texturally mature to super-mature (Figure 5.13). They are generally well lithified and well cemented with silica or carbonate cement; rarely, they are porous and friable. Hence, due to their stability (quartz content; see also Figure 5.8), quartz arenites are common in the geologic record (particularly in the Paleozoic–Mesozoic). They make up about one-third of all sandstones. Quartz arenites typically occur in association with assemblages of rocks deposited in stable continental cratonic environments such as eolian, beach, and shelf environments (see Figure 5.13). Thus, they tend to be interbedded with shallow-water carbonates, and in some cases, with feldspathic sandstones. Cross-bedding is a characteristic feature, and ripple marks are commonly noted. Fossils are present but rarely abundant. In some shallow-marine deposits, trace fossils such as the burrows of the *Skolithos* ichnofacies (i.e., those with simple vertical burrows and J or U-shaped vertical burrows) are locally abundant.

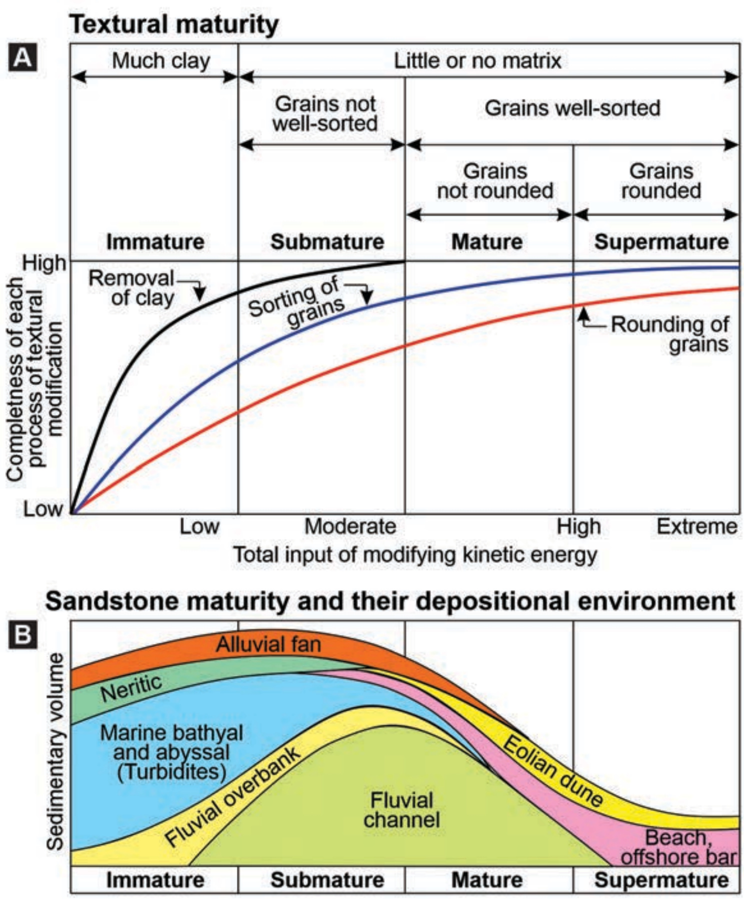


FIGURE 5.13 Textural maturity of sandstone and their depositional environments. (Modified after Folk, 1951.) A: Textural maturity. As sands are transported, they lose clay and become better sorted and better rounded. B: Sandstone maturity and their depositional environments. Using the three criteria (clay content, sorting, and roundness), a sand can be characterized as immature, sub-mature, mature, or super-mature.

5.2.6.2 Feldspathic Arenites

Feldspathic arenites or arkosic arenites or arkose, contain <90% quartz (monocrystalline grains), 10% feldspars (more feldspars than rock fragments; feldspars are mostly orthoclase and microcline rather than plagioclase) and minor amounts of micas and heavy minerals (Figure 5.11). They are medium- to coarse-grained with high percentages of subangular to angular grains. The matrix ranges from negligible amounts to >15%. Sorting varies from moderately well sorted to poorly sorted, thus, they are texturally immature or sub-mature, i.e., wackes (see Figures 5.11 and 5.13). They may be structureless (massive) to parallel laminated or cross-laminated. Fossils are noted in marine deposits. They make up ~15% of all sandstones and are noted in sedimentary successions of all ages, more so in the Mesozoic and Paleozoic. Feldspathic arenites are often associated with conglomerates, shallow-water quartz or lithic arenites, carbonate rocks, or evaporites, and occur in cratonic or stable shelf settings. Less common occurrences include sedimentary successions deposited in unstable basins or other deeper water settings.

5.2.6.3 Lithic Arenites

Lithic arenites, in contrast to feldspathic arenites, have >10% rock fragments, i.e., more rock fragments than feldspars (see Figure 5.11). They also have a high content of unstable rock fragments (such as volcanic and metamorphic clasts) with some stable clasts (such as chert) (Figure 5.11). They are mostly poorly sorted with framework grains (such as quartz) that are poorly rounded with large amounts of secondary matrix. Hence, they are texturally immature to sub-mature (lithic wackes) (see Figure 5.11). Lithic arenite occurs as irregularly bedded, cross-stratified fluvial deposits to evenly bedded, graded, marine turbidites associated largely with fluvial or deeper water marine conglomerates, pelagic shales, and cherts. Lithic arenites include graywackes (wackes; see Figure 5.11), i.e., lithic arenites that are dark gray to dark green in color, well lithified, and with a matrix of secondary chlorite. Lithic arenites and graywackes make up nearly one-half of all sandstones.

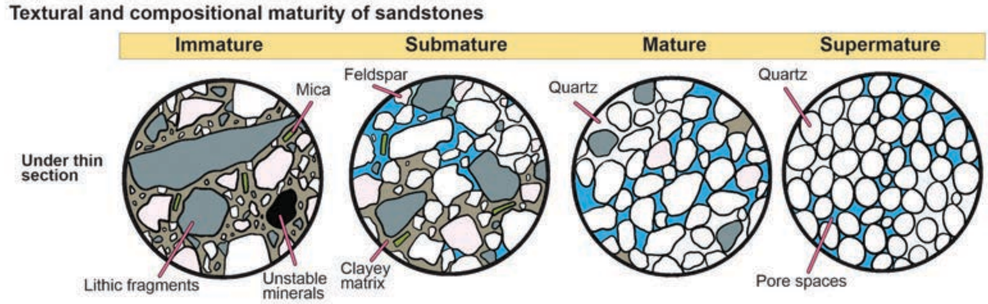
5.2.6.4 Wacke (Graywacke)

Monocrystalline quartz is the most abundant framework component and makes up 25 to 50% of the rock. The feldspar clasts are often angular to subangular. Grains of chert, mudrock, limestone, polycrystalline quartz, and volcanic rocks are commonly noted. Clasts of detrital muscovite, biotite, and chert occur as accessory components. The SiO_2 content varies from 50% to 70%, reflecting the moderate amount of quartz and feldspar. The wackes have 15% or more matrix, hence they are immature; the matrix is rich in clay minerals and chlorite, so, they also have high Al_2O_3, MgO, and FeO + Fe_2O_3 content. Some wackes have a sand framework which is also poorly sorted and angular or subangular; many consist of well-rounded, well-sorted sands. Many wackes were deposited by waning turbidity currents. Graded bedding, sole markings, and the systematic upward changes in sedimentary structures and grain size (characteristic of turbidites) also characterize wackes. The wacke sandstone sequences also have deep-water abyssal and bathyal body fossils, pelagic fauna and flora, and re-transported shallow-water organic remains.

5.2.7 Provenance of Sandstones

Provenance is to trace the origin of sedimentary rocks (see Figures 5.14 and 5.15) (see also Folk, 1951, 1974; Dott, 1964; Pettijohn et al., 1987; Garzanti, 2017). The provenance of sandstone enables us to decipher the history of a sedimentary basin, and to infer the characteristics of source areas from its compositional and textural properties (Pettijohn et al., 1987).

Dickinson (1985) modified Dott's (1964) classification and proposed a model using ternary plots of quartz (Qt and Qm; polycrystalline, monocrystalline quartz, respectively), feldspar (F; K-feldspar, plagioclase) and lithic fragment/rock fragment (volcanic/metavolcanic rock fragments,



	Immature	Submature	Mature	Supermature
Interstiatil material	abundant clayey matrix; minor pore space/cement	moderate clayey matrix; variable pore space/ cement	pore space/cement; minor clayey matrix	pore space/cement; almost no clay
Grain composition	abundant lithics, feldspar, micas, unstable minerals	diverse range of compositions	mostly quartz with minor feldspar, lithics and /or mica	almost entirely quartz +/- heavy minerals
Sorting and Rounding	poorly sorted; angular	moderately sorted; sunangular/ subrounded	well sorted; subrounded	very well sorted; well rounded
Provanance, Processes	derived from an arc, crystalline basement, or orogenic flysch/molasse; minor weathering & transport	diverse source areas and processes		derived from a stable craton or recycled orogen; intense weathering, prolonged transport
		winnowing	sorting	rounding
Common environ-ments	alluvial fans, turbidites, overbank	proximal rivers, overbank, turbidites,	rivers, some beaches, eolian dunes	beaches, eolian dunes

FIGURE 5.14 Summary of textural and compositional maturity of sandstones. (Modified after Folk, 1951, 1974; Dott, 1964; Garzanti, 2017.)

sedimentary/metasedimentary rock fragments) (L) end members to display sediment and source rock characteristics i.e., the source area of the sandstone framework grains (Figure 5.15A) (see also Dickinson and Suczek, 1979; Dickinson et al., 1988).

The main provenance types for the sandstones are: continental block (stable craton, basement uplift), recycled orogens and magmatic arc (Dickinson and Suczek, 1979; Dickinson, 1985; Boggs, 2010) (Figures 5.15B and 5.16). The Qt-F-L plot (Figure 5.15A) uses sediment maturity where increased amounts of quartz are considered more mature; the less stable components like feldspar and lithics are removed by chemical and mechanical weathering during sediment transport. The Qm-F-L plot (Figure 5.15B) uses source rock as a provenance parameter where lithics (detrital mineral composition) are used as indicators of reworked orogenic provinces (Figure 5.15B).

The erosion of sedimentary and volcanic rocks produces increased proportions of rock fragments. The stable continental craton-derived sandstones are sourced from low-lying granite and gneissic rocks and recycled platform sediments (Dickinson and Suczek, 1979; Dickinson, 1985) (see Figure 5.16A). The sand from continental block accumulates either in continental interior or passive continental margin or continental flanks of foreland basin (Dickinson, 1985). Recycled orogenic provenance occurs as a consequence of plate collision; in other words, the collision of the major plates results in the upliftment of orogenic belts, thus creating source areas for adjacent depression ones (Dickinson and Suczek, 1979; Dickinson, 1985; Boggs, 2010) (Figure 5.16B). These uplifted sedimentary and metamorphic rocks constitute the main type of source, thus, the framework of the accumulated sands have high amount of sedimentary and metasedimentary rock fragment, moderate quartz content, and a high quartz to feldspar ratio (see Boggs, 2010). Magmatic arc provenance constitutes the source for the sands in the zones of plate convergence (Figure 5.16C). The volcanoclastic rocks are rich in volcanic lithic material and plagioclase feldspar; these were erupted and eroded, from stratovolcanos shed (undissected) in forearc, and backarc, can even reach up to

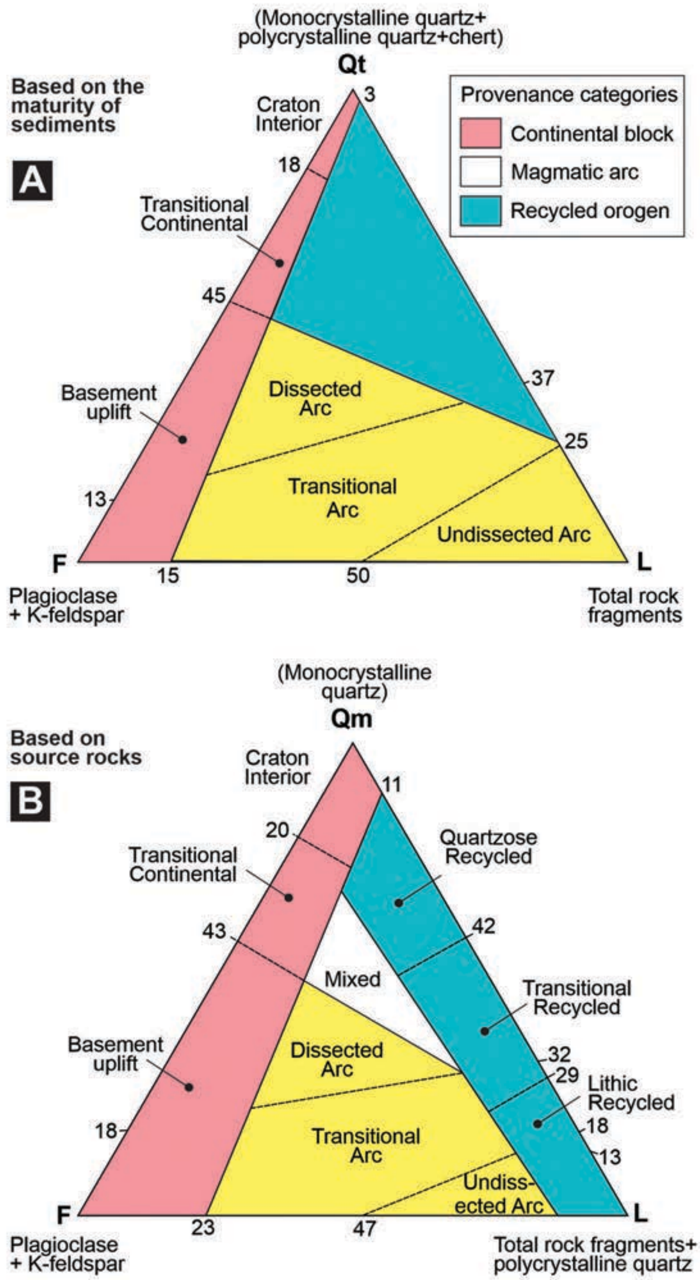


FIGURE 5.15 Ternary plots of sediment provenance. (Modified after Dickinson et al., 1983; see also Dickinson and Suczek, 1979; Dickinson, 1985.) A: Qm-F-L provenance diagram (provenance based on the maturity of sediments). This diagram is used to interpret the ingredients of sandstone in terms of the plate-tectonic setting of the rocks that produced the sand. Q is quartz, F is feldspar, and L is lithics (rock fragments). This diagram is best used for sediment that does not have many quartz grains. B: Qm-F-L provenance diagram (provenance based on the source rocks). This diagram is best used when the sandstones contain a lot of chert or polycrystalline quartz grains. Qm is monocrystalline quartz, F is feldspar, and Lt is total lithics. By assigning lithic quartz to the lithics category, it makes it easier to differentiate sediments coming from the recycled rocks of mountain ranges.

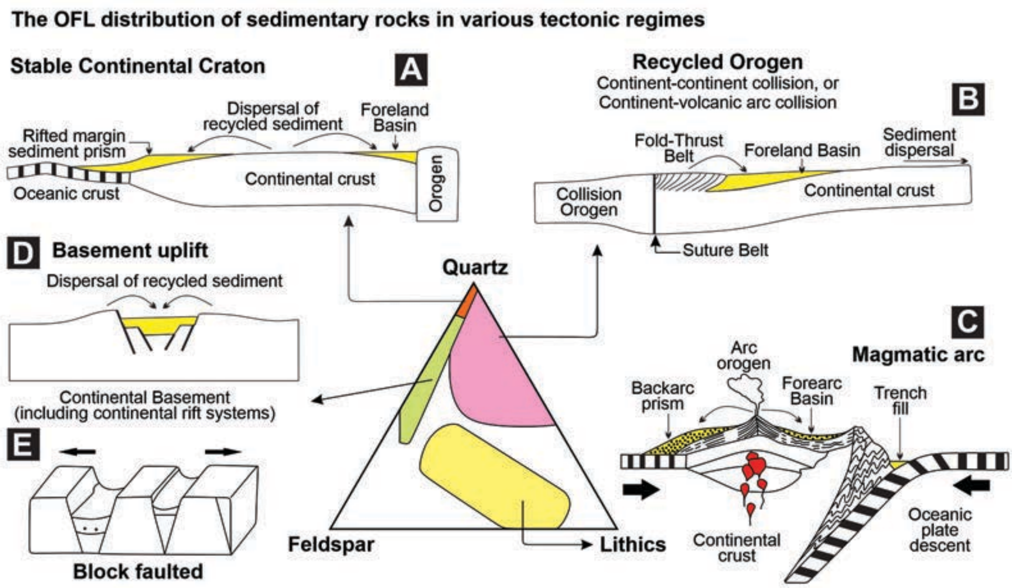


FIGURE 5.16 The QFL distribution of sedimentary rocks in various tectonic settings (Modified after Dickinson and Suczek, 1979; Dickinson, 1985.) A: Stable continental craton. B: Recycled orogen. C: Magmatic arc. D–E: Basement uplift.

the foreland basin (Dickinson and Suczek, 1979; Disckinson, 1985; Boggs, 2010). Basement uplift in correspondence to faulting forms horst-graben systems noted in continent interiors and pull apart basins (Figures 5.16D–E). Sands eroded from these uplifted areas are generally rich in feldspar (arkose) due to the absence of long transportation (Dickinson and Suczek, 1979; Dickinson, 1985). The above mentioned plots are highly simplified but are a useful means of documenting systematic changes as uplift and erosion exposes deeper crustal rocks.

5.3 CONGLOMERATES

Conglomerates make up 1–2% by weight of the total sedimentary rock mass (Garrels and McKenzie, 1971). Conglomerate is often used as a general term for sedimentary rocks that contain a large fraction (~30%) of gravel-sized (>2 mm) clasts with a matrix of finer-grained sediments of sand, silt, or clay, that fill the interstices between the clasts (Figure 5.17). Thus, a sedimentary rock consisting largely of well-rounded to subrounded lithified gravel is a conglomerate (see Table 5.5; Figure 5.17). Grain roundness (or angularity) is measured using grain outlines (see Figure 5.12). These very coarse clastic rocks (also called *rudites* in Latin) or rudaceous sedimentary rocks or *psephites* (in Greek) are also useful proxies for inferring provenance, depositional environments, paleogeography, and tectonic settings.

Conglomerates, within the sedimentary rock record are important for two reasons: (a) they are mostly associated with high energy systems, and (b) they are good proxies for inferring provenance of the hinterland, as they consist of sediments that generally have not been transported far, and are often linked to nearby tectonic activity. In terms of origin and depositional mechanisms, they contain similar sedimentary structures (such as tabular and trough cross-bedding, graded bedding, among others) are thus, closely related to sandstones. However, it must be mentioned here, that for now, conglomerates and breccias are treated together as “conglomerates.” But if the major clasts are angular, the rock is called a breccia. This fact is important when considering its origin, as breccias

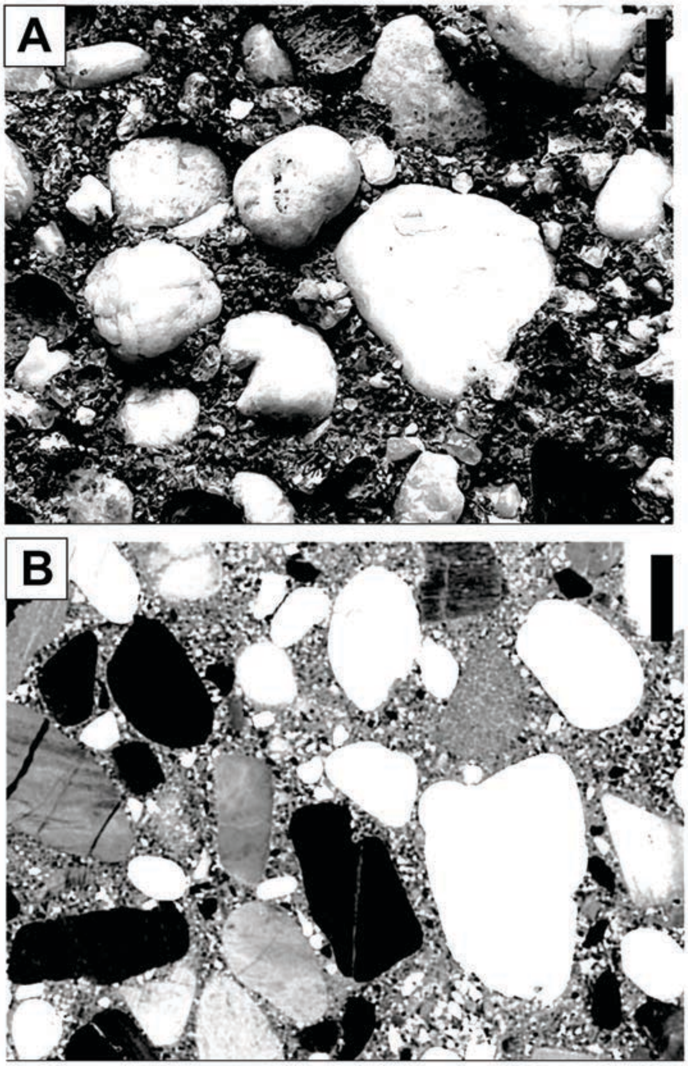


FIGURE 5.17 Conglomerate. A: Showing the assortment of various rounded clasts of different sizes. B. Rounded clasts in thin section. Bar in B is 10 mm.

TABLE 5.5
Classification of conglomerates and diamictites based on clast stability and fabric support

	Type of fabric support	
Percentage of ultrastable clasts	**Clast-supported**	**Matrix-supported**
>90	Quartzose conglomerates	Quartzose diamictite
<90	Petromictic conglomerates	Petromictic diamictite

are not necessarily sedimentary. But in general, lithified gravel and rubble are called conglomerate and breccia, respectively.

5.3.1 Composition of Conglomerates

Conglomerates largely contain gravel-sized quartz grains and clasts (framework grains) such as rock fragments (clasts of K- and plagioclase feldspars, and micas: muscovite and biotite), and such heavy minerals (olivine, pyroxene, amphibole, zircon, magnetite, and hematite). The conglomerate matrix is made up of sand- or mud-sized grains (see Figure 5.17). Rare conglomerates are also noted that only have stable clasts such as quartzite, chert, or vein quartz, among others. Additionally, based on source rocks and depositional conditions, a conglomerate may also contain rock fragments of igneous, metamorphic, or sedimentary rocks.

5.3.1.1 Classification of Conglomerates

Conglomerates can be classified based on texture (i.e., the amount and chemical composition of the matrix) as orthoconglomerates (meaning "true") and paraconglomerates (Pettijohn, 1957) (see Figure 5.18). Rock in which the clasts (i.e., gravel-sized framework grains) touch each other (i.e., very little matrix; often <15%) are called orthoconglomerates, whereas if the clasts do not touch each other (lots of matrix), the rock is called a paraconglomerate (see Figure 5.18A). Hence, orthoconglomerates possess a grain/clast-supported framework (Figure 5.18A). However, depending upon the spaces between clasts filled by matrix or fine sediments, they can be open-framework (no matrix between clasts) or closed-framework (Figure 5.18A). The paraconglomerates have matrix between 15–50% of sand and finer clasts (see Figure 5.18A).

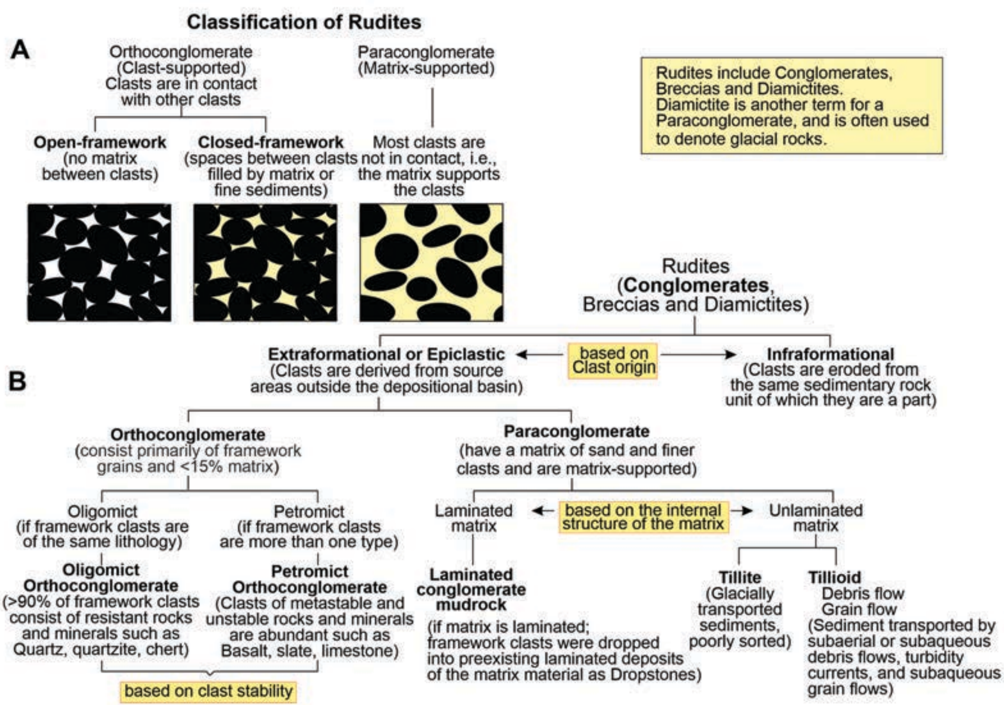


FIGURE 5.18 Classification of conglomerates. (Modified after Pettijohn, 1957.)

Pettijohn (1957) proposed a classification (later modified by Boggs, 1992) based on visible textural and compositional features (Figure 5.18B). Initially, based on the origin of clasts, two types of conglomerates are noted: extraformational (i.e., derived from rocks located outside the depositional basin and thus the framework material is exotic) and intraformational (i.e., derived from rocks located within the depositional basin) conglomerates (Figure 5.18B). Thus, the intraformational conglomerates have framework grains and matrix of the same composition, i.e., the limestone mud clasts float in a similar limestone mud matrix.

The extraformational conglomerates (or epiclastic conglomerates) are subdivided into two: orthoconglomerates and paraconglomerates based on percentage of matrix (Figure 5.18B). Orthoconglomerates are matrix-poor (i.e., consisting primarily of framework grains and <15% matrix) with a grain-supported fabric (i.e., the framework grains contact and support one another), whereas paraconglomerates are matrix-rich with a fabric supported by matrix grains. The matrix-supported conglomerates are also called diamictites (cf. Boggs, 1992). However, the term diamictite is often used for poorly sorted glacial deposits, a term applied to non-sorted or poorly sorted siliclastic sedimentary rocks that contain larger particles of any size within a muddy matrix (see Figure 5.18B).

On the basis of framework grain composition (clast stability), the conglomerates are further divided into oligomict and petromict types. In oligomict (orthoquartzose) conglomerate (also referred to as oligomict orthoconglomerate), the framework clasts (such as granule and coarser grains) are >90% and consists of resistant rocks and minerals such as metaquartzite, vein quartz, and chert, whereas petromict (or polymict orthoconglomerate) contain clasts of metastable and unstable rocks such as basalt, slate, and limestone, are abundant. In oligomict orthoconglomerates, the framework clasts are of the same lithology, whereas in petromict orthoconglomerates, they display more than one type (containing a mixture of largely unstable or metastable clasts such as basalt, limestone, shale, and metamorphic phyllite). The petromict orthoconglomerates are much more common and abundant than the oligomict orthoconglomerates (Figure 5.18B).

The paraconglomerates, based on the internal structure of the matrix, are of sand and finer clasts and are matrix-supported. They are further divided into laminated and unlaminated matrix types (see Figure 5.18B). The paraconglomerates, with a matrix of finely laminated mudrock and where the coarser framework grains float, are called laminated conglomerate (pebbly or cobbly, or bouldery) mudrock (Figure 5.18B). A good example of this type are dropstones, where the post-depositional compaction of the plastic matrix around the more resistant clasts produces a drape-like pattern of overlying laminations. In general, laminations imply that the framework clasts were dropped into pre-existing laminated deposits of the matrix material as dropstones. Paraconglomerates in which the matrix is disorganized and unlaminated are either tillite (of glacial origin; the sediment deposited by melting glaciers is called a till; poorly sorted), or tilloid (if deposited by mass movement, i.e., transported by subaerial or subaqueous debris flows, turbidity currents, and subaqueous grain flows) (Figure 5.18B).

5.3.1.2 Types of Conglomerates

5.3.1.2.1 Quartzose (Oligomictic) Conglomerates

These conglomerates are derived from metasedimentary rocks containing quartzite, igneous rocks (containing quartz-filled veins), and sedimentary successions, particularly limestones with chert beds. The metasedimentary rocks are a type of metamorphic rock that was formed first through the process of deposition and solidification of sediments and then underwent high pressures and temperatures, causing the sediments to recrystallize. In this, the less stable rock types were removed by weathering, erosion, and sediment transport, resulting in the concentration of stable clasts. The quartzose clasts occur as thin, pebbly layers or lenses of pebbles within dominantly sandstone units. Hence, the quartzose conglomerates are either clast-supported or matrix-supported. They are common in the geologic record ranging from the Precambrian to the Tertiary. Most quartzose conglomerates are of fluvial origin, being deposited mainly in braided streams. Some marine, wave-worked quartzose conglomerates were deposited in the littoral (beach) environment.

5.3.1.2.2 Petromictic Conglomerates

These conglomerates consist largely of metastable clasts consisting of or derived from plutonic igneous (such as granite, granodiorite), volcanic (andesite, basalt), metamorphic (schist, gneiss, quartzite), and sedimentary (limestones) rocks. The clasts within a conglomerate may be dominated by one or another type of clasts, hence, may be called a limestone conglomerate or a basalt conglomerate or a schist conglomerate. Conglomerates dominated by plutonic clasts are rare, probably as granites tend to disintegrate into sand-sized fragments rather than remain as larger blocks. In general, in the geological record, the volume of petromict conglomerates is greater than that of quartzose conglomerates; their thicknesses may reach thousands of meters. The petromict conglomerates are often transported by fluid-flow and sediment-gravity flow mechanisms and deposited in fluvial to shallow and deep marine environments.

5.3.1.2.3 Intraformational Conglomerates

These are composed of within depositional basin clasts as opposed to outside the basin clasts for extraformational conglomerates. Intraformational conglomerates originate by penecontemporaneous deformation of semi-consolidated sediment and re-deposition of the fragments fairly close to the site of deformation. The clasts are formed by penecontemporaneous breakup of sediment, ether subaerially, such as by drying out of mud on a tidal flat, or under water. Tidal currents, storm waves, or sediment-gravity flows cause subaqueous rip-ups of semi-consolidated muds. Shale rip-up clasts embedded in the basal part of sandstone units are very common reflecting deposition by sediment-gravity flows. The most common types of clasts (mostly angular or slightly rounded ones, suggesting little transport) noted in intraformational conglomerates are those of siliciclastic mud clasts and lime clasts. In some, flattened clasts are stacked on edge forming bed, caused by strong wave or current action; these are called edgewise conglomerates (Pettijohn, 1975). In general, the intraformational conglomerates, less abundant than the extraformational ones, are often laterally extensive forming thin beds (a few centimeters to a meter); they occur in all geological ages.

5.3.1.2.4 Edgewise Conglomerates

Edgewise conglomerates, also called stone rosettes, form on modern wave-washed beaches where there is a plentiful supply of platy rock fragments or shells (generally of bivalves). They form pavements where clasts are organized in crude radial or aligned patterns (also called edgewise conglomerate).

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6 Carbonate Rocks

6.1 LIMESTONES

6.1.1 Introduction

Carbonate rocks (mostly limestones and dolomites) make up 20–25% of all sedimentary rocks and occur from the Precambrian to the Quaternary (Figure 6.1A). Dolomite dominates in Precambrian and Paleozoic sequences, whereas limestone in Mesozoic–Cenozoic ones (Ronov, 1964) (Figure 6.1A). The carbonate rocks form chemically and biochemically where the dissolved ions are carried in solution from the source to sink (their depositional site) and precipitates to form solid minerals (rocks). The carbonate rocks are important as they (a) are repositories of past life forms (fossil record) and provide information about the origin and evolution of the earth; (b) contain fossils (many are made of exclusively of), hence, are best repositories of organic evolution; (c) contain structures and textures that are robust proxies for inferring paleoenvironmental conditions; (d) are of considerable economic significance, being used in a variety of agricultural (to neutralize soil acid) and industrial purposes (as a base to neutralize acids, and in the metallurgy industry as a flux for smelting iron and steel); (e) are used as construction material; (f) serve as reservoir rocks for >25% of world's petroleum reserves, 50% of the world's petroleum occur in carbonate reservoir rocks, and ~80% of the hydrocarbons in North America; and (g) are hosts to ore deposits due to their high porosity thus making them excellent host rocks for ore-bearing solutions such as the epigenetic deposits of lead and zinc.

6.1.2 Mineral Constituents

Carbonates are primarily composed of calcite ($CaCO_3$), aragonite ($CaCO_3$) and dolomite ($CaMg(CO_3)^2$). Although there are more than 60 carbonate minerals, only three are abundant in the earth's surface layer: calcite, aragonite, and dolomite; the first two are polymorphs of $CaCO_3$. Polymorphs are different crystal forms of the same compound, differing only because of their molecular arrangement within the unit cells of the crystalline lattice of each crystal form (i.e., an element that can crystallize into more than one crystal structure).

6.1.2.1 Calcite (Low- and High-Magnesian)

Two types of calcite are known: (a) low-magnesian calcite (or simply called calcite) that contain <4% mol $MgCO_3$ (such as the tests/shells of planktonic foraminifers, coccoliths, and brachiopods, etc.), and (b) high-magnesian calcite (still in calcite crystal structure) that contain >4% mol $MgCO_3$ (such as shells of echinoids, crinoids, benthonic foraminifers, red algae, etc.) (Figure 6.1B). It must be noted that the polymorphs of $CaCO_3$ have different susceptibilities, where normal calcite (i.e.,

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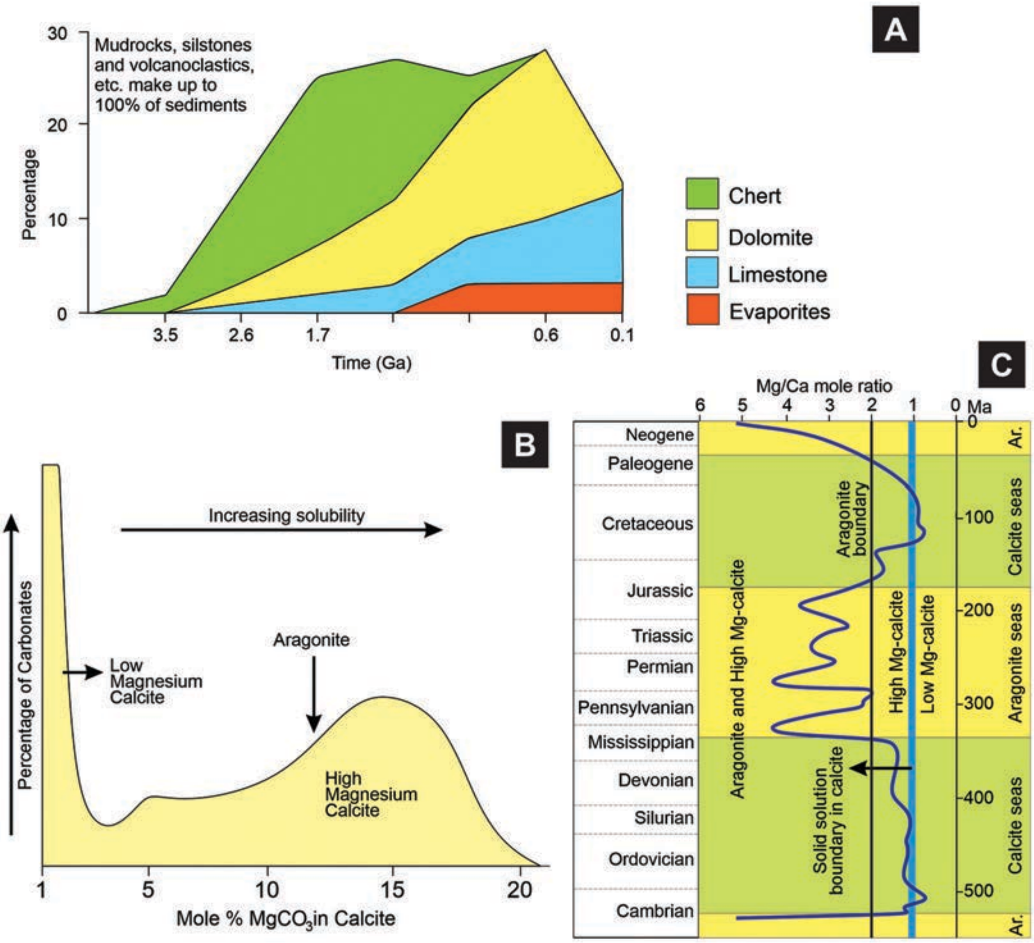

FIGURE 6.1 The relative abundance of non-siliciclastic sediments through time. A: Note that dolomites are more abundant than limestones during much of the Proterozoic, when microbial ecosystems dominated the biosphere. (Modified after Ronov, 1964; Eriksson et al., 2012.) B: Carbonate mineral relative solubility. The figure shows the relative solubility of calcite in freshwater with increasing Mg content. Note that calcite solubility is governed by its Mg content, i.e., the greater the amount of Mg entry in calcite structure the greater the distortion in lattice structure, which increasingly causes the calcite structure to be more structurally unstable. Thus, the most stable carbonate mineral phase is the low-Mg calcite. C: Changes in seawater composition through time. (Modified after Stanley et al., 2002.) The calcite sea is characterized by low-magnesium calcite as the primary inorganic marine $CaCO_3$ precipitate, whereas in an aragonite sea, aragonite and high-magnesium calcite are the primary ones (see also Palmer and Wilson, 2004).

low-magnesian calcite) is relatively stable, whereas aragonite and high-Mg calcite are metastable, i.e., at some point of time they will inevitably recrystallize to blocky calcite (Figure 6.1B).

6.1.2.2 Aragonite

Aragonite is a metastable polymorph (a polymorph has the same chemical composition but different crystal structure) of $CaCO_3$. Under aqueous conditions, aragonite gets rapidly converted to calcite (i.e., into low-magnesian calcite), hence, carbonates older than the Cretaceous age contain little aragonite (Figure 6.1B). Modern oceans have aragonite (mollusks, calcareous green algae, corals, etc.) and, to a lesser extent, high-magnesian calcite. In general, the production of skeletal and non-skeletal limestones during the Phanerozoic has fluctuated reflecting major changes in plate tectonics (Stanley et al., 2002) (Figure 6.1B). Such cyclicity was first recognized by Mackenzie and Morse (1992) who noted a weak correlation between high sea levels and high accretion rates (i.e., addition of new crust at mid-ocean spreading ridges). High rates resulted in the lowering of Mg/Ca ratios as Mg is removed from seawater and Ca is added during the hydrothermal alteration of hot new crust. The opposite, high Mg/Ca ratios, are more likely during low sea floor accretion rates (and lower sea levels), thus, promoting the precipitation of high-magnesian calcite and aragonite cements. Thus, first-order low sea level cycles coupled with high Mg/Ca ratios tend to favor aragonite seas and high sea levels with low Mg/Ca ratios, calcite seas (Figure 6.1B). The calcite sea is characterized by low-magnesium calcite as the primary inorganic marine $CaCO_3$ precipitate, whereas in an aragonite sea, aragonite and high-magnesium calcite are the primary ones (Figure 6.1C) (see also Stanley et al., 2002; Palmer and Wilson, 2004).

6.1.2.3 Dolomite (Stoichiometric Dolomite)

Dolomite (or stoichiometric dolomite) occurs in a few restricted modern environments, in certain supratidal settings and freshwater lakes, and is much less abundant in modern carbonate environments than aragonite and calcite.

6.1.3 Sediment Constituents

The mineralogy of carbonate rocks consists of various kinds of sand- and silt-size carbonate grains and various amounts of fine lime mud matrix and carbonate cements. Although limestones commonly contain only one or two dominant minerals, several kinds of carbonate grains are noted.

6.1.3.1 Carbonate Grains (Allochems)

Most carbonate grains are composite grains that are made up of a number of small calcite or aragonite crystals. These carbonate aggregates that make up the bulk of many limestones are called allochem (Folk, 1962) and include both non-skeletal (lithoclasts, ooids) and skeletal (fossil and fossil fragments) (see also Flügel, 2010). These grains typically range from coarse silt (0.02 mm) to sand (up to 2 mm), but larger particles such as fossil shells do also occur. Based on changes in shape, internal structure, and mode of origin, five major types of grains are noted: carbonate clasts, skeletal particles, ooids, peloids, and aggregate grains.

6.1.3.1.1 Carbonate Clasts (Lithoclasts)

Carbonate clasts are rock fragments derived either by the erosion of ancient limestones exposed on land or by the erosion of partially or completely lithified carbonate sediments within a depositional basin. Lithoclasts range in size from very fine sand to gravel, although sand-size fragments are most common. They generally show some degree of rounding, indicative of transport, but subangular or even angular clasts are not unusual. Lithoclasts have little environmental significance but yield information about the source area (provenance). In some cases, very small lithoclasts may be confused with large peloids. Two kinds of lithoclasts are recognized on the basis of their origin: intraclasts and extraclasts.

Intraclasts originate from within a depositional basin by the fragmentation of penecontemporaneous, and often from weakly cemented, carbonate sediments. Small pieces of this sediment are eroded from the seafloor and redeposited at or near the original area of deposition. Normal waves, storm waves, or currents cause the physical disruption of penecontemporaneous sediments, thus producing most intraclasts. But some intraclasts may also be produced by the organic activity on the surface of sediments, such as by burrowing or boring activity within sediments, and by the local sliding of weakly consolidated sediments (Flügel, 2010). Thus, intraclasts provide information about the state of the depositional basin, such as the presence of bottom currents and the consolidation state of the sea bottom.

Extraclasts, on the other hand, are lithoclasts generated by the erosion of much older, lithified carbonate rocks exposed on land (outside the depositional basin in which the clasts accumulate). They are simply carbonate rock fragments. It is often difficult to distinguish between extraclasts and intraclasts. If the lithoclasts lack recognizable fossils or show no evidence of weathering or lack recrystallized veins, it may be impossible to differentiate between extraclasts and intraclasts.

6.1.3.1.1 Skeletal Fragments / Skeletal Grains (Bioclasts)

In limestones, skeletal fragments occur as micro- or macrofossils or as broken fragments of larger fossils that sometimes make up most of the rock. The type of skeletal component depends on the age of the rock and the prevailing paleoenvironmental conditions under which the rock was deposited. Hence (generally speaking), in early Paleozoic rocks, trilobite skeletal remains dominate, whereas in Cenozoic rocks, abundant foraminifers dominate. Similarly, limestones formed in different paleoenvironments are characterized by different kinds of skeletal fragments. In limestones deposited in shallow waters (in high-energy environments where the water is agitated and well-oxygenated), fragments of colonial corals dominate; these build rigid, wave-resistant skeletal structures. However, in limestones deposited under quiet-water conditions, fragile forms such as the branching types of bryozoans which cannot withstand the rigors of high wave energy environments, dominate. Additionally, depending upon paleoenvironmental conditions, the skeletal remains within a given specimen of limestone may consist entirely or almost entirely of one species; however, mostly, several species are noted. Hence, fossils components provide important information about environmental conditions such as water depth, salinity, turbidity, and energy levels. Most skeletal grains are composed of aragonite, calcite or magnesian calcite (Flügel, 2004, 2010). Vertebrate remains, fish scales, conodonts, and the remains of a few invertebrate organisms such as inarticulate brachiopods are composed of calcium phosphate; diatoms and radiolarians, are composed of silica. It must also be kept in mind that the original composition of skeletal grains may be altered during diagenesis. Aragonite skeletons transform to calcite, whereas the high-magnesian calcite grains are altered to calcite or become dolomitized. Bioclasts may also undergo replacement by silica.

6.1.3.1.2 Non-Skeletal Fragments

***6.1.3.1.2.1* Peloids** Peloids are carbonate grains composed of microcrystalline or cryptocrystalline calcite or aragonite with no distinctive internal structures (Figure 6.2A). They are generally silt to fine-sand sized (0.03–0.1 mm; 30–100 μm) and rarely larger than that; they are smaller than ooids (discussed below). The most common type of peloids are fecal pellets, produced by organisms that ingest calcium carbonate-mud and extrude undigested mud as pellets; those that are well-rounded, symmetrical-shaped, are thought to be of fecal origin (fecal pellet) (Figure 6.2A). The fecal pellets are small, uniform in size, and oval to round in shape (Figure 6.2A). In ancient carbonate rocks, most fecal pellets are small, and are composed mainly of fine micrite with crystals ranging from 2 to 5 microns, but the pellets of some modern organisms may exceed 5 mm. The fecal pellets are opaque or dark colored due to the presence of fine organic matter, and may or may not have a thin, dark outer rim (Figure 6.2A).

The fecal pellets are produced by a variety of organisms such as crustaceans, holothurians, brachiopods, amphineurans, pelecypods, gastropods, ostracods, copepods, decepods, echinoderms, tunicates, worms, and even fish. These organisms ingest fine carbonate mud while feeding on organic-rich sediments. The pellets are shaped in their guts and then extruded out as pellets. Each kind of organism tends to produce pellets of a distinctive shape and size. The pellets can be differentiated from ooids by their lack of concentric or radial internal structure and from rounded intraclasts by their uniformity of shape, good sorting, and small size (see also Figure 6.2B). The peloids may also

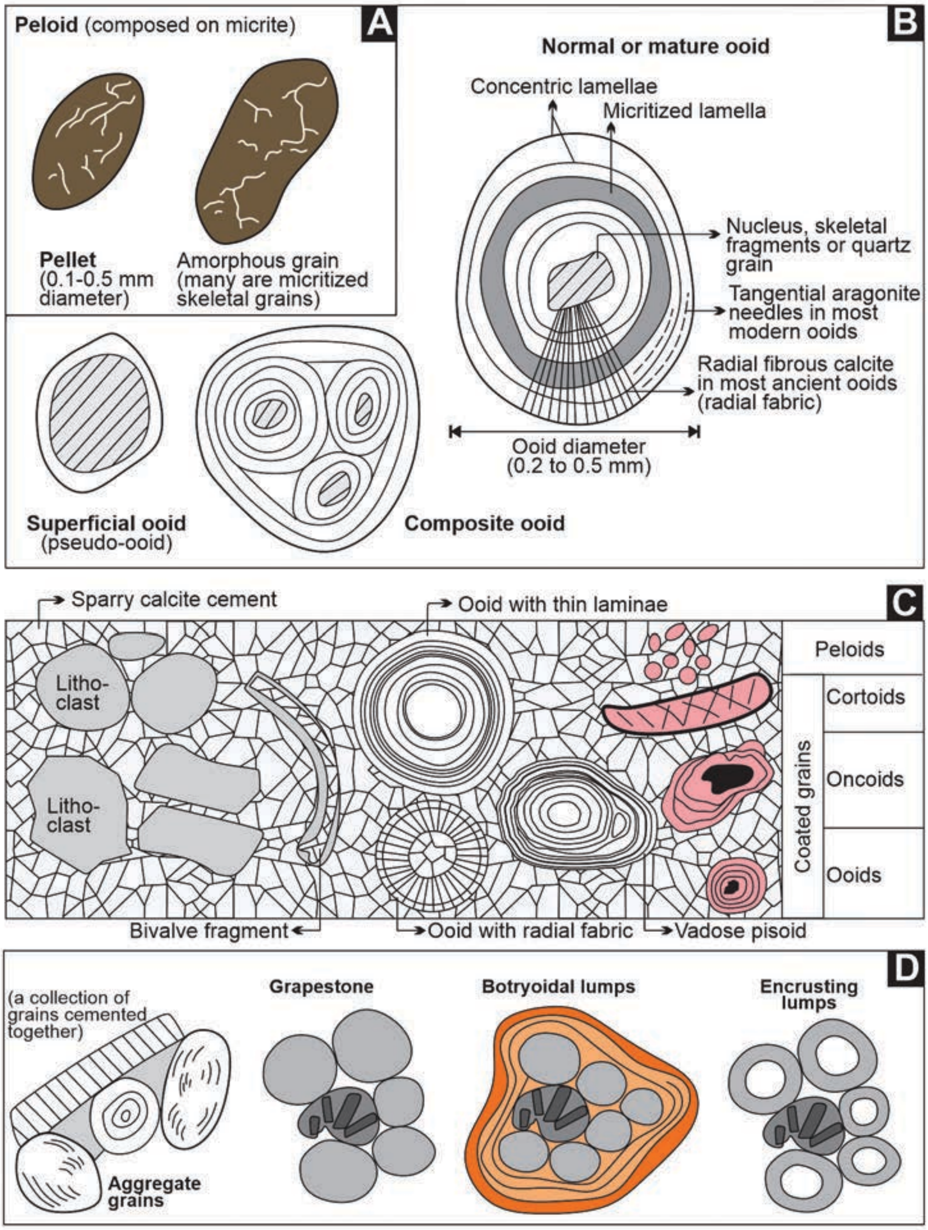

FIGURE 6.2 Types of coated grains. A: Peloids. B: Types of ooids (normal, superficial, and composite). C: Coated grains (ooids, oncoids, cortoids) and peloids. D: Aggregate grains (grapestone, botryoidal lumps, and encrusting lumps).

be produced by other processes, such as micritization of small ooids or rounded skeletal fragments caused by the boring activities of certain organisms, particularly endolithic (boring) algae. These boring activities convert the original grains into a nearly uniform, homogeneous mass of microcrystalline calcite. Although most fecal pellets are structureless, some do have a sieve-like internal structure, cylindrical holes parallel to the long dimension of the pellet. These holes may be filled with sparry cement, giving the pellets a speckled appearance in thin sections. Other pellets may have lamellae-like or ring-like internal structures. Most pellets appear to be produced by organisms living in quiet, marine waters with muddy bottoms. The pellets in ancient limestones occur most commonly in muddy limestones (micrites), and their presence is suggestive deposition in low-energy environments.

6.1.3.1.2.2 Coated Grains: Ooids, Oncoids, and Cortoids Coated grain is a general term used for all carbonate grains composed of a nucleus that is surrounded by an enclosing layer or layers called cortex (Peryt, 1983). On the basis of the structure of the cortex, various types of coated grains are noted such as ooids, oncoids, cortoids, and pisoids (see Figure 6.2C).

6.1.3.1.2.2.1 Ooids Ooids are coated, small, more or less spherical to oval carbonate grains that contain a nucleus, generally of a shell fragment, pellet, or a quartz grain, surrounded by one or more thin layers or coatings (the cortex) consisting of fine calcite or aragonite crystals (see Figure 6.2B–C). These coated grains are sometimes referred to also as ooliths, but ooid is the preferred term. Carbonate rocks formed mainly of ooids are called oolites. Most ooids are sand- to silt-sized particles and range in size from about 0.1 mm to more than 2 mm, with 0.5 mm to 1 mm being the most common size. Ooids are white to cream in color and commonly have a pearly luster if formed in agitated waters; the quiet-water ooids often have a dull luster. The spherical to subspherical ooids with several concentric layers whose total thickness is greater than that of the nucleus, are called normal or mature ooids (see Figure 6.2B). However, in some ooids, the coating only consists of one or two very thin layers, which have a total thickness less than that of the nucleus; these are called superficial ooids or pseudo-ooids (see Figure 6.2B) (see also Illing, 1954). Coated grains that have an internal structure similar to that of ooids but that are much larger (>2 mm) are called pisoids (a rock composed of pisoids is a pisolite). Pisoids are generally less spherical than ooids and are commonly crenulated. Some pisoids may be of algal origin, formed by the trapping and binding activities of blue-green algae (cyanobacteria) in the same way that stromatolites are formed. Spheroidal stromatolites that reach a size exceeding 1 to 2 cm are called oncoids (see Figure 6.2C). In general, most ooids display an internal structure of concentric layers, but some ooids show a radial internal structure, as well (see Figure 6.2B). Radial ooids that also display concentric layers (see Figure 6.2C) that were probably formed by the recrystallization of normal ooids; however, radial ooids may also be formed by primary sedimentation processes. Ooids are also formed by accretionary processes where $CaCO_3$ is precipitated onto the surface of a nucleus in waters saturated to supersaturated with calcium bicarbonate. Most modern-Holocene marine ooids are composed of aragonite, whereas ancient ooids are composed principally of calcite, owing largely due to the diagenetic alteration of aragonite or Mg-calcite see Figure 6.1B). Water agitation is important to the growth of ooids (Figures 6.3A–B). However, some ooids can also form under quiet-water conditions; these ooids are more likely have a radial structure. In general, ooids form where strong bottom currents and agitated-water conditions exist and where the saturation levels of calcium bicarbonate are high (Figure 6.3).

6.1.3.1.2.2.2 Oncoids Oncoids are coated grains that are more irregular in shape and laminae, and are much larger (from <2 mm to >10 mm) than ooids being formed in both non-marine and marine environments (see Peryt, 1983; Flügel, 2004) (see Figures 6.2C and 6.3B). The laminae (often wavy or crinkly) are composed mainly of fine micrite but that may contain silt- or sand-size non-carbonate detrital grains formed by the activities of organisms such as cyanobacteria (see Figure 6.2C). Many authors restrict the term oncoid to grains of cyanobacterial and bacterial origin (Peryt, 1983), while other also include carbonate grains encrusted by red algae and bryozoans (Richter, 1983). Also, irregularly layered, coated grains without organic structures (such as spongiostromate oncoids) and nodules in vadose environments (so-called vadoids in pedocretes) (Figure 6.2B) are considered by

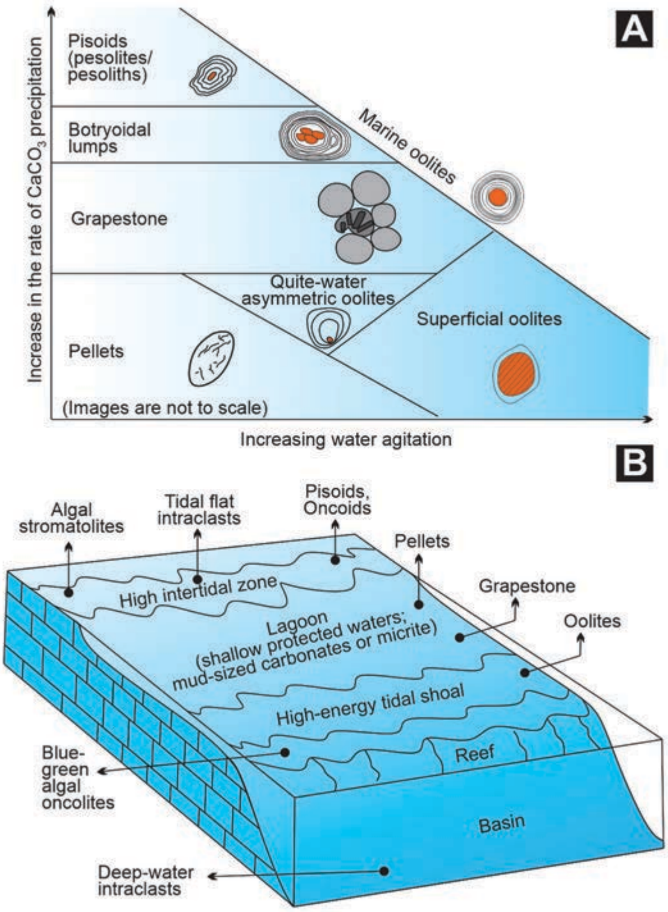

FIGURE 6.3 Spatial distribution of coated grains. A: Formation of pellets, peloids, and coated grains versus water agitation. Water agitation is important to the growth of coated grains. B: Depositional setting of various coated grains.

some workers to be oncoids. Most oncoids that form through the activities of encrusting organisms have been considered as a type of stromatolite (see also Logan et al., 1964).

6.1.3.1.2.2.3 Cortoids Some coated grains consist of fossils, ooids, or peloids coated with a thin envelope of generally dark colored micrite, commonly consisting of crystals ranging in size from 2 to 5 microns; these grains are called cortoids (Flügel, 1982) (see Figure 6.2C). The micrite envelope may originates by several processes, which can include destructive micritization related to the activities of microboring organisms, constructive development of envelopes by epilithic organisms (organisms that live on or attached to rocks or other stony matter), or by partial dissolution and recrystallization (Flügel, 2004, 2010). Cortoids resemble superficial ooids in that they have only a single layer around a nucleus; however, superficial ooids are characterized by a thin concentric layer (see also Figure 6.2B) that forms by accretionary processes rather than by micritization of the nucleus.

6.1.3.1.2.2.4 Pisoids Pisoids (pisolites or pesoliths) are coated grains that resemble ooids but differ in their internal structure (Figures 6.2C and 6.3). They are generally less uniform in shape (subspherical or irregular) (see Figure 6.2C), and are commonly larger (several millimeters to centimeters). Most are of non-marine in origin, and often with a non-biogenic nucleus. Those formed in the groundwater vadose zone are called vadoids or vadose pisoids (see Figure 6.2C), and those formed in caves are called cave pearls (see also Peryt, 1983; Flügel, 2004).

6.1.3.1.2.2 Intraclasts The term intraclast refers to sand-size or larger particles, texturally analogous to rock fragments, broken from consolidated or hardened materials in one locality and redeposited at another locality within the basin of deposition (see Figures 6.2 and 6.3). "Intra" means within, in this context within the basin of deposition, and "clast" denotes broken. Many intraclasts are recycled fragments of coherent sediment. Intraclasts are of various sizes and shapes. Many are angular and their diameters may exceed 2 mm.

6.1.3.1.2.3 Aggregate Grains These are irregularly shaped carbonate grains that consist of two or more carbonate fragments (bioclasts, ooids, peloids, pellets, intraclasts) joined together by a micrite or sparite cement that is generally dark-colored and rich in organic matter (see Figure 6.2D). The shape of an aggregate grain is often irregular globular or rounded but depends on the shape of the bounded grains and the amount of coating aggregating/binding these grains. Some aggregate grains closely resemble intraclasts and are regarded by some workers as a type of intraclast (Scholle and Ulmer-Scholle, 2003). Illing (1954) identified three major subtypes (see Figures 6.2D and 6.3): (a) grapestones – aggregates resembling in shape a bunch of grapes (the binding material forms a meniscus between the grains) (see Figures 6.2D and 6.3); (b) botryoidal lumps – grapestones with superficial ooid coatings (see Figures 6.2D and 6.3A); and (c) encrusting lumps – lumps smoother than grapestones but with hollow interiors (see Figure 6.2D). Aggregate grains in modern carbonate environments are composed mainly of aragonite, but such grains in ancient limestones are dominantly calcitic. Aggregate grains are particularly common in the modern ocean in the Bahamas, in shallow, brackish waters of the tidal zone and some sea-marginal hypersaline pools. Aggregate grains form under conditions where there is a supply of firm carbonate grains, uneven water turbulence, high water circulation, and very low sedimentation rates (Winland and Matthews, 1974).

6.1.3.1.2.4 Carbonate Muds (Lime Mud) The clay- and silt-sized particles (very fine-sized calcite crystals) of carbonate sediments are called lime mud (or carbonate mud) which largely consists of tiny needles and platelets of carbonate crystals (of aragonite, ~1–5 microns or 0.001–0.005 mm long) secreted by organisms. Lime mud may also form by the accumulation of tiny skeletal components secreted by algae. Lime muds may also contain small amounts of fine-grained detrital minerals such as clay minerals, quartz, feldspar, and fine-size organic matter. The carbonate mud has a grayish to brownish, sub-translucent appearance under the microscope. This is due to the extremely small crystal sizes of carbonate grains and sparry calcite crystals. Codiacean or green algae, especially the *Penicillus*, produces fine biocrystals (needles) of aragonite (<l5 μm) within the sheaths of their filaments. When the organism dies, the filaments disintegrates and releases the needles to the sea, where they accumulate as carbonate mud. Green algae (*Udotea* and *Rhipocephalus*) are also active contributors of the carbonate mud formation. The codiacean algae are known in rocks as old as the Ordovician, and they may have contributed aragonite needles to ancient carbonate muds. Coralline red algae secrete skeletons composed of high-magnesian calcite, probably derived from the breakdown of red-algal skeletons; it is abundant in carbonate muds in present-day Florida Bay and on the Great Bahama Bank west of Andros Island. Some carbonate mud also results from the physical breakage of sand-sized and larger aragonitic skeletal materials. Most modern carbonate mud originates through organic processes, from the breakdown of calcareous algae in shallow waters to yield aragonite mud, and by the deposition of carbonate nannofossils (<35 mm in size) such as coccoliths in deeper waters to yield calcite muds (chalks).

6.1.3.2 Microcrystalline Calcite

Micrite is the short from for microcrystalline calcite. It is a very fine-grained carbonate sediment (Folk, 1959). It is present as a matrix among carbonate grains, or it may make up most or all of a limestone. A limestone composed mostly of micrite is texturally similar to a siliciclastic mudrock or shale. Micrite in ancient limestones is commonly interpreted to indicate deposition under quiet-water conditions where little or no winnowing of the fine mud takes place (see Figure 6.3B). Micrite may also form by the inorganic precipitation of aragonite, which is later converted to calcite, from surface waters that are supersaturated with calcium bicarbonate.

6.1.3.3 Sparry Calcite

Besides carbonate grains (carbonate mud) and micrite (microcrystalline calcite), the third major constituent of limestones is sparry calcite (Figure 6.4). Many limestones contain large crystals (generally between 0.02 to 0.1 mm) of calcite called sparry calcite (Figure 6.4). Sparry calcite is distinguished from micrite by its larger size and clarity and from carbonate grains by its crystal shapes and lack of internal texture. Some sparry calcite fills interstitial pore spaces between grains or fills solution cavities as a cement. Sparry calcite can also form in ancient limestones by the recrystallization of primary depositional grains and micrite during diagenesis. Sparry calcite formed by recrystallization is difficult to distinguish from sparry calcite cement (Boggs, 1992). But this distinction is important as both recrystallized spar and sparry calcite cements have a strong bearing on limestone classification and environmental interpretation.

Much of the sparry calcite in limestones occurs as a cement that fills interstitial space among carbonate grains (Figure 6.4) and is common in grain-rich limestones, such as oolites that were deposited in agitated waters and that prevented fine-grained micrite from filling pore spaces. Hence, the presence of significant amounts of sparry calcite cement in a limestone is often suggestive of

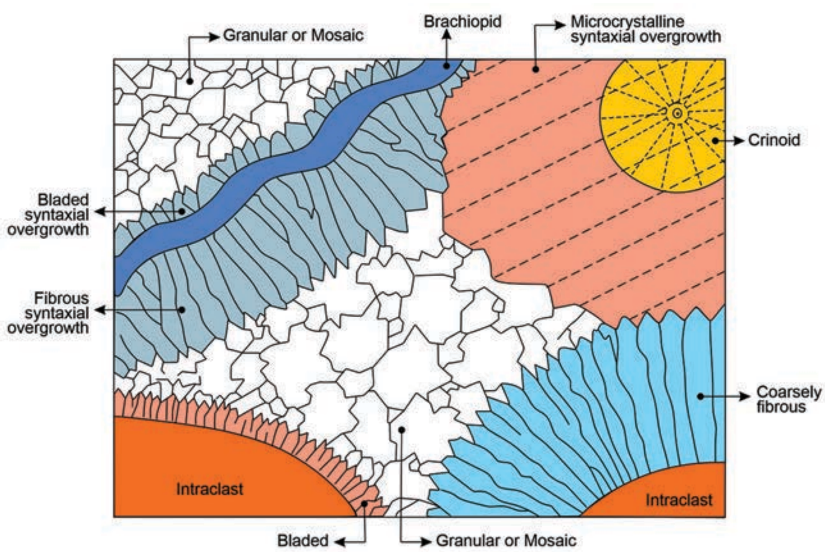

FIGURE 6.4 Sparry calcite cement fabric. Sparry calcite is the third major constituent of limestones after carbonate grains (carbonate mud) and micrite (microcrystalline calcite). Sparry calcite forms several types of cementation fabrics. The most common type is the granular or mosaic cement, composed of nearly equant crystals. Other types include fibrous cement, either coarsely or finely fibrous, bladed cement, and syntaxial cement (monocrystalline overgrowths) (most common). The figure shows the aspects of recrystallization in limestones. (Modified after Folk, 1965; Pray and Murray, 1965.)

deposition in agitated waters. But caution is needed as much of pore space fillings can also be secondary, i.e., produced by dissolution during diagenesis. Such sparry calcite cement that fills secondary pores neither has any relationship to depositional conditions nor any relevance for environmental interpretation.

Sparry calcite can form several types of cementation fabrics. The most common type is the granular or mosaic cement, which is composed of nearly equant crystals (Figure 6.4). Other types include fibrous cement, either coarsely or finely fibrous, bladed cement, and syntaxial cement (overgrowths) (Figure 6.4). The most common is the syntaxial overgrowth (monocrystalline overgrowth) noted around echinoderm fragments; this consists of plates composed of a single calcite crystal (Figure 6.4). The monocrystalline overgrowths are in optical continuity with these single-crystal echinoderm plates (Figure 6.4). These syntaxial cement rims are analogous to the overgrowths on quartz grains (see Figure 6.4, see also Chapter 5).

6.1.3.4 Dolomite Textures

Dolomite has a largely crystalline (granular) texture as opposed to limestone, which is characterized by the presence of grains, micrite, and/or sparry cement. Dolomites may be composed of crystals of nearly uniform size (unimodal size distribution) or crystals of various sizes (polymodal size distribution). Thus, based on crystal shape (more specifically crystal boundary shape), two types of dolomite are noted: planar (or idiotopic) dolomite that consists of rhombic, euhedral (well-formed) to anhedral (poorly formed) crystals; and non-planar (or xenotopic) dolomite, made up of nonplanar, commonly anhedral crystals (Figure 6.5). Each of these major kinds of dolomite can also be divided into subtypes (see Figure 6.5).

Dolomite occurs either as rhomb-shaped euhedral to subhedral crystals or as non-rhombic, commonly anhedral, crystals (see Figure 6.5). Rhomb-shaped dolomite commonly displays straight boundaries between crystals, referred to as planar dolomite (idiotopic dolomite of Gregg and Sibley, 1984) (Figure 6.5). Anhedral, non-rhombic dolomite is called non-planar dolomite (xenotopic dolomite of Gregg and Sibley, 1984) (Figure 6.5). Boundaries between crystals in non-planar dolomite are mostly curved, lobate, serrated, or indistinct (Figure 6.5). On the basis of textural characteristics, Gregg and Sibley (1984) and Sibley and Gregg (1987) divided the planar dolomite (or idiotopic dolomite) into four subcategories and non-planar dolomite (xenotopic dolomite) into three (Figure 6.5).

Planar-euhedral dolomite is made up of loosely packed but crystal-supported, well-formed rhombs (Figure 6.5). The intercrystalline spaces among crystals are filled with another mineral such as calcite, or the spaces may be empty (porous). The texture of porous dolomites of this type is sometimes referred to as sucrosic (sugary). Planar-subhedral dolomite has subhedral to anhedral crystals and very low porosity (Figure 6.5). Dolomites of this type have straight boundaries and large numbers of preserved crystal-face junctions.

Non-planar–anhedral dolomite is characterized by mainly anhedral crystals with curved, lobate, serrated, or indistinct intercrystalline boundaries (Figure 6.5). Inclusions may be abundant in the crystals, and they commonly display undulatory extinction. Non-planar-anhedral dolomite may be confused with planar-subhedral dolomite; however, the presence of irregular crystal boundaries and the scarcity of crystal-face junctions help to distinguish the two (Figure 6.5). Non-planar-porphyrotopic dolomite is similar to planar-porphyrotopic dolomite except that the crystals are mainly anhedral (Figure 6.5).

Gregg and Sibley (1984) proposed that dolomite texture is related to the temperature of formation. Below some critical temperature that lies between 50 to 100°C, called the critical roughening temperature, dolomite crystal growth produces dominantly euhedral crystals, resulting in planar textures (Figure 6.5). Growth of dolomite crystals above this temperature produces anhedral crystals, resulting in non-planar texture, although planar crystals can also form above this temperature for reasons not well understood (see Gregg, 2004; Machel, 2004).

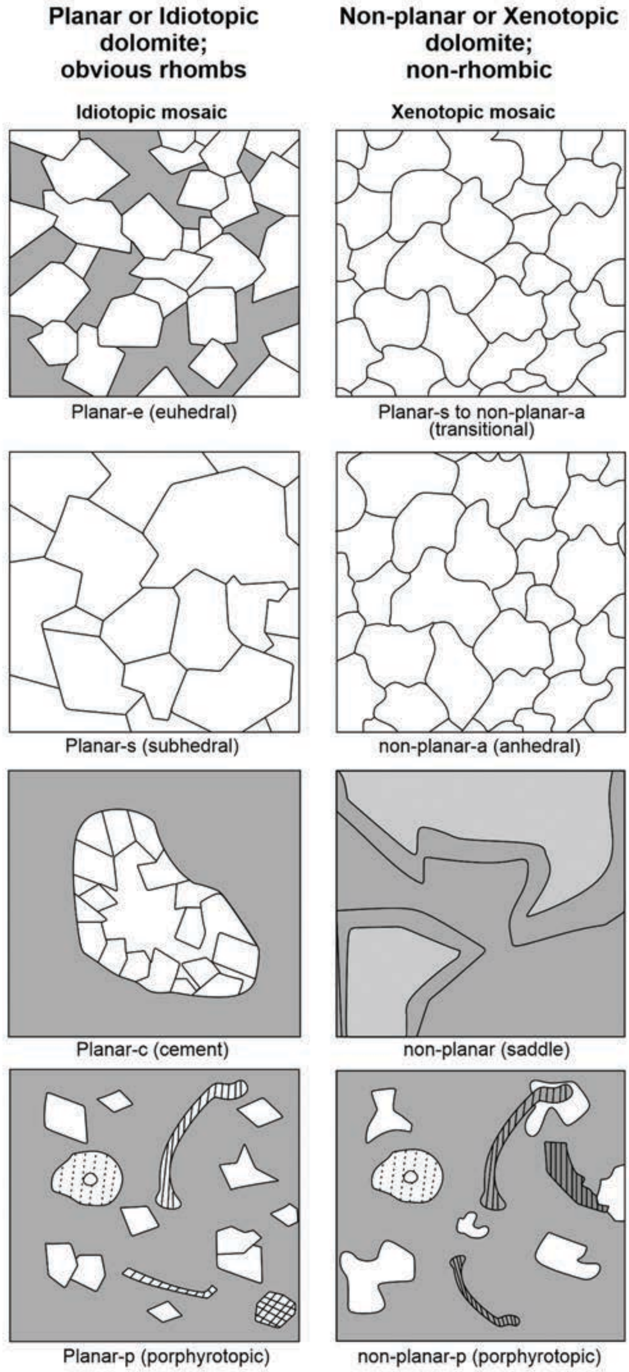

FIGURE 6.5 Dolomite textures. (Modified from Gregg and Sibley, 1984; Wright, 1984; Sibley and Gregg, 1987.) The figure shows common dolomite textural classification based on crystal-boundary shape (planar or idiotopic, non-planar or xenotopic). In the idiotopic texture, dolomite crystals are euhedral rhombs, crystal-supported frameworks with intercrystalline areas filled by another mineral or empty (porous). In xenotopic (non-planar) texture, the dolomite crystals are closely packed anhedral crystals with mostly curved, lobate, serrated, or irregular intercrystalline boundaries. Crystals often have undulatory extinction in crossed polarized light.

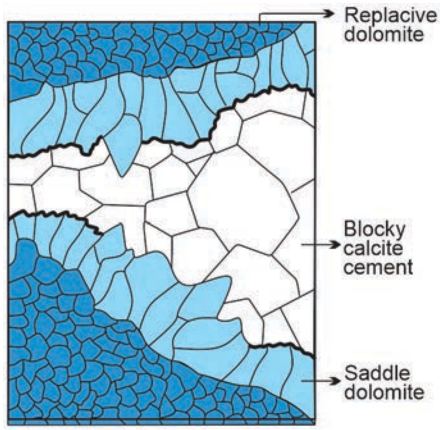

FIGURE 6.6 Saddle dolomite and blocky calcite cement in the bedding-parallel pore system. (Modified after Martín-Martín et al., 2018.) In saddle dolomite, the dolomite crystals exhibit a distinctive saddle-shaped morphology.

Some additional dolomite textures include coarse-crystalline, fine-crystalline, and saddle dolomite, among others. The coarse-crystalline dolomite is characterized by large, well-formed dolomite crystals that can be easily observed with the naked eye or under a microscope. Coarse-crystalline dolomite often forms in open spaces within vugs or fractures, where there is enough room for crystal growth. These crystals range in size from a few millimeters to several centimeters. The fine-crystalline dolomite consists of smaller dolomite crystals that are not easily visible to the naked eye. Under a microscope, these crystals appear as fine-grained or microcrystalline aggregates. These commonly occur as a replacement mineral in limestone or other carbonate rocks, where the original mineralogy has been altered by diagenetic processes. In saddle dolomite, the dolomite crystals exhibit a distinctive saddle-shaped morphology (Figure 6.6) (see also Martín-Martín et al., 2018). These crystals often form as replacements of pre-existing minerals, such as calcite, in carbonate rocks. The saddle shape is a result of the dolomite crystals growing around and enclosing the original mineral, leaving behind a void in the center. Saddle dolomite is commonly observed in hydrothermal environments or in rocks that have undergone burial diagenesis.

6.1.3.4.1 Zoned Dolomites

Many rhombic dolomite crystals are characterized by a cloudy, rhombic central zone with an outer clear rim and are referred to as zoned dolomites (Figure 6.7). The cloudy centers result from the presence of inclusions (of calcite or other minerals), whereas the clear rims are nearly inclusion-free. These zoned crystals may be formed by the replacement of a $CaCO_3$ precursor such as a micritic limestone, or they may grow into open pore space. Where they form within a precursor limestone, the cloudy centers represent replacement of the precursor $CaCO_3$. The clear rims must have been formed in empty pore spaces around the margins of the cloudy rhombs. The empty space may be created by dissolution of $CaCO_3$ from just beyond the limits of the cloudy replacement rhombs. The $CaCO_3$ dissolved from the immediately surrounding area is then reprecipitated syntaxially in the newly created space around the cloudy rhomb. Clear syntaxial rims may also form in optical continuity on dolomite crystals that project into voids.

6.1.3.4.2 Mottled and Zebra Structure

Some dolomites display a distinctive mottled structure, particularly on weathered surfaces ranging in shape from tubular to highly irregular, including anastomosing networks (Figure 6.8A). In

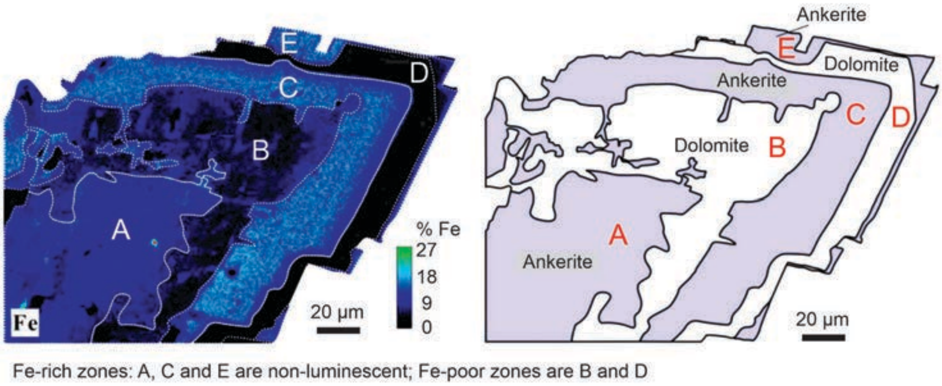

FIGURE 6.7 Cyclic zoned dolomite-ankerite cements in sandstones. (Modified after Ma et al., 2021.) Many rhombic dolomite crystals are characterized by a cloudy, rhombic central zone with an outer clear rim and are referred to as zoned dolomites.

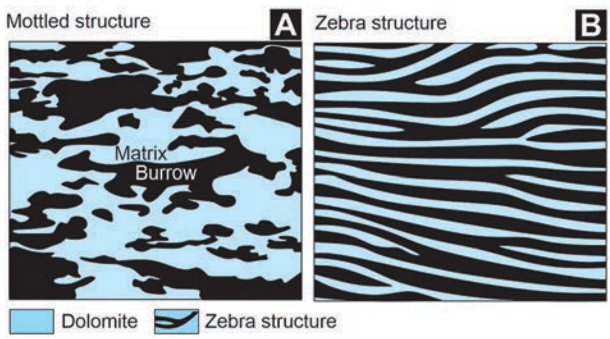

FIGURE 6.8 Dolomite textures. A: Mottled structure. Some dolomites display a distinctive mottled structure, particularly on weathered surfaces, ranging in shape from tubular to highly irregular, including anastomosing networks. B: Zebra structure. These consist of millimeter- to centimeter-scale regularly alternating light and dark bands, often consisting of dolomites.

general, mottling appears to be the result of incomplete dolomitization; however, various kinds of processes or conditions have been suggested to account for this differential dolomitization. Mottles oriented approximately parallel to bedding surfaces of the precursor limestone probably represent selective dolomitization along zones of higher porosity and permeability (Figure 6.8A). Tubular-shaped mottles may be relict borings or other organic traces. Very irregular mottles that cut indiscriminately across preserved stratification in limestones and that are not associated with obvious fracture systems are more difficult to explain. They may be diagenetic, but the mechanism responsible for isolation of the patches is not well understood. Kirkham (2004) suggests that mottled Jurassic-age dolomite in the Arabian Gulf is due to concentrations of microcrystalline iron sulfate owing to the activities of sulfate-reducing bacteria. For the formation of mottled structures (see also Yang et al., 2022, 2023), the beds probably were intensely bioturbated where the sediments and the burrow fill were coarse and loose with little clay, hence, easier to be dolomitized than the surrounding sediments resulting in partial dolomitization. This partial dolomitization occurred in the burrow system, where the degree of dolomitization was governed by the degree of bioturbation, and controlled by the occurrence and abundance of trace fossils such as those of *Thalassinoides horizontalis*, *Thalassinoides callianassa*, and *Planolites* (see Yang et al., 2022, 2023).

The zebra structure (Figure 6.8B) consists of millimeter- to centimeter-scale regularly alternating light and dark bands (Figure 6.8B) often consisting of dolomites (see Beales and Hardy, 1980; Diehl et al., 2010; Morrow, 2014; Wallace and Hood, 2018). These banded textures have major economic importance as they commonly occur associated with carbonate-hosted zinc-lead mineralization (see Sass-Gustkiewicz and Mochnacka, 1994) and in carbonate reservoir rocks, particularly in "hydrothermal dolomite reservoirs" (see Hiemstra and Goldstein, 2015). Their mode of origin is still not clear with various hypotheses: (a) formed by the replacement of sedimentary structures such as evaporite laminae (Beales and Hardy, 1980), replacement of dolomite (Sass-Gustkiewicz et al., 1982), fracturing (López-Horgue et al 2009), dissolution (Morrow, 2014), hydraulic overpressuring (Boni et al., 2000; Swennen et al., 2012) and grain growth during recrystallization (Kelka et al., 2017).

6.1.4 Classification of Carbonate Rocks

The carbonate rocks can be classified based on their mineral composition, texture, and depositional environment. One common classification scheme, though on a very broad level, is based on their texture and composition, which categorizes them into three broad types: marl, dolomite (dolostone), and limestone.

Marl is a carbonate-rich mudstone or claystone that contains a significant amount of calcium carbonate (usually calcite) (see Figures 6.9A–B). Marl can have a variable composition and texture, ranging from clay-rich to carbonate-rich layers (see Figures 6.9A–B). Quantitative classifications considerably vary in the proportions of clay and carbonate content (see Figures 6.9A–B); typically, they may contain 30–70% siliciclastics (clay, silt and biogenic silica) and 70–30% carbonates. Compton (1962), based on the proportion of siliciclastic and carbonate mud, provided another classification of marls (see Figure 6.9C). Marl shows overlapping properties of limestones and mudstones (mudrocks) (see Figure 6.9C). Marls are characterized by some fissility, due to the presence of platy clay minerals, which allows for these rocks to break along planar surfaces. Marl forms in shallow marine or lacustrine environments with high organic productivity.

Dolomite (also referred to as dolostone) is a carbonate rock that is primarily composed of the mineral dolomite ($CaMg(CO_3)_2$) and after calcite, is the second most abundant carbonate mineral. Dolostone forms through the diagenesis (post-depositional alteration) of limestone, where magnesium-rich fluids replace some of the calcium in the original limestone. Dolomites occur mostly as a diagenetic mineral (secondary mineral) in carbonate rocks. In rare cases, dolomite is also formed as a result of carbonate precipitation (like limestones) from saline waters in evaporitic environments, both coastal and continental. But, mostly, dolomite forms after deposition due to the replacement of calcite and aragonite in the presence of Mg-rich waters through a process called dolomitization. Dolomitization produces rhombohedral crystals of dolomite, thus, modifying in part or totally, the texture of the original limestone. Although, some dolostones do preserve fossils or other allochems, while others only show a crystalline texture, consisting of recrystallized rhombohedrons of dolomite (possibly representing primary dolomite or a completely dolomitized limestone). Dolomitization of carbonate sediments occurs either by the evaporation of seawater producing gypsum-rich brines that migrate in the underlying carbonate sediment, react with calcite to produce dolomite (evaporative reflux) or by the interaction between meteoric and marine groundwater (mixing of seawater and meteoric water). However, most dolostones contain calcite or aragonite along with dolomite, as the dolomitization process is often incomplete, thus, enabling to classify dolostones, based on the amount of dolomite with respect to calcite (see Figure 6.9D).

Limestone is a carbonate rock primarily composed of the mineral calcite ($CaCO_3$) and forms from the accumulation and lithification of carbonate sediments, such as shells, coral fragments, or microorganisms. Limestone has a range of textures, including fine-grained (micrite), coarser-grained (sparite), or a mixture of both and composition (detailed below).

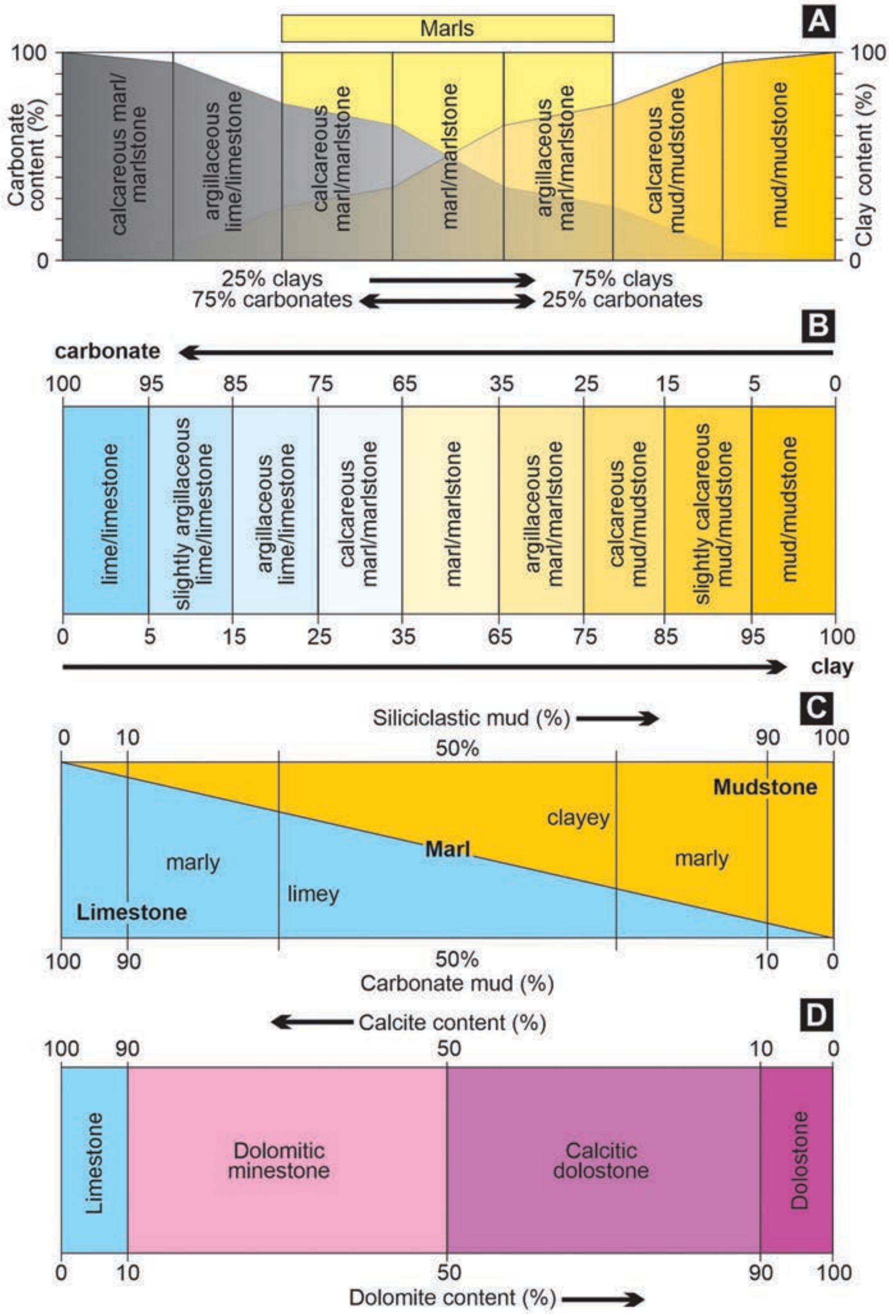

FIGURE 6.9 Classification of carbonate rocks based on texture and composition. A–B: Marl is a carbonate-rich mudstone or claystone with a significant amount of calcium carbonate (usually calcite). C: Classification of marls based on the proportion of siliciclastic and carbonate mud (after Compton, 1962). D: Dolostone classification based on the amount of dolomite with respect to calcite.

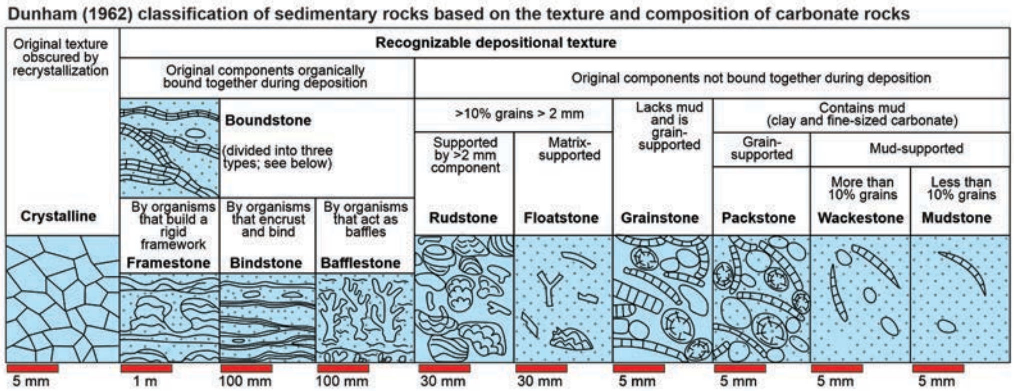

FIGURE 6.10 Dunham classification (based on depositional structures; field-based). (Modified after Dunham, 1962; Embry and Klovan, 1971.) Embry and Klovan (1971) divided Dunham's boundstone group into subgroups based on the nature of binding. This classification expansion reflects the ecological classifications of dominant reef organisms into guilds (constructors, binders, bafflers).

6.1.4.1 Dunham Classification (Based on Depositional Structures; Field-Based)

The Dunham classification is based on the texture and composition of carbonate rocks (Figure 6.10). Dunham (1962) recognized two broad limestone categories in which the original components are either bound together (collectively called boundstones) or not (no collective name) (see Figure 6.10).

In boundstones, the particles being deposited are bound by organisms or that they consist of frameworks constructed by organisms; this group includes reefs, stromatolites, and travertine (Figure 6.10). Later, Embry and Klovan (1971) divided Dunham's boundstone group into subgroups based on the nature of binding (Figure 6.10). Thus, limestones bound together with microbial laminae are called bindstones, those that originate in place (autochthonous limestones) as frame-built reefs are framestones and those that predominantly consists of sediments trapped by baffling organisms are bafflestones (Figure 6.10). This classification expansion reflects the ecological classifications of dominant reef organisms into guilds (constructors, binders, bafflers). The particle-supported rocks are further subdivided into two groups: (a) particle-supported limestones devoid of lithified lime mud, called as grainstones, and (b) particle-supported limestones containing some lithified lime-mud matrix, but not enough to keep the sand-size particles from touching one another, and called as packstones (see Figure 6.10). The matrix-supported limestones are also subdivided into two classes; (a) mud-supported limestones containing at most 10% sand-sized or larger particles, called as mudstones, and (b) mud-supported limestones containing at least 10% sand-sized or larger particles, called wackestones. In the Dunham classification, micrite is not used, where the fine-sized carbonate rock is called a mudstone (lime mudstone) (Figure 6.10).

The rationale of this classification is that it allows one to map changes in the rate of production of sand-sized particles relative to the rate of accumulation of lime mud / carbonate mud (Dunham, 1962). Thus, in calm waters, lime mud, if present, settles on the bottom and remains there. Hence, the limestone derived by lithification from the original lime mud is distinguished from the one derived by lithification of carbonate sediments devoid of lime mud. This relationship between sand-size particles and lime mud thus distinguishes an original sediment deposited in calm water from a sediment deposited in agitated waters.

Dunham's (1962) classification is based on depositional texture that are useful in classifying carbonate rocks. These include (a) the presence or absence of carbonate mud (particles less than 20 microns); (b) abundance of carbonate grains (particles larger than 20 microns); (c) whether the grains are mud-supported or grain-supported; and (d) evidence of organic binding during deposition. In general, the Dunham classification provides a useful framework for describing and categorizing carbonate rocks based on their texture and composition. It allows geologists to better understand the depositional environment, diagenetic processes, and potential reservoir properties of carbonate rocks.

6.1.4.2 Folk (Textural Classification; Laboratory-Based)

Folk (1959, 1962) subdivided the constituents of limestones into two broad categories: (a) allochems (sand-sized or larger particles) or allochemical constituents; and (b) orthochems or orthochemical constituents, and consisting of micrite (lithified original lime mud), and sparry calcite cement (Figure 6.11A). Four types of allochems are noted: (a) intraclasts, (b) ooids/oolites, (c) fossils/skeletal particles, and (d) peloids/pellets. The interstices among the allochems are filled with micrite or sparry calcite cement (the orthochemical constituents). This combinations of allochems and orthochems yields eight types of limestones (Figure 6.11B). In addition, micrite may lack sand-sized particles (dismicrite), and hence stands alone as a ninth kind of limestone (Figure 6.11B). A tenth kind is a reef rock (biolithite) (Figure 6.11B).

Folk (1959, 1962) assigned names to eight of these ten types of limestone by using composite words of two parts (see Figure 6.11B): (a) an initial abbreviated for the allochems, and (b) a word designating the orthochemical constituents. The orthochem words are (a) sparite (limestones with a cement of sparry calcite), and (b) micrite. The prefixes for the allochems are: intraclasts = intra; ooids = Oo; fossils = bio; pellets = pel. Thus, the names of these eight kinds of limestone are intrasparite, intramicrite, oosparite, oomicrite, biosparite, biomicrite, pelsparite, and pelmicrite (Figure 6.11B). If several kinds of particles are important constituents of a limestone, their abbreviations are strung together in order of increasing abundance. For example, a limestone composed of micrite and having allochems composed of 10% intraclasts, 20% skeletal debris (fossils), and 70% peloids, then it would be called an intrabiopelmicrite. The distinctive aspect of Folk's classification is its use of names for eight major kinds of limestone based on what lies between the particles (spar or micrite) (Folk, 1959).

6.1.5 Carbonate Sedimentary Structures

Carbonate sedimentary structures are features and textures that are commonly found in carbonate rocks, and composed primarily of calcium carbonate ($CaCO_3$) minerals such as calcite or aragonite. These structures provide important clues about the depositional environment, the processes involved in carbonate sedimentation, and aid in interpreting the geologic history of an area. Some common carbonate sedimentary structures include bedding, fossiliferous limestone, ooids, stromatolites, mud mounds, algal mats, and karst features. These are described at length in Chapter 4 Sedimentary structures; here they are briefly mentioned.

Bedding is layering or stratification observed in the rock. It varies in thickness and composition, reflecting changes in sedimentation conditions over time. Ooids are small, spherical grains composed of concentric layers of calcium carbonate formed in shallow, high-energy environments such as beaches or shallow marine environments. Ooid-rich rocks are known as oolitic limestone. Stromatolites are layered structures formed by the growth of microbial communities, such as cyanobacteria, in shallow marine or freshwater environments. They are often found in ancient carbonate rocks and provide evidence of early life on earth. Mud mounds, also known as carbonate mounds or bioherms, are large, mound-like structures formed by the accumulation of fine-grained carbonate mud and the growth of organisms such as algae, sponges, or corals. They can reach several meters in height and are useful proxies of past environmental conditions. Algal mats are layers or sheets of organic material formed by the growth and accumulation of algae in shallow marine or lacustrine environments. They are preserved in carbonate rocks and are often associated with

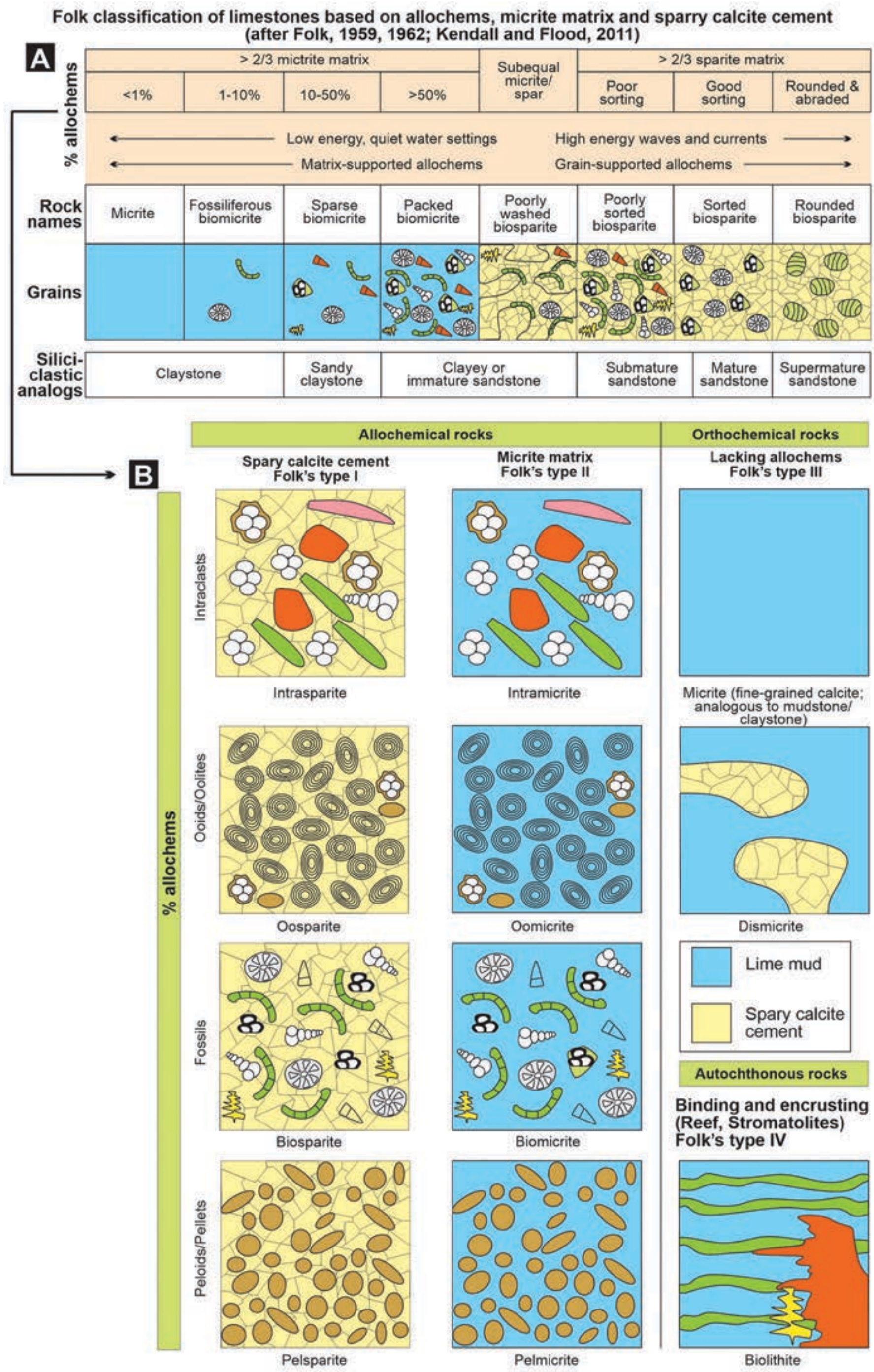

FIGURE 6.11 Folk (textural classification; laboratory-based). (Modified after Folk, 1959, 1962.) A: Folk (1959, 1962) subdivided the constituents of limestones into two categories: allochems (sand-size or larger particles), and (b) orthochems or orthochemical constituents, and consisting of micrite (lithified original lime mud), and sparry calcite cement. B: Four types of allochems (particles) were noted: (1) intraclasts, (2) ooids/oolites, (3) fossils/skeletal particles, and (4) peloids/pellets. The Folk classification provides a useful framework for describing and categorizing carbonate rocks based on their composition and texture.

carbonate buildups. Karst features, in carbonate rocks, are formed by the dissolution of calcium carbonate by groundwater; these features include the formation of caves, sinkholes, or underground drainage systems (see also Jain, 2014). These structures are characteristic of areas with extensive carbonate rock formations. These carbonate sedimentary structures are just a few examples of the diverse features that can be encountered in carbonate rocks. They provide important insights into the depositional environments, biological activity, and diagenetic processes that occurred during the formation of carbonate rocks.

6.1.5.1 Bedding Planes

Bedding planes, also known as bedding surfaces or stratification planes, are horizontal surfaces or layers that separate different sedimentary rock layers (see Figure 4.2 of Chapter 4). They are formed during the deposition and accumulation of sediments, and represent changes in sedimentary conditions over time. Sediments, such as sand, silt, or clay, settle out of water and accumulate over time. As each layer is deposited, it forms a distinct bedding plane. These layers vary in thickness, composition, and texture, depending on the prevailing sedimentary environment. Bedding planes exhibit various characteristics, such as horizontal orientation, color, and composition variations, and sedimentary structures, among others. Bedding planes are typically parallel to the earth's surface and are horizontal or nearly horizontal (i.e., have a largely horizontal orientation) due to the fact that the sediments settle under the influence of gravity, resulting in horizontal layering. Adjacent bedding planes may have different colors or compositions due to changes in the type of sediment being deposited or variations in environmental conditions, i.e., they exhibit color and compositional variations. Additionally, each bedding plane may contain distinct sedimentary structures, such as cross-bedding, ripple marks, or mud cracks that provide useful information about the prevailing depositional environment (see Chapter 4 for the types of sedimentary structures). Hence, bedding planes are essential for interpreting the history and processes involved in the formation of sedimentary rocks. They provide valuable information about the depositional environment, such as whether it was a marine, fluvial (river), lake (lacustrine) or aeolian (wind) environment. The characteristics of the bedding planes, including the sedimentary structures and variations in composition, can help geologists reconstruct ancient environments and understand the changes that occurred over time. Bedding planes also play a role in the mechanical properties of sedimentary rocks. They can influence the rocks' strength, permeability, and ability to store fluids, which has implications for engineering and resource exploration.

6.1.5.2 Hardgrounds

Hardgrounds are common sites of condensation. Examples include carbonate hardgrounds on shallow carbonate platforms that provide a home to boring and entrusting organisms (Figure 6.12). In general, hardground formation represents a gradual process and progresses via firmgrounds (Stage 1), to incipient hardgrounds (Stage 2), and then to fully lithified hardgrounds (Stage 3) (see also Pandey et al., 2018) (Figure 6.12). Condensed sections are a special kind of stratigraphic incompleteness (see Nicolaides and Wallace, 1997; Föllmi, 2016; Christ et al., 2015; Pandey et al., 2018). Thus, hardgrounds contain internal discordances and hiatuses caused by erosion or non-deposition. Bedload transport and erosion of sediment on the sea floor are typical of shoreface dynamics on siliciclastic shelves and carbonate platforms. Non-deposition may be due to sediment starvation (other than fine particles in suspension), for example during transgression when river supply is thwarted by rising base levels or situations where sediment bypasses a shelf or platform on its way to deeper waters.

Hardgrounds are surfaces within sedimentary rocks that are lithified (hardened) and resistant to erosion compared to the surrounding sediment (Figure 6.12). These surfaces are distinguished by their compact and cemented nature, often contrasting with the softer, unconsolidated sedimentary layers above and below. Hardgrounds typically form in marine environments, where they represent

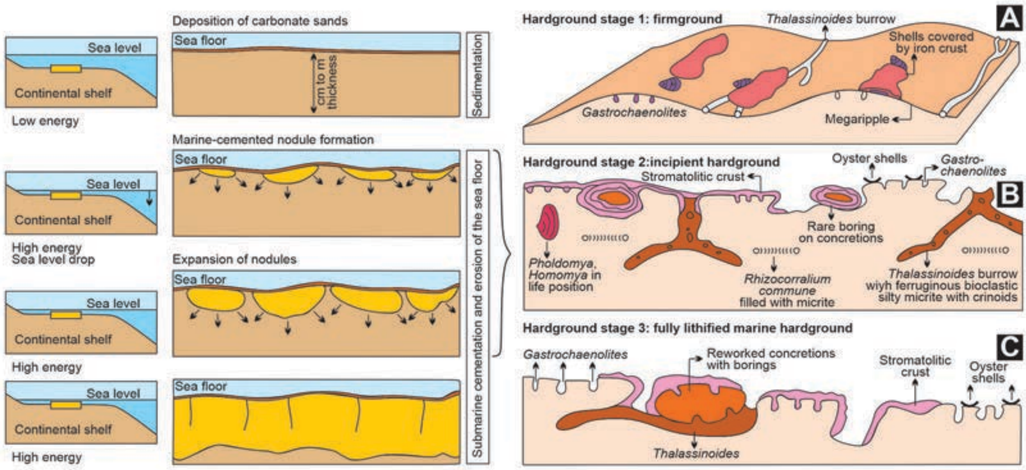

FIGURE 6.12 Conceptual model for hardground formation on a shallow-marine shelf. A–C: Hardground formation progresses via firmgrounds (Stage 1), to incipient hardgrounds (Stage 2), and then to fully lithified hardgrounds (Stage 3). (Left panel: modified after Nicolaides and Wallace, 1997; right panel: modified after Pandey et al., 2018.)

periods of relative stability and reduced sedimentation (Figure 6.12). They can develop in several ways such as cementation, biogenic activity, abrasion, and compaction, among others (Figure 6.12A). Cementation is the accumulation of cementing materials, such as calcium carbonate or silica that binds the sediment particles together, forming a hard layer. This process often occurs due to the precipitation of minerals from seawater or pore fluids within the sediment. Biogenic activity is the activity of organisms, such as burrowing or encrusting organisms, that disrupts sediment deposition and promotes the formation of hardgrounds (Figure 6.12B). Burrowing animals mix sediments and thus promote cementation, leading to the hardening of the substrate (Figure 6.12B). Abrasion and compaction are produced by physical processes such as wave action or currents that result in the compaction and consolidation of sediments, eventually forming hardgrounds (Figure 6.12C). These processes remove finer particles and leave behind a more cemented layer (Figure 6.12C). Hardgrounds often exhibit distinct features, such as borings, encrustations, or fossilized remains of organisms that colonized the hard surface (Figures 6.12B–C). These features provide valuable information about the past environments and the organisms that lived during the time of hardground formation (see Pandey et al., 2018). The presence of hardgrounds can have important implications for the interpretation of sedimentary rocks. They act as barriers to fluid flow, influencing the distribution of groundwater or hydrocarbon reservoirs. Hardgrounds also serve as horizons for correlation and stratigraphic analysis, as they often represent significant breaks or changes in sedimentation patterns.

6.1.5.3 Tepee Structure

A tepee structure is a distinctive type of geological formation characterized by the folding and tilting of rock layers into a conical or tent-like shape (Figure 6.13). These structures are typically found in sedimentary rocks, such as sandstones or shales, and are often associated with the presence of resistant layers or beds within the rock sequence. First named for their symmetrical, inverted V-shaped vertical profile in the 'Permian Reef Complex' of western Texas and adjacent New Mexico (Adams and Frenzel, 1950), these are conical shapes marked by low ridges (Figure 6.13). The ridges are often interconnected creating an irregularly polygonal network in plan view (Figure 6.13). Tepee

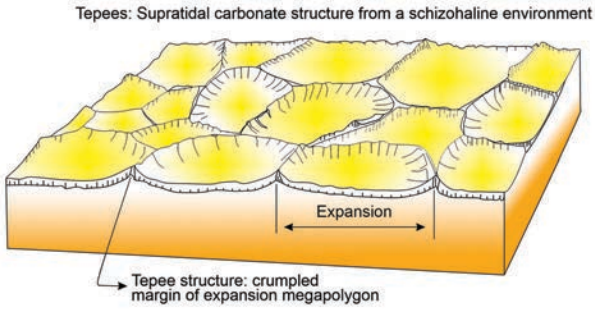

FIGURE 6.13 Tepee structure. This is a distinctive type of geological formation characterized by the folding and tilting of rock layers into a conical or tent-like shape.

structures are characteristic of ancient tropical tidal flats and arid coasts (Inden and Moore, 1983; Tucker and Wright, 1990; Pratt et al., 1992). They are common in many ancient carbonate tidal flat successions (modern analogs are known from tropical coastal areas), hence, are considered robust evidence for synsedimentary lithification of subaerially exposed carbonate sediments. Existing explanations for their formation involve desiccation, thermal expansion and contraction, fluctuating groundwater pore pressure or the diagenetic force of crystallization (see Pratt et al., 1992). Tepee structures are often seen in arid or semi-arid environments, where the combination of limited vegetation cover, dry climate, and high evaporation rates leads to the development of desiccation cracks. These cracks enhance differential compaction and stress within sedimentary layers, facilitating the formation of tepee structures. The process typically involves the interplay of several factors: differential compaction, confining pressure and moisture content. Sedimentary rocks are often composed of layers with varying permeability and porosity. When these rocks undergo compaction due to the weight of overlying sediments, the layers with lower permeability and porosity are less easily compressed. This differential compaction causes the resistant layers to buckle and fold, leading to the formation of the tepee structure. The presence of confining pressure, such as from overlying sediments or lithostatic pressure, also contributes to the formation of tepee structures. The pressure acts on the sedimentary layers, causing them to fold and tilt into a conical shape. Moisture content within the sedimentary layers plays an important role in the tepee formation. As water infiltrates the rock layers, it causes localized swelling and expansion. This expansion, combined with differential compaction and confining pressure, further promotes the folding and tilting of the layers. Tepee structures also act as natural traps for oil and gas, as the folding and tilting of the layers create structural traps that can hold hydrocarbon reservoirs.

6.1.5.4 Paleokarstic Surfaces

Paleokarstic surfaces (paleokarst) are ancient landscapes or topographic features that have been modified or sculpted by karst processes in the past. The karst processes are related to the dissolution of soluble rocks, such as limestone or dolomite, by acidic water, resulting in the formation of distinctive landforms and underground drainage systems (see Gabrovšek, 2002; Osborne, 2013) (see Figure 6.14). Paleokarstic surfaces form when karst processes occur over long periods of time and then become buried or covered by subsequent sedimentation. These buried karst landscapes are preserved and later exposed due to erosion or tectonic uplift. Thus, paleokarst features provide valuable insights into past hydrological conditions, paleoclimate, and the evolution of landscapes. Paleokarstic surfaces often exhibit several characteristic features, including solution cavities and conduits, karst valleys and depressions, karstic breccias and collapse structures, and paleokarstic fills, among others. These are briefly enumerated below.

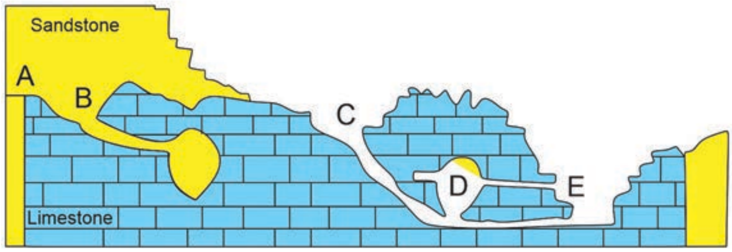

FIGURE 6.14 Paleokarst and unconformities. A: Unconformity surface. B: Doline in unconformity surface filled with sandstone. C: Young doline and cave invading ancient system. D: Ancient cupola, partly exhumed, intersected at base by modern cave. E: Ancient hall from cupola, intersected by valley incision, now forming cave entrance. (Modified after Gabrovšek, 2002.)

Solution cavities and conduits are void spaces formed by the dissolution of soluble rocks and range in size from small solution pockets to large cave systems. When exposed, these features are filled with sediment or preserve evidence of past water flow, such as speleothems (cave formations) or sedimentary deposits. Karst valleys and depressions are surface features that result from the collapse or subsidence of underlying karst cavities. They can take the form of sinkholes, dolines, or uvalas (see Figure 6.14). These depressions are filled with sediment or water, creating unique ecosystems or serving as important groundwater reservoirs. Karstic breccias and collapse structures are formed when the roof of a karst cavity collapses, creating a brecciated or fragmented rock mass. These collapse structures are preserved and exposed as paleokarstic features, providing evidence of past subterranean processes. Paleokarstic fills are sedimentary deposits that accumulate within karst cavities or depressions. They consist of clastic sediments, such as sand or clay, or chemical precipitates, such as calcite or gypsum. These fills provide valuable information about past environmental conditions and sedimentation processes. Paleokarstic surfaces also provide insights into past landscapes, paleoenvironments, and the history of groundwater flow.

6.1.5.5 Current and Wave Structures

Current and wave structures refer to distinctive features formed by the action of water currents and waves on sedimentary environments. These structures provide valuable information about the dynamics and processes of water flow in various aquatic environments, such as rivers, lakes, coastal areas, and deep-sea settings. The study of current and wave structures enables a better understanding of sediment transport, depositional environments, and the interpretation of ancient sedimentary rocks. Their analysis enables paleoenvironmental conditions to be reconstructed and the dynamics of ancient water bodies to be interpreted. Among others, some of the commonly noted current and wave structures include ripple marks, cross-bedding, flute marks, tool marks, bedforms, and storm deposits. These are described in detail in Chapter 4; they are briefly noted here.

Ripple marks are small-scale, wave-like patterns formed on the surface of loose sediment by the action of water currents or waves. They are classified into two main types: symmetrical ripple marks that have a symmetrical shape and form in oscillatory flow conditions, and asymmetrical ripple marks that have an asymmetric shape and form in unidirectional flow conditions. Cross-bedding is the inclined layers of sediment within a larger sedimentary unit. These inclined layers are formed by the migration of sand or sediment in response to water currents or wind. Cross-bedding provides information about the direction and intensity of the paleocurrents that deposited the sediments. Flute marks are elongated, streamlined depressions or ridges formed on the bed of a river or other water body by erosion or deposition of sediments. They are typically oriented in the direction of water flow and indicate the direction of paleocurrents. Tool marks are features formed by the impact or

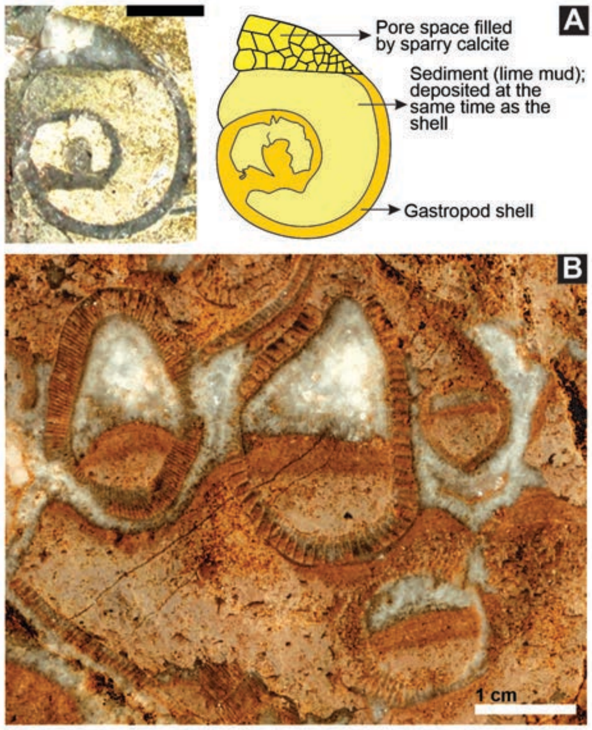

FIGURE 6.15 Geopetal structure. This provides clues about the original orientation of the sediment during deposition. For example, when snail shells are deposited in a bunch of sediment, they serve as tiny architectural elements, with a "roof" that protects their interiors. Any sediment mixed into the shell's interior will settle out (more or less horizontally), and then there will be empty space (filled with water, probably) above that. (Modified after Jago et al., 2020.)

abrasion of sediment particles carried by water currents or waves. They include grooves, scratches, or polished surfaces on rocks or other hard substrates. Tool marks provide evidence of the erosive power of water and the transport of sediments. Bedforms are larger-scale structures formed by the interaction of water currents or waves with sediments. Examples include dunes, which are asymmetrical mounds of sediments formed by the action of water currents, and ripples (smaller-scale features formed by the interaction of waves with sediment). Storm deposits are sedimentary layers or structures formed during intense storm events. These include storm beds, which are thick layers of sediment deposited rapidly during a storm, and storm shells, which are accumulations of shells or other marine organisms displaced by storm waves.

6.1.5.6 Geopetal Structures

Geopetal structures are sedimentary features that provide clues about the original orientation of the sediment during deposition (see Figure 6.15). The term "geopetal" comes from the Greek words "*geo*" meaning earth and "*petal*" meaning to fall or settle. These structures are formed when sediment accumulates in a way that preserves a record of the direction of gravity at the time of deposition. Geopetal structures can be observed in various sedimentary environments, such as marine, lacustrine (lake), or fluvial (river) settings. They are particularly useful in interpreting the depositional history of sedimentary rocks and understanding the paleoenvironmental conditions at the time of deposition.

For example, when gastropod shells are deposited, they serve as tiny architectural elements, with a "roof" that protects their interiors. Any sediment mixed into the shell's interior will settle out (more or less horizontally), and then there will be empty space (filled with water, probably) above that. As burial proceeds and diagenesis begins, that pore space may be filled with a mineral deposit, such as sparry calcite (see Figure 6.15A). The sediment (lime mud) is yellowish gray and granular (see Figure 6.15A). The sparry calcite is gray to white and crystalline, continuous with the shell material (see Figures 6.15A–B). This "cavity fill" structure (where the snail shell is the cavity being filled, partly by entrained mud and partly by diagenetic crystallization) serves as both a paleo-horizontal "level" and a geopetal structure (indicating which way was up when the sediment was deposited) (Figures 6.15A–B).

6.1.5.7 Fenestral Cavities or Birdseyes

The term "fenestral" comes from the Latin word "*fenestra*" meaning window, which describes the appearance of these cavities (see Figure 6.16). They resemble small windows or openings within the rock, hence the name "fenestral cavities" or "birdseyes" (Figure 6.16); in simple terms these cavities are calcite-filled voids that sparkle in the sunlight. The term "fenestral fabrics" was first used to characterize abundant voids (fenestrae) resulting from the partial or complete filling of fenestrae with sediments and/or cement (Tebbutt et al., 1965; Wolf, 1965). Fenestral cavities are irregular cavities found in muddy intertidal to supratidal carbonate sediments (Flügel, 2010). The fenestral cavities are isolated bubble-like vugs of 1–3 mm size range. They are often associated with microbial and algal mats, in supratidal, and in upper intertidal settings. Commonly these are associated with marshes, lagoons, or other shallow-water environments where fine-grained sediment is deposited. They are a type of growth cavity commonly found in sedimentary rocks, and characterized by irregular shape, and often filled with secondary minerals or sedimentary materials such as calcite, quartz, or clay minerals; the cavities sparkle as they are often filled with sparry calcite cement. Additionally, these can also be filled with sedimentary material, such as sand or silt, which are transported into the cavities by water or wind. The nature of the filling material can provide information about the sedimentary processes that occurred after the formation of the cavities, i.e., clues about the timing and sequence of diagenetic processes. Fenestral cavities typically form in fine-grained sedimentary rocks, such as shale or mudstone, which have a layered or laminated texture. These rocks often contain thin layers of different composition or porosity, such as organic-rich layers or layers with higher clay content. Over time, these layers undergo differential compaction or erosion, creating voids or cavities (Figure 6.16). The formation of fenestral cavities is often related to the presence of organic matter or the activity of burrowing organisms, which create voids

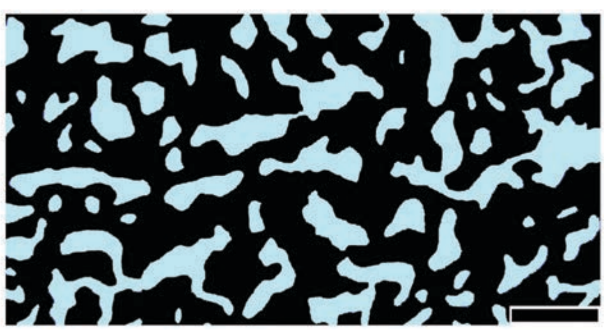

FIGURE 6.16 Fenestral cavities or birdseyes. These resemble small windows or openings within the rock, hence the name "fenestral cavities" or "birdseyes"; in simple terms, these cavities are calcite-filled voids that sparkle in the sunlight. (Modified after Tebbutt et al., 1965.)

or channels within the sediment. Fenestral limestones are present in reef-like sequences of different world regions and during various ages (Flügel, 2010).

The fenestral cavities take a number of forms: birdseye fenestrae (irregular, "birdseye"-shaped cavities, usually 1–5 mm across, formed by gas entrapment in the sediment); laminoid fenestrae (long, thin cavities, parallel to the sediment laminae, formed particularly in algal, laminated muds, and produced by the decay of organic material); and tubular fenestrae (cylindrical, near vertical tubes, formed by burrowing organisms or plant rootlets). Fenestral cavities may become filled with sparry calcite (sparite). If they remain unfilled, the fenestrae are responsible for the development of fenestral porosity in the sediment.

6.1.5.8 Laminoid Fenestrae

Laminoid fenestrae are characterized by their laminated or layered appearance. They typically form in fine-grained sedimentary rocks, such as shale or mudstone, which have a layered or laminated texture. The layers within the fenestral cavities can be composed of different materials, such as clay, silt, or organic matter. These layers can be parallel to the bedding plane or cut across them, creating a cross-cutting or oblique relationship with the surrounding rock layers. The layers within these rocks have different compositions or porosities, leading to the formation of fenestral cavities. The formation of laminoid fenestrae is often related to sedimentary processes, such as differential compaction or erosion. As the sediment is deposited and undergoes burial, some layers compact more than others, creating voids or cavities. Alternatively, erosional processes remove certain layers, leaving behind empty spaces. Laminoid fenestrae are also influenced by biological activity. Burrowing organisms, such as worms or clams, create voids or channels within the sediment, leading to the formation of fenestral cavities. These cavities, over time, become filled with sediments or secondary minerals. The presence of laminoid fenestrae provide information about the depositional environment and the processes that might have occurred during sedimentation. The layered nature of these fenestrae indicate periods of changing rates of sedimentation or variations in sediment composition. Additionally, the filling materials within the fenestrae provide insights into the diagenetic processes that might have occurred after the formation of the cavities.

6.1.5.9 Tubular Fenestrae

Tubular fenestrae are a type of fenestral cavity (vertical to subvertical) that are small (0.1 to 1 mm wide), cylindrical or tube-like shaped openings or cavities within the rock that are filled with secondary minerals or sedimentary material such as sand or silt that are transported into the cavities by water or other sedimentary processes. Alternatively, they can be filled with secondary minerals such as calcite or quartz that precipitate from fluids that infiltrate the rock. These cavities also vary in size, ranging from a few millimeters to several centimeters in diameter; they can intersect or cut across the bedding planes of the rock. Tubular fenestrae are cylindrical and straight to bifurcating, although U-, crescent-, and hook-shaped tubes also occur. Tubular fenestrae have rounded to irregular cross-sections and distinction between tubular and irregular fenestrae is sometimes difficult. Some tubular fenestrae appear to grade into laminoid fenestrae with increasing horizontal flattening. Locally, laminoid fenestrae are concentrically arranged about tubular fenestrae. The formation of tubular fenestrae is often related to several processes. One common mechanism is the dissolution of minerals within the rock. Certain minerals, such as calcite or gypsum, are soluble in water under specific conditions. As water infiltrates the rock, it dissolves these minerals, creating tubular voids or cavities. Another process is through the activity of organisms. Burrowing organisms, such as worms or clams, create tunnels or burrows within the sediment, which later become filled with sediment or secondary minerals. These burrows have a tubular shape, contributing to the formation of tubular fenestrae. The shape and orientation of these fenestrae indicate the direction and intensity of fluid flow within the rock. The filling materials within the fenestrae provide insights into the diagenetic processes that occurred after the formation of the cavities.

6.1.5.10 Vugs

Vugs are small to medium-sized cavities or voids found in rocks. Vugs vary in shape and size, ranging from microscopic to several meters in diameter. They may be spherical, elongated, irregular, or interconnected. The walls of the vugs are often lined with crystals or other minerals that precipitate from fluids filling the voids. Vugs can have economic significance when minerals fill in the vugs, they form valuable mineral deposits. Additionally, vugs serve as reservoirs for groundwater or hydrocarbons. They are formed by a variety of processes such as by dissolution, gas bubbles, and fossilization, among others. In some cases, due to chemical reactions with water or other fluids, minerals within the rock dissolve over time; this dissolution creates voids or cavities. Organic materials such as shells or plant matter decays and leaves voids in the rock (fossilization). These voids can become vugs when minerals fill in these empty spaces. Vugs provide useful information about the formation and history of the rock. The minerals lining the vugs indicate the composition of the fluids that filled them, which can help in better understanding the geological processes that might have occurred in the past.

6.1.6 Controls on Carbonate Deposition

The conditions that remove carbon dioxide from seawater and ultimately form limestone are analogous to uncapping a can of soda, resulting in the escape of gas. Uncapping can happen by three methods: uncap the can and thus release the pressure, heat it, or shake it. The CO_2 dissolved in seawater is similar to the fizz in the soda can. Broadly, seven parameters control the amount of CO_2 that can be dissolved; of which three are the most important. These are temperature, pressure (water depth), and degree of agitation. The other four parameters are organic activity, light, sediment masking and clogging, and the carbonate compensation depth.

A rise in seawater temperature promotes limestone deposition, hence carbonate sediments (modern and ancient) form more readily in tropical seas than in polar waters. Reduction in pressure (or depth) of seawater also influences limestone deposition; carbonate sediments form more easily in shallow waters than in the deep. The agitation caused by the breaking waves within the surf zone mixes seawater with air resulting in the additional absorption of CO_2 by the atmosphere, promoting limestone formation. Hence, modern fringe reefs grow faster in the direction that faces breaking waves.

Clams, snails, brachiopods, and zooplankton extract their calcareous skeletons from the seawater directly, whereas plants such as phytoplankton and algae promote $CaCO_3$ precipitation as they remove CO_2 from seawater by photosynthesis (combining it with water to produce organic tissue and energy). Thus, organic activity in the form of plants and animals precipitate $CaCO_3$ directly or change the geochemical environment for $CaCO_3$ precipitation to occur. As photosynthetic organisms such as calcareous algae and hermatypic corals need light for photosynthesis, hence, most large carbonate deposits form in shallow waters (i.e., in <20 m depth) where adequate light is available.

Additionally, besides available light, clear waters are also needed as muddy waters inhibit coral and algal growth. The solubility of $CaCO_3$ controls carbonate deposition, where calcium carbonate deposits in shallow, warm seawater, but dissolves in cold seawaters. This is because CO_2 easily dissolves in cold water, so $CaCO_3$ also dissolves in cold water. This depth of dissolution is called the calcite compensation depth (CCD), where the rate of $CaCO_3$ formation and sinking is equal to the rate the material dissolved (Figure 6.17). On modern abyssal plains the areal distribution of calcareous ooze (i.e., the unconsolidated shells of floating pelagic organisms that thrive in the photic zone; 0–200 m) is controlled by the temperature and pressure of very deep seawaters. Thus, below the CCD, no $CaCO_3$ is preserved; i.e., mostly below 4500 meters (5000 m in warm equatorial waters to 3000 m in polar waters). The depth of the CCD varies with water temperature; colder temperatures increase the rate of solution. Calcareous mudstones rather than limestones accumulate when clay

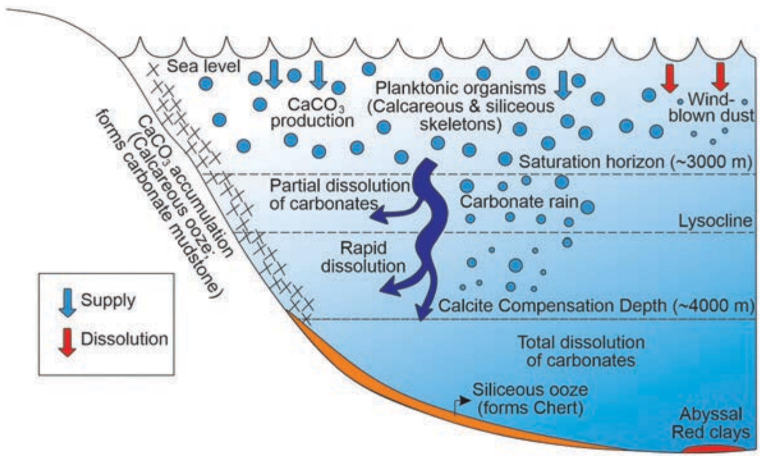

FIGURE 6.17 Calcite compensation depth (CCD). The solubility of $CaCO_3$ controls carbonate deposition, where $CaCO_3$ deposits are in shallow, warm seawaters, but dissolves in those that are cold. This is because CO_2 easily dissolves in cold water, hence $CaCO_3$ also dissolves in cold water. This depth of dissolution is called the calcite compensation depth (CCD), where the rate of $CaCO_3$ formation and sinking is equal to the rate at which the material dissolved. Hence, below the CCD, no $CaCO_3$ is preserved; i.e., mostly below 4,500 meters (5,000 m in warm equatorial waters to 3,000 m in polar waters).

and silt are supplied more rapidly as compared to carbonate sediments. Hence, thick limestone deposits occur when other kinds of sediments are deposited at very slow rates. Thus, the kind of sediment accumulating at any point in time and space reflects what isn't happening as much as what is; this is called the sedimentary masking effect. Additionally, this influx of mud clogs the filter-feeding apparatus and gills of many organisms, thereby critically affecting the survival of many carbonate-producing invertebrates, and consequently drastically reducing carbonate deposition (this process is called clogging).

6.1.7 Carbonate Diagenesis

Carbonate diagenesis refers to the processes that, over time, alters and transforms carbonate sediments and rocks such as limestone and dolomite that are primarily composed of calcium carbonate ($CaCO_3$) or magnesium calcium carbonate ($CaMg(CO_3)_2$), respectively. Diagenesis is the physical, chemical, and biological changes that occur in sediments or rocks after their deposition and burial and involves a variety of processes that modify the original composition, texture, and fabric of carbonate sediments and rocks (Figures 6.18A–B). These processes occur during burial and subsequent lithification (Figure 6.18C), as well as during exposure to fluids, pressure, and temperature changes. Thus, these processes are tied to continental sedimentary and meteoric environments (James and Choquette, 1990) that occur after the deposition or precipitation of a particle at the sediment/water interface, until the rocks enter into the domain of metamorphism (Figures 6.18A–B).

There are two main pathways in describing the diagenesis of continental carbonates. The first are the changes that follow a more or less continuous path from the initial deposition of the carbonates, or in some cases, siliciclastics (i.e., calcretes) and consists of progressive changes during the lithification of continental sedimentary deposits. Two domains are noted within this pathway: eogenesis (diagenetic changes that occur at or near the sediment surface), and mesogenesis (burial effects, where the pore-filling solutions are isolated from the overlying meteoric-derived waters) (Figure 6.19). The second pathway involves the effects of meteoric waters on carbonate rocks (both marine and continental), which have previously undergone diagenesis; this is the telogenetic stage, occurring after

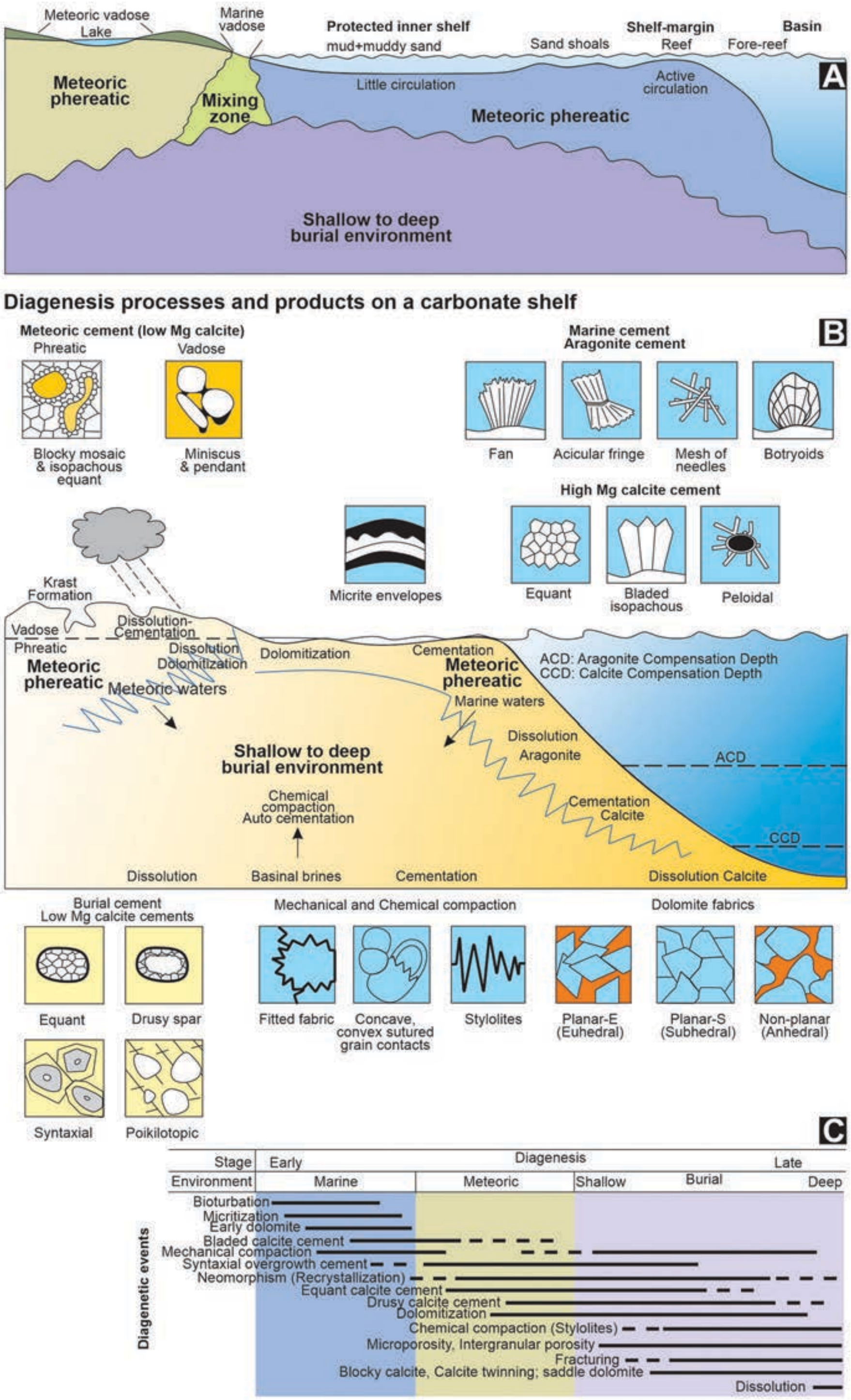

FIGURE 6.18 Carbonate diagenesis, illustrating various diagenetic environments and events. A: Carbonate diagenetic environments. B: Diagenetic processes and products on a carbonate shelf. (Modified after James and Choquette, 1990a, b; Tucker and Wright, 1990.) C: Generalized paragenetic sequence of diagenetic processes and phases.

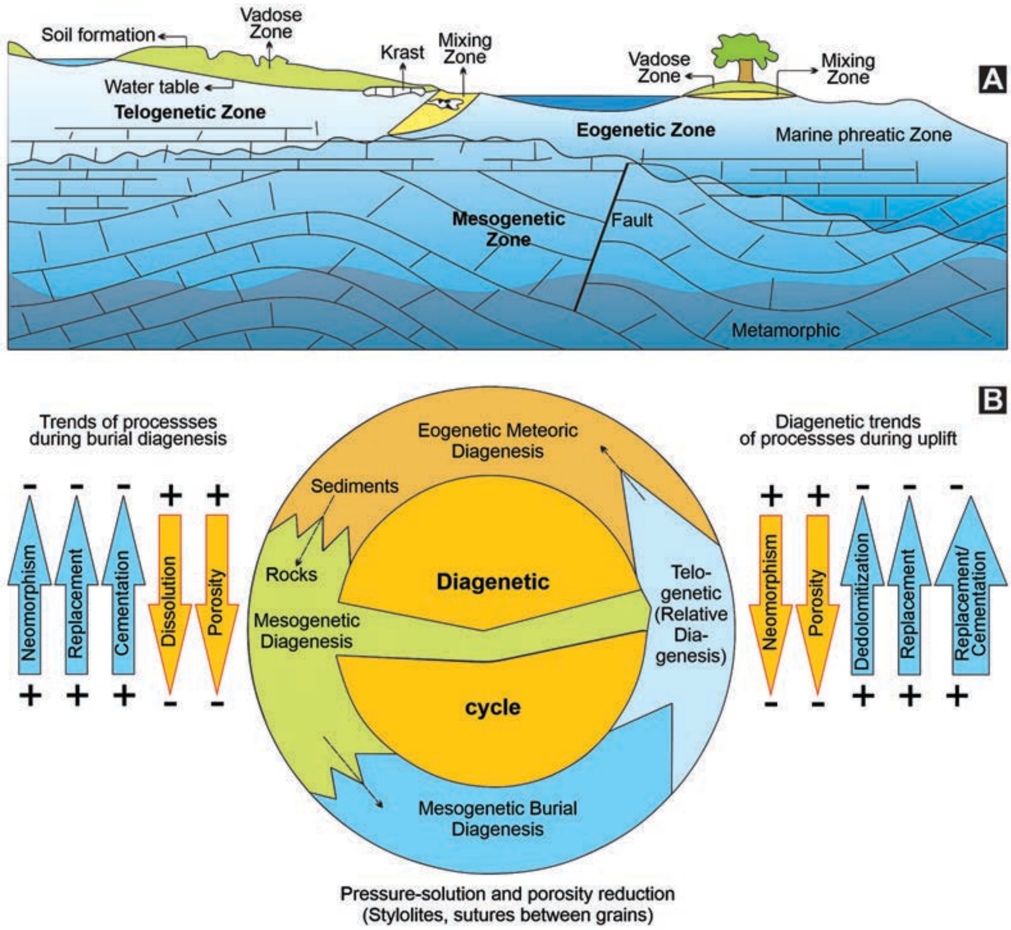

FIGURE 6.19 Relationship of carbonate diagenetic environments to zones of the burial diagenetic regime and diagenetic trends. A: Classical carbonate diagenetic subdivisions are into eogenetic, telogenetic, and mesogenetic domains (zones). (Modified from Choquette and Pray, 1970.) B: Changes in major diagenetic processes and parameters through burial diagenesis. Diagenesis is the physical, chemical, and biological changes that occur in sediments or rocks after their deposition and burial and involves a variety of processes that modify the original composition, texture, and fabric of carbonate sediments and rocks. (Modified from Armenteros, 2010.)

uplift and erosion (Figure 6.19). Pettijohn (1957) called it as the "reverse or retrograde diagenesis" as the intrusion of freshwaters produces a significant disruption of the continuity of the first diagenetic sequence, largely due to changes in the paleogeography of the depositional basin.

Some key processes involved in carbonate diagenesis include compaction, cementation, dissolution, recrystallization, dolomitization, burial and thermal alteration, and biogenic processes, among others; these are very briefly outlined here (see Figures 6.18 and 6.19; see also Adams and McKenzie, 1998). As the sediment is buried, the weight of the overlying layers causes compaction, reducing pore space and increasing the density of the sediment. Carbonate sediments are cemented by various minerals such as calcite or dolomite that fill the pore spaces and bind the grains together (cementation) (see Figure 6.18B). Carbonate minerals get dissolved in the presence of acidic fluids, such as groundwater, leading to the removal of carbonate material and the creation of pore spaces (dissolution) (see Figure 6.18B). During diagenesis, carbonate minerals undergo recrystallization, where existing minerals are dissolved and new ones are precipitated, resulting in changes in crystal size and shape (recrystallization) (see Figure 6.18B). Dolomite, a magnesium calcium carbonate mineral, forms from the alteration of limestones through the addition of magnesium-rich fluids (dolomitization) (see Figure 6.18B). With increasing burial depth, carbonate rocks experience increased pressure and temperature, which leads to the formation of new minerals, such as anhydrite or quartz, and the alteration of existing minerals (burial and thermal alteration) (see Figure 6.18B). Biological activity, such as the activities of organisms that produce carbonate skeletons or shells (e.g., corals, foraminifera), also influences carbonate diagenesis by creating porosity, altering mineralogy, and enhancing cementation (biogenic processes).

The specific sequence and intensity of these processes varies depending on factors such as the original composition of the carbonate sediment, the presence of organic matter, the availability of fluids, and the burial history of the rock (see Figure 6.19B). Thus, understanding carbonate diagenesis is important, particularly in petroleum geology, as it can affect reservoir quality and the migration of hydrocarbons. It also has implications for understanding the long-term preservation of carbonate fossils and the interpretation of past environmental conditions recorded in carbonate rocks.

6.1.7.1 Neomorphism (Calcitization and Recrystallization)

Neomorphism refers to the process of mineral alteration or replacement that occurs during diagenesis or metamorphism. It involves the transformation of one mineral into another mineral with a different crystal structure, composition, or both. The two common types of neomorphism in carbonate rocks are recrystallization and calcitization.

Recrystallization involves the dissolution and reprecipitation of carbonate minerals, resulting in the growth of new crystals. During recrystallization, the original carbonate minerals are dissolved, and the dissolved ions are transported and reprecipitated in new locations within the rock. This process leads to changes in crystal size, shape, and orientation, as well as the development of new textures (Figure 6.20A). Thus, recrystallization is the process by which crystal morphology is changed without major changes in the mineral composition (Ahr, 2008). This process is, sometimes, also called "neomorphism," a term coined by Folk (1965), who defined it to include both true recrystallization and mineralogical inversion (Ahr, 2008). True recrystallization is a change in crystal form without changes in mineralogical composition: for example, micrometer-sized crystals of calcite micrite going through transformation to millimeter-sized blocky calcite crystals (Figure 6.20A) during what Folk (1965) called aggradational neomorphism (aggrading; see Figure 6.20B). Mineralogical inversion is not strictly recrystallization; it is the process by which a metastable mineral such as aragonite or Mg-calcite undergoes both crystallographic and compositional change to become ordinary calcite (Ahr, 2008) (see Figure 6.20B).

Calcitization, on the other hand, is the process by which calcium carbonate minerals, such as calcite or aragonite, replace other minerals in a rock; this occurs when calcium-rich fluids percolate through the rock, thus dissolving the original minerals, and precipitating calcite in their place. Hence, calcitization leads to the preservation of the original fabric or texture of the rock, but with a different mineral composition.

Both calcitization and recrystallization occur during diagenesis, a process of physical and chemical changes that happens in sediment or rock after its deposition and burial. Factors such as temperature, pressure, fluid composition, and time influence the extent and nature of neomorphism. The study of neomorphism is important for understanding the diagenetic history of carbonate rocks. It also provides insights into the original mineralogy, the conditions under which the rocks formed, and the subsequent alterations that occurred over time (see Figure 6.20). This information is valuable for interpreting the depositional environment, diagenetic processes, and the potential reservoir quality of carbonate rocks in various geological settings.

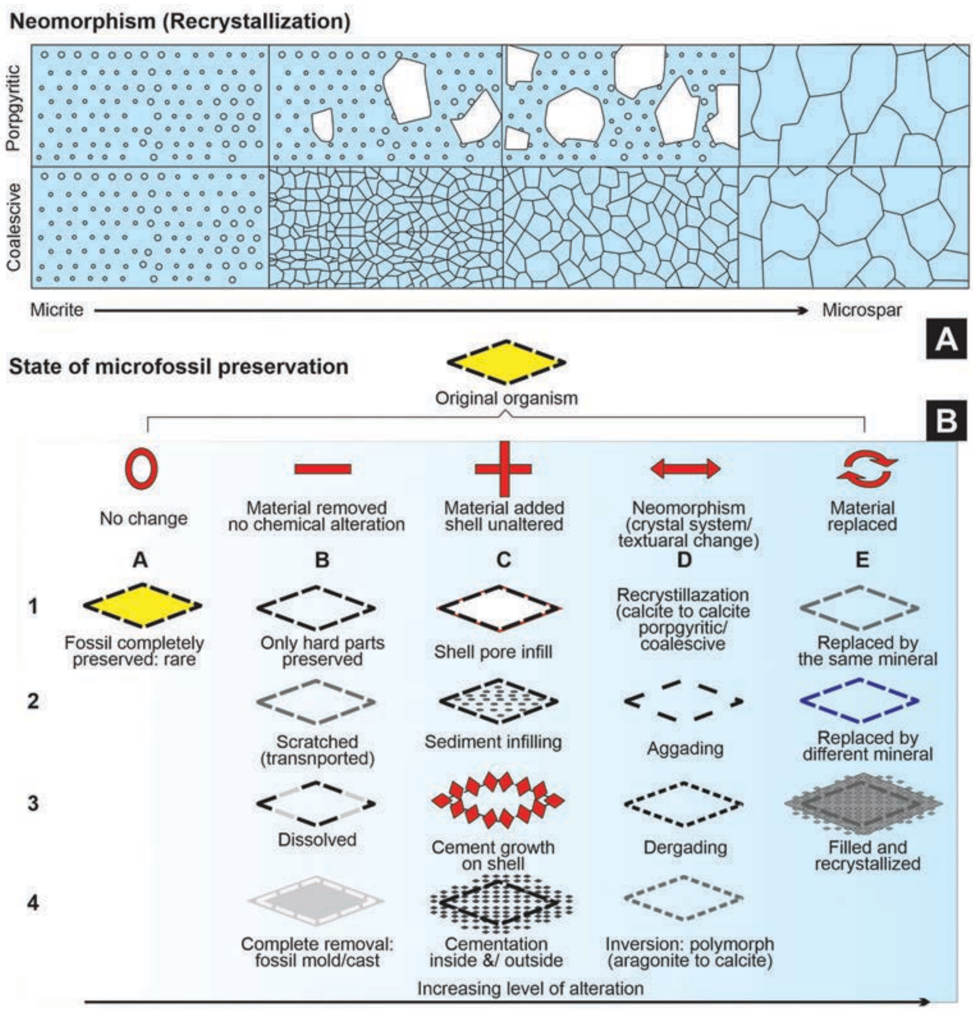

FIGURE 6.20 Neomorphism and the state of preservation. A: Neomorphism (recrystallization). (Modified from Ahr, 2008; see also Golreihan et al., 2018.) B. Illustration of the state of preservation of fossils in different conditions. Fossils can be fully preserved, including soft tissues (A), but this is very rare, occurring only upon immediate burial. Generally, changes (alterations) occur and remove materials (B), or include other materials (C), or bring changes to the crystal system by the present material, (D) or replace material by other materials that have either the same or different mineralogy (E). The level of change (alteration) increases from (A) to (E); each of these parameters are briefly enumerated. (B1) Only hard parts of the fossil are preserved without change, but the soft parts are removed. (B2) Scratches on the fossil tests are noted due to transportation, i.e., the physical removal of materials from the shell's surface. (B3) The changes in the diagenetic environment bring about chemical dissolution of the fossil. (B4) The fossil is completely removed but a mold or cast is preserved. (C1) Secondary minerals fill the pores of the test with the same or a different mineral. (C2) The normally soft sediment infilling is washed away, e.g., by ultrasonic treatment. (C3) Crystal overgrowth on the test. (C4) The fossil undergoes cementation, inside and/or outside. (D1–3) Change in crystal size or texture but preserving the same mineralogy leads to recrystallization (neomorphism). (D4) Inversion (i.e., neomorphism) occurs by changes in the crystal system or polymorph, i.e., aragonite to calcite. (E1) Replacement of the primary mineral by a secondary one; this change maintains the same mineralogy but has a different isotopic signature. (E2) A different mineral replaces the fossil test. (E3) Combination of adding material and alteration of the primary test.

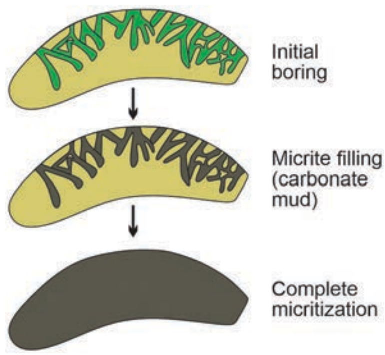

FIGURE 6.21 Micritization. This is a process of alteration of the original skeletal grain fabric to a cryptocrystalline texture, originally noted due to the repeated algal microborings and subsequent filling of the microborings by micritic precipitates.

6.1.7.2 Micritization

Micritization is a process of alteration of the original skeletal grain fabric to a cryptocrystalline texture, originally noted due to the repeated algal microborings and their subsequent filling by micritic precipitates (Bathurst, 1966) (Figure 6.21). Thus, micritization is a diagenetic process that involves the formation and accumulation of micrite, a fine-grained (<20 μm in size) carbonate sediments. Micritization occurs when carbonate mud or other carbonate particles are subjected to dissolution and reprecipitation processes. Micrite fills in pore spaces between larger grains, coat surfaces, and forms cementing materials. Micritization disrupts the original crystal structure, chemical composition and structure of the clasts in the sediment. The dissolution of larger carbonate grains releases calcium and carbonate ions into the surrounding pore waters. These ions then reprecipitate as micrite in areas with favorable conditions. Persistent micritization results in the formation of carbonate muds. Micritization can occur in a variety of environments, including marine, lacustrine, and even subaerial settings. Micritization is more common in lower-energy lagoons. Micritization is also commonly observed in carbonate rocks, such as limestone and dolomite, where it contributes to the overall texture and fabric of the rock. The presence of micrite has major implications for the porosity and permeability of carbonate rocks. Micrite is generally more compact and less porous than larger carbonate grains, leading to reduced permeability and, thus, decreased reservoir quality in petroleum reservoirs.

6.1.7.3 Cementation and Dissolution

Cementation and dissolution are two important diagenetic processes that significantly impact the properties and characteristics of sedimentary rocks (see Figure 6.20B).

Cementation is the filling of pore spaces between sediment grains with mineral cements (Figure 6.20B). The cements can be of various minerals, such as calcite, quartz, or clay minerals that are typically precipitated from fluids that flow through the rock (Figure 6.22). Cementation occurs during or after sediment deposition and is influenced by factors such as fluid composition, temperature, pressure, and the availability of dissolved ions. It significantly influences porosity and permeability of a sedimentary rock. As pore spaces are filled with cement, the interconnectedness of the pores decreases, reducing the permeability of the rock (Figure 6.22). But cementation also results in the lithification and strengthening of the rock, thus increasing its overall durability.

Dissolution, on the other hand, involves the removal or dissolution of minerals from the rock matrix (see Figure 6.20B). This process occurs when fluids, often acidic in nature, come into contact with the rock and dissolve certain minerals. In the case of carbonate rocks, the primary mineral that

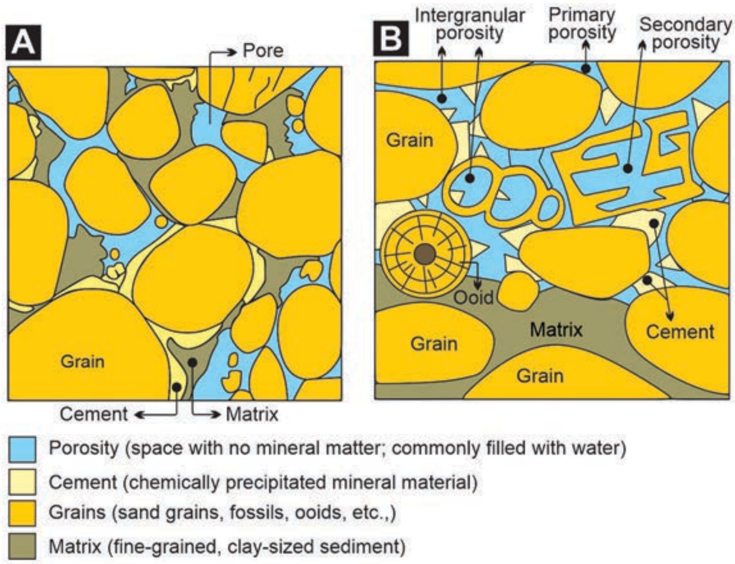

FIGURE 6.22 Types of cement. The cements can be of various minerals, such as calcite, quartz, or clay minerals that are typically precipitated from fluids that flow through the rock. A–B: The figures illustrate four fundamental parameters: grains, matrix (mud-sized sediments between the grains; it is primary, if deposited at the same time or soon after grains, or secondary, if formed by the diagenetic alteration of grains, cement (these are the chemical precipitates in pore spaces), and pore space (it could be primary or secondary).

is susceptible to dissolution is calcium carbonate (calcite or aragonite). Dissolution occurs in various environments, such as groundwater systems, oceans, or during burial diagenesis. Dissolution also leads to the creation of pore spaces and the development of secondary porosity within the rock (Figure 6.22B). This can increase the permeability of the rock and enhance fluid flow. On a larger scale, dissolution can also result in the formation of caves, sinkholes, and karst landscapes, where extensive dissolution has occurred.

The balance between cementation and dissolution processes in a sedimentary rock is crucial in determining its overall porosity, permeability, and reservoir quality (see Figure 6.20. Thus, understanding the factors that control these processes is important for evaluating the potential of a rock as a hydrocarbon reservoir or as a host for groundwater resources. Additionally, these processes provide valuable information about the diagenetic history and environmental conditions that the rock has experienced over time.

6.1.7.4 Replacement

Replacement refers to a diagenetic process where one mineral is replaced by another while maintaining the original shape or structure of the replaced mineral (see Figure 6.20B). This process occurs when fluids carrying dissolved minerals infiltrate a rock, causing the dissolution of the original mineral and the subsequent precipitation of a new mineral in its place. The replacement process occurs through several mechanisms, including ion-by-ion substitution, pseudomorphism, and metasomatism. Ion-by-ion substitution involves the exchange of ions between the original mineral and the infiltrating fluid, resulting in the gradual replacement of the original mineral's composition by the new mineral. Pseudomorphism, on the other hand, occurs when the new mineral grows within the crystal structure of the original mineral, preserving its external shape but with a different internal composition. Metasomatism refers to the complete replacement of the original mineral by a new mineral, often resulting in significant changes in the rock's chemical and mineralogical composition. Replacement can occur in various geological settings, such as hydrothermal systems, metamorphic environments, and sedimentary basins. Replacement is influenced by several factors such as temperature, pressure, fluid composition, and the availability of dissolved ions. The minerals involved in replacement vary widely, depending on specific geological conditions. For example, in hydrothermal systems, sulfide minerals like pyrite are replaced by chalcopyrite or bornite. Thus, understanding replacement is important for inferring the diagenetic history of rocks, and the processes and conditions that have influenced their mineralogical and chemical composition. Replacement has significant implications for the properties and characteristics of rocks, such as their porosity, permeability, and reservoir potential as well as the formation of ore deposits and the distribution of economic minerals, at large.

6.1.7.5 Stylolites

The term stylolite comes from the Greek words *stylos* (pillar) and *lithos* (stone); it was coined by Klöden (1828). Stylolites are geological features that occur in sedimentary rocks, particularly in carbonates such as in limestones and marbles (Stockdale, 1922, 1926, Bathurst, 1971; Rolland et al., 2012, 2014), cherts (Bushinskiy, 1961; Cox and Whitford-Stark, 1987), coal (Stutzer, 1940), and sandstones (Tarr, 1916; Stone and Siever, 1996). Stylolites are irregular, often wavy or serrated, surfaces that form due to pressure solution, a process in which minerals are dissolved under pressure (Figure 6.23). Stylolites show large teeth-like structures, a few mm to cm in size, pointing in the direction of largest compressive stress during their formation (Koehn et al., 2007) (see Figure 6.23A). They are found in limestones and dolomites, and are structures commonly found on planes of movement between beds, on contacts of chert and limestone, and along filled sinks and slumps in limestones (Bretz1, 1940, 1950). They bear close resemblance to stylolites.

Stylolites form when two adjacent layers of rock are subjected to pressure, causing the dissolution of minerals at the contact surface. This dissolution occurs due to the pressure exerted by the overlying rock (Figure 6.23A) or by tectonic forces (Figure 6.23B). Sorby (1862) was the first to attribute them to pressure solution, by describing them as "curious teeth-like projections with which one bed of limestone enters into another" (Figure 6.23 A). Pressure solution and cleavage are also observed in shales (Wright and Platt, 1982; Rutter, 1983). The dissolution of minerals at the contact surface leads to the creation of irregular, rough surfaces called stylolites (Figure 6.23A). These surfaces extend over varying distances within the rock, depending on the intensity and duration of the pressure. Stylolites have a jagged appearance, with teeth-like projections that fit together like puzzle pieces (Figure 6.23A). The minerals that are most commonly dissolved are calcite and dolomite, which are the main constituents of carbonate rocks. Stylolites are divided into three types, sedimentary stylolites (Figure 6.23A), tectonic stylolites (Figure 6.23B), and slickolites (Figure 6.23C).

Sedimentary stylolites are sub-parallel to bedding, i.e., horizontal at the time of formation (Figure 6.23A). Tectonic stylolites are perpendicular to the largest compressive principal stress axis, which can be horizontal, leading to vertical stylolites (Figure 6.23B). Slickolites develop on planes that are oblique to the largest principal stress direction (Figure 6.23C).

Studies suggest that stress plays a major role in the stylolite formation: stylolitic teeth direction is parallel to maximum compression (Stockdale, 1922; Eyal and Reches, 1983). However, stylolites have been noted to form at very shallow burial depths of tens of meters with very low stresses, also (Tada 1989). Several other studies attribute stylolite formation to textural (sharp changes in grain size and texture) and mineralogical (such as clays) heterogeneities (Rustichelli et al., 2012, 2015). Stylolites have significant effects on the properties of carbonate rocks. They act as pathways for fluid flow, enhancing permeability and facilitating the movement of fluids through the rock. Stylolites also contribute to the development of secondary porosity, as the dissolution of minerals creates void spaces within the rock. Thus, the study of stylolites is important for understanding the diagenetic history of carbonate rocks, as well as for interpreting the processes and conditions that might have

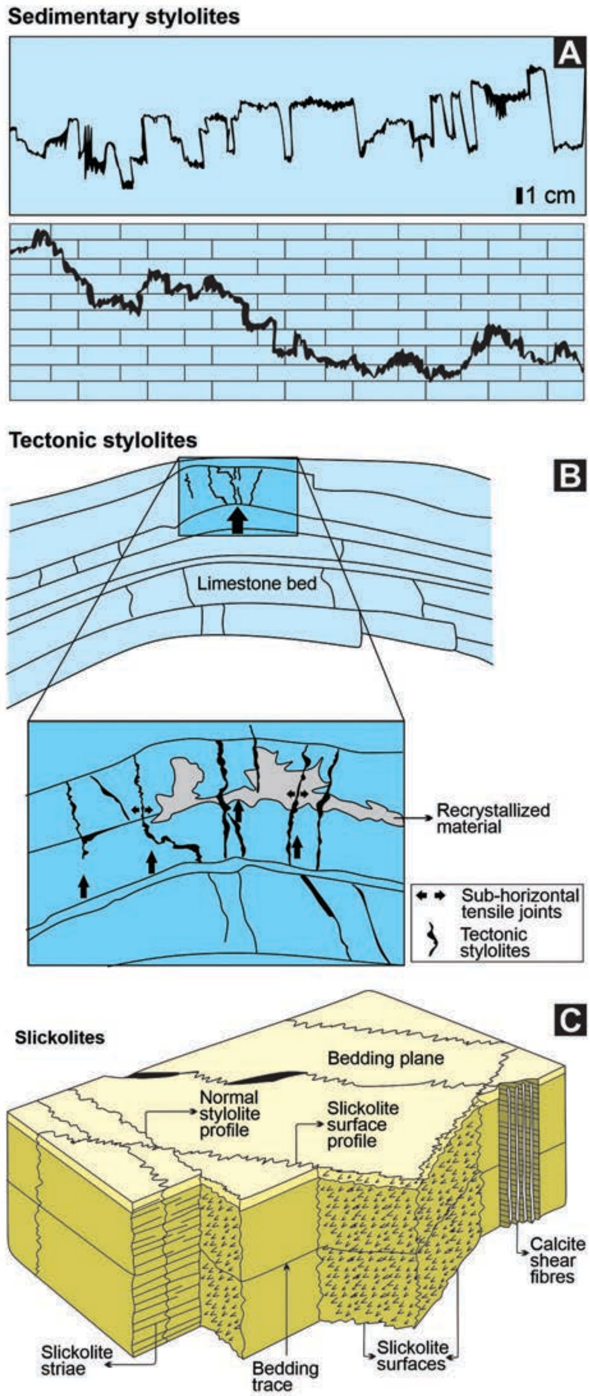

FIGURE 6.23 Types of stylolites. A: Stylolites have a jagged appearance, with teeth-like projections that fit together like puzzle pieces. The dissolution of minerals at the contact surface leads to the creation of such irregular, rough surfaces called stylolites. Stylolites are divided into three types: sedimentary stylolites (A), tectonic stylolites (B), and slickolites (C).

influenced their formation. Stylolites provide valuable information about the stress and pressure conditions that the rock has experienced, as well as the fluid flow patterns that might have occurred. They are also used as indicators of burial history and tectonic activity in sedimentary basins.

6.1.7.6 Dolomitization

Dolomitization is a diagenetic process in which calcium carbonate ($CaCO_3$) minerals, typically calcite, are replaced by dolomite, a mineral composed of calcium magnesium carbonate ($CaMg(CO_3)_2$). There are several mechanisms by which dolomitization occurs. A common mechanism is through the replacement of existing calcite by magnesium-rich fluids. This happens when magnesium-rich fluids infiltrate a carbonate rock and reacts with calcite, causing it to dissolve and be replaced by dolomite. Another mechanism involves the direct precipitation of dolomite from magnesium-rich fluids, bypassing the dissolution of calcite. Dolomitization results in significant changes to the mineralogy and fabric of the rock. Dolomite crystals often have a rhombohedral shape and are coarser-grained than the original calcite. The replacement of calcite by dolomite also leads to a decrease in porosity and permeability, as dolomite is less soluble than calcite. However, under certain conditions, dolomitization enhances porosity and creates secondary pore spaces, particularly if the dolomite crystals are well connected and have a fabric that promotes fluid flow. Dolomitization occurs in various geological environments, including marine and sabkha (coastal saline), and during burial diagenesis. Factors that influence the occurrence and extent of dolomitization include temperature, pressure, fluid composition, and the availability of magnesium ions. Thus, understanding dolomitization is important for inferring the diagenetic history of carbonate rocks, and for evaluating their reservoir potential in hydrocarbon exploration.

6.1.7.7 Dedolomitization

Dedolomitization is a diagenetic process that involves the partial or complete replacement of dolomite by other minerals, typically calcite. Dedolomitization typically occurs when fluids permeate a dolomite-rich rock and react with the dolomite, causing it to dissolve and be replaced by calcite. Thus, the dedolomitization process is divided into two steps: (a) the dissolution of dolomite, and (b) the precipitation of calcite. Dedolomitization is largely driven by the availability of calcium ions in the infiltrating fluid that reacts with the magnesium ions in the dolomite. As a result, the dolomite crystals are gradually dissolved, and calcite crystals grow in their place. This replacement of dolomite by calcite leads to significant changes in the mineralogical composition and fabric of the rock. Calcite crystals often have a rhombohedral shape and are coarser-grained as compared to the original dolomite. Dedolomitization also results in changes in the porosity and permeability of the rock, as calcite is more soluble than dolomite.

6.1.7.8 Silicification

Silicification is a process by which silica or SiO_2 (silicon dioxide) is deposited or replaced in an organism or mineral, resulting in the formation of a siliceous material. Thus, silicification process is a replacement of the original skeletal material enabled by the simultaneous dissolution of calcium carbonate and the precipitation of silica (Butts, 2014). This process occurs in various environments, such as marine sediments, hot springs, or volcanic ash deposits. In the context of organisms, silicification occurs in both plants and animals. For example, some plants, like grasses and horsetails, have silica in their cell walls, providing structural support. In some cases, the silica accumulates and forms opal or chert. In animals, silicification occurs in the form of fossilization (via permineralization, entombment, and replacement). Permineralization is the precipitation of silica within the natural cavities of porous materials, such as wood and bone, from silica-enriched percolating fluids. Silica entombment occurs in hydrothermal settings by the precipitation of silica on the external surfaces of organic objects. In general, when an organism dies and is buried, the silica-rich groundwater infiltrates tissues and replaces the organic material with silica, preserving the organism's original structure. This process results in the formation of petrified wood or fossilized shells. Silicification also occurs in minerals, where silica is deposited or is replaced in the crystal lattice of existing minerals resulting in the formation of various silicate minerals, such as quartz, opal, or chalcedony. Factors that control silicification are those that influence the dissolution/precipitation process: shell mineralogy, shell ultrastructure (and, therefore, surface area), the amount and location of organic matter, and the character of the enclosing matrix. Silicification occurs in deposits ranging from supratidal to basinal (Knoll, 1985).

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7 Chemical/Biochemical and Carbonaceous Sedimentary Rocks

7.1 EVAPORITES

7.1.1 Introduction

Evaporites (Figure 7.1) are layered crystalline sedimentary rocks that precipitate from brines in areas where the amount of water lost by evaporation exceeds inflow (inflow is the total amount of water from rainfall and influx via rivers and streams), and where the brine concentration process is driven by solar evaporation (Warren, 2006). Thus, evaporite formation needs three prerequisites: (a) a surface or near surface brine body that is saline enough to precipitate and preserve salt, (b) sufficient accommodation space within a sedimentary depression, and (3) a burial environment that does not allow undersaturated porewater throughflow to dissolve the buried salts. Evaporites occur in rocks of most ages, but are particularly common in Cambrian, Permian, Jurassic and Miocene successions (Ronov et al., 1980) (see Figure 7.2). Some individual evaporite deposits, such as the Miocene Messinian of the Mediterranean region (Figure 7.2A), reach thicknesses exceeding ~3 km (see Hsü, 1973; Meilijson et al., 2019).

The mineralogy of evaporite rocks is complex; around 80 minerals have been reported (Stewart, 1963; Warren, 1999), but less than a dozen are volumetrically important (Table 7.1). These include carbonates (such as calcite, dolomite, magnesite, and aragonite), sulfates (anhydrite and gypsum), and chlorides (particularly halite, sylvite, and carnallite), as well as various borates, silicates, nitrates, and sulfo-carbonates (Table 7.1). Halite, due to its volume and mobility, overshadows all other evaporite minerals and forms a group on its own. Those containing potassium and magnesium salts are called bittern salts and are lumped together by the industrial term, Potash (Table 7.1).

7.1.1.1 Marine Evaporites

Evaporites form under both marine and nonmarine conditions. The marine evaporites tend to be thicker and more laterally extensive than the non-marine ones; marine evaporites are also of greater geologic significance. The marine evaporite minerals are grouped by their chemical composition into chlorides, sulfates, and carbonates (see Table 7.1). If carbonate minerals, most of which are not evaporites, are excluded, the most common minerals in marine evaporites are the calcium sulfate minerals, gypsum and anhydrite. Halite is next in abundance, followed by potash salts, sylvite, carnellite, langbeinite, polyhalite, and kainite, and the magnesium sulfate, kieserite (see Table 7.1). Marine evaporites commonly contain mixtures of minerals, although gypsum (or anhydrite) and halite predominate. In modem evaporite deposits, gypsum is more abundant than anhydrite, but in ancient ones, anhydrite dominates, largely owing to the diagenetic alteration of gypsum to anhydrite. Marine evaporites may also contain various amounts of impurities such as clay minerals, quartz, feldspar, and sulfur. Evaporite deposits in modern marine environments are largely restricted to

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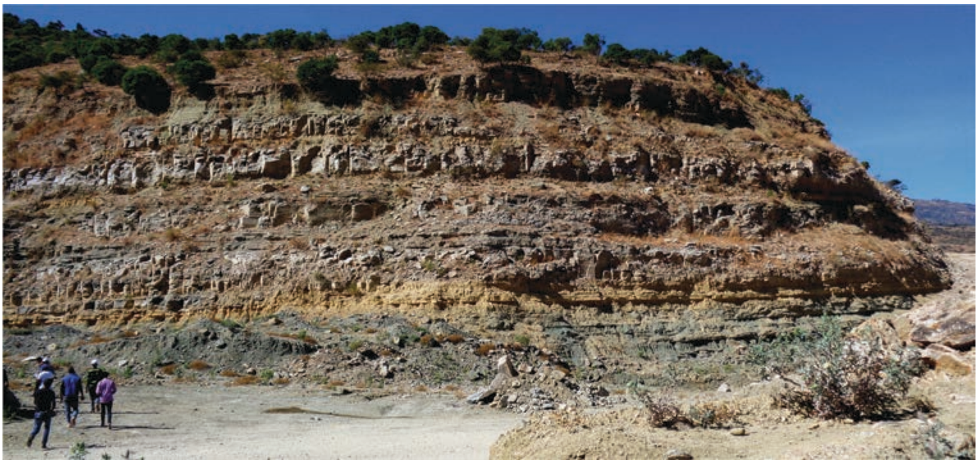


FIGURE 7.1 Gypsum beds intercalated with dolostones (yellow), Bathonian Gohatsion Formation, Jurassic of Ethiopia, exposed at Mugher. (Modified after Jain, 2019.)

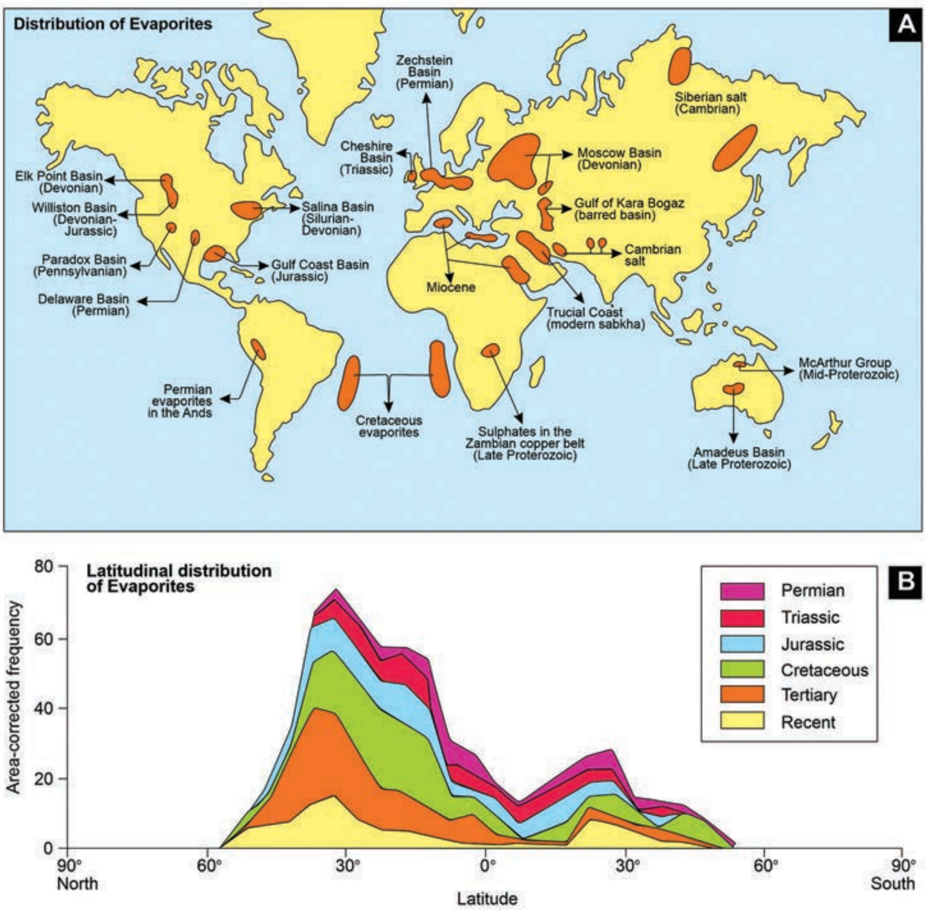


FIGURE 7.2 Distribution of evaporates. (Modified after Kendall 1992; Warren, 2016.) A: Age-based present-day geographic distribution of evaporites. B: Latitudinal distribution of evaporites through time.

TABLE 7.1
Properties of selected evaporites

Mineral	Class	Formula	Color	Taste	Density (kg/m³)	Mohs hardness
Halite	Chloride	$NaCl$	colorless, white	salty	2.04	2–2.5
Anhydrite	Sulfate	$CaSO_4$	white, gray, bluish, red	none	2.98	3–3.6
Gypsum	Sulfate	$CaSO_4.2H_2O$	white, gray, ochre, pink	none	2.35	2
Kieserite	Sulfate	$MgSO_4.H_2O$	colorless, white, gray, yellow	none	2.59	3–3.6
Epsomite	Sulfate	$MgSO_4.7H_2O$	white, reddish	sharp	1.71	2–2.5
Bischofite	-	$MgCl_2.6H_2O$	colorless, white	very bitter	1.56	1.5
Tachyhydrite	-	$2MgCl_2. CaCl_2.2H_2O$	yellow	sharp, bitter	1.70	1–2
			Potash salt			
Langbeinite	Sulfate	$K_2SO_4.MgSO_4$	colorless, white, pink	none	2.82	3–4
Polyhalite	Sulfate	$K_2MgCa_2 (SO_4)_4.2H_2O$	red, yellow, gray	none	2.79	2.5–3.5
Kainite	Sulfate	$4(KClMgSO_4).11H_2O$	white, red, yellow	salty, slightly bitter	2.12	2.5–3
Sylvite	Chloride	KCl	white, yellowish red	salty, bitter	1.86	2–2.2
Carnallite	Chloride	$KCl.MgCl_2. 6H_2O$	white, red, yellow	bitter	1.57	1–2.5

coastal regions, such as evaporitic lagoons and sabkha mudflats. However, evaporite successions within the stratigraphic record indicate that precipitation of evaporite minerals has at times occurred in more extensive marine settings (Warren, 1999, 2016).

7.1.1.2 Non-Marine Evaporites

Non-marine evaporites have minerals that are uncommon in marine evaporites as the water from which they precipitate, has more bicarbonate and magnesium with little or no chlorine. These non-marine minerals include bloedite ($Na_2S0_4.MgS0_4.4H_20$), borax ($Na_2B_4O_5(OH)_4.8H_20$], epsomite ($MgSO_4.7H_2O$), gaylussite ($Na_2CO_3.CaCO_3.SH_2O$), glauberite [$Na_2Ca(SO_4)$], magadiite ($NaSi_7O_{13}(OH)_3.3H_2O$), mirabilite ($Na_2SO_4.10H_2O$), thenardite ($NaS0_4$), and trona [$Na_3H(C0_3)_2.2H_2O$). However, the non-marine deposits also contain anhydrite gypsum, and halite and may well be dominated by these (see Smoot and Lowenstein, 1991). Another feature, beside mineral content, that identifies them is their pattern of deposited lacustrine minerals, inside out, called the bull's-eye pattern (Figures 7.3A–B; detailed below). In these bands of lacustrine evaporites, the most soluble mineral, halite, is concentrated near the center of the lake and occurs at the top of the evaporite sequence, whereas the least soluble minerals, carbonates, and sulfates, are concentrated around the edges of the lake and are at the bottom of the evaporite sequence (see Figures 7.3A–B). The non-marine evaporite deposits are commonly formed in modern arid closed basins, but few have been identified in the geologic record. The non-marine evaporites are important as their depositional setting, reflecting both climatic and tectonic conditions; they are also economically important mineral resources.

7.1.1.3 Lacustrine Evaporites

Most modern non-marine evaporites and almost all ancient ones occur in saline lake deposits or their associated facies. The term saline lake includes perennial, ephemeral, playa, salina, salar, chott, salt pan, inland sabkha, and any other terms used to refer to areas that at least intermittently hold standing bodies of water.

Saline lakes also hold minerals in large concentrations. However, the evaporite mineralogy of salt lakes, that depends largely on the local geology, varies considerably from one region to the other. A zonation of evaporite minerals is noted within a salt lake, with the least soluble minerals occurring around the edge and the most soluble ones being precipitated in the center of the lake, thus displaying a bull's-eye pattern, as mentioned above (see Figs 7. 3A–B). In a large barred basin with near-continuous inflow of seawater, the distribution of evaporites is of the teardrop-type pattern, with carbonates located near the barrier, gypsum in the central part, and halite-potash salts in the most distal end, the hypersaline peripheral part (Figures 7.3C–D). This arrangement contrasts with a basin totally cut off from the open ocean, where a bull's-eye pattern develops, with the most soluble evaporites deposited in the basin's center (Figures 7.3A–B).

In general, the minerals are precipitated out of a solution, as ions become more concentrated, and as the water evaporates (Figure 7.3E). On average, the seawater contains 35 gL^{-1} (35 parts per thousand) of dissolved ions that are mainly chloride, sodium, sulfate, magnesium, calcium, and potassium. The least soluble compounds are precipitated out of seawater, early on, so that calcium carbonate (such as calcite) is the first to be precipitated, and as the waters become more concentrated by calcium sulfate and sodium chloride (see Figure 7.3E). Potassium and magnesium chlorides only precipitate when the seawater has become very concentrated, i.e., where the volume of water is 5% (see Figure 7.3E). Salt lakes are commonly surrounded by inland sabkhas and saline mudflats. Saline pans are common in deserts. In terms of evaporite mineralogy of salt lakes, halite dominates, as in the Dead Sea and the Great Salt Lake (Utah, USA), and sodium carbonate and sulfate minerals dominate in Mono Lake (California, USA) and Carson Lake (Nevada, USA).

7.1.2 Kinds of Evaporites

7.1.2.1 Anhydrite

Calcium sulfates are deposited dominantly as gypsum, which can be altered into, and be pseudomorphed by anhydrite while the sediments are still in their general depositional environments. Gypsum is also dehydrated to anhydrite upon burial to a few hundred meters; the loss of water is accompanied by a 38% decrease in the solid volume of the gypsum. Due to this rapid dehydration with burial, most ancient calcium sulfate deposits are composed of anhydrite. Anhydrite can be hydrated back to gypsum upon uplift and exposure to low-salinity surface waters, with an accompanying increase in volume, thus distorting original depositional structures and textures; many calcium sulfate deposits are characterized by such distorted fabrics (Figures 7.4A–B). Irregular and contorted layers of gypsum form the so-called enterolithic texture (Figures 7.4A–B). Anhydrite is harder (hardness 3.5) and denser than gypsum; it is commonly white in the hand specimen, and is not easily scratched by a fingernail. On the basis of fabric, bedding, and the presence or absence of distortion, three fundamental structural types of anhydrites are recognized, nodular, laminated and massive.

7.1.2.1.1 Nodular Anhydrites and Chicken-Wire Structures

Nodular anhydrites are irregularly shaped lumps that are partly or completely separated from each other by a salt or carbonate matrix (Figures 7.4C–D). The term chicken-wire structure (common in many ancient sulfate deposits) is used for a particular type of nodular anhydrite that consists of slightly elongated, irregular polygonal masses of anhydrite separated by thin dark stringers of other minerals such as carbonate or clay minerals (Figures 7.4C–D). The formation of nodular anhydrite is initiated by the displacive growth of gypsum in carbonate or clayey sediments. Gypsum crystals subsequently alter to anhydrite pseudomorphs, which continue to enlarge by the addition of Ca^{2+} and SO_4^{2-} from an external source, and ultimately grow displacively into anhydrite nodules. Nodular anhydrites are noted in many modern coastal sabkha environments (shallow water: supratidal and intertidal flats) (see Figure 7.4E), but they can also form in deeper waters. The nodular and enterolithic textures are typical of a marine sabkha environment where other peritidal sediments are interbedded (such as microbial laminites/stromatolites, fenestral lime mudstones/dismicrites), or in a continental sabkha, where fluvial and aeolian sediments are associated. Anhydrite, gypsum, and dolomite are typical sabkha minerals.

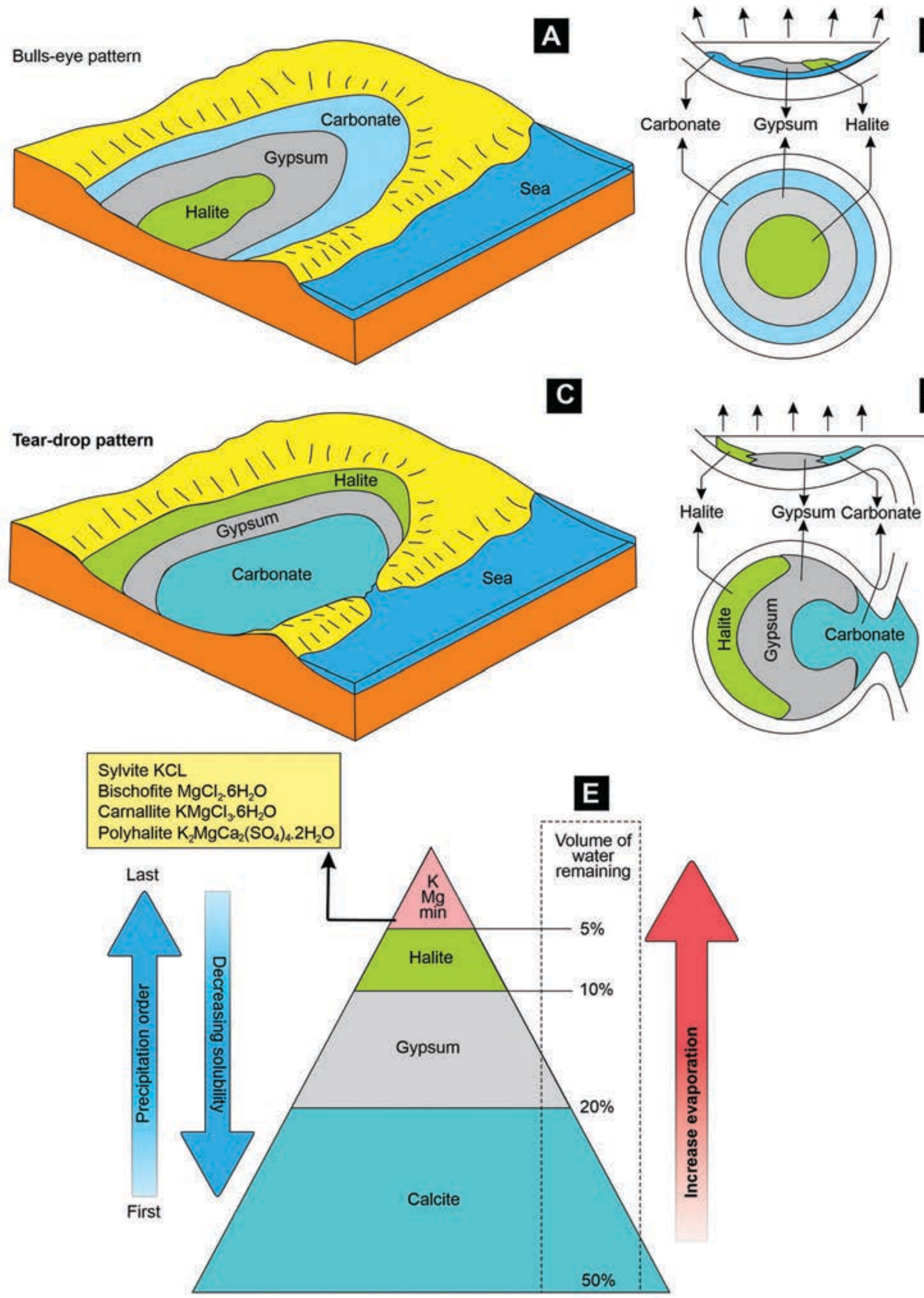


FIGURE 7.3 Models of evaporite deposition. A–B: Bulls-eye pattern. Most soluble salts in the center of the basin, typical of completely enclosed basins. Evaporation happens of the entire water body in an isolated basin. C–D: Teardrop pattern. Most soluble salts occur farthest away from the basin entrance. Typical of restricted basins with near-permanent connection to the open ocean. Evaporation increases across the semi-isolated basin. (A–D: Modified after Nichols, 2009.) E: Evaporation sequence. The triangular diagram shows the order in which evaporites precipitate in laboratory conditions. Based on percentage of evaporation of solution, the order of evaporite formation is as follows: 1. calcite (50% of water evaporation), 2. gypsum (80% of water evaporation), 3. halite (90% of water evaporation), and 4. potassium and magnesium salts (95% of water evaporation). The least soluble compounds precipitate first.

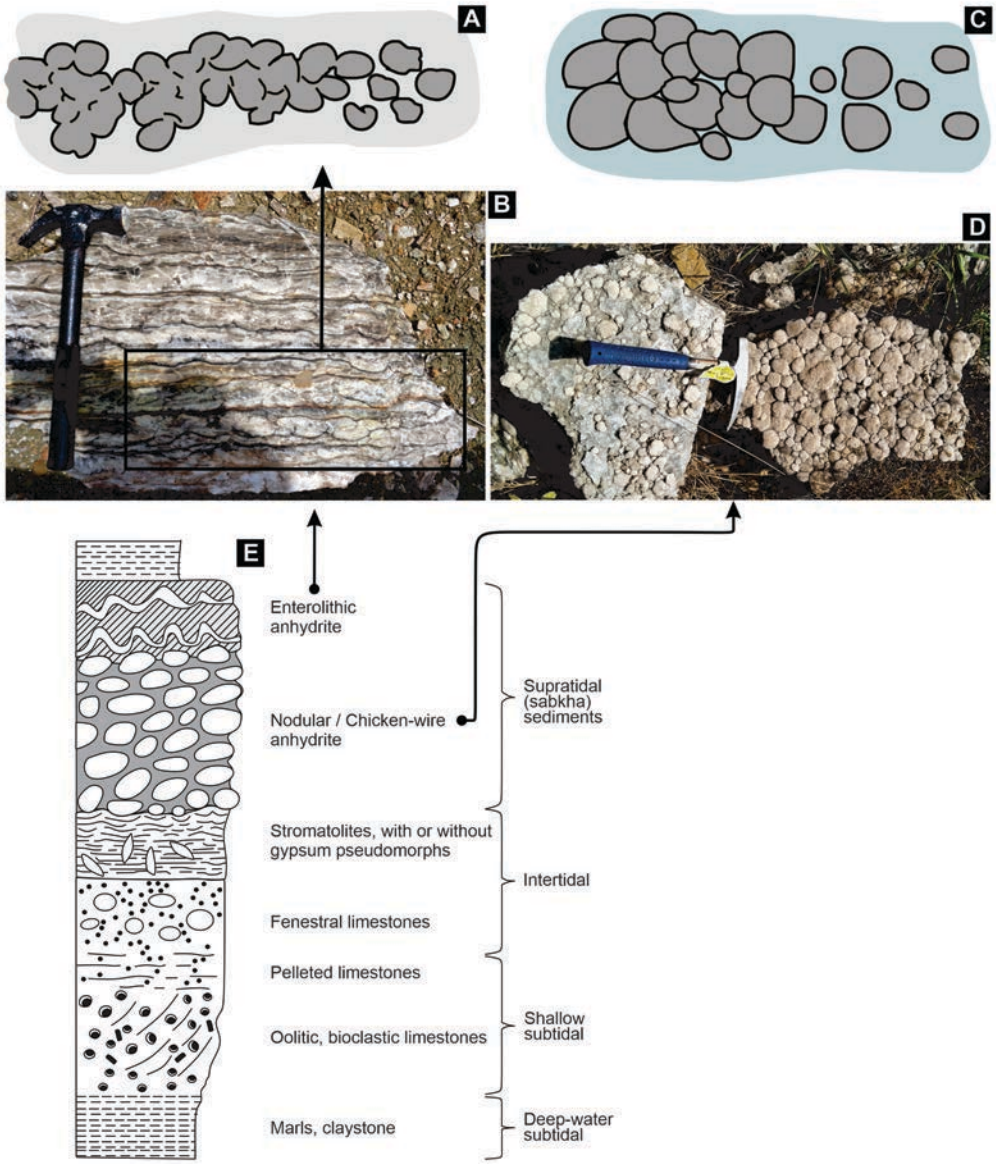


FIGURE 7.4 Enterolithic and chicken-wire structures. Bathonian Gohatsion Formation, Jurassic of Ethiopia, exposed at Mugher (after Jain, 2019). A–B: Enterolithic structures (enterolithic folds); irregular and contorted layers of gypsum. C–D: Chicken-wire structures; slightly elongated, irregular polygonal masses of anhydrite separated by thin dark stringers of carbonate or clay minerals. Hammer length is 33 cm. E: Typical vertical cycle of sabkha sediments. Such cycles range from several tens of meters in thickness. (Modified after Tucker, 1991.)

7.1.2.1.2 Laminated Anhydrites

These consist of thin, nearly white, anhydrite or gypsum laminations that alternate with dark gray or black laminae rich in dolomite or organic matter (Figure 7.5). These laminae are commonly only a few millimeters thick and rarely reach 1cm (Figure 7.5). Many of the thin laminae are remarkably uniform, with sharp planar contacts (Figure 7.5). They may constitute vertical successions hundreds

FIGURE 7.5 Laminated anhydrite. Bathonian Gohatsion Formation, Jurassic of Ethiopia, exposed at Mugher. (Modified after Jain, 2019.)

of meters thick. The laminae of anhydrite can also alternate with thicker layers of halite, producing laminated halite. Because of the lateral persistence of laminated evaporites, suggesting uniform depositional conditions over a wide area, the laminites are commonly interpreted to have been formed by the precipitation of evaporites in quiet water conditions, below wave-base. The laminated anhydrites could also form either in a shallow-water area protected from strong bottom currents and wave agitation, or in a deep-water environment.

7.1.2.1.3 Massive Anhydrite

Massive anhydrite is anhydrite that lacks perceptible internal structures. True massive anhydrite appears to be less common than nodular and laminated ones. Massive anhydrite represents sustained, uniform conditions of deposition. Haney and Briggs (1964) suggested that massive anhydrite forms by evaporation at brine salinities of approximately 200 to 275‰ (parts per thousand), just below the salinities at which halite begins to precipitate (the seawater has an average salinity of 35‰).

7.1.2.2 Gypsum

Gypsum is the hydrous form of calcium sulfate ($CaSO_4.2H_2O$); on burial, it may become dehydrated to anhydrite. Gypsum precipitates at the surface under the most arid conditions. Anhydrite has no water in its crystal structure ($CaSO_4$) and forms either by direct precipitation in arid shorelines or as a result of alteration of gypsum by burial; it becomes hydrated to gypsum if water is introduced. Primary gypsum occurs as elongate crystals of selenite when it forms from the precipitation out of water. If it forms as a result of the rehydration of anhydrite, it has a fine crystalline form as the nodules of alabaster (see Shearman and Fuller, 1969; Warren and Kendall, 1985) (Figure 7.6). In secondary veins, gypsum also occurs in a fibrous form. Gypsum can be reworked by waves and storms to form gypsarenite, which displays current structures. Most ancient gypsum exposed at the surface is actually secondary gypsum (alabastrine gypsum with centimeter-sized crystals) formed by the replacement of anhydrite or primary gypsum.

7.1.2.3 Halite

Halite (NaCl) precipitates out of seawater once it has been concentrated to 10% of its original volume (Figure 7.3E). Halite forms as crusts; in shallow waters as finely laminated deposits that may reach thicknesses of as much as 1000 m. The high solubility of sodium chloride (Figure 7.3E)

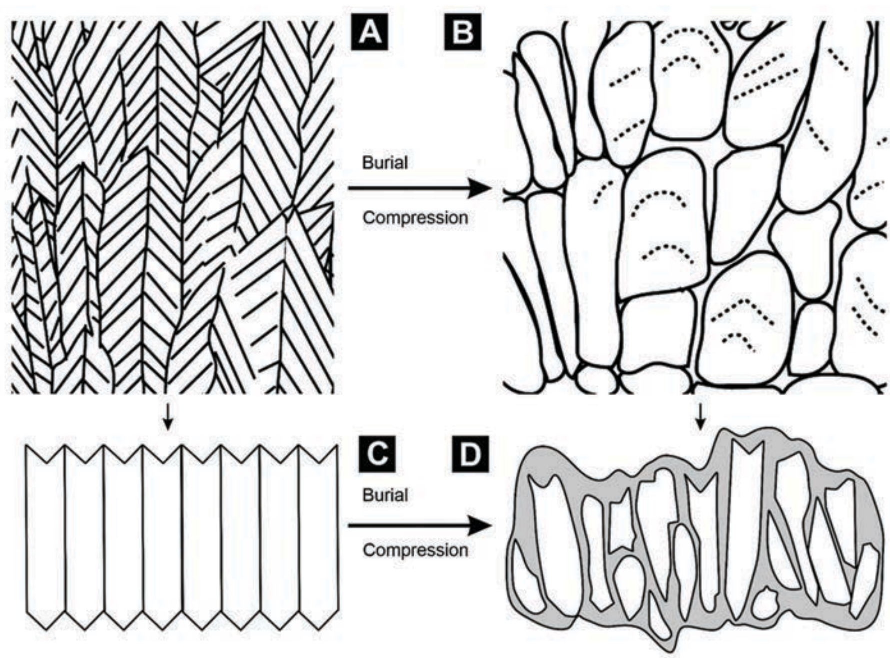


FIGURE 7.6 Characteristics of anhydrite. A–B: Preservation of gypsum ghosts as elongated anhydrite nodules created during burial. (Modified after Shearman and Fuller, 1969; Warren and Kendall, 1985.) A: Growth-aligned gypsum. B: Elongated anhydrite nodules ("ghost gypsum"). C: Vertical growth of bottom-nucleated selenitic gypsum. D: Replacement of gypsum by anhydrite during burial and compression forming "ghost gypsum."

means that it is only preserved in rocks in the absence of groundwater, which would otherwise dissolve it. Surface exposures of halite can be found in some arid regions where it has not been removed by rainwater. Halite deposited in relatively deep water (below wave-base) is typically well-bedded and laminated. Laminated halite deposits such as the Triassic salt deposits of Western Europe, commonly include anhydrite-carbonate laminae. Anhydrite along with other minerals such as dolomite, calcite, quartz, and clay may also be present in halite as inclusions. Halite deposits may also display sedimentary structures such as ripples and cross-bedding. Naturally occurring halite is rock salt; sylvite, a potassium chloride mineral, with which halite can be confused, has a more bitter taste than "normal salt" and is much less common. Halite is the major component of large evaporite-basin fills, and the main evaporite mineral of modern salt lakes and saline pans.

7.1.3 Textures of Evaporites

Many of the evaporite deposits, due to their high solubility and susceptibility to deformation are either diagenetically or secondarily altered during burial. Hence, very few "primary" evaporite beds older than ~25 Ma are known (Warren, 1999, 2006, 2016). Thus, evaporite deposits display textures that range from primary depositional features to diagenetically produced features, depending upon their ages and deformation histories (Figure 7.7) (Warren, 2016). Many ancient (subsurface) evaporites have undergone physical and geochemical diagenetic modifications or have undergone partial degrees of dissolution and fractional recrystallization, thus, altering the primary textures to secondary ones such as nodules and crystal pseudomorphs (Figure 7.7) (Warren, 2016). Primary features include a variety of crystal textures and bedding features, reflecting chemical precipitation

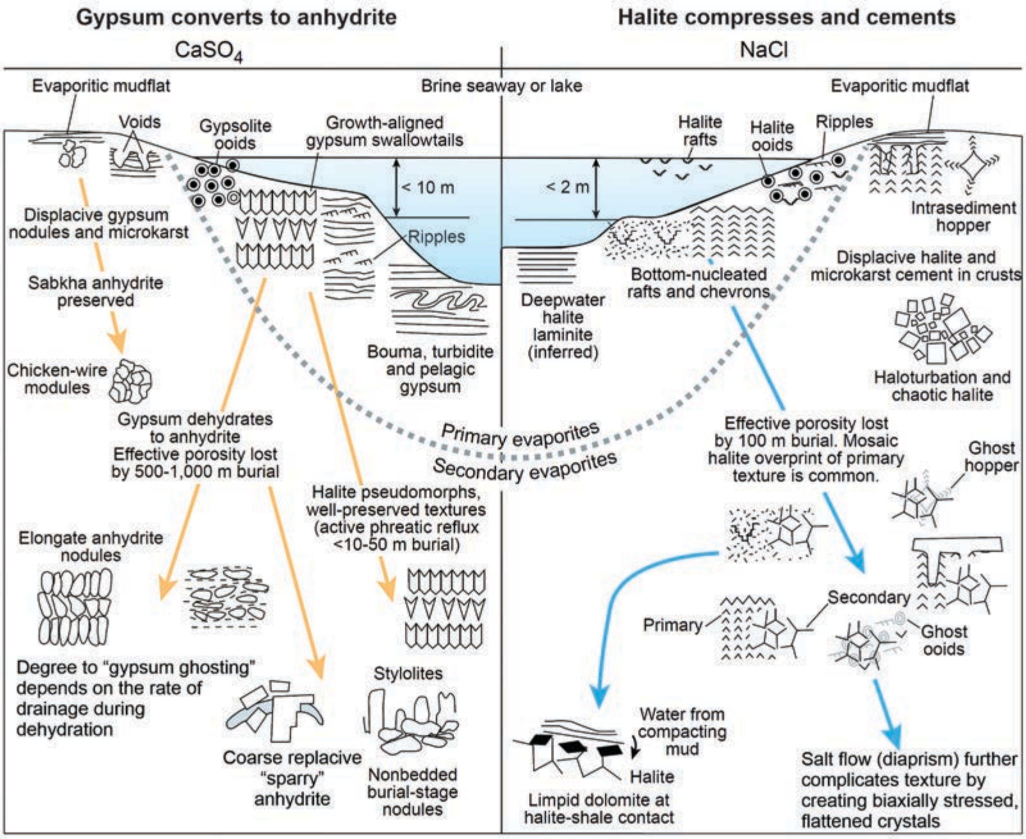


FIGURE 7.7 Secondary evaporite textures. (Modified after Warren, 2016.) The evaporite deposits are either diagenetically or secondarily altered during burial, due to their high solubility and susceptibility to deformation. Thus, evaporite deposits display textures that range from primary depositional features to diagenetically produced features, depending upon their ages and deformation histories. Wave and wind action reworks small crystals where rapid transition to gypsum saturation deposits small individual crystals (i.e., above the dashed line). Coarse crystal growth dominates in perennially saturated deeper brines (below the dashed line).

processes (such as crystal settling, bottom nucleation, among others) (Figure 7.7). Structures such as cross- or graded bedding and ripple marks reflect physical processes that indicate traction-current or turbidity-current transport (Figure 7.7).

7.1.4 Origin of Evaporite Deposits

7.1.4.1 Evaporation Sequence

When ocean water is evaporated in a laboratory, the evaporite minerals are precipitated in a definite sequence (Usiglio, 1848; Clarke, 1924) (see Figure 7.3E). Minor quantities of carbonate minerals begin to form when the original volume of seawater is reduced by evaporation to about 50% (Figure 7.3E). Gypsum appears when the original volume is reduced to about 20%, and halite forms when the water volume reaches approximately 10% of the original volume (see Kendall and Harwood, 1996) (Figure 7.3E). The precipitation of gypsum increases the Mg/Ca ratio in the remaining water, which favors the process of dolomitization. Dolomite occurs in association with evaporites in many ancient sedimentary successions as well as in some modern environments.

Magnesium and potassium salts are deposited when less than about 5% of the original volume of seawater remains (Figure 7.3E). The same general sequence of evaporite minerals occurs in natural evaporite deposits, although many discrepancies exist between the theoretical sequences as noted above, and the sequences observed in the rock record. Additionally, it must be noted that the proportion of $CaSO_4$ (gypsum and anhydrite) is greater than the proportion of Na-Mg sulfates in natural deposits than predicted from theoretical considerations (see Borchert and Muir, 1964).

7.1.4.2 Depositional Models

Modern evaporite deposits accumulate in a variety of subaerial and shallow subaqueous environments. Subaerial environments include both coastal and continental sabkhas, or salt flats, and interdune environments. Shallow subaqueous environments are present mainly in saline coastal lakes called salinas.

The basic requirements for the deposition of marine evaporites are a relatively arid climate, where rates of evaporation exceed the rates of precipitation, and the partial isolation of the depositional basin from the open ocean. Isolation is achieved by means of some type of barrier that restricts the free circulation of ocean water in and out of the basin. Under these restricted conditions, the brines formed by evaporation are prevented from returning to the open ocean, causing them to become concentrated to the point where evaporite minerals are precipitated. Although geologists agree on these general requirements for the formation of evaporites, considerable controversy still exists regarding deep-water vs. shallow-water depositional mechanisms for many ancient evaporite deposits.

Three possible models for the deposition of thick successions of marine evaporites are recognized, namely shallow-water, deep-basin (Figs 7.8A–B), shallow-water, shallow-basin (Figures 7.8C–D), and deep-water, deep-basin (Figures 7.8E–F).

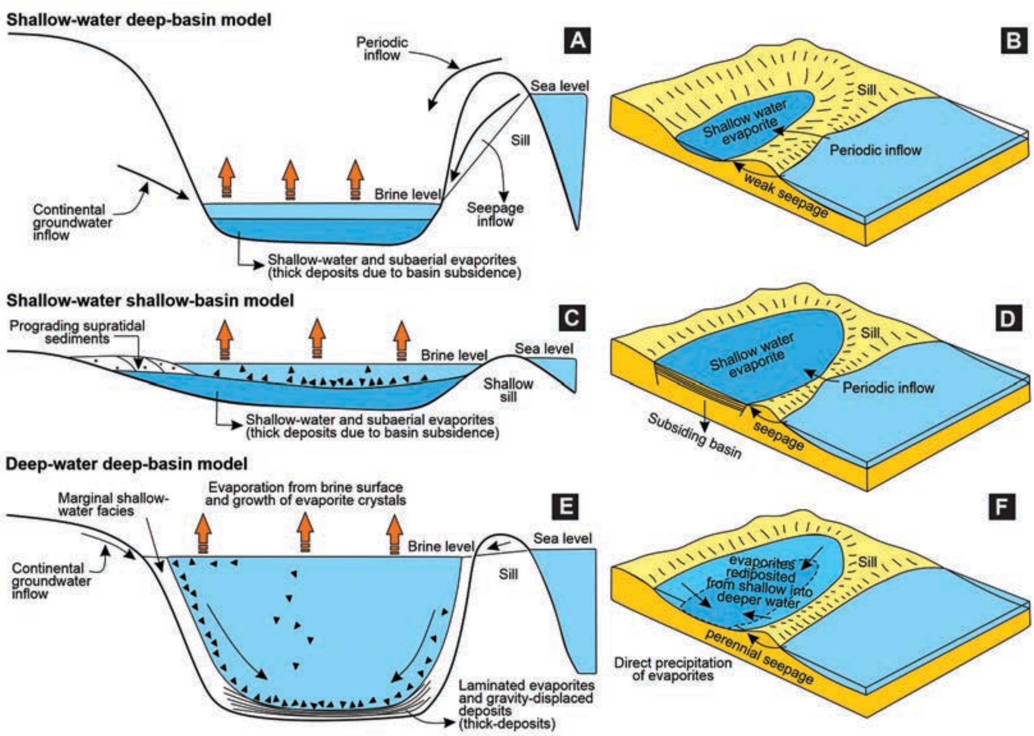


FIGURE 7.8 Evaporite depositional models. (Modified from Mann et al., 1999.) A–B: Shallow-water, deep-basin model. C–D: Shallow-water, shallow-basin model. E–F: Deep-water, deep-basin model.

The shallow-water, deep-basin model requires that the brine level in the basin be reduced well below the level of the sill, a process called evaporative drawdown; the recharge of water from the open ocean takes place only by seepage through the sill or by the periodic overflow of the sill (Figures 7.8A–B). Total desiccation of the floors of such basins could presumably occur periodically, allowing the evaporative process to go to completion and thereby depositing a complete evaporite sequence, including magnesium and potassium salts. Thick evaporite deposits accumulate under these conditions due to of continued subsidence (Figures 7.8A–B).

The shallow-water, shallow-basin model assumes concentration of brines in a shallow, silled basin, but it allows for the accumulation of great thicknesses of evaporites caused by continued subsidence of the basin floor (Figures 7.8C–D).

The deep-water, deep-basin model assumes existence of a deep basin separated from the open ocean by some type of topographic sill. The sill acts as a barrier to prevent the free interchange of basin water with the open ocean water, but it allows enough of the ocean water into the basin to replenish the one lost due to evaporation (perennial seepage) (Figures 7.8E–F). Seaward escape of some brine allows a particular concentration of brine to be maintained for a long time, leading to thick deposits of certain evaporite minerals such as gypsum.

7.2 SILICEOUS SEDIMENTARY ROCKS (CHERT)

7.2.1 Introduction

Siliceous sedimentary rocks are fine-grained, dense, very hard, and predominantly composed of SiO_2 (Silicon dioxide) minerals such as quartz, minor chalcedony, and opal (in young rocks) with impurities of siliciclastic grains and diagenetic minerals. Chert is a product of organic or inorganic precipitation. The amount of SiO_2 varies markedly in different types of cherts, ranging from more than 99% in very pure cherts such as novaculite to less than 65% in some nodular ones (Cressman, 1962). Aluminum is commonly the second most abundant element in cherts, followed by Fe and Mg or by K, Ca, and Na (Jones and Murchey, 1986).

Chert is the general term used for siliceous rocks as a group. These are common rocks in geologic successions ranging in age from Precambrian to Tertiary, but they make up only a minor fraction of all sedimentary rocks. They are particularly abundant in Jurassic to Neogene rocks, moderately abundant in Devonian and carboniferous rocks, and rare in Silurian and Cambrian deposits (Hein and Parrish, 1987). Cherts also have economic significance as silicon is used in the semiconductor and computer industries and for making glass and related products such as fire bricks, although much of this silica comes from quartz sand. Furthermore, siliceous deposits occur in association with important economic deposits of other minerals such as the Precambrian iron ores, uranium, manganese and phosphorite deposits and petroleum accumulations. Geologists are particularly interested in cherts as they provide information about paleogeography, paleooceanographic circulation patterns, and plate tectonics.

Cherts can be divided into three main textural types (Folk, 1974): (a) granular microquartz consisting of nearly equidimensional grains of quartz; average grain size is about 8–10 microns; size range <1 to 50 microns (Knauth, 1994); (b) chalcedony (fibrous silica), forming sheaf-like bundles of radiating, extremely thin crystals about 100 microns long; and (c) megaquartz, composed of equant to elongated grains commonly >20 microns in size.

The silica that makes up the tests of siliceous organisms is amorphous silica or opal, commonly called opal-A (Figure 7.9). As the remains of siliceous organisms contribute to the formation of chert, opal-A is present in some cherts, particularly those of Tertiary and younger age. Opal-A is metastable and crystallizes with time to opal-CT and finally to quartz (chert) (Behl, 2011). Hence, little opal is present in rocks older than ~60 Ma (Knauth, 1994). All gradations are present in siliceous deposits, from nearly pure opal to nearly pure quartz chert, depending upon the age of the deposits and the conditions of their burial.

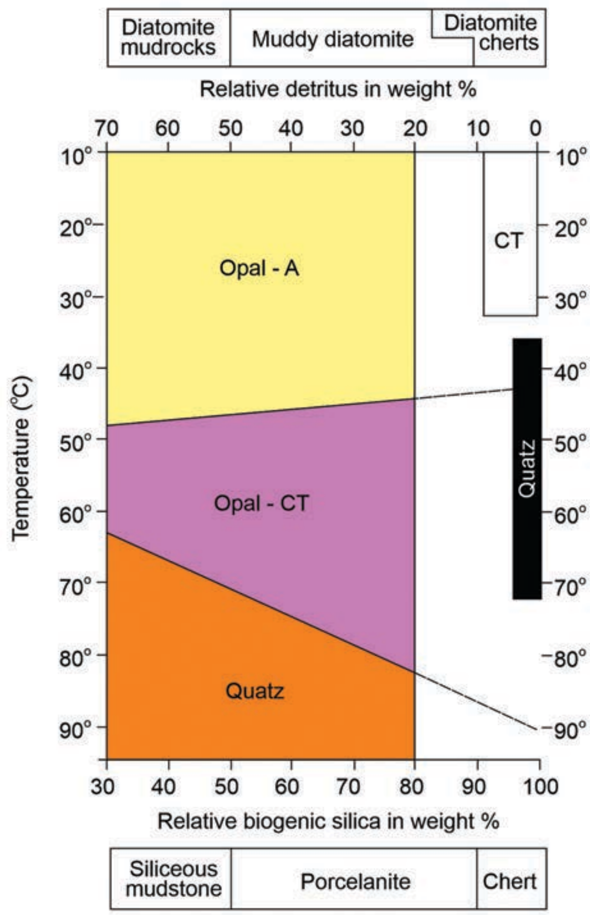


FIGURE 7.9 Silica phase diagram for the transformation of opal-A to opal-CT and subsequently to quartz. (Modified after Behl, 2011.)

7.2.2 Varieties of Chert

Several informal names are applied to chert, depending on their color, inclusions, and texture. Most cherts originate from the accumulation of siliceous ooze, consisting of fragments of some organisms (such as sponges) or shells of single-celled organisms (such as diatoms and radiolarians). Precambrian cherts contain microfossils of cyanobacteria. Cherts may also occur as nodules in limestones, where they form as a result of diagenetic processes. SiO_2-rich rocks may also form due to inorganic chemical precipitation, for example in the presence of silica-rich waters (i.e., geysers), but they are quite uncommon in the sedimentary record.

7.2.2.1 Major Varieties

Flint, Jasper, Novaculite, Forcelanite, Siliceous sinter, among others are major varieties of chert. Flint is a term used both as a synonym for chert and for a variety of cherts, particularly chert nodules. Flint is a cryptocrystalline form of quartz that occurs in chalk or marly limestones. Jasper is impure chalcedony or microcrystalline chert. It is colored red due to the inclusions of hematite and iron hydroxides and is typically associated with Precambrian iron formations and is called jaspilile. Novaculite is a very dense, fine-grained, even-textured, and microcrystalline to cryptocrystalline chert. Forcelanite is a fine-grained siliceous rock with a texture and fracture resembling that of unglazed porcelain. Siliceous sinter is porous, low-density, light-colored opaline or amorphous siliceous rock deposited by waters of hot springs and geysers. Although the most siliceous rocks predominantly consisting of chert, some also contain abundant detrital clays or micrite; these impure cherts grade into siliceous shales or siliceous limestones.

Chert is also divided on the basis of its gross morphology into two types: bedded cherts and nodular cherts. Bedded cherts are further distinguished by their content of siliceous organisms of various kinds. Cherts that originate from the lithification of siliceous ooze are classified according to the microfossils that make them, such as radiolarite (or radiolarian chert) if they are made up of radiolarians, or diatomite if they are made up of diatoms. Spiculite is the chert consisting of sponge spicules. In present-day oceans, radiolarian cherts forms near tropical regions, while diatomites are more common in the Arctic and Antarctic oceans. It must be noted that there are other organisms that produce important deposits of siliceous oozes, such as sponges and even cyanobacteria in the Precambrian.

7.2.2.1.1 Bedded Chert

Bedded chert, also called ribbon chert, consists of layers of nearly pure chert that are commonly interbedded with millimeter-thick partings or laminae of siliceous shale (Figure 7.10). Most chert beds lack internal sedimentary structures but graded bedding, cross-bedding, ripple marks, and sole markings have been noted. Many bedded cherts are composed dominantly of the remains of siliceous organisms that are commonly altered to some degree by solution and recrystallization. Most bedded cherts are formed when the silica-rich organic oozes deposited on the deep seafloor are

FIGURE 7.10 Bedded chert and nodules. Bedded chert are layers of nearly pure silica commonly interbedded with millimeter-thick partings or laminae of siliceous shale.

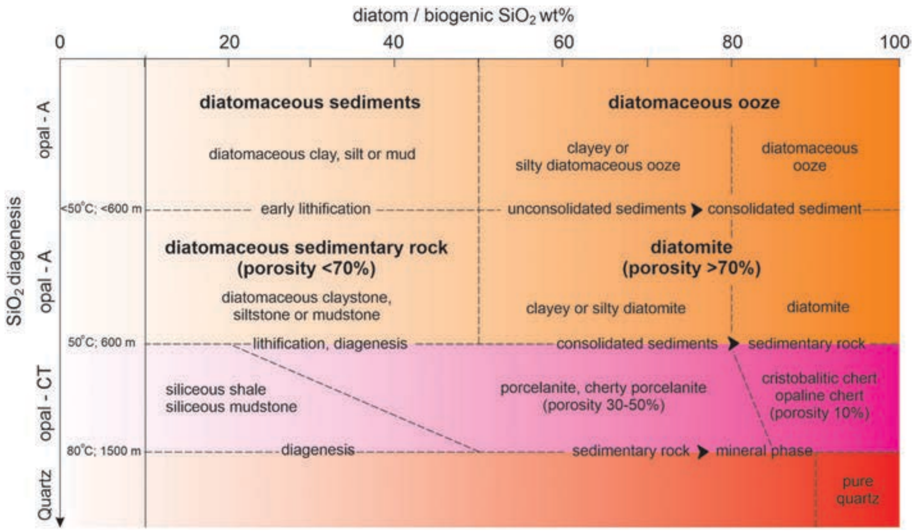


FIGURE 7.11 Categorization of diatomaceous sediments. The x-axis shows the weight percentage (wt%) of diatom/biogenic SiO_2 contained in sediments. The y-axis shows their corresponding transformations as a function of burial depth. (Modified after Aggrey et al., 2021; Zahajská, Petra, Sophie Opfergelt, Sherilyn C. Fritz, Johanna Stadmark, and Daniel J. Conley. "What Is Diatomite?" Quaternary Research 96 [2020]: 48–52. https://doi.org/10.1017/qua.2020.14.)

recrystallized. Most bedded cherts are found in relatively deep-water successions and are equivalent to the radiolarian and diatom siliceous oozes noted in the modern ocean floors. Bedded cherts can be subdivided into four principal types based on their type and abundance of siliceous organic constituents: (a) diatomaceous deposits, (b) radiolarian deposits, (c) siliceous spicule deposits, and (d) bedded cherts containing few or no siliceous skeletal remains.

7.2.2.1.2 Diatomaceous Deposits

Several organisms produce shells or skeletal fragments that consist of amorphous silica. The most important silica producers are unicellular planktonic organisms such as diatoms and radiolarians that live in the upper 200 m of the water column. Diatoms are single-celled algae (protists) that rely on photosynthesis and produce shells made of opaline silica between 2–200 micrometers (0.002–0.2 mm) in diameter. Diatomaceous deposits include both diatomites and diatomaceous cherts (Figure 7.11).

7.2.2.1.1.1 Diatomites Diatomite is a light-colored, soft, friable siliceous rock composed chiefly of the opaline (opal-A) frustules of diatoms, i.e., fossil diatomaceous oozes (see Figure 7.11) (Zahajsk et al., 2020). Diatomites of both marine and lacustrine (lake) origin are noted. Marine diatomites are commonly associated with sandstones, volcanic tuffs, mudstones or clay shales, clayey limestones (marls), and, less commonly, gypsum. Lacustrine diatomites are almost invariably associated with volcanic rocks.

7.2.2.1.1.2 Diatomaceous Chert Diatomaceous chert consists of beds and lenses of diatomite with a well-developed silica cement or groundmass that has converted the diatomite into dense, hard chert. Several hundred meters-thick cretaceous marine diatomaceous cherts have been reported, such as the Miocene Monterey Formation of California (Garrison et al., 1981); the diatoms evolved in the Cretaceous. Non-marine diatomaceous deposits have been reported in rocks as old as the Eocene

(Barron, 1987). When diatomaceous deposits are converted to quartz chert during diagenesis, the diatom tests are generally destroyed by dissolution and recrystallization.

7.2.2.1.1.3 Radiolarian Deposits These deposits consist predominantly of the remains of radiolarians, marine planktic protozoans with a lattice-like skeletal framework of opal. Radiolarian deposits are divided into radiolarite and radiolarian chert. Radiolarite is a hard, fine-grained, chert-like equivalent of radiolarian ooze, i.e., an indurated radiolarian ooze. Radiolarian chert is well-bedded, microcrystalline radiolarite that has a well-developed siliceous cement or groundmass. Radiolarian cherts are commonly associated with tuffs, mafic volcanic rocks such as pillow basalts, pelagic limestones, and turbidite sandstones suggesting a deep-water origin. These bedded cherts particularly those that exhibit a "pinch and swell" texture, are commonly called ribbon cherts. Some radiolarian cherts are associated with micritic limestones and other rocks suggesting deposition in water depths as much as 200 m (Iijima et al., 1979).

7.2.2.1.1.4 Siliceous Spicule Deposits Spicularite (or spiculite) is a siliceous rock composed mainly of the siliceous spicules of sponge. Spicularite is loosely cemented in contrast to spicular chert, which is hard and dense. Spicular cherts are mainly marine in origin and are associated with glauconitic sandstones, black shales, dolomite, argillaceous (clayey) limestones, and phosphorites. They are deposited mainly in relatively shallow waters, a few hundred meters deep.

7.2.2.1.1.5 Non-Fossiliferous Cherts Many bedded chert deposits contain few or no recognizable remains of siliceous organisms. Others contain few siliceous organisms and are associated with Precambrian iron formations, and with some Phanerozoic cherts.

7.2.2.1.1 Nodular Chert

Nodular cherts are sub-spheroidal masses, lenses, or irregular layers or bodies that range in size from a few centimeters to several tens of centimeters (Figure 7.12). They commonly lack internal structures, but some contain silicified fossils or relict structures such as bedding. Nodular cherts typically occur in shelf-type carbonate rocks and in some sandstones, shales, deep-sea clays, lacustrine sediments, and evaporites. Nodular cherts originate by diagenetic replacement.

FIGURE 7.12 Chert nodule. Nodular chert in limestones from the Antalo Limestone Formation, Upper Jurassic of Mugher, Ethiopia.

7.2.3 Origin of Chert

So, what were the sources of the silica, and what mechanisms operated in the past to extract silica from water, mainly from seawater, to form chert? It has been suggested that many nodular cherts in limestones (Figure 7.12) were formed in the ground water of mixed meteoric-marine coastal systems where dissolution of biogenic opal and mixing of marine and fresh waters formed waters that were highly supersaturated with respect to quartz and undersaturated with respect to calcite and aragonite (Knauth, 1979).

7.2.4 Silica Extraction from Seawater

Chert might be precipitated in some local basins where waters are saturated with silica, owing perhaps to the dissolution of volcanic ash or other processes related to volcanism (i.e., through the process of chemical extraction) (Ledesma-Vazquez et al., 1997). Also, some silica may be removed from seawater in the open ocean by precipitation or adsorption onto clay minerals or other silicate particles (see Mackenzie and Gees, 1971); however, such processes cannot account for the numerous bedded successions of nearly pure chert present in the geologic record.

The removal of silica from the ocean water by silica-secreting organisms to build opaline skeletal structures appears to be the only probable mechanism capable of large-scale silica extraction from undersaturated seawater (i.e., through the process of biogenic extraction). This biologic process has operated since at least the early Paleozoic time to regulate the balance of silica in the ocean. Radiolarians (Cambrian/Ordovician-Holocene), diatoms (Cretaceous-Holocene) and silicoflagellates (Cretaceous-Holocene) arc microplanktons that have built skeletons of opaline silica (opal-A). These siliceous microplankton (particularly diatoms and radiolarians) were abundant in the ocean during the Phanerozoic time, and extracted most of the silica, delivered to the oceans by weathering and other processes. Diatoms are probably responsible for the bulk of silica extraction from ocean waters in modern oceans and during much of the past 50 Ma (Calvert, 1974; Knauth, 1994); the radiolarians were important users of silica in the Phanerozoic Ocean, i.e., of Jurassic and older ages.

7.3 IRON-BEARING SEDIMENTARY ROCKS

7.3.1 Introduction

Some amount of iron is present in almost all sedimentary rocks ranging in average from ~0.4% in limestones, 2–4% in sandstones, ~4.8% in shales, and 5–6% in mudrocks (Blatt, 1982). Most iron-rich sedimentary rocks were deposited during three periods, the Precambrian, the early Paleozoic, and the mid–late Mesozoic (Jurassic-Cretaceous). The Precambrian banded iron formations (BIFs) are thick and laterally extensive deposits characterized by a fine chert-iron mineral laminations (Figure 7.13). Phanerozoic ironstones are mostly thin successions of limited areal extent, interdigitating with normal-marine sediments. Many such ironstones are oolitic and ooids are often composed of hematite (red/maroon in color) (Figure 7.14), and less commonly of berthierine-chamosite (green), goethite (brown) and, rarely, magnetite (black). Other common ironstones are hematitic limestones, where hematite has impregnated and replaced carbonate grains, and berthierine-chamositic, sideritic or pyritic mudrocks. The ironstones make up only a minor fraction of the total sedimentary record; however, they have great economic significance as iron ore deposits. They are the major source of iron mined for commercial purposes. Economically important iron deposits include the Lake Superior region-Labrador Trough (North America), the Transvaal-Griquatown region (South Africa), the Hammersley Range (Australia), the Krivoy Rog-KMA (Ukraine) and Minas Gerais (Brazil, South America).

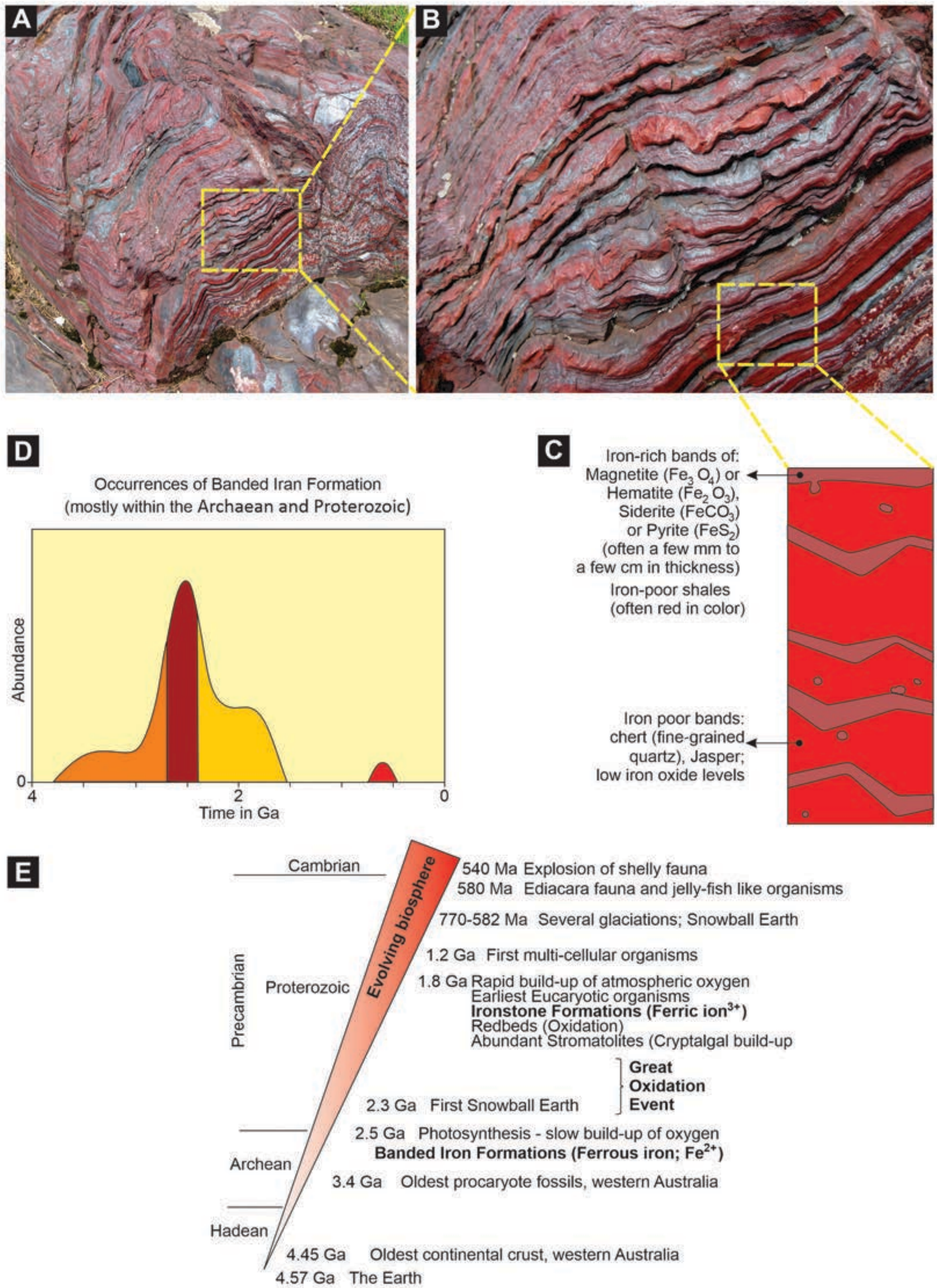


FIGURE 7.13 Banded ironstone formation (BIF). A–B: The Precambrian banded iron formations (BIFs) are thick and laterally extensive deposits characterized by fine chert–iron mineral laminations. C: Enlarged line diagram of B showing the two alternating iron-rich and iron-poor bands. D–E: Distribution of BIF through time. Note the Archean and Proterozoic spike.

7.3.2 Types of Iron-Rich Sedimentary Rocks

The major sedimentary iron deposits of the world are of two types: iron formation (for dominantly Precambrian, cherty iron-rich sediments) and ironstones (for dominantly Phanerozoic non-cherty iron-rich sediments) (see James, 1966). However, some workers do opine that it is incorrect to label ironstones as solely Phanerozoic deposits (Trendall and Morris, 1983). Kimberley (1994) noted that iron formation can either be cherty or non-cherty and that ironstone are rocks with >15% iron, whereas iron formation is a stratigraphic unit largely composed of ironstones. Here, ironstone is considered mainly as non-banded, non-cherty, commonly oolitic, iron-rich sedimentary rock and the term iron formation as mainly well-banded, cherty, iron-rich sedimentary rock (see Figure 7.13). Volumetrically, ironstones are much less important than iron formations. In general, the Precambrian BIFs and the Phanerozoic ironstones, constitute <1% of the total sedimentary rocks in the geologic record. Bog iron ores, iron-rich shales and iron-rich laterites are the other kinds of minor iron-rich sedimentary rocks (Dimroth, 1979).

7.3.2.1 Iron Formations

These are iron-rich deposits that range in age from early Precambrian to Devonian, although they are primarily of Precambrian age (James and Trendall, 1982). They consist of distinctively banded successions, 50–600 m thick, composed of layers enriched in iron alternating with layers rich in chert (Figures 7.13A–C). They are known as the Precambrian banded iron formations (BIFs). Most BIFs are of early Proterozoic age (1800–2400 Ma) but some are also know from the Archean, and from a much younger age, i.e., in the Cambrian (Figures 7.13D–E). Banding occurs on a scale ranging from millimeters to tens of meters (Figures 7.13A–C). Cherty iron formations are associated with dolomites, quartz-rich sandstones, and black shales and can grade locally into chert or dolomite. Iron formations can have a variety of textures including micritic, pelleted, intraclastic (rip-up clasts), peloidal, oolitic, pisolitic, and stromatolitic. Simonson (2003) suggests that the term granular iron formation (GIF) be used for iron formations that have (or originally had) coarse, granular textures and that the term banded iron formation be reserved for iron formations that have much finer-grained textures. Sedimentary structures reported from BIF include cross-bedding, graded bedding, load casts, ripple marks, erosion channels, shrinkage cracks, and slump structures.

7.3.2.2 Ironstones (Redbeds)

Ironstones are dominantly noted in early Paleozoic, Jurassic–Cretaceous, and early Cenozoic deposits, but they range in age from middle Precambrian to Holocene (Petranek and van Houten, 1997). They form bedded successions a few meters to a few tens of meters thick that are poorly banded or non-banded, in sharp contrast to the much thicker, well-banded iron formations. They commonly have an oolitic texture (Figure 7.14), and they may contain fossils that have been partly or completely replaced by iron minerals. Sedimentary structures include cross-bedding, ripple marks, scour-and-fill structures, clasts, and burrows, suggesting that mechanical transport of grains was involved in their formation. Ironstones are commonly interbedded with carbonates, particularly limestones, shales, and fine-grained sandstones of shelf to shallow-marine origin and they may grade locally to siliciclastic sedimentary rock units (van Houten and Bhattacharyya, 1982; Young and Taylor, 1989). Common sedimentary structures include cross-bedding, ripple marks, bioturbation, and scour-and-fill structures, and body fossils. Their association with shallow-marine shelf deposits of limestone, dolomite, mudrock, and sandstone suggests that they formed on nearshore shelf settings.

7.3.3 Mineralogy and Chemistry of Iron Formations and Ironstones

On the basis of the relative abundance of major kinds of iron-bearing minerals, James (1966) defined four different mineral clades in iron-rich sedimentary rocks: oxides, silicates, carbonates, and sulfides. The oxides and silicates are commonly the most important iron-bearing minerals; however,

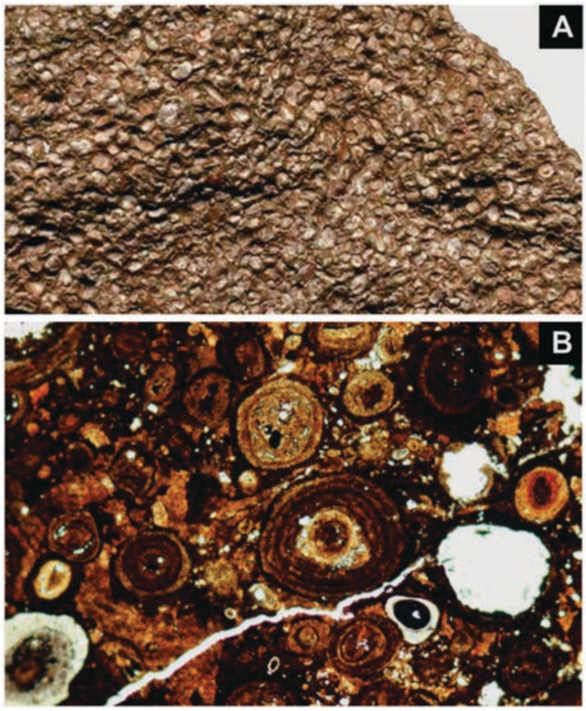


FIGURE 7.14 Ironstone. A: Many Phanerozoic ironstones are oolitic. B: In the Phanerozoic ironstones, the ooids are often composed of hematite (red/maroon in color).

sulfide minerals constitute major iron minerals but in thin beds. Iron formations consist mainly of SiO_2 and Fe, but the chemical composition of these rocks ranges widely depending upon the type of deposit. Although iron (Fe_2O_3, FeO or FeS) is the dominant chemical constituent in some iron-rich sediments, the iron content of many iron-rich sedimentary rocks is commonly exceeded by that of silica. The manganese concentration may reach considerable percentages in some iron formations. The average iron content of ironstones is similar to that of iron formations; however, ironstones typically have a higher concentration of aluminum and phosphorus and a lower silicon content (Appel and LaBcrge, 1987; Melnik, 1982; Trendall and Morris, 1983).

7.3.4 Iron-Rich Shales

Pyritic black shales occur in association with both iron formations and ironstones. They form thin beds in which sulfide content is as high as 75%. Pyrite occurs disseminated and may also be present as nodules, as laminae, and as a replacement of fossil fragments and other iron minerals. Pyriterich layers have been reported in some limestones, also. Siderite-rich shales (clay ironstones) occur primarily in association with other iron-rich deposits and are also present in coal measures. Siderite occurs disseminated in mudrocks or as flattened nodules and more or less continuous beds.

7.3.5 Miscellaneous Iron-Rich Sediments

7.3.5.1 Bog Iron Ores

Bog iron ores are minor accumulations of iron-rich sediments that occur particularly in small freshwater lakes of high altitude. Modern deposits contain iron-rich minerals such as goethite and siderite with minor amounts of manganese minerals such as psilomelane and pyrolusite. The bog iron deposits range from hard, oolitic, pisolitic, and concretionary forms to soft, earthy types. Iron-rich laterites are residual iron-rich deposits that form as a product of intense chemical weathering; these are highly weathered soils in which iron is enriched.

7.3.5.2 Manganese Crusts and Nodules

Manganese crusts and nodules (rarely more than a few centimeters in diameter) (Fig. 7.15A) are widely distributed on the modern seafloor in deeper parts of the Pacific, Atlantic, and Indian Oceans in areas where sedimentation rates are very low (Figure 7.15B). They have also been reported from ancient sedimentary deposits in association with such oceanic sediments as red shales, cherts, and pelagic limestones. Both iron-rich (15–20% Fe) and iron-poor (<6% Fe) varieties of manganese nodules are known.

The Fe-Mn nodules occur as individual rocks on the surface of or partially buried in sediments that blanket abyssal plains throughout the global ocean (Figure 7.15B). Nodules form where the rates of sedimentation are <10 mm per thousand years, such as at abyssal plains, in areas away from continents where biological productivity is low, and in areas that remain below the carbonate compensation depth but also near continental margins bordered by deep-sea trenches, arid climate, or with a wide continental shelf. Nodules are generally golf-ball size with average diameters of 1–12 cm (Figure 7.15A), but they can vary in diameter from millimeters to 20 cm. However, nodule size, density, and metal content varies greatly.

When the substrate, for millions of years, is been kept free of sediments on the ocean floor, the Fe-Mn crusts form pavements on rock surfaces and coats pebbles and cobbles (see Mizell and Hein,

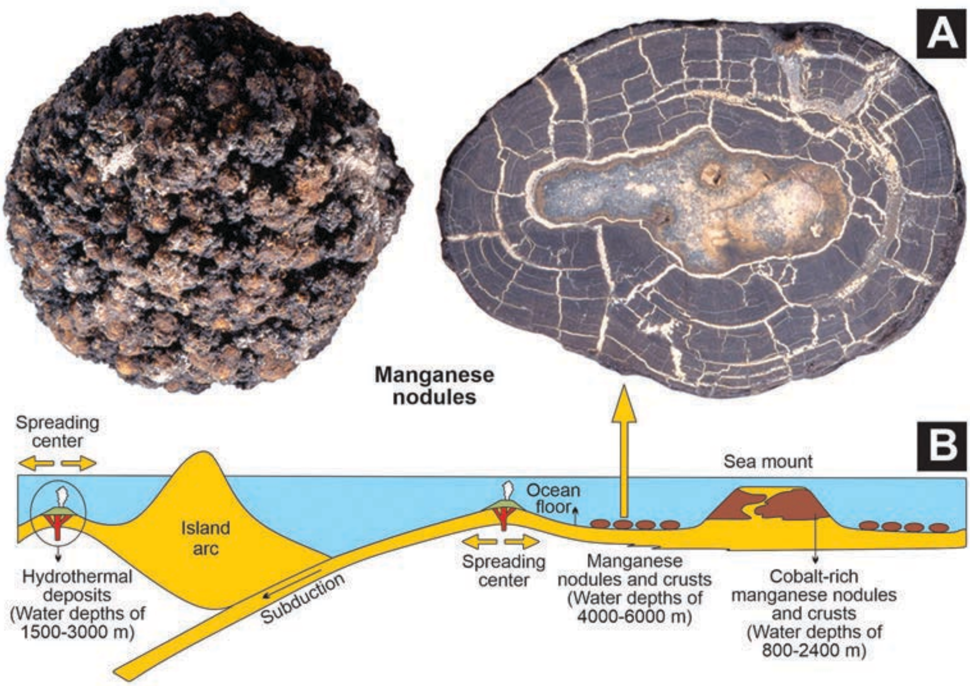


FIGURE 7.15 Manganese nodules and their occurrences. A: Manganese nodule and its thin section showing its internal geometry. (Modified after https://worldoceanreview.com. © Charles D. Winters/NatureSource/Agentur Focus.) B: Geological setting and water depth of the occurrences of manganese nodule. (Modified after Lee et al., 2013.)

2018). Thus, the geological setting for the formation of crust are the flanks of extinct submarine volcanoes (such as seamounts, guyots), ridges, and plateaus (Figure 7.15B). In the Atlantic and Indian Oceans, the crusts are also associated with oceanic spreading centers and their accompanying hydrothermal activity (Figure 7.15B).

7.3.5.3 Pyrite Crystals

Crystals of pyrite (iron sulfide) are noted in many carbonaceous black shales and limestones. Reducing (anoxic) environments (i.e., the lack of available oxygen) favors the development of pyrite. Due to the lack of replenishment of oxygen in the water column, the organic-rich muds accumulate on the sea bed as noted in the modern Black Sea. Hence, the organic matter is stopped from decaying and the iron that is present in the sediments is reduced and recrystallized as pyrite.

7.3.5.4 Iron-Rich Metalliferous Sediments

The iron-rich metalliferous sediments have been noted in several oceanic settings, particularly near active mid-ocean ridges. They form by the precipitation from metal-rich hydrothermal fluids that have become enriched through contact and interaction with hot basaltic rocks. These sediments are enriched in Fe, Mn, Cu, Pb, Zn, Co, Ni, Cr, and V. Metal-enriched sediments have also been reported from some ancient sedimentary deposits in association with submarine pillow basalts and ophiolite sequences of ocean crustal rocks.

7.3.6 Heavy Mineral Placers

These are sedimentary deposits that form by the mechanical concentration of mineral particles of high specific gravity, commonly in beach or alluvial environments. Magnetite, ilmenite, and hematite sands are common constituents of placers, particularly of beach and marine placers. Placers are local accumulations, generally less than 1–2 m in thickness, that occur mainly in Pleistocene-Holocene rocks.

7.3.7 Mechanism of Iron Deposition

The transport and deposition of iron-rich sediments are governed by the Eh and pH of the environment; Eh is more important than pH in determining which iron-bearing minerals will be deposited. The Eh–pH diagram (the Pourbaix diagram) illustrates the fields of stability of minerals in terms of the activity of hydrogen ions (pH) and electrons (Eh). However, as the iron geochemistry of natural systems is complex, such diagrams are of only limited use in interpreting the actual environment of iron deposition. Additionally, the mechanisms involved in the past for iron to be transported and deposited, forming iron formations and ironstones, are poorly understood.

The formation of iron-rich rocks involving the source of iron, its transport to the depositional basin, and precipitation within the basin, is also controversial. The following plausible scenarios are noted.

Scenario 1: Iron formations were deposited in continental shelf to upper slope marine environments and that the iron was derived by subaerial weathering of iron silicate minerals. But for land source, iron in the oxidized or ferric (Fe^{3+}) state is much less soluble than iron in the reduced or ferrous (Fe^{2+}) state. Thus, under oxidizing conditions (i.e., from the land source) iron would be precipitated rather than undergo into solution.

Scenario 2: Low atmospheric oxygen concentrations existed during the early Precambrian, allowing large quantities of iron to be transported from land in the soluble, reduced state (Fe^{2+}) to marine basins. But this does not explain the solution and transport of iron during later Precambrian and Phanerozoic times, when an oxidizing atmosphere existed.

Scenario 3: Iron was transported as colloids by physical processes rather than in true solution or that it was absorbed to clay particles or organic materials, and then transported along with these substances. Reducing conditions are required for transport of large amounts of iron in solution; this still cannot explain the late Precambrian and Phanerozoic time occurrences.

Scenario 4: The source of the iron lay within the depositional basin itself. The iron-bearing minerals were transported to the ocean where ferric iron was reduced by anoxic (no oxygen) bottom waters and the resulting ferrous iron (Fe^{2+}) was taken into solution.

Scenario 5: Iron was derived by the "exhalation" from oceanic rocks. Submarine reaction of outpouring lava and hydrothermal activity from hot springs located along mid-ocean ridges furnished iron to the ocean water, or possibly iron-rich fluids were also exhaled from source regions too deep within the crust or mantle for seawater convection. The ocean was stratified into an upper, oxygen-rich layer, and intermediate-deep oxygen-poor (anoxic) layers; the deep, iron-rich, anoxic oceanic waters moved upward toward the surface along continental shelves.

Among all the scenarios outlined above, there is a general consensus that a major source of iron in iron formations probably lay within the ocean itself, and some iron may have come from continental sources, particularly during the early Precambrian.

7.4 SEDIMENTARY PHOSPHORITES

7.4.1 Introduction

The phosphorus content of rocks is largely expressed as a percentage of P_2O_5 (phosphorus pentoxide) in the form of grains of apatite (a heavy mineral), bone fragments, or coprolites. The phosphorus-rich sedimentary rocks are variously called phosphate rock, rock phosphate, phosphates, and phosphorites. Sedimentary phosphorites are rocks that are significantly enriched in phosphorus and contain >15% P_2O_5 or 6.5% phosphorus; the average sedimentary rock contains <1% P_2O_5 or 1.5% phosphorus. Sedimentary rocks that contain <15% P_2O_5 are referred to as phosphatic such as the phosphatic shale. They contribute more than 80% of the world's production of phosphate and make up about 96% of the world's total resources of phosphate rock.

The total volume of sedimentary phosphates in the geologic record is very small. However, they have a special economic significance. Sedimentary phosphates occur in rocks of all ages from Precambrian to Holocene, but more so during the Precambrian and Cambrian times (see Figure 7.16A). They occur in central and southeast Asia (China, USSR/MPR, and Australia); the Permian in North America; the Jurassic and Early Cretaceous in Eastern Europe; the Late Cretaceous to Eocene in the Tethyan province of the Middle East and North Africa; and the Miocene of southeastern North America (see Figure 7.16B). Phosphorite nodules and phosphatic sediments occur also on the present ocean floor at shallow depths in the vicinity of coastlines. They are particularly common off the coasts of Peru and Chile, southwest Africa, eastern United States, southern and Baja California, the continental margins of India, and on some seamounts and atolls in the Pacific (see Glenn and Garrison, 1978).

7.4.2 Mineralogy and Chemistry

Sedimentary phosphorites (i.e., varieties of apatite) are composed of calcium phosphate minerals. The main varieties are fluorapatite [$Ca_5(PO_4)_3F$] (better known as carbonate-fluor-apatite, CFA), chlorapatite [$Ca_5(PO_4)_3Cl$], and hydroxyapatite [$Ca_5(PO_4)_3OH$]. Most are carbonate hydroxyl

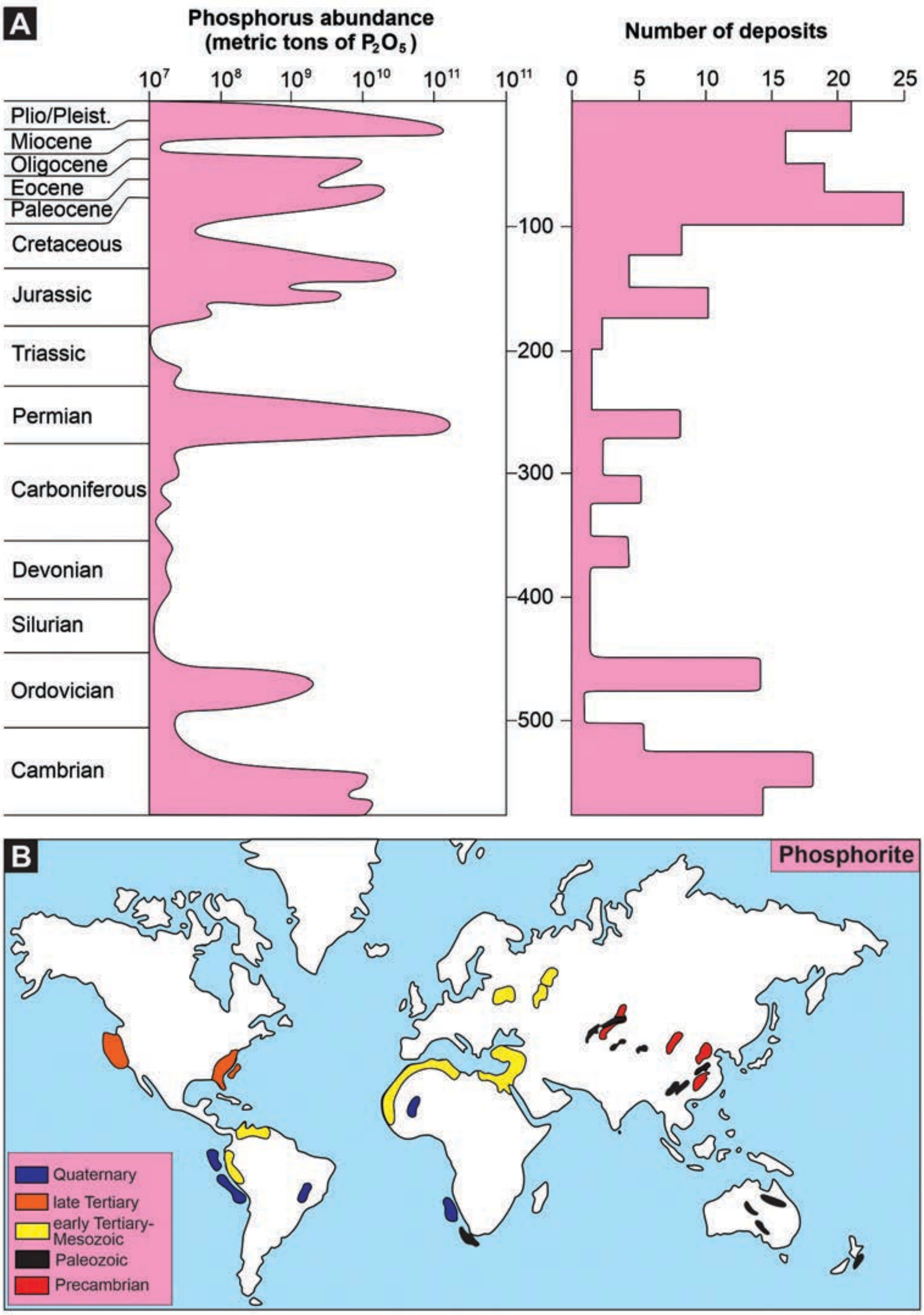


FIGURE 7.16 Distribution of phosphorites. A: Distribution through time. B: Geographic distribution (on a present-day map).

fluorapatites in which ~10% carbonate ions are substituted for phosphate ions to yield the general formula of $Ca_{10}(PO_4.CO_3)_6$; they are commonly called francolite. Collophane, a wastebasket term, is used for sedimentary apatites. Detrital quartz, authigenic chert (microcrystalline quartz), opal-CT, calcite, and dolomite are also common constituents of sedimentary phosphorites. Glauconite, illite, montmorillonite, and zeolites may also be present in some deposits; moderately abundant organic matter is a characteristic constituent of many phosphorites (Nathan, 1984). In general, the chemistry of phosphorites is dominated by three minerals – phosphorus, silicon and calcium (Nathan, 1984; Notholt et al., 1989).

7.4.3 Distinguishing Characteristics

Phosphate-rich sedimentary rocks occur in layers ranging from thin laminae a few millimeters-thick to a few meters-thick beds. Some phosphate successions such as the Phosphoria Formation of the Idaho-Wyoming area (USA) reach several hundred meters in thickness, although such successions are not composed entirely of phosphate-rich rocks. Phosphorites are generally interbedded with shales, cherts, limestones, dolomites, and, more rarely, sandstones. Phosphatic rocks commonly grade regionally into non-phosphatic sedimentary rocks of the same age. Phosphorites have textures that resemble those found in limestones such as peloids, ooids, fossils (bioclasts), and clasts that are composed of apatite. Peloidal or pelletal phosphorites are particularly common; oolitic phosphorites are somewhat less so. Most phosphorite grains are sand-sized, although particles greater than 2 mm may be present. These larger grains, referred to as nodules, can range in size to several tens of centimeters. The phosphatic grains may contain inclusions of organic matter, clay minerals, silt-sized detrital grains, and pyrite.

7.4.4 Phosphorite Deposits

Four kinds of phosphorite deposits are recognized: bedded, nodular, pebble-bed, and guano. The major phosphorite deposits are mainly bedded marine deposits.

7.4.4.1 Bedded Phosphorites

Bedded phosphorites form distinct beds of variable thickness, commonly interbedded and interfingering with carbonaceous mudrocks, cherts, and carbonate rocks. The phosphorite in bedded deposits occurs as peloids, ooids, pisoids, phosphatized brachiopods and other skeletal fragments, micrite-like apatite mud, and cements. Bedded marine phosphorites are common in Precambrian and Cambrian rocks of Australia, and the Cretaceous-Tertiary rocks of North Africa, among others. Bioclastic phosphorites are a special type of bedded phosphate deposits composed largely of vertebrate skeletal fragments such as fish bones, shark teeth, fish scales, and coprolites. Deposits composed mainly of invertebrate fossil remains such as phosphatized brachiopod shells are also known.

7.4.4.2 Nodular Phosphorites

Nodular phosphorites are brownish to black, spherical to irregularly shaped forms such as flat slabs to spherical nodules and irregular masses, ranging in size from a few centimeters to a meter or more. Internal structure of phosphate nodules ranges from homogeneous (i.e., structureless) to layered or with concentric bands. Phosphatic grains, pellets, coated grains (= ooids), shark teeth, coprolites and other fossils (such as vertebrate skeletal debris, especially those of fish) often occur within the nodules. Nodular phosphorites are particularly common in many Neogene to Holocene phosphatic deposits of the world (Burnett and Riggs, 1990). Phosphate nodules are also forming today in zones of upwelling in the ocean, such as on the Peru continental margin (Burnett and Froelich, 1988). Many ancient nodular phosphorites may have had a similar origin under conditions of marine upwelling; however, some ancient phosphorite nodules may be of diagenetic origin.

7.4.4.3 Pebble-Bed Phosphorites

Pebble-bed phosphorites are composed of phosphatic nodules, phosphatized limestone fragments, or phosphatic fossils that have been mechanically concentrated by reworking of earlier formed phosphate deposits. At places, the vertebrate skeletal fragments are concentrated locally to form bone beds, commonly with fish scales. Such concentrations are formed due to the current and wave reworking of sediments and the subsequent winnowing of finer material, leaving phosphatic grains as lag deposits. Favorable environments include transgressive and regressive shelf and shore zones and fluvial and intertidal channels.

7.4.4.4 Guano

Guano deposits are composed of bird and bat excrement that has been leached to form an insoluble residue of calcium phosphate. Guano occurs today on small oceanic islands in the eastern Pacific and the West Indies. Guano deposits are not important in the geologic record. Thick accumulations of bird guano are noted on some oceanic islands in the eastern Pacific, such as those of the Ocean Island, Nauru, Christmas Island, Makatea, and the Banabans, along the Pacific coast of South America, and in the West Indies.

7.4.5 Origin of Phosphorites

7.4.5.1 Chemical/Biochemical Processes

The principal phosphate minerals in sedimentary rocks are the various varieties of apatites, of which carbonate apatite ($Ca_{10}CO_3(PO_4)_6$) is economically important. Throughout geologic time, the weathering of phosphorous-bearing rocks on land have provided phosphorus to the oceans, through river runoff. The average concentration of phosphorus in river water is 20 parts per billion (ppb), as compared to 70 ppb in the ocean (Gulbrandsen and Roberson, 1973).

The biologic utilization of phosphate to build soft body tissue appears to provide the most feasible answer to the problem of phosphate concentration in sediments. Modern phosphate nodules are forming in areas of oceanic upwelling where a steady supply of phosphate brought from the large, deep-ocean reservoir allows continuous growth of organisms in large numbers. After death, these organisms and organic debris, if not consumed by scavengers, pile up on the ocean floor under reducing conditions where their decay is inhibited. These organic materials include the remains of phytoplankton and zooplankton, coprolites (feces), and the bones and scales of fish; all containing phosphorus. Under the reducing conditions of the seafloor, some of the soft body tissue is thus preserved long enough to be buried and incorporated into accumulating sediment. In this way, ~1–2% of the total phosphorus involved in primary productivity in upwelling zones is incorporated into the sediments (Baturin, 1982).

7.4.5.2 Physical Processes

A two-stage process has been suggested for the origin of ancient phosphorite deposits (Kolodny, 1980). In the first stage, apatite forms diagenetically in reducing basins by mobilizing phosphorus in interstitial waters (similar to the formation of modern phosphorites). The second and final stage involves the reworking and enrichment of these diagenetically formed phosphorite grains by mechanical concentration under oxidizing conditions. Concentration takes place in a high-energy environment, probably during sea-level low stands. During this stage, the phosphate grains are transported into a different depositional setting than in which they were formed. This method of phosphorite formation explains the clastic textures and primary sedimentary structures noted in ancient phosphorite deposits.

7.5 CARBONACEOUS SEDIMENTARY ROCKS

7.5.1 Coal, Oil Shale, and Bitumen

Most sedimentary rocks contain at least a small amount of organic matter, i.e., the preserved residue of plant or animal tissue. When this tissue decays, in an oxygen-poor environment, such as in a restricted basin, stagnant swamp, or bog, or if the supply of organic matter is so great that it overwhelms all available oxidants, decay-resistant organic matter is preserved long enough to become incorporated into accumulating sediments. Once buried, it persists for hundreds of millions of years.

7.5.2 Kinds of Organic Matter in Sedimentary Rocks

Three basic kinds of organic matter accumulate in subaerial and subaqueous environments: humus, peat, and sapropel. Peat consists of humic organic matter that accumulates in freshwater or brackish-water swamps and bogs where stagnant, anaerobic conditions prevent total oxidation and bacterial decay of the organic matter. Therefore, some of the humus that accumulates under these reducing conditions is preserved in sediments. Sapropel refers to fine organic matter that accumulates in lakes, lagoons, or marine basins where oxygen levels are low due to poor water circulation or where the supply of organic remains is high enough to suppress oxygen concentration levels. Sapropel consists of the remains of phytoplankton, zooplankton, and spores and fragments of higher plants. It must be mentioned that in ancient sediments, it is often difficult to differentiate with certainty between the types of organic matter found; both humic and sapropelic types are noted. Humic organic matter is the chief constituent of most coals, although a few are formed of sapropel. The organic matter in oil shales and other carbonaceous mudrocks and limestones originated from the sapropel, but it is so finely disseminated and altered that it is difficult to identify. This type of organic matter is called kerogen.

7.5.3 Classification of Carbonaceous Sedimentary Rocks

The dominant organic constituents of carbonaceous sediments are humic and sapropelic organic matter. Most coals are humic, although a few are sapropelic coals made up mostly of spores, algae, and fine plant debris. Cannel coals and boghead coals (see below) are sapropelic. The inorganic constituents are either siliciclastic grains or carbonate materials. Carbonaceous sediments are thus divided into three basic types of organic-rich rocks: coal, oil shale, and asphaltic substances. Each of these contains at least 10 to 20% of organic constituents.

7.5.3.1 Coals

Coal is defined as a readily combustible rock containing more than 50% by weight and more than 70% by volume of carbonaceous material formed from the compaction or induration of altered plant remains. Differences in the type (of plant material), rank (in the degree of metamorphism), and grade (range of impurities) characterizes different varieties of coal. Coals also contain impurities (such as ash) that are largely siliciclastic materials. Some very impure coals (bone coals) may contain 70–80% ash, but most coals have less than 50% ash by weight. The siliciclastic materials include quartz, clay, pyrite, and marcasite ("white iron pyrite"; iron sulfide, FeS_2), among others.

7.5.3.1.1 Classification

A common method of classifying coals is by rank, which is based on the degree of coalification or carbonification (the increase in organic carbon content) attained by a given coal owing to burial and metamorphism (Figures 7.17A–B) (see Bustin et al., 1985). Peat is not a true coal. It consists of unconsolidated, semi-carbonized plant remains with high moisture content (Figure 7.17A).

Lignite or brown coal is the lowest-ranked coal (Figure 7.17) characterized by high moisture content and it commonly retains many of the structures of the original woody plant fragments. They are dominantly Cretaceous or Tertiary in age. Bituminous coals are hard, black coals that contain fewer volatiles and less moisture than lignite and have a higher carbon content (Figures 7.17A–B). They commonly display thin layers consisting of alternating bright and dull bands. Subbituminous coal has properties intermediate between lignite and bituminous coal (Figures 7.17A–B). Anthracite is a hard, black, dense coal commonly containing more than 92% carbon (Figure 7.17C). It is a bright, shiny rock that breaks with conchoidal fracture, such as the fractures in a broken glass. Bituminous coals and anthracite are largely of Mississippian and Pennsylvanian (carboniferous) age. Coal classification by rank used in the United States is provided in Figure 7.17C. Higher rank coals have higher calorific value (see Trumbull, 1960; for details see American Society for Testing and Materials (ASTM), 1999). In general, higher-ranked coals are harder, contain less moisture and volatile matter, and have higher calorific values. Cannel coal and boghead coal are non-banded, dull black coals that also break with a conchoidal fracture; but have bituminous rank, and much higher volatile content than anthracite. Cannel coal is composed mainly of spores. Boghead coals are composed dominantly of non-spore algal remains. Bone coal is a very impure coal containing high ash content.

Coals are also classified on the basis of megascopic textural appearance and petrographic or microscopic constituents. On the basis of megascopic appearance, four types of coal (called lithotypes: containing millimeter-thick bands or layers of humic coal) are recognized: vitrain, clarin, durain, and fusain (see Stapes, 1919). Vitrain is brittle with a black luster. It is composed primarily of the maceral group vitrinite, which is derived from the woody tissue of large plants. It occurs in narrow, sometimes markedly uniform, bright bands that are about 3 to 10 mm thick. The thick layers break into conchoidal fractures. Vitrain is formed under somewhat drier surface conditions than clarain and durain. Clarain is characterized by alternating bright and dull black laminae that are <1 mm thick. The bright layers are composed of vitrinite and the dull ones are of the other maceral groups such as liptinite and inertinite. Clarain has a silky luster and is less brilliant than vitrain. Durain has a hard granular texture and is dull black to dark gray in color. It is composed of liptinite, inertinite, and large amounts of inorganic minerals. Durain occurs in layers between 3 to 10 mm thick. It is formed in peat deposits below water level. Fusain is a few millimeters thick and centimeters long, with silky and fibrous lenses. Most fusain is very soft and crumbles readily into a fine, soot-like powder. It is composed mainly of fusinite (carbonized woody plant tissue) and semi-fusinite from the maceral group inertinite, which is rich in carbon and is highly reflective. It closely resembles charcoal, both chemically and physically, and is formed in peat deposits swept by forest fires, or by fungal activity that generates intense heat, or by subsurface oxidation of the coal.

Under the microscope, coal consists of macerals (Stopes, 1935), i.e., organic units that are fragments of plant debris or fragments consisting of more than one type of plant tissue. The word maceral is derived from the Latin word *macerare*, meaning "to macerate." The starting materials for macerals are woody tissues, bark, fungi, spores, and so on; however, these materials are not always recognizable in coals. Coal macerals are identified on the basis of several characteristics: (a) reflectivity (the extent to which they reflect light), (b) degree of anisotopy (differences in reflectivity in different directions within a maceral) or isotopy as viewed under a petrographic microscope, (c) presence or absence of fluorescence when the specimen is irradiated with blue (ultraviolet) light, (d) morphology (shape), (e) relief, and (f) size (Bustin et al., 1985).

Three groups of macerals are noted: vitrinite, liptinite (formerly called exinite), and inertinite. The vitrinite group is the most abundant. They are derived primarily from the cell walls and woody tissues of plants. They show a wide range of reflectance values, but in individual samples these values tend to be intermediate as compared with those of the other maceral groups. Three major

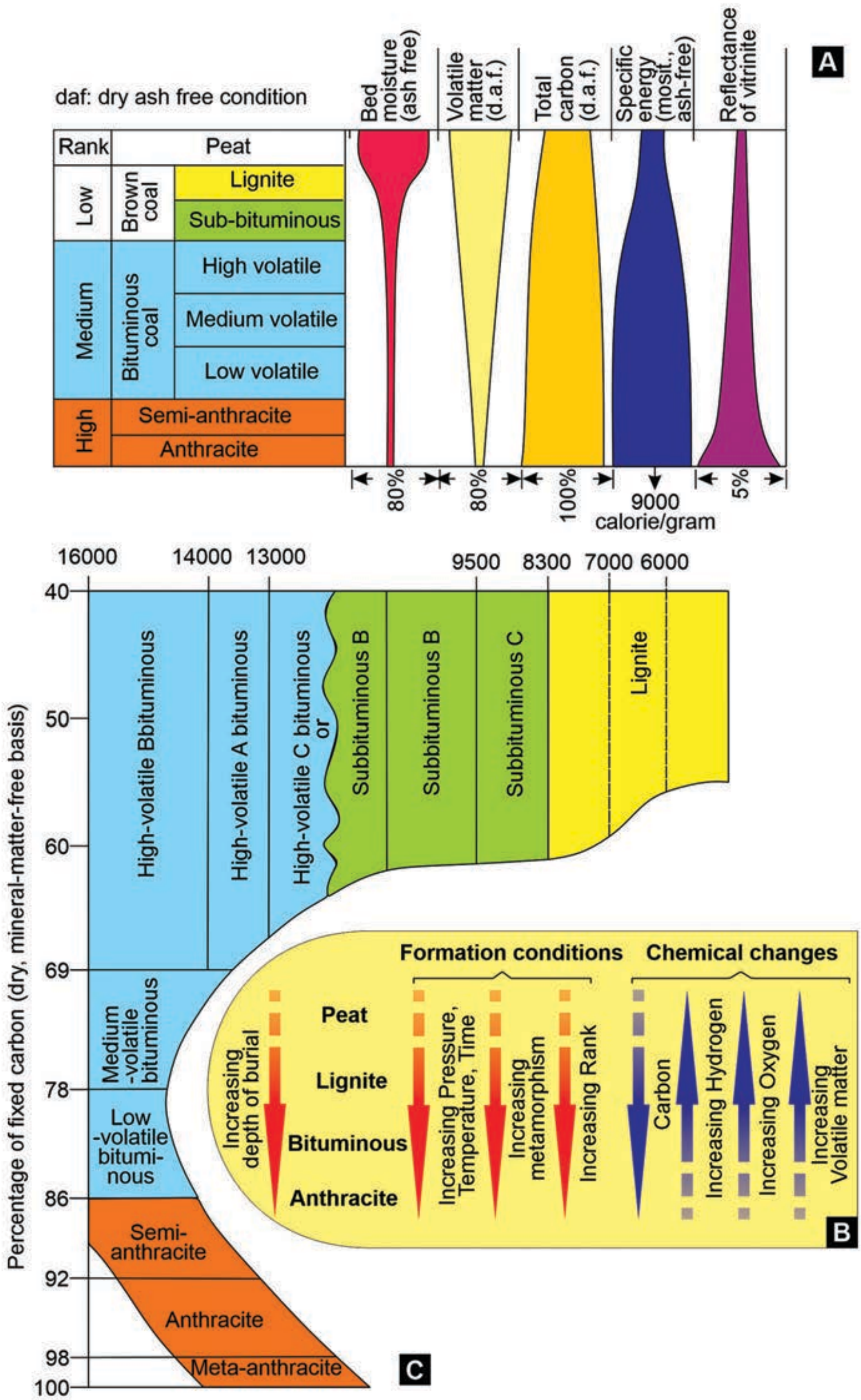


FIGURE 7.17 Coal properties and rank. A: Variation in some key coal properties with rank advance. (Modified after Ward, 1984.) B: Formation conditions and chemical changes as a function of depth of burial. C: Gross calorific value. Btu/lb is based on a moist, mineral-matter-free basis. Btu/lb: British thermal units per pound. (Modified from Trumbull, 1960.)

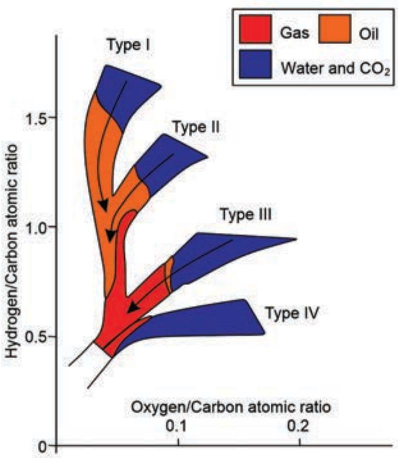
varieties are noted: telinite (the brighter parts of vitrinite that makes up the cell walls); collinite (clear vitrinite that occupies the spaces between cell walls); and vitrodetrinite, i.e., collinite in particulate form (<10 μm in size). The liptinite group makes up 5 to 15% of most coals. They are derived from waxy or resinous parts of the plant, such as cuticles, spores, and wound resins. Their reflectance values are usually the lowest. Several varieties are noted: sporinite (spores are typically preserved as flattened spheroids); cutinite (part of cross sections of leaves, often with crenulated surfaces); and resinite (ovoid and sometimes translucent masses of resin). The liptinites under ultraviolet light, show fluoresce (i.e., luminesce because of absorption of radiation); with increasing rank their optical properties are similar to those of the vitrinites, and the two groups become indistinguishable. The inertinite group makes up 5 to 40% of most coals. Inertinites are derived from strongly altered or degraded plant material that is thought to have been produced during the formation of peat; in particular, charcoal produced by a fire in a peat swamp is preserved as fusinite. Fusinite has a charcoal-like appearance with a cell texture. These cells may be empty or filled with mineral matter. The reflectance values of inertinites are usually the highest in a given sample.

7.5.3.2 Oil Shale (Kerogen Shale)

Oil shale is a fine-grained sedimentary rock from which large quantities (about 4–50% of the weight of rock) of oil are derived by heating to a temperature of ~350°C (through a process of distillation). Therefore, oil shales are important source rocks. The contained organic matter is largely kerogen (with minor amounts of bitumen).

Kerogen consists of inhomogeneous macromolecular aggregates that are insoluble in water, alkali, non-oxidizing acids and organic solvents (such as benzene, methanol, toluene, and methylene chloride). They make up 90% or more of the organic matter within the sedimentary rock and much of this organic matter is finely distributed and altered so that the organisms from which it was formed, cannot be identified. The remaining inorganic constituents include quartz, silt, and clay minerals. Kerogen is the most abundant form of organic carbon; three times more abundant than coal, petroleum, and gas. Kerogen consists of masses of almost completely macerated organic debris. Petrographically, kerogen is largely made up of liptinite macerals; vitrinite and inertinite occur only as minor fractions (Hutton, 1995). The organic matter is derived from three primary sources, terrestrial plants, lacustrine (lake) algae, and from marine organisms such as algae and dinoflagellates. Organic matter from terrestrial plants is less important those that from lacustrine algae and marine organisms. Broadly, there are two types of kerogens: sapropelic and humic. The sapropelic kerogens are rich in aliphatic compounds (i.e., lower molecular compounds) that are generally derived from aquatic and marine algae; these have good petroleum potential. The humic kerogens are rich in aromatic compounds (i.e., higher molecular compounds) that are largely derived from the remains of higher plants (i.e., vascular plants such as trees, bushes, etc.) and have poor petroleum potential. Little free oil occurs in oil shales, except in the form of small blebs, pockets or veins of asphaltic bitumens (commonly called asphalt).

Oil shales are characterized by distinct lamination caused by the alternations of millimeter-thick organic laminae and siliciclastic or carbonate laminae. Oil shales, based on their mineral content, are classified into three types: carbonate-rich, siliceous, and cannel. Carbonate-rich shales have high amounts of carbonate minerals. Thus, not all oil shales are shales but some are organic-rich siltstones, limestones, and impure coals. Cannel shale is an oil shale that consists predominantly of organic matter that completely encloses other mineral grains. It is sometimes classified as an impure cannel coal. Oil shale may contain between 60 and 90% of mineral matter (non-organic), while coal will contain, by definition, less than 40%. The kerogen within oil shale is also of different organic composition than coal, which has a more matured organic makeup – i.e., it is lower in hydrogen and higher in oxygen content, as in oil shale kerogen. The kerogen in oil shale is largely of Type I and III (see Figure 7.18). Type 1 has a high hydrogen/carbon (H/C) and low oxygen/carbon (O/C) atomic



Kerogen Type	Environment	Kerogen form	Origin	Hydrocarbon potential
I	Aquatic (or Estuarine)	Alginite	Algal bodies	
		Amorphous kerogen	Structureless debris of algal origin	Oil
			Structureless planktic material, mainly of marine origin	
II	Terres-terial	Exinite	Series of spores and pollen, cuticle of leaves and herbaceous plants	Waxy oil and condensate prone
III		Vitrinite	Fibrous and woody plant fragments and structureless colloidal humic matter	Gas, some oil
				Mainly gas
IV		Inertinite	Oxidized, recycled woody debris	"Dead carbon"

FIGURE 7.18 Kerogen types. Van Krevelen diagram, showing the four types of kerogen and how their elemental compositions change as they mature. Type I kerogen (algae and bacterial remains; rich in lipids, especially long-chain aliphatics; oil prone; H.-rich, O.-poor). Type II kerogen (planktic and bacterial remains with some higher plants; most common type; gas prone; O.-rich; H.-poor). Type III kerogen (land plants, i.e., vascular plants; rich in aromatic and poor in aliphatic structures; gas prone; poor oil potential). Type IV (carbonized material; hydrogen-poor constituents).

ratio and is derived mostly from the algal lipid matter (i.e., fats and fatty acids), rather than from carbohydrates, lignins, or waxes. Some kerogen in oil shale may also be of Type III, i.e., formed from vascular plant debris (see Figure 7.18). Metals like V, Ni, U and Mo are often enriched in oil shales.

Oil shales form in environments (both lacustrine and marine) where organic matter is abundant, and anaerobic or reducing conditions prevent its oxidation and total bacterial decomposition. Most oil shales are formed in stratified water bodies where oxygenated surface waters allow plankton to bloom, and where the anoxic bottom waters below, preserve the organic matter. The oil shales formed in lakes or swamps are often associated with impure cannel or bog head-type coals, tuffs and other volcanic rocks, or even evaporites. Many oil shales deposited in large lakes are carbonate-rich types and tend to have high oil yields, due to the enhanced preservation potential of organic material in lake environments. Oil shales deposited in marine environments are of the silica-rich type and have lower oil yields, although some Tertiary and Mesozoic siliceous oil shales have high oil yields. Marine oil shales deposited in shallow seas, on continental platforms, and basins are known from the Devonian of eastern and central North America, the Jurassic of Europe, and the Miocene of California.

7.5.3.3 Petroleum and Natural Bitumens

7.5.3.3.1 Petroleum

Petroleum is not a sedimentary rock but a carbon-rich, organic substance that accumulates predominantly in sandstones and carbonate rocks. Hence, it is included within carbonaceous sedimentary rocks. Petroleum forms from plant and animal organic matter by a complex process of maturation during burial that involves initial microbial alteration (diagenesis) and subsequent thermal alteration (catagenesis and metagenesis), thus forming a complex organic substance called kerogen. Kerogen undergoes thermal degradation (cracking) at burial depths exceeding ~1 km and temperatures of 50 to 120°C to form liquid petroleum, through a process called catagenesis. Liquid petroleum is

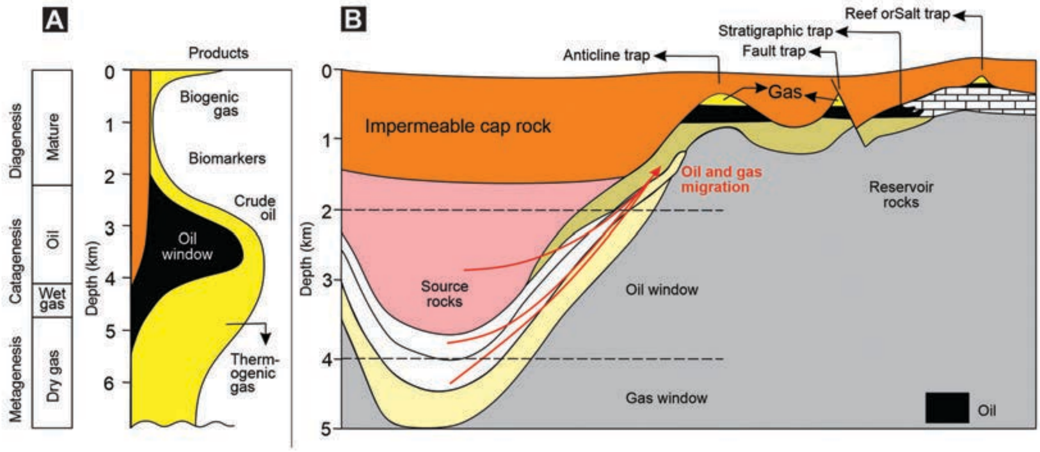


FIGURE 7.19 Formation of petroleum. Kerogen undergoes thermal degradation at burial depths exceeding ~1 km and temperatures of 50 to 120°C to form liquid petroleum, through a process called catagenesis. Liquid petroleum is subsequently cracked at temperatures ranging from ~150 to 200°C to form natural gas.

subsequently cracked at temperatures ranging from ~150 to 200°C to form natural gas (such as methane) (see Figure 7.19A).

The source materials for the organic matter that eventually converts to petroleum are contained primarily in organic-rich shales and carbonate rocks. After petroleum has formed from organic materials in these fine-grained source rocks, at substantial burial depths, it migrates out of the source rocks (through a process called primary migration) into coarser grained, porous, and permeable sandstones or carbonate rocks. It then migrates through water-filled pore spaces in these rocks (through a process called secondary migration) to structurally higher sites, where it eventually accumulates in traps such as anticlines. The rocks in which petroleum accumulates in these traps are primarily porous sandstones, limestones, and dolomites, called reservoir rocks (Figure 7.19B).

Petroleum is composed dominantly of carbon (about 85% weight) and hydrogen (about 13%), with sulfur, nitrogen, and oxygen (~2%) (Hunt, 1996). The molecular structure of petroleum is complex yielding several hundred different hydrocarbons. However, broadly three main series are noted: paraffins (alkanes), napthenes (cycloparaffins) and aromatics (arenes). Paraffins (alkanes) are open-chain molecules with single covalent bonds between carbon atoms. This includes most natural gases as well as many liquid petroleums. Napthenes (cycloparaffins) are closed-ring molecules with single covalent bonds between carbon atoms. This includes most napthene hydrocarbons. Aromatics (arenes) are one or more benzene ring structures with double covalent bonds between some carbon atoms. These are liquid petroleums with a strong aromatic odor. They make up only a small fraction of petroleums in natural crude oils.

7.5.3.3.2 Natural Bitumens

These include natural asphalts and mineral waxes that occur in a semisolid or solid state. They are black or dark brown with possess a characteristic odor of pitch or paraffin. Most natural bitumens are formed from liquid petroleums that were subjected to the loss of volatiles, oxidation, and biologic degradation after seepage to the surface. Bitumens occur as seepages, surface accumulations, or impregnations occupying pore spaces of sandstones or other sedimentary rock; they may also occur in veins and dikes. Natural bitumens, with respect to liquid petroleum, have lower percentages of carbon and hydrogen, and higher sulfur, nitrogen, and oxygen content. They are divided into two main types on the basis of their solubility in the organic solvent, carbon disulfide (CS_2): soluble bitumens and pyrobitumens. The soluble bitumens are further divided into three groups on the basis of their ease of fusibility or melting, into mineral waxes, natural asphalts, and asphaltites.

Mineral waxes are solid, waxy, light-colored substances that largely consist of paraffinic hydrocarbons of high molecular weight. Most are residuum of high-wax oils exposed at the surface. The most important native mineral wax is ozocerite (from the Greek word *ozokēros* meaning odoriferous wax; it is also called earthwax), and consists of vein-like deposits of greenish or brown wax in shales and sandstones. Montan wax (better known as lignite wax or OP wax) is an extract obtained from some kinds of brown coals or lignites. Asphaltites occur primarily in dikes and veins that cut sediment beds. They are harder and denser than asphalts and melt at higher temperatures and are soluble in carbon disulfide. Based on difference in density, fusibility, and solubility, the varieties of asphaltites are gilsonite, glance pitch, and grahamite. Pyrobitumins, like asphaltites, occur in dikes and veins but are infusible and largely insoluble in carbon disulfide. Based on the hydrogen/carbon ratio, two groups of pyrobitumins are noted. Those with H/C ratios more than 1 include elaterite and wurtzite (these are soft elastic substances). The indurated forms include albertite (a black, solid bitumen with a brilliant jet-like luster and conchoidal fracture) and ingramite. Those that have H/C ratios less than 1, include the metamorphosed and indurated pyrobitumins, impsonite, anthraxolite, and shungite.

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Section IVa

Depositional Systems: Continental Environments

Section IV (Chapters 8–21) has three subsections: continental environments (Chapters 8–12), marginal marine environments (Chapters 13–17), siliciclastic marine and pelagic environments (Chapters 18–19), and carbonate environments (Chapters 20–21).

The continental depositional environment subsection includes chapters on the following systems: fluvial, river, lacustrine, eolian desert, and glacial (Chapters 8–12, respectively).

The second subsection deals with marginal marine depositional environments and includes the following systems: deltaic, beach and barrier-island, estuarine, lagoonal, and tidal-flat (Chapters 13–17, respectively).

The third subsection deals with siliciclastic marine and pelagic depositional environments and includes shelf and oceanic (deep-water) environments (Chapters 18–19, respectively).

The fourth subsection deals with carbonate shelf environments (Chapters 20–21, respectively).

In all the chapters, a note on current classification/subdivisions, major characteristic features, associated sedimentary structures, and its environment of deposition are elaborated.

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8 Fluvial System

8.1 INTRODUCTION

The facies in continental (= terrestrial) environments are dominantly siliciclastic (i.e., a clay, silt, and sand-dominated lithology), characterized by the scarcity of fossils, and the complete absence of marine ones (Harvey et al., 2005). Most continental systems (Figure 8.1A) also have a low preservation potential as they are formed above the base level of erosion, the earth's surface above which sediments ultimately erode; in most cases, the base level is near or below the prevailing sea level. Thus, in general, the continental sedimentary rocks are less abundant than their marine and marginal marine counterparts. Nonetheless, they form an important part of the geologic record with immense economic significance such as the presence of natural gas and petroleum, coal, oil shale, and uranium; the alluvial fans are also important reservoirs of ground water (Figure 8.1). Four major types of continental environments are recognized – fluvial (alluvial fans and rivers/streams), lacustrine (lake), eolian (desert), and glacial (Figure 8.1A) (Boggs, 2006). The fluvial system is discussed in this chapter with a brief note on the fan delta, a subaqueous element (Figure 8.1B).

The Latin word "*fluvial*" means "*of, relating to, or occurring in a river*." Thus, fluvial deposits incorporate a very wide variety of sediments produced by the activities of rivers, streams, and associated gravity flow processes (see Blair and McPherson, 1994; Bogg, 2006; Nichols, 2009). The fluvial sediments are mainly deposited in river systems of humid regions, but they have also been noted in rivers within the eolian (desert) and glacial environments (see Harvey et al., 2005) (see Figure 8.1A). In general, arid environments are more fluvially dominant as compared to humid regions (see Baker, 1977). Although classified under many sub-environments, most fluvial deposits are assigned to two broad settings: alluvial fans and rivers (Figure 8.1); these two environments are often interrelated and overlapping (Collinson, 1996). This chapter is about alluvial fans (Figure 8.1).

8.2 ALLUVIAL FANS

Alluvial fans, a term introduced by Drew (1873), are more widespread in the drier parts of the world (Nemec and Postma, 1993; Gerson et al., 1993) as compared to in humid regions (Weissmann et al. 2010). However, they also occur in the arctic and alpine regions (Ritter and Ten Brink, 1986), in humid temperate regions (Kochel, 1990) and in humid tropics (Kesel and Lowe, 1987).

Ventra and Clarke (2018) noted that

Alluvial and fluvial fans are the most widespread depositional landforms bordering the margins of long-lived highland regions and actively subsiding continental basins, across a broad spectrum of tectonic and climatic settings. Their significance is relevant not only to the local morphodynamics of mountain regions and proximal basinal sectors, but also to the long-term

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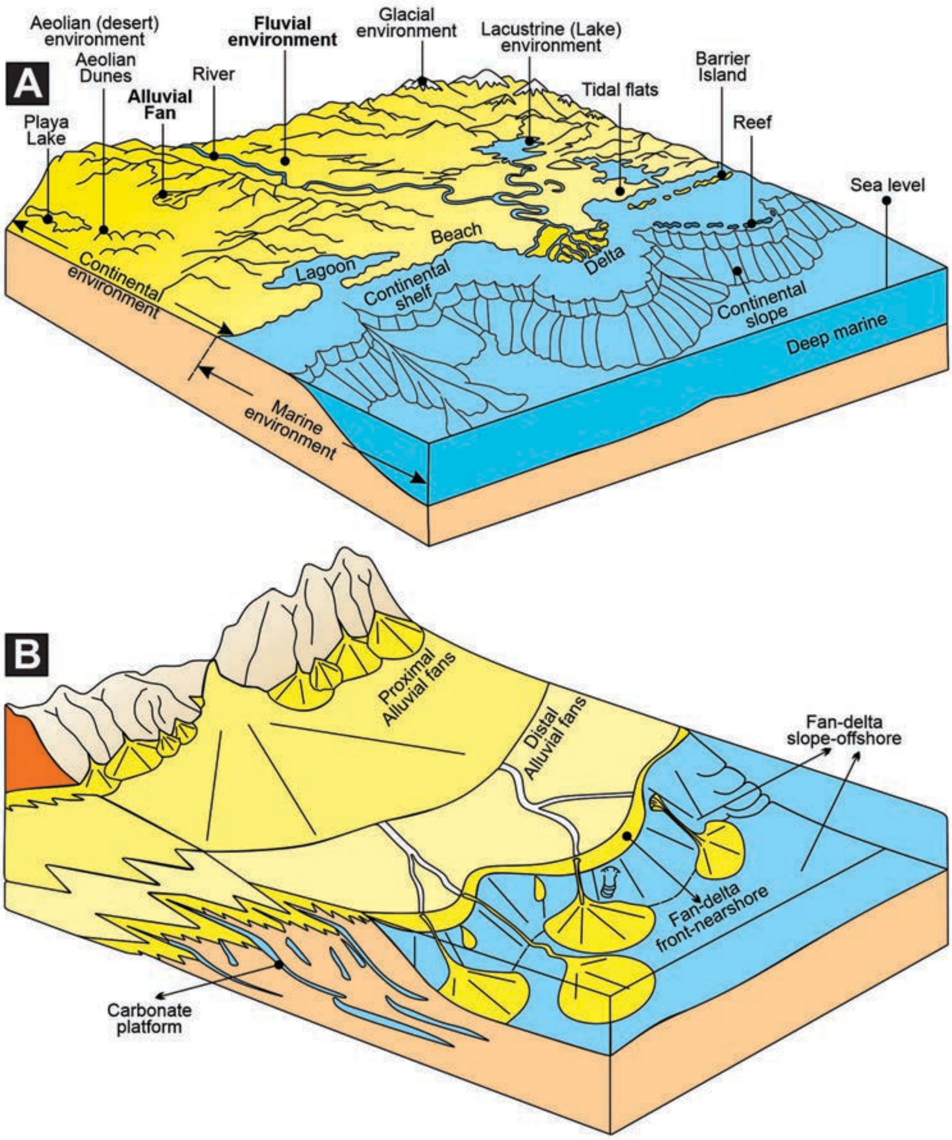

FIGURE 8.1 Continental environments and fans. A: Four major types of continental environments are noted – fluvial (alluvial fans and rivers), lacustrine (lake), eolian (desert), and glacial. The fluvial environment is discussed in this chapter (image modified, courtesy Brian Ricketts; www.geological-digressions.com). B: Location and depositional settings of alluvial fans and fan deltas discussed in the text.

evolution of sediment-routing systems, affecting the propagation of stratigraphic signals of environmental change and the preservation potential of stratal successions over much larger spatial scales than those they occupy.

The vast areal distribution, preservation potential, and varied architectural properties of alluvial fans also makes them valuable candidates with good reservoir potential and stratigraphic trapping mechanisms; they are also an important sedimentological element in exploring the geodynamics of continental basin margins (Moscariello, 2017). Hence, their better understating of the basin margin (where they occur) enables better predicting of the distribution and reservoir quality of plays and

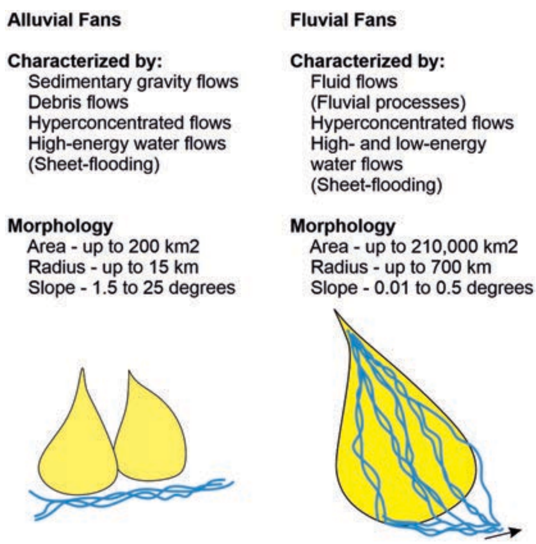

FIGURE 8.2 Major geomorphological and sedimentary characteristics of alluvial and fluvial fans. (Modified from Moscariello, 2017.)

their complex stratigraphic architectures in both mature (as in the southern North Sea, the American Midwest and South America) and in expanding oil provinces (such as in the plate interior basins of western and northern China).

However, having said that, the definition and distinction of alluvial and fluvial fans are still a matter of debate. Either they are regarded as distinct depositional landforms or as end-members of subaerial fan types, each distinguished by a set of characteristic sedimentary processes (see Blair and McPherson 1994; Hartley et al., 2010; Ventra and Clarke, 2018). Here, the alluvial fans are considered as landforms and sedimentary systems that are distinct from fluvial ones (see Figure 8.2). The much larger ones, the megafans, first adopted by Gohain and Prakash (1990), are loosely applied to fluvial fans that attain a radius in excess of ~30 km, and with a large areal extent (up to 105 km^2) (see Fontana et al., 2014). These are not considered here. The distinction between alluvial and fluvial fans is important as it directly impacts the assessment of sand body distribution, their permeability and porosity characteristics, connectivity and, ultimately in assessing hydrocarbon recovery (see Moscariello, 2017).

Based on hydrological and sedimentological processes and morphometric parameters, the two end member categories, alluvial and fluvial fans, can be distinguished (Figure 8.2). The alluvial fans commonly aggrade directly adjacent to (and abutting) their source relief and are fed by restricted catchments, often with high internal relief (Figure 8.3). They also have shorter radii (rarely up to several kilometers, typically from hundreds of meters to a few kilometers; see De Scally and Owens, 2004; Davies and McSaveney, 2008) and with higher gradients (especially in proximal areas, where they can attain slopes of up to several degrees; see Moscariello, 2005, 2017) (see Figure 8.2). Additionally, based on their origin from relatively small, poorly integrated catchments, the alluvial fans are generally affected by hydrological events of short duration, mostly independent of the climatic context in which they develop (see Figure 8.3). In sedimentological terms, this means that the runoff events dominate with highly concentrated bed loads and suspended loads (debris-flow deposits) (see Figure 8.3).

In contrast, the fluvial fans develop over much larger areas, attaining radii of several tens of kilometers and up to a few hundred kilometers (see Figure 8.4). The Pilcomayo River fan of central South America reaches ~700 km in radius and criss-crosses three countries, Bolivia,

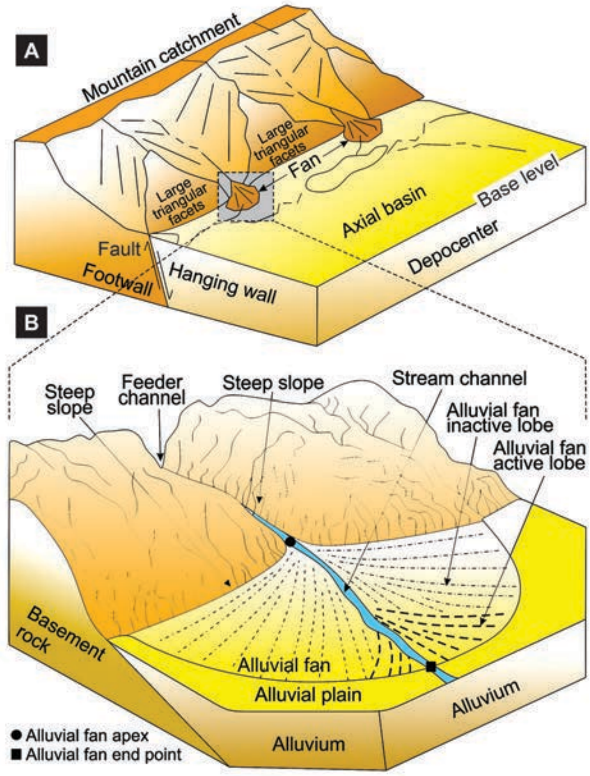

FIGURE 8.3 Location and depositional setting of alluvial fans. A: The alluvial fans commonly aggrade directly adjacent to (and abutting) their source relief and are fed by restricted (small and poorly integrated) catchments, often with high internal relief. B: Inset in A, showing major characteristics of an alluvial fan.

Paraguay and Argentina (see also Martín-Vide et al., 2014). It has a low gradient of a degree or so from the proximal to the distal areas (Hartley et al., 2010; Weissmann et al., 2010). These fans are commonly fed by extensive and well-integrated catchments, that develop over long time spans (Horton and DeCelles, 2001), thus giving rise to proper rivers along whose courses, a distinction between channels and overbank domains are well-defined, quite unlike those noted in alluvial fans (Figures 8.2 and 8.4).

8.2.1 Geomorphological Zones

The alluvial fan system is marked by three geomorphological zones, erosional, transfer, and depositional (Figure 8.5A). The erosional zone is marked by streams that actively downcuts, and removes bedrock from the valley floor and sides, into the stream bed (Figure 8.5B). The erosional zone contributes a substantial fraction of the clastic sediment needed for deposition in other sedimentary environments. In the transfer zone, due to the lower gradient, the streams and rivers neither actively erode, nor deposit sediments (Figure 8.5B). Thus, the zone becomes a conduit between the erosional and depositional zones (Figure 8.5). In the depositional zone, the sediments are deposited in river channels and on the floodplains of a fluvial system or on the surface of an alluvial fan (Figure 8.5). It must be noted that all the above zones may not always be present, as some systems are entirely erosional, while others may just have a transfer zone.

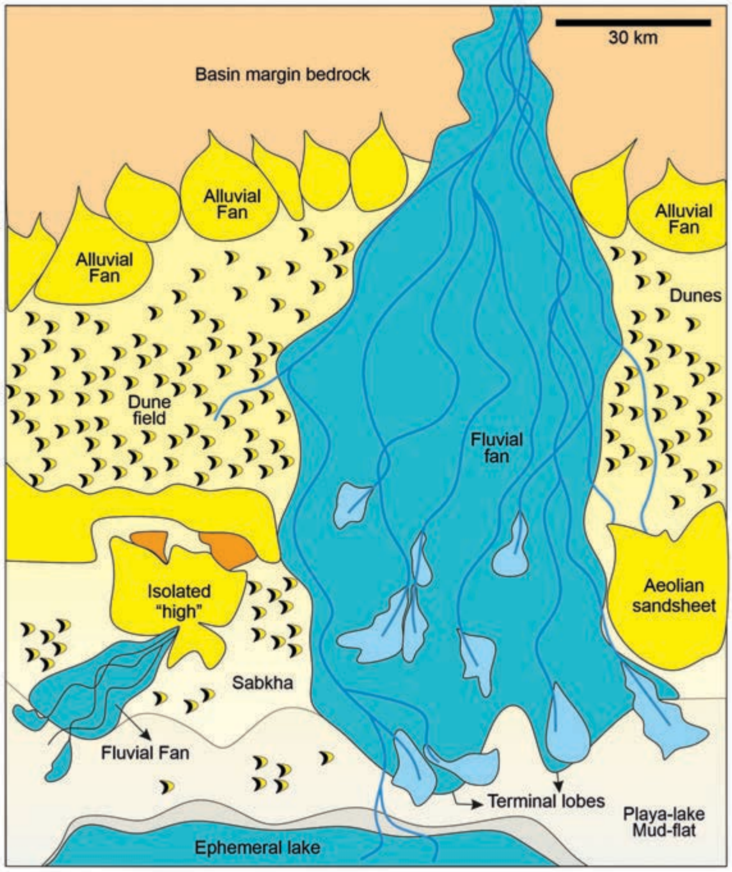

FIGURE 8.4 Depositional model, showing the interaction between typical desert depositional environment (dune field, sandsheet, sabkha, and playa) and fluvial fans discussed in the text. (Modified from Moscariello 2005.)

8.2.1.1 Alluvial Fan Zones

An alluvial fan is produced from a source area with a drainage network called a drainage basin (Figures 8.6A–B). The basin transports the erosional products of the source area to the fan in a single trunk stream that quickly loses energy and deposits the sediment load downstream or onto the axial basin system, called the toe (Figure 8.6B). This fluvial deposition by the stream is a cone-shaped deposit that radiates from the feeder canyon and forms a distinctive distributive triangular network (Figures 8.6A–B). Such branching systems are typically associated with the environments of net deposition (Figure 8.5A). This deposit, a product of repeated depositional events, in plan view, is commonly fan-shaped, where the contours bow downslope from the fan apex (Figure 8.6A). Thus, the morphology of an alluvial fan is a bridge between the erosive upland catchments (mountain catchments) where channels/streams (perennial, intermittent or ephemeral) are confined and the sediment is produced, and the down-fan basins, where channels can spread and the sediment is eventually deposited (Figures 8.6A–B).

Three zones are noted within the alluvial fan (Figures 8.6A–B). These are: fan head (variously termed as fan apex/proximal area/proximal zone), mid-fan, and the distal zone/lower (Figures 8.6A–B) (see also Blair and McPherson, 1994). Major characteristics of each of these zones are briefly enumerated below.

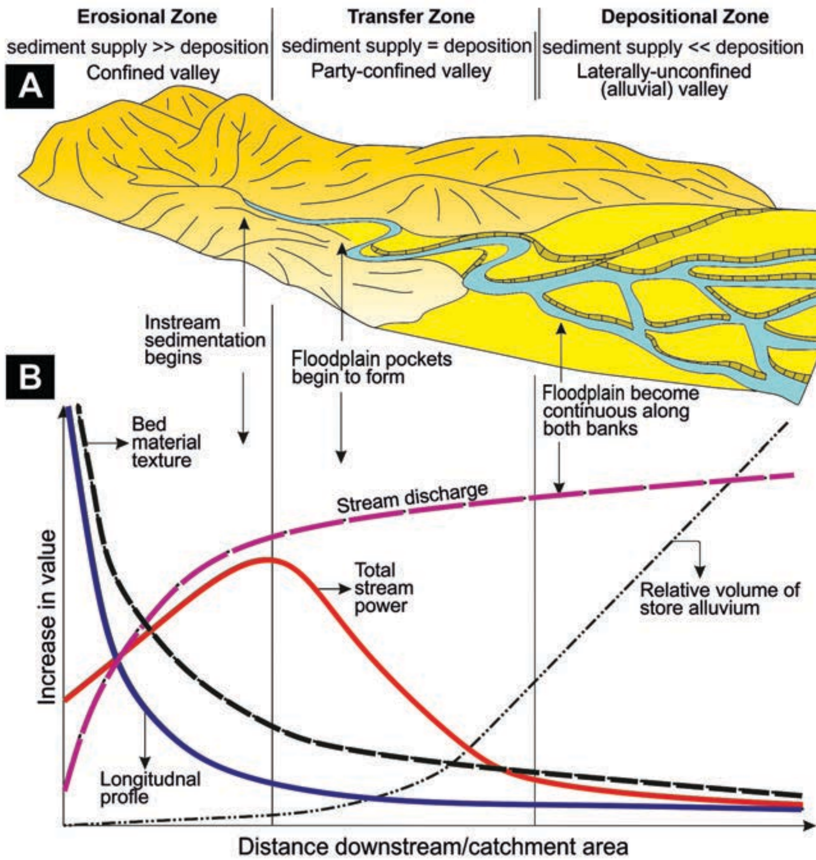

FIGURE 8.5 Geomorphological zones of the alluvial fan system and the relationship between downstream changes along a river and associated sediment processes. A: The alluvial fan is marked by three geomorphological zones – erosional, transfer and depositional. Erosional zone (sediment supply >> deposition). Streams downcut and remove bedrock from valley floor and sides by downslope movement of sediments into the stream bed. The erosional part contributes a major fraction of the clastic sediments needed for deposition in other sedimentary environments. Transfer zone (sediment supply = deposition). The gradient is lower so there is no active erosion, but also no deposition. Depositional zone (laterally unconfined (alluvial) valley). The sediment is deposited in the river channels and on the floodplains of a fluvial system or on the surface of an alluvial fan. B: Changes in bed material texture, stream discharge, stream power, and volume of alluvial across the three geomorphological zones of the alluvial fan system. (Modified from Breirley and Fryirs, 2005; Fryirs and Breirley, 2013; Church, 1992.) The relationship between downstream changes and associated sediment processes is shown as a graph between downstream distance/catchment areas vs. increase in stream power value.

8.2.1.1.1 Proximal Fan Zone

This is also referred to as fan head, fan apex, and proximal area. It is the highest point on the alluvial fan from where the surface spreads downslope, and is generally of a higher energy relative to the mid- and distal fan areas (Figure 8.6A-D). The fan head is typified by diverging channels (Figures 8.6A–B). The feeder trunk that enters the fan at the apex expands and branches into numerous parallel, subparallel, and braided channels (Figure 8.6B). The deepest entrenchments on fans occur near the apex and gradually shallow down-fan (Figures 8.6A–B). The braided-channel deposits are marked by coarse gravels with horizontal stratification, concave and planar cross-stratification, current ripples, and ripple cross-laminations (Figures 8.6C–D).

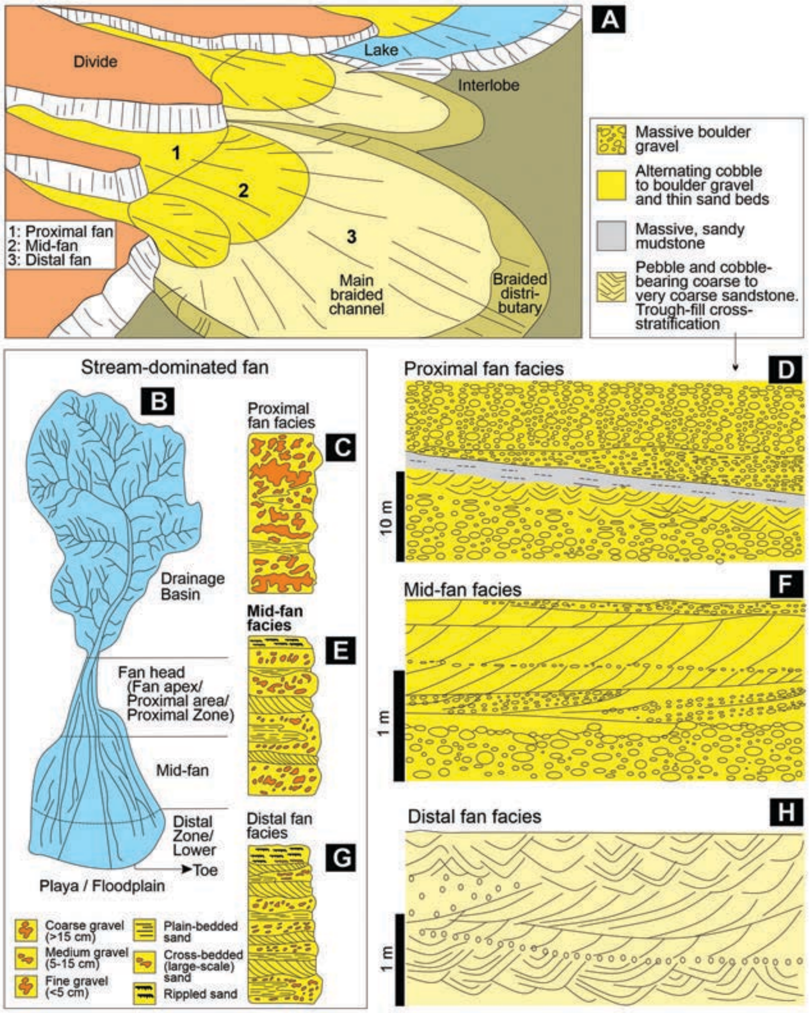

FIGURE 8.6 Alluvial fan zones, characteristic lithology, and sedimentary structures. (Modified from Collinson, 1996.) A: Alluvial fan zones and their characteristic lithology. Three zones are noted: proximal, mid- and distal. B: Drainage basin and alluvial fan zones of a stream-dominated fan. C–H: Fan facies. C–D: Proximal fan facies. This is characterized by thick, framework conglomerates that have flat bases and convex upwards top, elongated parallel to the current and flanked by cross-bedded pebbly sandstones. The longitudinal gravel bars dominate. E–F: Mid-fan facies. The conglomerates are inter-bedded with pebbly sandstones. The parallel-bedded gravels are interpreted as longitudinal bars that grow mainly by vertical accretion on the bar-top while the flanking sandstones show cross-bedding and are interpreted as transverse bars. G–H: Distal fan facies have gravels that are confined to thin beds and the lenses are scattered in tabular and trough cross-bedded sandstones, as also noted in modern outwash fans.

8.2.1.1.2 Mid-Fan Zone

The lower mid-fan to toe is the zone of deposition; it is a more stable zone (Figures 8.6A–B). The mid-fan is dominated by bifurcating, parallel, multiple flow paths and overbank sheet flow areas. The reorganization of lateral flows into a parallel braided network is typical of this zone. Repeated flow separation by longitudinal and transverse islands and bars with the coarsest gravel along their central axes has been noted (Figures 8.6E–F). In general, the gravel size is medium-grained and smaller than noted at the fan head zone with increased cross-stratification (Figures 8.6E–F).

8.2.1.1.3 Distal Fan

This is also referred to as the lower zone. This zone may grade into the top surface of an axial river terrace or a floodplain (Figures 8.6A–B). Axially derived fluvial deposits are often separated from the transversely derived alluvial fans, making two different sets of mineralogical compositions. The shallow channels are unable to effectively drain flash floods and thus, cause a wide spreading of overland sheetfloods. These sheetflood deposits are thin, relatively uniform flow veneers on the wide floodplain-like distal zone surface; typified by decreases in depth, gradient and grain size (see Figures 8.6G–H). The sheetflood deposits show a diminishing conveyance capacity, spreading layers of gravel and sand, and thus building a smooth apron of fine sediments on the fan's distal regions (Figures 8.6G–H).

8.2.1.2 Playa (Flood Plain)

It is a periodically flooded area in the axial basin, in front and down of alluvial fans (Figure 8.3A). It is the lowest-lying area and thus a local base level (Figure 8.3A). The playa may become an ephemeral lake that may completely dry up. Thus, the playa is characterized by layers of anhydrite, halite, and gypsum and with fine-grained detritus that represents the suspended fine fraction of the downstream floods, often intercalated with sand and some fine-sized gravels.

8.3 DEPOSITIONAL PROCESSES

The initial condition for depositing alluvial fans requires elevational differences between mountains and adjacent intermontane basins that operate as an accommodation space for sediment accumulation (Figure 8.3A). A fault system may control the accommodation space and cause tilting and differential subsidence of the basin (Figure 8.3A). Additionally, the rise of a marine or a lacustrine base level will also create accommodation for alluvial sedimentation. Here, it must be mentioned that the alluvial fans are subaerial features, but if they extend into water, then they are known as fan deltas (see Figure 8.1B; the fan deltas are dealt with at the end of the chapter).

Thus, the deposition on an alluvial fan (Figure 8.3) depends on (a) the availability of water, (b) the amount and type of sediment being carried from the feeder canyon, and (c) the gradient on the fan surface. The alluvial fan deposition operates by two processes: (a) sediment-gravity flows resulting in debris flows, mudflows and sieve deposits (see Figure 8.7A); and (b) stream deposits (or stream-flow/fluid-flow) resulting in sheetfloods and incised channel flows and braidplain flows) (see Figure 8.7B).

In the case of a dense mixture of water and sediments (i.e., a viscous slurry), both transport and deposition are by debris flow, that spreads out on the fan surface as a lobe (Figure 8.7A). In this scenario, the flow does not travel far and forms a small but steep alluvial fan cone (Figure 8.7A). When there is more water available, the viscous slurry gets diluted, and the deposition is then by sheetfloods (sheetflood deposition; see Figure 8.7B) or the flow is constrained to channels (stream-channel deposition; see Figure 8.7B). In this scenario, the dilute, water-laden fan deposits form fans with shallower slopes and with greater radial extent (~10 km). Galloway and Hobday (1996) provided a triangular classification diagram, categorizing fan types deposited by debris flow, stream flow, and sheetflood processes (Figure 8.8). The flow types and their operating processes are briefly enumerated below.

8.3.1 Sediment-Gravity Flows

8.3.1.1 Debris Flows

The sediment mixture has large amounts of detritus and a small quantity of water resulting in a flow that is a dense slurry similar to the consistency of a wet concrete mix (see also Blair and McPherson, 1994, 1998). This flow is laminar due to higher density and viscosity and it continues to flow as a

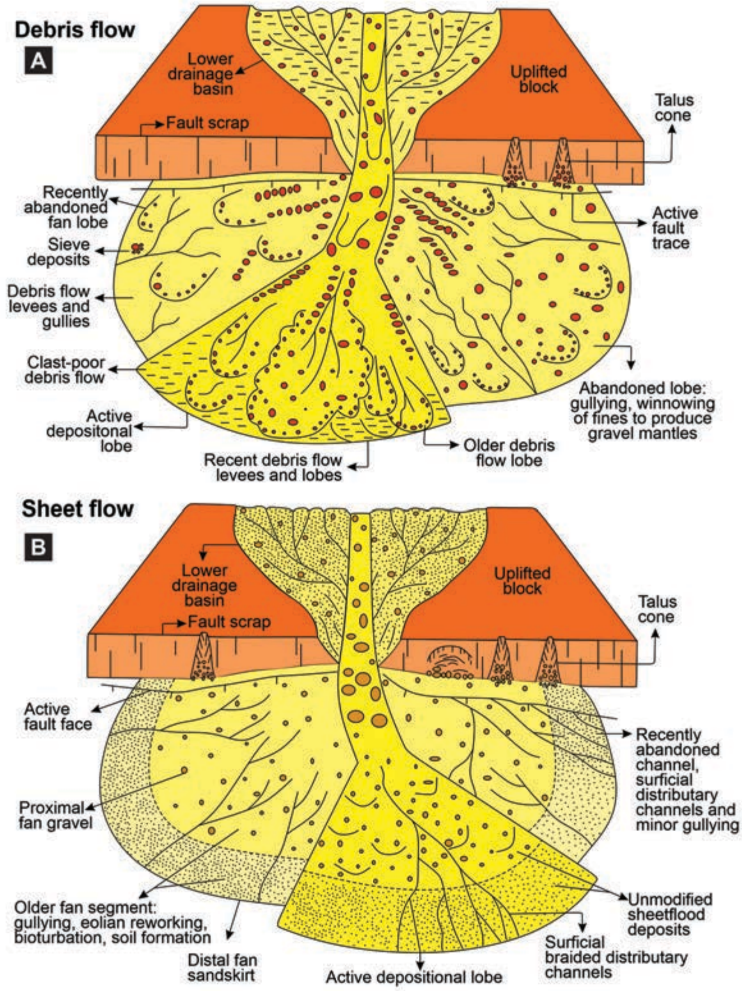

FIGURE 8.7 Drainage basin, and depositional features of alluvial fans. A: Dominated by debris-flow processes. B: Dominated by sheetflood processes (i.e., dominated by water flows). (Modified from Blair and McPherson, 1994.)

viscous mass until either the gradient decreases or the flow loses water (see Figures 8.7A, 8.8 and 8.9A). According to Nichols (2009), the major characteristics of a bed deposited by a debris flow (Figure 8.9A) are: (a) the conglomerate generally has a matrix-supported fabric, where the clasts are not in contact with each other and are almost entirely separated by a finer matrix; (b) sorting of the conglomerate is very poor due to lack of turbulence within the flow; (c) the clast-size difference is great, ranging from clay particles to meter-scale boulders; (d) the beds are structureless with clasts randomly oriented, but may have a crude alignment parallel to the flow as noted in the basal sheared layers; (e) the deposited beds display minor thinning in the downflow direction; and (f) the deposited beds range in thickness from tens of centimeters to meter-scale.

8.3.1.2 Mudflow

A mudflow is a type of debris flow that consists mainly of sand-sized and finer sediments (i.e., a fine-grained debris flow).

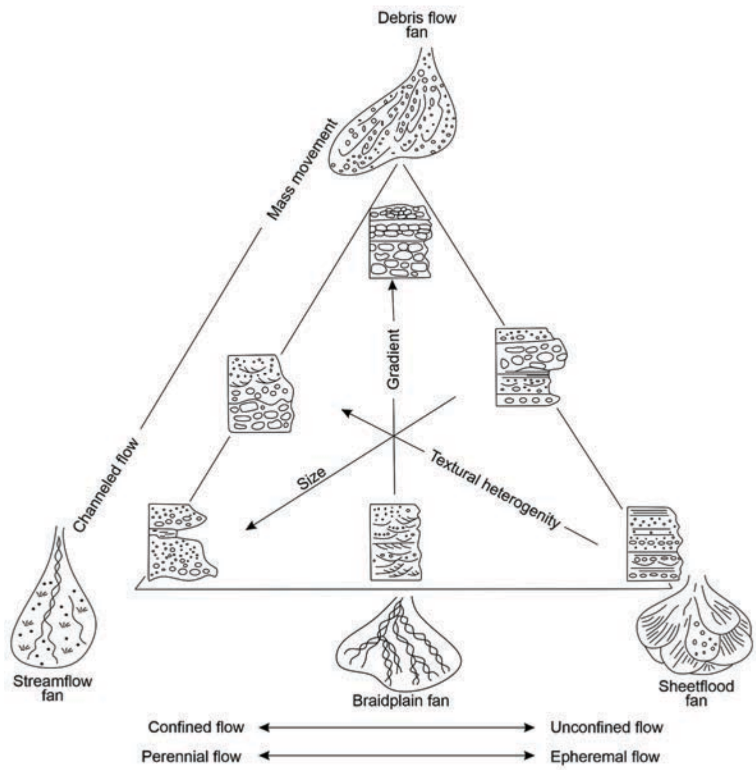

FIGURE 8.8 Triangular classification of alluvial fan types deposited by debris flow, stream flow, and sheetflood processes, as discussed in the text. Most fan systems are characterized by deposition which is a combination of these three processes. (Modified from Galloway and Hobday, 1996.)

8.3.1.3 Sieve Deposits

High rates of water percolation into the permeable fan gravel cause loss of discharge and initiate sediment deposition. When much discharge is lost, the outer parts of the fan system receive little or no water. The permeable fan surface acts as a sieve into which the water infiltrates, leaving a surface residue of coarse non-sorted, clast-supported, open-framework, matrix-poor deposit that also includes large cobbles and boulders (Hooke, 1967) (see Figure 8.7A). Sieve deposits may also form after the winnowing and removal of fine matrix out of debris flows. Sieve deposits form a characteristic hummocky surface.

8.3.2 Stream Deposits (or Stream-Flow/Fluid-Flow)

8.3.2.1 Sheetflood Deposition

When water, in the form of a heavy rainstorm, inundates the catchment area of an alluvial fan, the loose detritus is moved as bedload and in suspension, onto the fan surface (see Figures 8.7B and 8.9B). This flow spreads out as a sheetflood, a rapid, supercritical, turbulent flow (Blair, 2000), causing extensive overbank deposition in the middle and lower fan zone areas (Figure 8.9B). These laterally wide (planar) and unconfined sheet flows are shallow, a few decimeters deep and supercritical (i.e., hyperconcentrated flows; see also Blair and McPherson, 1994; Parker et al., 1998). The deposits are relatively well-sorted and clast-supported (Figure 8.9B) and consist of mud, sand, some granules, and fine pebbles; the grain size decreases downslope and away from the distributaries (see

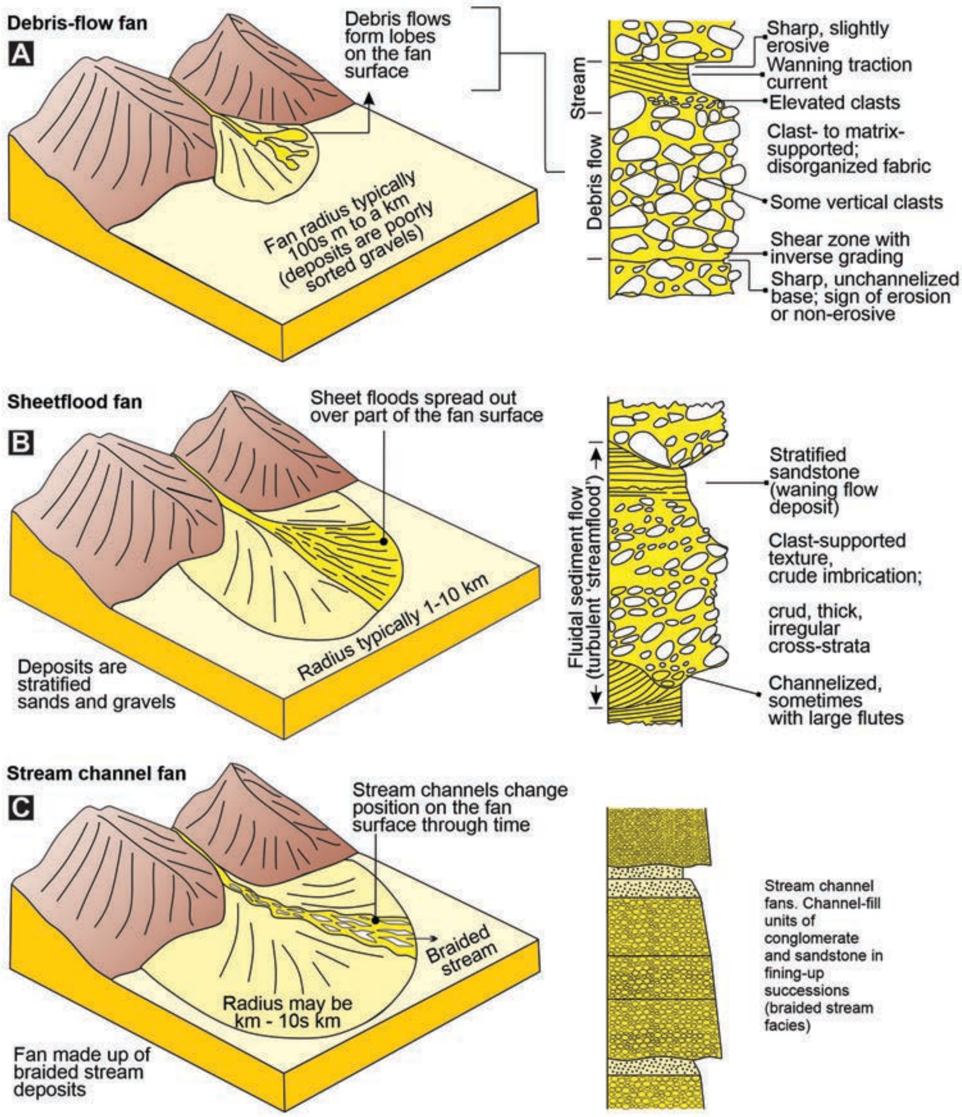

FIGURE 8.9 Alluvial fan types and their characteristic lithology. (Modified from Nichols, 2009.) A: Debris-flow-dominated. B: Sheetflood-dominated. C: Stream-channel-dominated.

also Costa, 1988). Most pebbles, cobbles, and boulders are carried as bedload, but the finer pebbles and granules are partially suspended along with sand and finer sediments.

The sheetflood deposits are composed of rhythmic couplets of alternating sand and clast-supported gravel, oriented subparallel to the fan surface with well-developed imbrication (see Figure 8.10). The most common style of bedding are the depositional couplets of coarse gravel (deposited as bedload), overlain by finer gravel and sand (deposited from suspension). These couplets are typically 5–20 cm thick and occur in packages of tens of centimeters to a couple of meters thick formed by individual flow events. Individual sheetflood deposits may be hundreds of meters wide and stretch from the apex to the toe of the fan, but individual couplets within them are typically only a few meters across.

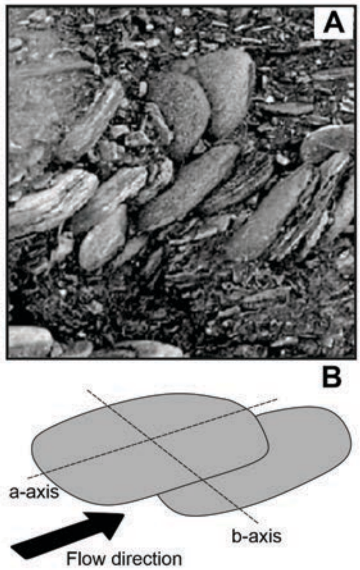

FIGURE 8.10 Imbrication. The sheetflood deposits are composed of rhythmic couplets of alternating sand and clast-supported gravels, oriented subparallel to the fan surface showing well-developed imbrication structures. A: Re-sedimented paraconglomerate deposited by density flow. B: Deposition from relatively high-viscosity fluids. The orientation results from clasts traveling with matrix and being forced by intergranular collisions into a position of least resistance to the surrounding flow.

The sheetflood deposits dominate the lower and distal parts of alluvial fans (see Figure 8.1B), with silt becoming dominant, downslope. According to Nichols (2009), the major characteristics of a sheetflood deposit are: (a) sheet geometry of beds are tens of centimeters to a couple of meters thick; (b) beds are very well stratified with distinct couplets of coarser gravel and sandy finer gravels; (c) imbrication of clasts is common; (d) sediment is poorly sorted, and silt and clay-sized particles are largely absent; and (e) beds show normal grading due to the waning flow.

8.3.2.2 Stream-Channel Deposition (Incised Channel Flow)

The abrupt reduction in gradient on the lower slope of the plain results in the deposition of gravel on bars within the channel, thus, forming a braided depositional form (Figures 8.8 and 8.9C). Repeated high-discharge events in the channel eventually choke the channel with sediments, and the active flow moves, either by a process of gradual lateral migration or by avulsion. Avulsion is the process in which flows are diverted out of an established channel into a new course at a lower elevation on the adjacent surface, resulting in a new channel, i.e., short-term flow switching within braided channels. With time, the position of the braided river channel migrates over the whole fan surface, depositing a more-or-less continuous sheet of gravel. The overall shape of the sediment body formed is similar to that of a fan formed by sheetflood deposits, but the radius is much larger, over 10s of km from apex to toe. Beds are sharp with clast-supported conglomerates fining up to sandstones (Figure 8.9C). Sedimentary structures are similar to those of a braided river, including imbrication (Figure 8.10) and cross-stratification in gravels and sandstones (cross-bedding).

It must be noted that the proportions of debris flow and water-laid deposits (stream deposits) vary greatly from fan to fan and may also change during a single fan's accumulation history. Where source area conditions are not favorable for debris flows, the water-laden sediments make up the fan deposits. Other fans consist mainly of debris flows. Most fans whose source areas produce debris flows also have flood events that result in the deposition of water-laden deposits (stream deposits). Thus, the deposits of many fans consist of inter-bedded deposits of debris and water flows in varying proportions.

8.4 MAJOR CHARACTERISTICS OF AN ALLUVIAL FAN

Major characteristics of an alluvial fan include tectonic setting, geometry, size, thickness, slope gradients, drainage, texture, facies (proximal fan, mid-fan, and distal fan facies), typical sequence, sedimentology, and fossils. These are briefly summarized below.

In terms of tectonic setting, the alluvial fans are typically noted in rifting continental grabens, foreland basins, collisional overthrust mountain belts, and highlands undergoing rapid uplift. They are associated with meandering fluvial valleys and playa lakes.

The geometry of an alluvial fan is wedge-shaped and limited in lateral extent (from only a few tens of meters to kilometers from the source highland).

The size of an alluvial fan ranges from 0.5 to 15 km in length. Megafans, the largest fans on the earth, may reach ~100 km in length and cover an estimated area of ~10,000 km^2.

The thickness of an alluvial fan ranges from a few tenths up to a few hundred meters. But, if subsidence continues, thickness can exceed up to 7000 m or more.

The slope gradients of most alluvial fans are in the 1 to 6° slope range. Discharge controls fan slope; the greater the discharge, the gentler and flatter the slope. Fans in humid regions have lower slopes than those in arid regions.

The drainage processes in an alluvial fan operate either along incised channels or in a sheet-like manner. The distributary system changes from a single channel at the apex (proximal zone; fan head) to a braided-channel pattern (distal zone; down-fan). Channel size, flow depth, and velocity reduce down-fan; channels become shallower, discharge decreases, and the sediment transport capacity is greatly reduced.

The alluvial fans have no definable texture. Due to the proximity to the source area and coupled with the short distance of transportation, textural immaturity is a characteristic feature of fan deposits. The clast size varies greatly, from clay-sized sediments to boulders, with a rapid size decrease downcurrent (i.e., size decreases in the down-fan direction). However, most fan sediments are typically coarse-grained and consist of pebble to meter-sized boulders, called fanglomerate. Both conglomerates (i.e., fanglomerates) and cross-bedded sandstones are very common in alluvial fans. Generally, the similarity in sediment texture from fan head (apex) to toe suggests lack of significant transport-related processes. Sorting slightly improves down-fan but, more often than not, remains moderate or very poor. Normal grading (i.e., fining upward), is commonly noted in clast-supported fan deposits; it results from a gradual decrease in sediment supply or from reducing flow velocities. Inverse grading (coarsening upward: i.e., the particle size increases stratigraphically upwards) is noted with stratigraphic thickening and corresponds to a prograding (forward moving) gravel front, i.e., when coarse lobes from the proximal fan zone prograde downward and buries parts of the mid-fan and the down-fan areas. The proximal fan zone (fan head) shows an immature, coarse texture, often with boulders several meters in size.

The alluvial fan facies is identified based on grain size, matrix composition, sorting and packing, clast or matrix support, stratification or massiveness, upward fining or coarsening, and clast imbrication (see Miall, 1978). Based on this, three sub-environments are noted – proximal, medial, and distal (see also Blair 1999) (see Figure 8.6). In the proximal fan facies (fan head), the sediments are stacked against a fault face and accumulate as a massive, structureless pile or with crude stratification. They are disorganized, poorly to extremely poorly sorted, and without any preferred orientation. Boulder to pebble-sized clasts dominate with occasional outsized boulders. An open-framework clast-supported mass of gravel is noted. Erosion and reworking dominate rather than deposition. In the mid-fan facies (mid-fan), the sediments are intermediate between the coarser fraction of the apex and the finer one of the down-fan. The alternations of coarse and fine deposits, and an increase in the finer-grained sheet deposits are noted. In the distal fan facies (distal zone), a transition from channel flow to shallow sheet flows is noted. A well-stratified, sheet-like, often laminated structure is commonly noted.

A typical sequence of an alluvial fans is composed of coarsening-upward sequences of cross-bedded sandstone, channel-lag conglomerates, and unsorted debris-flow deposits. Rarely, a fining-upward sequence forms during the decay of the fan.

In terms of sedimentology, the debris-flow deposits are unsorted, often showing reverse grading, and containing boulders. Sieve deposits form conglomerates with no matrix of finer particles. The sediments can be very immature and angular, with abundant coarse rock fragments and feldspars. Lenticular bodies composed of cross-bedded channel sand and pebble conglomerates formed by channel cut-and-fills are noted; they are most common near the apex, and decrease downslope. Paleocurrents radiate from the apex. Ripple marks and convolute laminations occur in finer-grained sheetflood deposits which are typically oxidized, so red beds are common.

8.5 FOSSILS OF ALLUVIAL FANS

The terrestrial environment has a poor potential for the preservation of flora and fauna (fossil plants, and animals, respectively) due to the scavenging by carrion or due to the fact that the tissue breaks down quickly by oxidation. Only very resilient parts of an organism (such as teeth and bones of vertebrates) are preserved or if the plant or animal is covered by sediment soon after its death. Faunal remains are therefore relatively rare, but plant fossils are more common and may even be locally abundant. Fossilized tree stumps may be preserved in situ (in place) in overbank deposits due to floods that partially bury the tree; pieces of branches and leaves occur within beds of both channel and overbank sediments. However, the most abundant plant fossil remains are those of pollen and seeds (palynomorphs); they are highly resistant to breakdown and survive long periods of transport before being deposited and preserved. This makes them very useful for dating and the correlation of terrestrial deposits. The footprints of animals in soft mud may be preserved if the mud dries hard and is later covered with sand; these are largely restricted to flood and alluvial plain deposits.

8.6 DATING OF ALLUVIAL FANS

Alluvial fans are the product of merging lobes of different ages. The age of a fan surface marks the end of the merging of lobes and the approximate time that had elapsed since the start of trenching and abandonment of the fan surface. A wide scatter of ages over one fan results from the time needed to gradually tie lobes together into one alluvial fan unit. Thus, fans are products of merging lobes, making the fan a diachronous unit. Most fan ages spread from younger than 10,000 to mid-Pleistocene. The absolute dating of an alluvial fan is done by four methods: (a) $^{230}Th/^{234}U$ of secondary carbonate ($CaCO_3$); (b) radiocarbon geochronology (on organic material); (c) terrestrial cosmogenic nuclide (TCN); and (d) optically stimulated luminescence (OSL). Relative dating techniques include dating by rock varnish on clast surfaces, and pavement properties.

Fan surfaces are dated either by $^{230}Th/^{234}U$ of secondary carbonate (i.e., carbonate precipitated from the soil solution rather than inherited from a soil parent material), or by radiocarbon geochronology, if datable organic material is available. Cosmogenic nuclides are rare radioactive isotopes created when high-energy cosmic rays interact with the nucleus of a solar system atom (see also Ivy-Ochs and Schaller, 2010). Rocks initially contain no cosmogenic nuclides. Over time, cosmogenic nuclides accumulate in the rock, and their concentration is proportional to the time of surface exposure (Lal, 1988). In hyper-arid regions where datable material, whether organic matter or pedogenic carbonate, is scarce, the optically stimulated luminescence (OSL) method is preferred (see Porat et al., 2010). When minerals are buried and shielded from light, the impact of natural radioactivity builds up a luminescence signal, proportional to the time spent in darkness, which is used for dating. The maximum age range of the OSL dating technique is limited to about 300,000 to 350,000 years (see Murray and Olley, 2002). The luminescence age of mineral grains underlying a fan surface represents their last exposure to sunlight before burial by the uppermost surface fan unit which is somewhat younger but close in time.

Relative dating is done by using rock varnish on the surfaces of the clast. With time, the gravel on the fan surface becomes coated with desert varnish which consists of layers of hydrous ferric manganese oxide, iron oxides, clay minerals, and trace elements. The Fe oxide typically amounts to less than 1%. Its percentage of cover, and the darkness of its varnish, increase systematically with age. Being time-transgressive, the fan surface provides useful features as pavement properties for estimating the relative fan age (Amit and Gerson, 1986). An approximate timescale of at least 100,000 years for the development of a mature smooth pavement surface with fully shattered small clasts and with a thick pedogenic carbonate phase has been noted. In active channels, recently abandoned bar sediments carried ages of 500–5000 years (Porat et al., 2010). Carbonate encrustation is another characteristic feature of a semi-arid fan surface. The higher and older the fan surfaces are, the better preserved the carbonate encrustations become.

8.7 CONTROLS ON ALLUVIAL FAN DEPOSITION

The development of alluvial fans is influenced by the setting of the fan and related processes. The setting of the fan includes site topography, which is a function of tectonics and its long-term geomorphic history, and thus, controls accommodation space. Related processes include: (a) the delivery of water and sediment from the mountain catchment; (b) its transport to and deposition on the fan; and (c) the potential for erosion of the fan surface. These processes are in turn controlled by the size and relief of the catchment and catchment area, bedrock geology of the catchment, climate, and the fan morphology itself, including its relationship to local base levels (see Figure 8.3). All these can be grouped into four interrelated factors: catchment and catchment area, bedrock (or basement rock) properties, tectonics, climate, and the base-level effect.

8.7.1 Catchment and Catchment Area

The alluvial fan morphology is often controlled by the characteristics of the contributing catchment area that changes very little over short time intervals (see Figure 8.3). However, short-term changes may occur in response to headwater stream capture (Mather et al., 2000) causing sudden changes in water and sediment supply, or in the long-term by changes in sediment availability related to the progressive erosion of the source area. The larger the area of the catchment, the greater the potential to store sediment within it and, therefore, the more likely a decrease in the total amount of sediment delivered to the fan per transport event. By contrast, smaller catchments have less potential for sediment storage and so are more likely to deliver sediment to the fan surface for any given transport event. However, the increased potential for discharge in larger catchments, given the larger surface area for precipitation to fall over, can also lead to the high delivery of sediment in some instances (Allen et al., 2013). Excess sediment supply leads to local sediment deposition and therefore fan aggradation, whereas excess power will lead to erosion and fan degradation (reduction). In tectonically active settings, where the source highlands are uplifted with respect to the adjacent basin, the alluvial fans tend to aggrade as accommodation space is continuously created, resulting in fans with a relatively small area with respect to their catchment area (Viseras et al., 2003). Hence, fan area-catchment area relationship is an important parameter that needs to be considered in the context of local tectonic settings (Whipple and Trayler, 1996; Allen and Densmore, 2000).

8.7.2 Bedrock Properties

The lithology and mechanical properties of the bedrock (or basement rock) underlying the catchment (see Figure 8.3) influence the volume and quality of sediment produced. Catchments underlain by rock types that are less resistant to erosion tend to produce alluvial fans that are larger in area than those produced by basins with more resistant rock types (Bull, 1962; Hooke, 1968). However, it has also been noted that larger fans are supplied by basins underlain by more resistant rocks as these have feeder channels that flow in steep, narrow canyons with little sediment in storage, such that most of the sediment is ultimately delivered to the aggrading fan surface (as noted in the White Mountains of California, USA; see Lecce, 1991). It has also been suggested that where the sediment production dominates, catchments with relatively more erodible lithologies produce more extensive fans, but, where the sediment storage factor is dominant, basins with less erodible lithologies produce smaller fans (Mills, 2000).

Bedrock lithology also influences sediment transport processes responsible for fan aggradation; where catchments produce significant volumes of clay- and silt-sized debris, sediment-water mixtures will reach the fan surface in the form of debris flows (Levson and Rutter, 2000; Moscariello et al., 2002) rather than unconfined water flows, affecting not only the primary architecture of the fan deposits, but also the overall geometry of the system (De Haas et al., 2016). It has also been noted that lithologies that weather to form a lot of mud will tend to generate muddy debris flows, whereas more resistant rocks will break down to sand and gravel, which is transported and deposited by sheetflood and stream-channel processes (see Blair, 2000; Nichols and Thompson, 2005). However, despite the various controls exerted by catchment morphometry and geology, over long timescales, these factors tend to be dominated by allogenic controls, such as tectonics and changes in base level and/or climate.

8.7.3 Tectonics

Tectonics causes uplift of the source area leading to an increase in sediment production, or change in fan gradient, by tilting (see Figure 8.3A). At the regional scale, uplift-induced dissection causes a change in the base level, thereby triggering increased fan-toe erosion. As alluvial fans develop at the margins of sedimentary basins that are also sites of tectonic activity; faults along the basin margin create uplift of the catchment area and subsidence in the basin (see Figure 8.3A). Being deposited atop a tectonically mobile zone, the alluvial fans undergo tectonic deformation. Tectonic uplift enhances bedrock erosion by supporting the capacity to detach and transport clasts. But, to ensure a continuous fan growth, a longtime trend of uplift or sinking is essential.

Tectonics determines the location of alluvial fans by creating sharp topographic differences. Strike-slip faults transfer their motion to the overlying fans, causing the fan segments to be shifted laterally. The amount of such horizontal slips can be determined by the offset of geomorphic markers. Any morphologic surface direction that is at variance with the radial channel pattern and significantly disturbs fan concentricity might be due to tectonic deformation. Hence, evidence of tectonic activity within an alluvial fan succession is noted, such as an influx of coarse detritus onto the fan, resulting from renewed tectonic uplift (see Nichols, 1987, 2009). Analysis of bed thicknesses and clast sizes within beds has been used as a means of identifying periods of tectonic uplift in the high ground adjacent to the basin.

The creation of accommodation space brought about by the continued basin subsidence is the most important role played by tectonics for basin-margin fans. When the rate of delivery from the hinterland exceeds the rate of creation of space to store the sediments, fan gravel progradation is enhanced. Sediment coarsening upward points to gravel progradation and is usually related to tectonic uplift events. For Quaternary fans, tectonics is an important control on both fan sedimentation (Calvache et al., 1997) and geomorphology (Bull, 1961; Silva et al., 1992); it is also considered the primary control in ancient sedimentological records (see Heyward, 1978).

Alluvial fans are often located in tectonically active areas and thus, tectonics is often considered to be the primary control in dictating both the location and morphology of fans, producing the setting, relief and accommodation space necessary for alluvial fan growth (Whipple and Trayler, 1996; Allen and Hovius, 1998). Additionally, on an alluvial fan, periods of rising base level generally leads to increasing accommodation and deposition, whereas lowering of the base level may lead to a reduction in the available accommodation space, resulting in erosion and/or bypass at the fan surface (see Harvey, 2012, 2013) (see also Figure 8.3).

8.7.4 Base-Level Effect

Base level is an imaginary horizontal surface to which the fluvial system strives to erode (see Figure 8.3). The base level of many alluvial fans is an inland, local, intermountain basin. Its margins, which are typical sites of fan deposition, are not base-level loci. A falling base level will trigger entrenchment only if the new exposed area is steeper than the threshold gradient for erosion. Most fans accumulate under relatively stable base-level conditions. A change in base level causes a switch from deposition to erosion in the fan-toe zone that progressively dissects the fan headwards. Base-level changes are either induced tectonically or as secondary effects of climatic change. These may be important on coastal fans following eustatic sea-level changes, or on fans at the margins of pluvial lakes following changes in lake level. However, the effects of base-level change are not necessarily straightforward. In the common case of a fall of base level, dissection will occur only if the newly exposed gradients are sufficient to trigger incision. On the other hand, a regional sea-level rise may cause fan dissection if it is accompanied by coastal erosion and fan profile foreshortening (see Harvey et al., 1999a). The effects are similar to erosional “toe-cutting” by a laterally migrating axial drainage (Leeder and Mack, 2001). The interaction between tectonic, climatic, and base-level factors is hotly debated but the consensus is that, at least for Quaternary fans, climate appears to play a primary role (Ritter et al., 1995).

8.7.5 Climate

The processes on the fan and the resultant morphology responds to climatically-led sediment supply and flood hydrology regimes (Harvey et al., 2005). Climate influences sediment availability within the catchment area through weathering, and sediment delivery from hill slopes to channels through slope failure and slope erosion processes. Climate also influences water and sediment supply to the fan through its influence on the flood hydrology of the stream system. A change in climate results in changes in the processes of deposition; for example, with an increase in rainfall, more water is available, and this results in the predominance of sheetflood and stream-channel processes, with fewer debris-flow events. The character of conglomerates deposited on the fan will reflect this climatic change, with more clast-supported and fewer matrix-supported conglomeratic beds.

Climatic variations within tectonic cycles also produces sufficient fluctuations in water and sediment discharge to change trends from aggradation to entrenchment. The sedimentary and the morphologic signatures following climatic events may be similar to those created by tectonics. However, regional synchroneity of the fan behavior is the signal implying climatic forcing. The climate also controls the vegetative cover and the water/sediment yield ratio that determine whether a fan will undergo aggradation or degradation. Vegetation in the sourcing area maintains slope stability by favoring infiltration and inhibiting runoff thus decreasing sediment supply delivery to the fan. Dry climates, deforestation, overgrazing, and soil tillage diminish soil moisture and vegetation density, thus, increasing slope erosion in the source area and thereby enhancing fan growth. Fan aggradation has also been linked to wetter conditions resulting in greater runoff generation and slope instability causing increased sediment supply. Significant changes in erosion-sedimentation regimes resulting from Quaternary climatic changes have been identified in many dry regions including the

American southwest (Dorn, 1994; Harvey et al., 1999b), and the drier parts of the Mediterranean region (Roberts, 1995; Harvey, 1996).

Climate change tends to act principally on geomorphic processes within catchments, controlling the spatial and temporal distribution of erosion, responsible for the primary sediment supply to the fan, as well as modulating the hydrological regime, and therefore, the discharge and stream power (increasing erosion). The climate also affects the vegetation cover in a catchment and on the fan surface, which, in turn, exerts a strong influence on the patterns of sediment yield and transport (Dorn, 1996). Debris flows are one of the most important formative processes for alluvial fans, capable of transporting large amounts of water and debris in very short time periods (D'Agostino et al., 2010). They can also result in significant modifications to alluvial fan topography, both during and after an event (Scheinert et al., 2012). Topographic changes, in turn, affect the magnitude, trajectory, inundation and runout length of subsequent sediment transport events (Volker et al., 2007; De Haas et al., 2016).

8.8 GROUNDWATER

Alluvial fans in piedmont areas are major reservoirs of groundwater. Fan aquifers are typically complex systems composed of multiple aquifers and aquitards. The highest infiltration rates on alluvial fans are at the apex zone. Farther down-fan, where the beds become thinner and sorting improves, there is decrease in hydraulic conductivity. The aquitards act as partial barriers to infiltration, confining vertical percolation, thereby directing the groundwater flow laterally into separate strata, thus, building individual pressurized horizons. As groundwater carries part of the granular load, withdrawal of groundwater from the fan aquifer causes compaction of the underground grain material and thus reduces pore volume. The removal of groundwater from the fan aquifer results in consolidation of deposits and causes ground subsidence. Changes in groundwater level have also been noted to accompany earthquakes.

8.9 PEDOGENIC PROCESSES

Soils in the fan's stratigraphic column testify to a period of surface exposure and stability during the depositional phase. This happens when deep entrenchment leaves parts of the fan surface inactive for longer periods of time. The paleosol thus marks a depositional hiatus, i.e., an inactive period within the depositional phase. In extremely arid to semi-arid climates, the typical product is the Reg soil that is marked by a vesicular, fine-grained Av horizon and a B horizon enriched in carbonate, gypsum, or salts. Gravel shattering by salts is found both on the fan surface and in the B and C horizons. For a further discussion on soils, see Chapter 1.

8.10 FAN DELTA

An event within the alluvial system, such as torrential rain in the hinterland may be felt almost instantaneously in the deep-water part. Hence, both the coastal and the deep-water systems are highly influenced by the alluvial system. More so, these systems are also closely associated in space, perhaps by only 1–10 km apart, and their associated facies are also closely stacked.

8.10.1 Fan Delta Classification

All coarse-grained deltas are called fan deltas, and are defined as coastal prism of sediments delivered by an alluvial fan and deposited either in a lake or in the sea (Figure 8.1B) (Holmes, 1965; McPherson et al., 1987; Nemec and Steel, 2009; Dart et al., 1994), i.e., they are mainly or entirely subaqueous, formed at the interface between the active fan and a standing body of water (Figure 8.11) (see also Nemec and Steel, 1988).

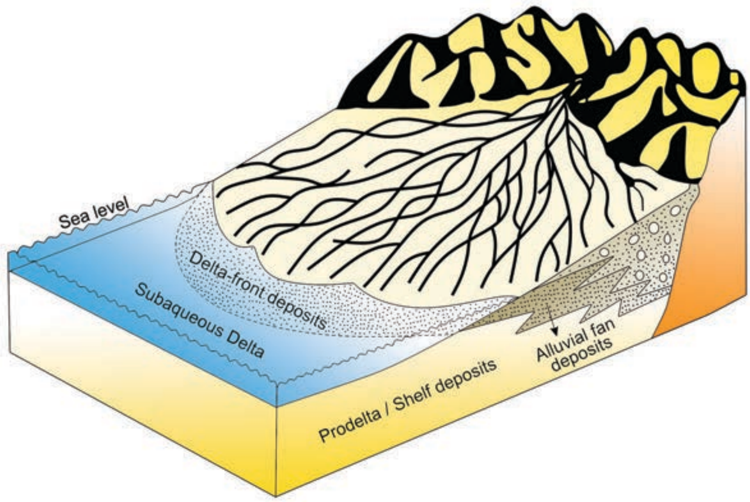

FIGURE 8.11 Fan delta. These are distinguished from alluvial fans by the presence of a basinal water body or by the evidence for the interaction of alluvial with marine or lacustrine processes.

The fan deltas are distinguished from alluvial fans by the presence of a basinal water body or by the evidence for the interaction of alluvial with marine or lacustrine processes (see also McPherson et al., 1987) (see Figure 8.11). The fan deltas are composed of: (a) alluvial fan deposits that constitute the subaerial proximal and mid-fan environments and, in most cases, make up the largest fraction of the fan delta; (b) a transition zone where the fluvial environment interfingers with coastal and littoral processes characterized by sediments with good sorting, high roundness, seaward imbricated gravel, swash lamination, horizontally bedded berm sands, or landward dipping backshore deposits; and (c) the subaqueous zone of the fan delta often covers an area several times larger than the subaerial fan; both aggradation and progradation increases the size of the fan delta.

The fan-delta system, based on the water depth of the basin and the gradient of the delta profile (see Ethridge and Wescott, 1984; McPherson et al., 1987; Massari and Colella, 1988; Postma, 1990), are of three types: (a) shelf-type, low-gradient, shallow-water deltas; (b) slope-type, deep-water deltas that have low to steep gradient, depending on the suspended load/total load ratio; and (c) Gilbert-type, steep gradient deltas that form in both shallow and deep waters. These are briefly described below.

8.10.1.1 Shelf-Type or Shallow-Water Fan Delta

Shelf-type or shallow-water deltas encroach on to low-gradient shelves with very shallow water depths at and near the river mouth, and hence, these types of fan deltas are very sensitive to fluctuations of the base level (Figure 8.12A). They have three physiographic zones: (a) delta plain or subaerial fan delta of alluvial setting (proximal to distal fan delta); (b) transition zone or delta front affected by waves; and (c) prodelta, occurring below wave base (Figures 8.12B–C), containing hemipelagites (Figure 8.13); in very shallow-water basins, the prodelta may not exist. A well-developed coarsening-upward sequence is noted (Figure 8.12C) (see Wescott and Ethridge, 1990) as in the Copper River fan delta, Gulf of Alaska, USA (see Galloway, 1976). In general, the shelf-type fan deltas gradually coarsen upwards from prodelta muds, either to shoreface and beach sandstones (if wave activity is high), or directly to distal fan-delta sandstones (if wave activity is low) (Figure 8.12C).

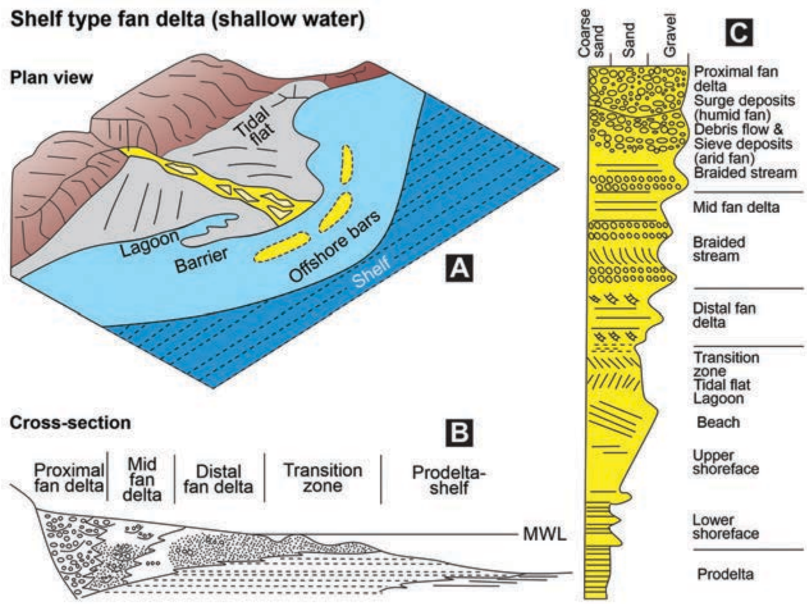

FIGURE 8.12 Shelf-type or shallow-water fan delta. A: Depositional setting. These deltas encroach onto the low-gradient shelves with very shallow water depths at and near the river mouth. B: Physiographic zones of shelf-type fan delta. C: Sedimentary sequence of a shelf-type fan delta.

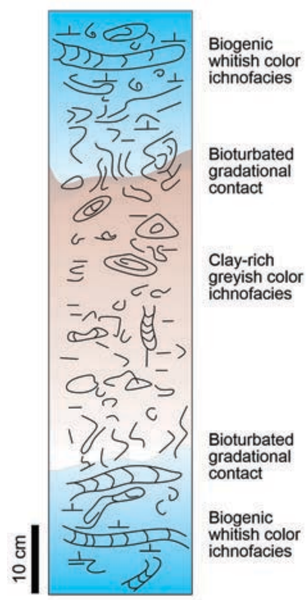

FIGURE 8.13 Hemipelagite facies. (Modified from Stow and Smillie, 2020.) The idealized model shows compositional cyclicity between clay- and biogenic-rich fractions. These variations depend on the input of different components. The hemipelagite are composed of fine-grained sediments, typical of marginal outer shelf and slope settings, and occur widely over the shelf, slope, and marginal basins. They are marked by indistinct bedding, gradational contacts, and cyclicity in their composition and color. There are no primary sedimentary structures but a pervasive bioturbation with a high diversity suite of trace fossils of the *Zoophycos-Nereites* ichnofacies (notably those of *Zoophycos*, *Chondrites*, *Planolites*, *Thalassinoides*, and *Phycosiphon*).

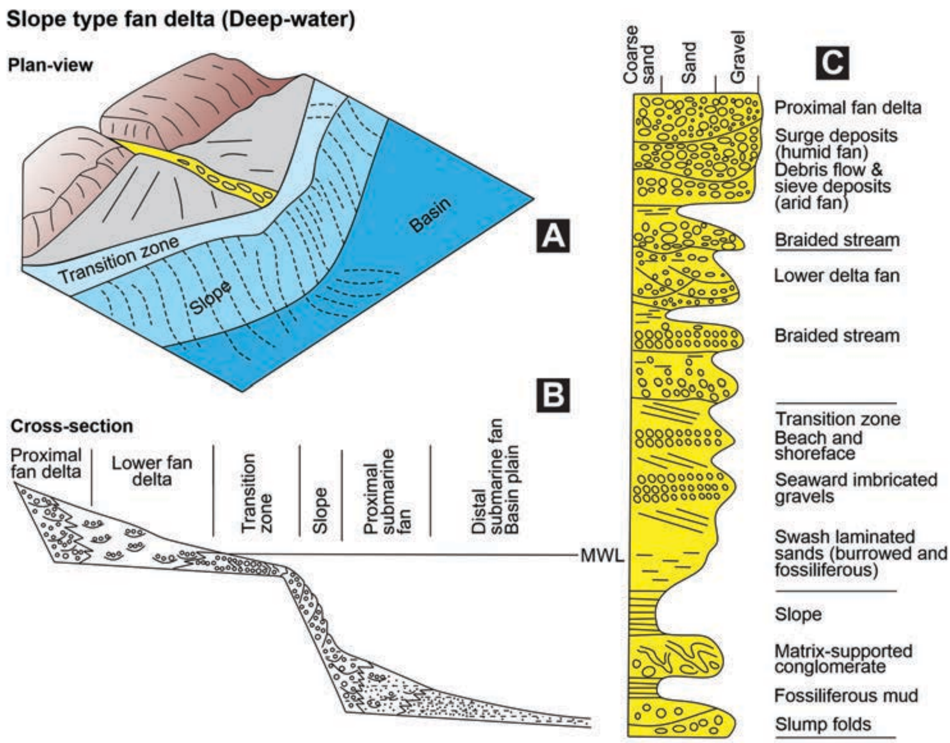

FIGURE 8.14 Slope-type, deep-water fan delta. A: Depositional setting. The deep-water fan delta fan has a slope separating a poorly developed delta front from the prodelta with a deep-water fan system that is often mud-dominated. B: A pronounced shelf/slope break is characteristic of a deep-water fan delta. C: Deep-water fan delta sedimentary sequence.

8.10.1.1.1 Hemipelagites

Hemipelagites are fine-grained sediments, typical of marginal outer shelf and slope settings (Figure 8.13). But in general, they are widespread over the shelf, slope, and marginal basins. They are a mix of biogenic pelagic (>10%) and terrigenous or volcaniclastic material (>10%) in which a large proportion of the terrigenous (or volcaniclastic) fraction (>40%) is silt-sized or greater (>4 μm) (see Stow and Smillie, 2020). They are also one of the principal marine sediment types covering large areas of continental margins and forming "background" facies of many deep-water successions (Stow, 1985; Stow and Smillie, 2020). Many black shale source rocks and organic-rich shale-gas reservoirs are largely of hemipelagic origin, although other processes may also be involved in their deposition (see Stow et al., 2001). An estimated 15–20% of the present-day seafloor, along continental margins, is composed of hemipelagites; volumetrically, they are more abundant due to their great thickness (see Stow and Tabrez, 1998 for a summary of modern examples). Limestone-marl cyclic sedimentation is commonly reported from ancient successions in which the marlstone units are hemipelagic and the limestones are pelagic in nature (see De Boer and Smith, 1994).

8.10.1.2 Slope-Type, Deep-Water Fan Delta

Slope-type, deep-water fan deltas have a slope separating a poorly developed delta front from the prodelta with a deep-water fan system that may be mud-dominated (see also Ethridge and Wescott, 1984) (Figure 8.14). The slope may be an inherent constructional element of the delta, a delta slope, as in deep-water fjords (Prior and Bornhold, 1990), or it may be a consequence of a faulted

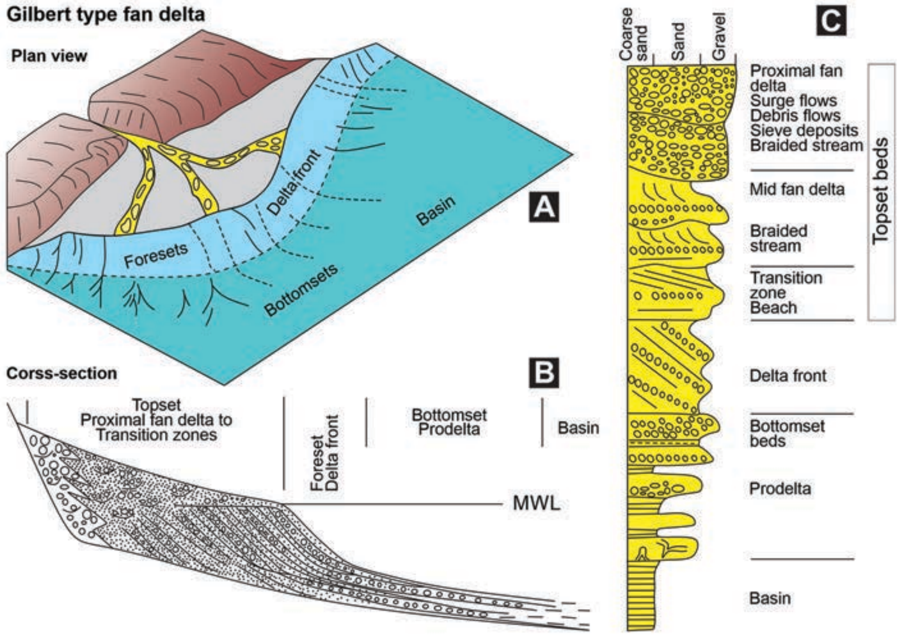

FIGURE 8.15 Gilbert-type fan delta. A–B: The delta possesses a steeply inclined profile characterized by conglomerates and large-scale high-angle delta front slopes; they occur in both shallow and relatively deep waters. C: Gilbert-type fan delta sedimentary sequence.

basin margin and therefore is separated from the delta front by a pronounced shelf/slope break (see Wescott and Ethridge, 1980, 1990) (Figures 8.14A–B). In such cases, subaerial fan gravels may sometimes pass seaward into a significant coastal transition zone of beach, shoreface and shelf gravels and sands before passing into deep-water mass flow deposits of the slope and base-of-slope settings (Figure 8.14C). Deep-water fan deltas are characterized by conglomerates in the slope facies and in the reworked transitional zone (Figure 8.14B).

8.10.1.3 Gilbert-Type Fan Delta

Gilbert-type deltas, occurring in both shallow and relatively deep waters (up to 150 m), are characterized by steeply inclined profiles, conglomerates, and large-scale high-angle delta front slopes (Figure 8.15A). Their development requires a high basin-margin gradient, and therefore they are confined to coarse-grained systems and consist of subaerial topset, subaqueous foreset, and bottomset beds (Nemec, 1990; Postma, 1990; Wescott and Ethridge, 1990) (Figures 8.15B–C). The topset beds are deposited by shifting channels and may be part of an alluvial fan, a braidplain, or a braided river. The foreset beds form where bed load, dropped at the river mouth, continues down the delta front as grain flow or as debris flow (Figure 8.15C). The slope gradients are commonly up to 20° and may reach 24–27° in sands and 30–35° in gravels (Nemec, 1990). The bottomset beds, deposited from a mixture of suspended load and gravity flows, form a low-gradient prodelta.

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9 River System

9.1 INTRODUCTION

River systems are important sediment transport conduits to oceans and lakes (Figure 9.1), rather than sites of sediment deposition (see Dalrymple et al., 1992). A river is a gravity-driven natural stream of water that flows regularly or intermittently in a channel or channels toward a receiving basin, commonly an ocean or a lake (Figure 9.1). Rivers deposit sediments (sediment load), acquired mostly by dissecting uplands or by reworking unconsolidated debris from previous erosional events, and some of this sediment is preserved under certain conditions to become part of the sedimentary record. But to understand this – i.e., the deposits of the river systems – it is important to first examine channel shapes (also referred to as channel forms or river forms) and sediment transport processes that deliver the sediments to basins.

9.2 CHANNEL FORMS

The channel form of rivers is described in terms of (a) the number of channels (single or multiple); (b) deviation from a straight path (i.e., sinuosity); (c) degree of channel subdivision by large bedforms (i.e., bars) and accreting islands around which the channel reaches diverge and converge (i.e., braiding); and (d) permanent distributive channel subdivision into stationary smaller channels (i.e., separated by floodplains) that each contain their own channels and point bars (i.e., anastomosing) (see Figure 9.2; see also Church, 1992). Some of these features, such as shape and size of bars, vary as a function of river levels; that is, they may appear differently at low-water stage than at the flood stage (= high-water stage). However, at the most basic level, it is useful to classify rivers according to (a) the number of channels they contain (see Figures 9.2A–B), and (b) sinuosity (see Figure 9.2C). The number of channels range from single-threaded to braided (with more than three interweaving channels that are frequently reorganized) to anastomosing (that typically have somewhat stable, vegetated islands between channel threads), to discontinuous streams (that have un-channelized reaches) (Figures 9.2A–B) (see Ashmore, 1991; Montgomery and Buffington, 1997).

Traditionally, on the basis of channel form, rivers are classified into three main types: (a) meandering (single-channel), (b) braided (multiple-channel), and (c) anastomosing (multiple-channel) (see Figure 9.1; see also Richards, 1982; Church, 1992). Leopold and Wolman (1957) proposed this tripartite geomorphological classification, i.e., a continuum of river patterns between end-members of straight, meandering, and braided (see also Figure 9.2) (see Bettess and White, 1983). Recent classifications have adopted this tripartite system but added additional types, such as the low-energy, fine-grained multiple-channel rivers, called anastomosing rivers (= also referred to as anabranching;

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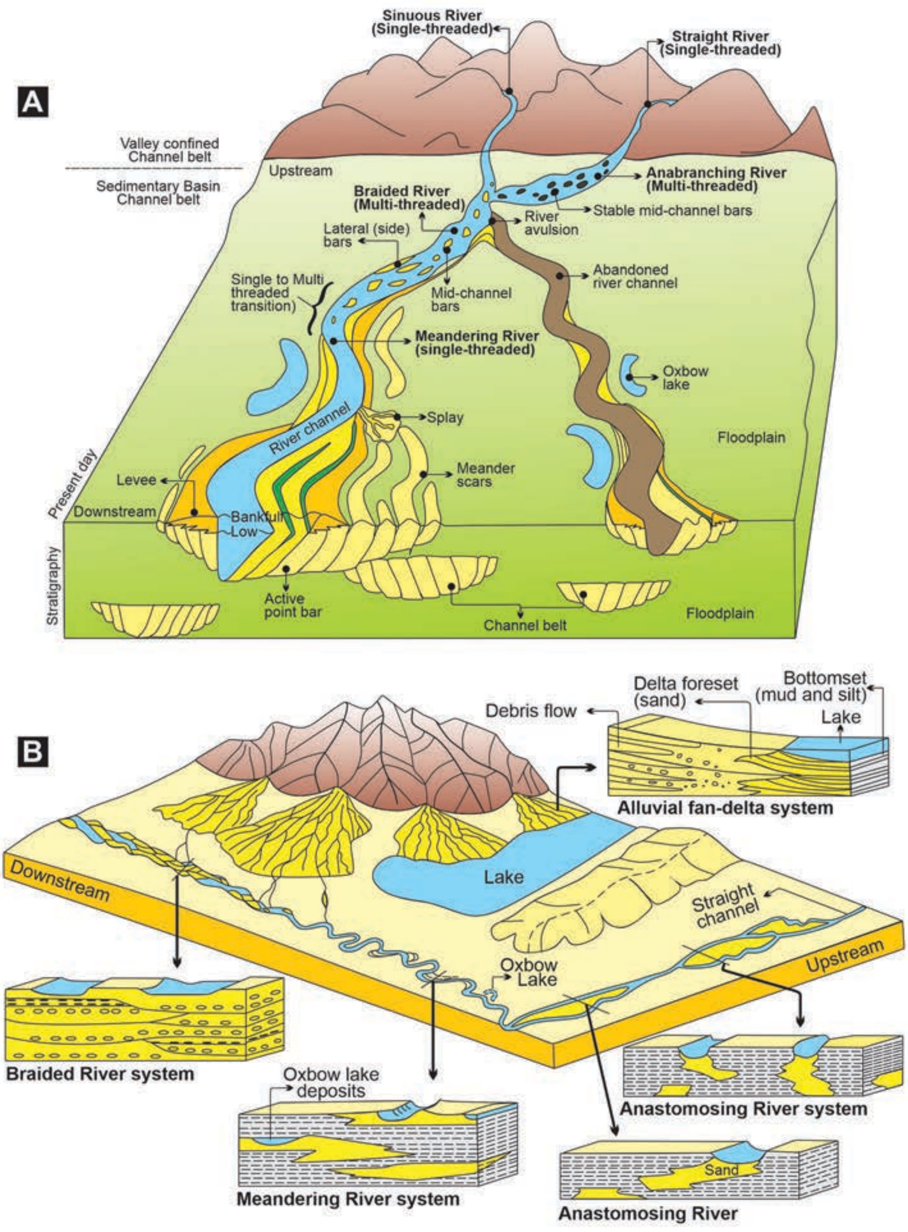

FIGURE 9.1 River system. A: Terminology used in the chapter illustrating straight, sinuous, meandering, braided, and anabranching river channels (active and abandoned) and their associated levees, bars, and overbank landforms. (Modified after Nyberg, B., Henstra, G., Gawthorpe, R.L. et al. Global scale analysis on the extent of river channel belts. Nat Commun 14, 2163 [2023]. https://doi.org/10.1038/s41467-023-37852-8. Available under CC by 4.0: https://creativecommons.org/licenses/by/4.0/). B: Depositional setting and lithofacies of meandering, braided, and anastomosing rivers. The braided river system is characterized by low sinuosity, movable channels, high-gradient, high flow intensity, dominated by bed load. The meandering river system is characterized by low-gradient, low flow intensity, dominated by suspension load and mixed load; bed load/suspension lead < 3. The anastomosing river system is characterized by channels, partly straight, partly sinuous, but stable and at places by a mud-dominated low gradient system. (Modified after Miall, 1999.) Note that different channel patterns have varying hydrodynamic conditions, and migration patterns, hence have differing lithology, grain size, sedimentary structure and facies associations, and vertical sequences.

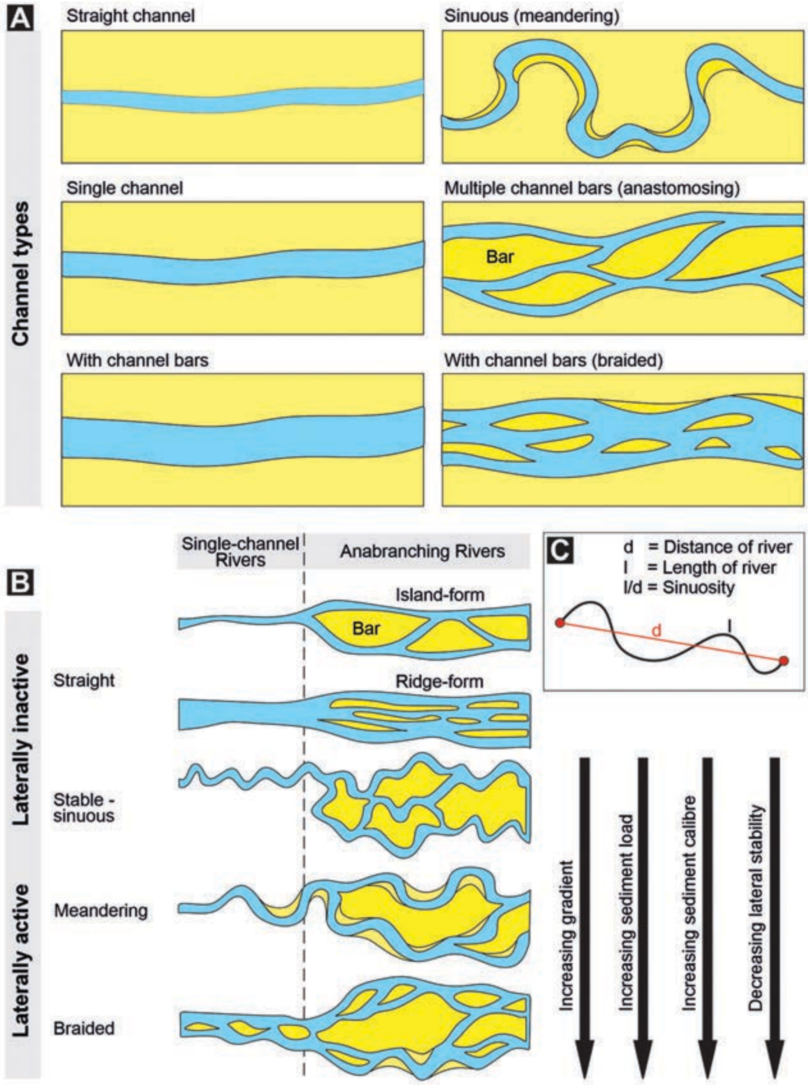

FIGURE 9.2 A: Channel forms. Several types of river can be distinguished, based on whether the river channel is straight or sinuous (meandering), has one or multiple channels (anastomosing), and has in-channel bars (braided). Combinations of these forms can often occur. (Modified after Nichols, 2009.) B: A classification of channel patterns linking single channel and equivalent anabranching planforms. Laterally inactive channels consist of straight and sinuous forms, whereas laterally active channels consist of meandering and braided forms. (Modified after Nanson and Knighton, 1996.)

see Nichols, 2009; Nyberg et al., 2023) (Figure 9.1A). Thus, in slope-discharge plots (i.e., in order of declining stream power and grain size), the rivers are separated into braided, meandering, straight, and anastomosing (Figure 9.3).

The term anabranching describes rivers that flow in multiple channels separated by stable, vegetated, alluvial islands that divide flows up to bankfull (= high-water stage) regardless of their energy or sediment size (Nanson and Knighton, 1996) (see Figures 9.1A and 9.2B). Individual anabranches can be straight, meandering, or braided. The term wandering river has become popular to describe anabranching rivers, an intermediate between meandering and braided (see Figure 9.1A).

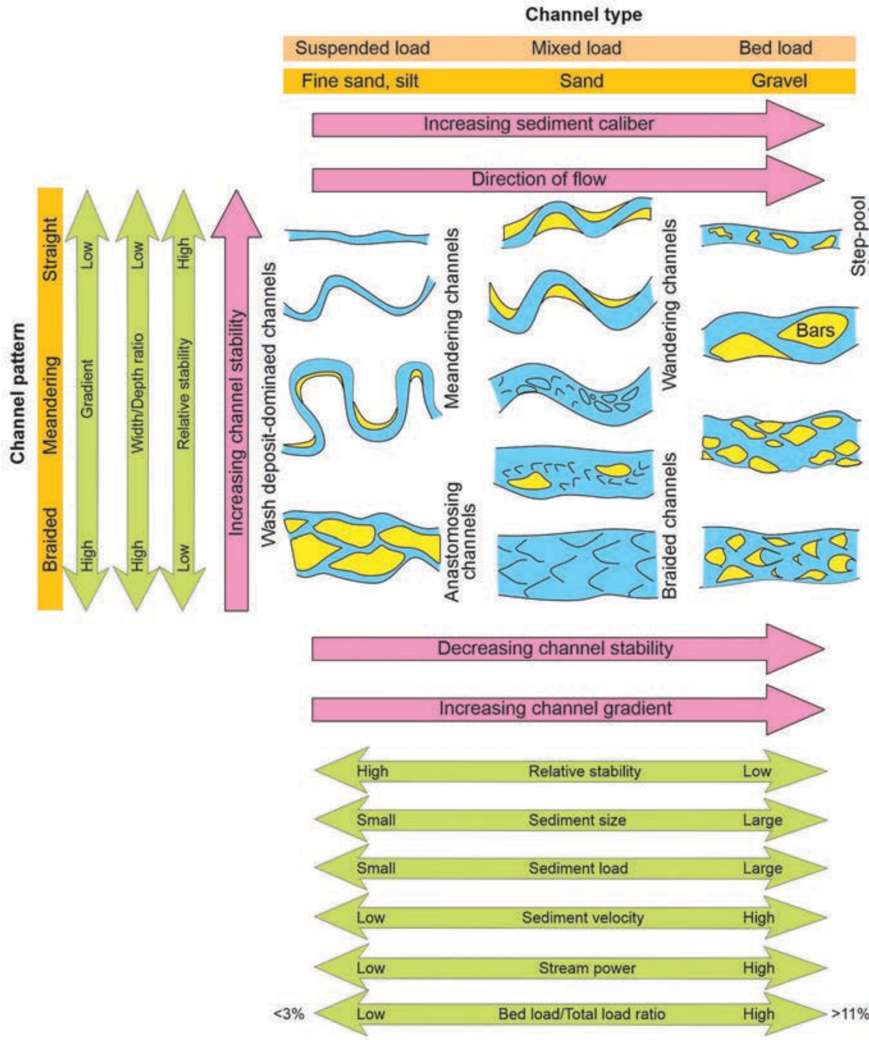

FIGURE 9.3 Dominant controls on channel morphology transforming from braided to meandering rivers, and vice versa, along their reach and through time. (Modified after Church, 2006; Hartley et al., 2010.)

However, this rigid classification is oversimplified and unsatisfactory as many rivers show combinations of sinuosity and braiding in their different reaches. Thus, even the same channel may show changing channel patterns along its length; or same channel reach can be meandering at bankfull stages (floods) and appear braided at low stages (see Russell, 1954). Factors (often complex, and interrelated) that influence channel sinuosity and braiding include (a) the magnitude and variability of stream discharge, (b) channel slope, (c) grain size of sediment, (d) bed roughnesss, (e) amount and type of sediment load (bed load *versus* suspended load), and (f) the stability of the channel banks (see Figure 9.3). A straight channel without bars is the simplest form but is relatively uncommon (see Figures 9.2 and 9.3). A braided river contains mid-channel bars that are covered at bankfull flow, in contrast to an anastomosing river, that consists of multiple, interconnected channels separated by areas of floodplain (see Figures 9.2 and 9.3) (see also Makaske, 2001). Both braided and anastomosing river channels can be sinuous. Sinuous rivers that have depositional bars only on the insides of their bends are called meandering rivers (see Figures 9.2 and 9.3). Both braided and meandering rivers are briefly enumerated below, but detailed later in the chapter.

9.2.1 Braided Rivers

Braided rivers are relatively high-energy systems that at low stage have multiple channels (i.e., successive divisions) divide and rejoin around alluvial islands (channel bars) (see Figure 9.4A). They tend to occur in settings with steep gradients, weakly cohesive banks, abundant coarse sediments, and with variable discharge (see Leopold and Wolman, 1957; Knighton, 1998). Channel bars or cross-channel bars (Figure 9.4A) that divide the stream into several channels at low flow are often submerged during high-flow stage. One or more alluvial islands or channel bars may be present in a given channel cross section (Figure 9.4A). Braiding is most well-developed

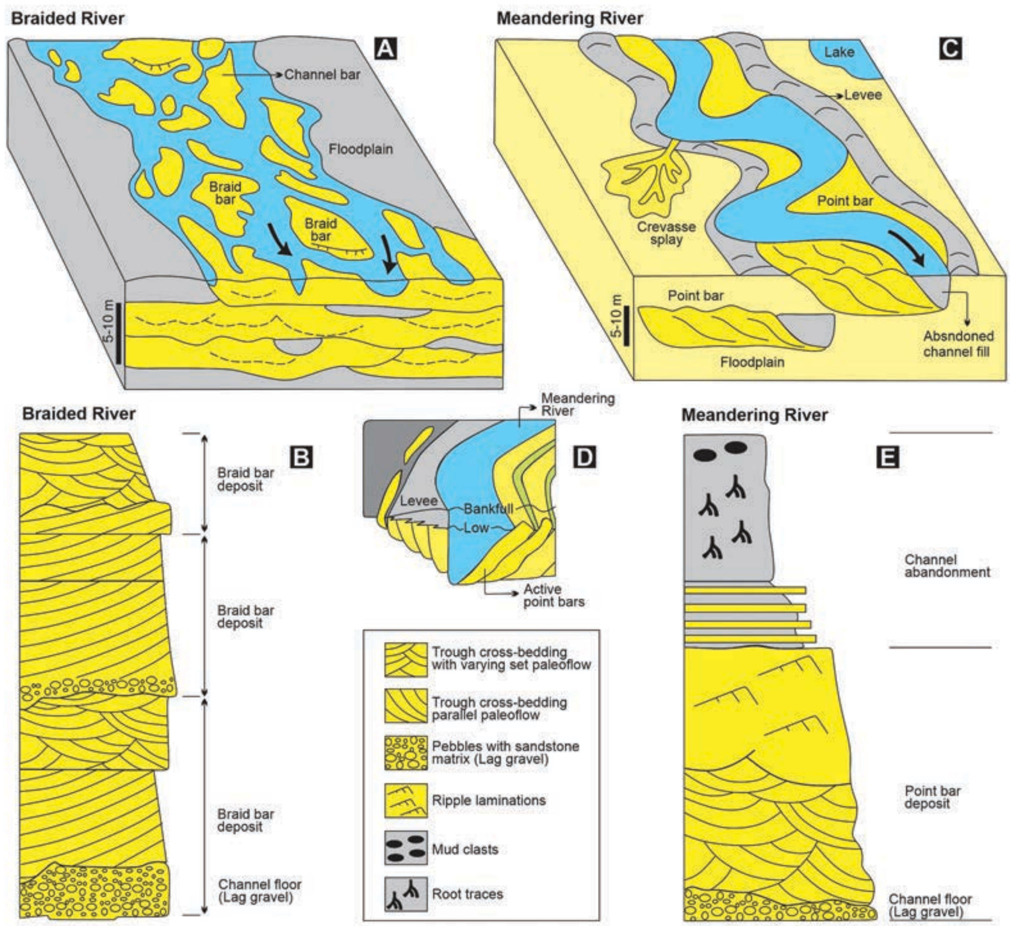

FIGURE 9.4 Comparison between braided and meandering rivers. A–B: Braided river. A: Depositional setting of a braided river showing braid bars (channel bars). B: Sedimentary log of a braided river. (Modified after Miall, 1978.) C–E: Meandering river. C: Depositional setting of a meandering river. (Modified after Donselaar and Overeem, 2008.) D: Bankfull stage of a meandering river. E: Sedimentary log of a meandering river. (Modified from Miall, 1978.)

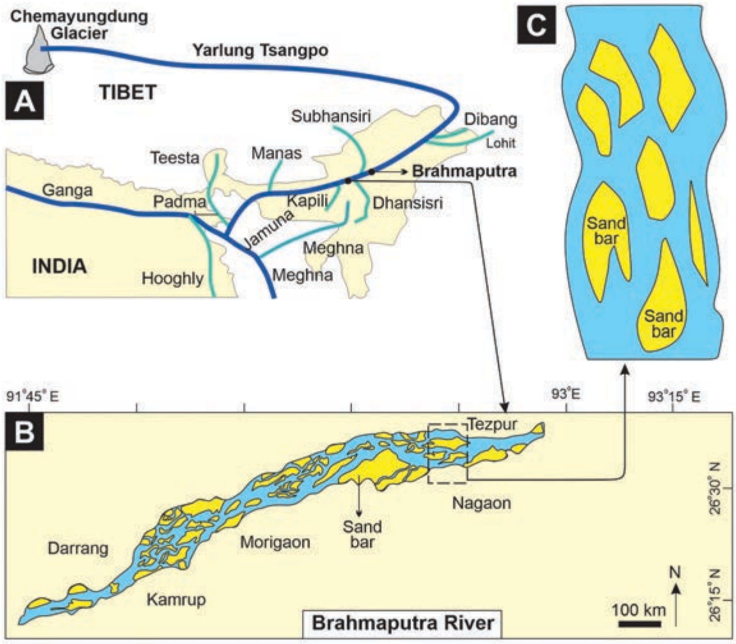

FIGURE 9.5 Brahmaputra River System. A: Stretch of the Brahmaputra River from its source in the Chemayungdung Glacier to its sink down south in the Bay of Bengal. B: Location and braided nature of the Brahmaputra River, showing island and sandbars of both the banks. The figure shows the middle Brahmaputra River basin (from Tezpur to Kamrup) with a total length of 340 km. C: Sand bars. The middle Brahmaputra River carries a huge amount of bed load.

in mountainous reaches of rivers, in streams of alluvial fans, and glacial outwash plains (see Blair and McPherson, 1994). Channel bars of such braided streams are commonly composed of gravelly material (Figure 9.4B). Bars tend to be built up by the addition of sediments at the downstream end, and on the lateral flanks; the upstream end is partly eroded. The channel bars are composed of coarser-grained lag deposits (Figure 9.4B) of the stream which could not be carried by the flow. Once such a channel bar is formed, it may become stabilized by the deposition of fine-grained sediments on top during high flows, and may become covered by vegetation. In general, sediments are typically dominated by gravel and sand (Figure 9.4B), but silt-dominated systems are also known. Braided rivers are prominent near mountains and glacial fronts, and occur from humid to arid regions. They commonly change down valley into finer-grained sediments and single-channel systems, although some feed directly into oceans and lakes as braided deltas. The Brahmaputra River, one of the world's largest braided systems, has a channel-belt width of up to 20 km, with individual channels several kilometers wide and maximum scour depths of up to 50 m (Figure 9.5).

Leopold and Wolman (1957) noted that the braiding (Figures 9.4A–B) or meandering (Figures 9.4C–E) of river channels depends mainly on the relationship of the channel slope to discharge. In case of the two rivers with the same discharge, braided channels develop on steeper slopes, and meandering channels develop on gentler ones. Steeper slopes cause larger sediment transport and bank erosion, and are often associated with coarser heterogeneous sediments. However, if discharge is rather high and banks are weak, braiding is common, even in rivers with fine-grained sediments (see also Swan et al., 2019). Braided rivers are also characterized by wide channels, and rapid and continuous shifting of both sediments and the position of channels. An extreme example

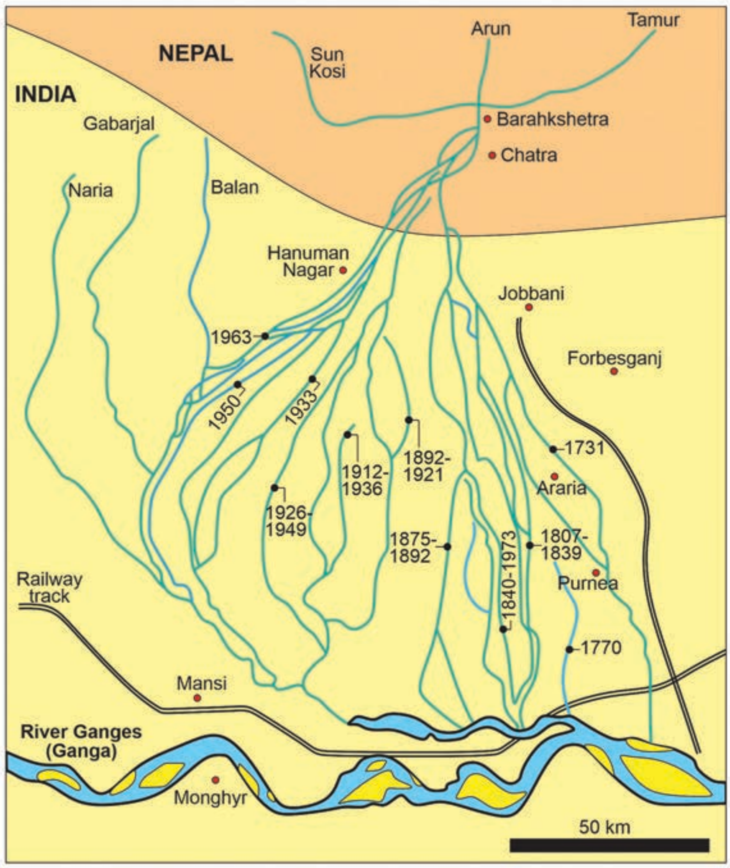

FIGURE 9.6 Westward migration of the Kosi River (India) through time. (Modified after Gole and Chitale, 1966.)

of lateral shifting of a braided stream is the Kosi River, a tributary of the Ganges River, India (Figure 9. 6). During the last two centuries, this river has consistently shifted its position westwards (Figure 9.6).

9.2.2 Meandering Rivers

Meandering channels (Figure 9.7) are usually single-channeled sinuous rivers with moderate energy, gradient, and width-depth ratio (Figure 9.7A). They tend to occur in settings that possess cohesive and/or well-vegetated banks, mixed loads of sand (sometimes gravel) and mud (Figures 9.7B–F), and with a perennial flow. Meandering channels are especially common in down valley (downstream), the lower-gradient reaches of rivers, and in estuaries. As the outer bank of each bend erodes, the point bar migrates systematically laterally and downstream, usually balanced by erosional retreat of the concave bank (Figure 9.7C). Where a meander tightens through time, the point bar may be partially truncated by erosion along chute channels (chute cutoff) or completely truncated with full isolation of the bend (neck cutoff), resulting in an oxbow lake (Figures 9.8 and 9.9). Repeated cutoffs and local channel migration result in a meander belt that is composed of many laterally juxtaposed channel segments (Figure 9.8) (see also Fisk, 1944). As the channel migrates, sediment eroded from the concave bank is transported obliquely across the channel and up the opposing point bar,

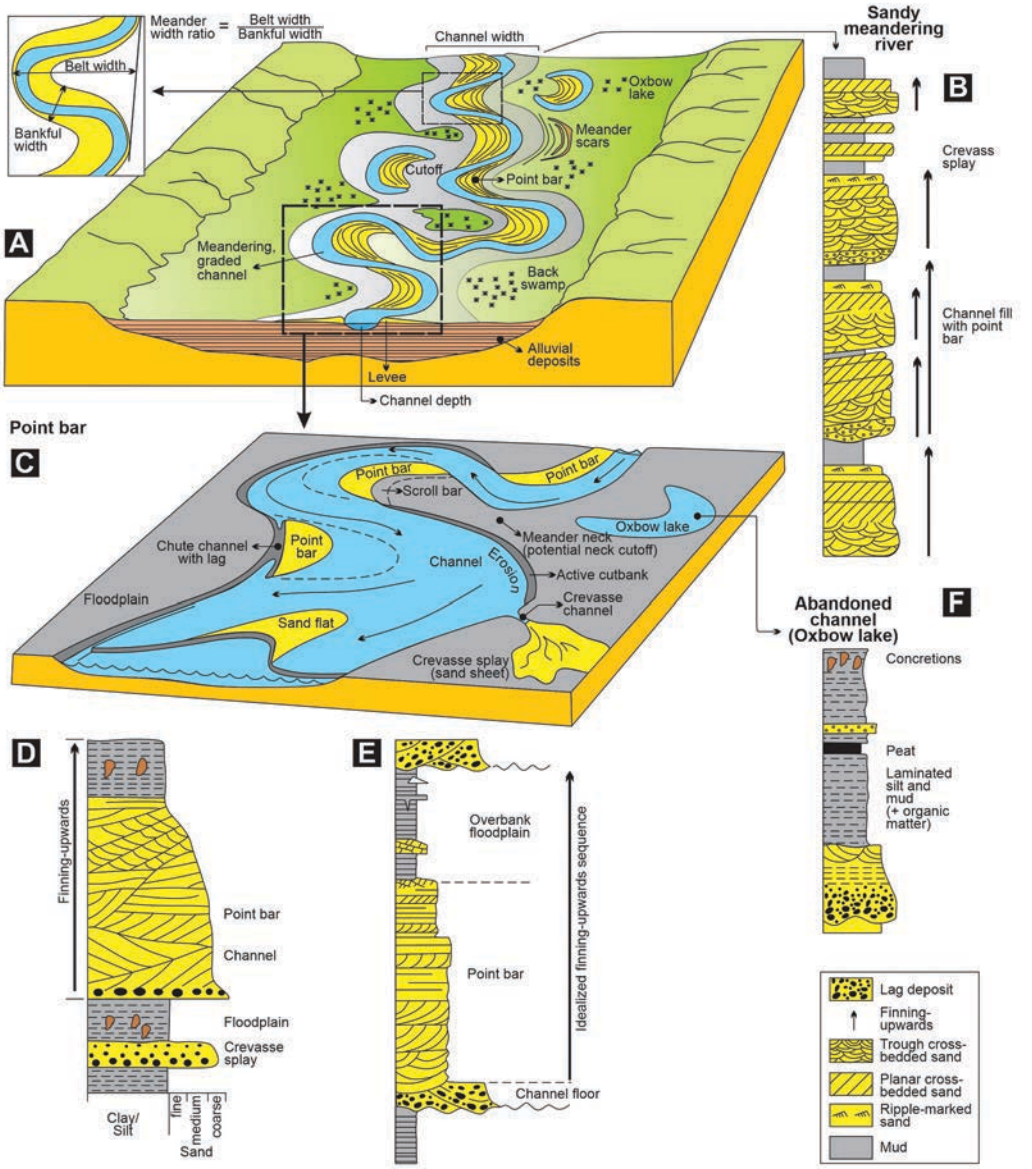

FIGURE 9.7 Channel landforms of a meandering river. A: Meandering channel subenvironments. B: Sedimentary log of a sandy meandering river. C: Point-bar deposits. D–F: Coarsening upwards cycle. D: Sedimentary log of a point bar. E: Sedimentary log of a point bar and overbank floodplain deposits. F: Sedimentary log of an abandoned channel (oxbow lake).

with flow waning progressively toward the bar top (Figure 9.9). Point-bar accretion is commonly linked to scroll bars that develop as a within-channel bar on the point-bar platform (Figure 9.9B) and through vertical and up and downstream growth, to become effectively a levee bank around the convex bank of a meander bend (Figure 9.9B). Meandering channels also possess well-defined pools and sediment bars joined by riffles (Figure 9.10). In general, meandering channels show slower rates of lateral shifting than braided ones. Rivers in their lower reaches are usually meandering. But if

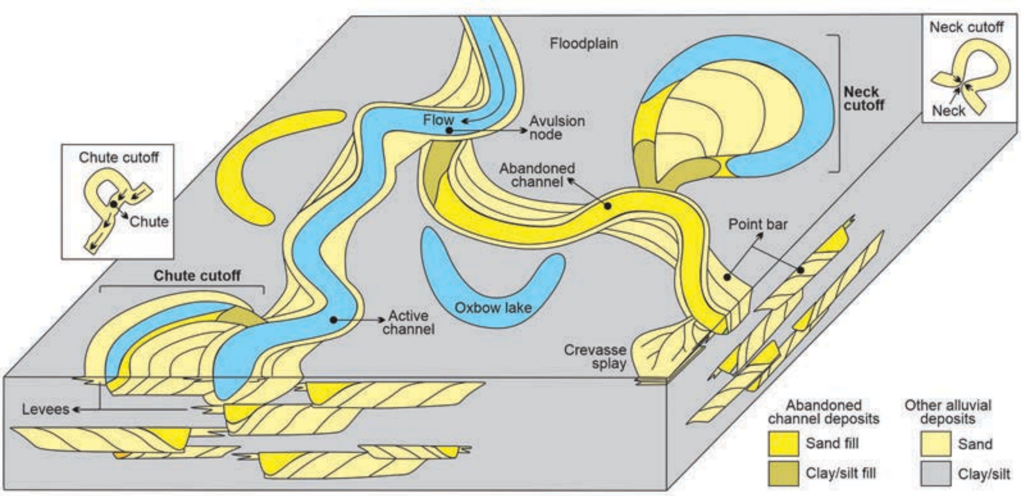

FIGURE 9.8 Channel abandonment in a meandering, graded channel. (Modified after Szewczyk et al., 2020.) Where a meander tightens through time, the point bar may be partially truncated by erosion along chute channels (chute cutoff) or completely truncated with full isolation of the bend (neck cutoff), resulting in an oxbow lake.

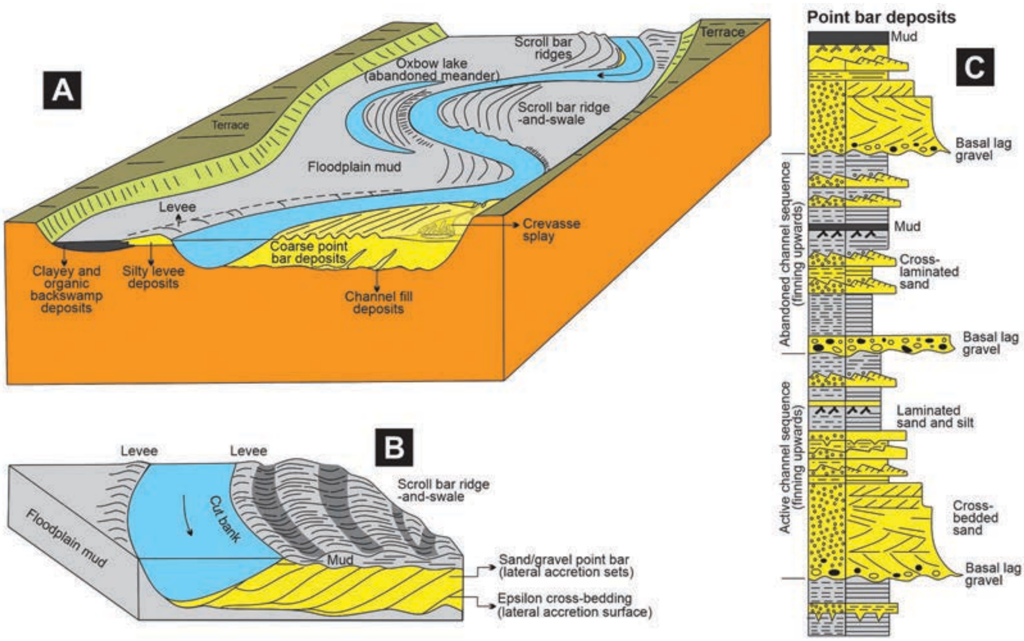

FIGURE 9.9 Point-bar deposits. A: Meandering river subenvironments. B: Formation of a levee. C: Sedimentary log of a point-bar deposit. As the point bar migrates laterally, basal gravels on the deepest part of the channel bed, or thalweg, are first covered by cross-bedded coarse sands produced by migrating dunes on the overriding lower part of the point bar. Continued lateral advancement of the point bar results in the burial of the dune deposits by progressively finer materials toward the top of the point bar, including cross-laminated medium to fine sands generated by migrating ripples. These cross-laminations may be interspersed with horizontally laminated fine sand and silt.

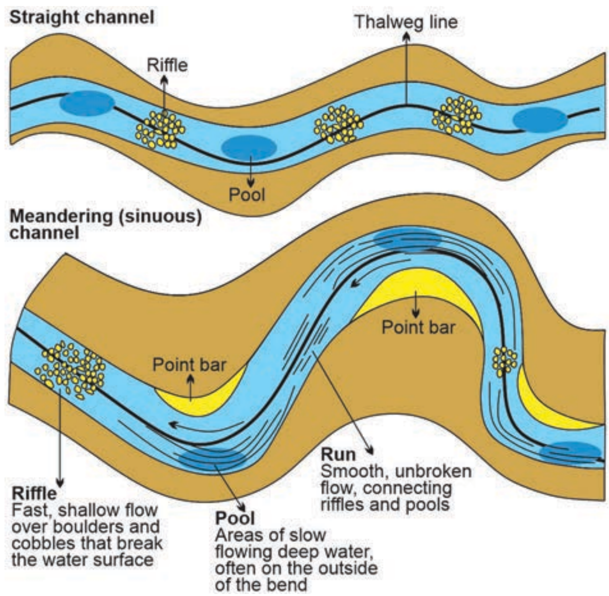

FIGURE 9.10 Channel characteristics showing thalweg, pools, and riffles. The thalweg of straight channels is sinuous, and shows deeper parts (pools) alternating with shallower parts (riffles). Erosion takes place along pools and deposition on sediment bars.

they are heavily charged with sediment and have high discharge, they show braiding, despite a fine-grained bed load.

9.2.3 Straight Rivers

Straight rivers rarely persist in an alluvial setting for a distance of more than 7–10 channel widths (see Leopold et al., 1964). Straight reaches are classed as those without significant bends for more than this distance (i.e., a sinuosity of less than 1.1) (see Figure 9.2C). Where compared to other patterns, they are at the low end of the flow-strength to bank-strength ratio, so are usually fine-grained (Figure 9.3). The thalweg of straight channels is sinuous, and shows deeper parts (pools) alternating with shallower parts (riffles) (Figure 9.10). Erosion takes place along pools and deposition on sediment bars (see Figure 9.10).

9.2.4 Anabranching/Anastomosing Rivers

Anabranching rivers are defined as multiple-channel rivers characterized by vegetated or otherwise stable alluvial islands (bars) that divide flows at discharges up to bankfull (see Nanson and Knighton, 1996) (Figure 9.11). Anabranching rivers are also known as anastomosing rivers (see Nichols, 2009); Schumm (1968) was the first to use the term "anastomosing channels." Anastomosing rivers is now used to designate a specific subset of relatively distinctive low-energy anabranching systems that are mostly associated with fine-grained or organic-rich floodplain deposits to high-energy gravel systems (see Smith and Smith, 1980; Nanson and Croke, 1992; Knighton and Nanson, 1993; Makaske, 2001). These systems are laterally stable, straight (most common) to highly sinuous, with low width-depth ratio and well-vegetated or have highly cohesive banks. Because of their tendency to accumulate sandy channels and fine-grained overbank muds and organics (Figure 9.11B), they are also important locations for the preservation of coal and hydrocarbons. They are also the dominant

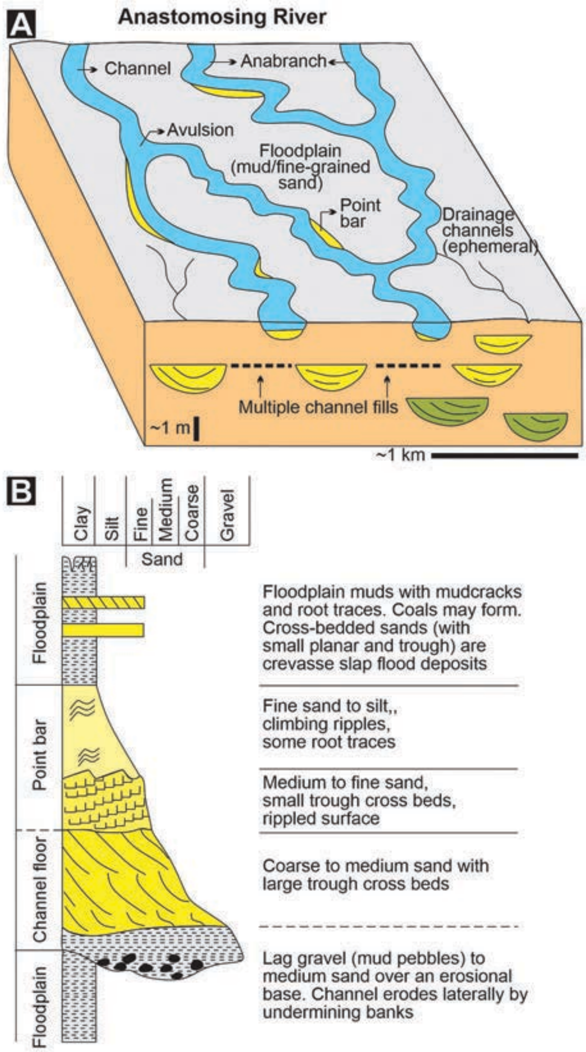

FIGURE 9.11 Anastomosing (anabranching) river. A: Depositional subenvironments. B: Sedimentary log of an anastomosing river.

style among the world's largest alluvial rivers (see Nanson and Knighton, 1996) but are less common as compared to meandering or braided ones. They occur widely from the subarctic to the tropics and from humid alpine to lowland arid regions (Smith, 1973; Smith and Smith, 1980). They are also known from more arid regions with sparse vegetation (see Gibling et al., 1998; Makaske, 2001). As their banks are stabilized by vegetation, lateral migration of channels is considerably reduced (Smith and Smith, 1980). In rapidly accreting humid regions, peats accumulate in floodplain lakes and swamps to form coal, and sandy paleochannels may act as reservoirs for hydrocarbons. However, not all anabranching rivers are rapidly vertically accreting and in arid environments they do not accumulate organics. Examples of anabranching rivers include the Columbia River of western Canada (Smith and Smith, 1980), Okavango Delta in Botswana (McCarthy et al., 1988), the semi-arid Australian Channel Country (Gibling et al., 1998), and the Ganges plains in the Himalayan Foreland Basin, a rapidly accreting anastomosing reach of the Bagmati River in Nepal.

9.3 CHANNEL SINUOSITY

Channel sinuosity (particularly useful for single-threaded channels) is calculated as the length along the river divided by the straight-line distance along the river valley i.e., the ratio of channel length to valley length (see Figure 9.2C). Rivers can have sinuosity ranging from 1 to 3 (i.e., the river length is three times longer than the valley). A river is considered to be sinuous if the distance measured along a stretch of channel divided by the direct distance between those points is greater than 1.5. The term "meandering" is used when this ratio is greater than about 1.3. Such rivers contain relatively few bars that are not bank attached, thus having braid-bar ratios close to 1 (see also Figure 9.2). Meanders are; self-similar over a wide range of scales and their planform (especially width and wavelength) can be related to channel discharge (see Knighton, 1998), and these relationships have been used to determine paleodischarges from ancient meander traces (Smith, 1987; Brierley and Hickin, 1991). Planform is the channel pattern as seen from an overflying aircraft (see also Figure 9.7A). Leopold et al. (1964) distinguished meandering from straight and braided rivers on the basis of sinuosity; rivers with a sinuosity of 1.5 or greater are meandering; those below 1.5 are straight and braided (see also Leopold and Wolman, 1957) (see also Figure 9.2).

9.4 SEDIMENT TRANSPORT PROCESSES

Rivers transport their sediment load in essentially four ways: (a) bed load or traction load which is almost constantly in contact with the bed; (b) saltating load that bounces or skips over the bed (intermediate between bed load and suspended load); (c) suspended load that is held in the water column by turbulence; and (d) dissolved load that is transported in solution (see Figure 9.12). The first three (bed, saltating, and suspended loads) are composed of detrital sediments, and the concentrations and relative proportions of the first two are highly dependent on the energy of the flow, which increases with water discharge. Suspended load is energy-dependent but can also be strongly influenced by the rate of sediment supply to the river. The concentration of dissolved particles depends largely on water temperature, catchment geology, groundwater chemistry, and vegetation. In contrast, dissolved load commonly decreases in concentration due to dilution effects with increasing flood discharge. These three detrital types (bed, saltating, and suspended loads) are the primary constituents of alluvial strata, with bed, and saltating loads commonly forming sedimentary flow structures diagnostic of variable flow conditions. Bed load is the coarsest fraction and moves mostly short distances during high magnitude flows, and is commonly the smallest proportion of transported sediments (often <10% of the total load) (Figure 9.12). For suspended or saltating sediments, sedimentation takes

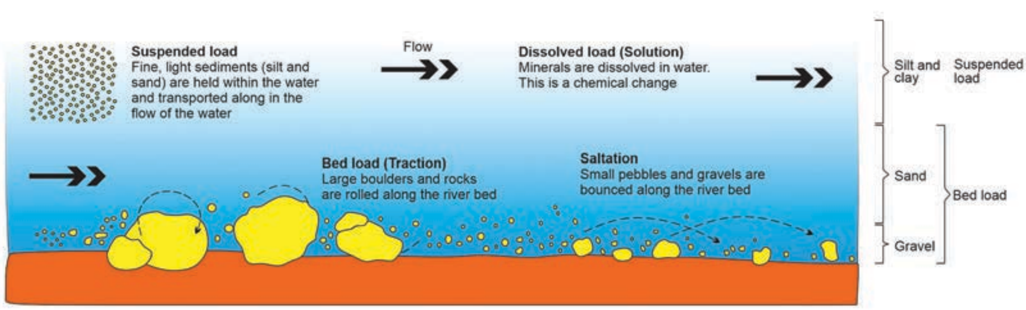

FIGURE 9.12 Sediment transport processes. Four processes occur: bed load or traction load which is almost constantly in contact with the bed; saltating load that bounces or skips over the bed (intermediate between bed load and suspended load); suspended load that is held in the water column by turbulence; and dissolved load that is transported in solution.

place when the flow velocity drops below that of the settling velocity, and for bed load when flow velocity drops below that is needed to maintain sliding or rolling of particles over the bed. When velocity decreases within the channel or on the floodplain surface, the coarsest fractions are deposited first. As a result, sediment sizes are sorted vertically, and laterally within the system. Suspension load is important in the deposition on natural levees and floodplains; whereas bed load is deposited as channel-lag deposits and at the lower part of point bars.

9.5 BRAIDED SYSTEMS (HIGH BED-LOAD CHANNELS)

Rivers with a high proportion of sediments carried by rolling and saltation along the channel floor are referred to as braided (bed load) rivers. Where the bed load is deposited as bars of sand or gravel in the channel, the flow is divided to give the river a braided form (Figures 9.4A–B). Flow is generally strongest between the bars and the coarsest materials are transported and deposited on the channel floor to form an accumulation of larger clasts, or as a coarse lag (lag gravel) (Figures 9.4A–B).

9.5.1 Point Bars

In general, in the higher gradient proximal reaches of a river, sediment transport (and erosion) occurs. Downstream, the channels form bars. Point bars (also called side bars and lateral bars) are attached to the river bank, where erosion occurs on the outside parts of bends and deposition on the point bars (Figures 9.4A–B; see also Figure 9.13). The point-bar sediments are characterized by cross-bedding and general fining upwards, towards the top of the bar (Figures 9.4A–B). Sediments accumulate on mid-channel bars in braided streams and on the inner banks of meandering river bends. Although "braided" and "meandering" (see Figure 9.4) are useful ways of categorizing ancient fluvial deposits, considerable variations in and combinations of these are present in both modern and ancient systems. Here, it must be mentioned that the anastomosing rivers do not possess bars, and this character separates them from meandering rivers.

9.5.2 Braid Bars

In braided rivers, braid bars (synonymous with channel bar, median longitudinal bar and transverse bars) are present in the mid-channel position (Figures 9.4A and 9.13). These are mainly controlled by processes of lateral and vertical deposition (migration) (Figures 9.4A–B), together with channel cutting and abandonment.

The channel bar migrates downstream by depositing sediment in front, like a delta, producing foresets. Channel bars also migrate laterally which also produces foreset laminae. However, channel bars are invariably covered with various bedforms such as giant ripples, megaripples, etc., and channel migration is mainly due to the migration of these bedforms. The main deposition in channel bars takes place during high-water stages of rivers (i.e., at the bankfull stage). The Brahmaputra River is a braided stream and it actively migrates laterally (see also Figure 9.5). Because of a high amount of sediment coming during the rainy season, the river builds up channel bars that migrate actively during floods, mainly due to migration of various bedforms (see Coleman, 1969). There seems to be no well-defined difference in the sequence of point-bar deposits and channel-bar deposits. This is especially true of the rivers with very fine-grained sediments. In fact, several rivers possess both point bars and channel bars at different times. Sometimes a point bar may even grow, become detached as a channel bar, and develop further (Sarkar and Basumallick, 1968).

In general, the braid bars can be thought of as double-sided point bars. As the current splits around the upstream end of the bar, helical flow causes lateral accretion on both sides of the bar.

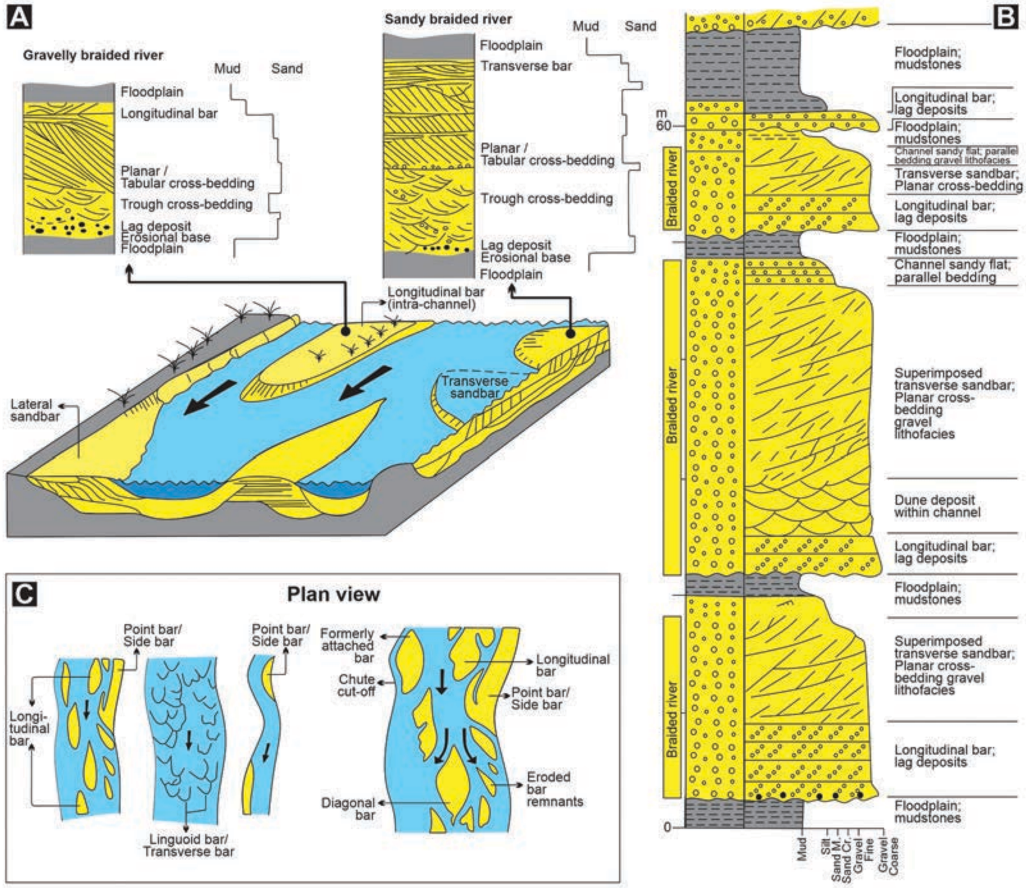

FIGURE 9.13 Braided river lithofacies and sand bars. A: Sedimentary sequence characteristics of a low sinuosity river with sand enrichment (braided river). The braided river lithofacies is broadly subdivided into a gravelly braided river and a sandy braided river dominated by sandy deposits. (Modified after Galloway, 1983.) B: Proximal gravel braided river sedimentary log showing planar cross-bedding, sand bars, and graded bedding in the lower part of the sequence in the Permian fluvial facies in North China. (Modified after Yu et al., 2018.) C: Types of bars.

Because braid bars are free to move, in contrast to point bars (Figure 9.13), scouring and subsequent delta-like deposition take place at the downstream end of the bar. Thus, braid bars migrate downstream. On the other hand, some braid bars that are colonized by vegetation, remain stable long enough to form channel bars (channel islands) (Figures 9.4A–B). In a gravelly braided river, the bar deposits commonly consist of cross-stratified granules, pebbles or rarely cobbles in a single set (Figures 9.4A–B).

A sandy bar will form a succession of cross-bedded sands (Figure 9.4B). Finer sands or silts on the top of a bar deposit represent the abandonment of the bar when it is no longer actively moving (Figures 9.4E). There is therefore an overall fining-up of this channel-fill succession (Figures 9.4B and 9.4E). The thickness may represent the depth of the original channel if it is complete, but commonly, the top part is eroded by the scour of a later channel.

The bars within the channel vary in shape and size (Smith 1978; Church and Jones 1982). The longitudinal bars are elongated along the axis of the channel (Figure 9.13A). The transverse bars are wider than they are long, spreading across the channel (Figure 9.13A). The linguoid bars are crescentic with their apex pointing downstream and the compound bars consist of sand, gravel or a mixture of clast sizes (Figure 9.13C). The longitudinal bars have low relief and their migration forms deposits that display a poorly defined low-angle cross-stratification in the downstream direction (Figure 9.13B). The transverse and linguoid bars have a higher relief and generate well-defined cross-stratification dipping downstream (Figure 9.13B). The deposits of a migrating gravel bar in a braided river therefore form beds of cross-stratified granules, pebbles or cobbles that later lithify to form conglomerates (lag deposits) (Figure 9.13B). In sandy braided rivers, the bars are comprised of a complex of subaqueous dunes over the bar surface that migrate over the surface of the bar in the direction of the stream current to build up stacks of cross-bedded sands (Figure 9.13B). Linguoid (arcuate) subaqueous dunes form trough cross-beddings, whereas the straight-crested ones may produce planar cross-bedded sands (Figure 9.13B). Compound bars comprise cross-stratified gravel with lenses of cross-bedded sand or there may be lenses of gravel in sandy bar deposits.

In summary, the major diagnostic features of a braided system include tectonic setting, geometry, typical sequence, sedimentology, and fossils. These are very briefly enumerated. In terms of tectonic setting, the braided systems occur in the upper reaches of alluvial plains, relatively near the upland source. Like alluvial fans, they are associated with rapidly down-dropping basins as they require an upland to provide the coarse material and high stream gradient (see Blair and McPherson, 1994; DeCelles et al., 1991). In terms of geometry, in braided rivers, elongate, fairly straight lenticular or sheet-like sand bodies grade laterally into finer deposits of an alluvial plain. A typical sequence in braided rivers displays a fining-upward sequence of channel-lag gravels, sandy trough cross-beds filling channels, and occasional tabular cross-beds migrating across channels, topped by laminated sand and mud with burrows and root casts (Figures 9.4 and 9.13). Unlike meandering rivers, braided systems are ephemeral and rapidly shifting, so the sequence may cross the channel and repeat several times (see Figure 9.4). Sedimentologically, in braided rivers, gravel is more common in longitudinal bars of the upper reaches of the system, but sand dominates, throughout (Figures 9.4 and 9.13). Unlike meandering systems, there is very little silt and mud (Figure 9.4). There is abundant tabular and trough cross-stratification; the vertically accreting plane beds are less common. Longitudinal bars show high-flow-velocity plane beds; ripples and dunes are common at lower flow velocities. In terms of fossil content, the braided systems are usually unfossiliferous except for root casts and burrows on vegetated sand flats.

9.6 MEANDERING SYSTEM

Leopold and Wolman (1957) divided rivers into three categories: braided, meandering, and straight. Braided channels have multiple-thread patterns as opposed to single-thread meandering and straight channels (see Figure 9.4). The meandering channels are distinguished from their straight counterparts by their sinuously winding course (Figure 9.4). Although in nature a continuum of river planforms exists, nevertheless, the meandering channel remains an important end member in all modern classifications and is also one of the most common patterns.

9.6.1 Floodplain Deposition

When the discharge exceeds the capacity of the channel, water flows over the banks and out onto the floodplain where overbank or floodplain deposition occurs (Figures 9.7–9.9). Floodplains are strips of land adjacent to rivers that are commonly inundated during seasonal floods (see Figures 9.7–9.9). Most of the sediment carried out onto the floodplain is suspended load, i.e., mainly clay- and silt-sized debris but may also include fine sands if the flow is rapid enough to carry sand in suspension (Figures 9.7–9.9). Thus, as the water leaves the confines of the channel it spreads out and loses velocity very quickly. The drop in velocity prompts the deposition of the sandy and silty suspended load, leaving only clay in suspension. These fine-grained sediments settle out of suspension from floodwaters and are carried into the flood basin, which may be a broad, low-relief plain, a swamp, or even a shallow lake (Figures 9.7–9.9). The fine sand, silt, and clay layers are deposited as thin sheets (from a few centimeters to several decimeters thick) over the floodplain (i.e., floodplain deposition in suspended-load rivers). These thin fine-grained deposits commonly contain considerable plant debris and may be bioturbated by land-dwelling organisms or plant roots. Floodplain deposits may show current ripples or horizontal laminations; rapid deposition results in the formation of climbing-ripple cross-laminations. Small-scale cross-bedding and horizontal bedding are commonly noted. Sand and silt units grade upward into finely laminated clayey sediments. Sheets of sand and silt deposited during floods are thickest near to the channel bank as coarser suspended load is dumped quickly by the floodwaters as soon as they start flowing away from the channel. Rapidly moving and shifting rivers do not have well-developed flood basins. In general, floodplain deposits are present along both braided and meandering rivers, although they are commonly noted along single-channel rivers.

Most fluvial deposits consist of sand and gravel, although mud is common in floodplain deposits of meandering streams (Figure 9.7). The sorting of most fluvial sediments ranges from moderate to poor. The deposits of point bars and braid bars generally display fining-upward grain size owing to the helical nature of sediment transport on bars (Figure 9. 7). Migration of meanders also produces a general fining-upward succession as channel-lag deposits (lag gravel) are overlain by fining-upward point-bar deposits and, in turn, silty and muddy floodplain deposits (Allen, 1970) (see Figure 9.9). Multiple episodes of meander migration produce vertical stacking of fining-upward successions in meandering-river deposits (see Miall, 1996). The lateral migration of braided rivers leaves sheet-like or wedge-shaped deposits of channel and bar complexes (Cant, 1982). Lateral migration combined with aggradation leads to deposition of sheet sandstones or conglomerates (lag deposits) that enclose very thin, non-persistent shales within coarser sediments.

Fluvial deposits commonly display abundant structures, such as planar and trough cross-bedding, upper-flow-regime planar bedding, and ripple-marked surfaces (see Figures 9.7, 9.9 and 9.13). Sedimentary structures yield unidirectional, downstream paleocurrent directions that tend to be more variable in meandering-river deposits than in braided-river deposits. Fluvial deposits may contain a variety of fossil hard parts of terrestrial animals as well as trace fossils created by both animals and plants (see Bridge, 2003; Jones and Schumm, 1999; Blum and Tornqvist, 2000). In general, the fluvial architecture is influenced by tectonics, climate, base levels, and channel types that control processes such as subsidence rates, slope changes, channel incision and aggradation, and channel migration and avulsion (Leeder, 1993). Floodplain deposits are economically important reservoirs of oil, natural gas, and water, and include significant coal reserves. They also provide detailed records of past and present geologic processes and continental environments.

In summary, the floodplain sediments are broadly grouped into two types: channel and overbank deposits. Channel deposits consist of channel-bar and channel-fill sediments. Channel-bar sediments are stratified sands and gravels that accumulate in response to river migration and changes in stream hydrology. Channel-fill sediments range from coarse-grained (active) to fine-grained (passive) deposits. Active fills are represented by interbedded sand and silt that accumulate in response to gradual channel abandonment largely by chute cutoff, whereas passive fills consist of silt and clay or organic sediments that are deposited within lakes or oxbows formed by neck cutoff (Fisk, 1947). Overbank deposits are fine-grained sediments consisting of clay, silt, and lesser amounts of sand with organic sediments (Flores, 1981). Overbank sediments accumulate during floods in natural levee, crevasse splay, and flood basins.

9.6.2 Natural Levees

When the stream floods and overtops its banks, deposition of fine sediment occurs on natural levees (Figure 9.14), in adjacent flood basins, and in oxbow lakes (see Figures 9.7–9.9). Deposition

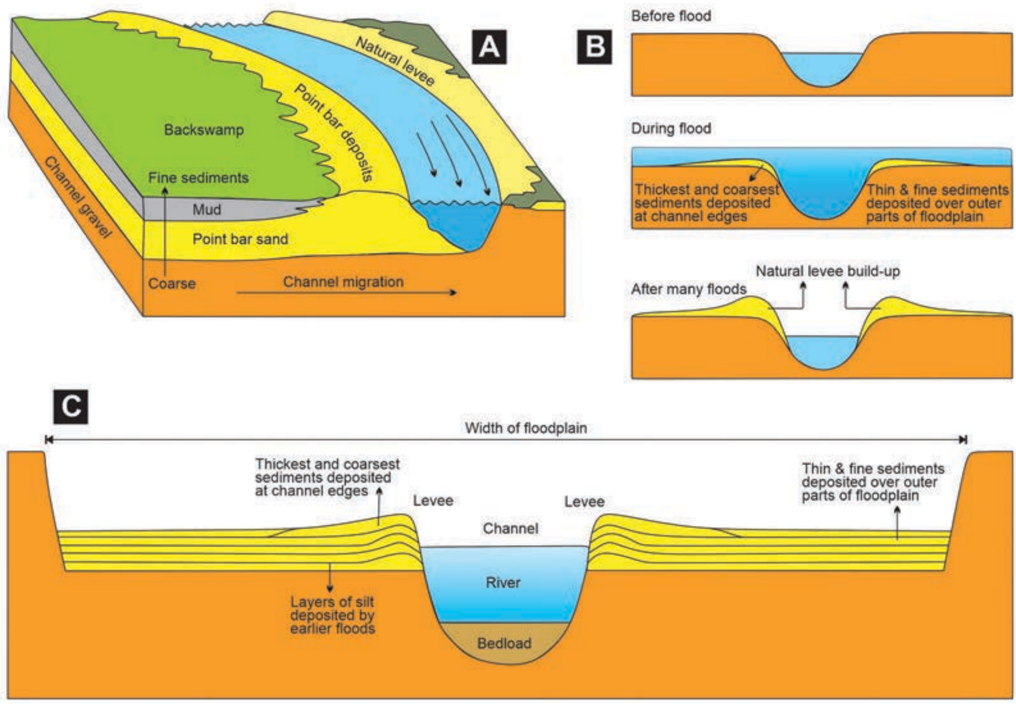

FIGURE 9.14 Levee formation. A: The repeated deposition of sand close to the channel edge leads to the formation of a natural levee. B–C: Natural levees are formed by the deposition of sediments when flood waters of a stream overtop its banks. As the velocity is reduced, much of the suspended sediments near the channel are deposited. Thus, the coarsest sediments are deposited near the channel and grain size decreases away from the channel.

from overbank waters results in upbuilding of the sediment surface and is called vertical accretion (Figure 9.14C), in contrast to the lateral accretion that takes place on point bars (Figures 9.8 and 9.9). Natural-levee deposits form primarily on the concave or steep-bank side of meander loops immediately adjacent to the channel as a result of sudden loss of competence, and they typically contain horizontally stratified fine sands overlain by laminated mud (Figure 9.14).

Thus, repeated deposition of sand close to the channel edge leads to the formation of a natural levee (Figure 9.14A), a bank of sediment at the channel edge that is higher than the level of the floodplain (Figure 9.14B), thereafter the levees gently slope from the riverbank into flood basins away from the channel (see Figure 9.14C). Hence, natural levees are wedge-shaped ridges of sediment bordering stream channels (see Figure 9.14C). Natural levees have widths of 2 to 3 km and are up to 5 m thick. Natural levees are formed by the deposition of sediments when flood waters of a stream overtop its banks (see Figures 9.14B–C). As the velocity is reduced, much of the suspended sediments near the channel are deposited. Thus, the coarsest sediments are deposited near the channel and grain size decreases away from the channel (see Figure 9.14C). Natural levees are best developed along the outer bends of meanders in low-gradient sinuous rivers with large suspended-sediment loads. The levee sediments are made up of somewhat finer-grained material than their corresponding point-bar sediments. However, the composition of the upper part of point-bar sediments is somewhat similar to the composition of levees as also the sedimentary structures that include small-ripple cross-bedding, climbing-ripple lamination, horizontal bedding, and parallel laminated mud layers. Natural-levee sediments are often rhythmically bedded sand and silt with lesser amounts of clay (Farrell, 1987). Sandy units (generally a few decimeters thick) are overlaid by muddy units (a few centimeters thick). Thus, in a vertical sequence, sandy and muddy layers alternate with each other. Surface features such as desiccation cracks or raindrop imprints, are commonly noted on the muddy surface. In ancient floodplain settings, natural-levee deposits consist of packages of interbedded fine sandstone, siltstone, and mudrock that commonly overlie channel-belt sandstones (Kraus, 1999).

9.6.3 Crevasse Splays and Avulsion

Crevasse-splay deposits may also occur on floodplains where rising floodwaters breach natural levees (Figures 9.4B and 9.7B) (see Kraus and Wells, 1999). Crevasse splays are tabular to lenticular sandy accumulations of sand and silt that accumulate in flood basins; they form when a levee breaks during high floods or flash floods (Figures 9.7C–D and 9.8). They are lobate to sinuous and contain distributary channel networks that may extend up to 10 km (Smith, 1986; Smith et at., 1989). The cone-shaped water-laden with sediments unit breaches the bank and moves out onto the floodplain; most overtopping occurs on concave banks (O'Brien and Wells, 1986). This narrow to broad tongue of water-laden sediments that cuts through the channel is called a crevasse splay (Figures 9.7C and 9.8). Later, these crevasses develop their own channel pattern and system.

Sometimes, crevasse channels may divert the main river discharge, causing a change in the river course (this process is called avulsion: the process where channels within crevasse splays develop into new river channels; see Figure 9.8). This breach at the bank (i.e., levee) does not occur instantaneously but is a gradually deepening and widening conduit for water to pass out onto the floodplain (Figure 9.8). Initially only a small amount of water and sediment passes through, but as the volume of water and the grain size of the carried detritus increases, overtopping occurs. Crevasse-splay deposits are typically lenticular, tens of centimeters to several meters thick, and are characterized by an initial upward coarsening, i.e., the crevasse-splay deposits are coarser grained than the associated natural-levee deposits in which they are embedded. The crevasse-splay deposits are coarser-grained than other types of overbank deposits. However, fine-grained crevasse-channel fills also occur (Smith and Perez-Arlucea, 1994). Sedimentary structures include small scale trough-cross-bedding and ripple lamination (Farrell, 1987). Upward-coarsening sequences reflect crevasse-splay progradation whereas deposition of overbank fines during waning flood stages produces upward-fining deposits (Smith and Perez-Arlucea, 1994).

9.6.4 Channel-Lag and Point-Bar Deposits

Sediments transported by rivers include silt and clay that are carried away much faster in suspension and deposited in flood basins (Figure 9.7). In the channel itself, the coarser sediments such as gravels and pebbles, lag behind, while the sand moves as bed load. These coarser sediments accumulate as discontinuous lenticular patches in the deeper parts of the channel as channel-lag deposits that are quickly covered by finer-grained sediments and preserved (see Figures 9.7D–F). In the lower reaches of rivers where coarser-grained sediments are not available, driftwood, unconsolidated sediments, mud pebbles, and dead organisms are concentrated as discontinuous and relatively thin-layered channel-lag deposits. These channel-lag deposits (basal lag gravel) occupy the lowest part of a channel or point-bar sequence and, if present, indicate the base of the channel (Figures 9.7D–F).

A channel moving sideways by erosion on the outer bank and deposition on the inner bank is undergoing lateral migration, the deposit on the inner bank is referred to as a point bar (see Figure 9.7C). The size and shape of point bars vary with the size of the river. In smaller streams, point bars are simple depositional features on the convex sides of the meanders dipping gently toward the channel, but in large rivers, they are composed of scroll-shaped ridges (scroll bars) alternating with depressions (swales) (see Figures 9.7A and C). Repeated channel migration during floods produces scroll bars (see Figure 9.7C). Swales are filled with fine-grained muddy sediments (silt and clay), or marshes may develop in them. Deposition on point bars results from lateral migration of a meandering river during flooding (see Figure 9.7B).

A point-bar deposit will show a fining-up from coarser material at the base to finer at the top with larger scale cross-bedding at the base and smaller sets of cross-laminations near the top (see Figures 9.7B and 9.9). As the channel migrates, the top of the point bar becomes the edge of the floodplain, and the fining-upward succession of the point bar is capped by overbank deposits. Stages in the lateral migration of the point bar of a meandering river are reflected as inclined surfaces within the channel-fill succession (Figure 9.9A). These lateral accretion surfaces (Figure 9.9A) are best developed when low discharge allows a layer of finer sediment to be deposited on the point-bar surface (Allen, 1965; Bridge, 2003; Collinson et al., 2006). These surfaces are low angle, less than 15°, and are inclined from the river bank towards the deepest part of the channel, i.e., they are perpendicular to the flow direction (Figure 9.9A).

In general, sedimentary units of point bars are discontinuous and lenticular in nature. Deposition at the point bar is very rapid. In an ideal sequence of point-bar, large-scale cross-bedding (megaripple bedding) is abundant in its lower part (Figure 9.9C). The thickness of units decreases upward (1 m or more at the lowermost part to only a few centimeters thick upwards). Above the megaripple bedding, small-ripple cross-bedding and climbing-ripple laminations are noted, followed by units of horizontal bedding. In many cases, small-ripple bedding is interbedded with horizontal bedding. On the top are silty and clayey layers rich in organic matter. Point-bar deposits also show accumulations of drifted plant material, freshwater mollusks, and mud pebbles, among others. In the lower part of the point-bar sequence, several channel-lag layers are noted. However, this ideal profile is not always well developed.

9.6.5 Channel-Fill Deposits (Cutoff Channels and Oxbow Lakes)

Channel-fill deposits represent sedimentation in a meandering channel that has been abandoned by a stream (Figure 9.8) due to (a) cutoff processes or avulsions, i.e., by the sudden abandonment of a part or the whole of a channel course, or (b) by filling up due to increased rate of sedimentation and the subsequent reduction in channel depth.

The recognition of lateral accretion surfaces (also known as epsilon cross-stratification; see Allen, 1965) (see Figure 9.9B) within the fining-up succession of a channel-fill deposit is a reliable indication that the river channel was meandering (Figure 9.9B). The outer bend of a meander loop is a bank made up of floodplain deposits of mainly muddy sediments (Figures 9.9A–B). Dried mud forms clasts that are carried by the river flow and are deposited along with sand in the deeper parts of the channel; they are preserved in the basal part of the channel-fill succession. During periods of high-stage flow, water may take a short-cut over the top of a point bar. This flow may become concentrated into a chute channel that cuts across the top of the inner bank of the meander (see Figure 9.8). Chute channels are semi-permanent features of a point bar, as they are only active during high-stage flow. They are identified as a scour that cuts through lateral accretion surfaces of a meandering-river deposit. The river flow may also take a short-cut between meander loops when the river floods: this may result in a new section of channel developing, and the longer loop of the meander built becomes abandoned; the abandoned meander loop becomes isolated as an oxbow lake (see Figure 9.8). The deposits of an oxbow lake are fine-grained, sometimes carbonaceous, sediments.

Two types of stream cutoff are noted: chute cutoff and neck cutoff (see Fisk, 1944) (Figure 9.8). The chute cutoff occurs when a stream in a meander loop shortens its course by cutting a new channel along a swale of a point bar (Figure 9.8). The neck cutoff occurs when a stream cuts a new channel through the narrow neck between two meander loops (Figure 9.8). The abandoned channels are slowly alluviated and sealed at both ends, isolating the old channel loop in the form of a cutoff lake, or oxbow lake (Figure 9.8).

It must be noted that accumulation and possible preservation of river channel deposits occur only if the river changes its position in some way, either by shifting sideways, or if the channel changes position on the floodplain; this process is called avulsion (Figure 9.88). When a river avulses, part of the old river course is completely abandoned and a new channel is formed (Figure 9. 8). An example of the abandonment of a short stretch is called an oxbow lake (Figure 9.8). When avulsion occurs, the flow in the old river course is reduced in volume and thus slows down, and the bedload is deposited. A decrease in the water supply reduces the capacity of the channel to carry sediments, making the flow sluggish, thereby depositing its suspended load. Abandoned and empty stretches of river channel are unlikely to remain empty for very long as when the river floods from its new course it will carry sediment across the floodplain to the old channel where sediment gradually accumulates. Hence, the final fill is a fine-grained overbank sedimentation related to a different river course.

Sedimentation is rather rapid at and near the ends of the cutoff lakes. Mainly clayey sediments and organic matter are deposited until filling is complete within the cutoff lakes. Clay and silt are most abundant with sand in minor amounts. Sandy units are cross-bedded and mud layers are laminated. The channel-fill succession in both meandering and braided rivers is built up as a result of sideways movement or lateral migration of the active part of the channel (Figures 9.4 and 9.8).

9.6.6 Flood-Basin Deposits

Flood basins are the lowest-lying part of a river floodplain. They are flat, poorly drained and featureless areas of little or no relief located adjacent to active or abandoned stream channels. They act as settling basins for suspended fine-grained sediments (mostly fine silt and clay) that settle from overbank flows after the coarser sediments have been deposited on levees and crevasse splays. In general, flood basin deposits are mostly finely laminated mud, interrupted by some sandy or silty intercalations. Organic debris, mottled structures, desiccation, and mud cracks are common. Bioturbation is noted when the rates of sediment accumulation are slow. The flood-basin sediments are finer than the corresponding natural levee, crevasse-splay, and point-bar deposits. A flood basin is often elongated, with low-lying areas running parallel to the channel. Braided streams with their rapid rates of lateral migration, and actively shifting meandering channels inhibit the development of thick flood basin deposits. Thick flood basin deposits are only produced if streams are fixed in their position, so that longer periods of deposition are available. A silt-clay layer 1 or 2 cm thick is usually deposited during one flood period. Large amounts of organic matter with shallow fluctuating water levels generally produce gray sediment colors and brown or red iron-oxide mottles.

9.7 DIAGNOSTIC FEATURES OF A MEANDERING SYSTEM

Meandering channels, regardless of the size of the river, are geometrically self-similar. For example, on average, the typically meander wavelength is 8–12 times the bankfull channel width and meander-bend curvature is about 2–3 times the channel width. As channel width (W) is related to discharge (Q), meander wavelength (L) is also related to river discharge as L= k/Q (where k ranges from 30 to 60, depending on the return period of Q) (see Figure 9.7A). This relationship is used in paleohydrologic studies of ancient rivers (see Dury, 1964). However, meandering channels also display non-periodic or random behavior where the regular meander-bend development is distorted by heterogeneities in the alluvium through which bends are migrating; meander bends can migrate laterally to form complex meander loops that are much larger than the associated meander bends, and high sinuosity meander bends can erode back on themselves to form meander cutoffs.

The meander geometry is controlled by its lateral migration. Meandering channels erode the outer banks of channel bends and maintain a constant channel width by achieving a matching rate of deposition on the point bar forming the inner bank (see Figure 9.7A). Rates of migration may be exceedingly slow (a small fraction of channel width/year) or very rapid (several channel widths/

year) depending on the stream power/bank-strength ratio, and on the degree of bend curvature (see Hickin and Nanson, 1975, 1984).

Meandering rivers transport and deposit a mixture of suspended and bed load (i.e., mixed load) (Schumm, 1981; Schumm and Ethridge, 1994). The bed load is carried by the flow in the channel, with the coarsest material carried in the deepest parts of the channel. Finer bed load is carried in shallower parts of the flow and is deposited along the inner bend of a meander loop where friction reduces the flow velocity. The deposits of a meander bend have a characteristic profile of coarser material at the base, becoming progressively finer-grained up the inner bank. The faster flow in the deeper parts of the channel forms subaqueous dunes in the sediment that develop trough or planar cross-bedding as the sand accumulates. Higher up on the inner bank where the flow is slower, ripples form in the finer sand, producing cross-lamination. The nature of sediments deposited by a meandering channel is related to (a) sediment availability for point-bar deposition, and (b) the process of lateral migration (see also Allen, 1963; but see also Miall, 1985). Within-channel lateral accretion deposits of gravel overlain by sand are capped by overbank fines deposited during flood discharges. Lateral accretion deposits consist of a basal gravel platform of bed material on which are deposited upward-fining sands and silts representing the declining energy of the depositional environment as the point bar aggrades to less frequently inundated elevations. Internal sedimentary structures vary from dune to ripple-related cross-beddings and trough cross-beddings in the lateral accretion deposits to horizontal (laminated) strata in the overbank deposits.

Meandering systems are associated with floodplain muds and lake deposits. They grade downstream into the deltaic systems and upstream into a braided system. Channel sequences typically form long, ribbon-like bodies of sand ("shoestring sands") within a thick sequence of shales. Channel sands may be scattered randomly throughout the sequence, depending on where the channel shifts after avulsion. As in the braided system, there is a fining-upward sequence, from a coarse basal channel-lag gravel to the finer sandy point-bar sequence of plane beds, trough cross-beds, and ripple drift. Unlike the braided system, however, meandering systems have a much larger fine-grained component of laminated muds formed in oxbow lakes, natural levees, crevasse splays, and floodplains. In meandering systems, the grain sizes range from coarse channel-lag gravels to finer floodplain muds. Laterally accreted point-bar sands show decreasing-flow-velocity sedimentary structures such as plane beds, trough cross-beds, and ripple cross-laminations. Floodplain muds are finely laminated and vertically accreted; climbing-ripple drift, mud cracks, raindrop impressions, soil horizons, organic matter, and fossils are noted. In the floodplain, organic matter and fossil wood are common. Land vertebrates and invertebrates occur in floodplain muds or in the channel sands. Freshwater mollusks are particularly characteristic. Table 9.1 provides a comparative account of major characteristics between meandering and braided channels.

TABLE 9.1
Properties of meandering vs. braided streams. Varying characteristics between meandering and braided channels

Characteristics	Meandering rivers	Braided rivers
Discharge (Q)	stable-moderately variable	highly variable
Sediment load	suspended > bedload	high bed load
Bank erodibility	low to moderate	high
Bank composition	clay, till, or silt	sand, gravel
Bank vegetation	good	poor
Channel gradient	moderate-to-low	moderate-to-high
Width/Depth	low	high

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10 Lacustrine System

10.1 INTRODUCTION

Lakes (Figure 10.1), numbering more than 117 million, cover about 4% of earth's total land surface. Finland, known as "The Land of a Thousand Lakes," is home to 187,000 lakes. The Caspian, located in central Asia, is the world's largest inland lake (Figure 10.2A), having an area of 386,000 square km with 40–44% of the total lake waters of the world. Lake sedimentation, consisting mainly of clastic material such as clay, silt, and sand-sized grains (Figure 10.2B) is more common in the present than noted in the geologic past (Lahijani and Tavakoli, 2012; Lahijania et al., 2018). The lacustrine depositional systems, or lakes, form in topographic lows and are generally low-energy systems. These lake sediments (Figures 10.2B–C) serve as important proxies for climatic changes, where sediment chemistry and mineralogy have been used to infer changes in episodes of climate (Xu et al., 2017; Jean-Philippe et al., 2019; Müller et al., 2020). Some lake sediments also contain economic deposits of oil shales, evaporite minerals, coal, uranium, and iron (Acosta-Góngora et al., 2018); many lake sediments have abundant organic matter that acts as source material for petroleum deposits (see Katz, 2001; Bohacs et al., 2000). The lake-associated rocks make up ~20% of worldwide hydrocarbon production (Calhoun, 1999), and lacustrine organic-rich rocks are significant sources of hydrocarbon deposits notably in Africa, South America, Southeast Asia, and China (Hedberg, 1968; Katz, 1995).

10.2 LAKE ZONATION

To understand lake types and the mechanisms operating within them, it is imperative to first understand the structure of a lake. Based on depth and distance from the shoreline, four lake zones are recognized: littoral, limnetic (pelagic), profundal, and benthic (Figure 10.3).

10.2.1 Littoral Zone

This is the topmost zone in a lake, and an area near the shore and includes coral reefs, rocky coasts, sandy beaches, and sheltered embayments (Figure 10.3A). It is also the most productive zone due to its availability of abundant energy from sunlight, nutrients from the land runoff, and dissolved oxygen. The zone is also characterized by high wave energies and water motion, with alternating submergence and exposure in the intertidal subzone.

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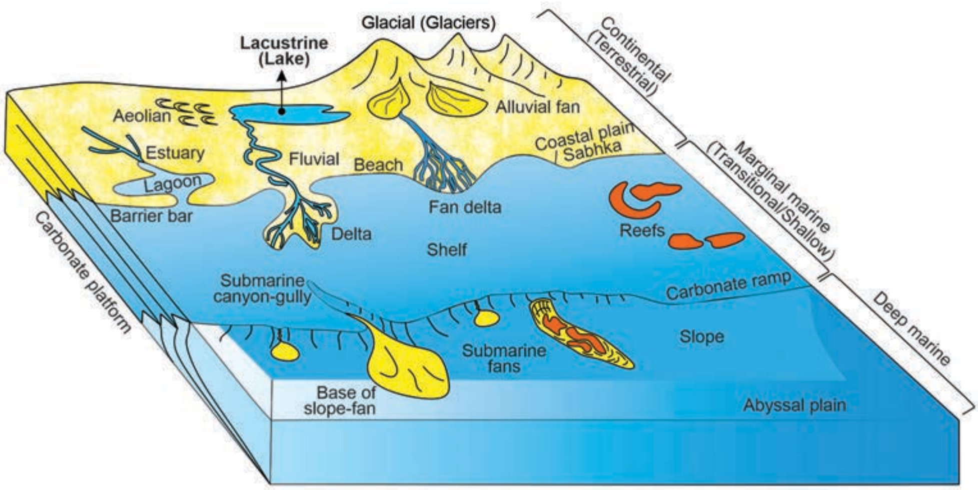

FIGURE 10.1 Depositional environments. The lacustrine depositional environment is highlighted.

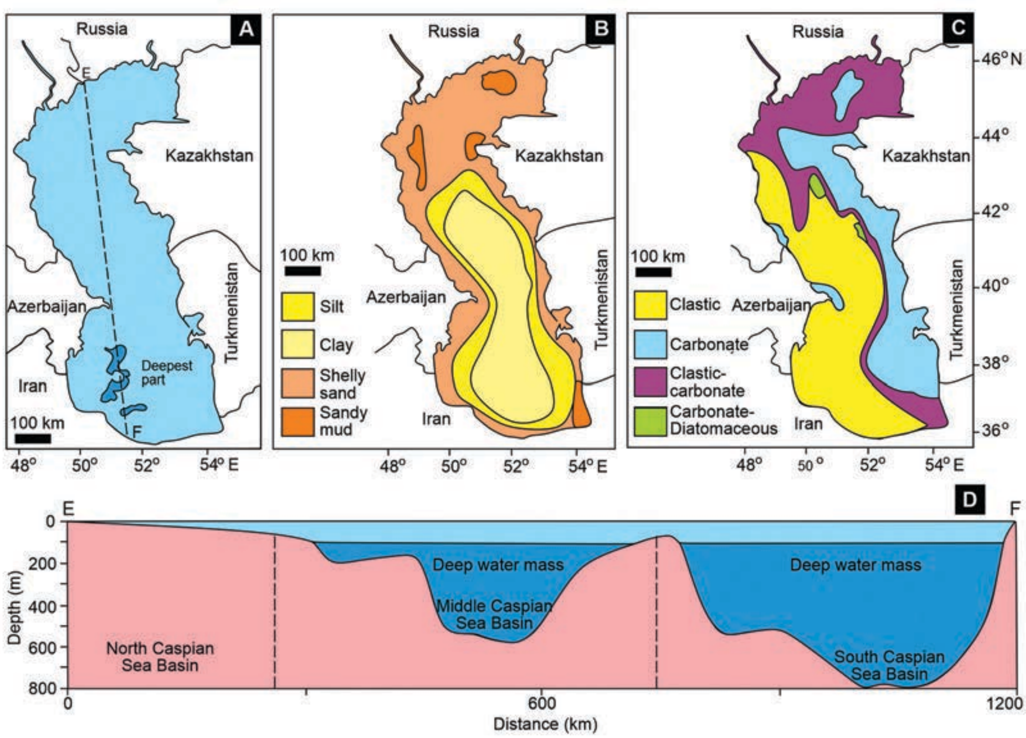

FIGURE 10.2 Caspian Sea. The distribution map of Caspian Sea surface sediments in terms of depth (A and D), grain size (B) and (C) composition. (Modified after Lahijani et al., 2018.)

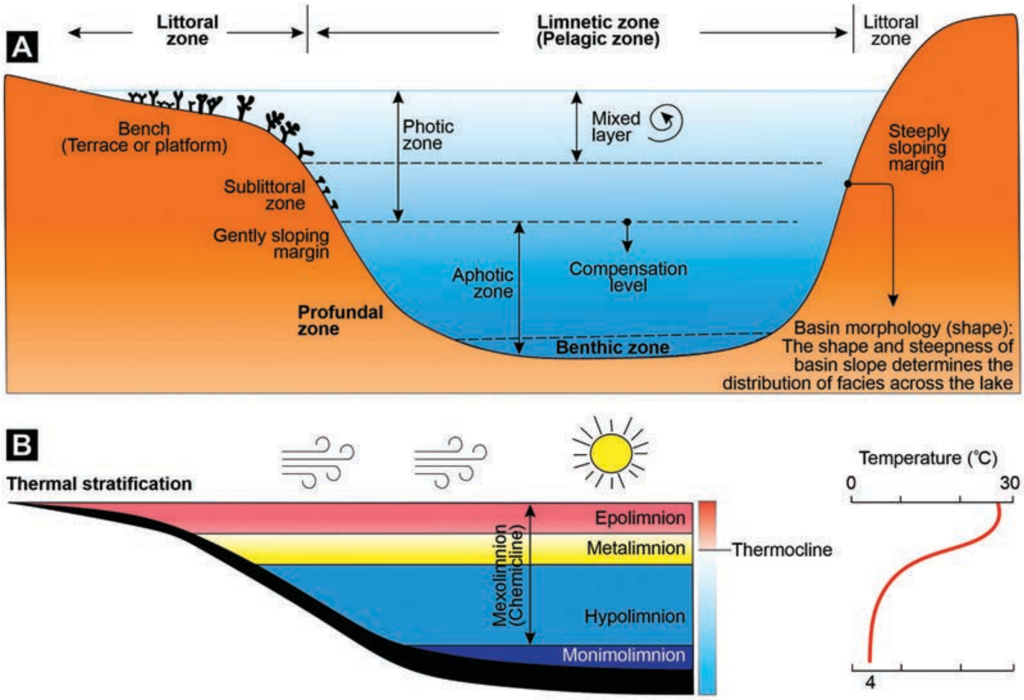

FIGURE 10.3 Lake morphology. A: Lake zonation. B: Lake stratification.

10.2.2 Limnetic (Pelagic) Zone

This is the sunlit part of the lake surrounded by the littoral zone and extends to depths where the sunlight can penetrate (Figure 10.3A). In deeper freshwater ecosystems, this zone lies above the light compensation level and beyond the littoral zone (Figure 10.3A). Most photosynthesis occurs in this part of the lake; hence it is often referred to as photic or sublittoral zones (Figure 10.3A). This zone is also populated by both phytoplanktons (microscopic organisms at the upper sunlit layer) and zooplanktons.

10.2.3 Profundal Zone

This zone is located beyond the range of sunlight penetration (i.e., part of the aphotic zone); it is much colder and denser than the previous zones (Figure 10.3A). It is located below the thermocline where the sunlight does not penetrate; the thermocline is where the temperature drops very rapidly (see Figure 10.3B). The process of photosynthesis is not possible in this zone; hence, low levels of photosynthesis lead to low levels of oxygen. Organisms in this zone are thus, adapted to cooler temperatures and lower levels of oxygen.

10.2.4 Benthic Zone

This zone is at the very bottom of a lake (Figure 10.3A). Organisms inhabiting this zone are called benthos. They live in a close relationship with the bottom substrate and some remain permanently attached to the bottom layer of the lake. The benthic boundary layer, which is the superficial layer of soil lining the body of water, contains the necessary nutrients to support an array of microscopic life. This zone is mainly inhabited by decomposers and insect larvae.

10.3 LAKE WATER MIXING AND THERMAL STRATIFICATION

Wind, waves, heat transfer, and in- and outflows to the lake enable lake water mixing. Water is most dense at 4° C (39 F) (Figure 10.3B). Denser water is heavier and sinks to the bottom of a lake while the less dense water is lighter and remains at the top of the lake (Figure 10.3B). This division into different layers of density due to differing temperatures is called thermal stratification (Figure 10.3B) (see Gorham and Boyce, 1989).

Deep lakes generally become physically stratified into three identifiable layers, known as the epilimnion, metalimnion, and hypolimnion (Figure 10.3B). The epilimnion is the upper, warm layer, and is typically well mixed (Figure 10.3B). Below it is the metalimnion or the thermocline region, a layer of water in which the temperature declines rapidly with depth (Figure 10.3B). The hypolimnion is the bottom layer of colder water, isolated from the epilimnion by the metalimnion (Figure 10.3B). The depth of mixing depends in part on the exposure of the lake to wind (its fetch), but is most closely related to the lake's size. Smaller to moderately sized lakes (50 to 1000 acres) stratify and are well mixed to a depth of 3–7 meters in north temperate climates. Larger lakes may be well mixed to a depth of 10–15 meters in summer (such as the Western Lake Superior near Duluth, Minnesota, USA). Note that although "thermocline" is a term often used synonymously with metalimnion, it is actually the plane or surface of maximum rate of decrease of temperature with respect to depth (Figure 10.3B). Thus, the thermocline is the point of maximum temperature change within the metalimnion (see Figure 10.3B).

In summer (Figure 10.4A), the sun heats the top layer of a lake, the epilimnion, and causes the water to become less dense (Figure 10.4A). The hypolimnion, the bottom layer of the lake, does not receive sunlight and therefore remains cold (Figure 10.4A). As the epilimnion is less dense, it floats on top of the hypolimnion and the two do not mix. Around the shoreline of a lake (i.e., within the littoral zone; see Figure 10.3A), the area where sunlight penetrates (and part of the epilimnion), algae, zooplanktons, and vascular plants grow. When they die, they sink to the bottom of the lake. Invertebrates and microbes living in the benthos recycle and decompose this dead material.

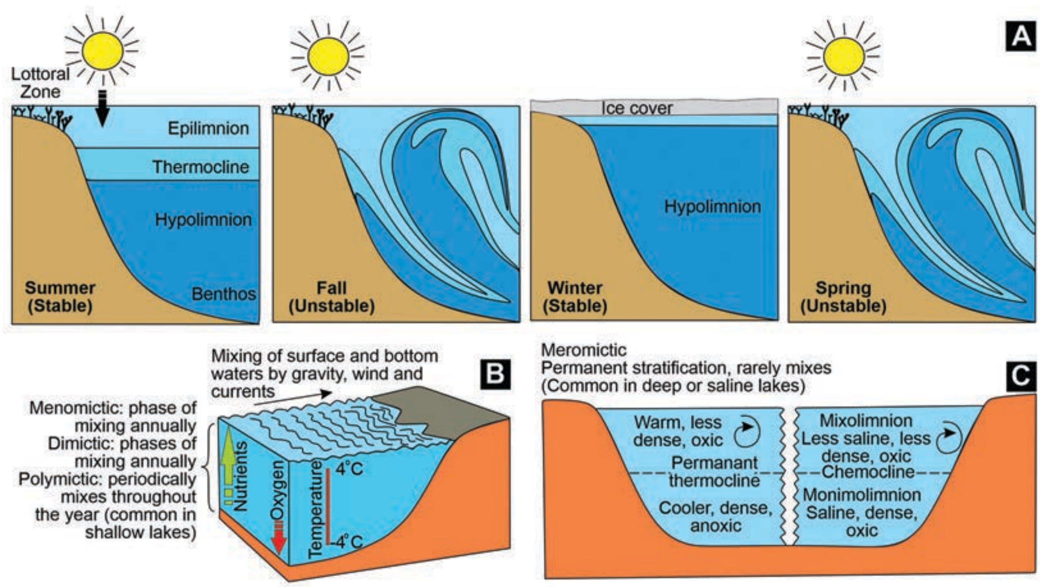

FIGURE 10.4 Lake stratification and mixing. A: Pattern of stratification for temperate lakes. B: Dimictic lakes are lakes with two mixing periods. C: Meromictic lakes are partially mixed lakes.

This recycling process uses up oxygen. As the lake does not mix during the summer, the hypolimnion is completely cut off from the epilimnion and hence does not receive a fresh supply of oxygen. Therefore, the hypolimnion becomes anoxic during the summer within a mesotrophic or eutrophic lake.

In fall (Figure 10.4A), the sunlight is not as strong and the nights are much cooler. This change in season allows the epilimnion to cool off. As the water in the epilimnion cools, the density difference between the epilimnion and hypolimnion is greatly reduced (Figure 10.4A). Wind can now mix the layers. In addition, when the epilimnion cools it becomes denser and sinks to the hypolimnion, mixing the layers (Figure 10.4A). This mixing allows oxygen and nutrients to be distributed across the whole water column.

In winter (Figure 10.4A), the lakes are covered with ice. Under ice, the water cannot mix as it is not exposed to wind. Most of the hypolimnion remains around 4° C (39 F) (Figure 10.4A). However, there is a thin layer of water under the ice that is colder than 4° C and therefore less dense. This thin layer of water floats on top of the hypolimnion throughout the winter (Figure 10.4A), but this stratification is not quite as stable as it is in the summer due to reduced density difference. This phenomenon is called inverse stratification as cooler water is sitting on top of warmer water (Figure 10.4A).

As in the summer, the hypolimnion is cut off from oxygen, so as decomposition takes place in the benthos, oxygen gets used up. When the hypolimnion becomes anoxic in the winter it is called winter kill as fish, and other living organisms that need oxygen, die. In addition, when the bottom of the lake is anoxic, chemical processes at the sediment/water interface cause phosphorus to be released from the sediments. When the ice melts in the spring and the lake mixes again, this increased phosphorus fuels algal growth (Figure 10.4A).

In spring, the ice melts, the wind picks up, and the lake mixes again (Figure 10.4A). This is called turnover or spring turnover. Oxygen and nutrients get distributed throughout the water column as the water mixes (Figure 10.4A). Then, as the weather becomes warmer, the surface water warms again and sets up the summer stratification. When lakes mix twice a year, i.e., spring and fall, such lakes are called dimictic (Figure 10.4B). Shallow lakes behave differently and can mix more often.

Thus, this pattern (summer stratification – fall turnover – winter stratification – spring turnover) is typical for temperate lakes (Figure 10.4A). Lakes with this pattern of two mixing periods are referred to as dimictic (Figure 10.4B). Many shallow lakes, however, do not stratify in the summer, or stratify for short periods only, throughout the summer. Lakes that stratify and de-stratify numerous times within a summer are known as polymictic lakes (see Figure 10.4B). Much less common are lakes that circulate incompletely, resulting in a layer of bottom water that remains stagnant. To distinguish them from the holomictic (mixing from top to bottom) lakes, these partially mixing lakes are referred to as meromictic (Figure 10.4C). They mix partially, in the sense that they may have extensive mixing periods which go quite deeply into the hypolimnion, but they do not turn over completely, and a layer of bottom water remains stagnant and anoxic for years at a time. The non-mixing bottom layer is known as the monimolimnion and is separated from the mixolimnion (the zone that mixes completely at least once a year) by a chemocline (Figure 10.3B). The chemocline is a sharp gradient in chemical concentration; the boundary in a meromictic lake separating an upper layer of less-saline water that can mix completely at least once a year (mixolimnion) from the deeper, more saline (dense) layer (monimolimnion) that is never mixed into the overlying layer (Figure 10.3B). Stagnant, and typically anaerobic, the monimolimnion has a high concentration of dissolved solids as compared to the mixolimnion. In general, meromictic lakes have large relative depths. These lakes are typically small and sheltered from the wind by the morphology of their basin. In this case, the density differences caused by temperature are smaller than density differences due to the high dissolved solids (salts) concentration of the monimolimnion. Large lakes that rarely freeze over are also typically monomictic, mixing throughout the fall, winter, and spring and stratifying in the summer.

10.4 CLASSIFICATION OF LAKES

Modern lakes occur in a variety of environmental settings, including glaciated inland plains and mountain valleys, non-glaciated inland plains and mountainous regions, deserts, and coastal plains (Figure 10.5; see also Table 10.1). They also exist in different climatic conditions ranging from very hot to very cold and from highly arid to very humid. Lakes are also associated with various depositional systems, such as glacial, fluvial, aeolian, and deltaic (Figure 10.5), and are classified in several ways; two major classifications are listed in Table 10.1. The major ones are briefly enumerated below.

10.4.1 Classification Based on Origin

Lakes are formed in basins or depressions as a result of: (a) glacial processes such as ice scouring, landslide-, ice-, and moraine-damming (Figures 10.5A–B); (b) tectonic movements such as faulting and rifting (Figure 10.5C); (c) volcanic activity such as lava damming or crater explosion and collapse (Figure 10.5D); (d) fluvial activity such as the formation of oxbow and levee lakes (Figure 10.5E); (e) man-made lakes, and (f) deflation by wind scour or damming by windblown sand (Figure 10.5G).

TABLE 10.1
Classification of lakes

Classification of lakes	
Classification 1	**Classification 2**
Classification based on size	**Classification based on type**
Small lakes	Temporary lakes
Medium lakes	Permanent lakes
Large lakes	Freshwater lakes
Great lakes	Saline lakes
Classification based on origin	**Lakes formed by earth movement**
Glacial lakes	Tectonic lakes
Tectonic lakes	Rift valley lakes
Volcanic lakes	
Solution lakes	**Lakes formed by glaciation**
Man-made lakes	Cirque lakes or tarns
	Rock-hollow lakes
Classification based on outflow of water	Lakes due to morainic damming of valleys
Open lakes	
Closed lakes	**Lakes formed by volcanic activity**
	Crater and caldera lakes
Classification based on trophic availability	
Oligotrophic lakes	**Lakes formed by erosion**
Mesotrophic lakes	Karst lakes
Eutrophic lakes	Wind-deflated lakes
Classification based on water chemistry	**Lakes formed by deposition**
Acidic lakes	Lakes due to river deposits
Alkaline lakes	Lakes due to marine deposits
Saline lakes	Lakes due to damming of water
Eutrophic lakes	Man-made lakes
Oligotrophic lakes	Crater lakes

Many present lakes are formed directly or indirectly by glacial processes (Figures 10.5A–B) (Picard and High, 1981); for example, the Great Lakes and most of the 10,000 lakes of Minnesota (USA). Some large modern lakes are also formed by tectonic (such as Lake Tanganyika in the East African rift system, and Lake Baikal in the Baikal rift system in Siberia) (Garcia-Castellanos, 2006) (see Figure 10.5C) and by volcanic processes (such as the Crater Lake, Oregon; volcanic calderas: Crater Lake) (Figure 10.5D). This is contrary to ancient ones that were largely formed by tectonic processes (Figure 10.5C). Of the twenty-five largest lakes by surface area today, ten are of glacial origin, seven occupy cratonic depressions, and four are in rift valleys (Smith, 1990). Lakes also form in karst sinkholes and meteorite craters, or are impounded in river valleys behind glacial moraines, lava flows, alluvium, or landslide debris.

10.4.2 Classification Based on Size

Modem lakes stretch from a few tens of square meters to tens of thousands of square kilometers. The largest modem lake is the saline, inland Caspian Sea with a surface area of 436,000 km^2 (Van der Leeden, 1975) (Figure 10.2). Other large lakes with surface areas ranging between 50,000 and 100,000 km^2 include Lake Superior, Lake Huron, and Lake Michigan in North America; Lake Victoria, located between Uganda and Kenya in east-central Africa; and Lake Aral, east of the Caspian Sea. Water depths of modern lakes range from a few meters in small ponds to more than 1700 m in the world's deepest lake, Lake Baikal, Siberia.

Small lakes are shallow and are typically <1 km^2 in size, often referred to as ponds. They have a maximum depth of ~6 m, and are found in a variety of environments such as forests, meadows, and wetlands. The medium lakes range in size from 1 to 100 km^2, with a maximum depth of ~30 meters. They are deeper than the small lakes and often have clearer waters. They are found in mountainous regions and coastal areas. The large lakes are typically >100 km^2 in size with a maximum depth of hundreds of meters. They are found in mountainous regions, coastal areas, and inland basins. The Great Lakes are a group of five large lakes located in North America (Figure 10.6). They are Lake Superior, Lake Michigan, Lake Huron, Lake Erie, and Lake Ontario; together, they form the largest group of freshwater lakes in the world, with a total surface area of around 244,000 km^2 (Figure 10.6).

There exists a confusion in the usage of the terms for the two water bodies – lakes and ponds. However, they differ in terms of size, depth, amount of light penetration, and the species of flora and fauna that inhabit them (see Table 10.2). But the main difference is their size. Lakes are larger than ponds, and they are usually deeper as well. However, there is no strict size requirement that defines a body of water as a lake or a pond, so the distinction remains somewhat subjective. Small lakes are often colloquially referred to as ponds. To make the distinction clearer, the general rule of thumb is that lakes tend to be larger than ponds.

10.4.3 Classification Based on Outflow of Water

10.4.3.1 Open Lakes

Open lakes are those that have an outflow of water with a relatively stable (fixed) shoreline and in which inflow and precipitation are approximately balanced (≥; equal to or greater than) by outflow and evaporation (Figures 10.7A–B). Siliciclastic sedimentation commonly predominates; however, chemical sedimentation can also occur in lakes that have a low supply of clastic sediments. In fresh-water lakes, a change in pH due to biological activity causes precipitation of calcite in the form of marls (chemical sedimentation).

The sediments of most hydrologically open lakes are dominated by siliciclastic deposits, derived mainly from rivers, but possibly including windblown, icerafted, and volcanic detritus. Much of this sediment is deposited near river mouths or along the shores of lakes. Gravelly sediments may

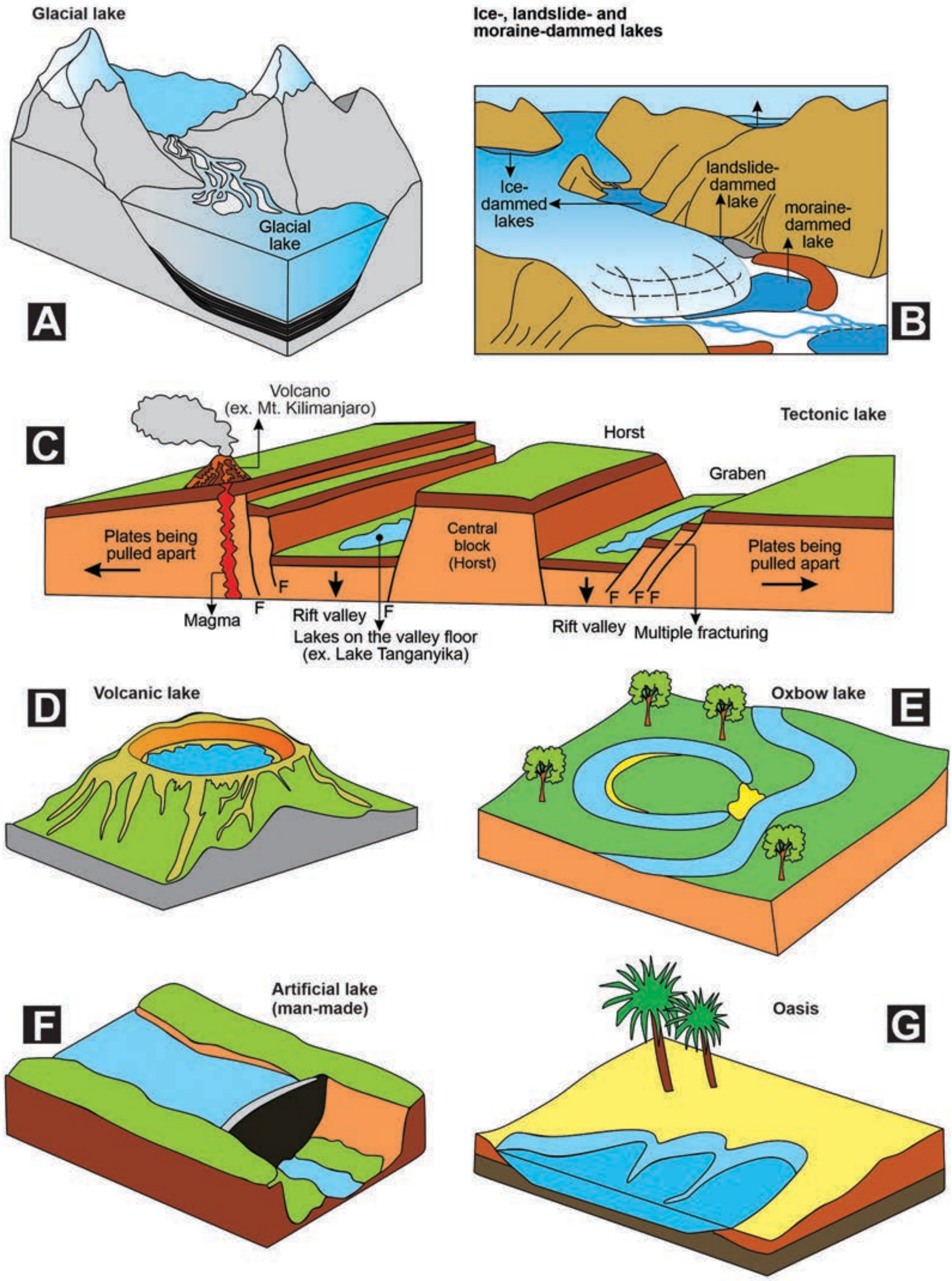

FIGURE 10.5 Lake classification based on their origin.

FIGURE 10.6 The Great Lakes. A: These are a group of five large lakes (Lake Superior, Lake Michigan, Lake Huron, Lake Erie, and Lake Ontario) located in North America. They form the largest group of freshwater lakes in the world, with a total surface area of around 244,000 km^2. B: Depth of The Great Lakes.

TABLE 10.2
Differences between a lake and a pond

Lake	Pond
A lake is a large body of water surrounded by land that is either open or closed.	A pond is a relatively small water body.
Lakes are thousands of meters deep.	Ponds are only a few feet deep.
Lakes have more diverse flora and fauna.	Ponds have less diverse flora and fauna.
The sunlight doesn't fully penetrate lakes.	The sunlight reaches throughout the area of a pond.

be present in the toes of alluvial fans or fan deltas that extend to the lake edge or into the lake. Sand accumulates mainly along the lakeshore in deltas, beaches, spits, or barriers. Sand may also be carried by turbidity currents into the middle of the lake; however, deeper parts of the lake are characterized by the presence of fine silt and clay.

In density-stratified lakes, muddy sediment is carried as a turbidity interflow above cold, denser lake waters. Coarser particles in such interflows settle fairly quickly and accumulate as silt layers. Finer particles settle more slowly to form clay layers. Thus, the siliciclastic deposits of open lakes consist of deltaic sands and muds (and possibly alluvial-fan gravels), turbidite sands and silts, and homogeneous to laminated muds.

In open lakes where the clastic sediment supply is low, chemical and biochemical processes predominate, resulting in the deposition of largely chemical sediments. Primary inorganic carbonate precipitation (caused by loss of CO_2 through plant photosynthesis and/or increase in water temperature or mixing of water masses) and the production of shells (by calcium carbonate- or silica-secreting organisms) account for most of the sedimentation. The invertebrate remains in lacustrine sediments include bivalves, ostracods, gastropods, diatoms, and charophytes and algae. Chemical

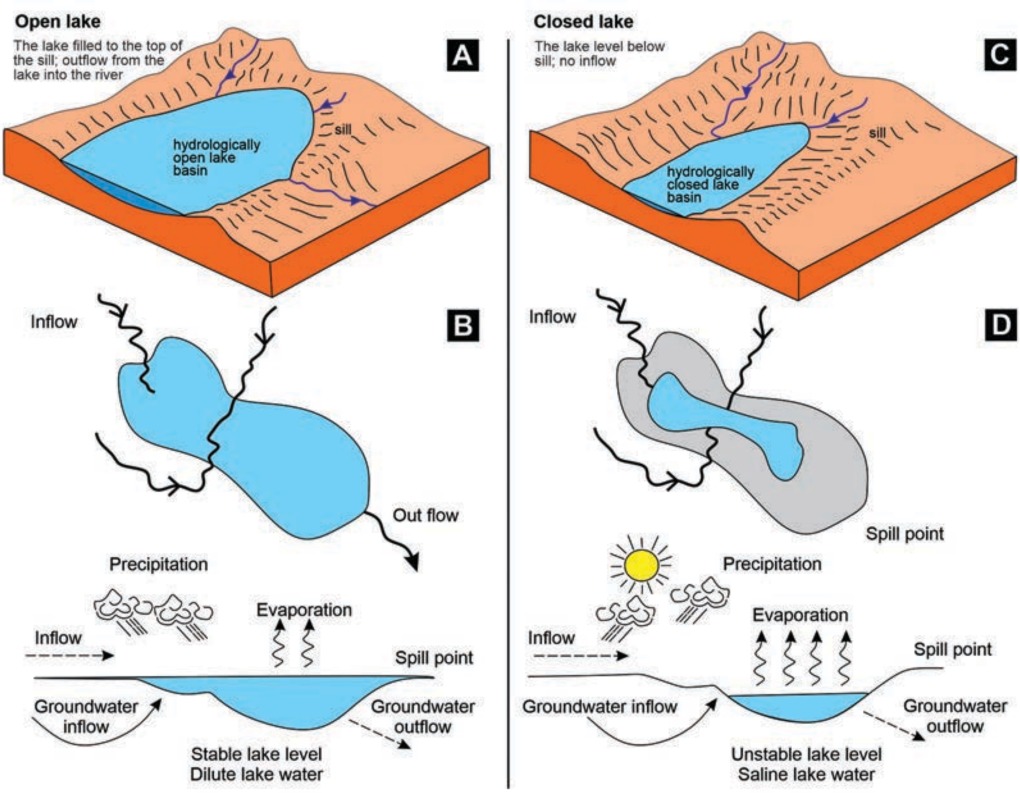

FIGURE 10.7 Open and closed lakes. A: Hydrologically open or open (through-flowing) lake. The lake is filled to the spill point and there is a balance of water supply into and out of the basin; characterized by a low concentration of dissolved salts and hence, low salinity. B: Example: Great Lakes (USA/Canada). Annual inflow + direct precipitation ≥ evaporation + annual outflow. C–D: Closed lakes or hydrologically closed (endorheic) or closed (terminal) lake. C: The rate of evaporation exceeds or balances the rate of water supply and there is no outflow from the lake; characterized by a high concentration of dissolved salts and hence, high salinity. D: Example: Great Salt Lake (Utah, USA). Annual inflow + direct precipitation < evaporation + annual outflow.

lake deposits consist mainly of carbonate sands and muds (less commonly siliceous diatom deposits). Stromatolites produced by blue-green algae (cyanobacteria) are also common in some lake deposits. Carbonate sediments may interfinger along the lake margin with siliciclastic deltaic or alluvial deposits.

10.4.3.2 Closed Lakes

Closed lakes do not have a major outflow and have fluctuating shorelines; inflow commonly exceeds evaporation and infiltration (Figures 10.7C–D). These conditions result in the concentration of ions in lake water and thus, a predominance of chemical sedimentation occurs; siliciclastic sediments may also accumulate. In a closed inland lake/basin or an isolated sea, four minerals (gypsum ($CaSO_4.2H_2O$), anhydrite ($CaSO_4$), halite (NaCl), and sylvite (KCl)) precipitate to form thick beds of evaporite minerals.

Hydrologically closed lakes occur in regions of interior drainage where lake levels may experience considerable fluctuation owing to seasonal flooding. Alluvial fans are commonly present around the borders of such lakes, and the sandy aprons (sandflats) of such fans may extend into the lake. During high water, the edges of these sandflats can be reworked by wave action, resulting in redeposition of wave-rippled sandy sediment along the lake edge. Most sedimentation in closed lakes takes place by chemical/biochemical processes in waters made saline by high rates of evaporation.

10.4.4 Classification Based on Trophic Availability

Lakes are also classified based on their trophic structure into three types: oligotrophic, mesotrophic and eutrophic (Figure 10.8).

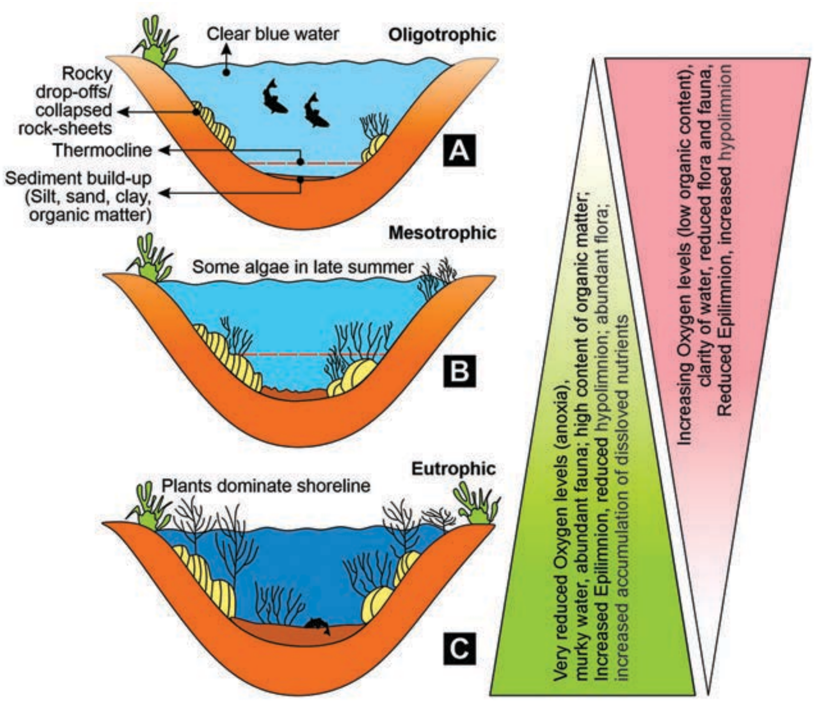

FIGURE 10.8 Lake classification based on trophic (nutrient) availability. A: Oligotrophic lake. B: Mesotrophic lake. C: Eutrophic lake.

Oligotrophic lakes (Figure 10.8A) contain small concentrations of nutrients required for plant growth; hence biodiversity is low. The phytoplanktons, zooplanktons, attached algae, macrophytes (aquatic weeds), bacteria, and fish are all present but in small populations. Thus, with little production of organic matter, there is also very little accumulation of organic sediments on the bottom of oligotrophic lakes.

Eutrophic lakes (Figure 10.8B), on the other hand, are rich in plant nutrients, especially phosphates and nitrates, thus, promoting increased algal growth. However, this with decomposing organisms, depletes the water of oxygen, causing death of other organisms, also. Thus, there is a high production of organic matter that drifts to the lake bottom as organic sediments. These sediments provide the food for high numbers of bacteria. The descending plankton and the bacteria, through their respiration, uses much or all of the available oxygen from the lower depths, causing death of other organisms.

Mesotrophic lakes (Figure 10.8C) have an intermediate level of productivity, greater than the oligotrophic, but less than the eutrophic lakes.

10.4.5 Classification Based on Water Chemistry

Lakes can also be classified based on water chemistry (Table 10.1); of them, saline and ephemeral lakes stand out and are discussed below.

10.4.5.1 Saline Lakes

Saline lakes (also called salt lakes) are permanent or temporary bodies of water with salinities >3 gm/liter and lacking any recent connection to the marine environment. While the use of 3 gm/liter to demarcate salt lakes is somewhat arbitrary, but it is generally accepted, and is also called the "calcite branch point," i.e., the salinity at which calcite is precipitated as a concentrate in natural waters. Hardie et al. (1978) defined a saline lake as a lake in which the water is made up of >5000 ppm of dissolved solute. These lakes occur in the hydrographically lowest areas of arid environments within closed-drainage basins (Figure 10.9) (see Hardie et al, 1978). Such conditions occur in arid and semi-arid regions (approximately one-third of the total world land area) (Figure 10.9).

Saline lakes are categorized into several categories: ephemeral, perennial, acidic, and alkaline; these occur in several depositional sub-environments ranging from alluvial fans, sandflats, dry mudflats, dune fields, and perennial and ephemeral stream floodplains (Hardie et al, 1978). The Caspian Sea is the largest salt lake on the earth with an area of 374 000 km^2 (Figures 10.2 and 10.9), whereas the Dead Sea is the lowest lake on earth, at about 400 m below sea level (Williams 1996, 1998) (Figures 10.2 and 10.9). Saline lakes also have economic value. They have long been mined for precipitate minerals such as sodium and potassium chlorides, borates, nitrates, sulfates of potassium, magnesium, calcium and sodium, calcium and sodium carbonates, and lithium chloride. Salt-lake biota is also of commercial value and includes algal products (such as *Spirulina*), *Artemia* brine shrimps and, in lakes of low salinity, fisheries. Both frozen and dried, *Artemia* and their cysts are important to aquaculture.

10.4.5.2 Ephemeral Lakes

Ephemeral lakes (Figures 10A–C) are a type of saline lake that is shallow and dries up every few years leaving behind a layer of salt(s) precipitated by the evaporating brine (Figure 10.10D) (see also Hardie et al., 1978). These lakes are cyclically recharged via stormwater runoff that gradually recedes as the water evaporates, leading to the formation of two sub-facies – a salt pan facies and a saline mudflat facies (Figure 10.10D) (Hardie et al., 1978). The salt pan facies occur in the lowest part of the lake bed and is characterized by layered salts, while the saline mudflat facies occur around the salt pan and is characterized by muddy clastic sediments that contain salt mineral crystals

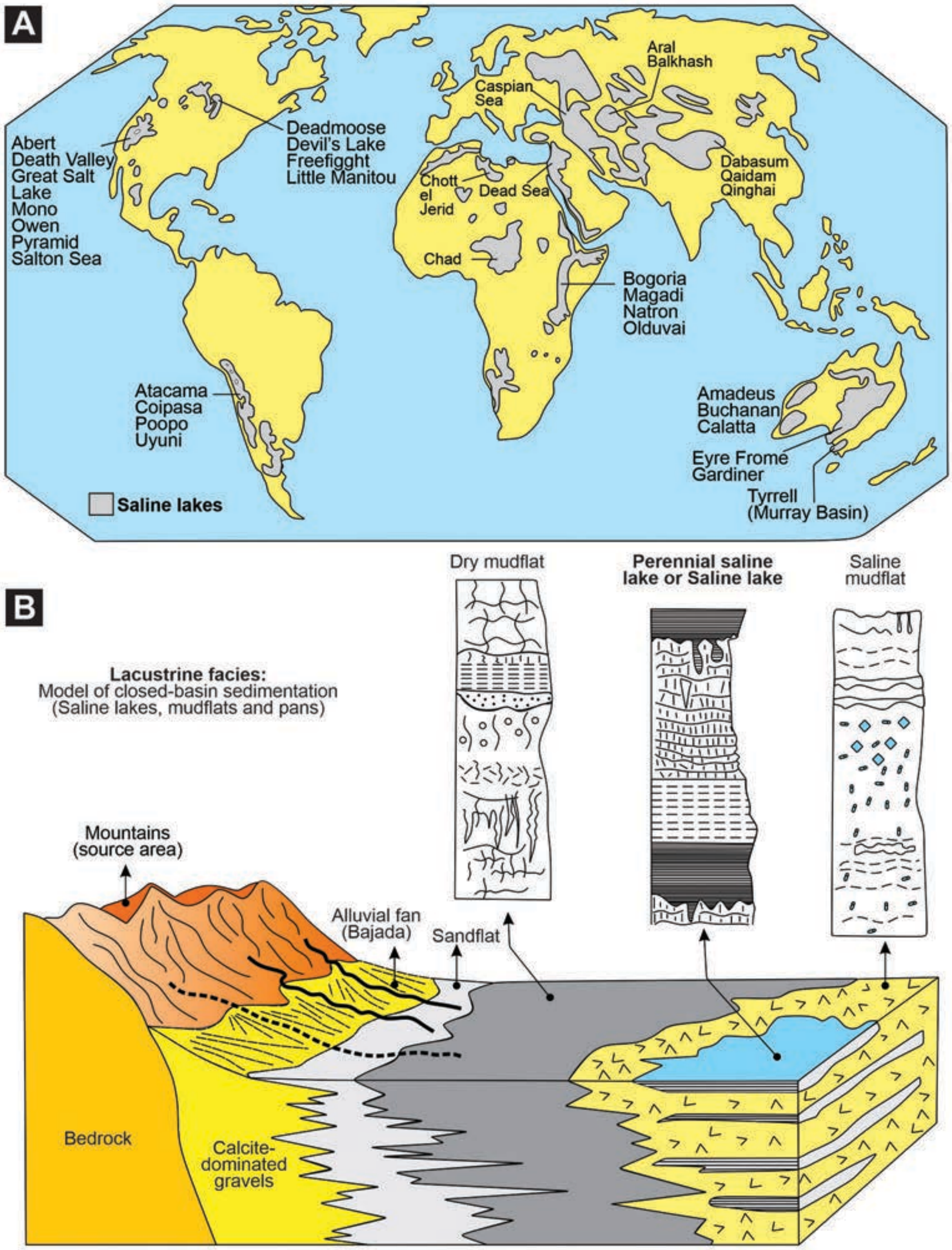

FIGURE 10.9 Saline lakes (closed lake basins). A: Distribution of saline lakes. B: Model of closed-basin sedimentation. (Modified after Eugster and Hardie, 1975, 1978.)

(Figure 10.10D) (see also Hardie et al., 1978). The saline mudflat sub-facies may have root casts (Figure 10.10D). When the turbulence from the flooding of the storm ceases, silt and clay-sized grains settle from suspension and create a thin lamina that extends across the recharged saline lake (Hardie et al., 1978). Wind-induced waves then cause the formation of wave ripples on the mud lamina or rework the mud lamina into a silt clay lenticular lamination (Hardie et al., 1978). In the

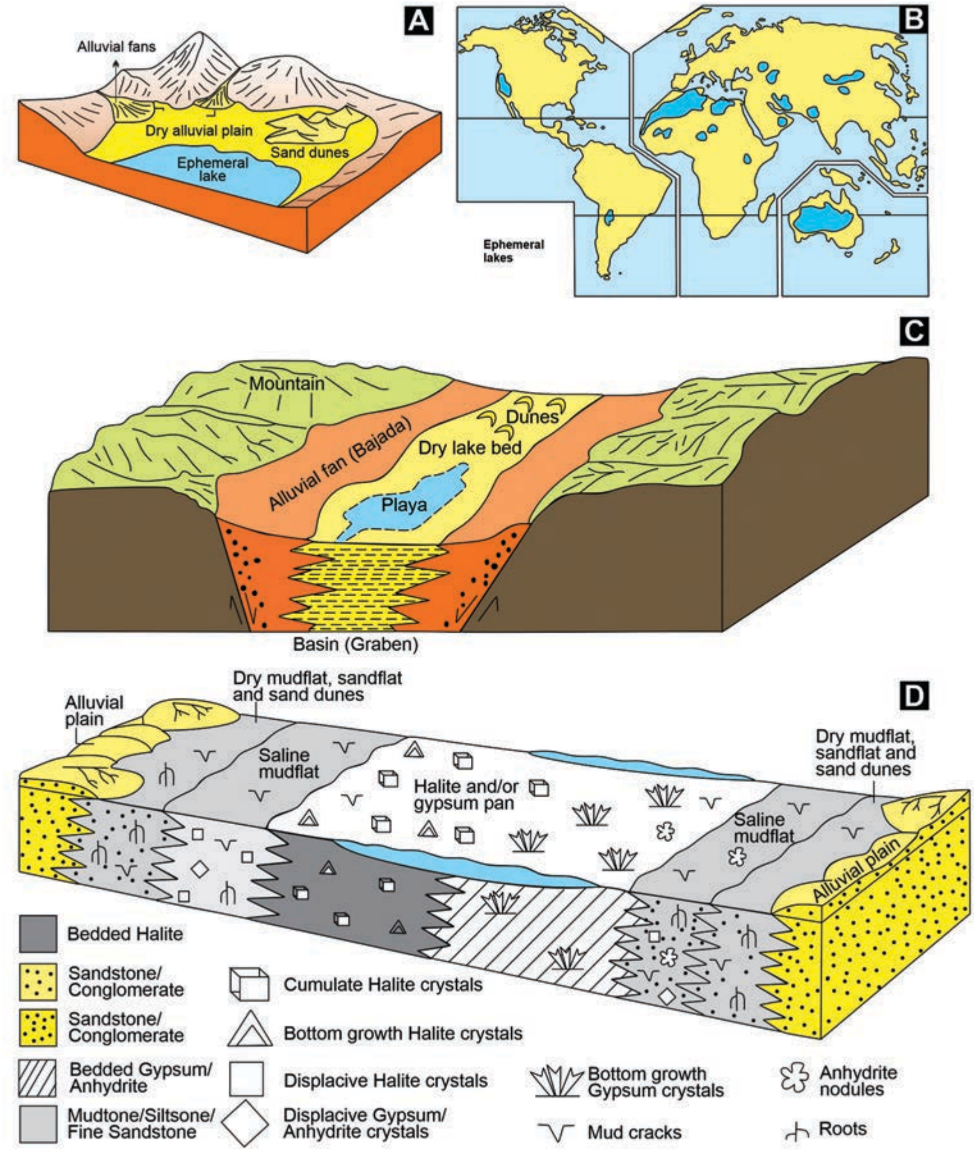

FIGURE 10.10 Arid region ephemeral lakes. A: Depositional setting of an ephemeral lake. B: Global distribution of ephemeral lake. C: Arid region depositional setting of ephemeral lakes. D: Lithofacies of an ephemeral lake. (A: Modified from Benison, Kathleen C. and Robert H. Goldstein. "Sedimentology of Ancient Saline Pans: An example from the Permian Opeche Shale, Williston Basin, North Dakota, U.S.A." *Journal of Sedimentary Research* 70 [2000]: 159–169. C–D: Modified after Hardie et al., 1978.)

salt pan facies, this thin mud lamina becomes the site of bacterial reduction of sulfate and creates iron sulfides and sulfuric acid, which gives the sediment a black color (Hardie et al., 1978). The salt pan facies is also characterized by couplets of layers following a storm flooding of the lake – the thin mud lamina that is black in color and has intergrowth of salt crystals and an overlying, thick layer of crystalized salts (Hardie et al., 1978) (see Figure 10.10D).

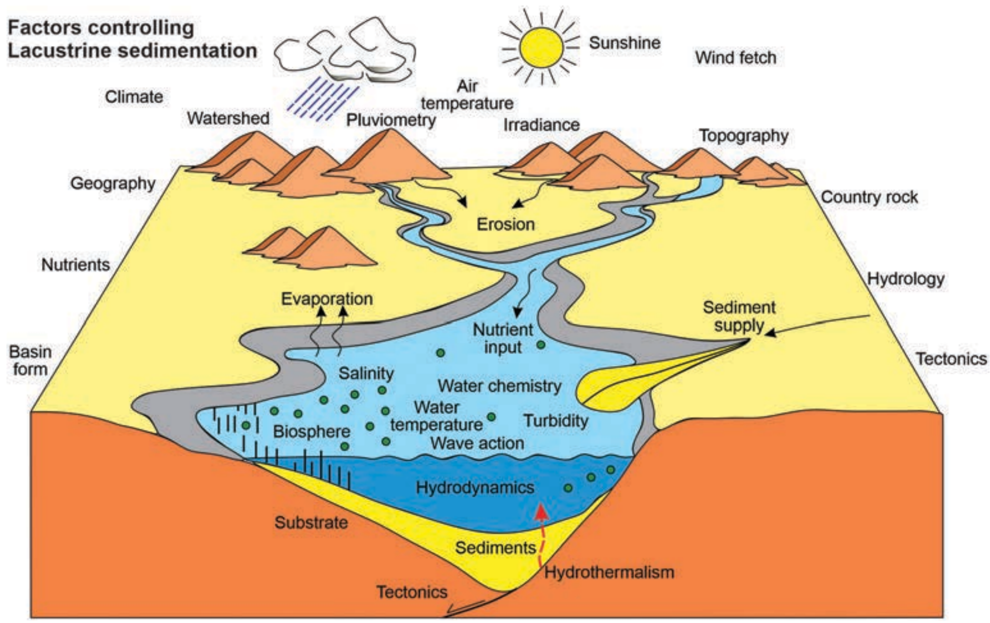

FIGURE 10.11 Controlling factors in lake sedimentation. (Modified from Lettéron A., 2018; with permission from Youri Hamon.)

10.5 CONTROLLING FACTORS IN LAKE SEDIMENTATION

The depositional processes in lakes are influenced by climatic, physical, chemical, and biological factors (Figure 10.11) (see Miall, 2014, 2015). Some of the attributes of the lacustrine depositional environment are similar to those of the marine environment (Figure 10.11), but there are major differences in basin size, water chemistry, and the operating physical (e.g., no tides in lakes) and biologic processes (see GierlowskiKordesch and Kelts, 1994b). These are briefly discussed below.

10.5.1 Climatic Factors

The water level in lakes is maintained by a delicate balance between evaporation and precipitation (Figures 10.7 and 10.11). Climate determines whether a lake is filled to overflowing (open) or acts as an internal drainage basin (closed) (see Figure 10.7). Climatic conditions strongly affect chemical sedimentation. For example, in arid regions, the lake is dominated by the precipitation of gypsum, halite, and various other salts (Figure 10.10C), but in humid climates, the chemical sedimentation is dominated by carbonate deposition.

Atmospheric heating, which is a function of climate, is responsible for density differences in lake water. These differences can cause stratification of water on the one hand (heating of surface water) or, under some conditions, generation of density currents (by cooling of surface water) that produce mixing and lake overturn (see Figure 10.3).

The vegetation cover in the drainage basin of the lake affects sediment input to lakes, being maximum in arid regions when the vegetation cover is lowest. In cold climates, seasonal drops in temperature lead to freezing of lakes, causing decrease in sediment input and the cessation of wave activity, allowing deposition of fine-grained suspended sediment during these quiet-water conditions. One type of lake sedimentation process that appears to be particularly characteristic of cold-climate lakes is the formation of varves, which are very thin, alternating light- and dark-colored

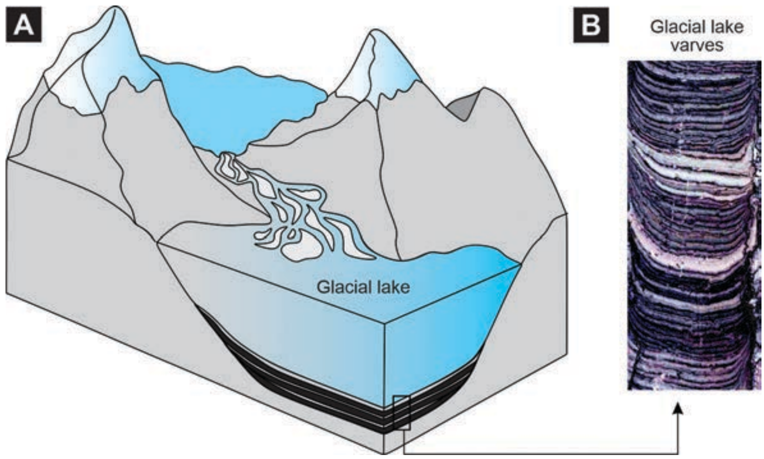

FIGURE 10.12 Glacial lake and varve. A: Glacial lake. The formation of varves is noted in the lake sedimentation process of cold-climate lakes. B. Glacial varves are made up of very thin, alternating light- and dark-colored sediment layers deposited in a glacial lake.

sediment layers (Figure 10.12). Thicker, light-colored, coarse-grained laminae accumulate suspension settling of fine sediment during summer conditions (Figure 10.12B). Thinner, finer grained, organic-rich, dark laminae form by slow suspension settling during winter months when lakes are frozen (Figure 10.12B). Another distinguishing characteristic of lake sediments is that individual lake beds tend to be thin and laterally continuous compared to associated fluvial deposits (although total lake sediments can be very thick). However, many sedimentary structures of lacustrine deposits are similar to those of shallow marine sediments. Varves are one of the more diagnostic characteristics of lake sediments, although light and dark laminae resembling varves have also been reported in non-lacustrine sediments (e.g., some laminated marine deposits).

Typically, a lake has periods of overturn (such as spring and fall), when the entire lake circulates, and periods when the water is density stratified (such as summer and winter) (see Figure 10.4). This regular alternation of stagnation and overturn produces fine lamination that can be very rhythmic and laterally extensive. As these laminations reflect both seasonal cycles and larger-scale climatic cycles, they are of great importance as paleoclimatic proxies. In proglacial lakes, the cycles of stagnant frozen sediments in winter and oxygenated sediments in summer overturn produce striking glacial varves (see Figure 10.12B).

The lack of oxygen from reduced circulation and from the excess nutrients leads to stagnant, reducing conditions (Figures 10.8B–C). Under these circumstances, a concentration of organic matter, producing black shales that are high in low-grade hydrocarbons, or kerogens are formed. These are also called oil shales.

10.5.2 Physical Processes

Physical processes such as wind, river inflow, and atmospheric heating affect sediment transport and deposition. Wind processes are important, as winds create waves and currents. River inflow may generate plumes of fine sediment that extend in surface waters far out into a lake, or the inflow may generate density underflows, or turbidity currents, that carry sediments along the lake bottom, toward the basin center. The deposition of siliciclastic sediments in the calmer, deeper portion of lakes takes place by settling of fine particles that were suspended in the water column owing to wave and current activity, or the deposition may occur from turbidity currents, generated when sediment-laden streams discharge into lakes.

10.5.3 Chemical Processes

Deposition of chemically formed sediment is common in closed lakes (see Figures 10.7C–D), where the chemistry of lake water varies from lake to lake, and is dominated by calcium, magnesium, sodium, potassium, carbonate, sulfate, and chloride ions. Thus, the most common chemical sediments in lakes of humid regions are carbonates, although in some lakes, phosphates, sulfides, cherts, and iron and manganese oxides are also present. In arid regions, where the rates of evaporation are high, chemical lake sediments are dominated by carbonates, sulfates, and chlorides. The evaporite deposits of lakes include many common marine evaporite minerals such as gypsum, anhydrite, halite, and sylvite, but they also include several minerals such as trona, borax, epsomite, and bloedite that are not common in marine settings. Although chemical sedimentation processes are most important in closed lakes, they may also dominate in some open lakes when the clastic sediment supply is low. Thus, where the clastic input into a lake is limited, chemical sedimentation can predominate.

Chemical precipitates are usually either saline or carbonate. Wherever evaporation exceeds inflow, the salinity can exceed 35,000 ppm dissolved solids, producing a saline lake (normal marine salinity is about 33,000 to 37,000 ppm). The most abundant chemicals are SiO_2 and ions such as Ca_2, Mg_2, Na, K, HCO_3, and Cl. As evaporation proceeds and increases the ionic concentration, dense brines sink to the bottom and precipitate evaporite minerals. Carbonates are the first to be produced, followed by gypsum. If the process continues, halite is precipitated, followed by a series of potassium and magnesium salts, many of which are unique to dry lakes.

Lacustrine carbonates, on the other hand, are produced where there is neither excess evaporation nor much clastic input. Unlike marine carbonates, freshwater limestones are produced mostly by inorganic precipitation. Fresh water typically contains abundant carbonate from dissolved atmospheric CO_2 and from dissolved bedrock carbonate. The carbonate ion concentration is strongly controlled by changes in pH, which fluctuates continuously in freshwater lakes. The precipitation of calcite is facilitated by two factors: plants use CO_2 and raise the pH, and warmer temperature lowers the solubility of calcite. Lacustrine carbonates are usually low-magnesium calcite, precipitated in finely laminated beds with mudstone and marl.

10.5.4 Biological Processes

Organisms play an important role in lake sedimentation by extracting chemical elements from lake water to build shells (by the extraction of CO_2 during photosynthesis, thereby aiding the precipitation of $CaCO_3$) and then the subsequent deposition of these shells. Thus, organisms living in lakes contribute their skeletal and non-skeletal remains to lake sediments. Diatoms carry out photosynthesis and are the only important type of lake organism that produces siliceous tests. Their remains form important diatomite deposits in many Pleistocene lakes. Others, such as bivalves, gastropods, calcareous algae, and ostracods are also common in many lakes and are important contributors of calcium carbonate sediments. The blue-green algae (cyanobacteria) carry on photosynthesis and trap fine sediments to form stromatolites. Under reducing conditions and high sedimentation rates, the remains of higher plants are partially preserved and eventually form peat and coal. Considering the small size of many lakes and their generally lower alkalinity and buffering capacity, compared to those of the open ocean, the assimilation of CO_2 by plants during photosynthesis is a much more important factor in controlling the pH of lakes than that of the ocean. Thus, increase in pH caused by photosynthetic removal of CO_2 exerts a dominant control in facilitating carbonate sedimentation in lakes.

Freshwater calcareous algae form a kind of stromatolite called an oncolite, which is a subspherical or tabular body formed when encrusting algae trap sediment. Some reach tens of centimeters in diameter and are known as algal biscuits. These freshwater fossils, though rare, are important as environmental indicators. In general, the oncolites are composed of oncoids, layered structures made by cyanobacteria. Hence, they are very similar to stromatolites, but approximately spherical structures rather than in columns.

10.6 CHARACTERISTICS OF LACUSTRINE DEPOSITS (BASIN TYPES AND SEDIMENTARY SEQUENCES)

Bohacs et al. (1999, 2000; see also Bradley, 1925; Carroll and Bohacs, 1995, 1999), on the basis of fill characteristics, divided lakes into three distinct types: overfill, balance-fill, and underfilled (see Figures 10.13–10.18; Table 10.3). These divisions are based on distinctive lithology, sedimentary structures, and biogeochemistry. Bohacs et al. (2000, 2003) proposed this tripartite classification of lithofacies associations and lake types which is relatively independent of age, water depth, and

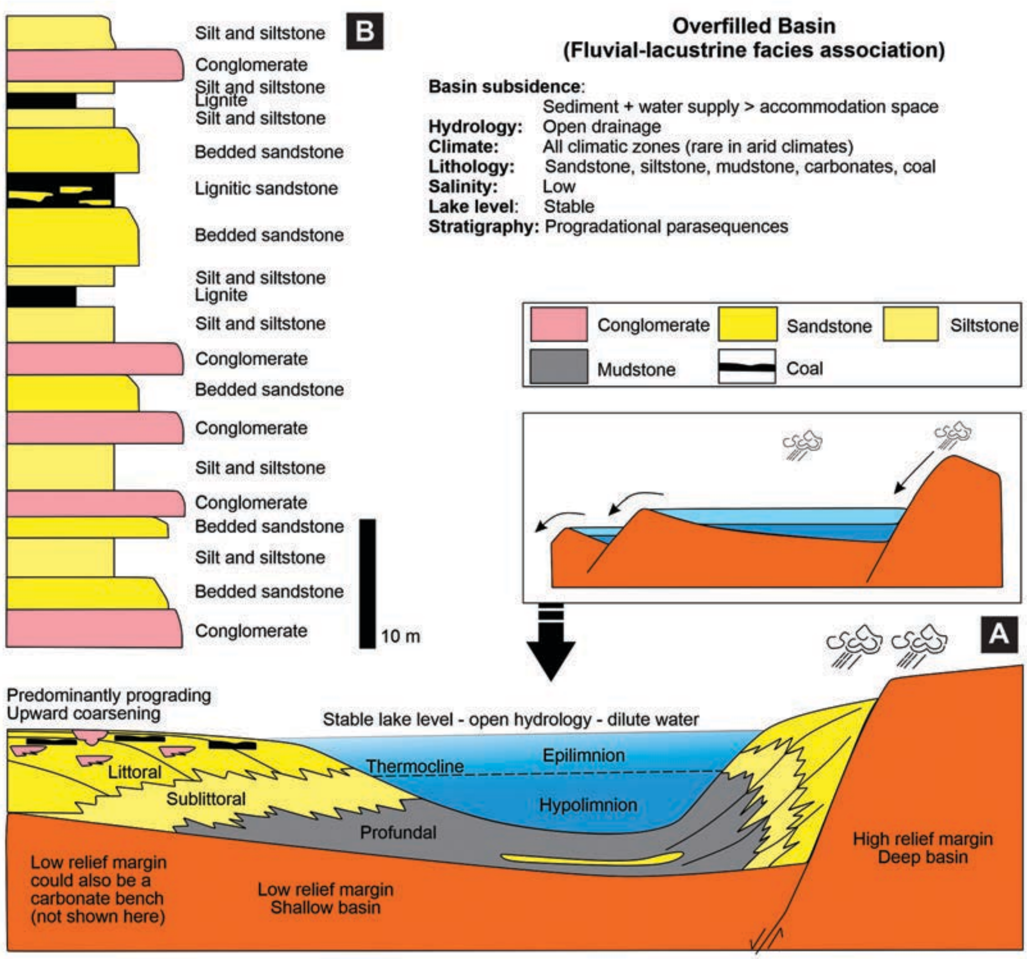

FIGURE 10.13 Overfilled lake basin. (Modified after Bohacs et al., 2000.) A. Diagram displays the major characteristics of an overfilled lake basin, such as persistently open hydrology, freshwater lake chemistry, high groundwater table, progradational shoreline architecture, close relation to fluvial systems, and interbedded fluvial deposits and coals. This lake-basin type occurs when the rate of supply of sediment+water consistently exceeds potential accommodation (usually when P/E is relatively high compared to rates of tectonic subsidence). P/E = precipitation/evaporation ratio. B. Lacustrine sequence in an overfilled lake basin.

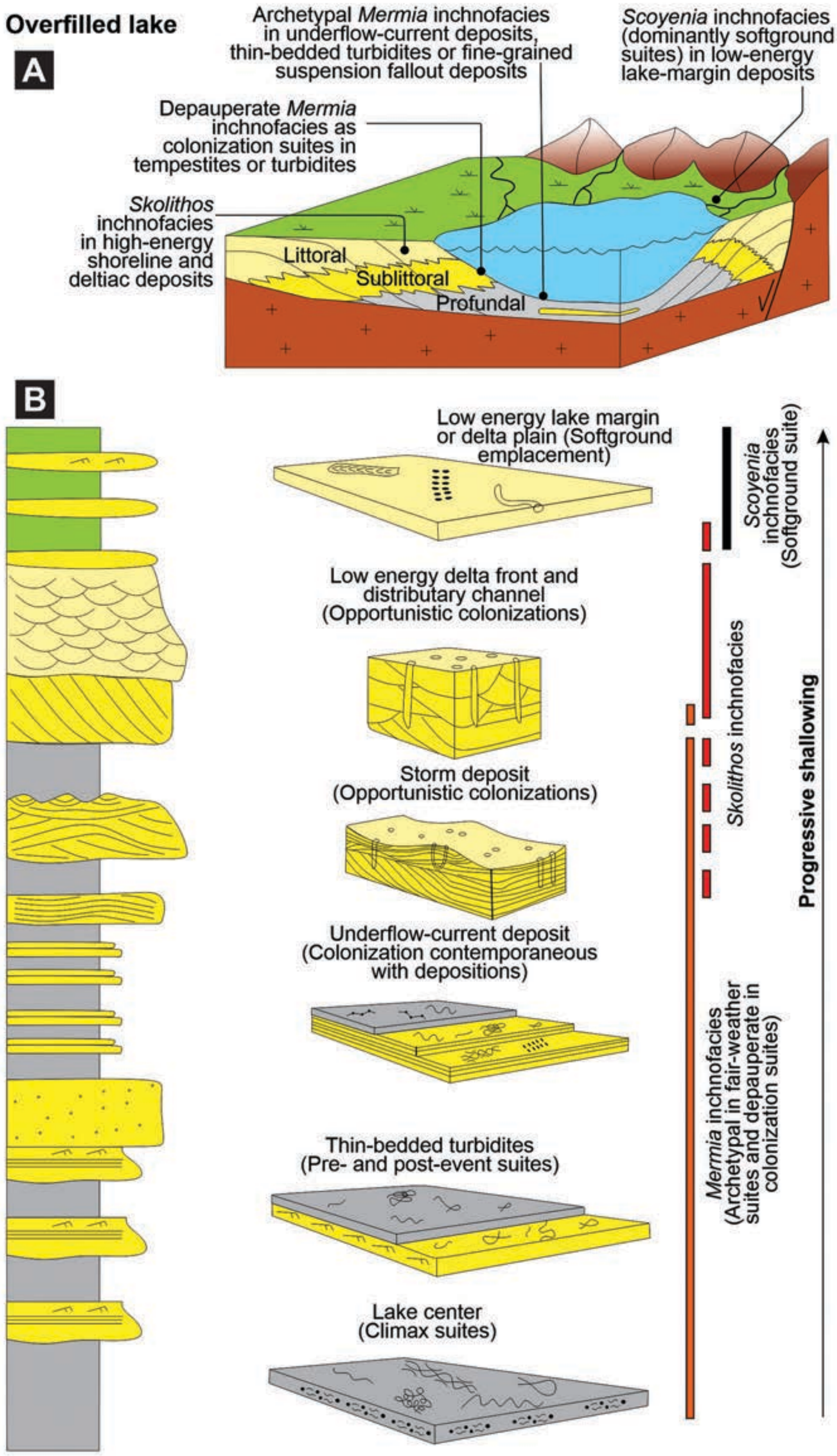

FIGURE 10.14 Overfilled lake basin. (Modified after Buatois and Mángano, 2004, 2009.) A. Depositional setting with the dominant trace-fossil assemblages. B. Lacustrine sequence in an overfilled lake basin with the dominant trace-fossil assemblages.

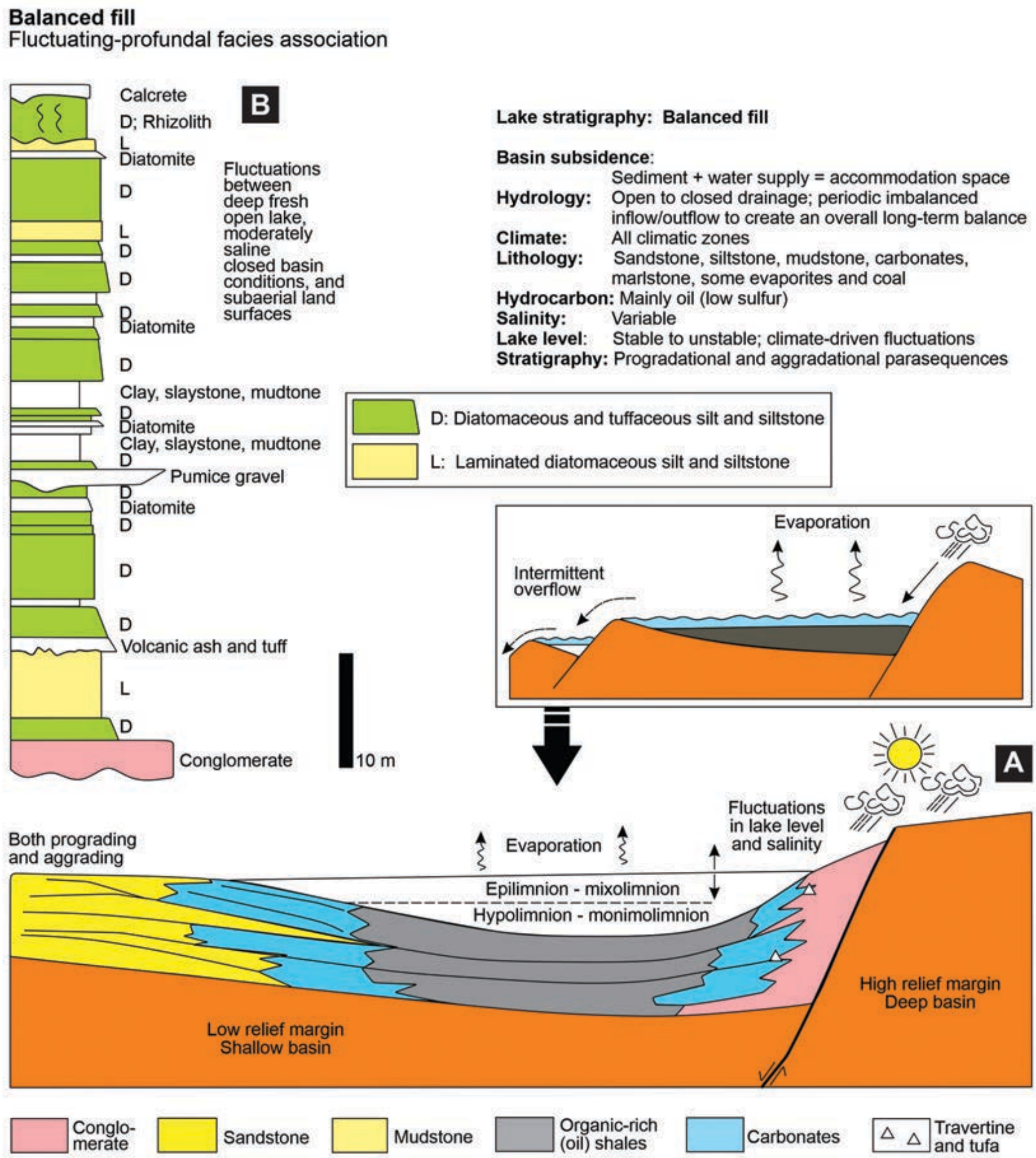

FIGURE 10.15 Balanced-fill lake basin. (Modified after Bohacs et al., 2000.) A. Diagram displays the major characteristics of a balanced-fill lake basin such as intermittently open hydrology, fluctuating groundwater table and lake water chemistry, thermal and chemical stratification, mixed progradational and aggradational shoreline architecture, and varied interbedding of clastic and carbonate strata. This lake-basin type occurs when the rates of sediment+water supply and accommodation are almost in balance over the time span of sequence development. Water inflows are sufficient to periodically fill available accommodation, but are not always balanced by outflow; hence, climatically driven lake level fluctuations are common. B. Lacustrine sequence in a balanced-fill lake basin.

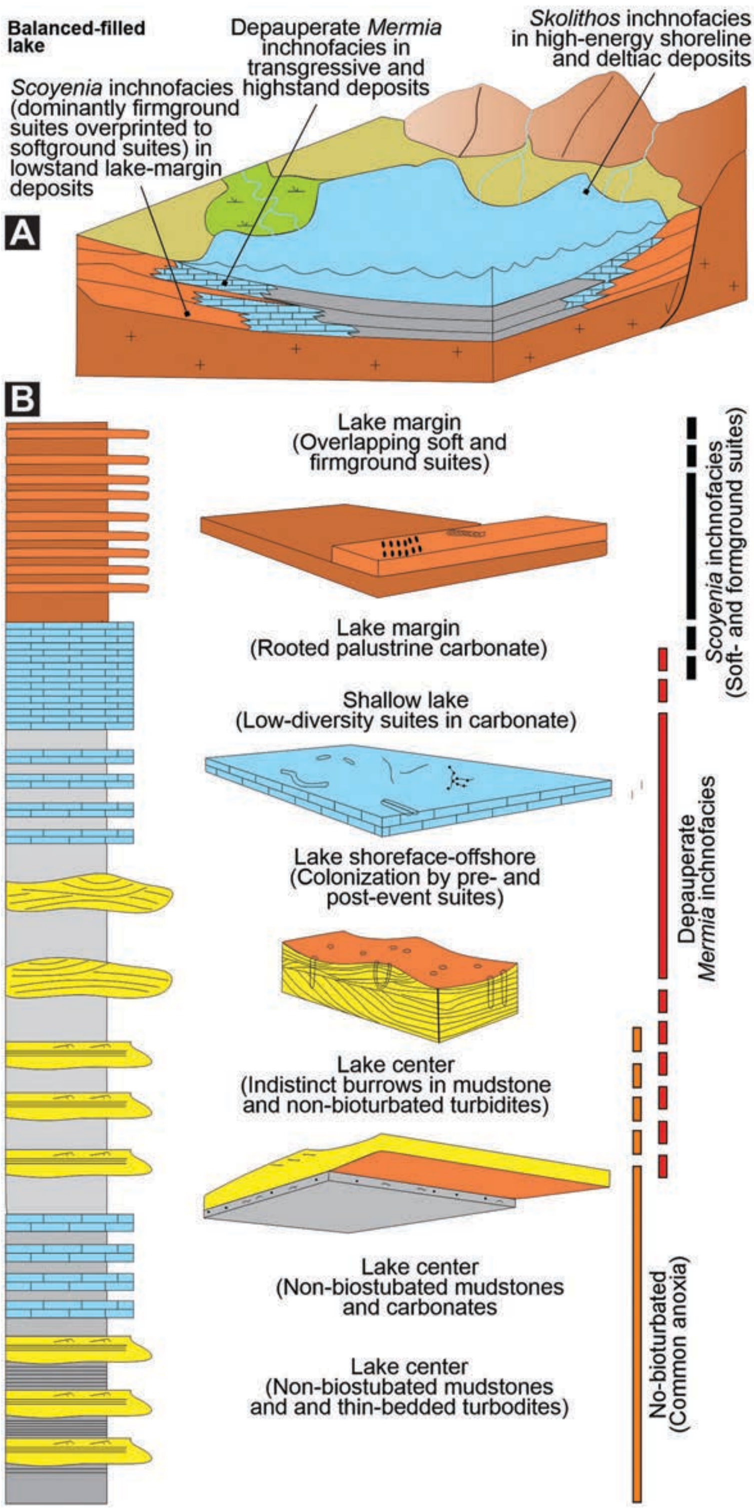

FIGURE 10.16 Balanced-fill lake basin. (Modified after Buatois and Mángano, 2004, 2009.) A. Depositional setting with the dominant trace-fossil assemblages. B. Lacustrine sequence in a balanced-fill lake with the dominant trace-fossil assemblages.

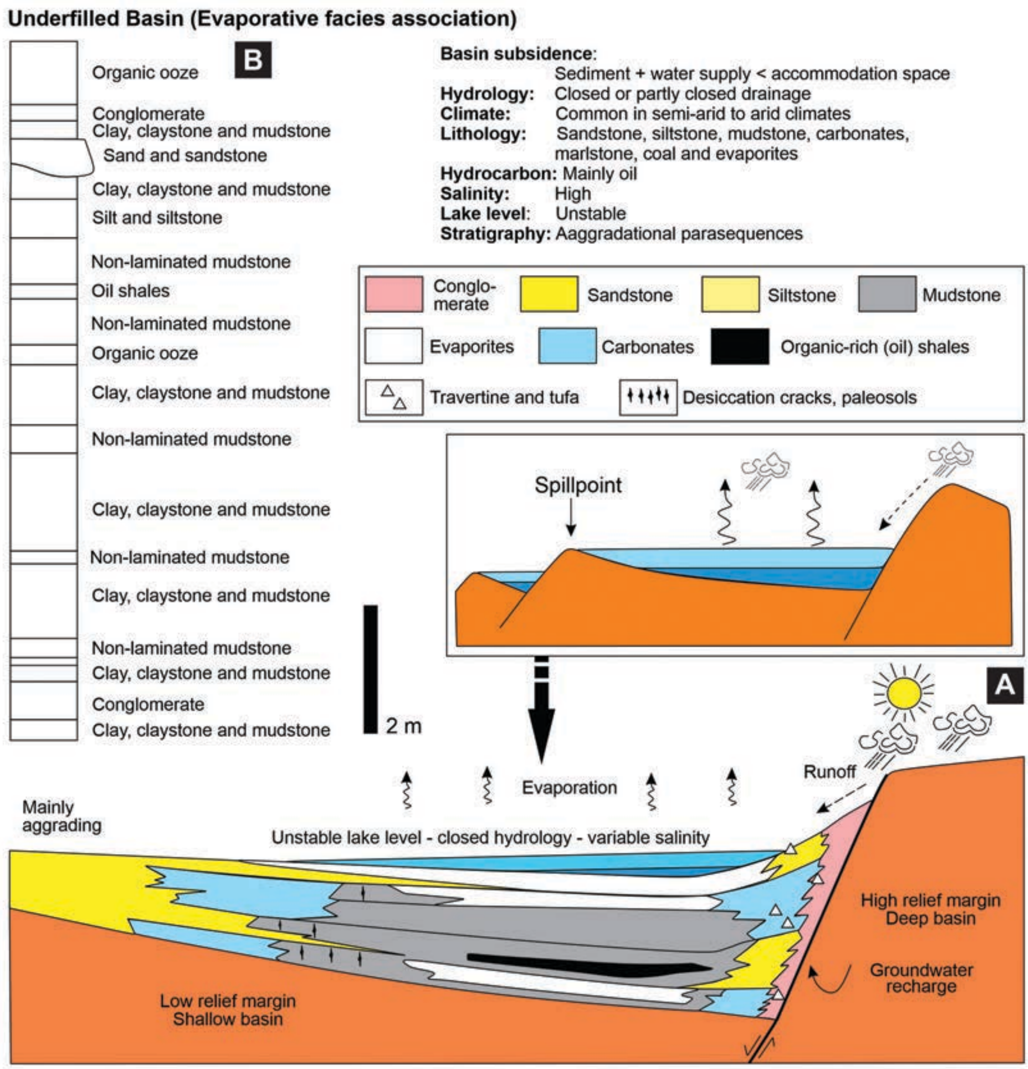

FIGURE 10.17 Underfilled lake basin. (Modified after Bohacs et al., 2000.) A. Diagram displays the major characteristics of an underfilled lake basin such as closed hydrology, chemical stratification, low groundwater table, high solute content of lake waters, extensive desiccation, highly contrasting lithologies, common association with evaporite deposits, and dominantly aggradational shoreline architecture. This lake basin occurs when the rates of accommodation consistently outstrip available water and sediment supply, resulting in a persistently closed basin with ephemeral lakes interspersed with playas or brine pools or both. B. Lacustrine sequence in an underfilled lake basin.

thickness. These lake types are controlled by two primary factors: accommodation space (related to geologic and tectonic setting) and the supply of water and sediment (related to climate) (see also Carroll and Bohacs, 1995, 1999). The association of the end member lithofacies with the lake classification scheme allows for the prediction of lake type based upon limited outcrop data and sedimentary structures and has proven to be a very effective framework for interpreting lacustrine basin fills (Johnson and Graham, 2004; Bohacs, 2004; Keighley, 2008). Details of these three lake basins are enumerated below.

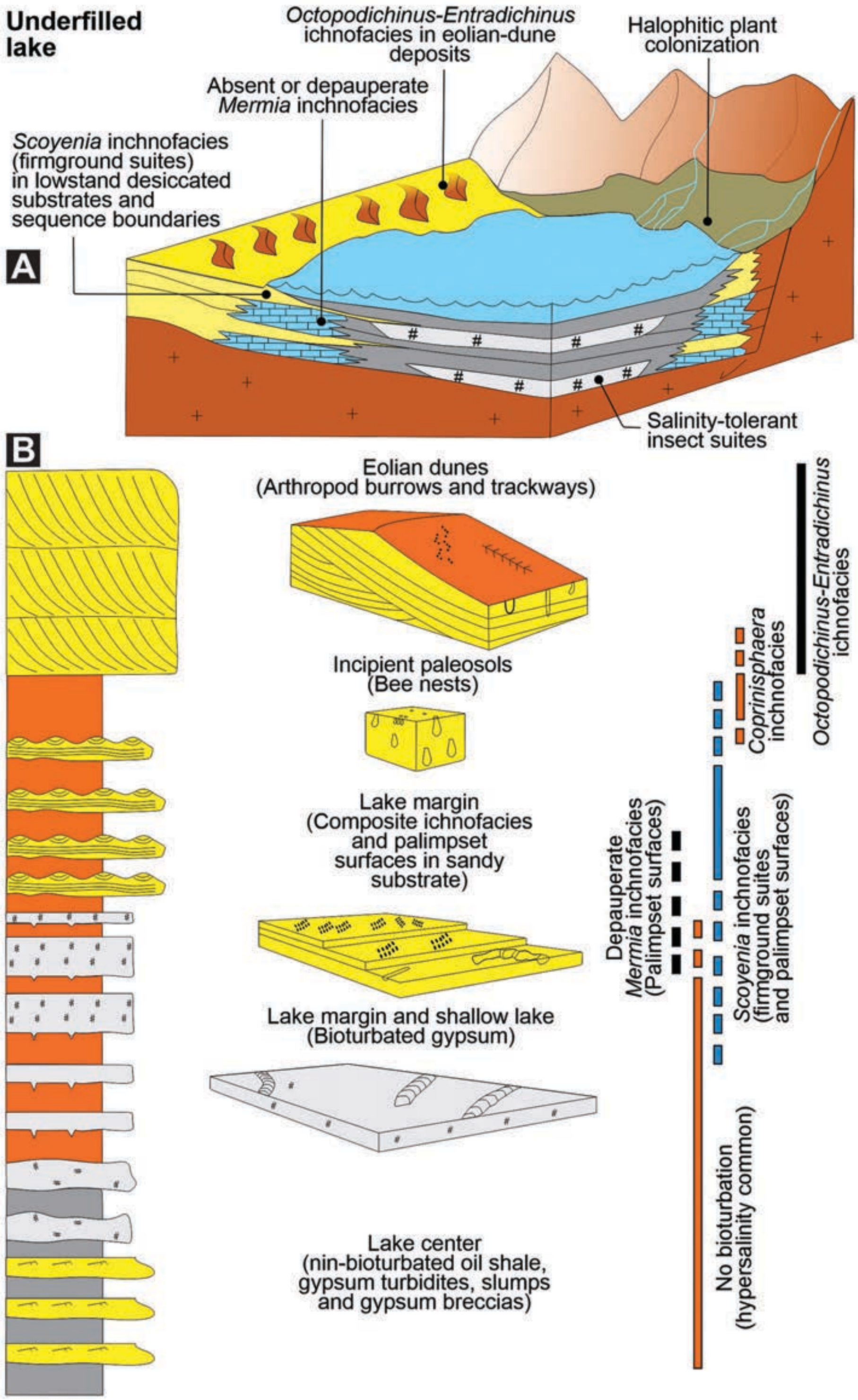

FIGURE 10.18 Underfilled fill lake basin. (Modified after Buatois and Mángano, 2004, 2009.) A. Depositional setting with the dominant trace-fossil assemblages. B. Lacustrine sequence in an underfilled lake with the dominant trace-fossil assemblages.

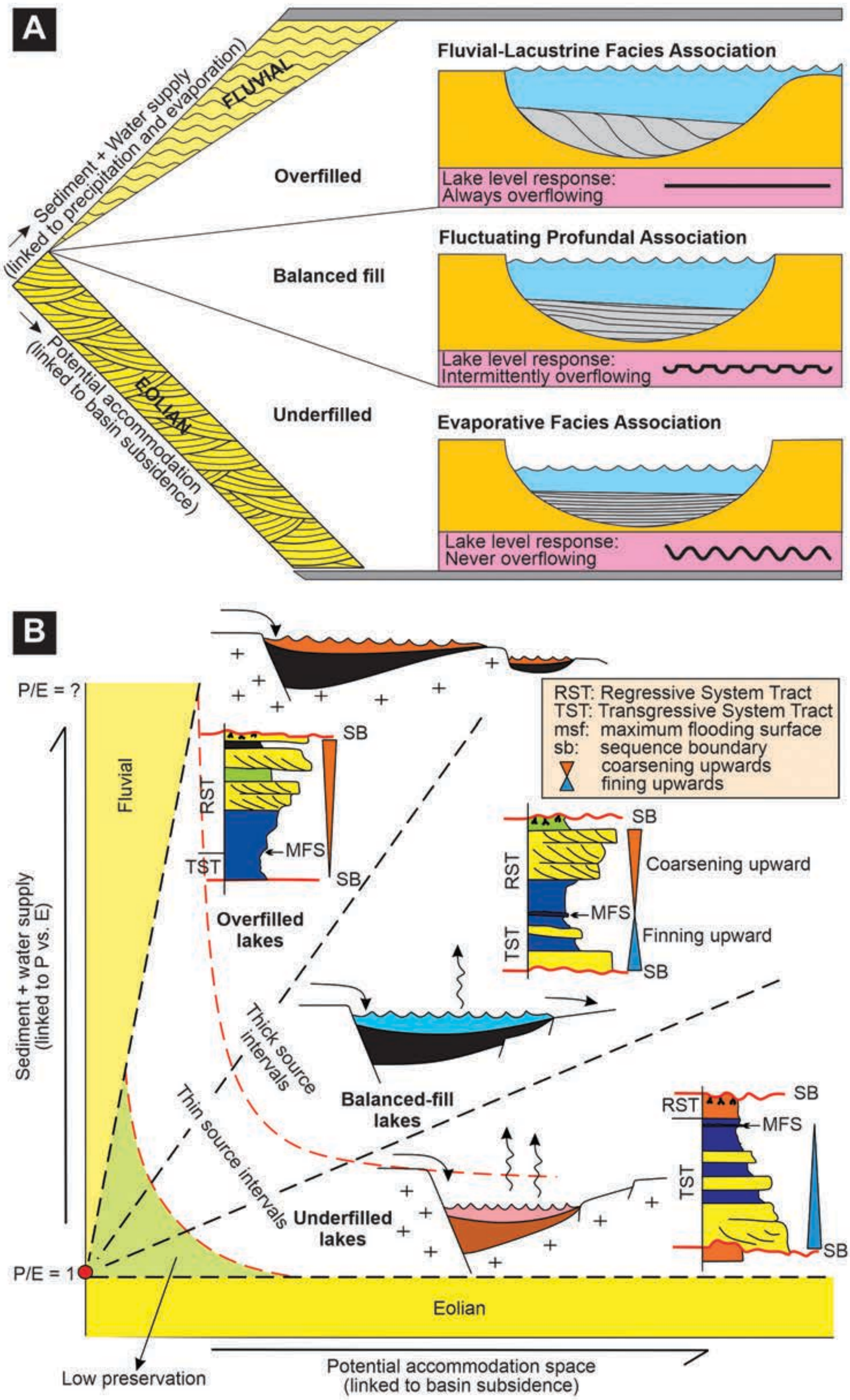
A
Sediment + Water supply
(linked to precipitation and evaporation)
FLUVIAL
Potential accommodation
(linked to basin subsidence)
EOLIAN
Overfilled
Balanced fill
Underfilled
Fluvial-Lacustrine Facies Association
Lake level response:
Always overflowing
Fluctuating Profundal Association
Lake level response:
Intermittently overflowing
Evaporative Facies Association
Lake level response:
Never overflowing
B
P/E = ?
P/E = 1
Fluvial
Sediment + water supply
(linked to P vs. E)
RST: Regressive System Tract
TST: Transgressive System Tract
msf: maximum flooding surface
sb: sequence boundary
coarsening upwards
fining upwards
SB
RST
TST
MFS
Overfilled
lakes
Coarsening upward
Finning upward
Thick source
intervals
Thin source intervals
Balanced-fill
lakes
Underfilled
lakes
Eolian
Low preservation
Potential accommodation space
(linked to basin subsidence)

FIGURE 10.19 Lake basin classification system. A. The three lithofacies associations (fluvial-lacustrine, fluctuating profundal, and evaporative) correspond to three basin types (overfilled, balanced-fill, and underfilled, respectively). These lake types are differentiated by two primary factors: accommodation space and the supply of water and sediment to the basin. (Modified after Carroll and Bohacs, 1999; Schon et al., 2012.) B. Lake basin-type model for classifying lacustrine deposits. (Modified after Carroll and Bohacs, 1995, 1999.) The horizontal axis is the potential accommodation space (tectonics), and the vertical axis is the sediment and water supply (climate) which links precipitation and evaporation. Accommodation is the space available for sediment accumulation below the basin's outlet or spillpoint, largely influenced by basin tectonics, along with sill uplift and erosion, and topography. Sediment+water supply is a function of climatic humidity, along with seasonality, local relief, and bedrock geology.

TABLE 10.3
Characteristics of the three types of lacustrine lake basins

Lacustrine Facies Association	Fluvial lacustrine (Overfilled Basin)	Fluctuating profundal (Balanced-Fill Basin)	Evaporative (Underfilled Basin)
Stratigraphy	Maximum progradation, parasequences related to maximum progradation (relatively stable), maximum fluvial input	Mixed progradation and desiccation, common distinct shoaling cycles, fluvial input variable	Maximum desiccation, high-frequency wet-dry cycles, high fluvial input
Stratal stacking patterns	Dominantly progradation, indistinctly expressed parasequences	Mixed progradation and aggradation, distinctly expressed parasequences	Dominantly aggradation, distinctly to indistinctly expressed parasequences
Sedimentary structures	Physical transport: ripples, dunes, flat bed, root casts, burrows, (infaunal and epifaunal)	Physical and biogenic: flat bed, current, wave, and wind ripples; stromatolites, pisolites, oncolites, mud cracks, burrows (epifaunal)	Physical, biogenic, and chemical: climbing current ripples, flat bed, stromatolites, displacive fabrics
Lithologies	Mudstone, marl, sandstone, coquina, coal, coaly shale	Marl, mudstone, siltstone, sandstone, carbonate grainstone, wackestone, micrite, kerogenite, algal biomarkers	Mudstone, kerogenite, evaporite, siltstone, sandstone, grainstone, boundstone, flat-pebble conglomerate
Organic matter	Freshwater biota, land plant, charophytic and aquatic algal organic matter, low to moderate total organic carbon, terrigenous and algal biomarkers	Salinity tolerant biota, aquatic algal OM, minimal land plant, moderate to high total organic carbon	Low-diversity, halophytic biota, algal-bacterial organic matter, low to high total organic carbon, hypersaline biomarkers
Source potential	Low to moderate total organic carbon, mixed type I to III kerogen, marked organic facies contrasts, distinct lateral changes in organic facies	Moderate to high total organic carbon, predominantly type I kerogen with types I–III mixtures near flooding surfaces, relatively homogeneous and laterally consistent organic facies	Low overall total organic carbon (with some high total organic carbon intervals), type I kerogen, minimum organic facies contrasts, laterally consistent organic facies
Hydrocarbon characteristics	Generate both oil and gas, very waxy, low sulfur oils, terrigenous biomarker assemblage dominant	Mostly oil generative, paraffinic but relatively non-waxy oils, low sulfur, algal biomarker assemblage dominant	Mostly oil generative, paraffinic oils, moderate to high sulfur, distinctive "hypersaline" biomarker assemblage

Source: Modified after Bohacs et al. (2000).

10.6.1 Overfilled Lake Basins

Overfilled lake basins have persistently open hydrology (Figure 10.13A), freshwater lake chemistry, progradational shoreline architecture (i.e., regression; the building forward or outward toward the sea of a shoreline or coastline), and commonly interbedded fluvial deposits (Figure 10.13B) (see also Bohacs et al., 2000). They occur when the rate of supply of sediment and water consistently exceeds accommodation space (i.e., the space available in which sediments accumulate). Buatois and Mángano (2009) noted that the overfilled lake basins have well-developed softground trace-fossil assemblages of *Mermia*, *Skolithos* and *Scoyenia* ichnofacies that are useful paleoenvironmental proxies and for delineating parasequences (see Figure 10.14). Fluvial discharge into overfilled lakes create density currents that oxygenate lake bottoms, facilitating the establishment of well-diversified biotic communities. Land-plant-derived organic matter is the prime source of nutrients, further enabling the development of deposit feeding benthic fauna in such permanently subaqueous, low-energy zones. In shallow overfilled lakes, distal facies commonly consist of delta-fed underflow-current and background fallout deposits hosting the *Mermia* ichnofacies (Buatois and Mángano, 2009) (see Figure 10.14). Intermediate facies contain wave-dominated delta-front and nearshore deposits, including storm-influenced hummocky cross-stratified and fair-weather wave- ripple cross-laminated sandstones (Buatois and Mángano, 2009) (see Figure 10.14). Under conditions of moderate to high wave energy, the *Skolithos* ichnofacies is noted (Buatois and Mángano, 2009) (Figure 10.14). Proximal (near shore) facies include distributary-channel, trough, and tabular cross-bedded sandstones with escape structures and vertical burrows, representing the *Skolithos* ichnofacies (Melchor et al., 2003) (see Figure 10.14). The upward shallowing successions due to delta and shoreline progradation have well-developed softground trace fossils of the *Scoyenia* ichnofacies (Buatois and Mángano, 2009).

10.6.2 Balanced-Fill Basins

Balanced-fill basins have intermittently open hydrology (Figure 15A), fluctuating lake water chemistry, thermal and chemical stratification, mixed progradational and aggradational architecture (i.e., the deposition process in which depositional area fills with the vertical stacking of sediment), and varied interbedding of clastic and carbonate strata (Figure 10.15B). This lake-basin type occurs when the rates of sediment + water supply are roughly in balance with the accommodation space. In balanced-fill lakes, carbonates are abundant (Figure 10.15A), hence their depositional sequences are quite similar to shallow marine carbonate or mixed carbonate-clastic settings (see also Bohacs et al., 2000). Buatois and Mángano (2009) noted abundant but low-diversity firmground trace-fossil suites of the *Scoyenia* ichnofacies; their low diversity is the reflection of their closed hydrology especially during lowstands, resulting in increased salinity, and causing stress on the lake biota (Figure 10.16). In general, the ichnofaunas from turbidite systems in balanced-fill lakes are less abundant and diverse than those from the overfilled lacustrine turbidites. Freshwater conditions are noted during transgression, but dysaerobic conditions may prevail, leading to greater stress on the benthic fauna.

10.6.3 Underfilled Lake Basins

Underfilled lake basins have persistently closed hydrology (Figure 10.17A), characteristic chemical stratification, high solute content of lake waters, extensive desiccation (drying) features, dominantly aggradational shoreline architecture and highly contrasting lithologies that are commonly associated with evaporite deposits (Figures 10.17A–B). This basin type occurs when rates of accommodation consistently outstrip available water and sediment supply, resulting in closed basins with ephemeral lakes (i.e., basins that remain flooded only for short periods of time during a year). Buatois and Mángano (2009) noted that the *Scoyenia* ichnofacies is associated with lowstand desiccated substrates in underfilled lakes (Figure 10.18) where the density of arthropod trackways is high, tracking omission surfaces. Some of these represent sequence boundaries expressed as co-planar surfaces of lowstand and subsequent flooding (Figure 10.18B). Rapid changes in depositional conditions reflecting desiccation during vertical aggradation lead to the formation of composite ichnofabrics that reflect successive bioturbation events (Figure 10.18B).

A summary of the three types of lake basins, overfill, balance-fill, and underfilled, with respect to rate of sediment supply, water availability (including precipitation and evaporation) and accommodation space, is provided in Figure 10.19. In general, the evolution of lacustrine depositional systems is dependent on basin tectonics, and the relationship between regional precipitation and evaporation (see Olsen, 1990). Thus, for the continued expansion of a lacustrine system, accommodation space needs to be greater than the sediment supply going into and out of the basin. Similarly, precipitation needs to be greater than evaporation or the lake will dry up (Figure 10.19).

10.7 LACUSTRINE AND MARINE SEDIMENTARY SEQUENCES

Large lakes and shallow seas produce fairly similar sedimentary sequences. In both cases, the dominant sediment is low-energy silt and mud with occasional carbonates. Although, on average, lake basins are much smaller than epicontinental seas, so their deposits tend to be much less laterally continuous than marine shales and limestones. Along the shore of a lake, there is a rapid change in facies, interfingering with a narrow belt of fluvial deposits and even alluvial fans that are less likely to occur along a marine coastline. Typically, lakes form a series of facies belts arranged concentrically from the mudstones or marls in the center to the coarsest sandstones on the margins. Because most lakes fill with sediment over time, they tend to show a sequence that is shallowing and coarsening upward (regressive) (see Platt and Wright, 1991) (Figure 10.20). Deep lakes that are

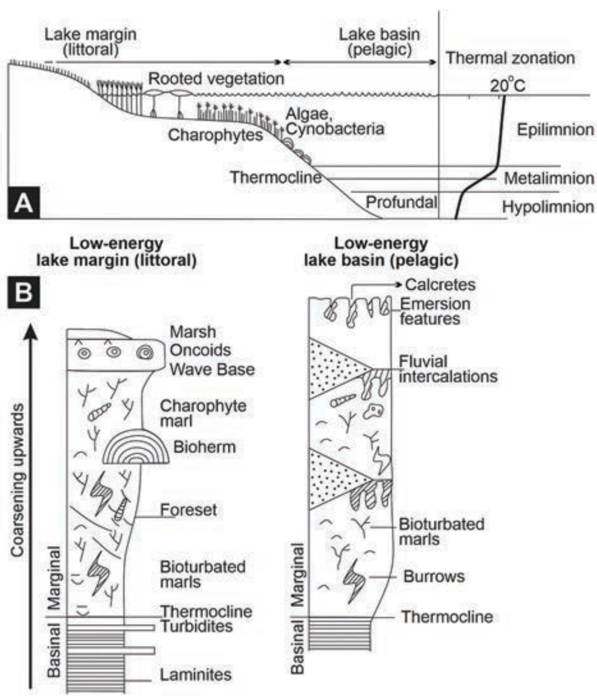

FIGURE 10.20 Lacustrine carbonates. (Modified after Platt and Wright, 1991.) A. Subdivisions of the lacustrine environment in a thermally stratified lake (summer). B. Coarsening-upward cycles in a low-energy lake margin (littoral setting) and a low-energy lake basin (pelagic setting).

affected by pulses of sedimentation during runoff peaks have turbidity currents that produce graded beds. The marginal fluvial sequences can prograde and fill in the lake, forming lacustrine deltaic deposits. Lacustrine deltas and turbidites are much smaller in scale than their marine analogs.

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11 Eolian System

11.1 INTRODUCTION

Eolian (mainly used in USA) or Aeolian (mainly used in UK) refers to the action of wind in shaping the landscape. The eolian desert system refers to the geological formations and processes associated with deserts that are mainly shaped by wind erosion and deposition. Wind plays a dominant role in shaping landforms (including the creation of dunes and sand sheets) and redistributing sediments (particularly sand, clay, and silt). The primary processes involved are erosion, transportation, and the deposition of sediments by wind. Deserts are arid regions characterized by low precipitation and sparse vegetation; they cover ~33% of the total land surface (Figure 11.1). Polar deserts (regions of ice cap climate) are also classified as deserts as they are characterized by low rainfall. However, they are distinguished from true deserts by their low annual temperatures and low evapotranspiration (see Figure 11.1).

Wind erosion in deserts occurs when strong winds pick up loose particles of sand, silt, and dust from the surface. These particles are then transported by the wind and cause abrasion and weathering of rocks and other landforms. Over time, wind erosion creates unique landforms such as sand dunes, desert pavements, and ventifacts (i.e., rocks shaped by wind abrasion). One of the most prominent features of eolian desert systems are sand dunes. These are mounds or ridges of sand that form as the wind transports and deposits sand grains. They have various shapes, such as crescent-shaped (barchan dunes), linear (transverse dunes), and star-shaped (star dunes). The formation and movement of sand dunes are influenced by wind direction, sediment availability, and topography. In addition to dunes, eolian desert systems also include other landforms such as yardangs (elongated ridges or hills formed by wind erosion), and desert varnish (a dark coating on rocks caused by windblown particles and chemical reactions). All the above-mentioned landforms are described later in the chapter.

Thus, understanding eolian desert systems is important for studying the desert ecosystem, sedimentary processes, and climate change. The study of eolian deposits and landforms also provides insights into past environmental conditions, including changes in wind patterns, aridity, and vegetation cover. In general, studying them also helps in assessing the impact of wind erosion on soil degradation and desertification (see Stoops, 2021).

11.2 TRANSPORT AND DEPOSITIONAL PROCESSES

Transport and depositional processes in deserts play a crucial role in shaping the landscape and creating distinctive landforms. These processes are primarily driven by wind and to a lesser extent by water, as deserts are characterized by arid conditions (i.e., areas that receive <250 mm/year of precipitation) and limited vegetation cover. It must be noted that water is a much more powerful transport agent than air due to its fluid properties; hence a lack of water is prerequisite for air to be

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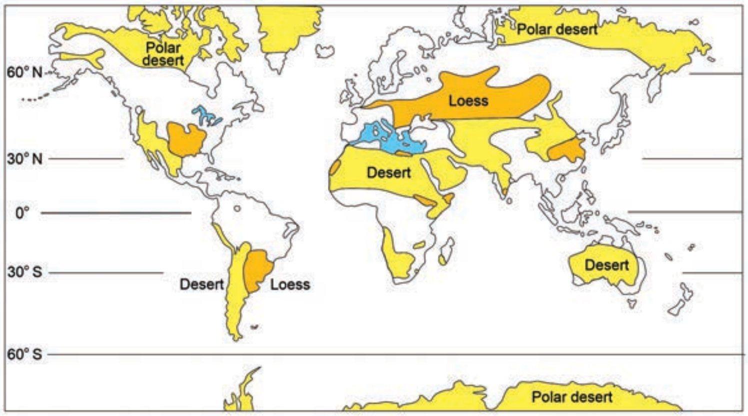

FIGURE 11.1 Global distribution of deserts and loess deposits. The loess deposits are accumulations of fine-grained, windblown sediments that are typically found in arid and semi-arid regions.

the dominant sedimentary force. Lack of water leads to scarce vegetation, which in turn facilitates wind-driven erosion as vegetation stabilizes sediments and prevents them from being easily moved around. Thus, the most common eolian environments are rocky deserts and ergs (the latter is the technical term for sandy deserts).

To better understand transport and depositional processes in deserts, it is important to have an idea of what controls wind erosion. Three parameters are important: wind velocity, surface cover, and grain size. The erosivity of the wind is an exponential function of wind velocity, i.e., if the wind velocity doubles, the wind is 8X more erodible or, if it triples, the wind is 27 times more erodible. This is given by the equation: $E = V^3$rho, where E is erosivity, V is velocity, and rho is air density. Hence, massive wind erosion (dust storms) is always accompanied by a significant increase in wind speed. Surface cover is an extremely important factor as there is negligible wind erosion on a vegetated surface. The threshold erosional velocity is related to the square root of grain size. Thus, larger grains resist erosion by virtue of their greater size (mass).

The key transport and depositional processes in deserts include wind erosion and deposition, saltation, suspension, and deflation (Figure 11.2). These are briefly described below.

Wind erosion occurs when strong winds pick up loose particles of sand, silt, and dust from the surface (Figure 11.2). This process is responsible for the formation of sand dunes and the removal of fine-grained sediments, leaving behind coarser ones. Most sand grains are transported by rolling or saltation, though clay- and silt-sized grains can be carried in suspension by winds that are strong enough, sometimes in quantities great enough to cause dust storms (Figure 11.2). Wind deposition occurs when the wind loses its energy and drops the sediment it was carrying, leading to the accumulation of sand, silt, and dust in specific areas. Contextually, it takes winds more than 60 mph just to move a grain 0.5 mm in diameter. Although deserts are arid, they can experience occasional rainfall or flash floods. During these events, water can cause erosion and deposition in desert environments. Flash floods can carve out channels and canyons, and sediment carried by the water can be deposited in alluvial fans (Figure 11.3A). The alluvial fans are fan-shaped deposits of sediment at the base of mountains or hills (Figure 11.3A).

Saltation is a process where wind transports sand-sized particles in a bouncing or hopping motion (Figure 11.2). The process transports sand grains in long (1 m or more) and low (within 1–2 m of the ground) trajectories as momentum is passed from grain to grain (Figure 11.2). Thus, as the wind moves across the desert surface, it lifts and carries these particles for short distances before they fall back to the ground due to gravity (Figure 11.2). At high wind speeds, saltation is more or less

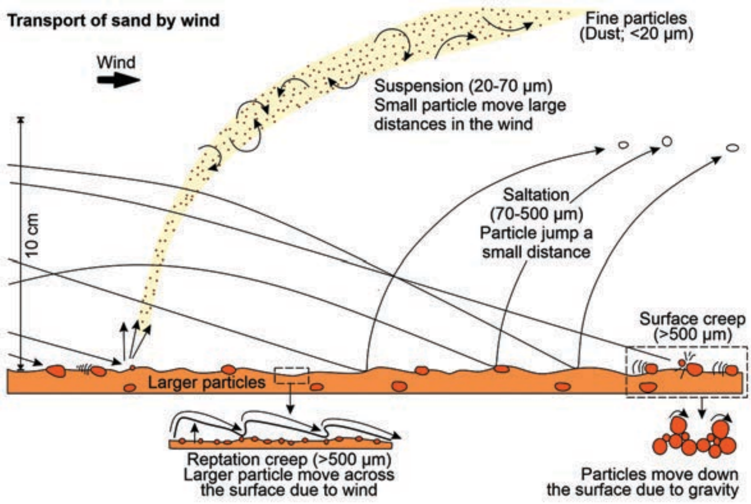

FIGURE 11.2 Sediment transport processes in deserts. In saltation, wind transports sand-sized particles in a bouncing or hopping motion. Surface creep (traction) is the movement of coarse sand and pebbles (up to 6x larger than saltating grains) as they slide and roll impacting one another and transferring momentum. In suspension, the smaller particles, such as silt and dust, are carried in the air by wind. Grains less than 0.2 mm in diameter are suspended in air as turbulent eddies, and are carried as dust for thousands of meters upwards, and 1000 km downwind, forming dust storms.

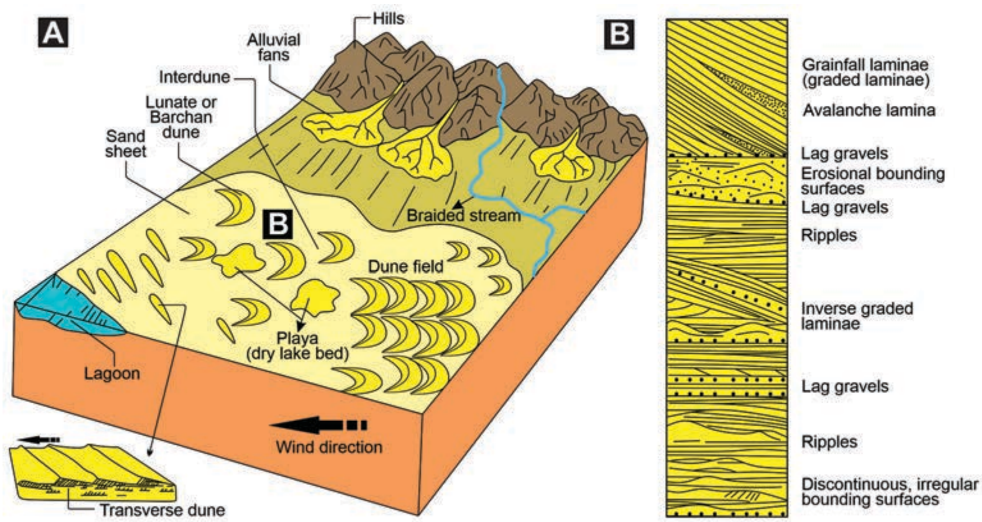

FIGURE 11.3 Eolian landforms. A: The transport and depositional processes interact with each other and contribute to the formation of various desert landforms, such as alluvial fans, various types of sand dunes, sheet sand, and playas (dry lake beds), among others. B: A typical eolian sheet sand deposit. (Modified after Ahlbrandt and Andrews, 1978.) B in Figure 11.3A marks the spatial location of the stratigraphic lithology illustrated in Figure 11.3B.

continuous and appears as a fuzzy layer next to the ground. Saltation is a significant mechanism for the transportation of sand in deserts.

Surface creep (traction) is the movement of coarse sand and pebbles (up to 6x larger than saltating grains) as they slide and roll impacting one another and transferring momentum (Figure 11.2). This process usually occurs with velocities greater than 4.5 m/sec.

Suspension is the process by which smaller particles, such as silt and dust, are carried in the air by wind (Figure 11.2). These fine particles remain suspended for long periods of time and are transported over long distances before settling back onto the ground. Air suspends grains less than 0.2 mm in diameter. Such grains are held in suspension by turbulent eddies, and are carried as dust for thousands of meters upwards, and 1000 km downwind.

Deflation refers to the removal of loose, fine-grained sediment by wind erosion (Figure 11.4). As the wind blows over the desert surface, it picks up and removes loose particles, leaving behind a deflation hollow or depression. Over time, deflation leads to the formation of desert pavements (elaborated below), where coarser particles are left behind to create a hardened surface (Figure 11.4).

These transport and depositional processes interact with each other and contribute to the formation of various desert landforms, such as desert pavements, alluvial fans, sand dunes, playas (dry lake beds), and wadis (dry riverbeds) (Figure 11.3). Understanding these processes is essential for studying desert geomorphology, sedimentary environments, and the impact of climate change on arid regions.

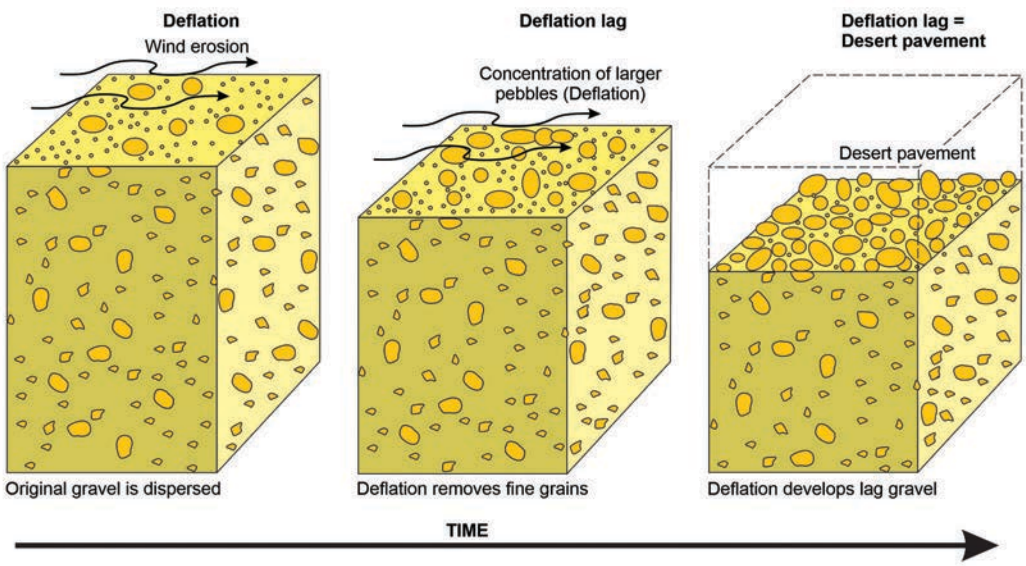

FIGURE 11.4 Deflation and desert pavements. Deflation is the removal of loose, fine-grained sediment by wind. Over time, deflation leads to the formation of desert pavements, where coarser particles are left behind and a hardened surface is formed. Desert pavements are erosive features and are also called reg or hammada. Hammada is an Arabic term meaning a rocky plain surface. The desert pavements are called by several names at different places: reg in western Sahara, serir in eastern Sahara, gibber plain in Australia and saï in Central Asia (Tarim Desert).

11.3 DESERT DEPOSITS

Desert deposits are primarily composed of sand, silt, and clay, and are often transported and deposited by wind and occasionally by water (see also Jain, 2014). A common desert deposit is the desert pavement, surfaces composed of closely packed pebbles, cobbles, and boulders (Figure 11.4). These pavements form as a result of wind erosion that removes fine-grained sediments and leaves behind coarser grains on the surface (Figure 11.4). Desert pavements are often found in arid regions with limited vegetation cover.

Blowouts occur when local areas are subjected to deflation forming blowouts (also called deflation hollows) (Figure 11.5). Blowouts are erosional hollows formed on vegetated (Figure 11.5A) to semi-vegetated areas in easily erodible sediments (such as sand, silt, etc.) (Figure 11.5B). They form where vegetation cover is either naturally low or absent, or reduced by various means (such as fire, animal or human activity, drought, coastal erosion, etc.), thus, allowing moderate to high-velocity near-surface winds to entrain and erode sediment forming a topographic depression (Figure 11.5B). These depressions may range from 3 meters in diameter and less than a meter deep to several kilometers in diameter and several hundred meters in depth. Blowouts are found in a variety of environments ranging from arid to semi-arid, glacial and arctic, temperate and tropical regions, and in coastal (Figure 11.5A) to continental environments (Figure 11.5B) (see Hesp and Smyth, 2019; Kooijman et al., 2021).

Alluvial fans are another type of desert deposit (Figure 11.6). These are fan-shaped accumulations of sediments that form at the base of a mountain or a hill (Figure 11.6). These are created by the deposition of sediments carried by intermittent streams or flash floods and are typically composed of a mixture of sand, silt, and gravel.

Playas are dry lake beds that form in desert regions (Figure 11.6). They are characterized by flat, clay-rich surfaces that are occasionally flooded during rare rainfall events. Playas often have evaporite deposits, such as salt and gypsum, due to the evaporation of water.

In addition to these deposits, desert environments also have various types of windblown sediment, such as loess (fine-grained, windblown silt) (see Figure 11.1) and dust. These fine particles are transported over long distances and deposited in other regions, including farmlands and even the deep ocean.

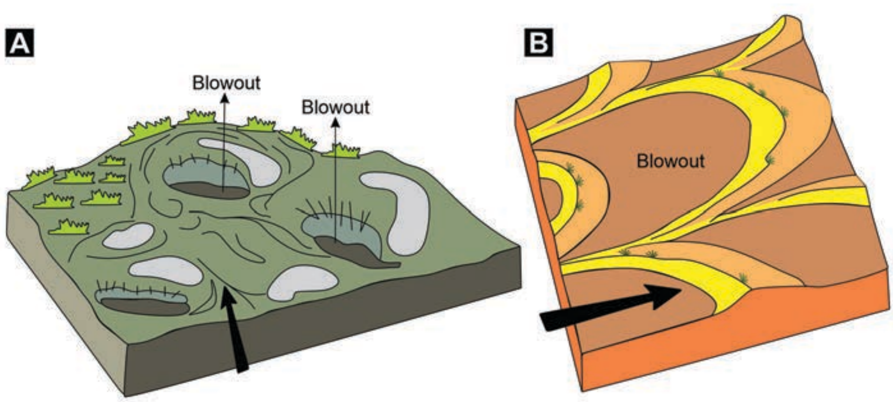

FIGURE 11.5 Blowouts. A–B: Blowouts are also called deflation hollows. They form where vegetation cover is either naturally low (A; as noted in coastal environments) or absent (B; as noted in continental environments), or reduced by various means (such as fire, animal or human activity, drought, coastal erosion, etc.), thus, allowing moderate to high-velocity near-surface winds to entrain and erode sediment forming a topographic depression.

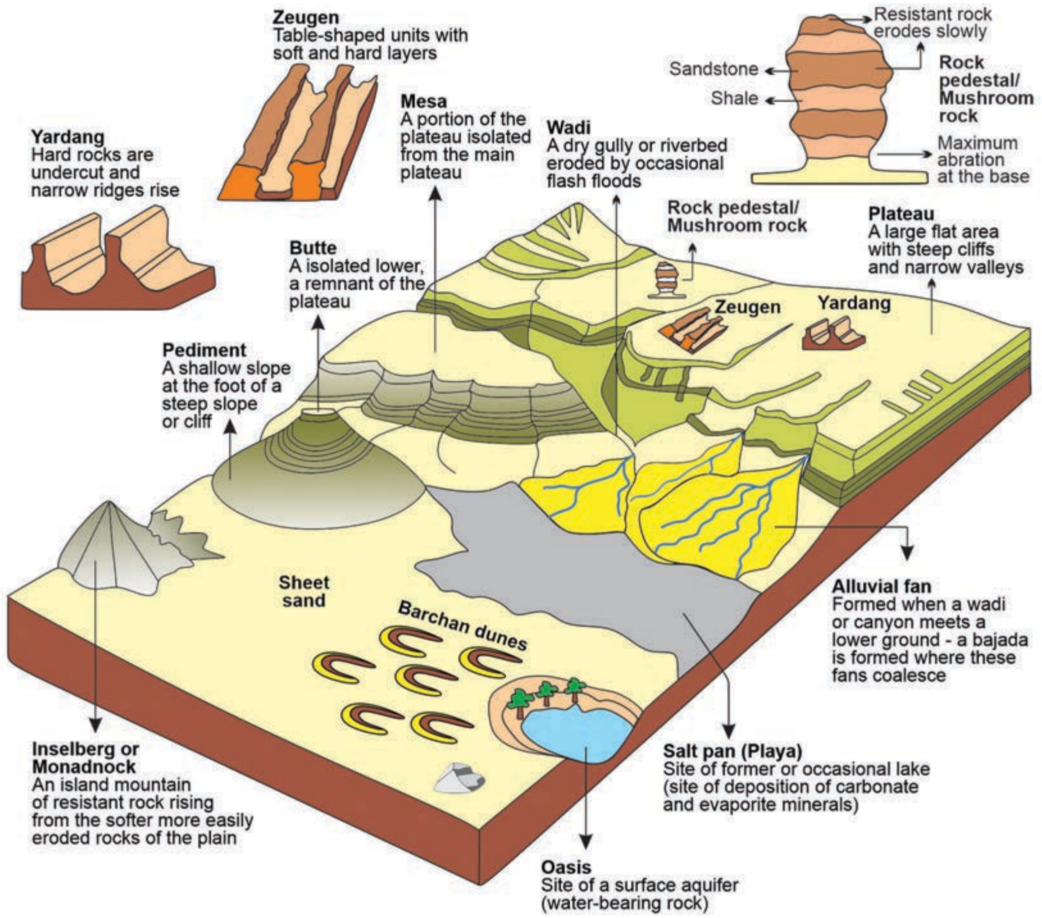

FIGURE 11.6 Desert depositional and erosional landforms. Among the depositional landforms, major ones include alluvial fan, salt pan (playa), pediment, and various types of dunes (barchan being the most prominent of them all). Among the erosional landforms, major ones are yardang and zeugen, mesa and butte, inselberg (monadnock or island mountain), and rock pedestal (mushroom rock).

The other desert deposit, and arguably the most common of them all, are sand dunes (Figures 11.3 and 11.6). These are mounds or ridges of sand that form due to wind transportation and deposition. They have various shapes such as crescent-shaped (barchan dunes), linear (transverse dunes), and star-shaped (star dunes). Sand dunes are typically composed of well-sorted, medium to coarse-grained sand. These are detailed below.

Thus, by studying desert deposits valuable information about past and present environmental conditions, including climate change, sedimentary processes, and landscape evolution is inferred. These deposits also contain important records of past ecosystems, human activities, and geological events.

11.3.1 Loess Deposits

Loess deposits are accumulations of fine-grained, windblown sediments that are typically found in arid and semi-arid regions (Figure 11.1). These deposits consist of silt-sized particles that are larger than clay but smaller than sand. Loess is usually light yellow or buff in color and has a distinctive ability to hold water. Loess deposits are formed when wind picks up fine-grained sediment from one area and deposits them in another. The particles are usually derived from the erosion of rocks and minerals, such as quartz, feldspar, and mica. The wind transports and deposits these particles over long distances, creating thick layers of loess. Loess deposits are important and they support fertile soils and have been extensively used for agriculture throughout history (see Stoops, 2021). The fine particles in loess hold water and nutrients, making it suitable for growing crops. Loess deposits also provide valuable information about past climates and environmental conditions. The layers of loess acts as a record of wind patterns and changes in vegetation over time. Scientists study these deposits to better understand past climate variations and their impact on ecosystems. However, loess deposits also pose challenges. The loose nature of the sediment makes it susceptible to erosion, which leads to extensive land degradation and desertification. In some cases, loess deposits are also prone to landslides and other geohazards.

11.3.2 Deflation Pavement

Deflation pavement refers to a land surface feature that is created by wind erosion where wind removes finer particles such as sand and silt from an area, leaving behind coarser materials like pebbles, gravel, and even larger rocks (Figure 11.4). Deflation (the entrainment of loose sediments) is commonly found in arid or desert environments where there is limited vegetation and loose, unconsolidated sediments. As the wind blows over the ground, it picks up and carries away the finer sediments, leaving behind a surface that is covered with larger, more resistant materials (Figure 11.4). Over time, this process creates a flat, barren surface with a layer of coarser sediments on the top called a deflation pavement (Figure 11.4). The pavement varies in thickness and composition depending on the local geology and the intensity of wind erosion (Figure 11,4). The deflation pavement also acts as a protective layer, preventing further erosion and stabilizing the ground surface (Figure 11.4). It can also influence other geological processes such as water runoff and sediment transport. The presence of a deflation pavement indicates past or ongoing wind erosion in an area and is an important feature for geologists studying desert landscapes and the effects of wind on landforms.

11.3.3 Yardang and Zeugen

Both yardang and zeugen are products of wind erosion. Wind erodes soft rocks through the abrasion process as they are more prone to wind erosion whereas hard rocks are less likely to get eroded; this differential erosion results in the formation of yardang and zeugen. Yardangs (a Turkish term) are long, elongated ridges or hills that are formed roughly parallel to each other and to the direction of the wind, thus a product of wind erosion (Figure 11.6). Zeugen, on the other hand, are tabular masses of resistant rocks that stand prominently in the desert and are composed of alternating layers of hard and soft rocks (Figure 11.6). Zeugens are horizontally eroded whereas yardangs are vertically eroded (i.e., vertically placed layers of soft and hard rocks).

11.3.4 Rock Pedestal (Mushroom Rock)

These structures are formed due to the erosion of an isolated rocky outcrop progresses at different rates at its bottom than at its top. Wind erodes the softer sediments from the bottom leaving the top hard resistant rock as a pedestal, thus, resembling a mushroom (Figure 11.6). These structures are commonly found in hot and arid regions such as deserts.

11.3.5 Mesas and Buttes

Mesa is a Spanish word for ‘‘table’’. Mesa is a flat, table-like landmass with a horizontal top layer made of resistant rocks and with steep sides (Figure 11.6). As erosion proceeds, columns of rock called buttes remain that jut out of the landscape (Figure 11.6).

11.3.6 Inselberg (or Monadnock)

An inselberg, also called an island mountain or bornhardt, is an isolated residual hill (knob, ridge, or a small mountain) rising abruptly from ground level, and consisting of steep slopes and rounded tops (Figure 11.6). It is often made of gneiss and granitic rocks.

11.3.7 Pediments

These are gently inclined (usually 2–3°, but may range up to 10–15°), and eroded bedrock surfaces near the mountain base formed by ephemeral streams by lateral planation or by sheetwash flooding, and developed on the bedrock (Figure 11.6).

11.3.8 Ergs

An erg (sand sea or dune sea, or sand sheet, i.e., vast sand sheet) is a flat area of a desert covered by windblown sand (fine-grained) (see Figure 11.6). This could be more than 125 km^2 of eolian or windblown sand with little or no vegetation; smaller areas are called dune fields. Ergs are typically found in arid or desert regions. They are also areas of actively shifting dunes, "fossilized" dunes, or extensive sand sheets. About 85% of all mobile sand is found in ergs, and they cover an area bigger than 32,000 km^2. Thus, ergs are large-scale features formed by wind-driven sediment transport and deposition, and are characterized by extensive sand dunes that can reach heights of tens of meters and cover hundreds or even thousands of square kilometers. The formation of ergs is closely tied to the availability of sand and the prevailing wind patterns within a particular region. The wind acts as a primary agent of erosion and transportation, picking up loose sand particles from exposed surfaces and carrying them across the landscape. When the wind encounters obstacles or changes in topography, it slows down and deposits the sand, resulting in the formation of dunes. Ergs vary in shape and size, ranging from linear dune fields aligned with the prevailing wind direction to complex networks of interconnected dunes. Thus, a large dune field is called an erg. The Skeleton Coast Erg in Namibia extends 2–5 km in length, and with a width of 20 km. Other examples include the Sahara Desert in Africa, the Arabian Desert in the Middle East, and the Simpson Desert in Australia. Geologists study ergs to understand the processes of wind-driven sediment transport, landform evolution, and the interaction between wind, sand, and the surrounding environment. These sand dunes also influence local climate patterns by altering wind flow and affecting temperature and humidity levels.

11.3.9 Desert Varnish

A dark coating on rocks caused by windblown particles and chemical reactions (Figure 11.7). Desert varnish is a surface stain, coating bedrock surfaces and surface sediments in arid regions around the world. It contains internal microscopic laminations (~<200 μm thick), each layer reflecting climatic conditions under which it was formed. These individual layers have been used to reconstruct local and regional climate history and the dating of geomorphic surfaces. Although the mineral and chemical components of desert varnish are windblown dust, atmospheric aerosols, and microbes, the role of microbial activity vis-à-vis desert varnish remains controversial (see also Dickerson, 2011).

11.3.10 Dunes

Dunes are landforms created by the movement of wind on loose sediments, such as sand or silt (Figure 11.8). They originate as a mound of free sand from a sandy surficial deposits (such as a beach or weathering sandstones) or from a blowout (Figure 11.5). As the mound grows, it develops the dune asymmetry characterized by a gentle windward slope and a leeward slip face at the angle

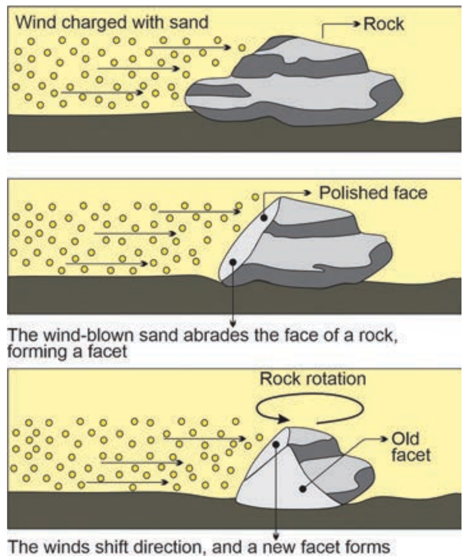

FIGURE 11.7 Desert varnish.

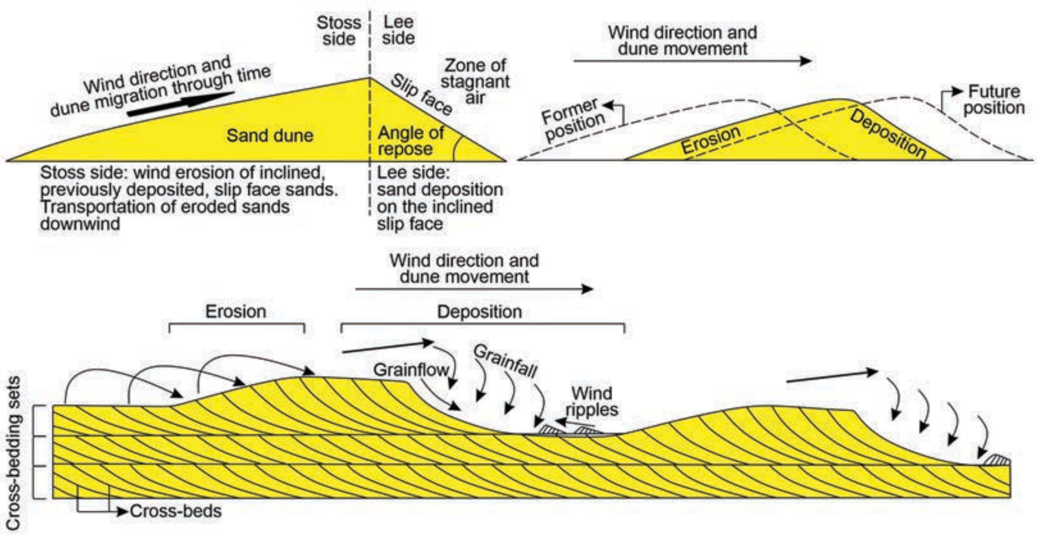

FIGURE 11.8 Dune terminology and the migration of dunes.

of repose for sand (Figure 11.8). Dunes possess the same longitudinal shape as a ripple but are several orders of magnitude larger in size. Dunes vary in size, shape, and composition, depending on the prevailing wind or water conditions and the available sediment. They are stable or advancing landforms of windblown sand, typically found in desert and coastal environments, and in some river and lake systems. Wind-formed dunes, known as aeolian dunes, are one the most well-known types (Figure 11.8).

Dunes (Figure 11.9A) typically form in areas with a steady wind direction and an abundant supply of loose sand (Figure 11.9B). The wind pushes the sand grains along the ground until they encounter an obstacle, such as vegetation, rocks, or changes in topography. The sand then accumulates and

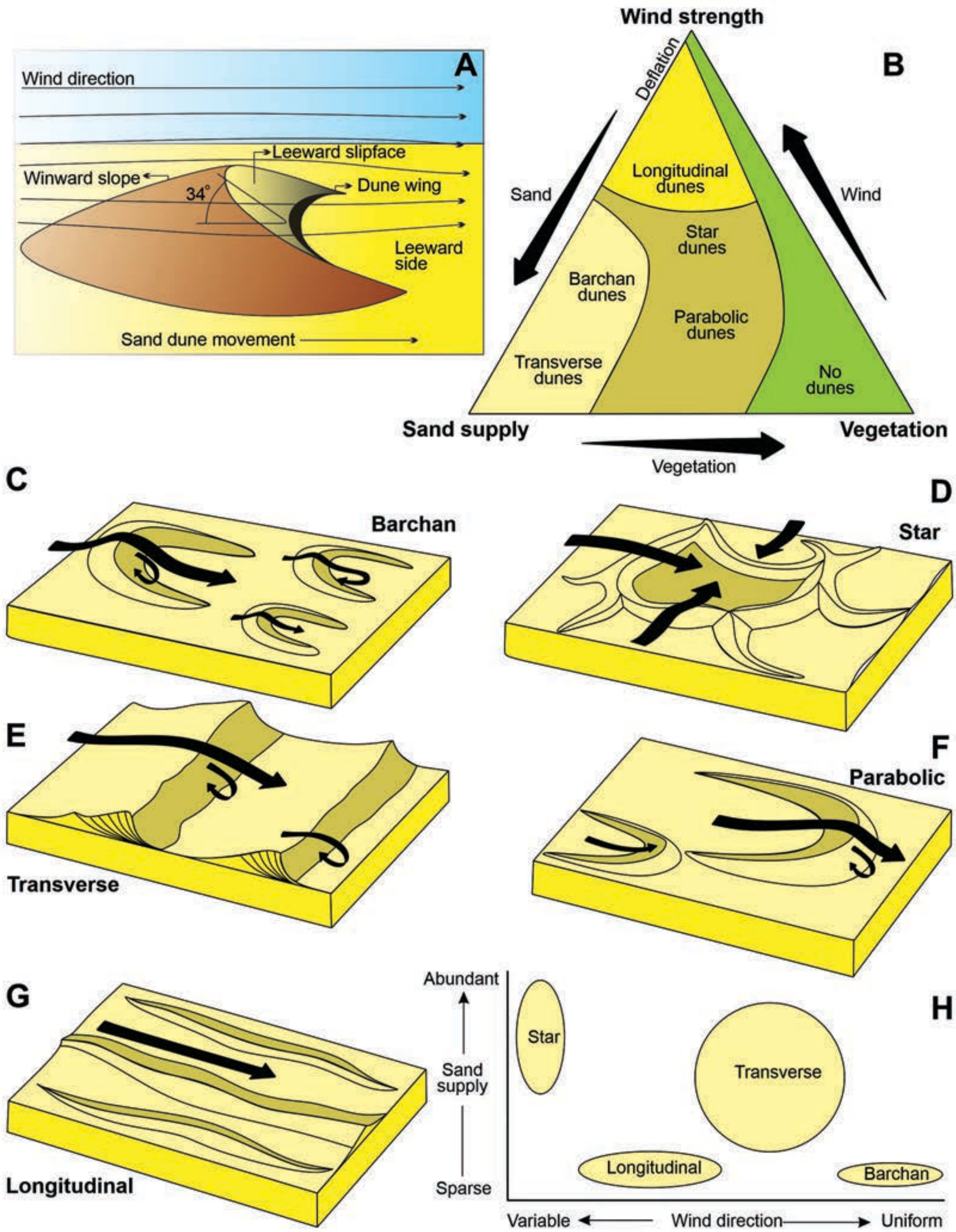

FIGURE 11.9 Sand dunes. A: Typical morphology of a dune. B: Triangular classification of dune, based on the availability of the three major components – sand supply, wind strength, and vegetation. C–G: Types of dunes. H: Dune types on a diagram of sand supply vs. wind variability. (Modified after McKee, 1979; Bishop, 2001.)

builds up, forming a dune. Thus, dunes are dynamic landforms that change over time due to erosion, deposition, and shifting wind or water patterns. There are several types of eolian dunes, such as barchan (Figure 11.9C), star (Figure 11.9D), transverse (Figure 11.9E), parabolic (Figure 11.9F), and longitudinal (Figure 11.9G). These are very briefly enumerated below.

Barchan dunes are crescent-shaped with horns pointing downwind (Figure 11.9C). They form in areas with limited sand supply and unidirectional winds (see Figure 11.9B). They are isolated, freely migrate across desert plains, and maintain their form. Star dunes have multiple arms radiating from a central point (Figure 11.9D). They form in areas with variable wind directions (see Figure 11.9B). Transverse dunes are linear and cuspate dunes that have a series of ridges perpendicular to the direction of the wind (Figure 11.9E). They form in areas with a steady wind and abundant sand supply (see Figure 11.9B). With stronger winds they evolve into barchans (Figure 11.9C). They usually occur on beaches, floodplain alluvium, or erodible sandy bedrock rather than in dry deserts. Parabolic dunes are associated with vegetation, so form in subhumid and semi-arid environments (rather than arid) where vegetation is nearby (such as beaches or grasslands) (Figure 11.9F). They originate as a blowout (see Figure 11.5). Longitudinal dunes are large (kms in length, ~ one km wide) linear forms, and form parallel to strong persistent winds (Figures 11.9G and 11.9B). They have elongated ridges parallel to the wind direction. They form in dry subtropical deserts with irregular sand supply. They are separated by lag gravel. The relative position of dunes discussed above with respect to the direction of wind and the availability of sand is provided in Figure 11.9H.

Large eolian bedforms such as complex and compound varieties of dunes of many sand seas are called draas (Figures 11.10A–B). These are well represented in the rock record, and noted in modern sand seas such as in the Namib, a coastal desert in Southern Africa. Draas are the largest elements within the hierarchical arrangement of bedforms (see Figure 11.10C) (see Wilson, 1972; Lancaster, 1988). Draas are characterized by the superimposition of smaller dunes of the same type (compound dunes) or different type (complex dunes) on larger structures. In general, draas are large-scale dunes (simple dunes) with smaller superimposed dunes and ripples on them (Figures 11.10A–B). They can reach a height of 20 to 450 m.

Dunes play an important role in shaping landscapes, providing habitats, and influencing local climate conditions. Geologists study dunes to understand sediment transport, landform evolution, and the interactions between wind or water and sediment.

11.3.11 Interdunes

Interdunes form in-between sand dunes (Figure 11.11). They are commonly found where wind is the main agent of sediment transport. Interdunes typically consist of flat or gently sloping areas of sand (low angle stratification; ~10°) that are not covered by dunes (Figure 11.11). Interdunes vary in size and shape, depending on the characteristics of the surrounding dunes and the local wind regime. They are elongated or circular in shape and range in length from a few meters to several kilometers. The formation of interdunes is influenced by various factors, such as wind patterns, sediment availability, and topography. When wind blows across a desert landscape, it carries sand particles and deposits them in areas of low wind velocity or turbulence. These areas are interdune areas, and are typically found in the troughs or depressions between dunes (Figure 11.11). The sand in interdunes is often loosely packed (structureless) and contains small ripples or dune-like features; incipient bioturbation may destroy stratification. The flat, stable surface of interdunes allows for the establishment of vegetation, which helps stabilize the sand and prevents further dune migration. In addition to their ecological significance, interdunes also have implications for human activities in desert regions. They affect the movement of sand and the stability of infrastructure, such as roads and buildings. Understanding their formation and dynamics is therefore important for land management and development in desert areas.

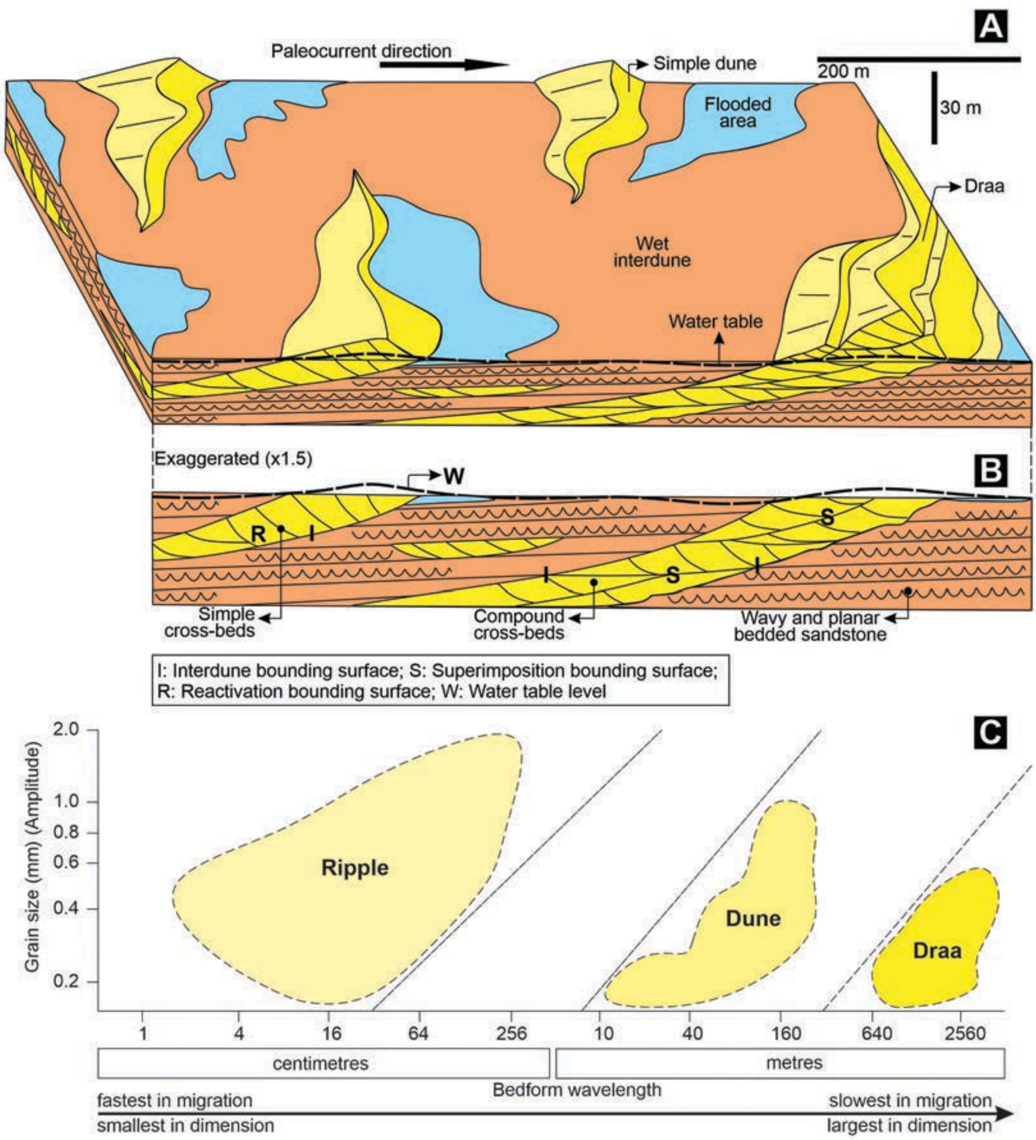

FIGURE 11.10 Draas. A–B: The draas are complex and compound varieties of dunes characterized by the superimposition of smaller dunes of the same type (compound dunes) or different type (complex dunes) on larger structures. C: Draas are the largest elements in a hierarchical arrangement of bedforms.

11.3.11.1 Dry Interdunes

Dry interdunes are also known as dry sand flats or interdune pavements (Figures 11.12A–B). These interdunes are characterized by their arid and barren nature, with little to no vegetation and minimal water presence. Dry interdunes form between active sand dunes and are typically flat or gently sloping areas of sand that lack the characteristic dune shape (Figures 11.12 A–B). They are often composed of loose, dry sand that has been deposited by wind processes. These interdunes vary in size and shape, ranging from small patches to extensive areas of several kilometers in width. The lack of vegetation is primarily due to harsh environmental conditions. The arid climate and limited water

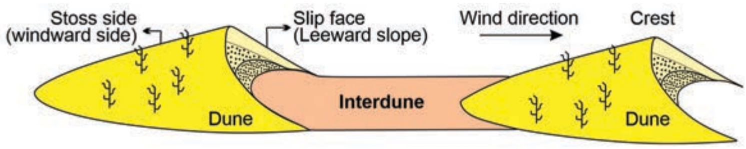

FIGURE 11.11 Interdune. They form in-between sand dunes and typically consist of flat or gently sloping areas of sand (with low angle stratification; ~10°). They are commonly found where wind is the main agent of sediment transport.

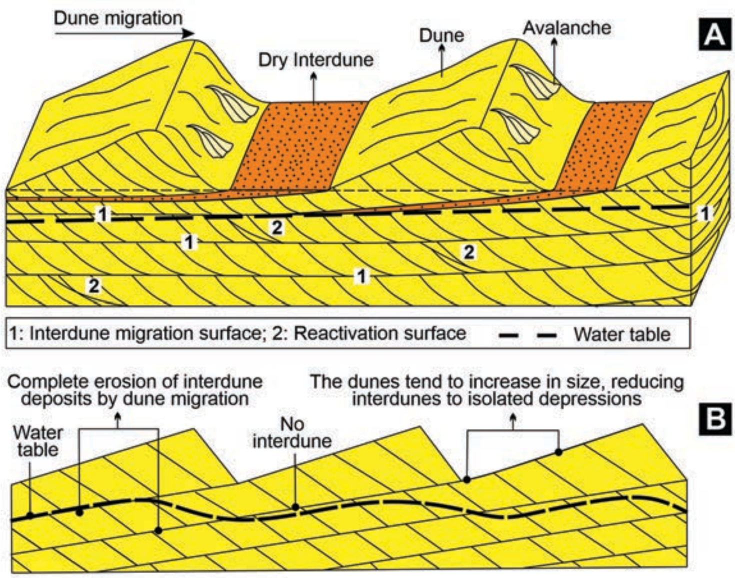

FIGURE 11.12 Dry interdunes. A–B: They are also known as dry sand flats or interdune pavements. These interdunes are characterized by their arid and barren nature, with little to no vegetation and minimal water presence (A: Modified after Kendigelen, Oguzhan. "Facies distribution, its implications for climate signals, and hydrocarbon potential of the Permian Lyons Sandstone, Front Range Basin, northern Colorado, USA." [2017].)

availability make it challenging for plants to establish and survive in these areas. The sandy substrate also tends to be unstable, making it all the more difficult for plant roots to anchor and access water. Hence, dry interdunes are often subject to intense wind erosion and deposition. Wind reshapes the sand surface, creating ripples, dunes, or even sand sheets. Dry interdunes are characterized by relatively coarse, bimodal, and poorly sorted grains in gently dipping, poorly laminated layers. The dry interdunes act as barriers to wind, helping to stabilize the dune system and preventing the migration of sand. The flat surface of dry interdunes also provides a pathway for water runoff, allowing it to infiltrate into the groundwater system and contribute to the desert's hydrological cycle.

11.3.11.2 Wet Interdunes

Wet interdunes, also known as damp interdunes, occur in sandy desert environments (Figure 11.13). These are characterized by the presence of water, either as surface water or groundwater, thus creating a moist environment within the dune system (Figure 11.13). The presence of water in wet interdunes is due to the accumulation of rainfall or runoff within the dune system. During periods of precipitation (rain), water collects in depressions or low-lying areas between dunes, forming

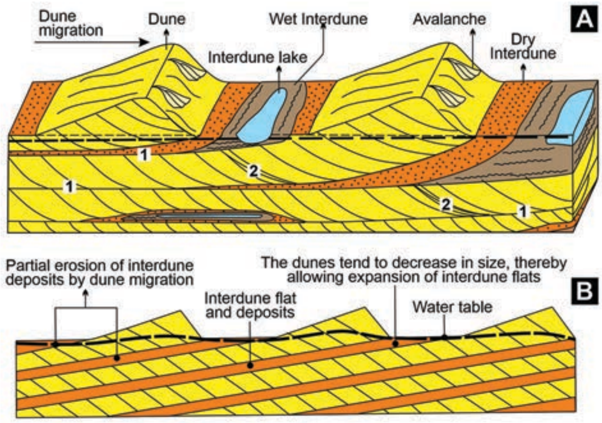

FIGURE 11.13 Wet interdunes. A–B: They are also known as damp interdunes and are characterized by the presence of water, either as surface water or groundwater, thus creating a moist environment within the dune system. (Modified after Kocurek and Havholm, 1993.)

temporary pools or wetlands. Another source of water in wet interdunes is groundwater. In some desert regions, underground water sources may intersect with the dune system, leading to the emergence of springs or seepage zones (Figure 11.13). This groundwater can sustain the moisture content within the interdunes, creating a unique habitat that supports specialized plant and animal species. Thus, these moist areas provide vital resources for desert flora and fauna, allowing for the establishment of vegetation and the survival of diverse organisms. Wet interdunes often support a higher biodiversity as compared to the surrounding dry dunes. Additionally, wet interdunes play a crucial role in the hydrological cycle of arid regions. The water stored within the dunes slowly infiltrates into the groundwater system, recharging underground reservoirs and contributing to the overall water balance within the desert ecosystem.

11.3.12 Sheet Sands

Sheet sands are characterized by extensive, laterally continuous layers of sand that commonly surround dune fields and are formed by the accumulation of sand grains transported by waves, currents, and tides (see also Fryberger et al., 1979). Sheet sands often exhibit well-sorted and well-rounded sand grains. The grains may be composed of various minerals, depending on the local geology. These deposits are typically characterized by low to moderately dipping cross-stratification (0 to 20°), gently dipping, curved, or irregular erosion bounding surfaces, gently dipping, poorly laminated layers resulting from grainfall deposition, and discontinuous, thin layers of coarse sand intercalated with fine sand (see Figure 11.3). The sheet sand deposits can extend over large areas and can be several meters thick. By analyzing the characteristics of sheet sand deposits, scientists can reconstruct past environmental conditions and track changes in coastal areas over time.

11.4 TYPES OF EOLIAN SYSTEMS

There are several types of eolian systems, that are characterized by processes and landforms associated with wind-driven sediment transport. These systems can be found in various environments, including deserts, coastal areas, and in some inland regions. The main types of eolian systems are desert (this chapter), coastal, inland, polar (polar deserts), and aeolian systems in other environments. These are briefly enumerated below.

The desert eolian system (this chapter) is found in arid and semi-arid regions, where wind erosion and transportation of sediments are dominant processes. The desert eolian systems are typically characterized by strong winds, limited vegetation, and abundant loose sand. The system includes features such as sand dunes, sand sheets, and desert pavements.

The coastal eolian system occurs in coastal areas, where wind plays a significant role in shaping the landscape. It includes features such as beach dunes, beach ridges, and sand spits. These systems are influenced by the interaction of wind, waves, and tides, and they often provide important protection against coastal erosion.

The polar eolian system (polar deserts) occurs in polar regions, such as the Arctic and the Antarctic, where wind-driven processes are important in shaping the landscape. It includes features such as snow dunes, snow drifts, and ventifacts (rocks shaped by wind erosion). Polar eolian systems are characterized by cold temperatures, strong winds, and a predominance of snow and ice.

The inland eolian system is found in non-desert regions, such as plains or plateaus, where wind is a significant agent of sediment transport. These systems are characterized by moderate to strong winds and lack significant vegetation cover. They include features such as loess deposits, sand sheets, and wind-eroded landforms.

Eolian systems in other environments include environments such as river valleys, lake shores, and volcanic regions. These systems are influenced by local topography, wind patterns, and the availability of loose sediments.

More importantly, the desert systems are also characterized as dry, wet, and stabilized eolian systems (see Kocurek, 1996). These are briefly described below.

11.4.1 Dry Eolian Systems

In dry eolian systems, the water table and its capillary fringe lie at a depth below the depositional surface (see Figure 11.14A). The system is characterized by the predominance of large-scale dunes, absence of interdunes, rare bioturbation and pedogenetic features, and the general lack of interbedded alluvial deposits. In general, dry interdune accumulations are uncommon in the geological record (Kocurek and Havholm, 1993).

11.4.2 Wet Eolian Systems

In the wet eolian system, the water table is at or near the depositional surface and hence deposition and erosion are controlled by the moisture content (see Figure 11.14B). These systems occur in environments where wind-driven processes occur in areas with a significant presence of water. These systems can be found in coastal regions, river floodplains, lake shores, and other wetland environments. Wet eolian systems exhibit unique characteristics and landforms due to the interaction between wind, water, and sediments. Some examples are coastal, fluvial, lacustrine, and wetland dunes. These are briefly enumerated below.

Coastal dunes are dunes formed along the coast in areas where wind and water interact. Coastal dunes are found in beach and barrier-island systems, where wind transports sand from the beach and deposits it inland, forming dune ridges.

Fluvial dunes form in river floodplains and are influenced by both wind and water. Fluvial dunes are typically elongated in shape and are formed by the wind carrying sediment from exposed riverbanks and depositing it on the floodplain.

Lacustrine dunes are found in lake shore environments and are formed by the interaction of wind and water. Lacustrine dunes are influenced by waves and currents; these dunes exhibit unique shapes and patterns.

Wetland dunes occur in wetland environments. These dunes are typically smaller in size and are formed by wind transporting and depositing sediments in areas with vegetation or with shallow water.

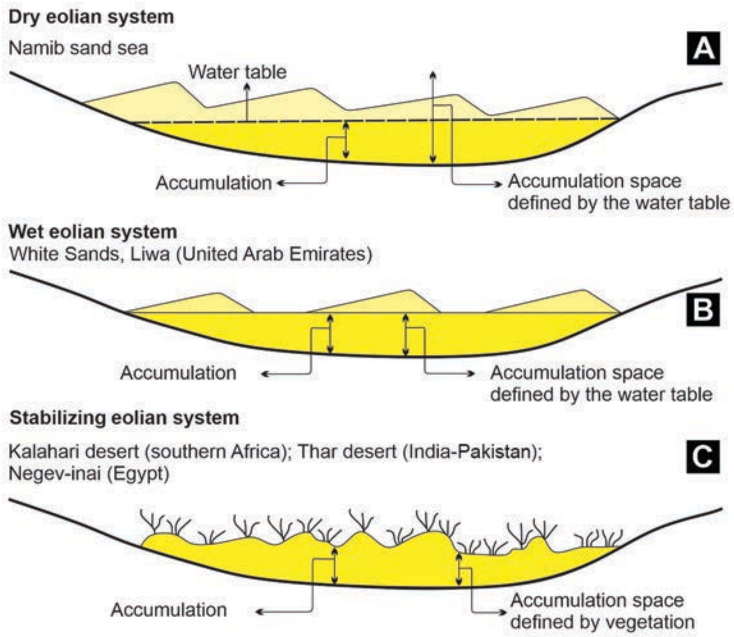

FIGURE 11.14 Types of eolian systems, showing the components of accumulation and preservation of a dune system. (Modified after Kocurek and Havholm, 1993.) A: Dry eolian system. The water table and its capillary fringe lie at a depth below the depositional surface. B: Wet eolian system. The water table is at or near the depositional surface and hence deposition and erosion are controlled by the moisture content. C: Stabilized eolian system. Vegetation and mud drapes play a stabilizing role and influence the behavior of the accumulating surface.

Wet eolian systems are important for understanding sediment transport, landform evolution, and ecosystem dynamics in these unique environments. They can provide insights into how wind and water interact and shape the landscape, as well as the ecological significance of these processes.

11.4.3 Stabilized Eolian Systems

In stabilized eolian systems, vegetation and mud drapes play a stabilizing role and influence the behavior of the accumulating surface (Figure 11.14C). In these systems, wind erosion and sediment transport are significantly reduced or halted, resulting in a relatively stable landscape (see Kocurek, 1996). These systems are found in various environments, including deserts, coastal areas, and inland regions. Some examples include stabilized desert dunes, vegetated coastal dunes, stabilized loess deposits, stabilized sand sheets, and stabilized volcanic ash deposits. These are briefly enumerated below.

Stabilized desert dunes are formed by windblown sand accumulating in specific shapes and patterns. Over time, dunes become stabilized through the establishment of vegetation or the accumulation of surface crusts. Stabilized desert dunes are characterized by fixed shape and limited sand movement.

Vegetated coastal dunes are formed by windblown sand in coastal areas. When vegetation, such as beach grasses or shrubs, takes root on the dunes, it stabilizes the sand and prevents further erosion. These dunes provide vital protection against coastal storms and erosion.

Stabilized loess deposits are fine-grained sediments composed of windblown silt and clay particles. In regions with loess deposits, vegetation cover stabilizes the sediment and prevents further erosion. Stabilized loess deposits form distinctive landforms, such as bluffs or terraces.

Stabilized sand sheets are extensive areas covered by windblown sands. In some cases, vegetation cover or surface crusts stabilize the sand and limit further movement. Stabilized sand sheets are found in both desert and non-desert environments.

Stabilized volcanic ash deposits are transported and deposited by wind, forming layers of fine-grained sediments. Over time, vegetation or other factors stabilizes these ash deposits, preventing erosion and maintaining a stable landscape.

Thus, the stabilized eolian systems are important for understanding the long-term evolution of landscapes and for assessing the stability and ecological significance of these areas. They provide insights into the processes of sediment transport, landform development, and the interaction between wind and the environment.

11.5 BOUNDING SURFACES IN EOLIAN DEPOSITS

Bounding surfaces are distinct layers or boundaries within the sedimentary sequence that separate different depositional units or episodes (Figure 11.15). These surfaces represent changes in sedimentation conditions, depositional environments, or erosional events. Bounding surfaces in eolian deposits are classified into various types based on their characteristics and origins. Some common types of bounding surfaces include erosional, deflation, lag, bedding, reworked, reactivation surfaces, and abrupt changes in sediment texture or composition. These are briefly enumerated below.

Erosional surfaces represent periods of erosion or non-deposition within the eolian system (see Figure 11.15). These can result from wind erosion, water erosion, or other erosional processes. Erosional surfaces are often characterized by a distinct change in sediment texture, color, or composition above and below the surface.

Deflation surfaces occur when wind removes fine-grained sediment from the surface, leaving behind coarser-grained material (deflation lag; see Figure 11.4). These surfaces can be reactivated or redefined when wind strength increases, leading to further erosion and deflation of the coarser grains. These surfaces typically have a smooth, flat appearance and are commonly found in sandy eolian environments.

Lag surfaces are formed when wind selectively removes finer-grained sediment, leaving behind a concentration of coarser particles (desert pavement; see Figure 11.4). They often represent a lag deposit of resistant grains, such as pebbles or cobbles, and can indicate a change in sediment supply or transport conditions.

Bedding surfaces are horizontal layers within eolian deposits that separate individual beds or laminae (see Figure 11.8B). Changes in sedimentation rates, wind direction, sediment supply, or other factors that influence sedimentation rates and patterns results in the formation of distinct beds or laminae above or below the bedding surface.

Reworked surfaces occur when previously deposited sediment is reworked or remobilized by wind, resulting in the mixing or reorganization of sediment layers. These surfaces are characterized by cross-bedding, scour marks, or other evidence of sediment reworking.

Reactivation surfaces (or redefinition surfaces or third-order bounding surfaces) are boundaries or surfaces within a sedimentary sequence that marks a change in the activity or behavior of the eolian system (Figure 11.15). These surfaces indicate a shift in sediment transport, deposition, or erosional processes, often associated with changes in wind strength, direction, or sediment availability.

Abrupt changes in sediment texture or composition: reactivation or redefinition surfaces can also be characterized by abrupt changes in sediment texture or composition. These changes can indicate a shift in sediment source or transport processes, often associated with changes in wind patterns or sediment availability.

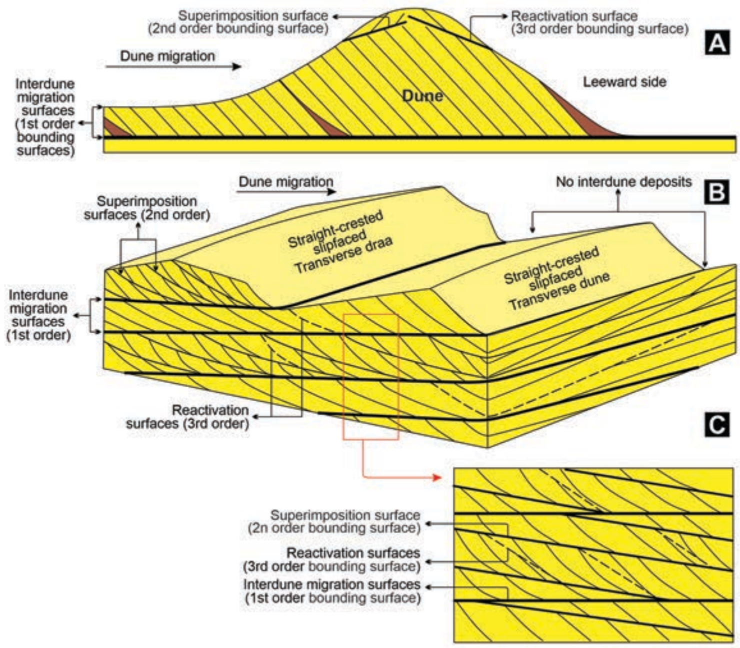

FIGURE 11.15 Bounding surfaces. Bounding surfaces are distinct layers or boundaries within the sedimentary sequence that separate different depositional units or episodes. Brookfield (1984) identified three bounding surfaces based on a hierarchical scheme including flat, first-order surface related to the migration of large bedforms; inclined, second-order surface that slopes downwards and encloses cosets of cross-strata deposited by smaller dunes superimposed on large bedforms; and third-order surface that is generated by the erosional modifications of the leeward faces of the migrating dunes.

Besides the aforementioned bounding surfaces, Brookfield (1984) and Kocurek (1996) suggested a different classification scheme. Brookfield (1984) identified three bounding surfaces based on a hierarchical scheme including flat, first-order surfaces related to the migration of large bedforms; inclined, second-order surfaces that slope downwards and enclose cosets of cross-strata deposited by smaller dunes superimposed on large bedforms; and third-order surfaces that are generated by the erosional modifications of the leeward faces of the migrating dunes.

However, Kocurek (1996) found this hierarchical scheme difficult to apply and proposed the following terminology. An interdune migration surface (or first-order bounding surface) is the interdune surface (left behind by stoss-side erosion of a passing dune, and possibly further eroded as an interdune surface) (see Figure 11.15B–C). Moist sand is more resistant to erosion than dry sand, so sometimes the water table controls the extent of interdune erosion. Interdune migration surfaces tend to be near horizontal. A superimposition surface (or second-order bounding surface) is created by the migration of one dune (usually a secondary smaller dune) over another (see Figure 11.15B–C). This can happen without erosion of the underlying dune. A reactivation surface (or third-order bounding surface) is created when a deposition process is temporarily interrupted by a change of wind direction causing erosion (see Figure 11.15B–C). The erosion surface becomes a reactivation surface if later deposition resumes in the original direction, laying down more deposits in the original orientation.

In general, the bounding surfaces are important for understanding the depositional history, environmental conditions, and sedimentary processes within eolian systems. They provide valuable information about the dynamics and evolution of eolian systems – changes in wind patterns, sediment availability, and the evolution of eolian landscapes over time. They can help identify periods of increased or decreased sediment transport, changes in depositional environments, and the influence of external factors on eolian processes. By studying these surfaces, scientists can gain insights into the history and behavior of eolian systems over time.

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12 Glacial System

12.1 INTRODUCTION

The downhill movement of a thick mass of ice (including air, water, and rock debris) under the pull of gravity is called a glacier. Glaciers make up about 10% of the land surface where the water is below freezing point; they also contain ~68% of the world's freshwater supply. Glaciers can be as small as a valley between two mountains or as large as a continent (such as the Antarctica) (Figures 12.1A–B). The polar regions, i.e., at latitudes above the Arctic and Antarctic circles (Iceland, Arctic, Alaska, and Antarctica,) house most of the world's glaciers (Figure 12.1).

12.2 FORMATION OF GLACIERS

In areas of high elevation and/or at high latitudes, when the snowfall in the winter exceeds the amount of snow and ice that melts in the following summer, glaciers form. Thus, their size and extent are governed by the prevailing regional climate. The balance between the accumulation (i.e., zone of accumulation) of ice, and the melting (i.e., zone of ablation) is called the glacier mass balance; the line that separates these two zones is called the equilibrium line (Figure 12.2). This division (i.e., the ratio between accumulation:ablation) has been used to subdivide glaciers; for example, the alpine glaciers range from 8:9 to 2:3 (see Meier, 1962; Deynoux et al., 2004). In the accumulation zone, snow doesn't melt even during summers but as glaciers move down, they eventually melt in the ablation zone (Figure 12.2). If a glacier has more accumulation than ablation for several years, then, the glacier advances downwards; however, if more ablation occurs, then, it retreats (Figure 12.2). The line above which, snow forms and remains round the year is called the snowline and glaciers form only at latitudes or altitudes above the snowline (Figure 12.3).

12.3 GLACIER MOVEMENT

In general, the movement of glaciers is slow; 0.01 to 0.1 m/day for large continental glaciers to 0.1–2 m/day for alpine ones. Flow rates are greatest in the center of the ice mass, and minimum at rock contacts. Glacier movement is by basal slip and plastic flow in the lower part (Figure 12.4), with the upper brittle part riding on the lower. Striations and chatter marks (horseshoe-like indentations in the bedrock) are evidence of basal slip. As long as ablation (loss) of ice is less than accumulation, glaciers advance. The flow of glacial ice produces vertical to nearly vertical, wedge-shaped cracks called crevasses (Figure 12.5). These range in size from a few cm to over 10 m in width and up to ~40 m in depth. Glaciers move to lower elevations under the force of gravity by two different processes: internal flow (= plastic flow) and basal slip (= sliding) (see Figure 12.4). The

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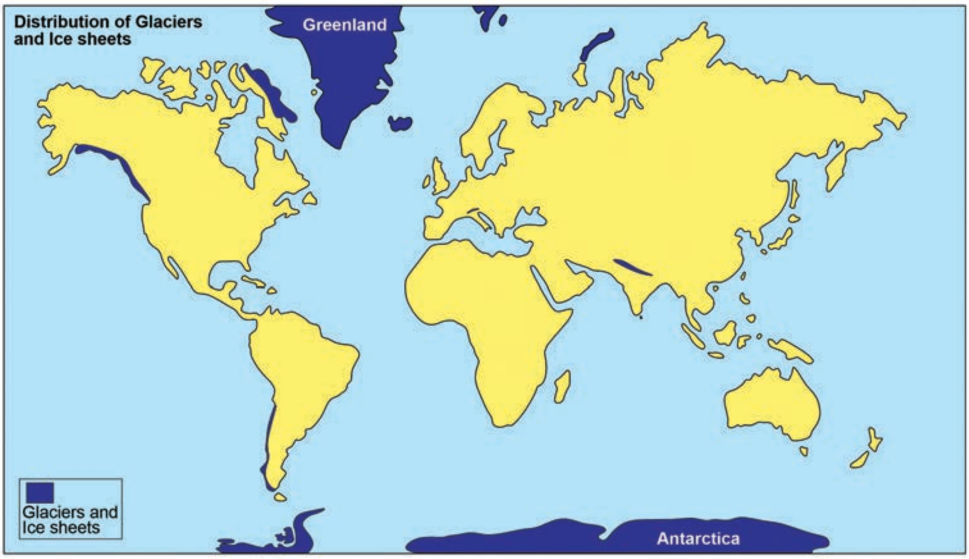

FIGURE 12.1 Distribution of major glaciers. Map shows the global distribution of major glaciers (in blue).

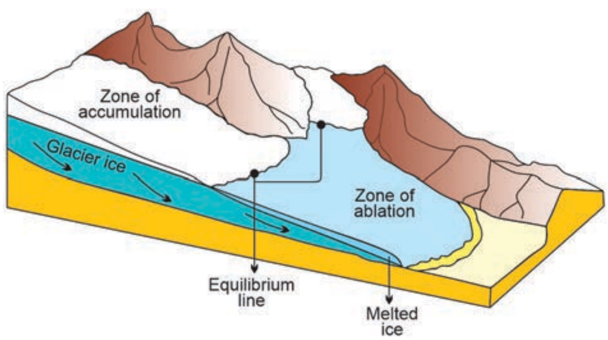

FIGURE 12.2 Zones of accumulation and ablation. Where the accumulation of snow occurs, it is called the accumulation zone, and where glaciers melt, it is called the ablation zone. The equilibrium line marks where accumulation is balanced by ablation over a one-year period.

basal melt water reduces friction by lubricating the surface and thus allows the glacier to slide across (basal slip) (Figure 12.4). The ice crystals slide over each other, i.e., there is displacement between them. The upper portion of a glacier is brittle; hence, fractures (or cracks) form called crevasses (Figure 12.4). Often, these occur when the lower portion of a glacier flows over a sudden change in topography or when accumulating snow and ice reaches a critical thickness of about 40 m; this increased stress induces internal flow (i.e., plastic flow) (Figure 12.4). Plastic flow is the dominant form of movement where all parts of the glacier are below the freezing point, including the base.

12.4 GLACIER TYPES

The classification of glaciers is based on their size and their relationship to topography. The smallest glaciers are confined to mountain valleys and are called valley glaciers or alpine glaciers (see Figure 12.5). Larger masses of ice may cover an entire mountain range or a volcano, and cover

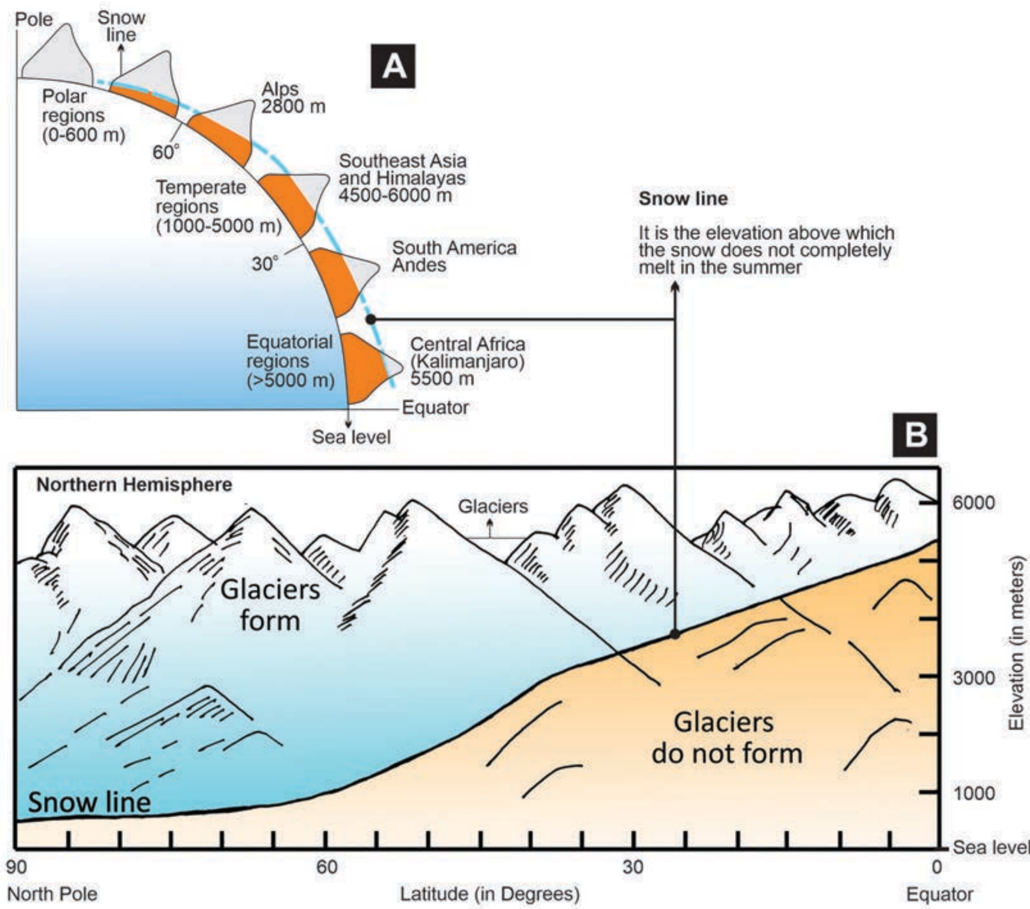

FIGURE 12.3 Snowline. A: Snowline is the line above which snow forms and remains so round the year. B: The glaciers form only at latitudes or altitudes above the snowline.

several thousand square kilometers; these are called ice sheets (larger ones) and ice caps (smaller ones) (Figure 12.6).

12.5 GLACIER CLASSIFICATION

Internal temperature, location, and size are used to classify glaciers. Based on this, two types of glaciers are noted: temperate and polar. In temperate glaciers, the ice is at a temperature near its melting point, whereas in polar glaciers, the ice always maintains a temperature well below its melting point.

Glaciers, based on their location and size, are also of two types: alpine (mountain/valley) and continental; these are further subdivided (see Figure 12.5). The alpine (mountain/valley) glaciers are much smaller than the continental ones, and originate on a mountain or in a mountain range (Figure 12.5). Broadly the alpine glaciers are of four types: cirque, valley, piedmont, and fjord (see Figure 12.5). Cirque glaciers are formed in a depression, occurring usually at the head of a valley, and occupy bowl-shaped depressions on the sides of mountains (Figure 12.5A). As the cirque glaciers grow larger, they spread out into the valley and flow down it as a valley glacier (Figure 12.5A). If a valley glacier extends down a valley and covers a gentle slope beyond the

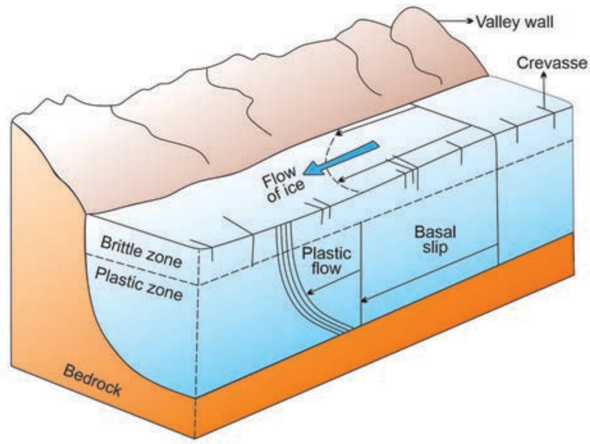

FIGURE 12.4 Basal slip (sliding) and plastic flow: the two mechanisms of glacier movement. Basal slip is sliding over the underlying surface. Plastic flow involves internal deformation within the ice. If a glacier is frozen to its bed, it moves only by plastic flow. The top of the glacier moves farther in a given time than the bottom does.

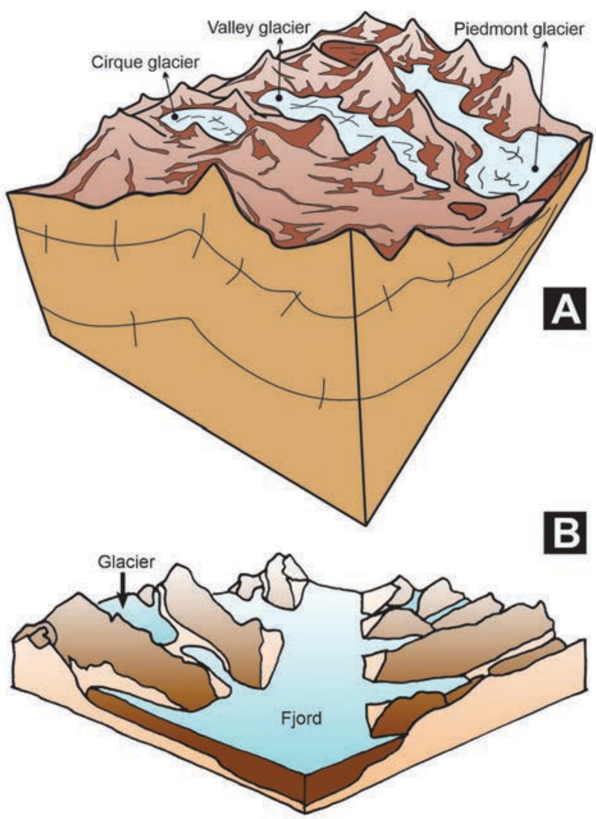

FIGURE 12.5 Valley glacier and its landforms. A: Types of valley glaciers. B: Fjord glacier. (Modified from Jain, 2014.)

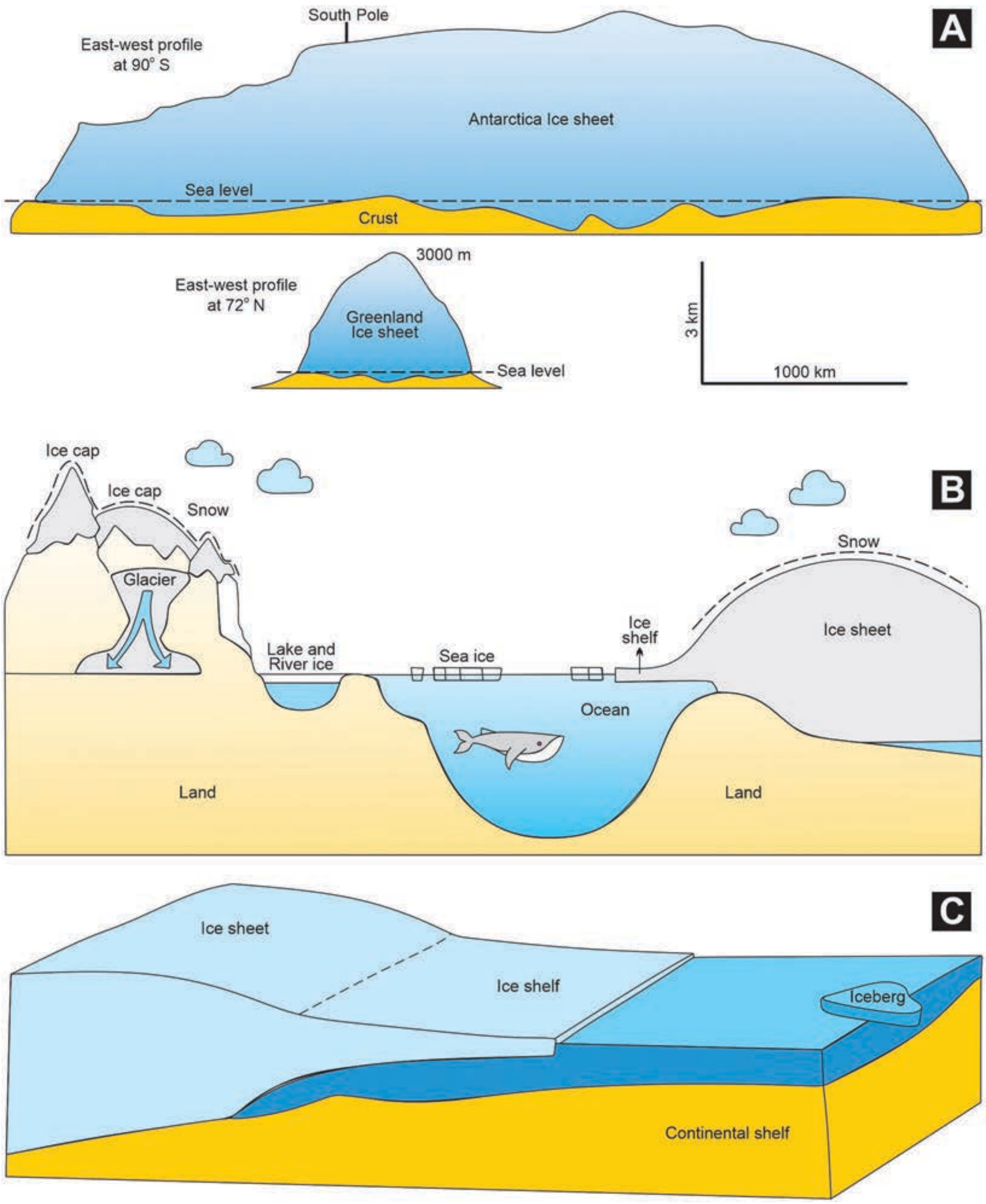

FIGURE 12.6 Continental glaciers. A: Comparative account of Antarctic and Greenland ice sheets. B: Spacial distribution of ice sheets and ice caps. C: Relative position of ice sheet and ice shelf.

mountain range, it becomes a piedmont glacier (Figure 12.5A); this is often formed by two or more coalescing alpine glaciers when entering a flatter area. Fjord glaciers are valley glaciers that extend down to the sea level, and carve a narrow valley into the coastline (Figure 12.5B).

12.5.1 Continental Glaciers

Continental glaciers (such as Greenland and Antarctica; see also Figure 12.1) have 99% of the world's ice and about three-fourths of earth's freshwater resources; they may be hundreds to thousands of meters thick (Figure 12.6A). The Greenland sheet (see also Figure 12.1B), at places, is >2.7 km thick and covers ~1.8 million sq. km, whereas the Antarctic ice sheet blankets ~13 million sq. km (Figure 12.6A). The continental glaciers, based on their size, are of two types: ice sheets (larger ones) and ice caps (smaller ones). An ice sheet covers more than 50,000 km^2 of land area, and may cover either an entire land mass or a continent, such as Antarctica (Figure 12.6A–B; see also Figure 12.1C). Ice sheets cover everything in the landscape, except at the margins, where they are relatively thin (Figure 12.6A). The Antarctic and Greenland ice sheets (Figure 12.6A–B; see also Figure 12.1) are the only existing ice sheets today and together make up ~99% of all the glacial ice. An ice cap covers less than 50,000 km^2 of land area and is not constrained by topographic features; thus, it can lie on top of either a mountain or a mountain range (Figure 12.6B). Ice caps are of two types: confluent glaciers (they merge into two or more glaciers), and outlet glaciers (they drain an ice cap or an ice sheet).

12.5.2 Ice Shelves

Ice shelves are sheets of ice floating on water and attached to land (Figure 12.6C). They usually occupy coastal embayments, and extend hundreds of kilometers from land and reach a thickness of ~1,000 m. The world's largest ice shelves are the Ross Ice Shelf and the Filchner-Ronne Ice Shelf in Antarctica (Figure 12.6C). An iceberg is a piece of ice that has broken from an ice shelf and that floats in the ocean (Figure 12.6C).

12.6 GLACIAL PROCESSES

Abrasion, plucking, and freezing-on are the mechanisms by which rocks and sediments are eroded and incorporated into a glacier (Figure 12.7). Abrasion happens when debris-laden ice slides over the bedrock and abrades it particle-by-particle, like sandpaper on a block of wood, and produces large amounts of silt-sized (0.002–0.0625 mm) sediments (Figure 12.7A). When the ice flows into or refreezes in fractures within the bedrock, plucking occurs (Figure 12.7A). Through this process, ice wedges, and removes blocks of bedrock and incorporates it into the glacier. Freezing-on usually occurs in the leeward (down side) of the bedrock and/or in the presence of other obstacles to the ice flow (Figure 12.7B). Glacier erosion is most rapid when (a) ice flows at high velocity, (b) the material is freezing to the bed, and (c) there is plenty of water. In most glaciers, this usually occurs downstream and below the equilibrium line (see Figure 12.2).

Glaciers that reach the ocean or a lake, calve off into large icebergs which then float on the water surface (as sea ice; see Figure 12.6B) until they melt. Upon melting, the rock debris is deposited on the seafloor or lakebed as unsorted chaotic deposits called dropstones (Figure 12.8). These sediments that are deposited on top of a glacier, and are later reworked by water and become a landslide, are called supraglacial sediments. Gravel, sand, and silt sorted by water and deposited by streams in front of the glacier is called an outwash or stratified drift (Figure 12.8). Outwash is meltwater-transported, and consists mainly of rounded sand and gravel, organized into distinct horizontal layers or bands. The unstratified drift (also called till), on the other hand, is ice-transported/deposited and not organized into distinct layers; it is highly variable, with particles ranging from clay to boulder-size (Figure12. 8B). Some of the finer outwash material is picked up by the wind and later deposited as loess. Both stratified and unstratified drifts are detailed later in the chapter.

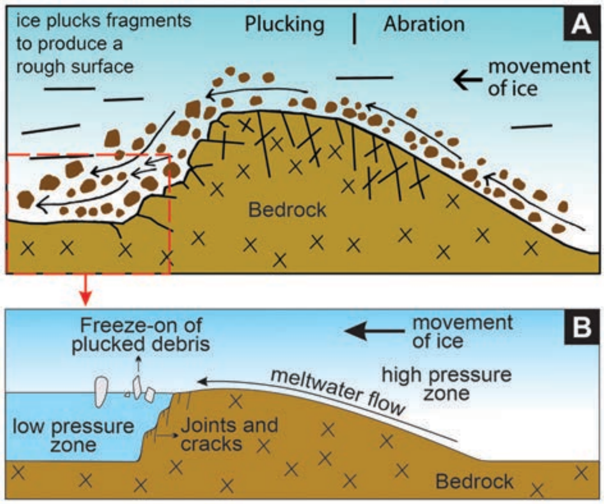

FIGURE 12.7 Abrasion, plucking, freeze-on, and the movement of ice. A: The processes of abrasion and plucking noted in a glacial environment. B: Freeze-on of the plucked debris.

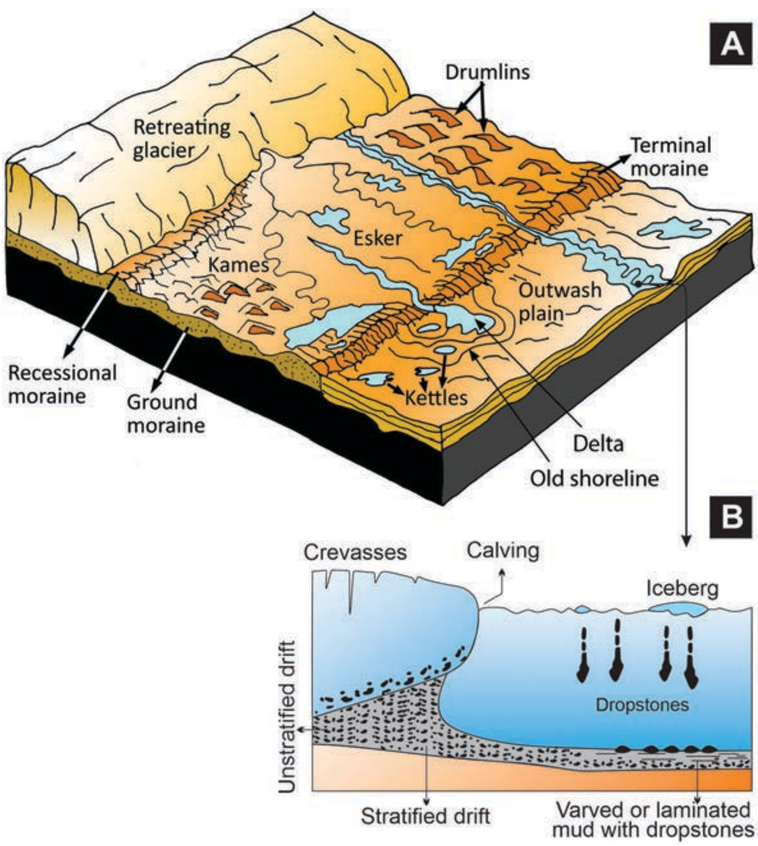

FIGURE 12.8 Dropstones and glacial drifts.

12.7 EROSIONAL LANDFORMS

Erosional landforms are formed by two types of glaciers, alpine/valley glaciers and by continental ice sheets and ice caps. In general, glacial erosion on mountains produces one of the most characteristic features, the U-shaped valley (Figure 12.9A). Other erosional features include horns, arêtes, cirques, hanging valleys, tarns, fjords, and roches moutonnées (Figure 12.9). In addition to these are glacial striations and polish**.** All these are enumerated below.

Horns are formed when three or more cirques are steeply carved out of a mountain producing a sharp peak (Figures 12.9B). The Matterhorn in the Alps is the most famous horn. Cirques are bowl-shaped depressions that occur at the heads of mountain glaciers resulting from a combination of frost wedging, glacial plucking, and abrasion (Figure 12.9C). Cirques are features where the side of the mountain has been scooped out, as if a giant ice-cream scoop has been taken out (Figure 12.9C). Arêtes are groups of horns in a row (Figure 12.9D). If two adjacent valleys are filled with glacial ice, the ridges between the valleys are carved into a sharp knife-edge ridge; this is called an arête (Figure 12.9D). Hanging valleys are formed where two or more glaciers or former valleys intersect at different elevations (Figure 12.9E). When a glacier occupying a smaller tributary valley meets the larger one, the tributary glacier usually does not have the ability to erode its base to the floor of the main valley. Thus, when the glacial ice melts the floor, the tributary valley hangs above the floor of the main valley and is then called a hanging valley (Figure 12.9E). Waterfalls occur where the hanging valley meets the main valley (Figure 12.9E). Tarns are small lakes that occur at the bottom of a cirque (Figure 12.9A).

Roches moutonnées are eroded and smoothed bedrock hills (Figure 12.10A). The gentle side faces up ice and has been smoothed by glacial abrasion. The down-ice side has a steep plucked face,

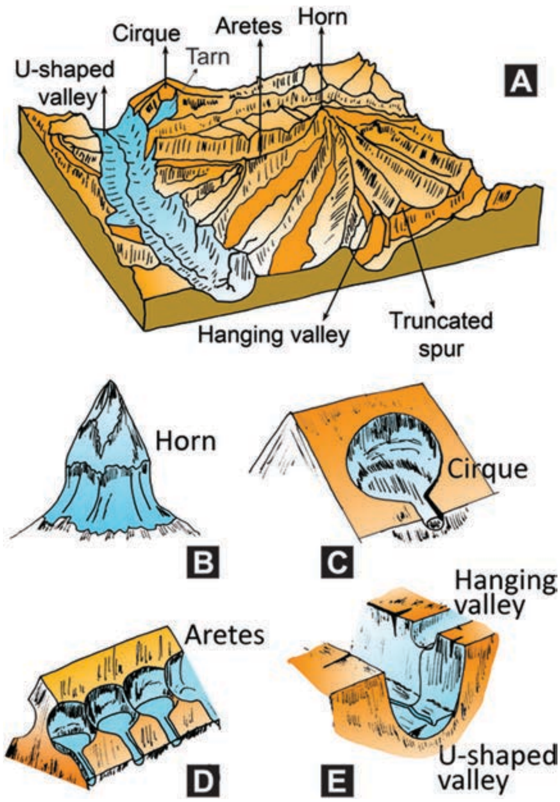

FIGURE 12.9 Alpine glacier landforms. A: Characteristic landforms of alpine glaciers. B–E: Details of alpine glacier landforms. B: Horn. C: Cirque. D: Arête. E: U-shaped valley (see text for further explanation).

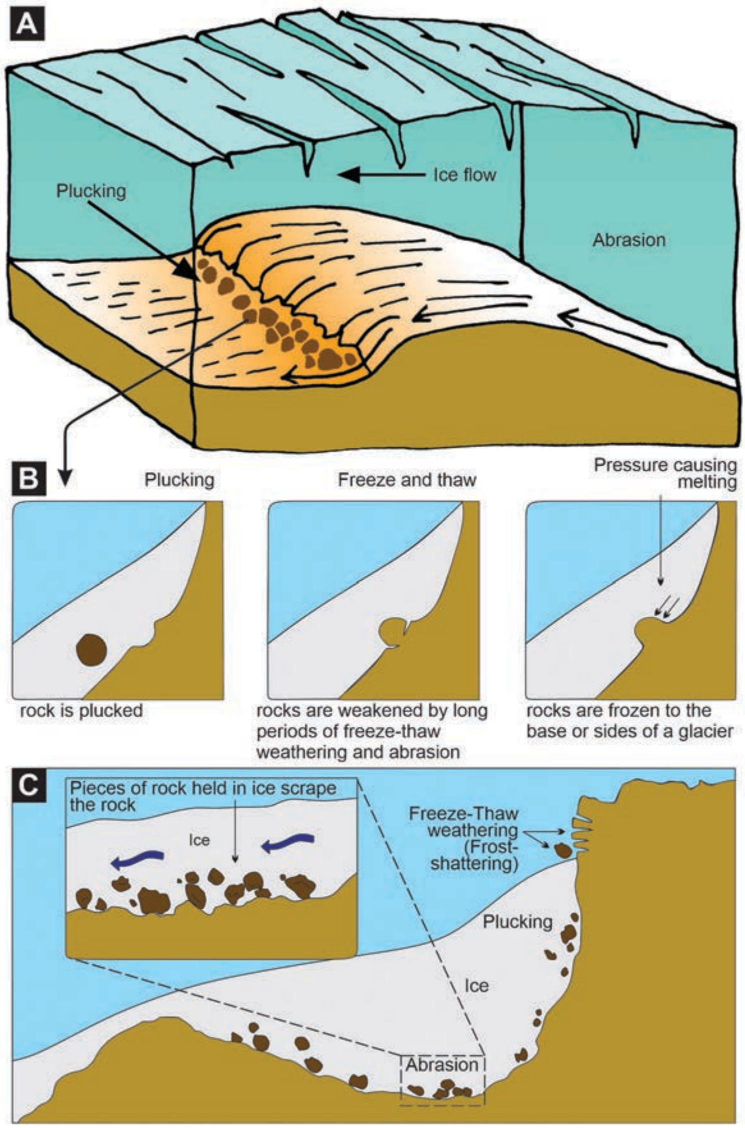

FIGURE 12.10 Roche moutonnée, plucking and abrasion. A: Roche moutonnée. B: The process of plucking. C: Abrasion.

where pieces of bedrock have been plucked out by the glacier ice (Figure 12.10B–C). Hence, roches moutonnées are smooth upstream, with rough and steep downstream resistant rocks (Figure 12.10). Both alpine and continental glaciers form roches moutonnées. Fjords are narrow inlets along the seacoast that were once occupied by a valley glacier (Figure 12.5B).

Glacial striations are long, deep, parallel scratches and grooves made at the bottom of glaciers by rocks embedded in the ice that scrap against the bedrock. These erosional features enable the direction of ice movement (i.e., the flow directions of glaciers) to be inferred. Glacial polish are small-scale features where the glacier acts like sandpaper on the underlying surface, resulting in a smooth surface. In temperate environments, beneath ice caps and ice sheets, both these abrasional features, striations and glacial polish, are noted.

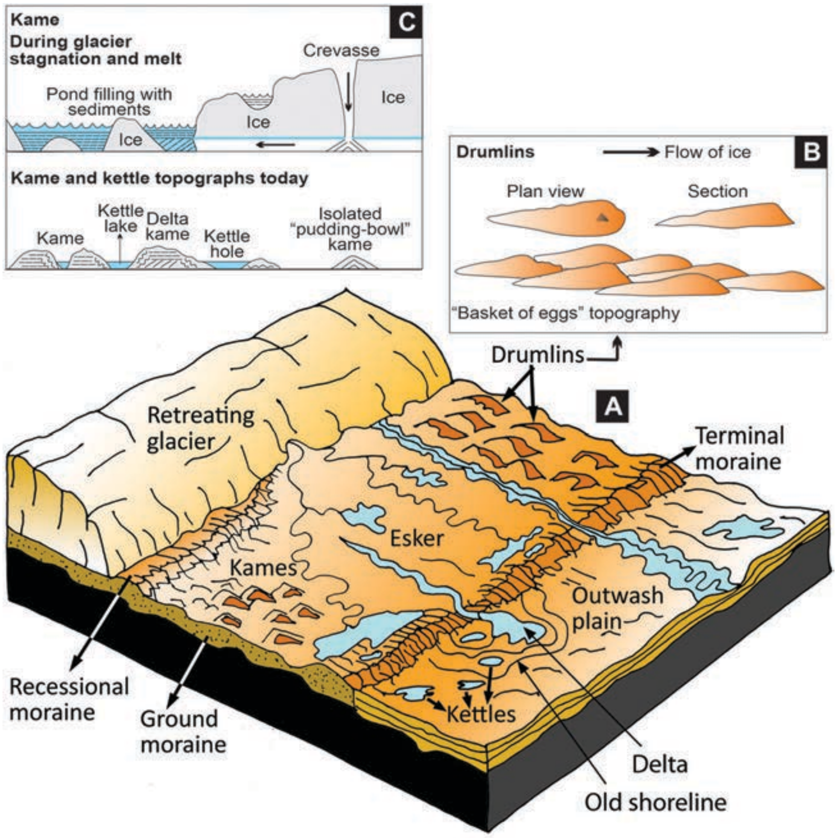

FIGURE 12.11 Drift deposits. A: Types of drift deposits, such as kettles and kettle lakes, kames, eskers, and outwash plains. B: Drumlins (see also Figure 12.8A for its special position within the glacial environment). C: Formation of kames and kettle lakes.

The land surface beneath a moving continental ice sheet is transformed into smooth streamlined elongated hills called drumlins (Figure 12.11A–B). These form when a glacier flows over a mound of sediment and the flow of the ice creates the streamlined shape, elongated in the same direction as the glacial flow (Figure 12.11A–B). Drumlins are usually about 1–2 km long and about 15–50 m high. Most are made of till and some partly of bedrock.

12.8 DEPOSITIONAL LANDFORMS

There are two types of depositional landforms: till (unstratified drift) and stratified drift deposits (Figure 12.8).

12.8.1 Till (Unstratified Drift) Deposits

Till is a non-sorted glacial drift and is made up of a random mixture of different-sized fragments of angular rocks in a matrix of fine-grained, sand-to clay-sized fragments, produced by abrasion within the glacier (Figure 12.12). A till that has undergone diagenesis and has become indurated is called a tillite.

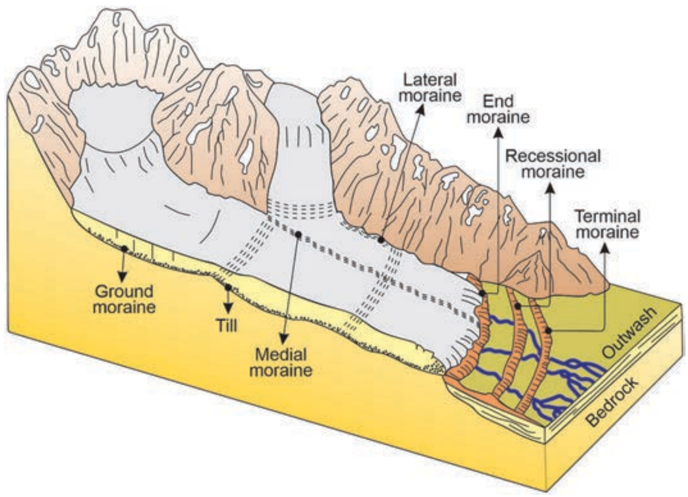

FIGURE 12.12 Depositional glacial landforms, showing the moraine types.

Moraines are the glacier-formed accumulations, i.e., sediments transported by a glacier and then deposited (Figure 12.12). These are till deposits and are different from the underlying bedrock. Depending on their position of formation in relation to the glacier, there are ground, end, terminal and recessional, lateral, and medial moraines (Figure 12.12). The ground moraines are deposited beneath the glacier; the material is dragged under the base of the glacier and deposited over a wide area on the valley floor (Figure 12.12). End, terminal, and recessional moraines are piles of material dropped by the glacier at its terminus/end (Figure 12.12). Hence, their presence marks the furthest advancement of a glacier. If there are warmer conditions, then a glacier recedes. If the glacier stabilizes again during its retreat and the terminus remains in the same place for a year or more, a new end moraine, called a recessional moraine, is formed (Figure 12.12). Thus, a recessional moraine is a ridge of moraine marking a temporary halt in the general retreat of a glacier, whereas a terminal moraine marks the furthest extent of the glacier (Figure 12.12). Lateral moraines are deposited along the sides of mountain glaciers (Figure 12.12.). Thus, they are a ridge of till that builds up at the edge of a glacier. They are typically made up of weathered material from the valley sides that has fallen onto the glacier. Medial moraines occur when two valley glaciers meet to form a larger glacier, the rock debris along the sides of both glaciers merges to form a medial moraine (Figure 12.12). These occur as black streaks in an active glacier.

Erratics are large boulders transported by glaciers (sometimes up to hundreds of miles), and often deposited at a considerable distance from their origin (Figure 12.13). These are left behind when the ice melts, and thus do not conform to the geology of the bedrock, i.e., surrounding landscape (Figure 12.13).

12.8.2 Stratified Drift Deposits

Stratified drift deposits are sediments deposited by glacial meltwater that are sorted and layered. These include kettle and kettle lakes, kames and kame terraces, eskers, and outwash plains and terraces (see Figure 12.11). Kettles are depressions that form underneath a glacier and remain so after the glacier has melted; in these depressions, water fills and they become small lakes (Figures 12.11A and 12.11C). In these lakes, fine-grained sediments are deposited. Melted chunks of ice left by a retreating glacier forms kettle lakes (Figures 12.11A and 12.11C).

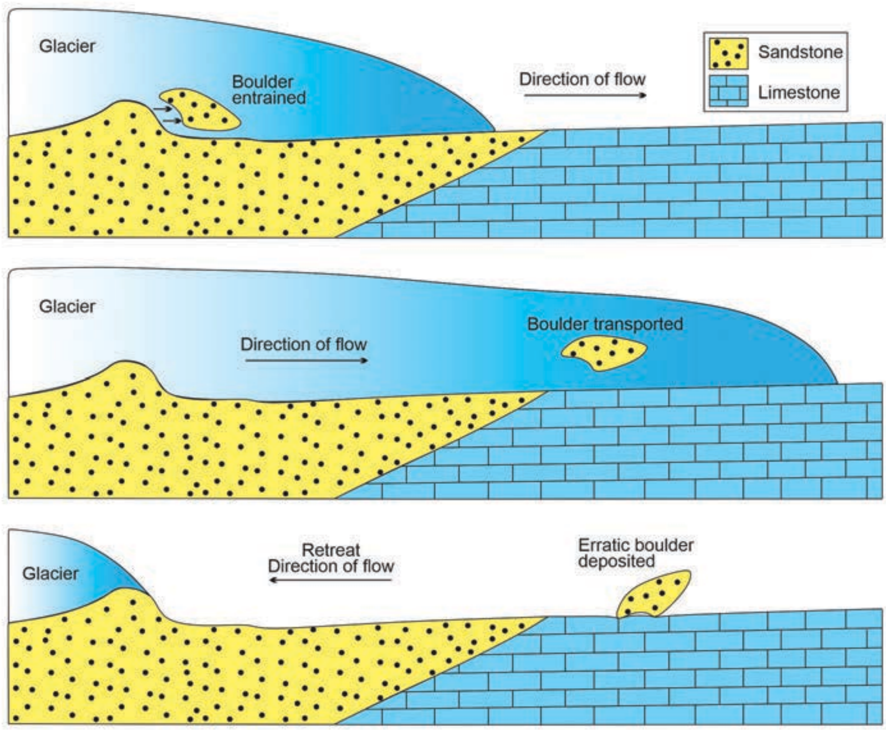

FIGURE 12.13 Development of an erratic.

Streams and lakes formed on top of stagnant ice may deposit stratified sediments on top of a glacier. When the glacier melts these deposits are set down on the ground surface. These former lake deposits become kames, and former stream deposits become kame terraces (Figure 12.11C). Thus, kames are hills of sediment that are commonly found near kettles (Figures 12.11A and 12.11C).

Eskers (Figure 12.11A) are long sinuous ridges of sediment (sand and gravel) deposited by streams that run under or within a glacier (Figure 12.11A). These stream-deposited sediments become eskers after the ice has melted. Eskers are often broad and flat-topped, or may have a single crest or they may split into parallel ridges. Eskers are also sinuous, with a height only of a few to several tens of meters, although some are known to continue for several kilometers; most are shorter or discontinuous. They are often formed at the margin of glaciers or ice sheets during their retreat.

Outwash plains form in front of a glacier as the meltwater carries silt, sand, and gravel away from it (Figure 12.11A). Streams running off the end of a melting glacier are usually choked with sediment and form braided streams (i.e., flowing in multiple channels), which deposit poorly sorted stratified sediment in an outwash plain (see Figure 12.11A). These silt-rich deposits are often referred to as outwash (see Figure 12.11A). These deposits are reworked by wind to form loess deposits. Outwash plains are characteristics of continental glaciers. If the outwash streams cut down into their outwash deposits, the banks form river terraces called outwash terraces.

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Section IVb

Depositional Systems: Marginal Marine Environments

The marginal marine environments lie between continental and marine depositional settings (Figure IVb.1) and are characterized by a narrow zone, dominated by river, wave, and tidal processes (Figure IVb.2) (see also Boyd et al., 1992; Dalrymple et al., 1992). In these settings, the salinities vary in different parts of this system, ranging from fresh to brackish to supersaline, depending upon river discharge (the dominant force) and climatic conditions (Figure IVb.2). Some marginal marine environments are characterized by intermittent to nearly constant subaerial exposure while others are covered by shallow waters. Some are characterized by high-energy waves and currents, whereas some lagoonal and estuarine environments are dominated by quiet-water conditions (Figure IVb.2).

The marginal marine environments also display a wide variety of sediment types such as conglomerates, sandstones, shales, carbonates, and evaporates. The main depositional settings for marginal marine sediments include deltas, beaches, strand plains, barrier bars, estuaries, lagoons, and tidal flats (Figures IVb.1 and IVb.2). Estuaries and lagoons are mostly characterized by transgressive coasts, whereas deltas are features of prograding coasts (Figure IVb.2). In Chapter 13, deltas (including fan deltas) are elaborated, followed by succeeding chapters on beach and barrier-island systems, estuaries, and lagoons (see Figures IVb.1 and IVb.2).

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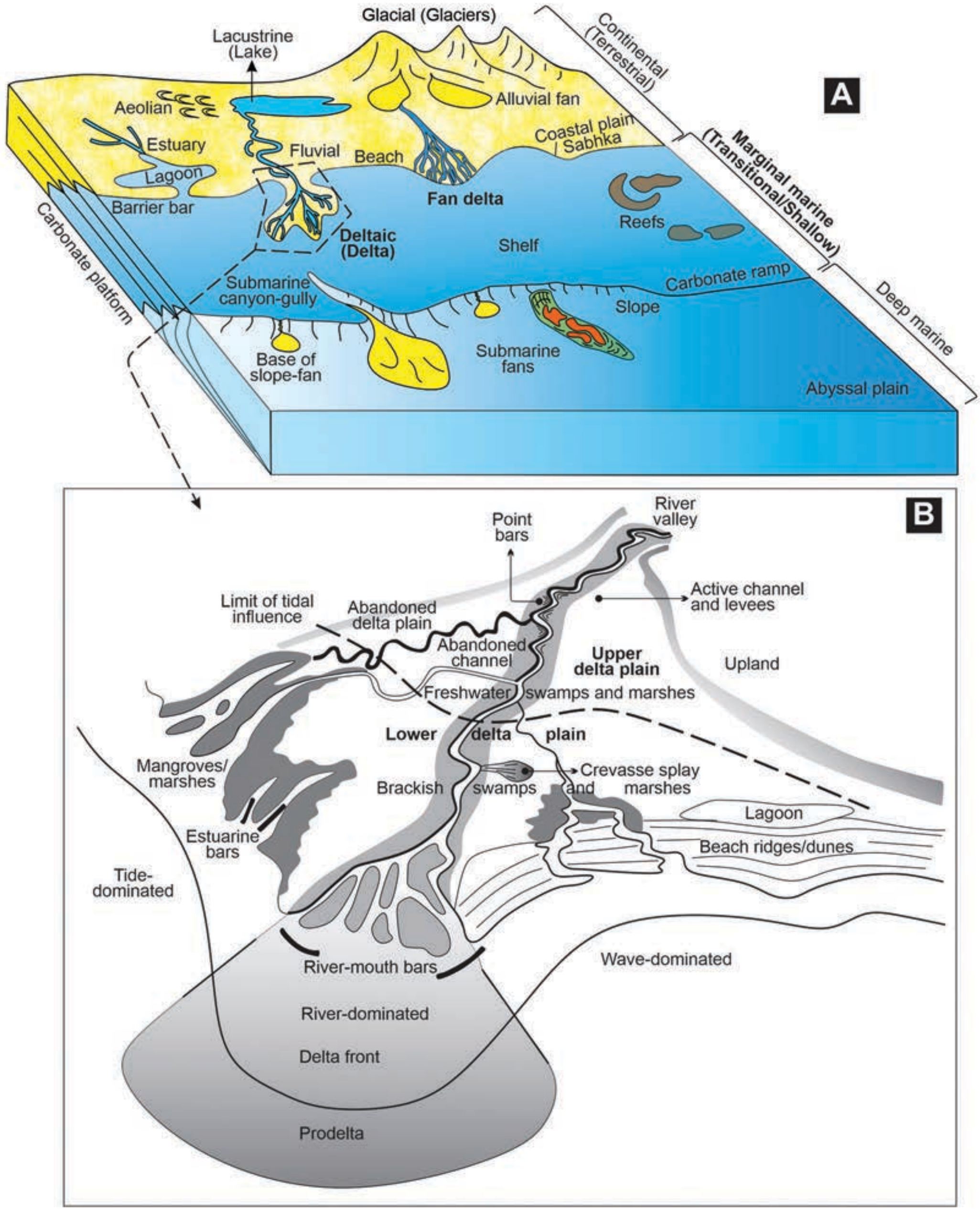

FIGURE IVb.1 Marginal marine environments. A: Delta system. In bold are topics under discussion in this chapter. (Image modified from www.geological-digressions.com/; courtesy Brian Ricketts.) B: Nearshore morphology showing the respective influence of the three domains, river, tides, and waves. (Modified after Anthony, 2015.)

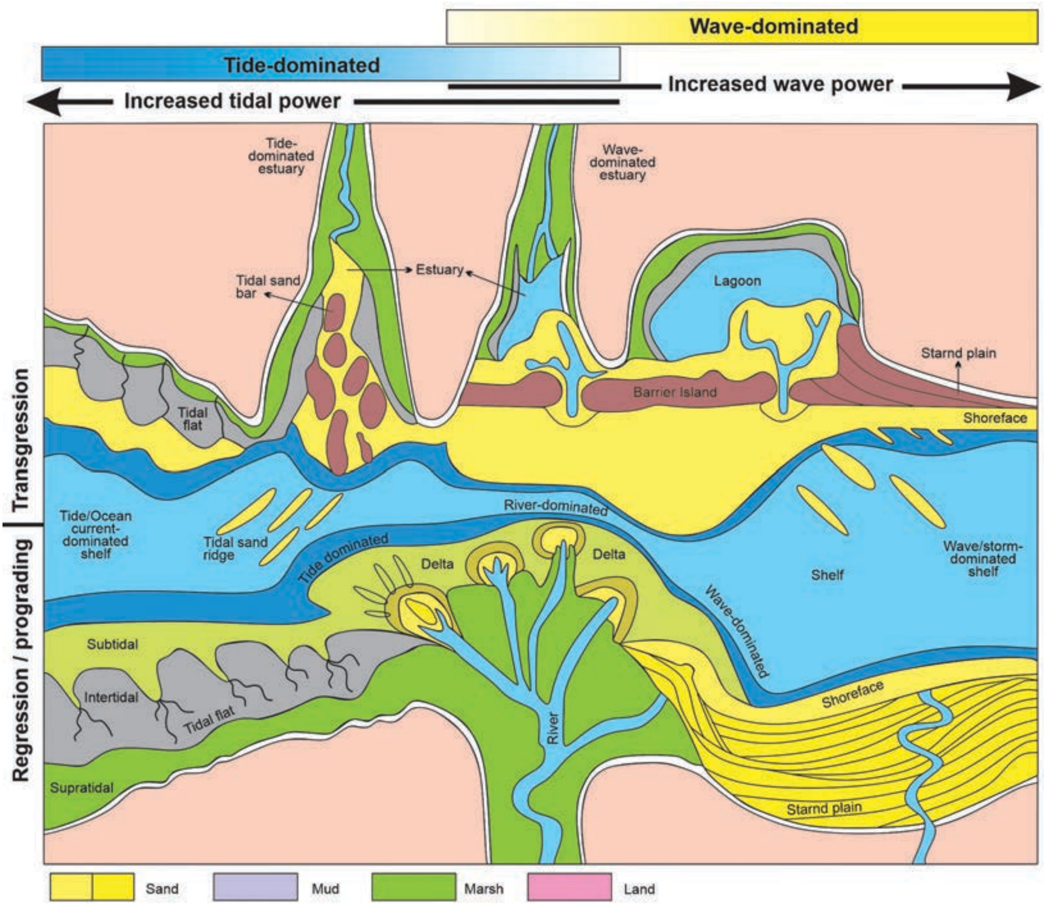

FIGURE IVb.2 Major coastal environments with emphasis on wave- and tide-dominated environments. (Modified after Boyd et al., 1992.)

13 Deltaic System

13.1 INTRODUCTION

Herodotus, the Greek philosopher, in 490 BCE used the word "*delta*" to describe the triangular-shaped alluvial plain formed at the mouth of the Nile River (Egypt) and its distributaries (Figure 13.1). Elliott (1986) described deltas as discrete shoreline protuberances formed where rivers enter oceans, semi-enclosed seas, lakes or lagoons, and where sediment is supplied more rapidly than it can be redistributed by basinal processes (see also Zavala et al., 2021). Thus, deltas can form in lakes and inland seas and most importantly, in the open ocean, as well (see also Figure IVb.1). Deltas are essentially regressive in nature, where their deposits record a seaward migration or progradation of the shoreline (see Figure IVb.2). It must be kept in mind that most sediments in a delta come from rivers in contrast to estuaries that receive much of the sediments from the marine realm and in which deposits are essentially transgressive (see Figure IVb.2). Throughout geologic time, much of the siliciclastic sediments (such as sand, silt and mud) transported to coastal zones are deposited in deltas (see Figure 13.1B). These deltaic sediments are also important hosts for petroleum and natural gas, coal, and some minerals, such as uranium.

In general, deltas form where large, active drainage systems with heavy sediment loads exist as in the trailing-edge coasts (or passive coasts) such as the east coasts of Asia and the Americas where tectonic activity is low (see Figure 13.2). Fewer than 10% of major modern deltas occur in collision coasts, where tectonic activity is high and drainage divides are close to the sea (Figure 13.2) (Wright, 1978). Under such conditions, large drainage systems, necessary to supply heavy sediment loads, are not developed.

In the broadest sense, deltas can be defined as depositional features (subaerial and subaqueous), formed by fluvial sediments. In many instances, the deposition of fluvial sediments is strongly modified by marine forces such as waves, currents, and tides; therefore, the depositional features display a high degree of variability. The depositional features include distributary channels, river-mouth bars, interdistributary bays, tidal flats, tidal ridges, beaches, eolian dunes, swamps, marshes, and evaporite flats.

13.2 COMPONENTS OF A DELTA

There are three components of a deltaic body: (a) delta plain (characterized by the presence of distributary channels); (b) delta front (the steepest part of the system); and (c) prodelta (where fine deposits are accumulated over a sub-horizontal surface; nearly flat) (Figure 13.3). The delta plain is further divided into an upper (steep to gently dipping) and a lower delta plain (nearly flat and with a distributary mouth bar containing coarser sandy deposits) (Figure 13.3A); the prodelta is also

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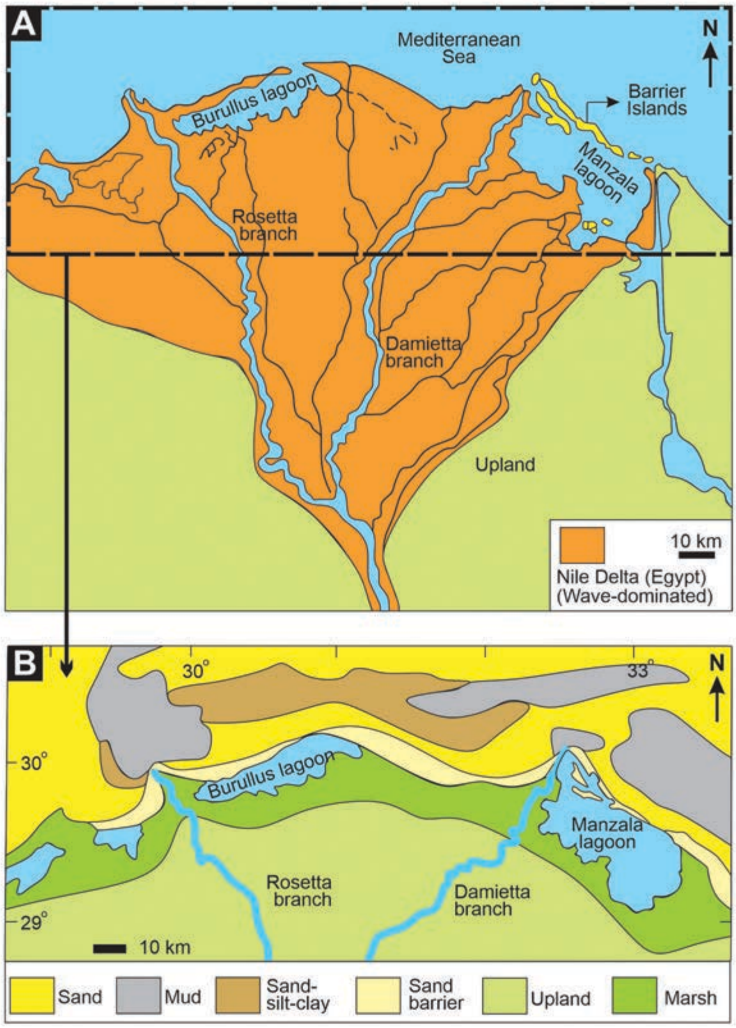

FIGURE 13.1 Nile Delta. A: Wave-dominated (Type 2) delta. B: Sediment distribution patterns in the Nile Delta. (Modified after Coleman, 1981; Sestini, 1989.)

divided into a proximal (characterized by heterolithic bedding; see Figure 13.4) and a distal prodelta with mud (see Figure 13.3). The upper delta plain comprises braided and meandering channels, distributary channels, levee formation, and overbank flooding (see Figure 13.5). The lower delta plain straddles the area of the river-marine interaction and extends landward from the low-tide mark to the limit of tidal influence. It includes bifurcating or anastomosing channels, actively migrating tidal channels, and bay-fill deposits (such as interdistributary bays, crevasse splays, natural levees, and marshes) (Figure 13.5).

13.3 DELTA CLASSIFICATION AND SEDIMENTATION PROCESSES

Deltas are landforms that form at the mouth of a river where it meets a body of water, such as a lake or an ocean. They are characterized by their triangular or fan-shaped appearance and are composed of sediments that are deposited by the river as it enters still waters (see Figure IVb.1).

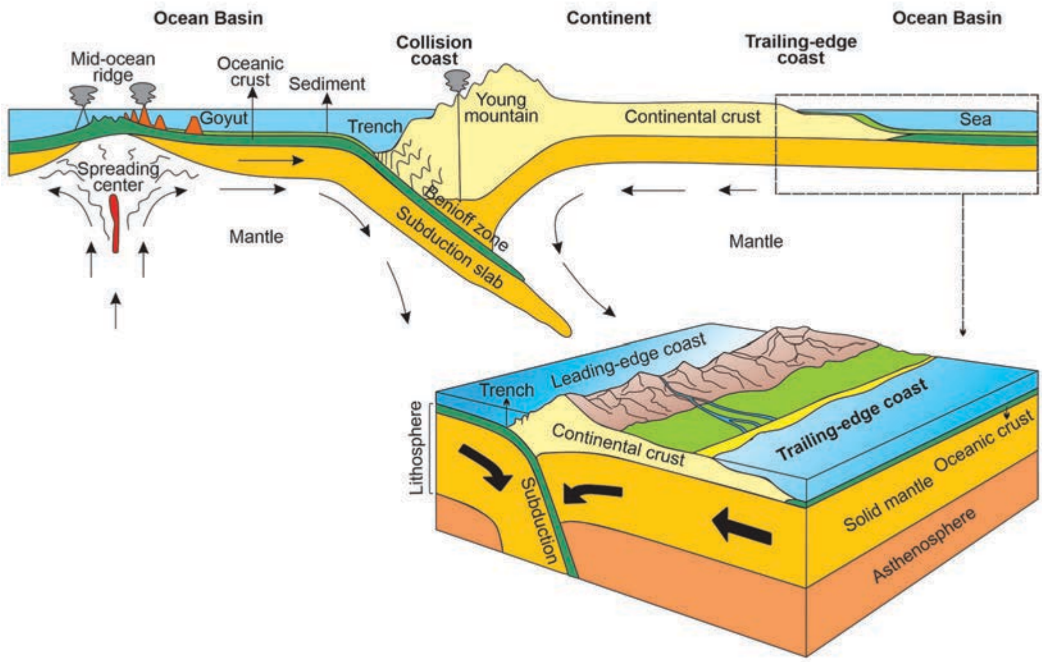

FIGURE 13.2 Plate tectonics and the placement of modern deltas in collision coasts (trailing-edge coast). (Modified after Inman and Nordstrom, 1971.)

Based on sedimentation processes, the deltas are classified as: (a) fluvial (river)-dominated, (b) wave-dominated or (c) tide-dominated (see Figure 13.6). Each of these kinds of deltas are further subdivided on the basis of dominant grain size (see Orton, 1988; Orton and Reading, 1993), i.e., mud/silt, fine sand, gravelly sand, or gravel (Figure 13.6B). River or fluvial (sediment) input, wave-energy flux, and tidal flux are the most important processes that control the geometry, trend, and internal features of the progradational sand bodies of deltas (Galloway and Hobday, 1983). Thus, the distribution and characteristics of deltas are controlled by a complex set of interrelated fluvial and marine/lacustrine processes and environmental conditions. The different types of deltas are briefly summarized below.

13.3.1 Fluvial-Dominated Deltas

The fluvial-dominated deltas, such as the Mississippi Delta in the United States (Figure 13.7), the Ganges-Brahmaputra Delta in Bangladesh and India, and the Nile Delta in Egypt (Figure 13.1), typically have a lobate or bird's foot shape, with distributary channels that branch out and carry sediments into the receiving body of water, a basin. These deltas are marked by high sediment supply from the river that dominates the sedimentation processes and plays a key role in shaping delta morphology and sediment distribution patterns. Thus, the sedimentation patterns in fluvial-dominated deltas are influenced by factors such as river discharge, sediment load, and the rate of sea-level rise. High river discharge and sediment load results in rapid delta growth, whereas low river discharge and sediment load leads to delta retreat or erosion.

The river carries a large amount of sediments (including sand, silt, and clay) and water that is deposited directly at the river mouth. This discharge of river water and sediments into a lake or ocean (= basin) is called a jet (Bates, 1953). The jets (i.e., sediment-laden river water; also referred to as plume) as it enters the basin (lake or ocean) are grouped into three categories: (a) jets that are

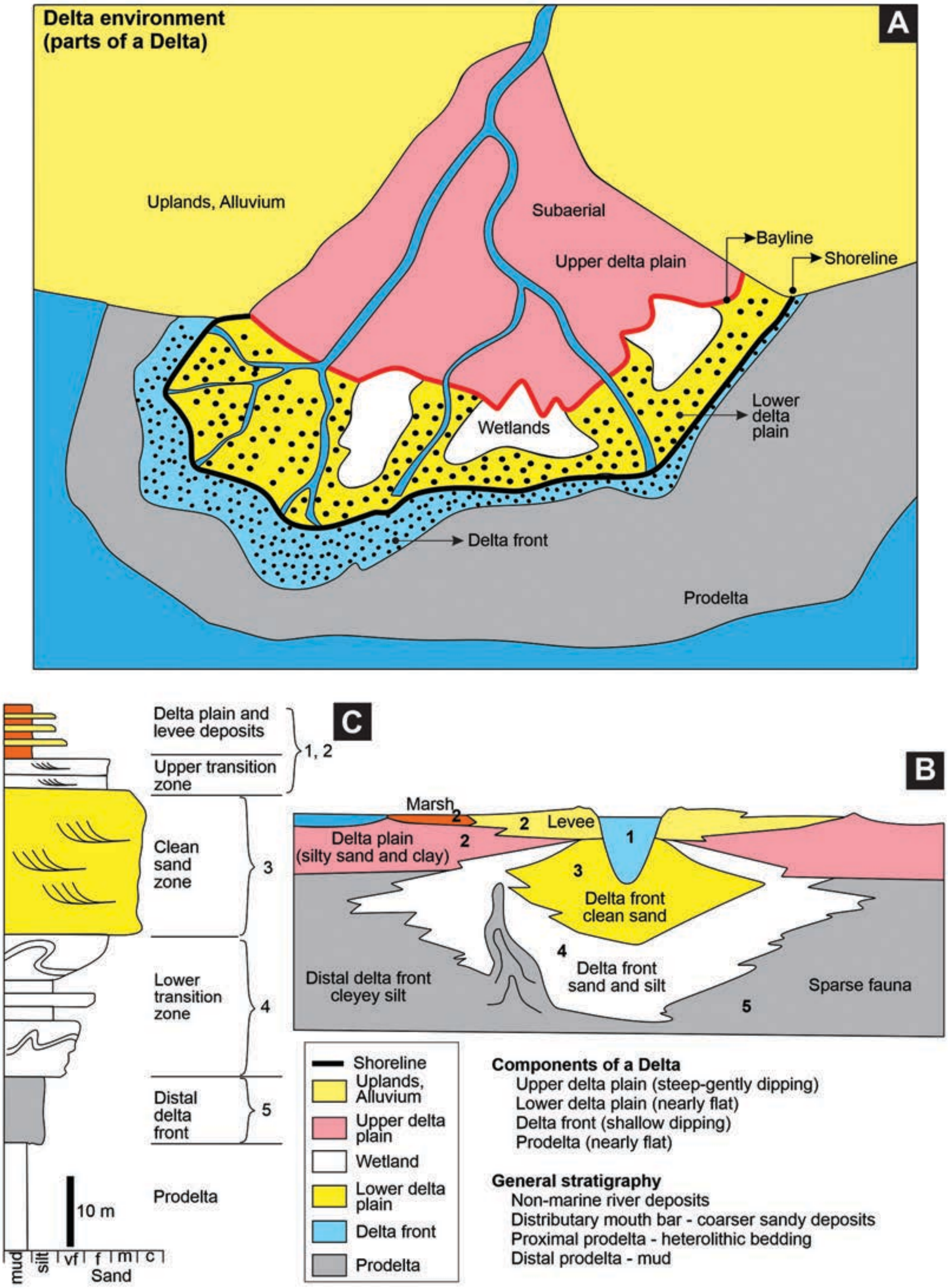

FIGURE 13.3 Components of a delta. A–B: The three components of a deltaic body include the delta plain (with distributary channels), delta front (the steepest part of the system) and a prodelta (fine sediment deposition over a sub-horizontal surface). C: Characteristic delta zone facies.

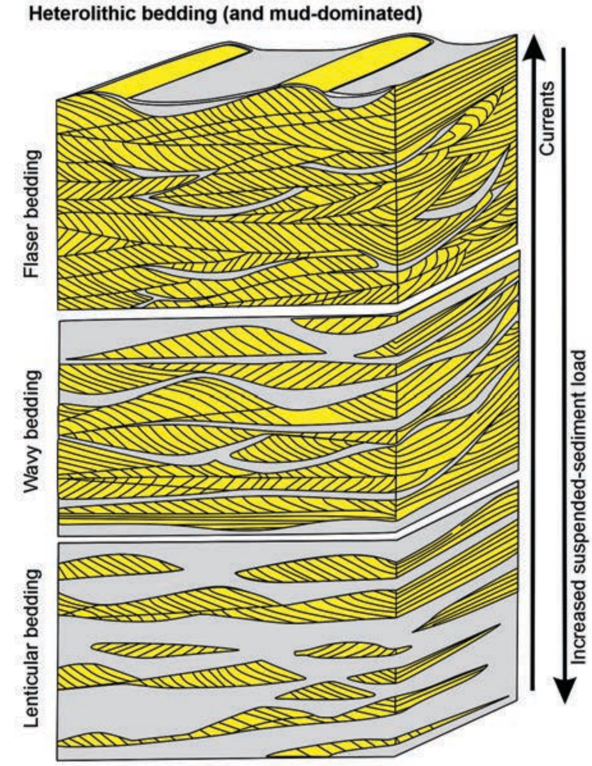

FIGURE 13.4 Heterolithic bedding characteristic of a proximal prodelta.

equally dense as the basin water (called homopycnal flow) (Figure 13.8A); (b) less dense than the basin water (hypopycnal flow) (Figure 13.8B); and (c) denser than the basin water (hyperpycnal flow) (Figure 13.8C) (see Bates, 1953; see also Zavala et al., 2021 for the modification of Bates's classification). These types of flow are briefly summarized below.

The river water entering the basin water is of almost equal density, the homopycnal flow (Figure 13.8A). This results in a rapid, thorough mixing and abrupt deposition of much of the sediment load, deceleration of outflow jet and, hence, the outflow jet travels far into the basin (Figure 13.8A). This is commonly noted at the mouths of coarse-grained rivers, forming Gilbert-type deltas (i.e., inertia-dominated deltas; hyperpycnal flow) (Figure 13.9). These deltas display a topset, foreset, and bottomset arrangement of beds, formed as sediment deposition progrades basinward (Figure 13.9).

If the river outflow is less dense than the basin water, as noted in rivers that flow into denser seawater or a saline lake, it flows outward on top of the basin water as a horizontally oriented plane jet called hypopycnal flow (buoyancy-dominated deltas) (Figure 13.8B). Hence, the lower-density outflow enters the higher-density receiving basin water, such as when fresh water enters into a marine basin; lighter outflow water floats like a buoyant plume above the salt wedge. Fine sediments are carried in suspension outwards to some distance from the river mouth before they flocculate and settle (Figure 13.8B). The hypopycnal flow tends to generate a large, active delta-front area, typically dipping at 1° or less, as compared to 10°-20° dip of most Gilbert-type deltas (see Miall, 1984). In such conditions, sediment deposits extend far in distance from the feeder channel. Lateral bars with subaerial and subaqueous levees are commonly noted. The hypopycnal flow is the most important type of river outflow in marine basins.

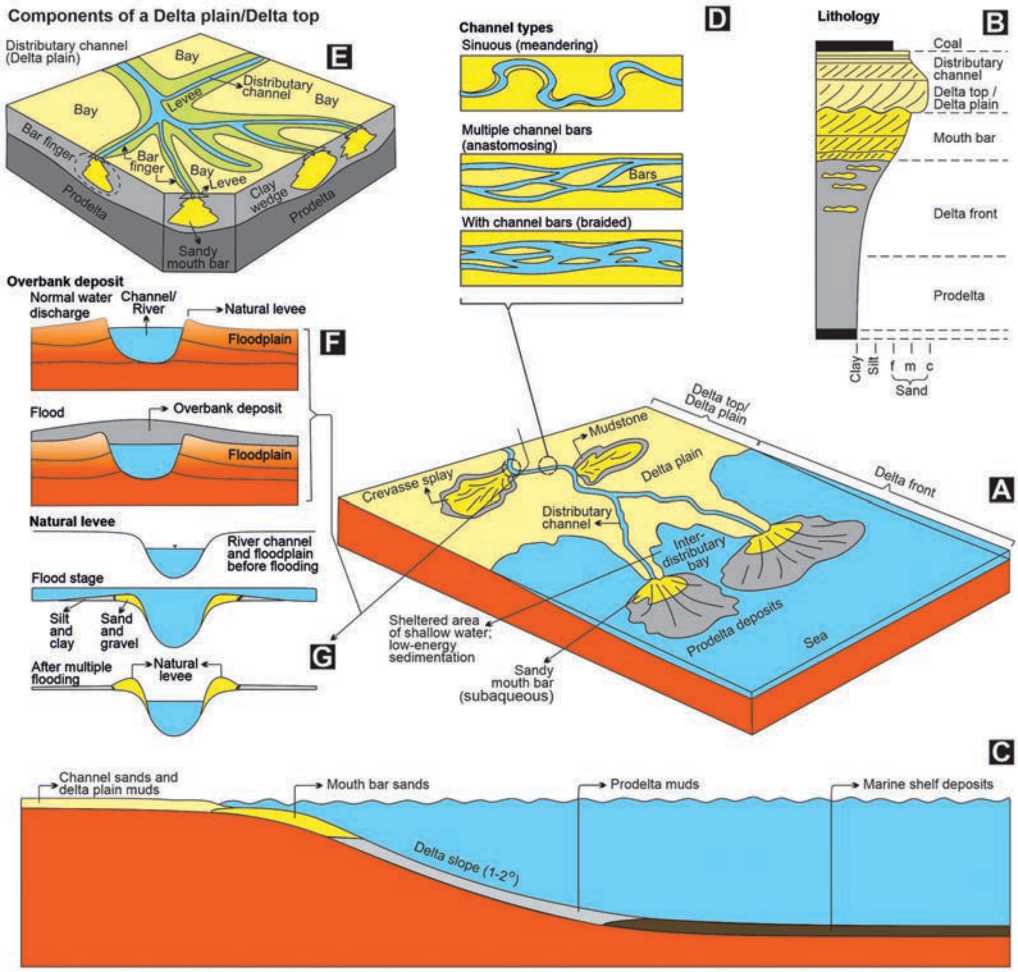

FIGURE 13.5 Delta components and associated sediments. A–C: The two subenvironments of a delta, delta front and delta top/delta plain and associated sediments. Progradation results in a coarsening-up sequence (B). D: Channel types noted in the delta plains. E: Bar-finger and distributary channels in a delta plain. F–G: Formation of overbank deposits and levee formation.

The river water that has a higher density than the basin water flows beneath the basin water, i.e., where higher-density outflow enters into lower-density shallow nearshore waters. The outflow jets move as underflows, get decelerated and spread rapidly and laterally, forming bifurcating channels with middle-ground bars (mouth bars). This is commonly noted during floods, and thus generates a vertically oriented jet flow called the hyperpycnal flow (friction-dominated deltas) (Figure 13.8C). This type of jet flow moves along the bottom as a density current and deposits its load along the gentle slopes of the delta front, forming turbidites (Figure 13.8C).

13.3.1.1 Bar-Finger Sands

In fluvial-dominated deltas, sedimentary structures called bar-finger sands are commonly noted, as in the Mississippi Delta (see Figures 13.7A–B). They are characterized by elongated, finger-like sand bodies that extend perpendicularly to the shoreline or river channel (Figures 13.7A–B). The

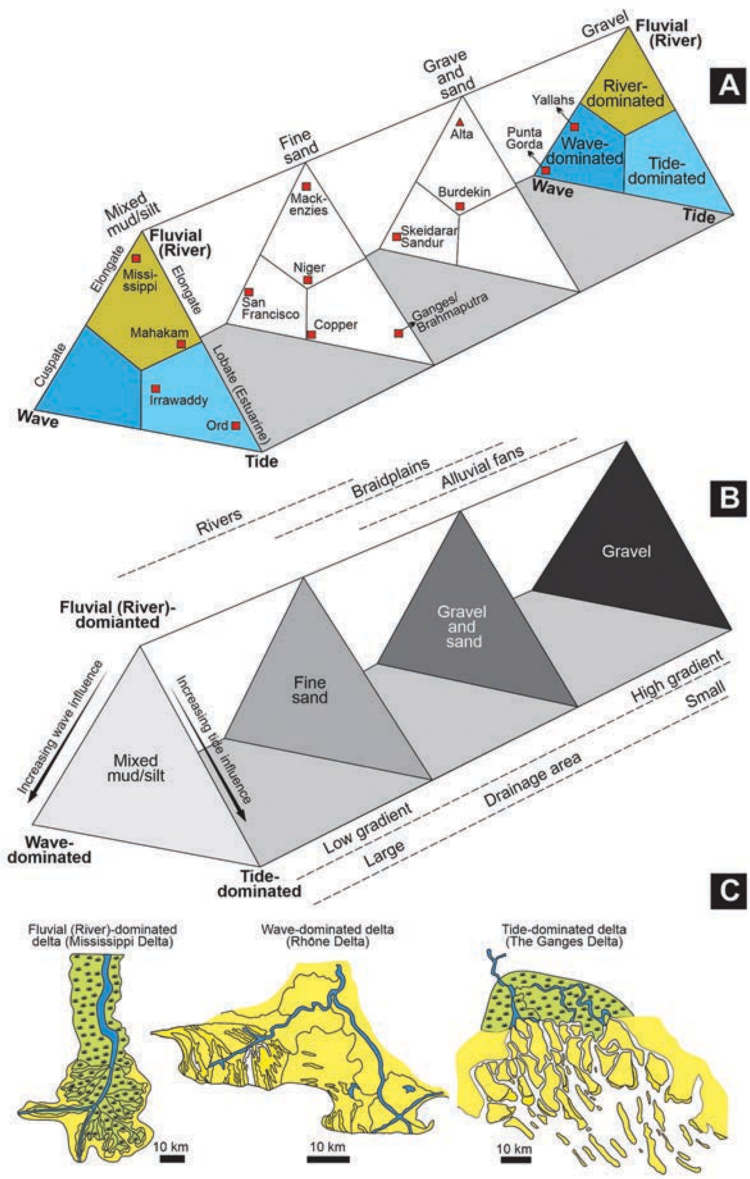

FIGURE 13.6 Delta classification. A–B: Based on sedimentation processes. The deltas are classified as fluvial-, wave-, or tide-dominated. Each of these is further subdivided on the basis of dominant grain size, i.e., mud/silt, fine sand, gravelly sand, and gravel. (Modified after Orton, 1988; Orton and Reading, 1993.) C: Examples of three major delta types.

bar-finger sands form as a result of sediment deposition and migration by water currents. In fluvial environments, they typically develop in meandering river channels or braided river systems (see Figures 13.5D–E). As the river transports sediment, they form bars or elongated sand bodies that protrude into the channel (Figures 13.7A–B). These bars can have a finger-like shape, with narrow, elongated ridges of sand extending across the channel. The formation and preservation of bar-finger sands are influenced by factors such as sediment supply, water currents, waves, and

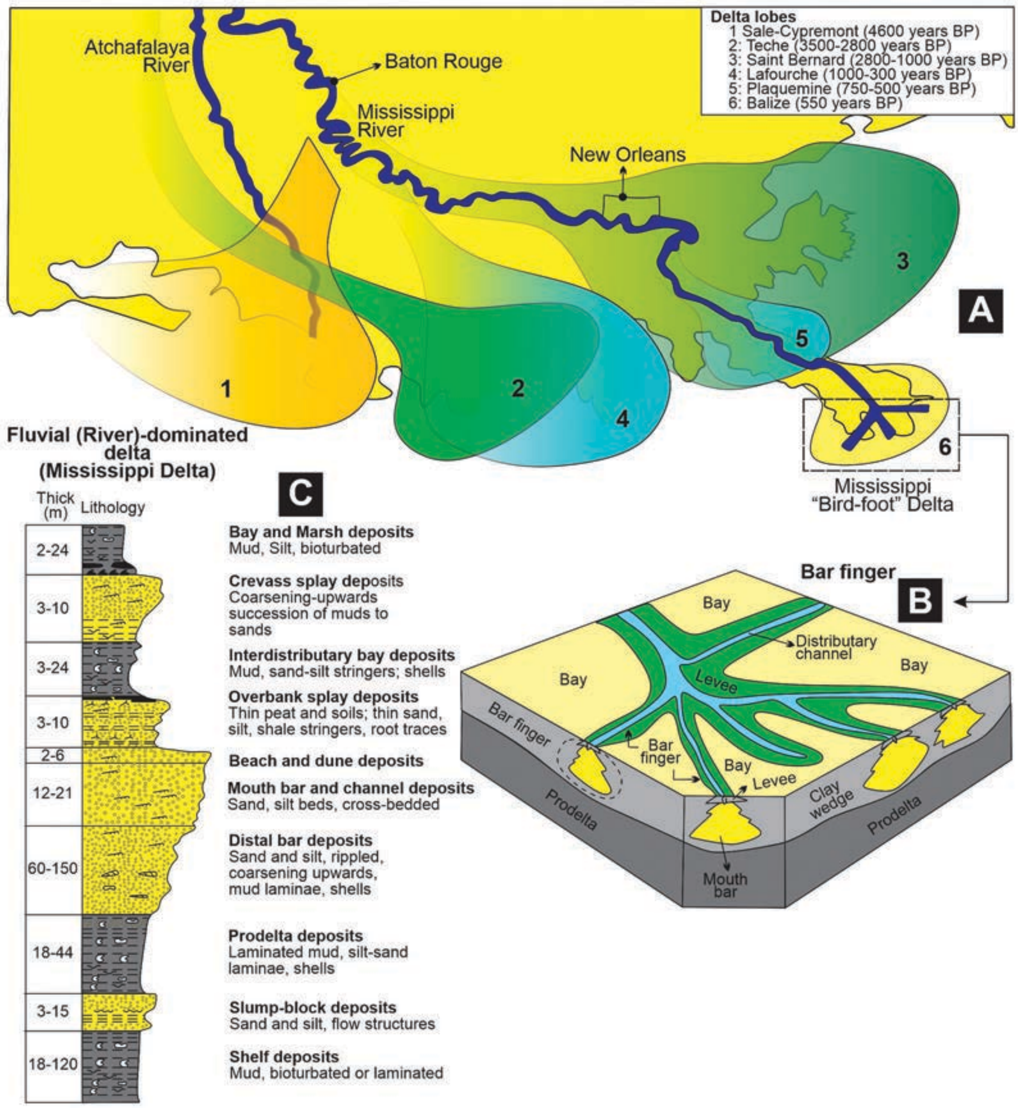

FIGURE 13.7 Mississippi Delta (USA) (fluvial-dominated delta) and bar-finger sands. A: It typically has a lobate or bird's foot shape, with distributary channels that branch out and carry sediments (sand, silt, and clay) into the receiving body of water, a basin. B: Bar-finger sands. These are elongated, finger-like sand bodies that extend perpendicular to the shoreline or river channel and are formed as a result of sediment deposition and migration by water currents. C: Fluvial-dominated delta facies. The common sediments include marsh and natural-levee deposits, delta-front silts and sands, and prodelta clays.

sea-level changes. Thus, they are indicators of a dynamic sedimentary environment and provide insights into past depositional processes and the evolution of fluvial or coastal landforms. In coastal environments, bar-finger sands can form as a result of longshore drift, which is the movement of sediment along the shoreline due to wave action. Waves approach the shoreline at an angle, and as they break, they transport sediments along the coast. This leads to the formation of sandbars or ridges that extend perpendicular to the shoreline, creating finger-like appearances.

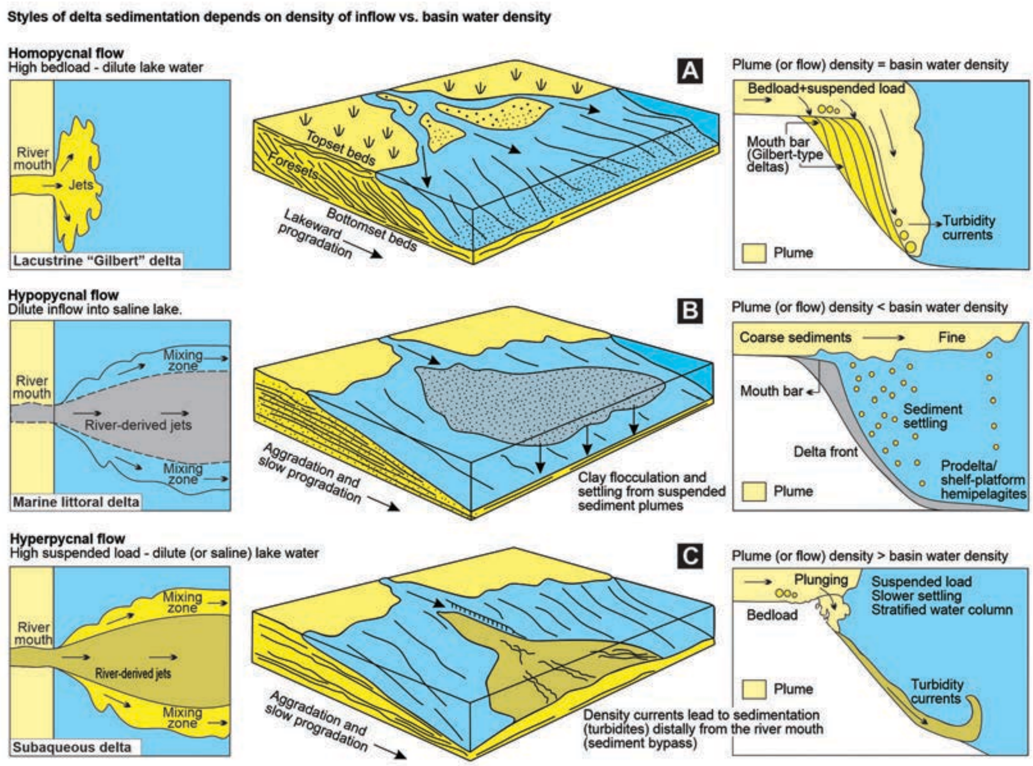

FIGURE 13.8 Styles of delta sedimentation. (Modified after Bates, 1953; Zavala et al., 2021.) The river carries a large amount of sediment (including sand, silt, and clay) and water that is deposited directly at the river mouth. This discharge of river water and sediments into a lake or ocean is called a jet. These jets as they enter the basin (lake or ocean) are of three types. A: Jets that are equally dense as the basin water (Homopycnal flow). B: Jets that are less dense than the basin water (hypopycnal flow). C: Jets that are denser than the basin water (hyperpycnal flow).

The modern Mississippi Delta is a classic example of a birdsfoottype delta and the formation of bar-finger sands (Figure 13.7). The Mississippi Delta consists of distinct sedimentary lobes that have been active during the past 5,000–6,000 years, suggesting that periodic channel or distributary abandonment is a common process (Figure 13.7A). The Mississippi Delta system displays a well-developed birdsfoot distributary system, typical of the delta, with bar-finger sands developed at the mouths of the distributaries (Figure 13.7B). Common sediment include marsh and natural-levee deposits, delta-front silts and sands, and prodelta clays (Figure 13.7C). Other modern deltas that are largely fluvial-dominated include the Mackenzie Delta (Canada, Beaufort Sea) and the Alta delta (Norway, Alta Fjord).

13.3.2 Wave-Dominated Deltas

In wave-dominated deltas (see Figure IVb.2), wave energy controls sediment distribution and delta morphology (see Anthony, 2015). The waves break along the shoreline, causing sediments to be transported along the coast, and deposited in the delta. The sediment deposition in wave-dominated deltas typically forms a triangular or lobate-shaped delta, where sediment accumulates along the shoreline and gradually extends out into the receiving body of water (Figure 13.10). The wave energy also erodes the delta, especially during storms or high-energy wave conditions. This erosion

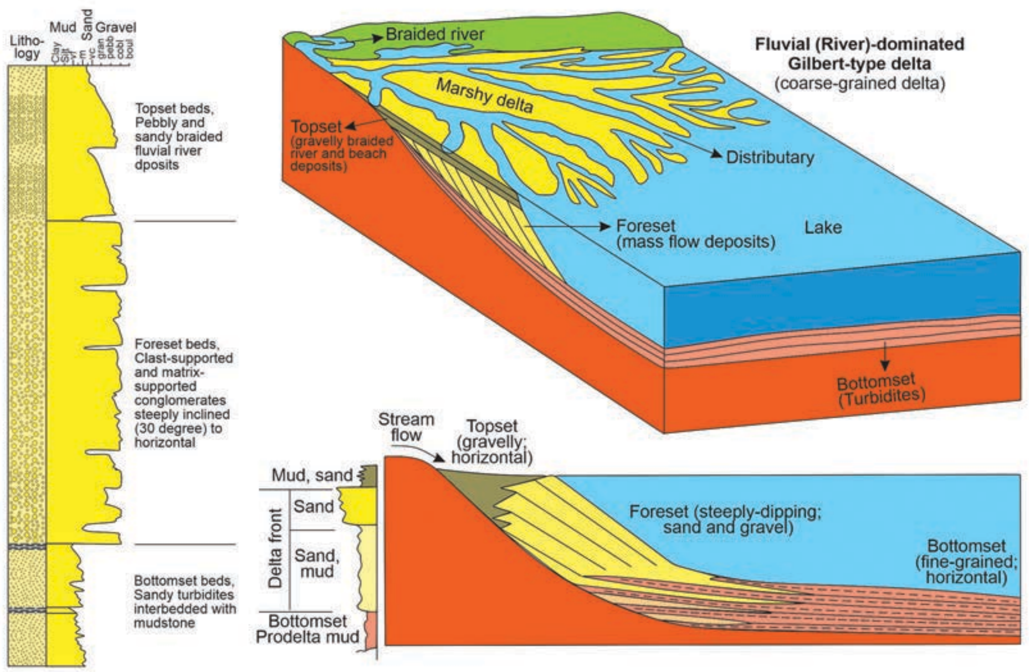

FIGURE 13.9 Characteristics of a Gilbert-type delta (inertia-dominated delta; hyperpycnal flow; see also Figure 13.8). These deltas display a topset, foreset, and bottomset arrangement of beds, formed as sediment deposition progrades basinward.

reshapes the delta and redistributes sediments, leading to changes in delta morphology over time. Some major examples of wave-dominated deltas include the Nile Delta in Egypt (Figure 13.1), the Rhone Delta in France (Figure 13.10A) and the Paraiba do Sul Delta in Brazil (Figure 13.10B). These deltas are influenced by strong wave energy, which shapes their morphology and sediment distribution patterns.

The distributary-mouth deposits (middle-ground bars) are reworked by waves and are redistributed along the delta front by longshore currents to form wave-built shoreline features such as beaches, barrier bars, and spits (Figure 13.10B). The Paraiba do Sul Delta of Brazil is a good example of a wave-dominated delta (Figure 13.10B). The coastal plain consists of littoral marine sediments (Pleistocene-Holocene) and fluvial and lagoonal sediments (Holocene) (Martin et al., 1987). Tidal range over the delta is mesotidal (moderate) but the wave energy is very high. The sediment discharge from the river is also high leading to the extensive development of mouth bars (middle-ground bars) at the mouth of the river (Figure 13.10B). The sediment is transported at high rates across the mouth, leading to the formation of amalgamated, sandy beach ridges (see Bhattacharya and Giosan, 2003) (Figure 13.10B). Owing to this extreme wave energy, the delta is dominated by high-energy environments in which sand deposition takes place (Figure 13.10B). Muds accumulate locally in lagoons, but the interdistributary bay mud deposits characteristic of the Mississippi Delta are absent. The strong sediment drift suppresses the formation of a backbarrier lagoon but produces a more amalgamated and sandier beach ridge plain (Figure 13.10B). Other examples of modern wave-dominated deltas include the Skeidarar Sandur (Iceland, North Atlantic), the Punta Gorda (Belize, Gulf of Honduras), the Tseng-wen (southwest Taiwan; Liu et al., 2013), the Sao Francisco (Brazil; Dominguez et al., 1987; Dominguez, 1996).

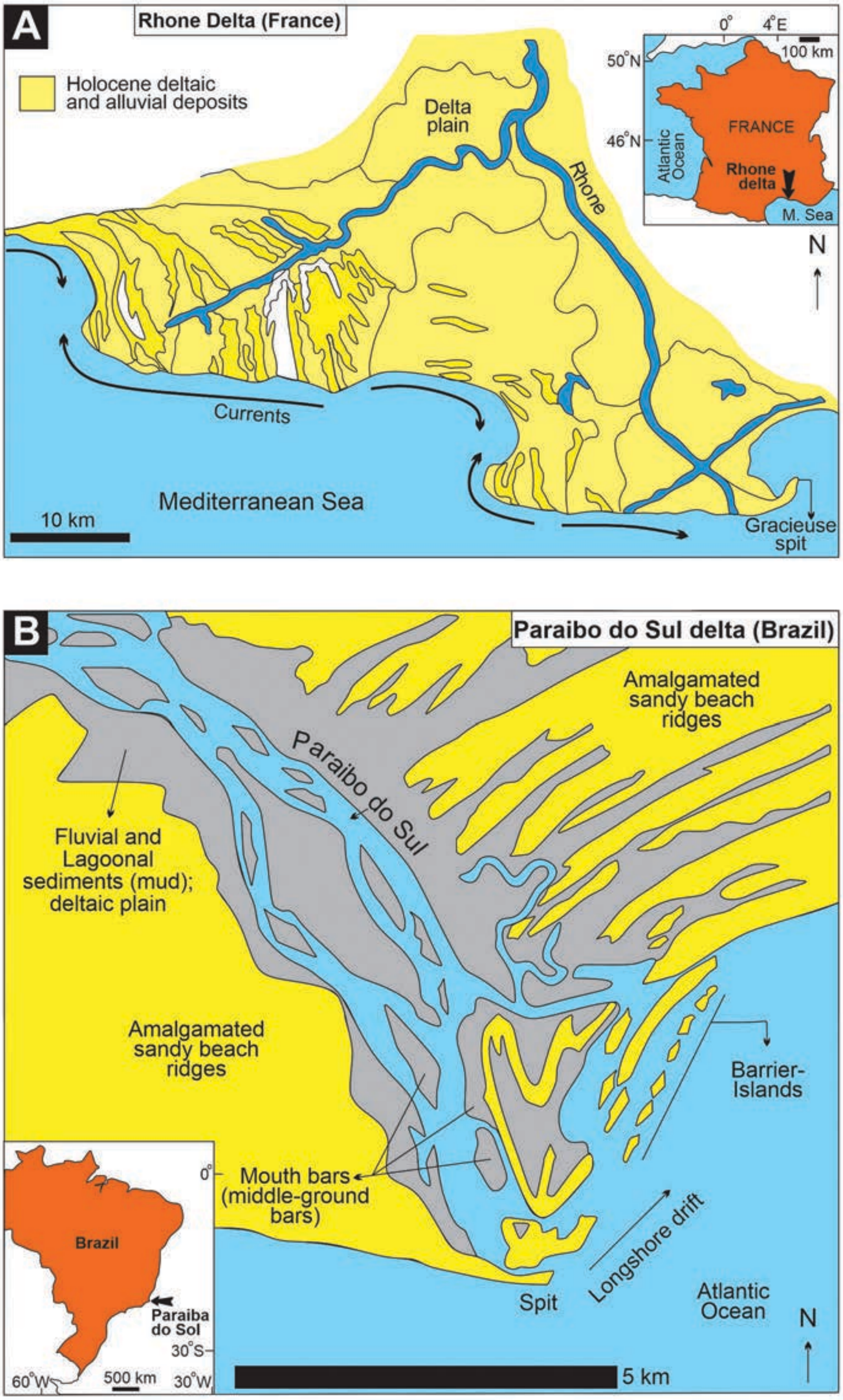

FIGURE 13.10 Wave-dominated deltas. In this type of delta, wave energy controls sediment distribution and delta morphology. The deposited sediments typically form a triangular or lobate-shaped delta, where they accumulate along the shoreline and gradually extend out into the receiving body of water. A: Rhone Delta in France. B: Paraiba do Sul Delta in Brazil.

13.3.3 Tide-Dominated Deltas

Tide-dominated deltas are an end member of the river-wave-tide ternary delta classification (see Figure 13.6). These deltas are large systems that extend hundreds of kilometers across and along continental margins. They are not only characterized by strong tidal influences, but also by fluvial and marine processes that play a critical role in defining the character and behavior of these margin systems (Figure 13.11). Tidal forces control sediment distribution and delta morphology where ebb and flood tides influence flow dynamics and sediment transport. During flood tide, tidal currents bring in sediment from the open ocean and deposit it in the delta whereas during ebb tide, currents reverse direction and carry sediment out of the delta and into the open ocean (Figure 13.11A). The sediment deposition typically forms elongated and branching channels and sand-filled, funnel-shaped distributary networks. The tidal currents also create an intricate pattern of channels and tidal flats, with sediment accumulating in areas of reduced flow or at the mouths of distributaries. The tide-dominated deltas, due to a large fluvial influence, are also very variable and difficult to characterize as rivers differ widely in their discharge, sediment load, seasonality, and grain size. Additionally, these deltas display varying sediment transport energy, hence, the sedimentary successions formed are heterolithic with interbedded sands, silts, and clays (Figure 13.4), and encompass both fining- and coarsening-upward facies associations (Figure 13.11B). The strong tidal influence is manifested by a network of tidal sand bars and channels oriented roughly parallel to the direction of tidal current flow (Figure 13.11A). A variety of sediment types thus accumulate such as tidal-bar or tidal-ridge sands; braided, channel-fill sands, natural-levee, tidal-flat, and flood basin muds. Examples of tide-dominated deltas include the Ganges-Brahmaputra Delta in Bangladesh, Ord Delta (Timor Sea, Australia), and the Mekong Delta in Vietnam.

The tide-dominated delta system is broadly divided into subaerial and subaqueous components of a compound clinoform (Figure 13.11C) (see Goodbred and Saito, 2012). A clinoform, a dominant architectural component of a deltaic-to-continental-slope succession, is a basinward-dipping, chronostratigraphic stratal surface (see Gilbert, 1885; Rich, 1951; Bates, 1953; Patruno and Helland-Hansen, 2018). The subaerial part is further subdivided into a lower delta plain (influenced by tides and other marine processes) and an upper delta plain (a part that is above the tidal influence and is dominated by fluvial processes) (see Figure 13.11C). The offshore subaqueous delta is also subdivided into delta front and prodelta. However, Goodbred and Saito (2012), based on morphology and sediment facies, subdivided the clinothem into delta-front platform (or the subtidal delta plain), the delta-front slope, and the prodelta (see Figure 13.11C). Beyond the rollover point (i.e., the topset-foreset transition), the 'foreset' and 'bottomset' regions of the clinoform correspond to the delta-front slope and prodelta, respectively (see Figure 13.11C). The paired subaerial and subaqueous delta clinoforms are referred to as delta-scale compound clinoforms (see Goodbred and Saito, 2012; Patruno and Helland-Hansen, 2018). In the subaerial delta, the fine-grained sediments form a mud-dominated delta system (characteristic of the intertidal to shallow subtidal zones), salt marshes, mangroves, muddy tidal flats, tidal channels, and channel-mouth bar with laminated to thinly-bedded sand-mud alternations (i.e., heterolithic stratification) with rhythmic climbing-ripple cross-lamination, flaser, lenticular, and wavy laminations or beddings (see Figure 13.4). Tidally influenced deposition also includes bidirectional features of sand-layer stacking and cross-laminations, and mud-drapes. On the outer delta-front platform towards the rollover point (i.e., the topset-foreset transition), sediments typically coarsen. Sedimentary structures include fine to medium-scale bedding with wave ripples, hummocky and trough cross-stratification, and frequent sharp erosional contacts formed by storm-wave scour. At the delta-front slope, coarsening-upward succession of alternating sand and mud deposits or laminated to bioturbated muds are noted. Individual bedding units often comprise of graded (upward fining) and finely laminated sand-silt layers with sharp basal contacts.

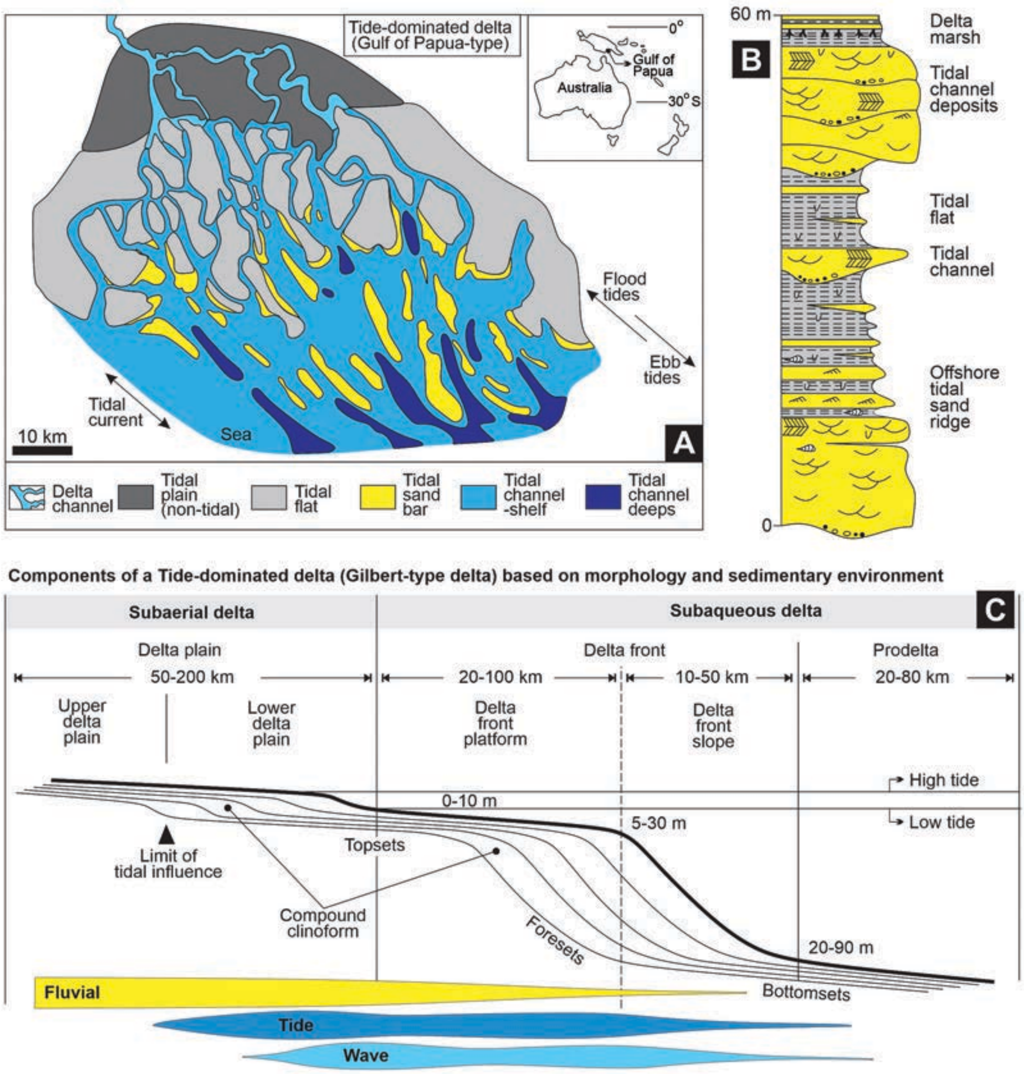

FIGURE 13.11 Tide-dominated delta. A: Gulf of Papua. In general, the tide-dominated deltas are large systems, and an end member of the river-wave-tide ternary delta classification. They are characterized by strong tidal influences along with fluvial and marine processes. B: Major facies associations. These deltas display varying sediment transport energy, hence, the sedimentary successions formed are heterolithic with interbedded sands, silts, and clays, and encompass both fining- and coarsening-upward facies associations. C: Components of a tide-dominated delta (Gilbert-type delta) based on morphology and sedimentary environment. The tide-dominated delta system is broadly divided into subaerial and subaqueous parts of a compound clinoform; each represented by a well-developed prograding clinoform.(Modified after Goodbred and Saito, 2012.)

In general, an idealized stratigraphic succession for a tide-dominated delta can be subdivided into two major parts. The lower portion is characterized by an upward-coarsening succession from the prodelta to delta-front slope; the outer platform deposits are marked at their top by a sharp-based wave and current scours (see also Figure 13.11B). The inner delta-front platform and shoaling to subaerial delta-plain facies is marked by an upward-fining succession of prograding deposits (see Figure 13.11B).

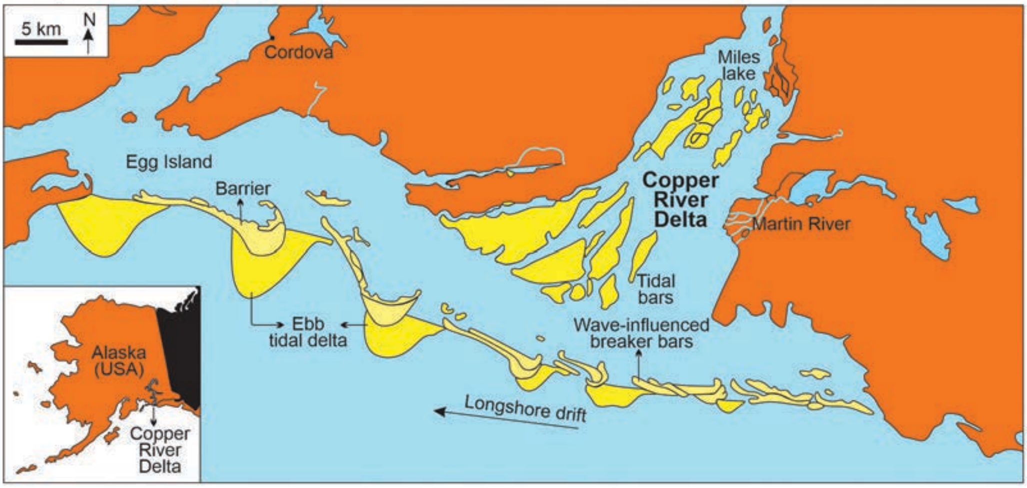

FIGURE 13.12 Mixed-process delta. The Copper River Delta in the Gulf of Alaska (USA) is a good example of a delta strongly influenced by tides and high wave power, and thus, the delta falls somewhere between a tide-dominated and a wave-dominated delta. (Modified from FitzGerald et al., 2014.)

13.3.4 Mixed-Process Deltas

Mixed-process deltas refer to a type of deltaic system where multiple factors and processes contribute to sediment distribution and delta morphology. These deltas are influenced by a combination of wave energy, tidal forces (tidal currents), and riverine processes (river flow). The relative importance of each process varies depending on factors such as the strength of waves, tides, and river discharge. For example, in areas where waves are dominant, sediments are transported along the coast and deposited in the delta, resulting in a wave-dominated deltaic system. In other areas where tides are stronger, tidal currents play a more significant role in shaping the delta, leading to a tide-dominated deltaic system. In some cases, the river flow may be the primary driver of sediment transport, resulting in an inertia-dominated deltaic system.

The mixed-process deltas often exhibit complex and diverse morphologies, with a combination of lobes, distributaries, tidal flats, and channels. The sediment distribution also varies spatially and temporally, depending on the interaction of the different processes. Examples of mixed-process deltas include the Mississippi Delta in the United States and the Ganges-Brahmaputra Delta in Bangladesh and India. These deltas experience a combination of wave energy, tidal forces, and riverine processes, resulting in a complex interplay of sediment transport and deposition. The Copper River Delta in the Gulf of Alaska (USA) is a good example of a delta that is strongly influenced by tides and high wave power, and thus, the delta falls somewhere between a tide-dominated and a wave-dominated delta (see Figure 13.12) (see Galloway, 1976; FitzGerald et al., 2014). Tidal range may exceed 3 m, and tidal currents are up to 2 m/s. Strong, wind-driven swells, coupled with the westerly marine current, set up a net westward longshore drift. Thus, deltaic sediments are modified by both tidal and wave-related processes.

13.4 PHYSIOGRAPHIC AND SEDIMENT CHARACTERISTICS OF DELTAS

The depositional features of a delta are variable due to varying sediment input, outflow velocity, and wave and current energy. But, despite their variable nature, they can be broadly divided into two components, subaerial and subaqueous (see Figure 13.13A) (see also Coleman and Wright, 1975; Galloway and Hobday, 1996; Goodbred and Saito, 2012).

The subaerial component is the deltaic plain, and is divided into an upper delta plain that largely lies above the high-tide level, and a lower delta plain that is between the low-tide mark and the upper limit of tidal influence, dominated by fluvial processes (Figure 13.13A). The lower delta plain, in contrast, is exposed during low tide but is covered by water during the high tide, and thus, influenced by both fluvial and marine (wave) processes (Figure 13.13A). The deltaic plains are characterized by their unique geomorphology, which is shaped by the interplay of sediment deposition, erosion, and the dynamics of the river, and the receiving water body. It is a low-lying, flat or gently sloping area that is created by the accumulation of sediments carried by the river and deposited as the river

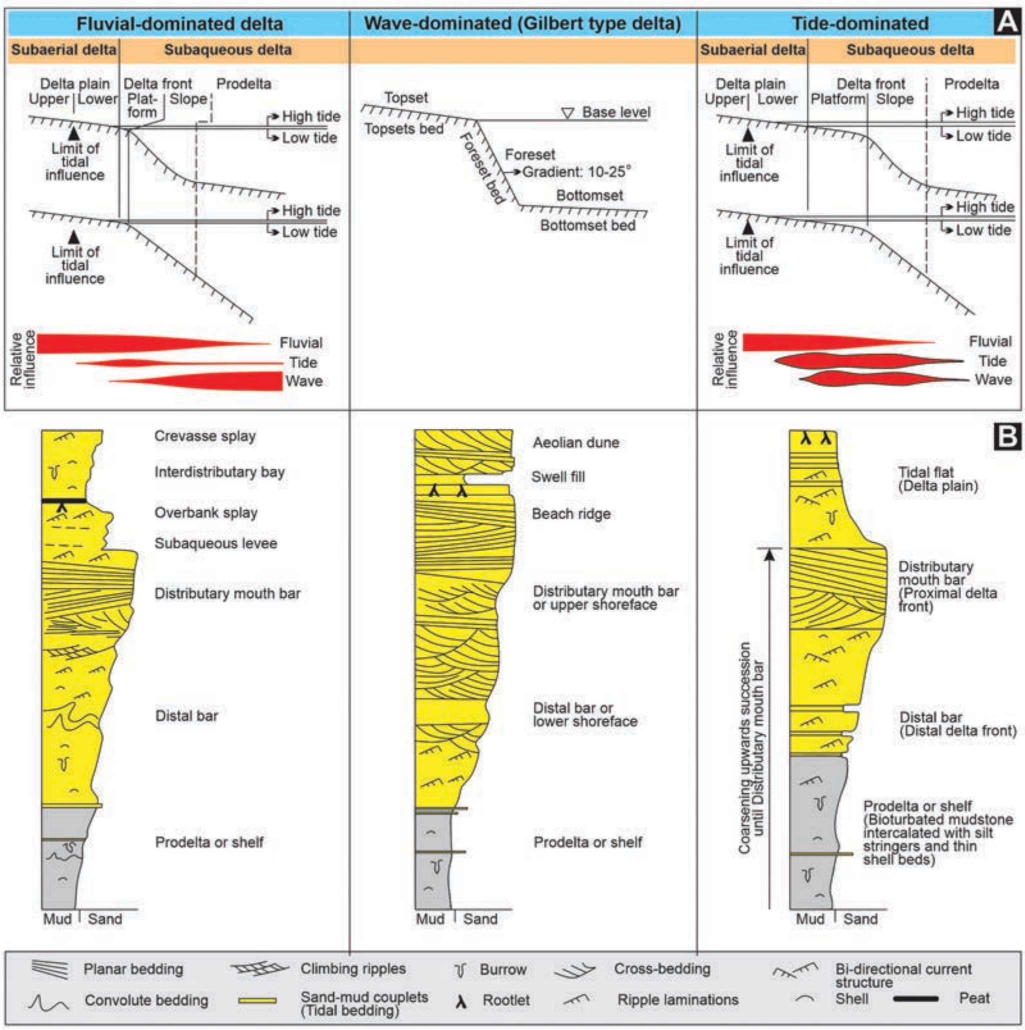

FIGURE 13.13 Delta cycle based on the Mississippi Delta, showing the changes associated with delta growth and abandonment. The figure shows the processes and responses in both the river-dominated regressive and marine-dominated transgressive phases of development. (Modified from Roberts, 1997; Coleman et al., 1998.) A–B: The delta cycle incorporates (1) the initiation and rapid progradation of a delta, (2) the systematic loss of flow efficiency and sediment dispersal, (3) the abandonment of the delta by diversion of flow to a more favorable course, and (4) the marine reworking of the delta perimeter during a local transgression, driven primarily by the combined effects of subsidence (local sea-level rise).

flow slows down upon entering the still water, resulting in the formation of a fertile and often marshy landscape. The shape and size of a deltaic plain depend on various factors such as sediment supply, river discharge, wave and tidal energy, and the slope of the river channel. The deltaic plain (the subaerial component) ranges from small, narrow formations to large, expansive ones that extend for many kilometers but, in general, is less spaced (in length and width) than the subaqueous component (Figure 13.13A).

The upper delta plain, largely influenced by fluvial processes, is characterized by distributary-channel migration and associated fluvial sedimentation processes such as channel and point-bar deposition, overbank flooding, and crevassing into lake basins (Figure 13.13B). Also included are braided and meandering channels, back swamps, and floodplain environments such as swamps, marshes, and freshwater lakes that are predominantly characterized by fluvial sands, gravels, and muds (Figure 13.13B). In deltas where the tidal range is large, the lower deltaic plain shows maximum width with an active distributary system and abandoned distributary fill deposits, flanked by marginal-basin or bay-fill deposits. Distributary channels are numerous; the environments between channels make up the major part of the lower delta plain and includes actively migrating tidal channels, natural levees, inter distributary bays, bay fills (crevasse splays), marshes, and swamps (see Coleman and Prior, 1982).

The subaqueous delta plain lies seaward of the lower deltaic plain below low-tide water level, and extends outward for distances from a few kilometers to tens of kilometers (Figure 13.13A). The subaqueous delta plain forms when a sediment-laden river enters a body of water and deposits its sediment load beneath the water surface. The sediment is transported and distributed by various processes, including river flow, waves, and tides. These processes influence the shape and size of the subaqueous delta, as well as the distribution of sediment within it. The morphology of a subaqueous delta plain varies depending on sediment supply, river discharge, wave and tidal energy, and the slope of the river channel. In areas with high sediment supply and low wave or tidal energy, subaqueous deltas build up and extend outward, creating a prograding delta plain. In areas with lower sediment supply or stronger wave and tidal energy, the delta is more subdued and/or even erosional. The uppermost part of the subaqueous delta, lying at water depths down to 10 m or so, is called the delta front (Figure 13.13A). The remaining seaward part is called the prodelta (slope) (Figure 13.13A); the prodelta may extend up to water depths of 200–300 m.

On fluvial-dominated deltas, deposits typically consist of sand, and possibly gravel, deposited near river mouths, forming distributary-mouth-bar deposits (Figure 13.13B). On the other hand, the delta front may be dominated by high-energy marine processes, including waves, longshore currents, and tides. On wave- and tide-dominated deltas, sediment is reworked and winnowed by high-energy marine processes (such as waves, longshore currents, and tides), creating well-sorted delta-front sheet sands that are cross-bedded (Figure 13.13B). The finest silts and clays (muds) are transported still farther seaward and settle on the prodelta on the outermost part of the subaqueous delta (Figure 13.13B). Previously deposited sediments may be re-entrained, transported, and redeposited farther downslope on the subaqueous delta by gravity-driven mass-movement processes such as landslides, slumps, turbidity-current flows, and mud flows. The major physiographic and sediment characteristics of deltas are briefly enumerated below.

The physiographic characteristics of deltas are their shape, size, topography, channels (and lobes), and sediment composition (size, sorting, stratification, and supply). These are very briefly enumerated. Deltas have different shapes, including fan-shaped, bird's foot, cuspate, or arcuate, depending on factors such as the sediment supply, river discharge, and wave or tidal energy. Deltas vary in size from small formations to large, expansive landforms that extend for many kilometers. Deltas are typically low-lying and flat or gently sloping areas, with elevation gradually increasing from the water's edge towards the inland areas. They often have distributary channels that branch out from the main river channel and distribute sediment and water across the delta plain. Some deltas have multiple lobes, where sediment is deposited in distinct areas, creating a complex deltaic landscape.

The sediment characteristics include sediment composition, size, sorting, stratification, and supply. Deltas are composed of a mixture of sediment types, including sand, silt, clay, and organic matter that are transported and deposited by the river. The size of sediment particles in deltas varies, with coarser sediments typically found closer to the river mouth and finer sediments farther away. Sediments in deltas also exhibit varying degrees of sorting, with well-sorted sediments indicating efficient sorting and transport processes, while poorly sorted sediments indicate limited sorting. Sediments in deltas are often stratified, with distinct layers or beds formed by different sediment deposition events over time. Deltas receive sediment from the river, which depends on the geology of the river basin, erosion rates, and human activities such as damming or land use changes.

Thus, understanding the physiographic and sediment characteristics of deltas is important for studying their formation, evolution, and dynamics.

13.5 DELTA CYCLES

Delta cycle refers to the repetitive patterns of sediment deposition and erosion that occur in deltaic environments over time. These cycles are driven by changes in the relative sea level, sediment supply, tectonics, climate variation, major diversions of rivers upstream, and other environmental factors. Smaller-scale changes may result from processes such as switching of delta lobes, distributaries, or tidal channels.

Delta cycles typically consist of four main stages: progradation, aggradation, retrogradation, and abandonment (Figure 13.14). During the progradation stage, sediment is delivered to the delta and is deposited at the delta front (Figure 13.14A). Thus, the deltaic deposits prograde seaward, leading to the generation of a coarsening-upward succession of facies as delta-front sands advance seaward over prodelta silts and clays; the delta extends seaward as new sediment is added to the system (Figure 13.14A). In the progradation stage, the rate of deposition is equal to the rate of accommodation (Figure 13.14A). This stage is characterized by the formation of delta-front deposits, including foresets, the inclined layers of sediment that builds outward from the shoreline. A good example is the progradational lobes of the fluvial-dominated Mississippi Delta (see Figure 13.7).

In the aggradation stage, sediment continues to accumulate at the delta front, causing the delta to grow vertically (Figure 13.14B). This stage is characterized by the deposition of delta-top deposits that are either horizontal or gently inclined layers of sediments that accumulate on the top of the delta (Figure 13.14B). These deposits include distributary channels, i.e., channels that distribute sediments across the delta top.

During the retrogradation stage, the rate of sediment supply decreases or the relative sea level rises (Figure 13.14C). As a result, the delta front retreats landward, and erosion occurs. The sediments are reworked and transported away from the delta front, resulting in the erosion of foresets and the formation of bottomset deposits (the horizontal or gently inclined layers of sediments that accumulate behind the delta front).

The abandonment stage occurs when the river feeding the delta jumps to a new channel and empties its sediment load somewhere else along the coast, and thus shuts of the sediment supply to the abandoned delta (Figures 13.14D–E). This abandoned delta subsides below sea level, and coal beds develop on top, as a cap (Figure 13.14F). Delta abandonment also occurs when the delta can no longer keep pace with rising sea levels or the sediment supply decreases significantly. The delta ceases to grow and may be partially or completely submerged. Erosion and reworking of sediment continue, and the delta is modified by wave and current action.

These above mentioned four stages can repeat multiple times over the lifespan of a delta, resulting in the formation of stacked sedimentary sequences (Figure 13.14F). Sediments that make up a complete delta cycle range in thickness from 50 to 150 m (see Miall, 1984). This ‘‘delta cycle’’ is

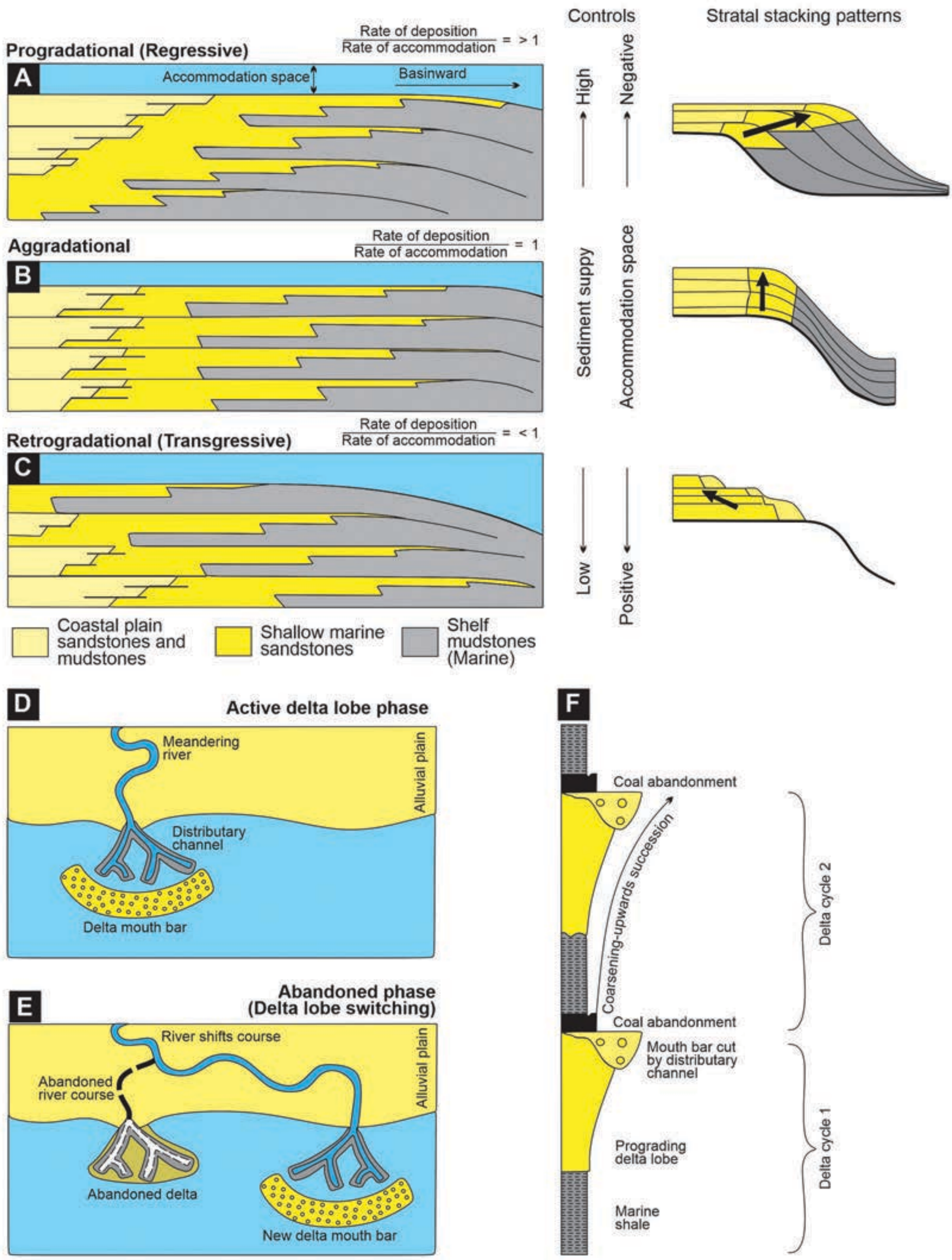

FIGURE 13.14 Delta cycles. Delta cycles typically consist of four main stages: progradation (A), aggradation (B), retrogradation (C), and abandonment (D–E). F: These four stages repeat multiple times over the lifespan of a delta, resulting in the formation of stacked sedimentary sequences.

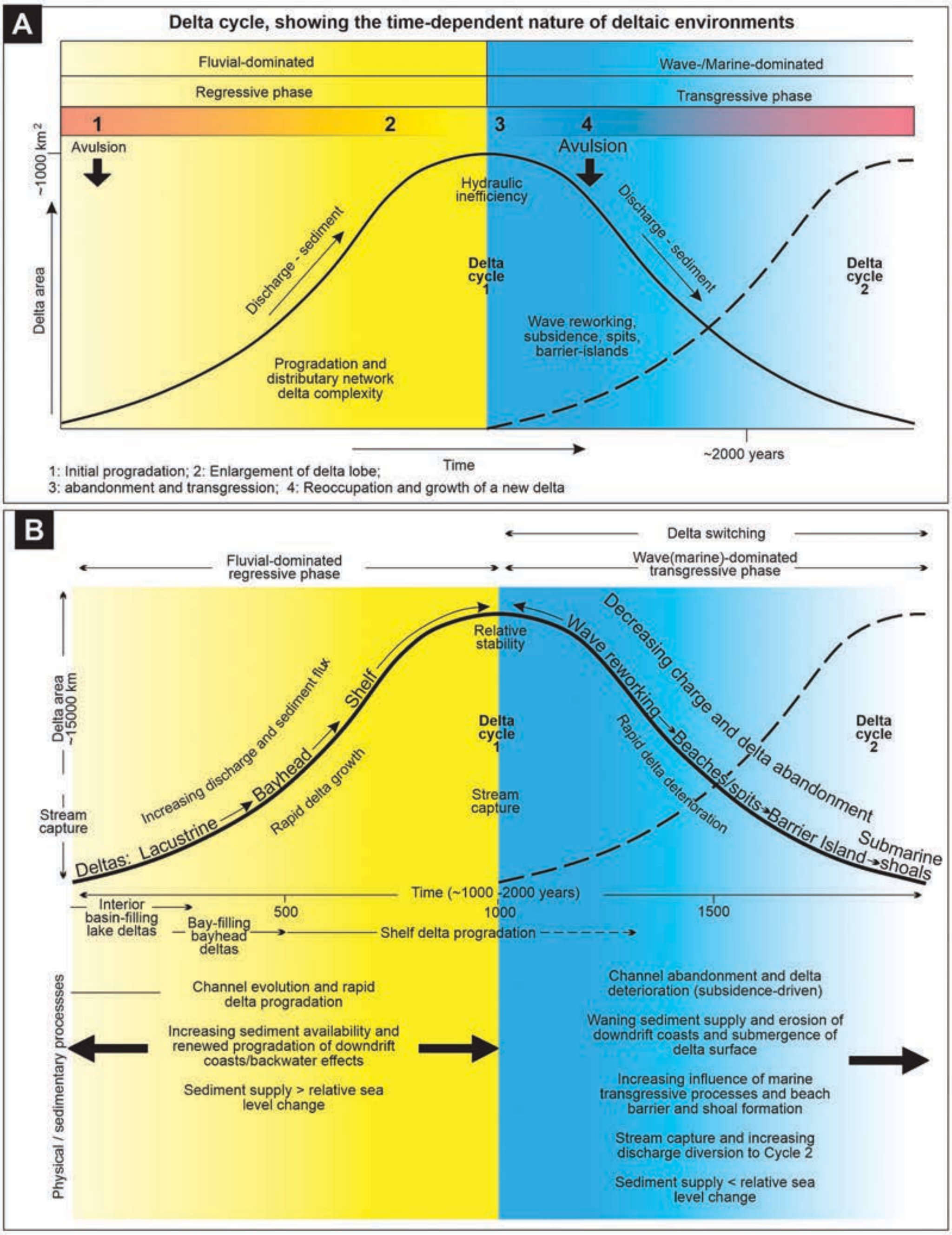

FIGURE 13.15 A: Physiographic and sediment characteristics of deltas. B: Idealized facies successions of fluvial-, wave, and tide-dominated deltas. (Modified from Coleman and Wright, 1975; Galloway and Hobday, 1996.)

associated with distinct sequences of change (see Roberts, 1997), including: a) an initial and rapid progradation of the delta; b) enlargement of delta lobe / sediment dispersal; c) abandonment of the delta by the diversion of flow and transgression; and d) reoccupation and growth of a new delta; marine reworking of the delta during a local transgression driven primarily by subsidence (local sea

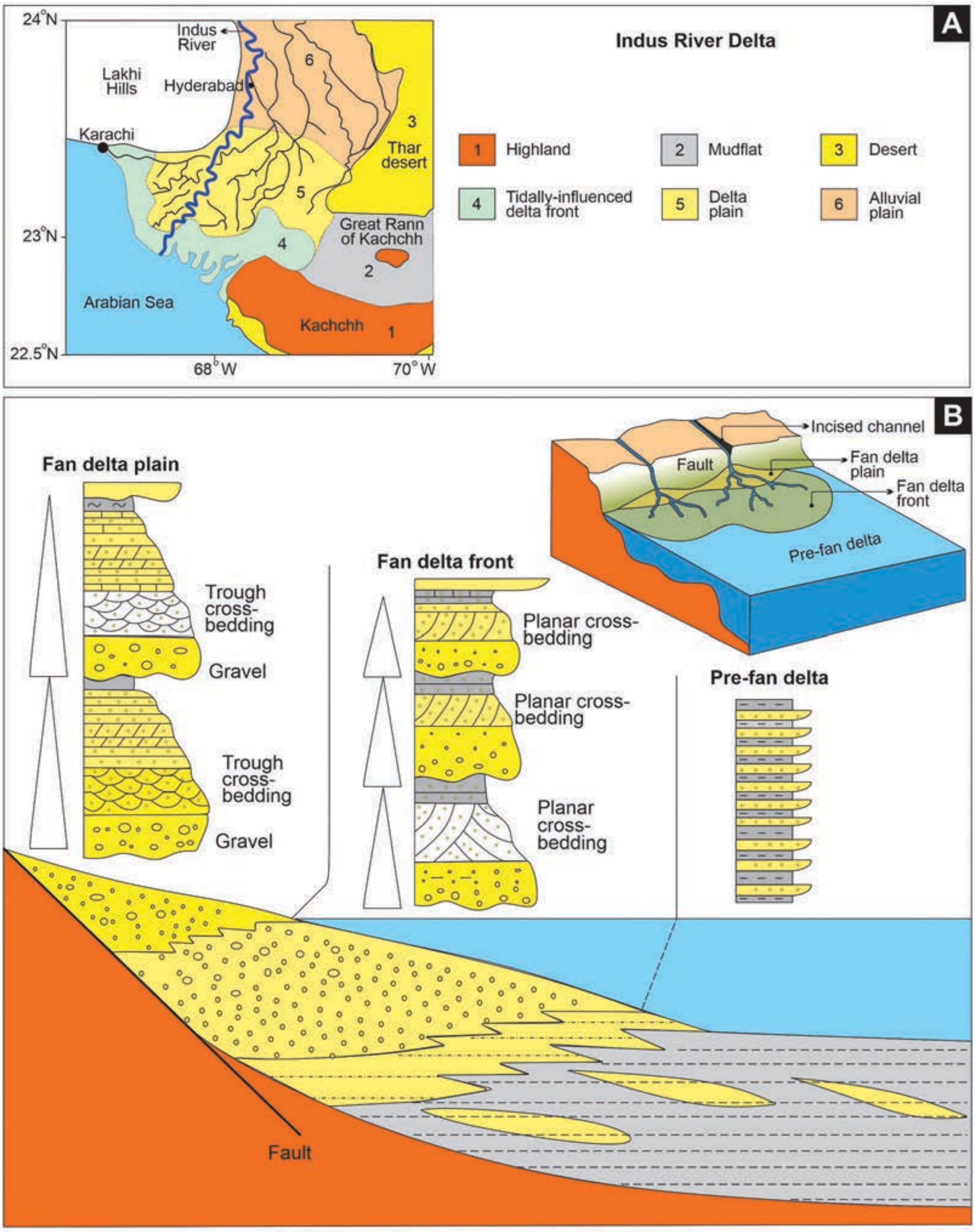

FIGURE 13.16 Fan delta. A: Indus fan delta. B: Facies association of a fan delta in the south of Albert Rift, Uganda. (Modified after Xu et al., 2023.)

level rise) (Figure 13.15). As noted in the Mississippi Delta, Roberts (1997) proposed this model for the delta cycle that begins after delta switching (Figures 13.15D–E) (see also Coleman and Wright, 1975; Coleman et al., 1998). The model incorporates both regression (fluvial-dominated phase) and transgression (marine-/wave-dominated phase) (Figure 13.15). Regression progresses through basin filling by a lake delta, bay filling by a bayhead delta, and progradation of a shelf delta during a time when sediment supply provided by the river outpaces relative sea level. Transgression takes place because of an increasing influence of marine processes wherein the delta deteriorates and is abandoned by its distributary stream. Regression and transgression of a delta lobe occurs on a time scale of approximately 1,000 to 2,000 years.

13.6 FAN DELTAS

Holmes (1965) defined a fan delta as a coastal prism of sediment delivered by an alluvial-fan system (subaerial component) and deposited, mainly or entirely subaqueously, at the interface between the active fan and a standing body of water (see also Nemec and Steel, 1988). Examples of fan deltas include the Nile Delta in Egypt and the Indus Delta in Pakistan (see Figure 13.16A). These deltas are characterized by their fan-shaped morphology and the deposition of sediment from rivers. The Indus River originates high on the Himalayas (at an elevation of 5500 m), and thereafter flows more than 2900 km through India and Pakistan before emptying into the Arabian Sea forming a fan-shaped delta with an extensive system of swamps, mudflats, creeks, estuaries, marshes and mangroves forests (Figure 13.16A).

Fan deltas typically form in areas where the river's sediment load exceeds the transport capacity of the water body it enters. As the river enters the still water, it slows down, causing the sediment to settle and deposit. This sediment is then distributed in a fan-shaped pattern, with coarser sediments deposited closer to the river mouth and finer sediments, farther away (Figure 13.16B). The shape and size of fan deltas vary, depending on factors such as sediment supply, river discharge, and the slope of the river channel. In some cases, the sediment load may be large enough to build up the delta above the water level, creating a prograding delta that extends outward into the water body. In other cases, the sediment load may be smaller, resulting in a submerged or partially submerged delta. This subaqueous portion may be fluvial-dominated, wave-dominated, or tide-dominated. Sediments are deposited downslope in the subaqueous part of a fan delta by processes such as slumping and debris avalanching, turbidity-current flow, and inertia (hyperpycnal) flow that takes place particularly during flood stages.

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14 Beach and Barrier-Island Systems

14.1 INTRODUCTION

The beach and barrier-island systems are dynamic coastal landforms formed by the interaction of waves, currents, and sediments (Figure 14.1). They are typically found along the shoreline, providing a transition zone between land and sea and consist of sand, gravel, and pebbles (or even cobbles) that are deposited by the wave action (i.e., forming wave-dominated landforms). Finer-sand beaches have very low gradient/slope (~1°) compared to those with pebbles and cobbles (~20°). Beaches also vary in size and shape, depending on factors such as wave energy, sediment supply, sea-level changes, and coastal morphology. In the mid-latitudes, beaches have siliceous or quartz sand grains derived from erosion, in topics, carbonate sediments (coral reef detritus and shells) dominate, whereas in higher latitudes, physical weathering produces coarse rock fragments and gravels. Hence, sediment composition greatly varies on a beach; the sediments are derived from land and delivered via rivers, glaciers and shoreline erosion, and from marine organisms (such as sea shells).

The coastal sand dunes (or barrier dune systems) (see Figure 14.1) are found above the high-water mark of sandy beaches and occur on ocean, lake and estuary shorelines, and river mouths. They are most common along coasts exposed to strong winds and with abundant sediment supply, as noted along the Atlantic East Coast. But they can also occur in any climatic zones, from the Arctic to the equator, and hence are globally distributed.

14.2 BEACH AND BARRIER-ISLAND SYSTEMS

Beaches range from low-energy systems, where small waves lap against the shore, to those with high waves breaking across several hundred meters of the surf zone. They are exposed to micro (<2 m), meso (2–4 m), macro (4–8 m) and mega tidal (>8 m) ranges (see Figure 14.2A). Davis and Hayes (1984) noted that the relative influence between wave processes and tidal processes is a major factor in the overall development of coastal morphology including, but not restricted to, barrier islands. Hence, beaches, based on the relative tide range (Figure 14.2B), are categorized into three types: wave-dominated, tide-dominated, and mixed wave-tide-dominated (discussed later in the chapter) (see also Short, 2006) (see Figure 14.2C). The relative tidal range is the ratio of tide range to breaker height or in simple terms, the difference in height between high and low tides (Figure 14.2B). Davis (1994) noted that the wave-dominated coasts (Figure 14.2C) typically have well-developed barrier islands (Figure 14.1A), and the tide-dominated coasts (Figure 14.2C) are marked by their absence (see Davis and Hayes 1984; Davis 1989), whereas the mixed-energy (mixed wave-tide-dominated) coasts (Figure 14.2C) have barrier islands, where both waves and tides have a significant influence.

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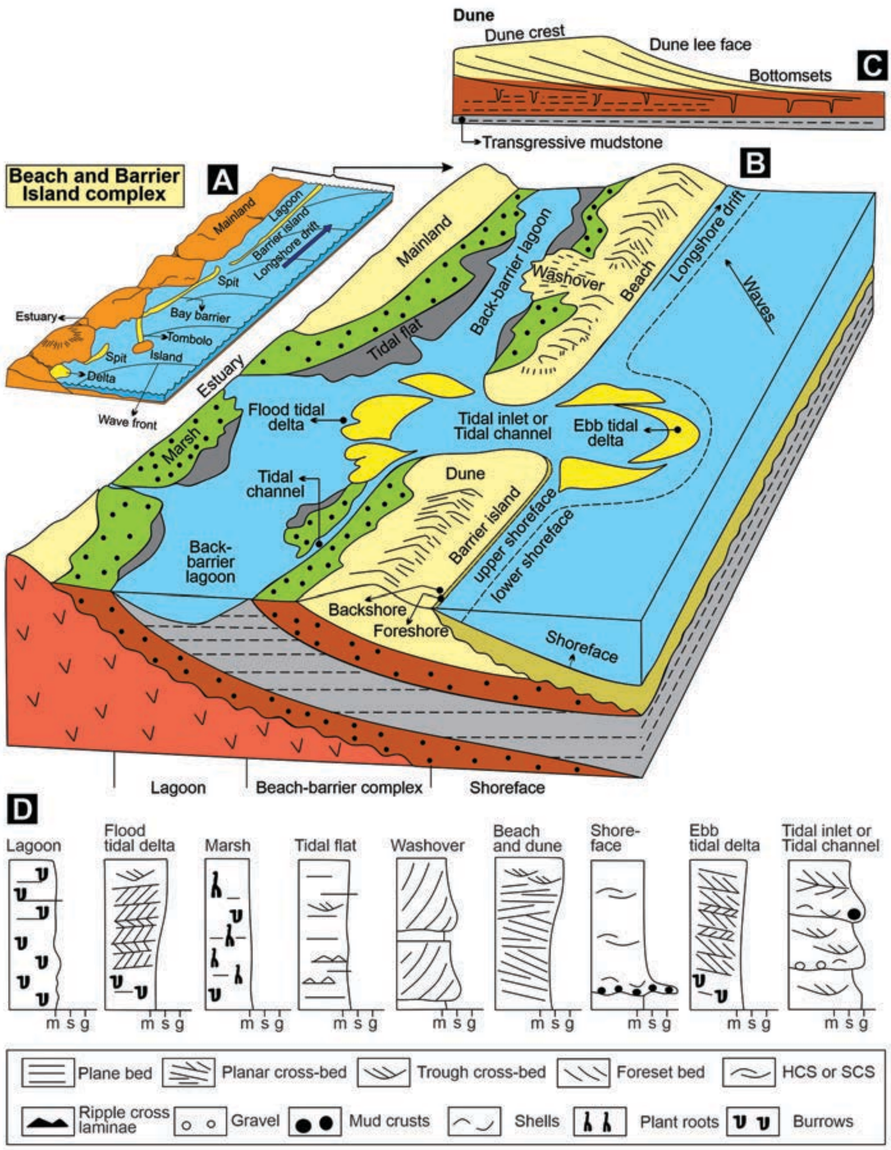

FIGURE 14.1 Subenvironments of a beach and barrier-island system. A: Beach and barrier-island complex. B: Subenvironments. (Modified after Walker, 1979, 1992.) C: Dune system. D: Lithosections of each subenvironment mentioned in A–B. (Modified after Reinson, 1992.)

Barrier islands, elongated and narrow landforms that run parallel to the mainland coast, account for 10 to 15% of the world's total shorelines (Figures 14.1A–B). They are separated from the mainland by a lagoon (back-barrier lagoon) and are formed by the accumulation of sand and sediments carried by waves and currents (Figure 14.1). Thus, barrier islands act as a natural buffer, protecting the mainland from the full force of ocean waves and storms. Barrier islands often develop along low profile coasts, characterized by gentle offshore gradients, low to moderate tidal range with an abundant supply of sediments. Important chains of barrier islands are found along the coast of Alaska (the USA), the Colombian Pacific coast, Ecuador, Peru, Brazil, Africa, the Netherlands, Germany,

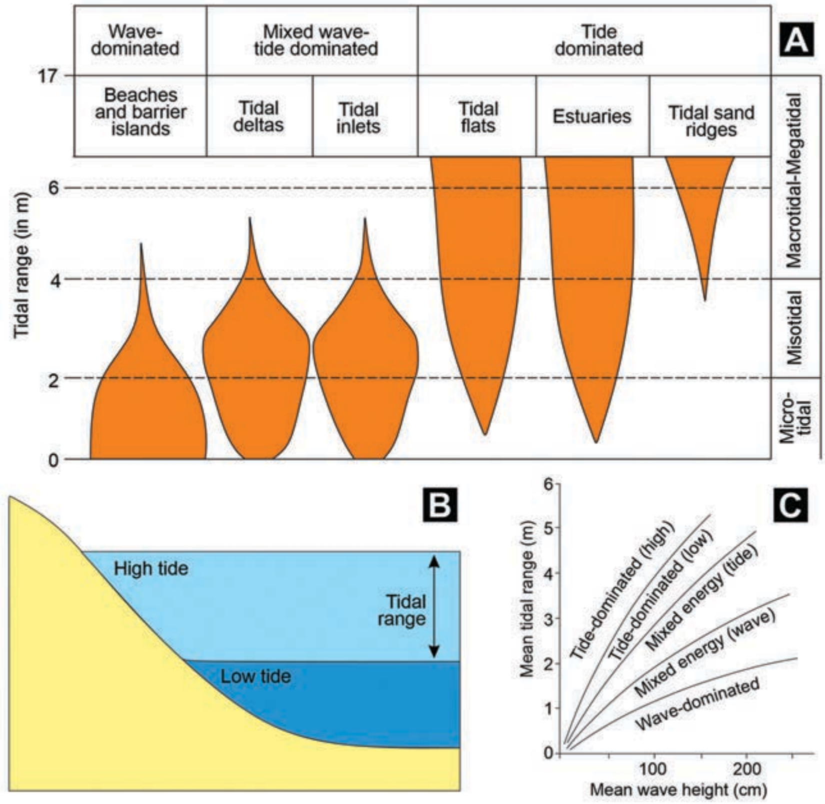

FIGURE 14.2 Tidal ranges. A: Beaches are exposed to micro (<2 m), meso (2–4 m), macro (4–8 m) and mega tidal (>8 m) ranges. (Modified from Davis and Hayes, 1984.) B: Tidal range. C: Coastal types based on the relationship between mean tidal range and mean wave height. (Modified after Hayes, 1979; Davis and Hayes, 1984.)

Denmark, Siberia, India, Sri Lanka, the southern coast of China, and Australia (see also Short and Woodroffe, 2009). The longest stretches, with a total length of 4500 km, are located along the eastern and Gulf coasts of North America.

Barrier islands are particularly vulnerable to erosion and changes due to their exposure to wave action and storms (Figure 14.3A). Waves continuously shape and reshape the beach by eroding sediment from some areas and depositing them in others through coastal erosion and deposition, respectively. The wave cuts out the rock at the base, forming a sea arch that may collapse to isolate the point as a stack (Figure 14.3B). Rocks behind the stack may be eroded away, and the sand eroded from that point collects behind it forming a tombolo, a sand strip that connects the stack to the shoreline (Figure 14.3C). Where sand supply is low, wave energy erodes a wave-cut platform across the surf zone, thus, exposing the bare rock with tidal pools, at low tides (Figures 14.3A–B). Over time, barrier islands may migrate landward or seaward, depending on the balance between sediment supply and erosion. They are also subject to changes in their stability and shape, due to changes in sea levels. In general, three agents transport sediments across a natural barrier: storm-driven water (overwash), tidal-induced currents (inlets), and windblown sand (dune migration). These processes, based on their relative importance, contribute to barrier beach migration.

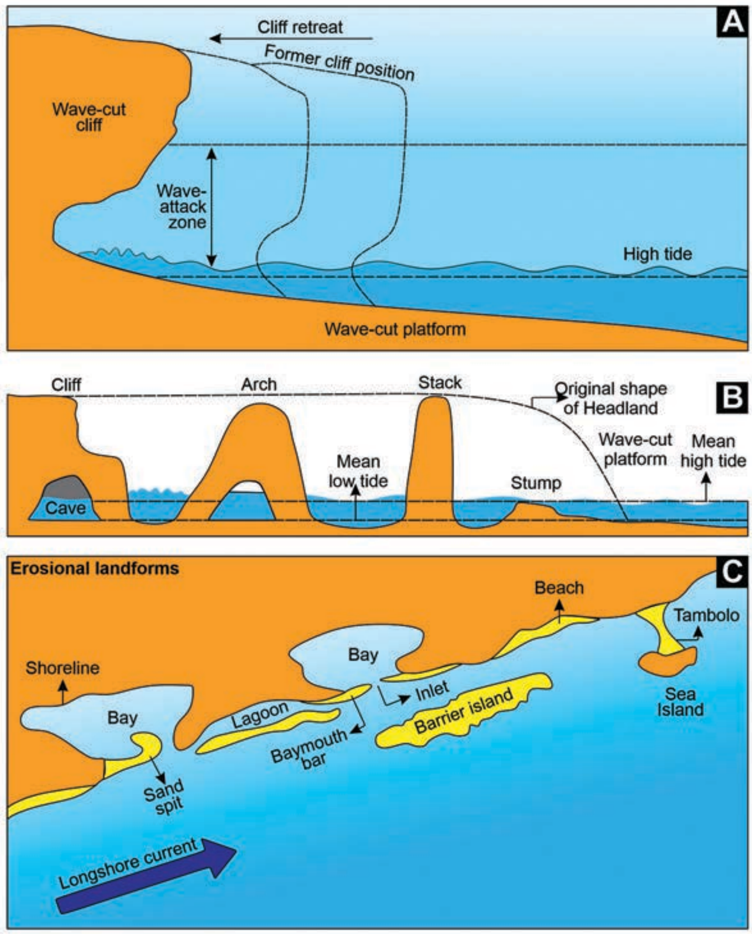

FIGURE 14.3 Coastal landforms. A: Formation of wave-cut platform. B: Landforms made by erosion. C: Landforms made by deposition. (A–B: Modified from Short and Woodroffe, 2009.)

14.3 BEACH MORPHOLOGY

Beach morphology refers to the physical characteristics and features of a beach, including its shape, profile, and sediment composition. It is influenced by various factors, such as wave energy, and sediment supply. A beach extends from the fair weather wave base (FWWB) where the waves begin to feel the sea bottom and shoal, through the surf zone, and to the upper limit of the wave swash (swash or swash zone) (Figure 14.4).

An ideal beach profile is divided into four zones: backshore, foreshore, nearshore, and offshore (Figure 14.4). The offshore is below any shoreline-derived processes; it is geologically active due to the cascading sands of turbidites and deeper currents (including contourites).

The nearshore is affected by the waves, i.e., that part of the shore where water depth is a half wavelength or less (i.e., wave shoaling: the change in the shape and behavior of waves as they move into waters of decreasing depth; from circular to oval at shallow depths) (see Figure 14.4A). Thus, the width of this zone depends on the maximum wavelength of the approaching wave train and the slope of the seafloor (Figure 14.4A). The nearshore area is typically called the shoreface and

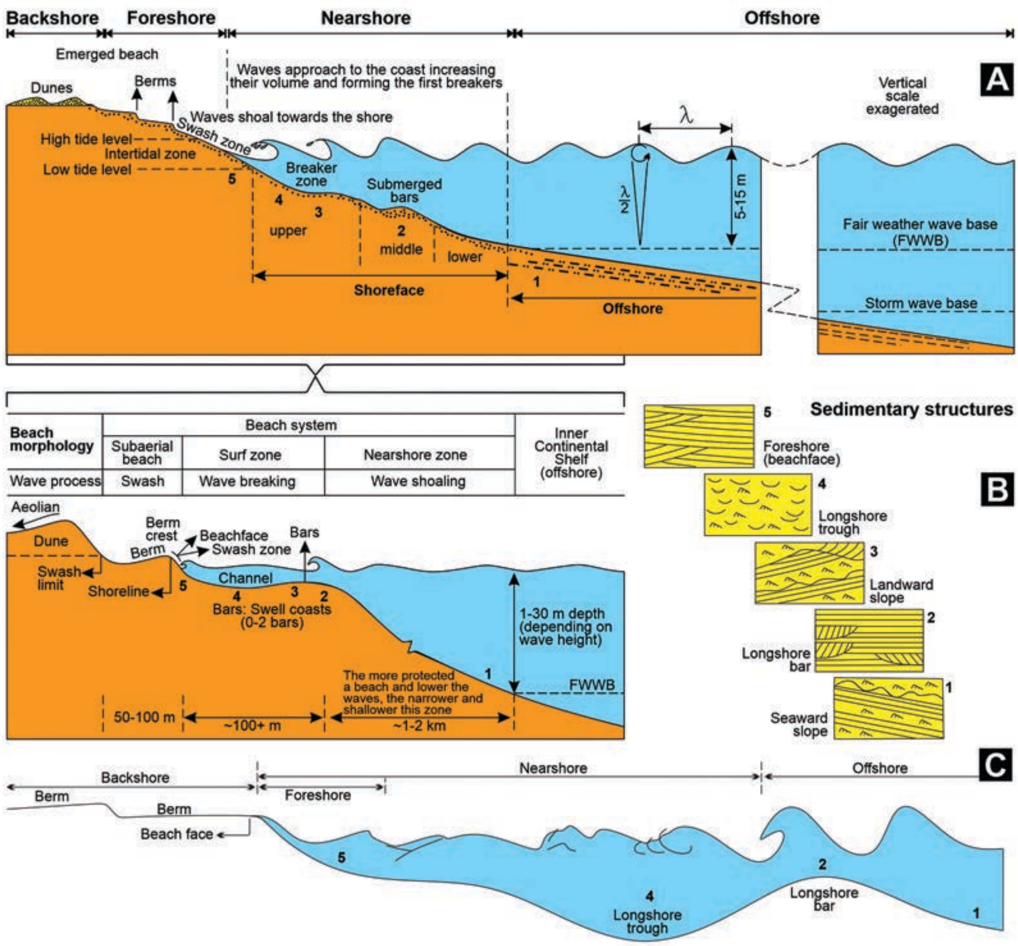

FIGURE 14.4 Beach morphology. A: Characteristics and features and zones. B. Associated sedimentary structures. (Modified after Davidson-Arnott and Greenwood, 1976.) C: Morphological features of a nearshore environment.

is broken into three segments: upper, middle and lower shoreface (see Figure 14.4A). The upper shoreface (see Figure 14.4A) is affected by everyday wave action and typically consists of finely-laminated and cross-bedded sands, whereas the middle and lower parts are marked by hummocky cross-stratified sands, HCS (Figure 14.5). The foreshore (Figure 14.4) is the area (mostly overlapping the surf zone) that is periodically wet and dry, due to the action of waves and tides; it displays planer-laminated, well-sorted sands (beach foreshore; see Figures 14.4 and 14.5). The surf zone is where the waves break (see Figure 14.6). The beach face is where the swash of the breaking wave runs up, and the backwash flows back down (see Figure 14.6). Above the beach face are low ridges called berms (Figure 14.4). A berm is the high point on the beach; it is the backshore-foreshore boundary that changes seasonally (Figure 14.7). During summer, the wave energy is typically lower, and this allows sand to be piled onto the beach (Figure 14.7A), whereas higher winter storm energy moves the summer berm sand off the beach, and piles it in the near from which it will be replaced next year as it is moved back onto the summer berm (Figure 14.7B). This causes deposition and erosion (respectively), seasonally and leads to the berm migration. Behind the berm and the beach, there may be a zone of dunes representing sand-blown particles by onshore winds

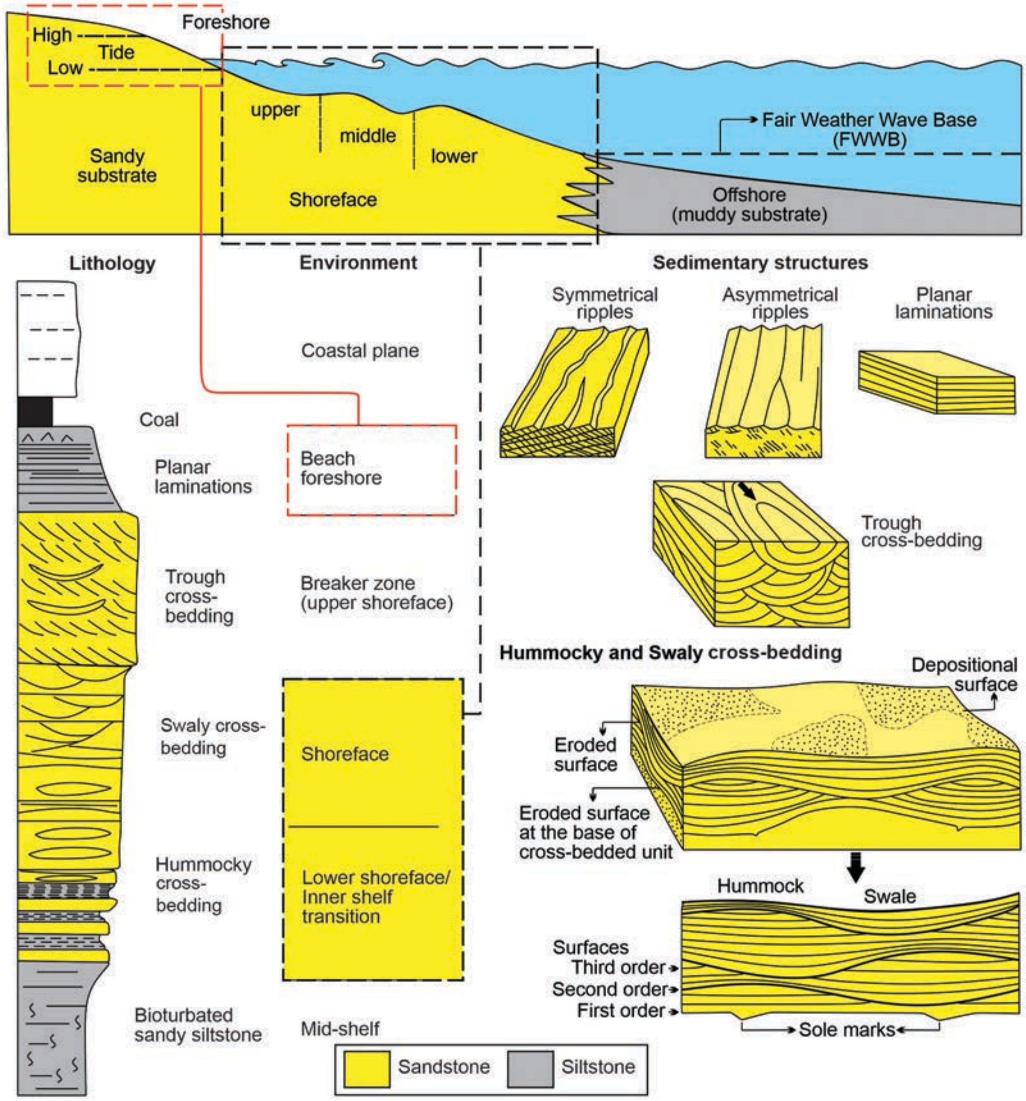

FIGURE 14.5 Shoreface depositional environments, sedimentary structures, lithofacies, and zones. Symmetrical ripples, common in upper shoreface, reflect bi-directional flow. Asymmetrical ripples, common in upper shoreface and foreshore settings, reflect unidirectional flow. The trough cross-bedding, i.e., beds that cut across troughs of other beds; the cut is sharper in hummocky than in swaley cross-bedding.

(see Figures 14.1D–E and 14.4A). This area behind the berms that is always above the ocean in normal conditions is known as the backshore (Figures 14.1A and 14.4A). Table 14.1 provides major characteristics of backshore and foreshore areas.

The shape of a beach is determined by the balance between erosion and deposition processes. Low-energy waves, such as those found in sheltered bays, enable sediment deposition, resulting in a wider beach with a gentler profile, whereas high-energy waves, in exposed coastlines, erode the beach, thus, forming a steep profile with a narrow beach width (Figure 14.7). The profile of a beach refers to its elevation or slope from the shoreline to the backshore; it varies from being gentle and a wide profile (i.e., dissipative beach; Figure 14.7A) to a steeper, and narrower profile (i.e., reflective

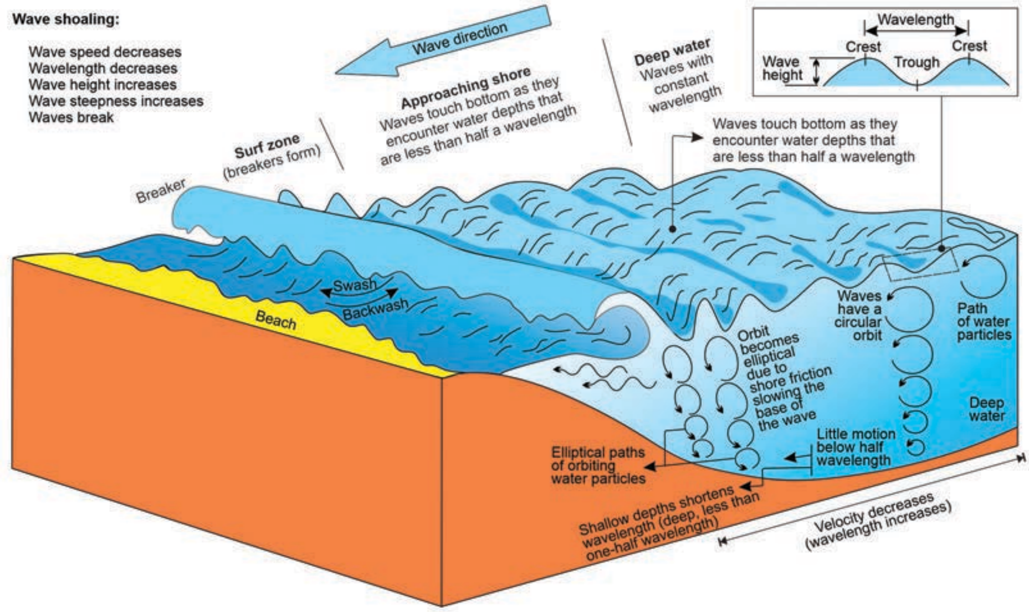

FIGURE 14.6 Wave shoaling and wave characteristics.

TABLE 14.1
Differences between a backshore and foreshore

Feature	Backshore	Foreshore
Location	The backshore is located landward of the high-tide mark and is above the normal reach of the waves	The foreshore is the zone between the high-tide mark and the low-tide mark, regularly affected by waves and tides
Elevation	The backshore is situated at a higher elevation than the foreshore	The foreshore is at a lower elevation than the backshore
Wave influence	The backshore is less affected by wave action and typically remains dry except during extreme high tides or storm events	The foreshore experiences direct wave action and is often submerged during high tide
Sediment characteristics	The backshore may have drier, coarser sediments, including sand and gravel, which are less prone to water saturation	The foreshore generally consists of wet, fine-grained sediments such as sand, silt, and clay that are influenced by wave and tidal processes

beach; Figure 14.7B). The dissipative beaches have a lower wave energy and allow waves to dissipate their energy as they break on the shore (Figure 14.7A), whereas reflective beaches have a higher wave energy and tend to reflect waves back into the ocean (Figure 14.7B).

Waves shape the coast by means of swash (i.e., erosion: the breaking down and carrying away of sediments by the sea) and backwash (deposition: when sediments are carried by the sea and are deposited or left behind on the coast) (see Figure 14.8A). Two types of waves do this: constructive and destructive waves (Figures 14.8B–C). The constructive waves have low wave height and long wave length with low frequency, between 6 and 8 waves per minute. They are associated with weaker backwashes and stronger swashes (Figure 14.8B), allowing the build-up of wide flat beaches

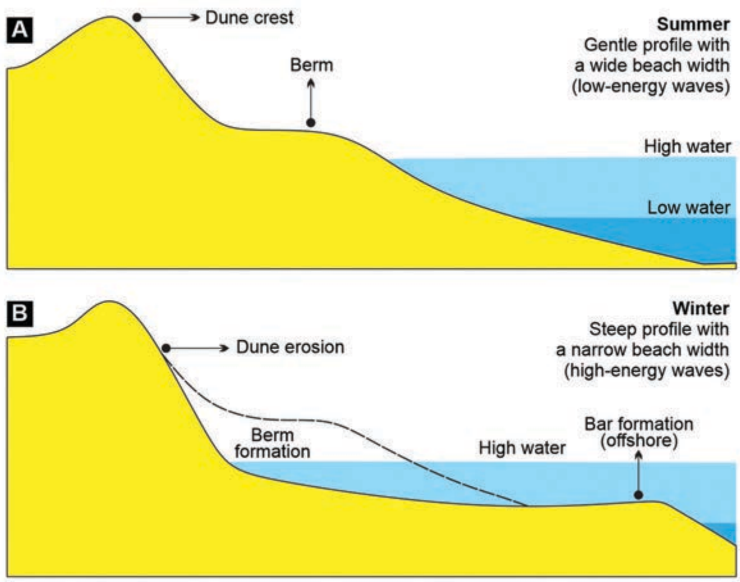

FIGURE 14.7 Beach profile. A: Gentler waves deposit sand from offshore bars onto the beach, resulting in wide and higher (elevation) beaches. B: Stronger waves with more energy pick up those particles deposited, and carry them back offshore in bars, thus resulting in narrower and steeper (profile) beaches. The offshore bars buffer the beach from erosion as they cause waves to break further offshore. (Modified from Friends of IBSP – https://friendsofibsp.org/breaking-news/science-of-the-shore-a-tale-of-two-beaches-winter-summer-beach-profiles/).

and thus are associated with sediment deposition; constructive waves tend to form sandy beaches. Large oceans with large fetch form large waves, called destructive waves. These waves have large wave height and short wave length and are characterized by tall breakers that have high downward force and a stronger backwash than swash (Figure 14.8C). This downward energy helps erode beaches and cliffs by pulling the sediments away from the shoreline (Figure 14.8C). In addition, due to a dominant backwash, they erode the beach, thus resulting in a narrow and steep beach profile; destructive waves tend to form pebbly beaches.

Coastal processes, such as longshore drift, tidal currents, and storm events, also play a significant role in shaping beach morphology (Figure 14.9). The longshore drift is the process by which sediments are transported along the coast by the movement of waves and currents (Figure 14.9A-B). It results in the formation of spits, tombolos, and other coastal landforms (Figure 14.2C). These processes are enumerated below in more detail.

14.4 DEPOSITIONAL PROCESSES

Beach depositional processes refer to the various mechanisms by which sediment is transported and deposited on a beach. These processes are driven by the action of waves, currents, and tides. These play a crucial role in shaping the morphology and composition of a beach.

14.4.1 Wave Action

Waves are the primary driver of sediment transport on a beach. As waves approach the shore, they break and generate a swash and backwash motion (Figure 14.8A). During the swash phase, waves

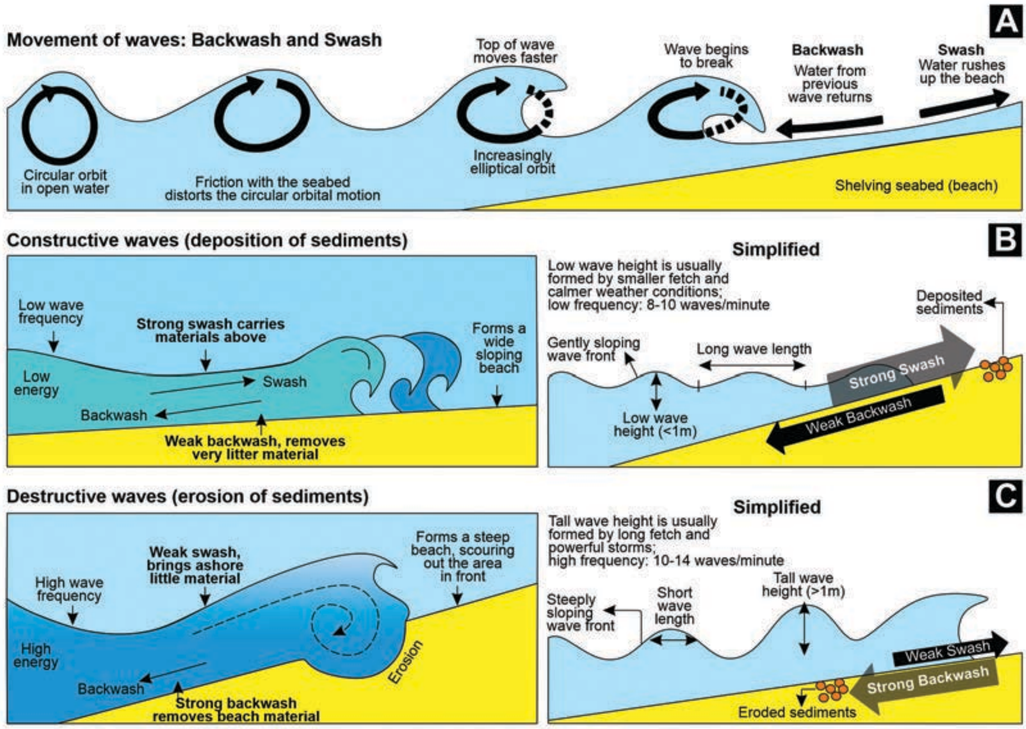

FIGURE 14.8 Movement of waves. A: Swash and backwash. Swash (i.e., erosion: the breaking down and carrying away of sediments by the sea), and backwash (deposition: when sediments are carried by the sea and are deposited or left behind on the coast). B: Constructive waves. The constructive waves have low wave height and long wave length, and low frequency. They are associated with weaker backwash and stronger swash, allowing the build-up wide flat beaches, and thus are associated with sediment deposition. C: Destructive waves. These waves have large wave height and short wave length and are characterized by tall breakers that have high downward force and a stronger backwash than swash. This downward energy helps erode beaches and cliffs by pulling the sediments away from the shoreline.

carry sediment up the beach, while during the backwash phase, sediment is returned back to the ocean (Figures 14.8 and 14.9A). This back-and-forth motion of sediment is known as the beach drift.

14.4.2 Longshore Drift

Longshore drift is the lateral movement of sediment along the shoreline (Figure 14.9A). This occurs when waves approach the beach at an angle, causing sediment to be transported parallel to the coast (Figure 14.9A). This process is driven by the combined action of wave swash and longshore currents (Figure 14.8A). Longshore drift results in the formation of sandbars, spits, and other landforms (Figure 14.2).

14.4.3 Wave Refraction

Waves usually approach the shoreline at an angle (Figure 14.9A), with one end of the waves of the train slowing down first, thus causing the waves to bend toward the beach (Figure 14.9B). Such bending of waves as they enter shallower water is called wave refraction. Refracted waves approach the shoreline at an angle creating a slight difference between the swash as it moves up the

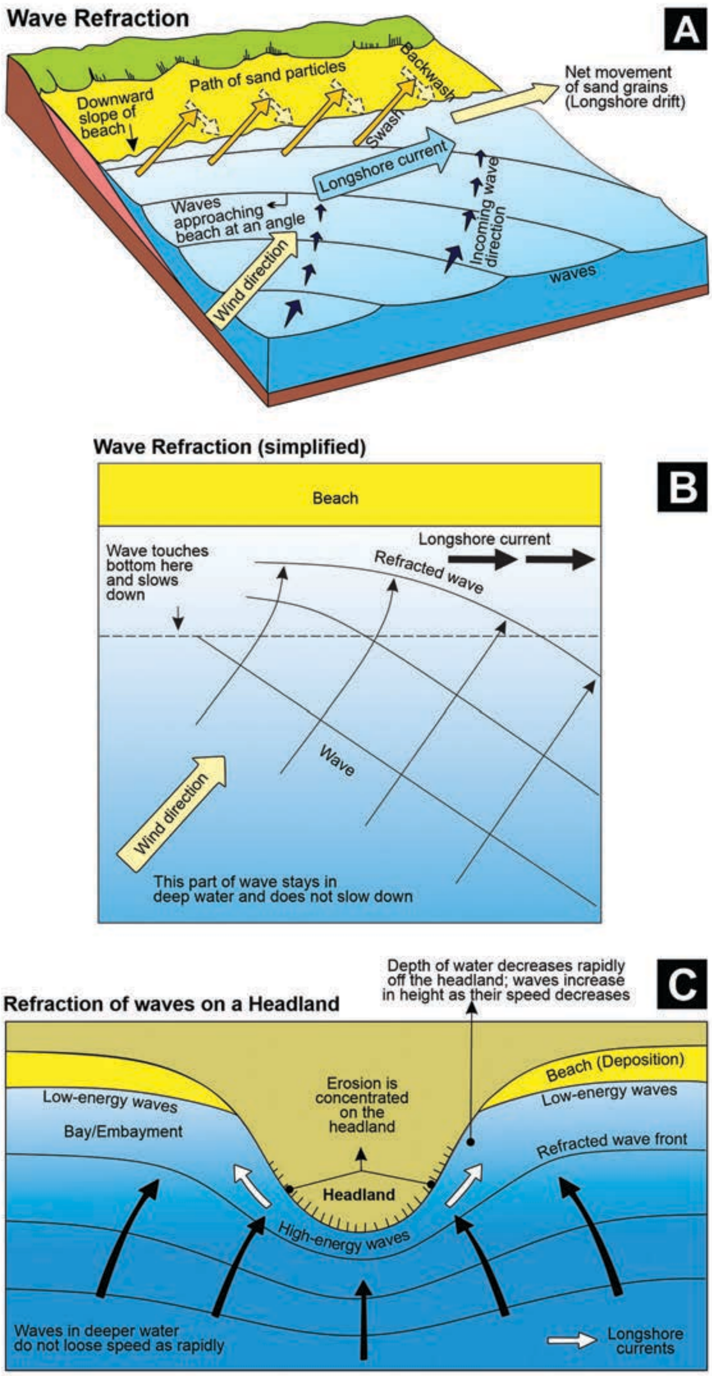

FIGURE 14.9 Wave refraction. A–B: Waves approach the shoreline at an angle with one end of the wave slowing down first, thus causing the waves to bend toward the beach. Such bending of waves as they enter shallower water is called wave refraction. C: The effects of wave refraction tend to concentrate wave energy on headlands that stick out of the coastline and tend to defocus or diverge the energy in the bay (or embayment). Erosion at headlands removes material and transports it to bays/embayments, where wave energy is less. The effect of wave refraction is that it tends to straighten out a coastline over time and smoothens out irregularities.

beach face (i.e., foreshore), and the backwash as it flows back down (Figure 14.9A). This results in a net movement of the water along the beach creating a current called the longshore current (Figures 14.9A–B). Sand stirred up by waves in the surf zone is thus moved along the shore by longshore drift (Figure 14.9A). Friction with the sea bottom as waves approach the beach causes the wave front to become distorted or refracted as velocity is reduced resulting in the waves becoming increasingly parallel to the shoreline. This is because, as waves enter shallower waters, they slow down and bend (Figure 14.9B). The effect of this is that the waves strike the shoreline at a lesser angle than the angle of approach; in other words, the waves strike more nearly parallel to shore, but not completely parallel (Figure 14.9B). Thus, waves in the middle of the bay, where the water is deeper, do not lose velocity as rapidly. Since the paths of the waves in the bay are diverging or spreading out, wave energy is much reduced, allowing deposition to occur. In contrast, the effects of wave refraction tend to concentrate wave energy on headlands or protrusions that stick out of the coastline and tend to defocus or diverge the energy in the bay (or embayment) (Figure 14.9C). This is important as wave refraction affects erosion, transportation, and deposition along the coastline/beach. Erosion at headlands removes material and transports it to bays/embayments, where wave energy is less (Figure 14.9C). The effect of wave refraction is that it tends to straighten out a coastline over time and smoothens out irregularities. Thus, as the coastline matures, it loses all irregularities and becomes increasingly straighter and smoother.

14.4.4 Tides

Tidal currents also contribute to sediment transport on a beach. Tides are controlled by the gravitational pull of the moon, and to a lesser extent by that of the sun (Figure 14.10A). The moon pulls the water in the ocean towards it, creating a bulge, and thereby causing a high tide (Figure 14.10A). Additionally, the moon not only pulls the water but also pulls the earth towards it, thus, causing a second bulge of water and the second high tide on the other side of the earth (Figure 14.10A). Twice a month, the earth, moon, and sun are aligned: this puts an extra gravitational pull on the tidal bulge, to produce an extra high tide called a spring tide (Figure 14.10B). When the sun and moon are at right angles to each other, neap tides occur, when the tidal range is lowest (Figure 14.10B). Thus, during high tide, waves and currents push sediment landward, while during low tide, sediment is deposited as the water recedes (see Figure 14.10B). This tidal action leads to the formation of tidal flats and tidal channels (Figure 14.1A).

14.4.5 Storms

Storm events have a significant impact on beach deposition. High-energy storm waves erode sediments from the beach, causing beach erosion and offshore sediment transport. Conversely, after a storm, sediment can also be re-deposited on the beach, resulting in accretion and the formation of new sandbars and dunes.

14.4.6 Rip Currents

Another coastal phenomenon related to longshore currents is the presence of rip currents (Figure 14.11). These involve the nearshore configuration of the seafloor and/or the arrival of wave trains straight onto the shore. These are narrow currents that move seaward through the surf zone (Figure 14.11). In areas where wave motion pushes water directly toward the beach face, or the shape of the nearshore seafloor refracts and focuses the water movement toward a point on the beach, the water that piles up there, must find an outlet back to the sea (Figure 14.11). The outlet is provided by relatively narrow rip currents that carry the water directly away from the beach (Figure 14.11). During high seas, they act as a mechanism for returning the water back out to sea, and a conduit

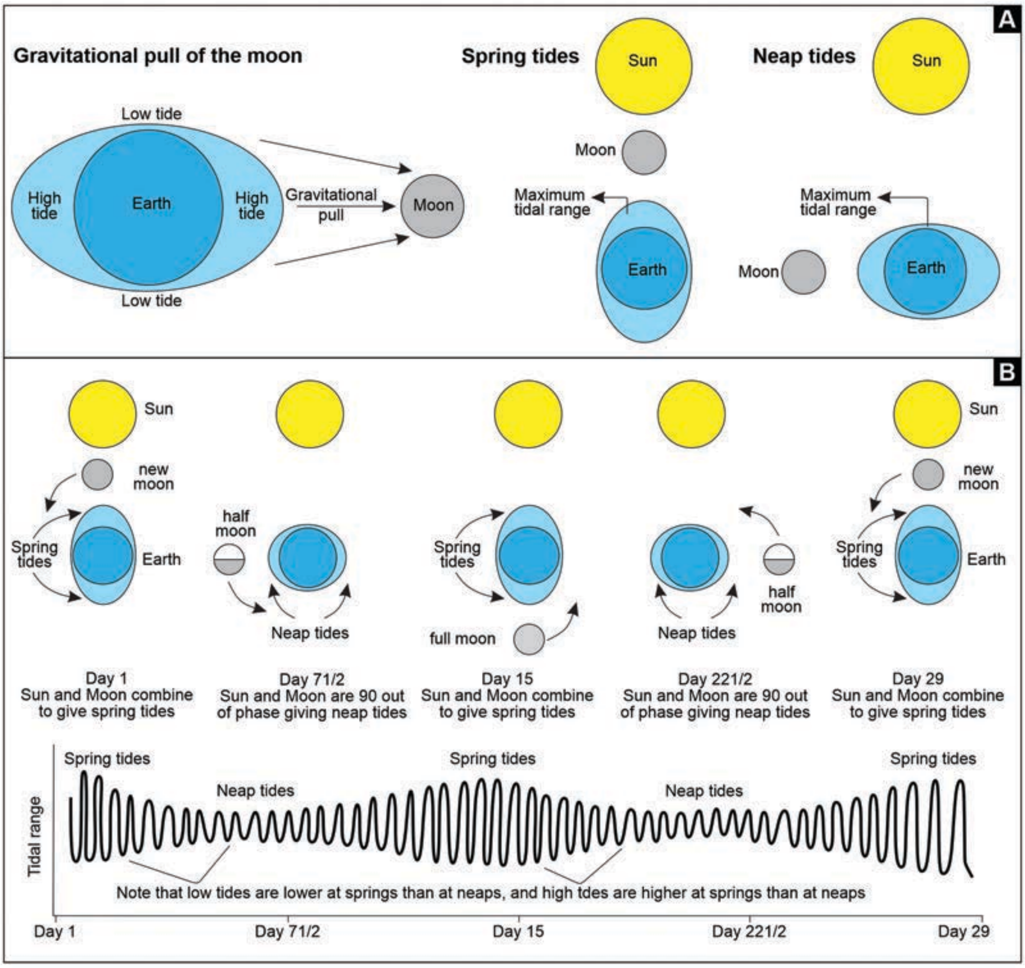

FIGURE 14.10 Tides, and the gravitational pull of the moon, and the sun. Spring tides happen when the sun, moon, and earth are lined up. This alignment causes regular high tides and low tides to be higher. Neap tides, on the other hand, happen when the sun, moon, and earth are at a right angle, thus causing regular high tides and low tides to be lower than usual. Thus, during high tide, waves and currents push sediments landward, while during low tide, sediments are deposited as the water recedes.

to transport seawards, the eroded beach sediments. They are also a major hazard to beachgoers, as swimmers caught in such currents find themselves being carried out to sea (Figure 14.11). Swimmers may attempt to return to shore by swimming directly against the current but this is fruitless as they quickly tire against the strong current. A better solution is to ride it out to where it dissipates, then swim around it and return to the beach, or swim laterally, parallel to the beach, until one is out of the current, and then return to the beach.

14.5 WAVE PROCESSES

Beach wave processes are the interactions between the waves. These interactions play a major role in shaping the morphology and dynamics of coastal areas. The processes are driven by the energy and characteristics of the waves, and the beach's topography and sediment composition. Some key

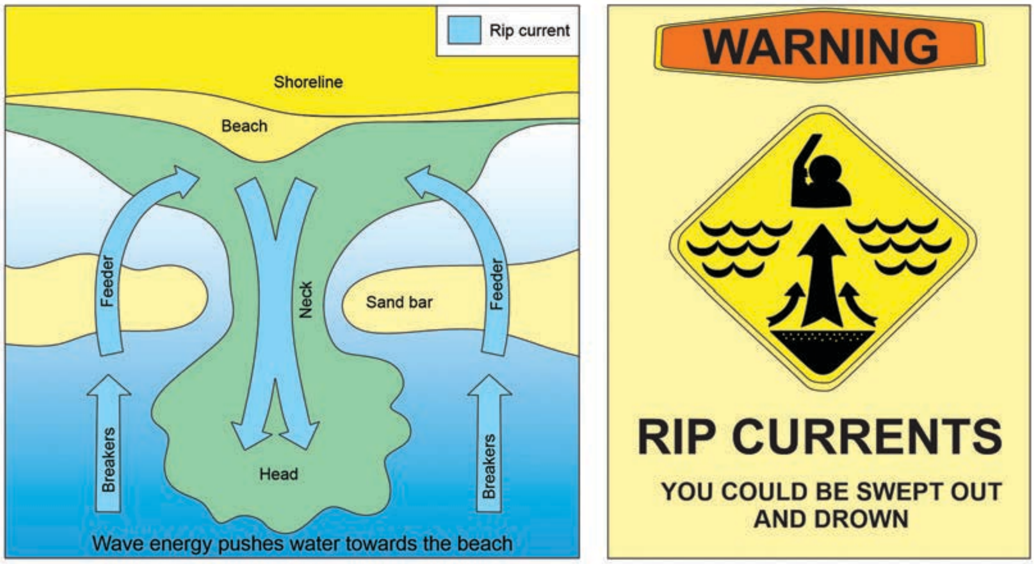

FIGURE 14.11 Rip currents. These are narrow currents that move seaward through the surf zone. In areas where wave motion pushes water directly toward the beach face, or a point on the beach, the water that piles up there, must find an outlet back to the sea. The outlet is provided by relatively narrow rip currents that carry the water directly away from the beach. Thus, they are a major hazard to beachgoers, as swimmers caught in such currents find themselves being carried out to sea.

processes include wave breaking, swash and backwash, sediment transport, wave refraction, wave erosion and deposition, and wave impact. These are briefly enumerated below.

Waves approach the shore and eventually break as they encounter shallow waters; this is called wave shoaling. Breaking waves release their energy, causing turbulence and sediment suspension. The type of wave breaking (including spilling, plunging or surging) depends on the type of beach slope and wave characteristics. In spilling waves, the top of wave crest "spills over" and wave energy is released gradually across the entire surf zone; the slope of the beach is gentle (Figure 14.12A). In plunging waves, the wave crest "curls over" the front of the wave and wave energy dissipates quickly; the slope of the beach is moderately steep commonly noted at shorelines (Figure 14.12B). In surging waves, the wave never breaks as they never attain critical wave steepness, common along upwardly sloping beach faces or seawalls; the wave energy is released seawards (Figure 14.12C).

As waves break, they generate a swash and backwash motion (Figure 14.8). The swash is the forward movement of water and sediment up the beach, while the backwash is the return flow of water and sediment back to the ocean (Figure 14.8). This back-and-forth motion is known as the beach drift and is responsible for sediment transport along the shoreline.

Waves transport sediments along the beach through a combination of processes. The swash carries sediments up the beach face, while backwash carries them back down (Figure 14.8). This results in a net movement of sediments along the beach, known as longshore drift (Figure 14.9A). Sediment can also be transported offshore or onshore depending on wave energy and direction.

When waves approach the shore at an angle, they undergo refraction, causing them to bend and align more parallel to the shoreline (Figures 14.9A–B). This refraction concentrates wave energy on headlands and disperses it in bays, influencing erosion and deposition patterns along the coast (Figure 14.9C).

Waves erode beaches and coastlines through various processes such as via hydraulic action (the force of water against the beach), abrasion (the grinding action of sediment particles), and corrosion

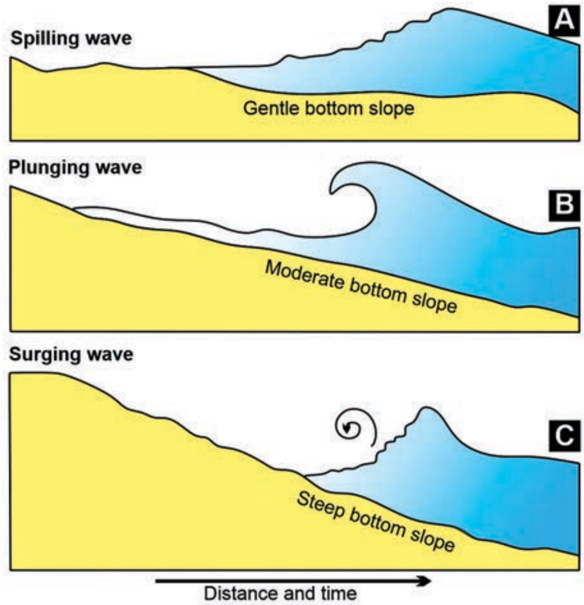

FIGURE 14.12 Types of wave breakers. As the waves approach the shore, they break as they encounter shallow waters. This is called wave shoaling, which is of three types. A: Spilling. The top of the wave crest "spills over" and wave energy is released gradually; the slope of the beach is gentle. B: Plunging. The wave crest "curls over" the front of the wave and wave energy dissipates quickly; the slope of the beach is moderately steep. C: Surging. The wave never breaks as they never attain critical wave steepness; the wave energy is released seawards.

(chemical dissolution of rocks). Wave erosion results in the formation of cliffs, sea caves, and other coastal landforms (Figure 14.3).

Waves also deposit sediments on the beach, contributing to its formation and growth. When wave energy decreases, the sediment settles out of suspension and is deposited, leading to the formation of berms, dunes, and other beach features (Figures 14.7 and 14.8).

As the water gets shallower, the seabed interrupts the circular motion of the water making the waves more elliptical (Figure 14.8A). This causes the crest of the wave to rise up and eventually collapse on to the beach. The water that rushes up the beach is called swash and the water that flows back towards the sea is called backwash (Figure 14.8A). The force of breaking waves has a significant impact on the beach (Figures 14.8B–C). The energy of waves crashing onto the shore causes erosion, sediment displacement, and structural damage to the coastal infrastructure. The intensity of wave impact depends on wave height, period, and the angle of approach.

14.6 BREAKER/SURF ZONE

The breaker zone, also known as the surf zone or surf line, is an area along the shoreline where waves break as they approach the shore (Figure 14.4). It is the most dynamic part of the beach and extends from the breaker zone to the beginning of the foreshore (Figure 14.4A). It is the region where the energy of the waves is dissipated and is transformed into turbulent water motion. When the water depth is ~1.5 times the wave height, the waves break. The breaker zone is characterized by several key features that include breaking waves, turbulence, swash and backwash, sediment transport, and wave energy dissipation. These are briefly enumerated below.

The waves break when the wave crest becomes unstable and collapses, releasing its energy. The type of wave breaking depends on the beach slope, wave characteristics, and water depth. Common types of wave breaking include spilling (breaker on low gradient slope), plunging (breaker on moderate slope), and surging (breaker on steep slope) (Figure 14.12). As waves break, they create turbulent water motion within the breaker zone (Figures 14.4 and 14.6). This turbulence is caused by the interaction between the breaking waves and the irregularities of the seabed. The turbulent flow within the breaker zone results in the suspension and transport of sediment, leading to erosion or deposition along the shoreline (see Figure 14.8). Within the breaker zone, the swash and backwash motion of the waves play an important role as it transports sediments along the shoreline due to its back-and-forth motion (Figure 14.4). As stated above, the swash is the forward movement of water that facilitates sediments up the beach, whereas the backwash is the return flow of water and sediment back to the ocean (Figure 14.8).

Thus, the breaker zone is a significant site of sediment transport. As waves break, they generate strong currents that move sediment along the shoreline (Figure 14.8). The direction and magnitude of sediment transport within the breaker zone depend on wave angle, wave energy, and beach slope. Sediments are transported either parallel to the shore (longshore drift) or offshore (Figure 14.9A). The breaker zone also plays a crucial role in dissipating the energy of the incoming wave. As waves break, their energy is dissipated through turbulence and friction. This energy dissipation helps to protect the beach from excessive erosion and reduces the impact of waves on coastal structures.

14.7 SWASH ZONE

The swash zone is influenced by the movement of water during the swash phase of wave action (Figure 14.6). It is the region where waves wash up onto the beach and move up the slope of the beach face. The swash zone plays a role in shaping the morphology of the beach. The movement of water and sediment during the swash phase leads to the formation of berms, ridges, and other beach features. The swash zone also contributes to the overall profile of the beach slope. Some major characteristics of the swash zone are swash motion, sediment transport, wave run-up, and backwash. These are briefly enumerated below.

The swash zone is defined by the forward movement of water and sediment up the beach (Figure 14.6A). During the swash phase, waves rush up to the beach face, carrying water and sediments with them. The swash motion varies in intensity and this variation depends on wave height, wave angle, and beach slope. As waves rush up the beach, they carry sediment with them, moving it landward. The swash motion thus contributes to the deposition of sediment on the beach, and eventually builds the shoreline (see Figure 14.8B). After the swash motion, there is a backwash phase where water and sediment flow back towards the ocean (Figure 14.8). This backwash motion carries sediments back down the beach face and thus leads to erosion and the sediments being transported, offshore (Figure 14.8C). The swash zone is also characterized by the distance that waves run up the beach face (Figure 14.13). Wave run-up is the maximum distance that waves reach on the beach during the swash phase (Figure 14.13). It can be influenced by factors such as wave height, wave period, beach slope, and the presence of obstacles on the beach.

14.8 WAVE-INDUCED CURRENTS

Wave-induced currents are generated by the action of waves in the nearshore environment. These play a significant role in sediment transport, coastal erosion, and the overall dynamics of coastal areas. Some key characteristics of wave-induced currents include longshore and rip currents, and swash and backwash currents; these have been described above but are briefly mentioned in context of wave-induced currents. Added to this list is the cross-shore current, enumerated below.

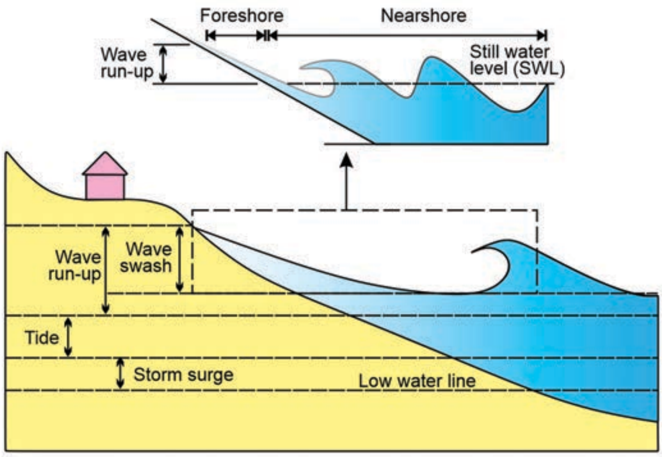

FIGURE 14.13 Wave run-up. This is the maximum distance that waves reach on the beach during the swash phase.

14.8.1 Longshore and Rip Currents

The oblique approach of waves to the coast generates longshore currents (Figure 14.9A). As waves break and rush up the beach, water is pushed along the coast in the direction of wave propagation (Figure 14.9A). These currents transport sediments along the shoreline, leading to the formation of sandbars, spits, and other coastal features (Figure 14.3). Rip currents are narrow, fast-moving channels of water that flow seawards from the shore; these are typically formed when waves break strongly along a shoreline that has a large slope (Figure 14.11).

14.8.2 Swash and Backwash Currents

Swash currents are generated by the swash motion of waves, i.e., the forward movement of water up the beach during the wave's approach (Figure 14.8A). These currents contribute to sediment transport and sediment deposition on the beach (Figure 14.8B). In contrast, backwash currents are the return flow of water and sediment back to the ocean after the swash motion (Figure 14.8C). As waves recede, water and sediments flow back down the beach face. The backwash currents contribute to sediment transport offshore and the erosion of the beach (Fig .14.8C).

14.8.3 Cross-Shore Currents

Cross-shore currents are water movements that occur perpendicular to the shoreline (Figure 14.14). They are influenced by the breaking and reforming of waves as they approach the shore (Figure 14.14). These currents transport sediments landward or seaward, depending on wave conditions and beach morphology (see Huisman et al., 2019).

14.9 BARRIER-ISLAND SYSTEMS

Barrier-island systems are coastal landforms that consist of a long, narrow strip of sand or sediment that runs parallel to the coast and is separated from the mainland by a lagoon or estuary (Figure 14.15). Broadly, the barrier-island system has three subenvironments: (a) barrier-island chain, (b) back-barrier lagoon, and (b) channels that connect the back-barrier lagoon to the open sea

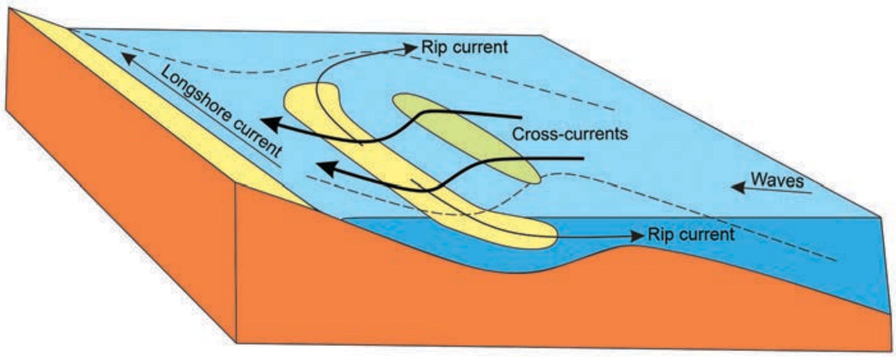

FIGURE 14.14 Cross-shore currents. These water movements occur perpendicularly to the shoreline. Cross-shore currents transport sediments landward or seaward, depending on wave conditions and beach morphology.

(Figure 14.15A) (see also Davis, 1994). Thus, the barrier-island system includes the barrier island itself, a back-barrier lagoon or estuary, and a series of smaller islands or spits (Figure 14.15A). These landforms act as a natural barrier between the open ocean or a large body of water and the mainland, protecting it from the full force of waves, storms, and erosion (Figure 14.15A). They are typically found along low-lying coastlines and are formed through a combination of sediment deposition, wave action, and sea-level changes.

14.10 CHARACTERISTICS OF MODERN BEACH AND BARRIER-ISLAND SYSTEMS

Modern beach and barrier-island systems are dynamic coastal environments that are characterized by several distinct features and processes. Some of the key characteristics include: sand-dominated beaches, dune systems, sediment transport, and coastal processes. These are briefly enumerated below.

Modern beaches and barrier islands are typically composed of fine to medium-grained sand that is constantly being shaped and reshaped by waves and currents. The sand is often well-sorted and can exhibit a range of colors, depending on the mineral composition. These sands originate on the beach shoreface, foreshore and backshore (see Figures 14.15B and 14.16); such deposits are often hundreds of meters wide, hundreds of kilometers long, and 10–20 meters thick (see also Frey et al., 2011).

Beaches and barrier islands often have dune systems that are formed by windblown sand (see also Figure 14.1C–D). Dunes provide a natural barrier against erosion and help to stabilize the coastline. They vary in height and shape depending on the prevailing wind patterns.

Waves and currents play a crucial role in the movement of sediment along the coast. Sediment is transported alongshore, driven by the prevailing wave direction, and can accumulate in certain areas, resulting in the formation of beach and barrier-island systems (Figure 14.15A).

Modern beach and barrier-island systems are subject to a range of coastal processes, including erosion, deposition, and sediment reworking. These processes are influenced by factors such as wave energy, tidal currents, sea-level fluctuations, and storm events.

14.11 BEACH DEPOSITS

Beach deposits are accumulations of sediments, such as sand, gravel, and shells, along the shoreline of a beach or beach face or foreshore (see Figures 14.5 and 14.16). This area extends from the low to the mean high tide (swash zone; see Figure 14.4A). The formation of beach deposits

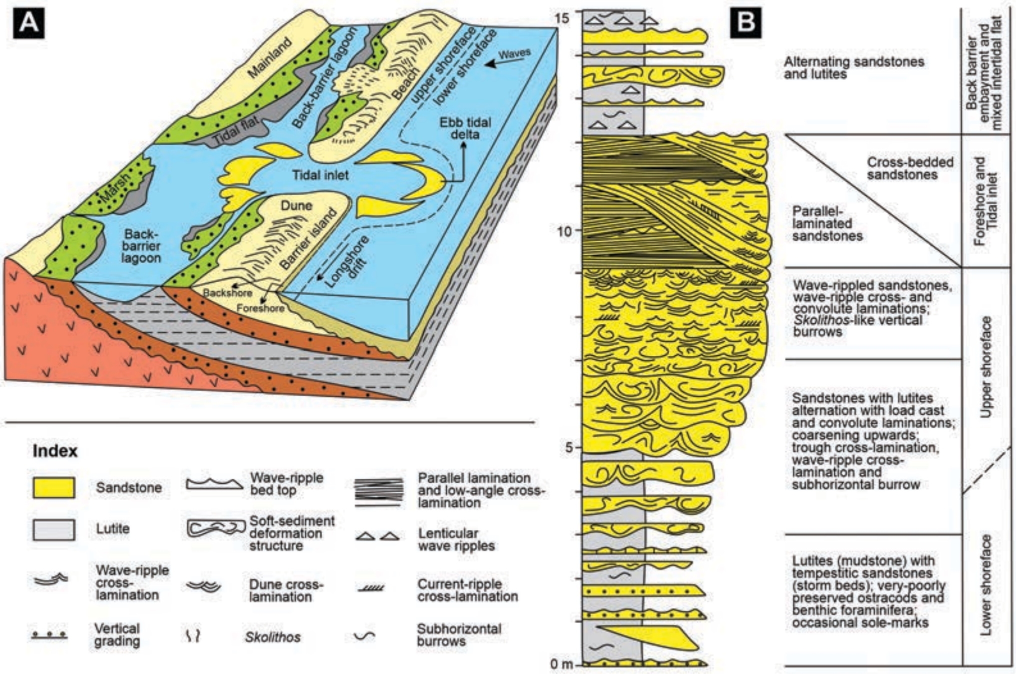

FIGURE 14.15 Barrier-island system, subenvironments, and deposits. A: Associated coastal landforms. (Modified after Walker, 1979.) B: Vertical succession of facies, inchospecies and sedimentary structures of a sandy beach deposit. (Modified after Payros et al., 2022.)

begins with the erosion and weathering of rocks and sediments in the hinterland. These particles are then transported by rivers and streams to the coast, where they are further transported along the shoreline by waves and currents. As the energy of the waves decreases near the shoreline, the sediments settle and accumulate, forming beach deposits. The shape and characteristics of beach deposits vary depending on various factors, such as wave energy, sediment supply, and coastal morphology. For example, high-energy beaches with strong waves tend to have coarser sediments and steeper beach profiles, while low-energy beaches with weaker waves have finer sediments and gentler slopes. Beach deposits are typically composed of well-sorted and well-rounded particles, as they have been subjected to the constant agitation and sorting action of waves. The size and composition of the sediments vary, depending on the local geology and the energy of the waves. In general, foreshore sediments are fine to medium-grained sands with some pebbles and gravel layers (Figure 14.16). The sedimentary structures include parallel laminations that gently dip (2–3°) seawards (see Figures 14.4B and 14.16).

14.11.1 Backshore Deposits

Backshore refers to the land area on a beach that lies between the high-tide mark and the dunes or coastal vegetation (see Figures 14.4A and 14.16). It is the part of the beach that is typically dry and above the reach of normal wave action (see Figures 14.4A and 14.16). The backshore is characterized by its relatively stable and less dynamic nature as compared to the foreshore and nearshore zones of the beach. The backshore is influenced by a combination of factors, including wave action, tides, and wind. While it is generally above the high-tide mark, it can still be affected by storm surges and exceptionally high tides during extreme weather events. Backshore deposits

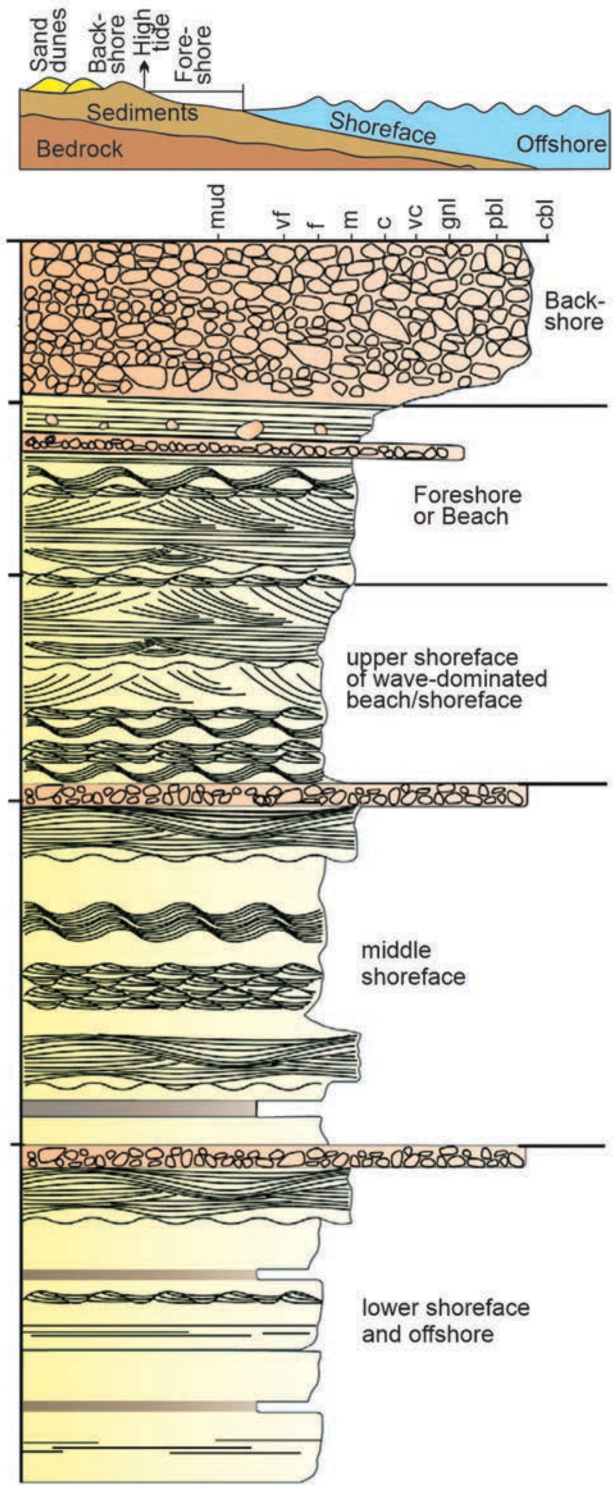

FIGURE 14.16 Sedimentary structures and lithology of a modern beach with its subenvironments. (Modified after Frey and Dashtgard, 2011.) Abbreviations: vf: very fine-grained sand; f: fine-grained sand; m: medium-grained sand; vc: very coarse-grained sand; gnl: granule; pbl: pebble; cbl: cobble.

consist mainly of interbedded gravel and sand, with crude, low angle seaward-dipping bedding (see Figure 14.16) (see also Payros et al., 2000; Frey et al., 2011).

14.11.2 Shoreface Deposits

Shoreface deposits are sedimentary deposits that form in the shallow waters of the ocean, just offshore from the beach (Figures 14.5 and 14.16). This zone is typically located between the low-tide mark and the point where waves break and become turbulent (= breaker zone). Shoreface deposits are composed of a mixture of sand, silt, and clay that are transported and deposited by waves and currents (Figures 14.5 and 14.16). These sediments are often well-sorted and well-rounded due to the sorting action of the waves. The size and composition of the sediments varies depending on factors such as wave energy, sediment supply, and the geology of the coastal area. The shoreface is categorized into upper, middle and lower shoreface (Figures 14.5 and 14.16). Upper shoreface deposits mainly consist of trough cross-beds with planar laminations, whereas the middle-lower shoreface deposits are formed under relatively lower-energy conditions with planar laminations and hummocky cross-stratification (Figures 14.5 and 14.16).

14.12 BACK-BARRIER DEPOSITS

Back-barrier sediments are formed in the area between a barrier island or spit and the mainland (Figure 14.15). This region is typically a low-lying, marshy area located behind the primary coastal dune or beach. The back-barrier sediments are composed of a mixture of fine-grained materials, such as silt, clay, and organic matter (Figure 14.15B). They are often deposited by tidal currents, riverine inputs, and wind-driven processes. These sediments can accumulate in the form of mud flats, marshes, and tidal channels within the back-barrier zone (Figure 14.15A). Tidal currents play a significant role in transporting and depositing fine-grained sediments within the back-barrier, while riverine inputs can introduce additional sediment and nutrients into the system. Wind-driven processes, such as aeolian transport, also contribute to the deposition of sediments within the back-barrier. In general, there are several subenvironments and deposits within the back-barrier such as washover, tidal-channel, tidal-delta, tidal-flat and lagoon and marsh (Figure 14.15A). These are briefly enumerated below.

14.12.1 Washover Deposits

Washover deposits, also known as washover fans or washover sediments, form when waves or storm surge overtop a barrier island and deposit sediments on the landward side (Figure 14.17A). Washover deposits often form fan-shaped or lobate deposits that extend landward from the breach point of the barrier (Figure 14.17B). These deposits vary in thickness, composition, and grain size depending on the characteristics of the storm event and the sediment available for transport. During a storm event, high-energy waves and storm surge breach the natural barrier of a coastal system, such as a barrier island or dune, and flow onto the landward side. As the waves and surge recede, they leave behind sediments that they have transported and deposited during the overwash process. This sediment includes sand, shells, organic matter, and other debris. Thus, washover results in the deposition of (a) sand on and landward side of the beach crest, (b) large fan-shaped deposits on back-barriers, (c) large sheet-like deposits over an entire barrier, (d) sand deposition into back-barrier waterways, or (e) may even lead to breaching of coastal barriers. The morphological deposits occurring during washover include (a) washover fan, (b) washover terrace, and c) sheet wash deposits (see Figure 14.17B) (see also Donnelly et al., 2004). Thus, washover deposits contribute to the accretion and growth of barrier islands by adding sediments to the landward side, while also providing evidence of past storm events and sea-level changes. These deposits also provide valuable information about the dynamics of coastal systems and their response to storm events.

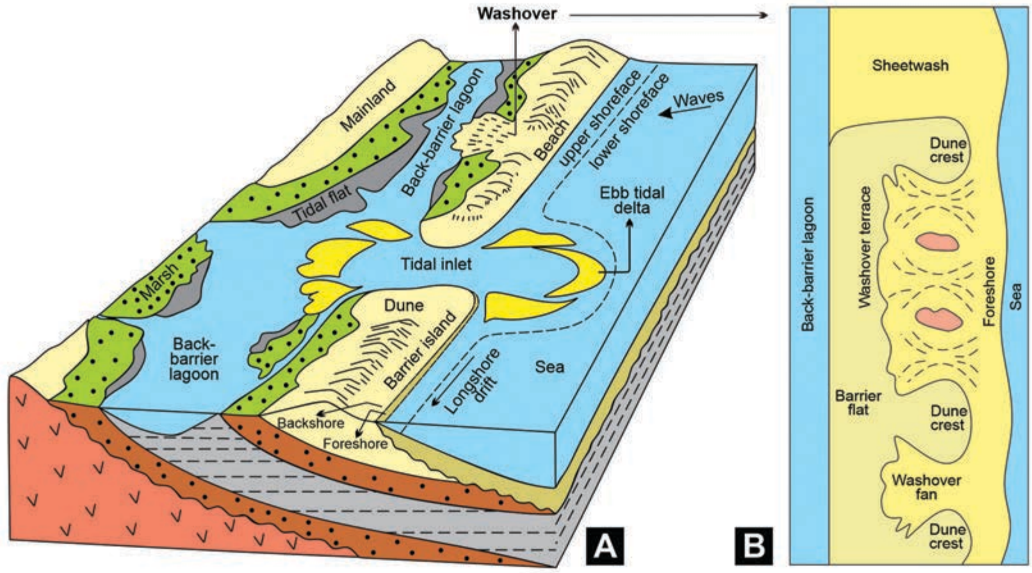

FIGURE 14.17 Barrier-island subenvironments and washover deposits. A: Barrier-island subenvironments showing the position of washover. B: Types of washover deposits. These deposits form when a wave or storm surge overtops a barrier island and deposits sediments on the landward side.

14.12.2 Tidal-Channel Deposits

Tidal-channel deposits form within tidal channels; these are dynamic, narrow, and elongated channels influenced by regular tidal fluctuations/currents (see Figure 14.18). These deposits are typically composed of a mixture of sand, silt, clay, and organic matter. As tidal currents flow through these channels, they transport and deposit sediments, leading to the formation of tidal-channel deposits. In areas with strong tidal currents, the deposits may be coarser and better well-sorted, consisting primarily of sand. In contrast, areas with weaker currents have finer-grained deposits, including silt and clay. Tidal-channel deposits often exhibit distinct sedimentary structures, such as cross-bedding, ripple marks, and mud drapes. These structures provide evidence of the dynamic nature of tidal currents and can be used to interpret the flow direction and the intensity of currents within the channel. Thus, tidal-channel deposits serve as archives of past environmental conditions, recording changes in sea level, sedimentation rates, and the evolution of tidal channels, themselves, over time.

14.12.3 Tidal-Delta Deposits

Tidal-delta deposits form at the mouth of a tidal inlet or estuary, where the interaction between tidal currents and fluvial (river) flow leads to the accumulation of sediment (Figure 14.18). These deposits are characterized by their distinctive sedimentary structures and composition, which are influenced by the dynamics of tidal and fluvial processes. In the ebb-tidal discharging flow, leaving the inlet, the flow velocities decrease and sediment is deposited, thus creating the ebb-tidal delta (Figure 14.18). In contrast, the flood-tidal delta is created by sediment deposited by the inflowing flood current leaving the inlet; the flood-tidal delta may be fan- or horseshoe-shaped (Figure 14.18). Consequently, basins with large tidal ranges (and limited wave influence) have well-developed ebb- and flood-tidal deltas (Figure 14.18). The overall morphology of the tidal deltas, especially the ebb-tidal delta, depends on the combined action of waves and tides. The dominant wave action moves sediment onshore and limits the area over which the ebb-tidal delta can spread out. Hence,

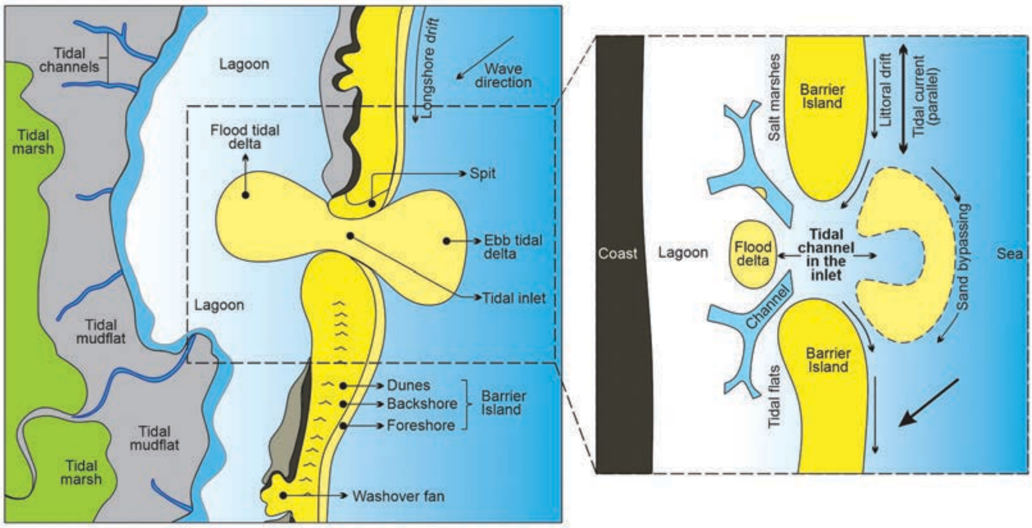

FIGURE 14.18 Characteristics and depositional setting of a lagoon within the barrier-island system. (Modified after Davis, 1994.)

for tide-dominated entrances, viz. with limited wave action, ebb-tidal deltas tend to be large relative to the flood-tidal delta. For wave-dominated entrances, the ebb-tidal deltas tend to be relatively small, whereas the flood-tidal deltas are well-developed with flood shoals that can be emergent at low tide. The sedimentary structures found in tidal-delta deposits are cross-bedding, ripple marks, and channel-fill deposits that reflect the movement and deposition of sediments by tidal and fluvial currents. Tidal-delta deposits may also exhibit mud drapes that form when fine-grained sediments settle out of suspension and accumulate on the delta surface.

14.12.4 Tidal-Flat Deposits

Tidal-flat deposits form in intertidal areas that are exposed during low tide and submerged during high tide (Figure 14.19A). Tidal flats are dynamic environments that experience the influence of both tidal and fluvial processes (Figure 14.19B). Three zones are noted in a tidal flat (Figures 14.19 and 14.20): a) the supratidal zone, located above high-tide mark, b) the intertidal zone, located between high- and low-tide marks, and c) the subtidal zone which occurs below low-tide mark and is rarely exposed (see Figures 14.19 and 14.20). The intertidal and supratidal zones lie landward of the subtidal zone (Figures 14.19 and 14.20) (see also Flemming, 2005; Plint, 2010).

In general, the tidal-flat deposits are typically composed of fine-grained sediments, such as mud, silt, clay and sand (Figure 14.19B). If wave energy is low, then fine-grained sediments such as mud is deposited in its upper reaches (mud flats; see Figures 14.19B and 14.20). If wave energy is higher, coarse-grained sediments such as sand is deposited in the lower reaches (sand flats; see Figures 14.19B and 14.20). In general, the sediments on tidal flats are often well-sorted and fine-grained due to the low-energy environment and the settling of particles out of suspension. The sedimentary structures found in tidal-flat deposits are indicative of the processes at work in these environments. These structures include mud cracks, ripple marks, cross-bedding, lenticular bedding, and lamination, which reflect the alternation between wet and dry conditions during tidal cycles (see Figures 14.19B and 14.20B). Additionally, tidal-flat deposits may also contain distinctive suits of trace fossils that characterize various zones of a tidal flat (see Figure 14.21) (see also Mángano and Buatois, 2004; Mángano et al., 2002; Desjardins et al., 2012). Desjardins et al.

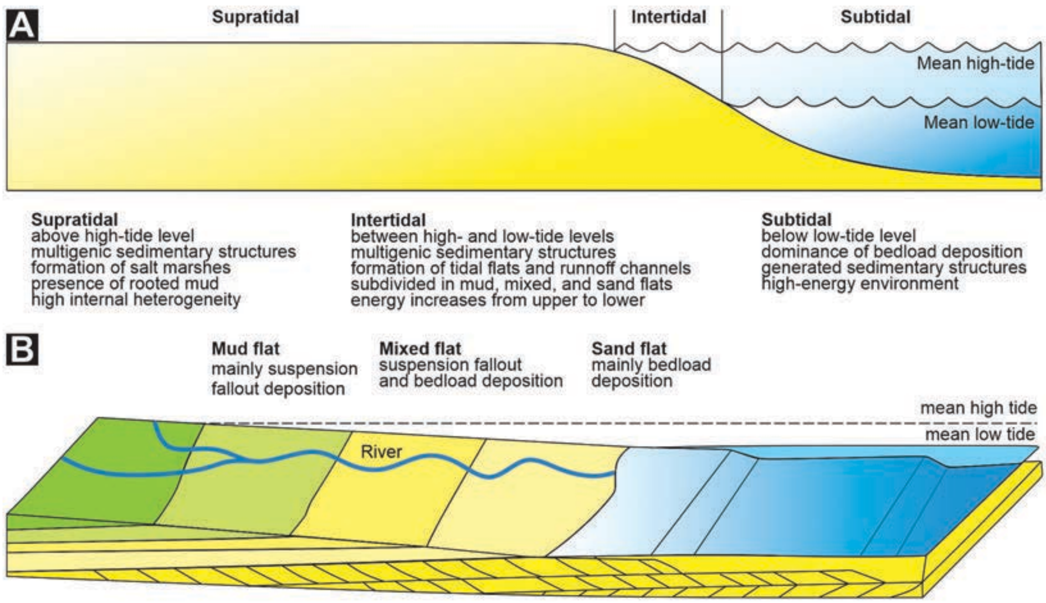

FIGURE 14.19 Morphology of tidal flat and depositional setting. A: Tidal-flat components and their associated facies. B: Distribution of sand bodies in a tidal flat. (Modified after Desjardins et al., 2012.)

(2012) noted that in a fining-upward sequence, the replacement of the *Skolithos* ichnofacies by the *Cruziana* ichnofacies reflects tidal-flat progradation (Figure 14.21). Generally, in shallow tide-dominated settings, *Skolithos* ichnofacies dominate lower intertidal and shallow subtidal deposits, whereas the *Cruziana* ichnofacies is best noted in middle and upper intertidal environments, and in low-energy settings of the inner shelf (Mángano and Buatois, 2004; Desjardins et al., 2012) (see Figure 14.21).

14.12.5 Lagoonal and Marsh Deposits

Lagoonal and marsh deposits form in lagoons and marshes, which are shallow, brackish water environments found along coastlines (see Figures 14.1A–B and D). Lagoonal deposits typically consist of fine-grained sediments, such as mud, silt, and clay that settle out of suspension in the calm, low-energy lagoon environment. These sediments are often well-sorted and exhibit laminations and ripple marks, reflecting the alternating wet and dry conditions of tidal cycles. Lagoonal deposits may also contain biogenic structures, such as shell fragments, burrows, and organic matter (Figure 14.1D).

Marsh deposits (see Figure 14.1B), on the other hand, are composed mainly of organic material, including plant remains and peat as they are characterized by the growth of vegetation, such as grasses and reeds that trap and accumulate organic matter (see Figure 14.1D). Over time, this organic material undergoes partial decomposition, leading to the formation of peat. Marsh deposits may also contain fine-grained sediments, such as mud and silt that are carried into the marsh by tidal or fluvial processes.

Lagoonal and marsh deposits preserve records of past environmental conditions, including changes in sea level, sedimentation rates, and the influence of climate. These deposits also serve as reservoirs for hydrocarbons, such as oil and gas, as the fine-grained sediments and organic matter provide favorable conditions for their accumulation and preservation.

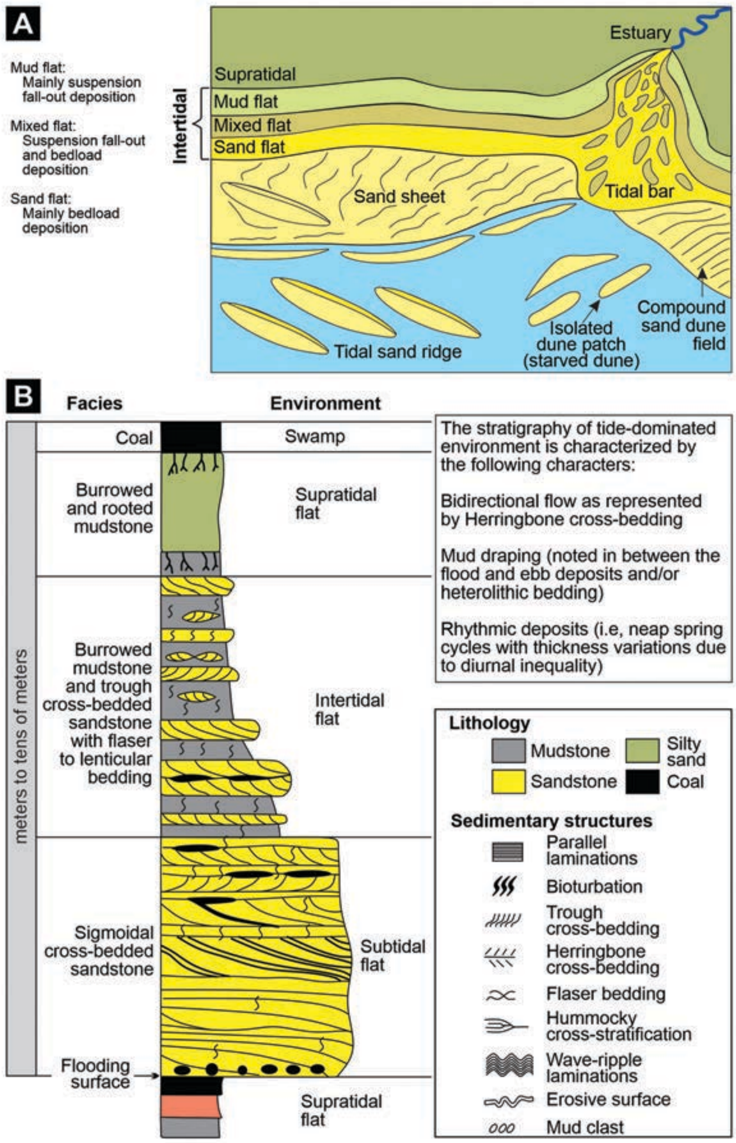

FIGURE 14.20 Tidal-flat environment and distribution of sand bodies. A: Tide-dominated shallow marine environments and the distribution of sand-body in it. B. An upward fining tidal deposit with deposition facies and sedimentary structures of subtidal flat, intertidal flat and supratidal flat. (Modified after van Wagoner et al., 1990.)

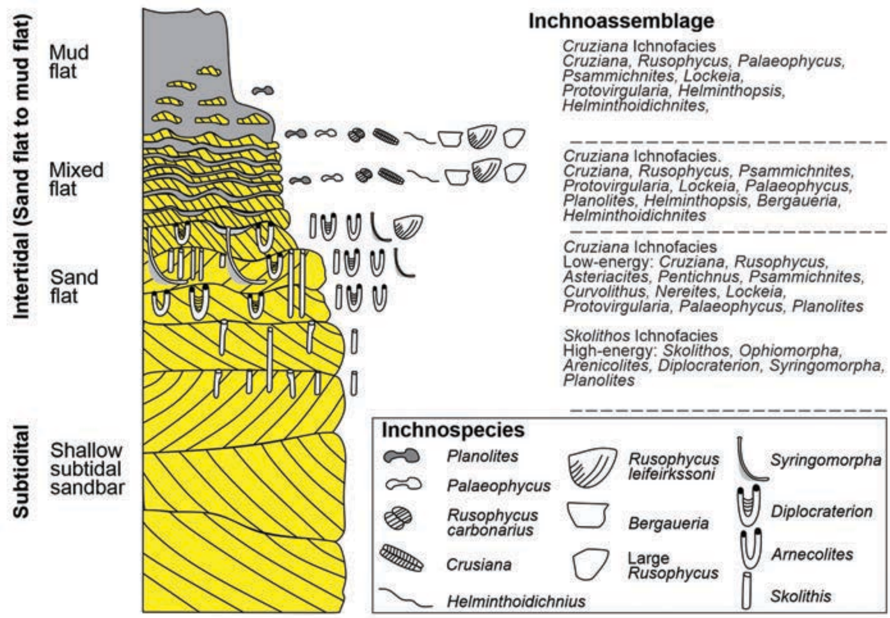

FIGURE 14.21 Fining-upward sequence and associated ichnospecies. The figure shows the replacement of the *Skolithos* ichnofacies by the *Cruziana* ichnofacies. *Skolithos* assemblage is the dominant ichnofacies in shallow subtidal and lower intertidal settings whereas in middle and upper intertidal setting by the *Cruziana* assemblage within the shallow tide-dominated settings of the inner shelf. (Modified after Mángano and Buatois, 2004; Desjardins et al., 2012.) This change in ichnofacies (from *Skolithos* to *Cruziana*) also reflects tidal-flat progradation.

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15 Estuarine System

15.1 INTRODUCTION

Estuaries are semi-enclosed coastal bodies of water where freshwater from rivers and streams meets and mixes with saltwater from the ocean (Figure 15.1) (see also Boyd et al., 1992; Dalrymple et al., 1992). Estuaries, in a general sense, are considered as the lower courses of rivers open to the sea (Figure 15.1). Hence, this unique environment is influenced by both marine and terrestrial processes, resulting in a wide range of geological features and dynamics. One of the primary geological processes in estuaries is sedimentation. Estuaries receive large amounts of sediment from rivers and streams, which is transported and deposited within the estuarine system forming mudflats, sandbars, and tidal channels (Figure 15.1). Tidal dynamics also plays a significant role in estuarine geology. The rise and fall of tides in estuaries create strong currents and water flow patterns. These tidal currents erode and shape estuarine bottom, creating channels and shoals. Tidal forces also influence the movement and distribution of sediment within the estuarine system.

The term "estuary" comes from the Latin word *aestus*, meaning tide, and from the adjective *aestuarium*, meaning tidal. Many authors view estuaries as tidal mouths of rivers (see Pethick, 1984) (Figure 15.1). Estuaries have been defined and classified variously (see Cameron and Pritchard, 1963; Pritchard, 1967; Caspers, 1967; Fairbridge, 1980; Dalrymple et al., 1992; Cooper, 1993; Dyer, 1997; Perillo, 1995; Townend, 2005). Some authors consider "estuary" to be synonymous with "inlet," "bay," "sound" or "lagoon"; "freshwater estuary" has been used to describe the transition zone where a river enters a lake, and where mixing of freshwater with the slightly differing chemical characteristics takes place (Figure 15.1B). Dalrymple et al. (1992, p. 1132) defined an estuary as "the seaward portion of a drowned valley system which receives sediment from both fluvial and marine sources and which contains facies influenced by tide, wave, and fluvial processes. The estuary is considered to extend from the landward limit of tidal facies at its head, to the seaward limit of coastal facies at its mouth" (see also Dalrymple and Choi, 2007). Thus, estuaries form only in the presence of a relative sea-level rise (i.e., transgression) (see Figure 15.1) (see also Boyd et al., 1992; Dalrymple et al., 1992). Progradation fills and destroys estuaries, causing them to change into deltas (Figure 15.1B). Dalrymple et al. (1992) suggest that the net landward movement of sediment derived from outside the estuary mouth distinguishes estuaries from deltas.

15.2 CHARACTERISTICS OF ESTUARIES

Major physiographic, hydrologic, and sediment characteristics of estuaries are briefly summarized below. The physiographic characteristics include: (a) they are shallow and wide-mouthed shaped with broad intertidal areas and gradually sloping bottoms; (b) they vary in size from small coastal inlets to large, complex systems; (c) they have irregular shorelines with numerous bays, coves, and

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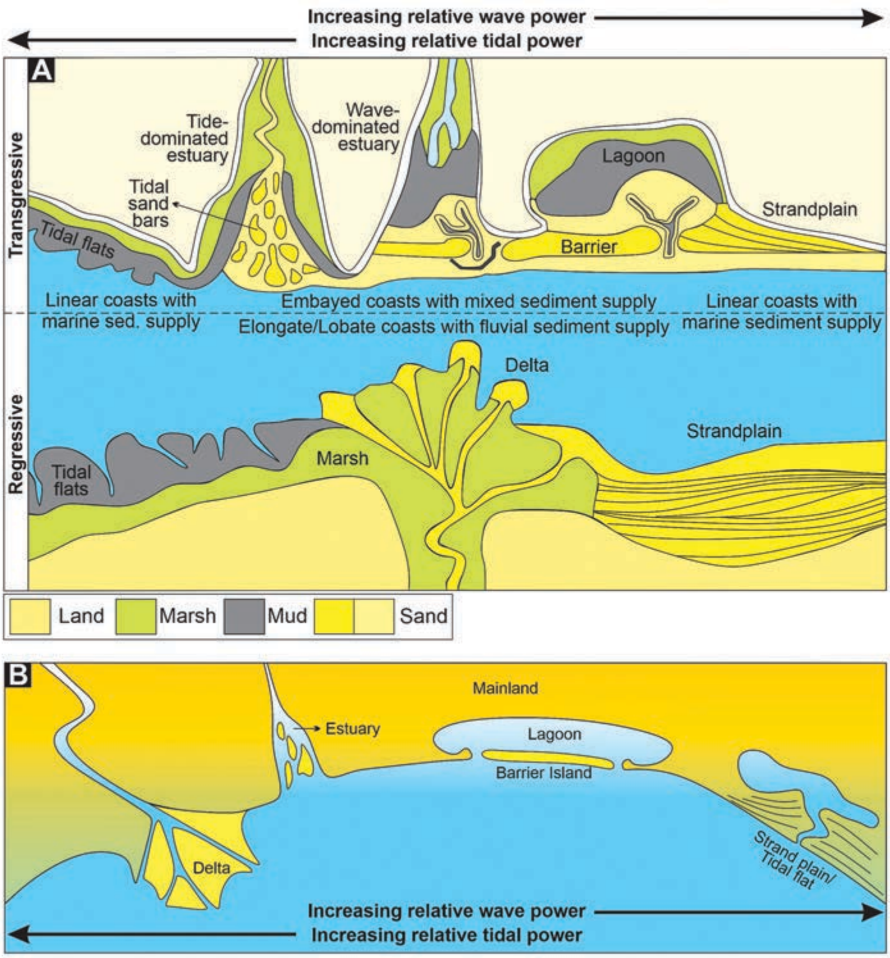

FIGURE 15.1 Coastal depositional environments within the marginal marine setting. A: Figure shows the increasing influence of tides (to the left) and waves (to the right). (Modified after Boyd et al., 1992.) B: Coastal depositional environments within the marginal marine setting showing the increasing influence of waves (to the left) and rivers (to the right).

tidal creeks; and (d) the shape and size of an estuary influence water circulation patterns, sediment deposition, and the distribution of habitats within the system.

Major hydrologic characteristics include: (a) estuaries experience a dynamic flow of water due to the interaction between freshwater and saltwater; this mixing is influenced by tides, river discharge, and wind; (b) tides play a crucial role, causing the rise and fall of water levels; (c) the tidal range varies depending on the location and the shape of the estuary; (d) river discharge brings freshwater and sediment into the estuary, affecting salinity levels and nutrient inputs; (e) they have stratified water columns, with fresher water on top and saltier water at the bottom, thus, impacting the distribution of organisms and the transport of sediment within the system.

Major sediment characteristics include: (a) estuaries are sediment-rich environments, receiving inputs from rivers, coastal erosion, and offshore sources; (b) the sediment varies in grain size, ranging from fine silt and clay to coarser sand and gravel; (c) hydrodynamic forces in the estuary, such as tidal currents and waves influence sediment deposition and erosion patterns; (d) they often have distinct sedimentary environments, including mudflats, sandbars, salt marshes, and tidal channels, and (e) sediment composition and characteristics affect water clarity, nutrient cycling, and the distribution of benthic organisms in the estuarine system.

15.3 ESTUARY CLASSIFICATION

15.3.1 Based on Geomorphology

Pritchard (1952) identified four types of estuaries (Figures 15.2A–D): (a) coastal plain, (b) fjords, (c) bar-built, and (d) tectonic estuaries (Figures 15.2A–D) (see also Fairbridge, 1980). Coastal plain estuaries (Figure 15.2A) are low-relief, funnel-shaped drowned river valleys that are open to the sea and set within a wide coastal plain such as the Chesapeake Bay and Delaware Bay in the eastern USA (see also Valle-Levinson, 2010). Fjords (Figure 15.2B) are high-relief estuaries with a U-shaped valley profile that are formed from marine flooding (drowning) of glacially over-deepened (eroded) troughs (valleys) during the Holocene sea-level rise such as noted in northwestern USA, Canada, and northern Europe. Fjards (also called firths) are related to fjords but have much lower relief. Bar-built estuaries are submerged valleys or bays that have been partially enclosed by the development of a bar or spit across the entrance (Figure 15.2C). These are low-relief estuaries that are L-shaped in plan-view with lower courses parallel to the coast (Figure 15.2C). Blind estuaries, on the other hand, are seasonally blocked by longshore drift or dune migration. Both are very similar to lagoons and some workers even consider them as lagoons (Figure 15.2C). Tectonic or compound estuaries (Figure 15.2D) result from faulting and land subsidence, such as noted in the San Francisco Bay, Manukau Harbour in New Zealand, and the Rias of northwest Spain. In general, these are flask-shaped, high-relief ria backed by a low-relief plain created by tectonic activity. Ria (Figure 15.2E) are moderate relief estuaries that develop in winding valleys; these are drowned river valleys that remain open to the sea (Figure 15.2E). However, Pye and Blott (2014) noted that many of the features included within the Pritchard (1952) classification are not estuaries according to the strict definition which requires significant river influence, not only in terms of salinity but also sedimentary processes and morphological development. Pye and Blott (2014) regarded tidal inlets, bays, lagoons, straits, and sounds, which receive little or no freshwater, as separate types of water bodies. However, they also noted that a clear distinction between estuaries and tidal inlets, bays, lagoons and deltas is always not possible as a continuum of process domains and landforms exists (see also Davies, 1977; Dalrymple et al., 1992; Pye and Blott, 2014). Hence, Pritchard's (1952) classification of estuaries is followed here (Figure 15.2).

15.3.2 Based on Physical Processes

Dalrymple et al. (1992), in contrast to the previously proposed salinity-based tripartite division of estuaries by Rochford (1951) and Fairbridge (1980), proposed a similar model but based on physical processes such as hydrologic characteristics and the kinds of sediment and sediment bodies formed within an estuary (see Figure 15.3). Dalrymple et al. (1992) proposed three zones: a) an outer zone dominated by marine processes (waves and/or tidal currents); b) a low-energy central zone, where both marine energy (generally tidal currents) and river currents play a role; and c) an inner, river-dominated zone (Figure 15.3). The tripartite division of Dalrymple et al. (1992) corresponds with the patterns of net bedload transport, where the river-dominated inner zone has the coarsest sediments and the central zone typically contains the finest-grained bedload sediments present in the estuary, regardless of whether the estuary is wave- or tide-dominated (see Figures 15.3C–D).

Dalrymple et al. (1992) noted a coarse-fine-coarse distribution of lithofacies in most wave-dominated estuaries (Figure 15.3G). A marine sand body accumulates at the mouth in the high wave-energy domain (transgressive subtidal shoals and/or washover deposits). Sand and/or gravel is also deposited at the head of the estuary by the river, forming a bay-head delta. The low-energy central

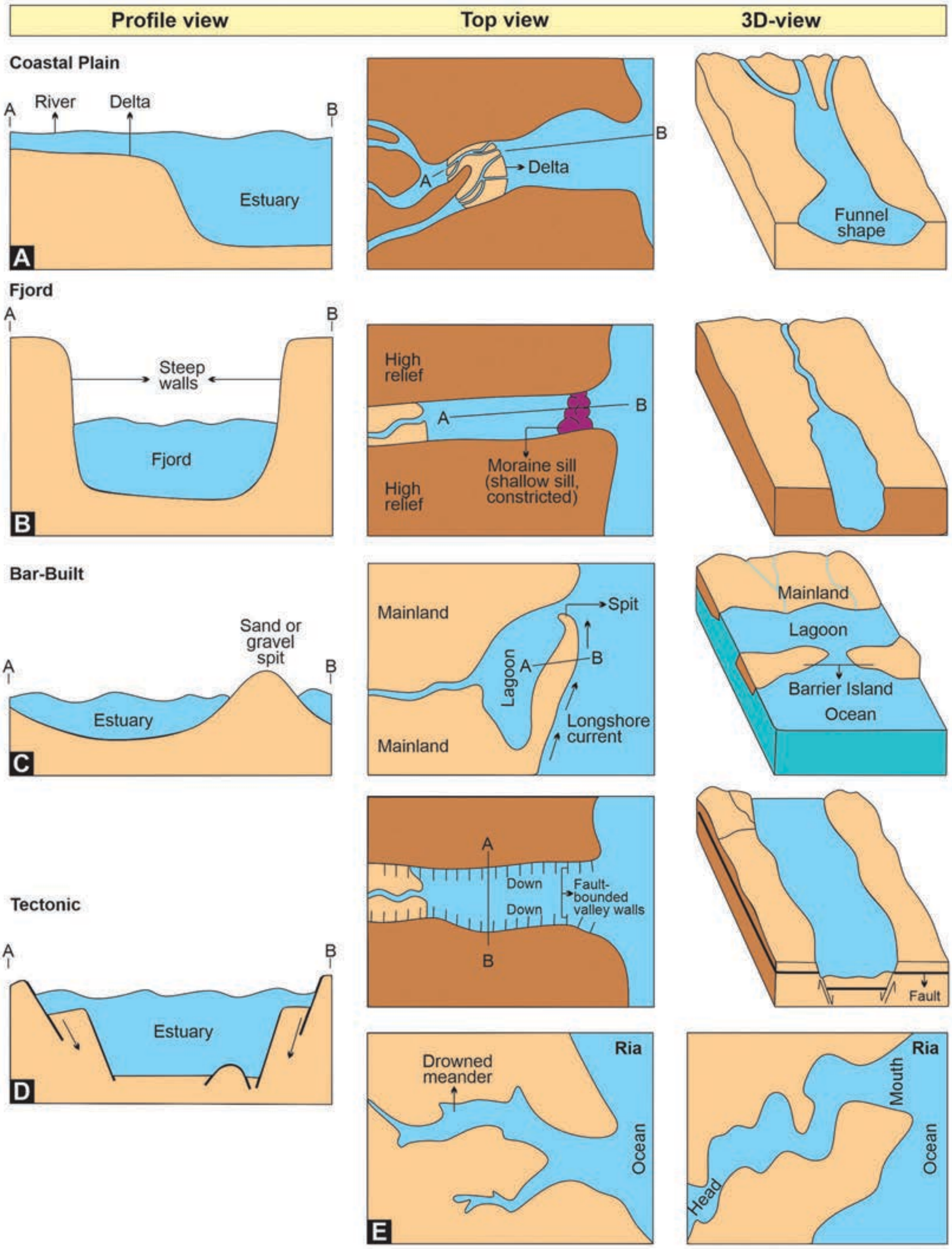

FIGURE 15.2 Types of estuaries.

part (central basin) acts as the prodelta region where fine-grained organic muds accumulate. The equivalent area of shallow (nearly filled) estuaries contains extensive salt marshes and is crossed by tidal channels that pass directly into the river channels. The funnel-shaped tide-dominated estuaries (Figures 15.3A–C) are less well known than their wave-dominated counterparts (Figure 15.3E–G).

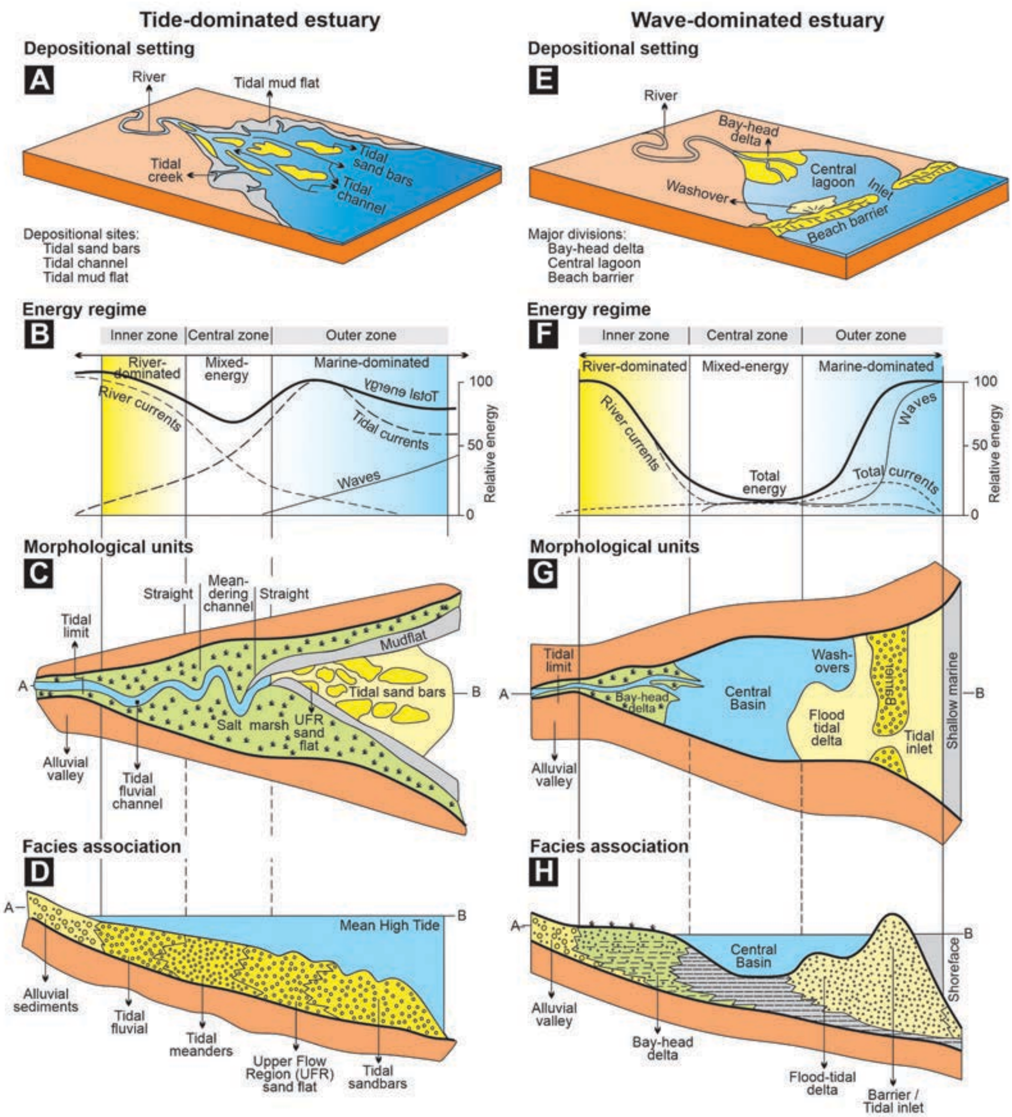

FIGURE 15.3 Depositional setting, energy regime, morphological units, and facies associations. (Modified after Pritchard, 1967; Dalrymple et al., 1992; Boyd et al., 1992; FitzGerald et al., 2015.) A–D: Tide-dominated estuary. E–H: Wave-dominated estuary.

Tidal-current energy exceeds wave energy at the mouth of tide-dominated estuaries, and elongated sand bars are developed (Figure 15.3B). Fluvial (river) energy decreases seaward as in wave-dominated systems. The tripartite facies distribution is not as obvious as in wave-dominated estuaries; the sands occur in the tidal channels that run along the length of the estuary (Figure 15.3C). Nevertheless, the central zone (energy minimum) is the site of the finest channel sands. Muddy sediments accumulate in tidal flats and marshes along the sides of the estuary (Figures 15.3B and D).

The second facies, which coincides with the tidal-energy maximum, consists of upper-flow-regime (UFR) sand flats that display a braided channel pattern where the estuary is broad but becomes confined to a single channel further headward (Figure 15.3B). The facies consist of parallel-laminated fine sands. The low-energy central zone includes the "straight-meandering-straight" channel pattern

(Figure 15.3B). The outer straight reach is tidally dominated, and contains alternate, bank-attached bars and some mid-channel bars (Figure 15.3B). The inner straight reach also contains similar bar types. The region between the two straight reaches contains tight meanders that commonly exhibit symmetrical point bars. This meandering zone is the lowest-energy portion of the system and grain sizes in the channel become finer toward this area from both directions.

In summary, Dalrymple et al. (1992) noted that coarse sediments supplied by marine and river processes accumulate in the outer, marine-dominated and inner, river-dominated portions of the estuary, respectively, while finer sediment is present in the central zone. The nature of the facies within each of the zones depends on the relative influence of waves and tides; thus, estuaries can be divided into wave- and tide-dominated types (Figure 15.1A). In the ideal wave-dominated estuary, the tripartite facies distribution is clearly expressed (Figure 15.3): a marine sand body that consists of barrier-related deposits including flood-tidal delta sediments; a typically muddy central basin; and a bay-head delta formed by river discharge. An analogous three-fold subdivision is also present in tide-dominated estuaries but is not as clearly developed as tidal currents penetrate into the inner estuary more effectively than waves. The marine sand body consists of elongate sand bars and broad sand flats. Headward of this, the channel narrows and shows a straight to meandering to straight progression of sinuosities. The meandering reach contains the finest channel sediments and is the location of bedload convergence. It is dynamically equivalent to the central basin of wave-dominated systems.

15.3.3 Based on Dominant Coastal Processes

Boyd et al. (1992), based on dominant coastal processes and applicable to deltas, estuaries, lagoons, strand plains, and tidal flats (see Figure 15.1), improved the ternary classification of Dalrymple et al. (1992) and included both morphologic and evolutionary components (Figure 15.4). They defined estuary as the seaward part of a drowned valley that receives sediment from both fluvial and marine sources and exhibits facies that is influenced by tide, wave, and fluvial processes. They followed Dalrymple et al. (1992) who considered an estuary to extend from the landward limit of tidal facies (estuary head) to the seaward limit of coastal facies (estuary mouth). Boyd et al. (1992) defined a barrier as a transgressive elongated, parallel-to-shore sand body that consists of several sandy units, such as beaches, dunes, tidal deltas, washovers, and spits. The barrier separates lagoon and estuary embayments from the marine domain and are components of the estuary and lagoon systems. They may be connected to the mainland at either end and penetrated by tidal inlets, forming barrier islands. Boyd et al. (1992) defined strand plains as shore-parallel sand bodies with beaches and dunes, and which are found along prograded linear coasts and not associated with embayments (Figure 15.1). The strand plains preserve multiple shoreline positions and are underlain by more seaward facies such as the shoreface (Figure 15.1). Boyd et al. (1992) extended the classification of Dalrymple et al. (1992), and added a third dimension, relative time (expressed in terms of transgression and progradation), to form a triangular prism (Figure 15.5). In this model, progradation changes (such as estuary filling) are shown by the movement toward the back of the prism, whereas changes associated with transgression (i.e., progressive flooding of estuaries) are represented by the movement toward the front face (Figure 15.4). This model, thus, links depositional environments with shoreline behavior through time (Figure 15.4) (see also Dalrymple et al., 1992, Boyd et al., 1992; Harris et al., 2002; FitzGerald et al., 2015).

15.3.4 Based on Dominant Hydrologic Characteristics

Dalrymple et al. (1992) proposed two classes of estuaries based on the dominant hydrologic characteristics and the kinds of sediment and sediment bodies formed: tide-dominated, wave-dominated, and mixed wave- and tide-dominated.

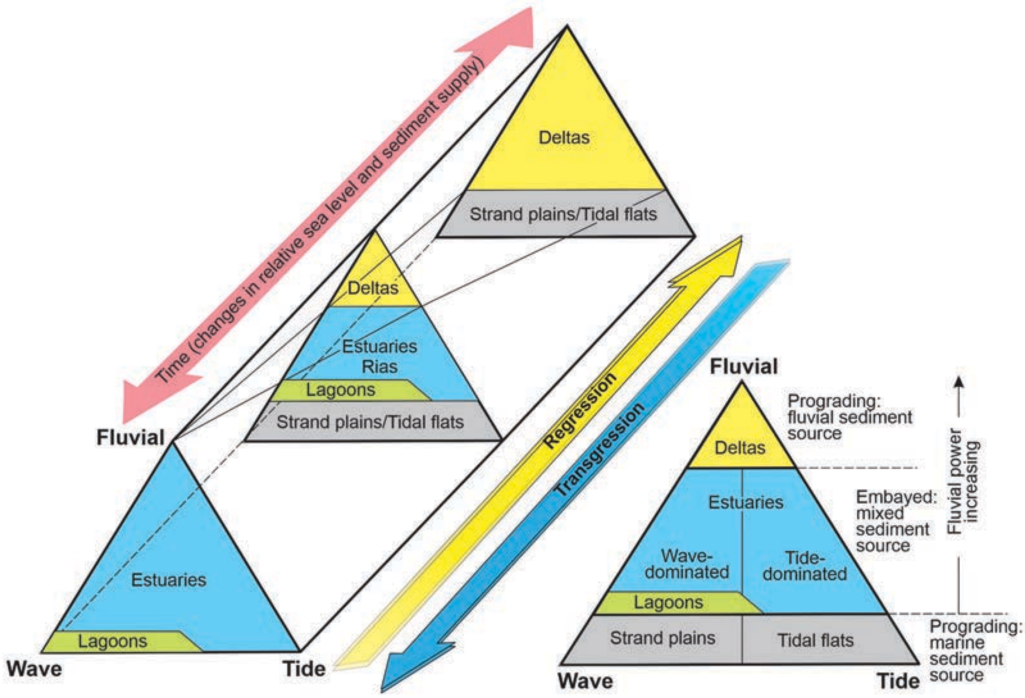

FIGURE 15.4 Triangular conceptual morphological model. (Modified after Dalrymple et al., 1992, Boyd et al., 1992; Harris et al., 2002; FitzGerald et al., 2015.) The left side of the diagram shows the types of depositional systems for transgressive and regressive regimes with respect to wave versus tidal dominance. It shows the evolution of estuaries, deltas and strand plains vis-à-vis sediment supply and sea-level changes. Through time, with an abundant sediment supply, estuaries are filled, ultimately leading to delta formation. The bottom right side of the triangular diagram shows the ternary process-based coastal classification. The uppermost part of the triangle is the delta, the middle trapezoid is the estuary, and the bottom area belongs to a spectrum of prograding, straight coastlines, ranging from tidal flats to strand (beach-ridge) plains. Lagoons occupy a linear field between estuaries and prograding, straight coasts. The coastal environments are based on the relative strength of wave versus tide power (from left to right), and increasing fluvial power, upward from the lagoon. The fields are also differentiated on the basis of the degree of coastal embayment, and sediment source.

15.3.4.1 Tide-Dominated Estuary

A tide-dominated estuary has a funnel-shaped entrance (mouth) where the hydrodynamics are primarily influenced by tidal currents rather than by wave action, thus, shaping estuarine morphology and sediment transport processes (Figures 15.1 and 15.5A). The associated subtidal channels are flanked by widespread and diverse intertidal and supratidal habitats (such as shoals, mangroves, saltmarsh, and salt flats) (Figure 15.5). Tidal inundation of these habitats results in the trapping and deposition of terrigenous and resuspended sediments. The tide-dominated estuaries occur mainly on meso- to macrotidal coasts, more so on macrotidal ones (see Figure 15.5B) where tidal-current energy exceeds wave energy at the mouth of the estuary, creating energy conditions in the estuary higher than those noted in typical wave-dominated estuaries. Water in the estuary is commonly well-mixed. Elongated sand bars (sand banks) develop parallel to the length of the estuary from the sand carried into the estuary from marine sources (Figure 15.5A). These tend to dissipate tidal energy. On the other hand, constriction of incoming tidal currents between the tidal bars causes an increase in their velocity for some distance up the estuary. Figure 15.3A–D shows the hydrologic conditions and sediment bodies developed in tide-dominated estuaries.

In addition to forming tidal bars/banks/ridges in the mouth of the estuary, sand may be transported landward through the estuary in a tidal-fluvial channel. Bedforms ranging in size from ripples to large dunes develop on sandy sediments in bars and tidal channels. Cross-bedding generated by the migration of these bedforms dip in either a landward or a seaward direction. Flaser bedding forms during slack water conditions owing to the deposition of suspended mud over sand ripples. Muddy sediments are also deposited in the lower energy parts of the estuary floor and in salt marshes/flats adjacent to the channel along the edges of the estuary (Figure 15.5A). Muddy sediments are characterized by nearly planar alternations of silt, clay, very fine sand, and carbonaceous (plant) debris. Bioturbation by burrowing and feeding organisms locally mix and homogenize these layers. Estuarine sediments typically contain a brackish-water fauna that includes oysters, mussels, other pelecypods, and gastropods. In tide-dominated estuaries, bioclastic debris is common and is deposited as a lag on the channel floor. Dune bedforms are created that migrate with the tidal currents to form cross-bedded sandstones. Other sedimentary structures include mud drapes, reactivation surfaces, and herringbone cross stratification. Examples of tide-dominated estuaries include Cook Inlet (Alaska, USA), Ord River (Australia), Gironde Estuary (France), and the Severn River (United Kingdom).

In summary, the major characteristics of tide-dominated estuaries are (a) large tidal range, (b) strong tidal currents, (c) wide mouth and shallow channels, (d) sediment deposition, and (e) saltwater intrusion. These are briefly enumerated here. The tidal range is large but variation in tidal range also depends on estuary location, shape, and its proximity to the open sea. The tidal currents are generally strong, affecting water circulation and sediment transport within the estuary; the direction and strength of the currents change with the ebb and flow of the tides. The tide-dominated estuaries have a wide mouth (entrance), thus allowing for the inflow and outflow of large volumes of water during tidal cycles. The channels are relatively shallower as compared to those in wave-dominated estuaries. The tidal currents transport and deposit fine-grained sediments in intertidal areas and deeper channels as mudflats and tidal flats. Strong tidal currents also cause saltwater to intrude into the estuary, thereby mixing with freshwater inputs from rivers resulting in the formation salinity gradient, which affects the distribution of both estuarine organisms and habitats.

15.3.4.2 Wave-Dominated Estuary

Wave-dominated estuaries (Figure 15.6A) are formed on high-energy microtidal coasts (Figure 15.6B) where oceanic waves bring unconsolidated sands towards the shore, but the riverine flow is enough to maintain an open, but rather restricted (by mobile sand shoals in the lower estuary part) connection with the sea. Mature forms tend to be linear in shape, but may have significant side embayments, depending on their evolutionary stage. Tidal currents are not significant drivers due to their microtidal range (Figure 15.6B) (Dalrymple et al. 1992). Thus, in a wave-dominated estuary, the hydrodynamics are primarily influenced by wave action rather than tidal currents (see Figures 15.1 and 15.6). In these estuaries, the energy from ocean waves plays a significant role in shaping estuarine morphology and sediment transport processes. Hence, the tidal influence is small, and the mouth of the estuary experiences high wave energy. Sediments tend to move alongshore and onshore into the mouth of the estuary, where a subaerial barrier/spit or submerged bar is formed (Figure 15.6A). This barrier prevents most of the wave energy from entering the estuary, thus, only internally generated waves are present behind the barrier. Depending upon tidal range and current velocity, a small number of inlets may be kept open in this barrier. Alternatively, the barrier may close the estuary entirely at times to produce a blind estuary or a coastal lake.

The relative influence of marine and river processes and the kinds of sediment bodies formed in different parts of a wave-dominated estuary are illustrated in Figures 15.3E–H. Muddy sediments, supplied mainly by the river, accumulate in the middle part of the estuary where total energy is lowest (Figure 15.3F). Deposition of mud (or muddy sediments such as biogenic debris such as molluscan shells, and wood fragments) is enhanced by the mixing of river and marine waters, which

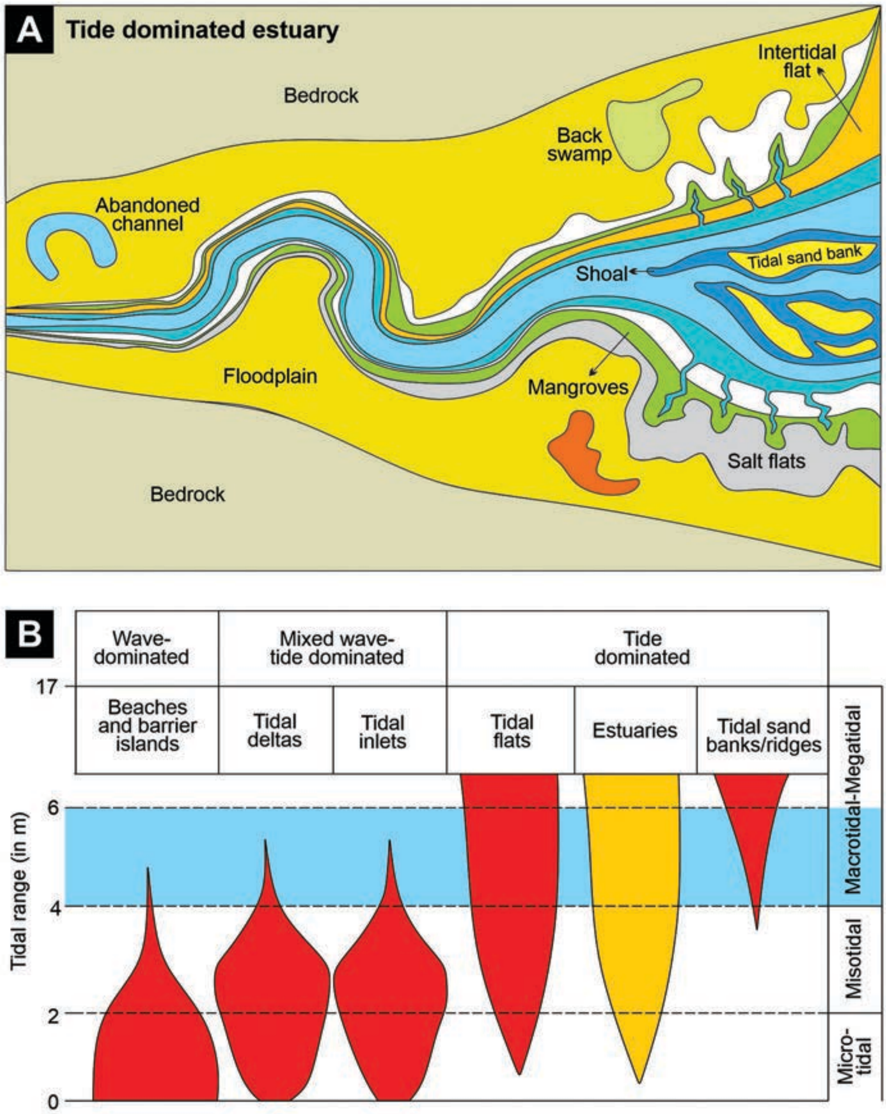

FIGURE 15.5 Tide-dominated estuary. A: Depositional setting. B: Types of tidal depositional systems and relative tidal ranges, highlighting the range of tide-dominated estuaries. Most wave-dominated estuaries are typical of macrotidal range. (Modified after Hayes, 1979.)

causes flocculation of clay particles owing to positively charged ions in seawater that neutralize the negative charges of the clay particles. Horizontal and subhorizontal stratification and bioturbation traces are commonly noted. In the head of the estuary, coarse, river-derived sediments are deposited in channels. Examples of modern wave-dominated estuaries include San Antonio Bay (United States), Miramichi River (Canada), and Hawksbury Eshlary (Australia).

In summary, the major characteristics of wave-dominated estuaries are: (a) strong wave energy, (b) limited tidal influence, (c) narrow mouth and deep channels, (d) sandbars and barrier islands, and (e) sediment transport. These are very briefly enumerated. Strong wave action from the open ocean

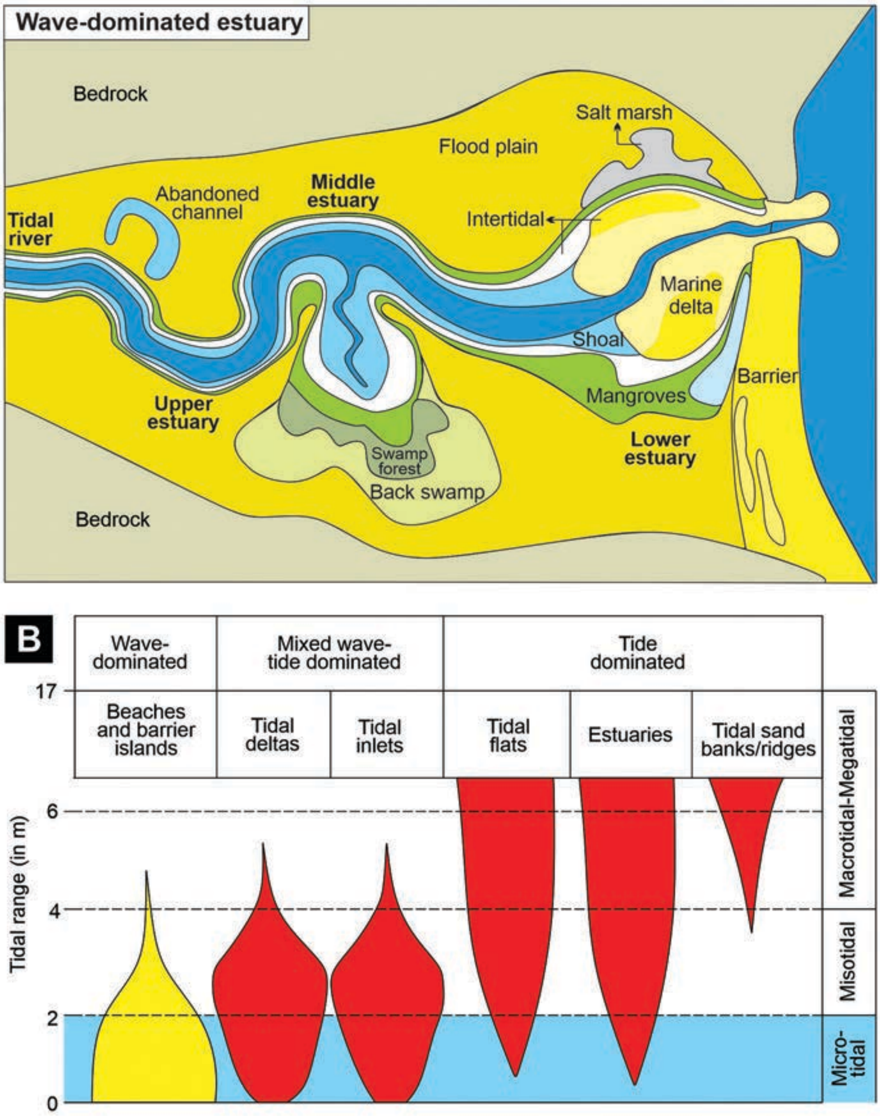

FIGURE 15.6 Wave-dominated estuary. A: Depositional setting. B: Types of tidal depositional systems and relative tidal ranges, highlighting the range of wave-dominated estuaries. Most wave-dominated estuaries are typical of microtidal range (Modified after Hayes, 1979.)

is characteristic, hence, the waves have a significant impact on the hydrodynamics and sediment transport within an estuary. The tidal currents are weaker, the tidal range is very small, and has very little influence on water circulation and sediment movement. A narrow entrance (mouth) restricts the inflow and outflow of water; generally, the channels are deeper, allowing waves to propagate further inland. The wave energy forms sandbars and barrier islands along the shoreline which protects the estuary from wave action and influences sediment dynamics. Finally, waves transport sediments from the open ocean into the estuary; this sediment is fine-grained and is deposited in deeper channels or along the shoreline, leading to the formation of mudflats or sandy beaches.

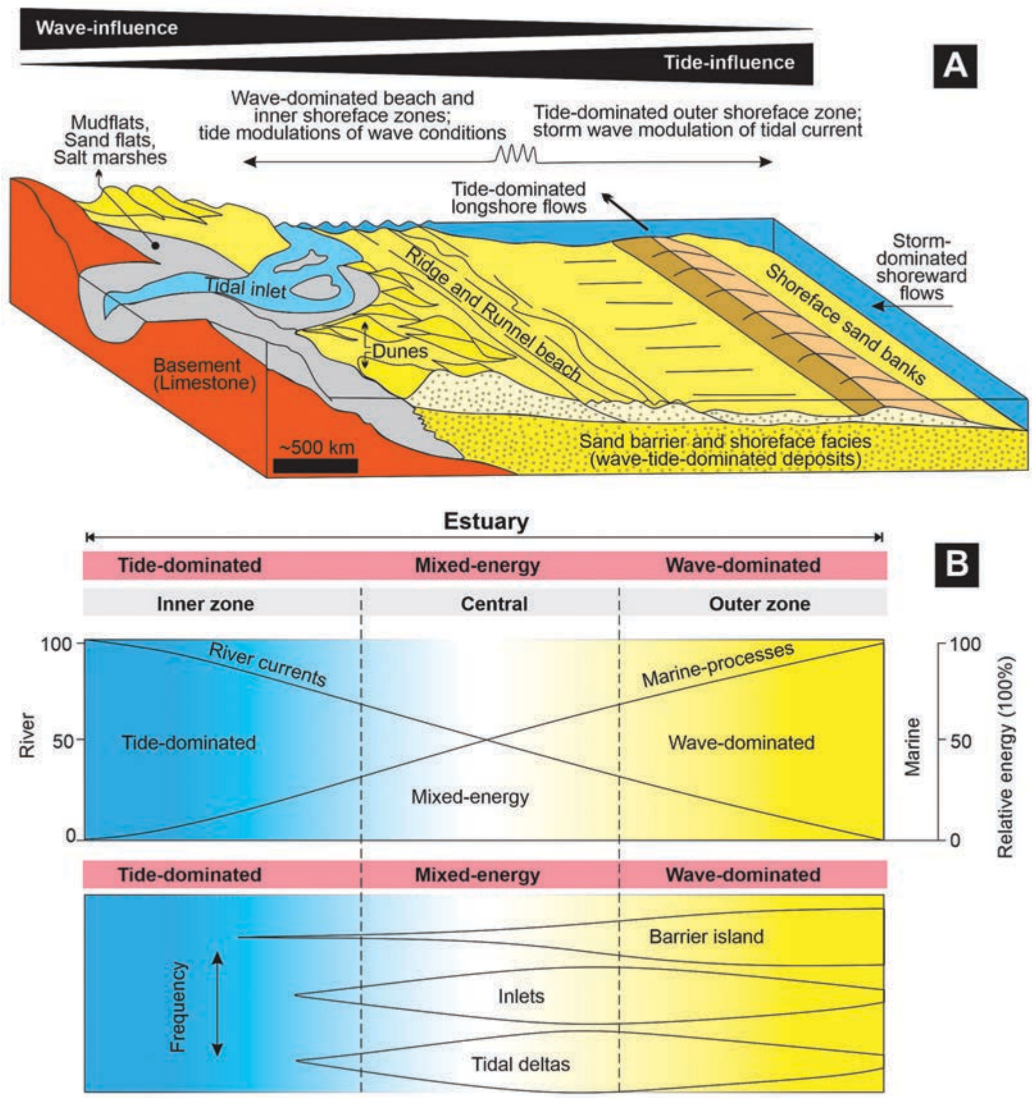

FIGURE 15.7 Mixed wave- and tide-dominated estuary. A: Coast with shoreface sand banks that are driven alongshore and onshore by storm waves and tidal and wind-induced currents. (Modified after Anthony, 2018.) B: Energy regimes of an estuary (Modified after Dalrymple et al., 1992.)

15.3.4.3 Mixed Wave- and Tide-Dominated Estuary

These estuaries have characteristics that are intermediate between wave- and tide-dominated types (Figure 15.7A). In these, tidal energy increases relative to wave energy, the barrier system of a wave-dominated estuary becomes progressively more dissected by tidal inlets and elongated sand bars develop in locations previously occupied by barrier segments and linear bars of ebb-tidal deltas (Dalrymple et al. 1992). Marine-derived sand is transported to greater distances up the estuary, and the generally muddy central basin is replaced by sandy tidal channels, flanked by marshes. Thus, in such estuaries both wave action and tidal currents play a significant role in shaping the hydrodynamics and sediment transport processes within an estuary (Figure 15.7B). Hence, the energy from both waves and tides influence estuarine morphology and sediment dynamics (Figure 15.7B)

(Dalrymple et al. 1992). Examples of mixed-energy estuaries include the St. Lawrence River, Canada; Willipa Bay, USA; and Oosterschelde Estuary, the Netherlands.

Major characteristics of a mixed wave- and tide-dominated estuary are: (a) variable hydrodynamics, (b) tidal range and wave energy, (c) morphological features, and (d) sediment transport and deposition. These are very briefly enumerated here. The estuarine hydrodynamics depends on the relative dominance of waves and tides, resulting in complex flow patterns, including tidal bores, wave-driven currents, and turbulence. These mixed estuaries experience a combination of tidal range and wave energy; the variation in tidal range depends on the location of the estuary, its shape, and its proximity to the open ocean, whereas wave energy is influenced by wind strength, fetch, and bathymetry. These estuaries also exhibit a combination of characters that are associated with both wave-dominated and tide-dominated estuaries such as wide entrances (mouths) with sandbars, barrier islands influenced by wave action, and deeper channels and tidal flats influenced by tidal currents. The combined energy from waves and tides in mixed estuaries results in complex patterns of sediment transport and deposition. Sediment is transported and deposited in different areas of the estuary, as influenced by the interaction between waves, tides, and the estuarine morphology.

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16 Lagoonal System

16.1 INTRODUCTION

A coastal lagoon is defined as a shallow elongated body of water, such as a sound, channel, bay, or saltwater lake, near or communicating with the open ocean or sea and partly or completely separated from it by a low, narrow elongated strip of land, such as a coral reef, barrier island, sandbank/sandbar, or spit (Bates and Jackson, 1980; Van Wagoner et al., 1990) (see Figure 16.1A). A lagoon is typically characterized by calm, brackish or saline water conditions and is influenced by a combination of marine, freshwater, and terrestrial processes. Lagoons commonly extend parallel to the coast, in contrast to estuaries, which are oriented approximately perpendicularly to the coast (Figure 16.1A). Many lagoons have no significant freshwater runoff; however, some do receive river discharge. The main movement of water within lagoons is in the form of tidal currents (that move in and out through narrow inlets between barriers) or wind-forced waves. Lagoons occur in close association with river deltas, barrier islands, and tidal flats (Figure 16.1A).

Lagoons are formed by the deposition of sediments along a coastline, resulting in the formation of a barrier island or sandbar that eventually separates the lagoon from the sea (Figure 16.1A). Lagoons can also form in coastal depressions or behind coral reefs. Besides their varied processes of formation, lagoons also vary in size and shape. They can be relatively small and narrow or large and elongated. The shape of a lagoon is often influenced by the geological and geomorphological characteristics of the coastline. The salinity of the water in a lagoon varies depending on factors such as tidal influences, freshwater input from rivers or groundwater, and evaporation rates. The lagoonal systems are also sites of significant sedimentation, with sediment being transported and deposited by both marine and freshwater processes. Fine-grained sediments such as silt and clay/mud are often found in the lagoon, while coarser sediments such as sand and gravel accumulate along the barrier island or spit (Figure 16.1B). The latter two are often associated with lagoonal systems, and act as a natural barriers separating the lagoon from the open ocean (Figure 16.1B). These landforms are typically composed of sand and are dynamic, constantly changing due to the action of waves, tides, and currents

Lagoons are largely areas of low water energy, except within tidal channels/inlets that extend into the lagoon (see Figure 16.2). At the ends of tidal inlets, both within the lagoon and on the ocean sides, tidal deltas develop and sandy sediments are deposited (see Figure 16.2). The deposited lagoonal sediments include sources from rivers, ocean, shores, and barriers or internally such as by organic production, chemical precipitation, and erosion of older deposits (see Nichols and Boon, 1994). But, in general, lagoons are dominated by silt and mud deposits; storms may cause washover of coarser sediments from the barrier (see Figure 16.2).

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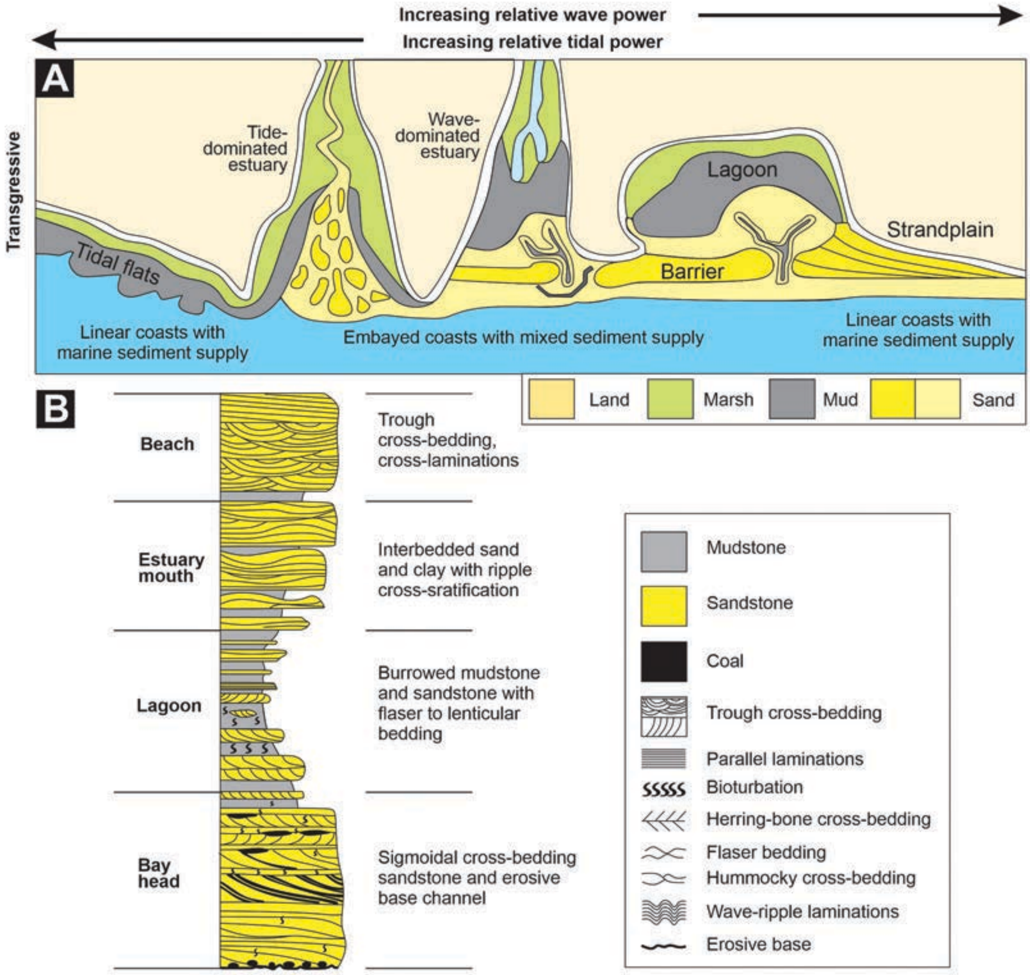

FIGURE 16.1 Depositional setting of a lagoon, stratigraphic sequence, and sedimentary structures. A: Depositional setting of a lagoon. (Modified after Dalrymple et al., 1990.). B: Lagoonal stratigraphic sequence of a wave-dominated estuary with progradational succession, and associated sedimentary structures. (Modified after van Wagoner et al., 1990.)

Lagoon deposits differ from estuarine ones; for those in which freshwater discharge from rivers into a lagoon is non-existent, the sediments are largely from marine sources. Additionally, due to their low-energy environment, lagoonal deposits are mainly of fine-grained sediments. Sandy sediments are largely confined to: (a) tidal deltas at the mouth of tidal inlets; (b) tidal channels that extend into the lagoon; (c) washover lobes behind barriers; and (d) lagoonal beaches (i.e., parts of the lagoonal shoreline) (see Figure 16.2). Tidal channels show current ripples and small-scale cross-bedding, whereas the low-energy lagoonal bottom is characterized by bioturbated silty or muddy sediments, commonly with thin intercalations of sand. This sand is horizontally laminated, and may display ripple cross-laminations. In very arid settings, such as the lagoons in the Persian Gulf, lagoons are characterized by the deposition of evaporites (mainly gypsum with minor halites and dolomites). In less hypersaline conditions, such as behind barrier reefs (as in Australia), carbonates

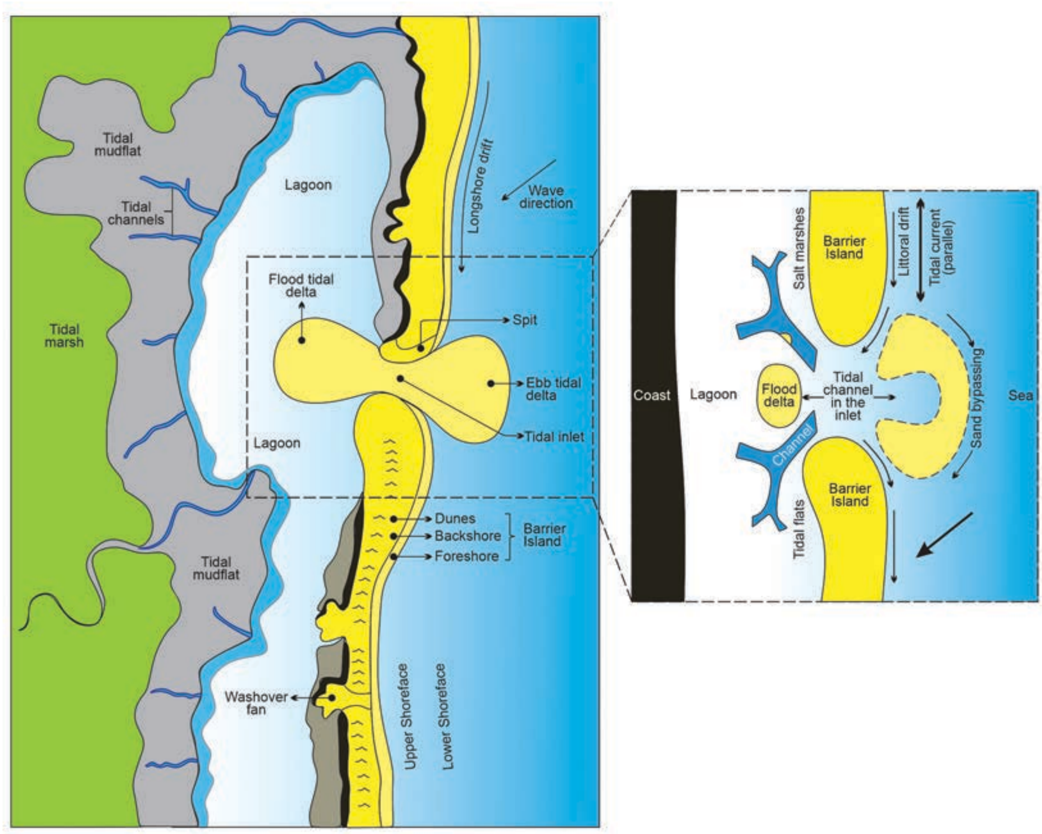

FIGURE 16.2 Depositional setting of a lagoon, with emphasis on tidal channels/inlets that extend into the lagoon and the associated sediments.

dominate and are associated with carbonate mud and ooids in more high-energy settings; algal mats commonly develop in shallow settings (supratidal and intertidal zones) forming stromatolites.

16.2 TYPES OF LAGOONS

Kjerfve and Magill (1989), based on geomorphology and the nature of water exchange with the coastal ocean, identified three types of lagoons: choked, restricted, and leaky (Figure 16.3).

16.2.1 Choked Lagoons

Choked lagoons, also known as choked estuaries or choked tidal inlets, occur when the entrance of a lagoon or estuary becomes partially or completely blocked such that the normal exchange of water between the open sea and the lagoon is hindered (Figure 16.3A). This blockage can be caused by various processes, such as sediment deposition, sandbars, or coastal landforms. Coastal erosion or sediment transport leads to the accumulation of sediments at the entrance of a lagoon, gradually blocking it. The blockage causes the lagoon to become stagnant, with reduced water circulation and increased sedimentation. Reduced water circulation leads to the accumulation of pollutants, nutrients, and organic matter, affecting water quality and potentially leading to eutrophication.

Choked lagoons occur along coasts with high wave energy and significant alongshore drift (such as the Coorong Lake in southern Australia).

16.2.2 Restricted Lagoons

Restricted lagoons, also known as barrier lagoons or coastal lagoons, occur when a lagoon is partially or completely separated from the open sea by a barrier, such as a sandbar, barrier island, or spit, thus, restricting the exchange of water between the lagoon and the open sea (Figure 16.3B). They commonly exhibit (a) two or more entrance channels or inlets, (b) have a well-defined tidal circulation, (c) are strongly influenced by winds, and (d) are generally vertically mixed (such as Lake Pontchartrain, Louisiana, USA). Restricted lagoons are typically associated with longshore sediment transport, wave action, and sea-level fluctuations. Over time, sediment deposition and growth of barriers create a blockade separating the lagoon from the open sea. This barrier can be temporary or permanent, depending on the dynamics of sediment transport and coastal geomorphology. The limited exchange of water with the open sea leads to reduced wave action and tidal influence. As a result, restricted lagoons often have calmer waters and lower salinity levels as compared to the adjacent marine environment. Thus, restricted lagoons are characterized by a relatively low-energy environment as compared to the open sea.

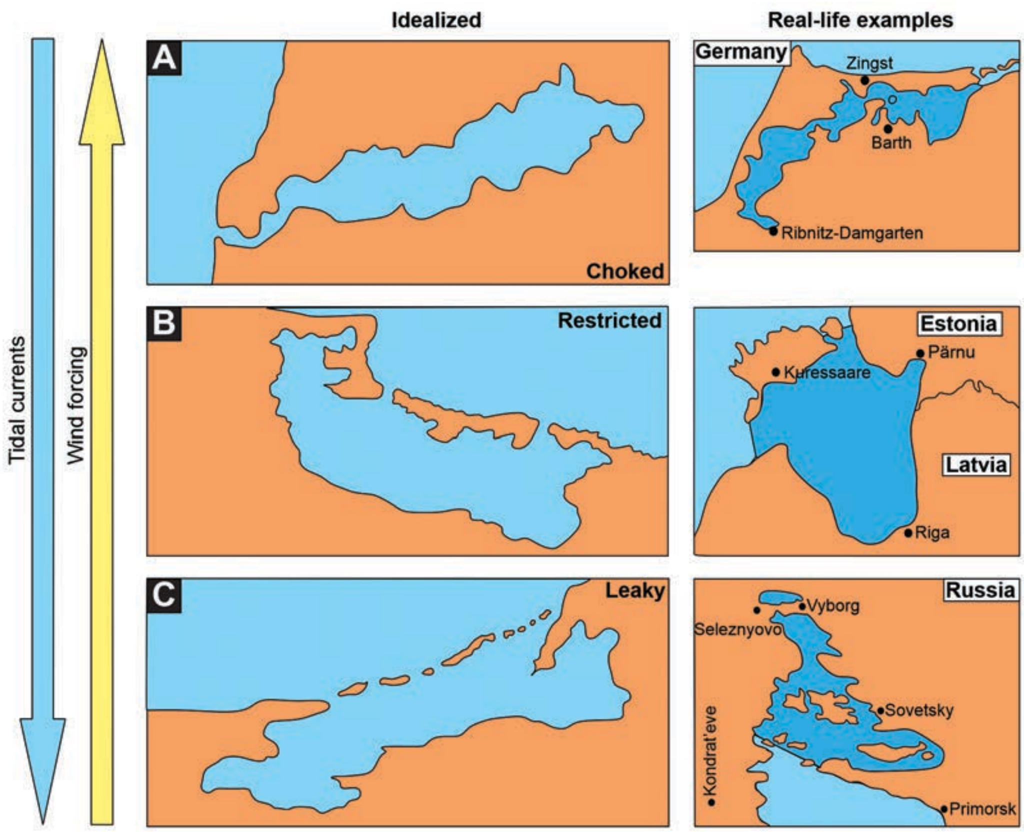

FIGURE 16.3 Types of coastal lagoons.

16.2.3 Leaky Lagoons

Leaky lagoons, also known as leaky-barrier lagoons, occur when a lagoon or coastal water body has permeable sediments or underlying geology that allows water to seep or leak into or out of the lagoon. Leakage in lagoons occurs through various pathways (Figure 16.3C). One common pathway is through the sediments themselves, where water can infiltrate or exfiltrate through permeable layers or channels. Another pathway is through the underlying geology, such as karstic formations or fractures that allow water to flow in or out of the lagoon. Leakage can also occur through connections with groundwater systems or nearby rivers and streams. Leaky lagoons typically occur along coasts where tidal currents play a greater role in sediment transport than the wind waves (such as the Belize Lagoon, Belize). Leaky lagoons may stretch for more than 100 km along coasts but are commonly no more than a few kilometers wide and are characterized by (a) wide tidal passes, (b) efficient water exchange with the ocean, (d) strong tidal currents, and (e) sharp salinity and turbidity fronts. Water levels in the lagoon fluctuate with tides, precipitation, or groundwater levels. The exchange of water with the surrounding environment is significant, leading to changes in salinity, nutrient concentrations, and water temperature within the lagoon. The leakage in lagoons can have both positive and negative impacts on the ecology of the system. On one hand, the exchange of water introduces nutrients, sediments, and organic matter into the lagoon, which enhances productivity and supports diverse ecosystems. On the other hand, excessive leakage leads to the loss of water from the lagoon, resulting in reduced water levels, increased salinity, and potentially adverse effects on the organisms.

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17 Tidal-Flat System

17.1 INTRODUCTION

Tidal flats occur in coastal areas of flat, exposed mud or sand that are alternately covered and uncovered by tides, and are characterized by the absence of strong wave activity (Figures 17.1A–B) (see Semeniuk, 2018, and references therein). The tidal flats are marshy and muddy to sandy areas partially uncovered by the rise and fall of tides, i.e., they experience regular tidal inundation and exposure (Figure 17.1B). They are almost featureless plains dissected by a network of tidal channels and creeks that are exposed during low tide (Figure 17.1A) (see also Mauz and Bungenstock, 2007).

The tidal flats develop either along open coasts of low relief and with relatively low wave energy (Figure 17.2A) or behind barriers on high-energy coasts protected from waves by barrier islands, spits, reefs, and other structures (see Figure 17.2B). In general, tidal flats form in sheltered areas, such as estuaries, bays, and lagoons, where the energy of waves and currents is relatively low (Figure 17.2B). They can also develop in the intertidal zone of open coastlines (Figure 17.2A), particularly in areas with large tidal ranges (i.e., meso- to macrotidal; see Figure 17.1D). The vertical distance between the high- and low-tide line in most modern tidal environments ranges from 1 to 4 m (mesotidal coasts), although tidal ranges of 10–15 m or more (macrotidal coasts) also occur (Figure 17.1D), such as at the Bay of Fundy (Canada). The tidal flats may range from a few kilometers to as much as 25 km. Thus, the tidal flats environment is confined to the shallow margins of the ocean, where its flat topography allows for the deposition and accumulation of fine-grained sediments such as mud or sand that are often rich in organic matter (Figure 17.1C).

As tides rise, the flood waters move into channels until at high tide they are overtopped and water spreads over and inundates the adjacent shallow flats. The ebb tide again exposes the channels and intervening flats. Thus, the sedimentary processes on tidal flats are influenced by the ebb and flow of tides. During high tide, water inundates the flats, transporting and depositing sediment. As the tide recedes, the flats are exposed to air, and sedimentation continues as the water drains away. This alternating process of sediment deposition and erosion creates tidal channels, sandbars, and mudflats. In temperate regions, salt marshes commonly cover the upper parts of tidal flats, and muds and silts accumulate near high-water level (Figure 17.1C). At the same time, mixed mud and sand are deposited in the middle tidal-flat region, and sands accumulate in channels and on the lower parts of the tidal flat (Figure 17.1C). In arid to semi-arid regions, tidal flats become desiccated, forming mud cracks and gypsum and halite crystals in the mud. The surface of tidal flats in subarctic regions is marked by surficial scars, caused by ice-pushed boulders, and ice-rafted pebbles and cobbles. Modern tidal flats are primarily sites of siliciclastic deposition; however, carbonate sediments and,

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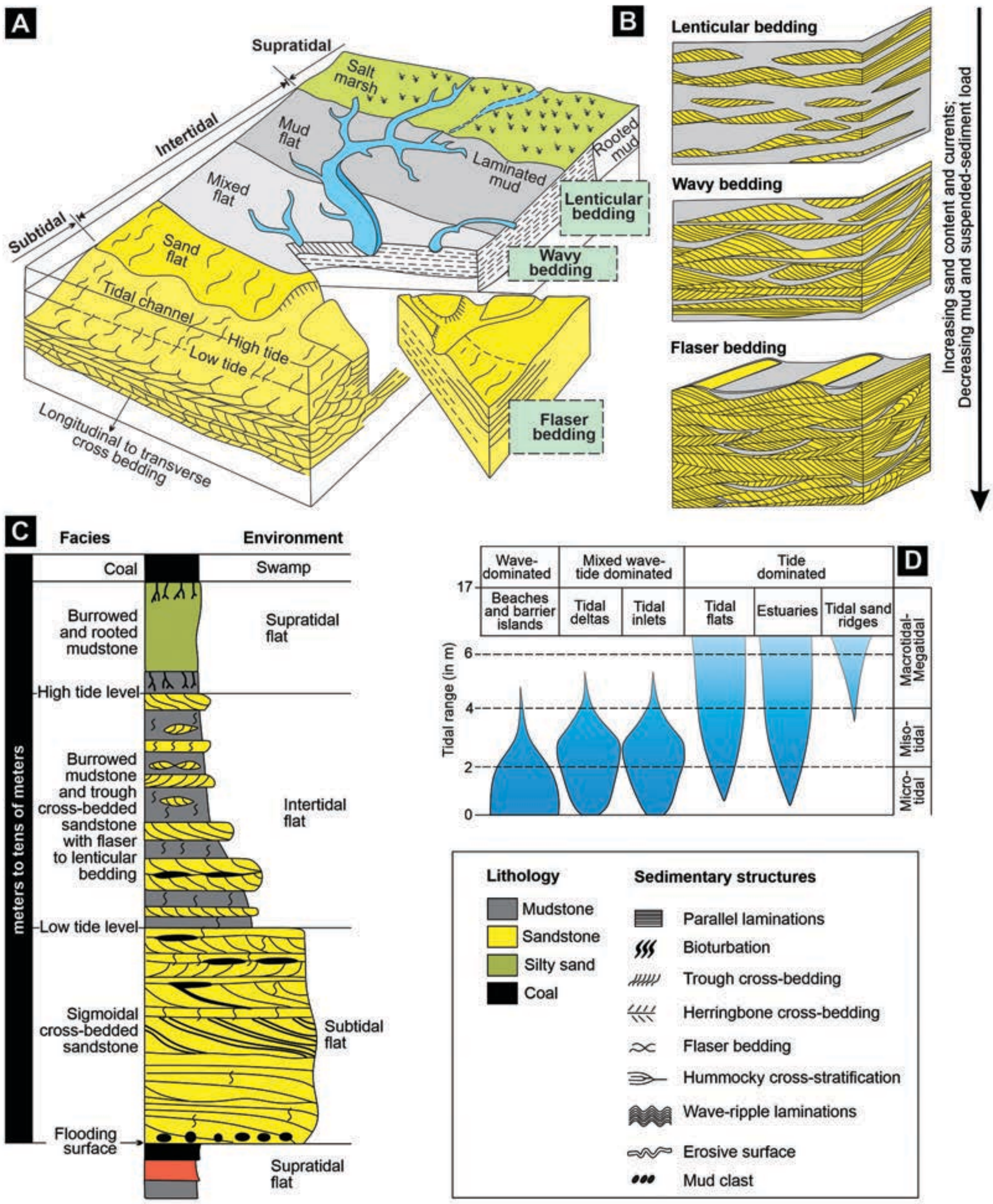

FIGURE 17.1 Tidal-flat system. A: Depositional setting of a tidal flat with its major subdivisions. (Modified after Walker and Jemes, 1992.) B: Various sedimentary structures (bedding) noted in a tidal-flat environment. C: Idealized vertical facies sequence of a tidal flat environment. Common sedimentary structures noted in subtidal to intertidal zones are mud-drapes, reactivation surface and herring-bone cross-bedding. (Modified after Van Wagoner et al., 1990.) D: Tidal ranges (the vertical distance between the high and low tides; see Figure 17.1A). (Modified after Hayes, 1975.)

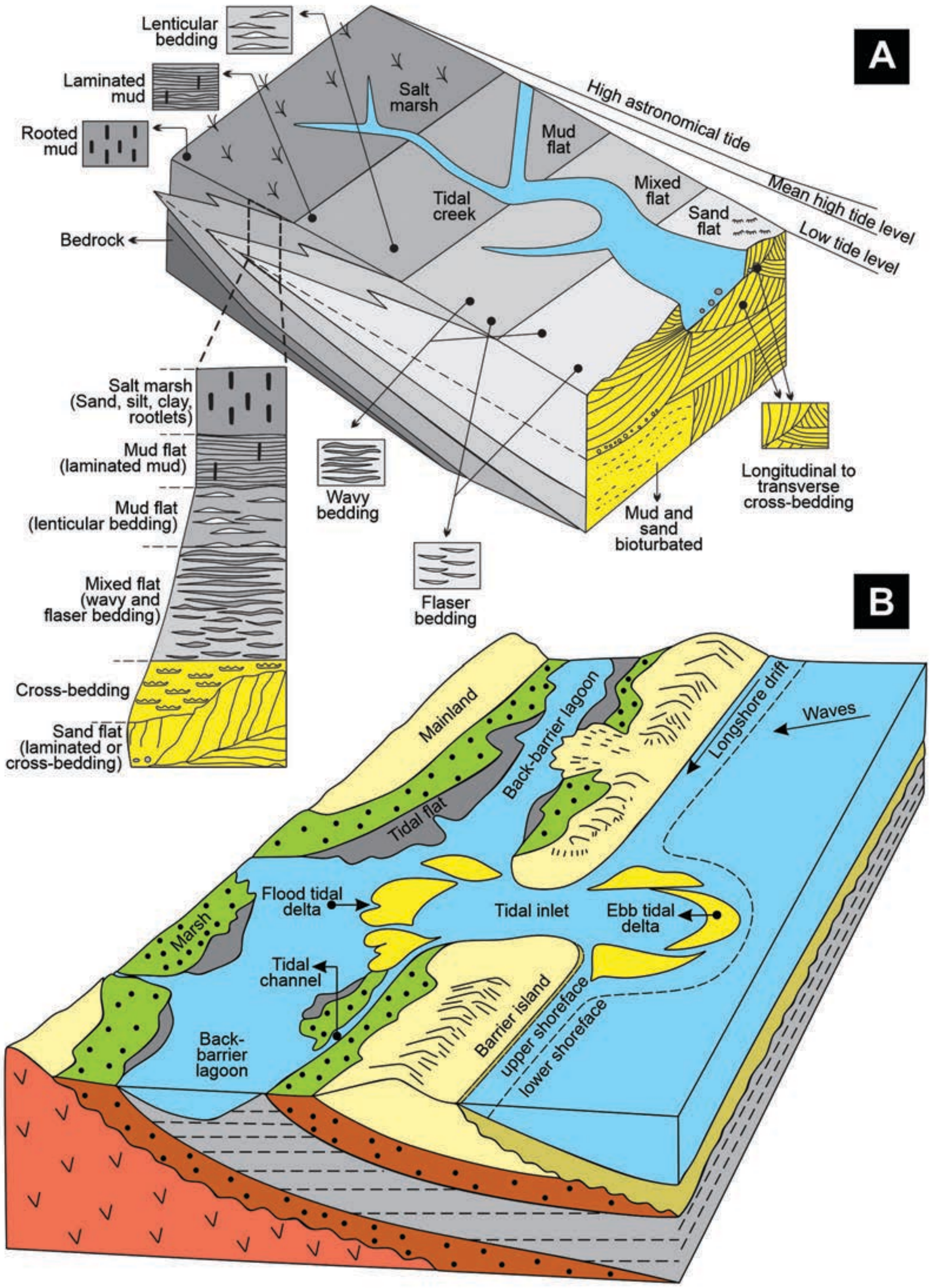

FIGURE 17.2 Depositional setting of tidal-flat systems. A: Tidal-flat area showing intertidal (sand flat, mixed and mudflat) and supratidal setting (salt marsh) in a prograding coastline. (Modified after Mauz and Bungenstock, 2007.) B: Development of a tidal flat protected from waves by barrier islands, and other structures.

in a few areas, evaporites also accumulate such as those in the Bahamas, the Persian Gulf, Florida Bay, and the western coast of Australia (see Hardie and Shinn, 1986). The tidal deposits have economic significance as oil and gas deposits have been discovered in both siliciclastic and carbonate tidal facies, and uranium is present in some sandy tidal facies.

17.2 DEPOSITIONAL SETTING

The tidal-flat environment is divided into three zones: subtidal, intertidal, and supratidal (Figure 17.1C). The subtidal zone which is inundated with water most of the time, encompasses the part that normally lies below 10 m and is subjected to the highest tidal-current velocities (Figure 17.1C). In this part, bed-load transport and deposition dominate, although wave processes also play a part. The intertidal zone lies between mean high- and low-tide levels (Figure 17.1C) and is subaerially exposed either once or twice each day. Both bed-load and suspension sedimentation take place in this zone. The supratidal zone lies above normal high-tide level (Figures 17.1A and 17.1C) and is exposed to subaerial conditions most of the time but may be flooded by spring tides twice each month or by storm tides at irregular intervals; it is incised by tidal channels (Figure 17.1A). The sedimentation is dominantly from suspension. On some tidal flats, the supratidal zone is a salt marsh incised by tidal channels and in mid or semi-arid climates, evaporite deposition (referred to as a sabkha) is commonly noted.

Thus, the depositional setting of a tidal flat is influenced by several factors, including tidal range, sediment supply, hydrodynamics, and topography. The hydrodynamics plays a crucial role in shaping sedimentary features, as the energy of the tides determines the transport and sorting of sediment, as also the formation of erosional and depositional features. In areas with higher energy, such as tidal inlets or channels, coarser sediments are transported and deposited. In contrast, in more sheltered areas, fine-grained sediments accumulate. Topography also influences sedimentation. Tidal flats with a gentle slope, allow sediments to settle and accumulate. However, the presence of channels or other topographic irregularities also results in localized erosion or deposition.

17.3 SEDIMENTARY PROCESSES AND SEDIMENT CHARACTERISTICS OF TIDAL FLATS

Tidal processes and waves influence sedimentation on siliciclastic tidal flats producing sediments with characteristic grain-size (sand, silt, and mud). Although sedimentation is dominated by tidal currents, wind-driven waves and the currents generated by these waves also play a role in deposition on the flats (Ridderinkhof, 1998). Tidal currents move up the gentle slope of the tidal flat during flood tide and back down during ebb tide. Tidal currents can reach velocities of 1.5 m/s or more, and velocities on the flats range from 30 to 50 cm/s (Reineck and Singh, 1980) causing transport of sandy sediments and producing ripples, cross-beddings, and plane beddings. Sand deposition dominates in both shallow subtidal and lower intertidal zones (see Figure 17.1C). Channel sands are characterized by ripples and internal cross-bedding that sometimes display bimodal directions of foreset dip, generated by reversing tides. The sands thus display herringbone cross-stratification, i.e., the cross-laminated sediments deposited during flood tide dip in the opposite direction to those formed during ebb tide (see Figure 17.1C). During rising tide. as water fills tidal channels, it spills out from the channels and spreads across intervening flats between channels. Both fine sand and mud are deposited on these low-energy, flat areas. Sandy and muddy sediments deposited on the flats between channels are characterized by small-scale ripple cross-lamination, flaser bedding, wavy bedding, lenticular bedding, and, more rarely, finely laminated bedding (Figures 17.1B–C and 17.2A). The supratidal zone, the zone of lowest energy on the tidal flat, is marginally affected by tidal currents and waves. The deposits are mainly mud but may include abundant plant debris in supratidal

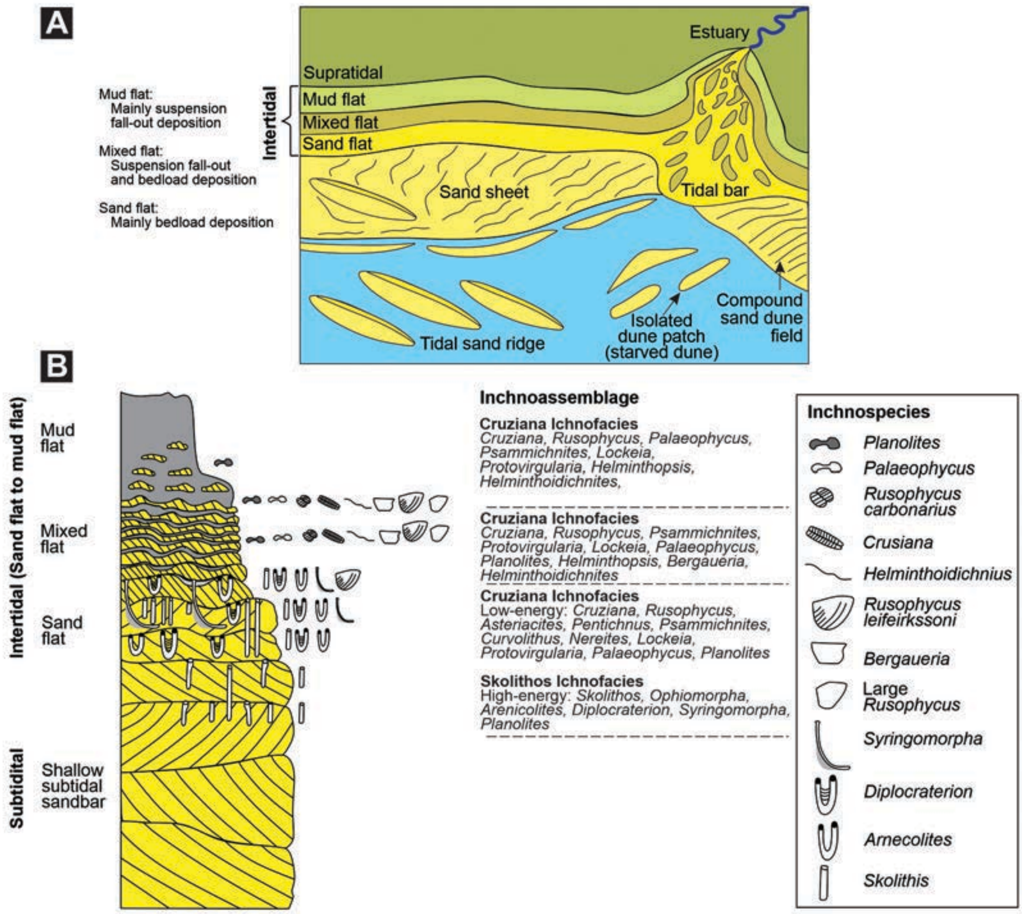

FIGURE 17.3 Tidal zones and ichnofacies. A: Tidal zones and sand bodies. B: Ichnofacies noted in a tidal flat system. (Modified after Desjardins et al., 2012.)

marshes that eventually forms peat (Figure 17.1C). Desiccated, cracked muds (mud cracks) and rooted mud, among others, are characteristic features of the supratidal zone (Figure 17.1C).

Although most tidal flats are sites of siliciclastic deposition, some tidal flats are dominated by carbonate sediments. Lime muds and sand-sized skeletal fragments generated within the subtidal zone may be transported into the intertidal and supratidal zones by waves and currents. In arid and semi-arid climates, gypsum, anhydrite, and dolomite are chemically precipitated in the supratidal and upper intertidal zones due to strong evaporation.

In supratidal and intertidal zones (Figure 17.3A), blue-green algae trap and bind fine sediments to produce stromatolites. Organisms such as bivalves, crustaceans, polychaete worms, foraminifers, diatoms, and blue-green algae (cyanobacteria) inhabit tidal flats and produce fecal pellets; they cause extensive bioturbation of sediment, and generate burrows belonging to *Cruziana* and *Skolithos* ichnofacies (Figure 17.3B) (see also Desjardins et al., 2012).

Transgression and regression form deposits of laterally adjacent tidal-flat environments to become superimposed, generating a characteristic successions of vertical facies (Figure 17.4) (see also Chough et al., 2004). Progradation produces a generalized fining-upward succession that begins with subtidal and lower intertidal cross-bedded sands, followed upward by mixed sand and mud in the middle intertidal zone, and mud and peat in the upper intertidal and supratidal zones (Figure 17.4A).

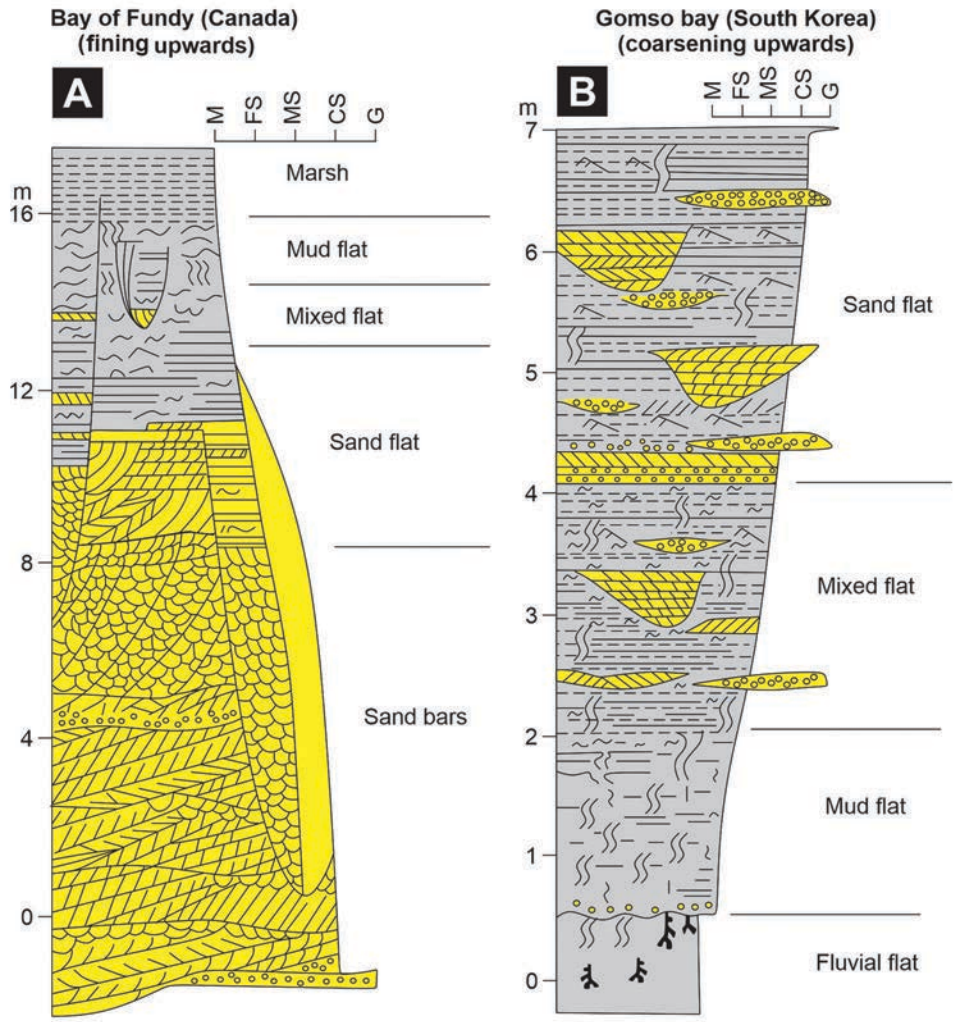

FIGURE 17.4 Tidal flat facies models. A: Bay of Fundy, showing fining-upward succession by coastal progradation. B: Gomso Bay (South Korea), showing coarsening-upward succession by coastal retrogradation with sea-level rise. (Modified after Klein, 1977; Dalrymple et al., 1990; and Chough et al., 2004.)

Transgression generates a coarsening upward succession that displays the same general facies but in reverse order (Figure 17.4B). Similar patterns of subtidal, intertidal, and supratidal carbonate facies develop on coasts characterized by carbonate tidal flats (see Dalrymple et al., 1990; Hardie and Shinn, 1986).

In summary, some key sedimentary processes that control the depositional features and sediment characteristics of a tidal flat are tidal currents, erosion and deposition, sediment sorting, sediment composition, sedimentary structures, organic matter accumulation, and bioturbation and biogenic structures. These are very briefly enumerated below.

Tidal currents are the primary driver of sediment transport and deposition on tidal flats (Figure 17.1). During flood tide, water flows onto the flats, carrying sediment with it. As the tide recedes during the ebb tide, the currents reverse, causing sediment to settle and deposit. The alternating tidal currents result in erosion and deposition of sediment on tidal flats. During high tide, sediment is transported and deposited in areas of reduced current velocity, such as channels. During low tide, the exposed flats undergo erosion as water drains away, carrying sediment with it. Tidal flats often exhibit distinct sediment sorting patterns. Coarser sediments accumulate in areas with higher energy, such as tidal channels or inlet mouths, where the currents are stronger. On the other hand, finer sediments, such as mud, settle in more sheltered areas, such as mudflats or flood basins (see Figure 17.2). Tidal flats are typically characterized by the deposition of fine-grained sediments, such as mud or silt (Sediment composition) (Figures 17.1 and 17.2). In ancient tidal-flat deposits, mudstones or siltstones are commonly noted, indicating a low-energy environment where fine-grained sediments settle and accumulate (Figures 17.1 and 17.2A). In general, the sediment composition varies depending on the source of sediment and local geological factors. Common sediments include clay/mud, silt, sand, and organic matter (Figures 17.1 and 17.2A). Tidal flats exhibit characteristic sedimentary structures that result from alternating processes of deposition and erosion. These structures include tidal channels, sandbars, mudflats, ripple marks (both symmetric and asymmetric) and cross-bedding, among others (Figures 17.1 and 17.4). Tidal channels are formed by the scouring action of tidal currents, while sandbars and mudflats are deposited in areas of reduced current velocity (Figures 17.1 and 17.2A). Laminations, or thin layers of sediment, can also be present, reflecting the cyclical nature of tidal-flat deposition (Figure 17.2A). Tidal flats are often rich in organic matter due to the accumulation of plant and animal debris (organic matter accumulation) (Figure 17.1C). The fine-grained sediments, low energy, and the sheltered nature of tidal flats create conditions suitable for organic matter preservation. The presence of organic matter can be observed through dark-colored layers or by the occurrence of fossilized remains (Figure 17.1C). This organic matter contributes to the fertility of the flats and thus supports diverse ecological communities. Over time, the sediments on tidal flats undergo diagenetic processes, such as compaction, cementation, and lithification. Burial of sediments under subsequent layers leads to lithification and the formation of sedimentary rocks, such as mudstones or sandstones (Figure 17.1C). The diagenetic transformation of tidal-flat sediments can alter their physical and chemical properties, affecting their porosity, permeability, and overall rock fabric. Tidal flats support diverse ecological communities, and their activities leave behind distinctive structures in the sediment (bioturbation and biogenic structures) (Figure 17.3). Examples include burrows, tracks, or trails created by organisms such as worms, crabs, or bivalves (Figure 17.3) (see also Desjardins et al., 2012). These biogenic structures are preserved in ancient tidal-flat sediments, providing evidence of past biological activity.

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Section IVc

Depositional Systems: Siliciclastic Marine and Pelagic Environments

18 Shelf Environment

18.1 INTRODUCTION

The shelf environment is a shallow, relatively flat area that extends from the coastline to the continental rise (Figures 18.1A–B). It encompasses the shallow-water neritic zone, the area of the ocean lying shoreward of the shelf break (Figure 18.1B). Hence, it is typically located between the shoreline and the shelf break, which is the point where the seafloor drops off steeply towards the deeper ocean (Figure 18.1A). The shelf break on modern shelves lies at an average depth of about 130 m, but it ranges between 18 m to 915 m (Bouma et al., 1982). The width of the shelf also varies greatly, ranging from a few kilometers to hundreds of kilometers, depending on factors such as tectonic activity, sediment supply, and changes in sea level.

The shelf environment is characterized by relatively shallow water depths, typically less than 200 meters. It is influenced by both terrestrial and marine processes. Both siliciclastic and carbonate sediments accumulate in marine shelf environments, although most modern continental shelves are covered by siliciclastic sediments. Carbonate sediments are restricted to a few shelves, largely in tropical areas.

18.2 TYPES OF SHELVES AND DEPOSITIONAL SETTING

In modern oceans, the shallow-marine environment occupies mainly the continental shelf area around the margin of the continents, called the pericontinental or marginal sea (Figure 18.1C).

18.2.1 Pericontinental or Marginal Sea

A pericontinental or marginal sea is a type of sea that is partially enclosed by a coastline and located adjacent to a continent (Figure 18.1C). It is typically shallower than the open ocean and is influenced by both terrestrial and oceanic processes. Terrestrial inputs, such as from rivers and streams, deliver sediments, nutrients, and organic matter to the sea, supporting diverse ecosystems and contributing to the formation of sedimentary deposits. Oceanic processes, such as tides, currents, and upwelling, play a major role in shaping the pericontinental sea environment. Pericontinental seas may result from the flooding of low-lying coastal areas due to rising sea levels or by tectonic activity, or they may be remnants of ancient oceans trapped between continents during the formation of supercontinents. Examples of pericontinental seas include the Mediterranean Sea, the Gulf of Mexico, and the South China Sea. These seas are characterized by relatively shallow water depths, typically less than 200

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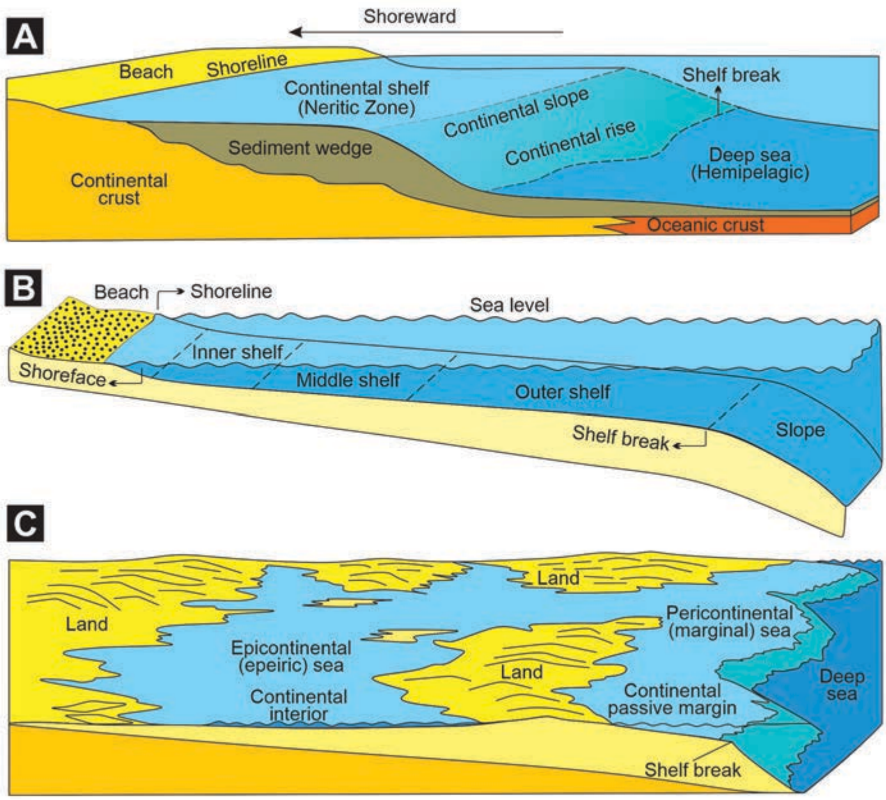

FIGURE 18.1 Divisions of the shelf environment. A: The shelf environment extends from the coastline to the continental rise. B: Subdivisions of the shelf environment. C: Distinction between the pericontinental and the epicontinental seas.

meters, and are often connected to the open ocean through narrow channels or straits. The depth and shape of a pericontinental sea vary, depending on factors such as tectonic activity, sedimentation, and sea-level changes.

18.2.2 Epicontinental Sea

An epicontinental sea, also known as an inland or epeiric sea, is a type of sea that existed in the past and was located on or near a continent (see Figure 18.1C). Unlike pericontinental seas, which are partially enclosed by a coastline, the epicontinental seas were fully surrounded by landmasses (i.e., within the continental interior) (Figure 18.1C). They are typically formed during periods of high sea levels, or tectonic activity that caused the flooding of low-lying coastal areas. They are relatively shallow, with depths ranging from a few meters to a few hundred meters. The epicontinental sea is influenced by both terrestrial and marine processes. Terrestrial inputs, such as from rivers and streams, deliver sediments, nutrients, and organic matter and contribute to the formation of sedimentary deposits. Marine processes, such as tides, currents, and upwelling, also play a role in shaping the epicontinental sea environment. Many shallow-marine deposits preserved in the geologic record were deposited in epicontinental seas, for which there are no similar modern analogs (Figure 18.1C). Examples of ancient epicontinental seas include the Western Interior Seaway in North America during the Late Cretaceous period and the Tethys Sea during the Mesozoic era.

18.3 PHYSIOGRAPHY AND DEPOSITIONAL SETTING OF SHELF ENVIRONMENT

The shelf environment (largely siliciclastic as in the present day) is divided into three broad divisions: shallow inner shelf (dominated by tidal, wind-driven, and storm-wave processes), middle shelf, and the deeper-water outer shelf (Figure 18.1B). Their boundaries are not well defined as their positions fluctuate with changing sea levels. The width of shelves also varies based on their plate-tectonic settings (Shepard, 1973; Eisma, 1988). Shelves along the fore-arc of convergent continental margins are very narrow, whereas the shelves are very broad in the back-arc basins of convergent margins, on divergent or trailing-edge continental margins, and on cratonic downwarps that open to the sea.

The physiography of the shelf environment refers to the physical features and characteristics of the seafloor. It varies greatly depending on factors such as tectonic activity, sediment supply, and sea-level changes. The shelf environment typically consists of a gently sloping seafloor, extending from the coastline to the shelf break, which is the point where the seafloor drops off into deeper water (see Figures 18.1A–B). The slope of the shelf varies from gradual to steep, depending on factors such as the proximity of sediment sources and the intensity of wave and current activity. The shelf environment is often characterized by high sedimentation rates, as it receives a significant amount of sediment from terrestrial sources. These sediments are transported and redistributed by waves, tides, and currents, leading to the formation of various depositional features. Common depositional features on the shelf include sandbars, sand waves, and tidal deltas. These features vary in size and shape depending on the intensity of wave and current activity, as well as the availability and composition of sediments. The shelf environment is an important site for the accumulation of organic-rich sediments, such as mud and silt. These sediments are derived from terrestrial and marine sources and support diverse ecosystems, including benthic communities and habitats for various marine organisms.

The shelf itself is divided into different zones such as the inner shelf (closest to the coastline and generally shallower), middle shelf, and the outer shelf, which is farther offshore and typically deeper (Figure 18.1B). The shelf break marks the transition from the shelf to the deeper oceanic basin. These three zones are briefly enumerated below.

18.3.1 Inner Shelf

The inner shelf (see Figure 18.1B) is characterized by relatively shallow water depths, typically ranging from a few meters to tens of meters. It is an important area for sediment deposition and erosion. It is influenced by various geological processes, including wave action, tidal currents, and sediment transport. These processes result in the formation of distinct geological features and sedimentary deposits such as sandbars, sand waves, and ripples. Sediment transport in the inner shelf is typically driven by longshore currents, which moves sediments along the coastlines, parallel to the shore. The inner shelf is also influenced by tidal currents, which causes sediments to be re-suspended and transported. Tidal currents create tidal deltas; the latter are accumulations of sediment at the mouth of estuaries or river mouths. The inner shelf is often characterized by a mixture of sediment types, including sand, silt, and clay. The composition of the sediment varies depending on factors such as the proximity of sediment sources, the intensity of wave and current activity, and the geology of the adjacent coastline.

18.3.2 Middle Shelf

The middle shelf is located between the inner shelf and the outer shelf (Figure 18.1B) and is characterized by intermediate water depths, typically ranging from tens to hundreds of meters. It is influenced by wave action, tidal currents, and sediment transport resulting in the formation of distinct geological features and sedimentary deposits. Wave energy and wave-driven currents, cause erosion of the seafloor and the transport of sediments resulting in the formation of sandbars, sand waves, and ripples on the seafloor. Sediment transport is typically driven by longshore currents, which move sediment along the coastline parallel to the shore. Tidal currents also play a role in shaping the middle-shelf environment. They cause sediment re-suspension and transport, contributing to the overall sediment dynamics of the area. Tidal currents create tidal deltas and tidal channels, important features on the middle shelf.

18.3.3 Outer Shelf

The outer shelf is located between the middle shelf and the shelf break (see Figure 18.1B) and is characterized by relatively deeper water depths, ranging from hundreds to thousands of meters. The outer shelf is often subject to wave energy and wave-driven currents, which cause erosion of the seafloor and the transport of sediments leading to the formation of sandbars, sand waves, and ripples. The outer shelf is often characterized by a mixture of sediment types, including sand, silt, and clay.

18.4 SHELF SEDIMENT TRANSPORT AND DEPOSITION

Shelf sediment transport and deposition are driven by a combination of factors, such as waves, currents, tides, and gravity. The deposition of sediment on the continental shelf occurs when the energy of the transporting forces decreases, allowing sediments to settle out and accumulate. This happens in areas of reduced wave or current energy, such as behind sandbars, in embayments, or in areas sheltered from strong currents. Sediment deposition on the shelf contributes to the formation of various sedimentary features, including sandbanks, mud flats, and submerged deltas. Wave action is the primary driver of sediment transport on the continental shelf. Waves approaching the shore generate currents that move sediment along the shoreline and parallel to it; this is called the longshore sediment transport. As waves break and dissipate energy, they also cause sediments to be suspended in the water column (suspension transport) or to be transported along the seafloor (bedload transport). Currents, both tidal and non-tidal, play a significant role. Tidal currents are generated by the gravitational pull of the moon and the sun and cause sediments to be transported. Non-tidal currents, such as those driven by winds or density differences, influence sediment transport by carrying sediments along the shelf, by eroding the seafloor, and by depositing sediments in certain areas. Tides also contribute to sediment transport and deposition; they cause sediments to be re-suspended and transported during ebb and flood tides. The rise and fall of tides form tidal deltas that are accumulations of sediments at the mouth of estuaries or rives. Gravity-driven processes, such as mass wasting and downslope sediment transport, occur on the continental shelf; these move the sediments downslope due to gravitational forces. Mass wasting results in the formation of submarine landslides and turbidity currents that transport large volumes of sediments downslope.

18.4.1 Palimpsest

A palimpsest refers to a geological feature that has been shaped by multiple events or processes over time, leaving behind evidence of its complex history. These features often bear the imprint of past geological events, such as erosion, deposition, or tectonic activity. A palimpsest also refers to layers of rock or sediment that have been exposed and subsequently covered by new layers, preserving a record of the past. Palimpsests unravel the history of a particular area and throw light on the processes that might have shaped it over time.

18.5 TYPES OF SHELVES

The shelf is of two types: wave- and storm-, or tide-dominated (see Figure 18.2A). However, most shelves are influenced by a mixture of these two processes. Approximately 80% of modem shelves are wave- and storm-dominated, and 17% are tide-dominated (see Swift et al., 1986). About 3% of shelves are dominated by intruding ocean currents (i.e., circulating cells or gyres; see below). These types of shelves are briefly enumerated below.

18.5.1 Wave- and Storm-Dominated Shelves

These shelves are shaped by the dominant processes of wave action or storm events and typically occur in areas where wave energy is high and consistent, such as along open coastlines (see Figure 18.2A). The waves erode the shoreline, transport sediments along the shelf, and deposit them in specific areas. As a result, wave-dominated shelves often have a gently sloping profile with sediment being transported offshore and deposited in deeper waters.

Storm-dominated shelves, on the other hand, are shaped by intense and episodic storm events; these are typically found in regions where storms are frequent and powerful, such as in the vicinity of tropical cyclone-prone areas. Storms generate strong currents, large waves with power to do significant sediment transport, thus leading to the reshaping of the shelf. As a result, these shelves often have a more irregular and dynamic morphology (as compared to wave-dominated shelves), with sediments being distributed in a patchy manner due to the episodic nature of storm events.

Fair-weather waves, swells, storm waves, wind-driven surface currents, river-generated plumes, and density currents form part of the complex spectrum of transport processes operating on wave- and storm-dominated shelves. These are briefly enumerated below. But to better understand how waves transport sediments, it is imperative to first understand how waves operate and their basic terminology (see Figures 18.2B–C).

18.5.1.1 Basic Wave Terminology

The orbital motion of water generated by the passage of orbital waves dies out downward at a depth equal to about one-half of the wave length (wave length is the distance between two adjoining wave crests); this depth is called the wave base (Figure 18.2B). As the orbital motion in deep waters is unobstructed by the sea bottom, the wave orbits are nearly circular (Figure 18.2B). As a wave moves into shallower waters, where depth is less than one-half the wave length, the sea bottom begins to interfere with the orbital motion of the wave, and thus affects the shape of the wave orbit (Figure 18.2B). In very shallow waters, where water depth is less than about 1/20 the wave length (Figure 18.2C), the motion of the particles is strongly affected, and the orbits become more elliptical (Figure 18.2B). They become progressively flatter downward below the surface, until near bottom they are essentially linear, generating a to-and-fro oscillating motion as the wave passes producing bidirectional flows along the seafloor. The velocity of this bottom flow is called the orbital velocity and is commonly greater in one direction than the other. When the velocity flow exceeds the threshold of grain movement, net transport happens.

18.5.1.2 Fair-Weather Waves

Fair-weather waves, also known as wind waves or ordinary waves, are the regular, small-scale waves that occur on the surface of the ocean or other bodies of water bodies under normal, non-stormy weather conditions (see Figure 18.2B). The fair-weather waves are generated by the wind blowing across the water surface. The fair-weather wave base is shallow, so the movement of sediments takes place largely on the innermost shelf in the onshore direction (Figure 18.2B). The fair-weather wave base is commonly in the order of 10–15 m. The breaking of waves in very shallow waters generates currents (wave swash), in an onshore direction, and longshore currents, directed laterally

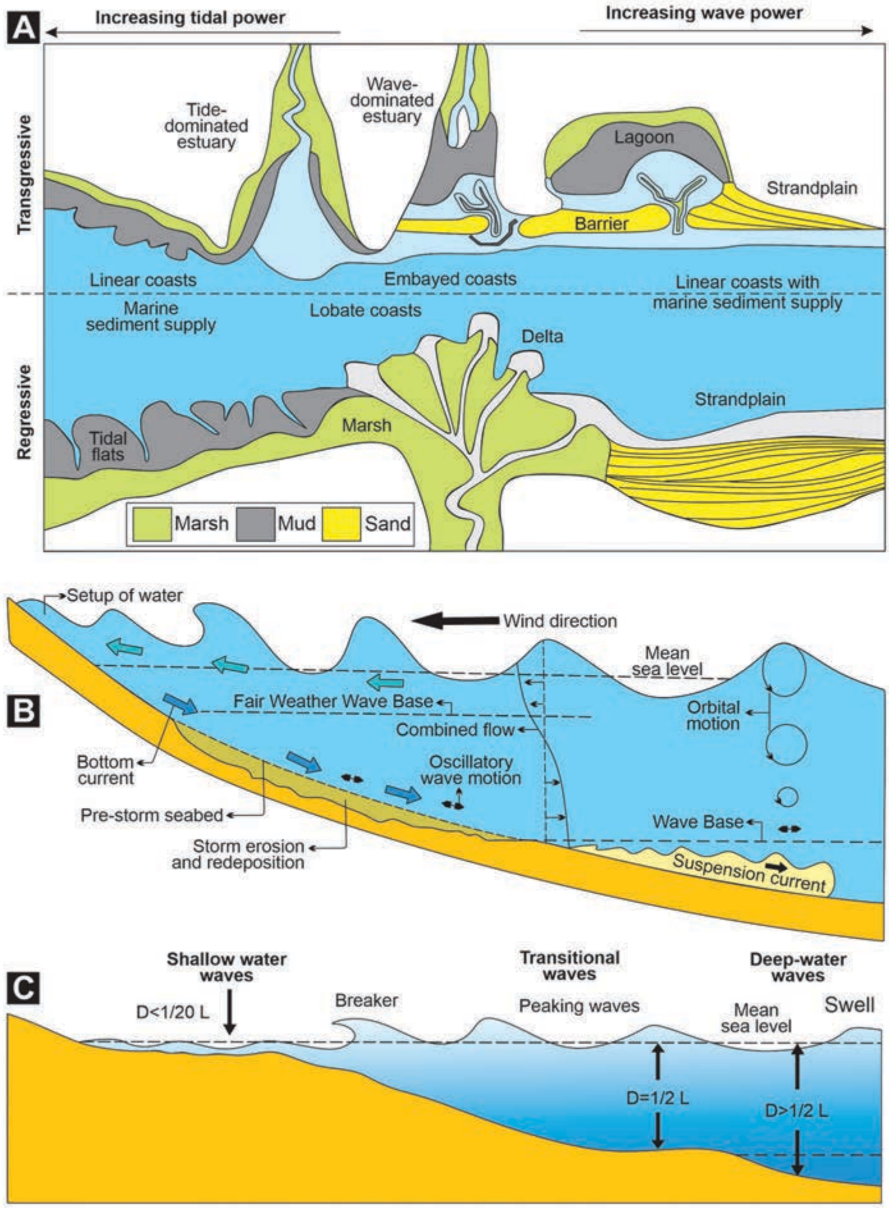

FIGURE 18.2 Depositional setting and wave terminology. A: Characteristic features of wave- and tide-dominated shelves. (Modified after Boyd et al., 1992; Dalrymple et al., 1992.) B: Wave characteristics and orbital motion. C: Wave length. In very shallow waters, where water depth is less than about 1/20 the wave length, the motion of the particles is strongly affected, and the orbits become more elliptical (B).

along the shore. This overall shoreward movement of water in the nearshore zone above the fair-weather wave base traps sediments in the nearshore zone. In general, fair-weather waves are typically smaller in size and less powerful as compared to waves generated during storms or extreme weather events; they also have a shorter wavelength and lower wave height. The size of fair-weather waves is influenced by factors such as wind speed, duration, and fetch (the distance over which the wind blows uninterrupted) (see Figure 18.3A).

18.5.1.3 Swells, Wind-Forced Currents, Geostrophic Flows, and Storm Waves

Swells are low-relief (low wave heights), long-period, regular and uniform wave patterns, and long-wavelength waves generated by storms that may originate far out at the sea (Figures 18.3B–C). They travel across the ocean without the direct influence of local winds. They can travel thousands of kilometers before reaching the coastline and are often a result of strong winds blowing over a large area for an extended period of time. (Figure 18.3A). Swells are typically more organized and less

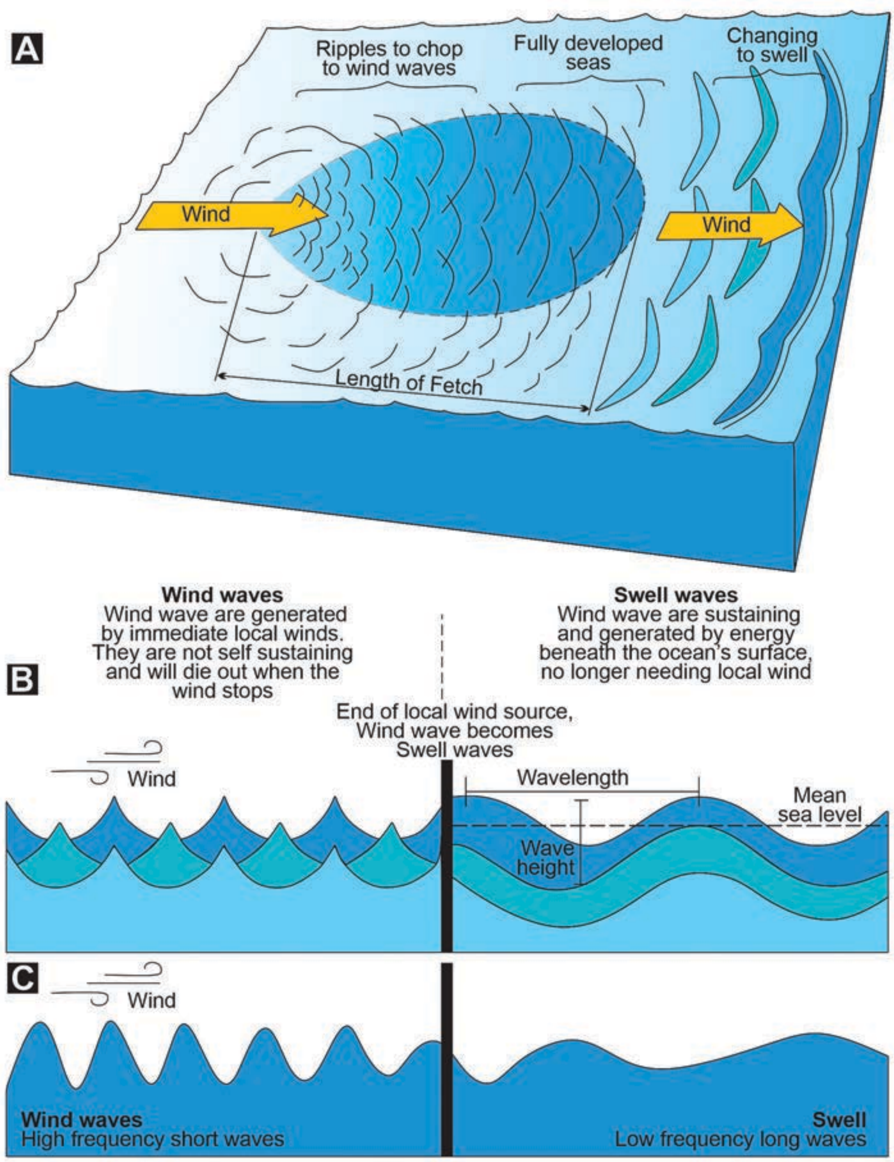

FIGURE 18.3 Swells, wind-forced currents, and storm waves. A: The size of fair-weather waves is influenced by factors such as wind speed, duration, and fetch (the distance over which the wind blows uninterrupted). (Modified after Novo, 2018.) B: Difference between wind and swell waves. C: Amplitude variation between wind and swell waves.

chaotic than storm waves. When swells move onto the shelf they tend to "stir" the bottom to greater depths than do fair-weather waves and have an even greater effect on shelf transport and deposition.

Wind-forced currents are the result of the wind blowing over the surface of the ocean, creating a force that pushes the water in a particular direction. These currents are influenced by factors such as wind speed, duration, and fetch (the distance over which the wind blows uninterrupted) (Figure 18.3). Wind-forced currents are unidirectional currents generated by wind shear stress as wind blows across the water surface, gradually putting into motion deeper and deeper layers of water (i.e., Ekman transport).

The deeper layers of water are deflected by the Coriolis force, so that their direction of movement diverges from that of surface layers (see Figure 18.4A). The Coriolis force is generated by the earth's rotation, causing moving objects to be deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere (Figure 18.4B). If the velocity and duration of wind are great enough, then the water movement may extend to the seabed with enough velocity to transport sediments. Highly energetic storm waves in the shore zone vigorously erode the beachface and upper shoreface and the sediment is flushed seaward, mainly by a combination of storm-enhanced rip currents and downwelling currents that evolve from wind-driven or wind-forced currents (Figure 18.4A).

Strong winds commonly create wind-forced currents that flow parallel to shore and therefore do not provide much offshore sediment transport. If, however, currents moving along the shoreline are deflected landward owing to the Coriolis force, an onshore pile-up of water takes place. Piling up of water onshore creates an elevation of the water surface, i.e., a coastal setup (Figure 18.5). The different water levels at the coast and offshore result in a hydrostatic pressure difference on the ocean floor that drives a bottom flow seaward (downwelling) (Figure 18.5). As the bottom water flows seaward, it is deflected laterally to form a geostrophic current (Figure 18.5). These geostrophic flows achieve velocities at water depths of 10–20 m to as much as 60 cm/s (Walker and Plint, 1992). Flows of this magnitude are not capable of transporting much sandy sediments unless they are accompanied by strong wave-driven oscillatory motion at the sea bed. Oscillatory wave motion provides the necessary shear stress needed to lift grains off the bottom which are then transported by the geostrophic currents (see Snedden et al., 1988) (see Figure 18.5). Unidirectional and oscillatory currents operating together are called combined flows (see Figure 18.2A). However, it must be noted that sediment moves obliquely offshore, and some sand moves from the beach shoreface into offshore settings. But geostrophic currents tend to be parallel to isobaths (parallel to the shore) and are not directed seaward. Hence, sediment is not transported by such currents to any great distance outward onto the shelf.

Storm waves affect sediment movement in deeper waters on the middle and outer shelf. Because of their longer wave length and period, storm waves moving across a shelf are able to rake the seafloor to depths of as much as 200 m (Komar et al., 1972). The orbital velocities generated by these storm waves may be nearly equal so that no net seaward transport of sediment occurs; however, the waves re-suspend bottom mud and tend to spread or dissipate it around the seafloor (see Figure 18.2A). Thus, a net sediment transport occurs either seaward or landward.

18.5.1.4 Sediment Plume

It is the suspension and transport of sediment particles in water, and creating a visible cloud-like or turbid appearance. These plumes occur in various aquatic environments, including rivers, estuaries, coastal areas, and even in the open ocean. They are typically formed by natural processes such as riverine input, wave action, currents, and tides. When sediment is disturbed or re-suspended from the seafloor or riverbed, it becomes suspended in the water column and forms a plume. The sediment discharged at river mouths into the ocean is carried onto the shelf either as buoyant plumes or underflows (bottom plumes) (see Figure 18.5). Buoyant (hypopycnal) plumes commonly develop when the low-salinity water coming from rivers and estuaries flows out on top of a higher-salinity

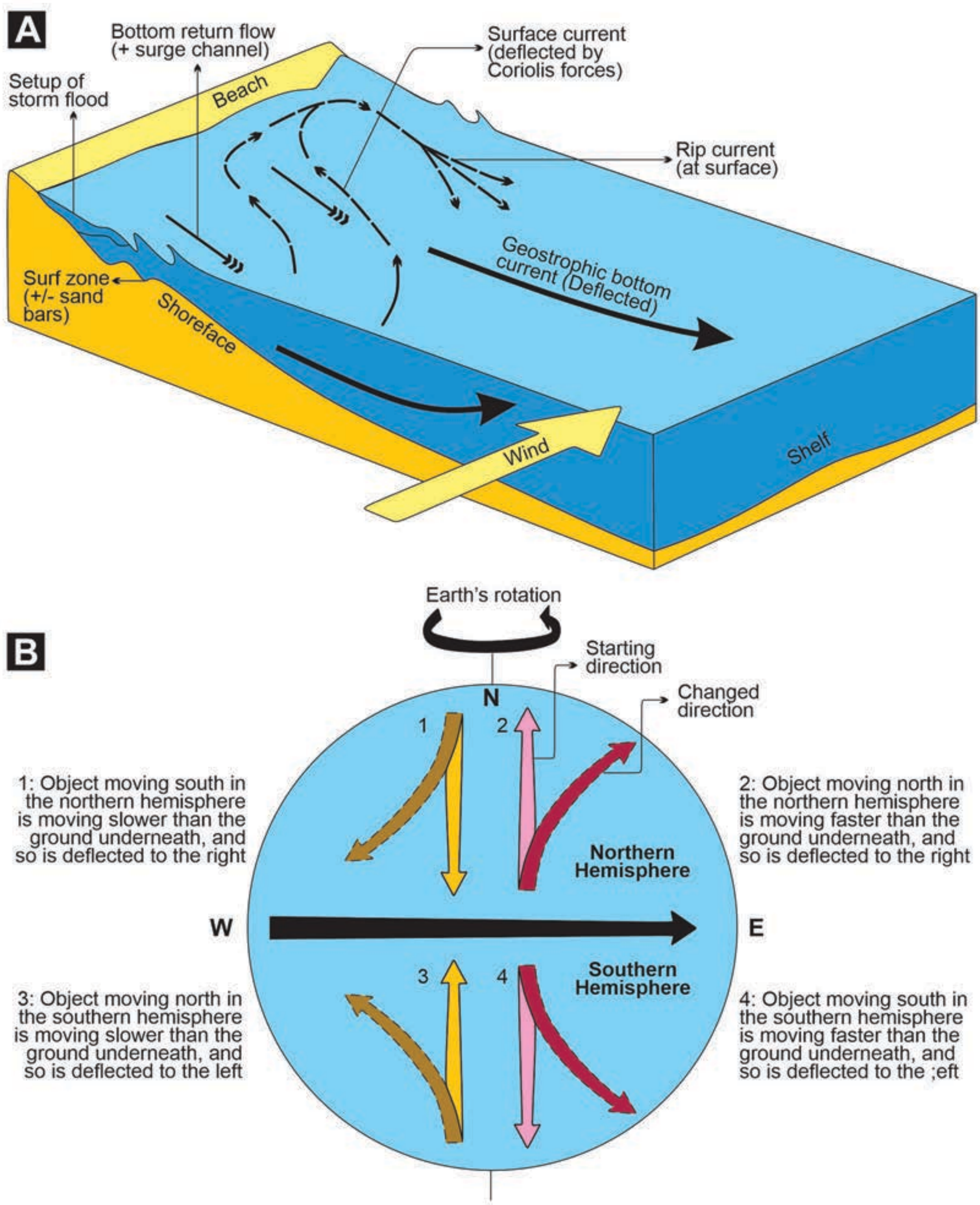

FIGURE 18.4 Geostrophic currents. A: Energetic storm waves in the shore zone vigorously erode the beachface and upper shoreface and the sediment is flushed seaward, by a combination of storm-enhanced rip currents and downwelling currents that evolve from wind-driven or wind-forced currents. B: Coriolis force and currents movement. The Coriolis force is generated by the earth's rotation, causing moving objects to be deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere.

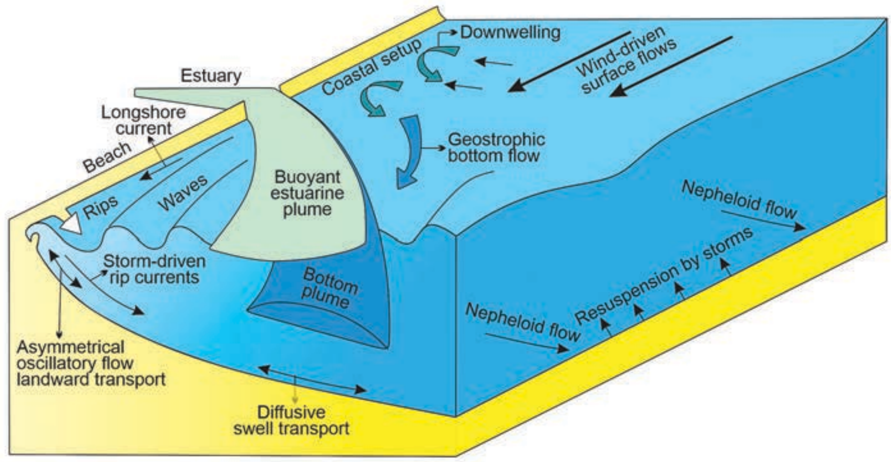

FIGURE 18.5 Major shelf processes for sediment transport. (Modified after Nittrouer and Wright, 1994.)

seawater (see Figure 18.5). These plumes generally do not reach farther than the inner or middle shelf before being carried parallel to the coast owing to the Coriolis force (see Nittrouer and Wright, 1994). There, the suspended fine sediment (the coarser sediment drops out near the river mouth) gradually settles to the shelf floor. Thus, the size, composition, and behavior of a sediment plume depends on several factors, including the characteristics of the sediment particles (such as grain size, density, and settling velocity), water currents and turbulence, and the duration and intensity of the disturbance. Fine-grained sediments, such as silt and clay, tend to stay suspended in the water for longer periods and travel long distances, while coarser sediments settle more quickly.

18.5.1.5 Nepheloid Flow

Nepheloid flow refers to the movement of suspended sediment particles in the water column of oceans or other bodies of water. It is a type of turbidity current, i.e., a downslope movement of sediment-laden water (Figure 18.5). The nepheloid layer may reach heights of several hundred meters above the seafloor. This layer is denser than the surrounding ambient water but not dense enough to sink rapidly. Owing to low settling velocity, the fine sediments remain in suspension within the nepheloid layer for periods ranging from days to weeks in the lowest 15 m of the water column and from weeks to months in the lowest 100 m (Kennett, 1982). The movement of nepheloid flows is driven by gravity and the density difference between the sediment-laden water and the surrounding water. As the sediment-laden water flows downslope, it transports and deposits sediments over large distances, thus shaping the seafloor and influencing sediment distribution.

18.5.2 Sediment Characteristics of Storm-Dominated Shelves

Storm-dominated shelves are characterized by high-energy wave and current conditions that result in the continuous re-suspension and transport of sediments. Storm-dominated shelves predominate on most of the world's coasts. Modern examples of storm-dominated shelves include the Atlantic shelf off the eastern coast of the United States, the Pacific shelf off Oregon and Washington, and the Bering Sea. The storm-dominated shelves are characterized by low tidal current velocities (<25 cm/s), with shallow fair-weather wave base (~10 m). Hence, little coarse sediment moves on these shelves except during intense storms. The sedimentation pattern depends on the extent to which the shelves are mantled with – either relict sediments or modern sediments. Shelves with abundant relict sediment, such as the Atlantic, are characterized by sand bodies (shelf sand ridges) (detailed below). Shelves with a greater fraction of modern sediments, such as the Pacific shelf off Oregon and Washington (USA), are characterized by a greater proportion of finer-grained and thoroughly bioturbated sediments (mudstone/shale). Additionally, the intense wave action and currents on storm-dominated shelves result in poor sorting. The sediments of different sizes and densities are mixed and transported together, resulting in poorly sorted deposits.

Storm-dominated siliciclastic shelves are also characterized by coarse-grained storm layers and hummocky cross-stratification (see Figure 18.6A). The storm layers are thin and interlayered or embedded in finer-grained muds (Figure 18.6A). The coarser material typically consists of coarse silt, fine sand, shell fragments or, less commonly, gravel. The storm layers characteristically show vertical size grading (see Figure 18.6A). Hummocky cross-stratification consists of curving, gently dipping laminae, both convex-up (hummocks) and concave-up (swales) that intersect at low angles (Figure 18.6A). The Middle to Upper Cambrian of the Helan Mountains, northwest China provides a good example of storm-influenced shelf carbonate deposits (Figure 18.6B; see also Chuanmaol et al., 1993). Four major kinds of storm-influenced tempestite associations are noted (see Figure 18.6B). These are 1) supra-intertidal stromatolite, 2) barrier skeletal-oolitic, 3) upper shelf proximal micrite, and 4) lower or distal shelf shale-micrite tempestite association (see Figure 18.6B). They are also highstand facies tracts and transgressive facies tracts. Each parasequence is characterized by a coarsening-upward vertical facies associations, and a gradual shallowing is followed by an abrupt deepening, producing maximum flooding surface at the parasequence boundary (Figure 18.6C).

18.6 TIDE-DOMINATED SHELVES

Tide-dominated shelves are primarily shaped and influenced by tidal processes and hence are typically found in areas with a large tidal range, such as estuaries, bays, and coastal lagoons. Tidal processes, such as tidal currents and tidal fluctuations, play a significant role in shaping the morphology and sedimentary deposits of tide-dominated shelves. The dominant tidal currents transport sediment along the shelves, causing erosion in some areas and deposition in others forming various landforms such as tidal flats, tidal channels, and tidal deltas (see Figure 18.2A). The sediment is often well-sorted, and fine-grained, consisting of silt and clay particles.

18.6.1 Tidal Processes

Tidal processes include the movement and fluctuation of ocean waters caused by the gravitational forces exerted by the moon and, to a lesser extent, by the sun (see Figure 10; Chapter 14). These processes result in periodic changes in sea level, known as tides that occur twice a day in most coastal areas. The rise and fall of tides is accompanied by horizontal movements of water called tidal currents. These currents generated on the shelf are bidirectional but asymmetrical with respect to velocity, i.e., the flood- and ebb-tide velocities are different. Such asymmetrical currents result in sediment transport in the direction of the stronger current. If both current phases are able to transport sand, herringbone cross-stratification (the foresets in successive sets are directed in opposite directions due to reversing currents, resembling somewhat the bones of a fish) and reactivation surfaces (erosional surface within a set of cross-beds), form (see Figures 18.7A–B). In some tidal environments, couplets of sand-mud laminae, with mud deposited on top of sand during a standstill are also produced; their stacking produces tidal successions called tidal rhythmites or tidalites (detailed below; see Figure 18.7C). Such beds show cyclic changes in thickness that represent neap-tide and spring-tide variations in tidal current velocity (see also Dalrymple et al., 1991; Liu et al., 2022).

There are two main types of tides: spring tides and neap tides (see Figure 14.10; Chapter 14). Spring tides occur when the gravitational forces of the moon and the sun align, resulting in higher high tides and lower low tides (Figure 14.10; Chapter 14). Neap tides, on the other hand, occur when

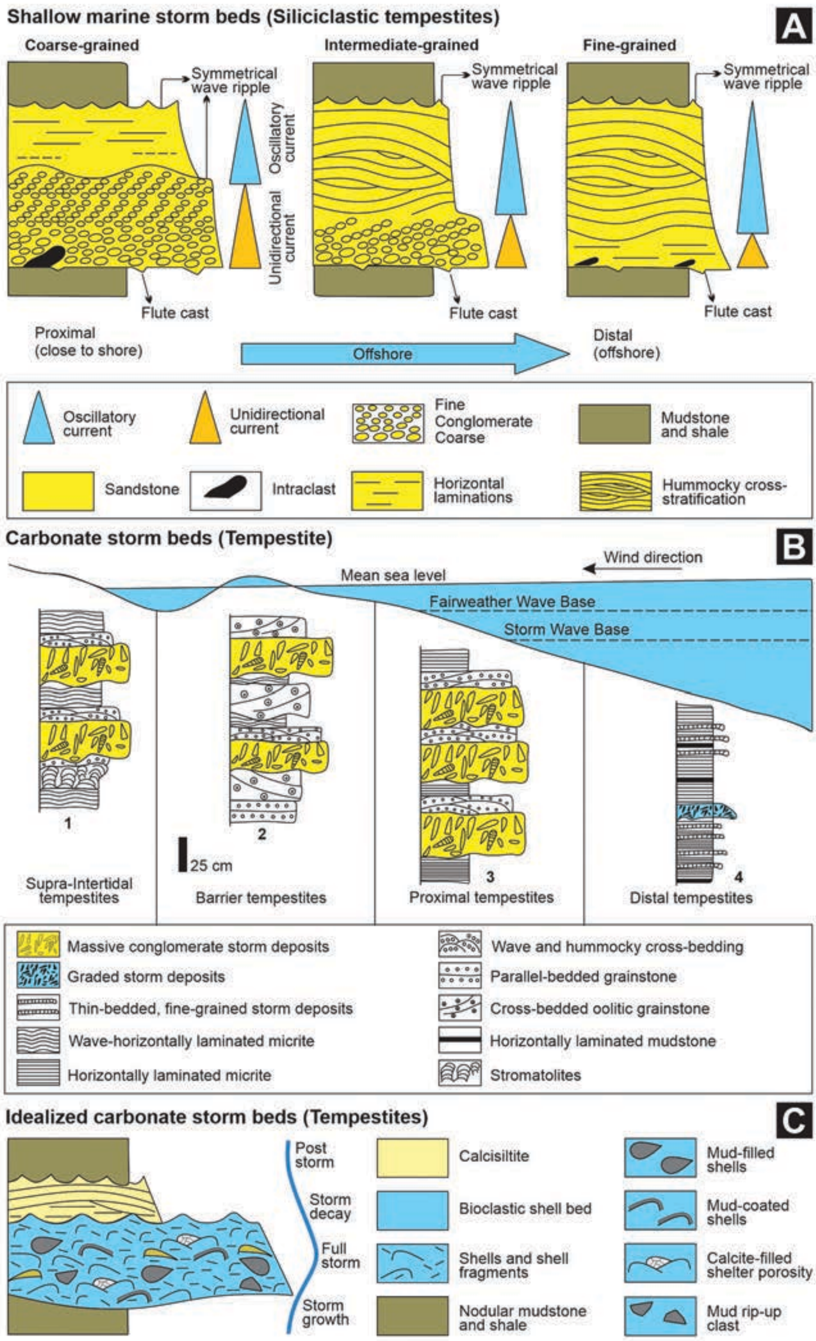

FIGURE 18.6 Storm-dominated siliciclastic shelf facies, sedimentary structures, and tempestites. A: Storm-bed characteristics in a shore (coarse-grained) to offshore (fine-grained) sedimentary sequence and structures. (Modified after Cheel and Leckie, 1992.) The swaley stratification in sharp-based very fine sandstones topped by wave ripples is typical of tempestite deposits. B: Relationships between the four kinds of carbonate tempestite associations. (Modified after Chuanmaol et al., 1993.) C: An idealized carbonate storm bed, showing how sea level affects storm-dominated siliciclastic shelf facies as reflected in an upper shelf proximal tempestite association (see A; modified after Chuanmaol et al., 1993.)

the gravitational forces of the moon and the sun are perpendicular to each other, resulting in lower high tides and higher low tides (Figure 14.10; Chapter 14). As tides move across the ocean, they generate tidal currents, the horizontal movements of water caused by the rising and falling tides. Tidal currents can be quite strong, particularly in narrow channels and estuaries, and have a significant impact on sediment transport and coastal erosion. It must be noted that much of sediment movement by tidal currents occurs when they are aided by wave action. The orbital motion of waves is sufficient to lift grains off the seafloor, which are then transported some distance by the currents (Komar, 1976). The tidal processes are influenced by several factors, such as the shape of the coastline, the depth of the water, and the topography of the ocean floor.

18.6.2 Tidal Rhythmites or Tidalites

Tidal rhythmites, also known as tidalites, are sedimentary rock formations characterized by rhythmic layering or bedding reflecting the influence of tidal processes (see Figure 18.7C). These rhythmites are typically found in coastal environments and are formed as a result of the regular and cyclical nature of tides. They exhibit distinct alternating layers or laminae that vary in thickness and composition (see Figure 18.7C). The layers are typically composed of fine-grained sediment, such as silt and clay, that settle out of suspension during periods of low energy during tides. These fine sediments are deposited in a horizontal or gently inclined manner (see Figure 18.7C). During each tidal cycle, there are two high and two low tides (see Figure 14.10; Chapter 14). The high tide brings in water and sediments, while the low tide results in reduced water energy and sediment settling (deposition). This cyclical pattern leads to the formation of alternating layers in the sediment, with each

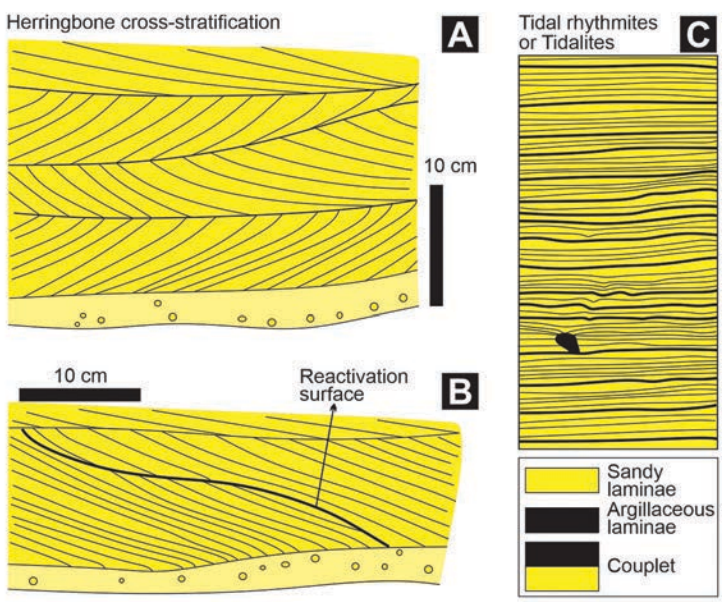

FIGURE 18.7 Tidal sedimentary structures. A: Herringbone cross-stratification (HCS). The foresets in successive sets are directed in opposite directions due to reversing currents, resembling somewhat the bones of a fish. B: Reactivation surface (erosional surface within a set of cross-beds). C: Couplets of sand-mud laminae. The mud is deposited on top of sand during a standstill; their stacking produces tidal successions called tidal rhythmites or tidalites. (Modified after Liu et al., 2022.)

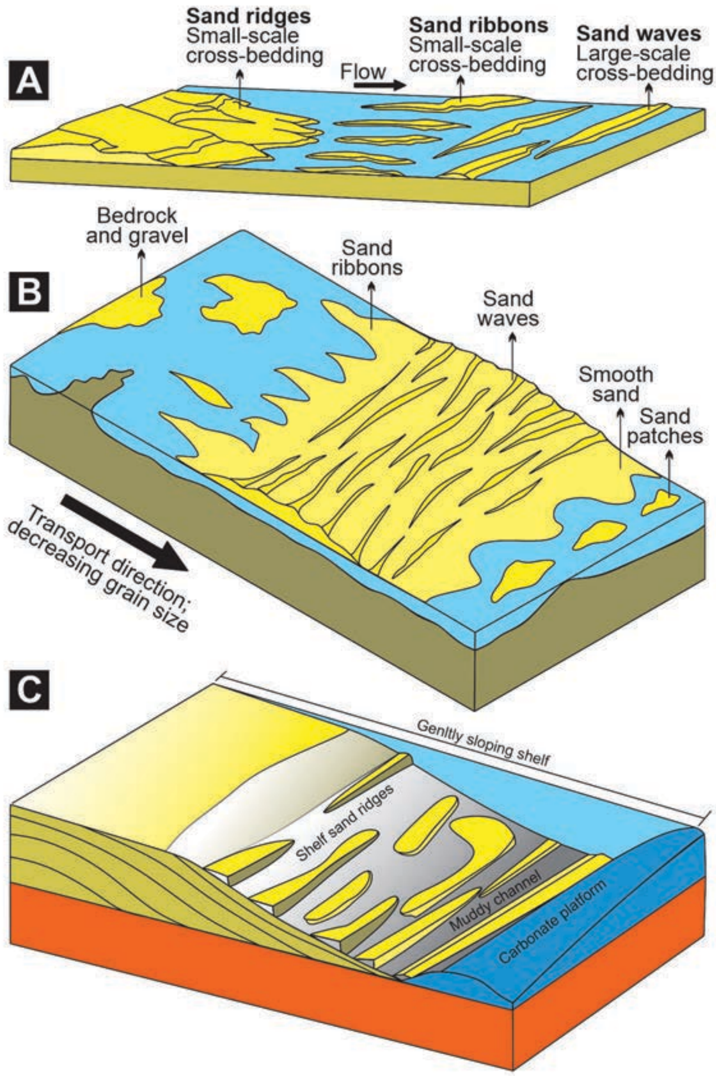

FIGURE 18.8 Sand bodies in a tide-dominated shelf. A: Bedforms formed on a tide-dominated shelf. B: At high tidal velocities, the seabed is eroded, leaving furrows and gravel waves. With reduced velocities, the eroded sediments are deposited to form flow-parallel sand ribbons, large and small dunes, a rippled sand sheet, and sand patches. C: The shelf sand ridges similar to tidal sand ridges commonly occur on wave- and storm-dominated shelves.

layer representing one tidal cycle, thus providing valuable information for reconstructing paleotidal patterns, estimating tidal ranges, and changes in sea level over time.

18.6.3 Sediments of Tide-Dominated Shelves

The sediments of tide-dominated shelves are typically fine-grained and well-sorted, consisting of silt and clay, often deposited in layers or beds, reflecting the cyclical nature of tidal processes (see Figure 14.10; Chapter 14). In areas with higher tidal ranges, thicker sediment layers accumulate, while areas with lower tidal ranges, thinner layers are formed.

Tide-dominated shelves are characterized by sand bodies such as large sand waves (i.e., large-scale cross-bedding; see Figure 18.8). These can be symmetrical (if produced by tidal currents with equal ebb and flood peak speeds) or asymmetrical (caused by unequal ebb and flood velocities). The tidal sand ridges (see Figure 18.8) similar to the shelf sand ridges (see Figure 18.8C) commonly occur on wave- and storm-dominated shelves. In addition to sand waves and shelf sand ridges, tide-dominated shelves also have sand patches, sand sheets, and gravel sheets, all characterized by small-scale bedforms, and patches of bioturbated muds in areas sheltered from tidal currents and waves (see Figure 18.8) (see also Stride et al., 1982). At high tidal velocities, the seabed is eroded, leaving furrows and gravel waves (Figure 18.8B). With reduced velocities, the eroded sediments are deposited to form flow-parallel sand ribbons, large and small dunes, a rippled sand sheet, and sand patches (Figure 18.8). Sand ridges form in the dune belt if enough sand is present (see Figure 18.8). Most tidal shelf sands are characterized by cross-bedding (Figure 18.8). Small-scale cross-bedding and ripple cross-lamination, produced by the migration of ripples and small dunes, and large-scale cross-bedding generated by the migration of dunes and sand ridges are commonly noted (see Figure 18.8).

18.7 SHELVES AFFECTED BY INTRUDING OCEAN CURRENTS

The flow of major surface ocean currents is driven by prevailing winds around the earth. These winds, in conjunction with the Coriolis force arising from the earth's rotation, and drive the ocean currents into gigantic circulating cells, or gyres, that rotate clockwise (deflected to the right) in the Northern Hemisphere (North Pacific and North Atlantic) and counterclockwise in the Southern Hemisphere (South Pacific, South Atlantic, Indian Ocean) (see Figure 18.4B). Shelves affected by these intruding ocean currents are often referred to as boundary current-dominated shelves.

Boundary currents are large-scale ocean currents that flow parallel to the coastline and are driven by the earth's rotation and wind patterns. These ocean currents intrude onto some shelves (mostly in the outer shelf) with enough bottom velocity to transport fine sandy sediments such as the north-western Gulf of Mexico, affected by the Gulf Stream system. Some have bottom velocities large enough (such as Kuroshio and Agulhas currents) to transport sandy sediments and form sand waves, sand ribbons, and coarse sand and gravel lag deposits (see Boggs et al., 1979) (see Figure 18.8). The Taiwan Strait between Taiwan and China is a broad shelf invaded by ocean currents forming extensive fields of sand waves (Boggs, 1974) (see Figure 18.9). Boundary current-dominated shelves

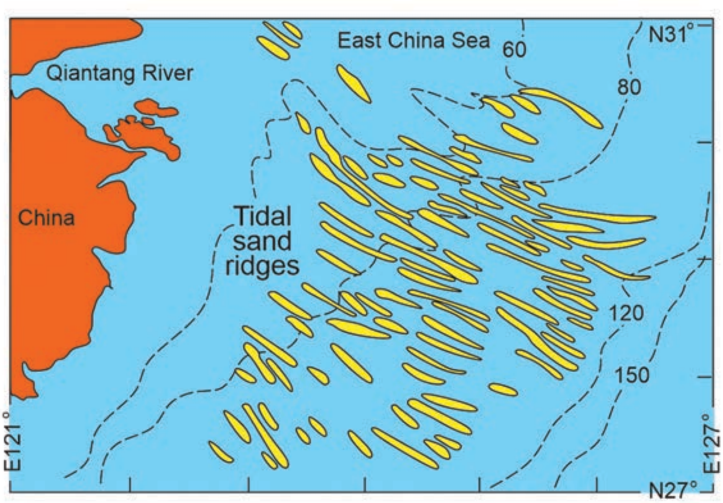

FIGURE 18.9 Tidal sand ridges in the East China Sea. (Modified after Liu et al., 2007.)

typically experience strong and persistent currents that can transport large amounts of sediment along the coast. These currents can erode, transport, and deposit sediments, shaping the morphology and sediment distribution of the shelves. The sediment transport occurs in various ways, including alongshore transport, cross-shelf transport, and offshore transport.

The interaction between boundary currents and shelves also results in the formation of distinct sedimentary features. For example, longshore bars and ridges develop parallel to the coastline due to the alongshore sediment transport by the boundary current. These shelves are often characterized by a mixture of sediment types, including both coarse and fine-grained sediments. Coarser sediments, such as sand and gravel, are typically found in areas of high energy and along the outer edges of the shelves where the boundary currents are strongest. On the other hand, finer-grained sediments, such as silt and clay, are more likely to accumulate in areas of lower energy, such as embayments and estuaries. The sedimentary characteristics of boundary current-dominated shelves varies depending on the strength and direction of the boundary current, the local geomorphology, and the availability of sediment sources. Other processes, such as waves, tides, and river input, also influence sediment dynamics and composition of these shelves.

18.8 SHELF TRANSPORT BY DENSITY CURRENTS

Shelf transport by density currents is the movement of sediments along the seafloor of continental shelves driven by differences in water density. Density currents occur when there is a contrast in the density of water masses, typically caused by variations in temperature, salinity, or sediment concentration. Density currents are initiated by various processes, such as river plumes, tidal mixing, or density differences in coastal waters. When these currents reach the continental shelf, they influence sediment transport and deposition patterns. Some density currents such as buoyant plumes, nepheloid flows and underflows (dense brines that flow seaward along the sea bottom), driven by density differences arising from suspended sediments (see Figure 18.5), are not significant agents of shelf transport and no modern shelf is dominated by such processes. The sediment transported by density currents varies in size and composition depending on the source and characteristics of the sediment. Coarser sediments, such as sand and gravel, are more likely to be transported by density currents with higher energy and velocity. Finer-grained sediments, such as silt and clay, can remain suspended in the water column for longer periods and be transported over longer distances. As density currents flow along the shelf, they pick up sediments from the seabed, suspending them in the water column. This suspended sediment is transported downslope, creating sediment gravity flows called turbidity currents. These are a type of current that carries sediments in suspension, creating a dense, sediment-laden flow. Such currents are able to transport large amounts of sediment downslope, eroding the seafloor and depositing them in submarine canyons, channels, and basins on the continental shelf.

18.9 EFFECTS OF SEA-LEVEL CHANGE ON SHELF TRANSPORT

Sea-level changes have a significant effect on shelf transport processes. As sea level rises or falls, the water depth over the continental shelf changes, altering the dynamics of sediment transport and deposition. Thus, changes in sea-level affect both erosional and depositional processes, and the kinds of sediments deposited on shelves, thus, reshaping the stratigraphic architecture of shelf sediments. During periods of sea-level rise, the increased water depth leads to the inundation of previously exposed areas of the shelf. This results in the reworking and redistribution of sediments as they are transported by currents and waves. Sediment that was once located on the shelf may be transported offshore, deposited in deeper water, or reworked along the coastline. Sea-level rise can also cause changes in the strength and direction of coastal currents. As the shoreline moves inland, the circulation patterns of coastal currents are modified, potentially leading to changes in sediment transport pathways. Sediment is transported alongshore or cross-shelf in different directions, thus, impacting the distribution and deposition of sediments on the shelf. During periods of sea-level fall, the exposure of previously submerged areas of the shelf facilitates erosion and the removal of sediments. As water depth decreases, waves and currents become more effective in eroding and transporting sediments. This results in the formation of erosional features, such as submarine canyons or incised valleys; the sediment is removed from the shelf and transported to deeper waters. Sea-level change can also influence the formation and evolution of coastal features, such as barrier islands and estuaries. As sea level rises, barrier islands migrate landward, with the sediment being transported and deposited on the shoreward side. Conversely, during periods of sea-level fall, barrier islands migrate seawards as the sediment is eroded and transported offshore.

18.10 BIOLOGICAL ACTIVITIES ON SHELVES

Biological activities on shelves include various processes, such as bioturbation, bioerosion, and biomineralization. These play a crucial role in shaping the geology of shelves and the interactions between organisms and sediments on the seafloor, thus influencing sediment dynamics, deposition, and erosion processes. Bioturbation is the mixing and reworking of sediments by burrowing organisms, such as worms and bivalves. This process enhances sediment mixing, nutrient cycling, and oxygenation of the seafloor, thus affecting the physical and chemical properties of the sediments. Bioerosion occurs where organisms such as mollusks, sponges, and sea urchins physically erode and break down hard substrates like rocks and shells. Bioerosion leads to the formation of borings, pits, and grooves on the seafloor that have the potential to modify the topography and composition of shelf sediments. Biomineralization is the process by which organisms produce minerals, such as shells and skeletons, through biological processes. Many marine organisms, including corals, mollusks, and foraminifera, build calcium carbonate structures that contribute to the formation of carbonate sediments on shelves. These carbonate sediments accumulate and form reefs, banks, and other carbonate features on the shelf. The presence and activities of organisms on shelves also impacts sediment stability and resistance to erosion. Dense seagrass meadows and kelp forests help stabilize sediments by reducing wave energy and trapping sediment particles with their roots. This action contributes to the formation of cohesive sediments and thus protects the shoreline from increased erosion.

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19 Oceanic (Deep-Water) Environment

19.1 INTRODUCTION

The oceanic environment is characterized by deep waters, far from the influence of the coastline found typically beyond the continental shelf, where the water depth exceeds 200 meters (Figure 19.1A). This area, between the continental slope and the deep ocean floor, occupies ~65% of the earth's surface and forms the deep-water environment (see Figs. 19.1A–B). In general, in the rock record, the deep-water deposits (a) are far less abundant than the shallow-water ones as sedimentation rates are considerably lower in deeper waters resulting in the accumulation of thinner sediment record, (b) are very less studied vis-à-vis the shallow-water deposits due to their greater depth and relatively less economic potential for petroleum exploration (except for turbidites), and (c) are also subjected to destruction (of the deep seafloor) by subduction in trenches (Figure 19.1C). However, in recent times, increased interest in the deep ocean basins has been noted due to (a) the need for more fossil fuel reserves, i.e., for petroleum exploration in deeper waters, and (b) the possibility of mining manganese nodules and metalliferous muds from the ocean seafloor (see Figure 19.1B).

19.2 DEPOSITIONAL SETTING

The oceanic (deep-water) depositional setting is characterized by some fundamental features such as the accumulation of fine-grained sediments, pelagic sediments, biogenic remains, abyssal plains, turbidity currents, and water depth.

The deep ocean is predominantly composed of fine-grained sediments, such as clay and silt that settle slowly through the water column and accumulate on the seafloor over long periods of time. The pelagic sediments are fine-grained sediments that slowly settle from the water column in the deep ocean (Figure 19.1B). These sediments include clay minerals, siliceous ooze (composed of microscopic silica-based organisms), and calcareous ooze (composed of microscopic calcium carbonate-based organisms) (Figure 19.1B). The oceanic depositional setting is also characterized by the presence of biogenic remains, such as the shells and skeletons of marine organisms. These remains contribute to the composition of pelagic sediments and are preserved as fossils; the accumulation of microscopic planktonic organisms forms chalk deposits. The abyssal plains are vast, flat areas of the ocean floor found in the deepest parts of the oceanic depositional setting, and characterized by the accumulation of fine-grained sediments, including clay and silt (Figs. 19.1A–B). The turbidity currents are sediment-laden currents that flow downslope along the seafloor (Figure 19.1B). They are responsible for transporting and depositing sediment in the deep ocean. The oceanic (deep-water) depositional setting is also characterized by its significant water depth, typically exceeding 200 meters; in the open ocean; at places, the depth can reach up to several kilometers.

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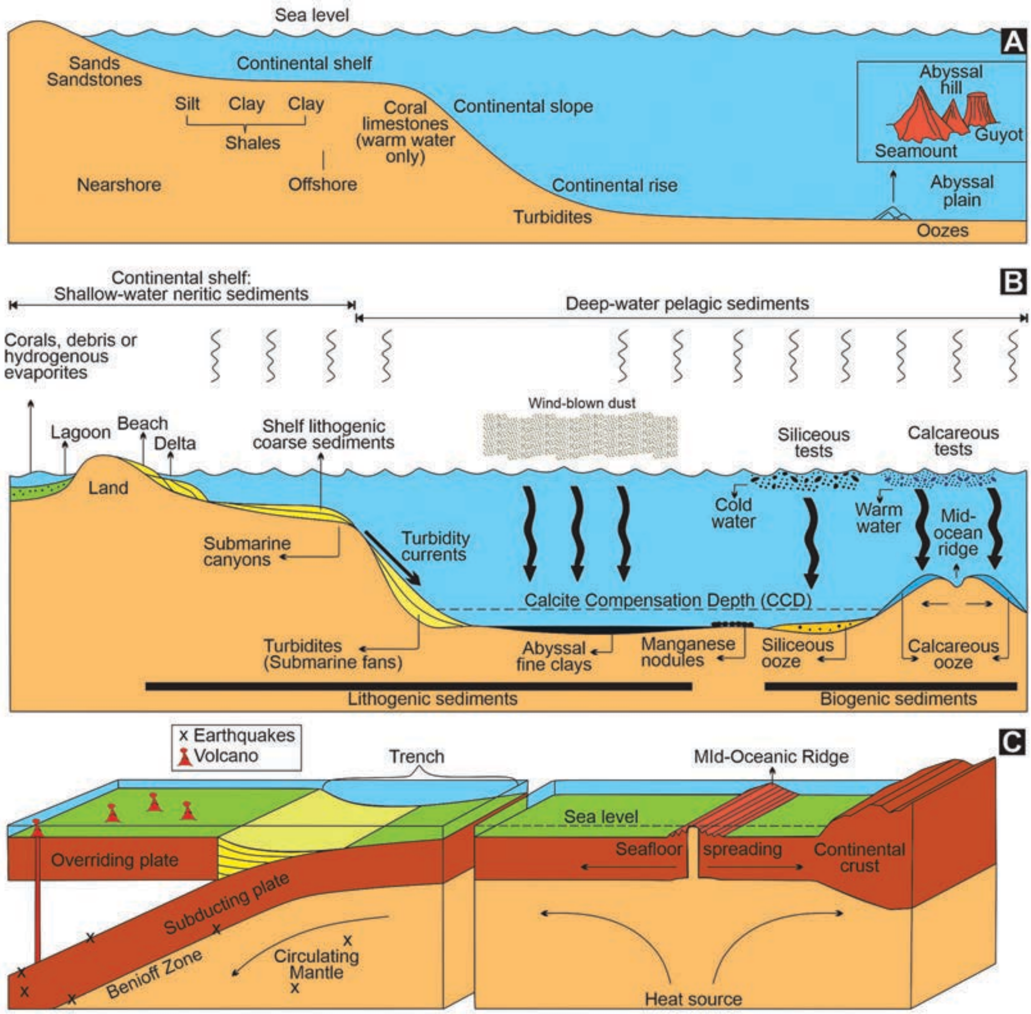

FIGURE 19.1 Divisions of the oceanic (deep-water) depositional setting. A–B: The area between the continental slope and the deep ocean floor forms the deep-water environment. This area is characterized by accumulation of fine-grained sediments, pelagic sediments, biogenic remains, abyssal plains, turbidity currents, and manganese nodules and metalliferous muds at the ocean seafloor (B). C: Oceanic ridges (mid-ocean ridge; MOR) and the deep-water depositional setting. The MOR is a long, underwater mountain range formed by tectonic plate movements where two plates move away from each other. As they move apart, the magma from the mantle rises to fill the gap, creating new crust through a process called seafloor spreading. This results in the formation of a continuous mountain range that extends for thousands of kilometers.

19.2.1 Continental Slope

The continental slope extends from the shelf break, which occurs at an average depth of about 130 m in the modern ocean, to the deep seafloor (see Figs. 19.1–19.2). In general, the continental slopes are comparatively narrow areas (10–100 km wide), and those that dip much more steeply than the continental shelf (Figs. 19.1–19.2). The average dip of modern continental slopes is about 4°, although slopes may range from <2° off major deltas to >45° off some coral islands.

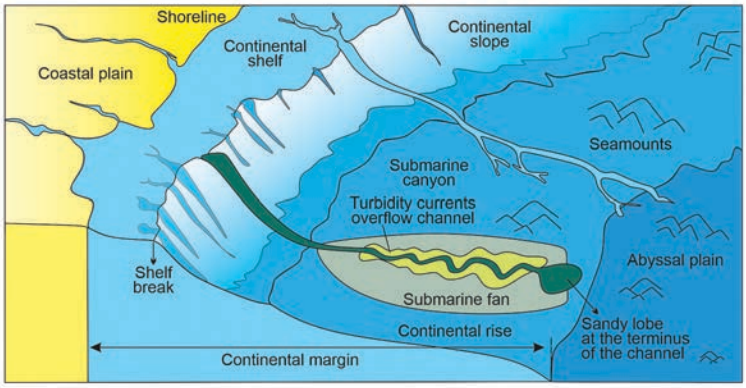

FIGURE 19.2 Continental slope and rise, showing submarine canyons.

The continental slope is characterized by a series of submarine canyons, channels, and gullies formed by erosional processes, such as turbidity currents or underwater landslides that transport sediments and carve out the seafloor (Figure 19.2). The submarine canyons are oriented approximately normal to the shelf break and have their heads near the slope break, and generally do not cross the continental shelf (see Figure 19.2). Some large canyons extend seaward beyond the base of the slope to form deep-sea channels that may meander over the nearly flat ocean floor for hundreds of kilometers (see Figure 19.2). Turbidity currents are the main agents of canyon cutting on the slope and the deeper seafloor, initiated by local slope failure (slumping), followed by the headward growth of erosional scars. The sediment on the continental slope varies in composition, ranging from coarse sands and gravels near the shelf break to finer silts and clays further down the slope. The sediment is often transported by gravity-driven processes, such as turbidity currents or debris flows that rapidly transport large volumes of sediment downslope. Thus, the continental slope serves as a pathway for the transport of nutrients and organic matter from the continental shelf to the deep ocean, supporting diverse ecosystems and marine life. The slope can also be a site for the accumulation of mineral resources, such as oil and gas deposits or manganese nodules.

19.2.2 Continental Rise

The continental rise is the gently sloping transition zone between the continental slope and the deep ocean basin (Figure 19.1A). It is characterized by a thick accumulation of fine sediments such as silts and clays that settle out of the water column and gradually accumulate over time. These are derived from various sources, including rivers, coastal erosion, or biological productivity in the surface ocean. Thus, the continental rise is an important depositional environment, as it serves as a major sink for eroded sediments transported from the continental shelf and slope and deposited onto the deep ocean basin floor.

Off passive continental margins, a continental rise is present at the base of the slope (see Figure 19.2). It is a gently sloping transition zone between the continental slope and the deep ocean basin, and is characterized by a thick accumulation of sediments (fine silts and clays) that are transported from the continental shelf and slope by various sedimentary processes, such as turbidity currents or sediment gravity flows (Figure 19.2). The continental rise is built in part from submarine fans extending seaward from the foot of the slope (Figure 19.2). Thus, the continental rise serves as a major sink for sediment eroded from the continents and is a transition zone where sediments from the continental shelf and slope are transported and deposited before reaching the deep ocean basin floor. The continental rise is also a site for the accumulation of mineral resources, such as oil and gas deposits and manganese nodules.

19.2.3 Deep Ocean Basin

The continental rise and deep ocean basin include the part of the ocean that lies below the base of the continental slope, and make up ~80% of the total ocean seafloor (Figs. 19.1A–B). Thus, the deep ocean basin is a vast and flat expanse of the seafloor that lies beyond the continental slope and rise (see Figure 19.1A). Within the deep ocean basin, the abyssal plains are the dominant feature where the sediments are typically fine-grained such as clay and silt, transported by various processes, including turbidity currents, ocean currents, or settling from the water column (see Figs. 19.1A–B). In general, the deeper part of the ocean beyond the continental slope is divided into two physiographic divisions: the deep ocean floor, characterized by the presence of abyssal plains, abyssal hills (volcanic hills <1 km high), and seamounts (volcanic peaks > 1 km high) (Figs. 19.1A–B), and the deep-sea trench and oceanic ridges (Figure 19.1B-C).

19.2.3.1 Abyssal Plains and Hills

Abyssal plains are extensive, nearly flat areas punctuated here and there by seamounts, gutots, and abyssal hills (see Figure 19.1A). Some abyssal plains are also cut by deep-sea channels (Figure 19.2). Mid-ocean ridges extend across some 60,000 km of the modern ocean and overall make up about 30–35% of the area of the deep ocean (see Figure 19.1C). Mid-ocean ridges (enumerated below in more detail) are particularly prominent in the Atlantic, where they rise about 2.5 km above the abyssal plains on either side. Rocks on these ridges are predominantly volcanic; in general, these are not active areas of sedimentation.

19.2.3.2 Deep-Sea Trenches

A deep-sea trench (oceanic trench or trench) is a long, narrow, and steep depression on the ocean floor formed by the convergence of tectonic plates where one plate is subducting, or diving beneath, another plate (see Figure 19.1C). The denser oceanic plate sinks beneath the less dense continental or oceanic plate, and forms a trench (see Figure 19.1C). Trenches extend for hundreds or even thousands of kilometers. They are often associated with volcanic activity and earthquakes (see Figure 19.1C). The Mariana Trench in the western Pacific Ocean is the deepest known trench on earth, reaching a depth of approximately 10,994 meters at the Challenger Deep.

19.2.3.3 Oceanic Ridges (Mid-Ocean Ridges)

A mid-ocean ridge is a long, underwater mountain range that runs through the center of the ocean basins (Figure 19.1C). It is formed by volcanic activity and tectonic plate movements (at divergent plate boundaries, where two tectonic plates move away from each other) (Figure 19.1C). As the plates more apart, the magma from the mantle rises to fill the gap, creating new crust through a process called seafloor spreading (see Figure 19.1C). Over time, this process results in the formation of a continuous mountain range that extends for thousands of kilometers. The mid-ocean ridge is characterized by a central rift valley, where the plates are moving apart, and flanked on both sides by mountain ranges (see Figure 19.1C). The ridge can reach heights of several kilometers above the surrounding seafloor. The ridge is also marked by numerous volcanic features, such as rift valleys, fissures, volcanic cones, and hydrothermal vents, where hot water and minerals are released into the ocean.

19.3 TRANSPORT AND DEPOSITIONAL PROCESSES TO AND WITHIN DEEP WATER

Most sediments deposited in deeper waters, other than windblown ones, originate on the shelf and make their way across the shelf onto deeper waters. Across-shelf, the sediment movement includes transport of coarser sediments by turbidity currents and the seaward advection of fine sediments by sediment plumes and nepheloid flows (see Figure 19.3).

In summary, below are some major (but not restricted to) processes involved in the transport and deposition of sediments in the deep water. These include turbidity currents, sediment gravity flows, suspension settling, pelagic sedimentation, and hemipelagic settling. These are briefly enumerated below:

The turbidity currents are powerful underwater currents that transport sediments down the continental slope and into the deep sea (Figure 19.2). They occur when sediment-laden water becomes denser than the surrounding water and thus rapidly flows downslope. Turbidity currents carry a wide range of sediment sizes, from fine silt to coarse sand, and can create submarine canyons and deep-sea fans as they deposit sediment on the seafloor.

The sediment gravity flows are similar to turbidity currents. They are the downslope movement of sediments due to gravity and can take various forms, including debris flows, mudflows, and slumps. These flows typically occur in areas with steep slopes or where there are unstable sediments, such as in submarine canyons (Figure 19.2).

Suspension settling in deep waters occurs when fine-grained sediments remain suspended in the water column for extended periods of time due to low energy levels. Thus, fine silt and clay particles are transported over long distances before settling on the seafloor.

Pelagic sedimentation occurs when the deep-water environment receives sediments from the overlying water column (Figs. 19.1B and 19.4). This process involves the deposition of fine-grained

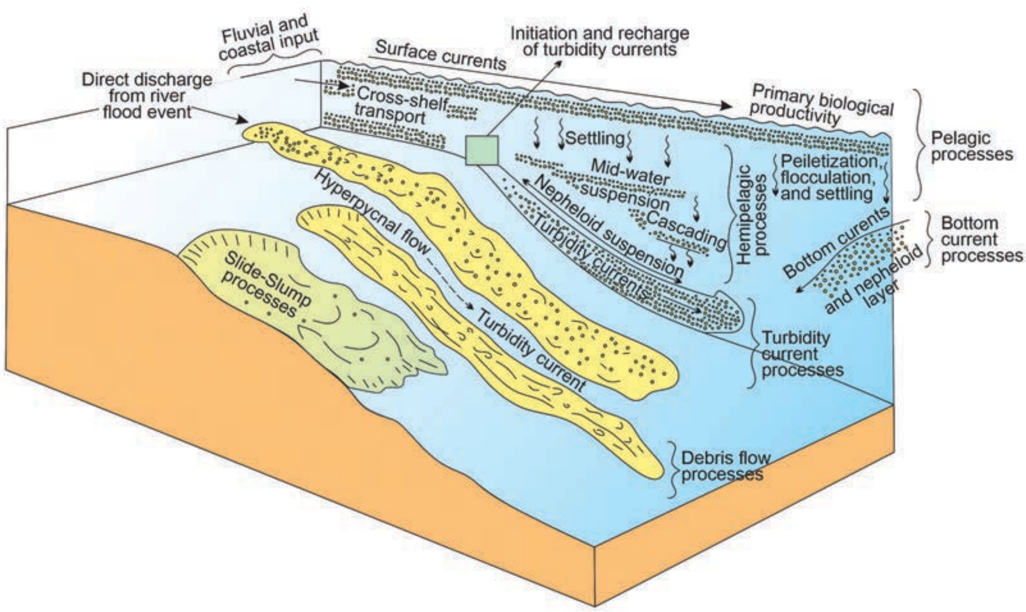

FIGURE 19.3 Transport and depositional processes to and within deep water.

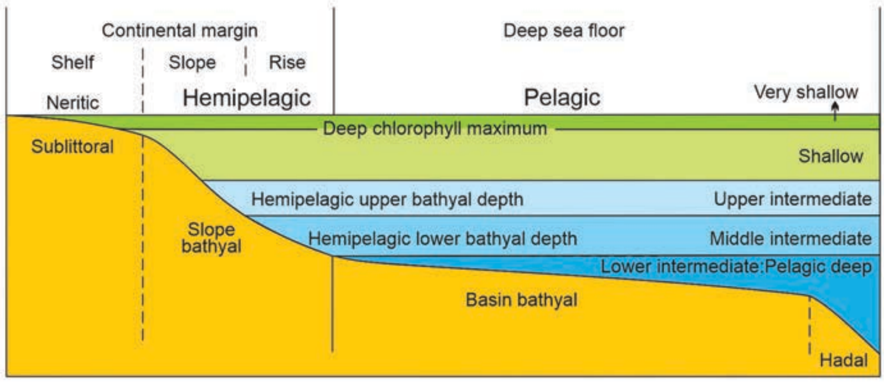

FIGURE 19.4 Components of pelagic and hemipelagic depositional environments. The pelagic sedimentation occurs when the deep-water environment receives sediments from the overlying water column resulting in the deposition of fine-grained particles, such as microscopic planktonic organisms or volcanic ash that settles slowly through the water column. These particles, over time, accumulate to form pelagic sediments, such as oozes, diatomaceous earth, or volcanic ash layers. The hemipelagic settling marks the deposition of sediment that falls slowly through the water column and settles on the seafloor. It occurs in areas where fine-grained sediment is continuously supplied, such as from turbidity currents or suspended sediment. Such sediments are typically well-sorted and accumulate as thin, continuous layers on the deep-sea floor.

particles, such as microscopic planktonic organisms or volcanic ash that settles slowly through the water column. Over time, these particles accumulate to form pelagic sediments, which include oozes, diatomaceous earth, or volcanic ash layers (Figure 19.1B).

Hemipelagic settling marks the deposition of sediment that falls slowly through the water column and settles on the seafloor (Figs. 19.1B and 19.4). It occurs in areas where fine-grained sediment is continuously supplied, such as from turbidity currents or suspended sediment. Such sediments are typically well-sorted and accumulate as thin, continuous layers on the deep-sea floor.

A variety of other minor processes also transport and deposit sediments within the deep ocean, such as wind transport from continents, airfall and submarine settling of pyroclastic particles generated by explosive volcanism within and outside ocean basins, sediment plumes, floating ice, mass-flow processes, various kinds of bottom currents, surface currents, and pelagic settling (see Stow, 1994; Stow and Mayal, 2000). Some of the major ones are enumerated below.

19.3.1 Wind Transport, Sediment Plumes, Ice Rafting, and Nepheloid Transport

Winds, particularly in coastal areas or shallow regions, play a role in the transport of sediments into deep-water environments and enable the transport of fine suspended dust particles seaward, where they settle out over the ocean hundreds of kilometers from the shore. Wind transport is the primary mechanism by which clay-sized siliciclastic sediments are transported to the deep ocean (see Figure 19.1B). Strong winds generate waves and currents that also move sediments along the seafloor or suspend them in the water column. Thus, this wind-driven transport moves sediments from shallow areas to deeper waters, thereby contributing to the deposition of sediments in the deep.

Sediment plumes are turbulent, sediment-laden water masses generated by various processes such as river runoff, turbidity currents, or as re-suspension from the seafloor. In narrow continental shelves, freshwater surface plumes carry fine sediments to considerable distance across the shelf into deeper waters, before mixing and flocculating, thus causing clay particles to settle (Figure 19.3).

These plumes extend vertically or horizontally in the water column, carrying sediment particles of various sizes. Sediment plumes can travel long distances before settling on the seafloor, contributing to the deposition of sediments in deep-water environments (Figure 19.3).

Fine sediments re-suspended by storms on the outer shelf move down the slope in near-bottom nepheloid suspension (see Figure 19.3); these flows can move sediments for hundreds of kilometers and to water depths of 6000 m or more. The turbidity flows moving downslope also inject fine sediment into the water column (see Figure 19.3). The deep bottom currents help in re-suspending sediment into nepheloid layers.

Nepheloid transport refers to the suspension and transport of fine-grained sediment particles in the water column (Figure 19.3). These particles remain suspended due to turbulence or currents, thereby creating a cloudy or turbid appearance in the water known as a nepheloid layer. Nepheloid transport occurs over long distances and contributes to the deposition of fine-grained sediments on the seafloor in deep-water environments.

Ice rafting at high latitudes (Arctic and Antarctic regions) is the movement of ice that transports sediment from the land to the sea. Thus, the melting of the floating ice dumps mixed-sized sediments (glacial-marine sediments) onto the shelf and into the deep ocean floor. Icebergs and sea ice carry sediments entrained within them and deposit them as they melt or break apart over long distances, contributing to the formation of sedimentary layers on the seafloor.

19.3.2 Currents in Canyons

Currents in canyons play a significant role in the transport and deposition of sediment in deep-water environments (Figure 19.2). Canyons are steep-sided valleys or channels that cut through the continental slope and extend into the deep sea (Figure 19.2). They are formed by various processes, including erosion by rivers, turbidity currents, or mass wasting events. Tidal currents measured in submarine canyons are also capable of transporting silt and fine sand even at depths exceeding 4200 m (see Shepard et al., 1979). Two types of currents are noted in submarine valleys: ordinary tidal currents (rarely exceeding velocities of ~50 cm/s; these flow alternately up and down the valley in response to tidal reversals), and occasional surges of strong downcurrent flow with velocities up to 100 cm/s. These currents help to winnow canyon deposits and keep them free from fine sediments.

Some key processes and characteristics of currents in canyons include down-canyon flows, channelized flows, hydraulic jumps, eddy formation, and sediment bypass and deposition. These are briefly enumerated below.

Down-canyon flows are gravity-driven flows, such as turbidity currents or sediment gravity flows that travel down the canyon, carrying sediments from the continental shelf or slope into the deep sea. These down-canyon flows are powerful and can transport large volumes of sediments over large distances.

Channelized flows in canyons often have well-defined channels that guide the flow of water and sediments. These are narrower and deeper than the surrounding seafloor, thus creating a pathway for currents to flow through. Channelized flows enhance the transport of sediment within the canyon, as the confined space increases both the velocity and erosive power of the currents.

Hydraulic jumps occur when the velocity of the current abruptly decreases, causing a change in the flow regime. This results in the deposition of sediments due to the sudden decrease in velocity, causing sediments to settle out of the flowing water. These depositional features contribute to the formation of sedimentary fans or layers within the canyon.

Eddy formation occurs when currents flow through canyons, and they encounter obstacles or irregularities in the canyon floor, forming eddies or vortices. Eddies trap and suspend sediments within the water column, thus, enhancing sediment transport within the canyon. Eddies also promote the mixing of water masses, redistributing nutrients and influencing biological productivity within the canyon ecosystem.

Sediment bypass and deposition occur in canyons. The latter act as pathways for sediment transport, allowing sediment to bypass the continental shelf and slope and reach the deep sea. But canyons also serve as sites of sediment deposition, particularly at canyon heads or within channelized sections. The abrupt decrease in velocity and change in flow regime at these locations causes sediments to settle and accumulate, forming deep-sea fans or aprons.

19.3.3 Contour Currents

Temperature or salinity variations cause density differences in the surface ocean waters resulting in vertical circulation of water masses referred to as thermohaline circulation. Due to density stratification, the bottom currents adjacent to continental margins tend to flow parallel to depth contours or isobaths and thus are called contour currents (Figure 19.5). Deep-water contour currents, also known as deep ocean currents or thermohaline currents, are large-scale oceanic currents that flow horizontally along the contours of the seafloor in the deep ocean. The Coriolis force affects their movement and deflects them (left in the Southern Hemisphere and right in the Northern Hemisphere) into paths parallel to depth contours; hence, they are sometimes also called geostrophic contour currents. As contour currents are best developed in areas of steep topography where the bottom topography extends through the greatest thickness of stratified water column, they are important transporters of sediments on the continental slope and rise.

19.3.4 Pelagic Rain

Pelagic rain, also known as marine snow, is the continuous shower of organic and inorganic particles that sink slowly from the surface waters of the ocean to the deep sea floor through a process called marine snowfall (also referred to as pelagic rain or pelagic settling) (see Figs. 19.1B and 19.3). These particles include dead plankton, fecal matter, detritus, mineral particles, and other organic materials. The transport of pelagic rain particles occurs through a combination of sinking due to gravity and the action of marine organisms that consume and excrete particles, thus enhancing their sinking rate. The calcareous- and siliceous-shelled planktic organisms settle through the ocean water column to the seafloor upon death (Figs. 19.1B and 19.3). As shells settle onto the ocean floor, they are re-transported by turbidity currents or contour currents and form extensive biogenic deposits

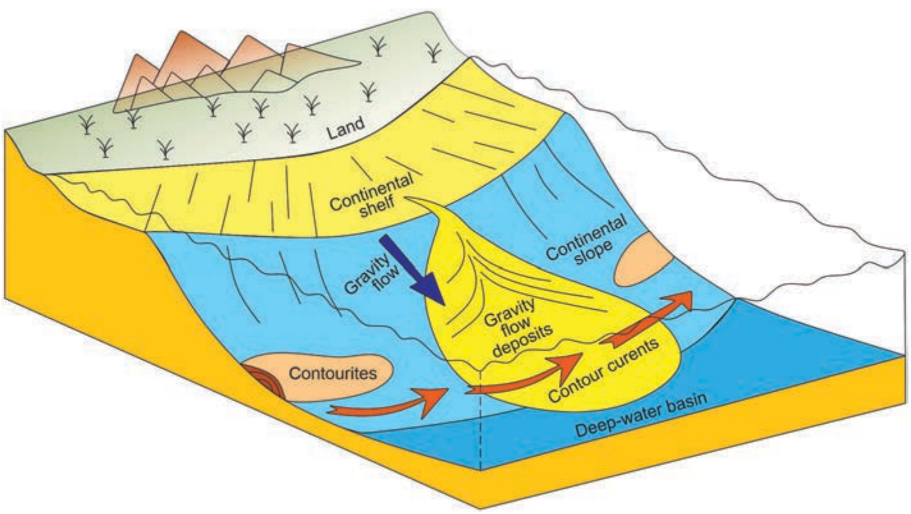

FIGURE 19.5 Contour currents and contourites.

or oozes (Figure 19.1B). These oozes are important contributors to ancient deep ocean sediments, particularly in Jurassic and younger rocks, forming thick deposits of chalk, radiolarian chert, and diatomite. Fine-sized siliciclastic particles (such as clay minerals, quartz, feldspars) are transported from land by surface plumes, wind, or by floating ice and settle along with organic remains to form mixed biogenic and siliciclastic sediments referred to as hemipelagic deposits via hemipelagic processes (Figs. 19.3–19.4). The pelagic rain is a crucial source of food and energy for deep-sea organisms. As the particles sink, they provide a continuous supply of nutrients to the deep-sea ecosystem. Bacteria and other microorganisms decompose the organic matter, releasing nutrients that support the growth of deep-sea organisms. Pelagic rain also plays a significant role in the carbon cycle and the sequestration of carbon dioxide (CO_2) from the atmosphere. As organic particles sink, they carry carbon with them to the deep sea, effectively removing it from surface waters. This process helps regulate the concentration of CO_2 in the atmosphere and influences the overall carbon budget of the planet.

19.3.5 Explosive Volcanism

Deep-water explosive volcanism refers to volcanic eruptions that occur underwater, typically at depths greater than 200 meters. These eruptions happen along mid-ocean ridges, subduction zones, or volcanic seamounts and result in the formation of volcanic structures such as seamounts, volcanic cones, and calderas (see Figs. 19.1A and C). These structures are created when the magma rises to the surface and erupts, releasing gases and forming explosive eruptions. Additionally, explosive volcanism also results in the formation of volcanic ash, lapilli, and bombs that are ejected both subaerially and subaqueously. Coarse material ejected subaerially tends to be deposited by air fall close to the eruption column on all sides of the vent. This explosive activity generates plumes, pyroclastic flows, and even tsunami waves. Deep-water explosive volcanism also creates hydrothermal systems formed when water interacts with hot volcanic rocks, creating high-temperature fluid vents known as hydrothermal vents. These vents release mineral-rich fluids that support unique ecosystems and are important for the cycling of elements in the deep ocean. Deep-water explosive volcanism also plays a role in the creation of new seafloor along mid-ocean ridges, the recycling of oceanic crust at subduction zones (Figure 19.1C), and the growth of volcanic seamounts (Figure 19.1A). Volcanic eruptions also release large amounts of volcanic ash, gases, and chemicals into the water column, affecting water chemistry and potentially harming marine life. However, these very eruptions also create new habitats and provide nutrients that support the growth of deep-sea organisms.

19.3.6 Turbidity Currents and Other Mass-Transport Processes

High-velocity turbidity currents generated on the shelf or on the upper continental slope are the most important mechanism for transporting sands and gravels to deeper waters through submarine channels (see Norrnark and Piper, 1991) (see Figure 19.3). On passive margins, in particular, sediments spread out from the mouths of the canyons onto the deep seafloor to form deep-sea submarine fans (see Figure 19.2). Under some conditions, downslope, the submarine debris flows are transformed into turbidity currents (see Piper et al., 1999). Additionally, other mass-transport processes, such as creep, gliding (sliding), and slumping, are also responsible for large-scale re-transport of sediments on the steepened continental slopes.

19.4 MAJOR KINDS OF MODERN DEEP-SEA SEDIMENTS

Four broad classes of deep-sea sediment, terrigenous and pelagic (biogenic), volcanic, and authigenic sediments, are noted. These are briefly enumerated below.

19.4.1 Terrigenous Sediments

The terrigenous sediments are derived from the erosion and weathering of land-based sources (transported by rivers, wind, or ice to the deep sea) and are composed primarily of mineral particles, such as clay, silt, sand, and gravel. These sediments are most abundant near continental margins and include a wide range of particle sizes depending on the proximity to the source and transportation processes. Terrigenous deposits include gravel, sand, and mud derived from land and transported within the more proximate parts of the deep ocean by a variety of processes (such as turbidity currents, contour currents, ice rafting, etc.). The deep-sea siliciclastic sediments include hemipelagic muds (abyssal clays), turbidites and other sediment gravity-flow deposits, contourites, glacial-marine sediments, and slumps and slides, ranging in size from clay to gravel. Some deposits exhibit well-developed bedding while others have a disorganized fabric with poorly developed bedding. The terrigenous sediments are classified into different types based on their particle size and origin: hemipelagic muds, abyssal clay, turbidites, contourites and ice-rafted debris (glacial-marine sediments). These are briefly described below.

19.4.1.1 Hemipelagic Muds

Hemipelagic muds are fine-grained sediments composed of clay to silt, sandy clay with some sand, and organic matter mixed in; they are a major type of sediments deposited on continental slopes. Typically, these muddy deposits contain >5% biogenic remains and >40% silt (terrigenous component) (see Stow and Piper, 1984) deposited under very low current velocities. Hemipelagic muds form through the settling of fine particles from the water column, derived from various sources such as terrigenous material transported by rivers, windblown dust, volcanic ash, and biogenic particles (the remains of planktic organisms) (see Figure 19.3). The fine terrigenous component incudes quartz, feldspar, micas, and clay minerals and/or volcanogenic sediments such as ash, fine pumice, and palagonite. Granules and pebbles of pumice may occur as isolated fragments, pockets, or distinct layers associated with hemipelagic muds. Hemipelagic muds may also contain the remains of siliceous organisms, particularly diatoms, and calcareous organisms, such as foraminifers and nannofossils, as well as fine lime muds swept off carbonate platforms into deeper waters. Hemipelagic muds range in color from gray to green and, more rarely, to reddish brown. They are typically deposited in relatively low-energy environments, such as the deep ocean floor and continental slopes (see Figure 19.3). The deposition of hemipelagic muds is influenced by factors such as ocean currents, bottom water oxygen levels, and the availability of sediment from nearby sources. Hemipelagic muds are poorly laminated to massive, and are generally moderately to highly bioturbated.

19.4.1.2 Turbidites

Turbidites are coarser-grained terrigenous sediments that result from turbidity currents (see Figure 19.6). Turbidity currents are powerful underwater flows that transport sediments downslope from continental slopes to deep-sea basins (Figure 19.6A). These currents are triggered by various factors, such as earthquakes, landslides, or the oversaturation of sediment within the water column. Turbidity currents are characterized by their high sediment concentration, which gives them a dense, muddy appearance (Figure 19.6B). As they flow downslope, turbidity currents lose energy and deposit their sediment load on the seafloor. This results in the formation of distinct layers of sediment known as turbidites (Figure 19.6C). Thus, turbidites are typically composed of a mixture of sand, silt, and clay-sized particles, with the grain size decreasing as the current loses energy (Figs. 19.6B–C). The sand layers within turbidites are often well-sorted and can show characteristics of sediment gravity flows, such as cross-bedding or graded bedding (Figs. 19.6B–C). Sole markings, such as flute casts, groove casts, and load casts, are common on the base of many turbidite sequences. The deposition of turbidites can occur in various deep-sea environments, including submarine canyons, continental

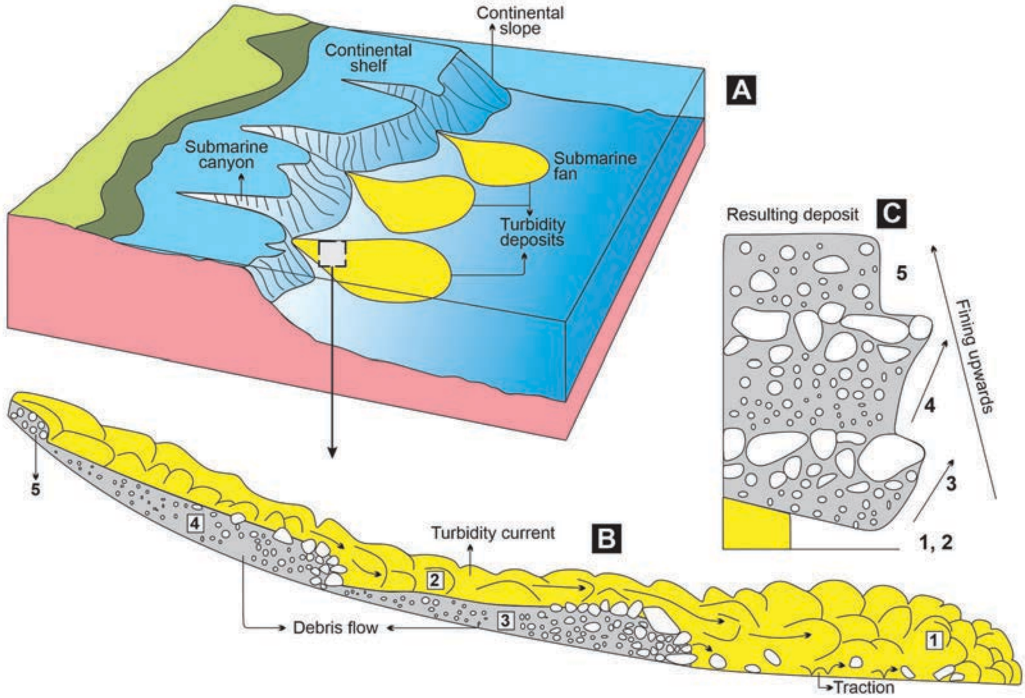

FIGURE 19.6 Turbidites. A: Geological setting of a turbidite deposit. B: Characteristic of a turbidite current. (Modified from Sohn, 2000.) C: Graded bedding in a turbidite deposit.

slopes, and deep-sea fans (Figure 19.6A). Turbidites are widely distributed in the modern ocean on passive margins and in both back-arc and the fore-arc regions of active margins.

19.4.1.3 Contourites

Contourites are a type of sediment that accumulate along the contours of the seafloor due to the action of bottom currents (Figure 19.5). These currents can be driven by various factors, such as tides, winds, or density differences in the water column. Contourites are typically found in areas with strong and persistent bottom currents, such as submarine canyons, channel systems, or along continental slopes (Figure 19.5). Contourites are composed of a mixture of clay, silt, and sand, with the grain size varying depending on the strength of the bottom currents. In areas with weaker currents, finer-grained sediments dominate, while coarser-grained sediments accumulate in areas with stronger currents. Contourites also show negative grading (coarsening upward) from muddy through silty to sandy contourites, followed up by positive grading (fining upward) back through silty to muddy contourite facies (see Stow et al., 1998). Muddy contourites are more homogeneous, poorly bedded, and highly bioturbated, whereas the silty ones have thin, irregular layers, commonly displaying a mottled appearance; they are also bioturbated. Contourites display both sharp and gradational contacts. The sediment composition of contourites also varies depending on the proximity to terrestrial sources and the presence of biogenic material. The deposition of contourites occurs through a combination of processes, including erosion, transport, and deposition. Bottom currents erode sediments from the seafloor and transport them along the contours of the seafloor (Figure 19.5). As the currents lose energy, they deposit the sediments, forming distinct layers or mounds on the seafloor. Thus, the characteristics of contourites vary depending on the specific environment in which they are deposited. For example, in submarine canyons, contourites may be characterized by erosional features, such as scours or channels, while on continental slopes, they may form elongated mounds or drifts. In modern oceans, contourites occur in numerous places, such as the continental rise of eastern North America, the continental margin off northwest Britain, and the southern Brazil margin.

19.4.1.4 Ice-Rafted Debris (Glacial-Marine Sediments)

Glacial-marine sediments are deposited in deep-sea environments as a result of glacial processes and their interaction with the marine environment (Figure 19.7). These sediments are typically found in areas where glaciers or ice sheets have extended into the ocean, such as fjords, continental margins, and deep-sea basins. Thus, they are derived from various sources, including glacial erosion, iceberg calving, and meltwater runoff and are composed of a mixture of rock fragments, sand, silt, clay, and organic material. The composition and grain size of these sediments vary depending on the proximity to the glacier, the type of rock being eroded, and the transport and depositional processes involved. Hence, they are poorly sorted gravelly sands or gravelly muds with crude to well-developed stratification containing angular, faceted, and striated pebbles as the coarse fraction. The deposition of glacial-marine sediments is influenced by a combination of processes, including ice rafting, turbidity currents, and suspension settling. Icebergs and ice sheets can transport and deposit large amounts of sediment as they melt and interact with the marine environment. Turbidity currents, triggered by glacial meltwater or sediment-laden plumes, also transport and deposit sediment downslope. Glacial-marine sediments can exhibit distinct characteristics that reflect their glacial origin, such as dropstones (large rocks dropped by icebergs) (Figure 19.7), rhythmic layering (resulting from seasonal variations in sediment input), and glacially striated grains (indicating glacial erosion). These sediments also contain evidence of past glaciations, such as fossilized marine organisms and isotopic signatures that provide insights into past climate conditions. The modern ocean floor at high latitudes is covered by these glacial-marine sediments, as in the subpolar North Atlantic, and the circum-Antarctic, among others.

19.4.2 Slump and Slide Deposits

Slump and slide are previously deposited pelagic or terrigenous deposits that form due to mass wasting processes within the submarine environment (Figure 19.8). Mass wasting is the downslope movement of sediment or rock under the influence of gravity. In deep-water settings, slumps and slides occur on the seafloor and can result in the deposition of characteristic sedimentary features (Figure 19.8). Slump is characterized by the rotational movement of sediment or rock along a curved

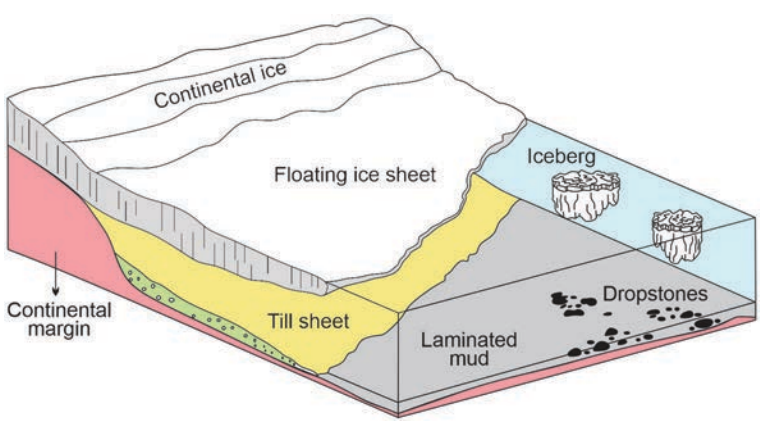

FIGURE 19.7 Dropstones.

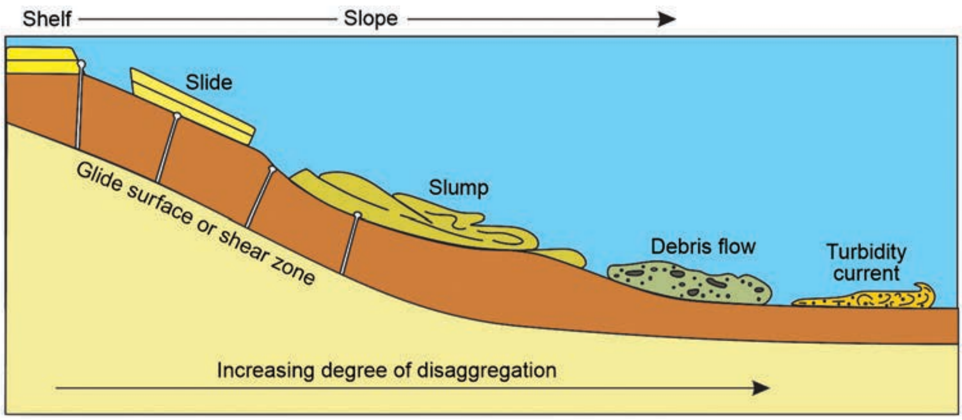

FIGURE 19.8 Depositional setting of slide-and-slump deposits. (Modified from Shanmugum et al., 1995.)

failure surface. It typically occurs on gentle to moderate slopes and results in the formation of slump blocks or lobes (Figure 19.8). During the transport process, the consistency of the slump mass is disturbed, resulting in faulted, contorted, and chaotic bedding and internal structure (Figure 19.8). Slump deposits are characterized by chaotic, jumbled sediment layers with disrupted bedding and folding.

Slide deposits, on the other hand, represent a more catastrophic and rapid type of mass-wasting event and involve the detachment and downslope movement of a large mass of sediment or rock along a failure plane (Figure 19.8). Slide deposits are characterized by a relatively undisturbed bedding orientation but may exhibit folding or faulting due to the intense forces involved. Slide deposits form distinct lobes or tongues of sediment on the seafloor. Both slump and slide deposits display a range of sediment sizes, including sand, silt, clay, and even larger boulders or blocks. The composition and grain size of these deposits depend on the source material and the processes involved in the mass-wasting event. These deposits are common on continental slopes with high rates of deposition, such as off the Mississippi and Rhone deltas, and on slopes with glacial-marine deposits (see Nardin et al., 1979; Schwab et al., 1993).

19.4.3 Pelagic Sediments

Pelagic sediments accumulate in the open ocean, far away from the influence of land-derived sediment sources (see Figs. 19.1B and 19.4). These sediments are composed primarily of fine-grained particles, such as clay and silt, and are typically deposited slowly over long periods of time. They are derived from a variety of sources, including biogenic material, volcanic ash, and windblown dust. In general, pelagic sediments are typically fine-grained and well-sorted, with particles that are often smaller than 0.06 mm in diameter. They can exhibit distinct layering or banding, known as varves, that reflect seasonal variations in sedimentation. These sediments also contain microfossils, such as foraminifera, radiolarian or diatoms, which provide valuable information about past oceanic conditions and climate.

19.4.3.1 Pelagic Clays

The pelagic clays are derived from land but are deposited by slow settling in the more distal parts of the ocean. They are commonly red to red-brown in color, and are siliciclastic muds that contain clay minerals, zeolites, iron oxides, and windblown dust or ash, covering vast areas of the deep ocean below ~4500 m.

19.4.3.2 Ooze

Biogenic materials, such as the remains of marine organisms like plankton and shells, are a major contributor (see Figure 19.1B). These organisms produce calcium carbonate shells or silica-based skeletons which settle to the ocean floor upon death and accumulate as sediments forming calcareous (composed mainly of $CaCO_3$ tests such as foraminifers and nannofossils) and siliceous (composed mainly of siliceous tests largely of diatoms and radiolarians but may include silicoflagellates and sponge spicules) oozes, respectively (see Figure 19.1B). These are briefly enumerated below.

19.4.3.2.1 Calcareous Ooze

These are widespread in modern deep oceans, particularly in the Atlantic Ocean, at depths shallower than ~4500 m, i.e., above the calcium carbonate compensation depth (see Figure 19.1B). Calcareous oozes are typically found in areas with high biological productivity and relatively low rates of sedimentation. They are most common in warm, shallow waters such as the tropical and subtropical regions. The calcium carbonate shells of the organisms are preserved in sediments due to the low solubility of calcium carbonate in seawater. Calcareous oozes play an important role in the global carbon cycle as they sequester CO_2 from the atmosphere.

19.4.3.2.2 Siliceous Ooze

Siliceous ooze is typically found in areas of high biological productivity and relatively low rates of sedimentation, similar to calcareous ooze (Figure 19.1B). It is most common in colder waters, such as the subpolar and polar regions of the ocean, and in upwelling zones where nutrient-rich waters rise to the surface. The silica shells of diatoms and radiolarians are preserved in the sediment due to the low solubility of silica in seawater. Siliceous oozes are often found in regions with higher levels of dissolved silica in the water, which allows for the abundant growth of these organisms. Like calcareous oozes, siliceous oozes also play a role in the global carbon cycle by sequestering CO_2 from the atmosphere. The burial of silica shells in the sediment removes carbon from the surface waters and stores it in the deep ocean, helping to regulate the earth's climate.

The lithified equivalent of a calcareous ooze is called chalk and those of siliceous oozes are transformed during burial into bedded cherts. Volcanic ash and dust from terrestrial sources also contribute to pelagic sediments. The volcanic eruptions release ash and other particles into the atmosphere that are then transported by wind and eventually settle in the ocean. Similarly, windblown dust from deserts and other arid regions is also transported over long distances and deposited in the open ocean.

19.4.4 Chemical Sediments

Chemical sediments form in the deep ocean through chemical processes rather than by the accumulation of biological remains. They are typically composed of minerals that precipitate out of seawater (such as manganese nodules; see Figure 19.9) or are formed through chemical reactions within the water column. The manganese nodules are rounded lumps or concretions that contain high concentrations of manganese and other elements such as iron, nickel, and copper (see Figure 19.9). They form on the seafloor over long periods of time as metal-rich minerals precipitate out of seawater and accumulate around a nucleus, such as a shell fragment or rock. They are typically found in abyssal plains and deep-sea basins (see Figure 19.9).

Phosphorites are also formed through precipitation. They are rich in phosphorus and form from the accumulation of organic matter and phosphate minerals, and hydrothermal deposits, which form around underwater hydrothermal vents through the precipitation of minerals from superheated water (for more details see Chapter 7).

Evaporites are sedimentary rocks that form when saline water evaporates, leaving behind minerals that were dissolved in the water (for more details see Chapter 7). Common evaporite minerals

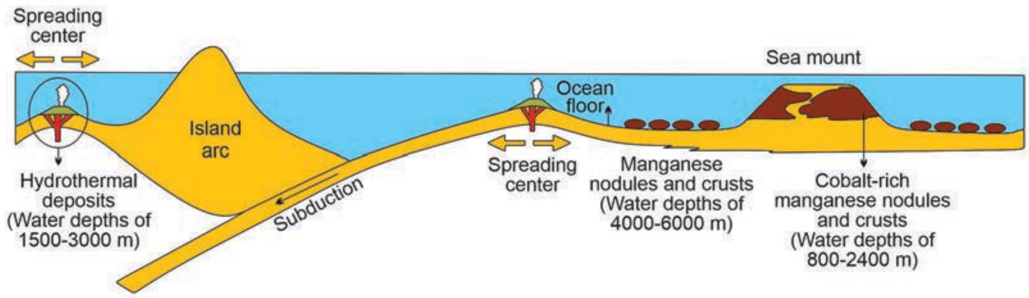

FIGURE 19.9 Depositional setting of manganese nodules.

include gypsum, halite (rock salt), and anhydrite. These sediments are typically found in areas with restricted circulation, such as enclosed basins or lagoons, where evaporation rates are high (for more details see Chapter 7).

These deep-water chemical sediments help to reconstruct the history of seawater chemistry, the movement of tectonic plates, and the formation of mineral resources. Additionally, these sediments also serve as a record of past climate changes and provide insights into the evolution of marine ecosystems.

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Section IVd

Depositional Systems: Carbonate Shelf Environments

20 Carbonate Depositional Environment

20.1 INTRODUCTION

Carbonates make up a quarter of sedimentary rocks within the geologic record. They provide information about the earth's history and its environments and are hosts for water, petroleum, and some metallic elements. Additionally, as carbonates are associated with deposition in relatively shallow seas, they are highly fossiliferous and preserve remains and host almost 25% of marine life (Knowlton et al., 2010). They are also a major carbon sink as they account for nearly 25% of the global CO_2 sink into the marine sediments (Jones et al., 2015). Thus, the knowledge of the marine carbonate system is critical for the study of the global carbon cycle (Falkowski, et al., 2000) and more so for the understanding of past and future climate changes. The carbonate system is also very important for understanding marine organisms and changes in sea level in the geologic past. The chemical signatures of carbonate systems have been used to answer tectonic questions from the earth's past (MacDonald et al., 2009). Additionally, our understanding of carbonate textures and the basic processes of carbonate deposition comes only from studying modern carbonate environments. In the geologic past, broad epeiric seas, hundreds to thousands of kilometers wide, were sites of extensive carbonate deposition.

Modern carbonates are deposited on a few continental shelves that are located mainly at low latitudes in clear, shallow, tropical to subtropical seas with little terrigenous (siliciclastic) input (Figure 20.1) (Tucker et al., 1990; Whalen, 1995; James and Jones, 2015a). The carbonate deposits are the most dominant sediment cover, even though most modern shelves are covered by siliciclastic sediments. Most of these tropical carbonate-producing shelves, such as Florida Bay and Western Australia, are attached to the mainland (Figure 20.1) (Vacher and Quinn, 1997). Besides these, carbonate sediments also form at some higher latitudes (30 to 60° N and S; cool-water shelves); these predominantly consist of shell remains (Nelson, 1988; James and Clarke, 1997). A few carbonates form in non-marine environments, such as in lakes, streams, caves, soils, and dune settings. Although these carbonates are proxies of the prevailing environment, their volume in the ancient past is negligible; hence, they are not considered further.

20.2 CARBONATE SHELF (NON-REEF) ENVIRONMENTS

20.2.1 Depositional Setting

The marine carbonate sediments are deposited primarily on shallow shelf (Figures 20.2A–B) and broad epeiric platforms (Figure 20.2C) covered by shallow waters (see also Simo et al., 1993; Tucker et al., 1990; Markello et al., 2008; Wilmsen et al., 2018; Michel et al., 2019; Laugié et al., 2019). The carbonate platforms occur on the margins of cratonic blocks, in intracratonic basins, across the tops

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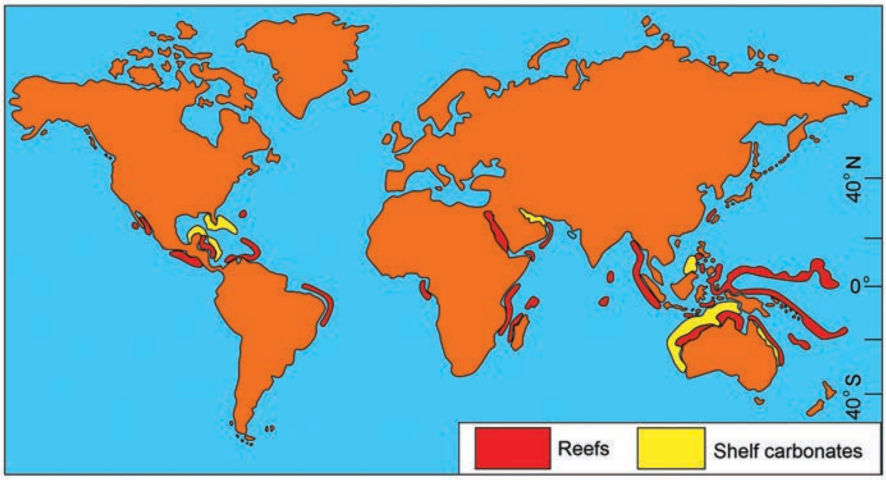

FIGURE 20.1 Global distribution of shelf carbonates and reefs. The distribution map of modern-day carbonate environments shows that their deposition is largely 30° north and south of the Equator, an area of warm temperatures. (Modified after Bosence and Wilson, 2003.)

of major offshore banks, and on localized positive features on wide shelves (Figure 20.2C) (Wilson and Jordan, 1983). Carbonate environments are also present in some parts of marginal marine environments, such as beaches, lagoons, and tidal flats. Although platform carbonates (Figure 20.2C) are a major component of the earth's system, their spatial distribution through geological time is difficult to interpret due to the (a) incompleteness of geological records, (b) sampling heterogeneity, and (c) their intrinsic complexity; the latter is due to variations in climate, currents, organic matter cycle, topography, eustasy, and tectonics (Tucker et al., 1990; Markello et al., 2008; James and Jones, 2015b). However, in modern oceans, the basic types of carbonate platforms or shelves are (Figure 20.3): (a) rimmed platforms, (b) unrimmed (open shelf) platforms, (c) ramps, (d) isolated platforms, and (e) epeiric platforms (see Harris, Moore, and Wilson, 1985; James and Kendall, 1992; Read, 1982, 1985). In the case of ancient environments, they are epeiric platformss (see also (Opdyke and Wilkinson, 1993; Pohl et al., 2019).

20.2.1.1 Rimmed Carbonate Shelves

The rimmed carbonate shelves are shallow platforms that are marked at their outer edges (margins) by a pronounced break in slope into deeper waters (Figure 20.3A). They have a nearly continuous rim or barrier along the platform edge that consists either of a reef buildup or a skeletal/ooid sand shoal that absorbs wave energy and restricts water circulation, thus, creating a low-energy shelf environment, called a lagoon, positioned landward of the shelf-edge barrier (Figure 20.3A). The lagoon commonly grades landward into a low-energy tidal-flat environment rather than a high-energy beach zone. These rims can be formed by a variety of processes, including tectonic activity, sea-level changes, or the growth of reef-building organisms.

One common type of rimmed carbonate shelf is the barrier reef (detailed later in the chapter). These are long, narrow formations that parallel the coastline and are separated from the shore by a lagoon. They are typically formed by the growth and accumulation of coral reefs along the outer edge of a shallow shelf. The coral reefs provide a barrier that protects the lagoon from the open ocean, creating a unique and diverse marine environment (shelf-edge reef). Another type of rimmed carbonate shelf is an atoll (detailed later in the chapter). These are circular or oval-shaped formations

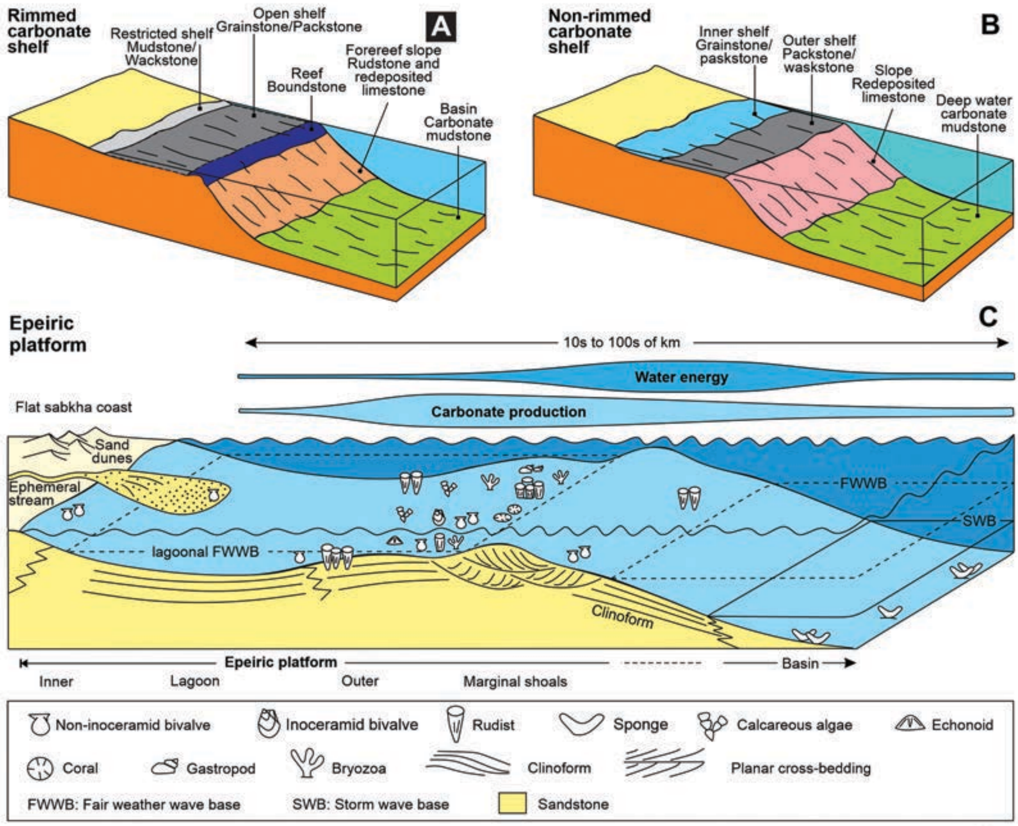

FIGURE 20.2 Carbonate depositional settings. A: Rimmed carbonate shelf. B: Non-rimmed carbonate shelf. C: Epeiric carbonate platform, showing the distribution of major facies associations and fauna. (A–B: Modified after Nichols, 2009; D: Modified after Wilmsen et al., 2018.)

that consist of a central lagoon surrounded by a ring of coral reefs, typically found in areas where volcanic islands or seamounts have subsided or eroded over time, leaving behind a circular or semi-circular reef structure. Rimmed carbonate shelves are also formed by tectonic activity. For example, in areas where the continental shelf is uplifted or tilted, a rimmed shelf forms along the edge of the uplifted area, creating a distinct boundary between the shallow shelf and the deeper oceanic waters. Thus, the presence of a rim or barrier in rimmed carbonate shelves influences sedimentation patterns and the ecological dynamics of the environment. The barrier creates a protected lagoon or inner shelf environment, which has a different sediment type, water depth, and ecological communities as compared to the outer shelf or the open ocean.

Modern examples of rimmed shelves include South Florida Bay, the Bahama Platform (an isolated platform), the Belize Shelf in the western Caribbean off Guatemala, and the Great Barrier Reef area of Australia (see also Wilson and Jordan, 1983; Sellwood, 1986; Jones and Desrochers, 1992; Wright and Burchette; 1996). The Permian (Guadalupian) carbonate deposits of the Delaware Basin, western Texas, and southern New Mexico are examples of ancient rimmed shelf deposits (Ward, Kendall, and Harris, 1986; Saller et al., 1999). The rimmed carbonate shelves also serve as reservoirs for hydrocarbons, making them economically useful, besides being valuable proxies of past environmental conditions and the evolution of marine life based on their geological features.

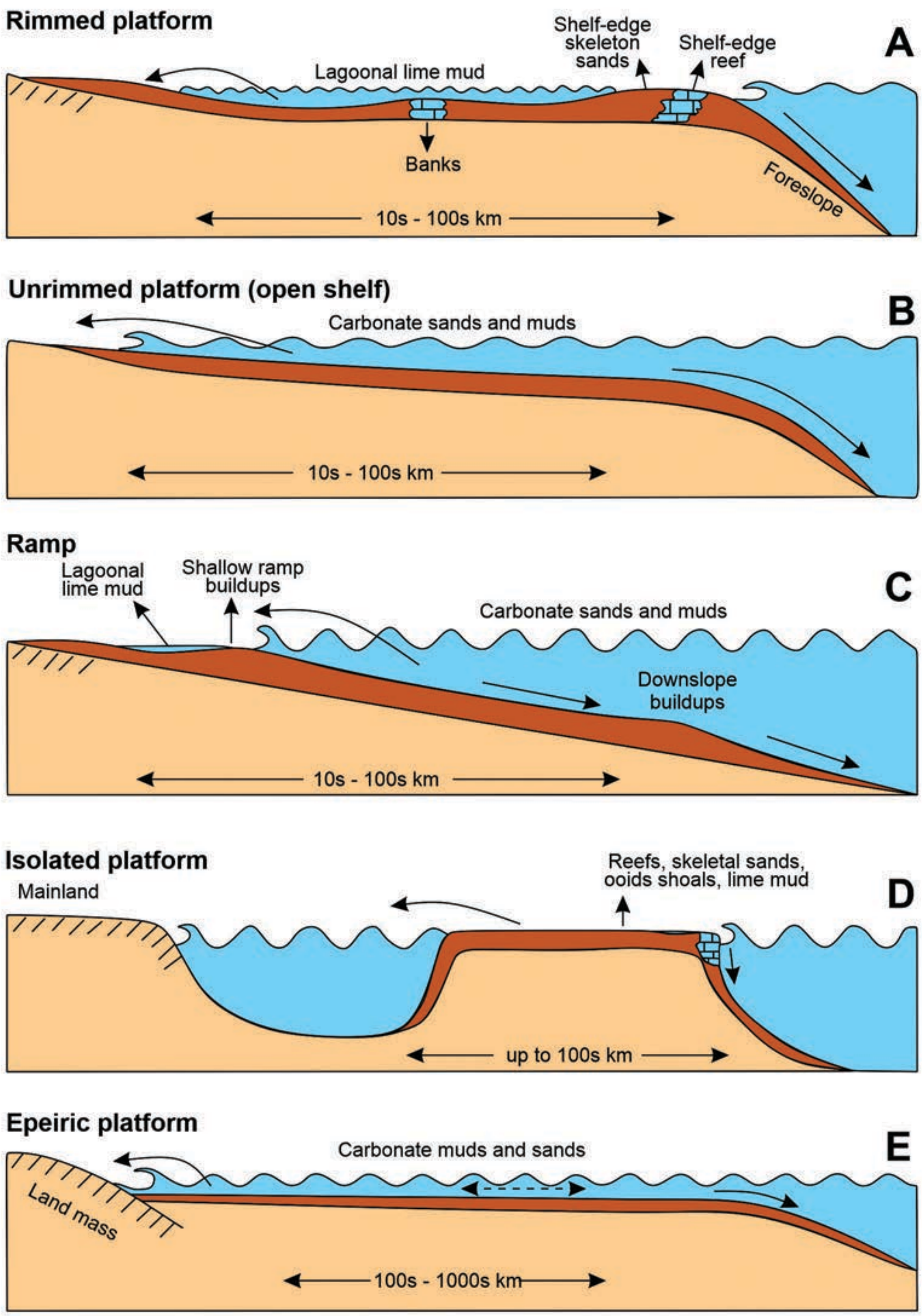

FIGURE 20.3 Types of carbonate platforms or shelves. (Modified after Harris, Moore, and Wilson, 1985; James and Kendall, 1992; Read, 1982, 1985.) A: Rimmed platform. B. Unrimmed (open shelf) platform. C: Ramp. D: Isolated platform. E: Epeiric platforms.

20.2.1.2 Unrimmed Platforms

An unrimmed platform is a broad, flat or gently sloping area on the continental shelf that lacks a distinct rim or barrier (Figure 20.3B). The platform is characterized by the deposition of carbonate rocks, such as limestone or dolomite, formed by the skeletal remains of marine organisms and the precipitation of calcium carbonate from the seawater. The unrimmed platforms are noted in a variety of settings, including tropical, subtropical, and temperate regions, typically in areas with warm, clear, and shallow marine waters, as these conditions favor the growth and accumulation of carbonate-producing organisms. In the present day, the unrimmed platform occurs on the leeward side of large tropical banks and in all cool-water carbonate settings (see James and Kendall, 1992). Most cool-water carbonates accumulate on unrimmed (open) shelves.

The water circulation across an unrimmed platform allows for the formation of a moderately high-energy beach zone that develops alongshore and skeletal or ooid-pellet sand shoals that form along shelf edges. The sedimentation processes on unrimmed platforms are influenced by a combination of biological, physical, and chemical factors. Organisms such as algae, corals, mollusks, and foraminifera play a significant role in the production of carbonate sediments through their skeletal remains. These sediments accumulate as loose sediment or become lithified into solid carbonate rocks over time. Thus, unrimmed carbonate platforms are affected by much the same physical processes as siliciclastic shelves. Physical processes, such as wave action, currents, and tides, also contribute to sedimentation on unrimmed platforms by transporting and redistributing carbonate sediments, creating various depositional features, including sand waves, ripples, and sediment drifts. Chemical processes, particularly the precipitation of calcium carbonate from seawater, also contribute to the formation of carbonate sediments on unrimmed platforms. This process occurs when the concentration of dissolved calcium and carbonate ions in seawater exceeds their solubility, leading to the formation of solid calcium carbonate crystals. Modern examples of tropical unrimmed shelves or carbonate ramps include the eastern Gulf of Mexico off the Florida coast; the Yucatan Shelf, Mexico, in the southern part of the Gulf of Mexico; and the Trucial Coast of the Persian Gulf. Unrimmed platforms also have economic significance, as the deposited carbonate rocks serve as reservoirs for oil, gas, and other valuable resources. Their study also provides valuable insights into past environmental conditions, as well as the evolution and dynamics of carbonate-producing organisms.

20.2.1.3 Ramps

A ramp is a gently sloping (<1°; i.e., a very low depositional slope gradient) unrimmed platform on which shallow-water deposits pass downslope with only a slight break in slope into deeper-water facies (Figure 20.3C). The break in slope is not marked by a pronounced reef, but by the presence of discontinuous sand shoals along the shelf edge where water energy is high (Figure 20.3C). Thus, carbonate ramps commonly form a shallow-water shoreline or lagoon to a basin-floor continuum – i.e., they typically extend from the shoreline to deeper marine environments, and gradually transition from shallow to deeper waters (Figure 20.3C).

A carbonate ramp is composed mainly of carbonate sediments, such as limestone or dolostone, and is often associated with warm, tropical environments, common in areas such as the present-day Bahamas, the Persian Gulf, and the Great Barrier Reef (Australia). A large proportion of carbonate successions in the geological record were also deposited in ramp-like settings. The sedimentation on carbonate ramps is primarily driven by the accumulation of skeletal debris from marine organisms like corals, algae, mollusks, and foraminifera. These organisms produce calcium carbonate skeletons that accumulate over time to form limestone or dolostone deposits. The gradual slope of carbonate ramps allows for the continuous deposition and preservation of these carbonate sediments. Carbonate ramps are important reservoirs for hydrocarbons.

It must also be mentioned here that there is a lot of confusion in the usage of terms, carbonate "shelf," "ramp," and "platform." Following Wright and Burchette (1998), carbonate shelves are shallow, flat-topped structures with a clearly defined margin determined by a steep slope down to the adjacent basin such as the modern east Florida shelf. A ramp is a low-gradient submarine slope, particularly on continental shelves, where the dominant sea-floor sediments are of carbonate mineralogy (see Figure 20.3C). Carbonate "platform" is applied to any thick, more or less flat-topped carbonate depositional system and distinguishes such features from the much more general and widely applied concept of a "shelf." It must be noted that many ancient ramp systems appear to have developed into flat-topped "rimmed" carbonate platforms. A prominent school of thought holds that ramps might represent the incipient, catch-up stages of a rimmed platform (where carbonate supply had not filled the available accommodation space) while carbonate shelves represent true keep-up systems (where the carbonate pile has built up to sea level and keeps pace with subsequent changes in sea level).

20.2.1.4 Isolated Platforms

Isolated platforms (also known as Bahama-type platforms) are shallow-water carbonate shelf environments that are physically separated from adjacent carbonate shelf systems (Figure 20.3D). Unlike rimmed carbonate shelves, which are bounded by barriers or rims, isolated platforms are not connected to other carbonate shelf systems. They are tens to hundreds of kilometers wide, commonly located offshore of shallow continental shelves, and surrounded by deep water that may range from several hundreds of meters to a few kilometers deep (Figure 20.3D). The platforms may have gently sloping, ramp-like margins or a more steeply sloping margin as in rimmed shelves (Figure 20.3D). Isolated platforms are generally free of clastic sediments.

The isolated platforms are characterized by the accumulation of carbonate sediments, including limestone and dolomite, formed from the skeletal remains of marine organisms and the precipitation of calcium carbonate from seawater. The isolation of these platforms has resulted in a unique pattern of sedimentation, distinct carbonate facies, depositional environments and ecological dynamics. Without the influence of nearby carbonate shelf systems, the isolated platforms have different water depths, sediment types, and ecological communities as compared to the surrounding areas. The sedimentation processes are influenced by a combination of biological, physical, and chemical factors, similar to other carbonate shelf environments.

Isolated platforms occur in various geological settings such as the tectonically stable regions or areas where subsidence has created isolated basins. They can also form on top of volcanic islands or seamounts that have subsided or eroded over time, leaving behind a platform of carbonate sediments. Although the Bahamas are probably the best-studied example of a modern isolated carbonate platform, numerous other "carbonate islands" are noted in the modern ocean, such as Bermuda, Barbados, and the Cook Islands, where carbonate sediments are presently accumulating or were deposited during the Pleistocene time (see Vacher and Quinn, 1997). An example of an ancient isolated platform is the early–middle Triassic carbonates of the Dolomite Alps of northern Italy similar to the modern-day Bahama Banks (Bosellini, 1984). The Dolomite Alps were deposited under moderately low-energy, open to restricted, shallow, and subtidal conditions (Bosellini and Rossi, 1974).

The study of isolated platforms provides an insight into past environmental conditions, evolution and dynamics of carbonate-producing organisms in isolated settings, besides being economically useful, as the carbonate rocks serve as reservoirs for oil, gas, and other valuable resources.

20.2.1.5 Epeiric Platforms

Epeiric platforms, also known as epicontinental shelves or epicontinental seas, are relatively flat, shallow marine basins that are partially or completely enclosed within a continent (Figure 20.3E). Hence, they are influenced by both marine and terrestrial processes, resulting in unique sedimentary environments and diverse ecosystems. The sediments deposited in epeiric platforms are often fine-grained, such as mud or silt, due to the limited wave energy and strong influence of river input.

Epeiric platforms are formed when sea levels rise and flood low-lying areas of a continent, creating a connection between the ocean and the continental interior (Figure 20.3E). Epeiric platforms are characterized by their extensive coverage, typically spanning large areas of continents, and by their relatively shallow depths (usually <200 m). These platforms form when portions of continents become submerged due to changes in sea level or tectonic activity. They are often associated with periods of high sea level, such as during times of global warming or when ice caps melt, causing the sea level to rise. As a result, seawater floods the low-lying areas of the continents, creating vast, shallow marine environments.

The epeiric platforms occur in various geological settings, including rift basins, cratonic basins, and coastal plains. They are found in both tropical and temperate regions, and their sedimentary deposits range from carbonate rocks to clastic sediments, depending on the local geology and environmental conditions. As such, the epeiric platforms are characterized by low-energy environments, calm water conditions, and limited wave and current action, allowing for the accumulation of thick sedimentary sequences, including muds, sands, and carbonates. Hence, the sedimentation processes on epeiric platforms are influenced by a combination of factors, including the input of sediment from rivers, biological activity, and chemical processes. Rivers transport sediments from the continent into the platform, contributing to the deposition of clastic sediments. Biological activity, such as the growth and accumulation of marine organisms, facilitates the formation of carbonate sediments. Chemical processes, such as the precipitation of minerals from seawater, also contribute to the formation of sedimentary deposits. Epeiric platforms also support diverse ecosystems, such as coral reefs, seagrass beds, and benthic communities that inhabit the sediment surface or burrow within it. The shallow-water depths and relatively stable environmental conditions also make epeiric platforms important habitats for a wide range of marine life. The epeiric platforms also have economic significance, as the sedimentary deposits on epeiric platforms serve as reservoirs for oil, gas, and other valuable resources and provides valuable insights into past sea-level changes, paleoclimate, and the evolution of marine life.

No modern examples of carbonate epeiric platforms exist; however, such platforms were common in the past, particularly during the Paleozoic and parts of the Mesozoic periods (Wright and Burchette, 1996; Wilmsen et al., 2018). The bulk of carbonate sediments formed throughout geologic time were deposited on epeiric platforms, particularly during the late Precambrian (as in China), Cambrian–Ordovician (as in North America and the Middle East), Mississippian, Triassic–Jurassic (as in Western Europe), Permian, and Tertiary (as in the Middle East) (see Wright and Burchette, 1996; Tucker et al., 1990).

Although carbonate environments extend from the supratidal zone to deeper parts of the basin, off the shelf; the shallow platform basin (that constitutes the middle and outer shelves) is also the primary site of carbonate production, and the so-called “subtidal carbonate factory” of James (1984) (see also Laugié et al., 2019) (Figure 20.4). The sediments produced in this carbonate factory are deposited mainly on the shelf; however, some sediments are eventually transported landward onto tidal flats and beaches and into subtidal settings. Others are transported seaward off the shelf onto the slope and into the deeper basin (Figure 20.4). Little carbonate sediment is generated in the deeper-water basin environment off the shelf except for the fallout of calcium carbonate secreting plankton from near-surface waters.

20.2.2 Slope/Basin Carbonates

The carbonate sediments are largely shallow-water deposits (as outlined above), however, deeper-water carbonates are also noted in modern oceans, such as the Bahamas-Florida region, Belize, Jamaica, Grand Cayman, the northeast Australian coast, and several atolls in the Pacific and Indian Oceans (see Coniglio and Dix, 1992). They have also been reported in the ancient past from many Phanerozoic successions, deposited primarily on the shelf, largely derived from the shelf by transport

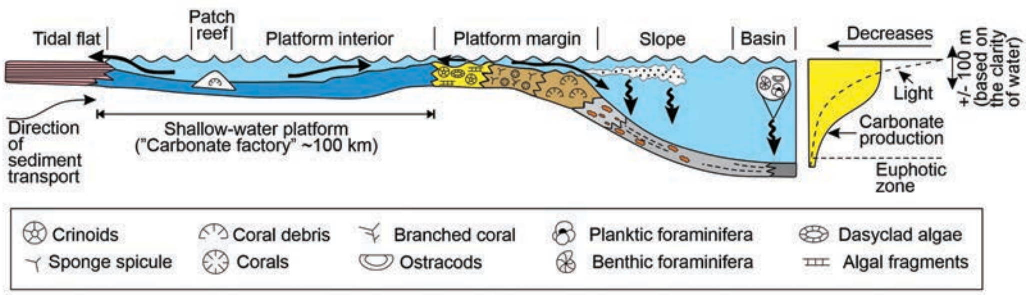

FIGURE 20.4 Slope/basin carbonates (subtidal carbonate factory). The carbonate environments extend from the supratidal zone to deeper parts of the basin, off the shelf; the shallow platform basin is also the primary site of carbonate production. (Modified after James, 1984.)

processes that include storm waves, turbidity currents, debris and grain flows, slumping, sliding, and rock falls. It must be noted that modern slope carbonates consist mainly of pure carbonates, in contrast to many ancient slope deposits that include a large percentage of terrigenous clastic rocks, as well.

20.2.2.1 Chemical and Biochemical Processes

Seawater pH, temperature, and CO_2 control the solubility of calcium carbonate ($CaCO_3$). Loss of CO_2 due to increased temperature, decreased pressure, or plant photosynthesis greatly influences the inorganic precipitation of $CaCO_3$. Although, in both modern and ancient oceans, the relative importance of chemical (inorganic) precipitation of $CaCO_3$ as compared to its organic production is not well-known (see Shinn et al., 1989). In modern oceans, carbonate deposition by organisms capable of extracting $CaCO_3$ from the seawater to build their shells, is far more important than inorganic processes.

Chemical processes involving carbonates include carbonation, precipitation, dissolution, and acid-base reactions, while the biochemical processes include calcification, photosynthesis, respiration, and carbonate buffering. All these are very briefly enumerated below. Carbonation is the process by which carbon dioxide gas dissolves in water to form carbonic acid, which then reacts with carbonate minerals to form bicarbonate ions. Precipitation occurs when the concentration of carbonate ions in water is high, and they can react with calcium or magnesium ions to form insoluble carbonate minerals such as calcite or aragonite. Dissolution is the opposite of precipitation, and occurs when carbonate minerals are broken down into carbonate ions and dissolved in water. In acid-base reactions, the carbonate ions act as a base and react with acids to form CO_2 gas, water, and salt.

Biochemical processes are carried out by organisms such as corals, mollusks, and algae. Many marine organisms, including corals and mollusks, use carbonate ions to build their shells or skeletons (calcification). They take up dissolved carbonate ions from seawater and convert them into solid calcium carbonate structures. Some algae and plants, such as seagrasses, use carbon dioxide from the water to carry out photosynthesis. This process converts carbon dioxide into organic carbon compounds, releasing oxygen as a byproduct. Organisms that live in aquatic environments, including bacteria and animals, respire by converting organic carbon compounds into carbon dioxide (respiration). This process releases carbon dioxide back into the water. The presence of carbonate ions in water helps to regulate the pH by acting as a buffer (carbonate buffering). They can absorb excess hydrogen ions, preventing the water from becoming too acidic. Thus, through carbonate chemical and biochemical processes, besides playing a crucial role in the carbon cycle, they also regulate the pH in aquatic environments, and the formation of carbonate minerals.

Biogenic processes have been important throughout the post Precambrian, although some may have played a role in the carbonate production during the Precambrian as well. Organisms also contribute to the formation of carbonate sediments through their feeding and bioturbation activities, causing the breakdown of skeletal fragments and other carbonate materials. However, it must be mentioned that the organisms that are primarily responsible for carbonate production in the modern oceans are not necessarily the same as those that were once major carbonate contributors in the past, and that the principal carbonate formers have also changed through time (see Figure 20.5) (Wood et al., 2011).

Coccoliths and foraminifers are composed of low-magnesian calcite, whereas crinoids and echinoids are composed of high-magnesian calcite, and calcareous green algae and gastropods are composed of aragonite. Thus, some groups of organisms, such as corals, secreted different minerals at different times in their history in response to changes in seawater chemistry, primarily through changes in the ratio of magnesium to calcium (Stanley and Hardie, 1998, 1999; Wood et al., 2011). Corals formed calcite (low- or high-Mg) skeletons during much of the Paleozoic when a "calcite sea" characterized the world oceans; however, more recent corals, especially those living during the late Cenozoic, secreted aragonite skeletons (see also Sandberg, 1985) (see Figure 20.5). Although the

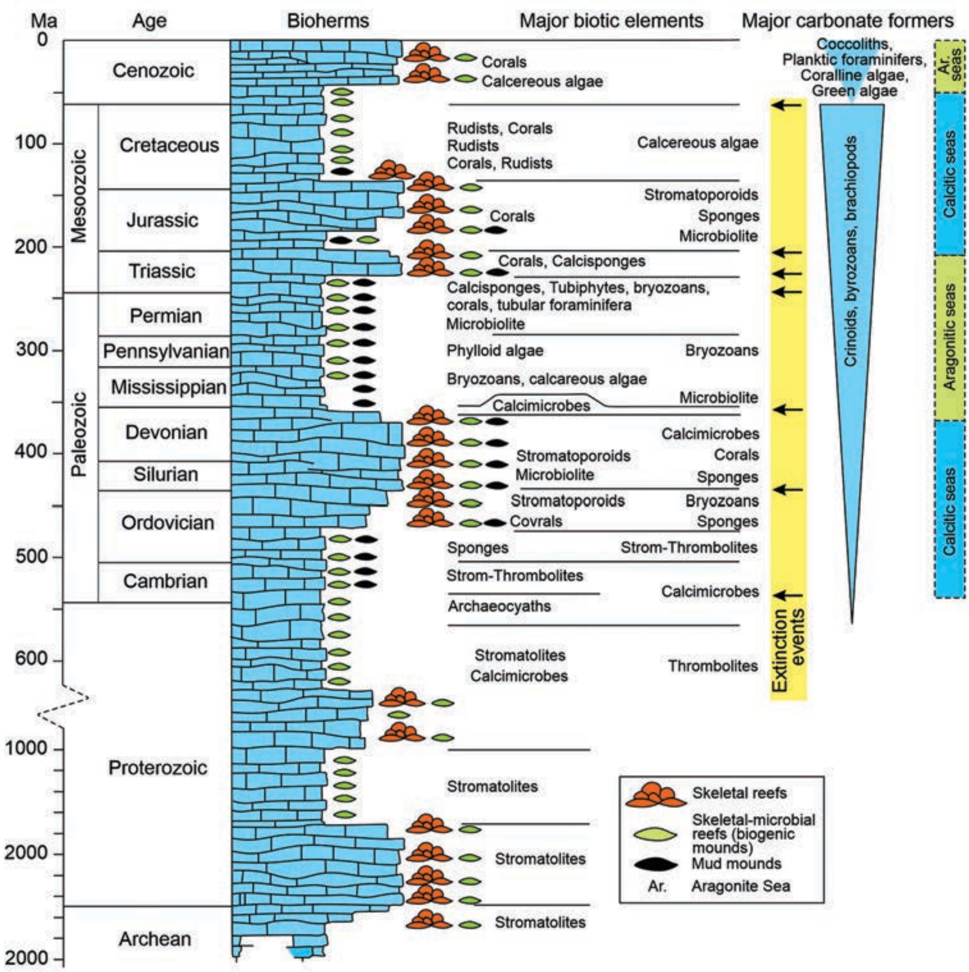

FIGURE 20.5 Major carbonate contributors through time. (Modified after James, 1984; Wood, 2011.)

skeletal mineralogy of many groups of organisms appears to parallel that of inorganic precipitates formed during times of calcite or aragonite seas, some other organisms, such as echinoids, crinoids, and brachiopods, secreted the same skeletal minerals throughout their history despite changing sea chemistry. Skeletal structures composed of aragonite are chemically less stable than calcite ones, and are thus more susceptible to dissolution and destruction during diagenesis.

20.2.2.2 Physical Processes

Physical processes are important in both the reworking and transport of carbonate materials on the shelf, and in aiding the production of carbonate sediments (Figure 20.6). Circulation of water onto the shelf brings fresh nutrients, necessary for organic growth, from deeper waters. Breaking waves against reef barriers on the outer shelf increases oxygen content in the water by interacting with the atmosphere and decreases CO_2 due to decreased water pressure. Thus, modern reefs are best developed in wave-agitated zones, and biogenic production of carbonate sediment, in general, is stimulated by strong water movements. On the other hand, strong waves crashing on the reef front break down reef rock, producing sand- and gravel-size bioclasts that subsequently undergo transport (both seaward and landward) from the reef. Agitated water is important for the formation of ooids, and currents help to generate and preserve grapestones and hardened fecal pellets by submarine accretion and cementation (grapestones are composite grains with an irregular shape that resembles a bunch of grapes). Waves and currents also winnow fine carbonate mud from coarser sediment and transport this mud off the shelf platform or into sheltered or protected areas of the shelf (Figure 20.6). Depending upon water energy, the coarser sediment itself may either remain as a winnowed lag deposit, forming sand- or gravel-covered flats, or be transported and deposited to create wave-formed bars and shoals, beaches, spits, or tidal deltas. Wave- and current-transported and -winnowed carbonate sand deposits are common along the outer edges of the shelf platform, where water energy is highest. Storms, as in siliciclastic shelves, also play an important part on carbonate shelves in re-suspension and transport of sediments. For example, storms transport most sediments from the subtidal shelf into the intertidal (tidal-flat) environment. Absence of wave and current activity on the shelf leads to stagnant circulation, leading to increased salinity, and possibly anoxic conditions. Such restricted environments constitute unfavorable habitats for many normal marine organisms, resulting in less carbonate production. Additional carbonate physical processes include diagenesis, lithification, metamorphism, weathering and erosion, and tectonic processes. These are very briefly enumerated below.

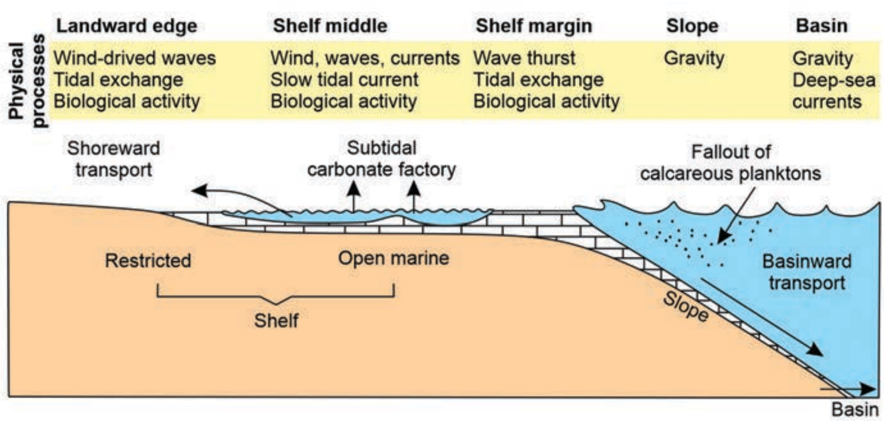

FIGURE 20.6 Carbonate shelf physical processes. These processes are important as they enable both reworking and transport of carbonate materials on the shelf, and in aiding the production of carbonate sediments.

Diagenesis is the physical and chemical changes that occur in sedimentary rocks after they are deposited. In carbonate rocks, on a very basic scale, diagenesis involves compaction, cementation, and recrystallization. Compaction occurs when the weight of overlying sediments compresses the carbonate particles, reducing pore space and increasing rock density. Cementation occurs when minerals such as calcite or dolomite precipitate and fill in the pore spaces, binding the sediment particles together. Recrystallization involves the dissolution and re-precipitation of carbonate minerals, leading to the growth of larger crystals. Lithification is the process by which loose sediments are transformed into solid rocks. In carbonate sediments, lithification involves the compaction and cementation processes mentioned above. Cementation creates a strong and durable rock, such as limestone or dolomite. Metamorphism refers to the changes that occur in rocks brought about by high temperatures and pressures. In carbonate rocks (limestone), metamorphism leads to the formation of marble. The heat and pressure cause recrystallization of carbonate minerals, resulting in a coarse-grained, crystalline rock. Weathering refers to the breakdown of the rocks at the earth's surface, while erosion involves its transport and removal. Carbonate rocks are susceptible to weathering, especially by acidic waters that dissolve carbonate minerals leading to the formation of karst landscapes, characterized by sinkholes, caves, and underground drainage systems. Tectonic forces, such as folding, faulting, and uplift, deform and reshape carbonate rocks resulting in the formation of folded and faulted structures, and the exposure of carbonate rocks at the earth's surface. Thus, carbonate physical processes play an important role in the formation, transformation, and erosion of carbonate minerals and rocks.

20.2.3 Skeletal and Sediment Characteristics of Carbonate Deposits

The deposition of carbonate sediments is favored in moderately shallow and warm waters that receive little terrigenous siliciclastic input. Some carbonate sediments are deposited in deeper waters, beyond the shelf edge. Most carbonate sediment deposited in deeper water results from the fallout of calcareous plankton (Figure 20.6) such as foraminifers, green algae (coccoliths), and tiny gastropods. These pelagic calcareous organisms evolved mainly in Jurassic and post-Jurassic times; therefore, deeper-water pelagic carbonates are not important in older rocks. In addition to pelagic carbonates, some shallow-water carbonate sediments may be swept off from carbonate platforms into deeper waters by storms or be transported by sediment gravity-flow processes (such as turbidity currents). Although carbonates form predominantly in warm-water settings, they can also accumulate in some cool-water, higher-latitude environments.

The cool-water carbonates, in some modern shelves, make an important contribution (see Farrow et al., 1984; Nelson, 1988; James and Clarke, 1997). These shelves are in middle- to low-latitude settings where cool-water currents intrude, and at high latitudes such as the Spitsbergen Bank in the Barents Sea. The cool-water carbonate shelf deposits have also been reported in ancient rocks ranging in age from Paleozoic to Tertiary in North America, Australia, and Europe (see James and Clarke, 1997).

In cool-water environments, the carbonate sediment is composed almost entirely of the skeletal remains of organisms, dominated by foraminifers and mollusks; such assemblages are referred to as foramol assemblages (Lees and Buller, 1972; Jones and Desrochers, 1992) (see Figures 20.7A–B). They are composed of benthic foraminifers, mollusks, barnacles, bryozoans, and calcareous red algae (Figures 20.7A–B). James (1997) suggests that heterozoan (named after organisms that feed through heterotrophic means) assemblage is a more appropriate term for foramol assemblage (Figure 20.7C).

Modern warm-water carbonates, especially reef carbonates, accumulate at a much faster rate than do cool-water carbonates. The warm-water (>20°C) carbonate assemblages of organisms, called chlorozoan (named from chlorophyta and zoantharia corals), are dominated by hermatypic corals (i.e., corals that live primarily within the photic zone) and calcareous green algae in addition to foramol components (Figures 20.7A–B). James (1997) proposed to replace the term chlorozoan by photozoan, to emphasize the light-dependent nature of the biotic constituents (Figure 20.7C). The warm-water carbonates contain, in addition to skeletal remains, substantial amounts of ooids, aggregate grains, peloids, and lime mud (see Figure 20.7A). The aforementioned assemblages are detailed below.

The skeletal and sediment characteristics of carbonate deposits also refer to the composition and physical properties of the materials that make up carbonate rocks, thus, providing information about the depositional environment and the processes that formed these deposits. Key characteristics include: skeletal composition, grain size, texture, porosity and permeability, and fossil content. Carbonate deposits often contain the remains of marine organisms, such as corals, mollusks, foraminifera, and algae that make up a significant fraction of a carbonate rock (skeletal composition). The composition of these skeletal fragments depends on the dominant organisms present within the depositional environment. Carbonate deposits range in grain size from fine-grained mud to coarse-grained sand. The grain size is influenced by factors such as the energy of the depositional environment and the availability of carbonate material. Generally, coarse-grained carbonate deposits are often associated with high-energy environments, such as coral reefs or shallow marine settings, while the fine-grained ones are found in low-energy environments like lagoons or deep-sea settings. The texture of carbonate deposits is the arrangement and shape of the grains within the rock. Carbonate rocks exhibit various textures, including mudstone, wackestone, packstone, grainstone, and boundstone. Porosity is the amount of empty space or voids within a rock, while permeability refers to the ability of fluids to flow through the rock. Carbonate deposits exhibit varying degrees of porosity and permeability, depending on factors such as the presence of pore spaces (voids), the degree of cementation, and the nature of the rock fabric. High-porosity carbonate rocks, such as reef complexes, serve as important reservoirs for oil and gas. The presence and diversity of fossils (including trace fossils such as burrows or tracks) in carbonate deposits provide insights into the paleoecology and paleoenvironment of the depositional setting. Thus, understanding skeletal and sediment characteristics of carbonate deposits enables better interpretation and reconstruction of the depositional environment, and in assessing the potential economic value of carbonate reservoirs.

20.2.3.1 Foramol Assemblages

Foramol assemblages are communities of foraminifera (single-celled organisms with a shell or test) that are found in carbonate deposits, specifically in foramol facies, i.e., carbonate facies that are primarily composed of foraminifera tests (Figures 20.7A–B). The composition and characteristics of foramol assemblages varies depending on factors such as temperature and salinity (water depth, nutrient availability, and sedimentation rates also affect foramol assemblages) (Mutti and Hallock, 2003) (Figures 20.7B–C). Different species of foraminifera have different environmental tolerances and preferences, so their presence or absence within a foramol assemblage provides clues about the prevailing paleoenvironmental conditions during deposition. For example, certain species of foraminifera are indicative of shallow-water, high-energy environments like coral reefs or near-shore settings. Other are indicative of deeper-water, low-energy environments like lagoons or open marine settings. Thus, by studying the foramol assemblages in carbonate rocks, past marine environment and changes in sea level, water chemistry, and other environmental factors over time, can be reconstructed. In addition to their environmental significance, the foramol assemblages also have economic significance as some foraminifera species are useful index fossils (biostratigraphic markers), enabling to date and correlate rock layers at different locations. This information is valuable in the exploration and production of oil and gas reservoirs, as well as in the study of geological history and paleoenvironments.

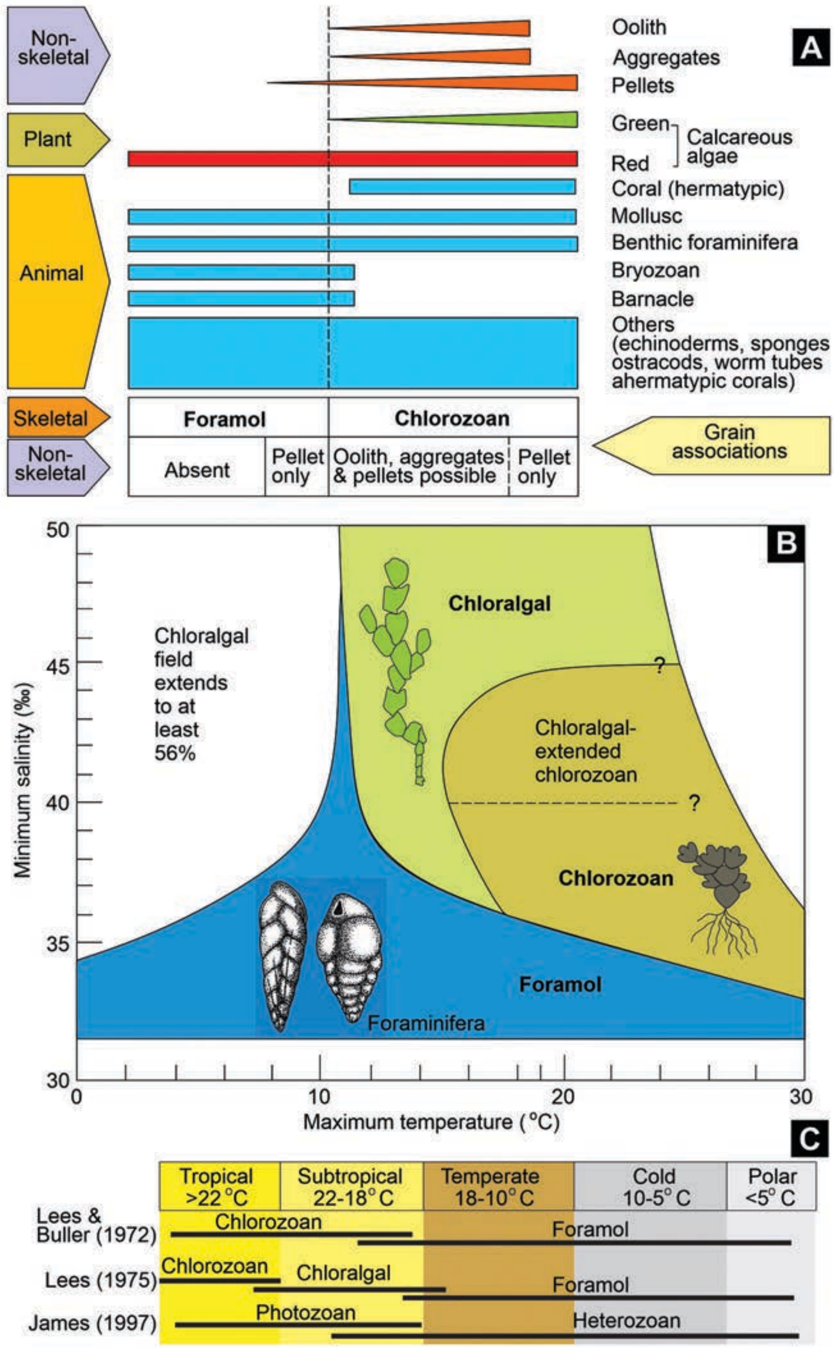

FIGURE 20.7 Carbonate assemblages. A: Skeletal grain associations in modern shelf carbonate environments. The figure shows the carbonate grain types associated with foramol and chlorozoan assemblages. Bars indicate the importance of grain type or a dominant constituent (Lees and Buller, 1972). B: Relationships between salinity–temperature annual ranges and occurrences of skeletal associations in modern shelf carbonate environments. (Modified after Lees and Buller, 1972.) C: The modern carbonates are influenced by ocean water temperature and salinity and are accordingly classified into distinctive biogenic associations along temperature and salinity gradients. (Modified after Mutti and Hallock, 2003.)

20.2.3.2 Heterozoan Association

James (1997) suggested that the foramol assemblage should be termed heterozoan (appropriately named after organisms that feed through heterotrophic means) (Figure 20.7). He first used the terms photozoan and heterozoan to remove temperature dependence between skeletal assemblages and associations; the skeletal assemblages are now related to their trophic requirement (light-based photosynthesis versus other sources) rather than water temperature alone (see Figures 20.7B–C) (James, 1997). A heterozoan association is a community of heterozoans representing a particular environment, ranging from terrestrial, freshwater, and marine environments and includes a wide range of organisms, such as animals, fungi, and some bacteria (see James 1997) (Figure 20.7A). These organisms rely on organic food sources rather than photosynthesis for their energy and nutrition; they are typically consumers or decomposers in the ecosystem. In the modern ocean, photozoan and heterozoan associations have distinct distribution patterns both with respect to latitude and ocean circulation. The heterozoan carbonate systems dominate in cooler eastern sides of the oceans and areas that are influenced by upwelled nutrient-rich waters or by nutrient-enriched terrestrial runoff, and in polar latitudes, whereas the photozoan association dominates in tropical and subtropical western ocean basins (see Figure 20.7C) (see also James 1997; Mutti and Hallock, 2003). The composition and characteristics of heterozoan association depend on the availability of organic food sources. In marine environments, heterozoan associations are found in areas with high organic productivity, such as nearshore habitats or areas with upwelling. These associations include filter-feeding organisms like bivalves and sponges, as well as detritivores and scavengers that feed on organic particles and detritus. Thus, by studying the heterozoan associations an insight into the functioning and dynamics of ecosystems is noted as this association plays an important role in nutrient cycling, energy transfer, and the decomposition of organic material.

20.2.3.3 Chlorozoan Association

Chlorozoan associations are communities of chlorozoans, microorganisms found in marine environments (Figure 20.7A). They are photosynthetic protists that belong to the phylum chlorophyta, commonly known as green algae. They are characterized by their ability to photosynthesize and produce chlorophyll. Chlorozoan assemblages are found in a variety of marine habitats, including coastal areas, coral reefs, and open ocean environments. They are primary producers, converting sunlight and nutrients into organic matter through photosynthesis. Hence, the chlorozoan assemblage is a key component of the marine food web, providing food and energy for other organisms. The composition and characteristics of chlorozoan assemblages depend on environmental factors such as temperature, light availability, nutrient levels, and water chemistry (Mutti and Hallock, 2003). Different species of chlorozoans have different ecological preferences and tolerances, leading to the formation of unique assemblages in different habitats (Figure 20.7A). For example, in shallow coastal areas, chlorozoans may form dense mats or blooms in response to high nutrient levels leading to eutrophication; extensive blooms also lead to oxygen depletion within the water column. In coral reef ecosystems, chlorozoans are important symbionts of corals, providing them with energy through photosynthesis. Changes in the composition or abundance of chlorozoans indicate shifts in environmental conditions or disturbances, such as climate change or pollution. In addition to their ecological significance, chlorozoans also have economic importance. Some species of chlorozoans are commercially cultivated for food, such as seaweed or algae products. They are also used in biotechnology, pharmaceuticals, and biofuel production.

20.2.3.4 Photozoan Association

A photozoan association is a community of organisms that relies on photosynthesis for their energy and nutrition needs. Photozoans are organisms that possess chlorophyll or other pigments that enable them to capture sunlight and convert it into chemical energy through photosynthesis. Photozoan associations are noted in various ecosystems, including terrestrial, freshwater, and marine environments; they dominate in tropical and subtropical western ocean basins (Figure 20.7C). These associations typically include plants, algae, and some bacteria that are capable of photosynthesis. They are the primary producers in the ecosystem, converting sunlight into organic matter and serving as the foundation of the food web. In marine environments, photozoan associations are abundant and diverse and include different varieties of marine algae, ranging from microscopic phytoplankton to large seaweeds. Photozoans play an important role by providing food and habitat for myriad marine organisms, ranging from tiny zooplankton to large marine mammals. In freshwater ecosystems, photozoan associations include various types of algae, such as diatoms, green algae, and cyanobacteria. These organisms float or attach themselves to surfaces in the water and use photosynthesis to produce energy. They are also an important food source for aquatic organisms, including zooplankton, small fish, and invertebrates. Studying the photozoan association is essential for understanding the cycling of nutrients and energy; changes in their composition or abundance also indicates shifts in environmental conditions.

20.2.4 Reefs

The precise meaning of the term reef is still a hotly debated topic. Carbonate workers are unable to agree on whether to restrict its use to carbonate buildups or bioherms that have a rigid organic framework or core, or to extend its definition to include carbonate buildups of other types that do not have a rigid framework or core. Here, a reef is a structure built by the in-place growth of organisms that acts as frame-builders. A reef is a wave-resistant, prominent structure on the sea floor and, therefore, influences and modifies sedimentation patterns in its vicinity.

20.2.4.1 Reef Depositional Setting

Most modern reefs form in shallow waters (Figure 20.8A). The most striking are the linear reefs located along platform margins, commonly called barrier reefs (Figure 20.8A). These reefs are more or less laterally continuous, and extend for hundreds of kilometers, such as the Great Barrier Reef of Australia that runs for some 1,900 km along the eastern shelf of Australia. In a few modern localities where shelves are very narrow, linear reefs are located up against the shoreline, with no intervening lagoon; these are called fringing reefs (Figure 20.8A). Isolated, doughnut-shaped reefs are called atolls; these occur around the tops of some Pacific seamounts that rise out of the deep waters (Figure 20.8B). These reefs form an outer wave-resistant barrier that encloses a shallow lagoon (Figure 20.8B). Small isolated reef masses commonly referred to as patch reefs (also pinnacle reefs, or table reefs) occur along some shelf margins or are scattered on the mid-shelf (Figures 20.8A–B).

20.2.4.2 Reef Structure

A reef consists of a central core, the reef framework, that grades seaward into the reef slope, and a loose accumulation of reef debris called the fore-reef talus (Figure 20.9A). The nearly flat, uppermost, shallowest part of the reef is called the reef flat, which grades landward into back-reef skeletal (coralgal) sands and subtidal lagoonal deposits (Figure 20.9A). Physiographically, the reefs are divided into fore-reef, reef-front, reef-crest, reef-flat, and back-reef zones (Figure 20.9A).

20.2.4.2.1 Reef Front and Reef Crest

The reef front extends from the highest point on the reef profile (reef crest) to a point below, where little or no skeletal frame-building occurs (Figure 20.9A). This depth varies according to local conditions but may extend to 70–100 m in modern reefs. The highest zone on the reef front, the reef crest, is the most exposed part of the reef, and is subject to wave activity (Figure 20.9A). The resulting reef morphology and composition depend on the prevailing energy regime (Adey, 1978; Tucker et al., 1990). In such high-energy zones, encrusting organisms dominate, especially coralline algae (Adey, 1975; Tucker et al., 1990). Crustose coralline algae predominate in such settings, as

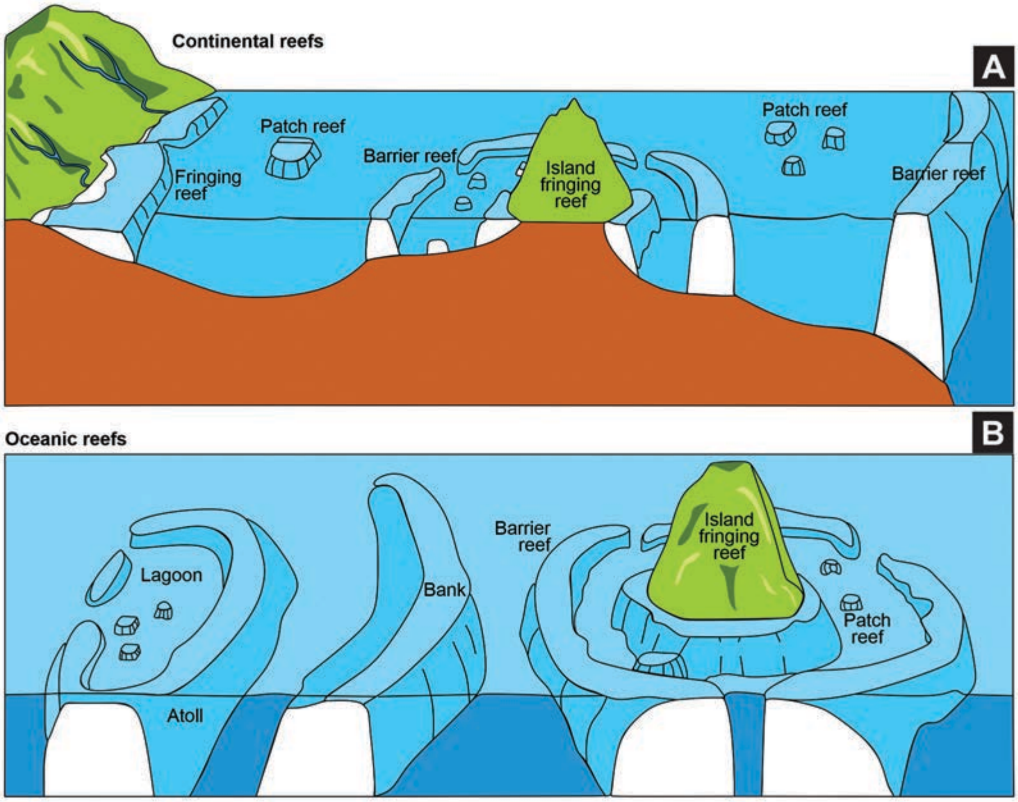

FIGURE 20.8 Types of reefs based on their shape. A: Continental reefs. B: Oceanic reefs.

they are able to withstand wave action and also have a high tolerance to light encountered in shallow waters (Bosence, 1983). In lower-energy crest zones, the hydrozoan *Millepora* occurs, or robust corals such as *Acropora palmata* (Geister, 1977). The crest zone also shows the highest levels of skeletal breakage and abrasion, thus, producing a rubble-strewn surface (Macintyre, 1984; Tucker et al., 1990).

20.2.4.2.2 Fore-Reef Slope

Seaward of the reef front is the fore-reef slope that grades down into the surrounding basin floor (Figure 20.9A). It is dominated by gravity-flow deposits and pelagic-hemipelagic sediments (Enos and Moore, 1983; Tucker et al., 1990). Two types of slopes are noted: depositional (or accretionary) reef margins, and bypass (escarpment) margins (McIlreath and James, 1984; Tucker et al., 1990). In the former, the reef front passes downslope continuously into the basin, whereas in the latter, a steep submarine escarpment separates the reef front from talus deposits, grading out into basinal sediments (Figure 20.9A).

20.2.4.2.3 Reef Flat

Behind and partially protected by the reef crest is the reef-flat zone, with two environments, the reef pavement and the sand apron (Figure 20.9A). The reef pavement is immediately behind the crest and is afforded some shelter by it. The pavement zone is variable in width, from a few meters to over 100 m as on the Belize barrier reef (James and Ginsburg, 1979; Tucker et al., 1990). The depth is typically only a few meters and the zone may be exposed at low tide.

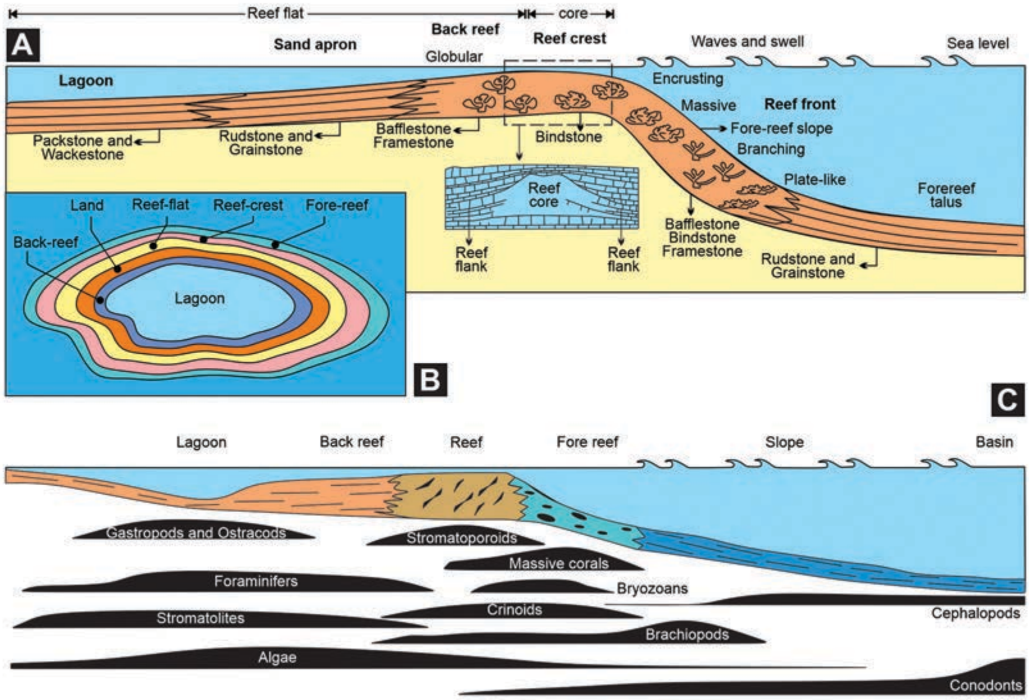

FIGURE 20.9 Reef and lagoon structure. A: Generalized structure of the reef with the associated microfacies. B. Generalized structure of a reef lagoon. C: The distribution of main fossil groups within the reef in the middle Devonian of the Carnic Alps. (Modified after Vai et al., 2002.)

20.2.4.2.4 Back-Reef Lagoons

Lagoons are protected, lower-energy areas behind the reef crest (Figure 20.9). Not all reefs have lagoons. Lagoons are variable in size, ranging from relatively small areas developed within atolls to larger zones behind major barrier reefs. Many reef lagoons are shallow (<10 m) some, like the Pacific atoll lagoons, are >70 m deep forming deeper-water deposits and are then, sometimes, misinterpreted as inter-reef sediments. The main characteristic of a reef lagoon is that it is protected and has restricted circulation (Figure 20.9). Thus, it has a lower-energy setting with finer sediments and different biota to other reef facies. Lagoons act as traps for sediment and the settling of fine sediment deters many sessile organisms (Figure 20.9C). Besides corals and sea-grasses, calcareous algae such as *Halimeda* and *Penicillus* are an important component and produce large amounts of carbonate sand and mud. Further sediments may come from terrigenous sources.

20.2.4.3 Bioherms and Biostromes

Bioherms are mound- or lens-shaped structures composed mainly of the skeletons or shells of carbonate-secreting organisms and enclosed in rocks of different lithology or character (Figure 20.10A). Some are true, in situ reefs, whilst others are formed as banks of loose, transported skeletal material. A bioherm may or may not have a rigid internal organic framework (Cumings, 1932).

Biostromes, on the other hand, are ribbon or sheet-shaped structures (Figure 20.10B), also made largely of skeletal material, either in growth position (reefs in the strict sense) or transported. Biostromes typically form in non-reef platform environments. Wilson (1975) uses the term carbonate buildup for a body of locally formed, laterally restricted, carbonate sediment that possesses

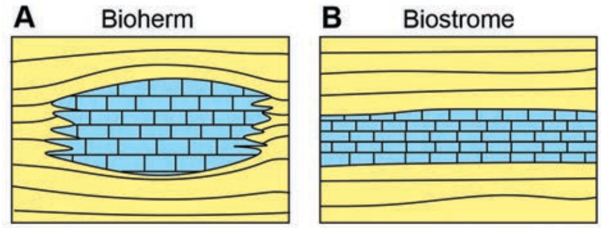

FIGURE 20.10 Bioherm and biostrome. A: A bioherm is a mound- or lens-shaped structure of skeletons or shells of carbonate-secreting organisms. B: A biostrome is also made of skeletal components, either in growth position (reefs in the strict sense) or transported, but is a ribbon or sheet-shaped structure.

topographic relief, without regard to the internal makeup of the buildup. Most reefs are built by larger organisms that are capable of thriving in energetic environments.

Thus, bioherm and biostrome are terms used to describe the outline shape of an organic accumulation, and not to denote any inherent internal structural organization or composition of the lens (Laborel, 2011). Biostromes are most usefully considered as single organic layers (i.e., beds; see Figure 20.10B) (Kershaw, 1994).

20.2.4.4 Mud Mounds

Biodetrital mounds, carbonate mounds, carbonate mud mounds, lime mud mounds, microbial mounds, mudbanks, reef mounds and stromatactis mounds are all synonymous terms for mud mounds (see Rodríguez-Martínez, 2011). Mud mounds are biosedimentary buildups (part of the reef system) (see Figure 20.11) that are dominated by fine-grained carbonates (>50% of rock volume), forming a matrix-supported fabric. Mud mounds have colonized the oceans since the Proterozoic period, from shallow platforms to deep aphotic basinal settings (Rodríguez-Martínez, 2011). They have varied external forms, from dome- to tabular-shaped. They also occur as isolated massive bodies or as stacked or amalgamated bodies, spreading over several kilometers. Dominant invertebrate macrofossils on mud mounds include pelmatozoans, bryozans, and sponges; others such as brachiopods, mollusks, arthropods, calcareous algae, and solitary and colonial corals are sporadically important elements (see Figure 20.11). Benthic foraminifera and polychaete worms also occur as part of the mud mound biota. In general, the biota of mud mounds is mainly dominated by heterozoan assemblages (see Figure 20.7). Mud mounds range in size from small (1–5 m high) to large structures that may be 100 m high (see Wendt et al., 1997).

20.2.4.5 Reef Organisms

Not all reefs are coral reefs; many organisms contribute to the formation of reefs, such as the blue-green algae (cyanobacteria), coralline red algae, green algae, encrusting foraminifera, encrusting bryozoa, sponges, and mollusks, among others (see Figure 20.5). In the geologic past, reef-building organisms also included some now-extinct groups, such as the archaeocyathids, stromatoporoids, fenestellid ioryozoans, and rudistid clams (see Figure 20.5).

In modern reefs, corals dominate (Figures 20.9 and 20.11A). Two types of corals are noted. In shallow-water reefs, the principal ones are hermatypic corals (such as zooxanthellae; hexacorals). Hermatypic corals have a symbiotic relationship with several kinds of unicellular organisms, mainly algae, referred to collectively as zooxanthellae. These algae live in or between the living cells of the corals and aid them in gaining energy by producing photosynthetic products (Cowen, 1988). They also facilitate the process of secreting calcium carbonate by removing CO_2 from the tissues during photosynthesis. As zooxanthellae require sunlit waters, hermatypic corals are restricted to living in very shallow waters that are sunlit.

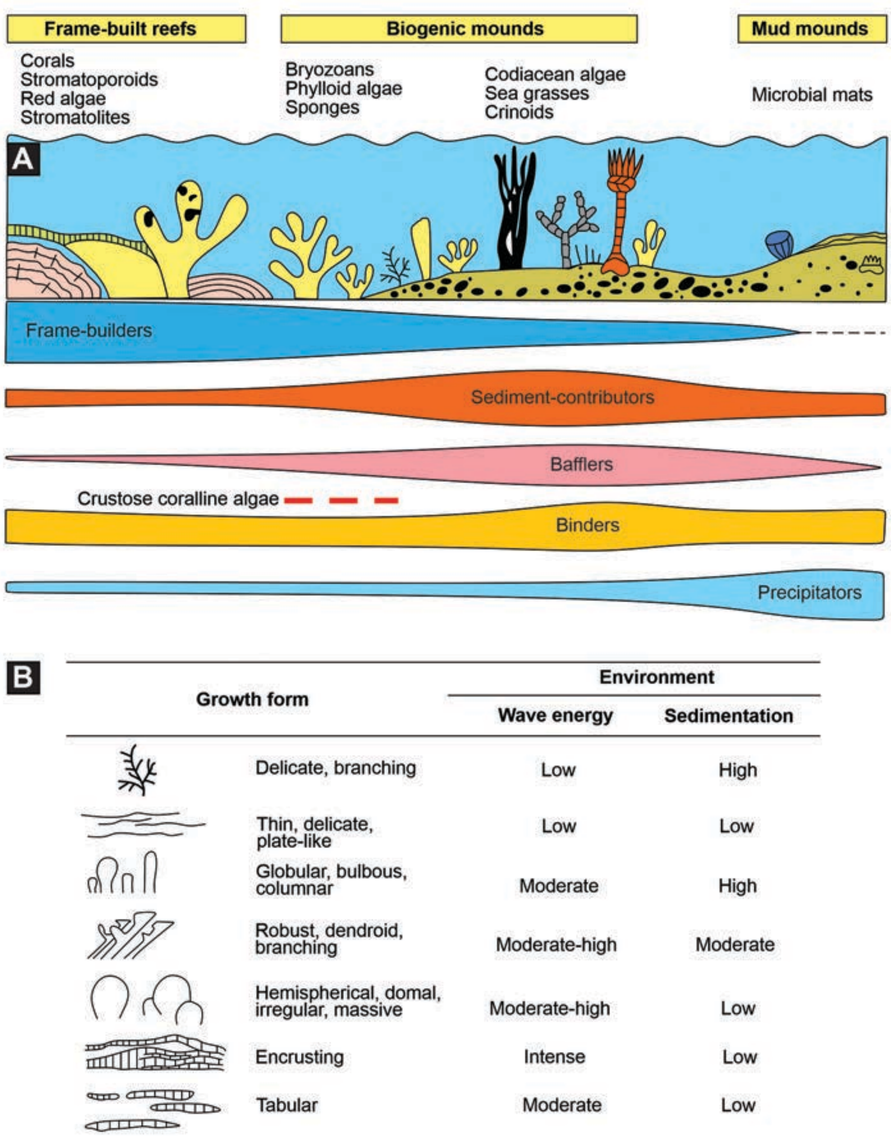

FIGURE 20.11 Mud mounds and their fauna. A: Reef-building organisms. B: Growth forms of reef-building organisms.

Ahermatypic corals (i.e., azooxanthelae), on the other hand, lack a symbiotic relationship and are not restricted to shallow water. They are one of the principal organisms today that form carbonate buildups in deeper waters. Their distribution ranges from shallow waters to depths greater than 2000 m (Stanley and Cairns, 1988).

Some reef-building organisms, such as corals and stromatoporoids (sponges), are important frame-builders (see Figures 20.5 and 20.11A), that construct wave-resistant cores of reefs. Others, such as crinoids and codiacian algae (such as *Halimeda*), whose skeletal elements may disintegrate into smaller fragments, are important sediment contributors (Figure 20.11A). Bafflers are organisms such as seagrass that provide a protective baffle against currents and thus generate a localized, low-energy environment in which fine sediments accumulate (Figure 20.11A). Binders such as cyanobacleria (that form stromatolites) trap and bind sediment (Figure 20.11A). Precipitators are primarily microbes, such as cyanobacteria, that help mediate precipitation of carbonate muds (Figure 20.11A).

The growth forms of reef-building organisms may range from delicate branching forms to globular or massive structures (Figure 20.11B). The form of the organisms is closely related to the water energy over the reef and, thus, varies over different parts of the reef. Organisms that live in low-energy parts of the reef tend to have delicate branching or plate-like forms (see Figure 20.11B). Those living in higher energy zones of the reef develop hemispherical, encrusting, or tabular forms that are better able to withstand strong wave action (Figures 20.9 and 20.11B).

20.2.4.6 Major Processes in Reef Formation

There are four processes that operate at varying degrees in the reef formation: constructive, destructive, cementation, and sedimentation.

20.2.4.6.1 Constructive Processes

The biological processes include the direct growth of calcareous organisms, or the effects of organisms such as sediment baffling or binding (Figure 20.11A). Heavily calcified forms or colonies act as the building blocks of the reef; they are the primary frame-builders. In modern reefs, these include scleractinian corals, crustose coralline algae and *Millepora*, a hydrozoan, whereas in ancient reefs, this role was taken by scleractinian, rugose and tabulate corals, stromatoporoids, calcareous algae and stromatolites (see also Figures 20.5 and 20.11A). The secondary frame-builders are encrusters, such as crustose coralline algae, serpulids, bryozoans, corals, foraminifers and gastropods in modern reefs. The sediment contributors include the calcareous alga, *Halimeda*; the sediment contributors are important in the formation of reef mounds where the accumulation of locally produced calcareous sediments results in reef growth (Figure 20.11A).

20.2.4.6.2 Destructive Processes

These processes damage and destroy growing reefs and include physical effects such as waves and biological destruction (bioerosion) (Figure 20.12). Physical destruction is constant and is caused

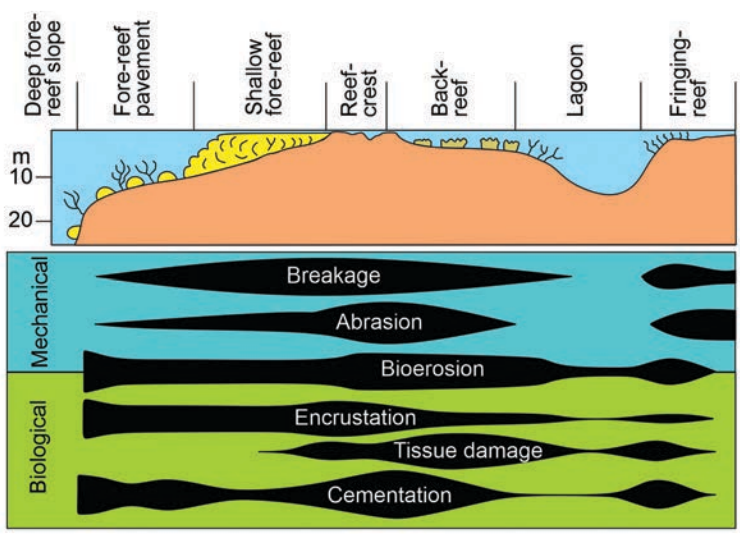

FIGURE 20.12 Major mechanical and biological processes noted in reef formation.

by wave and current activity on the reef. Storms and hurricanes are major influencers when major destruction of reef communities and sediment transport occurs (Hernandez-Avila et al., 1977; Tucker et al., 1990). Reef bioeroders are of four types: borers, raspers, crushers, and burrowers (Schroeder and Zankl, 1974). Borers include algae, cyanobacteria, fungi, sponges, sipunculids, polychaetes, mollusks, barnacles, and echinoids (Tucker et al., 1990). Raspers are browsing organisms, such as gastropods and echinoids that scrape the calcareous substrate to remove algal material (Stearn and Scoffin, 1977). Crushers include parrotfish (Frydyl and Stearn, 1978; Tucker et al., 1990). Burrowing is not an important bioerosive process in frame-built reefs, but is important in reef mounds.

20.2.4.6.3 Cementation

Extensive early cementation occurs directly from marine pore waters and greatly influences reef formation, and is largely responsible for the steep, wave-resistant profile of many reefs. It is pervasive in the reef-front and reef-crest zones where high-water flux occurs due to the pumping action by waves (Marshall and Davies, 1981; Tucker et al., 1990) (see Figure 20.12). In reefs with steep profiles, such as the modern walled reef complexes, a large surface area is available to wave and current action; hence, the force of seawater flux is high, and extensive cementation occurs (see the steep-fronted Devonian reefs of the Canning Basin; Playford, 1980; Kerans et al., 1986) (Figure 20.12). The reefs with low-angle profiles, present less of a barrier to the pumping action of waves and currents, and thus undergo much less cementation (Walls and Burrowes, 1985; Tucker et al., 1990).

20.2.4.6.4 Sedimentation

The high degree of biological activity on and around reefs leads to the accumulation of biogenic matter and reef-derived detritus. Such materials are supplied to reefs from three sources: by the mechanical breakdown of framework material by physical or biological processes, by material contributed by the decomposition of reef dwellers, and by the material supplied from outside the reef (Figure 20.12). The most important sediment contributors in modern reefs are *Halimeda*, coralline algae, corals, foraminifera, and mollusks (Figure 20.11) (Milliman, 1974; Tucker et al., 1990). In many ancient reefs, the larger benthic foraminifera are noted to have produced carbonates at rates comparable to those of corals, coralline algae and calcareous green algae, and are thus, important contributors to reef and near-reef sediments (Hallock, 1981; Tucker et al., 1990).

20.2.5 Major Reef Deposits

The types of carbonate materials formed in different reef zones include rudstone, bafflestone, bindstone, and framestone (Embry and Klovan, 1971 modified Dunham's 1962 limestone classification) (see Figure 20.13). Floatstone and rudstone are unbound carbonate grains, more than 10% of which are >2 mm in size; floatstones are mud-supported, and rudstones are grain-supported (see Figure 20.13). Bafflestones are carbonate components bound together at the time of deposition by stalked organisms that trapped sediment by acting as baffles (see Figure 20.13). Bindstones were bound during deposition by encrusting and binding organisms such as encrusting foraminifers and bryozoans, and framestones were bound by organisms such as corals, which build a rigid framework structure (see Figure 20.13).

These carbonate facies also reflect differences in water energy, rates of sedimentation, and the types of organisms. Water energy is highest on the reef crest which also contains the highest percentage of framework constituents (framestones) (see Figure 20.13A). As water energy decreases toward the fore and the back reefs, the percentage of framework constituents also decreases (see Figure 20.13). The non-framework fraction of the reef includes echinoderms, green algae, and mollusks, broken bioclasts (from the reef by wave action), and, in lower-energy zones of the reef, some lime mud. The fore-reef talus slope and back-reef coralgal sand zones are made entirely

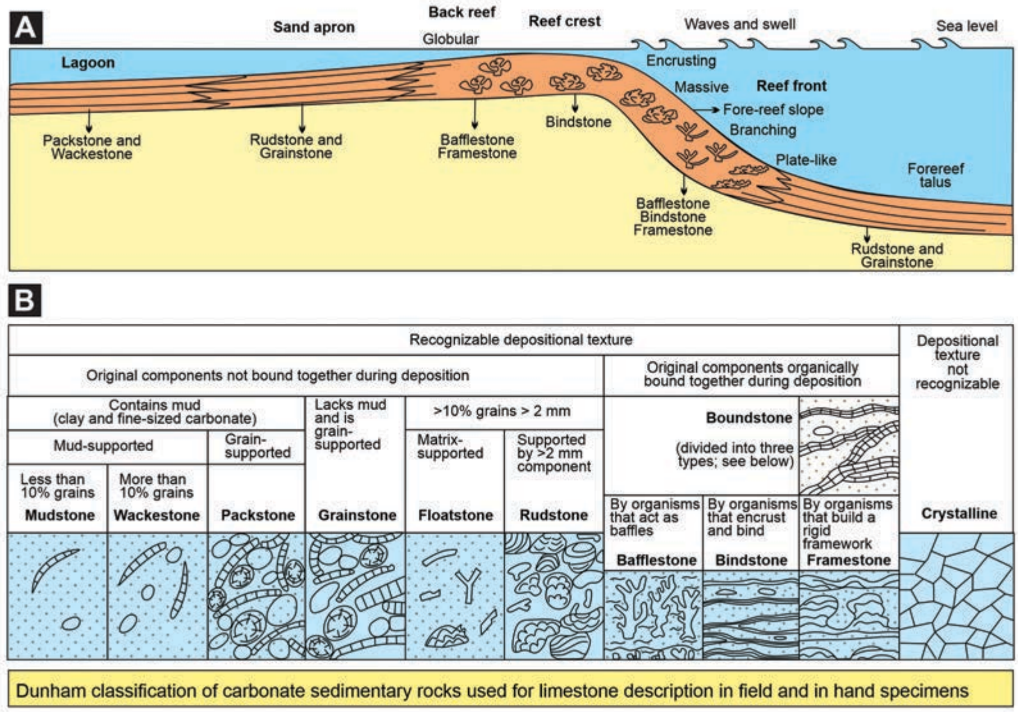

FIGURE 20.13 Reef structure and microfacies. A: Generalized structure of the reef with associated microfacies. B: Dunham's (1962) classification based on depositional-texture vis-à-vis the associated reef microfacies (as shown in A).

of non-framework components and consist principally of reef-derived bioclasts. Relatively few organisms live in these zones. The facies of modern, high-energy, platform margin-type reefs fundamentally consists of a central framework core largely made of corals and coralline algae; the core grades seaward through a zone of rubbly fore-reef talus to deeper-water lime muds or shales, and landwards through back-reef coralgal sands to finer-grained lagoonal deposits (see Figure 20.11A). Low-energy reefs do not develop the characteristic zoning as noted in high-energy reefs. Here, organisms growing on low-energy reefs are dominated by delicate and branching forms (see Figure 20.11B). Other low-energy buildups are composed largely of non-reef-type organisms. They consist of mound-shaped piles of skeletal fragments and/or bioclastic lime muds rich in skeletal organisms and minor amounts of organic boundstone; these structures are called reef mounds or mounds (see James and Bourque, 1992).

20.3 MIXED CARBONATE-SILICICLASTIC SYSTEMS

Mixed carbonate and siliciclastics sediments, also referred to as mixed carbonate-siliciclastic successions or carbonate-clastic transitional successions, are noted in many stratigraphic successions. Carbonate and siliciclastic sediments mix due to lateral facies mixing (due to spatial variations in environments) that produces lateral interfingering of carbonate and clastic sediments. Such carbonate-siliciclastic sediments may also be formed due to change in sea level and/or variations in sediment supply causing vertical facies variations within stratigraphic successions. Thus, siliciclastic facies may occur in a lateral interfingering relationship with carbonate ones or as distinct interbeds within carbonate successions. Carbonate-siliciclastic transitions are known in a variety of environments, such as coastal and inner to outer shelf (including reef), and the slope to basin environments (Lomando and Harris, 1991) or in temperate and tropical shelf environments (Haywick et al., 1992).

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21 Organic Reef Environment

21.1 INTRODUCTION

Organic reef environments (primarily coral reefs found in both shallow and deep marine settings) are unique and diverse ecosystems formed by the accumulation of calcium carbonate ($CaCO_3$) skeletons of marine organisms. Broadly, two types of organic reefs are noted: shallow- and deep-water reefs (see Figure 21.1A). Shallow-water organic reefs, largely built by corals that secrete $CaCO_3$ skeletons, are well-known and commonly studied. The carbonate skeletons accumulate over time, creating a framework that supports a wide variety of other organisms, such as sponges, algae, fish, and other invertebrates. The shallow-water reefs typically occur in tropical and subtropical regions where the water is warm, clear, and nutrient-rich, thus supporting the growth of coral colonies (Figure 21.1A). On the other hand, the deep-water organic reefs, also known as cold-water or mesophotic reefs (see Figure 21.1B), occur in deeper waters beyond the reach of sunlight, and typically noted in cooler temperate or polar regions (Figure 21.1B). These reefs are built by different types of corals and other calcifying organisms that are adapted to low light conditions. In general, the mesophotic reefs are typically found at depths ranging from 30 to 40 m and extending to over 150 m in clear waters (see Loya et al., 2019).

21.2 CARBONATE BUILDUPS

Carbonate buildups are the accumulation and growth of carbonate minerals, such as $CaCO_3$, produced by marine organisms (such as corals, mollusks, and algae), typically in marine or lacustrine (lake) settings. These organisms extract dissolved carbonate ions from the surrounding water and use them to build their skeletons or shells. Over time, the accumulation of these skeletal remains, along with other carbonate precipitates, results in the formation of a carbonate buildup.

Carbonate buildups have various forms such as bioherm, biostrome, reefs, mounds, banks, and platforms. The formation of carbonate buildups is influenced by several factors such as water depth, temperature, salinity, nutrient availability, and sedimentation rates. These factors determine the growth and distribution of carbonate-producing organisms and the conditions for carbonate mineral precipitation. For example, coral reefs typically thrive in warm, clear, and nutrient-rich waters (see Figure 21.1A), while microbial mats contribute to the formation of carbonate buildups in shallow, and hypersaline environments. The types of carbonate buildups are briefly enumerated below.

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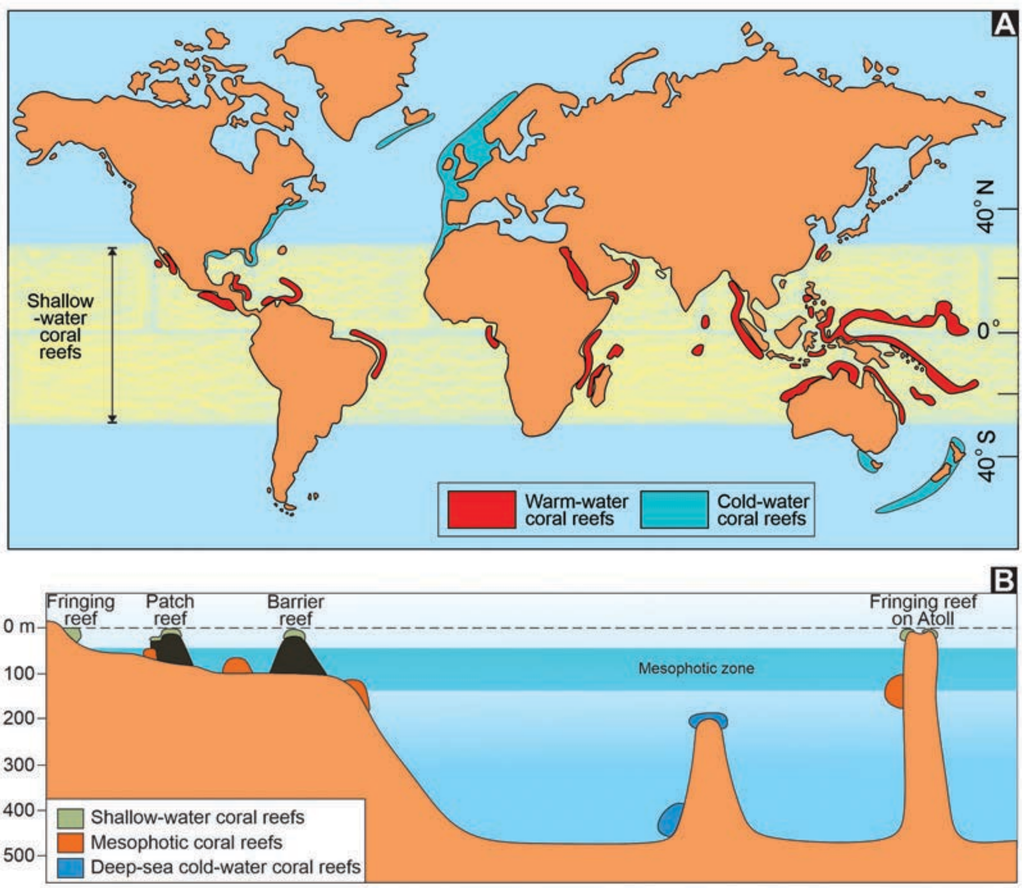

FIGURE 21.1 Distribution of coral reefs.

21.2.1 Bioherms

A bioherm is a type of organic reef or lens-like body that is enclosed in a rock of different lithology or character with or without a rigid internal organic framework (see Figure 21.2A). It is found in both shallow and deep marine waters; the shallow ones are most well-known and well-studied. They typically occur in tropical and subtropical regions where warm, clear, and nutrient-rich waters support the growth of coral colonies that build them. Their skeletons accumulate over time, creating a framework that supports a wide variety of other organisms, including sponges, algae, fish, and other invertebrates.

The deep-water bioherms are also known as cold-water or mesophotic bioherms (Figure 21.1B). They occur in deeper waters beyond the reach of sunlight (Figure 21.1B), and are typically found in cooler temperate or polar regions (Figure 21.1A). They are built by different types of corals and other calcifying organisms adapted to low sunlight conditions. Bioherms play a crucial role in the carbon cycle by sequestering carbon dioxide from the atmosphere and storing it in the form of calcium carbonate.

21.2.2 Biostromes

A biostrome is composed primarily of the remains of marine organisms (such as corals, bryozoans, brachiopods, mollusks, and algae), and is characterized by a dense concentration of skeletal

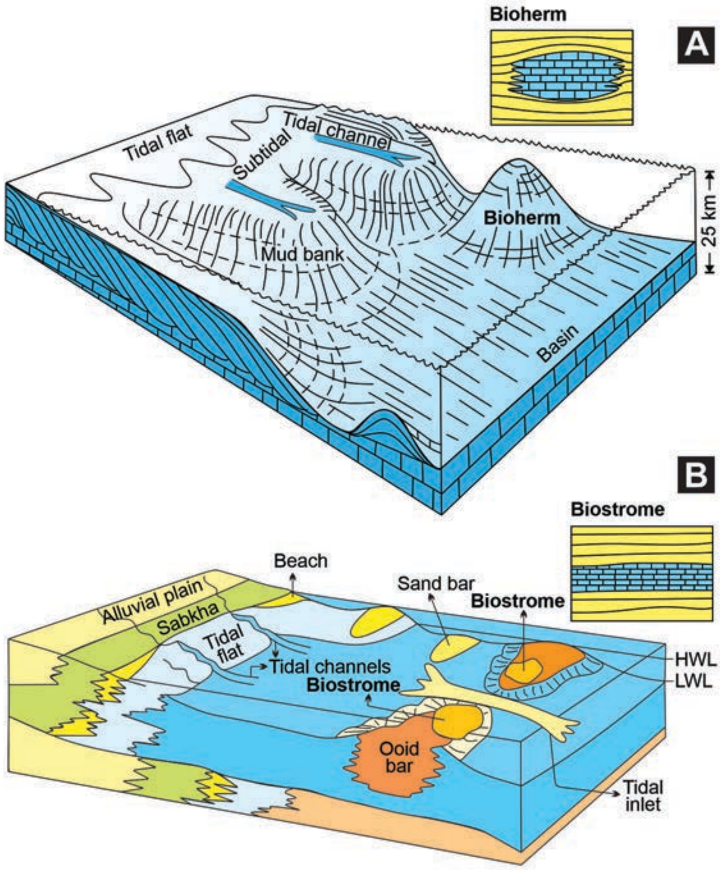

FIGURE 21.2 Bioherm and biostrome. A: Depositional setting of a bioherm. This is a type of organic reef or lens-like body that is enclosed in a rock of different lithology or character with or without a rigid internal organic framework. B: Depositional setting of a biostrome. These are composed of the remains of marine organisms (such as corals, bryozoans, brachiopods, mollusks, and algae), and are characterized by a dense concentration of skeletal fragments or shells that have accumulated and become lithified over time. These tabular bodies of carbonate rocks are typically formed in non-reef platform environments.

fragments or shells that have accumulated and become lithified over time (see Figure 21.2B). These organisms contribute their hard parts (shells, skeletons, or calcareous structures) that accumulate and become cemented together to form the rock. In general, a biostrome is a tabular body of carbonate rock typically formed in non-reef platform environments (Figure 21.2B). The formation of a biostrome is often associated with factors such as clear and warm waters, high nutrient availability, and suitable substrate for attachment and growth. Biostromes also have economic significance, as some of them serve as reservoirs for oil and gas.

21.3 MODERN REEFS AND REEF ENVIRONMENTS

21.3.1 Depositional Setting

The most common depositional setting of a reef is the shallow marine environment, typically in tropical or subtropical (temperate) regions, i.e., 66.5°N to 66.5°S (see Figure 21.1). These environments are characterized by warm waters, high light levels and with relatively low nutrients – optimal conditions for reef-building organisms, such as corals, to thrive. Water clarity (optimal light penetration) and sedimentation rates (the accumulation of $CaCO_3$ skeletons) greatly influence the reef depositional environment. Excessive sedimentation or high nutrient levels inhibit reef growth and lead to the dominance of other organisms, such as algae.

Reefs form in a variety of shallow marine settings, including barrier reefs, fringing reefs, patch reefs, pinnacle reefs, table reefs, reef mounds, microbial mounds, skeletal mounds, and mud mounds (see Figures 21.1B and 21.3). These are briefly enumerated below.

21.3.2 Fringing Reefs

Fringing reefs are one of the main types of reef formations (see Figures 21.1B and 21.3A). They are directly attached to the shoreline or coastline, with no lagoon or significant separation between the reef and the land (Figures 21.1B and 21.3). Fringing reefs typically grow along the shallow continental shelf, extending from the shoreline outwards. They are commonly found in tropical coastal areas. The formation of fringing reefs is influenced by several factors, including the availability of suitable substrate for coral growth, water depth, wave energy, and water quality. Fringing reefs require clear, warm waters with low sedimentation rates and high light penetration for the growth of coral polyps. Fringing reefs often have a distinct zonation pattern, with different coral species dominating different areas of the reef based on their tolerance to water depth, wave energy, and other environmental factors. The shallowest parts of the reef, closest to the shoreline, are typically dominated by branching corals, while the deeper parts may be characterized by massive or encrusting corals.

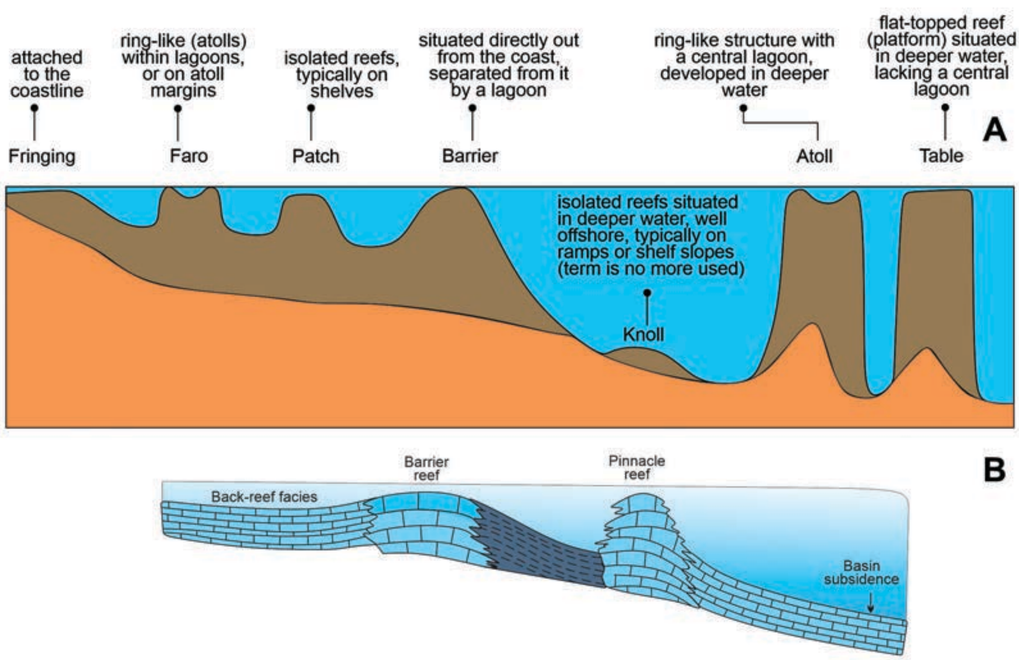

FIGURE 21.3 Reef types. A: Depositional setting of different reef types recorded in shallow marine environments. B: Depositional setting of barrier and pinnacle reefs.

21.3.3 Faro Reefs

This is a special type of small shelf atoll, formed on the rim or interior bank of a composite atoll or barrier reef (see Figure 21.3A).

21.3.4 Patch Reefs

These are typically circular or oval-shaped, small, isolated reef formations that are usually found within a lagoon or on the outer edges of a larger reef system (see Figures 21.1B and 21.3A). They vary in size from a few meters to several hundred meters in diameter. Patch reefs have a shallow, flat top and steep sides (see Figures 21.1B and 21.3A).

21.3.5 Barrier Reefs

Barrier reefs form parallel to the coastline (along platform margins) and are separated from the shoreline by a lagoon (see Figures 21.1B and 21.3) such as the Great Barrier Reef of Australia. They are long and narrow, and extend for many kilometers. Barrier reefs are typically found in tropical or subtropical regions with warm, clear, and nutrient-poor waters. Their formation is closely tied to the subsidence of the underlying seafloor and fluctuations in sea level. As the seafloor subsides, coral colonies grow upwards toward the water surface to maintain access to sunlight and thus, over time, the coral colonies build up and form a continuous barrier parallel to the coastline. The lagoon between the barrier reef and the shoreline is often relatively shallow and contains seagrass, mangroves, or other types of marine habitats with a diverse array of organisms, including corals, sponges, and algae. Barrier reefs also act as natural barriers, protecting the coastline from wave energy and erosion.

21.3.6 Table Reefs

Table reefs are also known as platform reefs or terraced reefs. These are flat, horizontal reef formations that extend parallel to the shoreline (see Figure 21.3A). Table reefs have a relatively uniform top surface, resembling a table or a terrace (see Figure 21.3A). They are typically found in areas with shallow, calm waters.

21.3.7 Pinnacle Reefs

Pinnacle reefs, also known as seamounts or underwater mountains, are tall and narrow reef formations that rise up from the seafloor (Figure 21.3B) (see McLaughlin et al., 2019). They can be found in both shallow and deep waters. They are characterized by their distinct vertical structure, with steep sides and a pointed or rounded top.

21.4 REEF MOUNDS

Reef mounds consist of large, rounded, or conical mounds formed by the accumulation and growth of corals, sponges, and other reef-building organisms over long periods of time. Reef mounds vary in size, ranging from a few meters to several hundred meters in diameter and can rise several meters above the seafloor. They are typically found in deeper waters, often beyond the reach of sunlight needed for photosynthesis. Unlike other reef formations, reef mounds are not directly attached to the seafloor but are instead built on top of previous generations of coral and other organic material. Their growth is influenced by water depth, currents, nutrient availability, and sedimentation rates. These factors influence the shape, size, and composition of reef mounds.

21.4.1 Microbial Mounds

Microbial mounds, also known as microbialites, are formed by the growth and accumulation of microbial communities in aquatic environments. In simple terms, these are microbially derived organosedimentary carbonate structure (Shapiro, 2004; White, 2020) and are composed of layers of microbial mats, which are complex communities of bacteria, archaea, and other microorganisms that interact with each other and the surrounding environment. Microbial mounds are found in a variety of aquatic habitats, including freshwater lakes, marine environments, and even in hot springs. They vary in size and shape, ranging from small, centimeter-scale mounds to large, meter-scale structures. Some microbial mounds have distinct shapes, such as cones, domes, or columns. In general, the microbialitic fabric is of four common types: laminated or stromatolitic, clotted or thrombolitic, branching or dendrolitic, and undifferentiated or leiolitic (Figure 21.4) (see also Shapiro, 2004; Riding 2011). The formation of microbialites is influenced by various factors, including water chemistry, temperature, nutrient availability, and water flow.

21.4.2 Skeletal Mounds

Skeletal mounds are large structures formed by the accumulation of skeletal remains of marine organisms, such as corals, sponges, and bivalves in both shallow and deep-sea environments. Their formation begins with the growth and accumulation of skeletal material produced by reef-building organisms. Over time, these skeletal remains accumulate and cement together, forming a solid structure. The mounds continue to grow vertically as new generations of organisms settle and build upon the existing structure.

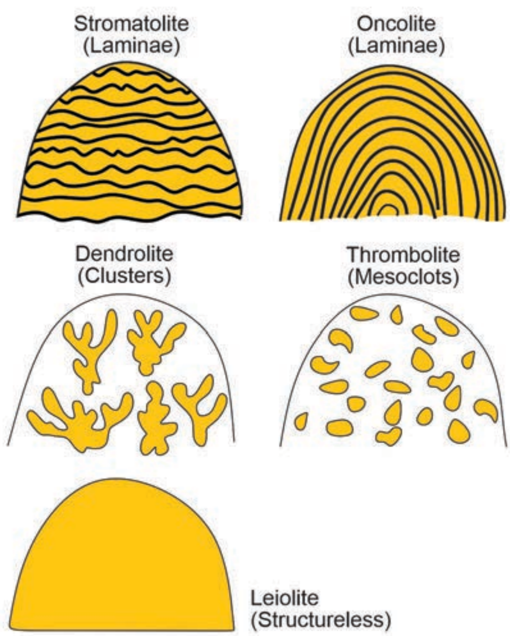

FIGURE 21.4 Major categories of microbialites. (Modified from Shapiro, 2004; Riding, 2011.)

21.5 REEF ORGANISMS

Coral reefs are diverse underwater ecosystems formed by the accumulation of calcium carbonate skeletons secreted largely by corals (Figure 21.5); corals are colonial organisms belonging to the Class Anthozoa. Reef-building corals are the primary organisms responsible for the construction of coral reefs. These corals have a symbiotic relationship with photosynthetic algae called zooxanthellae. The corals provide a protected environment and nutrients to the zooxanthellae, while the algae provide oxygen and organic compounds through photosynthesis. This symbiotic relationship allows corals to thrive in nutrient-poor waters and build their calcium carbonate skeletons.

In modern reefs, corals are the dominant constituents; they are of two types – hermatypic and ahermatypic. Hermatypic (zoanthellae) hexacorals abound in shallow-water reefs. They have a symbiotic relationship with several types of unicellular organisms, mainly algae, collectively referred to as zooxanfhellae. These algae live in or between the living cells of the corals and provide energy via

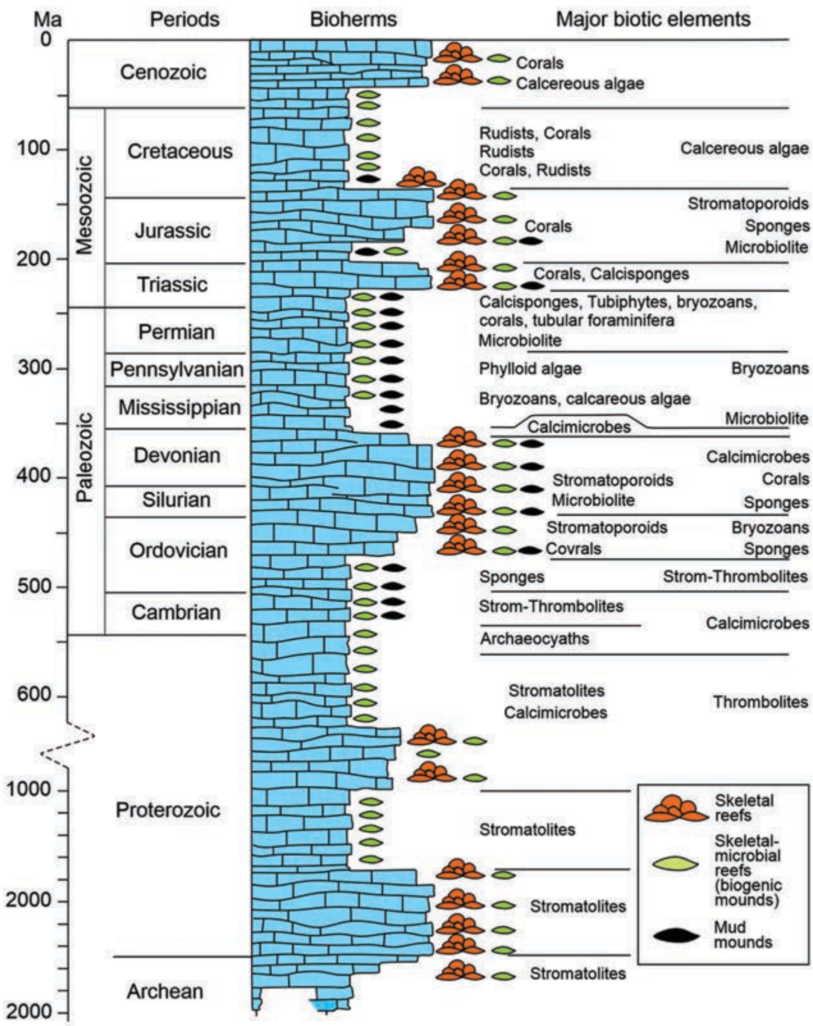

FIGURE 21.5 Major carbonate contributors through time. (Modified after James, 1984; Wood, 2011.)

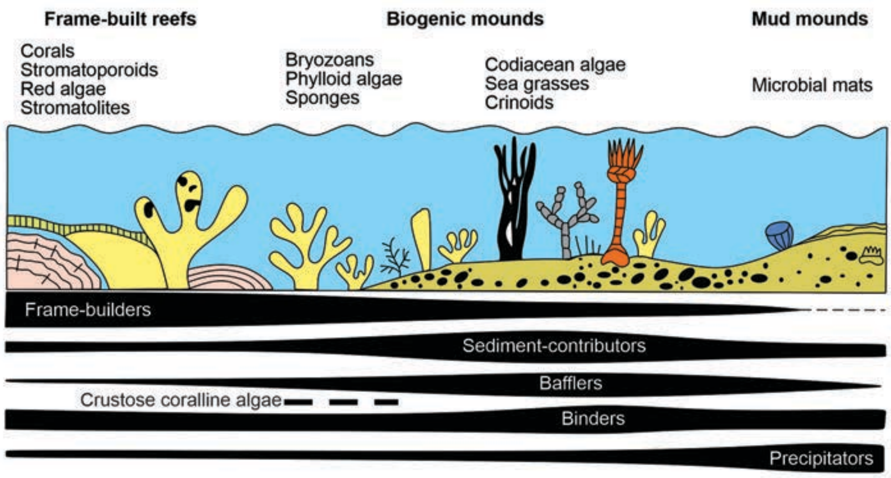

FIGURE 21.6 Coralline-crinoideal-reef-mounds and their reef-building fauna. These associations are noted in the inner seafloor of Florida (USA) and Shark Bay (Australia), Halimeda Mounds in the Indonesian Archipelago, and the Yucatan Sea (Mexico). (Modified after Foubert and Henriet, 2009.)

photosynthesis. They also facilitate the secretion of $CaCO_3$ by removing CO_2 from the tissues during photosynthesis. As zooxanthellae require sunlight, hermatypic corals are restricted to living in very shallow waters. Ahermatypic (azooxanthellae) corals, on the other hand, lack a symbiotic relationship (or do not require it) and thus are not restricted to shallow water (see Willison et al., 2001). Hence, their distribution ranges from shallow waters to depths >2000 m (Stanley and Cairns, 1988).

The reefs are synonymous with coral reefs, but many other organisms contribute to the formation of reefs such as the blue-green algae (cyanobacteria), coralline red algae, green algae, foraminifera, bryozoa, sponges, mollusks, echinoderms, sponges, sea fans, sea anemones, and other invertebrates and vertebrates (such as reef fish). Sponges filter water and provide habitats for other organisms, while sea fans and sea anemones contribute to the structural complexity of the reef. Algae (macro- and microalgae) are important primary producers, thus providing food and energy to other organisms.

In the geologic past, reef-building organisms included now-extinct groups such as the archaeocyathids, stromatoporoids, fenestellid bryozoans, and rudistid clams (see Figure 21.5). Some reef-building organisms such as corals and stromatoporoids are important frame-builders that construct wave-resistant cores of reefs (see Figure 21.6). Others, such as crinoids and codiacian algae (such as *Halimeda*), whose skeletal elements disintegrate into smaller fragments, are important sediment contributors (Figure 21.6). Bafflers such as seagrass provide a protective baffle against currents and thus generate a localized, low-energy environment in which fine sediment accumulate (Figure 21.6). Binders such as cyanobacteria (which form stromatolites) trap and bind sediments (Figure 21.6). Precipitators are primarily microbes, such as cyanobacteria, that help mediate precipitation of carbonate muds (Figure 21.6). In general, the growth forms of reef-building organisms range from delicate branching forms to globular or massive structures (see Figure 21.7). Organisms that live in low-energy zones often have delicate branching or plate-like forms while those living in higher energy zones develop hemispherical, encrusting, or tabular forms that better withstand strong wave actions (Figure 21.7).

21.6 REEF DEPOSITS

Reef deposits are primarily composed of the accumulated remains of reef-building organisms, such as corals, sponges, and algae in shallow, warm marine environments, typically in tropical or subtropical regions. The process of reef formation begins with the growth of these reef-building organisms that secrete $CaCO_3$ skeletons. Over time, these accumulate and form a framework or structure that provides habitat and support for a diverse array of organisms.

In general, fossil reefs differ in some respects from modern reefs. First is the nature of exposure; in fossil reefs, mostly either the core is available for analysis or part of the reef flank, and the reef facies is exposed in outcrops (i.e., exposing only a cross section of the reef). Secondly, and more importantly, the type of the dominant organism that formed the reef – the hermatypic corals dominate modern coral reefs but these only appeared in the Mesozoic and thus are not components of older reefs. Older reefs are dominated by other kinds of organisms, such as tubiphytes (problematic microstructures of possible algal origin), sponges, bryozoans, and microbes (such as cyanobacteria) (see also Fagerstrom, 1987; Stanley and Fagerstrom, 1988; Kiessling et al., 1999, 2002). Thirdly, the reef structure and nature of reef facies also depend upon the kind of reef available for study (such as barrier reef, fringing reef, or patch reef). Lastly, ancient reef facies undergo diagenesis, causing selective dolomitization or solution that either destroys parts of the reef complex or completely obliterates it.

In general, the reefs are divided into fore-reef, reef-front, reef-crest, reef-flat, and back-reef zones (see Figure 21.7). The reef consists of a central core, the reef framework that grades seaward into the reef slope, and a loose accumulation, of reef debris called the fore-reef talus (Figure 21.7). The nearly flat, uppermost, shallowest part of the reef is called the reef flat that grades landward into back-reef skeletal (coralgal) sands and subtidal lagoonal deposits (Figure 21.7). The core framework of skeletal bindstones and rudstones grades seaward into the fore-reef facies, consisting largely of rudstones and floatstones, broken from the framework by wave action (see Figure 21.7). The reef core grades landward into a high-energy, skeletal sand apron consisting largely of material broken off the reef core (see Figure 21.7). In turn, skeletal sands grade into a back-reef facies deposited under quieter water conditions containing skeletal sands and muds, pelleted muds, and micrites, and stromatolites (see Figure 21.7).

As water energy decreases toward both the fore reef and the back reef, the percentage of framework constituents (such as corals; bioclasts) also decreases (see Figures 21.6 and 21.7). The non-framework fraction of reefs consists of organisms such as echinoderms, green algae, mollusks (that do not build framework structures), bioclasts (broken from the reef by wave activity), and, some lime mud in lower-energy zones of the reef (Figures 21.6 and 21.7). The fore-reef talus slope and

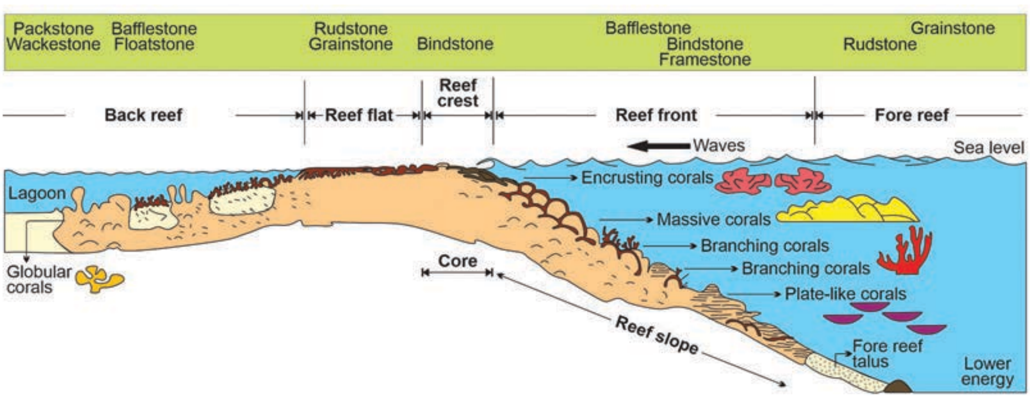

FIGURE 21.7 Reef structure and associated organisms (mainly corals). The growth forms of reef-building organisms range from delicate branching forms to globular or massive structures. Organisms that live in low-energy conditions possess delicate branching or plate-like forms while those inhabiting higher-energy conditions develop hemispherical, encrusting, or tabular forms that withstand strong wave action.

back-reef coralgal sand zone are entirely made up of non-framework constituents that consist principally of reef-derived bioclasts (see Figure 21.7). Evaporites may also be present in some back-reef facies, reflecting restricted water circulation in the back-reef environment. Reef growth eventually terminates due to the drowning of the reef or possibly, in some cases, due to burial by siliciclastic muds. The noted carbonate facies represent variations in water energy, dominant sedimentation processes, and types of organisms in each zone of the reef (Figures 21.6 and 21.7). Water energy is highest on the reef crest, which also contains the highest percentage of framework constituents (framestones) (Figure 21.7).

21.6.1 Low-Energy Reef Facies

The facies of modern, high-energy, platform margin-type reefs consists of a central framework core composed largely of corals and coralline algae; the core grades seaward through a zone of rubbly reef talus to deeper water lime muds or shales and landward through back-reef coralgal sands to finer grained lagoonal deposits (Figure 21.7).

The low-energy reef facies are characterized by relatively calm water conditions and minimal wave or current energy, typically noted in protected areas of the reef, such as lagoons, or back reefs (see Figure 21.7). This facies is characterized by the presence of fine-grained sediments, including mud, silt, and sand, usually deposited in areas where water movement is limited, thus allowing for the finer particles to settle and accumulate. The lack of strong wave action or currents in these environments prevents the transportation and reworking of sediments, leading to their improved preservation. Within the low-energy reef facies, various organisms and sedimentary structures are noted such as sea grass meadows in shallow, calm waters. These meadows contribute to the accumulation of organic matter and the formation of muddy sediments. In addition, microbial mats or algal mats also form in low-energy reef facies. These mats are composed of layers of microorganisms, such as cyanobacteria, that trap and bind sediments, creating laminated structures (stromatolites) (see Figure 21.4). Low-energy reefs are also characterized by more delicate and branching forms (see Figure 21.7). Some low-energy reefs are made of carbonate sands and muds built by organisms that are very similar in composition to reef-type organisms (James, 1984a, 1984b). Other low-energy buildups are composed largely of non-reef-type organisms consisting of mound-shaped piles of skeletal fragments and/or bioclastic lime muds rich in skeletal organisms with minor amounts of organic boundstone (see Figure 21.7); these are called reef mounds (James and Bourque, 1992).

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