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**Notes**

# Palaeoproterozoic seismites (fine-grained facies of the Chaibasa Formation, east India) and their soft-sediment deformation structures

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**Abstract:** Metamorphosed shales, heterolithic deposits and sandstones build up the Palaeoproterozoic Chaibasa Formation in east India. The shales (referred here to as the fine-grained facies) comprising mudstone (clay and silt size) with some minor amounts of very fine to fine sandstone were deposited below storm wave base in a deep marine basin that simultaneously underwent tectonic activity. This fine-grained facies contains strongly deformed layers, intercalated between undeformed layers. Sedimentological analysis of the deformations indicates that they formed while still in an unconsolidated or slightly consolidated state, partly during and after sedimentation, but before being covered by younger sediments. The types of deformation structures indicate an earthquake-induced origin. Thus, they should be considered as seismites. The soft-sediment deformation structures in the seismites show a wide variety of shapes and other characteristics that appear to depend on their relative position to the epicentre of the earthquake.

The entirely siliciclastic Chaibasa Formation in east India is 6–8 km thick. It rests partly on an Archaean granitic basement, partly on the terrestrial Dhanjori Formation and is overlain by the Dhalbhum Formation (Bose *et al.* 1997; Mazumder & Sarkar 2004; Mazumder 2005) (Fig. 1). The age of the Chaibasa Formation has not been established, but the underlying Dhanjori mafic volcanics are 2100 Ma (Roy *et al.* 2002*a*), whereas the minimum age of the Dalma Lavas that conformably overlie the Singhbhum Group (Bhattacharya & Bhattacharya 1970), to which the Chaibasa Formation belongs, are 1600 Ma old (Roy *et al.* 2002*b*). Thus, the Chaibasa Formation is of late Palaeoproterozoic age (Mazumder 2003, 2005).

The rocks underwent several post-depositional deformation phases and greenschist to amphibolite facies metamorphism that turned the sandstones into quartzites, and the shales, sometimes with alternating intercalations of very fine sandstones and shales that are interpreted as turbidites (Bose *et al.* 1997), into mica schists (Naha 1965; Saha 1994).

## Characteristics of the Chaibasa Formation

The Chaibasa Formation is built of repeated alternations of quartzites (metamorphosed sandstones), heterolithic units (very fine quartzite/schist intercalations) and mica schists (metamorphosed mudstones) (Bhattacharya 1991; Bose *et al.* 1997; Bhattacharya & Bandyopadhaya 1998; Mazumder 2005; Fig. 2).

The sandstone facies consists of very fine to fine sandstones that locally may be muddy. The individual sandstone units range from 5–45 m in thickness, but show considerable changes in thickness laterally (Fig. 2). The individual layers in these units are up to 2.5 m thick, but also show lateral thickness variations. Many sandstones show cross-lamination. The units show tidal rhythms (Bose *et al.* 1997; Mazumder 2004) and the sandstone units are interpreted as shallow-marine deposits reflecting periods during which the tectonically active basin was either uplifted or had been filled by muds to such a thickness that the previously deep-marine character had changed to shallow-marine. Thus,

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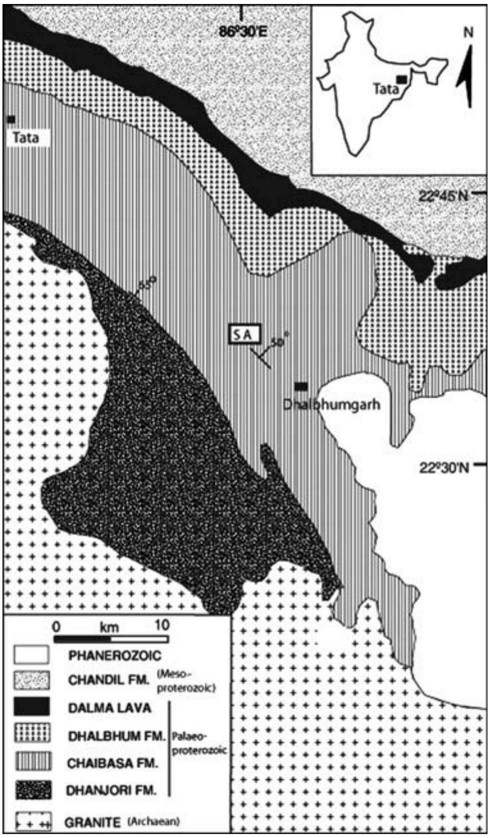


**Fig. 1.** Simplified geological map showing the disposition of the Chaibasa Formation and its adjacent formations (modified after Saha 1994; Mazumder *et al.* 2006); the Chaibasa and Dhalbhum formations together constitute the Singhbhum Group.

the deposition of the formation probably took several hundreds of millions of years to become several kilometres thick. It is therefore only logical that, under conditions of tectonic activity, shallow-marine and deep-marine nature intervals alternated (see Bose *et al.* (1997) and Mazumder (2005) for a discussion on the conditions under which the Chaibasa sequence was deposited).

The heterolithic facies consists of alternating layers of siltstone to very fine sandstone and shales. The relatively coarse layers of this unit range in thickness from 5–40 cm, most of them being 12–15 cm-thick. They show cross-bedding (including hummocky cross-stratification) or horizontal lamination and scour structures (25–30 cm deep, 1–1.5 m wide). The shale layers have variable thicknesses (2–60 cm) (Bose *et al.* 1997). We interpret this facies as transitional between the sandstone and the shale facies, being relatively distal but at a depth where the bottom was frequently affected by waves.

The fine-grained facies consists of metamorphosed shales (clay and silt fraction) with some minor amounts of quartzite (very fine sand fraction), often in the form of thin laminae but there are also some layers of quartzite. The individual units can reach thicknesses of over 100 m (Bose *et al.* 1997). No structures that might be ascribed to wave action have been found; in combination with the small grain size, this strongly suggests deposition of the sediments below storm wave base. Details of this facies, from which the soft-sediment deformation structures will be described and analysed, are provided in the next section.

Particularly in the upper part of the fine-grained facies, layers with soft-sediment deformation structures are abundant (Fig. 3) and show a wide variety of forms and complexity, including small clastic dykes, which Montenat *et al.* (2007) justifiably prefer to call 'sedimentary dykes', cone structures, load casts, pseudonodules, etc. (Fig. 4). The structures are interbedded with undeformed intervals, and have the same lithological composition as the undeformed layers. The lateral extent of the deformed layers is unknown because they cannot be traced beyond about a kilometre (which is, in fact, a remarkable distance for the occurrence of sedimentary deformations within one layer), because of limited exposure. It seems nevertheless that they have an even wider extent, because similarly deformed layers can be found far apart in more or less comparable stratigraphic positions, but lack of sufficient marker horizons makes precise correlation tentative.

## Characteristics of the fine-grained facies

The fine-grained facies of the Chaibasa Formation occasionally contains, apart from the predominant shales, some thin, very fine-grained sandstones. The shales mostly consist of units about 75 m thick without any differentiation in granulometry that can be observed with the naked eye, so that it is difficult to discern individual layers. The very fine sandstones rarely exceed 4 cm in thickness, and are more commonly measurable in millimetres (laminae). A few may reach a thickness of up to 50 cm. These relatively thick layers of very fine sandstone generally have an exposure-wide extent which gives them a sheet-like appearance. Probably they are turbidites (Bose *et al.* 1997) and show horizontal lamination and/or ripple cross-lamination (Fig. 5).

The shale without sandstone intercalations in this fine-grained facies is, in general, massive, mica-rich and dark in colour (Fig. 5). Although no true sandstone intercalations occur in this type of deposit, horizontal light-coloured sub-laminae of

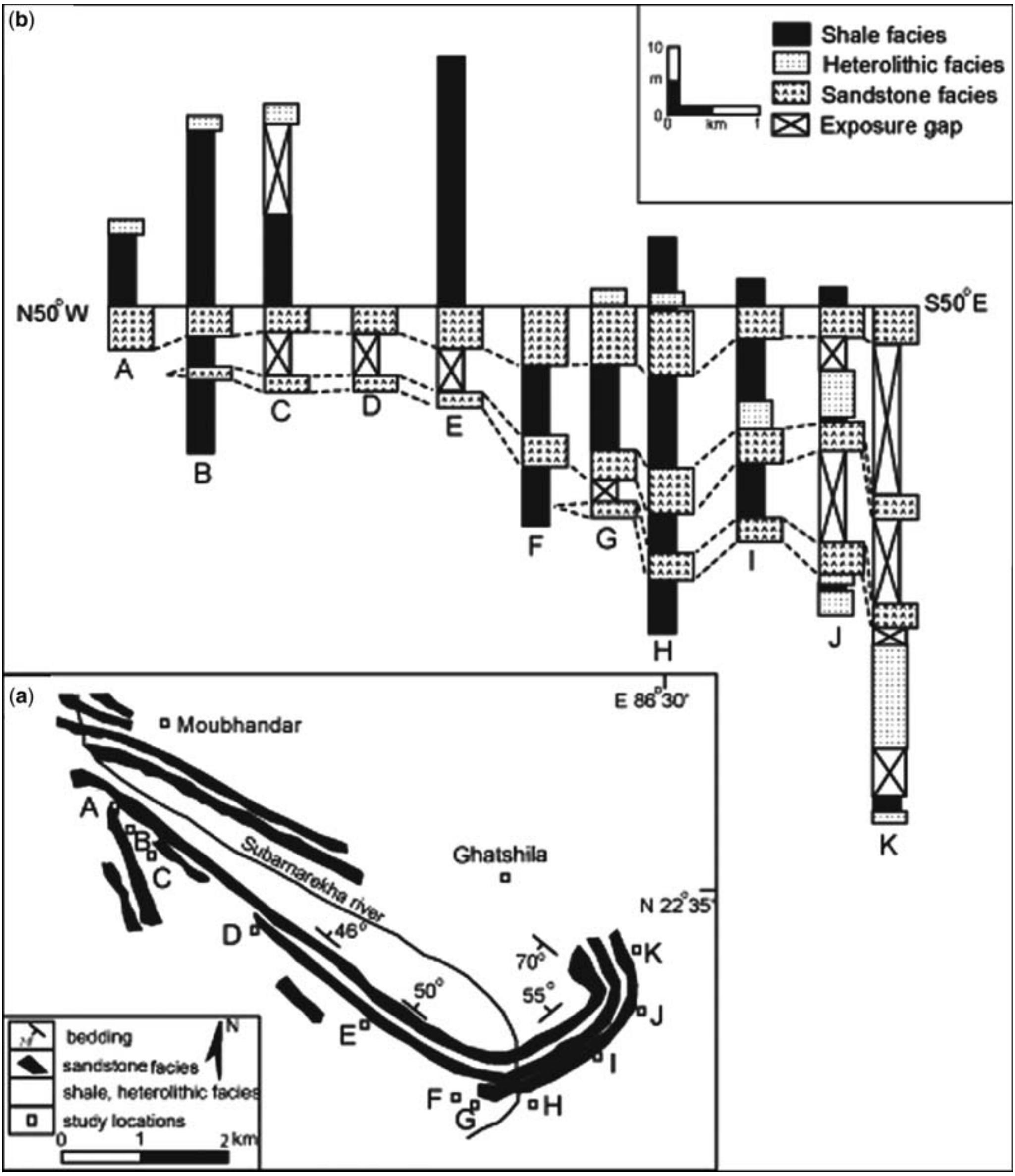


**Fig. 2.** The Chaibasa Formation. (**a**) Geological map showing lithological units of the Chaibasa Formation in and around Ghatshila (modified after Naha 1965). (**b**) Lateral and vertical lithofacies transitions of the Chaibasa sediments. Study locations are marked in Figure 2a (modified after Mazumder *et al.* 2006).

fine sandstone may be present locally. They show normal grading and light are in colour. These fine-grained units with sand/mud sub-laminae are separated from each other by sand-free, sheet-like mudstone beds with an average thickness of 3 cm, showing gradational bases and sharp tops. The ripple cross-lamination that is occasionally present (amplitude = 0.7–1.6 cm, wavelength = 2–6 cm) is distinctly asymmetric in profile with a tendency to climb (Fig. 6) and occurs in isolated ripple trains vertically separated by a set of horizontal laminae with an average thickness of 1.5 cm. Draping laminae on the lee faces of ripples are a common feature. These accretionary laminae tend to become horizontal, thus filling the ripple troughs completely and passing upwards gradually

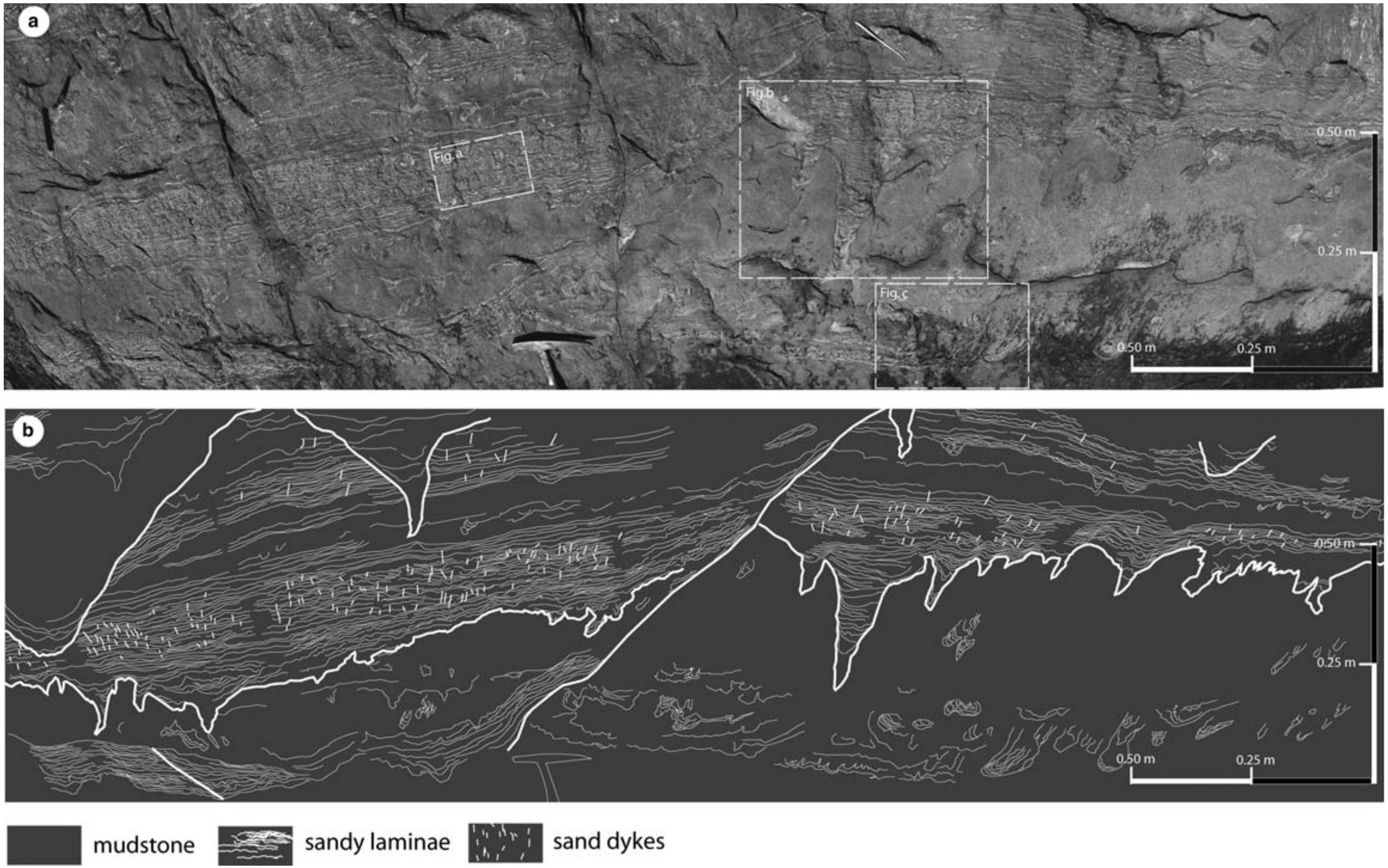


**Fig. 3.** Photo-mosaic (**a**) and derived drawing (**b**) of the soft-sediment deformation structures in part of an exposure of the Chaibasa fine-grained facies. The sequence of events resulting in the complex deformations is explained in Figure 16. The exposure (more or less perpendicular to the stratification) shows, among other deformations; a synsedimentary fault (Fig. 10); cone-shaped structures (Fig. 8); contorted mud balls (Fig. 11); pillow structures (Figs 7 & 14); load casts and pseudonodules (Figs 12 & 13). Details of the deformation structures in the areas marked by the rectangular boxes are presented in Figure 4.

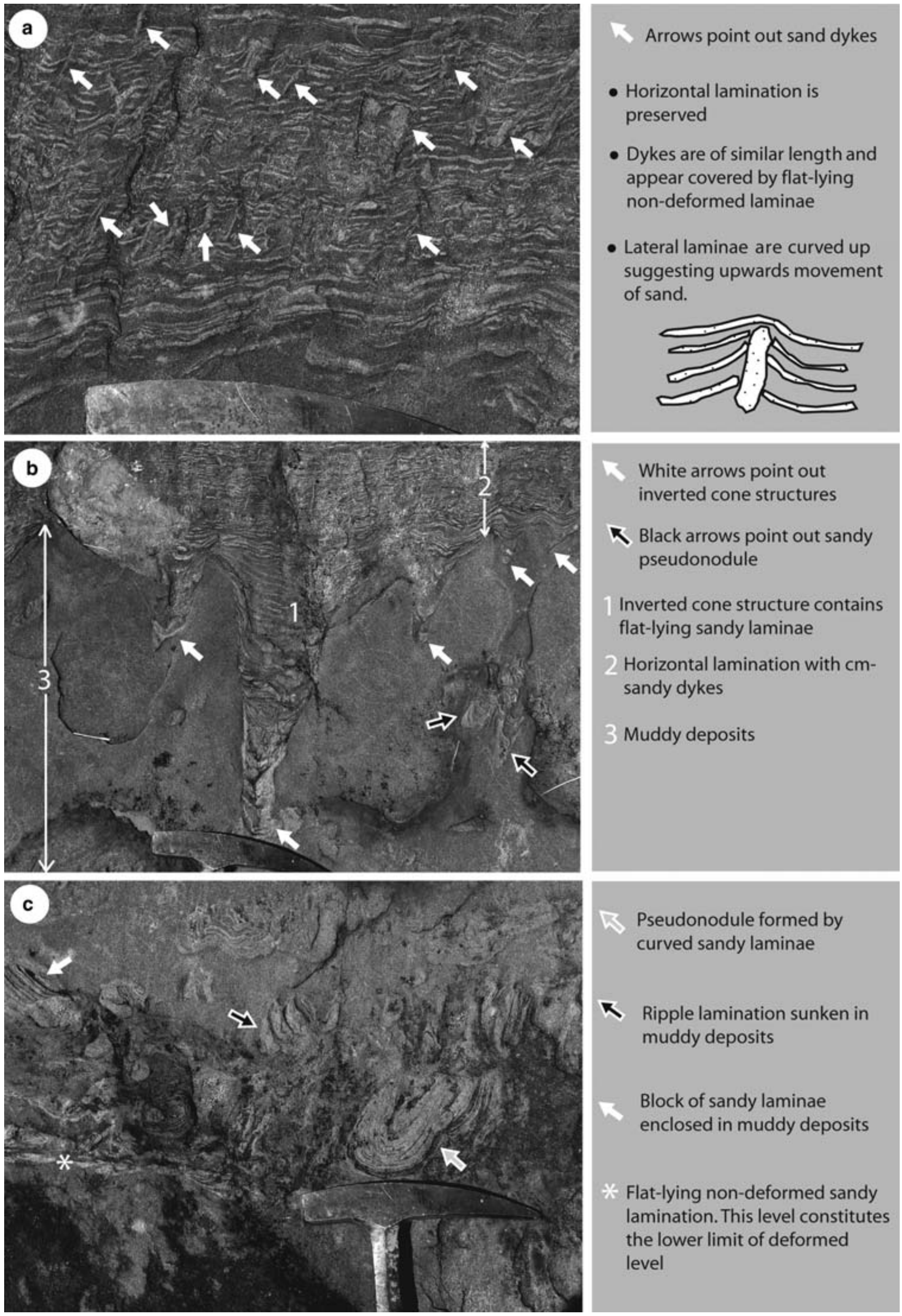


**Fig. 4.** Detailed deformation structures of the areas marked by rectangular boxes in Figure 3 (see hammer heads for scale). (**a**) Irregular lamination, contorted by numerous small-scale dykes of fluidized very fine sand; (**b**) Inverted cone structures with several other small-scale deformation structures; (**c**) Chaotic bed with sagged ripples and pseudonodule.

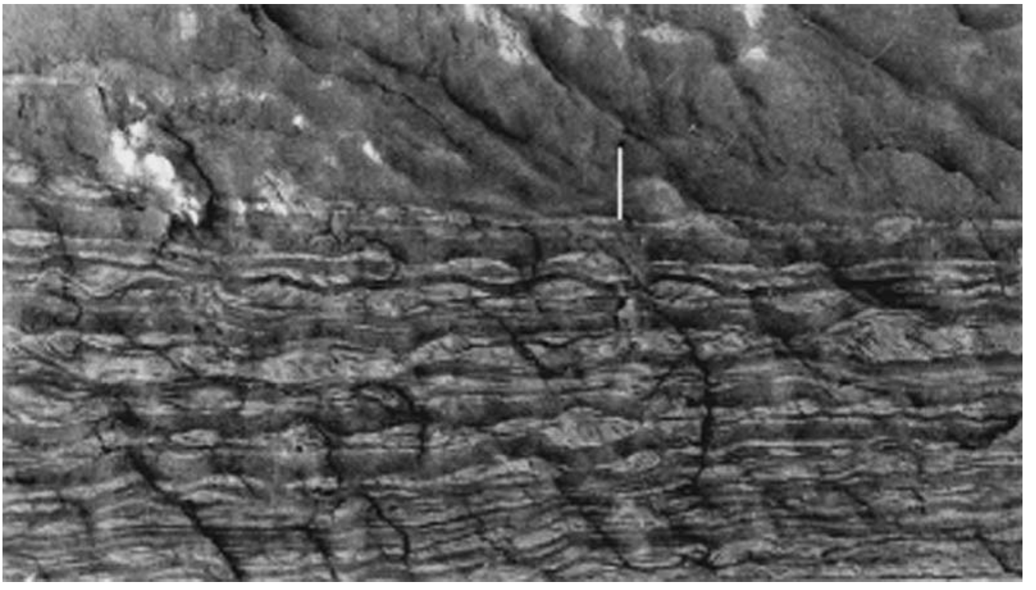

**Fig. 5.** Chaibasa fine-grained facies with intercalated thin laminae of very fine sand (light-coloured); the sandstone laminae show either horizontal or ripple cross-lamination. At places, the ripple trains are vertically stacked. Matchstick 4.2 cm.

into horizontal laminae. The individual layers, that in cross-section show ripple cross-lamination throughout, all have sharp, often fluted bases, and pass gradually into the horizontal laminae just mentioned (fig. 12d in Bose *et al.* 1997). The sets showing ripple cross-lamination at the base and horizontal lamination at the top range in thickness from 5 to occasionally 45 cm and are typically 15 cm thick. These thicker units are vertically separated from one another by sheet-like shale layers with an average thickness of 3 cm. These shale layers without coarser material are gradational but their tops are sharp.

The fine-grained facies is completely devoid of any wave-generated structure and is therefore likely to be a relatively deep-marine sediment formed below the storm wave base (Bose *et al.* 1997; Mazumder 2002, 2005). The relatively thick units that are separated by shales without thin sandy intercalations are probably products of episodic waning flows, possibly contour currents (an interpretation based on the various contributions to Rebesco & Camerlenghi 2008). Asymmetric ripples (Figs 5 & 6) imply traction. The draping laminae on the lee side of the ripples indicate weakening of the currents to the extent that ripple migration was no longer possible, but that sweeping of the suspension fall-out to the current shadow zone on the lee side of the ripples was still possible (Bose *et al.* 1997). The gradual upward transition from

**Fig. 6.** Strongly asymmetric ripples within the Chaibasa fine-grained facies; note vertically stacked and sagged ripple trains within alternating shale and fine-sandstone laminae (pen length 12 cm).

draping laminae to horizontal laminae indicates a further slow decline in tractive force, eventually allowing settling of the suspension load. Grading of the fine sandstone laminae corroborates this contention. With a deep-sea distal current under a persistent suspension cloud, the depositional system appears to be similar to that of the hemiturbidites described by Stow & Wetzel (1990). The sand-free shales in between the thicker units show normal grading and are considered as indigenous pelagic deposits settled from suspension between events that whirled up sediment from the sea floor, or that supplied suspended load by contour currents, low-density mass flows or any other type of hyperpycnal flow.

The sediment is very fine-grained, without any coarse material, ruling out supply by a nearby river. Possibly the sediments had their transient source in the distal shelf. Abundant slump scars, bevelled bed-edges and slump folds in the Chaibasa heterolithic facies, detailed in Bose *et al.* (1997) and Mazumder (2002), support this interpretation. The persistent alternations of finer- and coarser-grained laminae may be due to basal shear that prevented mud floccules to settle unless their diameter exceeded a threshold value (Stow & Bowen 1978, 1980).

## Objectives and approach

The objectives of the present study were to provide an inventory of the various types of soft-sediment deformation structures in the Chaibasa shales (fine-grained sediments deposited in a deep-marine facies contain much less, and much less variegated, soft-sediment deformation structures than the Chaibasa shales) and to analyse their genesis. This analysis was carried out following the approach suggested by Owen (1987), who pointed out that such a genetic analysis should be based on the reconstruction of: (1) the timing of deformation, and (2) the deformational mechanism(s).

### *Timing of the deformational processes*

It has been well-known for several decades that deformations cannot only take place when a sediment has been covered by younger layers (post-depositional deformations), but also during the depositional process (syndepositional deformations) or after deposition but before the sediment is covered by a younger layer meta-depositional deformations (Nagtegaal 1963; Allen 1982; Owen 1995).

Whether deformations are syndepositional or meta-depositional, is of great importance for their genetic interpretation. In the first case, seismic influence, may easily affect the subaqueous sediments that are still being formed and that have not started to consolidate. On the other hand, meta-depositional deformations, which develop most commonly within about half a metre from the sedimentary surface, but which may occasionally (depending on the deformational mechanism and the forces involved) also affect deeper sediments, are related to surficial sediments that had already undergone some degree of consolidation. Relatively few mechanisms are capable of meta-depositional deformation, but earthquake-induced shocks can generate recurrent horizontal shear strain triggering a thixotropic effect in unconsolidated deposits, resulting in an instantaneous segregation of the liquid from the solid sedimentary phase (Montenat *et al.* 2007).

### *Analysis of the deformational mechanisms*

Because the Chaibasa shales have been deposited in an active tectonic setting (Mazumder 2005), the analysis of the deformations (Mazumder *et al.* 2006) was directed first at the distinction between tectonic and non-tectonic (sedimentary) deformations. These may be interrelated, as tectonic shocks can lead to pressure gradients that induce the deformation of surficial, non-consolidated sediments. It is well-known that particularly the topmost sedimentary layer, if consisting of material that is susceptible to shock (earthquake)-induced deformations, can become strongly disorganized. Such layers, which tend to show more or less similar deformation structures over large distances, are known as ‘seismites’.

The term ‘seismite’ has been introduced by Seilacher (1969) for a layer (or a set of layers) with earthquake-generated soft-sediment deformation structures (Neuendorf *et al.* 2005). Seismites are probably the best proof of earthquakes in the history of Earth, although their recognition is difficult, partly due to the fact that unconsolidated sediments that are easily deformed by earthquake-induced shocks, are commonly also susceptible to disturbances not related to seismicity. Yet the detection of unmistakable fingerprints of seismic events in ancient successions is, however difficult, of paramount importance for insight into basin dynamics (Seth *et al.* 1990).

Sedimentological evidence of earthquakes other than in the form of seismites is scarce in the sedimentary record and the interpretation of such earthquake-induced phenomena is often even more questionable than the recognition of seismites (Seilacher 1984). For instance, faulting at a basin’s margin may result from earthquakes that also induce mass failure, but subsequent mass flows facilitated by the fault-created steep slope need in no way be directly related to seismic activity. Therefore it is difficult to separate the two genetic groups of mass-flow deposits within a single lithological unit.

Recognition of seismites is hampered by the fact that earthquakes are just one of the trigger mechanisms that induce soft-sediment deformation. Other mechanisms are: (1) pressure changes resulting from, for instance, the passage of storm currents, breaking waves or flood surges; (2) oversteepening of slopes; (3) gravitational density flows; and (4) rapid loading of sediments (Van Loon & Wiggers 1976*b*; Allen 1982; Owen 1987, 1995, 1996; Van Loon & Brodzikowski 1987; Maltman 1994; Rossetti 1999; Rossetti & Goes 2000; Jones & Omoto 2000; and refs therein). In addition to these processes, differential compaction of sediments can also create soft-sediment deformation structures, although the process may be slow.

The following combination of characteristic of layers with abundant soft-sediment deformation structures is commonly considered diagnostic for an earthquake origin (Sims 1973, 1975; Scott & Price 1988; Moretti & Tropeano 1996; Obermeier 1996; Alfaro *et al.* 1997; Rossetti 1999; Jones & Omoto 2000; Bose *et al.* 2001): (1) the restriction of deformational structures to discrete stratigraphic levels; (2) lateral continuity of the deformed character over large distances; (3) recurrence of such deformed layers through time; (4) a consistent deflection of palaeocurrent trends from their usual pattern across the deformed levels; (5) confinement between undeformed strata (or strata with a distinctly different origin of the deformations); and (6) a preferred association with wedges of intraclastic breccias, conglomerates and massive sandstones.

Our investigation was directed at finding out whether specific deformed layers in the Chaibasa shales should be considered as seismites. We analysed all types of soft-sediment deformation that were encountered in the Chaibasa shales, and interpreted their syn- or meta-depositionally nature. Finally, we reconstructed their genesis to find a possible seismic origin.

In the following sections, the soft-sediment deformation structures in the fine-grained facies of the Chaibasa Formation will be described. We group these structures according to their syndepositional or meta-depositional origin. As the difference between syndepositional and meta-depositional deformations, and the way in which they can be distinguished from one another, has been described extensively in the literature (Nagtegaal 1963; Allen 1982; Owen 1995), we will not detail the precise arguments for our interpretation of the timing here.

## Syndepositional deformation structures

Deformation structures in the Chaibasa shales that must have formed syndepositionally show a wide variety of shapes and other characteristics. On the basis of their forms they can be grouped into the following types.

### *Pillows*

Several levels in the fine sandy intercalations that are present within the fine-grained facies are characterized by pillow-shaped deformations that can be traced laterally over tens to hundreds of metres, and that occur within these levels without significant lateral interruptions. They are more common in the thicker sandy layers. The pillows have obtained a U-shape because they have sunk deep into the underlying mud, commonly having parallel vertical contacts, whereas their bottoms are rounded (Fig. 7). Their internal lamination tends to follow the outer shape. The size of the pillows (max. height 35 cm; max. diameter 12 cm) changes gradually laterally within a single level. Within individual structures, the degree of deformation decreases upward, and younger lamina sets may truncate the underlying older sets. This implies repeated sagging during ongoing sedimentation, which proves the syndepositional character.

Pillow structures like these have commonly been interpreted as due to loading of denser sediment in a two-layer system (Allen 1982 & references therein). The loading process, which should be considered as a response to unstable density contrasts or lateral variations in load (Owen 2003), may also be triggered by mechanisms such as differential

**Fig. 7.** Pillow beds consisting of predominantly clay- and silt-sized sediments, with thin laminae of very fine sand (light-coloured) that make the deformations better visible. Loading has resulted in plastic deformation, and very local fluidization has made the original lamination disappear in places.

compaction due to rapid sediment accumulation, wave activity inducing stresses and ice growth in glacigenic sediments (Ricci Lucchi 1995). A seismic origin for such pillow structures has also been suggested (Montenat *et al.* 1987; Cojan & Thiry 1992; Guiraud & Plaziat 1993; Rodríguez-Pascua *et al.* 2000).

The Chaibasa pillow structures developed in sediments that do not show sufficient density contrasts that could cause loading without a forceful trigger mechanism, because the density of water-saturated fine sands is barely higher than that of water-saturated mud (Anketell *et al.* 1970). The layers in which the pillows developed were below storm wave base (Bose *et al.* 1997), so that wave-induced stresses could not have been the cause. Ice growth in these deep-sea sediments can also be excluded, so a syndepositional seismically-induced origin is the most likely. The formation of the pillows was probably initiated by shock-induced liquefaction of some specific layers within sets of sandy layers, followed by an upward movement (into the direction of the lowest pressure) of the liquefied silty fine sand.

### *Cone-shaped contorted laminae sets*

A completely different type of soft-rock deformations consist of inverted cone-shaped structures (Fig. 8). The cones are filled with concave upward laminae of shale and silty sandstone in multiple sets. The younger sets truncate the older ones and the degree of concavity of the laminae decreases in an upward direction from one set to the next. These alternations of fine sandy and muddy laminae are generally absent outside the cones (Fig. 8).

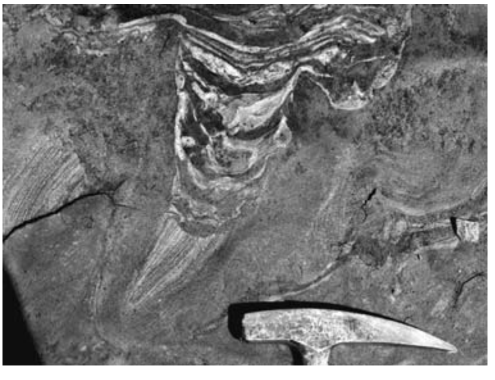

**Fig. 8.** Inverted cone-shaped contorted laminae sets, as seen in a section almost perpendicular to the bedding plane; note that the cones are filled with concave upward laminae of shale and silty sandstone in multiple sets. The morphological resemblance with ice-wedge casts (Montenat *et al.* 2007) makes it likely that cracks were formed in the seafloor (possibly by extension tectonics) that became filled with fine-grained sediment supplied by traction; this process must have taken place during widening of the cracks.

The configuration and the internal structure of the cones suggest that cracks formed on the subaqueous mud surface and that a current laden with mud and fine sand filled these cracks (Cozzi 2000). This can occur when a hydraulic jump is created in the underflow on top of the crack so that the particles can settle within the cracks.

The erosion at the base of each lamina set and the upward decreasing deformation rate suggest repeated enlargement of the cracks. Consequently, the crack-filling laminae sank plastically down into the cracks, thus creating new accomodation space. Because this process took place repeatedly, the oldest (lowermost) set of laminae was exposed to the largest number of deformation phases and thus became most deformed, whereas the youngest (uppermost) set suffered no deformation at all. The origin of the cracks is not clear. Cozzi (2000) mentions some cracks in a shallow-marine environment but does not provide a good explanation. In addition, the cracks described by Cozzi are somewhat different from those in the Chaibasa Formation. Considering the repeated enlargement of the cracks and the successive infillings, a series of successive events must have taken place within a relatively short time. The interpretation is that earthquake-induced shocks have been the trigger mechanism; the main reason for this interpretation is that no other mechanisms are known that might result in such cracks. Opening of cracks, filling up with sediment, reopening, etc. under subaquous conditions has been mentioned from other recently seismically active areas; the process has been detailed by Vachard *et al.* (1987) and Montenat *et al.* (2001).

An alternative explanation might be that the sea floor was deformed by synaeresis cracks (Plummer & Gostin 1981), but these structures are not well understood (Hounslow 2003) and apparently involve changes in volume of clay minerals induced by salinity changes (Collinson & Thompson 1982), expulsion of fluids from colloidal suspensions, or removal of water from mud layers that are in contact with brines produced by evaporation (McLane 1995). The Chaibasa shales do not show any indications of such conditions: on the contrary, the conditions seem to have been very constant over long periods (the whole succession most probably was deposited during a time span of several hundreds of millions of years), apart from some short-lasting events such as earthquakes. Tanner (2003) mentions that intrastratal synaeresis can be caused by earthquake-induced ground motion. This is consistent with Pratt (1998), who states that cracks in subaqueous argillaceous sediments may result from earthquake-induced dewatering. Another alternative is that the cracks are diastasis cracks (Cowan & James 1992), which have been described also from bathyal conditions (Shiki & Yamazaki 1996).

## *Infilled depressions*

One shale layer contains several depressions that all show at least one steep wall in cross-section (fig. 4 in Mazumder *et al.* 2006) The exposed depressions all have more or less identical sizes, on average 48 cm deep and 36 cm wide (the length cannot be established, as no 3-D exposures are present). The depressions represent erosional forms, as the laminae of the adjacent shales (and some very thin fine sandstones) are truncated by the margins of the depressions, which are themselves filled with comparable alternating laminae of shale and very fine sandstone (Fig. 9) that partly, particularly in the deepest part of the depressions, end abruptly against the depressions' walls and partly, and ever more clearly, follow the slope of the walls as they have a higher position in the infilling. None of the depressions has a symmetrical shape as the lower parts of the walls are steeper at one side than at the other side.

The laminae immediately underlying the structures do not exhibit any significant degree of deformation, so is must be concluded that the depressions are erosional features that later were gradually filled up; first with laminae that could not be deposited on the steep lower walls but (when the ongoing sedimentation had partly filled the depression until the level that the walls became more gently inclined) later following the then sedimentary surface within the depression (following the inclined walls and laterally continuing outside the depression). The individual sedimentary laminae/layers in the depressions are relatively thick in the centre of the depression and thin towards the margin, indicating that the depressions acted as a sediment trap.

**Fig. 9.** Cross-section through infilled depression within the Chaibasa shale (hammer length 42 cm). The infilling points at a synsedimentary filling of the depression.

The geometry of the structures resembles that of the meso-scale scours described by Sarkar *et al.* (1991) from a near-shore region, and described by Cheel & Middleton (1993) from an offshore environment. The formation of such erosional forms in a cohesive mud substratum is also known from turbidity currents (Mutti & Normak 1987). It is unlikely, however, that turbidity currents were involved in this case, as all turbidites found in this part of the Chaibasa Formation are thin and fine-grained, whereas the erosional structures have a depth of various decimetres.

Although the infilled depressions have many characteristics in common with 'normal' channels (the occurrence of one steep and one less inclined side seem to point at meandering channels), it is remarkable that they must be ascribed to relatively high-energy events, whereas such events left no other traces in the sediment. The infilling of the structures also lacks any sign of a lag deposit, which makes these depressions even more enigmatic. It was interpreted earlier that at least some of the depressions could have formed due to the lifting of sediment blocks by the successive phases of under- and overpressure during the passage of tsunami waves (Mazumder *et al.* 2006), but renewed analysis of the depressions showed features (e.g. the succession on the less steep side of these depressions) that contradict this view.

Since the sedimentary unit in which the depressions occur was not significantly deformed (only some minor deformations were observed but these could not unambiguously be ascribed to a seismic shock), this unit should, however, not be called a seismite.

## *Sagged ripple trains*

Vertically stacked and sagged ripple trains occur within a succession of alternating shale and fine-sandstone laminae (Figs 5 & 6). The sagging is generally more pronounced in the lower part of the stack. The laminae of very fine sandstone in the shale underneath the sagged ripple trains roughly follow the undulating base of the sagged ripples.

Ripple trains can repeatedly form and sag (Dzulynski & Kotlarczyk 1962; Dzulynski & Slaczka 1965; Dzulynski 1996). The rippled sediment under investigation has evidently been deposited in a number of phases, but before the underlying mud could release the excess pore water (Dzulynski & Slaczka 1965; Reineck & Singh 1980).

The sagging of the ripple trains must be due to partial liquefaction of the underlying mud. The liquefaction may have been caused by an earthquake-induced shock, but may, alternatively, be due to simple pressure resulting from differential loading (Lowe 1975; Jones & Omoto 2000).

## Metadepositional deformations

The typical metadepositional deformations in the Chaibasa shale show a wide variety of types. They include: (1) penecontemporaneous faults; (2) contorted mud balls; (3) pseudonodules; (4) chaotic layers; and (5) a collapse structure.

### *Penecontemporaneous faults*

Fault planes, commonly upthrusts, up to some 55 cm long and dipping 5–12°, occur in the upper part of the Chaibasa shales, without affecting the under- and overlying layers (Fig. 10). They show the following characteristics: (1) progressive squeezing of the laminae towards the fault planes; (2) a spreading in orientaton of the faults within each individual layer; (3) evidence of liquefaction, mostly at the top of the fault planes, and a decreasing degree of liquefaction downwards; and (4) planar or sigmoidal fault planes that consist of masses of shale, as commonly found along shear planes.

The position of the faulted layers intercalated with unfaulted layers is evidence of a metadepositional origin, and the faults themselves indicate that some compaction had taken place before the faulting, but differential compaction apparently did not play a role, as no indications for this process are present (the lateral changes in thickness of laminae in the direct neighbourhood of the fault are a result of shearing). The upthrusting of the slightly compacted sediments indicates that some lateral stress must have been exerted. A shock, presumably due to an earthquake, seems the most logical cause of such a stress, and thus a seismic shock must be considered as the most likely explanation for the faulting.

**Fig. 10.** Penecontemporaneous thrust within the Chaibasa shale (after Mazumder 2005). The fault plane shows in cross-section a characteristic, slightly sigmoidal shape.

### *Contorted mud balls in massive sandstone*

Some of the rare massive layers of very fine sandstone that are thicker (*c.* 30 cm) than most other sand layers in this fine-grained facies of the Chaibasa Formation, occasionally contain 'floating' balls of shale (Fig. 11) that internally show contorted laminae of silty to fine sandy material with upturned margins. The structures obviously consist of sediment that sank into the underlying layer and eventually became detached from the overlying parent mud layers (pseudonodules).

The loading of mud balls into sand is difficult to explain unless the sand also had a relatively low density. This may have been the result of a strong hydrostatic stress that developed within the sandy layers under an impermeable mud cover (Schwab & Lee 1988). The configuration indicates late-stage breakup of the overlying mud layers and also that liquefaction of the fine sand probably destroyed any primary sedimentary structures in the fine sands, reflected now by the massive fabric of the sandstone layers.

The structures do not provide a clue for the trigger that resulted in the liquefaction, but a suddenly increased hydrostatic stress seems the most likely. The quiet depositional environment, reflected by the fine-grained character of the sediments, does not provide any clue for such a sudden increase in

**Fig. 11.** Contorted balls of shale within layers of very fine sandstone of the fine-grained facies of the Chaibasa Formation. (coin diameter 2.5 cm). The balls of shale still show internal lamination indicating that the, then, slightly consolidated mud was deformed plastically, whereas the sandstone does not show any internal structure, suggesting that the sand was liquefied during the deformational process.

hydrostatic stress, particularly since the floating mud balls are the only deformation structures in this part of the section. An 'outside' trigger therefore seems the most probable; this may, considering the syn-sedimentary tectonics in the basin, well have been an earthquake-induced shock.

## *Levels with small-scale pseudonodules*

The original bedding in the shales is locally indicated by a high concentration of small crescent-shaped pseudonodules of very fine sandstone. The frequency of these pseudonodules diminishes downwards. Their lateral distribution is irregular: they form clusters with an average thickness of 8 cm (Fig. 12). The form of the pseudonodule clusters suggests lateral variability in the resistance of the depositional substratum. Lateral variation in thickness of the original sandy layer is highly unlikely, as such lateral variations are observed nowhere in this facies.

The pseudonodule levels can be explained satisfactorily only by assuming that a layer of very fine sand was deposited on top of a water-saturated mud unit of lower density. The sand started to sink down in the mud, forming load casts and eventually pseudonodules. The vertical distribution of the pseudonodules can be explained by increasing compaction and resistance to deformation of the muddy substratum with depth. This resistance is also reflected in the upturned edges of many pseudonodules. As the mud showed apparently some resistance, it may have been slightly compacted (which is also suggested by the shape of the pseudonodule clusters). Compaction, however little developed, would under quiet conditions have led to gradual loading of the sand layer, resulting (after continued loading) in a series of pseudonodules beside each other (Van Loon & Wiggers 1976*b*), but not in clusters of pseudonodules that contain two or even more pseudonodules on top of each other. It seems therefore likely that the loading process was not a quiet one but rather the result of a sudden, uncommon disturbance. Such a disturbance needs a trigger. This was most likely an earthquake-induced shock, as experiments (Kuenen 1965) have shown that vertical 'accumulations' of saucer-like pseudonodules can form as a consequence of vibrations affecting the sand layer and the overlying sediment (Reineck & Singh 1980).

**Fig. 12.** High concentration of small saucer-shaped pseudonodules of very fine sandstone within the Chaibasa shale; the distribution of the pseudonodules is irregular and they form elongated clusters. Pseudonodule clusters of this type have been produced experimentally, exclusively by vibration (Kuenen 1965).

## *Chaotic layers*

The internal structure of some shale layers, which are up to 70 cm thick, is chaotic. These layers look tabular in form in outcrop. They show, between their sharp lower and upper contacts with undeformed layers with the same granulometry (and thus porosity and, particularly, permeability) deformation structures (well visible because of the presence of laminae of very fine sandstone) such as convolutions, pillows (both mainly in the upper parts of these layers) together with load casts and pseudonodules (mainly in the lower part)

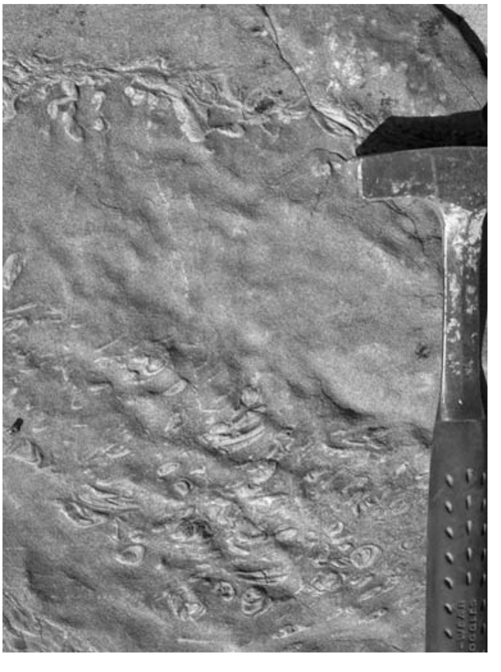

**Fig. 13.** Chaotic shale layer (section oblique to the bedding plane) with loadcasts and a pseudonodule cluster.

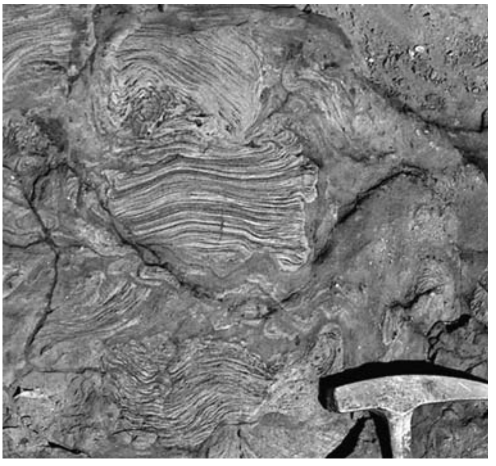

Fig. 14. Chaotic Chaibasa shale layer with exceptionally well-developed convolutions and pillows.

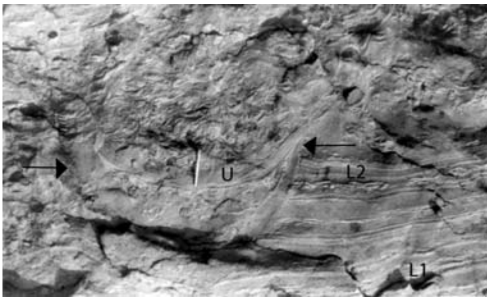


Fig. 15. Collapse structure (shown by arrows) within the fine-grained facies. Note the difference in deformation between the lower (units L1 and L2) and upper (U) beds. In both units, the intensity of deformation decreases upwards.

(Figs 13 & 14); these load structures have rotated in such a way that they are now upside down. Apart from these plastically deformed structures, liquefaction took place locally. Comparable deformations have been described frequently from seismites (see fig. 17a in Merriam 2005).

The individual layers with the chaotic deformations must have been affected by a stress that had the same intensity everywhere, as there are no significant lateral changes in the intensity of the deformations. The undisturbed layers below the deformed layers suggest that the deformation affected only the surficial sediments; the undisturbed layers on top of the disturbed layer indicate that the deformation must have taken place before the affected layer was covered with younger sediments. This suggests that the deformation took place because entrapped pore water was forced to find a way out (Williams 1966, 1969, 1970; Owen 1995). The mechanism that triggered this process cannot be reconstructed on the basis of the structures themselves but it cannot have been the weight of an overburden, as the overlying sediments are undeformed. It must therefore be deduced that, like in the case of the mud balls described above, an 'outside' mechanism, such as an earthquake-induced shock, has been responsible.

### *Collapse structure*

Two layers with a locally strange contact (Fig. 15) are present in places where the shale contains some fine sandstone (Mazumder *et al.* 2006). The lowermost layers are most intensely deformed, and their internal layering follows exactly the bending of the base of the overlying unit; this proves that the structure cannot be ascribed to erosion (scouring). The lower layer shows chaotic deformation, whereas the overlying layer shows no internal deformation, except for some vertical fractures/cracks (the origin of these fractures is still not clear, but the fractures are comparable to those described above about the penecontemporaneous faults and the block-shaped depressions). A slab of the upper layer penetrates the chaotic mass of the lower layer.

The chaotic character of the lower layer must be due to a stress field that forced water-saturated sediment to move laterally, thus creating some kind of underpressure that caused collapse of the overlying layer. As the intensity of the deformations in the lower layer diminishes in an upward direction, and as there is no internal deformation within the upper layer, is must be deduced that the force causing the deformation came from beneath. The force was apparently not strong enough to cause deformation upwards of the contact plane between the two layers, which seems to have served as a shear plane. This suggests that the chaotic deformation took place in a confined state, underneath the sedimentary surface.

## Discussion of the possible trigger mechanisms

Almost all structures described above can be generated under a wide variety of conditions. However, the clusters of pseudonodules, with several saucer-shaped specimens above each other, have only been produced experimentally as a result of vibrations (Kuenen 1965). Such vibrations occur under natural conditions almost exclusively as a result of earthquakes. These pseudonodule clusters are therefore strong evidence for earthquakes as the responsible mechanism. This interpretation is supported by the fact that the area at the time formed part of a tectonically active basin. The occurrence of synsedimentary upthrusts can be explained only by lateral pressure, for which – under the environmental conditions that seem to have lasted during sedimentation of the entire fine-grained facies without any noticeable interruption (apart from temporary non-sedimentation) – only seismic activity can be considered as a logical cause.

The other syndepositional and metadepositional soft-sediment deformation structures can all be explained as resulting from other deformational processes, but some of the possible genetic interpretations are mutually inconsistent, and other possible interpretations do not fit the environmental conditions (i.e. very quiet marine environment below storm wave base). In addition, most of the abundant soft-sediment deformation structures in the fine-grained facies cannot be explained by mass-flow transport (fine-grained turbidites occur, but high-density flows with sediment in a plastic state are absent). Moreover, sediments were deposited below storm wave base, which also excludes wave-related deformational mechanisms.

The only mechanism that can explain all types of soft-sediment deformation structures is a series of shocks (sometimes apparently at relatively short intervals, sometimes with longer intervals), sometimes in combination with repeated phases of hydrostatic under- and overpressure; in this context the two main earthquake-induced mechanisms for soft-sediment deformation are: (1) liquefaction of sediment generated by superficial shear waves; and (2) violent expulsion of pore water and liquefied sediment, generated by compression waves (Montenat *et al.* 2007). Considering the commonly very quiet sedimentary environment where only fine-grained sediment accumulated (the coarsest material being very fine sand), the shocks can be ascribed only to the tectonic activity of the basin.

The above does not imply that *all* soft-sediment deformations were triggered by an earthquake-induced shock. In almost all environments, even in tectonically quiet ones, some deformations may occur, particularly if the content of silt is high (as this grain size is very susceptible to deformations) and also if sediments with contrasting granulometry are present (Montenat *et al.* 2007). Because the fine-grained facies of the Chaibasa Formation, consists primarily of metamorphosed shales (derived from a combination of clay and silt), and because some sand was also present (although almost exclusively very fine sand), the sediment characteristics were favourable for soft-sediment deformation. The fact that apparently some consolidation of the muddy basin floor took place at some intervals (as a result of compaction and/or some interruption in sediment accumulation), thus providing a vertical change in physical characteristics after renewed sedimentation, made the topmost sediments even more prone to deformation. Although the above combination of characteristics favoured soft-sediment deformation, and although most soft-sediment deformation structures can be explained by different processes, it seems that a considerable part of the deformation was triggered by seismic shocks. This is even more likely because of the stratigraphic distribution of the soft-sediment deformation structures: they are concentrated in relatively few layers that are interbedded between undeformed layers, sometimes with several deformed layers within a short vertical succession. Although the deformed horizons cannot be traced without interruption over lateral distances of several kilometres (because of lack of exposures), the exposure-wide continuation of the deformed levels and the spatial distribution of the various types of deformation strongly favour their interpretation as layers with deformations that were triggered by earthquakes (Rodríguez-López *et al.* 2007), sometimes apparently with aftershocks (Seilacher 1984; Seth *et al.* 1990; Ricci Lucchi 1995; Bose *et al.* 1997, 2001; Jones & Omoto 2000). The deformed layers should therefore be considered as seismites.

Another strong argument for an earthquake-induced shock is the collapse structure. It must be deduced that a shear plane developed between the layer with a chaotic fabric and the overlying internally undeformed layer that collapsed. Earthquake waves, if not sufficiently strong, cannot propagate across shear planes (Schwab & Lee 1988) and this explains the contrasting degrees of deformation in the two layers.

Pillow structures have also often been attributed to seismicity (Montenat *et al.* 1987; Cojan & Thiry 1992; Roep & Everts 1992; Guiraud & Plaziat 1993; Rodríguez-Pascua *et al.* 2000). Experiments applying repeated jerks also resulted in similar pillows (Allen 1982; Owen 1996). An earthquake origin of these pillows in the Chaibasa shales is also suggested: (1) because the affected deposits formed below the storm wave base; (2) by their complex and multi-stage deformations revealing repeated seiches; and (3) by their lateral continuity and gradual lateral change in size and complexity. In addition, they are confined to certain stratigraphic levels, although the granulometric characteristics are the same throughout the entire fine-grained facies, which is prone to deformation.

## Conclusions

Several of the soft-sediment deformation structures show characteristics that indicate earthquakes as a

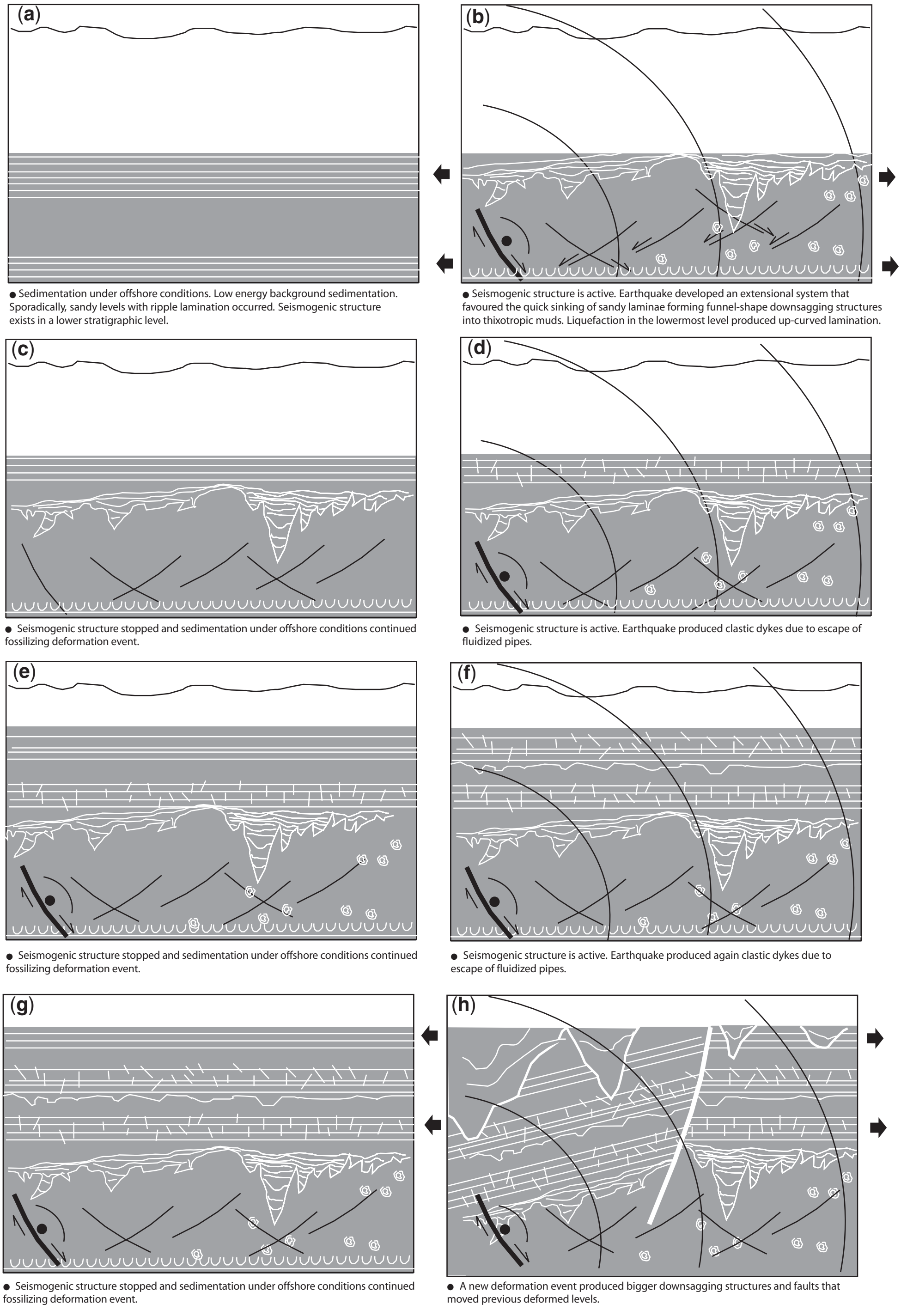


**Fig. 16.** Reconstruction of the sequence of events that can explain the complex soft-sediment deformation structures in the Chaibasa fine-grained facies as a result of seismic activity. Several phases of tectonic quiescence and of earthquake activity must have alternated.

trigger. The most logical reconstruction of the successive phases of seismic deformation as recorded in the Chaibasa fine-grained facies is as follows (Fig. 16). After an interval of tectonic quiescence (Fig. 16a), an earthquake develops faults and results in an extensional system that gives rise to down-sagging of sandy laminae into the underlying muds (Fig. 16b) that have become thixotropic due to the shock-induced stress. Then a new interval of tectonic rest (Fig. 16c) is followed by a new shock, resulting in the escape of pore water from the surficial sandy and muddy laminae. After the water has escaped, fluidized sand fills the escape pathways, thus forming sandy clastic dykes (Fig. 16d). During a third phase of tectonic quiescence the deformed layers become buried under a new cover of horizontal muddy and sandy laminae (Fig. 16e). A new earthquake then results in a new phase of clastic-dyke formation (Fig. 16f), comparable to the situation described before. Ongoing sedimentation of sandy and muddy layers during a next quiet interval (Fig. 16g) 'fossilizes' the buried deformation structures. Finally, a new, relatively heavy or nearby shock triggers large down-sagging structures (Fig. 16h, left-hand side) and results in faulting of the previously deformed levels. This sequence of events explains almost all soft-sediment deformation; no other sequence of events can do so satisfactorily.

In addition to the deformations that can be ascribed only to seismic activity, numerous structures can also be explained (if not better) due to earthquakes than as a result of other triggers. This all strongly supports the interpretation of most of the soft-sediment deformation structures in specific levels of the Chaibasa fine-grained facies as palaeoseismic features (although, obviously, not all deformations need necessarily be due to seismic shocks). Considering that the depositional setting of this formation, particularly during the later phase of sedimentation was in a tectonically active area, the seismite character of these layers becomes even more logical. The occurrence of turbidites in this succession further strengthens this hypothesis.

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