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Soft-Sediment Deformation in the Campanian-Maastrichtian Deltaic Deposits of the Afikpo Sub-basin, South-eastern Nigeria: Recognition of Endogenic Trigger

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Abstract

In this research, soft-sediment deformation structures are studied and analyzed from the Cretaceous deltaic deposits of the Afikpo Sandstone Member of the Nkporo Formation in Afikpo Sub-basin. These deformation structures are grouped into three morphological and deformational styles which include load structures (load casts, ball-and pillow-structures, flame structures, and pseudonodules), ductile disturbed structures (loop bedding, recumbent folding, and deformed lamination), and brittle deformation (synsedimentary fractures). The facies analysis revealed sedimentary structures that are indicative of storm and wave processes and rapid sedimentation. Though the trace fossils have a low to moderate diversity and abundance, the *Skolithos* ichnofacies, particularly the *Diplocraterion* and *Skolithos* burrows, are relatively frequent. These ichnofossils are indicative of high-energy conditions and rapid sedimentation. The recognition of the deformation processes and their triggers suggests that the soft sediment deformation structures in the study area are not tectonic in origin as they are associated with storm-generated sedimentary structures and structures formed due to rapid sedimentation. Furthermore, the deformational structures are restricted to a single stratigraphic layer, and are not laterally extensive suggesting an endogenic or non-seismic trigger mechanisms such as storm events, overloading, and rapid sedimentation.

Keywords: Cretaceous deltaic deposit, soft sediment deformation, deformation processes, non-seismic trigger.

1. Introduction

Soft sediment deformation structures (SSDS) occur in unconsolidated water-saturated sediments during or shortly after deposition, and before significant diagenesis (Owen et al., 2011). Interest in the study of soft sediment deformation structures and their triggering mechanisms have progressively developed considerably over the past decades (Kuenen, 1958; Sanders, 1960; Dott and Howard, 1962; Lowe, 1975; Allen, 1982; Mills, 1983; Owen, 1987; Moretti, 2000; Rodríguez-Pascua et al., 2000; van Loon, 2009; Oliveira et al., 2011; Owen and Moretti, 2011; Owen et al., 2011; Sarkar, et al., 2014; Shanmugam, 2017). Many authors have associated soft-sediment deformation structures with seismically-induced triggering mechanisms (Mohindra and Bagati 1996; Moretti, 2000; Rodríguez-Pascua et al., 2000; Ettensohn et al., 2002; Samaila et al., 2006; Bhattacharya and Bhattacharya, 2010; Martín-Chivelet et al., 2011; Lunina and Gladkov, 2016), whereas limited works refer such deformational structures to non-seismically-induced triggering mechanisms such as storms, waves, tidal surge, floods, overloading, rapid sedimentation, and groundwater movements (Allen, 1982; Owen, 1987; Kerr and Eyles 1991; Molina et al., 1998; Alfaro et al., 2002; Owen and Moretti, 2011; Põldsaar and Ainsaar, 2014). Furthermore, some soft-sediment deformation structures, such as loop bedding, seem to be specifically related with seismic trigger mechanisms (Calvo et al., 1998; Rodríguez-Pascua et al., 2000; Martín-Chivelet et al., 2011).

Detailed description and interpretation of the soft-sediment deformation structures (SSDS) in the Afikpo Sandstone Member, particularly in the Itigidi region, is not yet reported in any published work. This paper records for the first time the genesis and implication of soft-sediment deformation in the Afikpo Sub-basin. Published reports of detailed soft-sediment deformation structures in Nigerian sedimentary basins are mostly limited to the Bima Formation of the Northern Benue Trough (Jones, 1962; Benkhelil, 1980, 1989; Braide, 1992; Guirand and Plaziat, 1993; Samaila et al., 2006), the Ajali Formation of the Anambra Basin (Ladipo, 1988; Obi and Okogbue, 2004; Olabode, 2014), the Central and the Southern Benue Trough (Hoque, 1984; Benkhelil, 1986, 1987; Dim et al., 2016). Detailed facies analysis and ichnology of the Campanian-Maastrichtian deposits of the Afikpo Sandstone Member in the Itigidi region is discussed in Mode et al. (2018). This research concentrates on the recognition of the soft-sediment deformation structures (SSDS) in the Itigidi area.

In this study, sedimentology and a detailed description of the SSDS help establish some genetic relationships between ordinary sedimentary processes (endogenic triggering mechanisms) and the occurrence of deformed beds. Furthermore, from a regional point of view, this study describes and interprets the occurrence of soft-sediment deformation structures in the Afikpo Sandstone deposits of the Afikpo Sub-basin.

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2. Geological Setting

The Afikpo sub-basin commonly referred to as Afikpo Syncline is located in south-eastern Nigeria, and its origin is linked to the formation of the Benue Trough. The Benue Trough originated from the break-up of the Gondwana supercontinent which resulted in the separation of the African plate and South-American plate during the Late Cretaceous (Reyment, 1965; Murat, 1972; Nwachukwu, 1972; Olade, 1975; Kogbe, 1976; Petters, 1978; Wright, 1981; Benkhelil, 1982, 1989; Hoque and Nwajide, 1984; Maurin et al., 1986). The trough is part of a large West and Central African Rift System (WCARS) (Genik, 1992) which resulted from the crustal stretching of the African plate consequent upon the break-up of the Afro-American plate (Figure 1). The rift structure is generated by sinistral displacements along a pre-existing zone of northeast-southwest trending transcurrent fault (Benkhelil, 1982, 1989; Maurin et al., 1986; Popoff et al., 1986). The Benue Trough occurred as an elongate intracratonic structure over 1,000 km long and up to 250 km at its widest part. Olade (1979); Popoff et al., (1982); Benkhelil (1989) suggested a magmatic activity during the opening and filling of the Benue Trough which led to the deposition of the Abakaliki pyroclastics. However, Hoque (1984) argued that the pyroclastics are post-Santonian in age. Sedimentation in the Southern Benue Trough (Abakaliki Basin) commenced in the Aptian-Albian times, and was halted due to the Santonian thermo-tectonism (Figure 2). This Santonian compressional phase resulted in the folding, faulting and uplifting of the Abakaliki Basin to form an anticlinorium; displacing depocentres westward and eastward to form the Anambra Basin and Afikpo Sub-basin (Simpson, 1954; Benkhelil, 1986, 1989; Okoro et al., 2012a). Sedimentation commenced in the Afikpo Sub-basin during the Campanian to Early Maastrichtian with the Nkporo Shale and Afikpo Sandstone Members (Simpson, 1954; Reyment, 1965) which formed the Nkporo Formation; this is followed by the Maastrichtian Coal Measures which include the Mamu, Ajali, and Nsukka formations, with the Nsukka Formation terminating sedimentation in the basin (Nwajide, 2005).

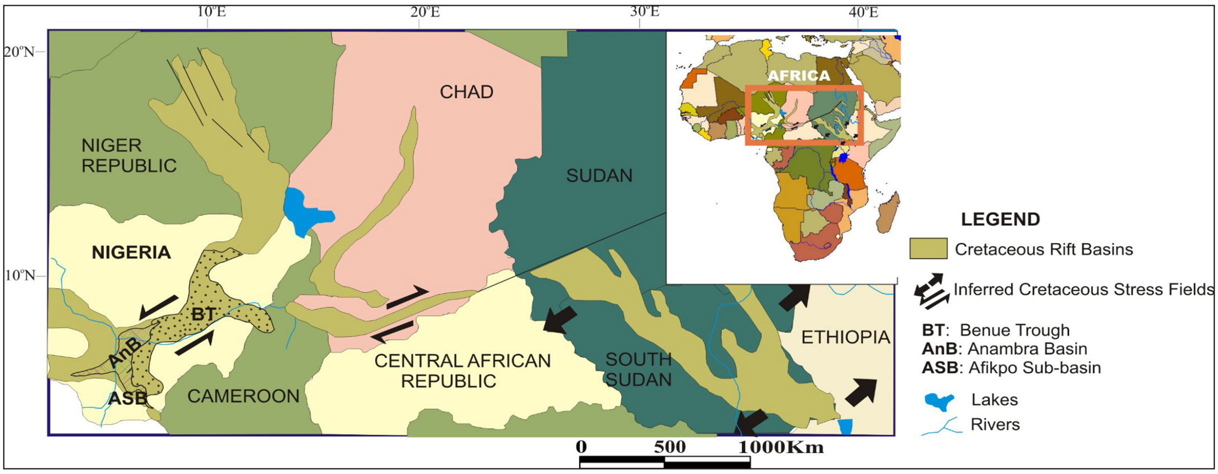


Figure 1. Tectonic map of the West and Central African Rift Systems (WCARS), showing the Benue Trough, the Anambra Basin, and the Afikpo Sub-basin (modified after Oha et al., 2016; Mode et al., 2018).

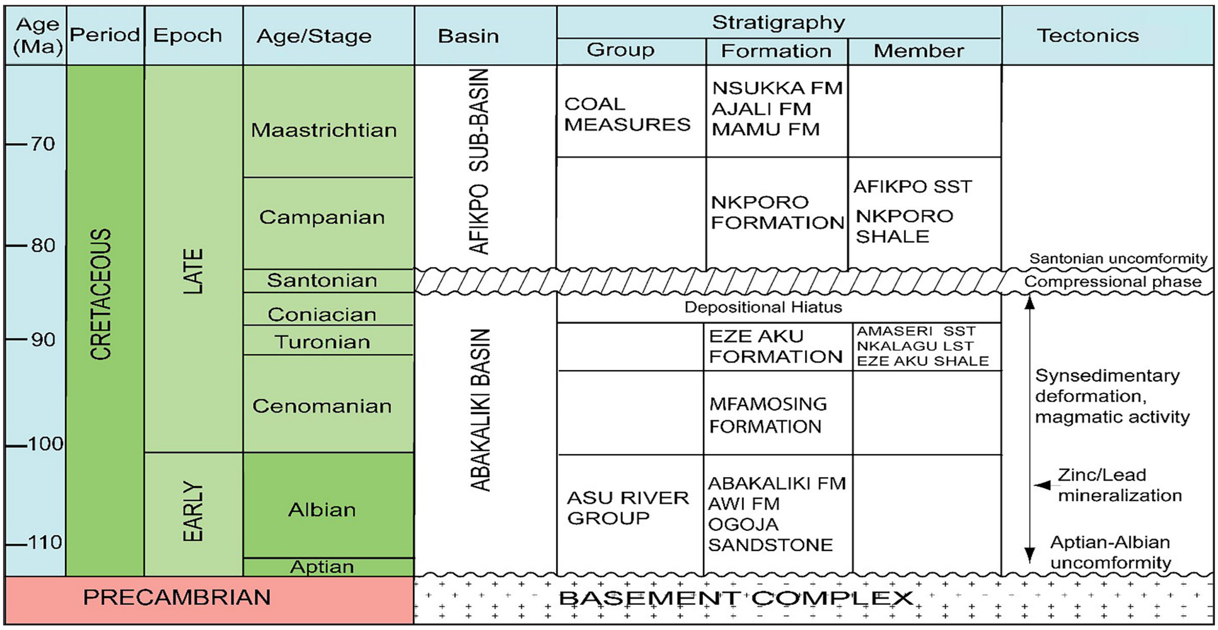


Figure 2. Tectonostratigraphy of the Abakiliki-Benue Trough and the Afikpo Sub-basin (Modified and redrawn from Okoro et al., 2012a; Nwajide, 2013; Ekwenye et al., 2015).

Note: SST stands for Sandstone; FM stands for Formation

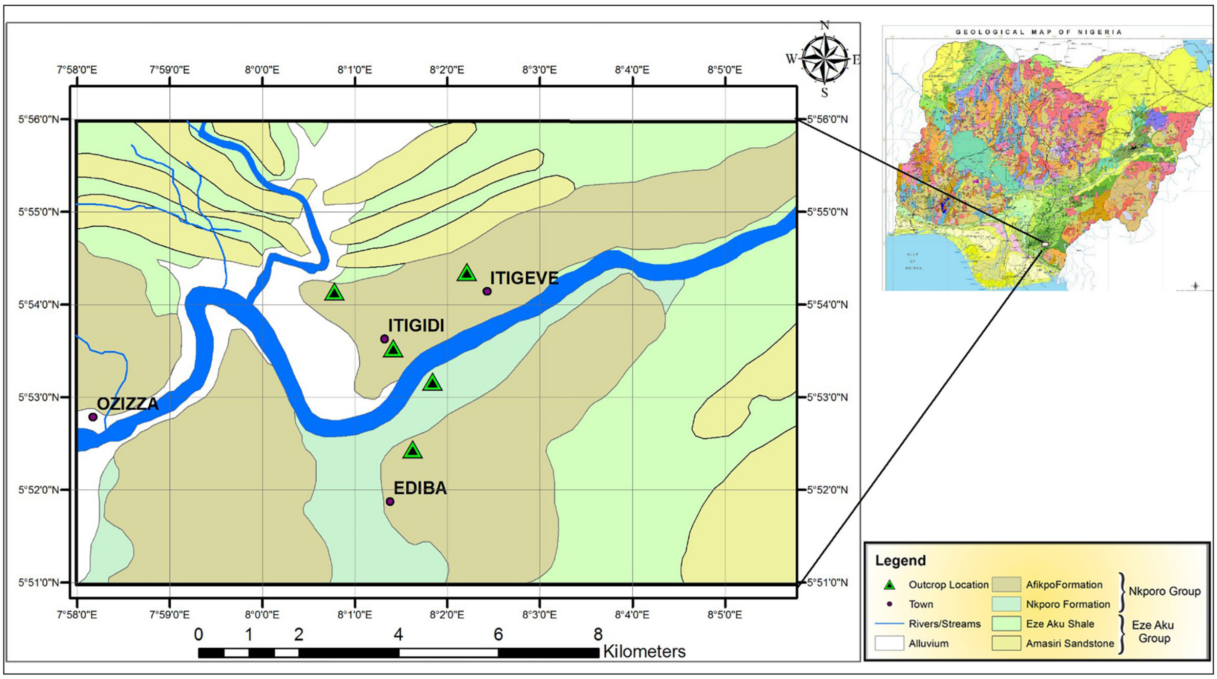


Figure 3. Geologic map of study area showing the major outcrop locations within Itigidi and environs.

3. Sedimentology

Several works have been carried out to investigate the depositional environments of the Nkporo Formation in the Afikpo Sub-basin, most of the studies are done on a regional scale (Simpson, 1954; Reyment, 1965; Banerjee, 1981; Benkhelil, 1986; Mode, 1991; Petters and Edet, 1996; Odigi, 2011; Okoro et al., 2012a, b), and there are limited data on the Itigidi-Ediba regions (Benkhelil, 1986; Mode et al., 2018). The study area is underlain by the Afikpo Sandstone Member of the Nkporo Formation (Figure 3). Mode et al., 2018 described the sedimentary units in the Itigidi region and grouped them into six facies associations (FA 1–6) depicting a prograding deltaic strata deposited on the shelf, prodelta, distal delta front, proximal delta front, distributary mouth bar, and bay-fill sub-environments. Abridged sedimentology of the study area is integrated with the SSDS to suite the purpose of this study.

The basal unit of the study area is a well-exposed fossiliferous shale and well-sorted fine to very fine grained sandstone with micro-hummocky cross-stratification (which has an average length of 5 cm and width of 2 cm), interpreted as shelf (FA 1) and the prodelta (FA 2) deposits (Mode et al., 2018) record no soft sediment deformation structures (Figures 4). Sets of coarsening-upward beds of structureless sandstone, siltstone, sandy heterolithic, current rippled laminated sandstone, thin muddy heterolithic, and bioturbated sandstone units, considered as distal delta front, and also known as distal bar (FA 3) (Mode et al., 2018), exhibit a number soft-sediment defomation structures. Load cast, ball-and-pillow structures, flame-like structures, pseudonodules were observed within the sandy heterolithic facies and sandstones which are underlain by muddy heterolithic facies and carbonaceous mudstone. Loop bedding occurs within the muddy heterolithic unit. Suite of trace fossils found within sandy units includes *Palaeophycus*, *Diplocraterion, Skolithos* and *Teichichnus*. The bioturbated sandstone facies which tops the distal prodeltaic deposits is intensely bioturbated consisting of *Skolithos, Cruziana,* and *Zoophycus* ichnofacies.

A coarsening upward succession of sandy and muddy heteroliths, followed by stacks of cross-bedded and horizontally-bedded coarsening upward sandstone and hummocky cross-stratified sandstone, dominates the proximal delta front (FA 4). FA 4 exhibits relatively fewer soft sediment deformation structures such as load casts and flame structures which occur locally within the sandy heterolithic and muddy heterolithic units. The load casts show low-diversity ichnofossils consisting of *Skolithos, Trichichnus* burrows.

The bioturbation intensity is low to moderate. Cruziana and Skolithos ichnofacies such as *Diplocraterion, Macaronichnus, Chondrites, Planolites, Rosselia, Conichnus, Skolithos? Cylindrichichnus* and *Trichichnus* were observed within the hummocky cross-stratified sandstone. *Diplocraterion parallelum* and *Conichnus* burrows are dominant at the basal part of the current rippled laminated sandstone facies.

The study are is capped by hummocky cross-bedded sandstone at the base; overlain by parallel to low-angle laminated sandstone and swaley cross-bedded sandstone facies referred to as a distributary mouth-bar deposits (FA 5). Here, the soft-sediment deformation structures, such as the load casts, ball and pillow structures, pseudonodules, flame structures, and recumbent fold, occur within the sandy mouth-bar deposits on top and the muddy heterolithic interpreted as bay-fill deposits at the base (FA 6). The degree of bioturbation is low within the mouth-bar and bay-fill deposits. *Diplocraterion, Skolithos, Conichnus* ichnofossils are common within the mouth-bar deposits.

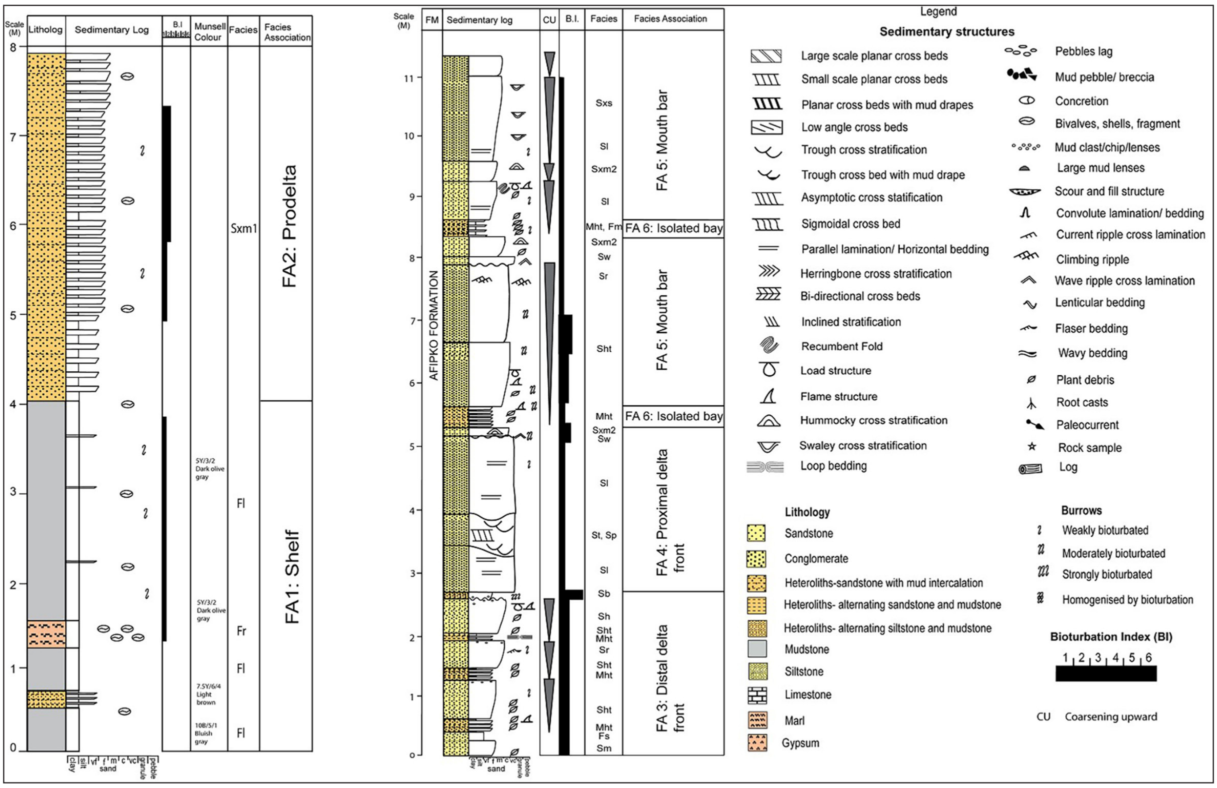


Figure 4. Representative lithostratigraphic log profiles of Ediba (a) and Itigidi (b) areas.

4. Results and Interpretation

4.1 Soft-Sediment Deformation Structure: Classification and Driving Forces

The soft-sediment deformation structures observed in the study area are dominant in the delta front, mouth-bar and bay-fill deposits. The deformation structures are contemporaneous with deposition or post-deposition, and they are common in medium to fine-grained sandstone, siltstone, heterolithic deposits, and carbonaceous mudstone in the study area. However, syndepositional fractures occur in coarse-grained sandstone. The deformational structures, especially the load structures, are commonly observed in the proximal and distal delta front deposits, where more competent materials (sandstone, sandy heterolithic sediment) overlie less competent materials (mudstone/shale, siltstone or muddy heterolithic sediment). These deformational structures are less common in sandy areas or units with less muddy sediments.

The soft-sediment deformation structures in the study area are classified into three groups based on morphological and deformational styles (Lowe, 1975; Mills, 1983; Owen, 1987, 2003; Moretti, 2000; Alfaro et al., 2002; Owen et al., 2011) as follows: 1) load structures which include load casts, pseudonodules, ball-and-pillow structures and flame structures 2) ductile disturbed structures such as recumbent folds, loop beddings, and deformed lamination, 3) brittle deformed structures which consist of syndepositional fractures. These morphological trends depend on the driving forces which acted on the sediments during deformation. The major driving forces that initiated these deformations include gravitational instability and shear stress by current (Figure 5).

Load Structures

The load structures encountered in the study area include load casts, ball-and-pillow structures, flame structures and pseudonodules (Figures 5 and 6). Most of the load structures occurred within the same horizon, and they are mostly formed by gravitational (vertical) forces associated with reversed density mechanism (Owen et al., 2011), or reduction in shear stress as in the case of flame structures (Mills, 1983). Some of the load structures are associated with or occur as deformed parts of climbing ripples, wave ripples, and hummocky cross-stratification.

Load casts are the commonest load structures and soft-deformation structures in the study area. Their maximum size is about 1.3–1.5 m in width and 50–80 cm in height, and they occurred in the medium to fine-grained sandstone and sandy heterolithic deposits (Figure 6a). Their internal structures are well-preserved. The load casts are large, undulating, with a bulbous protrusion; they slightly sink into the underlying muddy substrate, showing a concave profile (Figures 6a, c-e). Some load casts occur at the bottom of larger pillows (Figure 6e). The load casts are associated with water-escape structures formed between adjacent the load casts (see Figures 6c, f). The characteristics of these large-scale load casts are similar to those described by Alfaro et al. (2002) and Moretti and Sabato (2007).

Driving force: Load casts are characterized by the deposition of competent (denser) sandy sediment over less competent (less dense) muddy substrate, which results in gravitational instabilities due to a reversed density gradient (Mills, 1983; Moretti and Sabato, 2007; Owen et al., 2011). Reversed density deformation may also be produced by rapid deposition of weakly compacted mud, overlain by rapidly deposited medium-grained sands.

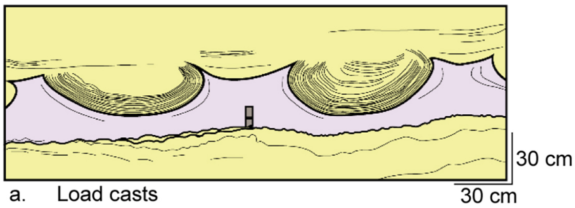
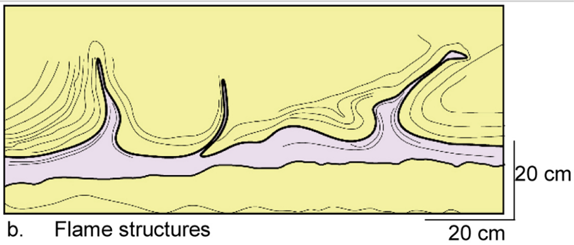
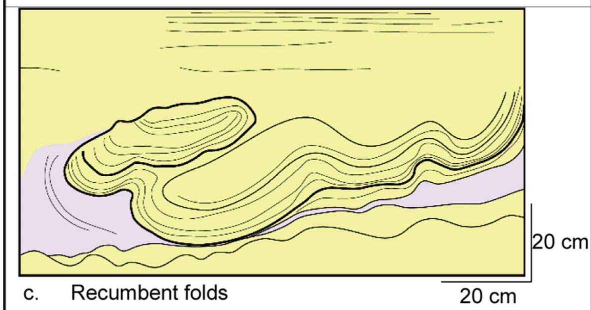
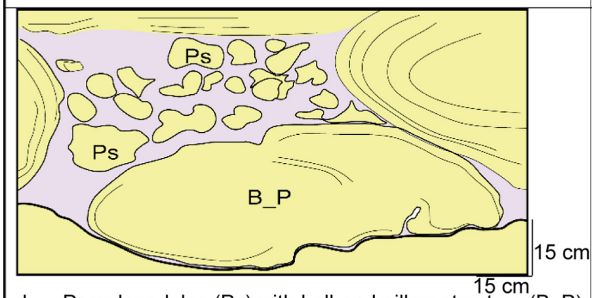
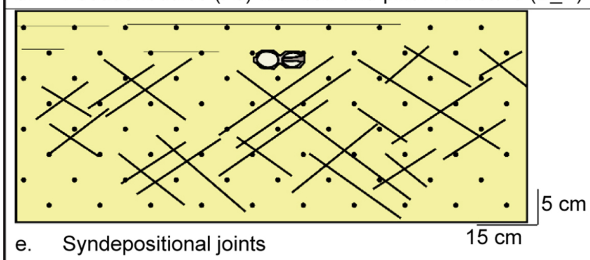
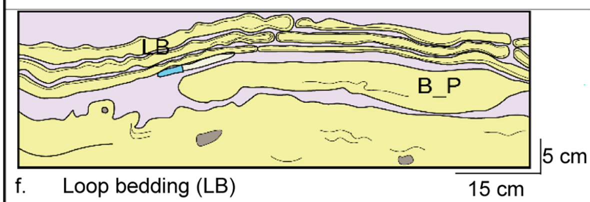
Soft sediment deformation structures	Driving Force	Depositional mechanism	Trigger mechanism
 a. Load casts	Gravitational instability due to reversed density gradient	Liquefaction and fluidization	Rapid sedimentation
 b. Flame structures	Vertical shear stress due to gravitational readjustment	Liquefaction and fluidization	Rapid sedimentation and storm
 c. Recumbent folds	Current shear or lateral drag of flowing fluid	Liquefaction	Rapid sedimentation and storm
 d. Pseudonodules (Ps) with ball-and-pillow structure (B_P)	Gravitational instability due to reversed density loading	Liquefaction	Rapid sedimentation and storm
 e. Syndepositional joints	Increase in pore pressure in cohesive material results to tensional stress	Liquefaction and intergranular shear	Rapid sedimentation or storm
 f. Loop bedding (LB)	Increase pore pressure results to tensional stress	Liquefaction and intergranular shear	Rapid sedimentation

Figure 5. A catalogue of the soft sediment deformation structures encountered in Itigidi, their driving forces, deformation mechanism and triggers.

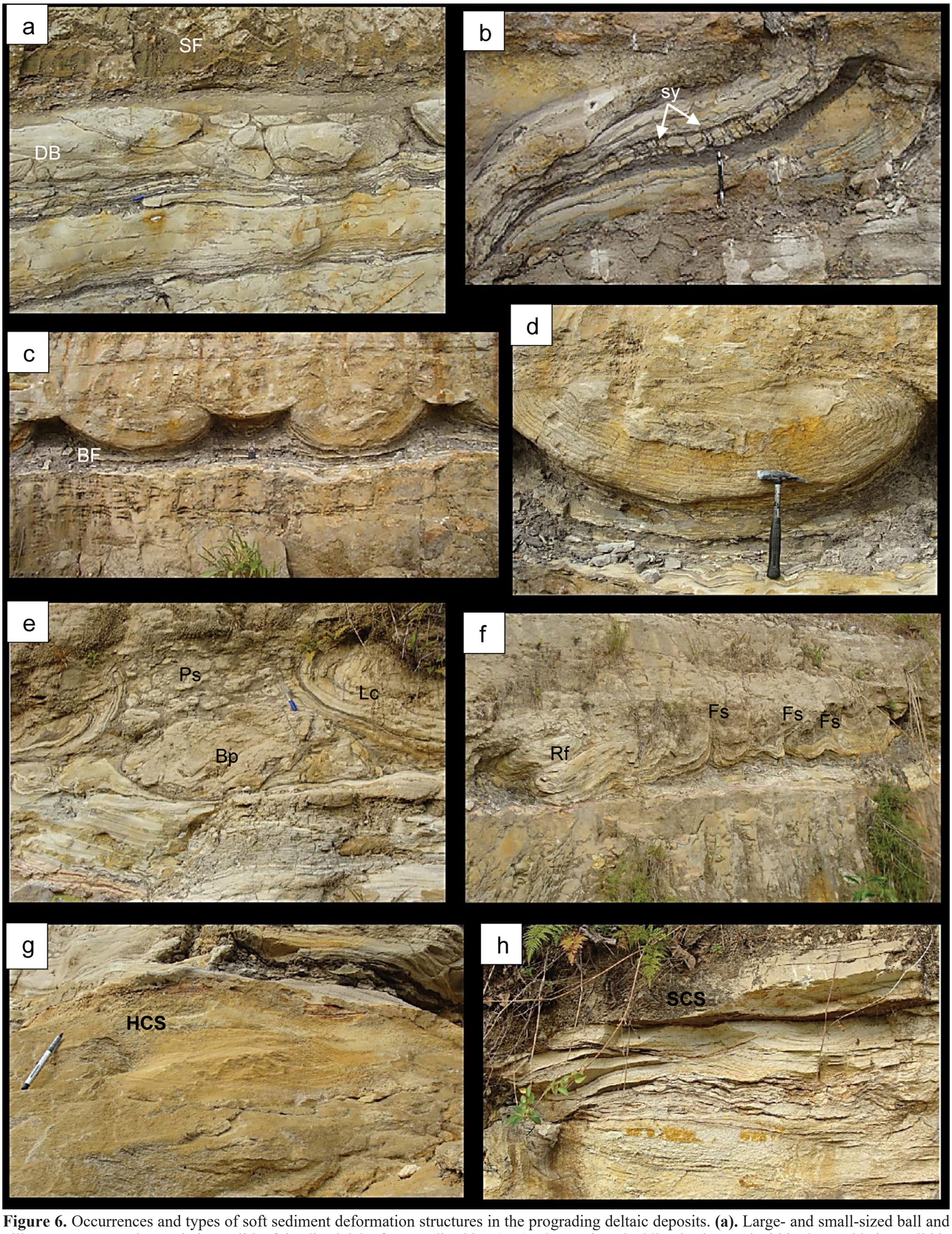


Figure 6. Occurrences and types of soft sediment deformation structures in the prograding deltaic deposits. **(a).** Large- and small-sized ball and pillow structures at the sandy heterolith of the distal delta front or distal bar (DB) whereas loop bedding is observed within the muddy heterolithic deposit. Syndepositional fractures (SF) occur in the horizontally bedded sandstone facies of the proximal delta front. **(b).** The flame-like structure shows syneresis cracks which may indicate changes in salinity. **(c).** Typical load cast structures with bulbous protuberances and water escape structures with upward directed crests formed within the bay-fill (BF) (muddy interval) and proximal delta front (sandy bed) deposits. **(d).** A close-up of the isolated load cast with deformed internal laminae forming a crescent shape that aligned with the morphology of the load cast. The interval is poorly burrowed, exhibiting low diversity, diminutive *Skolithos, Trichichnus* and *Planolites* ichnofossils. **(e).** Load structures consisting of load casts (Lc), ball and pillow structure (Bp) and pseudonodules (Ps) occurred within a mouth bar deposit. **(f).** Recumbent folds (Rf) in association with flame structures (Fs) are observed within the mouth bar deposit. The deformation displaced the muddy sediments of the bay-fill vertically or laterally. **(g).** Hummocky cross-stratified sandstone (HSC). **(h).** Swaley cross-stratified sandstone (SCS).

Flame structures commonly occur with load casts as upward-directed crests or tongue-like hydroplastic intrusions (Bhattacharya and Bhattacharya, 2010) of underlying muddy sediments into the load casts (Figures 6b-c, f). They are commonly referred to as water (fluid)-escape structures (Lowe, 1975; Alfaro et al., 2002). Flame structures are also formed in sandy units; they form large-scale isolated sandy flame-like structures with syneresis cracks as shown in figure 6b. Some poorly-developed flame structures are observed as convex-upward shaped flat crests (see Figure 6f). The flame structures consist of carbonaceous shale and alternating carbonaceous shale and siltstone. Internal lamination is observed in the muddy heterolithic layer. The well-developed flame structures are represented by vertical conduit, adapting to the external contour of the load casts (Alfaro et al., 2002). The flame structures have a maximum height of 80 cm and a width of 60 cm. They are similar to those described by Alfaro et al. (2002); Collinson et al. (2006).

Driving force: Flame structures are commonly associated with load casts. The reversed density loading which acts on a competent sediment creates a downward vertical force that controlled the formation of the load casts and simultaneously, an upward flow structure is formed by reduction in shear stress as the less competent substrate intrudes into the overlying layer (Mills, 1983; Oliveira et al., 2011) due to gravitational readjustment (Alfaro et al., 2002). Bhattacharya and Bhattacharya, 2010 suggested that the inclined nature of the flames (as observed in figure 6f), may indicate deposition along a gentle slope and/or subsequent modification of the flame structures by current (Brodzikowski and Haluszczak, 1987). The isolated sandy flame-like structure is formed from vertical displacement; the deformation is localized, and may have resulted from vertical shear (Owen, 1995).

Ball-and-pillow structures are very common in the study area, and are dominated by large-scale, isolated sandy structures with relatively undisturbed flat-tops and bulbous bases as described by Mills (1983). Small-scale ball-and-pillow structures are also observed (Figure 6a). The large-scale pillows have maximum width and height of about 80 and 40 cm, respectively, while the small-scale pillows have maximum width and height of about 30 and 16 cm, respectively. The pillows are embedded in silty sands, but are notably underlain by thin silty sands, unlike the pseudonodules. Most of the pillows have well-preserved internal structures and the laminations may be parallel, curved to concentric resembling the shape of pillows.

Driving force: Ball-and-pillow structures similar to other load structures resulted from the formation of reverse density mechanisms. Though, they are thought to be genetically different from pseudonodules and other load structures (Mills, 1983). Ball-and-pillow structures are distinguished from pseudonodules by the minor amount of finer sediments required in their formation (Blatt et al., 1980).

Pseudonodules are observed within the same bed layer with the other load structures. They occur as isolated masses of irregularly-shaped and fine-grained sand, embedded in an underlying deposit of contrasting density (Kuenen, 1958; Allen, 1982), which is of the finer matrix such as mudstone or siltstone (Figures 6k,l). Some of the sand nodules retain lamination; concentric and deformed laminations are observed. They are mostly small, but vary in sizes from 1–8 cm in height and 2–10 cm in width. (Mills, 1983).

Driving force: The formation of pseudonodules is similar to that of the loaded cast, but an advanced stage of it (Topal and Özkul, 2014). Here, the underlying fine- grained sediment is more viscous (Mills, 1983) or water-saturated (Kuenen, 1958) than that of the load cast. A reversed density loading is also postulated for the formation of pseudonodules (Blatt et al., 1980; Mills, 1983).

Ductile Disturbed Structures

Loop beddings are deformational structures characterized by bundles of laminae that are sharply constricted at intervals having shapes like loops or links of a chain (Cole and Picard, 1975; Gibling et al., 1985; Calvo et al., 1998). The loop bedding structure is commonly observed in finely-laminated sediments (Cole and Picard, 1975; Gibling et al., 1985; Trewin, 1986) and laminites (which consist of packages of diatomite laminae; see Rodríguez-Pascua et al., 2000) of lacustrine depositional environments, as well as carbonate microbialites (Martín-Chivelet et al., 2011).

In the study area, the loop bedding is observed in the muddy heterolithic units consisting of carbonaceous shale, siltstone, and a very fine-grained sandstone. This loop bedding is fairly exposed in about a 20 cm thick heterolithic unit (Figures 4, 5f); the length of the loops varies from 20 cm to 35 cm, and the thickness of the loop-bedded layer is about 5 mm thick. The loop bedding exhibits a boudinage-like morphology (which Martin-Chivelet et al., 2011 referred to as boudinage structures) and is not associated with any joints, faults, or microfaults.

Driving force: The loop bedding is formed as a result of ductile deformation (Rodríguez-Pascua et al., 2000) or reflects a plastic or hydroplastic deformation of partially lithified sediments (Martín-Chivelet et al., 2011). The loop bedding may be formed as a result of the stretching of unlithified or laminae bundles with low contrast incompetence, which is said to occur in response to seismic shocks due to the slow movement of extensional faults (Calvo et al., 1998; Rodríguez-Pascua et al., 2000; Martín-Chivelet et al., 2011). It can also occur because of the increase in the pore pressure of the unlithified sandy layers resulting in tension stress.

Recumbent folds occur in association with pseudonodules, load casts, and flame structures (Figure 6f). The folded sediments are developed in laminated sandy heterolithic units, and are characterized by isoclinal limbs and multiple circular hinges in a nearly horizontal axial state, exhibiting an antiformal and synformal geometries. The folds are not associated with faults or micro-faults. The height of the recumbent folds is between 30–60 cm thick, and the width is about 25 cm thick.

Driving force: Recumbent folds may be formed as a result of current (tangential) shear in semi-consolidated or liquefied sediments (Mills, 1983; Owen, 1996) or because of the lateral drag of a flowing sediment-rich fluid mass over water-saturated sands (McKee et al., 1962). The presence of alternating antiformal and synformal shapes, according to Owen (1995) suggests an interaction between a convection pattern of pore-water movement and upward movement of pore water that liquidized the bed.

Deformed laminations as explained by Moretti and Sabato (2007) include those deformed structures that are not categorized as load structures, loop beddings, recumbent folds, or brittle deformed structures in the study area. These structures are common in the sandy heterolithic units and exhibit varying morphologies. The deformed features may be folded, crumped, or stretched sediments. They occur in association with other soft-sediment deformation structures, either at the base or top of the bed unit.

Driving force: the deformed lamination may have been formed as a result of the current drag on partially-liquefied sand (Bhattacharya and Bhattacharya, 2010).

Brittle Deformed Structures

Syndepositional fractures occur at the lower and upper part of the very coarse to coarse-grained trough cross-stratified and massive sandstones. The highly dipping (>80°) joint sets display cross-cutting relationship; the major joints trend in a northwest direction (averaging 350°), whereas the minor joints indicate a westward direction (averaging 270°). No displacement is observed in the fractures (see Figures 5, 6a).

Driving force: Syndepositional fractures are the only soft-sediment deformation observed in the coarse-grained sandstone. No syndepositional faults, microfaults, or folds were observed. The cohesionless sandstone unit experienced only brittle deformation which suggests that the sediment was unsaturated with water during deformation (Mohindra and Bagati, 1996; Moretti, 2000). Brittle deformation arises from the cohesive behaviour of the sandstone when an increase in pore pressure is not strong enough to liquefy the sediment (Vanneste et al., 1999).

5. Discussion

5.1 Soft-Sediment Deformation Processes

A number of depositional environments such as shoreline, fluvial as well as the deep marine settings are commonly associated with soft-sediment deformation structures because of the slope gradient, high fluid discharge and rapid sedimentation (Pulham, 1989, Bhattacharya and Walker, 1991; Coleman, et al., 1983; Martinsen, 1989; Moretti et al., 2001; Oliveira et al., 2011). Figure 7 is a conceptual model (cross-section) showing the occurrence and distribution of soft-sediment deformation structures in a deltaic body (Nichols, 2009).

Soft-sediment deformation processes which include driving forces, deformation mechanisms and triggering mechanisms allow deformation to occur in a substrate (Oliveira et al., 2011; Owen et al., 2011). The driving forces are established for the various deformational structures discussed; they comprise reversed density gradient and shear stress. The shear stress may result in the current drag, increased pore pressure and pore-water movement in sediments. Liquefaction and fluidization are the main mechanisms of deformation in the study area; another mechanism is intergranular shear.

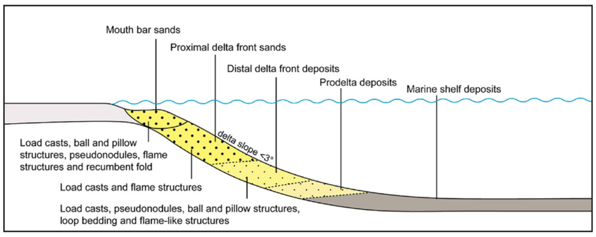


Figure 7. A cross-section across a delta lobe showing the occurrence and distribution of soft sediment deformation structures as observed in the study area (redrawn and modified after Nichols, 2009).

5.2 Deformation Mechanisms

Liquefaction and fluidization are the main deformation mechanisms in the study area. They are known to be the most common deformation mechanisms in unconsolidated, water-saturated sediments (Lowe, 1975; Allen, 1982; Owen, 1996; Moretti, 2000; Owen and Moretti, 2011). Liquefaction is known to occur when the grains are temporally suspended in a pore fluid either due to grain instability or excess pore water pressure (Lowe, 1975; Mills, 1983; Oliveira et al., 2011). Fluidization refers to the upward movement (i.e. fluid drag) of sediments' particles which results in dewatering or water-escape structures (Mills, 1983).

The load structures which include load casts, ball-and-pillow structures and pseudonodules are driven by gravitational forces, and are deformed mainly by liquefaction or reduction in shear stress of the sediments which make the sediments act as a viscous fluid. Flame structures are products of both liquefaction and fluidization. Fluidization occurs after liquefaction when the suspended grains have re-deposited, with a net movement of sediment particles downward. Subsequently, an upward flow of fluid through the fluidized sediments could result in the destruction of the original sediment fabric. Fluidization requires an external supply of fluid in an upward motion. The formation of recumbent folds is attributed to liquefaction (Mills, 1983) when the current shear (or lateral drag according to McKee et al., 1962) of a flowing fluid acts on the liquefied sand beds. Deformed laminations may be induced by liquefaction and loop beddings are probably induced by liquefaction and intergranular shear that occurred in cohesive, partially lithified, laminated sediments. Intergranular shear is a common deformation mechanism that results in brittle failure (such as syndepositional fractures) in cohesionless sediments (Owen, 1987) such as the coarse-grained sediments of the proximal delta front. Owen, 1995 suggested that the brittle failure occurs during a late stage in deformation when the deformation in the liquefied sediments is replaced by deformation through intergranular shear. Brittle failure is also common in partially liquefied sediment or sediment that was adjacent to liquefied sediment during deformation, such as sediment that was above a water table when liquefaction occurred (Bryant and Miall, 2010; Owen and Moretti, 2011).

5.3 Triggering Mechanisms

The triggering mechanisms for the deformation in the study are considered to be non-seismic or endogenic triggers (sensu Owen et al., 2011) based on the interpreted environment of deposition (EOD) and the relationship between sedimentary facies and the occurrence of soft-sediment deformation structures. The strata of the study area are interpreted as prograding river-dominated and wave-influenced deltaic deposits. Generally, deltaic systems are typically associated with soft- sediment deformation structures due to slope instability, rapid sedimentation, storm waves, and/or overloading mainly in the proximal region with a high-flow discharge (Pulham, 1989, Bhattacharya and Walker, 1991 and 1992; Coleman, et al., 1983; Martinsen, 1989; Bann and Fielding, 2004; Oliveira et al., 2011). The Itigidi and Ediba outcrops have a low palaeoslope gradient (<3º), so slope instability may not be an appropriate trigger for the deformation, and there is no evidence of slump structures in the outcrops. Overloading may be a triggering mechanism; it is induced by the rapid deposition of sands on a soft-substrate (Owen et al., 2011).

Although loop beddings have been considered as seismites by some authors (Calvo et al., 1998; Rodríguez-Pascua et al., 2000; Martín-Chivelet et al., 2011), for this research, the study area lacks evidence of tectonic signatures based on previous studies. Fractures are limited to joints; no structural displacements or folds occurred, neither is the basin known to be tectonically active. The soft-sediment deformation structures are not laterally extensive, and the deformed beds are not completely bounded above and below by undeformed units. The SSDS are only limited to the stratigraphic section at Itigidi.

6. Conclusions

This paper further reveals comprehensive soft-sediment deformation structures that are formed by endogenic triggers in a deltaic system. The encountered deformation structures are classified into three groups: i. load structure, which includes load casts, pseudonodules, ball-and-pillow structures and flame structures; ii. a ductile disturbed structure such as loop beddings, recumbent folds and deformed laminations; and iii. brittle deformed structure, which consists of syndepositional fractures. These deformations are formed mainly as a result of gravitational instabilities due to reversed density gradient and shear stress on liquefied or fluidized sediments. Non-seismic triggers such as rapid sedimentation, storm waves, and/or overloading are the major triggering mechanisms in the study area. The morphological, sedimentological, and stratigraphic characteristics of these soft-sediment deformations strongly indicate non-seismic or endogenic triggers. The presence of the ductile deformation structure – a loop bedding which has been regarded as a seismite (Calvo et al., 1998; Rodríguez-Pascua et al., 2000; Martín-Chivelet et al., 2011), occurs within an exclusively non-seismic triggered zone suggesting that the loop bedding is not exclusive to seismic trigger.

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