Submarine Mass Movements and Their Consequences
TABLE OF CONTENTS

FOREWORD................................................................................................................................... xi

Section 1 - Role of submarine slides in margin development

FRACTAL STATISTICS OF THE STOREGGA SLIDE
A. Micallef, C. Berndt, D.G. Masson, D.A.V. Stow ............................................................... 3

SUBMARINE PALEO-FAILURE MORPHOLOGY ON A GLACIATED CONTINENTAL MARGIN FROM 3D SEISMIC DATA
J.S. Laberg, K. Andreassen ......................................................................................................... 11

SLOPE INSTABILITY AND MASS TRANSPORT DEPOSITS ON THE GODAVARI RIVER DELTA, EAST INDIAN MARGIN FROM A REGIONAL GEOLOGICAL PERSPECTIVE
C.F. Forsberg, A. Solheim, T.J. Kvalstad, R. Vaidya, S. Mohanty ........................................ 19

REPEATED INSTABILITY OF THE NW AFRICAN MARGIN RELATED TO BURIED LANDSLIDE SCARPS
A. Georgiopoulou, S. Krastel, D.G. Masson, R.B. Wynn ..................................................... 29

ALONG SLOPE VARIATIONS IN MASS FAILURES AND RELATIONSHIPS TO MAJOR PLIO-PLEISTOCENE MORPHOLOGICAL ELEMENTS, SW LABRADOR SEA
M.E. Deptuck, D.C. Mosher, D.C. Campbell, J.E. Hughes-Clarke, D. Noseworthy…… 37

SUBMARINE LANDSLIDES ALONG THE NORTH ECUADOR – SOUTH COLOMBIA CONVERGENT MARGIN: POSSIBLE TECTONIC CONTROL
G. Ratzov, M. Sosson, J.-Y. Collot, S. Migeon,
F. Michaud, E. Lopez, Y. Le Gonidec .................................................................................. 47

THE SOUTHERN FLANK OF THE STOREGGA SLIDE: IMAGING AND GEOMORPHOLOGICAL ANALYSES USING 3D SEISMIC
J. Gafeira, J. Bulat, D. Evans .................................................................................................. 57

SUBMARINE MASS MOVEMENTS ON AN ACTIVE FAULT SYSTEM IN THE CENTRAL GULF OF CORINTH
M. Charalampakis, A. Stefatos, T. Hasiotis, G. Ferentinos ............................................... 67

ANALYSIS OF MULTIBEAM SEAFLOOR IMAGERY OF THE LAURENTIAN FAN AND THE 1929 GRAND BANKS LANDSLIDE AREA
D.C. Mosher, D.J.W. Piper .................................................................................................... 77

LANDSLIDE AND GRAVITY FLOW FEATURES AND PROCESSES OF THE NAZARÉ AND SETÚBAL CANYONS, WEST IBERIAN MARGIN
R.G. Arzola, R.B. Wynn, D.G. Masson, P.P.E. Weaver, G. Lastras ..................................... 89
Section 2 - Mass waste evolution: From slump to distal turbidites

EXPERIMENTAL STUDIES OF SUBAQUEOUS VS. SUBAERIAL DEBRIS FLOWS – VELOCITY CHARACTERISTICS AS A FUNCTION OF THE AMBIENT FLUID
H. Breien, M. Pagliardi, F.V. De Blasio, D. Issler, A. Elverhøi .......................... 101

THE GENERAL BEHAVIOR OF MASS GRAVITY FLOWS IN THE MARINE ENVIRONMENT
A.W. Niedoroda, C.W. Reed, H. Das, L. Hatchett .............................................. 111

SUBMARINE SPREADING: DYNAMICS AND DEVELOPMENT
A. Micallef, D.G. Masson, C. Berndt, D.A.V. Stow ........................................... 119

FLOOD-INDUCED TURBIDITES FROM NORTHERN HUDSON BAY AND WESTERN HUDSON STRAIT: A TWO-PULSE RECORD OF LAKE AGASSIZ FINAL OUTBURST FLOOD?
G. St-Onge, P. Lajeunesse .................................................................................. 129

UNDERWATER ROCKFALL KINEMATICS: A PRELIMINARY ANALYSIS
D. Turmel, J. Locat ................................................................................................. 139

ANTHROPOGENIC TURBIDITY CURRENT DEPOSITS IN A SEISMICALLY ACTIVE GRABEN, THE GULF OF CORINTH, GREECE: A USEFUL TOOL FOR STUDYING TURBIDITY CURRENT TRANSPORT PROCESSES
M. Iatrou, G. Ferentinos, G. Papatheodorou, D.J.W. Piper, E. Tripsanas .............. 149

Section 3 - New techniques, approaches and challenges in submarine slope instability analysis

PROBABILITY STUDY ON SUBMARINE SLOPE STABILITY
S. Yang, F. Nadim, C.F. Forsberg ................................................................. 161

MARINE DEEP-WATER FREE-FALL CPT MEASUREMENTS FOR LANDSLIDE CHARACTERISATION OFF CRETE, GREECE (EASTERN MEDITERRANEAN SEA) PART 1: A NEW 4000M CONE PENETROMETER
S. Stegmann, A. Kopf ......................................................................................... 171

Section 4 - Monitoring stress on submarine slopes and sediment physical properties

LINKING GEOTECHNICAL AND RHEOLOGICAL PROPERTIES FROM FAILURE TO POST-FAILURE: THE POINTE-DU-FORT SLIDE, SAGUENAY FJORD, QUÉBEC
J. Locat, P. Locat, A. Locat, S. Leroueil ............................................................. 181
RHEOLOGICAL PROPERTIES OF FINE-GRAINED SEDIMENTS IN MODELING SUBMARINE MASS MOVEMENTS: THE ROLE OF TEXTURE
S.W. Jeong, J. Locat, S. Leroueil, J.-P. Malet ............................................................. 191

MARINE DEEP-WATER FREE-FALL CPT MEASUREMENTS FOR LANDSLIDE CHARACTERISATION OFF CRETE, GREECE (EASTERN MEDITERRANEAN SEA) PART 2: INITIAL DATA FROM THE WESTERN CRETAN SEA
A. Kopf, S. Stegmann, S. Krastel, A. Förster, M. Strasser, M. Irving ......................... 199

RECURSIVE FAILURE OF THE GULF OF MEXICO CONTINENTAL SLOPE: TIMING AND CAUSES
R. Urgeles, J. Locat, B. Dugan ..................................................................................... 209

GEOTECHNICAL CONSIDERATIONS OF SUBMARINE CANYON FORMATION: THE CASE OF CAP DE CREUS CANYON
M. Sansoucy, J. Locat, H. Lee ..................................................................................... 221

Section 5 - Submarine slides in coastal areas, semienclosed seas (fjords, estuaries, gulfs) and lakes

SUBMARINE MASS MOVEMENTS IN THE BETSIAMITES AREA, LOWER ST. LAWRENCE ESTUARY, QUÉBEC, CANADA
G. Cauchon-Voyer, J. Locat, G. St-Onge ........................................................................ 233

SUBMERGED LANDSLIDE MORPHOLOGIES IN THE ALBANO LAKE (ROME, ITALY)
P. Mazzanti, F. Bozzano, C. Esposito ............................................................................. 243

DYNAMICS OF THE DELTAIC CANYON AREA OF THE RV. CHOROKHI, GEORGIA
K. Bilashvili .................................................................................................................. 251

THE 1990 SUBMARINE SLIDE OUTSIDE THE NIDELV RIVER MOUTH, TRONDHEIM, NORWAY
J.-S. L’Heureux, O. Longva, L. Hansen, G. Vingerhagen ........................................... 259

SUBMARINE SLOPE FAILURES NEAR SEWARD, ALASKA, DURING THE M9.2 1964 EARTHQUAKE

THE AD 1881 EARTHQUAKE-TRIGGERED SLUMP AND LATE HOLOCENE FLOOD-INDUCED TURBIDITES FROM PROGLACIAL LAKE BRAMANT, WESTERN FRENCH ALPS
H. Guyard, G. St-Onge, E. Chapron, F.S. Anselmetti, P. Francus ................................. 279
Table of Contents

MORPHOSEDIMENTOLOGY OF SUBMARINE MASS-MOVEMENTS AND GRAVITY FLOWS OFFSHORE SEPT-ÎLES, NW GULF OF ST. LAWRENCE (QUÉBEC, CANADA)
P. Lajeunesse, G. St-Onge, G. Labbé, J. Locat ......................................................... 287

SEDIMENT FAILURE PROCESSES IN ACTIVE GRABENS: THE WESTERN GULF OF CORINTH (GREECE)

Section 6 - Submarine landslides in volcanic island settings

HIGH FREQUENCY SEDIMENT FAILURES IN A SUBMARINE VOLCANIC ENVIRONMENT: THE SANTORINI (THERA) BASIN IN THE AEGEAN SEA
T. Hasiotis, G. Papatheodorou, M. Charalampakis, A. Stefatos, G. Ferentinos ........ 309

SEDIMENT STABILITY CONDITIONS WEST OF MILOS ISLAND, WEST HELLENIC VOLCANIC ARC
T. Hasiotis, G. Papatheodorou, G. Ferentinos ......................................................... 317

Section 7 - Submarine mass movements and tsunamis

MASS WASTING PROCESSES - OFFSHORE SUMATRA
D.R. Tappin, L.C. McNeil, T. Henstock, D. Mosher ............................................. 327

SLOPE FAILURES OF THE FLANKS OF THE SOUTHERN CAPE VERDE ISLANDS
T.P. Le Bas, D.G. Masson, R.T. Holtom, I. Grevemeyer ........................................ 337

TRIGGERING FACTORS AND TSUNAMIGENIC POTENTIAL OF A LARGE SUBMARINE MASS FAILURE ON THE WESTERN NILE MARGIN (ROSETTA AREA, EGYPT)
S. Garziglia, M. Ioulalen, S. Migeon, E. Ducassou, J. Mascle, O. Sardou, L. Brosolo ................................................................. 347

REASSESSMENT OF SEISMICALLY INDUCED, TSUNAMIGENIC SUBMARINE SLOPE FAILURES IN PORT VALDEZ, ALASKA, USA

TOWARDS THE MITIGATION OF THE TSUNAMI RISK BY SUBMARINE MASS FAILURES IN THE GULF OF CORINTH: THE XYLOCASTRO RESORT TOWN CASE STUDY
M. Charalampakis, A. Stefatos, K. Mpourdopoulos, G. Ferentinos ................. 367
# Table of Contents

PROBABILISTIC SMF TSUNAMI HAZARD ASSESSMENT FOR THE UPPER EAST COAST OF THE UNITED STATES  
S. Maretzki, S. Grilli, C.D.P. Baxter ……………………………………………………………………….. 377

ROLE OF SOIL BEHAVIOR ON THE INITIAL KINEMATICS OF TSUNAMIGENIC SLIDES  
A.S. Bradshaw, C.D.P. Baxter, O.-D.S. Taylor, S. Grilli …………………………………………. 387

REVISITING SUBMARINE MASS MOVEMENTS ALONG THE U.S. ATLANTIC CONTINENTAL MARGIN: IMPLICATIONS FOR TSUNAMI HAZARDS  
J.D. Chaytor, D.C. Twichell, U.S. ten Brink, B.J. Buczkowski, B.D. Andrews ……… 395

TSUNAMIGENIC LANDSLIDES IN THE WESTERN CORINTH GULF: NUMERICAL SCENARIOS  
S. Tinti, F. Zaniboni, A. Armigliato, G. Pagnoni, S. Gallazzi, A. Manucci, B. Brizuela Reyes, L. Bressan, R. Tonini …………………………………………. 405

TSUNAMIS GENERATED BY COASTAL AND SUBMARINE LANDSLIDES IN THE MEDITERRANEAN SEA  
G.A. Papadopoulos, E. Daskalaki, A. Fokaefs …………………………………………. 415

Authors Index ……………………………………………………………………………………………. 423

CD-ROM enclosed, containing full colour images which are printed in black-and-white in the book.
FOREWORD

Submarine mass movements and their consequences are of major concern for coastal communities and infrastructures but also for the exploitation and the development of seafloor resources. A tragic example of the vulnerability of coastal communities has been provided by the Indonesian tsunami of December 2004. Since 2005, as part of the scientific community efforts to minimize the impact of such natural disasters, the International Union of Geological Science (IUGS) and the United Nation, Educational, Scientific, and Cultural Organization (UNESCO) have sponsored an International Geoscience Program on Submarine Mass Movements and Their Consequences (IGCP-511). One of the main objectives of IGCP-511 members is to hold bi-annual symposia on these types of marine and coastal geohazards. The first symposium of this series was held in Nice (2003) and the second in Oslo (2005).

This 3rd Symposium on submarine Mass Movements and Their consequences provides an opportunity to review the state of the art in risk evaluation from submarine landslides, deposit characterization and its implication for coastal and offshore development. By bringing together professionals from the industry and academia with a range of different expertise, these proceedings hope to cover the full spectrum of aspects related to subaqueous mass movements and related consequences. The interdisciplinary views gathered in this book, arising from the conference, help identify future challenges, mitigation strategies and better management of the seafloor. To that effect, the Santorini is quite a unique venue for scientists and engineers interested in marine and coastal geohazards.

The book is organized in 7 sections from environmental settings along margins to mass movements and tsunamis. It also brings together our recent knowledge on submarine failure and post-failure analysis and in situ monitoring of stress and geotechnical properties. It also presents recent techniques for either in situ or laboratory analysis. Over the recent years new areas along the coast, fjords and estuaries have been investigated and are reported on herein. Finally, the venue of the symposium at Santorini provided a unique incentive to present various case histories of submarine mass movements and consequences around volcanic islands.

We want to offer special thank to Petra Van Steenbergen of Springer for her cooperation during the preparation of this book.

This series of symposia on Submarine Mass Movements and Their Consequences shall continue in the future to maintain the necessary momentum to keep our community vibrant. We are strongly convinced that this is only by meeting and sharing our views that we can hope for a better understanding and mitigation of the consequences of these catastrophic geohazards. And, for the readers of this book, we only hope that the enthusiasm and dynamism of our scientific community will transpire from the various papers.

Vasilis Lykousis, Dimitris Sakellariou and Jacques Locat
May 15th 2007
Section 1 - Role of submarine slides in margin development
FRACTAL STATISTICS OF THE STOREGGA SLIDE

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Abstract

The statistics of submarine mass movement inventories are poorly characterised in comparison to those of subaerial mass movements. In this study we investigate the aggregate behaviour of the Storegga Slide by carrying out a statistical analysis of its constituent mass movements. By using area as a proxy for mass movement magnitude, we demonstrate that the non-cumulative frequency-magnitude distribution of mass movements within the Storegga Slide is a power law with an exponent of 1.52. The Storegga Slide has the characteristics of a dissipative system in a critical state, where the input of sediment is continuous in the form of hemipelagic sedimentation and glacial deposition, and the output is represented by mass movements that are spatially scale invariant. We conclude that the Storegga Slide may be modelled as a large-scale geomorphic system that exhibits self-organised critical (SOC) behaviour. In comparison to subaerial mass movements, the aggregate behaviour of submarine mass movements is more comparable to that of the theoretical ‘sandpile’ model. The origin of SOC may be linked to the retrogressive nature of the Storegga Slide. Since SOC is an emergent feature, the large-scale behaviour of the Storegga Slide should be autonomous of the smaller-scale elements. A power law distribution also implies that incomplete submarine mass movement inventories may be extrapolated within the limits of power law behaviour, which is important in terms of hazard management.

1. Introduction

Concepts of non-linear dynamic systems, such as scale invariance and the fractal model, provide a powerful approach to the representation of a wide range of geoscientific data (e.g. fluvial systems (e.g. Pelletier 1999), coastal profiles (Southgate and Möller 2000). Scale invariant properties of data inventories are identified by demonstrating a single power law exponent in a frequency-magnitude distribution (Mandelbrot 1983). A power law distribution implies that when we compare the number of events of size $A$ or greater, with the number of events of size $\eta A$ or greater ($\eta$ is an arbitrary factor), the number always differs by the same factor $\eta^{-\beta}$, regardless of the absolute size of the events (Hergarten 2003). A power law distribution can be replaced with other measures of the size of the event (e.g. area, volume and thickness of mass movements are strongly correlated with each other, and a distribution can be converted between variables (Hovius et al. 1997)); thus a power law distribution is free of a characteristic scale and can be described as fractal (Mandelbrot 1983).

The Storegga Slide, located 120 km offshore Norway, is a mega-scale geomorphic system (Figure 1). Like most other submarine slides, the Storegga Slide has been investigated using an engineering approach. In subaerial geomorphology, the statistical characteristics of landslide inventories have become a recent focus of study (e.g. V. Lykousis, D. Sakellariou and J. Locat (eds.), Submarine Mass Movements and Their Consequences, 3–10. © 2007 Springer.
Guzzetti et al. 2002). In comparison, the statistics of submarine mass movement data are still poorly characterised. The extensive coverage and the excellent quality of the acoustic imagery from the Storegga Slide allow us to investigate the aggregate behaviour of the Storegga Slide and carry out a statistical analysis of its constituent mass movements. The objectives of this study are to assess whether the size statistics of the Storegga Slide mass movements exhibit scale invariance, and to explain the origin and implications of such behaviour.

Figure 1. Bathymetric contour map of the Storegga Slide (contour interval of 250 m). The headwalls that were extracted from the bathymetric data set are represented by solid black lines. The arrow indicates the direction of sediment mobilisation. The location of the Storegga Slide is shown in the inset.

2. Method

The study is based on a high resolution multibeam bathymetry data set covering the slide scar from the main headwall down to a water depth of ca. 2700 m (Figure 1). Most of the data have a horizontal resolution of 25 m or better. A mass movement is defined as a single episode of slope failure where sediment moves downslope under the influence of gravity. The area of the mass movement is delineated by a steep scarp at the upslope limit (headwall) and the distal point of the depositional section at the downslope limit. We use mass movement area as a proxy for magnitude. The estimation
Fractal statistics of the Storegga Slide

of the slide area is hindered by the difficulty in defining the boundaries of quasi-simultaneous slides in a retrogressive slope failure. Thus we try to estimate mass movement area using the length of the associated headwalls, which constitute easily identifiable and prominent features located at the upslope limit of the mass movement. Previous studies of the Storegga Slide have estimated the dimensions of sixty-three mass movements (Haflidason et al. 2004). We plot the headwall lengths against the mass movement areas from these published data (Figure 2a). $R^2 = 0.91$ implies a strong statistical dependency between area and length in the form:

$$A = 0.87l^{1.98}$$  \hspace{1cm} (1)

where

- $A$ is the area of mass movement (in m$^2$)
- $l$ is the length of headwall (in m)

We used a suite of geomorphometric techniques to extract the headwalls automatically from the bathymetric data set. A geomorphometric map, which is a parametric representation of a landscape decomposed into its elementary morphological units, was generated for the study area. The technique for producing a geomorphometric map is explained in more detail in Micallef et al. (2007). Headwalls are extracted as one-cell thick lineaments. Since the geomorphometric techniques delineate headwalls at the resolution of the bathymetric data, rather than at the scale at which the study area is being observed by an investigator, the techniques are more accurate than manual digitisation. Using geomorphometric mapping we were able to extract one hundred and five individual headwalls. The extent of a headwall is defined by the section of the headwall where sediment evacuation has occurred perpendicularly to the lineament. The length of each headwall was calculated using a Geographic Information System, and the area of the mass movement associated with each headwall was estimated using equation (1). A cumulative frequency-area graph was plotted for the mass movements. A non-cumulative distribution, defined in terms of the negative of the derivative of the cumulative distribution with respect to $A$, was then derived to enable comparison with previous studies (e.g. Guzzetti et al. 2002).
3. Results

The estimated areas of the mass movements range between 0.27 km² and 1174 km². The data in the non-cumulative distribution can be best correlated with an inverse power function (Figure 3a):

\[ \frac{dN}{dA} = 3900 A^{-1.52} \]  \hspace{1cm} (2)

where

\( N \) is the cumulative number of mass movements with an area > \( A \)

The exponent of this power function is 1.52. The range over which this function is valid is 0.3 – 100 km².

4. Discussion

The inverse power law distribution of mass movement areas, observed over ~2.5 orders of magnitude of the area, is evidence of fractal spatial statistics within the Storegga Slide system. Similar power law distributions have been identified in numerous subaerial mass movements of different types and sizes, occurring in a range of environmental
settings and triggered by a variety of mechanisms (Figure 3). A power law has also
also been detected in other natural phenomena, such as earthquakes (Turcotte et al. 2006).

Explaining the origin of this fractal distribution in geological terms is difficult. The
most prevalent explanation to this behaviour in subaerial environments has been self-
organized criticality (SOC) (Bak et al. 1987). SOC is a property of complex systems
whereby, in spite of heterogeneity at the small-scale of individual elements (e.g.
sediment grains), the large-scale, aggregate behaviour of the system exhibits order in the
form of a fractal distribution. This order is an emergent property of the system, which
occurs through autogenic dynamics and feedback mechanisms (Phillips 1995). In a self-
organised critical system, the “input” is nearly constant and the “output” is characterised
by a series of events. Self-organised critical systems are characterised by three
conditions (Bak et al. 1987): (i) the distribution of the ‘outputs’ is scale invariant; (ii)
the system is in a quasi-stationary (critical) state and (iii) the temporal behaviour of the
system is a 1/f (red) noise.

The Storegga Slide is a dissipative system, where sediment is mobilised or removed
from the slide area in the form of mass movements. The driving force of this system has
been the continuous deposition of glacially-derived material (during glacial maxima)
and hemipelagic sedimentation (during interglacials), for at least the last 3 million
years (Rise et al. 2005). This deposition resulted in a progressive increase in sediment
pore pressure, gravitationally-induced stress and surface slope gradient. Seismicity,
associated to glacially-induced tectonic movements, may constitute another driving
force as it enables the system to exceed thresholds. These are all characteristics of a
system in a quasi-stationary state. The distribution of mass movements within the
Storegga Slide is spatially scale invariant (Figure 3a). On the other hand, we are not
able to demonstrate temporal scale invariance of the mass movements due to a low
temporal resolution of the data. In consideration of the above, we conclude that the
Storegga Slide may possibly exhibit SOC.

SOC can be theoretically modelled using the ‘sandpile’ model, which is a simple
cellular automata model (Bak et al. 1988) (Figure 4). In this model, particles are
dropped randomly and continuously into a square grid of boxes. When a box
accumulates four particles, these are redistributed to the four adjacent boxes. This
redistribution may result in further instability, creating an avalanche. The non-
cumulative frequency-magnitude distribution of these avalanches was shown to satisfy a
power law, with an exponent of ~1 (Kadanoff et al. 1989) (Figure 3d).

The elements of the ‘sandpile’ model can be related to the components of the Storegga
Slide system. The dropping particles represent sediment deposition, the avalanches are
the individual mass movements, and the thresholds are associated with changes in slope
gradient, pore pressure and gravitationally-induced stress. The exponent associated with
the power law distribution of subaerial mass movements is generally >2.2 (e.g. Dai and
Lee 2002; Guzzetti et al. 2002; Malamud et al. 2004), whereas for the Storegga Slide,
the exponent is 1.52. The value of the exponents for mass movements is higher than that
of the theoretical ‘sandpile’ model.
Micallef et al.

within each cell of the grid. When a particle is added to the centre cell in this example, an avalanche of a size of 8 cells is triggered. In the ‘sandpile’ model, the frequency-magnitude distribution of these avalanches is power law.

The difference may be explained by the large number and variety of forces and controls associated with 3D ‘real’ mass movements, in comparison to the simpler 2D ‘sandpile’ model. The consideration of factors such as geological heterogeneity or soil moisture content tends to increase the exponent of frequency-size distributions in landslide models (e.g. Pelletier et al. 1997; Sugai et al. 1994). The fact that the exponent for mass movements within the Storegga Slide is considerably lower than that for subaerial slides could imply that, in comparison to subaerial mass movements, submarine mass movements are less complex and that the dynamics are more comparable to those of the ‘sandpile’ model. Submarine settings are characterised by gentler slopes, consistent geology and morphology over extensive areas (Shepard 1963), and therefore homogeneous boundary conditions. Subaerial settings, in contrast, consist of rougher landscapes where numerous driving forces, such as tectonic uplift and fluvial incision, interact with weathering and variable degrees of saturation, to generate a higher exponent for the power law distribution. An important role may also be played by cohesion. The sediments failing within the Storegga Slide are mainly clays. Mass movements in cohesive sediments were shown to exhibit lower exponents than those occurring in less cohesive material (Dussauge et al. 2003).

Some uncertainties do arise with the applicability of the ‘sandpile’ model to submarine mass movements, however. For example, the ‘sandpile’ model disregards aspects of inertia and cohesion, which are quite important in sliding within the Storegga Slide. SOC is not a sole property of cellular automata models. For example, Hergarten and Neugebauer (1998) developed a model of landsliding that exhibits SOC using partial differential equations. For our study area, another explanation of the fractal distribution of the mass movements and the potential SOC behaviour may be the fact that the Storegga Slide was a retrogressive slope failure (Haflidason et al. 2004). The slide was initiated close to the Faroe-Shetland Escarpment (Bryn et al. 2005). Large mass movements within this region destabilised neighbouring and upslope areas. The development of the Storegga Slide may be likened to a retrogressive cascade, because as the instability propagated upslope via the repeated collapse of the headwall, the mass

Figure 4. The theoretical ‘sandpile’ model based on a 5 × 5 grid. The dots indicate the number of particles within each cell of the grid. When a particle is added to the centre cell in this example, an avalanche of a size of 8 cells is triggered. In the ‘sandpile’ model, the frequency-magnitude distribution of these avalanches is power law.
movements became more numerous (Figure 2b) and smaller (Haflidason et al. 2004). The Storegga Slide extends over most of the continental slope, where topography is smooth, and boundary conditions are homogeneous (Shepard 1963). The extent of the Storegga Slide is in fact limited by changes in boundary conditions at its perimeter, in particular the decrease in slope gradient and the increase in the consolidation of sediments at the continental shelf (Gauer et al. 2005), as well as the presence of the North Sea Fan in the south and the Vøring Plateau in the north. The retrogressive cascade is also qualitatively similar to the activation of avalanches in the ‘sandpile’ model and may explain the fractal distribution of submarine mass movements. Other cascade models, such as the inverse cascade model, have been used to reproduce the self-organized critical behaviour of forest-fires (Turcotte et al. 1999).

The origin of the fractal distribution may also be attributed to factors that are unrelated to SOC. A power law distribution may be the signature of pre-defined geological structures (e.g. Hergarten 2003; Pelletier et al. 1997) or external mechanisms. Since we do not have detailed information about the spatial variation of geological structures within the Storegga Slide, we are unable to confirm the role of geological structures in relation to the observed fractal distribution.

5. Conclusions

Our results have direct implications relating to the modelling of submarine mass movements. SOC is put forward as the most likely origin of the observed power law distribution of submarine mass movements. The Storegga Slide may thus be modelled as a large scale geomorphic system in a critical state, incorporating dynamics of the ‘sandpile’ model. SOC is an emergent property of a system, and thus it is not built into the fundamental physical equations. This means that the aggregate behaviour of the Storegga Slide cannot be modelled using a reductionist approach based on the small-scale elements of the system. This also means, however, that limitations in data acquisition techniques can be circumvented when considering these emergent features. The retrogressive cascade, which is based on an open system where loss of support constitutes the threshold exceeding mechanism, fits the SOC behaviour well and emphasizes the importance of considering the interconnectivity of individual slides. The evolution of a retrogressive cascade on the continental slope, where boundary conditions are generally uniform, would explain the large size of the Storegga Slide. In fluvial systems, SOC has been associated with minimum energy dissipation (Rigon et al. 1994). We are not able to measure energy in a complex system such as the Storegga Slide, so we may only theorise that the Storegga Slide is a geomorphological system operating at the level of minimum energy dissipation, with SOC as an emergent feature. Another application of our results is in hazard management. Frequency-magnitude distribution of mass movements can be used to extrapolate incomplete inventories within the limits of power law behaviour (in our case, for mass movements ranging between 0.3 – 100 km² in area) and thus estimate event magnitude and total number of mass movements.

6. Acknowledgements

We would like to thank Norsk Hydro A.S. for providing the bathymetric data set.
7. References


SUBMARINE PALEO-FAILURE MORPHOLOGY ON A GLACIATED CONTINENTAL MARGIN FROM 3D SEISMIC DATA

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Abstract

The morphology of two lower – middle Pleistocene paleo-slope surfaces within a muddy glacigenic succession was studied on 3D seismic data. The lower surface is characterised by irregular relief. It terminates upslope by an escarpment that represents the upper part of a paleo-slide scar. The slide scar morphology is relatively similar to that of modern slide scars and failure is inferred to have occurred during or after a glacial maximum when the ice reached the paleo-shelf break. A large area immediately outside the paleo-slide scar was affected by sediment creep or sliding, thus the area of unstable sediments extends beyond the paleo-slide scar. The upper surface morphology is dominated by three straight to slightly meandering paleo-channels, at least one of them formed by mass wasting. Together, the two paleo-surfaces exemplify slope morphology that may result from sediment instability on glaciated margins.

Keywords: paleo-slide scar, paleo-channels, glaciated margin, Barents Sea

1. Introduction

Some of the largest submarine landslides have affected high-latitude glaciated continental margins, areas which now receive increased attention because of their hydrocarbon potential. On the Norwegian – Barents Sea – Svalbard margin submarine landslides have resulted in several slide scars including the Storegga (Bugge et al., 1987; Haflidason et al., 2004, 2005), Trenadjupet (Laberg and Vorren, 2000) and Hinlopen Slides (Vanneste et al., 2006; Winkelman et al., 2006). Paleo-slope records from these areas also reveal evidence of similar sized buried slide scars, indicating repeated large-scale sliding (Evans et al., 1996; Kuvaas & Kristoffersen, 1996; Laberg and Vorren, 1996; Solheim et al., 2005; Laberg et al., 2006). Another, less studied result of slope instability on glaciated margins is the development of large canyon – channel systems. Morphological studies of modern canyons have shown that sliding is an important process in their evolution and that they develop over a long period of time from the interaction of several processes which include repeated sliding (Laberg et al., in press).

In this study we have focused on the early – middle Pleistocene interval of a glacigenic depocentre in the south-western Barents Sea, the prograding wedge in front of the Bear Island Trough (Fig. 1). The objective of this study is to describe and discuss the morphology of two close-lying paleo-slope surfaces; the lower displays part of a paleo slide-scar, whilst the upper comprises several large paleo-channels, and to elucidate on the paleoenvironment during their formation.
seismic data and the red dot shows the location of well 7216/11-1. Contour interval is 100 m on the shelf and 500 m on the continental slope and in the deep sea.

2. Geological Setting

The Norwegian – Barents Sea – Svalbard continental shelf has a glacial morphology. Glacial erosion was most pronounced in the transverse troughs where ice streams were located during full glacial conditions when the Fennoscandian and the Barents Sea Ice Sheets extended to the shelf break along most of the Norwegian – Barents Sea shelf (e.g. Vorren, 2003). As a result of this spatially variable glacial erosion, some areas of the continental slope, i.e. the areas in front of the troughs received huge volumes of sediments. These depocenters have been called “Trough Mouth Fans” (Vorren et al., 1989). The largest trough mouth fan along the Norwegian – Barents Sea continental margin is the Bear Island Trough Mouth Fan (TMF) (Fig. 1) which developed due to deposition over repeated glacial episodes.

![Fig. 1. Bathymetric map of the western Barents Sea continental margin. The box outlines the area of 3D seismic data and the red dot shows the location of well 7216/11-1. Contour interval is 100 m on the shelf and 500 m on the continental slope and in the deep sea.](image)
The onset of the Bear Island TMF development has been dated to about 2.75 - 2.3 Ma (Eidvin et al., 1993; Sættem et al., 1992; Mørk and Duncan, 1993). Based on shallow boreholes in the northern outer Bear Island Trough (c. 150 km north of our study area) Sættem et al. (1992) suggested that the lower part of the fan was deposited by “a high sediment input onto a shallow, sand-dominated continental shelf in front of a grounded ice margin”. Where sampled during commercial drilling (mainly cuttings and small sidewall cores), the upper part of the fan comprises clast-bearing muddy sediments inferred to be glacimarine deposits which include Ice-Rafted Material (IRD). Data available from well 7216/11-1 located within the study area indicate that the paleo-slope surfaces studied in the present paper (Fig. 2) are located within muddy glacimarine sediments (Ryseth et al., 2003). Studies of 3D seismic data from the paleo-shelf indicated that grounded ice reached the shelf break from the level of the lower surface that is investigated here and upwards (Andreassen et al., 2004, 2007).

**Fig. 2.** Part of a seismic line showing the stratigraphic position of the lower and upper surface studied. Reflection R7 is the base of the late Pliocene – Pleistocene glacigenic sediments, Base Pleistocene the base of the Pleistocene sediments and R1 an upper regional unconformity of middle Pleistocene age.

### 3. Data Base

This study is based on a commercial 3D seismic data set covering an area of about 2900 km² (Fig. 1). The vertical resolution of the data is approximately 20-25 m. The theoretical limit for the horizontal resolution of 3D seismic data is ¼ of the seismic wavelet (Brown 2003), which here is ~ 20-30 m (using a seismic velocity of 2200 m/s). For the seismic interpretation the GeoFrame Charisma software was used.
4. Paleo-slope Morphology

4.1 THE LOWER SURFACE

The lower surface includes part of the upper paleoslope. It has been mapped from its upslope truncation by a semi-horizontal, slightly westward dipping “topset” reflection, downslope to the marginal high (Fig. 2). The central and northern part of this surface is dominated by irregular relief which terminates upslope by an escarpment (Fig. 3A). In some areas the escarpment is up to 50 ms (TWT) high and easily identified, in other areas it is more subdued probably because the height is below the vertical resolution of the seismic system. The area of irregular relief is separated into a northern and a southern part by a downslope oriented, steep-crested ridge which is up to 4 km wide and 100 ms (TWT) high (Figs. 3A-B). The ridge crest is irregular and dominated by amphitheatre-shaped depressions, except for in the lowermost part where a series of straight to curved lineations are seen (Fig. 3B). In an area north of the ridge a faint meandering pattern can be followed downslope from near the headwall (Fig. 3B).

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Fig. 3. A) Shaded relief time structure map of the lower surface with the vertical scale 2x exaggerated. Part of the headwall and sidewall is shown. R1 = downslope oriented, steep-crested ridge, R2 = a less-pronounced, downslope oriented ridge. Frame outlines Figure 3D. B) Three-dimensional perspective view from the northwestern end of (A). (1) = small-scale irregularities, (2) = area of curved lineations. Stippled lines: areas which show a faint meandering pattern. C) Three-dimensional perspective view from the north-west showing secondary escarpments (arrows). D) Volume amplitude plot and corresponding seismic profile showing a large area of curved lineations immediately south of the slide scar. The area is outlined by the stippled line on the amplitude plot. On the corresponding seismic line it is located between the green (lower surface reflection) and the black stippled lines (see Fig A for location).
Small-scale irregularities are seen, most pronounced near the escarpment (Fig. 3B). Several secondary escarpments are seen within the upper, southern part of the irregular area (Fig. 3C). They occur within the upslope part of a less-pronounced, downslope oriented ridge (Fig. 3A). A small escarpment marks the southern limit of the irregular area. To the south of this escarpment the paleo-slope has a relatively smooth relief. Immediately above this surface the reflections are discontinuous and a volume amplitude plot shows a large area of curved lineations (Fig. 3D), similar to the lower part of the northern ridge (Fig. 3B).

The areas of irregular morphology are inferred to represent the upper part of a paleo-slide scar. The upper escarpment probably forms part of the paleo-headwall while the downslope oriented ridge is an erosional remnant. The amphitheatre-shaped depressions on this ridge are likely the result of smaller-scale mass wasting, whilst the curved lineations in the lowermost part are the result of sediment creep. South of the ridge secondary escarpments occur. Such features are not found north of the ridge, instead a faint meandering signature is seen, originating from near the headwall. This difference may be related to sediment physical properties variations, with the more consolidated sediments south of the ridge more difficult to mobilise into flows. A large area south of the slide scar was probably also affected by sediment creep or sliding, from the present data base it is not possible to discriminate between the two alternatives. This formed a pattern of curved fractures separating rafts or ridges of sediments and was possibly part of the same event that resulted in the formation of the slide scar.

4.2 THE UPPER SURFACE

Stratigrapically, the upper surface is located slightly above the lower in a similar physiographic setting (Fig. 2). Its morphology is dominated by three straight to slightly meandering channels (indicated 1, 2, 3 in Fig. 4A). The upslope part of the southern two channels (1-2) could not be mapped because this part of the paleo-slope has been removed by subsequent erosion. The southernmost channel (1) originates as two channels, then merges into one which keeps its identity as a straight channel downslope (Fig. 4B). The channel is V-shaped, 50 ms (TWT) deep and has a width of about 250 m. The channel is visible on the amplitude plot (Fig. 4D), showing an increased acoustic contrast downslope (Fig. 4D).

The middle channel (2) is slightly meandering (Fig. 4B), it has a depth and shoulder width of about 25 ms (TWT) and 250 m, respectively and has a U-shaped cross-section. It is also well displayed on the amplitude plot (Fig. 4D), indicating contrasting sediments at the bottom of the channel. This could be due to erosion and subsequent deposition of more coarse grained sediments brought downslope from the area of flow origin. The third channel (3), is U-shaped, c. 500 m wide and terminates upslope in a headwall area (Figs. 4A, C). The headwall is amphitheatre-shaped, about 3 km wide, and is incised by second-order channels (Fig. 4C). Downslope from the headwall channel 3 is seen on the amplitude plot, although the contrast is not as clear as for the other channels (Fig. 4D).
5. Discussion

5.1 PALEO-SLOPE PROCESSES

Although only part of the upper slide scar was identified on the lower surface, its morphology is relatively similar to that of modern slide scars. Secondary escarpments downslope of the headwall form the upper boundary of subparallel paleo-surfaces as seen for instance in the upper Trænadjupet (Laberg and Vorren, 2000) and Nyk (Lindberg et al., 2004) Slide scars. This indicates that sediments at different stratigraphic levels were affected by the failure, that the failure may have been initiated at specific stratigraphic levels forming layers of weakness, and that this may have occurred during one major event followed by smaller, secondary events.

The morphology of the upper surface differs from the lower surface, being dominated by paleo-channels of various sizes. One channel terminates upslope in a headwall area with second-order channels. Several second-order channels may indicate channel formation by sliding over a longer period, as seen in modern canyons (Laberg et al., in press).
5.2 FACTORS PROMOTING LARGE-SCALE SLIDING

Within the glacigenic sediments studied, the lower paleo-slope surface represents the oldest level of large-scale sliding. Below this surface, 2D seismic data display mainly acoustically laminated sediments where intervals of single channels, channel systems and small-scale sliding have been shown. So what caused large-scale sliding in this area at this time? Studies of the late Pleistocene succession on the Norwegian – Barents Sea continental margin have shown that sliding events tend to occur during, or immediately after, glacial maximum periods (e.g. Solheim et al., 2005; Laberg et al., 2006).

The advance of an ice sheet to the shelf break results in sediment erosion below the ice. This erosion is most intense beneath fast-flowing ice streams and large volumes of sediment will be deposited in front of these on the upper continental slope. This rapid sediment loading affects the physical properties of the underlying sediments, makes them more prone to failure (Bryn et al., 2005; Laberg et al., 2003; Kvalstad et al., 2005). We therefore suggest that the submarine landslide which resulted in the slide scar partly displayed on our lower surface was the result of increased sediment input to this part of the continental margin. This was probably due to the advance of an ice sheet to or near the shelf break, in accordance with Andreassen et al. (2004, 2007).

5.3 FACTORS PROMOTING CHANNEL DEVELOPMENT ON GLACIATED MARGINS

The upper surface morphology is also related to sediment reworking but why this resulted in channel features and not a slide scar morphology as the lower surface event is not known. The northern channel was probably formed by sliding over a longer period. However, the southern two may have been formed by a similar process as envisaged for the channels identified by Sættem et al. (1992) slightly north of our study area. These channels may have been formed at the margin of ice caps or ice sheets, possibly by meltwater erosion. This interpretation is supported by the fact that they have a more pronounced acoustic contrast (Fig. 4D) compared with the northern; i.e. that they were not formed by the reworking of slope sediments as the northern channel but related to large input of meltwater, introducing more coarse-grained sediments to the channels which caused the acoustic contrast. Thus the upper slope morphology was most likely a result of channel formation both by mass wasting and glacial meltwater erosion.

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7. References


Abstract

The Krishna and Godavari rivers have formed the Krishna-Godavari delta where they enter the ocean. Both the rivers, which drain a significant part of the Indian peninsula, and the delta are influenced by a seasonal sediment supply controlled by the monsoon rains. The delta system receives nearly all its sediment during this annual flushing of the river system.

The slope instabilities on the delta are most likely driven by excess pore pressure induced by rapid sedimentation and therefore reflect sedimentation history and distribution. The overall geometry of delta shows that the Pleistocene subsidence on the inner part of the sub-aqueous delta is associated with toe-thrusts in more distal regions and that the morphology of the delta front is comprised of channel-levee systems with overbank and mass transport deposits. The climatically induced fluctuations in sea level have probably shifted the main depocentre through time with the more distal sedimentation occurring during sea-level lowstands.

This study is based on 2D and 3D-seismic data that has been combined with geotechnical information from boreholes. This investigation demonstrates the role regional geology has for the distribution and timing of mass movements.

Keywords: Submarine landslides, East Indian margin, delta deposits

1. Introduction

The eastern continental margin of India is an Atlantic type passive margin. It is characterized by a relatively narrow continental shelf, which widens toward the inner part of the Bengal Bay (Fig. 1), and a relatively wide area comprising the continental slope and rise. The present study area is located on the upper continental slope and consists two areas about 14 km apart. One with water depths between about 40 and 400 m, the other with water depths between about 300 to 1900 m (Fig. 2).

The Godavari River is the third largest river in India, after the Ganges and Brahmaputra. While the Krishna River drains a basin about two thirds of that of the Godavari, it only has half of the sediment flux. For both rivers a significant part of the annual discharge and therefore the sediment flux occurs during the relatively short monsoonal period. Together the Godavari and Krishna rivers deliver an extensive sediment load (255 million tonnes per year Sharma 2002, Sarin et al. 2002) to this part of the eastern Indian continental margin.
The present day sea level was established about 5000 years ago. Since that time, a near-shore sandy facies has formed bars, shoals and spits, whereas most of the shelf and slope has been covered by muddy deposits. A progradation of the Krishna-Godavari delta has also taken place since about 5000 years ago, with a seaward shift in the position of the shore line up until about 800 years ago (Bruckner, 1988). An important effect of sea level rise was a de-linking of the main rivers from slope valleys (Flood and Piper, 1997; Moscardelli et al. 2006; Piper et al. 1997; Posamentier and Kolla 2003), some of which terminate landward at around the 50 m isobath. Our main objective is therefore to demonstrate that there is a contrast between the history of shallow and deep water mass transport processes through our interpretations of the regional geology provided by 2D and 3D seismic data in combination with borehole information.

2. Seafloor Morphology

2.1. DEEP WATER AREA

The seafloor morphology of the deep water area exhibits several large channel systems and some smaller channels in regions shallower than about 750 m (Fig. 2 and Fig. 3). Between the channels in water depths greater than about 1000 m there are large areas with a hummocky seafloor. While these features shape the seafloor, seismic sections show that they are covered by a drape of acoustically laminated deposits (Fig. 3).

Numerous ENE-WSW trending synsedimentary listric faults offset the seabed in the northern part of the area. Evidence of toe thrust faulting can be seen in the south western and deepest parts (Fig. 2, right hand panel). The faults scarps on the seafloor provide slopes steep enough for slumps etc. to be triggered.

2.1.1. Hummocky seafloor and channel levee complex

The Channel B (Fig. 2) forms a channel levee complex that has developed through several avulsions. The adjacent hummocky seafloor is associated with units that have a chaotic seismic facies and are interpreted by us as overbank and mass transport deposits (Fig. 3). While the laminated levee deposits mostly underlie the mass transport deposits, there is also some inter-fingering on the levee flanks.
There is internal structuring within the mass transport units that demonstrate that they may be the result of several smaller episodes, perhaps inter-layered with or underlain by overbank deposits from the channel(s). The bases of the channels are sometimes associated with high amplitude reflectors (HARs) that probably consist of more coarse grained sediments (Flood and Piper, 1997; Piper et al. 1997). Because these partly underlie the levee deposits, they were probably formed before the channel levee system had matured and expanded into the area, perhaps after an avulsion, i.e. that they are analogous, but not identical to the turbidite deposits later formed at the mouth of the channels in the deep sea (Flood and Piper, 1997; Piper et al. 1997). The HARs having been formed on the slope whereas the deep sea turbidites deposits are on the relatively flat basin floor.

The hummocky seafloor terminates upslope along scarps that are interpreted to be the headwalls associated with the mass transport deposits (Fig. 2). In general there is a surface drape of acoustically laminated sediments across the whole deep water area. In several places this drape has erosional scars that only influence the uppermost sediment layers. These are therefore the youngest features in the area. They take the shape of elongated depressions with sharp upper and side boundaries and follow “valley” floors as would be expected for mass movements under the influence of gravity. The downslope termination is not as distinct and sometimes associated with a small mound. We interpret these features as having been formed by debris flows/slides/slumps and that can be seen to have originated along faults and other scarps or steeper areas on the sea floor. Their relief is about 2-5 m and they usually have quite a significant run-out (Fig. 4a).