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Shallow-water gravity-flow deposits, Chapel Island Formation, southeast Newfoundland, Canada

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ABSTRACT

A remarkable suite of shallow-water, gravity-flow deposits are found within very thinly-bedded siltstones and storm-generated sandstones of member 2 of the Chapel Island Formation in southeast Newfoundland. Medium to thick siltstone beds, termed unifites, range from non-graded and structureless (Type 1) to slightly graded with poorly developed lamination (Type 2) to well graded with lamination similar to that described for fine-grained turbidites (Type 3). Unifite beds record deposition from a continuum of flow types from liquefied flows (Type 1) to turbidity currents (Type 3). Calculations of time for pore-fluid pressure dissipation support the feasibility of such transitions. Raft-bearing beds consist of siltstone with large blocks or 'rafts' of thinly bedded strata derived from the underlying and adjacent substrate. Characteristics suggest deposition from debris flows of variable strength. Estimates of debris strength and depositional slope are calculated for a pebbly mudstone bed using measurable and assumed parameters. An assumed density of $2{\cdot}0\ \mathrm{g\ cm^{-1}}$ and a compaction estimate of 50% gives a strength estimate of $79{\cdot}7\ \mathrm{dyn\ cm^{-2}}$ and a depositional slope estimate of $0{\cdot}77°$.

The lithologies and sedimentary structures in member 2 indicate an overall grain-size distribution susceptible to liquefaction. Inferred high sediment accumulation rates created underconsolidated sediments (metastable packing). Types of sediment failure included *in situ* liquefaction ('disturbed bedding'), sliding and slumping. Raft-bearing debrites resulted from sliding and incorporation of water. Locally, hummocky cross-stratified sandstone directly overlies slide deposits and raft-bearing beds, linking sediment failure to the cyclical wave loading associated with large storms.

The gravity flows of the Chapel Island Formation closely resemble those described from the surfaces of modern, mud-rich, marine deltas. Details of deltaic gravity-flow deposition from this and other outcrop studies further our understanding of modern deposits by adding a third dimension to studies primarily carried out with side-scan sonar.

INTRODUCTION

Subaqueous mass-movement features occur in deep-sea fan, slope, and deltaic settings (Moore, 1961; Coleman & Garrison, 1977; Roberts, 1980; Embley, 1982; Prior & Coleman, 1982; Postma, 1984a,b; Prior *et al.*, 1984, 1986; Postma *et al.*, 1988a). Deltaic slides and gravity flows, however, '... have rarely been observed in ancient counterparts' (Coleman, 1981, p. 84), notable exceptions being those observed by Nemec *et al.* (1988) and gravelly mass flows described by Postma (1984a,b) and Postma & Roep (1985). A diverse suite of shallow-water gravity-flow deposits similar to those described from modern deltas are described in this paper from the Chapel Island Formation of Late Precambrian–Early Cambrian age. Characteristics of these gravity-flow deposits indicate that support mechanisms commonly varied both temporally and spatially during the deposition of individual beds. We attempt to summarize the possible transitions in dominant transport mechanisms and causes for sediment failure. For example, the debris-flow and slide deposits capped by hummocky cross-stratified sandstone described here provide a link between ancient gravity flows and the triggering mechanism of cyclical shear stresses associated with storms.

We believe that the geotechnical characteristics and bottom slopes of the deposits described in this study were very similar to those found along subaqueous fronts of modern, large, mud-rich, marine deltas. Discovery and study of other ancient analogues for this setting may help clarify transport processes for subaqueous, muddy gravity flows.

GEOLOGICAL SETTING OF THE CHAPEL ISLAND FORMATION

The Chapel Island Formation is the middle unit of three, dominantly siliciclastic formations that disconformably overlie thick Precambrian volcanic rocks exposed over most of the Burin peninsula in the Avalon Zone (Fig. 1) of the Appalachian Orogen (Williams, 1979) on Newfoundland. The oldest of these units, the Rencontre Formation, consists of conglomerates, sandstones and shales deposited in alluvial, fluvial and marginal marine environments (Smith & Hiscott, 1984). The Random Formation is the youngest unit in the sequence and consists of sandstones, shales and quartzites that were deposited in tidally dominated intertidal and subtidal environments (Anderson, 1981; Hiscott, 1982). The Chapel Island Formation is 1000 m thick and consists of sandstones, siltstones and mudstones with subordinant limestones (Fig. 2). At present, the formation is the focus of intense interest as a possible stratotype for the Precambrian–Cambrian boundary (Narbonne *et al.*, 1987; Landing *et al.*, 1988, 1989). The gravity-flow deposits described in this paper belong to member 2 of the formation, which consists of green siltstone and thin to medium bedded sandstones deposited in a storm- and wave-influenced deltaic setting in shallow subtidal and inner shelf environments (Myrow, 1987, 1992; Myrow *et al.*, 1988). Proximity to a major river delta is inferred from the 1 km thick mud-dominated section, stratigraphic position between fluvial and mid-shelf deposits (Rencontre Formation and member 3 of the Chapel Island Formation, respectively), and occurrence of slides and debris flows (this paper) indicative of rapid rates of accumulation (Hein & Gorsline, 1981). This report is based primarily on detailed analysis of the Fortune Head and Grand Bank Head localities (Fig. 1).

INFLUENCE OF STORMS AND PROXIMITY TRENDS

A companion paper (Myrow, 1992) outlines in detail the evidence for storms as a major control on deposition in member 2. Most of this member is composed of three facies: (i) a gutter cast facies characterized by siltstone with sandstone laminae to very thin sandstone beds and abundant pot and gutter casts similar to those described by Whitaker (1973) and others; (ii) a siltstone-dominated facies composed of siltstone with 30–50% sandstone laminae to medium-bedded graded sandstones (tempestites) and

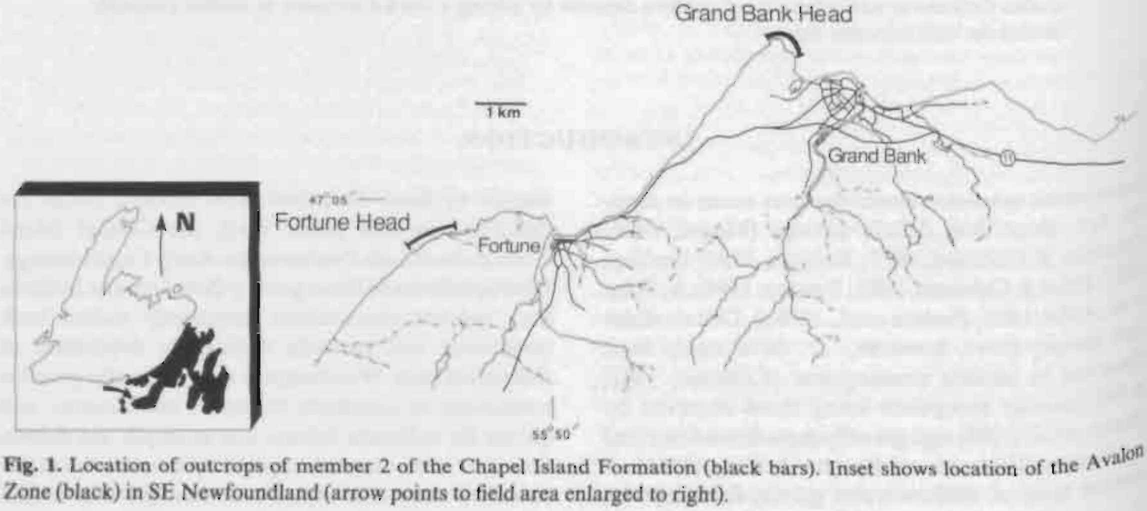

Fig. 1. Location of outcrops of member 2 of the Chapel Island Formation (black bars). Inset shows location of the Avalon Zone (black) in SE Newfoundland (arrow points to field area enlarged to right).

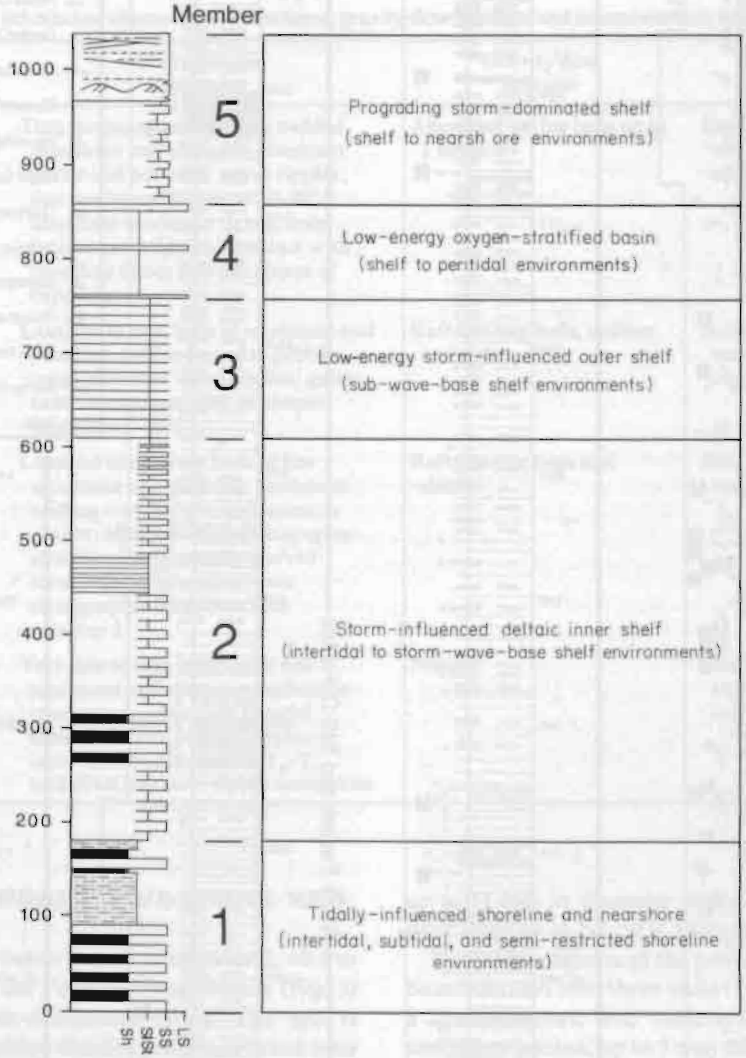

Fig. 2. Generalized stratigraphic column of the Chapel Island Formation showing the distribution of the five informally defined members. Brief palaeoenvironmental interpretations are given to the right.

few, if any, erosional structures; and (iii) a sandstone-dominated facies with thicker sandstone beds, some with hummocky cross-stratification. The vertical distribution of these facies within member 2 is shown in Fig. 3. Lithofacies, characteristic structures, and process interpretations are summarized in Table 1.

In the storm sedimentation model of Myrow (1992) the gutter cast facies occupies the shallow subtidal zone (Fig. 4), an area of sediment bypass or throughput across which high-velocity, sediment-laden flows erode deep, narrow scours (gutter casts). Little sand is deposited here; most of the sediment bypasses the very shallow subtidal zone and is deposited in deeper water. With deceleration, erosion of the sea floor ceases and continuous beds of more even thickness are deposited (siltstone-dominated facies). Further out on the shelf, bed thickness reaches a maximum (sandstone-dominated facies) and hummocky cross-stratification is most abundant. More distally, below storm wave base (represented by the thinly laminated siltstones of member 3 of the formation), bed thickness once again decreases.

The various gravity-flow deposits of member 2 are described separately below, and are subsequently related to sedimentary processes in this shallow-water depositional setting.

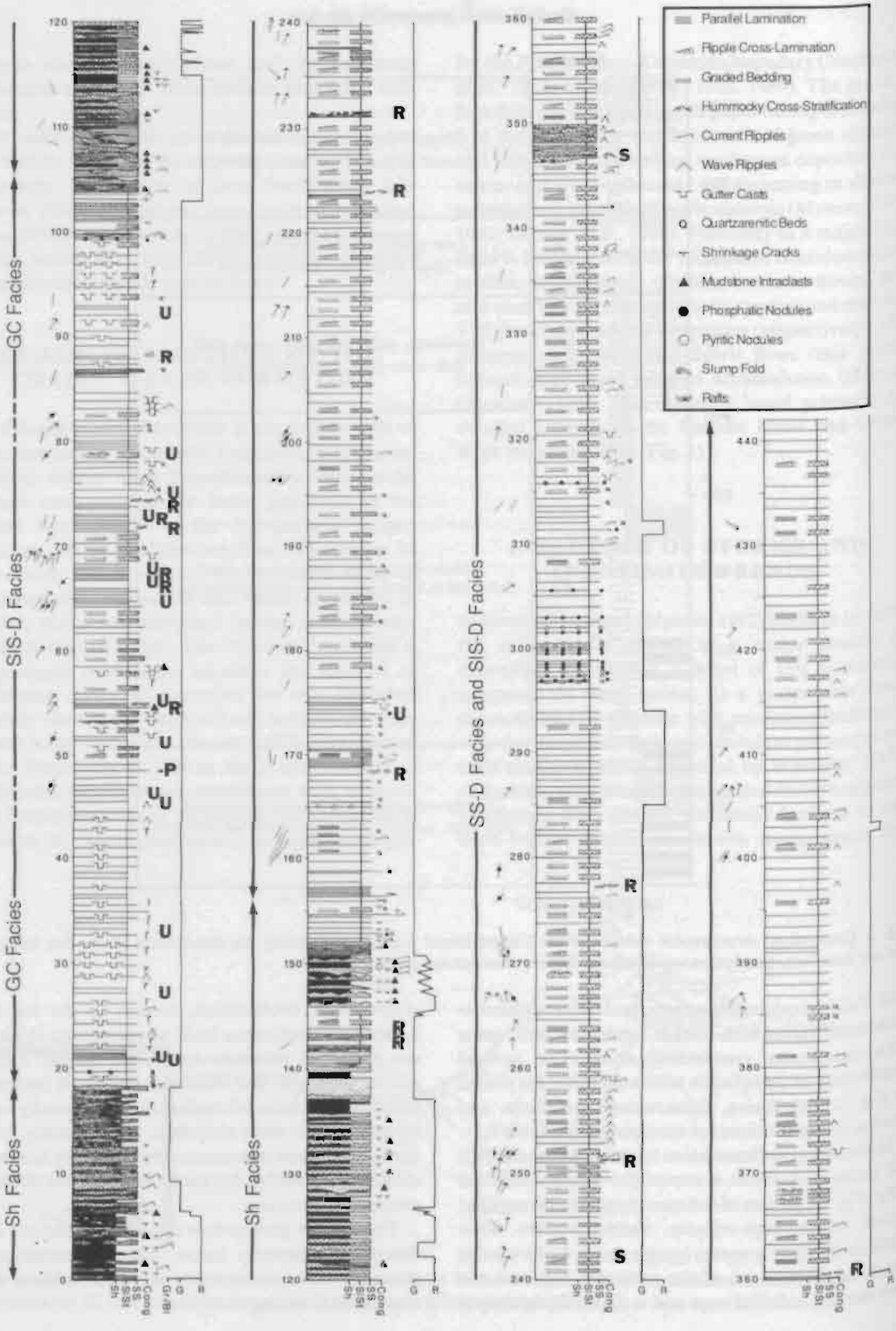
Parallel Lamination
Ripple Cross-Lamination
Graded Bedding
Hummocky Cross-Stratification
Current Ripples
Wave Ripples
Gutter Casts
Quartzarenitic Beds
Shrinkage Cracks
Mudstone Intraclasts
Phosphatic Nodules
Pyritic Nodules
Slump Fold
Rafts
Sh Facies
GC Facies
SIS-D Facies
SS-D Facies and SIS-D Facies
Sh
S/St
SS
Cong
Gr/Bl
G
R

Table 1. List of lithofacies, characteristic structures, gravity-flow deposits and interpretations for members 2 and 3.

Lithofacies	Characteristic structures	Gravity-flow deposits	Interpretation
Gutter cast facies	Thin laminae to very thinly bedded sandstone and siltstone; abundant gutter and pot casts; wave ripples; fine sandstone content = 10–40%; abundant sandstone dykes; only facies in stratigraphic contact with shoreline facies (with evidence of exposure)	Abundant unifite beds up to 1 m thick	Proximal deposits in nearshore zone of bypass and erosion
Siltstone-dominated facies	Laminae to thin beds of sandstone and siltstone; pebble lags; flat-pebble conglomerates; wave ripples; gutter casts uncommon; fine sandstone content = 5–40%	Raft-bearing beds, unifites and slides	Subtidal deposits of tempestites and storm lags swept from shoreline
Sandstone-dominated facies	Laminae to medium beds of fine sandstone and siltstone; sandstone content = 40–60%; conglomerates absent; abundant hummocky cross-stratification (generally starved forms); facies abundant near stratigraphic transition with member 3	Raft-bearing beds and slides	Shelf deposits (above storm wave base)
Member 3	Very thin to thin laminae of fine sandstone and siltstone; carbonate concretions; abundant parallel lamination and current ripples—individual beds resemble T_b–T_c turbidites; no wave ripple lamination	None	Mid shelf deposits (below storm wave base) of distal turbidite-like tempestites

ISOLATED PEBBLY MUDSTONE BED

The bed described below occurs in member 2, 48·5 m above the base of the Fortune Head section (Fig. 3) within the siltstone-dominated facies. The bed is normally graded, with a distinct grain-size break near the base that divides the bed into a lower clast-rich division and an upper fine-grained division. The total thickness of the bed is about 20 cm. The lower division is a matrix-supported, medium to very coarse sand- and pebble-bearing silty mudstone of variable thickness (3–4 cm on average). Clasts include quartz grains up to 11 mm in diameter and clasts of siltstone and dark shale up to 1·8 cm × 3 mm in cross-section.

The lower division of the pebbly mudstone bed can be subdivided into three units (Fig. 5). At the base is a discontinuous, and sometimes wispy, very fine sandstone lamina, up to 3 mm thick. Overlying this is 5–8 mm of interlaminated clayey siltstone and claystone with widely dispersed large pebbles. The third unit of the lower division consists of 2–3 cm of silty claystone with abundant, normally graded, sand- to granule-sized clasts. Ignoring the thin sand lamina at the base, the lower division is normally graded in

Fig. 3. Stratigraphic section from the Fortune Head locality (Fig. 1). The distribution of lithofacies is shown on the left of each column: Sh = shoreline facies, GC = gutter cast facies, SIS-D = siltstone-dominated facies, SS-D = sandstone-dominated facies. The full thickness of member 2 is represented in this section. The uppermost portion of member 1 is found at the base, and the contact between member 1 and 2 is at 18 m. The upper part of the measured section ends at the transition into member 3. The stratigraphic occurrences of gravity-flow deposits are shown to the right of the column: U = unifite beds, R = raft-bearing beds, S = slide horizons, P = pebbly mudstone bed. Arrows and other markings to the left of the column represent palaeocurrent measurements which are discussed in full by Myrow (1992).

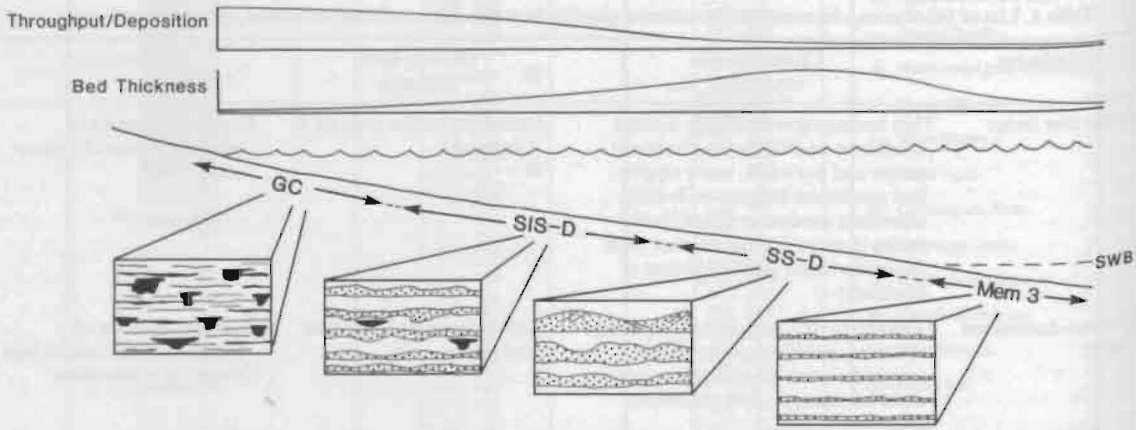

Fig. 4. The tempestite model proposed by Myrow (1992) for member 2 shows a bed-thickness trend that first increases then decreases away from the shoreline. The proximal setting is one of bypass and erosion (gutter cast facies: GC). Passing seaward, gutter casts die out and bed thickness increases (siltstone-dominated facies: SIS–D). Rare hummocky cross-stratification is formed in the thicker sandstone beds further out on the shelf (sandstone-dominated facies: SS–D). Below storm wave base (SWB), distal tempestites resemble classical turbidites (member 3).

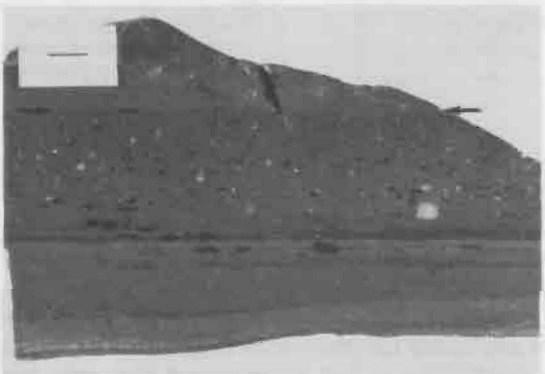

Fig. 5. Polished slab of pebbly mudstone bed. The contact between the lower clast-bearing division and upper claystone division is very sharp (arrows). The three parts of the lower division are clearly visible: (1) sandstone laminae, (2) laminated silt/clay with large clasts, and (3) clast-rich silty mudstone. Note flat to low-angle laminae within the clast-rich lower division that are defined, in part, by elongate shale clasts. Stratigraphic top is up. Scale bar = 1 cm.

maximum grain size (pebbles to granules to various grades of sand) and inverse to normally graded in terms of average grain size, as expressed in part by the percentage of muddy matrix. The clast-rich part of this lower division contains aligned elongate clasts that form cryptic horizontal or low-angle planar laminae that alternate with clast-poor silty mudstone laminae on a millimetre scale.

The upper division is a normally graded, green claystone with thin silt laminae at the base and 10–15% floating detritus of fine to coarse sand-sized quartz and sedimentary rock fragments with subhorizontal long-axis orientations. The thickness of this upper division is uncertain because the top of the bed is difficult to define in the outcrop, but it is at least 10 cm and possibly as much as 20 cm.

The lower contact of the upper division of the bed is sharply defined by 3–4 cm of light-coloured claystone with wispy, irregular, thin silt laminae. These laminae die out rapidly above into siltstone with only rare, discontinuous silt laminae. Normal grading in this division is defined by the percentage of silt and the abundance (but not size) of floating coarse grains.

Interpretation

Pebbly mudstones, with large clasts floating in a fine-grained matrix, like this bed, are generally interpreted as debris-flow deposits (e.g. Crowell, 1957—although Crowell did not refer to debris flow, he did specify origin from 'viscous sluggish slumps', p. 1004; Hampton, 1972). The matrix-supported texture of this mudstone bed suggests clast support by cohesive matrix strength. The style of lamination in the upper division, however, seems too regular and organized for debris-flow deposition (in the strictest sense), and yet sand is dispersed in the fine-grained matrix. These observations suggest that the depositing flow was probably of intermediate character, involving a number of different support mechanisms that changed with time (cf. Pierson, 1981; Shultz, 1984).

The basal sand laminae of the lower division may represent a lag deposited during steady-state flow, when the debris flow was more dilute and possibly turbulent. Later during flow deceleration and collapse, density and viscosity of the flow probably increased, turbulence became dampened and pseudo-laminar flow prevailed. Deposition from the transformed flow produced the second unit of the lower division, with its large clasts and interlamination of clay and silt. The inverse to normal grading in the two upper units of the lower, coarse-grained division is similar to that described by Aksu (1984) for Quaternary inferred debris flows in Baffin Bay. At the base of debris flows, a zone of strong shear is characterized by decreased strength and competence, accounting for the basal inverse grading (Hampton, 1975; Nemec & Steel, 1984). Stratification in debrites, with elongate grains parallel to flow, reflects pseudo-laminar flow conditions (Fischer, 1971). The lamination in the normally graded portion of the lower division of the bed described above is believed to be too fine to have formed from sorting along shear planes at the base of a downward thickening rigid plug during 'freezing' of the flow as described by Hampton (1975) and Aksu (1984). The lamination in the third unit of the lower division may instead indicate that the debris flow underwent a reduction in density and/or viscosity after deposition of unit 2, although we cannot provide a satisfactory explanation for formation of the lamination except by shear within the flow (see Stauffer, 1967; Carter, 1975). The interpretation we have presented for the units of the lower division of the bed involves possible flow transformations from more dilute (turbulent?) to less dilute (pseudo-laminar) to, again, more dilute to account for lamination. Such transformations suggest pulses of the flow, each with slightly different rheology. We favour locally surging flow components to account for fluctuations in properties, and not amalgamation of distinct flow events.

The upper division was probably deposited from an even more dilute part of the flow with significantly less strength. The silt laminae that characterize the lower part of this upper division are similar to Stow & Shanmugam's (1980) T_4 and T_5 laminae described for fine-grained turbidites. During deposition the upper division had attained some matrix strength, as indicated by the presence of scattered sand-sized clasts. Silty lamination and the very fine grain size are inconsistent with rapid mass settling. However, if this part of the flow was uniformly turbulent and dilute, the coarse grains would have settled to the base of the flow. The flow was therefore probably a moderately dilute, partly turbulent debris flow. The dilution, and therefore the degree of turbulence, may have been more important at the base of this upper division, where the laminae are best developed.

The contact between the upper and lower divisions is a rapid gradation. If these divisions were not deposited by two distinct flows, then there must have been two different density regimes within one flow (i.e. a stratified flow). Hampton (1972) showed experimentally that debris flows with moderate water content (70–75%) generate dense clouds of suspended material from erosion of the snout and mixing with overlying water. In a natural setting, this dispersion, moving more slowly than the head of the debris flow, would flow downslope over the top of the debris after its deposition. Depending on the quantity of sediment thrown into suspension, and the quantity of fluid inmixing, the resulting dispersion might be of variable character, but with turbulence a very likely attribute (Fisher, 1983). Postma *et al.* (1988b) describe experimental production of two-layer gravelly surge flows with a lower laminar inertial flow component and an upper turbulent suspension component. Their experimental flows were almost entirely granular (< 3% clay and added chalk powder) and generated on very high-angle slopes (25°), so they are not directly analogous to the deposits described herein. However, their deposits contained reversed grading and layers of oversized clasts, demonstrating that stratified flows are capable of generating such features. Whether a 'surface transformation' like that described by Hampton (1972) and Postma *et al.* (1988b) could yield a dilute, turbulent slurry with minor cohesive strength has never been demonstrated. Such a flow would account for features of the pebbly mudstone bed.

If the lower division of this bed was transported predominantly as a debris flow, then it may be possible to calculate the strength of the flowing debris and the slope on which it came to rest. Two parameters of the flow must be estimated in order to perform these calculations: density of the debris (for calculation of strength and slope) and thickness of the bed at the time of deposition (for calculation of slope). Modern subaerial debris flows have densities of 2·0–2·4 g cm^{-3}. Little is known about the densities of modern subaqueous debris flows, but an estimate of 2·0 g cm^{-3} seems reasonable (cf. Hiscott & James, 1985). A value of 6 cm, which is twice the present thickness of the upper units of the lower clast-rich division, is used as an estimate of original bed thickness. This value seems reasonable given the high mud/silt content, and agrees with estimates of compaction derived from thickening of laminae into carbonate concretions within the Chapel Island Formation (Myrow, 1987). These estimates are bracketed to cover a range of possible values, from 1·5 to 2·4 g cm^{-3} for density and 4·5 to 7·5 cm for bed thickness (1·5 and 2·5 times the observed thickness for the pebbly division). Debris strength can be calculated from maximum clast size (D_{max}) using the formula:

$$D_{max} = 8{\cdot}8\, k/[g(s-f)],$$

where k is yield strength, s is density of the clast, f is density of the matrix and g is gravitational acceleration (Hampton, 1970). Using a D_{max} of 1·1 cm and a value of 2·65 (quartz) for s, strength values for fluid densities of 1·5, 2·0 and 2·4 g cm^{-3} are given in Table 2. These values were used in estimation of slope (Hampton, 1970) using the formula:

$$T_c = k/(y \sin\theta),$$

where T_c is critical thickness, y is unit weight of debris and θ is slope angle. Table 2 contains the slope calculations using the upper and lower bracketing values of fluid density and bed thickness. The calculated range of slope is 0·17–3·66°. Using the assumed values of $T_c = 6$ cm and $f = 2{\cdot}0$ g cm^{-3}, the strength of the flowing debris is estimated at 79·7 dynes cm^{-2}, with deposition on a slope of 0·77°. Elevated pore fluid pressures may have significantly reduced strength and allowed transport onto much lower slopes (Pierson, 1981), so the slope values should be considered a maximum.

Table 2. Debris strength and slope estimates for pebbly mudstone bed using equations and values in the text.

f (g cm^{-3})	k (dyn cm^{-2})	T_c (cm)	sin θ	Slope (degrees)
1·5	140·9	4·5	0·0639	3·66
		7·5	0·0383	2·20
2·4	30·6	4·5	0·0050	0·29
		7·5	0·0030	0·17
2·0	79·7	6	0·0135	0·77

DISTURBED BEDS

The disturbed beds consist of slightly to moderately disrupted, thinly interbedded siltstone and sandstone (Fig. 6), underlain and overlain by horizontal strata of similar character. These disturbed zones are not bounded by shear planes, and therefore have cryptic boundaries. Internally, these units contain swirled and rolled laminae and small recumbent folds indicating plastic deformation of semi-consolidated sediment. There is minor evidence for brittle deformation in the form of tabular clasts and partially detached layers with relatively sharp terminations. Disrupted beds can be reconstructed using distinctive marker beds. Folds in the strata lack a consistent vergence.

Individual beds show varying degrees of deformation from thin zones of slightly undulose strata to highly disrupted zones. In one area, over 80 m of nearly continuous lateral exposure shows evidence of progressive deformation along the bed; this is illustrated in a line drawing of the outcrop in Fig. 7. This exposure has been divided into eastern and western sections (Fig. 7A–C & D–G, respectively). The fundamental feature of this exposure is a clearly defined lateral change from (i) well-bedded sandstone and

Fig. 6. Disturbed bedding in centre of photograph grades laterally into relatively undisturbed bedding at the top. Stratigraphic top to upper right. Scale is 15 cm long.

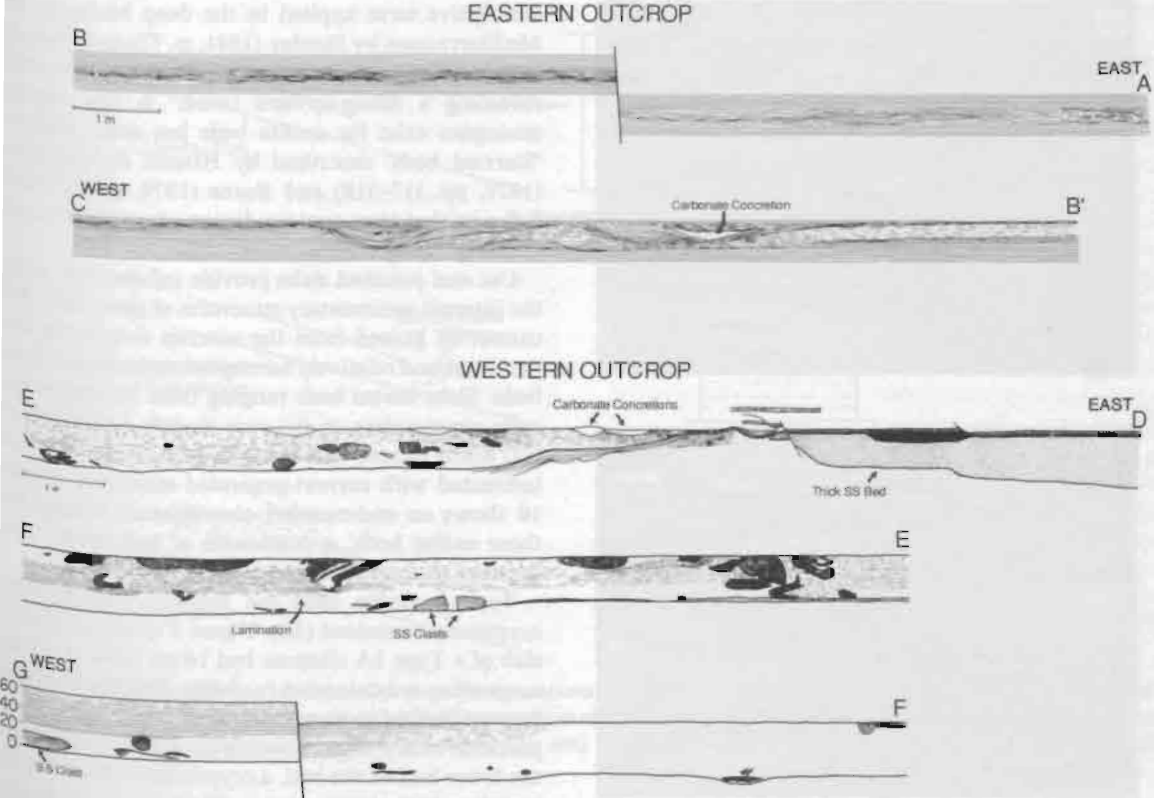

Fig. 7. Sketch of outcrop described in text. There is an 'eastern outcrop' of disturbed bedding and a 'western outcrop' with a raft-bearing bed and channellized sandstone bed. These two exposures are separated by a covered interval several metres wide. The exposure is otherwise continuous, with the exception of a covered interval of 1·8 m between B and B' in the eastern outcrop. The scale for each part of the outcrop is the same.

siltstone layers with disturbed bedding in the eastern section, to (ii) a clast-rich siltstone debrite bed in the western section (see section on 'raft-bearing beds' for a full description of the western section). Although a covered interval of 8–10 cm separates these sections, their correlation is ensured by marker beds that directly underlie and overlie this zone.

At the eastern limit of the eastern section, only the slightest hint of bedding disruption can be detected. From east to west there is a progressive increase in the intensity and depth of bedding disruption. To the east (Fig. 7A–B) the strata are rolled and bent into widely spaced synformal folds and small ball-and-pillow structures. At the western end of this eastern section (Fig. 7B'–C) the strata are strongly disrupted and, although there is no well-defined basal shear plane, the base of the deformed strata is better defined here than elsewhere. The complete eastern section may record a lateral transition from more or less *in situ* deformation to sliding, or downslope movement.

The lack of either well-defined slip surfaces or consistent vergence to the folds in the disturbed beds suggests that significant downslope movement did not occur. The disruption appears to have taken place by loading, possibly associated with high pore fluid pressures and partial liquefaction.

UNIFITE BEDS

Unifite beds are composed of graded to non-graded siltstone and silty mudstone characterized by a lack of obvious internal structure or lamination (Figs 8 & 9). These beds range from 10 to 70 cm thick (36 cm on average) and have very sharp bases and gradational to very sharp tops. These beds are termed unifites, a

Fig. 8. (a) Channel-like margin (outlined in ink) of macroscopically homogeneous unifite bed at FH-27·3. (b) Enlargement showing termination of surrounding, flat-lying beds against curved lower surface of unifite bed (arrows). Stratigraphic top is up. Scale divisions = 10 cm.

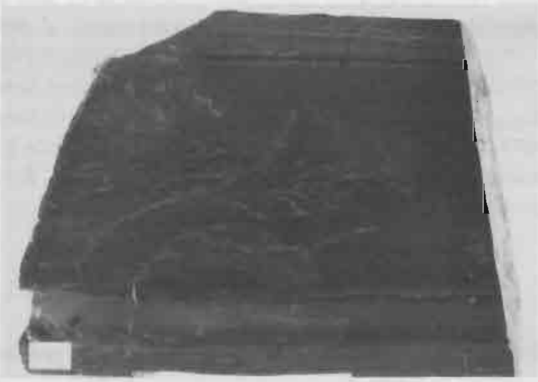

Fig. 9. Polished slab of Type 1A unifite. Close examination reveals no visible size grading or lamination. Note the extremely sharp base and top and small pseudonodules detached from base of the bed. Stratigraphic top is up. Scale bar = 1 cm long.

descriptive term applied in the deep basins of the Mediterranean by Stanley (1981, p. 77) to 'structureless or faintly laminated, often thick, mud layer[s] revealing a fining-upward trend.' A few ancient analogues exist for unifite beds but most, like the 'Slurried beds' described by Hiscott & Middleton (1979, pp. 317–318) and Burne (1970, pp. 221–226) differ in that they contain dispersed or graded sand grains or rip-up clasts.

Cut and polished slabs provide information about the internal sedimentary structures of these beds that cannot be gained from the outcrop due to the fine grain size and relatively homogeneous texture of these beds. Slabs reveal beds ranging from (i) megascopically structureless, to (ii) very slightly laminated, to (iii) subtly graded and laminated, to (iv) graded and laminated with current-generated structures. Figure 10 shows an end-member classification scheme for these unifite beds; a continuum of bed styles exists between the end-members.

Type 1 unifite beds are massive (1A) or very cryptically laminated (1B). Figure 9 shows a polished slab of a Type 1A siltstone bed 14 cm thick in which no grading or lamination is visible. The bed contains coarse silt-sized to very fine sand-sized, elongate shale particles with a slight bedding-parallel fabric, and in the lower half of the bed, a cryptic lamination due to variation in their abundance. The base of this bed is sharp and displays small flame and load structures. The upper surface is remarkably sharp and planar.

Type 2 unifite beds contain subtle delicate laminae but lack strong grading. The laminae, although not generally detectable on the outcrop, appear in cut and polished slabs as subtle changes in grain size, generally restricted to either the bases or tops of the beds. Subtle grading, if present, consists of a slight increase in clay content at the top of the bed.

Type 3 unifites are characterized by distinct lamination and grading, either of which may or may not be detectable in outcrop. Grooves, flute marks and small gutter casts have been noted on the base of a few beds. Typically, beds contain a lower division of silty mudstone with silt laminae and a thicker, light-green claystone cap. Silt laminae range from thick (up to 1 cm) and well defined to very thin, streaky and indistinct. These thicker laminae locally contain subtle, low-angle cross-laminae. A few beds contain thin coarse units at their base consisting of parallel-laminated and rippled medium and fine sandstone. Shale particles ranging in size from silt to fine sand are found dispersed throughout many of these beds.

Unifite beds are generally tabular over the full

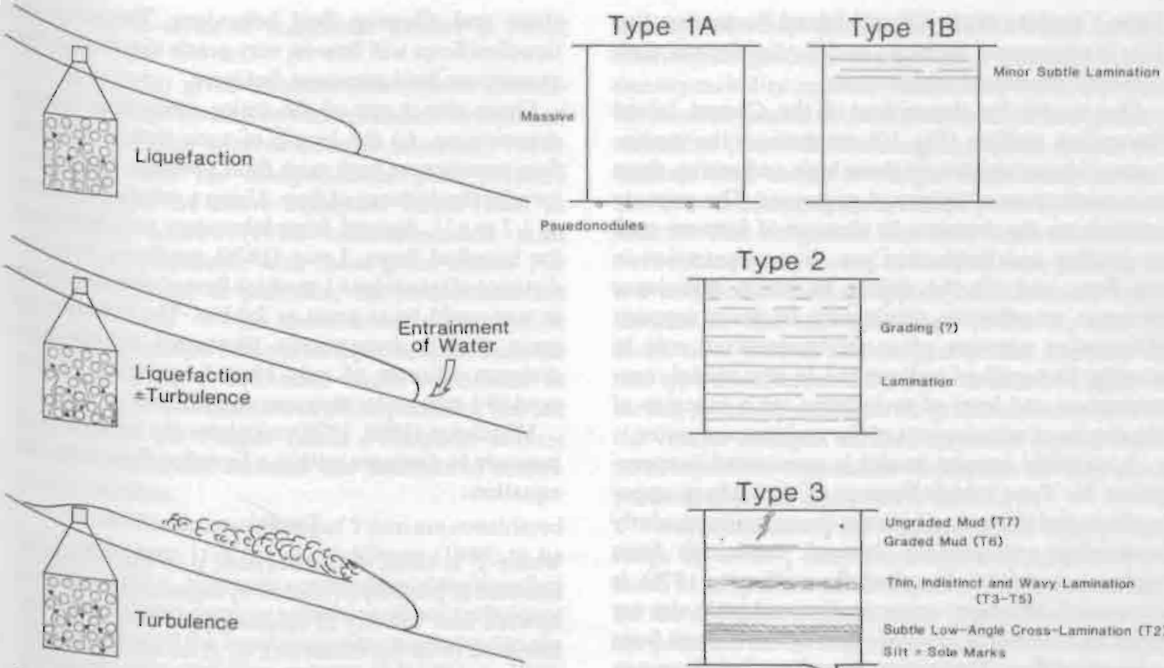

Fig. 10. Unifite classification scheme and process interpretations. Dominant characteristics of the three unifite types are given on the right. These represent a continuum of bed types whose characteristics are thought to be a function of degree of water entrainment and the extent of turbulence (represented on left). Turbidite divisions for Type 3 unifites are those of Stow & Shanmugam (1980).

extent of their outcrop exposure. A notable exception is a 35-cm-thick, macroscopically homogeneous siltstone bed, tabular over most of its exposure, that dramatically pinches to zero thickness over a distance of 1·4 m (Fig. 8). The geometry of the bed over this interval is that of a channel margin with a horizontal upper surface and gently curved lower surface that truncates underlying sandstone and siltstone beds.

Interpretation

Unifite beds were deposited by single events. This conclusion is based on their: (i) anomalous thickness relative to the beds in surrounding strata, (ii) relatively homogeneous grain size and texture, (iii) normal grading (some beds), and (iv) sequences of sedimentary structures (some beds). The thicknesses of these unifite beds are at least an order of magnitude greater than the average thickness of the stratigraphically underlying and overlying beds. Besides reworking of sediment by burrowing organisms, for which there is no evidence, there is no reasonable mechanism to generate such beds other than resedimentation of unlithified sediment. The differences in internal structure between the various types of unifites argues for variation in depositional processes, as outlined below.

These beds are similar to unifites described from the Mediterranean Sea by Rupke & Stanley (1974), Stanley *et al.* (1980), Stanley (1981) and Stanley & Maldonado (1981). Stanley's (1981, pp. 79 & 81) idealized unifite sequence consists '. . . at the base, of graded, faintly laminated, silt and silty clay . . ., and trends upward to more subtly or not graded, structureless and somewhat finer-grained mud . . .' Most of his unifites fall into one of two groups: (i) uniform, subtly graded muds, and (ii) faintly laminated and graded muds. In comparison with published 'ideal' fine-grained turbidite sequences, Stanley (1981) associates the first group with Piper's (1978) E_2 and E_3 divisions and Stow & Shanmugam's (1980) T_6 and T_7 divisions. The faintly laminated beds are compared to the E_1–E_2 and T_4–T_6 divisions of these workers, respectively (Stanley, 1981, p. 79). When compared with the fine-grained turbidite models of Piper (1978) and Stow & Shanmugam (1980), the massive, non-laminated

Type 1 unifites of the Chapel Island Formation (see Fig. 9) correspond to the E_3 and T_7 divisions of their idealized sequences, respectively.

Our model for deposition of the Chapel Island Formation unifites (Fig. 10) emphasizes the continuum of characteristics of these beds and relates them to a continuum of proposed processes. The primary controls on the presence or absence of features such as grading and lamination are: (i) concentration in the flow, and (ii) the degree to which turbulence becomes an effective mechanism of grain support. (Dispersive pressure plays an insignificant role in muddy, fine-grained sediments.) In this model, concentration and level of turbulence are a function of the degree of entrainment of the ambient sea water.

A turbidity current model is considered inappropriate for Type 1 beds because of their sharp upper surfaces and the lack of internal structure, particularly lamination and grading. Instead, deposition from high-concentration liquefied flows (Lowe, 1976) is advocated. Sediment concentration and grain size are primary controls on the character of the deposit from a liquefied flow. Deposition from liquefied flows can be described in terms of a hindered-settling model (Middleton & Southard, 1984, pp. 418–421) in which the interfaces between clear water and the dispersion, and between the dispersion and the deposited sediment, converge, with the dispersion maintaining constant density throughout. In high-concentration flows with limited size range, there will be limited size segregation and therefore little or no grading (Middleton & Southard, 1984, p. 89). High concentrations would certainly preclude any traction processes (Middleton & Hampton, 1973; Lowe, 1976). Such flows would be non-turbulent (Lowe, 1976). Lower concentrations and a more variable grain-size distribution would lead to turbulence and the development of grading, lamination and other internal sedimentary structures.

Terzaghi (1950, 1956) pioneered the study of spontaneous liquefaction and described natural occurrences in Holland and Norway. Liquefied flow results from gravity-induced movement of a liquefied sediment, or from liquefaction of a moving sediment slide (Lowe, 1976, p. 289). Spontaneous liquefaction occurs in loosely packed or metastable sediment in which vibrations or other stresses directly, or indirectly, increase the pore fluid pressure (Seed, 1968). Attainment of a stable configuration with closer packing and reduced pore volume requires the active displacement of water which creates high pore fluid pressures that suspend and separate the grains, counteracting normal stress and allowing fluid behaviour. Theoretically, liquefied flows will flow on very gentle slopes until the excess pore fluid pressures dissipate.

Grain size is one of the major controlling factors determining: (i) the length of time that a liquefied flow experiences high pore fluid pressure, and therefore (ii) the distance of flow. Using a velocity estimate of $1{\cdot}7\ \mathrm{m\ s^{-1}}$, derived from laboratory and field data for liquefied flows, Lowe (1976) predicted that the distance of travel for a 1-m-thick flow of silt 0·0625 mm in size could be as great as 2·0 km. The influence of grain size is dramatically illustrated by his flow distance estimate of only 19 m for medium–coarse sand (0·1 cm) under the same conditions.

Middleton (1969, 1970) calculates the time for pore pressure to dissipate within a liquefied flow using the equation:

$$T = dp/v,$$

where T is time, d is thickness, p is the fractional increase in porosity produced by liquefaction, and v is upward flow velocity of escaping pore fluid. For very fine sand (0·01 cm diameter), he gives estimates for p and v of 5% and $0{\cdot}01\ \mathrm{cm\ s^{-1}}$, respectively. Based on grain size and compaction data from carbonate concretions in the Chapel Island Formation (Myrow, 1987), a pre-compaction thickness of 75 cm (twice the average bed thickness) is considered a reasonable estimate for the unifites of this study. The time for dissipation of excess pore fluid pressure using these values would be approximately 6 min. Pore fluid expulsion times are a direct function of permeability, which means that the addition of small clay-sized particles, which clog pore necks, will cause a marked increase in these times. Because the above estimate is valid for well-sorted sand (0·01 cm), and the Chapel Island unifites are mostly clayey siltstones, it follows that the expulsion of pore fluid in the unifites must have taken considerably longer than 6 min.

Van der Knaap & Eijpe (1968) make a similar attempt to calculate 'relaxation' time using the equation:

$$T = 4L^2/c\pi^2,$$

where the relaxation time, T, is calculated from the sediment thickness, L, and a relaxation coefficient, c, is calculated from the permeability of the sediment. Formulas of Van der Knaap & Eijpe (1968) indicate a linear relationship between permeability and relaxation time. If the permeability of the unifites in this study was an order of magnitude smaller than the well-sorted fine sand used in their experiments (a reasonable assumption), then the relaxation time would be an order of magnitude greater at about 27 min.

Lowe (1976) gives resedimentation rates (resedimentation time divided by bed thickness) for liquefied beds of uniform spheres (from Andersson, 1961). Assuming that the dissipation of pore pressure calculated in the above equations would result in deposition, Lowe's (1976) resedimentation rates should be comparable with those given above. For grains 0·0125 cm in diameter, the resedimentation rate is 2·7 s cm^{-1}. This means a bed 100 cm thick would resediment 4·5 min after liquefaction. A similar thickness of silt (0·0625 mm) would resediment in about 20 min. The addition of a component of fine silt and clay in the Chapel Island Formation unifites would substantially increase this estimate of resedimentation time.

Even resedimentation rates of 5 min are considered by Morgenstern (1967) and Middleton (1970) to be large enough '. . . to permit acceleration of the liquefied sand mass down slope to velocities where turbulence and mixing with the overlying water might lead to the formation of a turbidity current' (Middleton, 1970, p. 267). On theoretical grounds, therefore, the transformations envisaged for the Chapel Island unifites (Fig. 10) are reasonable.

For the clay-rich flows that deposited the Chapel Island Type 1 unifites, elevated viscosities may have damped any turbulence, prolonging conditions of laminar flow. Alternatively, the beds may have been deposited prior to the onset of significant turbulence because of either: (i) the early dissipation of excess pore pressures, or (ii) a decrease in slope before a significant distance of flow. All these alternatives are plausible. The average slope on continental shelves is only 0·12° (Morgenstern, 1967), but much higher slopes are found in certain nearshore environments (Terzaghi, 1956; Moore, 1961). The reduction in slope necessary to cause deposition of Type 1 liquefied flows might have been fairly small, and are reasonably explained by local variations in nearshore or shelf topography.

The graded and laminated Type 3 unifite beds are believed to represent those liquefied flows that became turbulent (Fig. 10), and transformed into turbidity currents. This transformation was probably accomplished by the entrainment of water, causing reduction of density and viscosity and allowing the Reynolds number to increase above values for onset of turbulence. Structures in Type 3 unifites are similar to those in fine-grained turbidites (Piper, 1978; Stow & Shanmugam, 1980), and thick unifites described by Stanley and others (see Stanley, 1981). These include distribution grading, considered a feature of low-density turbidity currents (Middleton, 1967), and thin silt laminae similar to those described for fine-grained turbidites (Stow & Shanmugam, 1980). Those silt laminae with subtle low-angle micro-cross-laminae (i.e. Stow & Shanmugam's, 1980, T_2 and T_3 divisions) indicate that silt grains, deposited from suspension, were subsequently moved as traction load, possibly as low-amplitude climbing ripples. Indistinct and wispy laminae in this bed correspond to the T_4 and T_5 divisions of Stow & Shanmugam (1980). Laminae of this type are attributed by Stow & Bowen (1980) to sorting processes acting on silt grains and clay flocs in the viscous sublayer. Graded claystone caps in these beds represent suspension deposition of clay-sized particles from the dilute tail of the flow.

Visual estimates of the clay content of Type 3 beds range up to 50% (claystone caps may constitute as much as 15% of the thickness of the bed). Depending on the sediment concentration in the flows, there is the potential for the depositing flows to have had some degree of strength, assuming that at least some percentage of this clay was unflocculated. Whatever strength these flows may have had, it was not great enough to suppress turbulence—the well-developed grading and silt/mud laminae clearly reflect turbulence during deposition.

Type 2 beds exhibit characteristics transitional between Type 1 and Type 3 beds. These are therefore interpreted as the deposits of flows that reached only an intermediate stage in the evolution of a liquefied flow (Type 1) to a fully turbulent flow (Type 3). The development of lamination and minor grading indicate that some size segregation occurred. Development of these features is again considered a by-product of entrainment of water into a liquefied flow (Fig. 10). Grain support mechanisms in Type 2 flows probably included high pore fluid pressure and minor turbulence.

RAFT-BEARING BEDS

Raft-bearing beds consist of siltstone, similar in character to the unifite beds, that contain clasts, or 'rafts', of thinly interbedded sandstone and siltstone similar to the overlying and underlying strata of the siltstone-dominated facies. These raft-bearing beds also display laminae that are commonly swirled and deformed around clasts. Bedding thickness varies from 16 to 110 cm (45–50 cm on average). Grading, where present, is defined by a thin division (<2 cm thick) at the base only, which is composed of fine to coarse sandstone. Where lamination is absent or cryptic and rafts are particularly sparse, the distinction between these beds and unifite beds is difficult, suggesting a continuum in depositional processes. This is illustrated by a unifite bed, sketched in Fig. 11, in which eroded pieces of the underlying strata are incorporated at its base.

Rafts are found in a variety of sizes, from a few centimetres across to large ones 60 × 20 cm in cross-section. The rafts occur as contorted masses, angular fragments, and coherent slabs with relatively intact, gently folded to flat-lying bedding (Figs 12 & 13). These rafts are found at many positions within the beds, in some cases concentrated within the lower third of a bed, or within the central and upper parts of a bed.

Some raft-bearing beds show signs of incorporation of overlying strata by gravitational sinking. Thinly interbedded sandstone and siltstone are captured in various states of detachment and loading into underlying raft-bearing beds (Fig. 14).

Raft-bearing beds are generally tabular: one bed is traceable for over 120 m without any change in thickness. In other cases, the thickness and character

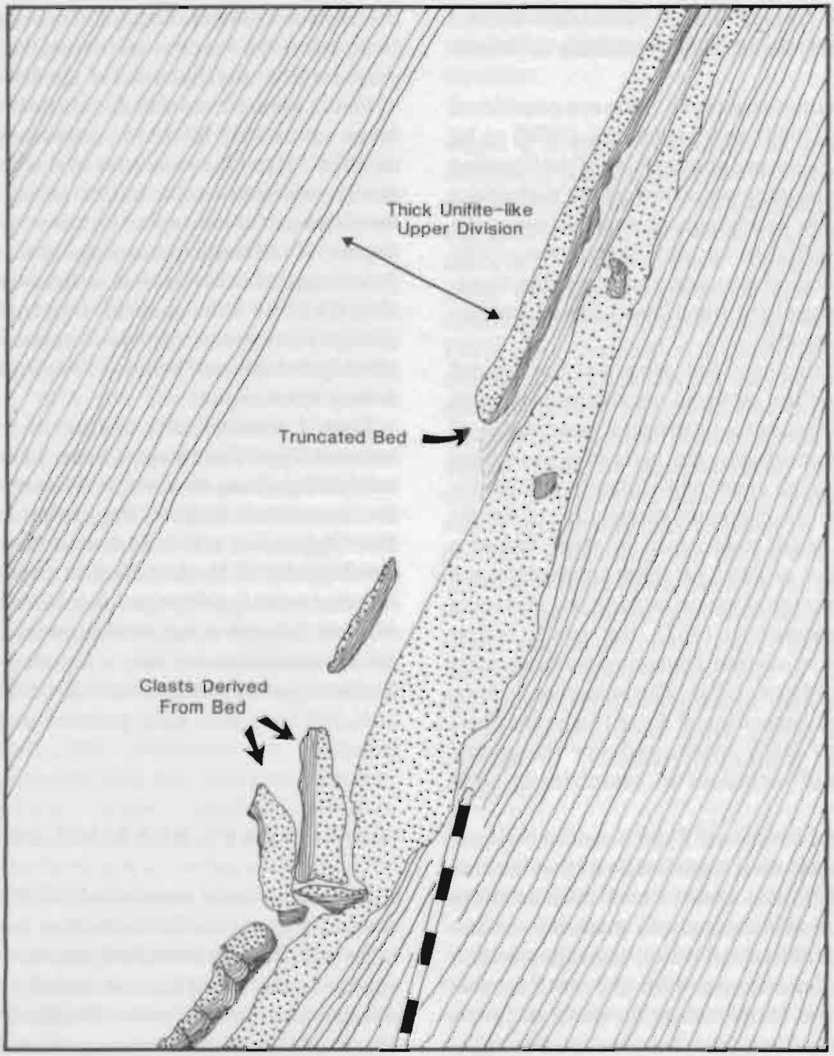

Fig. 11. Sketch of unifite bed with large blocks at base. Laminae in the sandstone clasts match those of the upper sandstone bed (on right). Scale divisions = 10 cm. Stratigraphic top is to the upper left.

Fig. 12. Bed illustrated in Fig. 7 showing a large folded raft at the top of the bed and two rounded sandstone clasts on the right (A and B). See text for details. Stratigraphic top is up. Upper and lower bed contacts arrowed. Scale divisions = 10 cm.

Fig. 13. Close-up of top of large raft from bed illustrated in Fig. 7. The laminae/very thin beds within the raft are sharply terminated at the top, including the carbonate nodule (N). Stratigraphic top is up. Scale is 15 cm long.

of these raft-bearing beds change along their length. The best exposed example is the 55–60-cm-thick raft-bearing bed in the western section of Fig. 7(D–G). Near the eastern end of this western section the raft-bearing bed pinches out. At the point of termination, a massive, grey, fine sandstone bed cuts downsection toward the east. This bed increases in thickness rapidly before the end of the outcrop.

The raft-bearing bed in Fig. 7 is noteworthy, in part, because the top of the bed is remarkably planar. One large raft of interlaminated sandstone and siltstone shows evidence of sharp truncation, with the upper bed surface cutting indiscriminately across the raft laminae at a high angle to the internal bedding of the raft (Fig. 13).

Interpretation

The presence of clasts at all levels in these beds indicates that the sediment had strength, a property of debris flows. The sedimentary structures/fabrics of debris flows include: matrix support, random fabrics, variable clast size, variable matrix, rip-up clasts, rafts, inverse grading and possible flow structures (Nardin *et al.*, 1979), many of which are typical of the raft-bearing beds.

Other workers have described large, variably deformed rafts of eroded underlying or adjacent sediment from a wide variety of debris-flow deposits. These include 'slurry' deposits (Wood & Smith, 1957; Burne, 1970; Morris, 1971; Hiscott & Middleton, 1979), pebbly mudstones (Dott, 1963; Stanley, 1975; Alvarez *et al.*, 1985; Hein, 1985), and clast- to matrix-supported conglomerates (Fisher, 1983). These large clasts can move within low-velocity flows over low slopes (Middleton & Hampton, 1973).

Debris flows commonly carry clasts that project above the top of the bed (Johnson, 1970). Many of the raft-bearing beds have rafts exclusively in their upper parts, but none project above the top of the bed. One possibility is that the flows were not dense enough to support projecting clasts: the clasts were at best neutrally buoyant. A second alternative is that the small percentage of these clasts that projected above the top of the bed at the time of deposition either sank down into the underlying sediment shortly after deposition, were erosively planed off, or both. The depression of the laminae below rafts supports some gravitational sinking after deposition. The erosional planation suggests, but does not prove conclusively, that clasts projected above the top of some beds after deposition.

Some of the features in the bed shown in Fig. 7 may be partly explained by rheological variation within the depositing flow. The upper part of this bed contains numerous large rafts, which indicates that the flow must have exhibited substantial strength. One raft appears, with slight reconstruction, to sample a stratigraphic thickness of approximately 25 cm. In muddy debris flows, strength and buoyancy effects, necessary for the support of large clasts, are provided by the matrix. Strength is divided into a cohesive component, from electrostatic attraction of clay minerals, and, more importantly (Trask, 1959; Pierson, 1981), frictional strength, due to grain-to-grain

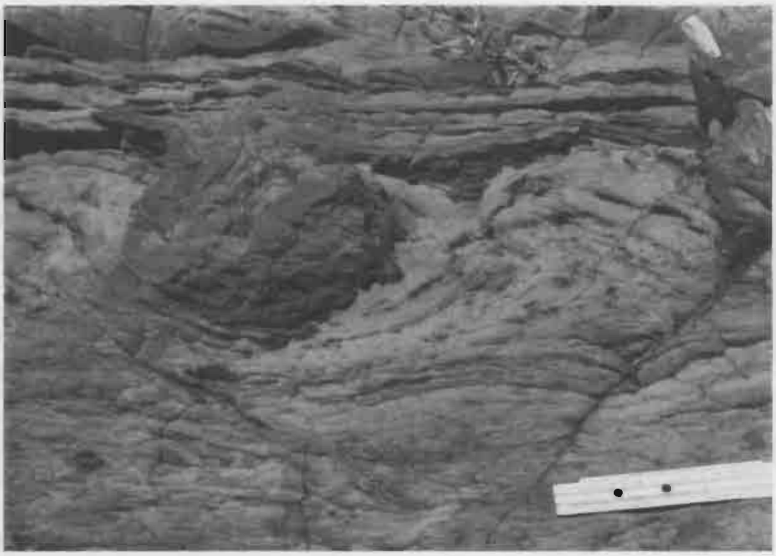

Fig. 14. Downward loading of overlying sandstone bed and depression of laminae directly underneath the load feature. Stratigraphic top is up. Scale is 15 cm long.

contacts. Buoyancy is enhanced with increasing concentration of fine-grained matrix. Across most of the outcrop, the bed in Fig. 7 contains laminae that are best developed in the central part of the bed. Laminae are curved downward and compressed directly under large rafts, and to a lesser degree above rafts as well. In the upper part of this bed, these are disrupted, particularly in zones where rafts are most abundant (see Fig. 7), suggesting that if the flow was laminar, the mixing and swirling of laminae was due to interactions among clasts. Raft-bearing debrites described by Burne (1970), Morris (1971) and Hiscott & Middleton (1979) also contain swirled and contorted internal structure. The lamination in these beds is thinner and less well defined than the lamination in the overlying and underlying strata, or that found in the rafts. The origin of this lamination is uncertain, and either represents lamination formed during debris flows or, more likely, original bedding that was distended and mildly deformed during transformation from sliding to debris flows.

The lower part of the bed is generally massive and contains scattered clasts, but most of the larger clasts can be seen to be lying directly on the lower bed surface. Restricted to this part of the bed are rounded, grey, faintly laminated sandstone clasts that are similar in colour, grain size, and texture to the thick sandstone bed at the eastern end of the western outcrop (Figs 7 & 12). The grey sandstone clasts are found *only* in this position within the bed (Fig. 7). The similarity in colour, grain size, etc., to the sand body shown in Fig. 7 suggests that these clasts were derived from the upslope extension of that sand body, or from a similar sand body, when it was in a semi-lithified state (the clasts have bent laminae). The fact that these blocks were resting directly on the underlying substrate at the time of deposition indicates they were too heavy to be carried by the flow. The presence of other, less-dense, rafts on the base of the bed indicates that, possibly as a result of liquefaction, elevated pore pressures or higher water content, the shear strength in this part of the flow may have dropped significantly in the late stages of deposition, although not completely (as indicated by the numerous clasts floating well above the base of this lower layer). This loss of strength might be attributable to (i) high shear stresses in this part of the flow and/or (ii) an increase in water content in the lower layer by incorporation of water below the nose of the flow ('surface transition', Fisher, 1983). The heavier, less-buoyant clasts could have then settled to the bottom, whilst others were maintained above the sea floor. Aksu (1984) infers a similar situation for some Quaternary submarine debrites in which the lower part of the flow apparently exceeded the plastic limit and behaved as a liquid, whilst the upper part of the flow continued to deform plastically. An alternative scenario would see deposition of the bed completed without the flow exhibiting any liquid behaviour, but with residual high pore fluid pressure in the lower layer facilitating partial, post-depositional liquefaction.

The middle and upper parts of the bed might have been riding as a passive semi-rigid plug at the time. Coleman (1981) describes mudflows from the Mississippi Delta in which the method of transport involves movement of a rigid plug over and within a zone of liquefied mud. He notes for these flows that 'the presence of partially disintegrated rafted blocks suggests laminar or plug flow rather than turbulent flow' (p. 74).

The raft-bearing bed in Fig. 7 was deposited in a channel. Of the original sediment that must have been excavated to form the shallow channel, only a small percentage can be accounted for by the rafts. The rest must have been carried downslope, either with the flow, or in part by processes acting prior to arrival of the flow (e.g. slumping).

It is remarkable that the stratigraphic position of the top of the raft-bearing bed in Fig. 7 is identical to the top of the corresponding slightly disturbed zone to the east. Assuming that, before failure and movement of the sediment, the sea floor was essentially planar on a large scale, then either the deposition of this bed resulted in no change in topography, or erosive processes subsequently acted on the new depositional surface to re-establish planarity. The erosional top to this bed indeed indicates that currents acted to reduce topographical irregularities after deposition. Anderton (1976) describes flat, extensive erosional surfaces from the shallow-marine environment. Slump horizons in the Cretaceous deltaic deposits described by Hubert (1972) also have bevelled upper surfaces that were cut by currents prior to renewed deposition. The nature of these eroding currents is unknown, and sediments associated with the currents have not been recognized.

If the disturbed horizon and the raft-bearing bed in Fig. 7 were formed simultaneously by a single trigger (e.g. storm, seismic shock), then this stratigraphic level might represent either the lateral development of a debris flow from an incipient slump, or a difference in deformation style (slide vs. slump) related to a difference in such factors as the local depositional slope, degree of liquefaction, or variation of lithification/sediment strength. Alternatively, a single trigger (e.g. storm, seismic shock) may have generated both slides and debris flows.

SLIDE DEPOSITS

Slide deposits are not common in member 2 and one well-exposed example from the Fortune Head locality is described below. The slump/slide horizon consists of 35–40 cm of folded thin and very thin beds of sandstone and siltstone that overlie a shear zone/surface (Figs 15 & 16). The folded strata are overlain by a hummocky cross-stratified sandstone bed of variable thickness. At the eastern end of the outcrop, a poorly defined concave-up shear zone cuts down through the 35–40 cm of strata at a high angle to bedding over a lateral distance of 2–3 m (Fig. 16). Towards the west, the shear zone becomes bedding-parallel and forms a well-defined *décollement* surface over which the strata are regularly folded into sharp anticlines and wide, gentle synclines (up to 40 cm in amplitude). Overlying the shear zone to the east, the folded strata are irregular, and small isolated synclines of folded strata are surrounded, laterally and below, by disrupted siltstone. Fine sandstone fills in over the

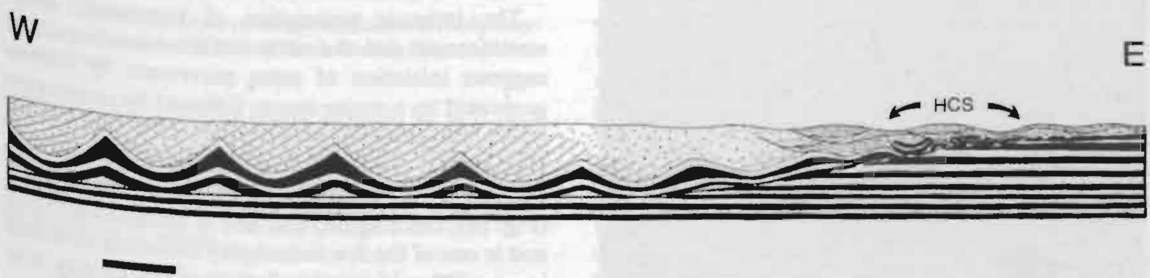

Fig. 15. Slide horizon covered by hummocky cross-stratification (HCS). Thinly interbedded sandstones and siltstones overlie the slide scar to the east and a well-defined décollement to the west. In the eastern part of the outcrop, hummocky cross-stratification is found in the overlying sandstone. To the west, large avalanche slip-faces bury the sharp anticlines of the underlying buckled beds.

folded strata as large slipfaces up to 40 cm thick in the deep synclinal depressions, recording sand migration laterally into these depressions. Above the curved slide scar to the east the sandstone bed thins to 8–10 cm, displays well-developed hummocky cross-stratification (Figs 16 & 17), and is capped with small two-dimensional symmetrical-crested ripples.

Interpretation

The overall structure of this slide deposit indicates that the layers essentially buckled, sliding along a fairly well-defined surface/zone, with the greatest amount of deformation occurring towards the west. The large folds in the western part of the outcrop formed as an accommodation of stresses created by the sliding mass. The deformation was compressional, and the sediment responded plastically, but with some integrity as evidenced by the sharp crests of the anticlines. A close temporal association of sand deposition with sliding is suggested by: (i) the lack of a fine-grained drape over the buckled surface (a clay drape would have been relatively easy to deposit and difficult to erode in the synclinal depressions), and (ii) the anomalous thickness of the overlying sandstone bed (an order of magnitude thicker than the average thickness of surrounding sandstone beds).

Fig. 16. Buckled horizon showing increased depth of disruption from top to bottom of photograph (outlined in ink), large synformal structures to the left of the notebook, and overlying hummocky cross-stratified bed (H). Stratigraphic top is to left. Notebook is 18·5 cm long.

Only a few examples of hummocky cross-stratified sandstone occur in the several hundred metres of strata below this horizon. Hummocky bedforms are considered to be a storm-generated feature, formed, at least partially, under the influence of long-lasting oscillatory currents (Southard *et al.*, 1989). Storm waves have been considered as triggers for slope failures by many authors (e.g. Dott, 1963; Henkel, 1970; Coleman, 1981; McGregor, 1981; Koning, 1982; Prior & Coleman, 1982; Saxov, 1982). Henkel (1970), studying the effects of Hurricane Camille in the Gulf of Mexico, provided a theoretical basis for understanding wave loading of submarine sediments. He described how fluctuations in bottom pressure associated with the passage of large surface waves create cyclical shear stresses that cause an increase in pore fluid pressure and associated decrease in strength, leading to failure and downslope movement (also see Clukey *et al.*, 1985; Prior *et al.*, 1989). According to Watkins & Kraft (1978), storm-induced sediment failure is plausible in water depths of up to 150 m. Henkel (1970) and others (Hampton *et al.*, 1978; Coleman, 1981; Cita *et al.*, 1982; Lindsay *et al.*, 1984; Kraft *et al.*, 1985) have attributed failures on the Mississippi Delta to wave-induced stresses. Storm waves off of the Huanghe Delta (Yellow River) reach heights of up to 7 m and are considered a probable cause for sediment failure on the delta (Prior *et al.*, 1986, 1989).

The intimate association of hummocky cross-stratification and the slide feature described above suggests initiation of mass movement by stresses generated by a major storm, followed by storm-wave remoulding of sand transported over the slide deposit. Hummocky cross-stratified sandstone also directly overlies a raft-bearing bed at Grand Bank Head (Fig. 18). Once again, this bed is anomalously thick and is one of the few hummocky cross-stratified beds in over 200 m of member 2 strata at this locality. Few examples of a link between sediment failure and storm processes have ever been documented in the rock record, probably because of the low probability of depositing sand directly on the surface of a slide,

Fig. 17. Close-up of Fig. 16 showing hummocky cross-stratified sandstone bed that overlies the buckled surface. Stratigraphic top is up. Scale is 10 cm long.

Fig. 18. Hummocky cross-stratified sandstone directly overlying raft-bearing bed at Grand Bank locality (Fig. 1). Pen is 13·5 cm long.

generating unequivocal storm features in the sand bed, and preserving the entire sequence. In addition, most ancient slides are described from deep-water settings (e.g. slopes), where deposition is well below the influence of waves.

ROLE OF LIQUEFACTION

Liquefaction (or partial liquefaction) is believed to have been an important triggering mechanism for many of the gravity flows that formed deposits in the shallow-water sediments of member 2. One factor that led to frequent liquefaction is the abundance of silt-sized sediment. Cohesionless fine sands and silts are highly susceptible to liquefaction because they have no intergranular electrostatic attraction and yet consist of grains that are too light to shift into stable packing at the time of deposition (Terzaghi & Peck, 1948; Andresen & Bjerrum, 1967; Keller, 1982). A second factor that predisposed the sediment of member 2 to liquefaction was a high rate of accumulation. Metastable sediments commonly occur in areas with high accumulation rates where accumulation exceeds the rates of consolidation and pore water reduction (Terzaghi, 1956; Middleton & Hampton, 1973; Hein & Gorsline, 1981). Evidence for high rates of accumulation in the deposits of this study include the abundance of gravity-flow deposits, paucity of amalgamated tempestites (Goldring & Bridges, 1973; Bourgeois, 1980; Kreisa, 1981), and the abundance of sedimentary dykes in the gutter cast facies. Calculated values of bottom slopes for deposits in member 2 indicate gentle gradients, probably less than 1°. Such slopes are more than sufficient for movement of liquefied flows (Morgenstern, 1967; Middleton & Southard, 1977; Prior & Coleman, 1982), and associated debris flows (Hein & Gorsline, 1981) and slides (e.g. Mississippi Delta: Shepard, 1955; Terzaghi, 1956; Moore, 1961; Embley, 1982).

We distinguish some deposits as liquefied flows (Type 1 unifites) since there is some reason to believe (e.g. grain size, geometry, lack of internal structure, etc.) that initiation was by liquefaction of a fairly homogeneous source (not the interbedded sands and silts that comprised the bulk of member 2). We recognize, however, that the bulk characteristics of these deposits overlap with those of non-cohesive debris flows (e.g. Schultz, 1984), and therefore may not have differed substantially in rheological behaviour. Additionally, we believe that most of the unifite flows were substantially different in rheology from flows that produced raft-bearing beds, although a spectrum exists between these bed types that would involve different degrees of visco-plastic behaviour.

The disturbed bedding of this study represents partially liquefied sediment in which little or no downslope motion took place. Flow of similarly failed, well-bedded material would have generated some of the other gravity-flow deposits described in this study.

FLOW CHARACTER AND TRANSFORMATIONS

Characteristics of the gravity-flow deposits described here indicate that the contribution of various support mechanisms commonly varied through time and in a downcurrent direction during the deposition of individual beds. Flows characterized by multiple support mechanisms (Middleton & Hampton, 1973) and flows in which temporal transformations in character occur have been discussed by, amongst others, Middleton (1970), Fischer (1983) and Schultz (1984). We have attempted in Fig. 19 to summarize the possible transitions in dominant transport mechanisms for beds in member 2 of the Chapel Island Formation.

Unifite bed types 1–3 (Fig. 10) represent products of transitions from liquefied flows (Fig. 19a) to turbidity currents (Fig. 19b). Some liquefied flows came to rest after relatively short transport. In other cases, flow transition to a turbidity current resulted from dilution and acceleration of the parent flow. All these unifite types were deposited in an area believed to have been relatively close to the palaeoshoreline.

The presence of large rafts of thinly bedded sandstone and siltstone in raft-bearing beds is compelling evidence for flow transitions that culminated in debris flows. The field relationships illustrated in Fig. 7 record contemporaneous partial liquefaction and incipient sliding with flow of debris. It may not be possible in cases such as these to know if sliding (Fig. 19c) or partial liquefaction (Fig. 19d), or a combination of these processes, led to debris flow. In either case, the material incorporated water and suffered strength reduction at the time of failure. There is no direct evidence that debris flows (raft-bearing beds) transformed downslope into turbidity currents (Type 3 unifites). The facies segregation of these gravity-flow deposits, namely abundant unifite beds and a near-absence of raft-bearing beds in the nearshore gutter cast facies and abundant raft-bearing beds (with a lesser percentage of unifite beds) in the more distal siltstone-dominated facies (Table 1), rules out such a transition. In fact, within member 2, an overall transgressive sequence, all but one unifite bed is found within the lower 75 m (Fig. 3), attesting to the proximity of these deposits.

The pebbly mudstone bed was emplaced by a two-layer flow in which an upper turbulent part was derived from erosion of material from the top of a moving debris flow (cf. Hampton, 1972). The lower pebbly division with mud-supported textures and concentrations of large grains above the base of the bed would represent a near-laminar portion of the flow. The upper muddy division with silty laminae similar to those developed in fine-grained turbidites would have been deposited from an overriding fully turbulent lower-density part of the flow. Such a flow transformation was produced experimentally by Postma *et al.* (1988b) for highly concentrated granular flows and presumably could operate in a thinner, more mud-rich flow.

Finally, slide deposits were formed from surficial sediment sliding that terminated without disarticulation of strata and transformation into more dilute flows (Fig. 19e). Field evidence suggests that for some beds, the cyclical shear stresses associated with passage of storm waves may have caused failure.

COMPARISON WITH MODERN DELTAIC GRAVITY-FLOW DEPOSITS

The palaeoenvironmental setting of the member 2 gravity-flow deposits is based on detailed facies analysis by Myrow (1987, 1992). The inferred near-shore position of the unifite beds has an analogue in the products of sediment failures at the fronts of modern deltas. Silt flow gullies described by Prior *et al.* (1986) from the Huanghe Delta are 100–500 m

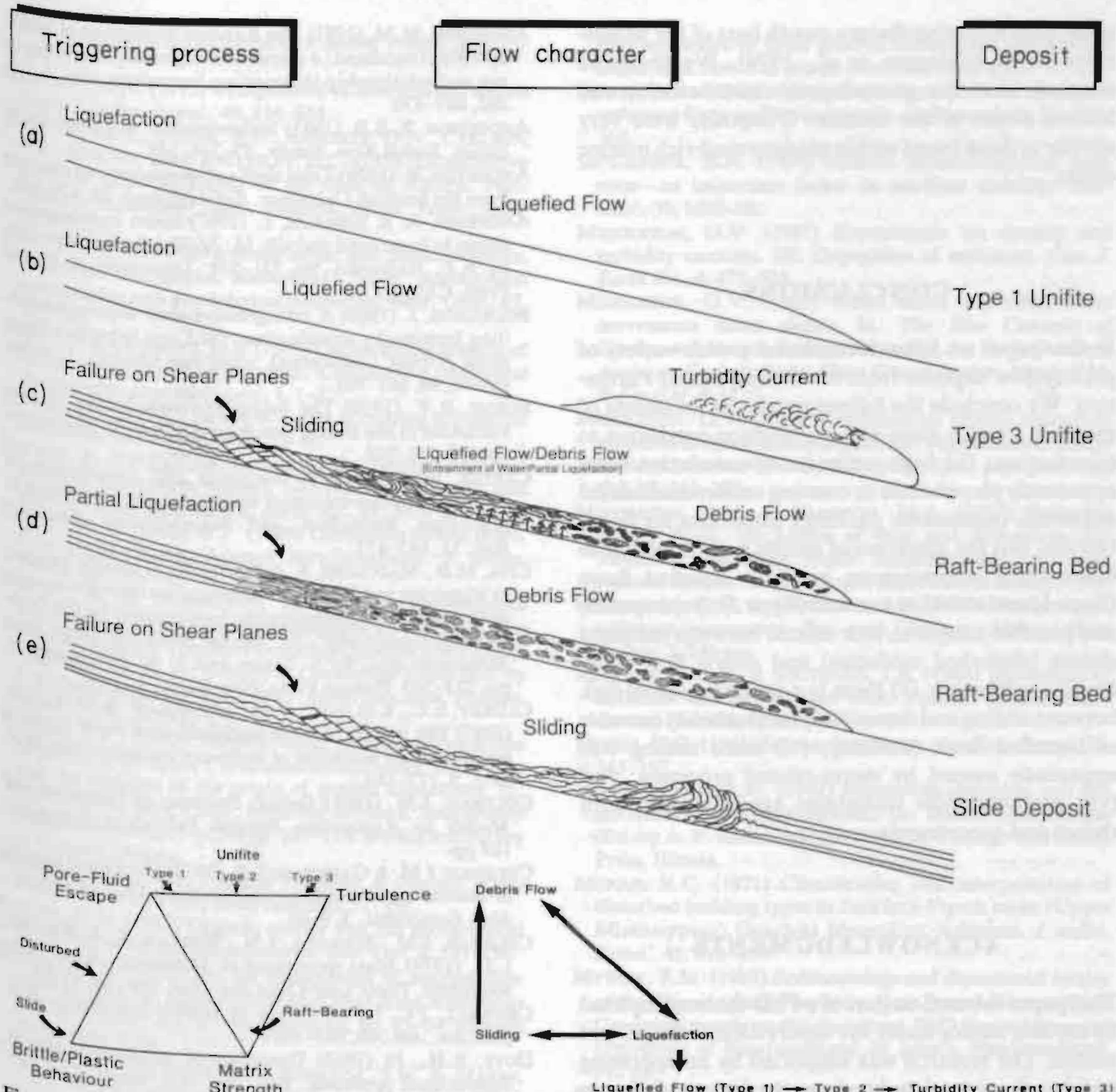

Fig. 19. Possible flow transitions (a–e) for the disorganized beds of member 2. Flow transitions are summarized with arrows in the lower right. Quadrilateral in lower left shows interpreted support mechanisms for the different bed types at the time of deposition.

wide but have a relief of only about 1 m (similar in scale to the Chapel Island deposits). The gullies are filled with acoustically transparent sediments that lack internal structure or bedding and are thought to have been generated by liquefaction. Slumps, slides and various gravity flows in the Mississippi Delta are very similar to the beds we describe from member 2, although they range to greater scales and are in somewhat greater water depths. The delta-front debris flows described by Prior *et al.* (1984) contain remoulded pieces of the sea floor, derived from faulting and sliding on the upper delta front, which are incorporated into water-rich sediments to form large lobes of block-bearing flows like the raft-bearing beds. Slump/slide horizons are also documented from modern deltaic settings, such as the rotational slumps associated with distributary mouth bars of the Mississippi River (Coleman *et al.*, 1974). We therefore conclude that the geotechnical characteristics and bottom slopes of the member 2 deposits were very similar to those found within modern mud-rich marine deltas.

CONCLUSIONS

In this paper we have documented a wide variety of gravity-flow deposits from the Chapel Island Formation. We conclude the following: (i) the sediments of member 2 had a grain-size distribution conducive to liquefaction; (ii) high sediment accumulation rates apparently played a role in creating underconsolidated sediments (metastable packing), important for liquefaction; (iii) the spectrum of unifite beds is related to flows along a continuum between liquefied flows (Type 1) and turbidity currents (Type 3); (iv) a spatial, and possible temporal, link existed between incipient sliding (disturbed subfacies) and debris flow (raft-bearing subfacies); (v) there is a possible genetic link between sliding and deposition from turbidity currents or liquefied flows (unifites); (vi) some sliding was apparently caused by storm-related processes; and (vii) proposed flow transitions are consistent with theory and experiments.

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