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Recognizing soft-sediment structures in deformed rocks of orogens

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ABSTRACT

Soft-sediment deformation structures are common on passive continental margins, in trenches at subduction zones, and in strike-slip environments. Rocks from all these tectonic environments are incorporated into orogens, where soft-sediment deformation structures should be common. However, recognizing soft-sediment structures is difficult where superimposed tectonic structures are present. In seeking characteristic features of soft-sediment deformation, it is important to separate questions that relate to physical state (lithified or unlithified) from those that address the overall kinematic style (rooted or gravity driven). One recognizable physical state is liquefaction, which produces sand that has much lower strength than interbedded mud. Hence structures which indicate that mud was stronger than adjacent sand at the time of deformation can be used as indicators of soft-sediment deformation. These include angular fragments of mud surrounded by sand, dykes of sand cutting mud, and most usefully, folded sandstone layers displaying class 3 geometry interbedded with mud layers that show class 1 geometry. All these geometries have the potential to survive overprinting by later superimposed tectonic deformation; when preserved in deformed sedimentary rocks at low metamorphic grade they are indicators of liquefaction of unlithified sediment during deformation.

1. Introduction

Soft-sediment deformation is a widespread phenomenon in a variety of tectonic settings including passive continental margins, subduction zones, and strike-slip environments. Because basins from these environments occur in orogens, soft-sediment structures would be expected to be equally common in the deformed rocks of orogens. However, separating soft-sediment deformation structures from those induced by later tectonism is challenging.

This article stems from a discussion in the early 1980s, between the senior author and Paul F. Williams, to whom this issue is dedicated, about the nature of structures in deformed clastic sedimentary rocks at low grade in central Newfoundland (Williams, 1983). Subsequent work by Paul and his students (e.g. Elliott and Williams, 1988) elsewhere in Newfoundland indicated that folds previously interpreted as synsedimentary had in fact formed much later in the deformation history. In the process, they showed that many of the features previously used as indicators of 'soft-sediment deformation' are invalid. Maltman (1994a) lamented the lack of clear criteria for recognizing the products of soft-sediment deformation where later tectonic overprints are present, a problem which has concerned both sedimentary and structural geologists for many decades. Since that time much has been learned about processes that deform sediment on present-day continental margins. The purpose of this article is briefly to review the occurrence of present-day soft-sediment deformation in environments that have the potential for preservation in future orogenic belts, and to suggest some geometrical criteria for the recognition of soft-sediment structures, particularly folds, in ancient orogens, where they have been overprinted by later tectonic events.

2. Soft-sediment deformation: definition

We define soft-sediment deformation, following Maltman (1984, 1994b) as any deformation, other than vertical compaction, of a sediment or sedimentary rock that is achieved by rearrangement of the original sedimentary particles, without internal deformation of those particles or of any interstitial cement. Deformation occurs primarily by the mechanism of grain-boundary sliding. Soft-sediment deformation, thus defined, passes imperceptibly into sedimentary processes such as debris flow. In general, processes like slumping, that leave some of the original bedding of previously deposited sediment, are included in 'soft-sediment deformation' whereas those that largely destroy pre-existing structures, such as debris flow, are generally regarded as

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sedimentation processes, but the distinction is arbitrary (Maltman, 1994b). In sandstones or conglomerates where grains have low sphericity, soft-sediment deformation typically produces no penetrative fabric at grain-scale. Where sedimentary grains have inequant shapes, however, it is possible for soft-sediment deformation to produce a fabric, though it is usually weak. Inequant grain-shapes are almost universal in fine-grained sediments (silt and mud) and indeed, fine-grained sedimentary rocks typically display a fabric (fissility), resulting from the preferred orientation of inequant grains during compaction. Because a fabric acquired during synsedimentary deformation might conceivably be emphasized mimetically during later tectonic deformation, the presence of a related fabric cannot unequivocally be taken as evidence that a structure is of tectonic origin.

3. Occurrence of soft-sediment deformation

Deformation of unlithified sediment occurs in numerous present-day environments, including unstable terrestrial slopes, areas of rapid marine and marginal marine sedimentation such as deltas, and sedimentary basins that are cut by active faults. In many such areas, soft-sediment deformation is a major hazard for human populations (e.g. Brunsden and Prior, 1984). To reduce the associated risk, significant research and engineering effort have been devoted to the prediction and even prevention of soft-sediment deformation (e.g. Hearn and Griffiths, 2001).

Actualism suggests that analogous structures must exist in ancient sedimentary rocks, and studies in undeformed sedimentary basins have typically been successful in identifying the products of soft-sediment deformation where contemporary tectonic structures are absent (e.g. Collinson, 1994; Strachan and Alsop, 2006; Strachan, 2008). In many such areas, soft-sediment deformation is associated with gravitationally driven movement of sediment down slopes that were formed during sediment deposition. The acquisition of deep seismic reflection profiles from continental slopes has enormously increased our knowledge of such structures on passive continental margins (e.g. Bilotti and Shaw, 2005; Morgan, 2003; Morley and Guerin, 1996; Morley, 2003).

Two challenging and distinct groups of questions arise in the discussion of soft-sediment deformation processes recorded in ancient rocks. The first group relates to the mechanical state of the sediments at the time of deformation: where they compacted, cemented, or otherwise lithified, and how strong were they? The second group of questions relates to the larger scale driving forces that led to deformation. This group includes questions like 'was deformation entirely due to gravitational instability of slopes or was part of the differential stress for deformation supplied by tectonic movements at depth?'. The second group of questions is often summarized as a dichotomous choice between 'gravity-driven' and 'tectonic' deformation. However, this statement of the dichotomy makes a false distinction, because gravitational forces acting on slopes are responsible for a large part of the differential stress even in clearly 'tectonic' deformation processes. For example, in foreland fold and thrust belts, surface slope is a large component of the 'critical taper' required for the self-similar growth of a tectonically driven deformed zone (Dahlen et al., 1984; Davis et al., 1983). A clearer distinction can be made on kinematic grounds between deformation that is 'superficial' because it occurs above a basal detachment that is linked to the surface in both up-slope and down-slope directions, and deformation that is 'rooted' in a shear zone or fault zone at depth (Fig. 1).

Even with this clarification, confusion between the two groups of questions is still common. Most practising geologists have a tendency, once it is shown that deformation occurred in unlithified sediment, to assume that it occurs by superficial, down-slope gravity-driven processes. However, soft-sediment deformation is clearly occurring at many present-day plate boundaries, where the driving force and overall kinematics are driven by rooted, tectonic processes. Furthermore, under some circumstances it is possible for pockets of unlithified sediment to become mobilized during deformation in otherwise lithified sedimentary packages undergoing tectonic deformation (Phillips and Alsop, 2000). Conversely, it is possible for quite large slabs of lithified rock to move, and even deform internally, in a scenario where movement occurs entirely down-slope, displaying 'superficial' kinematics. Fig. 1 presents an idealized, two dimensional representation of the spectrum of deformational processes, with four end-member environments. End-member A (superficial, unlithified) represents superficial down-slope movement of unlithified sediment in slumps and gravity slides. End-member B (rooted, unlithified) includes deformation of unlithified sediments in trenches and strike-slip fault zones. C represents the down-slope movement of lithified material, which, while arguably less common than A or B, is still capable of moving bodies of rock covering many square kilometres (Schultz, 1986). D represents the vast majority of classic rock deformation environments studied by structural geologists working in orogens, in which lithified materials are deformed by rooted processes.

Because answers to the two types of questions are independent, different evidence must be collected in order to answer each. For the second question group – whether the deformation is superficial or rooted – it is unlikely that observations at microscopic or outcrop-scale can provide answers. This can be simply demonstrated with a 'thought experiment'. Sediments at the toe of a continental slope gravity slide (case A) and at the leading tip of a subduction complex with a small 'critical taper' (case B) experience practically identical stresses. The rocks being deformed 'cannot tell' which environment they are in. Therefore, a geologist examining the rocks after deformation is unlikely to be able to tell from evidence in the outcrop, which process occurred. This is born out by the close geometric similarities between fold-thrust belts found at the base of the continental slope on passive continental margins, and those found in compressional orogens (e.g. Bilotti and Shaw, 2005). Only by mapping the basal detachment, to determine whether it links into an up-slope zone of extension in case A, or into a down-dip zone of high pressure metamorphism, in case B, is it possible to distinguish the two cases. Even where an up-slope extension zone is present, careful section balancing calculations are necessary to establish whether shortening at the base of a submarine slope is entirely balanced by up-slope extension, or whether a 'rooted' component is present (Hesse et al., 2009).

However, the distinctions represented by the first question group are based on the mechanical state ('unlithified' vs 'lithified') of the material being deformed; like other deformation mechanisms, soft-sediment deformation can potentially leave microscopic or outcrop-scale evidence of the deformation process. The remainder of this paper addresses the problem of identifying this evidence.

4. Recognition of soft-sediment deformation

Because orogens typically involve former passive margins, major strike-slip faults, and subduction zones, it is extremely likely that they contain numerous examples of soft-sediment deformation structures. However, overprinting by later deformation complicates the understanding of these structures enormously, to the point where few unequivocal criteria are available for their recognition (Maltman, 1994a). In some cases (e.g. Elliott and Williams, 1988; Karlstrom et al., 1982; Pajari et al., 1979; Pickering, 1987; Williams, 1983) major differences of regional interpretation have resulted from contrasting interpretations of the same structures by sedimentologists and structural geologists.

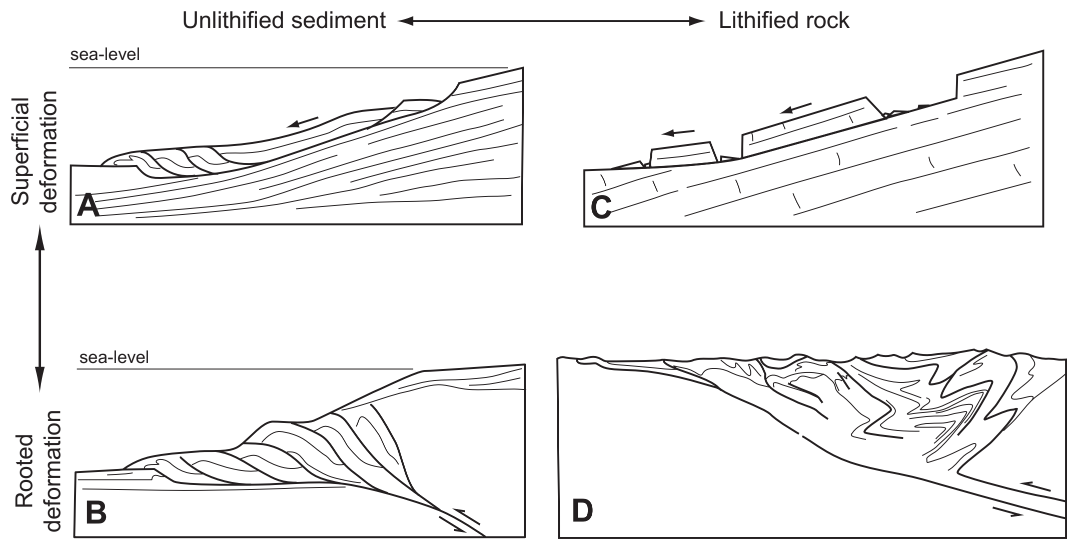

Fig. 1. Idealized, conceptual spectrum of deformation processes portrayed in two dimensions. Scale is arbitrary. A: Deformation of unlithified sediment by superficial processes (e.g. slumps, slides on passive continental margins). B: Rooted deformation of unlithified sediment (e.g. accretionary wedge deformation at subduction trenches). C: Superficial deformation of lithified rock (e.g. rockslides of exposed bedrock). D: Rooted deformation of lithified rocks (loosely based on cross-section of Escher et al., 1996).

4.1. Previous criteria

Many criteria have been suggested for the recognition of soft-sediment structures; in comprehensive reviews, Elliott and Williams (1988) and Maltman (1994a) examine many of these and conclude that few are reliable. For example, *welded contacts* have been proposed as a criterion for recognition by other workers (e.g. Horne, 1970; Pickering, 1987). Welded contacts are contact surfaces between deformed and undeformed rocks that show no indication of grain breakage, fabric development, veining, or other post-lithification deformation. Such contacts are commonly interpreted as depositional. In rocks without superimposed later fabric, the presence of welded contacts is a plausible criterion for exposure of the deformed zone at the sediment surface, and therefore for soft-sediment origin. However, when a penetrative tectonic fabric is superimposed, it is likely that an earlier fault contact could be effectively welded, and the criterion is unlikely to be robust.

One criterion that does seem to provide a clear indication of soft-sediment origin is the overprinting of deformation structures by sedimentary or organic structures that unequivocally show grain-by-grain sediment mobilization. Thus Alexander (1987) was able to show that plant roots had penetrated through slumped sediments in Jurassic sandstones from Yorkshire, UK. Trace fossils, sediment-filled dykes, and dewatering pipes are other examples of structures that may cross-cut deformational structures such as folds and faults. In most cases this is taken to demonstrate the pre-lithification origin. Nonetheless, Phillips and Alsop (2000) described sandstone-filled dykes that cross-cut F1 folds and S1 cleavage in the Dalradian Supergroup of SE Scotland and NW Ireland, and argued that regions within the Supergroup remained unlithified during part of the tectonic deformation history, supplying sand which was injected into other parts of the succession. However, these features are rare, and are thus of limited help in the vast majority of disputed cases.

4.2. Liquefied sand–mud geometries

Liquefaction is an important process in the generation of many structures in sedimentary rocks, including some that are typically regarded as products of the deposition process, and others that are clearly post-depositional and are typically discussed as products of deformation, rather than sedimentation. Pickering (1987) mentions liquefaction structure in fold hinges as a feature of soft-sediment folds, though without specifying how this is to be identified. Maltman (1994b) stresses that the division between sedimentation and deformation processes is somewhat arbitrary, but that liquefaction is often necessary for the mobilization of previously deposited sediment, whether in resedimentation or in deformation processes. In this section we examine a spectrum of structures involving liquefaction that spans the transition between deposition and deformation.

In the depositional category are conglomerates in which the clasts were clearly derived by erosion of recently deposited material within the general area of deposition (intraclasts). Some of the most spectacular examples of intraclast conglomerates are found in channelized deposits of submarine fans, where densely packed mudstone clasts are surrounded by sandstone. At first sight, some such deposits present a bewildering appearance, as the sandstones (which, after lithification, are more weathering-resistant) tend to stand out on weathered surfaces. In the example shown in Fig. 2 (Waldron and Jensen, 1985), from the Meguma Supergroup of Nova Scotia, regions of sandstone typically show concave re-entrants in their surfaces. Previous interpretations of outcrops like the one in Fig. 2 focused on the geometries of sandstone bodies, and suggested that they were remnants of beds deformed by 'slumping' processes (e.g. Smith, 1980). Upon closer observation, however, it becomes clear that the concave re-entrants in the sandstone surfaces are formed where sandstone accommodates the shape of angular mudstone clasts. The entire outcrop is better interpreted as the product of a single sediment gravity flow containing domains of sand and mud. Clearly, at the time of juxtaposition of sand and mud, the mudstone was more rigid, and the convex-outward, angular outlines of mudstone bodies indicate that they were formed at least in part by brittle deformation, whereas the sand must have behaved either as a fluid or as a plastic material with a yield stress much lower than that of the mud. Because recently deposited static sand is typically more deformation-resistant than contemporary mud (e.g. Kenney, 1984; Amundsen et al., 1985; Bell, 2000) it is reasonable to conclude that the sand was liquefied by tractions acting in the sediment gravity flow, resulting in a reversal of the 'normal' relative strengths of the two sediment types. This conclusion is apparent despite the presence of an overprinting

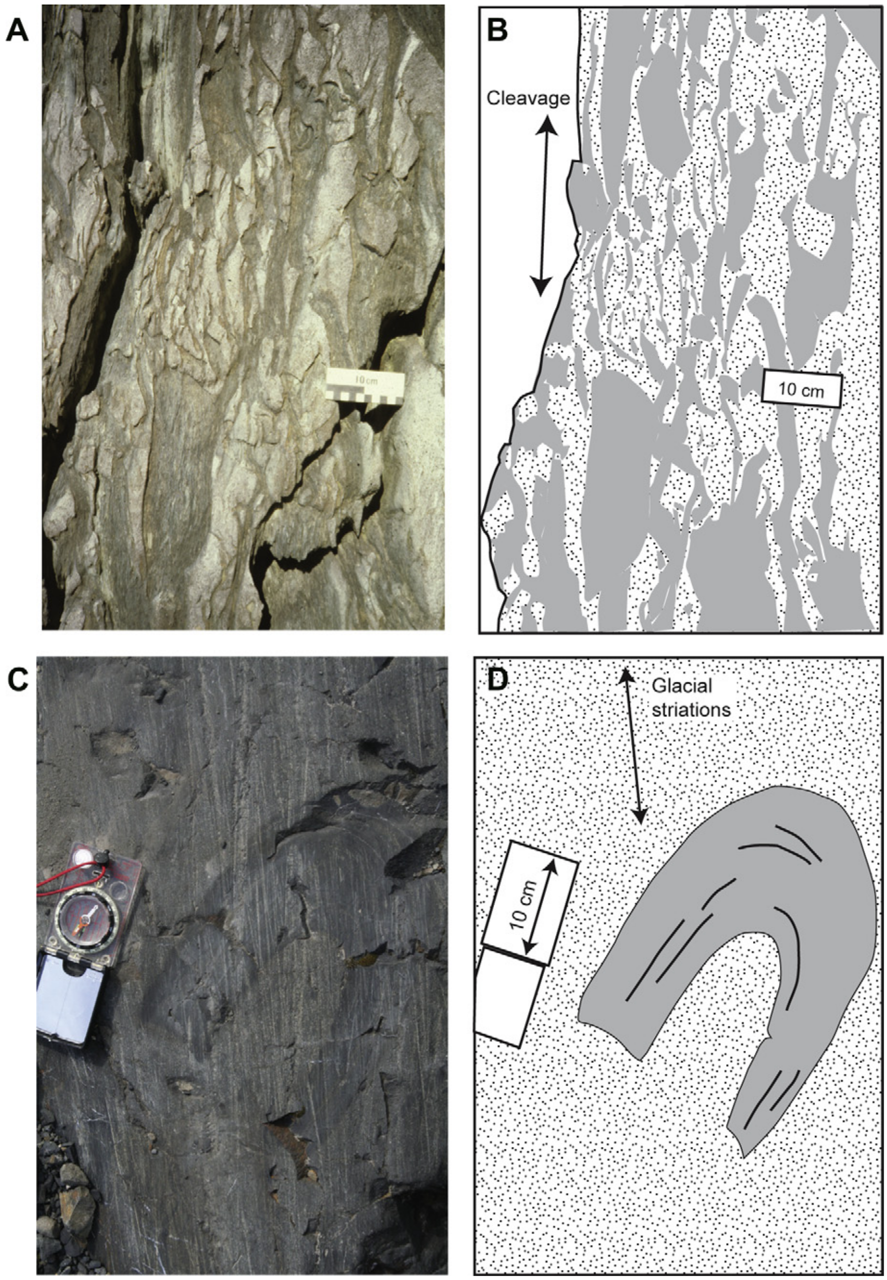

Fig. 2. Structures in intraclast conglomerates. A: Goldenville Group, Nova Scotia, Canada (after Waldron and Jensen, 1985). B: Interpretation of A showing intraclasts (grey) and concave re-entrants in margins of surrounding sandstone domains (stipple). C: Bowser Lake Group, Mount Dilworth, British Columbia. Intraclast shows tighter curvature on inner arc of fold than outer (class 1C fold) showing that it was stronger than the surrounding sand matrix at the time of folding. D: Interpretation of C, symbols same as B.

cleavage of purely tectonic origin, which has undoubtedly modified the shapes of the muddy domains.

Fig. 2b shows an isolated intraclast from a comparable bed of sandstone in the Jurassic Bowser Lake Group at Mount Dilworth, British Columbia (Elsewhere in this bed, intraclasts are packed in a configuration similar to Fig. 2a). The intraclast is clearly from a mud layer that was buckled during its incorporation into the sandstone bed. Note that the ends of the folded layer are angular, showing that the mud was sufficiently consolidated to show little mixing with the surrounding flow of sand. The layer is slightly thickened at the fold hinge, but the outer surface shows a greater radius of curvature than the inner surface, making this a class 1 layer in the dip-isogon classification of Ramsay (1967) (Fig. 3A and B). Class 1 layers are characteristic of the stronger units in folded multilayers (e.g. Ramsay and Huber, 1983), confirming the obvious conclusion from the setting of the clast that the mud had greater strength than the surrounding sand, which was probably a liquefied or turbulent flow.

Fig. 4, from the Mississippian Horton Group of Nova Scotia, shows another common geometry of mudstone and sandstone clearly produced by liquefaction of sandstone. A sand sheet oriented almost perpendicular to bedding connects with a lenticular sand body broadly parallel to bedding, that was clearly exposed at the sediment surface, because it contains laminations and other sedimentary structures produced by current traction of sediment. Though originally interpreted by Hesse and Reading (1978) as sand volcanoes, produced by upward expulsion of sand, these structures were shown by Martel and Gibling (1993) to be products of

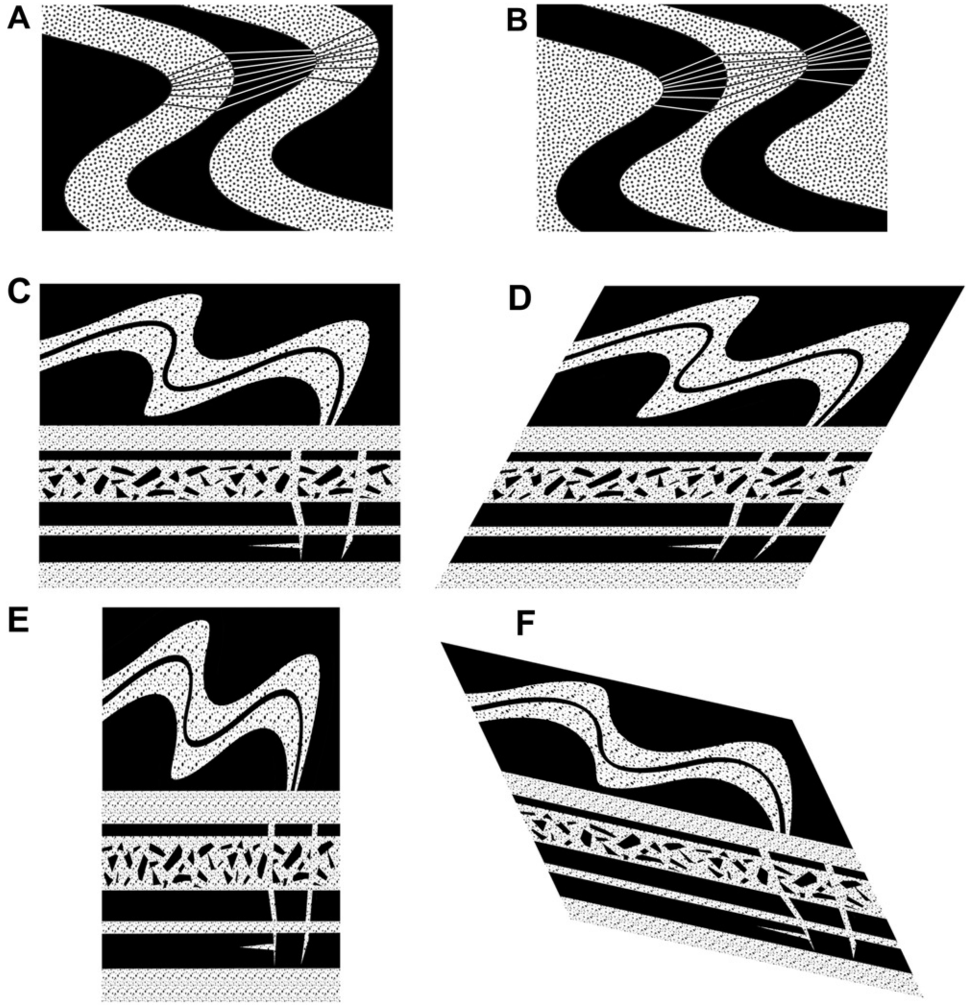

Fig. 3. Idealized diagrams of structures in deformed sedimentary rocks. Sand: stippled. Mud: black. A: Typical geometry of folds in clastic sedimentary strata deformed at low metamorphic grade, showing dip isogons (lines joining points of equal dip on successive surfaces). Sandstone layers typically show tighter curvature on inner arcs (class 1 geometry; Ramsay, 1967) while mud layers have tighter curvature on outer arcs (class 3). B: Reversal of normal geometrical relationships characteristic of folds formed while sand was liquefied. C: Undeformed configuration with: angular mud clasts surrounded by sand; sand-filled dykes cross-cutting mud layers; and folded liquefied sand layers. D: Diagram C with superimposed simple shear parallel to bedding. E: Superimposed pure shear parallel to bedding. F: Superimposed arbitrary strain.

Fig. 4. Sandstone-filled dyke below hummocky cross-stratified sandstone bed; Horton Bluff Formation, Nova Scotia, Canada.

downward injection of sand into fractures formed in recently deposited mud, probably as a result of overpressuring by wave loading. Martel and Gibling (1993) showed from the distribution of clasts that the sand behaved as a fluid while being injected, whereas the surrounding mud clearly behaved as brittle or semi-brittle solids, indicated by the sharp boundaries and matching sides of the fractures. Synsedimentary dykes are common in a variety of depositional environments. A spectacular example from the Alpine orogen is described by Parize and Fries (2003). The characteristic, angular geometries and cross-cutting relationships of synsedimentary dykes indicate that the dyke-filling material had much lower strength than the host sediment at the time of injection. Because of this, there is usually little difficulty in identifying them as features of sediment mobilization, even when overprinted by tectonic structures.

Fig. 5 shows structures at the base of a thick (300 m), approximately stratiform chaotic unit in Jurassic rocks of the Bowser Lake Group in the Canadian Cordillera (for a full description of this area, see Gagnon and Waldron, in press). Fig. 5A shows relationships in the contact zone between the chaotic unit and the underlying interbedded sandstone and mudstone. The lowest bed with clear deformational structures displays open folds in its upper surface

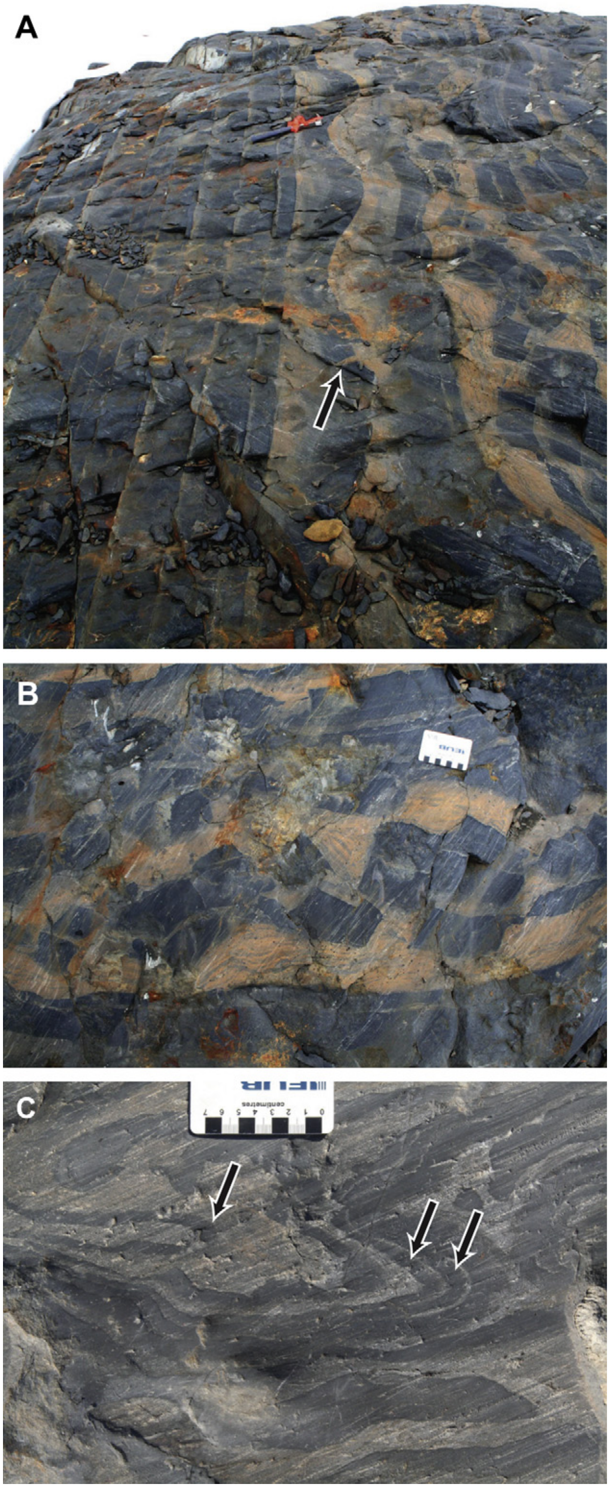

Fig. 5. Contact relationships at base of thick (>300 m) disorganized debris flow unit, Bowser Lake Group, Mount Dilworth, British Columbia, Canada (Gagnon and Waldron, submitted for publication). Dark rocks are mudstone; paler units are sandstone and siltstone. A: View of basal contact (younging direction to right); hammer (37 cm) lies on basal coarse bed of disturbed zone. Overlying folded mudstone unit contains sand-filled fissures (arrow). B: Higher mudstone unit in A is separated into angular blocks surrounded by sandstone. C: Folded sandstone and mudstone ~3 m higher in section. Mudstone layers marked by arrows show class 1 geometry (inner arc has sharper curvature than outer) indicating higher strength than surrounding sand.

with a wavelength of ~1 m. The base of the pale graded bed (hammer in Fig. 5A) is approximately planar, with the result that the bed varies in thickness from ~3 cm at the synclines to ~10 cm at the anticlines. The overlying mudstone bed shows roughly constant thickness (parallel fold – class 1B geometry) but deformation of the mudstone bed over the folds is accommodated by extensional fractures that are filled with sandstone from a bed above, presumably injected during deformation. Higher in the section (Fig. 5B), mudstone beds become divided into angular blocks surrounded by sand that has clearly flowed from adjacent beds, while sandstone beds show increasingly folded geometries. Continuity of bedding is progressively lost and folds become more abundant and tighter. Fig. 5C shows a fold in sandstone approximately 3 m above the basal contact. Note that despite the substantially higher strain, mudstone layers still show class 1 geometry whereas some intervening sandstone layers show class 3 geometry, indicating that the sandstone was weaker than the mudstone during folding. Comparable folds were described from the Carboniferous of New Brunswick by Wilson (2006), where they were also interpreted as products of soft-sediment deformation.

A comparable style of folding is shown in Fig. 6A, from deltaic units in lacustrine sediments of the Stellarton Formation, in the coal-bearing Pennsylvanian of Nova Scotia (Waldron, 2004). Within a folded and thickened sandstone unit, a very thin (~1 cm) layer of mudstone is folded in a series of near-parallel (class 1B) buckle folds surrounded by sandstone. The sandstone layers above and below the mudstone show extreme thickening into the hinges of the folds, indicating that the sandstone was much less competent than the mudstone due to liquefaction during deformation. In this example, there is independent evidence, mappable at the scale of a large, quarry-face outcrop, for up-slope extension approximately contemporary with folding (Waldron, 2004). The deformation is therefore inferred to be both soft-sediment and superficial.

5. Discussion – recognition of soft-sediment structures

In all the above examples, clear features of the geometry of sandstone and mudstone bodies indicate that sand was weaker at the time of formation of the structures than mud. However, studies of the mechanical properties of clastic sediments and sedimentary rocks show that under most circumstances sands are stronger than penecontemporaneously deposited shales. Textbooks (e.g. Ramsay and Huber, 1983) contain many examples of 'tectonic' folds in which sandstone beds display near-parallel geometry (class 1 in the classification of Ramsay, 1967) and intervening mudrocks show class 3 geometry and thickened fold hinges (e.g. Figs. 3A and 6B). Even in unlithified sediments, where geotechnical engineering measurements provide an indication of the relative strengths of different lithologies, sands are generally stronger than interbedded muds in laboratory tests (Bell, 2000; Kenney, 1984) and in the field (Amundsen et al., 1985). During diagenesis, grain interlocking, cementation, and pressure solution typically increase the strength of sands (Barton et al., 1986; Palmer and Barton, 1987) in the transition to sandstones. Thus it is clear that sands typically remain stronger than muds throughout normal conditions of burial and diagenesis.

However, liquefaction can reverse this pattern, by reducing the effective strength of a granular sediment to near zero (e.g. Taylor, 1984). Liquefaction can arise through a variety of processes, including wave loading, down-slope movement, earthquake shaking, and overpressuring during burial. Overpressuring of sands during burial may result from the inability of water to escape through overlying muds, which because of their low permeability form an effective seal. Hence uncemented sands may lose their strength through liquefaction, reversing the normal relative

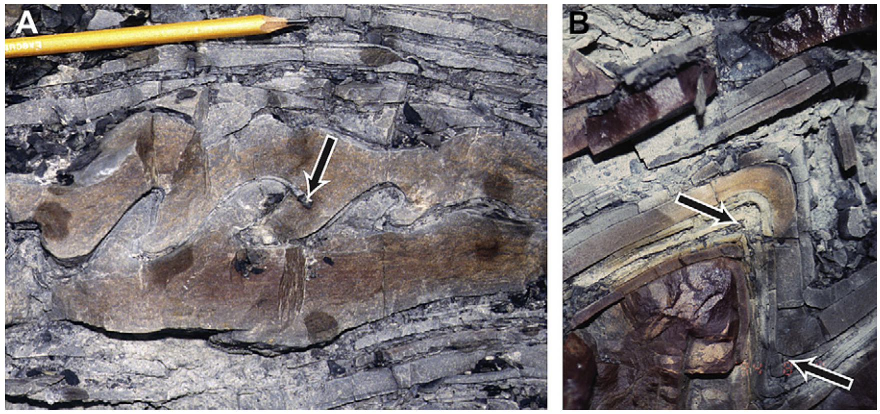

Fig. 6. A: Folds in very thin mudstone bed between two thicker sandstone beds interpreted to have been liquefied at the time of deformation (Waldron, 2004). Arrow marks break in mudstone layer. B: Fold more typical of tectonic folding styles from the same outcrop, showing sandstone layers (resistant) with class 1 geometry and mudstone layers with thickened fold hinges (arrows). Scale same as A.

strength relationship. This strength reversal, when it occurs, therefore provides an effective criterion for the recognition of soft-sediment folds at low metamorphic grade. At high metamorphic grades, it is likely that some quartzofeldspathic lithologies, having relatively low melting points, weaken more rapidly with temperature than pelites, so that folds formed at high temperature may show similar strength reversal to soft-sediment folds.

Three distinctive geometries occur in the above examples. They are: (1) angular, convex fragments of mudstone surrounded by sandstone bodies with corresponding concave angular re-entrants; (2) parallel-sided fractures through mudstone layers, filled by sandstone; and (3) folds in which mudstone beds display near-parallel-sided geometry in profile view (class 1), while intervening sandstone layers show class 3 geometry and/or strongly thickened fold hinges (Fig. 3B) in profile view.

The occurrence of any one of the above features in sedimentary rocks at greenschist facies or below indicates that the corresponding deformation occurred while sandstone was unlithified. The absence of characteristic liquefaction structures, however, cannot be used as an indication of post-lithification deformation, because liquefaction does not occur in all circumstances of soft-sediment deformation. Even where liquefaction does occur, it may be limited in time (as trapped fluids escape) and in space (where different units along strike may “cap” or seal fluids), so the relative strength of sand and mud may vary continuously. When liquefaction is not occurring, sand is likely to be stronger than freshly deposited mud, and deformation styles will be similar to those in lithified sedimentary rocks.

The potential complexity of liquefaction history is illustrated by the example in Fig. 7, from the Jurassic Sofular Formation, in the Antalya Complex of SW Turkey (Waldron, 1984, 1985). A folded zone between apparently undeformed beds has contacts that appear ‘welded’ and tectonic fabric is very weak. However, the folds occur within a thrust belt, and therefore could conceivably have formed by tectonic processes. The most convincing evidence for soft-sediment folding is the geometry at hinge ‘B’, where mudstone beds show near-parallel geometries and intervening calcarenite beds show extreme hinge thickening. Note that in this case, most of the visible layers show class 1 geometry, but the thickening of calcarenite layers into the hinge is much more pronounced than that of the mudstone. The overall geometry shows a fold interference pattern, in which an early fold (hinges A and C) has been overprinted by fold B. Because fold B clearly involved unlithified sediment, folds A and C must also have formed before lithification. However, folds A and C show more ‘normal’ layer geometries: calcarenite displays class 1 folds, and mudstone is thickened into some of the hinges in class 3 folds. The most reasonable hypothesis for the formation of these folds is that the A-C fold-pair formed as a result of down-slope movement in sediment that was ‘drained’ — its pore fluid was free to escape to the sea-floor. During continuing deformation the folded calcarenite became isolated from free pore-

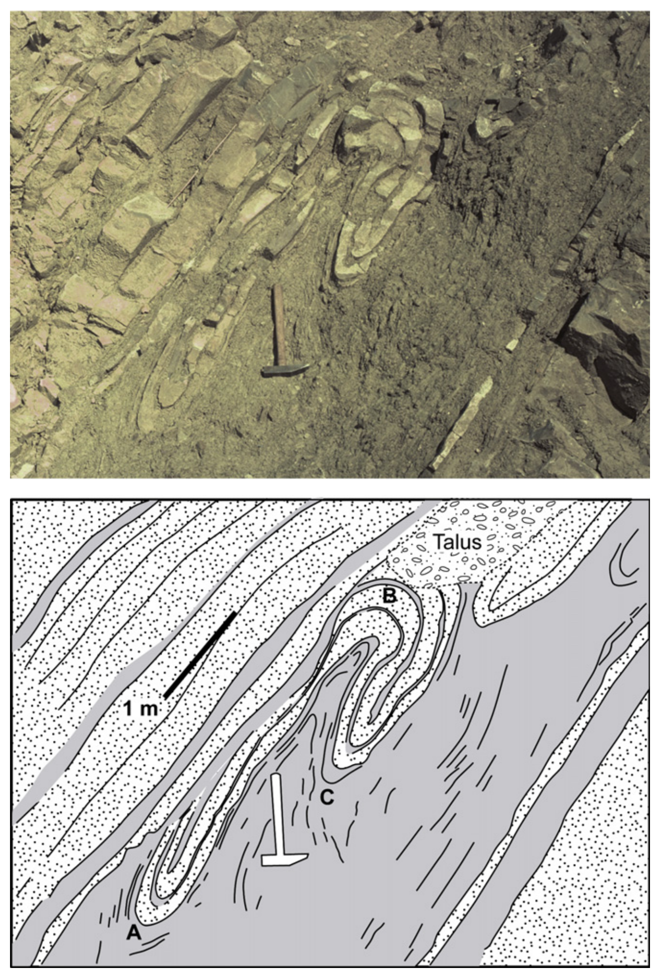

Fig. 7. Folded calcarenite and mudstone, Sofular Formation, Antalya Complex, SW Turkey (Waldron, 1984, 1985). Line drawing shows hinges A–C discussed in text.

fluid communication with the sea-floor, with the result that it became temporarily overpressured, allowing liquefaction to take place. Fold B developed during this transient phase of liquefaction.

Ramsay and Huber (1983) explore the relationship of fold style to layer thickness, viscosity contrast, and the relative proportions of strong and weak layers, in viscous materials. For isolated strong layers, they show that buckle folds have an initial wavelength proportional to the layer thickness and also dependent on the cube root of the viscosity contrast. In the case of multilayers, relationships are more complex, but scale with layer thickness at least for folds with wavelengths up to about 100 m (where gravity may become a significant factor). The relative thicknesses of strong and weak layers also control fold style, but in all the cases considered by Ramsay and Huber (1987) the stronger layers show class 1 shape, while the intervening less viscous layers fall into class 3. We anticipate therefore that characteristic liquefaction geometries should be observable over the whole range of scales and layer thicknesses observable in outcrop. The examples shown here include cases where thin mud layers are encased in thick sand (Figs. 2C, D and 6A) and where thinner sands are surrounded by muds (Fig. 5C).

Many claimed examples of soft-sediment deformation structures in orogens occur as pre-cleavage structures in areas where later deformation has produced an overprinting cleavage. How does overprinting by later deformation affect the three geometric criteria proposed? For a homogeneous overprinting deformation, each case can be considered in turn. Fig. 3C–F, shows various homogenous strains superimposed on an angular mosaic of blocks comparable to Fig. 2A, a parallel-sided sandstone dyke comparable to Fig. 3, and folded sand and mud layers comparable to Fig. 6A. Clearly, any such superimposed strain will change the angles between surfaces but the overall angularity and convex-outward shapes of the blocks will be preserved. Because homogeneous strains preserve parallelism, the distinctive features of the geometry of the dyke are also still recognizable after any superimposed homogeneous strain. A similar argument applies to the dip isogons that form the basis of fold layer classification (Fig. 3A and B). Lines that are dip isogons remain so after homogeneous strain, because they join points where layer surfaces are parallel. As demonstrated by Hobbs et al. (1976) in a discussion of the origin of similar (class 2) folds, layer geometries may be modified by superimposed homogeneous strain, and both class 1 and class 3 layers may approach similar (class 2) geometry, but the overall pattern of convergence and divergence of isogons survives. Thus distinction between more and less competent layers is preserved, even though careful measurements may be needed to detect it.

Outcrop surfaces that are not perpendicular to fold hinges provide distorted views of fold profiles, and are geometrically equivalent to the strained views shown in Fig. 3D–F. It is thus theoretically possible to make the same distinctions in cases, for example, where glacially smoothed planar outcrop surfaces prevent direct observation of fold profile planes. However, folds formed in soft-sediment are commonly highly non-cylindrical (e.g. Strachan and Alsop, 2006). Views oblique to the local profile plane should therefore be interpreted with caution.

Strength relationships characteristic of liquefied sediment, if present, can therefore survive a superimposed homogeneous strain (Fig. 3D–F), and may be visible even on oblique outcrop surfaces. Heterogeneous strain is of course less predictable, and it is likely, in cases where heterogeneous strain is controlled by contrasting mechanical properties, that superimposed post-lithification deformation may obliterate or reverse the contrasts produced by pre-lithification deformation associated with liquefaction. For this reason, it is emphasized that the absence of strength reversals related to liquefaction does not necessarily imply that deformation occurred in a lithified state. However, the presence of geometries showing sand weaker than mud, in sedimentary rocks at low or medium metamorphic grade, appears to be a reliable indication of soft-sediment deformation.

6. Conclusions

The recognition of soft-sediment deformation structures, especially folds, in the tectonically deformed sedimentary rocks of orogens presents special challenges. It is important to distinguish inferences about the lithification state of sediments at the time of deformation from those that relate to the overall geometry, kinematics, and driving mechanism, as unlithified sediments may be deformed by tectonic processes in 'rooted' structures. Some inferences about physical state may be made on the basis of geometrical criteria. In particular, liquefaction of sand can lead to a situation where sand is mechanically weaker than mud. The resulting structures, including angular convex mudstone domains surrounded by concave sandstone domains, and class 1 folds in mudstone layers interbedded with in sandstone layers displaying class 3 folds, are strong indicators of deformation of unlithified sediment.

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