Large Landslides on Passive Continental Margins: Processes, Hypotheses and Outstanding Questions

D.G. Masson, R.B. Wynn, and P.J. Talling

Abstract The volume, area affected, and runout of submarine landslides can exceed those of terrestrial events by two orders of magnitude. The Storegga Slide off Norway affected an area the size of Scotland and moved enough sediment to bury the entire country to a depth of 80 m. Modern geophysics provides a clear picture of large landslides and what their source and depositional areas look like. From this, we can deduce the processes that operated during downslope transport. However, our understanding of many aspects of landslide processes is based on hypotheses that are difficult to test. Elevated pore pressures are essential for landslide initiation on low continental margin slopes, yet understanding of how high pressures are generated or how fluid migration affects slope stability is limited. Sediments may be pre-conditioned for failure by the processes that originally deposited them, e.g., through creation of weak layers, but the processes and parameters that might control this are largely unknown.

Keywords Submarine landslide • slope instability • pore pressure • fluid migration • continental slope • bedding plane failure • trigger • preconditioning • methane hydrate • flow transformation

1 Introduction

Sedimentary deposits resulting from submarine mass movements (hereafter [submarine] landslides) occur on every passive continental margin, and are the dominant facies on many (e.g., NW African or US East Coast margins; Weaver et al. 2000; Hühnerbach et al. 2004; Twichell et al. in press). The scale of submarine landslides can exceed terrestrial events by two orders of magnitude and is almost impossible to comprehend. For example, the Storegga Slide off Norway (Fig. 1)

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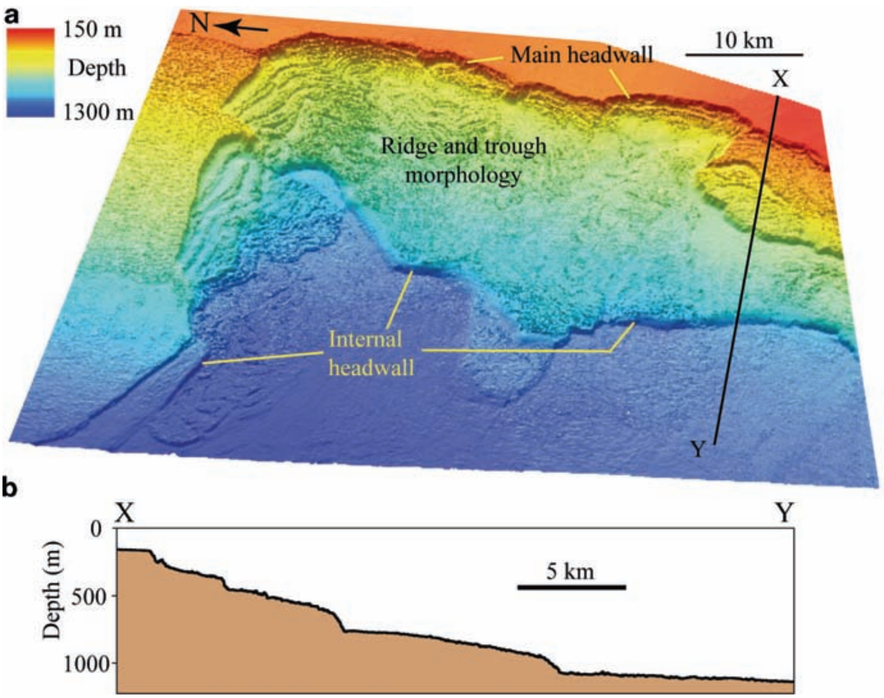

Fig. 1 (**a**) Shaded relief bathymetry image of the central part of the upper Storegga Slide, illuminated from the northeast, illustrating typical retrogressive landslide morphology. The ridge and trough topography is generated by lateral spreading, a landslide process in which a slab of sediment undergoes extension on a deforming softer layer (Micallef et al. 2007). The weak layers controlling failure progressively step up to shallower stratigraphic layers upslope. Areas controlled by different weak layers are separated by internal landslide headwalls, giving a characteristic step-like profile through the landslide scar (**b**)

affected an area similar to that of Scotland or the state of Kentucky (95,000 km^2) and moved sufficient sediment (~3,000 km^3) to bury either area to a depth of 80 m. The landslide occurred on a slope of about 1° (Haflidason et al. 2004).

Modern geophysical techniques provide a clear picture of where landslides have occurred and what their source and depositional areas look like. From this, the processes that operated during downslope transport (predominantly sediment slides, debris flows and turbidity currents) can often be deduced. For historical landslides, the trigger may also be known, with earthquakes, storms and anthropogenic effects among the most important. However, for many aspects of landslides our 'understanding' is based on hypotheses that are currently difficult to test in the real world. For example, it is generally agreed that elevated pore pressures are required to initiate landslides on typical low continental margin slopes, yet understanding of how these elevated pressures are generated or how fluid migration might affect slope stability is limited. To what extent are sediments pre-conditioned for failure by the processes that originally deposited them, for example through the creation of weak layers or by diagenesis? Is slope stability and landslide frequency affected by climate or sea-level change? Can changes in methane hydrate stability generate slope failure? Why do some landslides form slides or debris flows and others turbidity currents (while only a few are known to have formed both) and what controls transformations between flow types? How is several hundred cubic kilometers of sediment dispersed in an even larger volume of water during the formation of large turbidity currents? This paper presents the authors' view of current understanding of large landslides on passive continental margins, based mainly on observations of known landslide events.

2 Interpretation of Landslide Morphology

High-resolution bathymetry and sidescan sonar data show that many large landslides have similar morphology (Figs. 1, 2). This suggests a common failure process. In particular, landslides typically occur on a failure surface (or surfaces) conformable with the sediment bedding (O'Leary 1991).

2.1 Characteristics of Bedding Plane Parallel Landslides

Examples of large bedding plane parallel landslides on passive margins worldwide include the Storegga and Traenadjupet Slides offshore Norway (Bugge et al. 1988; Laberg and Vorren 2000; Haflidason et al. 2004; Micallef et al. 2007, 2009), the

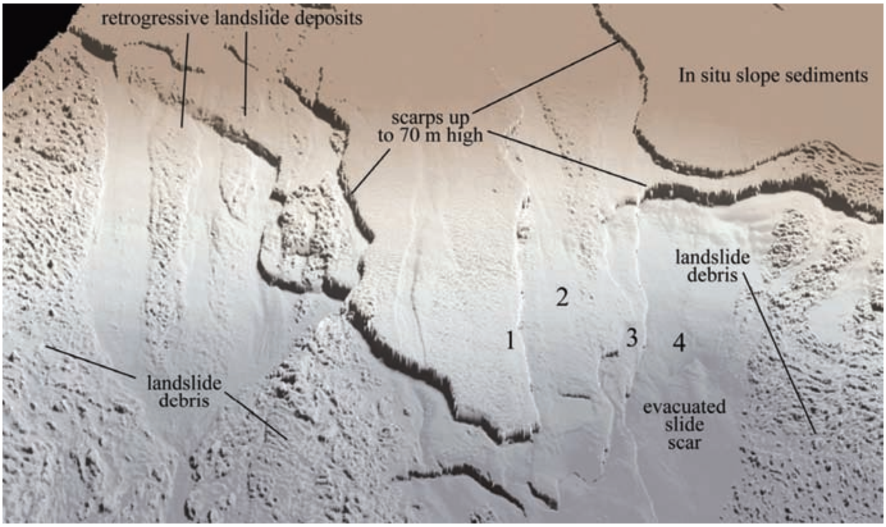

Fig. 2 Shaded relief image of part of a large landslide on the West African continental margin. Water depth varies from about 600 m (*top*) to 2,000 m (*bottom*). Image is approximately 25 km from *left* to *right*. Numbers indicate four distinct bedding plane parallel levels on which various sections of the landslide failed

Saharan and Mauritania Slides off West Africa (Georgiopoulou et al. in press; Krastel et al. 2006) and the Grand Banks and other slides on the eastern margin of North America (Piper et al. 1999; O'Leary 1991; Mosher et al. 2004). On the eastern flank of the Eivissa Channel in the Mediterranean Sea, four separate small landslides, all sharing the same failure surface, occur in parallel-bedded sediments (Lastras et al. 2004). These show continuity of a failure-prone layer over a distance >100 km.

The upper parts of bedding plane parallel landslides exhibit complex morphology that distinguishes landslide areas from adjacent areas of undisturbed smooth slope (Figs. 1, 2). Headwalls consist of multiple scarps ranging from a few meters to 100 m or more in height; internal headwalls separate different levels of failure surface (e.g., Bryn et al. 2003). This gives the landslide scar a staircase appearance, with the shallowest failure surfaces and thinnest failed section occurring farthest up slope (Figs. 1, 2). In some cases, tongues of debris that originated upslope in the landslide drape scarps downslope, indicating retrogressive landslide behavior (Fig. 2; Haflidason et al. 2004; Micallef et al. 2007). For most retrogressive landslides, the timescales over which retrogression occurs are not known. However, modeling of the tsunami associated with the Storegga Slide indicates that retrogression must have been rapid, with the entire slide mobilized in less than a few hours (Bondevik et al. 2005). *In situ* sediments adjacent to such landslides are characterized by parallel-bedded sequences (Fig. 3). Landslide failure surfaces are parallel to bedding and correspond to a limited number of key reflectors across the landslide (Figs. 1–3).

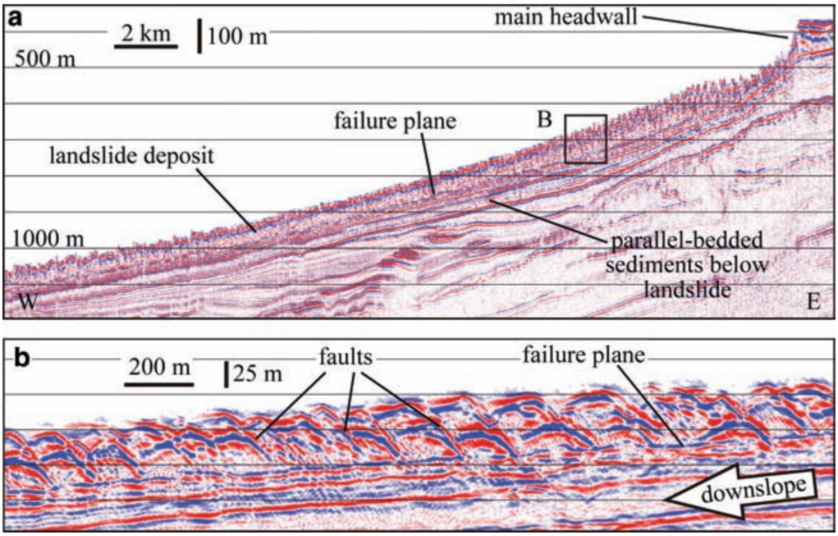

Fig. 3 Seismic profile from the Storegga Slide showing: (**a**) the steep headwall and apparently chaotic landslide deposit overlying the parallel-bedded *in situ* slope sequence. (**b**) Detail showing that the landslide deposit actually consists of coherent sediment blocks that preserve the pre-landslide stratigraphy, separated by distinct faults. Modified from Micallef et al. (2007)

Two distinctive end-member seafloor textures, one smooth and one rough, are typically seen within landslide scars (Figs. 1, 2). The smooth seafloor corresponds to evacuated landslide scar devoid of residual debris deposits. In contrast, the rough seafloor is covered by landslide debris up to a few tens of meters thick (Fig. 3). The structure of the debris, best illustrated by the bathymetry data, consists of closely packed blocks of only slightly deformed sediment separated by narrow depressions with a preferred alongslope orientation, interpreted as zones of more intense deformation (Figs. 1–3). The degree of fragmentation is highly variable, with the largest coherent blocks up to several kilometers across.

Because of the lower spatial resolution of seismic profiles relative to bathymetry, most landslide debris appears chaotic in profile, with individual blocks only poorly resolved (Fig. 3a). However, high-resolution profiles confirm the interpretation suggested by the bathymetry, imaging the coherent stratigraphy of the blocks interrupted by extensional normal faults (Fig. 3b). At the seabed, the faults correlate with seafloor depressions, confirming that the pattern of depressions seen in the bathymetry represents the plan view fault pattern (Micallef et al. 2007). The limited resolution of the seismic data restricts resolution of the thickness of any 'weak layer', suggesting only that it is less than a few meters thick. Cores that penetrate the failure surface typically show landslide debris overlying undisturbed older sediment; a distinctive layer may be seen at the level of the failure surface (Gee et al. 1999), but is not always recognized.

3 Failure Processes

3.1 Triggers and Preconditioning Factors

Many publications list the factors that contribute to initiation of submarine landslides (e.g., Hampton et al. 1996; Canals et al. 2004). Typically, such lists consist of two types of parameter that can be categorized either as 'triggers' or 'preconditioning factors'. Triggers are relatively short-period events that act to destabilize submarine slopes. Earthquakes, the trigger for most large historical submarine landslides (e.g., Hampton et al. 1996; Masson et al. 2006), are by far the best known, but low tides, storm wave loading, tectonic movements (e.g., tilting due to margin subsidence or salt movement), changes in methane hydrate stability and sealevel change also fall into this category. For the most part, the way in which these operate to trigger landslides, usually through rapid changes in sediment pore pressure, is understood, at least in principle. Preconditioning factors are related to properties of the sediments and are acquired during, or evolve from, the depositional process. Preconditioning includes aspects of the gross structure and stratigraphy of the sediments (e.g., presence or absence of sediment layers that might evolve into failure planes), water content, and pore pressure related to deposition (Canals et al. 2004). Slope might also be considered to be a preconditioning factor, but although a slope is clearly required, slope steepness does not seem to be important (Hühnerbach et al. 2004), although changes in steepness, e.g., due to tectonic movement, are important (e.g., Faroe Islands margin; van Weering et al. 1998). The importance of the overall fluid flow regime beneath the slope may also be important, as will be discussed later. Much less is understood about the role of preconditioning factors (in comparison to triggers) in the initiation, size and frequency of giant landslides.

3.2 Pore Pressure

Elevated sediment pore pressures, which act to reduce vertical effective stress in sediments (and thus the frictional resistance to landsliding), are among the key factors facilitating submarine slope failure (Canals et al. 2004; Kvalstad et al. 2005; Strout and Tjelta 2005). For example, it has been shown that the effective stress at the depth of the failure plane of the Storegga Slide needed to be reduced to only 10% of lithostatic stress (predicted on the basis of overlying sediment thickness) before failure could occur (Kvalstad et al. 2005). The required excess pore pressures were modeled by lateral migration of pore fluid from adjacent areas with much higher sedimentation rates and sediment thickness. This causes instability because fluid pressure can approach lithostatic stress where the overburden is thin. A similar process of lateral fluid migration has also been observed on the eastern continental margin of the United States (Dugan and Flemings 2000, 2002) and in the Gulf of Mexico (Flemings et al. 2008). At both these locations, pore pressure migration has been proposed as a cause of landslides.

4 Models for Large Bedding Parallel Landslides

The accepted view of continental slope sedimentation is that sediments derived from terrestrial erosion are transported over the continental shelf edge and deposited on the slope, building a seaward-thinning wedge-shaped deposit (Fig. 4a). Short-term sedimentation rates are highest on the upper slope, creating high pore pressures that lead to slope failure. The resulting landslides then propagate downslope through a combination of slope loading and erosion of the superficial slope sediments (Gee et al. 1999). Landslides that conform to this model are common on submarine deltas (e.g., those of the Nile and Mississippi rivers), where rapid deposition of muddy sediments on the upper slope predominates (Prior and Coleman 1982; Loncke et al. 2002). However, analysis of landslide morphology and water depth of landslide headwalls on continental slopes around the North Atlantic does not support the wide applicability of this model. Firstly, landslide headwalls are most abundant in mid-slope at 500–1,500 m water depth (some deeper than 3,000 m), rather than immediately seaward of the shelf edge (e.g., Saharan Slide and several landslides on the US East Coast margin (Hühnerbach et al. 2004;

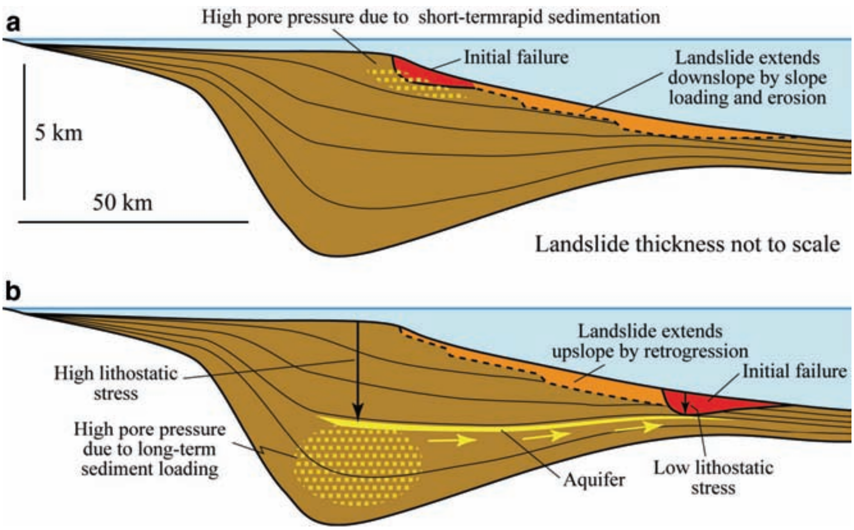

Fig. 4 Schematic illustration of a continental margin illustrating contrasting 'end member' initiation mechanisms for giant landslides. Landslide thickness is exaggerated to improve presentation. (**a**) Landslide initiated on the upper slope due to rapid sedimentation. (**b**) Landslide initiated on the lower slope due to lateral advection of high pore pressure from thicker sediment accumulation beneath the upper slope. In both (**a**) and (**b**), development of the landslide beyond the initial failure is controlled by weak layers within the parallel bedded slope sediment sequence

Georgiopoulou et al. in press; Twichell in press). Secondly, many large continental slope landslides are located on slopes characterized by moderate (when compared to submarine deltas) sedimentation rates (e.g., off NW Africa or the east coast of the USA; Weaver et al. 2000; McHugh and Olsen 2002). Thirdly, many landslides are retrogressive (Figs. 1, 2; Canals et al. 2004; Micallef et al. 2007). Observations of seafloor affected by retrogressive landsliding show that headwall retreat of tens of kilometers by this mechanism is common, indicating that the water depth at which the landslide initiated must have been considerably deeper than the final observed headwall water depth.

Our observations of headwall water depths and the retrogressive style of giant landslides are best explained by giant continental margin landslides that are initiated on the middle or lower continental slope. In particular, our proposed model (Fig. 4b) explains the depth distribution of landslide headwalls, the retrogressive nature of most giant landslides, and the progressively shallower depth of failure planes upslope. The lateral transfer of high pore pressure through the continental slope sediment sequence is key to landslide initiation on the lower slope. This requires (i) deposition of thick sediments on the upper slope, often on medium to long-term timescales, to generate initial high pore pressure, (ii) layering of relatively permeable and impermeable sedimentary horizons, to generate conditions where it is easier for pore fluid to migrate laterally over long distances along permeable horizons than to migrate vertically across the stratification, and (iii) good lateral continuity of the stratigraphy to ensure containment of the fluid flow system. For example, submarine canyons dissecting margin sediments may cut the stratigraphy and create a lateral escape route for high pore pressure, defusing the potential for large-scale landsliding. Thus areas of continental slope cut by canyons, and those affected by large-scale landsliding, are largely mutually exclusive (Weaver et al. 2000; Twichell et al. in press). In addition, the landslide initiation mechanism described above requires thick sediments rather than high sedimentation rates. This can explain landsliding in areas such as off NW Africa or on the eastern US continental margin, where sedimentation rates are moderate but have been continuous over long periods of time.

Once initiated, landslides can retrogress upslope because of loss of support for the sediment above the evolving headwall. This is most effective where specific sediment layers (weak layers) focus strain-softening behavior (strain concentration and loss of strength). Where such layers exist, the retrogressive failure mechanism can be modeled in accordance with conventional engineering principles (Kvalstad et al. 2005; Micallef et al. 2007). The sedimentology of weak layers is not well understood, mainly because they are destroyed in areas where landslides have occurred and it is not known how to identify them before they fail. High-resolution profiles from landslide scars show that failure planes almost always follow individual bedding planes, so it seems likely that weak layers correspond to specific sediment beds (Haflidason et al. 2003; Krastel et al. 2006) although “mechanical discontinuities” between beds of differing engineering properties have also been suggested (Haflidason et al. 2003; Canals et al. 2004).

Lateral propagation of high sediment pore pressure and the occurrence of laterally continuous, parallel-bedded sediments are two of the most important factors controlling occurrence of giant submarine landslides. Thus understanding the giant landslide potential of any continental margin requires a comprehensive understanding of the whole margin system. Geotechnical studies of the upper sediment sequence, as typically carried out by offshore industry, address only local slope stability issues, and on their own give little information on regional slope stability. Modeling of the pore pressure regime beneath the slope can be effective (Kvalstad et al. 2005), but full assessment of the fluid flow regime requires knowledge of the geology and physical properties of the sediment sequence beneath the continental slope, and is only possible where deep drilling information is available.

5 Outstanding Questions

Understanding of many aspects of landslide processes is based on hypotheses that are difficult to test. In many instances, we can infer how a landslide has behaved (e.g., in terms of flow processes or flow transformations) without understanding why it has behaved in this way. Some key questions to which, at best, we have only partial answers, are discussed below.

1. Why do some areas fail when adjacent areas don't? While it is recognized that bedding plane parallel landslides seem to occur where there is stratigraphic continuity over large areas of margin, it is also true that often only part of such an area fails, and a landslide area may change gradually into *in situ* slope through a zone of partial failure (e.g., northern margin of Storegga Slide; Haflidason et al. 2003). This is presumably due to subtle variations in pore pressure distribution, the spatially varying effects of the triggering event (e.g., an earthquake) or the mechanical properties of the basal 'weak' layer, but we have very little understanding as to the nature of these variations.
2. Climate and sea-level change appear to influence the distribution and frequency of continental margin landslides by changing the sediment delivery to margins, or more indirectly through changes in the gas hydrate stability zone or the frequency of earthquakes related to ice loading or unloading (e.g., Owen et al. 2007). Understanding of climatic influence on landslide frequency is particularly important for hazard analysis, since past frequency of landslides that may have occurred under a different climatic regime is often used as an indicator of future hazard. However, few submarine landslides have been dated with accuracy, limiting the effectiveness of slide frequency analyses. Most landslide ages are based on the age of sediments overlying the landslide deposit and should only be regarded as minimum ages; rough, elevated landslide topography may inhibit post-landslide sedimentation leading to erroneously young dates. Dating multiple cores from a landslide can overcome this problem, as in the case of the Storegga Slide (Haflidason et al. 2005). Alternative approaches include dating cores that penetrate the edge of a landslide deposit into the underlying sediments, thus giving a minimum and maximum age (e.g., Saharan Slide, Gee et al. 1999) or dating landslide-related turbidites (Weaver and Kuipers 1983) although establishing definitive landslide-turbidite correlation may be difficult.
3. Can changes in the methane hydrate stability zone on continental margins triggering submarine landslides? Such changes, leading to pore pressure increases or mechanical weakening of hydrate-bearing sediments have been suggested as a triggering mechanism for landslides on several continental margins including the Norwegian Sea and parts of the US East Coast margin (Maslin et al. 2004; Sultan et al. 2004; Hornbach et al. 2007 and references therein; Owen et al. 2007; Berndt et al. 2009). However, evidence supporting this theory is largely circumstantial and it remains controversial. Many landslides have occurred in areas where there is no evidence for the development of gas hydrates, although the difficulties in identifying hydrate bearing sediments in areas where bottom simulating reflectors (BSR) are not recognized on seismic profiles is acknowledged. Recent recognition that many large margin landslides were initiated in deep water (Hühnerbach et al. 2004; Twichell et al. in press) also mitigates against the involvement of hydrates in their triggering, since hydrates are most susceptible to dissociation at depths <1,000 m (Sultan et al. 2004; Hornbach et al. 2007). In the case of the Storegga Slide, the recognition that the slide probably initiated at water depths between 1,500 and 2,000 m appears to rule out hydrate dissociation as a triggering mechanism, since post-glacial warming of

the Norwegian Sea was limited to water depths <700 m, with deeper waters remaining below 0°C to the present day (Bryn et al. 2005).

4. Long runout, sometimes for many hundred kilometers on very low slopes (<1°), characterizes many continental margin landslides. This requires either a low viscosity flow (not generally supported by the cohesive nature and often only moderate deformation of the debris flow material) or a mechanism that lowers friction at the base of the flow. High pore pressure in a basal layer (Gee et al. 2001) and hydroplaning (Mohrig et al. 1998) have been suggested in this context. Are both these mechanisms possible, do they operate in similar or different settings, and how might the resulting deposits be distinguished?
5. Why do some landslides form slides or debris flows and others 'ignite' to form turbidity currents, and what controls transformations between flow types? On the West African margin, the Saharan Debris Flow, one of the largest landslides in the area (around 600 km^3; Gee et al. 2001), did not produce a significant turbidite (Georgiopoulou et al. in press), and is not recorded in the Madeira Abyssal Plain downslope (Rothwell et al. 1992). Yet other landslides on the same margin are the sources for the large turbidites (some >100 km^3) that form the fill of the same abyssal plain (Rothwell et al. 1992). All the landslides have essentially the same composition, yet behavior clearly differs significantly. Could the presence of thin sandy turbidite bases in some areas be important in the formation of turbidity currents? What is the influence of the topography over which the flow is moving?
6. How is several hundred cubic kilometers of sediment dispersed in an even larger volume of water during the formation of large turbidity currents? What causes the large-scale mixing that is required and how is the cohesion of the sediment overcome? Is it possible that hydroplaning facilitates injection of water into the base of the landslide, triggering flow instability and mixing?

6 Conclusions

Features that are common to many large landslides include:

(i) They occur in parallel-bedded sediment sequences that have stratigraphic continuity over large areas. Sediments are dominantly fine-grained.
(ii) Landslide headwalls typically occur at 500–1,500 m on the upper slope, although retrogression may extend the failure upslope to the shelf edge.
(iii) Failure occurs on one or more surfaces parallel to the sediment bedding, giving the landslide scar a staircase-like appearance. The failed section is typically 10–100 m thick.
(iv) The same layers fail over wide areas.
(v) Evacuated slide scars give areas of smooth seafloor. Residual debris within landslide scars consists of blocky, fragmented sediment.
(vi) Initial breakup of the landslide slab is by brittle deformation.

Taken together, these features suggest that bedding-plane parallel landslides result from failure of distinct ‘weak layers’ within sequences of otherwise stable sediments. True failure conditions only occur at the weak layer level.

Elevated pore pressure at the base of the landslide is clearly the key to failure. Such elevated pressure can be generated by rapid sedimentation, e.g., on submarine river deltas, and typically leads to landslides that initiate on the upper slope and propagate downslope. However, observations of many continental margin landslides suggest that they originate in areas with only moderate sedimentation rates and that they initiate in mid to lower slope and retrogress upslope. These are best explained by a model where high pore pressure generated deep beneath the upper slope migrates laterally into thinner sediment sequences beneath the lower slope, triggering failure of these sediments.

For many aspects of landsliding, the landslide deposits record how the landslide behaved, although we often don’t fully understand the underlying mechanisms. Why specific areas fail, how far the resultant landslide flows downslope, and how and why flows transform as they evolve during downslope transport, are among the more important outstanding questions.

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