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Geohazards and Ocean Hazards in Deepwater: Overview and Methods of Assessment

Dorrik Stow, Rachel Brackenridge, Urval Patel, Suzannah Toulmin
Institute of Petroleum Engineering, Heriot-Watt University, Edinburgh EH14 4AS, UK

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Abstract

Drilling for hydrocarbons in the deep marine environment provides a unique set of challenges for industry. Amongst these are the distinct hazards caused by natural geological and oceanic processes such as: (a) semi-permanent bottom currents, (b) episodic turbidity currents, slope instability and mass transport events (slides, slumps, debris flows), and (c) gas hydrate escape. We present data on the nature, effects and assessment of these deepwater hazards, including current velocities, transport/erosion capacities, recurrence intervals, and hydrate distribution. It is of upmost importance that the oceanographic conditions are carefully considered prior to deepwater operations to ensure the work can be carried out safely. Thorough risk assessment requires knowledge of existing bottom currents, an assessment of potential mass transport events and turbidity currents, and understanding of conditions likely to induce the formation and destabilisation of gas hydrates. Pipelines, cables, subsea installations, key connections such as the riser and any other seabed infrastructure are all susceptible to damage. Correct assessment of hazard will allow for the right equipment to be used for the operations including blowout preventers, riser size, vibration suppressors, and for the safest siting of cables and pipelines. It will also aid the subsea architecture design and planning operations with minimal downtime. All these considerations lead to a safer exploration and production while eliminating unnecessary cost.

Introduction

Deepwater is defined by different communities in slightly different ways. For the marine geologist and oceanographer, it is often taken as that area of the ocean beyond the shelf break, i.e. deeper than about 100-200 m water depth. For the sedimentologist, it is below storm wave base, which is also around 100-200 m. For the industry, deepwater exploration is considered to be that in water depths in excess of 500 m. For the purposes of this paper, we consider the shelf-slope transition (i.e. shelf break) as the key boundary between shallow and deep water.

This deepwater environment is far from tranquil. It is subject to a range of processes that will significantly affect drilling operations for hydrocarbon exploration and recovery. Bottom currents are everywhere present and in some areas they are especially active and with considerably elevated velocities. Here we first focus on deepwater bottom currents and the hazards they present. Their occurrence and distribution, nature and variability, and damage potential are discussed, followed by consideration of the methods of hazard assessment. We present a new method using the bedform-velocity matrix.

The second topic, mass transport events and turbidity currents, is discussed in a similar way. These processes occur on all types of continental margins and slopes, and are of great importance, both scientific and industrial. They are the dominant processes through which large amounts of sediments are transferred across the continental slope to the deep ocean. Enhanced knowledge of these processes and their depositional features is not only important for continued success of oil exploration in deep-water settings, but also for the potential impact on human life and settlements (i.e. tsunamis), and protection of offshore infrastructure (e.g. platforms, pipeline, telecommunication cables etc).

The third topic covered is that of gas hydrates in ocean sediments. These are nearly ubiquitous in near-surface sediments on continental margins, as a result of high pressures generated by the overlying water column. They too pose a serious hazard to deepwater petroleum operations.

Bottom Current Hazards

Occurrence and distribution

Currents are present throughout the world's oceans and seas. There has, to date, been no recording of any oceanic water mass that is completely calm with no movement (Zhenzhong and Eriksson, 1998). They occur at the surface (usually wind-driven) and throughout the entire water column (Gross and Gross 1996). Where wind is the driving force, intensity decreases significantly with depth. It is therefore unlikely that wind-driven currents will penetrate past the top few 100 metres of water except in some exceptional circumstances, for example, major storm events (Brooks, 1984). Little was known about deep ocean currents until just a few decades ago. Despite being first identified over seventy years ago (Wust, 1936), it was not until the 1960s that technological advances allowed for focused work on bottom and deep ocean currents to begin (e.g. Heezen and Hollister, 1963; Heezen et al., 1966). Direct current measurements, seabed photography, seabed and subsurface sampling and acoustic methods have been used to quickly advance our understanding of the hydrodynamics of the deep ocean (Shanmugam et al., 1995; Zenk, 2008).

It is now widely accepted that bottom currents are formed by various means: (1) wind-driven; (2) tidal-driven; (3) thermohaline-driven; (4) internal wave-driven (Stow et al., 2002; Shanmugam, 2006). Due to this range of current types, operations in all water depths can be affected. In deepwater, most bottom currents are generated as part of the thermohaline conveyor belt (Broecker, 1991; Rahmstorf, 2006), and form in the cold-water kitchens of Polar regions (Brackenridge et al., 2011). The global thermohaline circulation system is driven by intense cooling of highly saline surface waters in the Arctic and Antarctic regions. This leads to sinking of dense water masses and ventilation of the deep ocean. Subsequent mixing with overlying water masses and heating allows upwelling, and so the process is maintained. Two key ingredients are required to create sufficiently high water density to allow sinking: salinity and temperature. As a result, there are a limited number of source areas for bottom water generation; namely the Norwegian-Greenland Sea, Labrador Sea, Bering Sea, Weddell Sea, Ross Sea and other locations around Antarctica (Rahmstorf, 2006).

Warm-water kitchens, in which warm saline bottom water is generated by intense evaporation in low-latitude regions are relatively rare in today's oceans. The principal source is the Eastern Mediterranean Sea, with escape to the global ocean via the Gibraltar Gateway. Locally generation also occurs in the Red Sea and Persian Gulf, but escape to the ocean is prevented by shallow sills.

At depth it is also possible to encounter internal tides and waves. These commonly occur at the interfaces between water masses of different density and have the capacity to mix and transport large volumes of water and sediment (Zhenzhong and Eriksson, 1998). They are broadly comparable to surface waves in that they have highly variable amplitudes and wavelengths depending on the density gradient and other conditions.

Accelerated bottom currents are generated experienced along many of the continental margins under investigation by the hydrocarbon industry (Fig 1). These water masses form 'rivers' of highly vigorous water in the deep ocean along continental boundaries were Coriolis Forces and seabed morphology act to further increase velocity. In the Atlantic Ocean, deep oceanic currents affect many petroleum basins. The continental slope offshore Brazil is influenced by the Brazil Current, the South Atlantic Central Water, the Antarctic Intermediate Water, the North Atlantic Deep Water and the Antarctic Bottom Water (Viana et al., 2002). These water masses all contact with the continental slope at different depths and have various velocities ranging from sluggish to 40cm/s (Farrant and Javed, 2001). Some are northward flowing boundary currents, whereas others flow in a southward direction. The Gulf of Mexico is subject to highs of 70cm/s along the course of the Loop Current, and is generally seen to be around 50cm/s (Brooks, 1984). The current speeds of the North Atlantic Deep Water will pose additional challenges in the frontier regions offshore Greenland where large bottom-water sediment deposits have been identified (Hunter et al., 2008; Nielsen et al., 2011) as well as offshore the NW European margin, West of Shetland (Fig 1).

Where these currents reach sufficient velocities, they can erode, pick up, transport and deposit sediments. Great accumulations of sediments are named contourite drifts, and these can be covered in bedforms and erosional features (Stow et al. 2002; Rebesco and Camerlenghi 2008). Most commonly, any depositional or erosional feature influenced by bottom water currents will be orientated alongslope and can extend for many hundreds of kilometres. Where higher velocities are observed, regionally-extensive erosional surfaces or terraces cut into the continental slope are found (Hernández-Molina et al., 2008). Sediments range from fine muds to sands, and in some regions, gravelly accumulations can be found. Most contourite drifts are composed of mixed siliciclastic/biogenic material and may be closely associated with other deposits such as turbidite or glaciogenic material.

Nature and variability

Unlike episodic turbidity currents or continuous pelagic settling processes, contourites are deposited by a continuous to semi-continuous processes. They are semi-continuous since flow will wax and wane on numerous timescales and at numerous magnitudes, but the flow will be maintained over a geological timescale.

The key features of bottom currents have recently been compiled by Stow et al. (2008). Bottom currents (also known as contour currents) generally move in an alongslope direction, following the contours of the continental margin. They are capable of moving upslope and downslope, for example to move around morphological obstructions or sea floor features. The dominant direction will however, be parallel to the continental margin. Where they are present over low gradient slopes and abyssal plains, the bottom water will form a broad sluggish core moving at <10 cm s^{-1}. The water mass is be distinguished from other masses on account of temperature and salinity data, as well as its relative velocity. It is the sea floor morphology itself that is the chief control on the route of the bottom current core, its velocity and complexity (Zenk, 2008). Higher gradient slopes, such as the continental rise and slope, or some morphological obstruction such as a seamount, or submarine canyon can elevate the velocity of a bottom current. Where any water mass is constricted, its velocity will increase. Coriolis forces can further elevate velocity by pushing the water mass against the feature and further constricting it. Features such as subsea channels or gateways provide conditions for greatest increase in water velocity.

The velocity and nature of a current is highly variable over space and time. On examination of contourite deposits, changes in sedimentary structures, grain size and other properties represent different current intensities across the drift (e.g. Stow et al., 1986). Over space, currents decelerate over distance down current, until some morphological feature allows for further acceleration. Direct measurements of water velocities show highs of a few metres per second along the path of some water masses where they are constricted and accelerated (Gonthier et al., 1984).

The current velocity will also fluctuate over time, on various scales. Sedimentological work shows large scale (million year) changes, glacial-interglacial cycles on 100kyr scale and a smaller scale fluctuation on a 1000 yr timescale (Stow et al., 1986; Knutz, 2008). Direct measurements of bottom currents show further fluctuations of intensity on a seasonal and even daily cycle (Zhenzhong and Eriksson, 1998; Stow at al 2012). As a result, these currents are highly complex, difficult to study and challenging to predict.

Damage potential

Ocean currents provide additional risks in regions of petroleum exploration and production; Surface currents cause additional strains on automatic positioning equipment on drilling vessels and oil tankers. Intermediate currents cause huge amounts of additional strain on equipment connecting the rig with the seabed, and deep currents can affect sub-sea infrastructure. Here we examine the risks associated with vigorous bottom currents specifically. These pose a potentially great risk to deepwater operations including drilling, production and development (Chow et al., 2006). Key risks identified are: (1) triggering of downslope mass transport processes; (2) pipe-line walking and fatigue; (3) additional stress to risers and tethers.

Bottom water currents and mass-wasting. Both bottom currents and their resulting deposits (contourites) are capable of promoting large-scale slope instability (see also below). There are numerous examples where contourite and mass transport deposits co-exist in the geological record (e.g. Bryn et al., 2005), and it is now clear that contourite deposits can promote large-scale slope instability through numerous factors (Laberg and Camerlenghi 2008). Firstly contourites show characteristically high sedimentation rates when compared to many other deepwater processes (Stow et al., 2002a). This can lead to high fluid content and therefore high pore pressures and low slope stability. Secondly, the nature of contourite deposition results in moderately to well-sorted sediments being deposited. These are significantly weaker than their poorer sorted counterparts of higher internal shear strength. Finally, contourite accumulation can result is the formation of mounded drifts. These directly modify the shape of the continental margin, and locally increase the slope gradient which can lead to failure. Additional slope stability issues arise when vigorous currents come in contact with the seabed. Internal waves can promote slope instability. Where failure does occur along a contourite-dominated margin, the resulting deposit is impressively widespread. The majority of contourite-associated mass wasting events observed in the ancient record are considerably larger than the events experienced in recordable history.

Pipeline walking and fatigue. In shallow water operations, pipelines are laid in trenches to protect them from movement, however in deep water, infrastructure lies directly on the seabed in poorly consolidated sediments with extremely high clay and water content. This, combined with bottom water currents can result in movement of the pipeline and addition stress on connectors with tie-ins and risers (Carr et al., 2008). When combined with other hazards exerted on the equipment such as thermocycling of the pipeline (Rong et al., 2009; Whooley, A., pers. comm., 2011) and corrosion (McKinnel, 2011), premature failure is possible. Careful modeling and planning of pipeline routes is a highly complex procedure, requiring many variables to be considered, and the avoidance of localized regions of accelerated bottom water is important to avoid unnecessary strain on equipment. .

Stress to risers and tethers. Fatigue of risers, tethers and drill-strings is common in regions of constant ocean currents and can drastically increase production costs. Risers act as obstructions to flow and cause vortexes to form in the water column. This amplifies the strain on equipment – causing vibrations in the connections and increasing wear of the mechanism. Analysis on vortex-induced vibrations (VIV) has identified the cause to be due to water currents and wave actions. The main point of wear is close to the top of the riser or close to the touch down point, where bottom water currents are in operation (IntecSea, n.d.). Additional strain on connectors is common in deepwater regions swept by vigorous water currents due to increased riser curvature (Howells, 2000) and corrosion (McKinnel, 2011).

These are the main issues that arise from vigorous bottom currents and the ones that pose the greatest safety, environmental and economic cost. An additional problem is also re-entry procedures in deep water (Bullock et al., 1979). Such risks can result in premature fatigue of subsea infrastructure and increase the time and costs of operations. A detailed understanding of currents can allow engineers to design subsea structures to withstand necessary current conditions, without added costs. Current velocity data is essential for design of structures such as pipelines, manifolds and risers and can be particularly challenging in frontier areas swept by bottom water currents. This will be a future challenge in the West Shetland Region where bottom water velocities may exceed 1m/s (Kuijpers et al., 2002) and where 500 tonnes of subsea structures will be awarded for the Clair field development in 2012 alone (Sheehan, 2011).

Hazard Assessment

Currents are challenging to study and analyse. They can be directly identified on the basis of temperature, salinity and velocity anomalies in the water column. Subsea floats and tracers are widely used to examine the dominant routes of bottom waters (Richardson et al., 2000; Zenk et al., 2008) and velocities (Smith and Jacobs, 2005). From such measurements, the nature and characteristics of deepwater currents have been assessed. They are generally contained within an elevated velocity water 'core' that creates a boundary current, from which eddies and loop currents may peel off and may rejoin the main water core.

Current velocity measurements may be taken directly through the water column, using moored current metres and shipboard acoustic Doppler current profilers, but nature and variability of the current route and intensity make direct measurements unreliable. A study by Smith and Jacobs (2005) considered over 50,000 current measurements from various devices to produce a more reliable and realistic current circulation pattern in the Gulf of Mexico. Despite added boundary constraints and data smoothing, large errors were observed close to significant topographic features such as canyons and the shelf-break in addition to water mass boundaries. For hydrocarbon infrastructure to be safe and reliable for the duration of operations, the mean and maximum possible current intensity at the site of drilling is required. Discrete current velocity measurements should be recorded from moored tools in place for a number of months (e.g. Hamilton, 1990). These long timescale recordings are required to identify trends such as seasonal variations. Direct measurements on this magnitude would be very expensive and time-consuming to survey prior to operations beginning. Alternative methods of assessing current intensity over an area are possible, based on the knowledge of contourite depositional and erosional features.

Geotechnical companies can provide extensive acoustic datasets of the seabed in order to assess the regional geohazards prior to drilling operations commencing. Multibeam, side scan and sub-bottom profiler methods are used from ship-board or AUV (Autonomous Underwater Vehicle) instruments (Chow, 2006). When combined with seismic acquired at an exploratory stage, many potential hazards can be mapped such as irregular topography, faulting, gas hydrates and sand facies mapping. However, these methods have yet to be used to their full potential as they do not fully assess bottom current hazards. Here we outline the main ways in which the industry can utilise this data to further assess the geohazards relating to oceanic currents.

Damuth (1980) uses high frequency echograms of the sea floor to assess the seabed facies and distinguish between the different near-bottom processes. Contourite depositional and erosional features typically respond differently from downslope sediments such as turbidites and debrites. Migrating sediment waves and erosional furrows cause disrupted and hyperbolic acoustic responses at high frequency (Damuth, 1975). Although features such as sediment waves are present in some turbidite-dominated regions, acoustic character can be combined with existing knowledge of bottom water currents to positively identify regions most affected by currents. Another way in which acoustic data can be used to positively identify regions influenced by vigorous currents is to identify contourite features based regional trends of orientation and geometry. Seismic data can be used to identify contourite drifts on the basis on external and internal geometries, and the presence of laterally-extensive erosional discontinuities (Faugeres et al., 1999; Nielsen et al., 2008).

Bedform-Velocity matrix

Although the above methods are extremely useful in identifying bottom currents and acquiring a regional understanding of the hydrodynamic regime of the margin, they do not yield quantitative data on mean current velocity. Here we present a novel way of extracting this information using existing geophysical methods applied to the latest contourite research.

Heezen and Hollister (1971) first realised the succession of bedforms developing under ever-increasing bottom water velocities in deep water. Now, detailed sedimentological analyses of regions influenced by bottom water currents have led to the compilation of a bedform-velocity matrix (Stow et al 2009). From this, it is possible to estimate mean bottom water velocity from the sediment grain size and/or from bedforms on the sea floor (Fig 2). High-energy environments have the transport capacity for larger grain sizes and therefore will result in coarser grained deposits (Tucker, 1991; Reading and Levell 1996). Other considerations should be taken into account, such as grain density and distance from sediment source or sources, but on a whole, as the velocity of a bottom water mass decreases, so too does its transport capacity. The matrix is, in essence, very similar to other bed form-velocity matrix constructed for other depositional processes. It examines likely dominant grain sizes in addition to the expected bed forms on the sea bed formed by currents.

As noted above, bottom currents tend to show variability in flow speed over many different timescales. The bedform present on the seabed is, therefore, a record of the mean or dominant flow characteristics over the timescale of deposition. The grain size velocity matrix could therefore provide a robust means of identifying bottom water core routes and average velocities. The matrix can be used with high frequency, high resolution acoustic surveying methods which allow the seabed to be mapped in great detail. Bedforms can be identified using methods such as side-scan sonar (Kenyon and Belderson 1973), and images from Autonomous Underwater Vehicles (Chow, 2006). Bottom photographs can be used to gain yet more knowledge of the bedforms and features actively forming on the sea floor. If utilised to its full potential, the bedform-velocity matrix could provide a cost-effective method gaining detailed regional information on the risks posed by bottom water currents.

Mass Transport Events and Turbidity Currents

Nature and Variability

Slides and Slumps. Submarine slides and slumps involve the movement of coherent masses of sediments bounded on all sides by distinct failure planes (Mulder and Cochonat, 1996). The internal structure of the moving mass is largely undisturbed as most of the shear is localised along a basal failure surface. However, if the moving mass is unconsolidated, and depending on the strength and heterogeneity of the material, it may undergo complex internal deformation as it moves downslope. They range in size and volume from a few m^3 to several hundred to thousands of km^3 and are known to have run out distances in excess of 200km. Differentiation of slides and slumps is based on the values of the Skempton ratio *h/l*, where *h* is the depth of the slip surface and *l* is the length of failure: slides are translational with Skempton ratios of <0.15, whereas slumps are rotational and deep rooted with a h/l ratio between 0.15 and 0.33 (Skempton and Hutchinson, 1969).

Most submarine slides appear to be translational and are characterised by a fairly flat, slope-parallel basal failure surfaces (slope gradients <2°). Such low angle failure surfaces have been imaged on 2D seismic studies of the Storegga and Traenadjupet slides on the mid-Norwegian margin and the Afen slide from the Faeroe-Shetland channel (Canals et al., 2004). For these failures, the sliding surface is predetermined and normally corresponds to a discrete layer with low shear resistance, such as permeable sand layers or clay and sand interbeds (hemipelagic, pelagic and contouritic deposits) (Mulder, 2010).

However, with recent developments in high-resolution three-dimensional seismic surveying, significant topographic variations have been shown to occur along the basal slip surface (Gee et al., 2005). Such subsurface studies agree with well known outcrop examples, and suggest stepped, striated, concave or a combination of these slide surface topography can be present (Gee et al., 2005; Lee and Stow, 2007). As with translational slides, the slide surface is predetermined and relates to an interval of bedding-parallel weak layer. The topography is created as a result of the slide surface propagating upwards due to decreasing downslope driving stresses or resistive forces, such as sediment shear strength (Lee and Stow, 2007; Martinsen, 1994). These topographic features may initiate the development of internal deformation and dispersive pressure in the sliding mass, which could lead to the morphological transformation of the slide into a debris flow or turbidity current, as in the Gaviota Slide in the Santa Barba Channel (Greene et al., 2006), the Gondola Slide on the Southwestern Adriatic Margin (Minisini et al., 2006) and the 1929 Grand Banks event (Piper et al., 1999).

Slides are commonly associated with a variety of extensional and compressional features (e.g. normal, reverse and strike-slip faults etc). Extensional features (i.e. normal and listric faults) are most common in the upslope parts of the slide, especially in the headwall area (Martinsen and Bakken, 1990), where they are predominantly orientated perpendicular to the transport direction. In complex slides, where the motion of the initial mass transport event lead to instability of neighbouring area, such extensional features can generate large volume, linear troughs that control local sediment pathways until the relief is filled. Such features are common in shallow deltaic environments (e.g. Clark Fjord, Baffin Island) (Mulder and Cochonat, 1996; Syvitski and Farrow, 1989). These normal faults at the headwall scarp of slide gradually sole out at the level of the basal décollement (Stow et al., 1996). Downslope, at the front (often referred to as the 'toe') of the slide deposit, compression is dominant due to frictional freezing of the mass transport deposit, and is often characterised by imbricate thrust slices of chaotically deformed or coherently folded strata. The toe region is also known to have elevated topography created by the resistance to downslope movement. In addition to normal and reverse faults, strike-slip faulting may develop perpendicular to the maximum stress direction to accommodate the differential movement within the mass of sediments. Such features are difficult to recognise in outcrops and subsurface data, but may help explain longitudinal shear ridges observed on modern slide surfaces using side-scan sonar images (Hampton et al., 1996; Lee and Stow, 2007).

Submarine slumps exhibit many features of slides and are gradational with them (Stow et al., 1996). However, unlike slides, the basal failure plane of slumps is commonly concave upwards and the motion of the displaced material is rotational (Hampton et al., 1996). The exact reasons for the differing movement styles is still poorly understand, and as such, various authors have suggested that the poorly consolidated nature of the sediments is the controlling factor, while others have emphasised the importance of mechanical mixing due to the interaction of the moving mass with slide surface topography (Dykstra, 2005; Hampton et al., 1996; Mulder and Cochonat, 1996). This latter reasoning is borne out of the simple observation of more disturbed and chaotic features present in distal parts of a mass transport deposit. The internal structure of slumps is often chaotic and highly deformed, and the degree and style of deformation varies with position of the moving layer and the strength and heterogeneity of the material (Stow et al., 1996). Since slumps commonly involve plastic deformation, they will cease lateral motion once the applied shear falls below a critical value.

Slides and slumps are not isolated processes and often form complex structures with multiple phases of failures. The most common are multiple phases of retrogressive failures that form because of upslope propagation of the failure (Mulder, 2010; Mulder and Cochonat, 1996). Other less frequent complex slope failures are overlapping, additive and successive slumps and slides, where the term 'overlapped' is applied if the failure surface of the main body is merged with successive events, or 'additive' if the failure surfaces of induced events are not merged. Successive slides and slumps are generated when the initial failure of a mass of sediments triggers mobility in an underlying second material mass. Such slides have been inferred in the area around the Titanic wreck (Mulder, 2010).

As mentioned briefly above, slides and slumps evolve downslope into debris flows and turbidity currents through gradually increasing ambient water mixing and entrainment, and disintegration of coherent blocks. However, the exact nature and cause of this transformation is still poorly understood as landslides can travel hundreds kilometres without transformation into plastic flows or turbidity currents, while others transform close to the source. In reality slides and slumps are complex events, and elements of slides, slumps, debris flows and turbidity currents may all be present after a mass transport event (Masson et al., 2006).

Plastic Flows. Plastic flows are a type of mass transport events in which sediment and water are fully mixed such that no internal stratification is preserved. They are common in continental slope settings and are represented by debris flows, of which there are two types: cohesive flows (cohesive debris flows) and frictional flows (non-cohesive or cohesionless debris flows) (Dasgupta, 2003). Both cohesive and non-cohesive debris flows are characterised by finite yield strength due to either the cohesive strength of the material provided by high clay content (cohesive debris flows) or frictional strength due to interlocking of grains. The cohesive strength of the material imparts a pseudoplastic behaviour and an overall laminar state to the moving mass (Mulder and Alexander, 2001; Shanmugum, 1996). Debris flows keep on moving until the shear forces exerted by the downslope component of gravity falls below a critical value and the flow freezes *en-masse*. However, experimental and field observations suggest that deposition from debris flows can also occur from incremental aggradation of flow surges due to internal shearing throughout the moving mass. Consolidation of individual flow surges is not possible before the arrival of the next surge and thus the resulting deposit will be amalgamated (Dasgupta, 2003). Debris flows are known to move on low slope gradients, travel rapidly (speeds of several meters per second observed from subaerial debris flows) over long distances and are only very slightly erosional; characteristics which could be explained by the presence of a water layer beneath the moving mass (hydroplaning) that reduces the resistance due to drag on the seafloor (Mulder and Alexander, 2001; Pickering et al., 1989; Piper et al., 1999). Transformation of a debris flow to a more turbulent and dilute flow (i.e. turbidity currents) takes places if dispersive pressure and buoyant lift comes into existence as clast-support mechanisms.

Cohesive debris flows usually consists of fine-grained matrix with a significant amount of clay content, although movement as debris flows has been shown to occur in finest sand sized material where bulk clay content is as low as 2% (Mulder and Alexander, 2001; Stow et al., 1996). In addition, the high clay content is also responsible for the low rate of dilution of the flow by either deposition or entrainment of ambient fluid, thus the flow remain coherent for longer. The clay-water mixture constitutes the fluid phase in cohesive debris flows and provides the main clast-support mechanism for coarse grains and clasts during flow conditions. In contrast, in cohesionless debris flows the rheological pseudoplastic behaviour is due to the very high grain/water ratio. Truly cohesionless debris flows are not normally expected in nature as small amounts of clay impurities remain in the system. These types of debris flows develop primarily in well-sorted sand and gravel and occur on very steep slopes. The internal structure of both cohesive and non-cohesive flows is typically chaotic and disorganised, although reverse grading and coarse tail grading of larger clasts is known to occur.

Turbidity Currents. Turbidity currents are one of the most important ways by which fine, medium and coarse-grained material is transferred from shallow to deep water. They are turbulent suspensions of mud and sand in water, which are propelled by the downslope component of gravity acting on the excess density. They may occur as short-lived surge events that travel for only a matter of kilometres downslope, or go through a process of flow ignition such that an autosuspension process is generated in the flow. This permits very long distance transport over several thousands of km, both downslope and across flat abyssal plains. They can even travel a certain distance in an upslope direction before they come to a halt by a combination of frictional resistance, loss of sediment from the base of the flow and reverse gravitation pull.

Individual turbidity currents are discrete events with very variable recurrence intervals (10^0 - 10^5 y) and of very different sizes. The largest flows are known to overtop channel margins of 850 m in height. These are likely to be several kilometres in width and probably more in length. Much smaller turbidity currents also occur. Such currents can be channel confined or flow across open slopes with little apparent confinement. They can deposit beds from < 1 cm to > 10 m thick. Mean accumulation rates, therefore, are also very variable, typically from 10 cm to > 1 m/ky. The frequency of occurrence of turbidity currents ranges from 1/1000 years (approximately) for the distal Bengal fan, to one every few years for parts of the Amazon and Congo fan systems.

Occurrence and Distribution

Submarine mass transport events and turbidity currents are extremely widespread in both ancient and modern sedimentary environments, particular where fine-grained sediments predominate (Stow et al., 1996; Masson et al 2006). Hampton et al. (1996) used the term 'landslide territory' to designate the environments where they are most common, these include; open continental slopes, submarine canyon-fan systems, fjords, active river deltas, flanks of volcanic islands and along convergent margins. The largest known submarine mass movements have been known to occur on open continental slopes and on the flanks of volcanic islands. This could be in response to the unique morphological and geological conditions of these areas (Masson et al., 2006). Open continental slopes are often characterised by low slope gradients (<2°) and parallel-bedded, homogeneous, fine-grained sediments deposited over large areas (Bryn et al., 2005; Hampton et al., 1996). If mechanical inhomogeneities (i.e. parallel-bedding) control slope failures on open slope settings, it could potentially affect a significant area and produce a large translational slide (e.g. Storegga Slide on the mid-Norwegian Margin). Oceanic islands on the other hand are some of the steepest topographic features on the planet and volcanic processes tend to build, steepen, and overload these slopes with time, inducing slope failure (e.g. Canary and Hawaiian Islands) (Masson et al., 2002). The exact cause for failure on low angle (<2°) submarine slopes is still poorly understood, although excess pore water pressure is considered to be a major factor.

The factors that control the long term stability of subaqueous slope and therefore the occurrence of mass transport events and turbidity currents within these environments range from the obvious and short-lived processes, such as earthquakes, to those that are less obvious and operate on timescales of tens to thousands of years, such as climate change. In many cases the cause of the mass transport event is obvious, while in others the triggering mechanism of these and of turbidity currents can be cryptic. By considering the 'landslide territory' of Hampton et al. (1996), a number of temporally varying factors that control slope stability have been identified. These include: (1) the quantity, type and rate of sediment delivered to the continental margin, (2) sediment thickness of the depocentre, (3) changes in seafloor pressure and temperature, which can influence hydrate stability and the generation of free gas, (4) variations in seismicity, and (5) changes in groundwater flow conditions within the slope and shelf (Lee, 2009; Tappin, 2010). However, both Lee (2009) and Twichell et al. (2009) have argued that these controls may be secondary to the main driving force, namely climate change.

We suggest that it is exactly these factors that also influence the occurrence of turbidity currents, as many are derived directly from mass transport events. Other factors more specific to turbidity current generation include: (1) sudden excess sediment supply by rivers in flood (i.e. their generation from hyperpycnal flows); (2) rapid glacial discharge events; (3) re-suspension of shelf edge to upper slope sediment as a result of storm stirring and the incidence of internal tides/waves; and (4) storm build up of water across a continental shelf and its rapid discharge down submarine canyons. Taken together, these two sets of factors are considered as 'turbidity current territory'.

Damage potential

The hazards associated with submarine mass transport events are numerous and will depend on the scale, type and location of the movement. One of the more obvious consequences of submarine mass movements is the generation of tsunamis that may have devastating consequences for coastal areas. The Holocene Storegga Slide, one of the largest and well studied mass movements off the Norwegian continental margin, removed up to 2500-3500km^3 of sediments, affected an area of 90,000km^2 and had a run-out distance of about 800km (Bryn et al., 2005). Although the exact cause of the slide is still debated, both Kvalstad et al. (2005) and Bryn et al. (2005) have suggested that rapid sediment loading during interglacial periods resulted in the development of excess pore pressure and reduction of the shear strength in contourite marine clays and oozes. The slide was most likely triggered by a large earthquake with the initial failure occurring within the over-pressured marine clay and oozes.

The tsunami associated with this event caused widespread inundation in Norway, Scotland, Faroe Islands and NW Iceland (Smith et al., 2004). The run-up of the tsunami in the Shetlands exceeded 25m, while on the mainland it exceeded 5m locally. This compares with run-up heights between 2m and 30m for the Indian Ocean tsunami on December 26th 2004 that killed over 280,000 people, and destroyed coastal settlements and infrastructure across the Indian Ocean Basin (Synolakis and Kong, 2006). Another example tsunami hazards associated with mass failure occurred on April 1st, 1946 when a M 7.3 earthquake struck offshore the Aleutian Islands causing major slumping in the Alaskan Trench. The resultant tsunami had a run up of 30m along the Scotch gap area, killed 69 people and caused $25 million in infrastructure damage.

Tsunamis generated by landslides on this magnitude are not restricted to open continental shelves as evidenced by giant landslide scars on the flanks of volcanic islands such as Hawaiian-Emperor ridge, the Canary Islands, Reunion, the submerged flanks of Mt Etna and Stromboli and the Marquesas Island (Hampton et al., 1996; Masson et al., 2006; Masson et al., 2002). Along the Hawaiian-Emperor Ridge alone there are 68 major slumps and debris avalanches over 20km in length, with some over 200km long and volume in excess of 5000km^3 (Moore et al., 1989). On the Canary Island, 14 major slumps and debris avalanches, and genetically-related turbidites and debris flows, have been associated with the islands of El Hierro, La Palma, and Tenerife over the last one million years. On average, the typical submarine mass transport event on the Canary Island involves a volume of 50-200km^3, covers an area of a few thousand km^2, and has a run-out distance of 50-100km (Masson et al., 2006). Although research into landslide generated tsunamis is still in its infancy, many of the slides mentioned above may have had the potential to generate devastating tsunamis that could have affected distant coastal areas.

In the context of offshore oil and gas infrastructure, submarine hazards due to mass transport events and turbidity currents have the potential to cause significant loss of life or damage to the environment and field installations (Kvalstad, 2007). The triggering mechanisms for such events are primarily controlled by geological and climatic processes (see above) or by human activity. Offshore infrastructure at risk of damage by submarine mass transport events may include, but not limited to, exploration drilling rigs, piles, conductors, caissons, pipelines, flow lines, cables, manifolds, well heads and gravity base structures. The potential impacts of mass transport events on these structures are dependent on the scale, type, location and orientation of the movement. The structures can be subjected to loading, burial or erosion by gravity flows (i.e. debris flows and turbidity currents) or down-drag, uplift, rotation and translation by mass slides and slumps (Thomas et al., 2010b). For instance, seafloor erosion and sediment scouring due to the passage of a turbidity current or debris flow may result in the loss of support around foundations or beneath pipelines, while a rotational or translational failure will result in down-drag near the headwall, or uplift near the toe for the same structure.

Examples of damage to offshore infrastructure as a result of mass movement processes are numerous. Perhaps the most famous is the 1929 Grand Banks event on the continental slope south of the island of St Pierre. A series of small, thin skinned, regressive slumps where initiated as a result of a M 7.2 earthquake located 250km south of Newfoundland, in about 2000m water depth (Piper et al., 1999). These slumps underwent a series of surface and body transformations into debris flows and turbidity currents, and travelled at speeds up to $30ms^{-1}$ and deposited >$150km^3$ of sediments on to the Sohm abyssal plain (Mulder, 2010; Piper et al., 1999). The mass movement generated a tsunami that struck the southern end of Newfoundland's Burin Peninsula, killing 29 people and causing millions of dollars of damage. Perhaps more interestingly, the resultant turbidity current broke twelve transatlantic cables downslope, which allowed the speed of the current to be measured for the first time.

The 1979 Nice slide also generate a debris flow and turbidity current downslope that broke several telecommunication cables. The initial slide involved mud from the prodelta slope and occurred in response to the landfilling operations for the Nice international airport extension, but regressed onto land, removing part of the runway along with construction equipment and killing several people working on site (Mulder et al., 1997). In 1998, cable damaging submarine slides and turbidity currents were generated when a powerful earthquake struck offshore Papua New Guinea, killing over 2000 people (Masson et al., 2006). The resultant submarine slides and turbidity current travelled at least 280km and in water depths greater than 6000m. Other, more recent examples of cable breaks associated with submarine mass movements include the 2003 M 6.8 'Boumerdes' earthquake onshore Algeria that generated submarine landslides and turbidity currents, which subsequently damaged 6 cables and disrupted all submarine network in the Mediterranean region. And between 2003 and 2010, multiple cables breaks were reported as a result of earthquake triggered submarine mass movements offshore Taiwan.

One of the few published examples of submarine geohazards damaging oil and gas installations comes from wave and wind induced submarine slides and slumps during Hurricane Camille, which stuck the U.S Gulf of Mexico in August 1969 (Hampton et al., 1996). As a result of these mass movements, three oilfield platforms on the Mississippi River delta were displaced and damaged. Post storm surveys demonstrated that hurricane generated 20m high waves induced stresses on the seabed that were strong enough to cause foundation-disrupting (translation and uplift) slides in water depths greater than 100m (Bea et al., 1983).

In recent years, the passage of hurricanes Andrew, Katrina and Rita along the Gulf of Mexico has also damaged hundreds of pipelines when strong waves and currents trigger landslides and slumps in the weak under-consolidated clays issued from the Mississippi river. Presently, many offshore oil and gas prospects are located where slope instability can be a major geohazard. Examples of such prospects include the Atlantis and Mad Dog fields along the Sigsbee Escarpment in the Gulf of Mexico, numerous fields in the West Nile Delta offshore Egypt, and the Ormen Lange gas field on the Norwegian continental slope (Storegga headscar).

The prediction, assessment and/or mitigation of geohazards in offshore settings are therefore key to the continued successes of exploration and exploitation of hydrocarbons, particularly in deepwater areas.

Hazard Assessment

From the offshore oil and gas perspective, the Society for Underwater Technology estimates that the cost of damage to submarine structures caused by mass failure is more than $400 million annually. The preferred industry method of dealing with submarine geohazards is to identify and avoid them. However, for many locations their large extent may make them difficult to avoid, as is the case with the Storegga slide offshore Norway which overlies the Ormen Lange gas field, or the Gulf of Mexico where wave and current induced slides and slumps are frequent (Clayton and Power, 2002). In terms of risk assessment, it is essential to answer the following questions: (1) where did past and where will future mass movements and turbidity currents occur; (2) how frequently do they occur; (3) what are the triggering mechanisms; (4) what is the area of influence; (5) what is the nature and magnitude of the impact; and (6) can previous failures be re-activated (Locat and Lee, 2002; Thomas et al., 2010b). While not a detailed or comprehensive study, this section will outline some of the key tools used in the assessment of submarine geohazards.

For the determination and assessment of submarine geohazards a multidisciplinary approach is required that involves the acquisition of geophysical, geological and geotechnical data. However, one of the first steps is to undertake a comprehensive desk-based study incorporating all existing data (including site specific seismic data) and to compile a regional geohazard register or inventory to establish a 3D regional ground model (Bryn et al., 2007; Clayton and Power, 2002; Thomas et al., 2010b). The regional ground model allows attention to be placed on relevant geohazards, detailing their geomorphological, bathymetrical and geological character, while eliminating unrealistic geohazards, thus reducing the risk to an offshore development. The ground model progressively evolves from a desk-based conceptual model, to a geological/geotechnical model, and finally into a development specific engineering model as new data from the project specific site investigation becomes available. Key data considered for input into the 3D regional ground model are itemised below.

Seafloor Mapping and Seismic Data. To establish seabed topography and geomorphology, bathymetric data need to be acquired. The preferred tools used for bathymetric surveys are "multibeam" or "swath" echo sounder systems. Multibeam echo sounder systems are mounted to the ship's hull and collect depth data by transmitting acoustic signals in sixteen or more beams arranged in a fan pattern. They can be operated in water depths greater than 1500m, however the accuracy and resolution will decrease with increasing water depth due to beam spreading and increased foot print (Kvalstad, 2007). The use of long tethers in deeper water surveys is less productive and costly; therefore a recent development has been in the use of Autonomous Underwater Vehicles (AUVs) to provide high resolution bathymetric. AUV can now operate in water depths up to 4500m and provide invaluable information in the evaluation of geohazard risk assessment and site investigation, as shown by recent studies in the West Nile Delta and the Ormen Lange gas field within the Storegga Slide area (Bryn et al., 2007; Thomas et al., 2010a). Complimentary to the multibeam echo sounder survey is the use of side-scan sonar (SSS). Side-scan sonar, either tethered or mounted on an AUV for deepwater investigation, can provide information about seafloor morphology and reflectivity (backscatter) (Kvalstad, 2007). In this case, the variations in the reflectivity can be used to map out targets such as pock marks, seabed hard grounds, gas seeps and other geohazards, as well as sediment types.

In conjunction with seafloor surveys, it is also necessary to acquire 2D and 3D high and ultra high-resolution seismic data during development specific investigations. These techniques allow the identification of previous slide activity in the area, including mapping of features such as head scarps, side walls, topographic features on basal surfaces, extensional and compressional ridges and translated blocks. In addition, good quality seismic data provide an improved understanding of the frequency, magnitude and slope processes of previous mass movement events in the development area. Conventional 3D streamer seismic surveys give good penetration depth of about 5km, with vertical resolutions of 12m in the upper hundred metres of the sediments, and cover large areas effectively (Kvalstad, 2007). However, to assess the lateral and vertical extent of smaller mass movements and to allow seismostratigraphic mapping, ultra high-resolution (UHR), shallow penetration (50-75m), deep towed or AUV mounted CHIRP seismic survey profiles need to be acquired (Thomas et al., 2010b). CHIRP seismic data have vertical resolutions of less than 0.5m and allow integration with long piston and box cores for deriving frequency, magnitude and potential impact of geohazard on the development. Furthermore, due to the suitability of seismic facies to map-based interpretation, subsurface seascape evolution, sediment thickness distribution and deepwater geomorphology can now be viewed by means of horizon slices, strata slices, and interval attributes, further aiding the assessment of mass failure events (McConnell, 2004).

Geohazard Core Logging. Often, significant sections of the rock record are removed for geotechnical analysis, with the result that one or more events will be missed, creating uncertainties when deriving quantitative data (i.e. dates and sediment accumulation rates) for input into deterministic and probabilistic slope stability analysis (Thomas et al., 2010b). Therefore, an independent geohazard core logging strategy is required to determine event frequency and magnitude. By placing the mass movement events within a temporal context, estimations for frequency and magnitude can be derived. The most important geochronological dating methods for dating mass movements include: (1) biostratigraphic dating, (2) radiometric dating (^{14}C), (3) optically stimulated luminescence dating, (4) amino acid ratios in fossils, and (5) correlation to well dated standards such as the oxygen isotope record (Kvalstad, 2007). Furthermore, a well constrained chronostratigraphic framework allows you to correlate mass movement events to the climostratigraphic curve with the aim of reducing the perceived risk profile of the offshore development. Dating of various Atlantic margin slides indicates that the primary control on submarine slope failure is global climate change (Lee, 2009; Piper and McCall, 2003; Twichell et al., 2009). By correlating to the climostratigraphic curve, particular mass transport events dependent on certain conditioning factors (e.g. glacial periods) can be eliminated as they may no longer be present during the lifetime of the offshore development, thus reducing the risk profile of the development.

In addition, an independent geohazard core logging strategy allows integration with ultra-high resolution seismic data (e.g. AUV CHIRP seismic profile) and facies based frequency and magnitude estimation. For example, a recent study by Thomas et al. (2010a) in the West Nile delta demonstrates that, based on subsurface geophysical data alone, event frequency is often underestimated and magnitude overestimated. Multiple stacked mass movement deposits in the West Nile Delta appeared as a single seismic layer on CHIRP seismic records because individual layers where below the limits of resolution. However, integration with detailed facies based core logging highlighted the individual layers, thus increasing the frequency of events, but decreasing the magnitude (Thomas et al., 2010a). This last example further highlights the use of an independent facies based geohazard core logging in modern geohazard risk assessments.

Samples taken from turbidite beds within the cores studied can be anlaysed for grain size properties. We have established a simple link between maximum grain size (using the one percentile value) and inferred flow velocity of the turbidity current. This is shown in Figure 3 and can be used as a ready reckoning guide to assess the likely velocity (or relative energy) of future turbidity currents in the area. This is presented as a new compilation.

GIS (Geographical Information System). The use of GIS for data organisation, manipulation, graphic representation, regional analyses and slope stability assessments has become routine practice when investigating geohazards both onshore and offshore. In the simplest case, a GIS could be used as a database to compile, manage and present a mass movement inventory, and because associated slide parameters (e.g. location, area etc) are stored in a GIS attributes table, spatial interrelationships can be investigated. For example, Hitchcock et al. (2006) produced simple mudflow susceptibility maps in the Gulf of Mexico using GIS by ranking and assigning point scores for each contributing factor in map layer (i.e. maps of sediment accumulation rate, slope inclination, geology etc). The point scores are then summed up and depicted in a single mudflow susceptibility map. More recently however, the GIS platform has been implemented in studies dealing with predictive methods of mass movement occurrence. A recent study by Mackenzie et al. (2010) describe the use of deterministic slope stability assessment using GIS. A deterministic approach is used to relate the safety factor (FOS) as being dependent on parameters, such as shear strength, slope geometry, external loading or plane orientation, which are modelled probabilistically, enabling fully quantified risk assessment. In this case, GIS spatial analysis techniques could directly be used to derive safety factors against slope failure, with the additional advantage of providing quantitative outputs for subsequent geotechnical risk assessment (Mackenzie et al., 2010; Power et al., 2011). These as well as other examples emphasise the importance of GIS in modern slope instability assessments as it enables sophisticated, numerical-based mapping of slope failure susceptibility.

Gas Hydrates

Occurrence and distribution.

Gas hydrate is a near-surface, sediment cementing, temperature and pressure sensitive substance, with nearly ubiquitous presence on deepwater continental margins. It is a serious potential geohazard that must be assessed prior to deepwater petroleum industry operations.

Gas hydrate, or clathrate, is a solid ice-like substance made up of frozen water molecules that form a hollow cage and enclose a molecule of gas, most commonly methane (Kvenvolden, 1988). Gas hydrate formation requires low temperature, high pressure, water and a volume of gas in excess of solubility (Sloan, 2008). Where these conditions exist, hydrate can spontaneously form.

Gas hydrate was discovered in a laboratory 1810, detected in petroleum pipelines in the 1930s (Hammerschmidt, 1934; Sloan, 2008), but not proven to form natural, geological *in situ* deposits until the 1960s (Sloan, 2008). Ideal temperature and pressure conditions for hydrate formation occur in onshore permafrost regions, shallow water permafrost tongues, near-seafloor oceanic sediments, and in deep fresh water lake sediments; numerous geologically formed hydrate samples have been recovered from each of these environments. Recent estimates of the global volume of carbon sequestered in gas hydrate ranges from 0.83 $\times 10^{15}$ to 1.2 $\times 10^{17}$ (4.18 – 74,000 Gt of methane carbon), (Burwicz et al, 2011; Buffett & Archer, 2004; Klauda & Sandler, 2005; Davie & Buffett, 2001;

Milkov, 2004) with the consensus value on the order of 1 – 5 x 10^{15} (Milkov, 2004). With significant reserves of carbon stored in gas hydrate, it is potentially a viable economic future energy resource, but presents a concern for climate change and slope stability, in addition to presenting a hazard to offshore oil industry operations.

Gas hydrate in ocean sediments. While temperatures on the seafloor are usually above the freezing point of ice, high pressures generated by the overlying column of cool water are sufficient to create conditions suitable for gas hydrate formation. Figure 4 illustrates a phase boundary curve for [free gas + water]—[gas hydrate] system, and pressure-temperature conditions for an oceanic environment. Hydrate is stable at a given depth if temperature is lower than the hydrate/free gas equilibrium at that same depth. Gas hydrate can be stable within ocean water, and will often coat buoyant upwelling gas bubbles that dissociate to free gas when they pass the phase boundary on their upward journey. If conditions at or below the seafloor are conducive to hydrate formation, hydrate may form in place. The maximum depth beneath the seafloor at which hydrate is stable is determined predominantly by the geothermal gradient. The sub-seafloor interval that hydrate is stable over is termed the gas hydrate stability zone (GHSZ); the GHSZ extends on average to depths of 300 – 500 m beneath the seafloor (Milkov, 2004), although factors such high pore-water salinity and heat flow can retard the formation of hydrate.

Near surface sub-sea sediments are typically highly water saturated. If there is a volume of gas in excess of solubility and the seafloor is at appropriate pressure/temperature conditions, there is a good probability that gas hydrate will be present. Gas hydrate accumulations are nearly ubiquitous on continental margins (Fig. 5) and samples have been recovered from drill holes on most margins, including the Gulf of Mexico (Brooks et al, 1986), Prudhoe Bay offshore Alaska, MacKenzie Delta offshore Canada, the Mid-American Trench off Guatemala, both eastern and western US continental margins, Peru, New Zealand (Pecher et al., 2011; Bialas, 2011), offshore India (Winters et al., 2008), and in the Nankai Trough (Tsuji et al., 2009). Most new gas hydrate accumulations are initially identified through the manifestation of a unique signal in seismic data: the bottom simulating reflection (BSR). This is a reflection that parallels the seafloor and has a reverse polarity compared to the seafloor response. The BSR correlates with the base of gas hydrate stability and is thought to occur when gas hydrate at the base of the gas hydrate stability zone directly overlies free gas.

In addition to conducive temperature and pressure conditions, a volume of gas in excess of solubility is required to form gas hydrate. Gas can be of thermogenic origin (migrated upwards from deeper gas reservoirs); *in situ* biogenic origin (product of biomethanogenic processes within sediments within the GHSZ); or deeper biogenic origin (product of biomethanogenesis occurring below the GHSZ and carried up by fluid flow).

Nature and variability.

The structural relationship between porous, permeable sediment and gas hydrate at the grain scale takes one of four forms (Fig 6). Gas hydrates can be (a) disseminated freely within pore space; (b) coating sediment grains; (c) cementing and strengthening the bulk sediment at grain contact points; (d) cementing and strengthening the bulk sediment by forming part of the load-bearing matrix (Kleinberg and Dai, 2005; Jones et al., 2007; Worthington, 2010). For load bearing and cementing modes (b – d) hydrate contributes significantly to the strength of semi- and un-consolidated sediments (Dvorkin et al., 1999; Durham et al., 2003).

Natural gas hydrate accumulations can form as massive layers, thin laminae, nodules, veins and fracture fill, discrete grains disseminated through pore space, and as massive seafloor mounds. Vast quantities of hydrate form at low saturation, and are widely dispersed in the pore space of unconsolidated muds. The highest saturations of oceanic gas hydrate are found in highly permeable, high porosity shallow sands and sandstones, commonly interbedded with non-hydrate bearing finer-grained layers; marine sands with high saturations of gas hydrate recently drilled in the Gulf of Mexico (Boswell et al., 2010), Nankai Trough, Japan (Tsujii, 2009), India (Lee and Collett, 2009) and Cascadian Margin (Riedel et al., 2006). The recent Korean Ulleung Basin drilling program (UBGH2) revealed high saturations of pore-filling hydrate in the sandy layers of turbidites, with nodules, disseminated grains and vein and fracture filling hydrate in the interbedded finer grained muds (Bahk et al., 2011; Lee et al., 2011).

Methane is the most common hydrate forming gas that forms *in situ* (Kvenvolden, 1988); however, more than 130 different compounds have been known to form clathrate hydrates (Sloan, 2008) with ethane, propane, butane, isobutane and carbon dioxide all occasionally observed in nature. Whether a particular gas can form a hydrate is determined by the size of the cavity within the water-ice crystal lattice. The thermodynamic phase stability of hydrate will be different for different gases, with longer chained hydrocarbons stable at higher temperatures (lower pressures).

Damage potential.

Gas hydrates represent a genuine hazard to deepwater drilling, but documentation of specific incidents is scarce, partly because of the limited number of deepwater wells and partly because hydrates are not immediately recognised as the cause of an issue (Nimblett et al., 2005). However, hydrate related incidents in the Arctic permafrost include gas kicks, blow outs, gas leaks outside the casing, fires, stuck pipes and well subsidence (Yakushev and Collett, 1992). Traditionally, the oil industry has avoided the risk presented by gas hydrates by choosing to drill away from known hydrate deposits. With deepwater oil exploration expanding to ever greater depths where the seafloor is at stable hydrate forming conditions, it is becoming harder to avoid drilling through hydrate.

Hydrates represent a direct hazard to deepwater drilling in three primary modes: (1) Melting of *in situ* gas hydrate; (2) gas hydrate formation on subsea structures; (3) uncontrolled release of gas.

Melting of in-situ gas hydrate. Melting of gas hydrate occurs when the local temperature and pressure conditions are perturbed, or the gas hydrate is mechanically disturbed. Drilling itself will mechanically disturb hydrate, but can also generate heat through friction of drilling equipment against casing. Heat from drilling muds and hot fluids produced from below the GHSZ (e.g. water, hydrocarbons) is transferred into sediments surrounding the wellbore (or buried pipelines transporting hydrocarbons). Models of heat transfer show that the destabilization of hydrate can occur within a 100 m radius of seafloor and sub-seafloor structures, though the melting of hydrate will be concentrated and most rapid nearest to the structure (Peters et al., 2008). The rate of response of the hydrate reservoir to heating in a subsea component, and the radius of vulnerability will depend on the thermal conductivity of the component, as well as the permeability, porosity, gas hydrate saturation, water saturation, *in situ* stress state, shear strength, pore pressures and thermal conductivity of the gas hydrate and host sediments (Peters et al., 2008).

Melting of gas hydrate reduces the solid hydrate volume, releases water and methane into sediment pore-space, and decreases water salinity. These changes cause an increase in both the permeability and porosity of sediments. Additionally, if: (a) hydrate-bearing layers are sealed between low-permeability layers, the pore fluid pressure and effective stress will increase; or, (b) hydrate-bearing layers are not sealed and fluid flow carries excess water away, then sediment volume will expand (Dillon et al, 2002; Kayen and Lee, 1991; Waite et al., 2009). These changes combined with the development of interstitial gas bubbles, and the loss of cement and/or the load bearing framework have the potential to cause sediments to lose strength and become unconsolidated (Prior and Coleman, 1984; Paull et al., 2003). The loss of strength may cause localised fracturing or large-scale failure of sediments around subsea components triggering the failure of casing, wellbore, pipelines, rig supports and other ocean-bottom supported equipment (Dickens et al., 1997; Nimblett, 2005; Peters et al., 2008).

Gas hydrate formation on subsea structures. Methane, either dissociated from gas hydrate or escaping to the seafloor through vents or seeps, will form into gas hydrate if conditions are right; this presents a significant risk for seafloor components. When hydrate forms on the exterior of movable parts, including blow out preventers, valves and switches, it can hinder its performance with potentially serious consequences.

Uncontrolled release of gas. Uncontrolled release of gas may occur when in-situ gas hydrate melts and the release gas flows into the well, or when over-pressured gas immediately beneath the GHSZ sealed by an impermeable layer of gas hydrate is encountered. The rapid influx of gas into the well will cause drilling muds to become highly gasified, potentially triggering gas kicks, blow outs and, in some cases, fire (Collett and Dallimore, 2002).

Other modes of damage potential from hydrate. Hydrates can also be a drilling hazard where melting not-related to drilling has already occurred, or is occurring. Hydrate dissociation can be caused by an increase in seafloor and seawater temperature, resulting from a warming or directional change in bottom water currents (Westbrook et al., 2009), or to climatic affects. Dissociation can also be triggered by a decrease in seafloor pressure resulting from eustatic or tectonic sealevel changes. These changes have the potential to cause the dissociation of gas hydrate over extensive regions, potentially causing mass sliding or slumping events and damaging nearby, or distal subsea components. Gas hydrate dissociation has been suggested as an aggravating factor contributing to the triggering of the Storegga landslide on the Norwegian margin (Berndt et al., 2002; Bunz, et al., 2003; Berndt et al., 2004; Mienert et al., 2005) and in many slides on the United States Atlantic margin (Booth et al., 1994).

Hydrate can form spontaneously within the drill hole or production pipelines even when no hydrate zone exists. This will occur when methane mixes with water (drilling fluids) under favourable temperature and pressure conditions. If hydrate formation leads to a blockage of the pipe, it may lead to pressure build-ups behind the blockage leading to an explosion, or, forcing the movement of the plug through the pipeline, causing damage along the way. If hydrates decompose within a limited volume, sealed drill hole very high pressures can be generated, leading to potential blow out/rupture.

Examples Nimblett et al., (2005) report that deepwater wells in Southeast Asia suffered seafloor cracks adjacent to well sites and attributed this, at least in part, to hydrate melting in the zone around the wellsite. Nimblett et al. (2005) also noted operational issues attributable to gas hydrate dissociation in deepwater wells in Southeast Asia and North Africa, including gas flow out of wellhead ports between the surface and the casing. Casing failures occurred in the permafrost Messoyakha drill site in Russia (Goodman, 1979).

Rapid upwelling of gas bubbles (from dissociation or a blow out) over a wider area will a) drive water upwards, and outwards as it reaches the surfaces, leading to strong near-surface currents affecting the stability of rigs, ships and other vessels and, b) cause seawater to become aerated reducing vessel buoyancy (Garcia et al., 2008). Though recent studies indicate rigs and vessels should be able to withstand the loss of buoyancy and that high currents from mass hydrate dissociation or blowouts will have little effect in deepwater (Hammet, 1985; de Andrade Jr, 1997; Adams et al., 2003). Examples of vessels surviving massive gas releases from blowouts include the *West Vanguard* at Haltenbanken in the Norwegian Sea and *Actinia,* offshore Vietnam (Garcia et al., 2008).

Highly gasified muds and gas escaping to the surface, particularly between the casings, presents a large fire risk. Gas bubbling and fires attributed to gas hydrate have been documented, for example, at permafrost drill sites in Canada (Goodman, 1979).

Hazard assessment

In order to reduce the risk provided to oil industry infrastructure by gas hydrates, it is necessary to have a thorough knowledge of gas hydrate distribution, volume and saturation, reservoir temperatures, formation pore pressure, rock porosities and permeabilities (Collett & Dallimore, 2002) and, ideally, hydrate formation mode and bulk sediment strength before and after hydrate dissociation.

Petroleum industry pipelines, drill holes and infrastructure will only be at risk of suffering damage or downtime due to hydrate related problems if the structures themselves are located within water or sediments conducive to the formation of gas hydrate. This can be evaluated to a first order, by establishing the pressure, temperature and salinity conditions at a site of interest, using a combination of multibeam bathymetry, CTD casts (conductivity-temperature-depth), expendable bathythermograph (XBT), moored thermistor measurements and water chemistry measurements (particularly chlorinity, salinity and methane saturation).

The base of the gas hydrate stability is frequently correlated with the depth-converted position of a BSR observed in seismic data. A BSR is not present everywhere gas hydrate occurs, so if hydrate is expected to be present, the geothermal gradient is needed to predict the thickness of hydrate stability. Geothermal gradients can be measured by heatflow probes, or logging-while-drilling (LWD) tools, or by reconstruction from BSRs observed in adjacent locations (e.g. Yamano, et al., 1982; Minshull and White, 1989; Townend, 1997; Henrys et al., 2003).

Actual gas hydrate occurrence is more difficult to determine. As mentioned previously, accumulations are frequently identified through the appearance of a reverse-polarity BSR on seismic data, but where no BSR is visible, other methods can be used to identify gas hydrate accumulations. Seafloor observation methods, including deep camera tows, side-scan sonar, sub-bottom and water column profiles can be utilised to identify outcropping hydrate mounds. These techniques can also identify any chemosynthetic communities or bubble plumes associated with cold methane seeps, methane from hydrate dissociation or methane-rich fluids advecting through seafloor sediments. If the seafloor is at favourable hydrate forming conditions, then high concentrations of methane at the seafloor present a risk to ocean bottom structures even when there is no gas hydrate in the underlying sediments. Hydrate samples recovered from piston and gravity coring, seawater chemistry and pore water geochemistry can be analysed to determine the composition of hydrate forming gases. Recovered samples can also be inspected to ascertain the primary hydrate forming modes (see Fig. 6) and affect of hydrate on sediment strength.

The extent of the gas hydrate reservoir is delineated by combining seafloor observations and the observed BSRs with other geophysical observations. Seismic data can also be analysed in more detail to determine both the presence of gas hydrate and to calculate saturations, through high-resolution velocity of hydrate and non-hydrate bearing intervals (Wood et al., 1994, Mishra, 2004; Ryu et al., 2009), from full waveform inversion, e.g. (Dai et al., 2008; Crutchley et al., 2011), and by interpretation of seismic attributes from seismic cubes (Hato et al., 2006, Lee et al., 2009). Controlled source electromagnetic methods can be used to identify electrically resistive gas hydrate (e.g. Mehta et al., 2005; Schwalenberg et al., 2005; Weitemeyer et al., 2006a, b; Evans, 2007; Weitemeyer and Constable, 2011), and estimate gas hydrate saturations through application of Archie's Law (Archie, 1942).

If drilling has already commenced, wireline logs and LWD tools are useful for estimating hydrate saturations and measuring the petrophysical properties of hydrate bearing intervals. The most commonly used are acoustic velocity (e.g. Lee and Waite, 2008) and electrical resistivity tools, but recently developed integrated nuclear magnetic resonance logging and formation testing have proved invaluable for determining how hydrate is distributed through sediment pore space (Collet and Lee, 2011).

With a comprehensive suite of geophysical and borehole data it is possible to produce a good model of wellbore stability (e.g. Birchwood et al., 2006; Freij-Ayoub, 2007; Salehabadi, 2009; Khabibullin, 2011) and to plan for possible formation over-pressures, and gas release from the hydrate zone during drilling; additives can be added to the drilling fluids to inhibit hydrate formation in the pipelines, or chemically stabilize hydrates near the well bore.

Gas hydrates present a complex and dynamic risk to subsea structures, rigs and drill holes. With the expansion of petroleum exploration into ever deeper waters it is becoming increasingly difficult to avoid drilling through gas hydrate intervals to exploit deeper hydrocarbons. However, with an integrated analysis of commonly-acquired oceanographic, geophysical, geological and petrophysical data, with the addition of targeted LWD tools, a thorough understanding of the local gas hydrate system can be acquired. Such detailed knowledge of the local gas hydrate system will allow drilling engineers to anticipate and therefore mitigate the hazards presented by uncontrolled gas release from gas hydrate intervals, sediment destabilization, and hydrate formation within and on ocean bottom utilities.

Conclusion

In conclusion, it is worth re-emphasising that it is of upmost importance that the oceanographic conditions are carefully considered prior to deepwater operations to ensure the work can be carried out safely. Thorough risk assessment requires detailed knowledge of existing bottom currents, a thorough assessment of potential mass transport events and turbidity currents, and a clear understanding of conditions likely to induce the formation and destabilisation of gas hydrates. Pipelines, cables, subsea installations, key connections such as the riser and any other seabed infrastructure are all susceptible to damage. Correct assessment of hazard will allow for the right equipment to be used for the operations including blowout preventers, riser size, vibration suppressors, and for the safest siting of cables and pipelines. It will also aid the subsea architecture design and planning operations with minimal downtime. All these considerations lead to a safer exploration and production while eliminating unnecessary cost.

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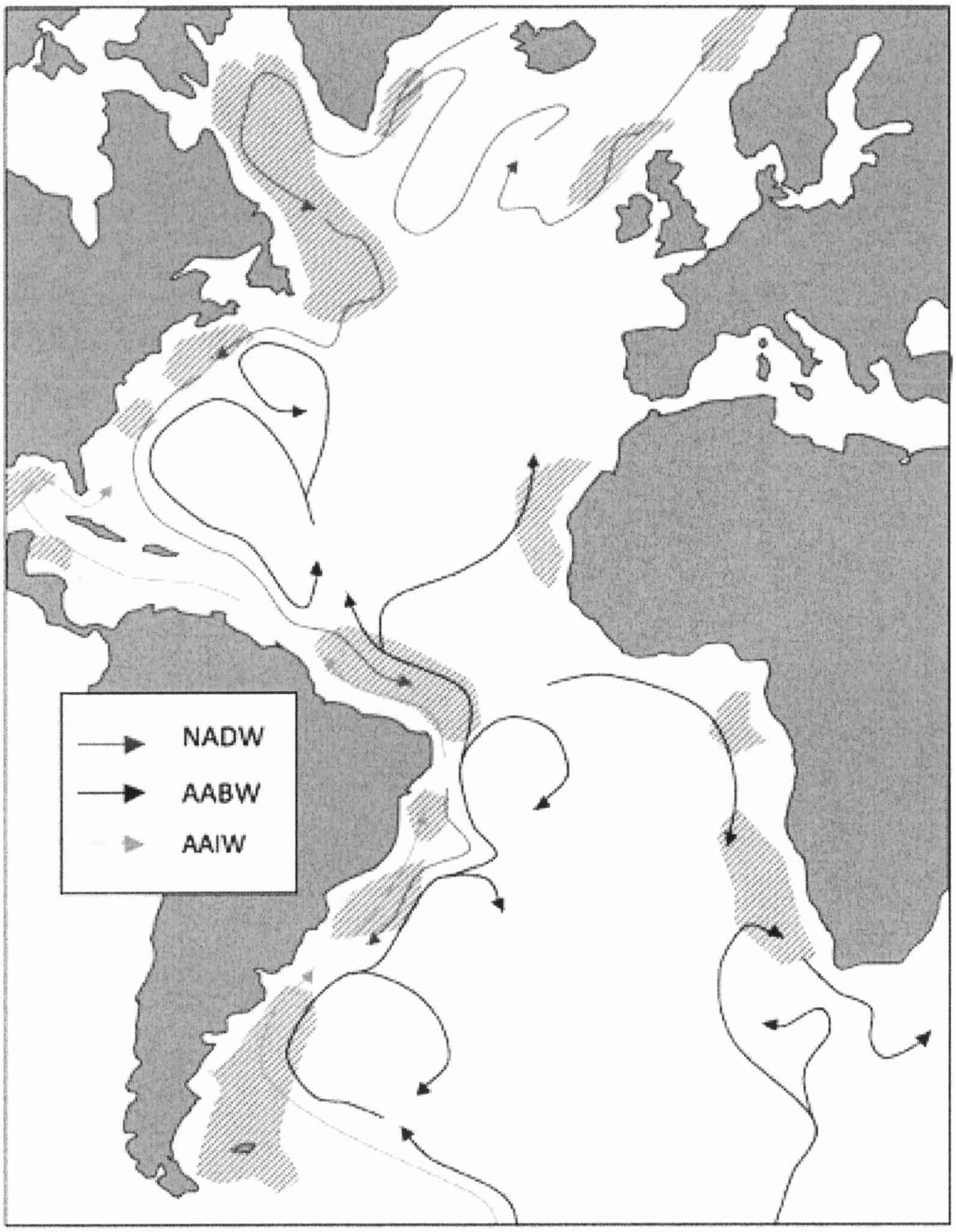

Figure 1. Principal bottom currents active in the North Atlantic. NADW = North Atlantic Deep Water, AABW = Antarctic Bottom Water, AAIW = Antarctic Intermediate Water. Shaded areas represent areas of maximum contourite deposition.

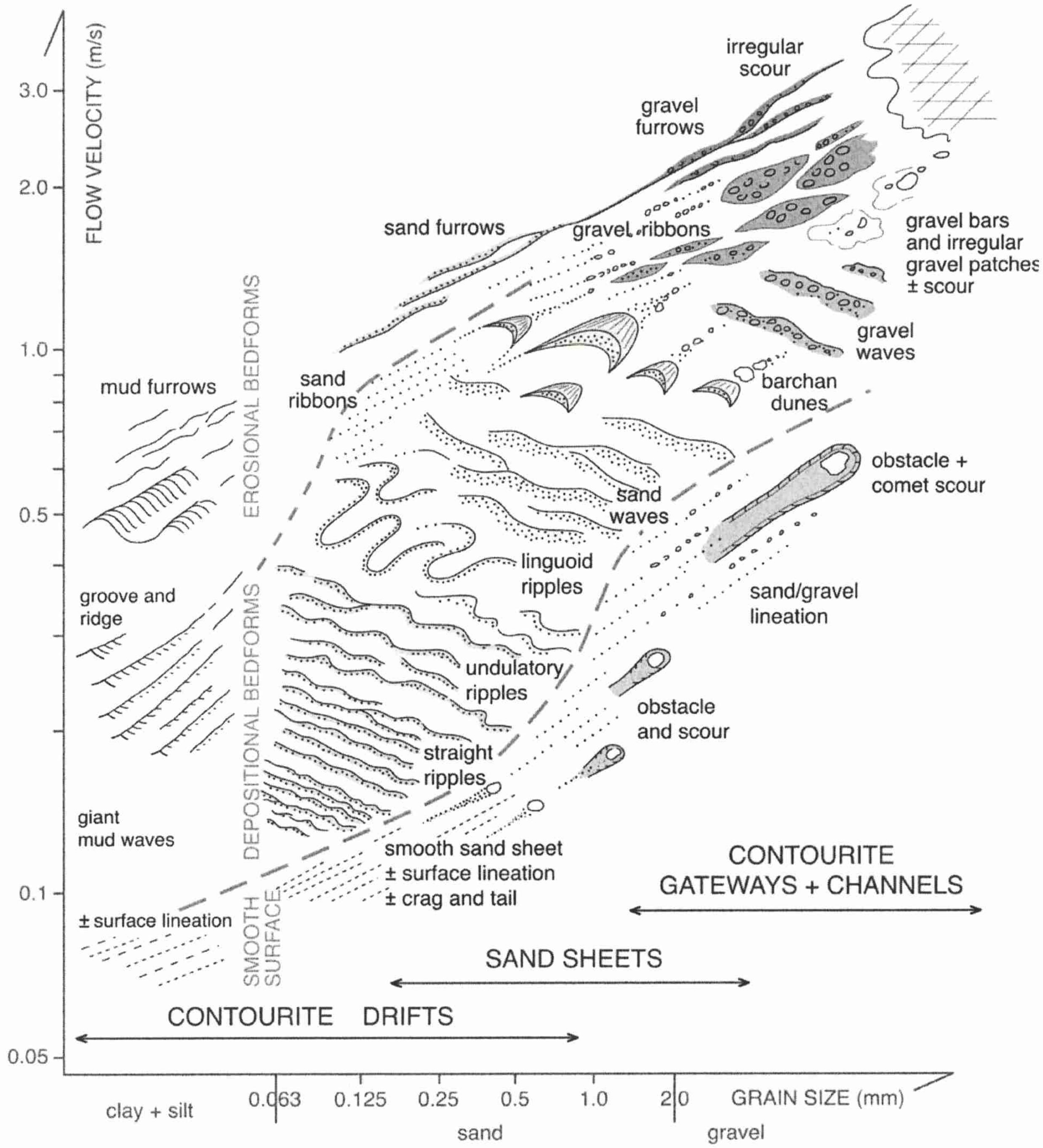

Figure 2. *Bedform-velocity matrix* for deepwater bottom current systems, showing mean grain size of sediment versus flow velocity at or near the seafloor, with schematic representation of the bedforms present under specific velocity-grain size conditions. Modified from Stow et al., 2009.

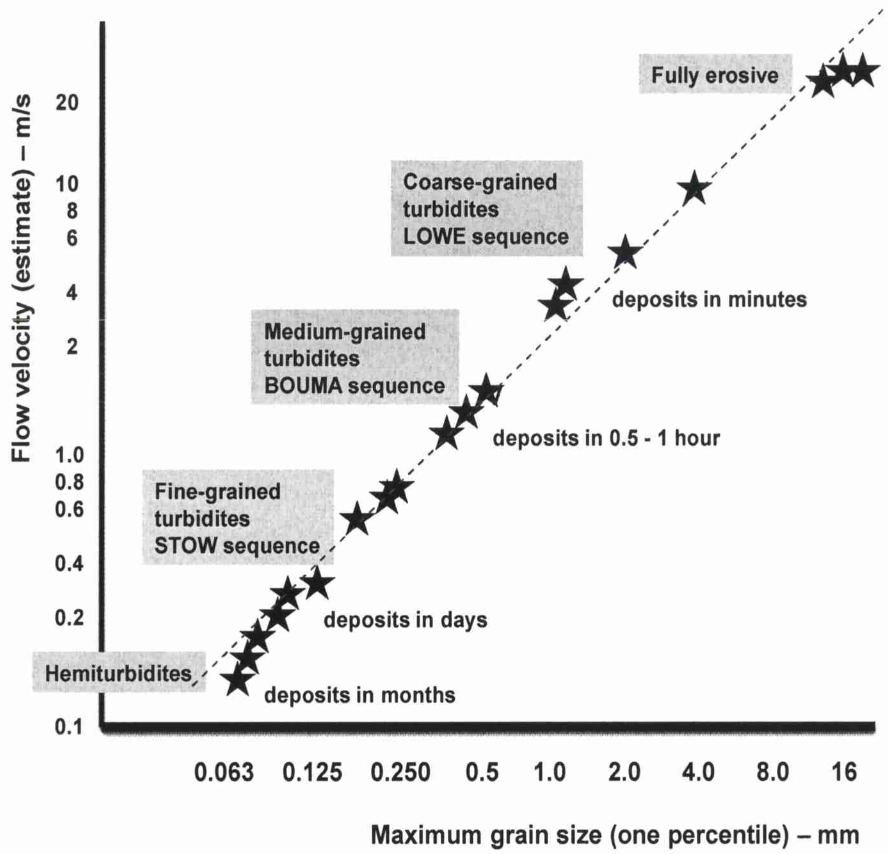

Figure 3. Grain-size versus flow-velocity plot for turbidites. All data from open literature. The grain size is the maximum observed at the one percentile level in the turbidite bed. The approximate grain size and velocity range for the principle turbidite types (models) are indicated, as well as the estimated time for deposition of a single bed to occur.

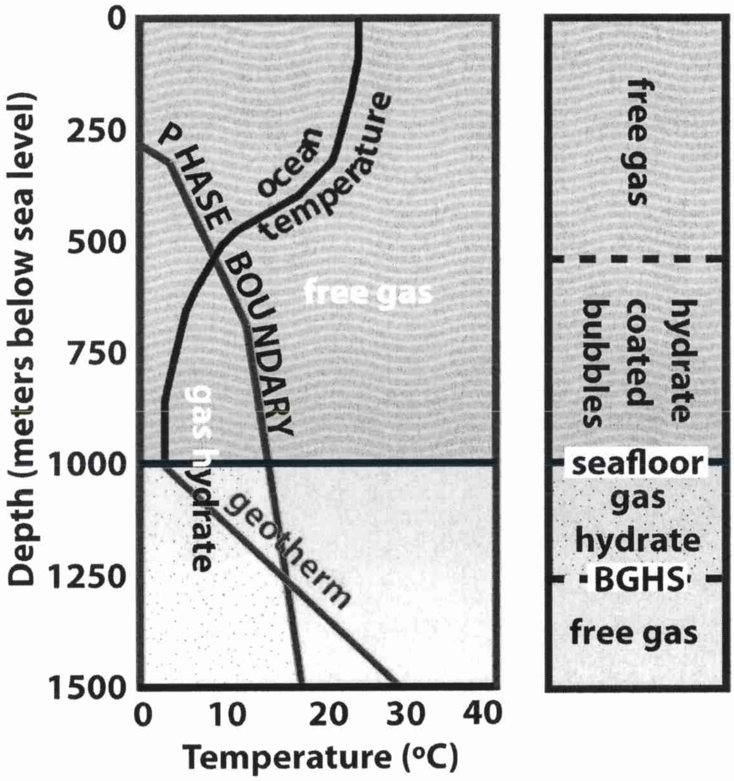

Figure 4. Gas hydrate stability diagram for an oceanic environment, with seafloor at 1000 m b.s.l (after Xu and Ruppel, 1999, modified from Ruppel, 1997). Gas hydrate is stable everywhere to the left of the phase boundary curve (hydrate – free gas equilibrium).

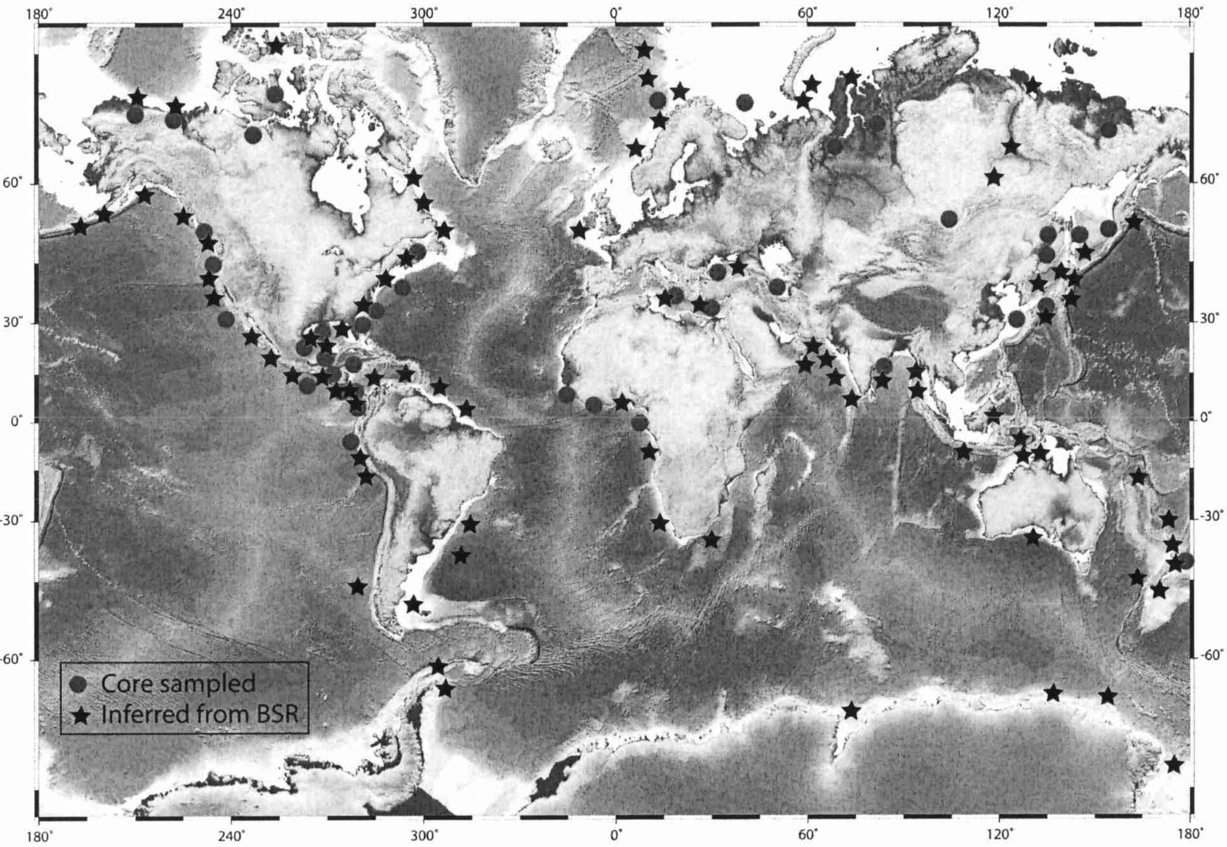

Figure 5. Global gas hydrate distribution; red dots = sampled in drill holes; black stars = inferred from BSRs. (Modified from Makogon, 2010)

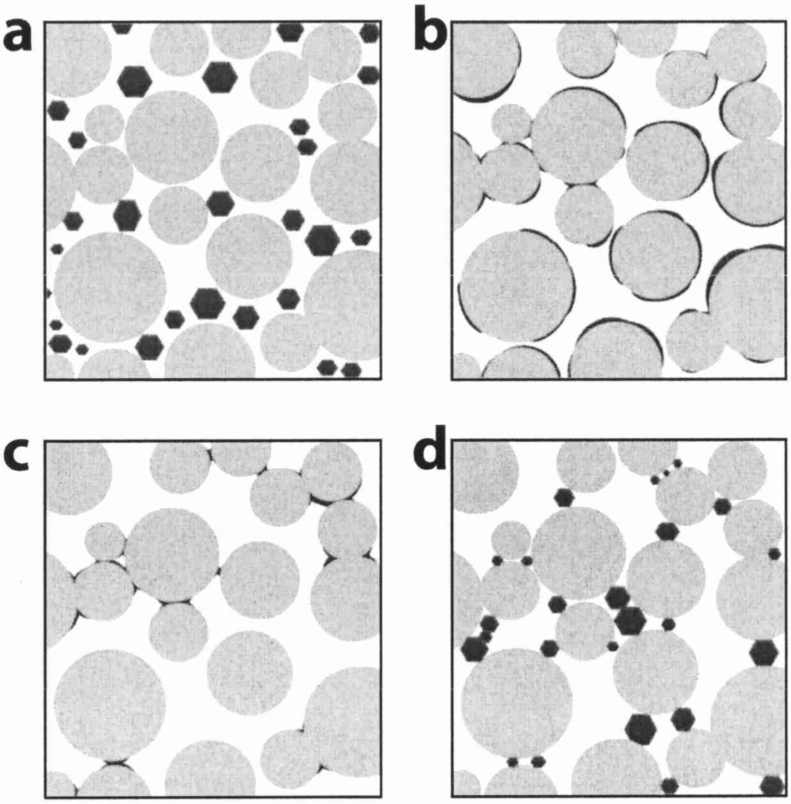

Figure 6. Hydrate forming modes at the pore scale – grey spheres = sediment grains; red = hydrate: a) hydrate disseminated throughout pore space; b) hydrate coating sediment grains; c) hydrate forming at contact points between mineral grains; d) hydrate forms part of the load-bearing matrix supporting sediment grains.