On the triggers, resulting flow types and frequencies of subaqueous sediment density flows in different settings

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1. Introduction

Turbidity currents, and other types of submarine sediment density flow (Talling et al., 2012a), are arguably the volumetrically most important process for transporting sediment across our planet (Table 1). They dominate sediment transport into many parts of the deep ocean, and form the most extensive sediment accumulations on Earth (submarine fans). Subsurface deposits from ancient flows contain major oil and gas reserves. Individual turbidity currents can on occasions contain more than ten times the annual sediment flux for all of the world’s rivers (Table 1), and be more than 200 km wide (Talling et al., 2007a). Some flows can travel at speeds of 3 to 19 m/s for hundreds of kilometres, and break networks of sea floor cables (Piper at al., 1999; Hsu et al., 2008; Carter et al., 2012; Cattaneo et al., 2012). These cables now carry 95% of transoceanic data traffic, including the internet and financial markets (Carter et al., 2009). Seafloor infrastructure for recovering oil and gas, in some cases worth tens of millions of pounds, can also be damaged by fast moving flows (Barley, 1999). Flow deposits (turbidites) potentially provide a record of even larger (up to > 3000 km$^3$) submarine landslides, which can produce damaging tsunamis (Masson et al., 2006). Some slope failures are triggered by earthquakes, and turbidites may provide a valuable long-term record of major earthquakes, but widespread slope failure is the only reliable criteria for inferring seismic triggering. However, not all major earthquakes trigger widespread slope failure, so that the record is incomplete in some locations.

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and their timing and triggers, is therefore important for understanding how sediment is moved globally, effective recovery of oil and gas resources, hazards to strategic cable networks, and the recurrence intervals of tsunamis and earthquakes.

This contribution starts (Section 2) by summarising how sediment is moved, well power and run-out distance. The summary is based mainly on the only information available from a submarine work makes this review timely. Hyperrpycnal river floods can coincide with large wave heights during storms, or failure of rapidly deposited sediment, making triggers difficult to isolate. General comments are therefore made on the evidence for triggering of submarine flows by hyperrpycnal floods. This is timely because previous studies have inferred that such flows are common, and sufficiently powerful to transport sand into the deep ocean (Mulder et al., 2003).

The deposits of flows triggered by different processes are also discussed in Section 2. This is important because in most situations the only information available from a submarine flow is its deposit (Talling et al., 2012a). It is particularly important to determine whether flow deposits can provide a reliable record of major earthquakes that generate tsunamis (Goldfinger, 2011; Atwater and Griggs, 2012).

Table 1
Volumes and frequencies of different types of event.

<table>
<thead>
<tr>
<th>Process</th>
<th>Volume</th>
<th>Average frequency</th>
<th>Some key references</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slope failures</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Storegga Slide offshore Norway</td>
<td>&gt; 3000 km³</td>
<td>–</td>
<td>Hauf and vcason et al. (2005)</td>
</tr>
<tr>
<td>15 non-volcanic slides in last 36 ka &gt; 100 km³; only well dated examples</td>
<td>&gt; 100 km³</td>
<td>Less than 2400 years</td>
<td>Uralab et al. (2013a) (global database)</td>
</tr>
<tr>
<td>1929 Grand Banks landslide-turbidite</td>
<td>-175 km³</td>
<td>–</td>
<td>Piper et al. (1999)</td>
</tr>
<tr>
<td>Turbidites in Madeira, Agadir, Seine, and Baoeic abyssal plains</td>
<td>-5 to -250 km³</td>
<td>1 per 1000 to 40,000 years</td>
<td>Clare et al. (2014), Talling et al. (2007a)</td>
</tr>
<tr>
<td>Volcanic island flank collapses of the Western Canary Islands</td>
<td>50 to 500 km³</td>
<td>1 per 150,000 years</td>
<td>Masson et al. (2006), Hunt et al. (2011)</td>
</tr>
<tr>
<td>Canyon head failures (can run out to deep ocean)</td>
<td>-0.01 to -0.001 km³</td>
<td>0.1 to 5 per year</td>
<td>Piper and Savoye (1993), Cooper et al. (2013), Carter et al. (2012), Hsu et al. (2008), Marshall (1978).</td>
</tr>
<tr>
<td>Delta-lip failures — Canadian fjords</td>
<td>-0.0001 km³</td>
<td>1 to 5 per year</td>
<td>Hughes Clarke et al. (2012), Hill (2012)</td>
</tr>
<tr>
<td>Individual river floods</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Largest floods of single rivers Single river in one flood in Taiwan</td>
<td>0.03 to 0.06 km³</td>
<td>1 to 50 years</td>
<td>Liu et al. (2013)</td>
</tr>
<tr>
<td>Single flood of Santa Clara River</td>
<td>0.04 km³</td>
<td>Decadal</td>
<td>Gorsline et al. (2000)</td>
</tr>
<tr>
<td>Single large flood of Var River</td>
<td>0.007 km³</td>
<td>Decadal</td>
<td>Mas et al. (2010)</td>
</tr>
<tr>
<td>Jokulhlaup in Iceland, 1996</td>
<td>0.07 km³</td>
<td>–</td>
<td>Maria et al. (2000)</td>
</tr>
<tr>
<td>Average annual river fluxes</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Var River</td>
<td>0.0006 km³</td>
<td>Annual</td>
<td>Khripounoff et al. (2012)</td>
</tr>
<tr>
<td>Rhone River</td>
<td>0.003 km³</td>
<td>Annual</td>
<td>Lambert and Giovanoli (1988)</td>
</tr>
<tr>
<td>Taiwan (total of all rivers on island)</td>
<td>0.1 to 0.2 km³</td>
<td>Annual</td>
<td>Liu et al. (2013)</td>
</tr>
<tr>
<td>Largest annual flux from a single river (Amazon)</td>
<td>0.4 km³</td>
<td>Annual</td>
<td>Milliman and Syvitski (1992)</td>
</tr>
<tr>
<td>All the rivers in world for a year</td>
<td>-6 km³</td>
<td>Annual</td>
<td>Milliman and Syvitski (1992)</td>
</tr>
<tr>
<td>Oceanographic or anthropogenic events</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Deep water cascading (sediment) in single canyon during a winter event</td>
<td>0.005 km³</td>
<td>–</td>
<td>Canals et al. (2006, 2009)</td>
</tr>
<tr>
<td>Trawling in Fonera Canyon (one year)</td>
<td>0.00024 km³</td>
<td>Annual</td>
<td>Puig et al. (2012)</td>
</tr>
<tr>
<td>Volcanic processes</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Largest caldera forming eruptions, e.g. Toba ~74 ka</td>
<td>2800 km³</td>
<td>1 per 800,000 years globally</td>
<td>Self (2006)</td>
</tr>
<tr>
<td>Super-eruptions (~450 km³) — e.g. Taupo 26 ka</td>
<td>&gt;450 km³</td>
<td>1 per 100,000 years globally</td>
<td>Self (2006)</td>
</tr>
<tr>
<td>Krakatoa in 1883</td>
<td>~12 km³</td>
<td>1 per 50 years globally</td>
<td>Self (2006)</td>
</tr>
<tr>
<td>Mount St Helens in 1980 (erupted)</td>
<td>~1 km³</td>
<td>1 per 10 years globally</td>
<td>Self (2006)</td>
</tr>
<tr>
<td>Dome collapse and pyroclastic flow on Montserrat in 2003</td>
<td>0.21 km³</td>
<td>–</td>
<td>Trofinovs et al. (2006)</td>
</tr>
<tr>
<td>Other processes</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Snow avalanches</td>
<td>Typically &lt; 0.001 km³</td>
<td>–</td>
<td>McClung and Shaerer (2005)</td>
</tr>
<tr>
<td>Lahars at Nevado del Ruiz in 1985 and Mount St Helens in 1980</td>
<td>0.1 km³</td>
<td>–</td>
<td>Pierson et al. (1990)</td>
</tr>
<tr>
<td>Sediment mobilised on land during a single major earthquake</td>
<td>5 to 15 km³</td>
<td></td>
<td>Parker et al. (2011)</td>
</tr>
</tbody>
</table>
whose threat was demonstrated clearly by the 2011 Japan and 2004 Indian Ocean tsunamis. Previous work has proposed that flows initiated by plunging hyperpycnal river floods produce distinctive deposits (Mulder et al., 2001, 2003; Plink-Björklund and Steel, 2004). The degree to which initiation process can be inferred from deposits alone, and the distinctive features of turbidites triggered by earthquakes and floods, are therefore outlined.

The timing and frequency of flows is then considered in different types of settings characterised by flows triggered in different ways (Table 2), and at increasing distance from source. This results in a series of generalised models for certain settings, which are presented in Section 3. It has been proposed that variations in sea-level are a dominant control on flow frequency, as in older sequence stratigraphic models (Vail et al., 1977), although this has been questioned by recent work (Burgess and Hovius, 1998; Carvajal et al., 2009; Covault and Graham, 2010). Section 3 continues by analysing the relationship between changes in flow frequency and sea level and sediment supply in different types of settings. This contribution finishes (Section 4) with a comparison of the frequency and magnitude (tempo) of sediment transport in fluvial and submarine fan systems, and its implications for the global tempo of sediment transport, and the resulting efficiency of organic carbon burial in different types of submarine settings.

1. Aims

The first aim is to summarise how submarine flows are triggered, and the flow types and frequencies that result, based on (mainly direct monitoring) data that are reasonably unambiguous. The second aim is to determine the extent to which processes that initiate flows can be inferred from their deposits. In particular, can the deposits from

<table>
<thead>
<tr>
<th>Trigger</th>
<th>Type of flow produced; frequency</th>
<th>Some key references</th>
</tr>
</thead>
<tbody>
<tr>
<td>Freshwater lakes and reservoirs</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Plunging hyperpycnal river flood discharge</td>
<td>Typically relatively slow (&lt;30 cm/s), dilute (&lt;0.01 vol% sediment) and prolonged flows; deposition is thin (mm to 2 cm), fine grained layers.</td>
<td>Crookshanks and Gilbert (2008), Lambert and Giovanoli (1988), Lambert and Hsu (1979),</td>
</tr>
<tr>
<td>2. Delta-front or delta-lip slope failures</td>
<td>Rare and powerful (&gt;3 m/s) events may occur in channel systems in some lakes.</td>
<td>Lambert and Giovanoli (1988).</td>
</tr>
<tr>
<td>Marine delta fronts</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Delta-lip failures triggered by low tides and river flood peaks</td>
<td>Volumes up to ~0.001 km$^3$; can plug delta front channels and produce relatively powerful flows that reach distal lobes. Triggers may be associated with very low tides or individual river flood peaks.</td>
<td>Hughes Clarke et al. (2012, 2013), Hill (2012).</td>
</tr>
<tr>
<td>2. Events that cause motion of crescentic bedforms</td>
<td>Many tens of events may occur each year, many of which are not associated with individual river flood peaks.</td>
<td>Hughes Clarke et al. (2012, 2013).</td>
</tr>
<tr>
<td>3. Plunging hyperpycnal river flood discharge</td>
<td>Only when rivers have hyperpycnal discharge. Very dilute; can follow density interfaces in water body, or form plumes along water surface</td>
<td>Mulder et al. (2003), and as seen in lakes.</td>
</tr>
<tr>
<td>River-fed canyon</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Slope failure in canyon-head or canyon margins</td>
<td>Rapid deposition of flood sediment may lead to delayed slope failure; earthquakes may trigger multiple slope failures. Both can lead to canyon flushing events that reach deep ocean.</td>
<td>Heezen et al. (1964), Hsu et al. (2008), Carter et al. (2012), Cooper et al. (2013).</td>
</tr>
<tr>
<td>2. Plunging hyperpycnal river flood discharge</td>
<td>Proposed here that these events form dilute and slow flows that deposit thin laminae (Fig. 8). Unknown whether they can reach deep ocean.</td>
<td>Khripounoff et al. (2009, 2012), Maa et al. (2010), Liu et al. (2012, 2013).</td>
</tr>
<tr>
<td>3. Large waves during storms.</td>
<td>Wave loading may re-suspend sediment in canyon – as for canyons fed by oceanographic processes.</td>
<td></td>
</tr>
<tr>
<td>Canyon systems fed by oceanographic systems</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Events in upper canyon that move crescentic bedforms; associated with large waves.</td>
<td>Most likely involves hydraulic jumps in dense liquefied flows associated with up-slope migrating breaches; but origin not well constrained.</td>
<td>Paul et al. (2010a), Paul et al. (2012), Cartigny et al. (2010).</td>
</tr>
<tr>
<td>3. Storm or internal tide resuspension of sediment</td>
<td>Dilute and slow moving flows, and tend to infill the canyon with fine sediment.</td>
<td>de Stiger et al. (2007), Martin et al. (2011).</td>
</tr>
<tr>
<td>4. Dense water cascades from shelf.</td>
<td>Associated with winter storms, and may coincide with wave resuspension in canyon head. Dilute flows transport large water masses, but tend to infill canyon with thin sediment layers.</td>
<td>Canals et al. (2006), Puig et al. (2008), Palanques et al. (2009).</td>
</tr>
<tr>
<td>5. Wave/tide modified flows that spill off shelf.</td>
<td>It is poorly known how these events continue beyond the continental shelf edge, but may transport large fractions of flood sediment.</td>
<td>Wright et al. (2001), Wright and Friedricks (2006), Puig et al. (2003).</td>
</tr>
<tr>
<td>6. Canyon flushing events.</td>
<td>Poorly documented as infrequent; most likely due to failures triggered by quakes or other processes.</td>
<td>Arzola et al. (2008), Talling et al. (2007a).</td>
</tr>
<tr>
<td>Open continental slopes</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Slope failures with a wide range of scales.</td>
<td>Infrequent, but can be extremely powerful and carry &gt;100 km$^3$ of sediment</td>
<td>Piper et al. (1999), Talling et al. (2007a).</td>
</tr>
<tr>
<td>2. Other processes.</td>
<td>Flows initiated by re-suspension of sediment by storms or tides in shallow water; wave modified cross shelf flows that spill off shelf, etc.</td>
<td>Dengler et al. (1984), Wright and Friedricks (2006)</td>
</tr>
</tbody>
</table>
submarine flows triggered by earthquakes and river floods be identified with confidence, and if so, how? This leads to some general comments on turbidite palaeoecology. The third aim is to summarise flow triggers, types and frequencies for a series of different settings. Generalised models are developed for each setting that seeks to understand their varied response to sea level and climate change. Finally, a comparison is made between the magnitude and frequency of sediment transport in submarine flows and rivers (Korup, 2012). This comparison aims to understand the general tempo of global sediment transport, its implications for organic carbon burial or loss.

1.3. Terminology

The term oceanographic system is used following Piper and Normark (2009; p. 348) to denote a system in which flows are triggered mainly by suspension of coastal, shelf or upper slope sediment by oceanographic processes including storms, tides and internal waves. Hyperpycnal river discharge is denser than the lacustrine or oceanic water body that it enters. Triggering of hyperpycnal flows by river discharge involves plunging of river water to continue to flow along the seafloor or lakefloor. If the sediment is deposited and stationary for a period of time, before being subsequently remobilised by slope failure or oceanographic processes, then this does not constitute a hyperpycnal flow. For example, if bedload on a fjord head delta is first deposited and later remobilised, it would not be termed a hyperpycnal flow in this contribution. The term canyon is used for a relatively deeply incised (typically > 100 m) conduit formed primarily by erosion. Gully is used for smaller-scale (typically a few tens on m deep) conduits, formed by erosion on a delta-front. Channel is used for a conduit formed mainly by deposition. Basin plain is used to denote a relatively flat area that lacks any channels or canyons. The terms used for various types of subaqueous sediment density flow (on occasions abbreviated to submarine or sublacustrine flow) are those defined by Talling et al. (2012a). Turbidity current is used only for flows that are relatively dilute and fully turbulent, and deposit sediment in a layer by layer fashion. Debris flow is a type of subaqueous sediment density flow that deposits sediment en-masse (Talling et al., 2012a,b). The term landslide is used here for other types of mass movement resulting from slope failure that are typically denser and have less internal deformation than debris flow, such as slumps or slides (Nardin et al., 1979; Schwab et al., 1993; Masson et al., 2006).

2. Flow triggers and the types of flow and deposit that result

The first part of this contribution analyses the processes that trigger turbidity currents, and discusses the types of flow and deposit that result, including new insights from recent studies that have monitored flows in action.

2.1. Flows generated by slope failure

Submarine and sublacustrine slope failures have an exceptionally wide range of scales, with volumes spanning many orders of magnitude. Three of the main types of slope failure are discussed here. It is important to understand how these slides are triggered in order to understand the origin and frequency of resulting subaqueous sediment density flows, whilst noting that not all slope failures disintegrate and mix with water to form such flows.

2.1.1. Open continental slope failures

The largest failures occur on open continental slopes and can be prodigious in scale, sometimes extending for several hundred kilometres and involving >3000 km³ of material (Fig. 1A; Table 1; Schwab et al., 1993; McDado et al., 2000; Hühnerbach et al., 2004; Hafldason et al., 2005). Events such as the Storegga Slide offshore Norway have produced damaging far-field tsunamis (Bondevik et al., 2005). Large landslides on the open continental slope typically have a stair-stepped morphology resulting from widespread failure along distinct weak horizons (Fig. 1A; Masson et al., 2006, 2010a), although they can occasionally involve much deeper failure (Winkelmann et al., 2008). The most remarkable feature of these slides is that they typically occur on seafloor gradients of <2°, implying that excess pore pressure must support a large fraction of sediment load before failure can occur (Dugan and Flemings, 2000; Flemings et al., 2008). Such low gradients are almost always stable on land (Masson et al., 2006, 2010a).

We are yet to monitor a major submarine landslide in action, and models for their preconditioning factors and triggers are poorly tested, leading to uncertainty. Most large continental slope landslides occur in water depths that are too deep for triggering by cyclic wave loading (Fig. 1B; McDado et al., 2000; Hühnerbach et al., 2004). It is likely that slopes are preconditioned to fail by rapid accumulation of impermeable sediment that generates high excess pore pressures (Fig. 1C; Flemings...
et al., 2008). This excess pore pressure may be transmitted to the toe of slope, where the overburden is less, thereby causing failure (Fig. 1D; Dugan and Flemings, 2000). However, large failures also occur in areas of slow sediment accumulation (Urlaub et al., 2012), and failure can occur several thousand years after rapid sedimentation has stopped (Fig. 1E; Leynaud et al., 2007; Urlaub et al., 2013a). It has been observed from the timing of cable breaks in several locations that continental slope failure can be triggered by major (>Mw 6.7) earthquakes (Heezen and Ewing, 1952, 1955; Piper et al., 1999; Cattaneo et al., 2012). Cable breaks document that fast sediment flows (3 to 19 m/s) were generated that ran out for hundreds of km beyond the continental slope (Fig. 2A, B). However, some large (Mw 8–9) earthquakes have failed to produce widespread slope failure (Völker et al., 2011; Sumner et al., 2013), and large failures may be more common in areas of rapid sedimentation than areas experiencing strong seismicity and slow sediment accumulation (McAdoo et al., 2000).

Dissociation of gas hydrate has been proposed as a trigger of large continental slope failures as it can weaken sediment (Maslin et al., 2005). However, gas hydrates are stable under deep ocean conditions. Summary models (Fig. 1C and D) summarising how rapid sediment loading can precondition a continental slope to failure, causing failure at the top (upper panel) or base (lower panel) of the slope. (E) Plot showing the age of the Storegga Slide (red star), nearby sedimentation rate on trough mouth fan (solid line), changes in sea level (dotted line). There is an ~10 ka delay from cessation of rapid nearby sedimentation to triggering of the Storegga Slide.

Fig. 2. Summary of important field monitoring datasets in different locations worldwide, with additional key references noted in this caption. (A) Events associated with earthquake triggered failures of open continental slopes. The events occurred offshore from the Grand Banks in 1929 (Heezen and Ewing, 1952; Piper et al., 1999) and Algeria in 2003 (Cattaneo et al., 2012). (B) Events associated with river-fed submarine canyons systems, such as Gaoping Canyon offshore SE Taiwan (Hsu et al., 2008; Carter et al., 2012; Liu et al., 2012), Zaire Canyon offshore West Africa (Khripounoff et al., 2003, Vangriesheim et al., 2009), and the Var Canyon offshore southern France (Piper and Savoye, 1993; Mulder et al., 1997; Khripounoff et al., 2012).
Fig. 2 (continued). (C) Events associated with river-fed submarine delta including the Squamish and Fraser River deltas (Hill, 2012; Hughes Clarke et al., 2012, 2013) and Bute Inlet (Prior et al., 1987; Conway et al., 2012) in British Columbia, Canada. (D) Events associated with river-fed freshwater lakes and reservoirs in a wide range of settings (summarised in Talling et al., 2013; e.g. Crookshanks and Gilbert, 2008), and in Lake Geneva (Lambert and Giovannoni, 1988; Girardclos et al., 2012). (E) Events in canyon systems fed by oceanographic processes that include Monterey Canyon (Paull et al., 2010a, 2010b; Xu et al., 2010), Eel Canyon and Santa Monica Basin offshore California (Gorsline et al., 2000; Xu, 2010, 2011), and Nazaré Canyon offshore Portugal (de Stiger et al., 2007; Arzola et al., 2008). (F) Events in shelf edge canyons that are fed by oceanographic processes, such as the Cap de Creus Canyon in the Gulf of Lions (Canals et al., 2006; Palanques et al., 2009), Eel Canyon and Shelf Offshore California (Wright and Friedricks, 2006; Puig et al., 2003), and Huanghe River and Gulf of Bohai (Wright et al., 2001; Wright and Friedricks, 2006). (G) Key to symbols used in parts A to F. Figure is based on monitoring data summarised in Talling et al. (2013a).
E  CANYON SYSTEMS FED BY OCEANOGRAPHIC PROCESSES

Monterey Canyon, California
- fed by sand from shelf
- multiple events each year - mainly coeval with winter storms
- upper canyon events comprise (i) dense liquefied sand flows associated with breaching
  (ii) dilute turbulent flows with speeds of <1.8 m/s and thickness of 50-80 m
- very active with crescentic bedforms (1.6°)
- 40-1000 kg blocks moved (7°)
- no flows within last 100 years
- powerful canyon flushing events every ~100 years
- enigmatic “flow slides”

Hueneme Canyon and Santa Monica Basin
- fed sand off shelf by waves and wave-modified flows
- multiple dilute flows each year - mainly coeval with winter storms. Up to 2.6 m/s speeds; 15-25m thick; < 2 hour long
- large decadal river floods contain ~0.01 km³ sediment
- flushing events every 300-450 yrs
- 7 large (~6.5 to 7 Ms) quakes in last ~200 years; and 1 of these quakes within basin
- 6 basinwide thin (~5 cm) silty turbidites in last 450 yrs; volumes of 0.01-0.1 km³
- IODP Site 1015 - events every ~300 yrs in last 7 ka; recent layers are thick (~1 m) sands
- Hueneme Fan (1 to 0.2°) with 5-40 m leveed channels
- Santa Monica Basin (< 0.1°)

Nazare Canyon, Portugal
- no river, fed sand by longshore drift
- subannual, very dilute flows, speeds of < 0.5 cm/s, mainly during storms
- crescentic bedforms in sandy canyon head
- Iberian abyssal plain
- laggan and gravel deposits - from rarer powerful flows?
- scours, sediment waves and incipient thalweg channel
- more open canyon mouth (< 0.5°)

F  SHELF-EDGE CANYONS FED BY OCEANOGRAPHIC PROCESSES

Cap de Creus Canyon, Gulf of Lyons
- 5-25 km wide shelf
- 400 kg weights carried 2 to 9 km down canyon axis in remobilised bed
- furrows
- dilute flows (<0.01 % vol) that last for up to several weeks; cascading dense water from shelf, assisted by resuspension by storm waves on shelf; speeds up to 0.85 m/s; an event can transport 0.005 km³ of sediment and run out beyond 1 km water depth. May trigger secondary remobilisation of bed to move 400 kg weights or cut furrows
- Cap de Creus Canyon (2°)
- Pyreneo-Laguedoc Ridge and Rhone Fan (< 0.5°)

Eel River Shelf and Canyon, California
- flood sediment transported across shelf due to resuspension by wave action
- wave-modified gravity flows form fluid mud layers < ~15 cm thick, travelling at up to ~50 cm/s on gradients of ~0.3°

Yellow River Shelf and Gulf of Bohai, China
- sediment transport due to resuspension by tidal flows with speeds of up to 1 m/s; sediment tends to be trapped on delta front, and moved parallel to contours.

Fig. 2 (continued).
Changes in sea level have been inferred to increase the frequency of submarine landslides through increased loads due to a thicker water column (Maslin et al., 2004). However, assuming sufficiently permeable sediment, changes in sea level only increase the hydrostatic pore pressure with the sediment, which would not lead to failure. Alternatively, eustatic sea level and climatic change affects rates of sediment supply from land, with more rapid sedimentation favouring slope failure (Fig. 1). However, it is not yet clear whether there is a statistically robust relationship between sea level and the frequency of large (>1 km³) slides (Urlaub et al., 2013a). The similar morphology of many continental slope failures suggests that they may have a common trigger (Masson et al., 2010a). Urlaub et al. (2013b) propose that such a global trigger may be sudden loss of structure in marine clays as they are buried, which modelling suggests can lead to excess pore pressures at the observed depths of failure planes. However, in general the preconditioning factors and triggers for failure of remarkably low angle continental slopes are not well understood in comparison to other types of failure, such as cyclic wave loading in shallow water.

2.1.2. Canyon margin failures

Slope failures along canyon margins have much smaller volumes (~0.01 to 0.001 km³) than large open-slope landslides (Table 1; McDaid et al., 2000; Hühnerbach et al., 2004), due to shorter hill-slopes, and they occur on much steeper gradients making them easier to trigger. Scalped sea-floor morphologies show that failures occur both at the canyon head, and along canyon walls in deeper water (Marshall, 1978).

Repeated mapping and cable breaks have constrained the timing and triggers of failures in four locations. Failure at the head of the Var Canyon in 1979 was due to extension of a runway for a Nice Airport (Fig. 3A; Piper and Savoye, 1993; Mulder et al., 1997), although seepage of fresh groundwater may have played a role in weakening the sediment (Dan et al., 2007). Wave action may often cause slope failure in sufficiently shallow water through cyclic loading, which may explain an event mapped in Scripps Canyon (Marshall, 1978). Failure of rapidly and recently deposited sandstone sediment generated a powerful, long runout flow in 2009 along Gaoping Canyon, offshore SE Taiwan (Fig. 4A–D; Hsu et al., 2008; Carter et al., 2012). The flow did not coincide with large wave heights (Fig. 4D), so that the final trigger is unclear. Earthquakes triggered canyon head or margin failure in the Gaoping Canyon in 2006 and 2010, which may have been accompanied by failures across a wider area. These events also generated powerful, long run out flows (Fig. 4E). Flows in the Zaire Canyon (Fig. 5) tend to occur during months of elevated river discharge, especially soon after peak river discharge (Heezen et al., 1964; Heezen and Hollister, 1971; Cooper et al., 2013). The Zaire River drains a very extensive low gradient basin, rather than a steep mountainous basin, and it is thus unlikely to reach sediment concentration that produce hyperpycnal flows (Milliman and Syvitski, 1992). This is confirmed by limited sampling that recorded average monthly suspended sediment concentrations of ~0.001 vol.% (26.3 mg/l; Couyel et al., 2005). These offshore flows are most likely triggered by slope failures. These studies show that canyon-head failures can be triggered by rapid addition of sediment, especially if it is poorly consolidated, and by wave action and earthquake shaking.

Although some of these failures had rather small initial volumes (~0.01 km³), they can generate fast moving flows (3–8 m/s) that travelled for several hundred km to reach the deep ocean (Figs. 3 to 5). These flows most likely eroded and flushed sediment from the canyon-channel system, and transported sand to deep sea fans, due to their fast speeds. Smaller canyon margin failures can also generate weaker flows that infill the canyon with sediment (Xu et al., 2013).

2.1.3. Prograding delta-front failures

Near-daily repeated mapping of the Squamish River-delta suggests that these events can be subdivided into two types (Figs. 2C and 6; Hughes Clarke et al., 2012, 2013), which is consistent with monitoring in Bute Inlet (Fig. 8A; Prior et al., 1987; Bornhold et al., 1994; Hill, 2012). The first type are relatively infrequent events that are more powerful and have long run-out, whilst the second type are much more frequent events with shorter run-out. Repeated mapping of the Squamish delta front suggests that the more powerful events are triggered by relatively large (~0.00015 km³) failures of the rapidly prograding, steep (40°) delta lip (Fig. 6C). Two of these lip-failures coincided with unusually low tides, whilst three later events coincided with surges in river discharge (Fig. 6E; Hughes Clarke et al., 2012, 2013). These events reached lobes at the end of the 2 km long channels on the Squamish delta-front, with the largest failure leading to burial of an acoustic Doppler current profiler (ADCP) moored on the lobe (Fig. 6A, E). Similar lip-collapse failures have been documented on the Fraser River-delta, where they occur in sediment that is loosely packed due to rapid deposition and are attributed to a combination of gas expansion and low tides (Christian et al., 1997; Hill, 2012). The occurrence of lip-failure during individual flood peaks in the Squamish delta suggests that increased rates of lip progradation, or increased shearing of bed by flood flow, can also induce slope failure.

The first type of more infrequent and larger-volume delta lips failure (Fig. 6C; Table 1), and may produce deposits that plug adjacent delta-front channels (Fig. 6B; Hill, 2012; Hughes Clarke et al., 2012, 2013). The failure itself, or reworking of channel plugging deposits, can generate flows that run out for several kilometres onto the prodelta (Hughes Clarke et al., 2013), with even longer run out favoured by better developed channel systems (Conway et al., 2012).

The second more frequent type of event on the Squamish delta-front involved smaller scale slope failures. The resulting flows generate trains of up-slope migrating crescentic bedforms with steep (40°) downslope faces, heights of ~2 m, and spacings of 20 to 40 m (Fig. 6D). Enigmatically, movement of these bedforms can also occur when there is no recognisable failure scar in repeat bathymetric maps (Hughes Clarke et al., 2013). Similar bedforms are seen on the Fraser delta-front (Hill, 2012), and may extend for much longer distances (~25 km) along the floors of channels in Bute and Knight Inlet (Conway et al., 2012). They also resemble the crescentic bedforms seen in Monterey Canyon (Fig. 7; Paull et al., 2010a), and other canyons fed by oceanographic processes along the Californian margin (Xu et al., 2010; Paull et al., 2012). Monitoring by Hughes Clarke et al. (2012, 2013) suggests that events which move crescentic bedforms comprise a relatively dense and thin (~2 m) near-bed layer that drives an overlying dilute cloud of sediment. The bedforms do not appear to flows generated by plunging hyperpycnal river floods, as the suspended sediment concentration in these rivers is too low, and because they often do not coincide with individual flood peaks (Figs. 6E and 8A; Prior et al., 1987; Bornhold et al., 1994; Ren et al., 1996; Hughes Clarke et al., 2012, 2013). The crescentic bedforms may be generated by a similar retrogressive breaching process that generates dense sediment flows with cyclic hydraulic jumps, as proposed for Monterey Canyon (Fig. 7; Talling et al., 2013; also see Paull et al., 2010a; Cartigny et al., 2010 for alternative models). However, in these Canadian fjords bedform motion does not coincide with large wave heights (Prior et al., 1987), as is the case in Monterey Canyon (Paull et al., 2010a). Crescentic bedforms may therefore be a rather general property of supercritical flows, or areas with loosely packed sediment formed by rapid deposition that can be disturbed by a range of triggers.
produced by slope failure may be prolonged over many hours, although they are not as prolonged as some river flood discharges that last for days to weeks. Cooper et al. (2013) described flows 150 km offshore in the Zaire Canyon that last for 1–10 days (Fig. 5C–D), which are most likely triggered by slope failure as the Zaire River is unlikely to reach sediment concentrations necessary for hyperpycnal flows (Mulder and Syvitski, 1995; Coynel et al., 2005). The front of an initial slide triggered flow may outpace its tail to generate prolonged flow further offshore. For instance, a flow front travelling at 2 m/s will reach a location 150 km offshore almost 8 days before its tail, if that tail travels at 20 cm/s. Both open slope and canyon margin failures can produce flows that deposit thick sand layers across submarine fans, and it is unclear whether other types of flow can also do this. As noted by others (e.g. Normark and Piper, 1991; Plink-Björklund and Steel, 2004; Piper
Fig. 4. Long run-out flows that broke cables offshore SE Taiwan associated with a very large flood in 2009 (A to D) and earthquakes in 2006 (E). (A) Map showing the path of two submarine flows associated with Typhoon Morakot in 2009. (B) Bathymetric profile along the flow path showing the location of numbered cables broken in 2009. (C) Discharge of the Gaoping River during Typhoon Morakot. A single measurement (circle) recorded a sediment concentration of 60 kg/m³ near the Gaoping River mouth that would cause hyperpycnal flow. Inset shows the canyon head during peak flood discharge in 2009. (D) Wave height recorded offshore SE Taiwan during Typhoon Morakot in 2009, from Hale et al. (2012). (E) Map showing the path of two submarine flows triggered by the Pingtung earthquakes in 2006. Parts A–C are modified from Carter et al. (2012), Panel E is from Hsu et al. (2008). Differences in the distance from shore (km) compared to (A) result from the different ways in which distances were measured by Hsu et al. (2008) and Carter et al. (2012).
Fig. 5. Powerful, sustained flows observed at 2000 m water depth in the Zaire Canyon by Cooper et al. (2013). (A) Map showing the Zaire Canyon–fan system. Box shows area in part b. Red and blue dots show locations of other monitoring studies (red — Khripounoff et al., 2003; blue — unpublished). (B) Detailed map of the Zaire Canyon showing the locations of moorings (red dots) in the study of Cooper et al. (2013). The two thalweg moorings contained ADPs located 85 m above the seafloor (thalweg short) and 225 m above the seafloor (thalweg tall). Contour interval is 25 m. (C) ADCP time series of the velocity of a turbidity current that lasted for ~6 days, which sustained velocities of ~1 m/s for ~5 days at heights of 5.2 m (red line), 7.2 m (black line), and 9.2 m (blue line) above the bed. Data from the Thalweg Short mooring site. (D) Contour plot of flow velocity profiles through the same turbidity current measured using a 300 kHz ADCP. The data gap (white) results from a lack of penetration from the ADCP, suggesting high near bed concentrations. Data from the Thalweg Short mooring site. All part of figure modified from Cooper et al. (2013).
Fig. 6. Field observations from the delta system fed by the Squamish River in British Columbia, Canada. The Squamish River does not reach hyperpycnal sediment concentrations. (A) Map of the three active channels associated lobes. Yellow star shows position of ADCP mooring. (B) Map showing the cumulative difference in height between surveys on May 2nd and August 22nd 2011. (C) Surveys on consecutive days, and change in elevation associated with a major delta-lip collapse. JD is Julian day. (D) Surveys on consecutive days showing migration of crescentic bedforms within the channels. (E) Timing of 93 surveys and events within the three delta-front channels and at the ADCP mooring site on the northern lobe (A), together with changes in river discharge. Blue lines denote flow event or survey timing. Events in the three channels are deduced from repeat bathymetric surveys. The ADCP was buried and unable to record events for a short period. Major delta-lip failures were associated with unusually low tides (blue stars) or individual surges in river discharge (red stars). All from Hughes Clarke et al. (2012, 2013).
Fig. 7. Observations from Monterey Canyon (A to C) and La Jolla Canyon (D) offshore California. (A) Map showing extent of crescentic bedforms in Monterey Canyon. (B) Current metre data from three moorings in Monterey Canyon (R1, R2 and R3 in Fig. 4A) during a dilute turbidity current in 2002. The profiles are hourly measurements (colour-coded black, red, yellow, green, cyan, blue, and purple) from the start to finish of the event. Negative flow speeds are in the down-canyon direction. (C) Map showing crescent shaped bedforms wavelengths of 20 to 80 m and heights of ~2 m, location of sediment, and initial positions of three 40 kg concrete blocks with beacons (inset photo) that moved episodically down canyon. (D) Still image from video footage taken from an ROV whilst vibracoring near crescentic shaped bedforms in La Jolla Canyon. An upper layer of cohesive mud cracked and underlying sand flowed down slope. The image is from a location with an average gradient of ~1°. See the supplementary material of Paull et al. (2012) for the entire video clip. Panel B is from Xu et al. (2004). Panel C is from Paull et al. (2010a).
and Normark, 2009) slope failure can produce deposits that often differ from the Bouma Sequence in key regards, and their internal structure is varied (Talling et al., 2012a).

More cores from monitored flows are needed to improve our understanding of deposits resulting from delta-lip failures, although cores from Bute Inlet (Zeng et al., 1991) suggest that these events and associated longer run-out turbidity currents can form massive sand intervals that are several tens of cm thick. Vibracore transects across Monterey Canyon suggest that events forming crescentic bedforms produce thick amalgamated massive sands that pinch out a few metres above the canyon floor, and comprise massive poorly sorted clean sand with dispersed mud clasts (Paull et al., 2010a). In general, it is problematic to distinguish between different types of slope failure from the turbidites they generate, or between turbidites resulting from slope failure and other trigger mechanisms (Piper and Normark, 2009).

2.2. Turbidite paleoseismology

Turbidites triggered by slope failure resulting from major earthquakes have the potential to produce a valuable record of past earthquakes, because it extends further back in time than records on lands. Turbidite paleoseismology is now being used to infer frequencies, extents and magnitudes of earthquakes (Goldfinger, 2011; and references therein), with this information incorporated into hazard assessment and civil planning in locations such as the Cascadia Margin (Atwater and Griggs, 2012). However, to apply turbidite paleoseismology it is necessary to be able to recognise turbidites caused by earthquakes, and distinguish them from turbidites caused by other triggers (Table 3). It is also necessary to understand whether some major earthquakes fail to generate distinctive turbidites, such that the turbidite record of major earthquakes is incomplete (Sumner et al., 2013). If this is the case, it is important to determine which submarine
settings and earthquake types tend to produce more or less complete records.

Major earthquakes have triggered large volume failures on continental slope (Piper et al., 1999; Cattaneo et al., 2012), but large volume failures might be triggered in other ways, such as by dissociation of gas hydrates. Conduit flushing can also increase the volume of sediment within flows triggered in other ways. Turbidite volume alone is therefore not a particularly reliable way of inferring paleo-earthquake triggering. The most reliable criteria is the turbidite results from widespread slope failure, associated with the widespread ground shaking that characterises major earthquakes (Table 3). Other causes of slope failure, including storms, tend to affect smaller areas of sea floor. Wide-spread slope failure can be inferred in two ways; from synchronous emplacement of turbidites in adjacent basins and from confluence tests (Table 3).

2.2.1. Synchronous emplacement of turbidites in adjacent basins

Demonstrating synchronous turbidite emplacement in adjacent basins can be difficult because it needs precise age control, and careful assessment of potential uncertainties in turbidite ages (Table 3). Dates typically come from samples of hemipelagic mud intervals between turbidites, as material within the turbidite is reworked, and may be much older than the flow event. This means that hemipelagic mud must be confidently distinguished from turbidite mud, which is not always easy (Talling et al., 2012a). Hemipelagic accumulation rates must then be inferred from these samples to calculate the age of hemipelagic sediment immediately overlying a turbidite, and hence the age of turbidite emplacement. Even in locations with closely spaced AMS radiocarbon dates, plausible variations in the accumulation rate of hemipelagic mud and depths of erosion beneath turbidites may lead to significant uncertainties in turbidite ages, which are greater than those involved in the AMS radiocarbon dates themselves. Synchronous failure in different locations can lead to variations in turbidite composition, such that correlations based on composition may not be reliable. Emplacement of the same number of turbidites during the same time period in two basins may suggest earthquake triggering, but it is possible the turbidites are not synchronous.

2.2.2. Confluence tests

The confluence test infers that earthquake shaking affects the headwaters of both tributaries of a submarine confluence, which are too widely spaced for them to be affected by a single storm (Shanmugam, 2008; Table 3). In very large storms, hurricane-force winds can extend for several hundred kilometres (Shanmugam, 2008). If the same number of turbidites are then seen in both the incoming tributaries, as the number of turbidites in the outgoing tributary, an earthquake trigger is inferred. This model is complicated if earthquake shaking only affects one of the headwaters, or if one of the flow paths is dominant and flows along the other path tend to die out before the confluence (Atwater and Griggs, 2012). It is also possible that some turbidity currents bypass sediment and leave no deposit in locations with steeper gradients. A further significant complication arises from potentially variable flow thicknesses along the flow path, and between different turbidity currents. This can lead to variations in the number of turbidites present in a given location, at varying heights up the sides of the channel. The number of turbidites in a core can therefore depend upon the height above the channel floor, as well as the total number of turbidity currents that have passed through that location. Cores lowered from a surface vessel can have significant uncertainties in their seafloor position (tens of metres) such that it is problematic to try and sample the same height above the seafloor in different locations. This issue could be addressed, but only with dense cross sections of precisely located cores, which may need a ROV-based coring system.

2.2.3. Internal character of deposits

The internal character of a turbidite cannot be used with confidence to infer that it resulted from slope failure triggered by an earthquake (Piper and Normark, 2009; Table 3). It is likely that slope failures triggered in different ways can produce rather similar deposits, sometimes with relatively prolonged flow (Section 2.1.4). It is highly unlikely that temporal variations in earthquake shaking (seismogram) are recorded as characteristic variations in grain size within turbidites from that event (Goldfinger, 2011). The rate at which sediment is released from a slope failure is unlikely to correspond exactly to the history of ground accelerations, with complex patterns of retrogressive failure and detachment of separate lobes within the landslide mass. Indeed, subaerial slope failure has sometimes begun after the earthquake shaking has ceased (Atwater and Griggs, 2012). The sizes of grains deposited can vary with both sediment concentration and flow velocity (Kuenen and Sengupta, 1970), and parts of the flow travelling at different speeds can catch up and amalgamate with each other (cf. studies of pulsing hyperpycnal flows generated by a single river flood peak; Best et al., 2005; Krippouff et al., 2012; Lamb et al., 2010).

It has been inferred that multiple stacked fining up sequences (perhaps each with a different composition and source) are characteristic of earthquake triggered turbidites, as failure occurs in many locations across a wide area (Shiki et al., 2000). However, multiple fining up sequences can also result from multi-stage slope failure (Hunt et al., 2011), flow reflection, or indeed pulsing in a hyperpycnal flow (Best et al., 2005). So this grading pattern is also not strongly diagnostic.

2.2.4. Do all major earthquakes produce extensive slope failure and turbidites?

Coring near the epicentre of the 2004 (Mw 9.1) and 2005 (Mw 8.7) Sumatran earthquakes suggests that they did not produce widespread turbidites within intraslope basins (Sumner et al. 2013). Indeed, five of the six cores from these basins lacked any turbidites in the last 100 to 150 years, despite the well documented occurrence of multiple
large magnitude (Mw > 7) earthquakes nearby (Sumner et al., 2013). Mapping of the seafloor in this area found few large slope failures (Tappin et al., 2007). Comparison of bathymetric surveys completed before and after the Mw 8.8 earthquake offshore Chile in 2010 also found no slope failures extending for > 1 km (Völker et al., 2011). These studies included some of the largest magnitude earthquakes in recent times. Therefore, it is not at all clear that all major earthquakes generate widespread slope failure and extensive turbidites. It is possible that earthquakes can rather cause consolidation and strengthening of slopes in some settings (Lee et al., 2004). Conversely, the 1929 Grand Banks earthquake (Mw 7.2), 2003 (Mw 6.8) earthquake in Algeria, and the 2010 (Mw 7.0) earthquake off Haiti all generated extensive turbidites (Piper et al., 1999; McHugh et al., 2011; Cattaneo et al., 2012). Further work is needed in locations where it is known that a major earthquake has occurred, to understand which locations provide the most complete turbidite record of major earthquakes.

2.3. Oceanographic processes

Oceanographic processes that trigger sediment density flows include waves, tides and internal waves that resuspend sediment forming down slope flows driven by their own excess density or continued sediment resuspension (Piper and Normark, 2009; Puig et al., 2014). These sediment flows can be subdivided into two types, depending on whether such flows traverse and cascade off the shelf, or are initiated locally in canyons or on the open continental slope.

2.3.1. Flow off the shelf

Transport of sediment across the shelf can occur through a range of mechanisms, including currents driven by surface winds or tides, longshore cells, and density currents in which sediment is resuspended by waves or tides (Fig. 2F; Wright and Friedrichs, 2006; Palanques et al., 2009). Large-scale field studies have shown that resuspension of sediment by waves or tides plays an important role in cross shelf transport, which can be more important than plunging river discharge (Traykovski et al., 2000; Wright et al., 2001; Wright and Friedrichs, 2006). Flood sediment can be resuspended by wave action and deposited on the mid shelf, before being remobilised and transported to canyon heads during later storms (Puig et al., 2003; Palanques et al., 2009). Such re-suspension processes are especially important near to river mouths that disperse sediment onto flat shelves, in locations away from steep canyon heads. Gradients on the shelf may be too gentle to allow long run of density flows without assistance from wave or tidal action (Wright and Friedrichs, 2006).

Generation of dense water on the shelf can lead to very low (<0.001 vol.%) sediment concentration flows driven mainly by differences in water density (Wilson and Roberts, 1995; Canals et al., 2006, 2009; Puig et al., 2008; Palanques et al., 2009). Such dense water cascades may be prolonged for days to several weeks, and reach speeds (up to 85 cm/s) that are capable of transporting sand (Canals et al., 2006). They may partly coincide with shorter duration storms, which help to re-suspended sediment in canyon heads. Prolonged dense water cascades can transport up to ~0.005 km³ of sediment off the shelf, but modelling suggests that they will lift off at depths of 500–1000 m unless they entrain sediment (Canals et al., 2009). Monitoring of canyons in the Gulf of Lions shows that dense water cascading can generate very dilute flows that continue and deposit <1 mm of sediment in deep (~2500 m) water beyond the continental slope (Palanques et al., 2009). Most storm or cascade triggered events have shorter run outs and infill canyons with thin and fine deposits. These flows may cut fields of furrows on the canyon floor, but do not appear to ignite and deposit thick layers of sand across submarine fans. Train wheels weighing ~400 kg used for moorings in the Cap de Creus Canyon were episodically moved for up to 9 km over a few days at the start of a dense water cascading event (Puig et al., 2008; Canals et al., 2009). This was either due to the train wheel sliding down-canyon due to drag imparted by the flow on the attached mooring (analogous to kite surfing; Puig and Canals, pers. comm., 2013), or slumping of loose canyon floor sand in a similar fashion to that in Monterey and La Jolla Canyons (Paull et al., 2010a, 2012). Such heavy objects could not be moved as bedload or suspended load in the slow (~1 m/s) flows.

Storm surges may also generate offshore-directed return flows due to relaxation of water previously set up against the coast. Again, it is unclear whether such a process can trigger sediment laden density currents that continue into deeper water. An event on the open continental shelf offshore Hawaii during the passage of Hurricane Iwa in 1982 was inferred to result from slope failure due to wave loading (Dengler et al., 1984). However, this offshore-directed flow may also have been caused by set up due to a storm surge, or sediment re-suspension due to very large wave heights.

Wave or tide re-suspension of sediment on the shelf tends to produce relatively thin (~20 cm), dilute (<1 vol.% sediment), slow moving (<50 cm/s), and fine grained sediment suspensions (Traykovski et al., 2000; Wright and Friedrichs, 2006), which then encounter steeper continental slopes or canyon heads. If the flows do not erode and accelerate, they would be expected to produce the relatively thin and fine grained deposits seen on the shelf (Wheatcroft and Borgeld, 2000). It is unclear whether some flows can ignite, and erode substantial amounts of sediment, to produce powerful flows that deposit sand layers across submarine fans.

2.3.2. Flows triggered within canyons

Oceanographic processes can generate very dilute (<0.001 vol.%) sediment density flows in submarine canyons (de Stiger et al., 2007; Martin et al., 2011). Sediment re-suspended into nepheloid layers by internal waves may become trapped near the canyon head, and then transported down canyon by internal tides and episodic dilute and slow moving turbidity currents that travel at up to 30 cm/s can last for several days (de Stiger et al., 2007; Martin et al., 2011). These turbidity currents typically (but not always) coincide with periods of large wave heights (de Stiger et al., 2007; Martin et al., 2011). These flows tend to produce thick muddy deposits in the upper-to-mid canyon (de Stiger et al., 2007), but do not appear to ignite and flush sediment from the canyon. Even when augmented by dense water cascades in the Gulf of Lions, they only occasionally ran out beyond the continental slope, and on these occasions formed deposits that were ~1 cm thick (Palanques et al., 2009).

Monitoring data suggests that these processes tend to produce dilute and slow moving flows that mainly comprise mud (de Stiger et al., 2007; Martin et al., 2011). This sediment tends to be deposited within the canyon, sometimes accumulating rapidly in the mid-canyon to form finely laminated or massive mud intervals (de Stiger et al., 2007; Martin et al., 2011; Masson et al., 2011).

2.4. Flows triggered by hyperpycnal river discharge

Theoretically, plunging of river freshwater should occur in some marine settings, although it will be somewhat less common than in freshwater settings due to the higher density of seawater. Originally it was thought that sediment loads of >36 m³ to 43 kg/m² (1.3 to 1.7 vol.%) were necessary for freshwater to plunge in marine settings (Mulder and Syvitski, 1995), but convective fingering can cause settling of sediment plumes at much lower (~1 kg/m²; ~0.04 vol.%) suspended sediment concentrations (Parsons et al., 2001; Mulder et al., 2003). Sediment concentrations in a significant number of rivers will reach these values, especially small rivers in mountainous settings, where hyperpycnal discharges may occur one or more times every century (Mulder et al., 2003; Dadson et al., 2005; Warrick et al., 2008). So it is likely that floods will on occasions generate discharges that plunge in the ocean. The question is then whether these plunging discharges can generate long run-out flows that reach the deep ocean, and whether they deposit thick layers of sand across submarine fans.
2.4.1. Evidence for hyperpycnal flows from direct monitoring

Submarine turbidity currents that were most likely triggered by plunging river floodwater have been monitored directly in two locations. Khripounoff et al. (2009, 2012) monitored a series of moderately slow (20–80 cm/s) and dilute (~0.07 vol.%) sediment flows in the Var Canyon, whose head lies next to the Var River (Fig. 3B, C). Hyperpycnal discharges from the Var River are expected to occur every few years (Mulder et al., 2001, 2003). Two of the flows did not coincide with river floods and were most likely generated by canyon margin failures (Fig. 3E). Three flows occurred up to ~36 h after the flood peak, and record failure of recently deposited flood sediment (Fig. 3C; Khripounoff et al., 2012). The remaining four events are coeval with flood peaks and may result from plunging flood discharge, although it is not stated whether they coincided with large wave heights that may have triggered canyon head slope failure (Fig. 3C; Khripounoff et al., 2009, 2012). Only one of these events reached the base of the continental slope, and these flows did not ignite despite the unusually steep gradients (4°–11°) in this canyon. Sediment traps and cores suggest that these flows tend to produce thin (1–10 cm) deposits within the canyon (Khripounoff et al., 2009, 2012; Mas et al., 2010).

Monitoring at the head of Gaoping Canyon offshore Taiwan documented prolonged flows associated with typhoons in 2008 and 2009 (Figs. 2B and 4; Carter et al., 2012; Liu et al., 2012, 2013). The Gaoping River reached sediment concentrations sufficient to plunge during both floods (Carter et al., 2012; Liu et al., 2012, 2013). The 2009 flow was extremely large and transported an estimate 0.06 km³ of sediment with peak water discharge of 27,000 m³/s, whilst the 2008 event transporting ~0.001 km³ of sediment (Table 1). Large (~4–10 m) wave heights were associated with both typhoons, so that wave-resuspension or wave-induced failure of the canyon could have contributed (Fig. 4). The powerful 2009 event transported sand tens of metres above the bed, and had a maximum flow velocity of 1.6 m/s, measured 56 m above the bed in the canyon head (Liu et al., 2012, 2013). The 2008 event failed to break cables lower down the canyon, but the larger 2009 event broke two cables located 110–160 km along the flow path (flow 1 in Fig. 4B; Carter et al., 2012). These breaks indicated an average frontal flow speed of 16.6 m/s, although cables located a further ~60 km along the flow path did not break in flow 1 (Fig. 4B). The 2009 typhoon also led to intrusions of relatively fresh water with very dilute sediment concentrations in the deep ocean (Kao et al., 2010).

Direct monitoring has yet to document a turbidity current generated by hyperpycnal flood discharge that has reached beyond the continental slope, with enough power to transport and deposit sand across submarine fans. A powerful long-run-out flow occurred two days after the 2009 Typhoon Morakot flood discharge had ceased (Fig. 2B and 4; Carter et al., 2012). This delay suggests that the flow was triggered by failure of rapidly deposited sediment in the canyon head, and not plunging river flood water. Powerful long-run-out flows occur in the Zaire Canyon–fan system (Fig. 5; Khripounoff et al., 2003; Vangriesheim et al., 2009; Cooper et al., 2013) but this system is led by a river that is not expected to generate hyperpycnal flood discharges because it drains an extensive low-gradient area (Milliman and Syvitski, 1992; Cynel et al., 2005).

2.4.2. Other lines of evidence for hyperpycnal flows

The other lines of evidence that have been used to infer that a turbidity current was generated by plunging hyperpycnal flood discharge have much greater uncertainties. The first method is to precisely date a turbidite and the associated flood. However, it is often difficult to date deposits with sufficient precise to be sure of the correlation, even for recently emplaced deposits. For instance, it has been inferred that a thick marine deposit in Saguenay Fjord in Canada is a hyperpycnite emplaced in 1663, after a major earthquake triggered a landslide that dammed a river (Syvitski and Schafer, 1996; Mulder et al., 2003). However, the age of this particular turbidite deposit is not known with precision. The turbidite might indeed be coeval with hyperpycnal flood
discharge due a landslide dam, but other processes could also be plausibly invoked, such as direct triggering of the turbidity current by submerged slope failure(s). A second method is based on the hypothesis that turbidites generated by hyperpycnal discharge have distinctive features (Fig. 9). Mulder et al. (2003) suggested that features such as inverse to normal grading, or abundant organic material, are diagnostic of hyperpycnal turbidites (Fig. 9A, B). However, as discussed in more detail in Section 2.4.4, other processes can generate turbidites with such features. Therefore, evidence other than deposit features must be used to demonstrate that a turbidite was triggered by hyperpycnal flood discharge, before deposit features can be used as diagnostic criteria. A third method is to compare the number of historical hyperpycnal floods and turbidites between two dated horizons. This approach was used by Mulder et al. (2001) for deposits in the Var Canyon. This study concluded that the frequency of the inverse-to-normal graded (and hence hyperpycnal) turbidites was similar to that of hyperpycnal river floods. However, relatively small changes in the threshold used to infer hyperpycnal discharge for the Var would lead to substantial changes in the estimated frequency of hyperpycnal floods. For instance, offshore monitoring suggests that hyperpycnal floods of the Var are much more frequent (Khripounoff et al., 2012), and occur at much lower discharges than those inferred by Mulder et al. (2001). So this method also has significant uncertainties.

2.4.3. General difficulties in proving that a submarine flow was triggered by hyperpycnal discharge

It can be difficult to be sure that a submarine flow that coincides with a river flood is indeed triggered by plunging hyperpycnal discharge. First, flood discharge tends to coincide with large wave heights during storms that can suspend sediment and trigger flows in a different way (Wright and Friedrichs, 2006). Although storm waves and flood discharge are decoupled in some locations, these tend to be rivers with larger-area and less-mountainous drainage basins that are less prone to hyperpycnal discharge (Milliman and Syvitski, 1992). Second, failures of the delta lip can also trigger flows during flood peaks (Fig. 6C, E; Hughes Clarke et al., 2012, 2013). Frequently repeated multibeam surveys are needed to show whether it is slope failure that triggers submarine flows during flood peaks (Fig. 6; cf., Hughes Clarke et al., 2012, 2013). This type of surveying cannot be undertaken when there are large waves, such as during typhoons off the Gaoping River (Liu et al., 2012, 2013). It is therefore difficult to distinguish events triggered by plunging hyperpycnal river floods from coeval delta front failures, or failures triggered by large wave heights. This does not mean that submarine hyperpycnal turbidity currents do not exist; it just means that it is problematic to document their occurrence and frequency. Monitoring does show that powerful submarine flows that reach the deep ocean can be associated with river floods, although they were triggered by failure of recently deposited flood sediment and not plunging flood discharge (Fig. 4A–D; Carter et al., 2012).

2.4.4. Deposits from flows triggered by hyperpycnal river discharge

We first summarise information of the deposits of monitored flows that coincide with river floods, before discussing three contrasting models for the character of flows and deposits triggered in this way.

2.4.4.1. Deposits of monitored flows. Hyperpycnal flows have been well studied in lakes and reservoirs. Although few of these studies have cored deposits from a monitored flow, coring has often recovered deposits from recent flow events, presumably of a similar general type. These deposits comprise thin layers that are mm to <10 cm thick, mainly formed of silt and clay (Fig. 9D, E; Lambert and Hsu, 1979; Umeda et al., 2006; Gilbert et al., 2006; Crookshanks and Gilbert, 2008; Talling et al., 2013). However, thin sand layers have been recovered from steeper delta fronts that can be up to a few cm thick (Best et al., 2005; Gilbert et al., 2006).

Flows in the Var Canyon that coincide with hyperpycnal river flood discharge were dilute and slow moving (Fig. 3B, C). Sediment traps recorded deposition rates of ~0.3 mm day⁻¹. These events most likely produced fine-grained deposits that were only a few mm thick, similar to those described in this system by Mas et al. (2010). It is unclear whether the more powerful flows in Gaoping Canyon that coincide with typhoons are triggered by plunging flood discharge, or by the action of large waves (Fig. 4D). These events could sometimes suspended sand at least 42 m above the bed (Liu et al., 2013), and recent deposits in the upper canyon include thick sand layers (Liu et al., 2012, 2013). However, these events failed to break cables beyond the upper canyon, and it is not clear whether they transport large volumes of sand into the lower canyon or deeper water.

2.4.4.2. Model 1: inverse-to-normally graded model of Mulder et al. (2003). Mulder et al. (2003) proposed that turbidites from events triggered by plunging hyperpycnal river discharge tend to record the rising and falling limbs of flood hydrographs in the form of inverse to normally grading (Fig. 9A). This model implies that faster flow equates to coarser deposition, and also suggests that an erosional surface may form at peak flood discharge (Fig. 9B). This type of grading pattern occurs within turbidites in locations such as the Var Canyon (Migeon et al., 2001; Mulder et al., 2001), and may be a result of varying flood discharge, but inverse to normal grading can also form in other ways. For instance, inverse grading may result from increases in flow concentration rather than flow speed (Kuenen and Sengupta, 1970; and see bed geometries in Talling et al., 2007b). Direct observations (Best et al., 2005; Khripounoff et al., 2012; Liu et al., 2012, 2013) and numerical modelling (Lamb et al., 2010) show that a single flood peak can produce turbidity currents with multiple pulses, which would form more complex grading patterns (Fig. 3C). Mulder et al. (2003) also inferred that abundant terrestrial organic matter was diagnostic of hyperpycnite turbidites. However, failure of deltaic sediment rich in terrestrial organic material could also produce flows with this composition.

2.4.4.3. Model 2: sandy deposits with traction structures from prolonged flows. Plink-Björklund and Steel (2004) proposed that plunging hyperpycnal floods produce sustained turbidity currents that deposited thick (up to several m) sand beds, which have abundant traction structures, and variable grading patterns or no grading (Fig. 9C). This model was based on well exposed outcrops in Spitsbergen where delta-front deposits can be traced laterally into basin floor turbidites. They presented several lines of evidence that these deposits were hyperpycnal turbidites, although none appears absolutely conclusive. This modelling implies that thick deposits tend to record prolonged flows, but slope failure can also produce sandy deposits that are ~1 m thick, and turbidite mud can pond to produce even thicker deposits (Talling et al., 2012a). Model 2 suggests that metre-thick planar laminated (T₈) and massive (T₆) sandstone intervals record prolonged flow over days to weeks, although these types of deposit can also form rapidly during a few minutes in laboratory experiments (Sumner et al., 2008; Talling et al., 2012a). As noted for Model 1, deposits rich in terrestrial organic remains can be produced by delta-front failure, which is also consistent with lateral connection between the deltaic and turbidite systems. The abrupt pinchout of hyperpycnal turbidite beds was attributed to looting of hyperpycnal flows, but abrupt pinoutsts of the same lithofacies can be produced by sandy debris flows generated by slope failure (Talling et al., 2012b). A low abundance of contorted slump deposits in the Spitsbergen outcrops favours triggering by hyperpycnal flood discharge. However, failure of prograding delta lips, or movement of crescentic bedforms of the type seen on the Squamish delta-front (Hughes Clarke et al., 2012, 2013), may not produce slump deposits that are ultimately preserved in the geological record. This is because the delta chutes in which the slumps occur can be flushed by later flows, and thus may be areas of net erosion over long time periods.

2.4.4.4. Model 3: fine-grained and thin (mm to a few cm) deposits. A third model is proposed and favoured here; that turbidity currents generated
by plunging hyperpycnal flood water tend to be slow moving (<50 cm/s) and dilute (<0.01 vol.% sediment), as seen in lakes and reservoirs, and also tend to produce thin (mm to ~2–10 cm) and fine grained deposits (Fig. 9D, E). Submarine hyperpycnal turbidites thus resemble hyperpycnal turbidites seen in lakes and reservoirs (Fig. 9E; Lambert and Hsu, 1979; Umeda et al., 2006; Gilbert et al., 2006; Crookshanks and Gilbert, 2008). They may include fine sand, sometimes with laminations, but are mainly mud, and can be rich in terrestrial organic material. Their limited thickness results from dilute sediment concentrations and slow aggradation rates, despite their prolonged duration. These deposits may display inverse to normal grading due to a current that promotes sediment failure, which generated a powerful, long runout turbidity current in 1979 (Fig. 3A; Piper and Savoye, 1993; Trofimovs et al., 2006, 2008, 2013). More dilute and even faster moving pyroclastic surges can also enter the ocean (Sparks et al., 2002), but their deposits have yet to be sampled in detail.

2.5. Glacial outwash megafloods and dam bursts

Large volumes of sediment may be carried into the ocean by jökulhlaups associated with glacier outburst flooding, producing offshore sediment plumes with hyperpycnal concentrations. A jökulhlaup from Vatnajökull, Iceland in 1996 transported ~0.07 km³ of sediment offshore during ~42 h (Maria et al., 2000). The submarine continuations of jökulhlaups are not well monitored, and it is unclear whether they generate long runout (~100 km) flows. Far larger glacial outwash floods have occurred in the past, and most likely played a role in global climatic change (Wright and Friedrichs, 2006; Warrick et al., 2008), so that hyperpycnal flood deposits are often later reworked, unless the flows run out below the wave base.

A second part to the model is that these slow hyperpycnal turbidity currents tend not to erode substantially, and do not incorporate large volumes of sand. Available direct monitoring observations suggest that hyperpycnal flows are slow (typically <0.3 m/s) moving (Talling et al., 2013) and hence not strongly erosive. Therefore, this model proposes that hyperpycnal flows usually flush large amounts of sand into deeper water, nor deposit metre-thick sand layers across submarine fans. All three models need to be tested more rigorously in the future by suitable field-scale observations.

2.7. Volcanic eruptions

Deep-seated collapse of volcanic island flanks produces the largest mass wasting events on our planet (up to ~5000 km³; Moore et al., 1989; Masson et al., 2006). They can generate turbidity currents that contain up to several hundred km³ of sediment and travel for hundreds of kilometres, sometimes up to 2000 km (Hunt et al., 2011). The turbidites leave behind produces key insights into collapse dynamics, including whether collapse occurred in a single stage or multiple stages. The way in which landslide material enters the ocean determines tsunami magnitude. A better understanding collapse dynamics is thus key for predicting landslide–tsunami hazards (Wynn and Masson, 2003; Hunt et al., 2011; Watt et al., 2012).

Shallow volcanic dome collapses, or collapse of eruption columns, can generate hot pyroclastic flows that enter the ocean. In many cases most of the eruption products end offshore from volcanic islands (Le Friant et al., 2009, 2010). They can generate dense submarine granular flows from which fines are elutriated efficiently, most likely by generation of steam by the hot flow. This elutriated material can form longer runout turbidity currents (Trofimovs et al., 2006, 2008, 2013). More dilute and even faster moving pyroclastic surges can also enter the ocean (Sparks et al., 2002), but their deposits have yet to be sampled in detail.

2.8. Anthropogenic triggers: mine tailings discharge, trawling and land reclamation

Direct monitoring has shown that discharge of mine tailing can generate dilute and slow moving turbidity currents, both in lakes and fjords (Hay et al., 1982; Hay, 1987; Normark, 1989). Trawling can generate dilute, slow moving turbidity currents that infill canyons (Puig et al., 2012). Extension of the Nice airport runway established conditions that promoted sediment failure, which generated a powerful, long runout turbidity current in 1979 (Fig. 3A; Piper and Savoye, 1993; Mulder et al., 1997).

3. Generalised models for different types of system: flow types and frequency

Generalised models are now presented that aim to capture the triggers, flow types and flow frequencies that occur in a series of different settings. Such models may prove more useful than a list of global triggers in some situations, for instance in assessing geohazards from flows at a particular field site. However, the models outlined here are based to a large extent on direct monitoring datasets (Talling et al., 2013), and there are relatively few locations where such monitor has been completed. Therefore, this is far from an exhaustive list of settings, and these models may be refined and added to in future contributions, especially as direct monitoring data becomes available for a wider range of locations. The classification adopted here is based on a combination of how sediment is mainly supplied (e.g. rivers or oceanographic processes) and system morphology (e.g. whether a large canyon is present).

Defining models inevitably place fixed boundaries within what is a continuum of real world situations. For instance, the proximal part of...
Fig. 10. Generalised models summarising flow types and triggers, runout distances, and flow frequencies in five different settings. See Fig. 2G for a full key to symbols used. (A – Model 1) Basin plains fed mainly by large open continental slope landslides. (B – Model 2) Submarine canyon–fan systems fed mainly by oceanographic processes (e.g. wave or tide resuspension of sediment — see Section 1.3). Canyon heads can either be close to shore, where they can intercept littoral cells (upper panel), or be located near the shelf edge (lower panel). (C – Model 3) River fed submarine canyon–fan systems, where the canyon head is located close to a major river mouth. (D – Model 4) Submarine delta systems associated with bedload dominated rivers, whose suspended sediment concentrations are too low to form frequent hyperpycnal flows. (E – Model 5) Freshwater lakes and reservoirs that are fed mainly by rivers that generate frequent hyperpycnal flows.
the present day Nile Fan represents an oceanographic canyon–fan system (type 2), whilst the Herodotus Abyssal Plain that lies at its termination more closely resembles a basin plain fed by large open slope slides (type 1). There may also be a continuum of marine river-fed systems from smaller delta-front systems and large canyon–fan systems (Talling et al., 2013).

### 3.1. Type 1 — basin plains fed by large open slope slides

The Seine, Madeira and Agadir Basin Plains offshore from the NW African margin, the Herodotus and Balearic Abyssal Plains in the Mediterranean, and the Sohm Abyssal Plain offshore Newfoundland are fed by relatively infrequent but very large volume flows, primarily generated by failures on the continental slope, which can sometimes involve canyon margins (Fig. 10E). Flows extended for hundreds of km across these basin plains, and had volumes in excess of 10–100 km$^3$ (Piper et al., 1999; Talling et al., 2007a). They occurred on average every ~1350 to 20,000 years (Hunt et al., 2013; Clare et al., 2014). In locations such as the Herodotus Abyssal Plain beyond the end of the Nile fan, there is a transition from distal basin plains fed by large landslides to channel–levee complexes formed by river-fed systems (type 3; Section 3.3). Locations on the continental slope can also be mainly fed by flows generated by slope failure further upslope (Poudertoux et al., 2012).

The distribution of recurrence intervals of flows in several different basin plains approximates a Poisson distribution (Clare et al., 2014). This Poisson distribution suggests that landslide occurrence is temporally random, with the time to the next event being independent of the time since the last event. A Poisson distribution also suggests that there is at most a weak control on flow frequency by sea level (Clare et al., 2014). Analysis of a global compilation of large (>1 km$^3$) landslide ages in the last 30 ka, suggests that there is little or no correlation with sea level, although this could simply result from large uncertainties in the dating of almost all landslides (Urlaub et al., 2013a). However, there may be a weak control on slide frequency by sea level in settings fed by major river systems, reflecting large increases in sedimentation rate that generate elevated excess pore pressures and increase the likelihood of failure (Urlaub et al., 2013b).

### 3.2. Type 2 — submarine canyon–fan systems fed by oceanographic processes

Sediment is supplied to these systems primarily from the shelf due to wave or tidal re-suspension of sediment at the canyon head or in cross-shelf flows, or longshore currents (Fig. 10D). Multiple events occur multiple times each year, mainly during periods with large wave heights (Fig. 7; Xu et al., 2004; Paull et al., 2010a; Paull et al., 2012). Such events infill the canyon and do not reach deeper water (Paull et al., 2011b). More dilute flows are also generated within the canyon by re-suspension of fine grained sediment by large waves or internal tides, and these flows infill the canyon (Martin et al., 2011). Canyons are flushed occasionally by very powerful events once every ~100 to ~1000 years caused by failure of canyon margins. Such failures could possibly be triggered by earthquakes but this has not been demonstrated unequivocally. The frequency of events in this type of system fed by oceanographic processes is strongly bimodal, with canyon filling and flushing events.

Examples of modern canyons fed by oceanographic processes include the Monterey, La Jolla, Hueneme, Mugu, Nazare, and Eel Canyons. They include canyon systems that have become disconnected from major river mouths, such as the Amazon, Rhone and Indus Canyons. Well developed submarine channels that extend from canyon mouths may be relics from periods in which the canyon was connected to the river mouth. In some cases, mainly where the canyon has never been connected to a major river mouth, there is no channel system beyond the canyon mouth. Changes in sea level will only affect strongly systems fed by oceanographic processes if it leads to reconnection with river mouths, otherwise flow frequency is mainly determined by the frequency of major storms and sediment supply.

### 3.3. Type 3 — submarine canyon–fan systems fed directly by large rivers

These systems are fed by large rivers that comprise canyon–channel systems extending into deep water, which build extensive submarine fans. They are very active when the canyon head is adjacent to the river mouth. Examples include the canyon–channel systems offshore from the Zaire, Magdalena, Bengal, and Gaoping Rivers (Figs. 2B, 4 and 5; Khrilpounoff et al., 2003; Heezen et al., 1964; Heezen and Hollister, 1971; Weber et al., 1997; Hsu et al., 2008; Carter et al., 2012; Cooper et al., 2013). Cable breaks suggest that powerful long run-out flows occur in these systems on average every 1 to 5 years (Heezen et al., 1964; Heezen and Hollister, 1971; Hsu et al., 2008; Carter et al., 2012). ADCP moorings in the Zaire canyon, located 150 km offshore in water depths of 2 km, recorded 11 events with speeds of up to 2.5 m/s during just 7 months (Carter et al., 2013). The 11 flows were prolonged with durations of 1–6 days, but their speeds did not wax and wane (Fig. 9A, B) in the manner attributed to hyperpycnal turbidity currents by the model of Mulder et al. (2003; Fig. 5). Peak flow velocities were observed only 1–3 min after arrival of the flow at the ADCP mooring (Fig. 5D; Cooper et al., 2013).

Powerful events in these river-fed canyon systems mainly occur during seasons of elevated river discharge, but they may not coincide with individual flood peaks (Figs. 4C and 8B, E). Flood discharges from rivers with large low-gradient drainage basins, such as the Zaire, are also unlikely to reach hyperpycnal sediment concentrations (Milliman and Switski, 1992; Cognel et al., 2005). This suggests that some of these flows are triggered by failure of recently deposited sediment in canyon heads (Heezen et al., 1964; Carter et al., 2012), but they may be triggered by other processes at any time of year. These canyon head failures may resemble those seen on delta-fronts in smaller river fed systems, which also tend to occur during periods of elevated river discharge, but they have longer run-out distances due to deeper water depths and longer flow paths. These fast moving flows flush sediment from canyons, transport it through channels, and produce deposits across distal submarine fans (Fig. 10C). It is inferred here that events triggered by plunging hyperpycnal river discharge, or resuspension of sediment in the canyon by waves or tides, are relatively weak and infill the canyon mainly with fine sediment. Their relative importance and frequency depends on the river’s suspended sediment concentration, and the frequency of large wave heights.

The frequency of flows is strongly affected by whether the river mouth remains connected to the canyon head. If the river mouth and canyon head become disconnected, for instance due to rapid flooding of a wide continental shelf, then the system is fed by oceanographic processes that sweep sediment off the shelf (Fig. 10D). This results in a change to much more infrequent flows, of the type described in Section 3.3. Examples of systems that have become disconnected during recent rapid sea level rise, and are now canyons fed mainly by oceanographic processes, include the Amazon, Rhone, and Indus fans (Piper and Deptuck, 1997; Prins and Postma, 2000; Ducassou et al., 2008; Dennielou et al., 2009). Connection between canyon head and river mouth depends not only on sea level, but also the rate of sediment supply and shelf morphology, and hence the time taken for the delta to prograde to the shelf edge (Burgess and Hovius, 1998; Carvajal et al., 2009). During ‘greenhouse’ periods with small magnitude fluctuations in sea level these systems are less likely to be disconnected due to rapid sea level rise. When river and canyon head are connected, the frequency of flows is determined by the changes in sediment supply from rivers. This means that the most frequent flows may occur during deglaciation rather than lowstands (Covault and Graham, 2010). The frequency of flows that
reach the distal fan is much greater in canyon–fan systems connect-
ed directly to river mouths, than in systems fed by oceanographic
processes or large open slope landslides. Flow recurrence intervals
and run out distances are not strongly bimodal (cf. Arzola et al.,
2008), and are not well described by a bimodal distribution of
canyon-filling and canyon-flushing events.

3.4. Type 4 – submarine delta-front systems fed by bedload dominated
rivers

Unusually well monitored examples of this type of setting include
the Squamish River, Fraser River, and Bute and Knight Inlet systems in
British Columbia, Canada (Prior et al., 1987; Bornhold et al., 1994; Hill,
2012; Hughes Clarke et al., 2012, 2013). This type of system lacks deeply
incised canyons, and has delta-front chutes that are less than a few tens
of metres deep (Fig. 10B). These chutes extend to feed channel systems
of variable length, sometimes up to 30 km in the case of Bute Inlet
(Conway et al., 2012). A characteristic feature of these systems is that
associated rivers have a substantial bedload component, and do not
reach the high suspended sediment concentrations needed for funneling
hyperpycnal flow. In some cases these delta-fronts occur at the head
of glacially-deepened fjords, although the Fraser River Delta is an
exception.

Tens of events occur each year in this type of system, almost exclu-
sively during periods of elevated river discharge (Figs. 6E and 8A; B;
Prior et al., 1987; Bornhold et al., 1994; Hughes Clarke et al., 2012,
2013). These events can be split into three types (Table 1). The first
type are relatively infrequent (5 to 7 events per year) and large
(−0.001 km³) failures of the delta lip that occur up to a few times each
year, which generate flows that rework sediment through the channel
system (Fig. 6C, E). Particularly large failures may only occur every
few years (Hill, 2012). Their frequency depends on sediment type,
progradation rate, and frequency of factors that trigger failure (river
flow surges, low tides or other factors that destabilise loosely packed
sediment). The second type of event is much more frequent with tens
of events occurring each year, and typically associated with motion of
crescentic bedforms (Fig. 6D, E). These events have shorter runout and
are less powerful than the delta lip failures. A third type of event is
generated by plunging hyperpycnal floodwater, whose occurrence and
frequency depends on whether the river can have elevated sediment
concentrations and its hydrograph. These events most likely deposit
thin and fine grained deposits, with similar deposits formed by surface
hypopycnal plumes, or plumes that intrude within the water body
(Marti et al., 2011). Hyperpycnal flood discharges occur less frequently
than in lakes and reservoirs because seawater is denser than freshwater.
However, hyperpycnal flood conditions characterise a significant num-
ber of oceanic river mouths, with frequencies of once every 1 to
100 years (Mulder et al., 2003).

Flow frequency is strongly linked to the rate at which sediment is
supplied by the river, especially bedload transport that drives delta-
lip progradation. This ensures that climatic or other factors that
cause variations in fluvial sediment supply have a strong influence of
the frequency of flows (it is ‘reactive’; Allen, 2008). Flow frequency
also depends on triggers for failure, such as individual river flood
peaks, but over longer periods this is less important than changes in
progradation rate. This type of system remains active during all
stages of sea level.

3.5. Type 5 – freshwater lakes and reservoirs fed mainly by river discharge

Direct monitoring shows that many lakes and reservoirs are charac-
terised by relatively frequent turbidity currents produced by
hyperpycnal discharge from river floods (Fig. 10A; Table 2). In some
locations, tens of flow events occur each year (Fig. 8D; Lambert and
Giovanoli, 1988), and some flows can be prolonged for periods of
up to several weeks (Crookshanks and Gilbert, 2008), with flow
frequency and duration determined by variations in river water dis-
charge and sediment concentration (Mulder et al., 2003). More power-
ful but infrequent flows in lakes can be triggered by slope failure
(Fig. 8D; Lambert and Giovanoli, 1988), which can produce thicker
and coarser-grained deposits that include debrites (Girardclos et al.,
2007).

Most lakes and reservoirs lack well developed channels, presum-
ably because they are dominated by weak flows generated by
hydropycnal river discharge, which are unable to erode the underly-
ing substrate. It is inferred that powerful events triggered by delta-
front slope failure are implicated in the formation of well developed
channel systems (Fig. 8C, D). This inference is based on two lines of
evidence. The first is occurrence of well developed channels on the
floor of Lake Geneva (Fig. 8C), where Lambert and Giovanoli (1988)
provide the only direct monitoring observation of a powerful
slope-failure triggered turbidity current in a lake (Fig. 8D). The sec-
ond is by analogy to well monitored river-fed submarine systems,
such as the Squamish River-delta, where delta-front slope failure is
associated with well developed channel systems (Fig. 6A). However,
flows generated by plunging flood discharges can potentially play a
role in sculpting channels (Forel, 1885), although their relative im-
portance is poorly constrained. Such erosion could either be due to
erosion by the submarine hyperpycnal flows themselves, or because
hyperpycnal flows deposit sediment that then fails. Lakes or reser-
voirs with channel systems may broadly resemble submarine sys-
tems fed by small rivers, such as those described in Section 3.4.

4. Tempo of sediment transport by submarine flows

Submarine flows and rivers are arguably the two volumetrically
most important processes for moving sediment across our planet. A
general comparison can be made between the volumes of sediment
transported, and the frequency–magnitude (tempo) of transport events
in these two types of system (Table 1).

4.1. Tempo and magnitude of sediment flux due to submarine sediment
density flows

It is apparent that the largest submarine events sometimes
transport far greater amounts of sediment than any event on land
(Table 1; Korup, 2012; Talling et al., 2013). The largest submarine
landslides contain >3000 km³ of material, which is ~500 times the
average annual sediment flux (~6 km³) from all of the world’s rivers
measured in the last few decades (Table 3; Milliman and Syvitski,
1992). Basin plain turbidites can contain >100 km³ of sediment
(Talling et al., 2007a). The largest landslides on land are far smaller
(Table 1), such that the Grand Banks submarine slide in 1929 in-
volved ten times the amount of material in the five largest subaerial
landslides in the last century (Piper et al., 1999; Korup, 2012). How-
ever, individual submarine flows transport a wide range of sedi-
ment volumes, and more frequent types of sediment flow do not
involve such extreme sediment volumes. The volume of sediment
transported by single floods of individual rivers can reach ~0.06 km³
(Table 3; Dadson et al., 2005), which is broadly similar to the sedi-
ment volume involved in large canyon head failures (~0.01 km³; Piper
and Savoye, 1993).

The tempo of subaqueous sediment density flows and rivers can be
compared using magnitude–frequency relationships, such as those cap-
turing the fraction of time taken to discharge a certain percentage of ei-
ther water or sediment. It is important to distinguish clearly between
tempos of sediment and water discharge, which can be very different
due to variable sediment concentration. For instance, Dadson et al.
(2005) reported that up to 80% of the water in some Taiwanese rivers
is transported during ~40% of the year, whilst 80% of the total sediment
transport occurs in ~1% of the time. It is apparent that discharge of water
within subaqueous sediment density flows is generally more episodic
than water discharge by rivers, as shown by Figs. 3B–E, 4C, 6E, and 8B, E that compare subaqueous flow timing and river discharge. However, there is considerable variability in the tempo of water discharge by rivers, with much steadier discharge from large rivers (e.g. Amazon or Congo; Milliman and Syvitski, 1992; Coyne et al., 2005) than from smaller rivers draining mountain belts in typhoon dominated (e.g. Taiwan; Davidson et al., 2005) or semi-arid settings (e.g. Southern California; Warrick and Milliman, 2003). The tempo of water discharge in subaqueous density flows may be even more variable, as average recurrence times of these flows can vary from days (Fig. 6E) to tens of thousands of years (Table 1; Hunt et al., 2013) depending on triggers and setting.

Sediment transport by submarine flows is often more episodic than by rivers. For instance, unusually large volume transport events reach some basin plains typically every few hundreds to thousands of years (Table 1; Hunt et al., 2013; Clare et al., 2014). Similarly, submarine flow events episodically flush sediment from canyons fed by oceanographic processes every ~300 to ~3000 years (Table 1). Conservatively assuming that each submarine flow lasts for ~1 day, this means that sediment transport occurs during just 0.001% to 0.0001% of the time. Sediment transport by some types of river can be episodic, but it is less episodic than for basin plain or canyon flushing flows. For instance, over 75% of suspended sediment discharge occurs in ~1% of the time in some Taiwanese rivers (Kao and Milliman, 2008), and ~50% of sediment discharge from Southern California rivers can occur in ~0.1% of the time (Warrick and Milliman, 2003). Future work may be able to make more detailed comparisons than those attempted here between tempo of water and sediment discharge in subaqueous density flows and rivers (cf. Fig. 6E).

A second question that can be posed is how efficient is the linkage between rivers and submarine systems, in terms of how long sediment is stored on the intervening shelf? This linkage can be very efficient where canyon heads are located next to large river deltas. For instance, around 80% of the sediment flux from the Gaoping River is flushed down the adjacent canyon each year, as measured over the last few decades (Hu et al., 2009). In other settings, especially where canyons and rivers have become disconnected, much of the sediment remains trapped on the shelf. This is linked to whether submarine systems react quickly to temporal changes in terrestrial sediment flux and climate, or whether there are large delays (Covault and Graham, 2010). These two types of situations have been termed reactive or buffered systems (Allen, 2008; Romans et al., 2009; Covault et al., 2010, 2013). Reactive systems tend to provide the best record of terrestrial climate change within submarine flow deposits (Romans et al., 2009; Covault et al., 2010, 2013). Oceanographic canyon–fan systems that are disconnected from rivers may tend to be buffered, although they may provide a record of storm-triggered events or major earthquakes. Systems in which river mouths are connected directly to canyon heads or delta-fronts are more likely to be reactive (Romans et al., 2009; Covault et al., 2010). Flow type may not react in the same way to external forcing as flow frequency. Romans et al. (2009) showed how the thickness of submarine flow deposits and their sand fraction in the Santa Monica Basin responded to external forcing in the last ~7 ka, whilst flow frequency showed little change over the same period.

4.2. Implications for the efficiency of organic carbon burial in submarine systems

The frequency of sediment transport has important implications for the efficiency of both terrestrial and organic carbon burial in submarine settings, and hence for the global carbon cycle. Large river systems can transport large amounts of terrestrial organic carbon to their deltas (Galy et al., 2007). For instance, the amount of terrestrial organic carbon transported by the Ganges–Brahmaputra system is greater than the amount of carbon drawn down from the atmosphere by silicate weathering in the Himalaya. Frequent submarine flows favour efficient burial of carbon in the deep ocean due to high accumulation rates on submarine fans. This is the situation for highly active canyon–channel systems fed by major rivers, such as the Bengal Fan (Galy et al., 2007), or the Nazare Canyon (Masson et al., 2010). There is relatively little oxidation and loss of terrestrial or marine organic carbon in these settings (Galy et al., 2007), although flushing of the Nazare Canyon may subsequently lead to oxidation and loss of organic carbon. However, burial of organic carbon is much less efficient in systems characterised by large but infrequent (~1 ka to ~10 ka) submarine flows. Slow long-term sedimentation rates favour oxidation of more labile types of organic carbon in the upper part of the turbidite mud cap. For instance, ~40% of the organic carbon in turbidite mud caps in the Agadir Basin Plain has been oxidised and lost. This is due to the slow sediment accumulation rate and long recurrence interval between flows that allows oxidation and loss of carbon from the bed tops. The efficiency of organic carbon burial is therefore strongly dependent on flow frequency, which is highly variable in different submarine settings.

5. Conclusions

5.1. Triggers and the types of flow that result

Monitoring studies that precisely constrain flow timing provide the most unambiguous information on how flows are triggered, although other types of information must be used to understand the most infrequent or powerful events. Processes that initiate flows are varied, and they generate flows of different types (Figs. 2 and 10; Table 2). Triggers can be grouped into different types of slope failure, oceanographic processes, and plungering of hyperpycnal river discharge (Piper and Normark, 2009).

Very large (~100 km3) failures of the continental slope and far smaller (0.008 km3) failures in canyon heads, can generate fast moving (>3.5 m/s) flows that travel for hundreds of kilometres in subaqueous flow from canyon and channel systems (Piper and Savoye, 1993; Mulder et al., 1997; Piper et al., 1999; Hsu et al., 2008). Canyon head failures sometimes involve recently and rapidly deposited sediment, and occur only a few days after a major flood (Fig. 4C; Carter et al., 2012). Prograding delta fronts can experience many tens of transport events per year, almost exclusively during periods of elevated river discharge (Figs. 6E, 8B; Prior et al., 1987; Hill, 2012; Clarke et al., 2012, 2013). Failure of the prograding delta-lip can occur several times each year. Lip failures coincide with surges in river discharge, which cause rapid lip progradation or increased shear of the bed, or low tides when failure may be favoured by expansion of gas (Fig. 6E). A second type of event is much more frequent and results in the upslope migration of crescentic bedforms (Fig. 7D, E; Hill, 2012; Clarke et al., 2012, 2013). Many tens of these events occur each year, and they are often not associated with surges in river discharge. Similar crescentic bedforms occur in submarine canyons, where bedform motion typically coincides with large wave heights (Fig. 5; Paull et al., 2010a, 2010b), and in lacustrine delta front channels (Fig. 8C; Girardclos et al., 2012). Such bedforms may be a widespread result of supercritical flows (Cartigny, 2010), and/or liquefaction of loosely packed sand that is disturbed by a variety of triggers (Paull et al., 2010a, 2012; Talling et al., 2013).

Sediment can be transported off the shelf by wave-modified or tide-modified flows that spill into submarine canyons or onto the open continental slope (Fig. 2f; Wright and Fredrichs, 2006; Palanques et al., 2009). Cascades of dense shelf water can also transport sediment off the shelf (Canals et al., 2006, 2009), and the action of large waves or internal tides can resuspend sediment within canyons (de Stiger et al., 2007; Martin et al., 2011). These ‘oceanographic’ processes tend to form dilute and slow moving flows that infill canyons, and these weak flows may not transport large volumes of sediment onto submarine fans.
Plunging river floods generate dilute and slow moving turbidity currents in lakes and reservoirs that are frequent on annual or shorter time scales (Lambert and Giovannoni, 1988; Crookshanks and Gilbert, 2008; Talling et al., 2013). A significant fraction of the sediment supplied to the ocean by some rivers occurs at hyperpycnal sediment concentrations, especially for small mountainous rivers (Milliman and Syvitski, 1992; Mulder et al., 2003; Dadson et al., 2005; Warrick et al., 2008). However, plunging flood discharges tend not to travel for long distances on the continental shelf, and re-suspension by storms or tides is needed to rework the sediment in stages to the shelf edge (Wright and Friedrichs, 2006). Dilute and slow moving submarine flows that reach the base of the continental slope (Fig. 3; Khripounoff et al., 2009, 2012), or more powerful events in canyon heads (Liu et al., 2012, 2013), have been associated with hyperpycnal river floods. However, these flows may also have coincided with large wave heights, and it is unclear whether they resulted only from plunging hyperpycnal freshwater.

5.2. Deposits

Weak and dilute flows generated by plunging hyperpycnal flood discharges most likely deposits thin (mm to ~10 cm) and fine grained sediment layers, similar to those documented for hyperpycnal flows in lakes and reservoirs (Fig. 8D, E). The available field observations suggest that they do not form metre-thick sand layers in deep water settings, as has been previously proposed (Mulder et al., 2003). A single flood peak may generate flow with multiple pulses (Fig. 3C; Best et al., 2005; Lamb et al., 2010; Khripounoff et al., 2012; Liu et al., 2012, 2013), so that deposit grading patterns may be more complex than simple inverse-to-normal grading. Failure of recently deposited flood sediment may generate flows with abundant organic matter, which is therefore not strongly diagnostic of hyperpycnal flows.

Slope failure may sometimes generate prolonged flows, and their deposits may differ significantly from the sequence of Bouma (1962). It is particularly important to be able to identify the deposits of submarine flows associated with slope failures that are triggered by major earthquakes. Such flow deposits potentially provide a long term chronology of major earthquakes for hazard assessment. The most reliable criteria are that the turbidite results from widespread slope failure(s) associated with extensive earthquake shaking. Precise dating is needed to establish synchronous turbidite emplacement in adjacent basins, whilst clastence tests are complicated by variable flow thickness and the potential for a dominant tributary (Atwater and Griggs, 2012). Mapping and cores of the sea floor in the vicinity of large recent earthquakes suggest that not all major earthquakes generate extensive turbidites, and the record of major earthquakes is incomplete in some locations (Völker et al., 2011; Sumner et al., 2013). Patterns of grading within turbidites are very unlikely to record the temporal pattern of earthquake shaking (seismograms), and irregular grading patterns can be formed in other ways (Hunt et al., 2011). Internal characteristics within turbidites are therefore not strongly diagnostic of earthquake triggering.

5.3. Flow frequency, and long term evolution of systems in different settings

Submarine flow frequency depends on initiation mechanism and distance from where the flows are triggered. The long term evolution of transport systems depends on changes in flow triggers and frequency, such as those due to changes in sea level and fluvial sediment fluxes. The frequency of flows in lakes and reservoirs depends on flow patterns, and tens of events can occur during each flood season (Fig. 8C; D; Lambert and Giovannoni, 1988; Talling et al., 2013). Submarine canyons fed by rivers may experience powerful flows that reach submarine fans every few years (Heezen et al., 1964; Heezen and Hollister, 1971; Hsu et al., 2008), or even more frequently (Cooper et al., 2013). There is a strong connection between the fluvial and submarine transport systems, although this may be due to failure of sediment in the canyon head rather than plunging flood discharge (Fig. 4D; Carter et al., 2012). Oceanographic processes tend to infill canyons, with much more infrequent events every century or millennium flushing this sediment into deep water. This concept of canyon filling and flushing is much less applicable to river-fed canyon systems (Fig. 10C, D). Rising sea level can cause a loss of connection between river mouths and canyon heads, producing a change from a river-fed canyon to a canyon fed by oceanographic processes. Activity in river-fed canyons is determined strongly by changes in fluvial sediment supply driven by climatic changes in the river drainage basin, such that the greatest activity may occur during either low-stand or rising sea level (Covault and Graham, 2010). Oceanographic fed systems can have activity that continues during high-stands in sea level (Covault and Graham, 2010). Deep-water basin plains are often fed by infrequent but very large flows generated by landslides on the continental slope (Talling et al., 2007a,b; Clare et al., 2014). The frequency of these flows can be almost random in time, such that the time since the last flow does not influence the probability of a flow occurring. There appears to be only weak (or no) dependence between the frequency of these large slides and changes in sea level (Urlaub et al., 2013a; Clare et al., 2014).

Together with river systems, submarine sediment density flows dominate sediment transport fluxes across Earth. The tempo of sediment transport in basin plain systems tends to be more episodic than fluvial systems. The largest submarine flows in these settings can transport over ten times the average annual sediment flux from all of the world’s rivers, as measured over the past few decades (Milliman and Syvitski, 1992). However, in river-fed submarine canyons the sediment fluxes appear more continuous, and more closely mimic those in the river system. It is hoped that this contribution stimulate novel future efforts to document the triggers and frequency of submarine sediment flows, and their highly variable character.

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