How sediments become mobilized

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Abstract: Geological sediments tend to strengthen during progressive burial but the interplay of porosity and permeability, strain and effective stress gives rise to numerous circumstances in which the strength increase can be temporarily reversed. The sediment becomes capable of bulk movement – sediment mobilization. Most explanations involve overpressuring, which results from additional loading being sustained by pore-fluid that is unable to dissipate adequately, leading to frictional strength reduction. The processes are highly heterogeneous, areally and with depth. The loads can be external ('dynamic') and both monotonic (e.g. a rapidly added suprajacent mass) and cyclic (e.g. the passage of waves), internal (e.g. the result of mineral reactions) and hydraulic (e.g. injection of external fluid). The sediments may become liquidized – that is, lose strength completely and behave as a fluid – through temporary fabric collapse (sensitive sediments) because loads are borne entirely by the pore-fluid (liquefaction), or by the grains becoming buoyant (fluidization), typically due to the ingress of externally derived fluids. In response to hydraulic gradients, buoyancy forces and reversed viscosity or density gradients, the weakened sediment may undergo bulk movement, though this requires failure of the enclosing material and sustained gradients. Mobilized but non-liquidized sediments retain some residual strength but can attain large shear displacements under critical state conditions.

Sediments undergoing burial tend to progressively increase in strength until they become rock. However, this strength increase can be temporarily reversed, with the sediment becoming so weak that it is capable of mobilization in numerous circumstances. In geological usage, the term mobilization (sometimes called remobilization) involves both rendering the sediment capable of motion and the bulk movement that commonly results. This paper is chiefly concerned with the former. Actually, most sediment mobilization probably takes place at the Earth's surface – due to water currents, mass movements, etc. – but we are here concerned with subsurface mobilization. For incompletely lithified sediments, this primarily depends on some form of what can generally be termed liquidization. Following Allen (1977, 1984) a liquidized sediment is one that behaves mechanically like a fluid (irrespective of the reason), even though it previously possessed a yield strength and hence had behaved as a solid (see also Owen 1987).

The present article reviews liquidization and other mechanisms by which incompletely lithified sediments – those that move chiefly by grain slippage – become capable of the bulk movements associated with subsurface mobilization (see Fig. 1). Although some of the principles outlined here can be extended to weak rocks, that is, where the mobilization involves the breaking of grains and inter-grain bonds and to intrinsically weak and ductile materials such as salt, here the primary concern is with sediments. The processes of movement, ranging from *in situ* mixing through the intrusion of clastic bodies to extrusion at the surface of the sediment sequence and the whole range of resulting structures, are not dealt with here.

Burial and related processes

Geological sediments are essentially mixtures of relatively strong grains with an intervening fluid, usually a brine. Of course, natural sediment systems are normally much more complex, with the fluid containing gases of changing solubilities and possibly immiscible hydrocarbon phases and the particles themselves changing in volume and shape, particularly where clay aggregates are involved. Burial tends to progressively strengthen the mixture (lithification), by displacing the fluid and packing the particles closer together (consolidation), hence promoting inter-grain contact (frictional strength) and chemical inter-reactions (diagenesis). The various processes of diagenesis – principally cementation, recrystallization and diffusion mass transfer, commonly called pressure solution where it is assisted by fluids – take place progressively. The ability of intact grains to slide past each other to allow bulk movement of the sediment is progressively curbed and supplanted by mechanisms that deform the grains, that is independent particulate flow is superseded by cataclastic flow, as the sediment is turned into rock (e.g. see Maltman 1994). Brown & Orange (1993) argued that this transition is likely to occur (in fine sands) at effective confining pressures between about 1 and 5 MPa, signifying,

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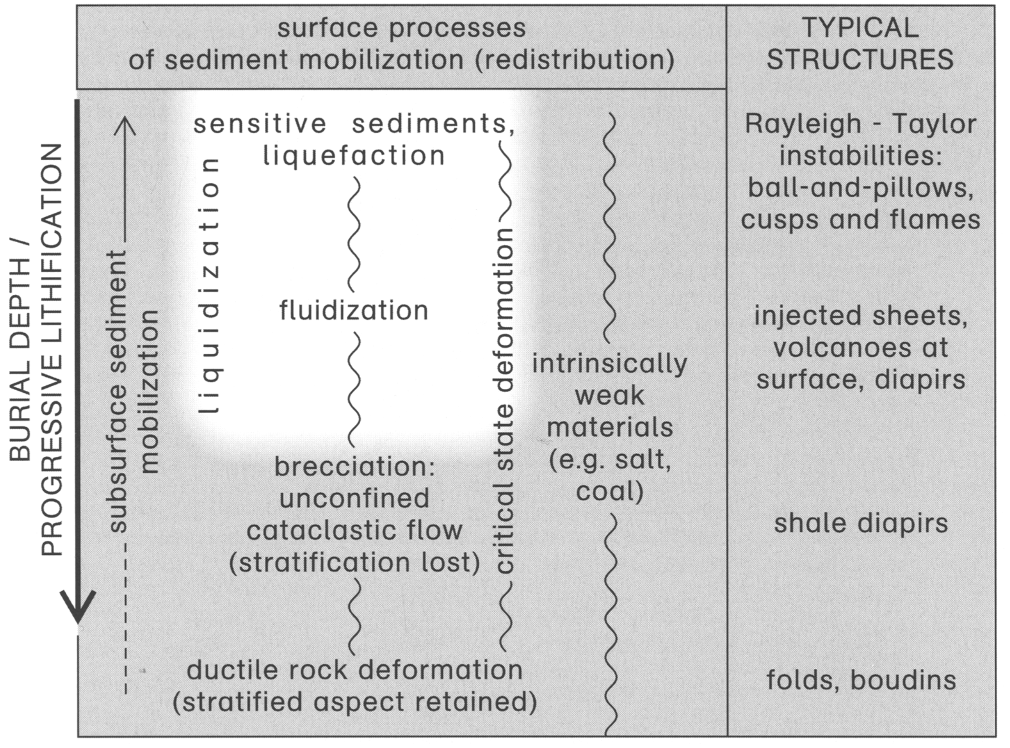

Fig. 1. Conceptual view of the range of processes involved in sediment mobilization. The untinted area indicates the scope of the present paper.

because elevated pore-pressures are typically involved, burial depths of several kilometres. We now begin looking at situations in which this progressive evolution is temporarily halted.

Consolidation

Consolidation is the chief mechanical aspect of burial. It is defined as time-dependent mechanical reduction in sediment volume, usually pore volume, in response to increased loading. In practice, consolidation normally takes place due to the weight of progressive additions of overlying sediment, that is, the increasing lithostatic load. In ideal circumstances the pore-fluid in the buried sediment only sustains the load imparted by the suprajacent fluid (pore-fluid plus any suprajacent water), but if the sediment is unable to drain adequately in response, the lithostatic load becomes proportionally added to the pore-fluid pressure, reducing the frictional contact between the grains. Where the entire load is borne by the pore-fluid, the sediment loses all its frictional strength and, assuming grain cohesion is negligible, therefore becomes liquidized. It becomes available for mobilization and remains in this state until the pore-pressure begins to re-equilibrate. The role of consolidation in mobilization is therefore intimately involved with the ability of the sediment to dissipate its pore water (e.g. Bitzer 1996).

The loss of porosity in order to maintain hydrostatic equilibrium is termed normal consolidation (e.g. path A–D on Fig. 2). Where the burial load is taken up by the pore-fluid, the sediment remains with fixed porosity despite its increasing burial depth. Such a sediment is overpressured and underconsolidated because it shows a greater porosity than expected. The situation has long been known in sedimentary basins such as the Gulf of Mexico (e.g. Stump & Flemings 2000) and is commonly recognised from positive porosity deviations from a normal depth profile.

Sediments have little elasticity, with the result that reductions in effective stress acting on a partially consolidated sediment do little to restore porosity and unloading paths tend to have little gradient (e.g. path D–C on Fig. 2). Decreases in effective stress (total stress minus pore-fluid pressure) can come about either through reduction in the burial load, say by erosion of some overlying material, or by increase in pore-fluid pressure, perhaps due to injection of pressured fluid from outside. The sediment enters the

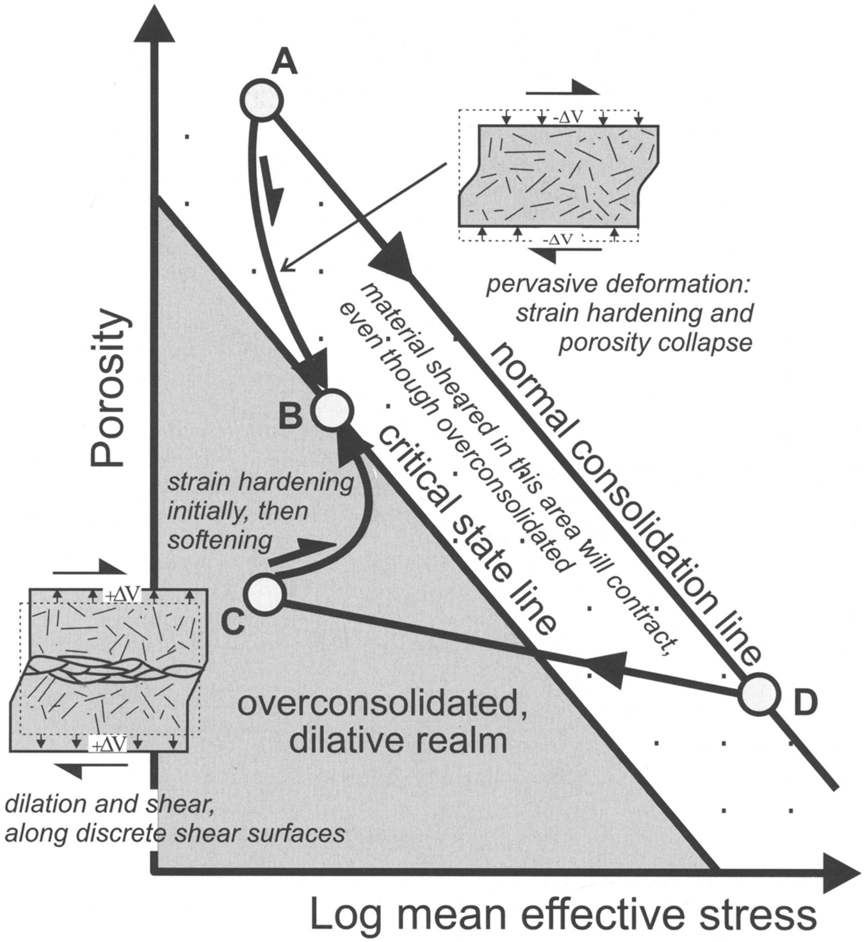

Fig. 2. Basic principles of sediment burial and deformation, relating porosity to changes in effective stress. Where drainage is sufficient to keep pace with increasing rises in total burial load, the sediment is normally consolidated, following a line such as A–D. If drainage is prohibited, the burial load will be taken up by the pore-fluid and the sediment will remain at point A, with fixed porosity despite its increasing burial depth. Line D–C represents a sediment-unloading path, with little porosity regain. Sediment at point C is overconsolidated, having experienced higher effective stresses. Increases in deviatoric stress (not shown on this two-dimensional diagram) will prompt deformation along a path dependent on the consolidation state. Overconsolidated material (e.g. at point C) will increase in volume during deformation and tend to shear along discrete surfaces until it achieves critical state (e.g. path C–B), whereas normally consolidated sediment loses volume and tends to deform pervasively (e.g. path A–B). However, all materials will reach the critical state, where they retain strength but are capable of large shear displacements.

realm of overconsolidation, occupied by materials that once experienced greater effective stresses. Note that late overpressuring, such as might arise from the injection of externally sourced fluids, gives a different porosity value from that due to drainage being curbed from the start. The implication of this is that the absence in a depth profile of marked porosity anomalies does not signify the absence of overpressured horizons. Wherever overpressuring arises during burial, the sediment is weakened and the possibility arises of it undergoing mobilization.

Critical state deformation

The overpressuring does not have to be sufficient for complete liquidization for the sediment to be involved in mobilization. Much sediment is likely to retain some residual strength during its movement, through grain cohesion or incomplete sealing of fluids and in any case liquidized sediment may dissipate pore-fluid during displacement and regain some frictional strength. In these situations, deformation at the sediment's critical state is important, as at that

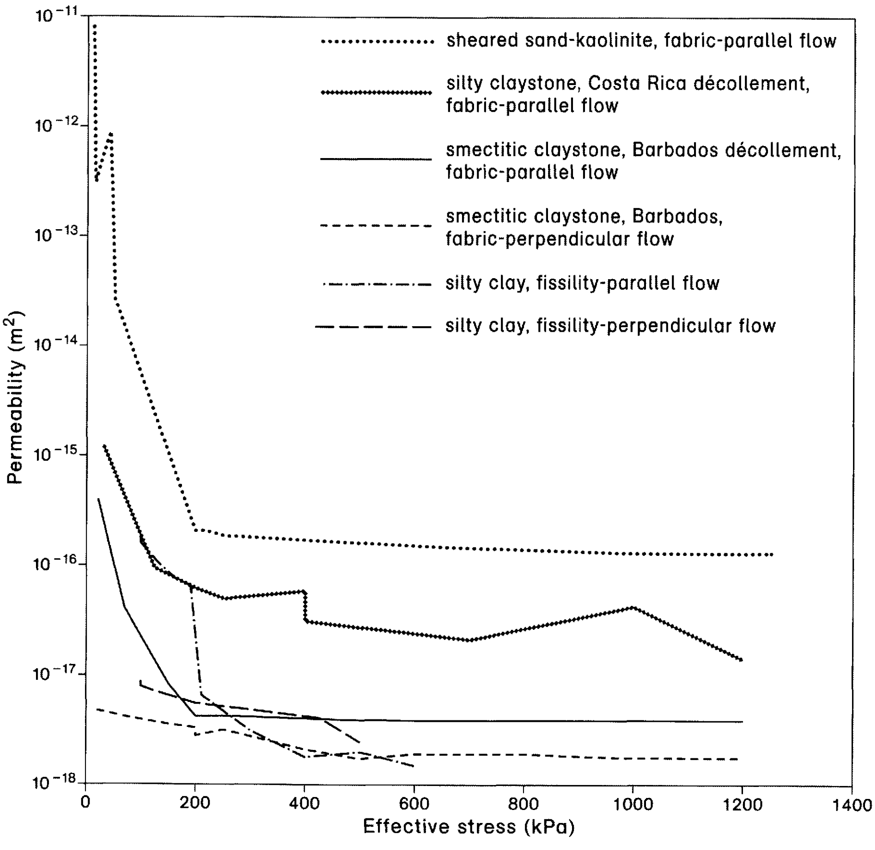

Fig. 3. Variation of permeability with mean effective stress for various anisotropic sediments. The permeability varies little with effective stress except below about 200 kPa, where the abrupt increases in permeability seen in fabric-parallel flow result from efficient dilation and connectivity of pores along the fabric to give a 'fracture permeability' (see Bolton *et al.* 2000). Apart from the sheared sand-kaolinite, which was generated in the laboratory, all the sediments were taken from Ocean Drilling Program cores, the fissile silty clay being from the Woodlark Basin.

particular combination of porosity, fluid-pressure and stress the sediment can undergo very large amounts of shear (e.g. Jones 1994), even at low deforming stresses. The sediment behaves overall in a weak, ductile manner rather than as a fluid; its actual response will depend on factors such as consolidation state (e.g. Yassir 1990).

Most overconsolidated sediments have to dilate at first, and deform along discrete shear surfaces (e.g. path C–B on Fig. 2), of which the scaly fabrics commonly described from clay diapirs and (non-cataclastic) deformation bands in sand dykes are manifestations. Sediments in other consolidation states undergo porosity collapse and a more pervasive style of deformation (e.g. path A–B on Fig. 2). Irrespective of the starting conditions, however, the deformation drives the sediment towards the critical state condition, represented as a line in Figure 2. The concept of a critical state only strictly applies to homogeneous, perfectly plastic materials and in practice, of course, some fluctuation in the parameters or geometrical constraint will limit the strain achieved. However, because many sediments reach a condition of weak residual strength more easily at high porosities and low effective pressures, a situation in which relatively low deviatoric stresses drive large shear displacements, the idea is very relevant to much sediment mobilization.

Permeability

The extent to which pore-fluid is dissipated in response to applied loads depends on the permeability of the sediment. Permeability is the capacity of a material to transmit fluid and is a crucial influence on the extent to which overpressures develop in a sediment. In general, high permeability sediments tend to become mobilized by overpressuring due to externally-pressured fluids entering and rapidly passing through the system, whereas lower permeability materials are more likely to be weakened by overpressures in trapped fluids. In order for a sediment to undergo critical state deformation, the permeability has to be sufficient to allow the appropriate porosity changes and variations in effective stress conditions to be reached.

Permeability is likely to differ from bed to bed in a sedimentary sequence, but is not fixed for any particular lithology. Active deformation affects the permeability behaviour (e.g. Stephenson *et al.* 1994), as do any fabrics imparted to the sediment by consolidation or by earlier deformation, especially where the effective confining pressures are low (e.g. Brown & Moore 1993; Clennell 1997). Recent laboratory work (e.g. Bolton *et al.* 1999) has emphasized how variable permeability can be in anisotropic sediments at different effective pressures (Fig. 3). In other words, although sedimentary sequences are often assigned a single representative permeability value, in fact the drainage will vary intricately throughout – and through time, especially where burial or deformation fabrics are involved. From time to time at various levels in a sediment pile, there may well be materials with inadequate drainage, which therefore become overpressured and hence prone to mobilization.

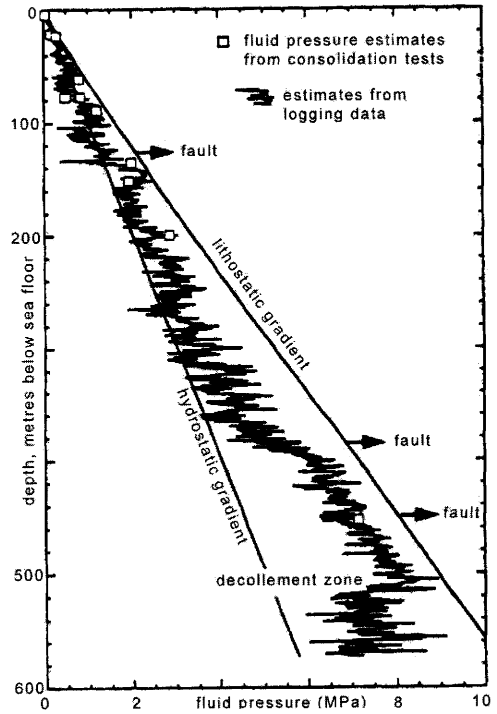

Fig. 4. Fluid pressure variations with depth in the Barbados accretionary prism, estimated from laboratory consolidation tests and well-log data. From Moore *et al.* (1995).

Complexity of relationships

For the reasons outlined above, simple relationships between burial loading, porosity and fluid pressures are unlikely in a consolidating sedimentary sequence. At their simplest, porosity gradients will be exponential and fluid pressures linear (e.g. Bahr *et al.* 2001), but even this is unlikely (Maltman 1994). A smooth gradient of fluid pressure is only possible where the sediment is able to drain continuously to maintain equilibrium conditions. If it is not, the lithostatic gradient will also be affected, as the sediment density will be reduced to less than normal. Moreover the rate of burial-loading due to sedimentation may well vary temporally, temperature effects become increasingly relevant with depth (e.g. Graham *et al.* 2001) and various other processes – discussed below under loading – may arise. Such complexities operate differently in different lithologies, so that in a layered sequence there may well be numerous pressure anomalies corresponding to different sedimentary horizons.

As an illustration, Figure 4 shows fluid pressure variations with depth offshore Barbados. The values are indirect estimates rather than actual measurements, but show well the intricacies that probably arise in nature, here in a fairly homogeneous sequence of mudstones, though with the complicating effects of marked variations in smectite content (Brown & Ransom 1996). In this active tectonic region, deformation is adding a further complication to the consolidation history of the sediment. Figure 5 also illustrates fluid pressure heterogeneity at Barbados, but in an areal perspective. Note the marked lateral variations in fluid pressure (interpreted from differences in seismic polarity), though perhaps the degree of heterogeneity shown by this active low-angle fault may be greater than that expected in a subsiding sedimentary layer. Evidence from rocks (e.g. Vannucchi & Maltman 2000) also reveals the complex fluctuations in fluid pressure that may affect sediments early in their burial history.

In summary, while the consolidation of a sediment tends to progressively strengthen it, in a sequence of bedded lithologies the fluid pressures are likely to be evolving through time in a complex way. There may well be numerous situations at different times, at different depths, in different places where a sediment is vulnerable to enhanced or reduced drainage and hence anomalously high fluid pressures. At such times, the sediment is reduced in strength, even to the extent that it behaves like a fluid

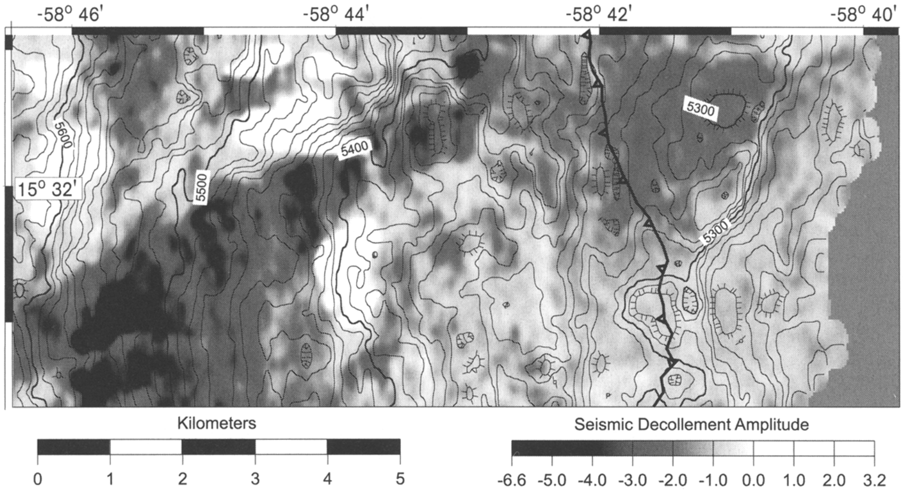

Fig. 5. Lateral fluid pressure variations on the basal décollement of the Barbados accretionary prism inferred from seismic amplitude. The dark areas indicate negative seismic polarity, thought to represent areas of high pore-fluid pressure. Based on Shipley *et al.* (1996).

and consequently becomes available for wholesale movement.

Anomalous sediment loading

As outlined above, weakening of a sediment usually comes about through the addition of some load with which pore-fluid dissipation cannot keep pace. Lithostatic loading naturally arises as sediments accumulate, but a host of additional loads are possible (Fig. 6). Most mobilization comes about through combinations and mutual interactions of several effects rather than a single loading process.

Dynamic loading

A mechanical load that arises from a source external to the sediment is called dynamic loading. Such a load tends to be transmitted rapidly through the sediment and, depending on the stiffness of the sediment framework, partitioned into the pore fluids. Being virtually incompressible, the fluids typically sustain most of the excess load. The most obvious example of dynamic loading is lithostatic, the suprajacent sedimentation summarised in the preceding section. However, this varies from the slow, steady deposition of fines through to the instantaneous emplacement of major allochthonous masses. The latter process can be tectonically driven rather than sedimentary and although the rates will be lower, emplacement of nappes and thrust sheets is thought to be important for triggering overpressures in active orogenic belts (Yassir, pers. comm.). Moreover, tectonically driven shear in conditions of reduced drainage can induce over-pressuring (Yassir 1990). Other examples of dynamic loading in addition to continuing sedimentation and the effects of tectonism are the introduction of igneous material above or within the sediment sequence (e.g. Elsworth & Day 1999) and for near-surface sediments, a whole host of surficial processes ranging from footsteps (Lewis & Titheridge 1978) to meteorite impact (Read 1988). Each of these processes can be regarded as an individual event – although some are clearly repetitive – and hence such discrete loading is described as monotonic. The loads can be very large and sediments may be instantaneously strained. Poorly permeable sediments may generate pore pressures sufficiently large to prompt hydrofracture, which will then allow rapid drainage, at least temporarily. Monotonic loads are therefore unlikely to generate over-pressuring for long periods.

Two external sources of additional loading that are relatively long-lived and therefore not strictly monotonic are artesian conditions (e.g. Massari *et al.* 2001) and the application of tectonic stresses. The first pressurizes the pore-fluid directly whereas the second, in addition to tending to deform the sediment, will be partitioned into the fluid according to the drainage conditions. McPherson & Garven (1999) argued that tectonic loading is the primary reason for overpressuring in the Sacramento Basin (although not in this case leading to mobilization) and G. Westbrook (pers. comm.) invoked tectonic loading to explain mud volcanism offshore Barbados. In both these examples, the loading is ulti-

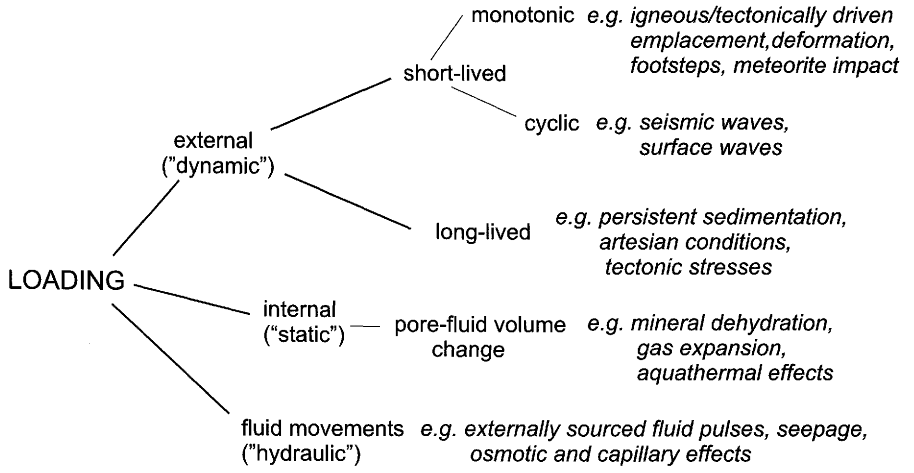

Fig. 6. Conceptual representation of the range of loading processes that may affect sediments in addition to progressive burial due to continued sedimentation. Note that the representation is neither comprehensive nor a rigorous classification.

mately the result of compressive stresses due to plate motion.

Repeated stressing of a sediment, arising from the oscillatory passage of waves, for example, is known as cyclic loading (e.g. Grozic *et al.* 2000; Pestana *et al.* 2000). Here also the duration of the effects is likely to be short, geologically speaking. The catastrophically weakening effects of seismic waves, especially shear waves (Youd *et al.* 2001), are well-known. Subaqueous sediments buried to a few hundred metres can experience pressure pulses from waves travelling at the sea surface (Seed & Rahman 1978), in association with winds and tides (Wang & Davis 1996). The stress rise from each of these pulses may well be much less than in a monotonic event, but if each fluid pressure response is incompletely dissipated before the next pulse, the incremental accumulation of overpressure can be large. Experimental work by Sassa & Sekiguchi (2001) demonstrated that progressive wave loading is more effective in liquefying sands than loading from standing waves. A variant on this behaviour, called cyclic mobility and arising where the periodic loads oscillate in nature between compression and tension, can induce strength loss even in relatively dense sediments (Castro 1987). Wang & Davis (1996) showed how the effects of tidal cyclic loading depended on the permeability of the sediment and the strength moduli/compressibility of the grain framework and pore fluid (see also Bachrach *et al.* 2001), the latter being very sensitive to the presence of gas (Wang *et al.* 1998). Lithostatic and even super-lithostatic loads are possible if the sediment lacks stiffness completely, at least in some layers within a sedimentary sequence (e.g. Zhao *et al.* 1998).

Static loading

Long-lived overpressures are generally more likely to be due to processes that operate within an evolving sediment – collectively referred to as static loading – rather than those imposed externally. Static loading tends to act directly on the pore-fluids, reducing the frictional strength of the sediment by an equivalent amount (e.g. Cobbold & Castro 1999). The sediment strength depends on any remaining frictional resistance and the amount of inter-grain bonding (cohesion) that diagenesis has imparted. The common processes of static loading are likely to become more important at greater depths, as lithification proceeds. The processes include aquathermal pressuring (Luo & Vasseur 1993) and hydrocarbon generation and maturation, including the expansion of rising gases such as methane (Osborne & Swarbrick 1997). Mineral dehydration, and especially the smectite-illite transformation, may also become important. The smectite content of clayey sediments can exceed 50%, so that large volumes of water, which can comprise up to 25% of the mineral, is made available when it dehydrates to illite. Although the transition is usually taken to occur when temperatures during burial reach around 60–80°C, Fitts & Brown (1999) have argued that stress can trigger the reaction, even where it is as low 1.3 MPa. Moreover, growing packets of illite progressively coalesce and restrict dewatering, further enhancing the overpressuring effect (Freed & Peacor 1989).

The role of gas-hydrates is currently unclear. Their prevalence is increasingly being recognized, together with the enormity of the gas and water they store and trap, which can be released if the stability field changes, say in response to sea-level change. Kennett & Fackler-Adams (2000) argued that dissociation as the hydrates respond to even small T-P changes would prompt widespread sediment deformation, even hundreds of metres below the sea floor. Such gas-hydrate driven overpressuring was invoked by Cherkis *et al.* (1999) to account for the strength reduction and consequent sediment instability offshore Spitzbergen, though such effects are disputed (e.g. Bouriak *et al.* 2000).

Hydraulic loading of sediments ranges from effects that are subtle but widespread, to more localised but dramatic processes associated with vigorous fluid movements. Any excess hydraulic head will drive pore-fluid to move and the resulting osmotic and capillary forces will tend to weaken the intergrain friction. Hydrocarbons in the pore-fluid will promote buoyancy of the grains. Such effects may be common, although Osborne & Swarbrick (1997) calculated that the effects would be small. Where the permeability allows rapid movement of the pore fluids, the drag force exerted on the sediment particles, termed the seepage force, can further add to the pore-fluid pressure and even exceed the weight of the grains. In this situation, the particles are buoyant and are readily entrained by the moving fluid, which may be a mixture of liquids and gases. This is the basis of fluidization, discussed in the following section.

The relative importance of the loading sources outlined above to sediment mobilization is still much debated and no doubt varies between different geological situations (e.g. Hall 1994). For example, Osborne & Swarbrick (1997) considered that stress-related mechanisms are the most likely cause of overpressuring in many sedimentary basins whereas theoretical calculations by Wangen (2001) suggested that they chiefly result, at least in more deeply buried sandy sediments, from cementation of the pore-spaces. Current work (e.g. Lonergan *et al.* 2000) suggests that fluids entering the sediment system from outside are increasingly being seen as a trigger for sediment mobilization, at least in hydrocarbon provinces.

Liquidization

Sensitive sediments

A sediment becomes mobilized because it is in a condition of insufficient strength to resist the forces driving it to move. Usually the weakening is temporary and is related to fluid pressures, as outlined above, but there are other processes (e.g. see Owen 1987). Some sediments are intrinsically vulnerable to abrupt, albeit slight, loading and become instantaneously destabilised: they are thixotropic. For example, sediments with unusually high pore-water contents, perhaps through having a honeycomb or 'house-of-cards' arrangement, collapse easily. Normally, disturbance of such sediments leads to a more stable arrangement on recovery: the remoulded sediment gains in strength. However, reflocculation of the particles may generate a framework that is weaker than before. The situation can arise, for example, through slight changes in pore-water chemistry and where chemical leaching has reduced grain cohesion, so that the restored framework is less rigid (Torrance 1999). Such sediments are termed 'sensitive' (Torrance 1983) and where the ratio of undisturbed to disturbed strength is large, the sediment is regarded as 'quick'. Torrance (1983) reviews schemes for quantitively defining the quick condition and the relevant mechanisms. The structures produced by sensitive sediments appear indistinguishable from those formed in sediments liquidized by other mechanisms, though with the high porosities that are normally required they are only likely to form in situations of shallow burial.

Liquefaction

Where any additional load on a sediment is wholly sustained by the pore-fluid and cohesion is negligible then the sediment loses strength completely and effectively behaves as a fluid – a state known as liquefaction. Such a situation can readily come about in relatively near surface sediments, where they are susceptible to the processes of particularly rapid loading and diagenesis may not have proceeded far. The sediment will remain in this condition until the pore pressure is reduced and some inter-particle frictional strength is restored. Volumetrically, most sediment mobilization appears to be ascribed to liquefaction, although recent work is emphasising the importance of fluidization (see below). Vaid & Sivathayalan (2000) summarized the kinds of variables that affect the susceptibility of sands to liquefaction, which includes the fabric adopted by the grains following previous disturbance. Oda *et al.* (2001) argued that reductions in resistance to repeated liquefaction events are due to increased void connectivity, which promotes sensitivity to future stresses.

Fluidization

Where the sediment strength is lost through moving interstitial fluids buoying the particles, the state is called fluidization. The fluid-drag force balances or exceeds the particle weight, a situation normally requiring rapid ingress of external fluids and usually in subsurface sediments operating in an upward direction. Sands and other coarse, permeable sediments are probably more vulnerable to these effects of rapid pore-fluid flow than low permeability sediments such as clays. Jolly & Lonergan (in press) discuss how theory predicts that sands with well-sorted, rounded grains should fluidize most easily, although in some natural examples the coarsest sediments have preferentially mobilized. Hovland & Judd (1988) have discussed the movement of gas in sediments and Nicholls *et al.* (1994) discussed how in fluidized near-surface sediments the form of the resulting structures depends on factors such as permeability and strength of the sediment, and the rate of the pore-fluid movement. There is a vast amount of civil engineering literature on liquefaction and its mitigation (e.g. Seed *et al.* 2001) and on fluidization and its engineering applications (e.g. Yang 1999).

Sediment mobilization

Liquefaction and fluidization relate to the state of the sediment and in themselves are not sufficient to give bulk movement of the sediment. However, such liquidized sediments are behaving, by definition, as a fluid and are, therefore, subject to fluid pressure gradients. The elevation head, as in a fluid at rest, no longer balances out the pressure head in overpressured fluid and so the fluid will attempt to move to lower values of the potential gradient. For this to occur, the hydraulic gradient must also be able to trigger movement of the weakened sediments and the gradient has to be sustained long enough to allow the movement to be accomplished. For example, rupture of a low permeability layer that enabled liquefaction of the underlying, sealed sediments may suddenly allow their escape along the pressure gradient. The sediments can then only move as long as they are sufficiently overpressured to still be liquefied. Fluidized sediments require vigorous fluid movements but sediment can only be displaced along with the moving fluid as long as the pressure gradient is sustained. In other words, mobilization processes tend to be self-terminating.

The pore fluid in a fluidized sediment is similarly impelled in the direction of lower fluid potential, but in this case the sediment particles are moving by entrainment. In both the fluidization and liquefaction cases, the fluid gradient will be upwards overall (Hovland & Judd 1989), but can locally be in any direction, including downwards (e.g. Nicholls 1995). This is why, in addition to any constraints presented by the host material, mobilized sediments do not necessarily intrude upwards. Dasgupta (1998), for example, reported liquefaction structures recumbent in orientation, presumably due to locally horizontal hydraulic gradients and Huang (1988) described clastic sediments intruding downwards.

In recent years, the injection of externally pressurized liquids and gases into high permeability sediments, reducing the strength of the aggregate and driving rapid fluid movements that buoy the sediment particles, has been invoked for the mobilization of sands in the North Sea (e.g. Dixon *et al.* 1995; van Balen & Skar, 2000). Such fluidization will be very largely directed upwards and may be occurring in hydrocarbon provinces on a scale not fully realized, with major implications for reservoir geometries (e.g. Lonergan *et al.* 2000; Jolly & Lonergan in press).

Sands appear to be most amenable to fluidization, especially where porosities are high. Coarser sediments normally have greater porosities and permeabilities, but these factors seem to be offset by the greater flow velocity need to buoy the heavier clasts. In addition, it would seem likely that larger apertures would be needed to allow coarser sediments to move. Jolly & Lonergan (in press) have recently reviewed the factors that determine the structures that result from sand mobilization. In general, clays are too impermeable to allow the bulk fluid movement necessary for fluidization and their greater cohesion will resist disaggregation. Both sands and clays can liquefy at shallow burial depths, where loading processes easily exceed the sediment strength, and mobilization is facilitated by the steep hydraulic gradients resulting from the nearby free surface. However, porosity loss with burial soon increases the frictional resistance of sand and progressively improves its resistance to liquefaction. Hence, at depth relatively steeper potential gradients are required to mobilize sand. This may be one reason why sand reaches surface to be extruded as volcanoes less commonly than clays.

An additional driving effect can operate. Sediments that are weakened or liquidized may well be less dense and less viscous than the overlying material and the resulting reversed gradients will be highly unstable. Because gravitational potential results from the product of height above datum and mass, the overlying layer has a higher potential energy. It will tend to sink while the buoyant liquefied layer mobilizes upwards, in an attempt to produce the more stable, greater-density-with-depth, configuration. Any perturbations at the interface between the overlying, denser material and the underlying less dense sediment will, where both are behaving as fluids, act as Rayleigh-Taylor instabilities, causing the irregularities to amplify until the gravity-driven overturn can be achieved (e.g. Ronnlund 1989; Harrison 1996).

Thixotropic behaviour and Rayleigh-Taylor instabilities in layered sediments are probably confined to shallow levels of burial and the resulting structures to sizes of no more than tens of metres or less. Liquefaction and fluidization structures occur at these scales also, but range up to the kilometre sizes now being documented on seismic sections. It is important to note that most sediment mobilization, certainly at the larger scale, probably involves more than one mechanism. Liquefaction and fluidization probably often work together, sediments with sensitive fabrics may be involved, while associated material may retain significant residual shear strength and be deforming at or near its critical state.

Diapiric melanges provide a good example of material in different mechanical states being intimately related (e.g. Barber *et al.* 1986). Clay diapirs are essentially driven by excessive fluid pressures, which also act to liquidize much of the sediment. Liquid mudflows result if there is extrusion. Other parts of the intruding diapir may be shearing under critical state conditions, forming the scaly clays typical of such material (e.g. Vannucchi *et al.* in press). Whether the sediment moves as a fluid, through liquidization, or as a ductile solid through critical state deformation, will depend on factors such as lithology and the physical conditions, which will vary with time and position in the diapir. For example, material at the diapir margin or near to a potential conduit will drain more efficiently and retain/regain some residual strength. Orange (1990) has described how blocks in a diapiric melange remain solid during intrusion, and Brown & Orange (1993) have documented how various mechanisms operated at different parts of a diapiric melange and how they varied through time. Critical state deformation probably increases in importance with increasing depth, as progressive lithification reduces the ability of the sediment to liquidize and begins to promote cataclasis. Jones (1994) explained how the concept relates to bonded sediments such as shales and the initiation of shale diapirs.

Conclusion

Irrespective of their size, mobilization structures in incompletely lithified sediments owe their origin to the same basic cause: temporary weakening of the material as it progresses towards becoming rock. The above review has summarized how the sediments – which in contrast to rocks deform primarily by independent particulate flow – undergo such weakening. Some sediments are inherently prone to sudden weakening and are referred to as being sensitive, or in extreme cases, quick. Some sediments attain the stress/porosity combination that allows them to deform at their critical state where, typically in overpressured conditions and at low deviatoric stresses, the large shear strains associated with mobilization can be achieved even though the material still has strength.

Most sediment mobilization, however, comes about through the two main processes that are capable of liquidizing the sediment, making it behave as a fluid. Both involve overpressuring. These are: (1) liquefaction, where non-equilibrium additional loads are borne by the pore-fluid, so that friction between the cohesionless grains is effectively removed and the sediment loses its frictional strength; and (2) fluidization, where pore-fluids, commonly aided by injection of further liquid and/or gas from outside, can buoy the sediment particles. It is likely that both processes often work together to liquidize a sediment. The kinds of loads involved, additional to those due to normal sedimentation, range from monotonic (e.g. suprajacent emplacement of allochthonous masses) and cyclic (e.g. oscillatory passage of storm or seismic waves) dynamic loading to the results of internal processes such as mineralogical changes in the sediment. Because of the over-pressuring, the liquidized sediment is subject to a hydraulic gradient, normally but not necessarily upwards. However, only if the overpressuring seal is ruptured, say by hydraulic fracturing, while retaining some driving hydraulic gradient, can there be bulk movement. As long as the sediment remains liquidized or deforms at critical sate conditions, it is capable of undergoing very large displacements, even on the scale of kilometres now being reported from mobilized sediments.

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