Slope failure dynamics and impacts from seafloor and shallow sub-seafloor geophysical data: case studies from the COSTA project


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Abstract

Holocene and slightly pre-Holocene submarine landslide are found both in high-latitude glacial-dominated margins and in lower latitude, river-dominated margins. This paper constitutes a major assessment on some of the best-studied submarine instabilities in the world. We review and update from original data and literature reports the current state of knowledge of Storegga, Traenadjupet and Finneidfjord slides from the mid-Norwegian margin, Afen Slide from the Faeroe-Shetland Channel, BIG’95 Slide and Central Adriatic Deformation Belt (CADEB) from continental slope and inner continental shelf settings off the Ebro and Po rivers in the Mediterranean Sea, Canary Slide west of the westernmost, youngest Canary Islands and Gebra Slide off the northern tip of the Antarctic Peninsula in the southern hemisphere, i.e. those studied in the Continental Slope Stability (COSTA) project. The investigated slides range in size from the gigantic 90,000 km² and almost 3000 km³ Storegga Slide to the tiny 1 km² and 0.001 km³ Finneidfjord Slide. Not only do individual submarine landslides rarely involve processes precisely fitting with pre-established categories, mostly based on subaerial research, but also they display complex mechanical behaviors within the elastic and plastic fields. Individual events can involve simultaneous or successive vertical to translational movements including block detachment, block gliding, debris flow, mud flow and turbidity currents. The need for an in-depth
revision of the classification criteria, and eventually for a new classification system, based on the new imaging capabilities provided by modern techniques, is more than obvious. We suggest a new system, which, for the moment, is restricted to debris flows and debris avalanches.

Volume calculation methods are critically reviewed and the relations between some key geomorphic parameters are established for the selected slides. The assumed volume missing from scar areas does not necessarily match the actual volume of sediment remobilised by an individual event since in situ sediment can be remoulded and eventually incorporated during the slide downslope journey. CADEB, a shore-parallel prodelta detached from its source, is the exception to the good correlation found between across slope width and alongslope length with slide area. Height drop measured from the headwall upper rim to its foot correlates with the debris deposit maximum thickness unless the slide moves into restricted areas, which prevent farther forward expansion of the deposit, such as Gebr and BIG’95. In such cases, “over-thickened” deposits are found. A particularly loose and fluid behavior can be deduced for slides showing an “over-thinned” character, such as the Canary Slide that traveled 600 km.

Scar areas and slip planes have been investigated with particular emphasis. Although slide headwalls might present locally steep gradients (up to 23° for Storegga Slide), the slope gradients of both the failed segment margins and the main slip planes are very low (max. 2° and usually around 1° and less). An exception is the Finneidfjord Slide (20°–5°) that occurred in 1996 because of a combination of climatic and anthropogenic factors leading to excess pore pressure and failure. Mechanically distinct, low permeable clayey “weak layers” often correspond to slip planes beyond the slide headwall. Since not only formation of these “weak layers” but also sedimentation rates are climatically controlled, we can state that slide pre-conditioning is climatically driven too.

Run-out distances reflect the degree of disintegration of the failed mass of sediment, the total volume of initially failed material and transport mechanisms, including hydroplanning. Commonly, specific run-outs could be attributed to distinct elements, such as cohesive blocks and looser matrix, as nicely illustrated by the BIG’95 Slide. Total run-outs usually correspond to matrix run-outs since the coarser elements tend to rest at shorter distances. Outrunner blocks are, finally, a very common feature proving the ability of those elements to glide over long distances with independence of the rest of the failed mass.

In addition to pre-conditioning factors related to geological setting and sedimentation conditions, a final trigger is required for submarine landslides to take place, which is most often assumed to be an earthquake. In high latitude margins, earthquake magnitude intensification because of post-glacial isostatic rebound has likely played a major role in triggering landslides. Although it cannot be totally ruled out, there is little proof, at least amongst the COSTA slides, that gas hydrate destabilisation or other processes linked to the presence of shallow gas have acted as final triggers.

Keywords: COSTA project; BIG’95 Slide; mid-Norwegian margin

1. Introduction

A tremendous effort has been made in the last few years to characterize and better understand seafloor failures in European margins (Mienert and Weaver, 2002 and references therein) and elsewhere (Locat and Mienert, 2003 and references therein). The more that is known about continental margins, the clearer it becomes that submarine slides are a widespread phenomenon (Canals, 1985; Hütherbach et al., 2004). The interests of the oil industry have triggered their study, jointly with other deepwater geohazards, mostly during the last decade as related to the exploration and exploitation of hydrocarbon resources in the deep sea (Campbell, 1999). A variety of slides, often related to fluid escape, is known to occur in the most important offshore oil provinces such as the Norwegian margin, the Gulf of Guinea, the Gulf of Mexico and the Caspian Sea (Barley, 1999). The second largest gas field discovery off Norway, the Ormen Lange field, is located within the scar created by the Storegga Slide, possibly the largest submarine slide in the world ocean (Bryn et al., 2003a).

Seafloor failures represent a major threat not only to the oil and offshore industries but also to the marine environment and coastal facilities. It is well known that large historical seafloor failures have engendered destructive tsunamis. Recent results indicate that the large tsunamis that devastated Lisbon and struck the Gulf of Cadiz and North Atlantic coasts both in Europe and Africa in 1755 following a magnitude ~8.5 earthquake probably had a landslide contribution.
Seismicity in the Southwestern Iberian margin results from tectonic activity along the Europe–Africa plate boundary connecting the Azores Triple Junction to the west to the Gibraltar Strait to the east (Zitellini et al., 2001). The Lisbon event, as it is known, represents the largest natural catastrophe in Western Europe since the Roman period, which resulted in about 60,000 casualties in Portugal alone (Baptista et al., 1998). The destruction of Lisbon, at the time one of the main capitals in Europe, terrified European society. At present, there is an important international on-going effort off Portugal to further investigate the source area of the Lisbon earthquake and related submarine landsliding (Zitellini et al., 2001; Gracia et al., 2003).

The breaking of submarine telegraph cables during the Grand Banks event following an earthquake in 1929 is also an outstanding case that had a profound impact on deep sea sedimentological research. The water depth of the area affected ranges from 650 to about 2800 m, and the distance between the scar rim and the most distal deposit is $\sim$850 km. It appears that the Grand Banks mass movement may have reached a maximum velocity of about 70 km/h according to Heezen and Ewing (1952). The thickness of the turbidity current was of the order of hundreds of meters as determined from erosional trimlines (Piper and Aksu, 1987). Detailed descriptions of the Grand Banks event and resulting deposits can be found in Rupke (1978) and Piper et al. (1999). This event occurred at a time when the now widely accepted concepts of turbidity currents and the continuum of submarine mass gravity flows (from slumps to debris flows to erosional features cut by turbidity currents, to turbidite deposits) had not yet been conceived. Heezen and Ewing (1952) and Heezen and Hollister (1971) shook the scientific community after convincingly identifying the Grand Banks slumps and turbidity current as the cause of deep sea sediment failures. Their works followed famous earlier papers by Kuenen (1937) and Kuenen and Migliorini (1950) where these authors demonstrated the existence of turbidity currents and showed some of their properties after conducting a series of classic flume experiments. Other key pioneer papers that greatly helped in establishing the current background on mass gravity flows were those of Morgenstern (1967), Hampton (1972) and Middleton and Hampton (1976) to cite just a few. As correctly pointed out by Rupke (1978), these theories revolutionized the study of clastic sediments and enormously stimulated research on deep-sea sedimentary processes.

Now, we know that sediment failure around the epicenter of the 1929 Grand Banks earthquake shows a downslope transition from retrogressive thin-skinned rotational slumps, through debris flows, to erosional features cut by turbidity currents, to turbidite deposits (Piper et al., 1999). The 1929 turbidity current was thus triggered by prolonged numerous relatively small failures nourishing it over a period of about 11 h (Hughes Clarke, 1988). While limited deep-towed side scan sonar imagery, high resolution seismic reflection profiles, sediment cores, in situ shallow geotechnical measurements and submersible observations are available (Hughes Clarke et al., 1989; Piper et al., 1985, 1999), multibeam mapping of the continental slope area disturbed by the 1929 Grand Banks earthquake has not been completed, a surprising situation. The need for new data including swath bathymetry has steered an international consortium of research teams, which has advanced plans to deploy there the best geophysical tools available for deep seafloor and sub-seafloor imaging. The benefits from such an endeavor are anticipated to be of major importance.

A third recent submarine landslide that had a major impact both on coastal facilities and on the scientific community in Europe occurred off the French town of Nice in the Northwestern Mediterranean the 16th of October 1979. The source area was the prograding prodelta of the Var River that accumulated on a very narrow shelf. The nearby Monaco observatory registered no earthquake that could have triggered the slide (Malinverno et al., 1988). Because of the very steep nature of the seafloor off Nice, undercutting cannot be excluded as a concurrent potential triggering mechanism. In addition, sediment failures off Nice are favored by the common occurrence of underconsolidated, meter-thick sediment layers (Cochonat et al., 1993; Klaucke and Cochonat, 1999) although ridge-forming normally consolidated to overconsolidated sediments could also be involved (Mulder et al., 1993, 1994). Three types of sediment failure have been distinguished by Klaucke and Cochonat (1999): superficial slumping, deep-seated failure often associated with successive rotational slides and gullyng of the canyon walls.
The shelf and upper slope 1979 slide evolved into a turbidity current, which, as in the Grand Banks case, broke submarine communication cables. The calculated peak velocity of the mass movement was 40 km/h according to Gennesseaux et al. (1980). The suction effect of the downslope-moving sediment mass generated first a retreat of the sea and, second, a several meters high tsunami wave (Groupe ESCYA-NICE, 1982; Malinverno et al., 1988). As a consequence of the event, part of a land filled area reclaimed to the sea to enlarge the airport of Nice was destroyed, bulldozers were dragged deep into the sea and various people were killed (Savoye, 1991; Mulder et al., 1997). In addition to the work already carried out, the stability of the Nice offshore area is being actively investigated with a priority for observations with highly capable imaging tools, in situ measurements, laboratory tests and modeling (Savoye et al., 2004; Sultan et al., 2004). New in situ instruments such as IFREMER's flexible penetrometer (Penfeld) have been first deployed off Nice.

In a date as recent as July 1998, a tsunami most probably generated by a submarine slump hit the Sissano coast in northwestern Papua-New Guinea (Tappin et al., 1999). Wave heights of 10 m were observed along a 25-km stretch of coastline with maximum heights of 15 m and overland flow velocities of 54–72 km/h. The death toll was over 2200, surpassed in the XXth century only by a tsunami on the coast of Sanriku, Japan, in 1933 (Kawata et al., 1999). The tsunamigenic submarine slump occurred 25 km offshore and was itself probably triggered by an estimated 7.0 magnitude earthquake. The Sissano tsunami is the first that has been comprehensively investigated very soon after its occurrence by seabed and sub-seabed imaging, sediment coring, ROV and manned submersible observations, measurements of potential fields and computer simulations. The approximately 760-m-thick, 5–20-km³ slump took place in an arcuate, amphitheatre-shaped structure made of fine grained, cohesive and stiff sediments that failed by rotational faulting. Fissures, brecciated angular sediment blocks, vertical slopes, talus deposits and evidence of active fluid expulsion have been found in the amphitheatre area. A failure plane with at most a 100-m high exposed scar has been identified on the slump headwall. Also, the occurrence of several events of different ages in the same source area has been postulated. Local seabed morphology resulted in focusing the magnitude and wave-height distributions of the tsunami along the coast (Tappin et al., 2001).

The most recent submarine landslide generating a tsunami that we are aware of, took place on the flanks of the volcanic island of Stromboli, Thyrrhenian Sea, while writing the present paper (December 30, 2002). According to an oral account by S. Tinti from the University of Bologna, Italy, two successive slides, one subaerial and submarine and the other only subaerial, affected an area prone to instability known as Sciara del Fouco. The total volume of rock and debris remobilised was about 28.5 millions of m³ (Bosman et al., 2004). The first slide was responsible for the observed tsunami, which flooded part of the lowlands to the north of the island. The observed height of the wave was up to 10 m at specific locations. There were no casualties. The Stromboli tsunami wave was recorded by tide gauges in nearby islands and also in Milazzo, north of Sicily, where tankers were displaced during oil transfer operations, and oil depots on the coast were close from being hit by the wave.

The above accounts only represent a small part of all the known occurrences of submarine slides in historical times. To illustrate our points, we have deliberately chosen a few slides that generated tsunamis since these are the ones that have a stronger social, economical and scientific impact. Many other submarine landslides are known to have occurred not only during the historical epoch but also throughout the Holocene (Canals, 1985; Hühnerbach et al., 2004). Note that submarine landslides, eventually associated with tsunamis, might be rather frequent along European and North Atlantic margins, even on segments that can be considered tectonically quiet (i.e., Lisbon, Grand Banks and Nice slides and tsunamis).

One of the major advantages of studying geologically recent or historical seafloor mass movements is that they can be much better constrained than older events in terms of resulting morphologies, deposits, dynamics, impacts and ages. To achieve such knowledge, state-of-the-art high resolution geophysical tools (i.e., swath bathymetry systems, deep-towed side scan sonars, high to ultra-high resolution 2D and 3D seismic reflection profiling) are required to provide seafloor and sub-seafloor images of unprecedented quality that can then be used to investigate the above points. The
enormous improvement in surveying equipment during the last few years is bringing to the surface events and impacts, jointly with their fine-grained details, that could not be resolved previously. Coring is a necessary complement to get datable samples for events that have occurred in pre-historical times or whose timing is not well known even if historical.

One of the main tasks within the “Continental Slope Stability” (COSTA) project has been to investigate slope failure dynamics and impacts from seafloor and sub-seafloor shallow geophysical data with the aim to assess:

- External morphology and internal structure of slope failures and resulting deposits
- Slip plane geometries for small, medium and megaslide events
- Run-out distances and flow pathways
- Triggering mechanisms
- Ages of slide events, either single-phased or multi-phased, and recurrence intervals

To achieve the above aims, which overall could illustrate the variability of submarine sediment failures, the research effort focused on eight pre-Holocene to present case studies representing the variety of submarine instabilities that can be found along ocean margins. Seafloor instability events in this paper are now among the best studied in the world. Describing the main results achieved through their study, extracting overall conclusions and distilling implications are the primary goals of the current paper, which also includes a review and summary of previously published data.

2. Setting of the studied slides

Six of the slides studied are located in Europe’s margin and have been systematically and intensively investigated within the COSTA project. These are from north to south Traenadjupet, Storegga and Finneidfjord Slides, off Norway, Afen Slide from the Faeroe-Shetland Channel, and BIG’95 Slide and the Central Adriatic sediment deformation belt from the Mediterranean Sea (Fig. 1). The Canary Slide affecting an ocean island flank has been added as an end member not represented by the European case studies. A slide that occurred off the Northern Antarctic Peninsula, Gebra Slide, has been included for comparison purposes with slides on the Norwegian margin. These case studies cover from shallow to deep settings, from glacial-dominated to river-dominated margins, from giant to tiny instabilities and from long run-outs to almost in situ deformation (Table 1). Several of the studied instabilities on glacial-dominated margins, such as Traenadjupet, Storegga and Gebra, develop off or at a short distance from the mouths of ancient glacial troughs, which were occupied by fast moving ice streams during glacial times. Ice stream-related basal erosion and till transport led to high sediment inputs at the tidewater terminus of the glacial systems off glacial trough mouths (Canals et al., 2002).

The considered slides extend from 69°10’N (Traenadjupet) to 61°15’N (Afen) along Europe’s Atlantic margin and from 1°00’E (BIG’95) to 16°10’E (Adriatic deformation belt) along the northern Mediterranean margins. The Canary Slide is comprised between the crosses of 27°48’N and 31°18’N with 18°30’W and 24°06’W respectively. Finally, the Gebra Slide lies at the crosses of 62°14’S and 62°38’S with 57°40’W and 58°06’W west of the Antarctic Peninsula (Fig. 1; for specific location figures, see different articles in this volume). Their specific geological settings are provided below.

The Traenadjupet Slide is located east and north-east of the marginal Voring Plateau and extends from the shelf break at 400 m to more than 3000 m water depth in the Lofoten Basin abyssal plain. The first published mention to the Traenadjupet Slide, based on 3.5 kHz profiles, was by Damuth (1978) who reported an area of sediment removal on the Norwegian continental slope off Traenadjupet. The name of the slide comes from the close-lying large Traenadjupet glacigenic trough on the shelf. Traena is an island close to coast, while “djupet” means “deep”, in this case a trough separating shallower bank areas.

The Traenadjupet Slide developed in a passive continental margin setting with a continental shelf more than 200 km wide. The shelf has experienced various phases of glacial erosion and sediment bulldozing by the advancement of ice to the shelf break during glacial epochs (Ottesen et al., 2001). Shelf and slope bedrock in the area consists of Tertiary and Mesozoic sedimentary rocks (Sigmond, 1992). Quaternary glacigenic debris flow deposits and glacimarine sediments capped
by a <2-m-thick Holocene hemipelagic drape cover the continental slope. On the shelf, the Quaternary succession is formed by till units interbedded with stratified glacimarine sediments (Laberg and Vorren, 2000). The slide scar is off Traenadjupet glacial shelf trough where the Quaternary succession is comparatively thin (King et al., 1987). The Traenadjupet Trough is the most pronounced on the mid-Norwegian continental shelf, reaching a water depth of more than 450 m.

The Storegga Slide ("great edge" slide, in Norwegian, according to Evans et al., 1996, or simply "shelf edge" after geographic location) is located immediately south of the Voring Plateau in the mid-Norwegian margin. Discovered in 1979, the first report to describe the giant Storegga Slide was by Bugge (1983). The 290-km long headwall of the Storegga Slide is situated in water depths of 150–400 m along the shelf break, 100 km off the Norwegian nearest coast. The distalmost deposits passed through 3800 m of water depth northwest of the Aegir Ridge in the southern Norway Basin (Bugge et al., 1987).

It would be more appropriate to refer to the "Storegga slides" or "slide complex" since what is identified as the Storegga Slide is actually the result of

Table 1
General depositional settings for the submarine landslides studied in detail within COSTA. CADEB: Central Adriatic Deformation Belt

<table>
<thead>
<tr>
<th>TARGETS</th>
<th>GLACIAL-DOMINATED</th>
<th>RIVER-DOMINATED</th>
<th>OTHER</th>
</tr>
</thead>
<tbody>
<tr>
<td>SHALLOW (shelf)</td>
<td>small</td>
<td>Afen, Gebra, Trænaadjupet, Storegga</td>
<td>Finneidfjord</td>
</tr>
<tr>
<td>DEEP (slope)</td>
<td>large</td>
<td></td>
<td>CADEB</td>
</tr>
<tr>
<td>WATER TEMP.</td>
<td>COLD</td>
<td></td>
<td>WARM</td>
</tr>
</tbody>
</table>

Fig. 1. Location of the studied instabilities. Gebra Slide is located off the northern tip of the Antarctic Peninsula.
a succession of events extending through several tens of thousand of years according to Bugge et al. (1987, 1988) or almost concurrent and younger according to new AMS $^{14}$C datings reported in Haflidason et al. (2003a,b) (see Section 10 below). The southern headwall of the Storegga Slide cuts a prominent outbulge formed off the mouth of the Norwegian Channel glacial trough, in a setting that is similar to the Gebra Slide (see below) except for the size of the area and the volume of sediment involved. Other smaller glacial troughs containing ice-streams may also have converged towards Storegga’s head.

The Storegga Slide area partly coincides with a Cenozoic depocentre characterized by a thick prograding sediment wedge developed during the late Pliocene–Pleistocene period (Jansen et al., 1987; Rokoengen et al., 1995; Eidvin et al., 2000; Haflidason et al., 2003a; Evans et al., 2002). Most of the mobilised materials are normally consolidated, stratified and relatively soft fine grained Plio–Quaternary sediments of glacimarine and hemipelagic origin which grade to ice-proximal or diamicton-type sediments towards the upper slope and shelf (Haflidason et al., 2003a). However, more consolidated sediments as well as older sediments are also involved locally (Bouriak et al., 2000). The undisturbed acoustic character of the autochthonous sediments is commonly disrupted by a wide variety of features, which are interpreted to result from fluid expulsion (Evans et al., 1996). Contour current-deposited soft clays have been identified in the two large margin embayments where the Storegga and Traenadjupet slides originated (Bryn et al., 2003a) and could have played a fundamental role as weak layers (see Sections 7 and 9 below).

The so-called Second Storegga Slide, dated at about 7200 yr BP (Bondevik and Svendsen, 1994, 1995), engendered a 10–11-m high tsunami wave that impacted most of the Norwegian coastline and reached the eastern coasts of Scotland and Iceland at least (Dawson et al., 1988, 1993; Svendsen and Mangerud, 1990; Harbitz, 1992; Bondevik et al., 1997).

Last but not least, a major gas discovery, the Ormen Lange field, has been made in a depth of 800–1200 m close to the steep back wall left by the Storegga Slide. A major program has been funded by oil companies to “evaluate large scale margin stability, identify slide release mechanisms, evaluate the risk of possible reservoir subsidence as a result of production, evaluate possible measures to reduce the risk in the event of a development, as well as map the seabed to identify good pipeline routes out of the slide area” (www.offshore-technology.com/projects/ormen/). Production is expected to start in 2007 following completion of a gas liquefaction plant onshore. Tanker vessels will carry most of the liquefied gas to the United Kingdom market.

The Gebra Slide off the Trinity Peninsula, Antarctica, also occurred off a glacial trough in a passive margin. It was discovered in 1993 during the Gebra-93 cruise aboard the Spanish research vessel Hesperides. First called Gebra Valley, it took the name from the cruise acronym, which means “Geological Evolution of the BRAnsfield Basin” (Canals et al., 1993, 1994).

The source area is located in the middle and lower slope, whereas the resulting deposit mostly accumulated in the flat-bottomed King George Basin of the Central Bransfield Basin between the Trinity Peninsula and the South Shetland Islands. The Trinity Peninsula has a total width of about 80 km and includes an inner and an outer continental shelf, a slope and a continental rise that extends down to the basin floor (Gracia et al., 1996b). The inner continental shelf is up to 250 m deep and is incised by four large glacial troughs that behaved as main sediment pathways to the outer shelf during glacial times (Canals et al., 2002). The troughs merge into the outer continental shelf, where the shelf edge varies in depth from about 600 to 750 m in front of the Gebra Slide headwall. The shelf edge is made of basinward-convex till lobes formed at the prolongation of the glacial troughs. It is assumed that the ice was grounded on the shelf edge during glacial maxima. The total water depth range for the Gebra Slide is from 900 m for the uppermost scarp to 1950 m in the deep basin. The slide scar is cut into the toe of the glacial-period continental slope prograding strata (Imbo et al., 2003).

The Gebra Slide itself has not been cored but there are numerous cores from the surrounding non-failed areas (Table 2). The lower section in the King George Basin is covered by a thick late-glacial and post-glacial unit named U8 by Prieto et al. (1999) as illustrated by Imbo et al. (2003). The continental slope around Gebra Slide is formed by alternations of diamictons and hemipelagic/turbiditic layers accumulated during glacial and interglacial periods, respectively (Prieto et al.,...
Overconsolidated muddy and silty sand glacial tills have been sampled by vibrocoring in the outer continental shelf and the shelf edge (Dingle et al., 1998; Canals et al., 2002). Muddy sands and sandy muds cover most of the continental slope (Barcena et al., 1998 and references therein). Recent massive muds resulting from hemipelagic sedimentation, graded sediments deposited by turbidity currents and siliceous oozes related to productivity blooms have been reported from lower slope and basinal settings by Yoon et al. (1994) and Fabres et al. (2000). Various authors have also identified layers including variable amounts of sand-sized volcanic ashes. Late-glacial and post-glacial sedimentation rates are noticeable high in the Bransfield Basin, from 60 to 490 cm ka\(^{-1}\) (Harden et al., 1992).

The Finneidfjord Slide represents slides in the innermost part of formerly glaciated margins that later have been modified by riverine processes. It is the smallest and youngest of all the studied slides. It was observed to happen in June 1996 (see Sections 9 and 10 below). The slide was named after the geographic location, Finneidfjord, which is part of Soerfjorden, Hemmes Commune, Nordland County in Norway. The first report to describe the slide in detail was by Janbu (1996). The slide itself is centered at 66\(^\circ\) 11’ N and 13° 48’ E with the outer boundary of the resulting depositional lobe lying at less than 60 m of water
depth. The slide developed within the slope of a submarine shore face consisting of glacimarine and marine clayey sediments and retreated inshore and onland. As reported by Longva et al. (2003), ground investigations prior to the slide, in connection with public works inshore, showed that the beach sediments comprised soft sensitive clays with layers of quick clay (Sultan et al., 2004) and silt, overlain by up to 5 m of sand. Rockhead sloped towards the fjord with the clay layer therefore thickening downslope towards the shore. At the shoreline, bedrock was encountered at a level of about −15 m. The role of free gas in the activation of submarine slides in Finneidfjord has been examined by Best et al. (2003).

The Afen Slide occurs on the west Shetland slope of the >1000-m deep, NE–SW oriented, glaciated Faeroe-Shetland Channel, 87 km northwest of the Shetlands Islands, and is centered around 61°18′N and 02°27′W. The Channel is a bathymetrically complex narrow passageway in the exchange of deepwater from the Arctic to the Atlantic. The influence of modern hydrodynamics on seafloor bedforms and sediment distribution on the west Shetland slope has been described by Kenyon (1986), Long and Gillespie (1997) and Masson (2001). Much of the shelf and upper slope is characterized by relict glacigenic features such as morainal ridges and iceberg plough-marks (Long, 2001). The imprint of the glacial/interglacial cyclicity is of paramount importance for margin development, although the sedimentary style of the margin is believed to have been initiated in the early Pliocene (Leslie et al., 2003). During the Plio–Pleistocene, locally extensive slope sedimentation shifted the shelf edge seaward more than 50 km (Long et al., 2003a). The thickness of the Quaternary deposits on the west Shetland slope typically is less than 200 m, which contrasts with the >0.5 km known in nearby margins such as the mid-Norwegian margin (Long et al., 2003b).

The Afen Slide is the best-studied instability event in the Faeroe-Shetland Channel, where many other failure events dating back to 200 ka ago have been identified (Long et al., 2003a,b). It was first observed in 1996 from TOBI sidescan and pinger data collected during an environmental survey commission by AFEN, the “Atlantic Frontiers Environmental Network” from where it took its name. The Afen Slide moved northwest from a source area bounded by the 825-m isobath and flowed downslope to 1120 m of water depth. The continental slope in the area is almost devoid of significant topography and displays a gentle uniform gradient of around 2° diminishing to less than 1° beyond 1050 m of water depth (Wilson et al., 2003a,b).

The Afen Slide area is mostly made of clayey and silty glacimarine sediments modified by the interaction of along-slope and down-slope transport processes, with a dominance of the first since at least the second glacial stage of the Pleistocene epoch dated at 0.9 Ma ago (see above) (Stoker et al., 1993, 1994; Wilson et al., 2003a,b). Clean sandy layers and pockets are known or have been inferred to exist below and within the slide lobe, as well as sandy contours are located directly above the slide and within the stable sediment packages surrounding the failed area (Masson, 2001; Sultan et al., 2004). The Afen Slide headwall is cut into an elongated contourite mound field, as illustrated on Fig. 5 of Bulat and Long (2001).

A 3D exploration seismic data cube provided by Shell UK Exploration has been examined to determine the regional setting of the Afen Slide by looking in detail at the seabed pick and examining key subsurface reflectors. This has revealed the presence of faulting and minor downslope channels within the Quaternary sediments in the slide area. There is co-location of these faults and features in the seabed outline of the Afen Slide (Long and Bulat, 2001a). The potential linkage of the Afen Slide with a main structural feature, the Victory Transfer Zone, is under investigation (Long and Bulat, 2001b).

The BIG'95 Slide occurred on the prograding, river-dominated continental slope of the Ebro margin, east of the Iberian Peninsula, in the Northwestern Mediterranean Sea, an area with fewer earthquake activity if compared to other Mediterranean areas (Grüntthal et al., 1999). The Ebro margin forms the western side of the Valencia Trough, a late Oligocene/early Miocene–Pleistocene extensional basin between the Balearic Promontory and the Iberian Peninsula, which was almost totally opened at 10 Ma BP (Fernandez et al., 1995; Gueguen et al., 1998). Several volcanic structures related to the Neogene and Quaternary extension are known in the Valencia Trough, including the Ebro margin where the Columbretes Islets are the emerged expression of a 90×40-km, mostly buried volcanic
The Ebro margin is mainly fed by the 900-km long Ebro River, draining almost one-sixth (~85,000 km$^2$) of the Iberian Peninsula. The Ebro is the fourth largest sediment source to the Mediterranean Sea, discharging annually five to six million tons of sediment, a volume that probably was about three times larger during Quaternary lowstands when most of the progradation took place (Nelson, 1990). The Plio–Pleistocene thick progradational sequence forming the Ebro continental margin is known as the Ebro Group and includes the lower Ebro Clays, a clayey unit of Pliocene age and the upper Ebro Sands, a Pleistocene clastic shelf complex (Soler et al., 1983). This sequence overlies the erosional Messinian unconformity created when the Mediterranean dried up (Clavell and Berastegui, 1991; Maillard et al., 1992; Escutia and Maldonado, 1992). The growth pattern and sedimentation rates of the Ebro margin were controlled by glacioeustatic sea-level oscillations, a relatively strong subsidence and climatically driven changes in sediment supply (Farran and Maldonado, 1990).

In terms of physiography, the Ebro margin consists of an up to 70-km wide continental shelf, one of widest in the entire Mediterranean, a 10-km narrow slope with a mean gradient of 4° and a smooth continental rise that progressively deepens till encountering the Valencia Channel. While the upper course of the Valencia Channel is now hardly recognizable since it is buried under the deposits resulting from the BIG’95 Slide (Canals et al., 2000c). Core samples taken both within the BIG’95 remobilised mass of sediment and in adjacent non-failed areas allowed to identify: (i) an upper unit made of rather homogeneous brownish clays and clayey silts with abundant foraminifera of hemipelagic origin; (ii) an intermediate unit with convoluted laminations of interbedded sands, silts and silty clays and low fossil content; and (iii) a lower unit composed of grey massive, sometimes laminated silty clays with little fossil content (Urgeles et al., 2003). It is considered that these three units correspond to the post-, syn- and pre-slide materials, respectively (Wilmott et al., 2001). Relatively coarse black sands attributed to a very recent turbiditic event have been identified on top of the upper unit in the uppermost, filled course of the Valencia Channel. While the upper unit is also identified in the non-failed areas, materials below it consist either of massive clays and silty clays in the upper slope, or interbedded clayey silts and sands of turbiditic origin in the levede lower slope (Urgeles et al., 2003). Massive medium sands have been also recovered in the uppermost continental slope by vibrocoring.

The Canary Slide occurs in a setting, a volcanic ocean island flank, which is unique amongst those of the other slides considered in this paper. It extends for about 600 km from the northwestern lower slope of El Hierro Island in the Canary Archipelago, at 3900 m of water depth, to the eastern edge of the Madeira Abyssal Plain (MAP), at about 5400 m water depth. It can be considered that the Canary Slide is centered at 30°N and 21°W. It was identified as a separate event in 1989 since previously it was believed to be part of the Saharan Slide, discovered in the late 1970s of the XXth century (Embley, 1976).

The Canary Archipelago is made of seven, roughly E–W-oriented large islands and a few islets in the Atlantic Ocean offshore northwest Africa. The basaltic-dominated intraplate Canary oceanic-island volcanoes have been related to an upwelling mantle plume or hotspot, now most probably located close to the island of El Hierro. Notwithstanding its proximity to the African margin, all the islands have been built up on oceanic crust fractured by a WNW–ESE-oriented Atlantic system, and an ENE–WSW- to NNE–SSW-oriented system related to the Atlas Range inland in
west of the abyssal plain, respectively (Weaver et al., 1992; Lebreiro et al., 1998; Alibes et al., 1999).

At least 18 large lateral collapses and landslides originating from the flanks of the various islands in the Canaries are known, with a majority being partly subaerial and partly submarine (Canals et al., 2000b; Urgeles et al., 2001; Krastel et al., 2001; Masson et al., 2002, and references therein). The occurrence of those landslides seems to be related to volcanic rift zones forming star-shaped three-arm alignments. But the Canary Slide is submarine only (see Section 9 below) (Masson et al., 1998). Onshore, landslide headwalls are most often expressed as arcuate embayments with steep cliffs (Cantagrel et al., 1999; Navarro and Coello, 1989; Ridley, 1971). The discovery of landslide deposits offshore the Canaries proved to be fundamental for an integrated seascape/landscape integrated evolutionary model (Urgeles et al., 1998).

The pattern of sediment influx observed west of the islands of El Hierro and La Palma, the second youngest island (2 Ma BP) shows that the E–W migration of the major volcanic-forming episodes controls the sedimentary processes and the location of volcaniclastic depocenters (Urgeles et al., 1998). Above a lower unit believed to consist mostly of pelagic sediments, the intermediate and upper units filling the MAP are dominated by turbidites (Duin et al., 1984; Searle, 1987). MAP turbidites are of three types, volcanic-rich, organic-rich and calcareous, reflecting source areas in the Canary Islands, the NW African continental margin and seamounts to the west of the abyssal plain, respectively (Weaver et al., 1992; Lebreiro et al., 1998; Alibes et al., 1999).

Several studies indicate that the flanks of El Hierro and nearby La Palma are covered predominantly by fragmented sedimentary material, including pelagic sediments, turbidites and volcaniclastic products (Simm et al., 1991; Masson et al., 1992, 1997, 1998). Slope gradients vary from 1° at the source to virtually 0° at the edge of the abyssal plain. This suggests a highly mobile flow. The deposit itself consists of a mixture of clasts and matrix of the above lithologies. Large slabs up to 300 cm across have been identified too (Masson et al., 1997). Similarly to the BIG’95 Slide, the slide deposit is capped by a thin layer of hemipelagic sediment post-dating it and partly filled channels have been also identified within the area of the flow (Masson et al., 1992). These channels can be up to 10 km wide and are typically 10–30 m deep. The Canary Slide deposit correlates with a prominent turbidite in the MAP known as the “b” turbidite (Weaver et al., 1994).

The origin of the Central Adriatic Deformation Belt (CADEB) is object of intense scientific debate. The reader should be aware that, although studied as a submarine landslide within the COSTA project, some authors think that it is not a failure but a depositional feature similar to sediment waves, resulting from hyperpycnal flows out of the Appenine rivers and possibly the Po river (e.g., Lee et al., 2002). This paper will consider CADEB to be an end-term landslide from here onwards, because it differs from the other study cases because downslope sediment displacement is very limited or null, and because the along-slope dimension of the affected sediments is several times larger than the across-slope dimension. CADEB occurs as a narrow deformation fringe parallel to the isobaths between 43°N and 42°N, from offshore Ortona and the northern Gargano Promontory, at water depths of 30–110 m. CADEB is thus, jointly with Finneidfjord Slide, the shallowest of the studied instabilities. CADEB and BIG’95 (see above) provide a view on seafloor instabilities in river-dominated margins covering both deep water and shallow water, and passive and active margin settings. CADEB was first imaged locally in 1989 and at a regional scale in 1992 (Hovland and Curzi, 1989; Correggiari et al., 1992).

The Adriatic Sea is a narrow (92–220 km), NW–SE-elongated (800 km), shallow semi-enclosed basin that communicates with the deep Ionian Sea in the Western Mediterranean through the Otranto Strait. The
Adriatic Sea constitutes the latest Apennine foreland basin, which is segmented according to lithospheric thickness, state of deformation and rates of subduction (Ciabatti et al., 1987; Royden et al., 1987). The movement of the westward-dipping Adriatic microplate triggers a shallow distributed seismicity characteristic of the area (Argnani et al., 1993; Doglioni et al., 1994). Substantial Quaternary uplifting has been observed both inland and offshore the Gargano Promontory, where several historical strong earthquakes are also known (Tinti et al., 1995). The northwestern third of the Adriatic Sea is occupied by the largest epicontinental shelf in the Mediterranean Sea, constructed with the sediment input from the Po River and Appenine rivers during and subsequently to the post-glacial sea level rise. The modern shelf overlaps a former glacial alluvial plain (Trincardi et al., 1994).

The CADEB deformed unit belongs to the Central Adriatic shelf mud wedge lying on top of the late Holocene maximum flooding surface and is, therefore, part of the late Holocene highstand systems tract (Correggiari et al., 2001). The modern mean suspended load entering from the western side of the Adriatic Sea, constructed with the sediment input from the Po River and Appenine rivers during and subsequently to the post-glacial sea level rise. The modern shelf overlaps a former glacial alluvial plain (Trincardi et al., 1994).

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Perhaps surprisingly, the main sediment source is the ensemble of Apennine rivers with a total drainage area of \(23 \times 10^3 \text{ km}^2\), a mean suspended load of \(24 \times 10^9 \text{ kg yr}^{-1}\) and a sediment yield exceeding \(10^6 \text{ kg km}^{-2} \text{ yr}^{-1}\). The Po River, which enters the Adriatic Sea at about 45°N forming one of the largest deltas in the entire Mediterranean Sea, has a drainage area of \(54 \times 10^3 \text{ km}^2\), a mean suspended load of \(15 \times 10^9 \text{ kg yr}^{-1}\) and a sediment yield of \(0.28 \times 10^6 \text{ kg km}^{-2} \text{ yr}^{-1}\) (Frignani et al., 1992; Milliman and Syvitski, 1992; Bartolini et al., 1996).

The general cyclonic circulation carries fine-grained particles supplied by the Po and Apennine rivers south-eastward along the eastern Italian coast. The sediment accumulates as shore-parallel muddy prodeltaic wedges detached from their river source. The muds are layered, display high water and clay contents, low density and low shear strength (Correggiari et al., 2001; Cattaneo et al., 2003a). CADEB is, therefore, a good representative of the tens of meters thick extensive mud-dominated coastal wedges common in most Mediterranean margins. River inputs resulted in progradation after the present sea-level highstand was reached ca. 5.5 cal ka BP (Correggiari et al., 1996; Cattaneo and Trincardi, 1999).

According to several reports, seafloor crenulations are common in these mud-dominated prodelta slopes offshore river mouths all around the Mediterranean and elsewhere. They generally occur under seafloor gradients of tenths of a degree, display a variety of internal geometries and seem to be associated to high sedimentation rates (Correggiari et al., 2001 and references therein). These have been estimated to be higher than 1.5 cm yr\(^{-1}\) for the western Adriatic shelf. In the CADEB area, the mud prism shows a sigmoid section with an almost horizontal topset region (0.02°) and a foreset region inclined about 0.5°, locally up to 1°. The depocenter of the up to 35-m-thick CADEB muddy wedge is at 35–40-m water depth while in the modern Po Delta, further north, it is located at the shoreline (Cattaneo et al., 2003a). Widespread diffused impregnation by biogenic gas in the shallowest topsets has been inferred from acoustic masking in very high resolution (VHR) seismic reflection profiles offshore Ortona and north of Gargano Promontory (Correggiari et al., 2001).

### 3. Methods

High resolution state-of-the-art geophysical methods were applied to the case studies considered within the COSTA project. These include both seafloor and shallow sub-seafloor imaging tools deployed both near the sea surface and near the ocean bottom. Sediment cores provided materials to groundtruth geophysical interpretations and to perform age analyses of the events. Table 2 summarizes the methods used to investigate the slope failure dynamics and impacts of each of the eight landslides of the COSTA project.

It is beyond the scope of this paper to describe in detail the various methods used, and all of them are widely known by the scientific community. However, we have added at the end of each paragraph below a brief selection of references that are easy to read, handbooks or scientific articles that will allow the interested reader to expand their knowledge of the principles and practicalities of the various groups of techniques mentioned in Table 2. Those interested by
8. Triggering mechanisms

A triggering mechanism, or a combination of triggering mechanisms is required to destabilise sedimentary packages already prone to failure because of a set of preconditioning factors. Failure occurs when the downslope oriented shear stress (driving shear stress) exceeds the shear strength (resisting stress) of the material forming the slope, as expressed by the well known Mohr–Coulomb failure criterion:

\[ \tau_f = c' + (\sigma - u)\tan \phi' \]  

where \( \tau_f \) is the shear strength (equivalent to the shear stress at failure), \( c' \) is the effective cohesion, \( \phi' \) is the friction angle, \( \sigma \) is the total stress acting normal to the failure surface and \( u \) is the pore water pressure. The term \( (\sigma - u) \) is the effective normal stress, \( \sigma' \). Generically speaking, processes that reduce the strength include earthquakes, tidal changes and sedimentation as the most important, but also wave loading, weathering and presence of gas. Those that increase the stress are wave loading, earthquakes, tidal changes, diapirism, sedimentation and erosion (Hampton et al., 1996). The equilibrium of a slope will finally depend on the relation between resisting forces and driving forces. For most sediment, the effective cohesion is low or even negligible and the friction angle generally is about 20–45° depending on compositional variations. More significantly, expression (2) shows a linear relation between shear strength and effective stress, i.e., a reduction in the effective stress leads to an equivalent reduction of the shear strength.

Gravity, seismically induced stress and storm-wave induced stress are considered the most significant downslope driving stresses with respect to submarine landslides. Mechanically, a landslide occurs when the downslope driving stresses nearly always involving gravity and other factors, exceed the resisting strength of the slope-forming material. While many studies of limiting equilibrium of submarine slopes have been conducted, it is beyond the scope of this paper entering into detailed descriptions on that subject. A good summary could be found in Hampton et al. (1996), which also includes relevant references. The same authors identify five submarine environments where slope failure is most common since they often match the necessary conditions for landsliding to occur, such as rapid accumulation of thick sedimentary deposits, sloping seafloor, and other types of high environmental stresses. The environments that could be designated as “submarine landslide territory” are: (1) fjords, (2) active river prodeltas, (3) submarine canyon-fan systems, (4) the open continental slope and (5) oceanic volcanic islands and ridges. The COSTA project includes representatives of all but one (the submarine canyon-fan systems) of these environments: Finneidfjord (fjords), CADEB (active river prodeltas), Storegga, Traenadjupet, BIG’95, Gebra and Afen (open continental slope), and Canary (oceanic volcanic islands).

A substantial number of factors that could explain large scale slope instabilities along continental margins can be found in the literature. They would include (1) high sedimentation rates leading to build-up of excess pore pressure (overpressurized layers) and underconsolidation (weak layers), (2) loading and crust flexing by a static weight such as a grounded ice sheet, (3) fast loading by a dynamic weight such as a landslide mass released from upslope, (4) destabilization of gas hydrates, (5) fluid seepage including seepage of shallow methane gas, (6) bubble-phase gas charging, (7) presence of diagenetic fronts, (8) oversteepening of the margin, (9) erosion at the base of the slope, (10) seismic loading due to earthquakes, (11) low tides, (12) storm-wave loading, (13) sea-level change, (14) volcanic growth and dyke injection, (15) faulting, (16) tectonic compression, (17) diapir and mound formation, (18) biologic processes and (19) human activities on or affecting the seafloor. Canals (1985) grouped the triggers into external (i.e., seismic loading or storm-wave loading) and internal (i.e., weak layers or diagenetic fronts). Locat and Lee (2002) report several case studies illustrating how most of the factors listed above favored or directly triggered submarine landslides. When interpreting past submarine landslides, combinations of some of the triggers listed above are often invoked. It is worth mentioning that, though seismic loading and oversteepening have been considered as triggers since the early work of Morgenstern (1967), it has been also demonstrated that repeated seismic shaking could lead to “seismic strengthening” of the sediments if drainage is allowed between successive events (Boulanger et al., 1998; Boulanger, 2000).

Because of their varied settings, the submarine landslides studied within the COSTA project illustrate how different pre-conditioning factors and triggers could interact to finally lead to sediment failure at...
<table>
<thead>
<tr>
<th>Landslide</th>
<th>Margin type</th>
<th>Seismic activity</th>
<th>Volcanism Recent activity, dyking</th>
<th>Volcanic structures nearby</th>
<th>Presence of gas and diagenetic fronts</th>
</tr>
</thead>
<tbody>
<tr>
<td>Storegga</td>
<td>Passive</td>
<td>Intermediate</td>
<td>Lacking</td>
<td>Ocean basalt/continental margin boundary under the intermediate slide segment</td>
<td>Gas hydrates confined to a small zone along the northern flank of the slide and the slide area itself. BSR present</td>
</tr>
<tr>
<td>Canary</td>
<td>Ocean island flank</td>
<td>Low, related to volcanism</td>
<td>Last, 37 ky old volcanic edifice still growing</td>
<td>Flank of volcanic island less than 1.12 My old</td>
<td>Not known</td>
</tr>
<tr>
<td>Traenadjupet</td>
<td>Passive</td>
<td>Intermediate</td>
<td>Lacking</td>
<td>Lacking</td>
<td>BSR detected</td>
</tr>
<tr>
<td>CADEB</td>
<td>Active</td>
<td>High ($M_{\text{L}}$6.6)</td>
<td>Lacking</td>
<td>Lacking</td>
<td>Presence of shallow gas</td>
</tr>
<tr>
<td>BIG’95</td>
<td>Passive</td>
<td>Low to intermediate ($M_{\text{L}}$4.9)</td>
<td>Fluid escape likely detected from the feet of seamounts in the Valencia Trough</td>
<td>Volcanic dome beneath the scar area, Columbretes Islets volcanic archipelago upslope the scar, and several seamounts nearby</td>
<td>Likely presence of shallow gas in former prodeltaic sediments on the shelf edge and upper slope</td>
</tr>
<tr>
<td>Gebra</td>
<td>Passive</td>
<td>Intermediate to high ($M_{\text{L}}$6.7)</td>
<td>Subrecent to modern volcanic activity both submarine and subaerial (Deception Island mostly), associated with seismicity</td>
<td>Young volcanic lineaments along basin axis. Incipient seafloor spreading</td>
<td>Gas hydrates known in nearby areas (South Shetland Islands margin). Diagenetic fronts also known in the region Distinct BSR close to the area representing a diagenetic front associated with opal-A/opal-C transition</td>
</tr>
<tr>
<td>Afen</td>
<td>Passive</td>
<td>Low ($M_{\text{L}}$2.9)</td>
<td>Lacking</td>
<td>Lacking</td>
<td>Presence of biogenic gas</td>
</tr>
<tr>
<td>Finneidfjord</td>
<td>Passive</td>
<td>Intermediate ($M_{\text{L}}$5.8)</td>
<td>Lacking</td>
<td>Lacking</td>
<td>Presence of biogenic gas</td>
</tr>
</tbody>
</table>
various time and size scales. In Table 8, we summarize preconditions factors and final triggers of the instabilities investigated within our project. While five of them, including the largest and the smallest of the slides studied (Storegga, Traenadjupet, BIG’95, Afen and Finneidfjord slides), occurred on passive margins, there are also representatives of ocean island flanks (Canary Slide), back arc basins (Gebra Slide) and active margins (CADEB). Curiously, CADEB, the instability from the active West Adriatic margin is the only one dominated by plastic deformation with no disintegration nor flow of the sediment mass.

The Storegga Slide, the largest of the studied instabilities, is according to the latest interpretations based on a wealth of data partially from the oil industry (Table 2), the last of a series of slides occurring in the same area during the last 500 ka. Such a succession of relatively similar events follows a repeated cycle of climatically controlled sedimentary processes leading every time to about the same result. The cycle starts with the deposition of fine-grained marine clays on the slope and outer shelf during interglacials and transitions. These clays accumulate under the influence of energetic currents. Contourite drifts reaching more than 100 m of thickness developed during these periods of reduced or non-existing grounded ice on the continental shelf. Contourite sediment may partly fill in old slide scars and other sea floor depressions. Afterwards, as glaciation progresses, the sedimentary regime changes dramatically. Subglacial bulldozing-like transport is greatly enhanced, especially under ice streams whose imprint is still clearly visible on the modern mid-Norwegian continental shelf as glacial troughs and megascalar streamlined structures (Figs. 7 and Canals et al., 2000a, 2003). During glacial maxima, the grounding line is pushed forward to the edge of the continental shelf and fast flowing ice streams actively carry deformable subglacial till which is quickly deposited on the outer margin and subsequently transported downslope as glacial debris flows to form glacial fan systems as the North Sea Fan and Skjoldryggen glacial depocenter (Dowdeswell et al., 1996; Vorren et al., 1998; Solheim et al., 1998; Dimakis et al., 2000; Bryn et al., 2003b; Dahlgren and Vorren, 2003). A main depocenter of glacial clays has been proposed for the area now occupied by the upper Storegga Slide scar, close to the North Sea Fan (Bryn et al., 2003b, their Fig. 2). Under such a situation, excess pore pressure tends to rise. In turn, effective strength of sediments barely increases with sediment burial, being thus lower than expected. Clayey units deposited during interglacials and transitions have then a great potential to behave as slip planes or weak layers. As Bryn et al. (2003b) point out, in these circumstances, permeability is a key factor in trapping excess pore pressure and also in transferring excess pore pressure laterally to areas with less overburden, reducing the strength where slopes are in addition steeper because of the lack of smoothening by fast sediment accumulation. It must be noted that the Storegga Slide is located in the reentrant between the North Sea Fan and Skjoldryggen outer margin glacial depocenters, in a situation which is similar to that observed for the Gebra Slide and along the Pacific margin of the Northern Antarctic Peninsula (Canals et al., 2002; Imbo et al., 2003; Amblas et al., submitted for publication). Reentrants in between shelf edge glacial lobes and outer margin depocenters would then behave as instability corridors because of lateral transfer of excess pore pressure generated beneath thick nearby sediment piles but also because of the weight of grounded ice. Thicker ice occurs in the main ice streams, thus reinforcing the excess pore pressure effect, while ice is thinner in smaller ice streams and intervening areas where slide headwalls would tend to locate.

A final external trigger is, however, required. For the Storegga Slide and its cousin Gebra Slide, this external trigger was likely earthquake activity, a hypothesis that seems to be confirmed by modelling and reports on seismic activity (Bryn et al., 2003b; Imbo et al., 2003). Note that failure of the Storegga and Gebra Slides did not necessarily occur synchronously with glaciation maxima. In fact, enhanced seismicity during initial deglaciation because of glacioisostatic rebound may well be the required final trigger, as strongly suggested by the ages of both Storegga and Gebra Slides (see Section 10 below). Earthquake swarms may progressively prepare the

Notes to Table 8:

a For landslides whose occurrence has not been observed, this column refers to the most likely final triggers.

b Refers to distance from headwall upper edge.

c Last Glacial Maximum.
sediment pile for sliding through the formation of creep structures, open cracks and initial block detachment until final and major failure occurs. The Gebra Slide in particular lies in a highly seismically active area where earthquakes with a magnitude of up to 6.7 have been measured recently (Pelayo and Wiens, 1989; Ibanez et al., 1997; Jin et al., 1998). In addition, young volcanic edifices and lineaments with associated hydrothermal activity occur nearby (Suess et al., 1987; Schlosser et al., 1988; Klinkhammer et al., 1995; Gracia et al., 1996a,b; Bohrmann et al., 1999). The possibility that layers of volcanic ash behaved as weak layers and developed into slip planes in the Northern Antarctic Peninsula region and for the Gebra Slide in particular has been suggested by Imbo et al. (2003). These layers would add to marine clays and oozes as potential mechanical discontinuities favoring landsliding.

The Traenadjupet Slide lies 300 km north of the Storegga Slide and the general setting and boundary conditions are very similar. The shallowest part of the Traenadjupet Slide headwall is just in front of the mouth of a former ice stream whose path is observed on the modern continental shelf (Fig. 7). This location determined the formation of a main depocenter of glacial clays on the area presently occupied by the upper slide scar and likely resulted in the generation of excess pore pressure during glacial times. Glacial and early deglacial high sedimentation rates on the slope promoted instability but, most important, prevented fluid escape from the relatively thin layers (<10 m) of interglacial and interstadial sediments due to the low permeability of the glacigenic clays (Table 8). Layers made of interglacial and interstadial sediments would then behave as weak layers and slip planes. The Traenadjupet Slide could differ from the Storegga and Gebra slides in that the initial failure might be located near the present headwall, although there is also the possibility for it to have occurred downslope from a large escarpment now lying at 1800 m of water depth (Laberg and Vorren, 2000). A second major difference is that Traenadjupet Slide occurred several thousand years after the withdrawal of the ice sheet from mid-Norway, whereas Storegga and Gebra took place at earlier times (see Section 10 below).

The Afen Slide occurred in a low seismicity region and, as with the BIG’95 Slide, similar size and shape slides are buried less than 100 ms below it, thus indicating a recurrence of failures. The Afen Slide headwall and the headwalls of older slides buried beneath are located within a contourite drift deposit comprising very well sorted, low cohesion silty sand. Sub-horizontal slip planes also show evidence of contourite activity and represent similar previous deposits. Sandy contourites are less cohesive than clayey sediments and are also susceptible to liquefaction under dynamic loading, thus having the potential for raising the pore pressure of the surrounding cohesive sediments (Wilson et al., 2003a,b). Contourite units are, therefore, of particular relevance since they have acted as slip planes not only in Afen Slide but also along the North Atlantic margin of Europe, as illustrated by Traenadjupet and probably some of the Storegga phases. Sedimentation rates are, however, very low in the Faeroe-Shetland Channel, where the Afen Slide is, with less than 200 m of Quaternary deposits (Long et al., 2003b). This implies the lowest sedimentation rates amongst all the instability areas studied within the COSTA project, jointly with those from the BIG’95 Slide (Table 8).

Vertical faults occasionally reaching the seafloor, with offsets of several meters, have been identified on high resolution seismic reflection profiles from the Afen Slide area. There is a set of faults with traces that mimic the edges of the slide scar 100 ms below the seafloor. The COSTA project has shown that a distinct BSR has been identified in the Afen Slide area but geophysical evidences and cuttings from an exploration well drilled through it indicate that the reflector represents a diagenetic front associated with opal-A to opal-C transformation (Table 8). In addition to the above-described preconditioning factors, an external localized trigger is required to explain the Afen Slide. Indirect evidence from several authors show that seismicity in the Afen area might have been enhanced because of post-glacial isostatic rebound, as it has likely been the case for the Storegga and Traenadjupet Slides (Muir-Wood, 2000; Stewart et al., 2000). Since the Victory Transfer Zone passes directly beneath the Afen Slide, renewed activity along such a transfer zone could have triggered slope failure (Rumph et al., 1993; Wilson et al., 2003a,b). Nevertheless, historical data and five years of active seismic monitoring with detection capabilities of magnitude 2 have shown that seismicity is negligible in the entire Faeroe-Shetland Channel (Musson, 1998; Ford et al., 2002). Therefore,
it can be concluded that the area of the Afen Slide is presently stable in the short term (Hobbs et al., 1997).

The Canary Slide is one of the world’s best studied debris flows affecting the flanks of an oceanic island. The overall morphology of the failure area suggests removal of a slab-like sediment body that started to disintegrate almost simultaneously with the onset of downslope transport. According to Masson et al. (1998), failure was triggered by sudden loading of the lower island slope by a debris avalanche deposit, known as El Golfo Debris Avalanche that removed part of the island and upper slope. The upper part of the 4700 m high scar left by the debris avalanche is perfectly visible nowadays and forms the 1500 m high El Golfo cliff inshore, one of the most spectacular slide headwalls in the world. The two events, debris avalanche and debris flow, must have been about synchronous because the avalanche fills and disguises the seafloor expression of the Canary Slide headwall, which implies the avalanche deposits were still mobile enough when the debris flow was triggered. Roberts and Cramp (1996) considered both the effects of ground accelerations related to earthquakes and of loading by the debris avalanche, and concluded that the loading mechanism was more likely to trigger a disintegrative failure leading to a debris flow. The avalanche itself would likely have been triggered during an eruptive phase, or shield phase, when the development of the volcanic rift zones on the three-arm shaped El Hierro Island was at a maximum. Tensional stresses accumulate in rift zones during shield building and, when acting on oversteepened piles of recently formed volcanic material, can suddenly lead to large landslides (Urgeles et al., 1997). Sea level changes could also have contributed to triggering instabilities on El Hierro Island (Weaver and Kuijpers, 1983; Masson et al., 1993).

The BIG’95 Slide is the best representative COSTA project landslide occurring in river-fed, siliciclastic, progradational continental slopes. It most likely occurred because of a combination of external and internal factors and a final trigger (Table 8). Such a combination would include enhanced local sedimentation on the upper and mid-slope in association with depocenters from the paleo-Ebro River during lowstands, differential compaction of sediments as related to a volcanic dome beneath the main scar, oversteepening of the margin due to dome intrusion and lowstand presumed high sedimentation, low-to-moderate seismic activity and postglacial sea level rise (Farran and Maldonado, 1990; Grünthal et al., 1999). Gas hydrates are not known in the Ebro margin, but gassy sediments exist in the modern Ebro prodelta and could have existed in former outer shelf and upper slope depocenters. That the BIG’95 area is prone to failure is demonstrated by the vertical stacking of several acoustically transparent, lens-shaped bodies separated by stratified intervals observed in mid penetration high resolution seismic reflection profiles (Lastras et al., 2004c). The final trigger for the main event might have been an earthquake because one earthquake of magnitude 4 to 5 occurs statistically every 5 years, according to the USGS/NEC PDE instrumental records for the last 30 years. Larger earthquakes in the mid and northern Mediterranean Iberian margin are also known from historical pre-instrumental reports and eyewitnesses. The largest earthquake measured in the instrumental (\( M_{\text{wag}}=4.6 \)) epoch took place offshore in May 15, 1995, several miles north of BIG’95 and had the potential to induce significant ground motions. However, except for perhaps fluid escape features, no other newly formed structures or remobilisation events were observed at that time, thus evidencing that the slope is now stable at least for seismic events of that magnitude.

As previously stated, the CADEB mostly represents a pre-failure state where sediment is deformed but not failed yet (Fig. 2A). CADEB is situated in the foredeep/foreland basin of the Apennines and is affected by frequent seismic activity and associated tsunamis (Ciabatti et al., 1987; Royden et al., 1987). The maximum historical shock took place on the July 30th, 1627 in the Gargano Promontory, with an intensity-derived magnitude of 6.1. Recurrence times of 84 and 228 years have been found for both tsunamis of any size and for large events produced by earthquakes of magnitude \( \geq 6.6 \), respectively, after simulations (Tinti et al., 1995 and references therein). The active tectonics of the area is also expressed by the Quaternary uplift of older rocks forming structural highs. Accumulation rates for the late Holocene are the highest (1000–10,000 cm ka\(^{-1}\) with 636 cm ka\(^{-1}\) as averaged maximum) amongst all the study areas and, as a result, a 35-m-thick mud wedge formed over the last 5.5 ka during the present sea level highstand (Correggiari et al., 2001 and Table 8). This 5.5 ka to
present mud wedge, on top of the seaward-dipping downlap surface (maximum flooding surface) acting as weak layer and slip plane, in between Ortona and Gargano, constitutes, strictly speaking, the CADEB. However, sediment accumulation rates have not been linear and noticeable variations have occurred during the late Holocene both in time and space. The main sources of sediment are the Po River and the eastern Apennine rivers with modern discharges of $15 \times 10^6$ and $32 \times 10^6$ t yr$^{-1}$, respectively (Frignani and Langone, 1991; Cattaneo et al., 2003b).

The geostrophic circulation in the western Adriatic Sea favors mud accumulation south of riverine sources parallel to the coast of the Italian Peninsula thus resulting in a shore-parallel depocenter which overall displays crenulations over as much as 40% of its extent. Shearing planes characterize the shallower head region, whereas compressional pressure ridges and mud diapirs become dominant in the toe region, expressing the downward push of the entire muddy wedge. However, nowhere has deformation evolved into disintegration and flow, which could be at least partially attributed to the plasticity of the sediment that can thus accommodate deformation without evolving into failure. The extremely low slope angles (less than a half degree, Table 6) and the presumed escape of fluids, which relieves excess pore pressure generated by the very high sedimentation rates could have played an additional role in preventing disintegration. In addition to gas at very shallow levels, which hardly could have played a role in the deformation processes, gas venting and gas-charged sediments have been reported from other shallow stratigraphic units beneath the late Holocene mud wedge (Conti et al., 2002). A lowering in sedimentation rates for the last century measured from $^{108}$Pb activity offshore Ortona opens the question as to whether the CADEB deformations are relict, slowed down or still active. Comprehensive descriptions and discussions of CADEB and nearby late Holocene muddy prodeltas in the western Adriatic Sea can be found in Correggiari et al. (2001) and Cattaneo et al. (2003a,b).

The small Finneidfjord Slide is the best known of the COSTA slides in terms of preconditioning factors and triggering mechanisms since it occurred in 1996 as previously stated, over an area previously affected by an older slide. That Finneidfjord is prone to sliding is proven by the eight slides known along its shores, which occurred during the last 2000 years. The Finneidfjord area displays intermediate seismic activity with a 5.8 estimated magnitude maximum historical shock dated for the 31st of August 1819. Glacio-isostatic rebound, which has been related to an enhancement of the seismic activity in Northern Europe, is estimated at 3.5 cm yr$^{-1}$. Biogenic gas is known to occur within the sediment, where a stratigraphic unit including a weak layer often caps it. Sedimentation rates are rather high, from 150 to 250 cm ka$^{-1}$. Initial sliding is thought to start at half tide by hydrostatic overpressure along a weak layer. The slide punctured quick clay pockets in the shore ramp and developed as a quick clay slide (Table 3). Changes in groundwater flow were registered some time before the slide event. The spring before the slide was particularly wet and led to high hydrostatic pressure (excess pore pressure) in ground previous to the event. Anthropogenic factors include a nearby main road with heavy traffic causing tremors over the onshore part of the slide and construction of a tunnel with many detonations close to the slide area, and rock debris dumped on shore next to the slide scar which could contribute to changing groundwater flow. The initial slide started below sea level at the steepest part of the fjord slope, about 50–70 m from a highway running parallel to the shore. Afterwards, the slide developed retrogressively landwards and in less than 5 min took away successively the beach below the road, 250 m of the road, one car and a nearby house that sank into the sea. Four people died. Several minor mass movements occurred afterwards but 1 h later everything was quiet again (Janbu, 1996; Longva et al., 2003). The pre-slide beach slope already had a low safety margin, possibly less than 10%, which was easily exceeded because of the combination of natural, and anthropogenic causative factors.

From the detailed descriptions above and from the summary in Table 8, a picture of the preconditioning factors and final triggers for the COSTA instabilities appears with clarity. To what extent these findings can be extrapolated to other failures that will eventually be investigated with the same degree of detail is something that will be unveiled in the near future. Seismic activity is an important factor, even in areas such as passive margins where seismic quietness can be expected. Activation of seismicity because of post-glacial isostatic rebound is a major, widespread
process in high latitude margins. However, there are slides in low seismicity areas too, such as Canary, Afen and to a lesser extent BIG’95. Volcanic activity and volcanic structures could also be relevant, since they may lead to oversteepening during growing phases, induce seismicity, generate ash-rich weak layers, create mechanical discontinuities and cause slope changes either because of volcano growth or crustal overloading.

Gas, and especially gas hydrates and their dissociation, have often been associated with slope instability (Hampton et al., 1996). It must be possible, however, that silica and other diagenetic fronts, still poorly known, could also play a significant role in slope destabilisation. BSRs could both be caused by gas hydrate and diagenetic front boundaries. An important quality of gas is that its absence nowadays does not imply at all that amounts of it were not present when destabilisation occurred, as hypothesized for the source area of the BIG’95 Slide. Shallow gas, essentially of biogenic origin, is common in shallow unstable areas fed by rivers such as CADEB and Finneidfjord. The relationship between pockmarks and other gas escape features and sliding is controversial. While fluid venting would relax excess pore pressure and thus diminish the risk of failure, it could also be an indication that overpressures exist and this would favor failure. It should be also expected that after unloading of part of a slope because of a large landslide, fluid escape should reactivate since the overburden pressure has been dramatically reduced. A similar phenomenon has been hypothesized for the reactivation of volcanic emissions following giant landslides in ocean islands. In fact, active pockmarks are thought to be very rare.

Sedimentation rates and the nature of the sediments are of crucial importance since they determine not only the rate of generation of excess pore pressure, but also their eventual relaxation as a function of the permeability of the sedimentary strata. The “sedimentation factor” plays, in addition, a role over the development of mechanical discontinuities and weak layers. To that respect, climatically controlled strong bottom currents, forming contourite drifts and layers, hold a great potential to behave as slip planes and have been of enormous relevance in some ocean margins such as the eastern North Atlantic. It is important to consider the sources of sediment and the distance of the failed zone from the point or line sources since both parameters control the development of properties that directly relate to instability. In the same way sedimentation is climatically driven, instability also responds to climatic cycles as demonstrated by the repeated occurrence of sliding events at about the same place once a specific set of sedimentary conditions is achieved. This is particularly true for instabilities in glaciated margins, where slide occurrence is controlled by the location of the grounding line and ice streams carrying basal lodgment till to the upper slope. The more we could learn about how climate influences slope failure, the better we will be able to perform sound slide forecasting.

Finally, after the cumulative effect of pre-conditioning factors, a final, most often external, trigger is required for failure to occur. Earthquake activity is by far the most common external trigger invoked to account for the COSTA slides.

9. Ages of slide events

The available information about the ages of the submarine landslides studied within the COSTA project is summarized in Table 8, where information about dating techniques, number of datings, sedimentary units dated and, in some cases, recurrence intervals is provided too. It is best reading Table 8 jointly with Tables 2 and 3 where information on the number and type of sediment cores, and on the retrogressive and/or multi-staged character for each of the landslides can be found. In any case, dates in Table 8 refer to main failure events unless otherwise indicated.

Due to its complexity and significance, dating of the Storegga Slide has been a major task within the COSTA project. Over 100 cores of various types have been collected to date within both major and minor sliding events (Table 2). Dating proceeded in various steps including lithological analyses and detailed stratigraphy of selected sediment cores, handpicking of the foraminifer N. pachyderma at key intervals and age analyses by 14C AMS complemented by tephrochronology, magnetic susceptibility and gamma density. According to the latest results, the main Holocene Storegga Slide occurred approximately at 7300 14C yr BP or 8200 calendar years BP, and is the last of a series of giant slides which succeeded each other at semi-regular intervals during the last 500 ka
Of course, not every type of instability has been considered within COSTA, mostly because of the limited resources and time available to perform the committed researches. While deep glaciated margin settings have been reasonably well covered (five case studies out of eight), in connection with the oil industry interest, river-dominated settings have been insufficiently investigated, and carbonate-dominated settings have not been addressed at all. Other than CADEB, which is not a unique but a rather specific case of muddy detached prodeltaic system, studies of instabilities in shallow prodeltas and delta fronts have not been performed within COSTA. Failure events affecting the levees of deep-sea fans and channel-levee complexes, and those destabilising inner canyon and channel walls in slope settings, which could result in sediment plugs favoring avulsions and formation of neochannels and lobes, wait to be investigated using a common, integrated approach. The well developed carbonate margin segments common in the Mediterranean hold a wide spectrum of mass movement types about which, in addition to their location and general features, little is known.

We intend for the present paper to be not a final ending but an intermediate stage contributing toward bringing consistency and solving some of the key questions formulated during the last 10 years around the topic of submarine landsliding. If this paper helps in clarifying some concepts, in opening new questions and in promoting novel research approaches, the many scientists co-authoring it will feel their task accomplished.

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