Trench-slope controlled deep-sea clastics in the exposed lower Surma Group in the southeastern fold belt of the Bengal Basin, Bangladesh

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Abstract
Deltaic to shallow marine (neritic) depositional settings have until now been the accepted interpretation for the Neogene Surma Group of rocks exposed throughout the southeastern fold belt of the Bengal Basin. The present study revises the earlier views and proposes some new insights. On the basis of detailed field studies carried out in the Sitapahar anticline, Rangamati area, and in the Mirinja anticline, Lama area, it is here proposed that the Surma Group succession represents an overall basinward progradation from deep marine to coastal marine depositional settings. Sedimentological evidence strongly suggests that the lower part of the Surma Group represents a slope apron, the growth of which is thought to have been governed by a westward-migrating accretionary prism complex within the active margin setting of the Indo–Burmese plate convergence. Thin packages of distinct turbidity-current-generated deposits together with some slump and debris-flow deposits contained within thicker intervals of mudstone characterize the lowermost exposed unit of the Surma Group in the Sitapahar anticline. The overlying slope deposits are essentially mudstone-dominated. Comparable deep-sea clastics with thicker intervals of sandstone turbidites, contained within a submarine canyon, are present in the Mirinja anticline. This new model suggests that it may be timely to re-evaluate the existing stratigraphic and tectonic framework of the eastern Bengal Basin. © 1999 Elsevier Science B.V. All rights reserved.

Keywords: Bengal Basin; Surma Group; accretionary prism; slope apron

1. Introduction
The geological evolution of the Bengal Basin covering Bangladesh and part of eastern India started in the late Mesozoic with the breakup of Gondwana-land and is still continuing (Alam, 1989). The Bengal Basin is well known for the development of one of the thickest (about 20 km) sedimentary piles in the world. The Chittagong–Tripura Fold Belt (henceforth called CTFB) (Fig. 1a) bordering the eastern side of the Bengal Basin has exposed a considerable thickness of the Neogene Surma Group of clastic sedimentary rocks (Table 1). The status of the lower Surma Group (Bhuban Formation) of rocks from the standpoint of facies analysis has not yet been firmly established. However, in most published accounts (e.g. Ganguly, 1983; Roy, 1986; Johnson and Alam, 1991; Shamsuddin and Abdullah, 1997) the deltaic to shallow marine depositional settings (neritic environment) inferred for the whole of the Surma Group succession has been seen as a paradigm for the stratigraphic, tectonic and sedimentological history of the Bengal Basin. For the purpose of the present study,
detailed fieldwork was carried out in two anticlinal structures within the CTFB, namely the Sitapahar anticline in the Rangamati area and the Mirinja anticline in the Lama area (Fig. 1b). Facies analysis of the exposed Surma Group rocks in both these anticlines suggests that the Group is characterized by
an overall progradation from deep marine to shallow marine depositional settings.

In this paper, we first discuss the regional tectonic evolution of the Bengal Basin, with particular emphasis on the CTFB region, in order to infer the paleogeographic setting that prevailed during Miocene time, then we describe the facies and facies associations of the studied sequence in an effort to interpret the depositional environments, and finally stimulate future lines of research for a better understanding of the sedimentological evolution of the Tertiary succession in the CTFB.

2. Regional tectonic evolution

Before discussing the regional tectonic evolution, the present status of regional stratigraphic understanding of the eastern part of the Bengal Basin is briefly presented here. Evans (1932) established a lithostratigraphic classification for the Tertiary sedimentary succession exposed in the Lower Assam Basin, which is situated northeast of the Bengal Basin. Since then, without any detailed regional correlation, this stratigraphic nomenclature (Table 1) has been used uncautiously for the CTFB and surrounding re-
Table 1
Traditional stratigraphic classification for the CTFB

<table>
<thead>
<tr>
<th>Age (approx.)</th>
<th>Group</th>
<th>Formation</th>
<th>Thickness (approx.) (m)</th>
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<td></td>
<td>Baraichari Shale</td>
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<tr>
<td></td>
<td></td>
<td>Bhurban</td>
<td>3000+</td>
<td></td>
<td></td>
<td>Sitapahar Sand-shale</td>
<td>830+</td>
</tr>
</tbody>
</table>

For convenience, local stratigraphy of the Sitapahar anticline, with study section indicated, is compared at the right. The Barail Group (Oligocene) and the Disang Group (Late Cretaceous to Eocene), whose spatial distributions are shown in Fig. 2, lie in stratigraphically lower position below the Surma Group (see text for discussion).

a Based on Evans (1932).
b Based on Chowdhury (1982).

The tectonic framework and structural setting of the Bengal Basin has been discussed by several workers, including Bakhtine (1966), Raju (1968), Guha (1978), Murphy and staff BOGMC (1988), Lohmann (1995) and Alam (1997). In the context of the present study the following discussion concentrates on the tectonic and structural evolution of the CTFB (Fig. 1).

Various independent lines of evidence demonstrate that the Indian plate was subducting below...
both the Tibetan and the Burmese plates during the greater part of the Tertiary. The process of formation of the CTFB is a mere westward extension of that of the Indoburman Range and is closely related to the easterly subduction of the Indian plate in an arc–trench tectonic setting (Fig. 1) displaying all the common elements of that setting (Fig. 2). However, it is important to note that most geological features of the Himalayas are not correlatable with those of the Indoburman arc and that the gross tectonic setting is also significantly different (Dasgupta and Nandy, 1995). The former provides a full record of collisional orogeny, whereas the Indoburman record is incomplete. The entire Indoburman Range represents an accretionary prism development and migration due to the continuous eastward subduction of the Indian plate beneath the west Burma block from the Late Cretaceous up to the end of the Oligocene (Fig. 1a), so that the prism complex as a whole youngs towards the west (Hutchison, 1989; Varga, 1997). As a result the flysch deposits of the Disang and Barail Groups are exposed in the Indoburman Range (Fig. 2). There is no reason to believe that this westward migration of the prism complex had stopped at this stage; rather it was still continuing to migrate further westward through the CTFB region, which Dasgupta and Nandy (1995) have termed a ‘Neogene accretionary prism’, formed above the subducting Indian lithosphere (Figs. 1 and 2).

Because of the counter-clockwise rotation of the Indian plate, at some time after the initial collision with Tibet (Curray, 1990), the ocean basin in the east gradually started to close from north to south (oblique subduction). As a result, the CTFB region was left in the early Neogene as a remnant ocean basin of Dickinson and Suczek (1979), and included the migrated trench-slope bathymetry from the east. In the eastern part of the basin, the subduction complex of the Indoburman arc had already emerged above sea level, the modern analogue of which may be the ridged forearcs (eastern Aleutian Trench, Cook Inlet) of Dickinson and Seely (1979).

This remnant ocean basin received sediments from two directions: (1) collision-derived sediment from the Barail and Naga Hills entering the basin from the north along the trench axis, and (2) Indoburman arc-derived sediments shed transversely to the trench from the east (Roy, 1986; Shamsuddin and Abdullah, 1997). Due to this heavy sediment loading, the trench axis has shifted further westwards and southwards, giving rise to the Neogene accretionary prism (Figs. 1 and 2). According to Dasgupta and Nandy (1995) the present flexure of the Indian lithospheric plate lies somewhere below the Bengal foredeep. More specifically, Khan (1991) has placed this present subduction line along the border between the Bengal foredeep and the CTFB. Along this line consumption of the oceanic plate is still continuing under a smothered trench-fill whose bathymetric expression is gradually revealed southward as the Andaman Trench (Fig. 1a) (Roy, 1986). It is remarkable that in the CTFB region the trench-slope
bathymetry has been smoothed out to shallow-water conditions probably sometime in the Late Miocene. At present the deep-water subduction zone exists only southwards from approximately 19°N as the Andaman–Sunda Trench (the solid portion of the ‘line of present subduction’ shown in Fig. 1a), along which the Bengal deep-sea fan sediments are subducting eastward beneath the Burmese plate. A good account of the tectonic and sedimentary history of this region is given by Curray et al. (1982), from which readers can get an approximate analogy of the depositional setting for the lower Surma Group proposed in this paper. It should be noted that the ophiolitic, exotic blocks found in the Indoburman ridge would be less common in the CTFB due to the increasing distance of the then trench axis from the initial line of subduction (in other words, the increasing width of the arc–trench gap); if such blocks are present, they would be at a lower stratigraphic horizon than that exposed in the study area.

The structural developments in the CTFB may have been largely controlled by the accretionary prism and involve east-dipping, high-angle thrust faults produced by off-scraping of oceanic sediments. Within individual thrust sheets, the sediments in the upper part may be deformed by the process of structural infolding (subduction kneading) giving rise to the long, shallow, almost N–S-trending folds in this region. It is also likely that the accretionary basins (trench-slope basins) of Dickinson and Seely (1979) developed as structural depressions between the thrust masses.

Duplex accretion is another tectonic feature of the subduction complex. Lohmann (1995) and Sikder (1998) have pointed out some duplex structures in the western part of the CTFB. These workers have also suggested thin-skinned detachment, shear-off tectonics to explain the structural style of the CTFB, but did not relate these processes directly to a subduction complex. We believe that the tectonic and structural development of this region is more readily explained by accretionary prism formation. Compressive wrench tectonics (as a result of convergent–oblique movement of the Indian plate, mentioned earlier) have also influenced the structural style of the CTFB (Murphy and staff BOGMC, 1988). In future, the identification of major individual thrust sheets (accretionary wedges) would be of vital significance to determine the chronological order of the rock strata, since it is well known that the complex as a whole youngs towards the west.

3. Stratigraphic sequence of the lower Surma Group in the Sitapahar anticline

In the Sitapahar anticline fieldwork was carried out along the Rangamati road section. The oldest 570 m of this exposed section (which is the main focus of the present study) is shown in a graphic log (Fig. 3). The entire sequence is divided into three major units. (1) Unit C (lower unit, 71 m thick) is characterized by several thin packets of turbidites and slumped beds contained within a muddy sequence. The unit is thought to have been deposited in a setting not far basinward from the base-of-slope. (2) Unit B (middle unit, 291 m thick) is a monotonous muddy slope deposit that contains some localized zones of very thin-bedded turbidites. (3) Unit A (upper unit, 208 m thick) represents the progradation of the first shoreface sand body on a 176-m-thick shelfal mud. Detailed bed by bed measurements have been carried out in all these units.

On the basis of the overall regional tectonic setting (Fig. 1a) it can be assumed that the trend of the paleo-coastline in the CTFB was almost north–south. Alam (1995) has documented a similar paleo-coastline trend from the paleocurrent analysis of the Baraichari Shale Formation (Table 1).

3.1. Sedimentary facies and depositional environments

3.1.1. Unit C

The overall sand–shale ratio of this unit is 30:70. For the convenience of discussion the unit is further divided into two subunits which are discussed below in stratigraphic order.

Subunit C1. At the base of this subunit there exists a 1.5-m-thick slump bed with isolated and enclosed slump blocks (up to 1 m long) within a muddy matrix of distorted nature. The lower part of this subunit is not organized and develops two Bouma sequences of the types T_{ab} and T_{a-c}. The upper part shows repetitive occurrence of eleven medium-bedded Bouma T_{ab} sequences (Fig. 4a). Thickness of
Fig. 3. General litho-stratigraphic column of the lower part of the Surma Group exposed in the Sitapahar anticline with detailed sedimentological logs of salient portions.
Fig. 4. Outcrop photographs of selected facies from the lower part of the Surma Group in the Sitapahar anticline, Rangamati area — (a–f) see Fig. 3 for stratigraphic positions of the photographs — and in the Mirinja anticline, Lama area — (g, h). (a) Repetitive occurrence of thickening-upward Bouma T_ab sequences, upper part of subunit C (hammer in the center for scale). (b) Upper part of slump bed in division C2a; note small-scale channelized sand bodies within muddy slump (arrowed) (stick for scale is 2 m long). (c) Alternation of massive sandstone beds and wavy–lenticular beds in the upper part of C2a, terminating upwards with hemipelagic mud of subunit C2b (scale division in cm). (d) Typical Bouma T_e–c turbidite in division C2b. Divisions T_d and T_e broke off this specimen (scale is in cm).
Fig. 4 (continued). (e) Laterally persistent very thin beds of calcified very fine sand and silt within slope mud in subunit B2 (stick at left is 2 m long). (f) Zoophycos burrows at the base of unit A. (g) Coarse-grained turbidite displaying strong basal scouring and inverse-to-normal grading at the base, from the axial part of the submarine channel (coin is 2.7 cm in diameter). (h) Turbidite facies with Bouma divisions $T_a$, $T_b$, and $T_c$, from the marginal part of the submarine channel (coin is 2.2 cm in diameter).
individual beds varies from 12 to 50 cm. The base of the $T_a$ division (fine sand) is sharp to slightly erosional with associated rip-up mud clasts and some micro-loading. Almost all the $T_d$ divisions characteristically show convolution and micro-injection structures. The uppermost two sand beds are amalgamated along an erosional surface (Fig. 3, C1). The beds of this overall thickening-upward subunit C1 show onlap termination onto local basin relief in the southwards direction.

The subunit C1 shows no evidence of major channeling and is interpreted here as a small-scale depositional lobe in which the thickening-upward trend results from a compensation cycle (Mutti and Sorrino, 1981) due to the progressive smoothing of subtle depositional relief or the progressive lateral shifting of the turbidity current axis. The same types of small-scale (5 to 25 m thick) thickening-upward depositional sandstone lobes with onlap termination within an active margin setting have also been described from the geological record (Cazzola et al., 1985).

Subunit C2. This subunit is further divided into three divisions: C2a, C2b and C2c (Fig. 3).

Division C2a. This division begins with an even larger (nearly 7 m thick) slump bed containing dispersed slump blocks within mud. The lower part of this slump bed contains large blocks (>1 m), some with preserved bedding characteristics, whereas the upper part (Fig. 4b) is more muddy, enclosing a few small channelized sand bodies (85 cm long, 20 cm thick) that nearly retain their original attitude. These channelized sand bodies are filled with very fine to coarse sand with some mud clasts and coal streaks, and both their lower and upper boundaries are concave with abrupt thinning of channel margins (Fig. 4b). The top 2.5 m of division C2a is characterized by the alternation of massive sandstone beds, with normal grading in the lowermost sand bed, and wavy–lenticular fine sand–mud beds (Fig. 4c). This upper part of C2a (above the slump bed) begins with a distinct scoured base containing numerous rip-up mud clasts. It contains three sand beds that show an overall fineing- and thinning-upward trend.

Two alternative explanations can be offered for the generation of these wavy–lenticular beds (mentioned above), which do not accord with the divisions of the Bouma sequence. The origin of these beds can be easily comprehended when the top part of division C2a is interpreted as a small channel–levee system. A similar small-scale (6 m thick) association of massive sand beds and wavy–lenticular beds from the lower part of the Halifax Formation, Nova Scotia, has been interpreted as a channel–levee system (Stow et al., 1984). The structural scheme of some parts of the wavy–lenticular (rarely flaser) beds is thought to match with the fine-grained turbidite model of Stow and Shamugam (1980). Alternatively, some of the tractional characteristics of ripple cross-lamination (coupled with? bi-directional cross-lamination observed in one dislocated block) of the wavy–lenticular beds suggest similarities with the ‘variant model’ (bottom-current-reworked turbidite) suggested by Stanley (1987). It is to be noted that the paleo-Bengal Basin was a deep-water embayment opening to the south (see tectonic discussion). In this type of basin internal waves and tides commonly play a major role in bottom-current reworking processes (e.g. McCave et al., 1980).

Divisions C2b and C2c. These divisions are self-explanatory in the graphic log (Fig. 3). The random association of Bouma sequences $T_a$, $T_{abc}$ and $T_c$ (Fig. 4d) with debris-flow (mostly mud-flow and rarely grain-flow types) deposits characterizes these divisions. Two small channels (50 cm wide, 15 cm deep) (Fig. 3) with conglomeratic fill were encountered in division C2b. The orientations of the channel walls are east–west. Pebbly mudstone beds in division C2c, formed by cohesive freezing, contain large (up to 40 cm) rolled-up sandstone clasts, and show considerable flowage of matrix mud. $T_a$ divisions in C2b and C2c are frequently load-casted with some dish structures and vertical water escape pipes. The sand beds in division C2c pinch out laterally after about 150 m in a northwards direction.

Division C2b is characteristically dominated by highly carbonaceous, massive, homogeneous, non-burrowed, and grayish black mudstones of hemipelagic origin. This thick sequence of hemipelagite indicates basin starvation, probably due to a short-term sea-level rise. The turbidite beds within the divisions C2b and C2c are thought to be mainly of uncon fined sheet-flow origin. Two down-to-basin synsedimentary, high-angle normal faults with sand intrusion along the fault planes appear to have affected parts of divisions C2a and C2b.
3.1.2. Unit B

This unit is a predominantly muddy sequence having a typically laminated character, sometimes with silt streaks and lenticles, and occasionally showing a massive and carbonaceous nature. The unit contains four subunits: B1, B2, B3 and B4 (Fig. 3). In subunits B1 and B2 very thin (1.5 to 3 cm), very fine-sand to silt beds (calcified) are repeated at 15 to 35 cm intervals (Fig. 4e). Primary internal structures within these beds might have been obliterated due to calcification (early diagenetic origin?). Subunit B4 also contains eight calcified, sharp-based, thin (5 to 8 cm) fine-sand beds. Similar laterally persistent thin sand beds within slope mud have been described from the Eocene Hecho Group, Spain (Mutti, 1977). Subunit B3 is a large-scale slide-body, thought to be of a synsedimentary origin. No evidence of channeling was observed within unit B (perhaps through lack of exposure).

3.1.3. Unit A

This unit begins with a nearly 5-m-thick zone of extensive Zoophycos burrowing (up to 13 cm long) (Fig. 4f) with some Rhizocorallium burrowing, which is thought to indicate a shelf–slope transition zone. Except for the upper subunit A1 (Fig. 3), the remaining part of unit A is dominantly muddy with silt streaks and very fine-sand lenticles, the frequency of which gradually increases upwards. Unlike the muddy portions of units C and B, no carbonaceous mud is present and the burrows of Rhizocorallium and spiral Rosselia occur occasionally in this unit. Some thin (<10 cm), sharp-based very fine-sand beds with bidirectional cross-lamination are found sporadically at the upper part of this muddy sequence (below subunit A1). These beds probably formed during unusually high tides within the inner shelf environment.

Subunit A1. A detailed graphic log of subunit A1 is presented in Fig. 5. The lower 21 m of this subunit show a distinct coarsening- and thickening-upward trend, erosionally overlain by an upper 9 m showing an indistinct fining-upward trend. The lowermost 6.5 m of A1 contains 34 small-scale (5 to 26 cm thick) fining-upward cycles together with a few random and fining-then-coarsening-upward cycles. Fig. 5b shows a representative (15 cm thick) fining-upward cycle. In each cycle the sand is gray and the mud is bluish gray in colour. The very fine-sandy lower part of a cycle is characterized by bipolar cross-lamination with hair-thin mud drapes, the middle part contains a wavy–lenticular structure of silty very fine sand and mud, which is culminated by the upper laminated mud. The thickness of the sandy portion of the cycles increases upwards. This lower part of subunit A1 grades up into a 14.5 m thick, apparently massive di-
vision of a fine-sand bed containing faint parallel- to low-angle cross-lamination, above which a thin (15 cm) bed of profuse mud clasts is present as erosional lag deposits (Fig. 5a). The upper 9 m of subunit A1 is characterized by monotonous wavy (‘tidal’) bedding (Reineck and Wunderlich, 1968) with the proportion of mud increasing upwards. The fine to very fine sand of this part of subunit A1 is yellowish brown in colour, probably indicating intermittent subaerial exposure during the tidal cycles.

The depositional environment of subunit A1 (Fig. 5) is inferred to be a tidal sand ridge progradation on the shelf mud followed by tidal flat progradation, within a regressive shelf and open coast setting (e.g. Meckel, 1975; Johnson, 1977) at the time of highstand sea-level. The absence of tidal signatures in the proximal ridge deposits may be due to wave reworking, the indication of which is preserved as an erosional lag at the ridge crest. The overlying tidal flat deposits, with wavy-bedding and bipolar cross-lamination, are thought to have been deposited in the mixed-flat environment with slight progradation (Fig. 5).

Filed evidence from both the Sitapahar and Mirinja anticlines suggests that tide-dominated shallow marine shelfal to coastal depositional settings, with alternating transgressive and regressive pulses, characterize the upper Surma Group rocks. For example, Alam (1995) described the tide-dominated sedimentation in the Baraichari Shale Formation (Table 1). The Tipam Group and the Dupitila Formation, overlying the Surma Group, represent a continental–fluval depositional environment that prevailed at the final stage of regional regression aided by tectonic upheaval of the CTFB (Alam and Ferdous, 1995).

3.2. Comparable deep-sea clastics in the Mirinja anticline

Alam and Karim (1997) have interpreted a transition from a shallow marine muddy shelf (below wave-base) to tide-dominated coastal settings for the Surma Group rocks exposed in the Mirinja anticline. However, on the basis of our detailed field work we re-interpret the lower Surma Group succession in the Mirinja anticline as showing evidence of progradation from deep marine to shallow marine settings. Rather than offering a detailed interpretation of the lower Surma Group in this anticline we emphasize here some salient features.

A submarine canyon, encased within the slope mud and breaching the shelf edge, is laterally traceable (in a N–S-oriented cross-sectional view) for about 400 m as a multiple-stepping erosional feature with an exposed erosional relief of 15 m. The orientation of the channel wall is roughly east–west, and the current direction within the channel-fill turbidites (about 50 m thick) is towards the west. Repetitive occurrence of thick-bedded (average 60 cm) coarse silt to fine sand with numerous characteristic water-escape marks, such as internal convolution, network of dish and pillar structures, water-escape pipes, etc., resulting possibly from liquefied flow, characterize the lower part of the channel-fill deposits. Some of these beds are typically capped by a thin (2–3 cm) sand layer with tractional ripple cross-lamination, the depositional mechanism of which can be explained by the late-stage reworking of the aqueous currents generated within the ambient water above the liquefied flows due to the shearing at the flow surface (Lowe, 1982). The upper part of the channel-fill deposits consists of an alternation of thick-bedded (average 1 m), fine to medium, massive sandstones (Fig. 4g) containing some water-escape marks at the top. These beds are thought to have been deposited by high-density turbidity currents. A few of these beds show partial amalgamation. Some thin-bedded (10 to 20 cm) partial to complete Bouma sequences (Fig. 4h) occur at the channel margins. The tide-dominated shoreface sand unit occurs above a nearly 150-m-thick muddy upper slope to shelfal sequence overlying the channel-fill turbidites. It is notable that from the turbidite-fill and for about 140 m downwards, grazing traces of gently meandering forms of Nereites (probably Scolicia sp.) have been observed on the bases of several thin-bedded sandy turbidites.

4. Discussion

There is no direct evidence for the water depth in which the inferred deep-water clastics of the lower Surma Group in the study area have been deposited. No macrofossils have been encountered in these rocks during the fieldwork. Likewise, a quest for trace-fossils in the rock units C and B was not successful.
It appears that the Surma Group rocks are sparsely fossiliferous in terms of macrofossils. On the other hand, micro-paleontological work on the lower Surma Group in the study area has not yet been conducted. In fact, no data is available on the bathymetrically sensitive microfauna from the lower Surma Group even in the CTFB region, to verify our interpretation of a deep-water origin of these rocks. We have, therefore, relied heavily on the sedimentary features and facies associations presented in Fig. 3, in interpreting these rocks (units C and B) as deep-sea clastics. The following discussion explains why we think that the lower Surma Group rocks in the studied sequences are not of a shallow marine origin.

Generation of turbidity currents on open shelf or deposition of discrete shallow marine ‘turbiditic’ sandstones have long been debated (e.g. Hamblin and Walker, 1979; Nelson, 1982; Duke, 1990; Tokuhashi, 1996). Nelson (1982) suggested an equivalent turbiditic terminology — $S_b$, $S_c$, $S_d$ and $S_e$ — for the storm-generated sand layers in an open shelf setting. It is noteworthy that no counterpart for a Bouma $T_a$ division (i.e. $S_a$ layer) exists in Nelson’s scheme. Hamblin and Walker (1979) and Tokuhashi (1996) described turbidity-current-generated sandstone beds in shelfal settings below storm-wave base, in which $T_a$ beds are very rare and, if present, are only less than a few tens of centimeters thick. Also, if a discrete rock sequence, characterized by traction-generated structures like $T_b$ and $T_c$ beds, is not associated with a $T_a$ bed, its true turbiditic origin may be questioned (Duke, 1990; Shanmugam, 1997).

With respect to the discussion above, our reasons for attributing the rocks of units C and B to deep marine, rather than shallow marine depositional settings are as follows.

1. The sandstones of units C and B do not show any hummocky cross-stratification (HCS) or wave ripple forms.
2. The highly carbonaceous and homogeneous mudstones encountered in units C and B are not present in unit A.
3. Trace fossils of *Rhizocorallium* (with protrusive spreiten) and spiral *Rosselia*, belonging to the *Cruziana* ichnofacies that seem to occur in shallow-water shelfal environments, are occasionally encountered within the mudstones of unit A; but these trace fossils are not encountered in units C and B.
4. The Bouma $T_a$ division is typically present within the turbiditic beds of unit C, with some of them passing upwards into nearly complete classical Bouma sequences (e.g. Fig. 4d). Thicknesses of individual $T_a$ divisions vary from 10 to 150 cm, with an average of 40 cm. Shallow marine shelfal turbidity currents, even if they occur, do not seem to be strong enough to deposit thick $T_a$ beds (e.g. Tokuhashi, 1996). High-energy/high-density turbidity currents originating at the slope base can easily explain the deposition of these $T_a$ beds encountered in unit C.
5. The deposits of all the end-member types of sediment gravity flows — turbidity current flows (that predominate in our case), liquified flows, grain flows (very rare in our case) and cohesive debris flows (Lowe, 1982) — with associated slump beds are present in the studied sequence. However, some sort of gravity-related deposits (e.g. slumps, debrites, turbidites, etc.) may exist in front of a rapidly prograding river delta within a shallow marine setting; but in the rock record these deposits should be directly associated with the advancing delta front sand. The overall rock context (Fig. 3) suggests that this is not the case in the studied sequence, since above unit C a 465-m-thick sequence of predominant mud occurs before the advancement of the first shoreface sand unit (which is also not a delta front sand). Furthermore, there is no modern or ancient published example of the occurrence of the whole spectrum of gravity-flow deposits within the distal part of an open shelf.

Therefore, it seems logical to conclude that the rocks of units C and B have been deposited beyond the shelf–slope break, within the deep-water settings of a bathyal environment.

At our present level of understanding, it would be premature to fit the above-described sequence into a specific morphological class of turbidite systems. However, the closest match appears to be with the ‘mud/sand-rich slope apron’ in the classification scheme of Reading and Richards (1994). Our slope apron deposits are characterized by intimate association of turbidites with slump and debris-flow deposits, indicating greater slope instability, with slope failure triggering the turbidity currents. In unit C, three discrete morphological elements (that may be considered as architectural elements) are inferred (C1, C2a and C2c). Subunit C1 represents a small
submarine depositional lobe occurring near the foot of the slope, where the gradient of the slope decreases significantly. The upper part of division C2a indicates a small channel–levee complex resulting, perhaps, from the lateral shifting of the channel on the low-gradient part of the basin. Division C2c probably represents a debris-mass deposited close to the slope base. These types of small-scale morphological features are common in mud/sand-rich slope aprons (Reading and Richards, 1994). Geological records (e.g. Stow et al., 1985) show that slope aprons along active margins are subjected to considerable slumping and other mass movements similar to those described from unit C.

In the studied sequence there is no evidence for more distal basin plain or abyssal plain deposits, such as distal turbidite sands, extremely fine laminated silt within mud, etc. Consequently, the 291-m-thick muddy sequence of unit B is interpreted as slope deposits overlying the base-of-slope deposits of unit C. The heterogeneous muds of unit A show shallow marine characteristics described earlier, and the first tide-dominated shoreface sands (subunit A1) (Figs. 3 and 5) occur at the uppermost part of this unit. Therefore, the shelfal deposits of unit A expectedly overlie the slope deposits of unit B as part of a basin-ward progradational sequence. The considerable thickness of muddy slope and shelf deposits is probably indicative of the paleogeographic setting in which a wide shelf separated the coastal clastic system from the deep marine clastic system. This point also demonstrates that our studied sequence is different from the deposits of a delta-fed submarine ramp system, in which progradational sequence shows abbreviated slope–shelf deposits directly overlain by delta front sands (Heller and Dickinson, 1985).

It is notable that no tectonic breaks have been encountered within the studied sequence (570 m thick). This unbroken nature of the sequence, deposited in environments ranging from base-of-slope to the shoreface, probably indicates a late-stage trench-slope bathymetry, in which the structural reliefs (e.g. structural infoldings) on the inner-slope wall (similar to the present-day youthful Andaman trench-slope) have been masked by heavy sediment influx from two directions (see tectonic discussion). Alternatively, but less likely, unit C could be ponded turbidites deposited in the uppermost slope basin formed by structural infolding. For the validation of this inference, more data would be needed from the rock sequence lying below unit C, i.e. the lowest part of the Surma Group, which are still not available.

We believe that small-scale deep-sea fans, associated with the transverse slope channels or gullies from the east (e.g. the submarine canyon in the Mirinjia anticline, described earlier), and/or large-scale elongate deep-sea fans, associated with the longitudinal submarine canyons (inferred) from the north, could have been generated in the more axial sectors of the basin, beneath the slope apron deposits described here. Such a possibility may also need to be borne in mind in future hydrocarbon exploration, in order to encompass the stratigraphic traps commonly associated with gravity-flow-generated sandstone bodies enclosed within basinal mudstone.

On the basis of facies analysis, Dasgupta et al. (1991) have interpreted the Barail Group (3500 m thick), in the Lower Assam Basin, north Cachar, as deposits of a basinward migrating progradational submarine fan complex in the active margin of the Indo–Burmese plate convergence, and suggested that sedimentation appears to have continued into the overlying Surma Group. It should be noted that our studied sequence includes only part of the lower Surma Group, and that the rocks in the still lower part of the group remain undescribed. Therefore, it is of crucial importance to study the lowermost part of the Surma Group both from the exposed sections in the remote eastern anticlinal structures of the CTFB and from the subsurface data, to have a better understanding of the basinward progradational nature of the Group inferred in this study. In addition, detailed paleocurrent analyses of the Surma Group rocks in different areas of the CTFB would be necessary in order to understand the pattern of facies changes within this part of the Bengal Basin. Finally, a micro-paleontological study of the Tertiary succession in the CTFB would enhance the understanding of the stratigraphical and sedimentological evolution of this region.

5. Conclusions

A deltaic to shallow marine (neritic) depositional setting is the accepted traditional interpretation for
the entire Neogene Surma Group rocks exposed in the southeastern fold belt of the Bengal Basin. On the basis of a detailed facies analysis, the present study revises the earlier views and concludes that part of the lower Surma Group represents a slope apron, deposited in deep marine (bathyal) environments. The study further suggests that the whole of the Surma Group represents an overall basinward progradation from deep marine to coastal marine depositional settings, and the growth of this progradational sequence has been governed by a westward-migrating accretionary prism complex within the active margin setting of the Indo–Burmese plate convergence.

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