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Sedimentary Geology 186 (2006) 67–88
Sedimentary Geology
www.elsevier.com/locate/sedgeo

Soft-sediment deformation in a tectonically active area: The Plio-Pleistocene Zarzal Formation in the Cauca Valley (Western Colombia)

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Received 8 February 2005; received in revised form 5 October 2005; accepted 11 October 2005

Abstract

The Plio-Pleistocene Zarzal Formation corresponds to fluvio-lacustrine sediments deposited in an intramontane depression within the Colombian Andes, associated with the Cauca-Romeral Fault System. It crops out mainly in the Cauca Valley where numerous field sections have permitted the mapping of the vertical and lateral lithological variations. Lacustrine deposits of sands, silts, clays and diatomites are interbedded with fluvial sand and gravel beds and fluvio-volcanic mass flows derived from the volcanic Central Cordillera.

Numerous soft-sediment deformation structures are encountered in this formation, particularly in fine- to medium-grained sands, silts and clays: load structures (load casts, flame structures, pseudonodules), water escape structures (water escape cusps, dish-and-pillar and pocket-and-pillar structures), soft-sediment intrusions (clastic sills and dykes), disturbed laminites, convolute laminations, slumps and synsedimentary faulting.

Deformation mechanisms and driving forces are related essentially to gravitational instabilities, dewatering, liquidization and brittle deformations. Field and regional geological data show that most of these deformations are related to seismicity and can be interpreted as seismites. This area has a geological and recent seismic history and outcrops show both syn- and post-depositional faulting related to the transpressional regime of this part of the Colombian Andes, which generates strike-slip faults and associated local normal faults. The drainage pattern within the Zarzal Formation shows the signature of neotectonics. Moreover, the fine to coarse-grained sands of the Zarzal Formation are lithologies prone to liquefaction when affected by seismic waves. The intercalation of the deformed intervals within undisturbed strata confirms the catastrophic nature of the events. Finally, the large areal extent of the deformations and the type of structures are compatible with seismites.

Consequently, the existence of seismites in the Zarzal Formation represents corroboration of tectonic activity in this area during the Pleistocene. Earthquakes with a magnitude higher than 5 can be postulated, based upon the proximity of active faults and the types of deformations.

Keywords: Cauca-Romeral Fault System; Dewatering; Injection; Liquidization; Load structures; Seismites; Synsedimentary faulting

1. Introduction

The Colombian Andes comprise three SSW–NNE trending, sub-parallel cordilleras (Fig. 1). The com-

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doi:10.1016/j.sedgeo.2005.10.009

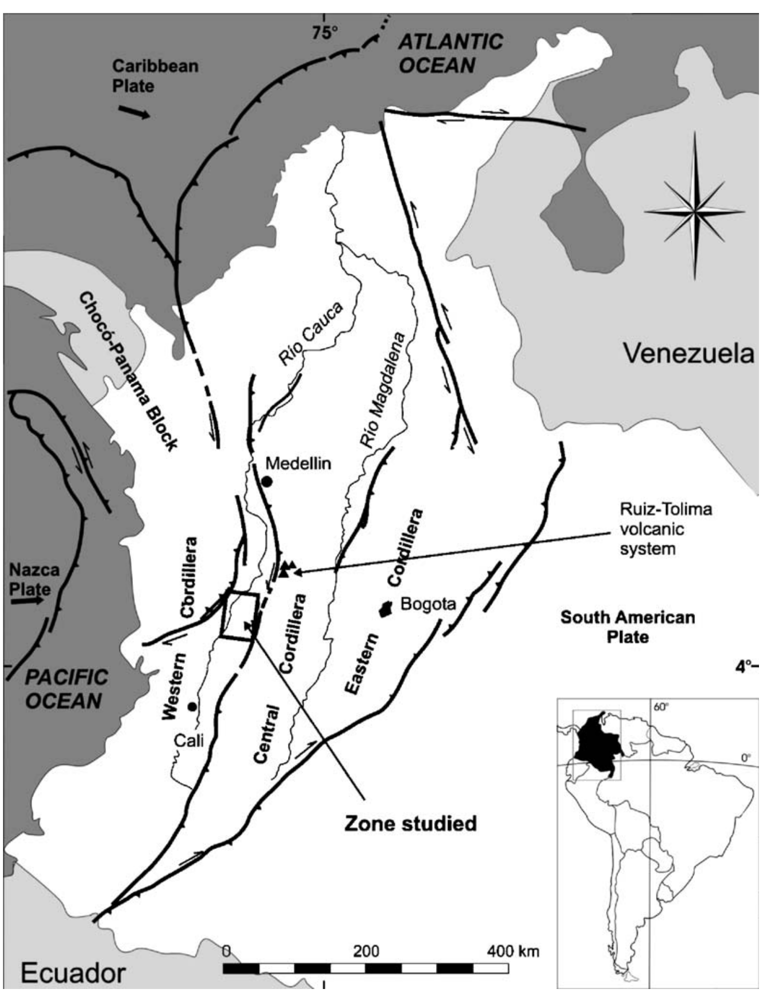

Fig. 1. Megatectonic framework and location of study area.

plex tectonic framework of this area results from the interaction between three tectonic plates (Nazca, Caribbean and South American) and the Panama–Costa Rica microplate (Pennington, 1981; Taboada et al., 2000). This interaction induced a compressive tectonic regime in the north-western part of Colombia, particularly in the studied area located between the Western and Central Cordilleras, north of the city of Cali (Fig. 1).

The studied area covers parts of three Colombian departments (Fig. 2): Quindío (city of Armenia), Risaralda (city of Pereira) and Valle del Cauca (cities of Cartago and Zarzal). This zone comprises two intramontane depressions where Plio-Pleistocene sediments were deposited: the Cauca Valley to the west displays numerous outcrops of the Pleistocene Zarzal Formation between the towns of Zarzal and Cartago. It is separated by folded Tertiary outcrops from the Quindío–Pereira depression to the east where the Plio-Pleistocene fluvio-volcanic fans of Quindío, Pereira and Cartago accumulated. These two sedimentary units interfinger in the western part of the Cartago, Pereira and Quindío Fans. This tectonically active zone is dissected by several major SSW–NNE trending fault lineaments such as the Romeral Fault System (Fig. 3). In particular, the rupture at shallow depth of the Armenia fault caused the dramatic earthquake of Armenia in January 1999 (INGEOMINAS (Instituto nacional de investigaciones geológico-mineras), 1999). The Ruiz-Tolima volcanic system in the Central Cordillera is associated with these faults and formed the source of the fluvio-volcanic fans deposited to the west (Fig. 2).

The Zarzal Formation encountered in the Cauca Valley was deposited in a fluvio-lacustrine environment. It contains numerous soft-sediment deformation structures, particularly in the Cartago area (Fig. 4). The aims of this paper are to describe the various types of

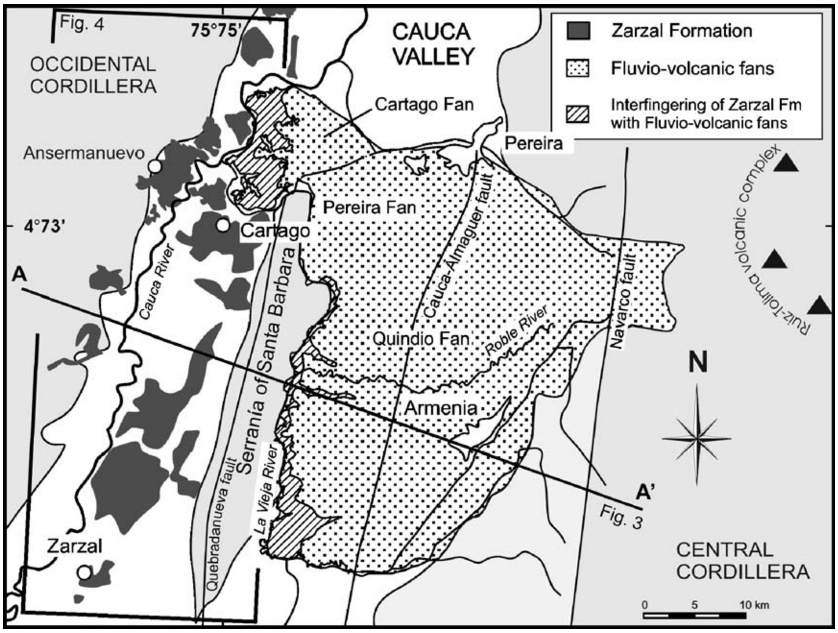

Fig. 2. Regional distribution of Plio-Pleistocene sediments in the Valle del Cauca, Risaralda and Quindío Departments. Location of Figs. 3 and 4.

deformations encountered and discuss their potential triggering mechanisms.

2. Geology of Zarzal Formation

The Zarzal Formation has been so far poorly studied. Boussingault (1903) was the first to describe siliceous deposits intercalated with sand and sandy clay beds in the Cartago area, where they form low-relief hills. He considered these sediments as the infill of a lake. The name Zarzal Formation was attributed in 1955 to these deposits of diatomite, clay and volcanic sand (Keiser, Nelson and Van der Hammen, unpublished, in Van der Hammen, 1958; see also De Porta, 1974).

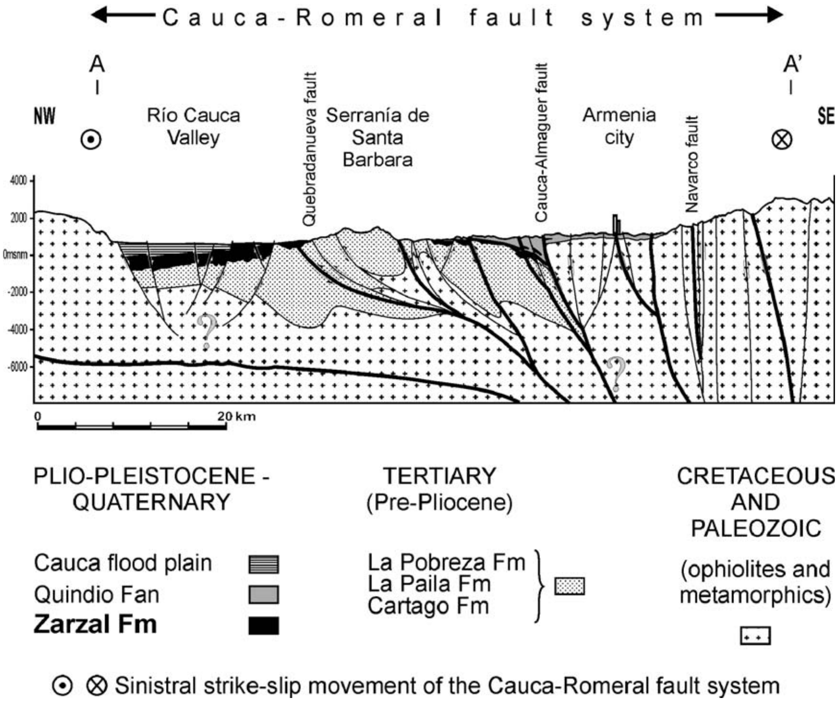

Fig. 3. Geological cross-section across the Valle del Cauca and Quindío Departments. See Fig. 2 for location.

Cardona and Ortiz (1994) were the first to analyze in detail the depositional environment of these sediments. They interpreted three types of facies: braided-stream channel deposits, floodplain sediments and lake deposits. They also noted the geomorphological expression of this unit: low-relief hills dissected by a well-marked

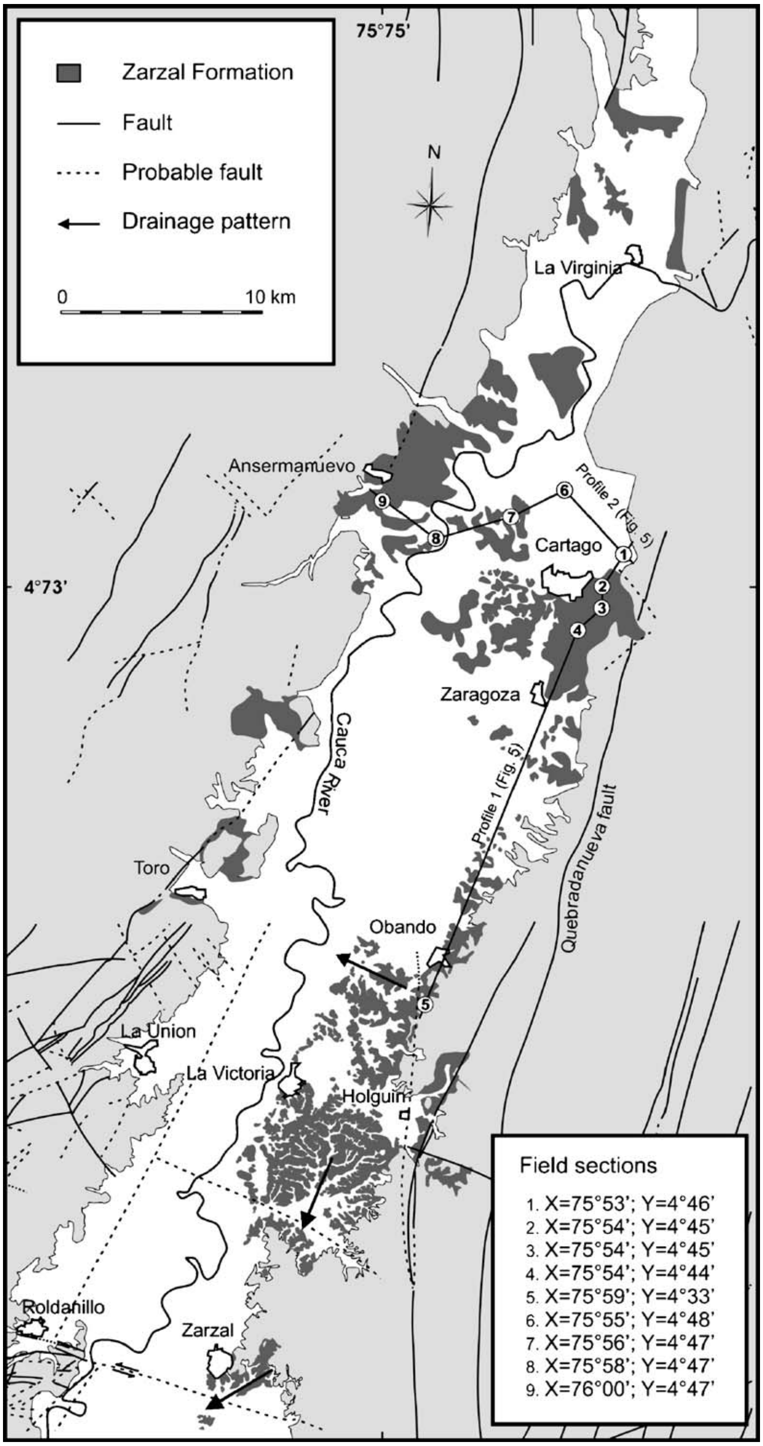

Fig. 4. Detailed geological and location map of study area. See Fig. 2 for location.

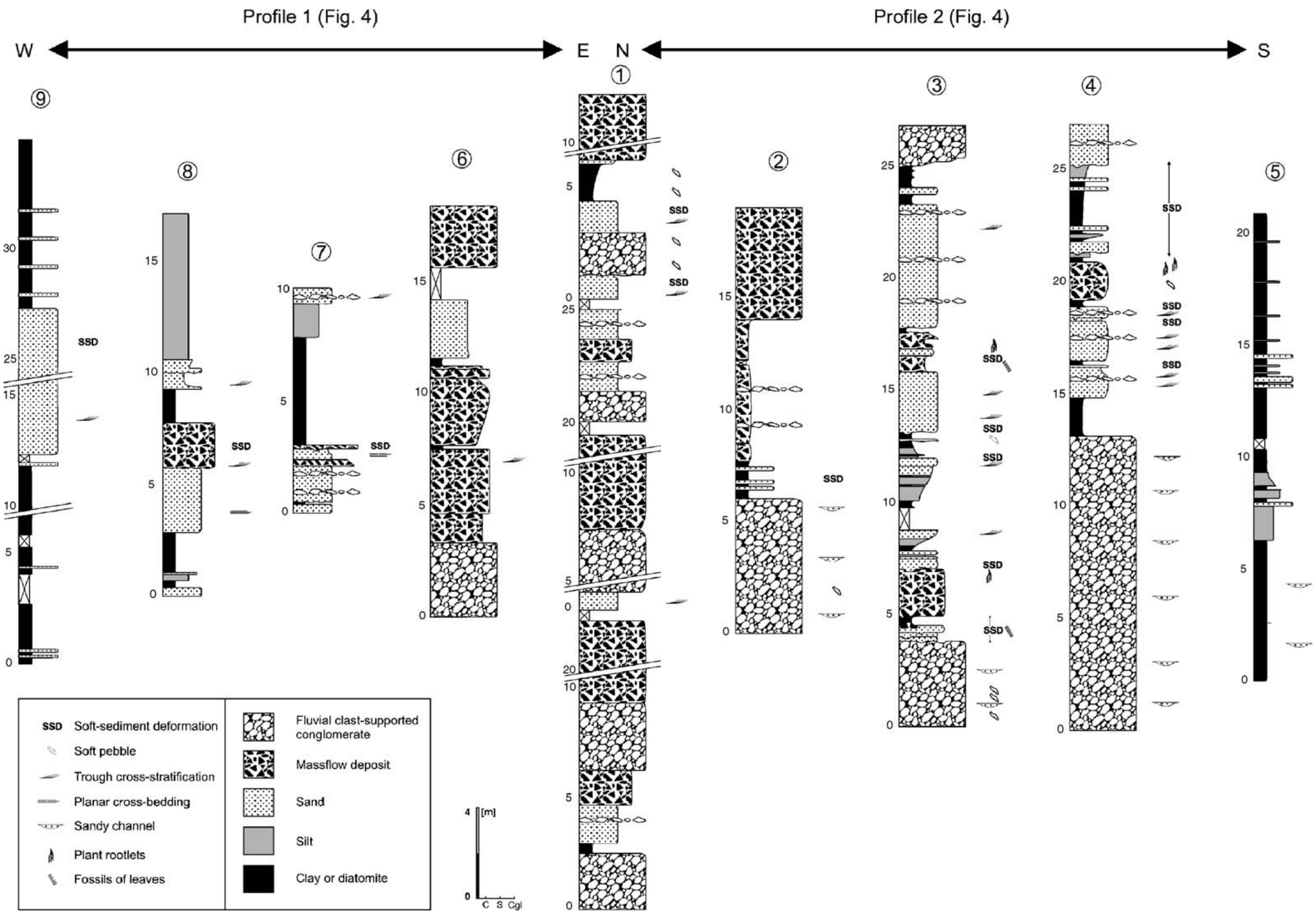

Fig. 5. E–W and N–S trending correlations of field sections in the Zarzal Formation. See Fig. 4 for location of field sections.

drainage pattern and numerous surface fractures. They recognized traces of block tectonics younger than the Pliocene Andine orogeny and also described the interdigitation of the Zarzal Formation with the fluvio-volcanic fans of Cartago and Pereira to the east.

So far, no precise age dating has been achieved for the Zarzal Formation. According to Van der Hammen (1958), it assumes a probable Pliocene age for the Zarzal Formation without any scientific evidence. The Zarzal Formation disconformably overlies the Miocene La Paila Formation on the western flank of the Serranía de Santa Barbara (Fig. 2; McCourt, 1984; Nivia et al., 1995). In the Cauca Valley, the Zarzal Formation is disconformably overlain by gravels of alluvial fans fed by the surrounding reliefs, by grey palaeosols rich in volcanic materials and by recent alluvial sediments (Nivia et al., 1995). In fact, prior to the present research, the only certainty that exists is that the Zarzal Formation postdates the Pliocene Andine compression, i.e. it has a late Pliocene to Pleistocene age.

The dramatic earthquake of Armenia in January 1999 prompted the detailed geological study of the Plio-Pleistocene deposits in the Quindío Department, especially of the fluvio-volcanic fans and their relation with the Zarzal Formation in Quindío and the Cauca Valley (Figs. 2 and 3; Guarin, 2002; Guarin et al., 2004; Gorin et al., 2003; Suter, 2003; Suter et al., 2003). Detailed field studies confirm the observations of Cardona and Ortiz (1994). The Zarzal Formation consists of autochthonous lacustrine sediments (diatomites) with a variable degree of interfingering with volcanic, fluvio-volcanic and fluviatile influxes derived from surrounding sources. In the Cauca Valley, where numerous sections can be observed (Figs. 4 and 5), the degree of interfingering increases towards the east and ongoing studies show that the Cartago Fan (Fig. 2) probably supplied most of the allochthonous material. At the western edge of the Cartago Fan, volcanic mass flows are intercalated with Zarzal fluvio-lacustrine sediments and can be observed westwards up to the Ansermanuevo area (Figs. 4 and 5). Moreover, thin intercalations of volcanic ash are encountered in the diatomites. In the La Vieja and Roble valleys, east of the Serrania of Santa Barbara (Fig. 2), lacustrine deposits of the Zarzal Formation can also be observed underlying fluvio-volcanic sediments of the Quindío Fan (Suter, 2003). Preliminary palynological data from clays of the Zarzal Formation show a significant presence of *Alnus* pollen. Because the first record of this tree in Colombia dates back to 1 my (Hooghiemstra and Cleef, 1995), a large part of the Zarzal Formation is probably of Pleistocene age. These data are awaiting confirmation from ongoing radiometric dating in volcanic ashes.

Field work and aerial photographs have confirmed the evidence of recent tectonic activity, particularly the diverging drainage pattern of Zarzal outcrops in the Cauca Valley (Fig. 4). Most of the soft-sediment deformation structures described below are within the sections shown in Fig. 5.

3. Overview of soft-sediment deformation and classifications

Although different authors have proposed classifications, no universally accepted scheme exists for various reasons. First of all, the merely descriptive classifications do not take into account the processes which formed the structures. On the other hand, the genetic classifications are based on assumptions about the inferred processes and parameters that have acted during the deformation. In many cases, the relationships are not clear and their application to field work is difficult. Although some classifications have been proposed in order to link morphology and genetic processes, they do not answer all questions. Another difficulty is the large variety of descriptive terms. Although this paper does not aim at elaborating a new classification, a review of some proposed classifications is necessary, in order to clarify the terminology that will be used (Table 1). It refers mainly to Lowe (1975), Brenchley and Newall (1977), Mills (1983) and Owen (1987, 2003).

Lowe (1975) proposed a classification for water escapes structures. It is quite comprehensive but confusion may exist for some structures. Nevertheless, Lowe's classification has been used in recent papers (Rossetti, 1999). Brenchley and Newall (1977) proposed a classification for contorted bedding, which has been updated by Mills (1983). Although the diagnostic features are relatively easy to establish, the direction of movement is difficult to infer. Moreover, it does not include soft-sediment structures such as dykes and water escape structures. More recently, Owen (1987, 2003) proposed two attractive classifications encompassing respectively all soft-sediment deformation structures and only load structures. That of 1987 intends to establish a practical scheme taking into account genetic aspects. It can be considered as an extended version of the classifications proposed by Elliott (1965) and Allen (1982). Owen (1987) distinguishes three types of deformation mechanisms and five classes of driving forces. Simultaneously, based upon the same criteria, he suggests a classification that can be applied

Table 1
Comparison between some classifications of soft-sediment deformation (SSD) structures showing nomenclatures and classification criteria

	Lowe (1975)	Brenchley and Newall (1977)	Owen (1987)	Owen (2003)
Type of classification	Classification of consolidation-related SSD	Classification of contorted bedding	Expanded classification of SSD and working classification for SSD	Classification of load structures
Classification criteria	Morphology, solid–liquid kinematics and origin of stresses causing flowage	Morphology and dissection of movement	Deformation mechanisms and driving force	Morphology
Proposed nomenclature	Convolute lamination	Convolute lamination	Convolute lamination	Convolute lamination
	Load structures	Load casts	Load casts	Simple load casts
		Asymmetric load casts		Pedulous load Cast
			Pseudonodules	Attached pseudon
				Detached pseudon
		Ball and pillow	Ball and pillow	Ball and pillow
	Heave structures			
		Slumps	Slumps	
	Dykes		Clastic dykes	
	Sills			
			Sand volcanoes	
			Dish structures	Water-escape cusps
			Cusp	
	Hydroplastic mixing layers			
	Liquefaction layers and pockets			
	Fluidization channels and layers			
		Sand rolls	Recumbent folds	Wrinkle marks
		Multilayer folds		Contorted heavy mineral laminae
	Deformed cross-bedding	Recumbently folded cross-bedding		Preglacial involutions

principally to deformation in sands. This genetic classification is very broad and covers all of the commonly encountered soft-sediment deformation structures. The second classification (Owen, 2003) is exclusively designed for load structures. It differentiates between load casts (including flame structures) and pseudonodules, which can be further subdivided. The different kinds of load structures are linked to a deformation series.

This short overview of classifications illustrates that they are either incomplete, difficult to apply in the field or include overlapping definitions. Consequently, most authors have grouped structures into morphological categories (e.g., Rossetti, 1999; Rossetti and Goes, 2000; Rodríguez-Pascua et al., 2000) or into a mixture of morphological and genetic categories (e.g., Alfaro et al., 1997).

In the Zarzal Formation, most of the soft-sediment deformation structures reported by Lowe (1975), Mills (1983) and Owen (1987, 2003) can be observed. However, some complicated structures could not be related precisely to a classification. Consequently, the structures will first be described morphologically and grouped into four groups, further subdivided into 14 categories (Fig. 6). Subsequently, the processes potentially associated with their genesis will be discussed.

4. Soft-sediment deformations in the Zarzal Formation

Soft-sediment deformation structures in the Zarzal Formation are encountered mainly in the region between Cartago and Ansermanuevo (Fig. 4). Although the most frequently deformed lithologies are fine to medium-grained tuffaceous sands and clays, deformations are present in other lithologies, including diatomites and tuffaceous gravels. The following types of structures have been observed (Fig. 6).

4.1. Load structures

4.1.1. Load casts

The classification used here follows essentially that proposed by Owen (2003). Load casts are the most frequent structure encountered in the studied area. Although the term “cast” is inadequate, because the structure cannot really be associated with a cast, the term is

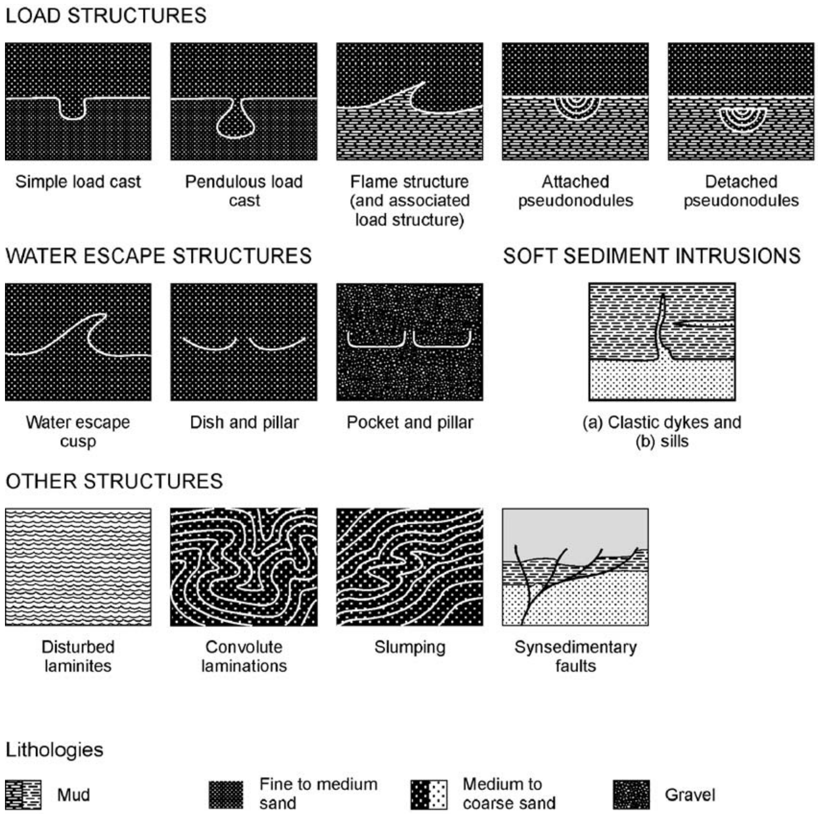

Fig. 6. Types of soft-sediment deformation structures observed in the Zarzal Formation.

kept in order to avoid confusion in the nomenclature. Both categories proposed by Owen (2003) have been encountered.

4.1.1.1. Simple load casts. Their size varies from centimeters to meter. They occur in different lithologies, but mostly in tuffaceous sands and gravels overlying silty clays and diatomites (Fig. 7A). They show a concave profile and slightly penetrate into the underlying bed. Laminations are usually gently deformed, although in some cases this structure appears associated with more pronounced deformation, such as convolute lamination and water escape structures (Fig. 7A). It is similar to the “sagging load cast” of Alfaro et al. (1997).

4.1.1.2. Pendulous load casts. Their size fluctuates from a few up to 50 cm (Fig. 7B). They occur in various lithologies but, similarly to the simple load casts, they are more frequently encountered in tuffaceous sands and gravels overlying clays and diatomites. Occasionally, such deformations may be observed in fine-grained sands overlying medium to coarse-grained sands. Their shape is very variable: generally they become narrower upwards (Fig. 7B), but this is not always the case (Figs. 7C and 8A). They show features similar to the “drop structures” of Anketell et al. (1970). Their base is generally sub-planar and resembles convex downward lobe, the thickness of which shows frequent lateral variations. The internal laminations are generally slightly deformed, but can also locally be strongly deformed and associated with convolute lamination and water escape structures.

4.1.2. Flame structures

Their size varies from centimeters to decimeters (Fig. 8A). In most cases, this structure is poorly developed and only rarely does it correspond to so-called “mud diapirs” (Owen, 2003). This structure affects only clays and diatomites. They appear always in association with load structures (Fig. 6).

4.1.3. Pseudonodules

This structure is less frequently encountered than load casts. Of the three types described by Owen (2003), the most frequently observed is the attached pseudonodule. Detached pseudonodules have been observed only locally. They comprise a single row of

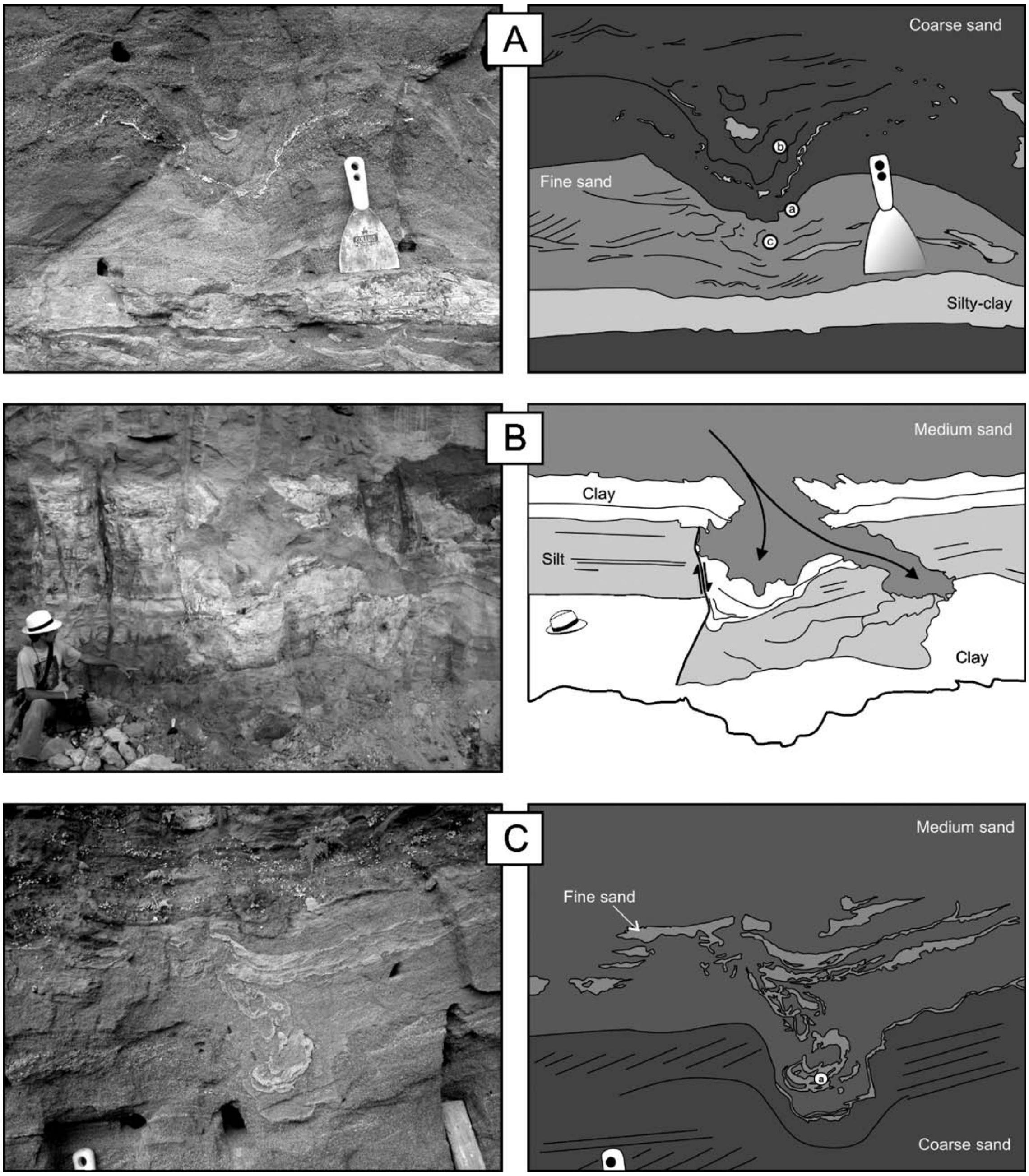

Fig. 7. Load structures. (A) Simple load cast (a) (section 8, see Figs. 4 and 5 for location) associated with convolute laminations (b) and water escape structure (c). (B) Pendulous load cast (section 8, see Figs. 4 and 5 for location). This structure is associated with a subvertical synsedimentary fault. Part of the deformed fine sand–clay is liquefied, probably as a result of slumping. (C) Pendulous load cast (section 8, see Figs. 4 and 5 for location) showing internal deformations (a) associated with gravity loading.

pseudonodules overlain by matrix, whereas ball-and-pillow structures correspond to vertically stacked pseudo-nodules. The latter have not been encountered.

4.1.3.1. Attached pseudonodules. Their size varies between 10 and 30 cm. They occur in medium and coarse-grained tuffaceous sands overlying fine and medium-grained tuffaceous sands and, sometimes, clays and diatomites. Locally, medium-grained sands overlying coarse-grained sands can be observed (Fig. 8B). Their shape is variable but generally corresponds to a concave profile, sometimes slightly deformed (Fig.

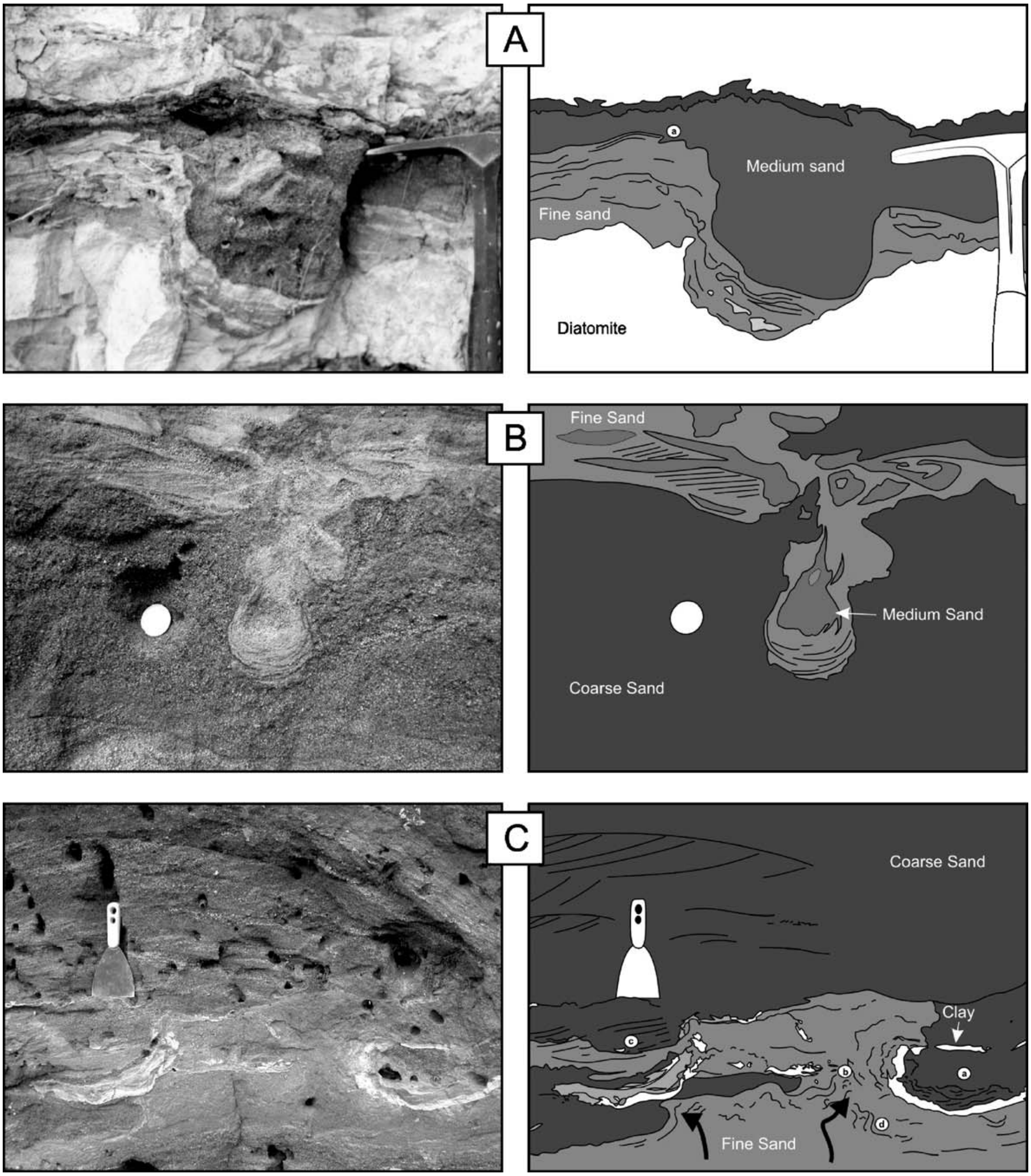

Fig. 8. Load structures. (A) Pendulous load cast (="drop structure", Alfaro et al., 1997), near section 8 (see Figs. 4 and 5 for location). It forms a pocket of medium-grained sands overlying deformed, finely laminated sands. The upper part of the latter is affected by flame structures (a). (B) Attached pseudonodule made of fine-medium-grained sands, which sank into coarse-grained sands (section 8, see Figs. 4 and 5 for location). (C) Attached pseudonodule (a), water escape structure (b), simple load cast (c), convolute lamination (d) (section 8, see Figs. 4 and 5 for location). The attached pseudonodule (a) displays slightly deformed laminations. Note the clay layer within (a) which seems to have been dislocated by the water escape.

8B and C). The internal laminations are not totally concentrical but show a synclinal form, sometimes slightly deformed. They are often associated with load casts, convolute laminations and soft sediment intrusions.

4.1.3.2. Detached pseudonodules. Their dimension varies between 5 and 20 cm. They are associated with fine to medium-grained sands floating in clays or fine-grained sands and display a concave-upward shape. Laminations are diffuse and slightly deformed, similar to those described by Rodríguez-Pascua et al. (2000) and thereby differing from classical pseudonodules. In some cases, this structure can be interpreted as lighter sediments sinking into denser sediment and associated with attached pseudonodules (Fig. 9A). This structure has been observed in association with load casts, soft sediment intrusions and convolute laminations.

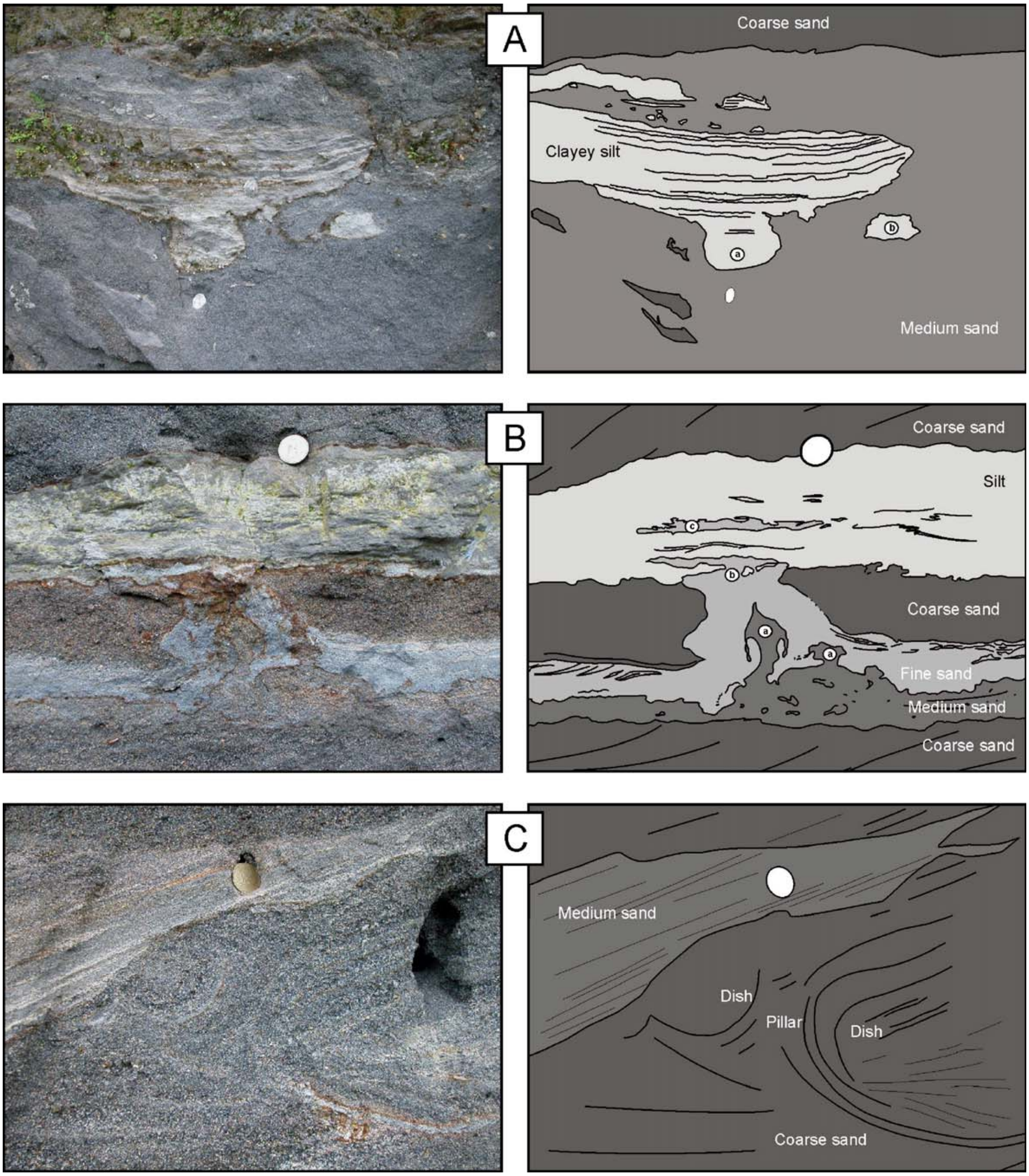

Fig. 9. Load and water-escape structures. (A) Attached (a) and detached (b) pseudonodule (section 8, see Figs. 4 and 5 for location). Note that the clayey silt has a greater density than the liquefied sand. (B) Water escape cusps (a) (section 8, see Figs. 4 and 5 for location) formed by medium-grained sands intruding fine-grained sands. Observe the laccolith shape of the fine-grained-sand intrusion into medium-grained-sands and silts (b). This structure is capped by hardly deformed silts, which have been penetrated by a sandy sill (c). (C) Dish and pillar structure (section 8, see Figs. 4 and 5 for location). The undisturbed laminations of the overlying medium-coarse-grained sand lense proves that the underlying deformation occurred prior to its deposition and is not related to loading.

4.2. Water escape structures

4.2.1. Water escape cusps

This structure is similar to that described by Rossetti (1999) and appears as bodies of sands that penetrate into overlying sand beds. They are morphologically similar to flame structures which occur only in mud, but differ in that cusps represent masses of underlying deformed sands that intrude into the overlying sand beds (Fig. 6). They display a curved shape without evidence of internal deformation (Fig. 9B). Their size generally varies between 20 and 30 cm.

4.2.2. Dish-and-pillar structures

They show the typical appearance described by many authors (e.g., Lowe and LoPiccolo, 1974): fine- to medium-grained sands with shapes similar to concave-upward saucers, separated by subvertical columns of medium to coarse-grained sands (Fig. 9C). They are up to 15 cm wide. They do not show evidence of internal deformation, although in some cases the pillars may be slightly deformed.

4.2.3. Pocket-and-pillar structures

They display features similar to those described by Postma (1983). Pockets have a flat base and are concave-upward. Their height and diameter measure between 10 and 15 cm. They are filled with coarse-grained sands and fine gravels (Fig. 10A). Pillars are columnar, generally straight and vertical with a height between 5 and 15 cm. Their fill consists of undeformed to gently deformed coarse-grained sands.

Dish-and-pillar and pocket-and-pillar structures have different origins but a similar morphology. They are abundant and are associated principally with convolute lamination and soft-sediment intrusions.

4.3. Soft-sediment intrusions

These structures defined by Lowe (1975) show variable morphology, composition and size. Because of this, it is hard to classify them. Together with load structures, they are the most common deformation encountered in the study area. In general, one can differentiate clastic sills from dykes.

4.3.1. Clastic sills

In most cases, sills are arranged like beds and can locally be confused with them. Normally they do not show internal structures, except for slightly deformed laminations. They consist predominantly of fine to medium-grained sands and occasionally coarse-grained sands. Their thickness fluctuates from 2 to 50 cm, most of them being some 10 cm thick. Their length varies from a few centimeters to 5 m. Some sills are ptygmatic, some bifurcate and others are broken up into smaller sills (Fig. 10B). Some are interconnected and others deformed by load structures.

4.3.2. Clastic dykes

Their composition is variable, consisting of predominantly coarse-grained and lesser medium-grained sands. They are internally massive. Dykes crosscut different types of lithologies, predominantly sands but also clays and diatomites. Their thickness varies between 5 and 30 cm. Their length may reach up to 1 m, but generally fluctuates between 10 and 50 cm. Their shape is variable, they may be contorted or fractured. One particular type of dyke (Figs. 10C and 11A) shows characteristics similar to those described by Rodríguez-Pascua et al. (2000) for dykes originating from the liquefaction of a basal sand bed. In this case, a set of dykes approximately 2 m from each other shows a clear interconnection with the basal sand feeding the intrusion. Dykes are composed of medium-grained sands. These vertical structures have a constant width of some 10 cm and their length reaches up to 1 m. Fractures seem to have locally guided the intrusion (Fig. 10C). They crosscut laminites, diatomites and thin beds of fine-grained sands. In the same outcrop, one can observe a set of lateral dykes not connected to a basal sand layer but laterally related to a connected dyke. This type of disconnected dyke crosscuts silts that may be contorted (Fig. 11C). The dimension and arrangement of disconnected dykes are similar to those of the connected dykes.

Because of their abundance, soft-sediment intrusions are related to almost all types of soft-sediment deformations, but most frequently with load structures, convolute laminations and dish-and-pillar structures.

4.4. Other structures

4.4.1. Disturbed laminites

This structure is associated with varve-like laminites, which consist of alternations of diatomites, clays and very fine-grained sands. The latter show a slight deformation but keep their thickness and continuity. These deformations are between 2 and 10 cm thick (Fig. 11B) and do not appear to be associated with other structures, except for some small clastic dykes.

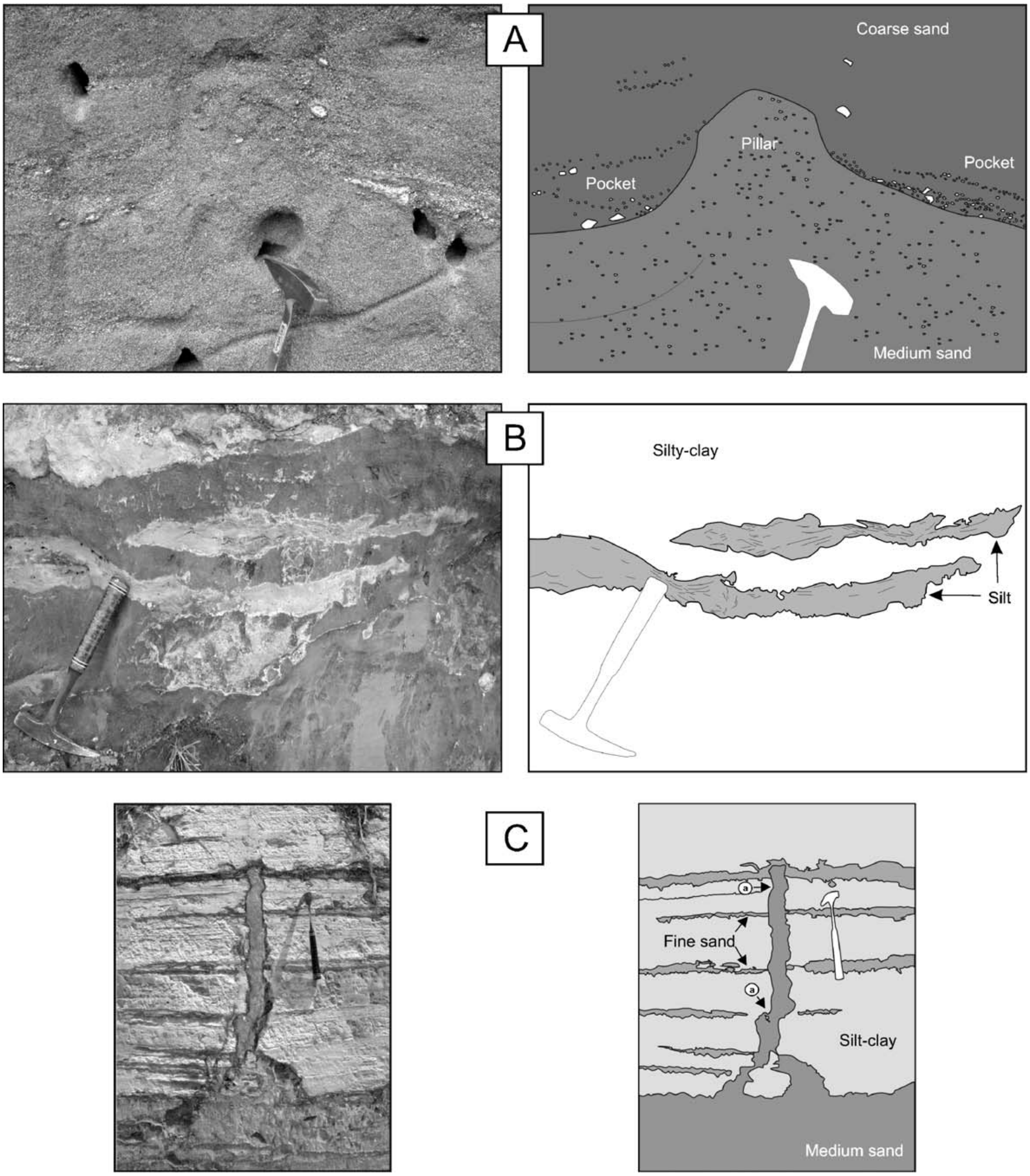

Fig. 10. Water escape and soft-sediment intrusion structures. (A) Pocket and pillar structure (section 8, see Figs. 4 and 5 for location). (B) Ptigmatic and bifurcated clastic dyke (section 6, see Figs. 4 and 5 for location). (C) Medium-grained sand, rooted vertical dyke intruding silty clays (near section 8, see Figs. 4 and 5 for location). This intrusion seems to be partially controlled by fractures (a).

4.4.2. Convolute laminations

This structure is not abundant. Sediments affected correspond in general to fine to medium-grained tuffaceous sands. The deformation is generally some 10 cm thick, but may reach up to 20 cm. The stratification is highly contorted and shows shapes similar to deformed synclines and anticlines (Fig. 11C). This structure is associated mainly with load structures, soft-sediment intrusions and slumps.

4.4.3. Slumps

This type of deformation is observed only sporadically, but can be locally quite significant. The lithologies involved are fine and medium-grained tuffaceous sands.

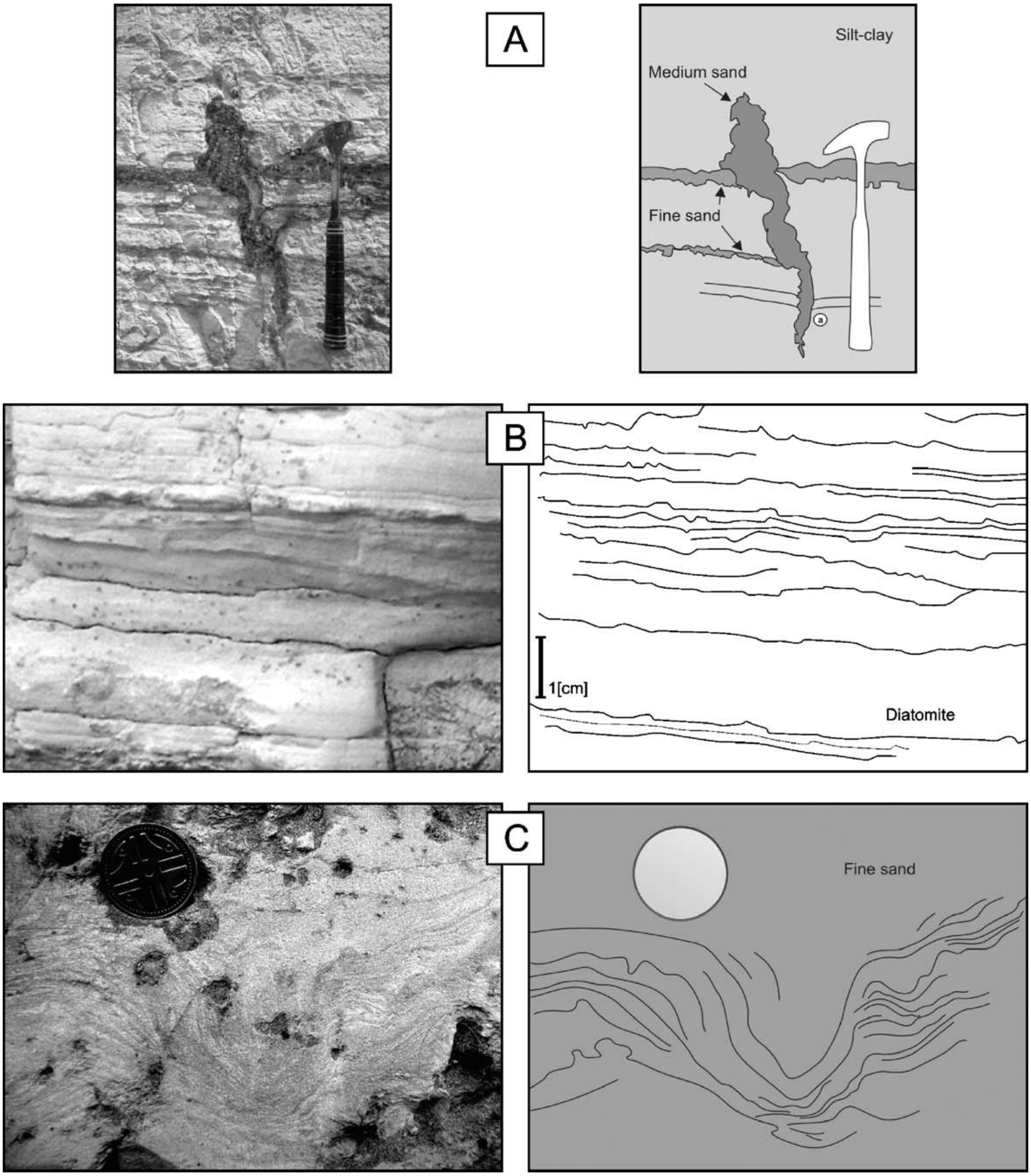

Fig. 11. Soft-sediment intrusion and other deformation structures. (A) Medium-grained sand, subvertical disconnected dyke (near section 8, see Figs. 4 and 5 for location) showing downward bending of intruded, fine-grained sediments (a) in the lower part. This feature indicates the lateral movement of the injection. (B) Disturbed laminites (near La Victoria, see Fig. 4 for location). (C) Convolute laminations (section 6, see Figs. 4 and 5 for location).

The thickness of the deformation varies between 10 cm and 1 m (Fig. 12A). They may show features similar to an intraformational fold. The axial plane of the folds is generally subvertical but may also be subhorizontal (recumbent). They are associated with load structures, convolute laminations and synsedimentary faults.

4.4.4. Synsedimentary faults

This type of brittle structure affects intervals some 20 cm to 1 m thick. Several types of faults exist in the area studied, namely high-angle planar normal faults (Fig. 7B), listric normal faults and reverse faults (Fig. 12B). Offsets vary between 2 and 20 cm. Lithologies

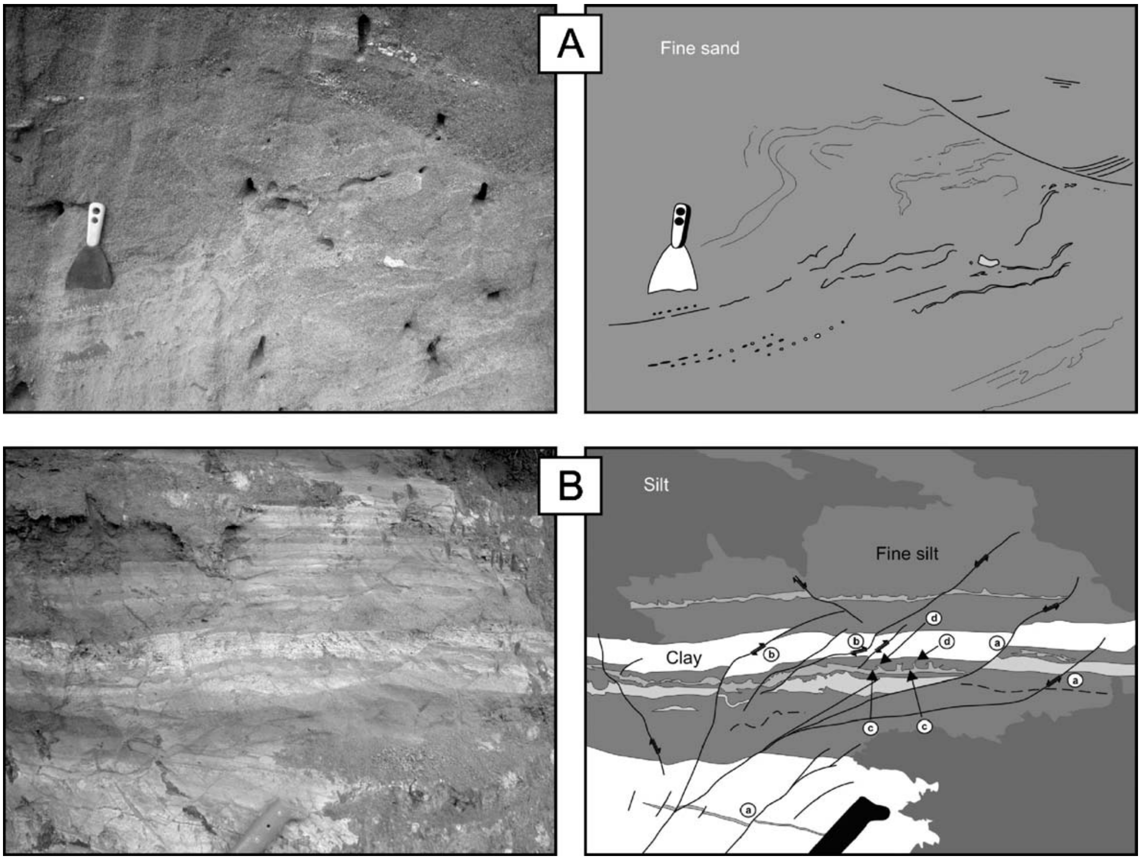

Fig. 12. Other soft-sediment deformation structures. (A) Slump with subhorizontal axial plane (section 8, see Figs. 4 and 5 for location). (B) Synsedimentary faulting showing both listric (a) and reverse (b) faults (section 8, see Figs. 4 and 5 for location). This structure can be assimilated to a flower structure indicative of strike-slip movements. Small-size load cast (c) and flame (d) structures are also observed.

involved vary from one site to the other, but generally consist of fine to medium-grained tuffaceous sands and locally clays. Synsedimentary faults are associated mainly with load structures.

5. Discussion

5.1. Deformation mechanisms and driving forces

Soft-sediment deformation is the disruption of unlithified sedimentary strata (Mills, 1983). This disruption occurs in response to a deformation mechanism, a driving force and a triggering mechanism (Allen, 1982, 1986; Owen, 1987). The deformation mechanisms and driving forces have been reviewed by various authors (e.g., Allen, 1977; Lowe, 1975; Mills, 1983; Owen, 1987, 2003; Maltman, 1994; Maltman and Bolton, 2003). The deformation occurs before significant compaction of the sediments has taken place (Mills, 1983). If a driving force such as a reverse density gradient, slope failure, slumping or shear stress is acting, the sediment strength can be significantly reduced as a consequence of a process known as liquidization (Allen, 1977, 1982). As stated earlier (Anketell et al., 1970; Mills, 1983; Owen, 1987; Rossetti, 1999), various driving forces can act simultaneously during deformation, so that in many cases there is no unique cause of deformation. When sediment is liquidized, it can be deformed in response to relatively weak stresses, which under normal conditions would not affect it. Four types of liquidization can be differentiated: thixotropy, sensitivity, liquefaction and fluidization (Owen, 1987). A sequence of soft-sediment deformation structures originates from these processes.

In many cases, deformation mechanisms are initiated by an external agent, i.e. a triggering mechanism. Some of these mechanisms have been identified: artesian groundwater movement, earthquakes, storm currents and gravity flows (e.g., Lowe, 1975; Sims, 1975; Owen, 1987, 1996).

Deformation mechanisms and driving force are interpreted below for each structure. Discussion of the triggering mechanisms follows.

5.1.1. Load structures

These structures are formed in response to gravitational instability (Moretti et al., 1999). The origin of the *simple load casts* in the Zarzal sediments is mostly related to a reverse density gradient (Anketell et al., 1970). The driving force is associated with the difference in densities between sands/gravels and the underlying sands/silty clays and diatomites. When the liquidization occurs in both layers, a Rayleigh–Taylor instability forms (Selker, 1993). The gravitational readjustment leads simultaneously to a descent of the denser sediment and an ascent of the lighter sediment. The resulting deformation depends upon the contrast of dynamic viscosities (Anketell et al., 1970; Alfaro et al., 1997). In the case of the Zarzal Formation, the differences were minimal or the lower layer had a higher dynamic viscosity favouring the formation of simple load casts.

In some cases uneven loading probably acted as a driving force, where a marked difference in densities did not exist, as for example between fine-grained sands and silty clays. The force is associated with lateral variations in the distribution of sediment load when the substrate is liquidized and loses its capacity of support (Owen, 2003). In some cases, the presence in the lower layer of convolute laminations formed by the liquidization process seems to support this interpretation.

The *pendulous load casts* have a similar origin but are associated with a more advanced stage of deformation (Anketell et al., 1970; Allen, 1982; Alfaro et al., 1997, 1999; Owen, 2003). In the study area, this structure is present mainly where the difference in density is marked, for example between tuffaceous sands and clays or diatomites (Fig. 7B). Therefore, it is probable that a simple load cast evolves into a more deformed structure, the pendulous load cast (Owen, 2003) or into a drop structure (Alfaro et al., 1997, 1999). In some cases, the structure seems unrelated to a difference in density, for example when fine-grained sands overlie coarse-grained sands. In this case, it is likely that the deformation is linked to uneven loading, a similar mechanism to that postulated by Rodríguez-Pascua et al. (2000) for the origin of some pseudonodules. The liquefaction of the underlying layer produces a decrease in the bulk density and shear strength, thereby favouring the genesis of the structure.

Flame structures have been attributed to a high difference of dynamic viscosity between the interacting layers: the lower one has a dynamic viscosity notably lower than the upper one (Anketell et al., 1970) and consequently diapiric intrusions of fine-grained sediments take place and form flame structures (Mills, 1983). However, Owen (2003) suggests that the influence of relative viscosity on load structure morphology should be reconsidered and that the amplification rate plays an important role in the genesis of antiforms (flame structures) and synforms (load casts).

Attached pseudonodules are formed by various processes (Owen, 2003), as can be deduced from experimental data (Anketell et al., 1970; Owen, 1996; Moretti et al., 1999). In the Zarzal Formation, the most probable origin is uneven loading, because there is no marked difference between the lithology of pseudonodules and that of the associated matrix and, moreover, these deformations may be associated with convolute lamination (Fig. 8C). Where this structure corresponds to a normal density gradient, the genesis is similar to that postulated by Rodríguez-Pascua et al. (2000): the liquefaction of the underlying layer generates a decrease in the bulk density and shear strength which allows the development of the pseudonodules (Fig. 8B).

Detached pseudonodules in the Zarzal Formation are related to the sinking of load casts in water-saturated fine-grained sediments as established by Kuenen (1958). The coexistence of load casts and detached pseudonodules supports this interpretation. However, in some cases, when the detached pseudonodule is composed of lighter material (Fig. 9A) than the surrounding sediment, its origin probably relates to uneven loading (Rodríguez-Pascua et al., 2000).

Some authors (Anketell et al., 1970; Owen, 2003) have proposed the existence of a deformation series for load structures. In the study area, one can observe a transition from simple load casts to pendulous load casts, attached pseudonodoles and detached pseudonodules, but without ball and pillow structures. This means that the deformation reached a relatively advanced stage.

5.1.2. Water escape structures

5.1.2.1. Water escape cusps. This structure has been interpreted by Owen (1996) as associated with local fluidization of the lower sand layer and a water escape process. The deformation produced by the water escape is superimposed on the deformation associated with a gravitationally unstable density gradient.

5.1.2.2. Dish-and-pillar and pocket-and-pillar structures. These structures have been interpreted by Lowe and LoPiccolo (1974) as originating during the compaction and dewatering of unconsolidated sediments. Pillars are formed during compaction associated with an explosive escape of water along vertical or subvertical columnar flow paths. Dishes are associated with dewatering and involve a complex interaction between escaping water, sediments and sedimentary structures. According to Lowe and LoPiccolo (1974), the concave-upward morphology of the dishes would indicate that the deformation was relatively important. The model proposed by Cheel and Rust (1986), whereby these structures make part of a series including the ball-and-pillow structures, does not seem appropriate in the studied sediments where no relation exists between ball-and-pillow and dish-and-pillar structures. According to some authors (e.g., Lowe and LoPiccolo, 1974; Lowe, 1975; Allen, 1982), the triggering mechanism of these structures is related to overloading and sediment gravity flows. However, other authors (e.g. Plaziat and Ahmamou, 1998; Moretti et al., 1999) associate them with seismicity. In the study area, the latter activity seems to be the principal triggering factor (see below).

According to Postma (1983), the pocket-and-pillar structures are fluid escape structures resulting from the local liquefaction and fluidization associated with high rates of sediment deposition and the presence of less permeable and texturally immature sediments. In the Zarzal Formation, the latter condition exists, whereas rapid deposition rates have not been established.

5.1.3. Soft-sediment intrusions

These structures have been generally interpreted as originating from the injection of liquidized sands into the surrounding, mostly overlying strata (e.g., Walton and O'Sullivan, 1950; Potter and Pettijohn, 1977; Lowe, 1975). Most of the intrusions in the Zarzal Formation are interpreted as liquefied intrusions (Lowe, 1975), because they are structureless and associated with deformation of the adjacent strata. Fluidized intrusions are also present in a lesser proportion and correspond to sands injected along fractures and bedding planes (Fig. 10C). The presence of ptygmatic intrusions (Fig. 10B) indicates that the injection is associated with a higher degree of compaction in the surrounding sediments (Kuenen, 1968).

The presence of dykes vertically connected to a sand layer reflects an upward movement. The lateral dykes are associated with a lateral sand flow which produces the upward/downward bending of the intruded strata (Fig. 11A). In both cases, the deformation is linked to the liquefaction of the basal sand (Rodríguez-Pascua et al., 2000). Some authors (e.g., Martel and Gibling, 1993; Parize et al., 1987; Parize and Fries, 2003) relate clastic dykes to sediment gravity flows and storm waves. However, in the Zarzal Formation, there is no sedimentological evidence supporting the latter mechanisms (see below).

5.1.4. Other structures

5.1.4.1. Disturbed laminites. A similar structure to this has been described by Rodríguez-Pascua et al. (2000) and interpreted as the product of ductile deformation, although they did not observe changes in thickness, but only a bending of the laminites. Probably the sediments kept some residual strength favouring a weak, ductile behaviour (Maltman and Bolton, 2003).

5.1.4.2. Convolute laminations. Although no unique interpretation exists with respect to the generating mechanism, this structure has generally been considered as an extremely complex form of load structure (Dzulynski and Smith, 1963; Lowe, 1975; Mills, 1983; Rossetti, 1999). Interpretation of the driving force varies among authors: some relate it to current drag or bed shear and others to slumping (Mills, 1983; Owen, 1996; Plaziat and Ahmamou, 1998). In the Zarzal Formation, these structures are not associated with clays and the probability of thixotropic behaviour is low. Consequently, the mechanism is probably linked to fluidization contrasts which create gravitational instabilities (Brenchley and Newall, 1977; Owen, 1996; Rossetti, 1999). In a few cases, the association of convolute lamination with soft-sediment intrusions suggests that a slope in the substrate may have been involved and that the structure is related to a hydroplastic deformation (Plaziat and Ahmamou, 1998).

5.1.4.3. Slumps. These structures are associated with the downslope movement of underconsolidated sediments under the influence of gravity. The failure occurs when the sediments are steepened beyond the stable angle of repose (Mills, 1983). Considering the minimal mixing of sediments and the general conservation of the bedding, the slumps can be classified as coherent (sensu Dzulynski, 1963) and resemble the "contorted" slumps mentioned by McKee et al. (1962).

5.1.4.4. Synsedimentary faults. This brittle deformation corresponds to a cohesive behaviour (Owen, 1987; Vanneste et al., 1999), whereby the increase in pore water pressure is not enough to liquefy the sediments. The presence of faults and their association with undeformed strata correspond to a brittle deformation when sediments are either unconsolidated or partly consolidated (Rossetti and Goes, 2000). The coexistence of structures associated with ductile and brittle deformations can be related to differential compaction, which determines the pore pressure within the sediments: the more compacted and less saturated sediments have a brittle behaviour, whereas the less compacted and more saturated sediments show a ductile behaviour (Mohindra and Bagati, 1996; Rossetti, 1999; Rossetti and Goes, 2000).

5.2. Triggering mechanisms

Several mechanisms can trigger synsedimentary deformation. The best known are sediment loading (e.g. Anketell et al., 1970; Lowe and LoPiccolo, 1974), storm currents (e.g., Dalrymple, 1979; Molina et al., 1998; Alfaro et al., 2002) and seismicity (e.g., Seilacher, 1969; Lowe, 1975; Sims, 1975; Martel and Gibling, 1993; Mohindra and Bagati, 1996; Calvo et al., 1998; Lignier et al., 1998; Rossetti, 1999; Alfaro et al., 1999; Vanneste et al., 1999; Jones and Omoto, 2000; Rodríguez-Pascua et al., 2000; Bowman et al., 2004).

Sediment loading seems to be of minor importance in the Zarzal Formation. Such processes could be associated with the rapid deposition of tuffaceous sands and gravels originating form the reworking of pyroclastic material and gravity flows. The weight of these sediments on water-saturated sediments could be a triggering mechanism (Owen, 1996; Jones and Omoto, 2000). However, no direct relationship between the loading of tuffaceous sediments and soft-sediment deformations has been observed in the field.

Storm currents can be a triggering mechanism for soft-sediment deformations. The Zarzal Formation shows no evidence of any sedimentary structure associated with high energy (for example swaley and hummocky cross-stratification , see Molina et al., 1998). On the other hand, the vast lateral extent of the soft sediment deformation observed suggests a more regional triggering mechanism than one related to the local action of storms. Moreover, the liquefaction of sediments requires a storm wave height in excess of 6 m (Alfaro et al., 2002), something impossible in a lake like that in which the Zarzal Formation was deposited.

Seismicity is the most probable triggering mechanism of the soft sediment deformation structures encountered in this study. Seismicity can cause the fluidization of granular solids (Allen, 1982, 1986; Lowe, 1975). Although all the structures characterizing seismites (Seilacher, 1969) have not yet been encountered, a sufficient number of criteria (Bowman et al., 2004 and references therein) confirming the role of seismicity as triggering mechanism are established.

The studied area has a well-established present-day seismicity (CRQ (Corporación autonoma Regional del Quindío) and Univ. del Quindío, 1999) related to the Cauca-Romeral Fault System. The latter represents a Cretaceous paleosuture and has been active since then (McCourt, 1984). In recent times, it has been related with the earthquakes that affected the cities of Manizales, Pereira and Armenia. Field evidence confirms that tectonic faulting (Figs. 13 and 14) and tilting (see drainage pattern, Fig. 4) have affected the Zarzal For-

Fig. 13. Post-depositional extensional tectonics in the Zarzal Formation (section 7, see Figs. 4 and 5 for location).

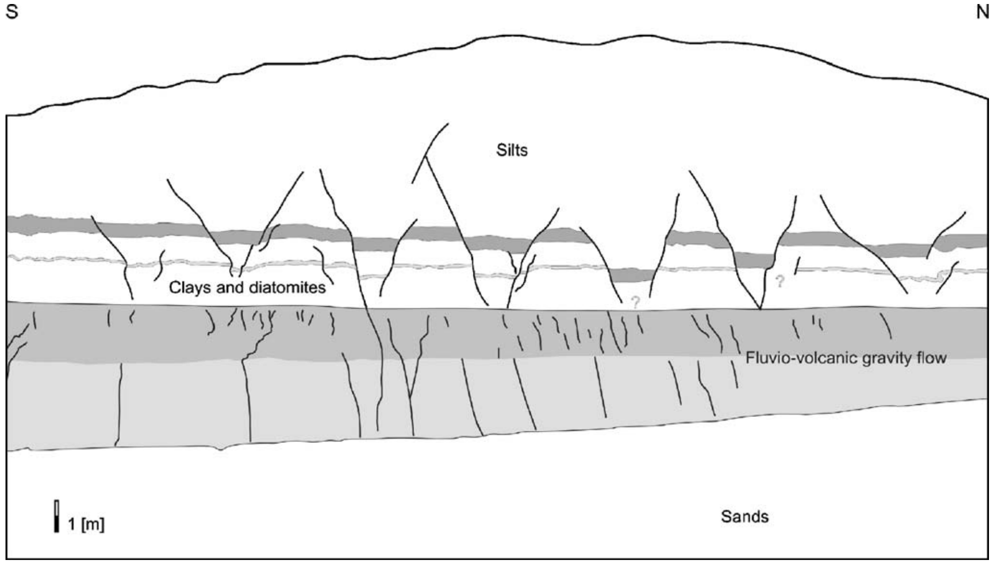

Fig. 14. Post-depositional strike-slip tectonics in the Zarzal Formation (section 8, see Figs. 4 and 5 for location).

mation. Moreover, the lithology of the Zarzal Formation corresponds to sediments quite to deformation when exposed to seismic waves.

The structures encountered in the Zarzal Formation are similar, both in size and shape, to those described in the field as seismites (e.g., Sims, 1975; Calvo et al., 1998; Rossetti, 1999; Vanneste et al., 1999; Rossetti and Goes, 2000; Lignier et al., 1998; Jones and Omoto, 2000; Greb and Dever, 2002; Bowman et al., 2004) or those generated experimentally (e.g., Kuenen, 1958; Owen, 1996; Moretti et al., 1999). The deformed intervals are intercalated within undeformed strata, thereby reflecting catastrophic events followed by periods of relative stability. Moreover, the deformed strata can be observed throughout a fairly large area (Figs. 4 and 5) although, so far, the lack of precise dating does not permit the correlation of the events in the different studied sections. Moreover, the classification of Wheeler (2002) provides further criteria pointing to the presence of seismites: among the six criteria (or tests) identified by Wheeler (2002) for evaluating the occurrence of seismites, Zarzal sediments fulfill four tests, the other two being inconclusive.

The abundance of soft-sediment deformation structures indicates the proximity of an active fault zone. However, further detailed field investigations are needed to precisely determine which were the active faults within the system. The soft-sediment deformation in the Zarzal Formation can be interpreted as seismites related to the activity of faults associated with the Cauca-Romeral Faults System. Field evidence demonstrates synsedimentary tectonic activity with both strike-slip (Fig. 12) and extensional faults (Fig. 7B). Similar tectonic activity continued after the deposition of the unit, as demonstrated by the occurrence of post-depositional strike-slip and extensional movements (Figs. 13 and 14). This area of Colombia is in a transpressional regime (Taboada et al., 2000), which generates strike-slip faults and associated local normal faults (pull-apart basins).

The relation between soft-sediment deformations and the magnitude of earthquakes has received much attention (e.g., Seed and Idriss, 1971; Sims, 1975; Allen, 1986; Scott and Price, 1988; Marco and Agnon, 1995; Obermaier et al., 2002). Although some authors (e.g., Seed and Idriss, 1971) consider that magnitudes from 2 to 3 are enough to trigger liquefaction, most scientists estimate that the magnitude of an earthquake should be higher than 4.5 (e.g., Marco and Agnon, 1995) to be registered in the sedimentary record. Scott and Price (1988) suggest that a magnitude of less than 5 does not cause a significant sediment liquefaction beyond 4 km from the epicenter and that a magnitude of 7 does not significantly affect sediments beyond 20 km. Considering these values and the distance to the faults that might have been active during the sedimentation of the Zarzal Formation in the Cauca Valley (e.g., some 10 km to the Quebradanueva Fault, Fig. 4), it can be tentatively postulated that the magnitude of the earthquakes having generated the seismites was between 5 and 7. The record of recent instrumental seismicity confirms the magnitude of earthquakes in this area (CRQ (Corporación autonoma Regional del Quindío) and Univ. del Quindío, 1999). This range of

Fig. 15. Injection dykes (section 8, see Figs. 4 and 5 for location) are a proof of extension (Rodríguez-Pascua et al., 2000).

postulated magnitudes corresponds with the scale established by Rodríguez-Pascua et al. (2000) for a series of soft sediment deformation structures: they postulate that the existence of sand dykes and pseudonodules indicates magnitudes between 5 and 7 (Fig. 15).

6. Conclusions

The Plio-Pleistocene Zarzal Formation in the Cauca Valley was deposited in an intramontane depression between the Western and Central Cordilleras of Central Colombia, a zone affected by the movements of the Cauca-Romeral Fault system. Fine-grained lacustrine sediments (sands, silts, clays and diatomites) alternate with fluvial-dominated coarser sands and gravels and fluvio-volcanic mass flows derived from the Central Cordillera to the east.

Sands, silts and clays in the Zarzal Formation preserve intervals with abundant soft-sediment deformation interbedded with undeformed strata. Deformation structures can be classified into four groups: load structures (load casts, flame structures and pseudonodules), water escape structures (water escape cusps, dish and pillar, pocket and pillar), soft-sediment intrusions (clastic sills and dykes) and other structures (disturbed laminites, convolute laminations, slumps and synsedimentary faults).

Deformation mechanisms and driving forces of these deformations correlate with those known in the literature: load structures are the result of gravitational instabilities related to density differences or uneven loading; water escape structures are associated with dewatering and soft-sediment intrusions with the injection of liquidized sands into surrounding strata; disturbed laminites are the result of ductile deformation, convolute laminations are linked to gravitational instabilities associated with inverse density gradients and slumps to gravitational downslope movements. Synsedimentary faulting is a brittle deformation that takes place when the pore pressure is not sufficient to liquefy the sediments.

Field evidence and regional geological criteria point to seismicity as the most probable triggering mechanisms for the deformations, which are consequently interpreted as seismites. There is no field evidence of the influence of sediment loading phenomena or of deformation linked to storm activity. The studied area has a history of geological and recent seismicity related to the Cauca-Romeral Fault System. The sandy lithology of the Zarzal sediments is prone to liquefaction when exposed to earthquakes and structures encountered are similar in size and shape to those encountered in seismites. Moreover, the disturbed intervals have a large areal extent and their intercalation within undeformed strata reflects the catastrophic nature of these events.

In summary, the Plio-Pleistocene Zarzal Formation represents a new case history of soft-sediment deformations related to earthquakes.

Acknowledgements

The authors are indebted to Dr. A. Espinosa for providing regional geological information and help in the field. The authors are grateful to the two anonymous reviewers for their helpful comments. This research is supported by the Swiss National Science Foundation (grant no. 21-67080.01). Part of the field work was supported by the Swiss Academy of Sciences.

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