

## Chapter 42

# Shallow Gas and the Development of a Weak Layer in Submarine Spreading, Hikurangi Margin (New Zealand)

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**Abstract** Submarine spreading is a type of mass movement that involves the extension and fracturing of a thin surficial layer of sediment into coherent blocks and their finite displacement on a gently sloping slip surface. Its characteristic seafloor signature is a repetitive pattern of parallel ridges and troughs oriented perpendicular to the direction of mass movement. We map ~30 km<sup>2</sup> of submarine spreads on the upper slope of the Hikurangi margin, east of Poverty Bay, North Island, New Zealand, using multibeam echosounder and 2D multichannel seismic data. These data show that spreading occurs in thin, gently-dipping, parallel-bedded clay, silt and sandy sedimentary units deposited as lowstand clinofolds. More importantly, high-amplitude and reverse polarity seismic reflectors, which we interpret as evidence of shallow gas accumulations, occur extensively in the fine sediments of the upper continental slope, but are either significantly weaker or entirely absent where the spreads are located. We use this evidence to propose that shallow gas, through the generation of pore pressure, has played a key role in establishing the failure surface above which submarine spreading occurred. Additional dynamic changes in pore pressure could have been triggered by a drop in sea level during the Last Glacial Maximum and seismic loading.

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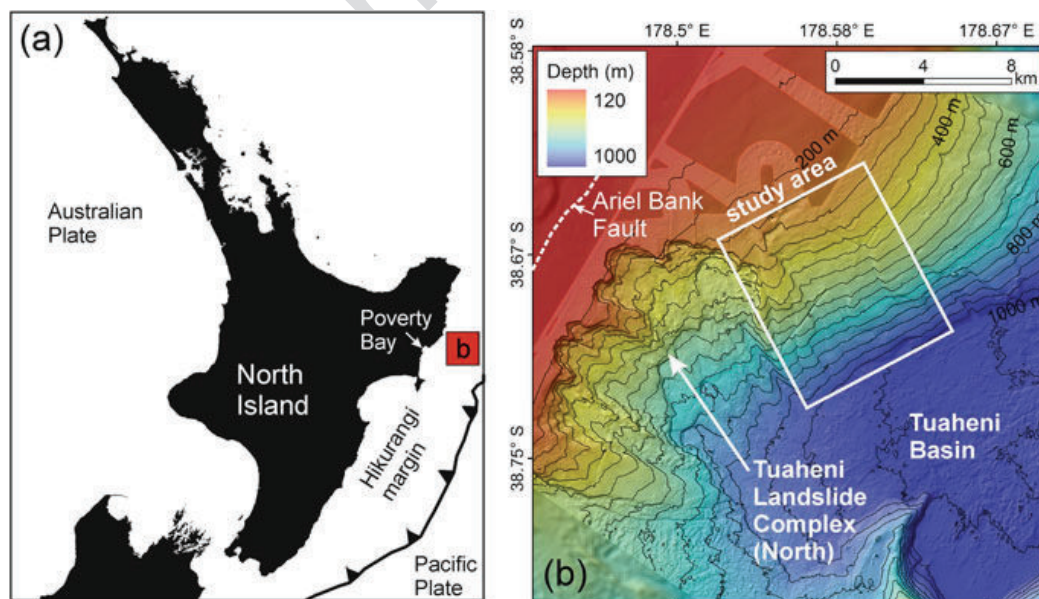
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## 25 42.1 Introduction

26 Spreading entails the finite and downslope surficial displacement of rock/sediment on  
 27 gently sloping ground, and the fracturing of the displaced mass into coherent blocks  
 28 (Varnes 1978). Displacement mostly occurs along a shear zone (Rohn et al. 2004), and  
 29 the deformation may involve subsidence, translation, rotation and disintegration of the  
 30 upper coherent units (Dikau et al. 1996; Varnes 1978). The ground deformation  
 31 associated with spreading comprises the extensional fissuring of the surface units in  
 32 the form of alternating ridges and troughs (Dikau et al. 1996). The literature on  
 33 spreading is not as extensive and exhaustive as for other types of mass movement,  
 34 and little is known about the mechanics of the failure process. Deformation in a spread  
 35 is known to be driven by a combination of transient and static shear stresses, attributed  
 36 to a loss of shear strength of the underlying sediment, which allows the overlying  
 37 material to slide downslope as intact blocks. The geological conditions conducive to  
 38 spreading are usually those where consolidated rocks or sediments overlie a ductile  
 39 substratum (Dikau et al. 1996; Rohn et al. 2004). In terrestrial environments, spreading  
 40 is inextricably linked to the build up of pore pressure and associated liquefaction,  
 41 which may occur in shallow underlying deposits either during an earthquake or due to  
 42 changes in the height of the water table (Kanibir et al. 2006).

43 In submarine settings, numerical and mechanical models have indicated that,  
 44 similarly to terrestrial environments, an increase in pore pressure may be a key  
 45 preconditioning factor and trigger of spreading (Kvalstad et al. 2005; Micallef  
 46 et al. 2007). In this paper we address the hypothesis that, by influencing pore pressure  
 47 in sub-seafloor sediment, shallow gas can promote the development of a weak layer  
 48 above which submarine spreading can occur. We do this by analysing geophysical  
 49 data acquired from offshore the east coast of North Island, New Zealand (Fig. 42.1).



**Fig. 42.1** (a) Location map. (b) Bathymetric map of the continental slope offshore Poverty Bay, showing location of study area. Isobaths at 50 m intervals

In comparison to terrestrial spreading, submarine spreading has received very little attention. First reported from offshore California (Field et al. 1982), most of what we know about submarine spreading comes from studies of the Norwegian passive continental margin (Baeten et al. 2013; Gauer et al. 2005; Kvalstad et al. 2005; Micallef et al. 2007, 2009). Nevertheless, the characteristic submarine spreading morphology, in the form of a recurring pattern of ridges and troughs, can be observed in numerous submarine landslides around the world (Lastras et al. 2003, 2006; Micallef et al. 2013; Piper et al. 1999; Vanneste et al. 2006). This means that submarine spreading is a widespread type of mass movement that has played an important role in the development of submarine landslides in different settings, and which therefore merits more detailed investigation.

## 42.2 Study Area

Our study area is located on the upper slope of the Hikurangi margin, 45 km east of Poverty Bay, North Island, New Zealand (Fig. 42.1). The east coast of the North Island straddles the boundary between the Pacific and Australian tectonic plates. This margin is characterised by the westward subduction of the Pacific Plate beneath the North Island, at a rate of about 4.5–5.5 cm year<sup>-1</sup> (Beavan et al. 2002). Across the continental shelf in the region of our study area, active eastward verging splay faults from the plate boundary mega-thrust are known to project to the seafloor (Mountjoy and Barnes 2011). On the mid- to upper-slope, however, there is a lack of active tectonic deformation, which results in a relatively simple facies geometry. The upper continental slope of the Hikurangi margin is underlain by Miocene to recent slope basin sequences with possible Cretaceous and Paleogene sedimentary rocks at depth (Barnes et al. 2002; Mountjoy and Barnes 2011). Overlying these sequences at the shelf break are lowstand clinoforms deposited during the Quaternary glacial cycles (Barnes et al. 2002; Pedley et al. 2010). These deposits are formed of gently dipping, parallel-bedded clay, silt and possibly sandy sedimentary units (Alexander et al. 2010). Modest size (0.01 km<sup>3</sup>) to very large (3,000 km<sup>3</sup>) submarine landslides have occurred on the Hikurangi margin (Barnes et al. 2010; Kukowski et al. 2010). Some of the best preserved examples of these occur on the upper continental slope directly off Poverty Bay and to the south-west of the study area (Fig. 42.1b). These include the ~30 km<sup>3</sup> Poverty Debris Avalanche, and the ~10 km<sup>3</sup> Tuaheni Landslide Complex (Mountjoy et al. 2009).

## 42.3 Data and Methods

Our study is based on two types of data. The first is a multibeam echosounder dataset covering 700 km<sup>2</sup> of seafloor (Fig. 42.1b). These data were acquired using a hull-mounted Kongsberg EM300 multibeam system during two cruises (TAN1114

87 in 2011 and TAN0810 in 2008). The bathymetry data were processed with CARIS  
88 Hydrographic Information Processing System (HIPS) by accounting for sound  
89 velocity variations, tides and basic quality control. A bathymetry grid with  
90  $25 \times 25$  m bin size was derived. The second dataset comprises high resolution 2D  
91 multichannel seismic reflection data acquired during the TAN1404 cruise in 2014  
92 (Fig. 42.2). The acquisition system entailed a 0.7 t GI Gun and a 150 m long  
93 streamer with 96 channels. Processing included crooked line common midpoint  
94 (CMP) binning (CMP spacing of 1.5 m), frequency filtering (Butterworth filter with  
95 low-cut corner frequencies of 25 and 55 Hz), normal move-out correction, stacking  
96 and 2D Stolt migration. All cruises were carried out on board the *R/V Tangaroa*.

## 97 **42.4 Results**

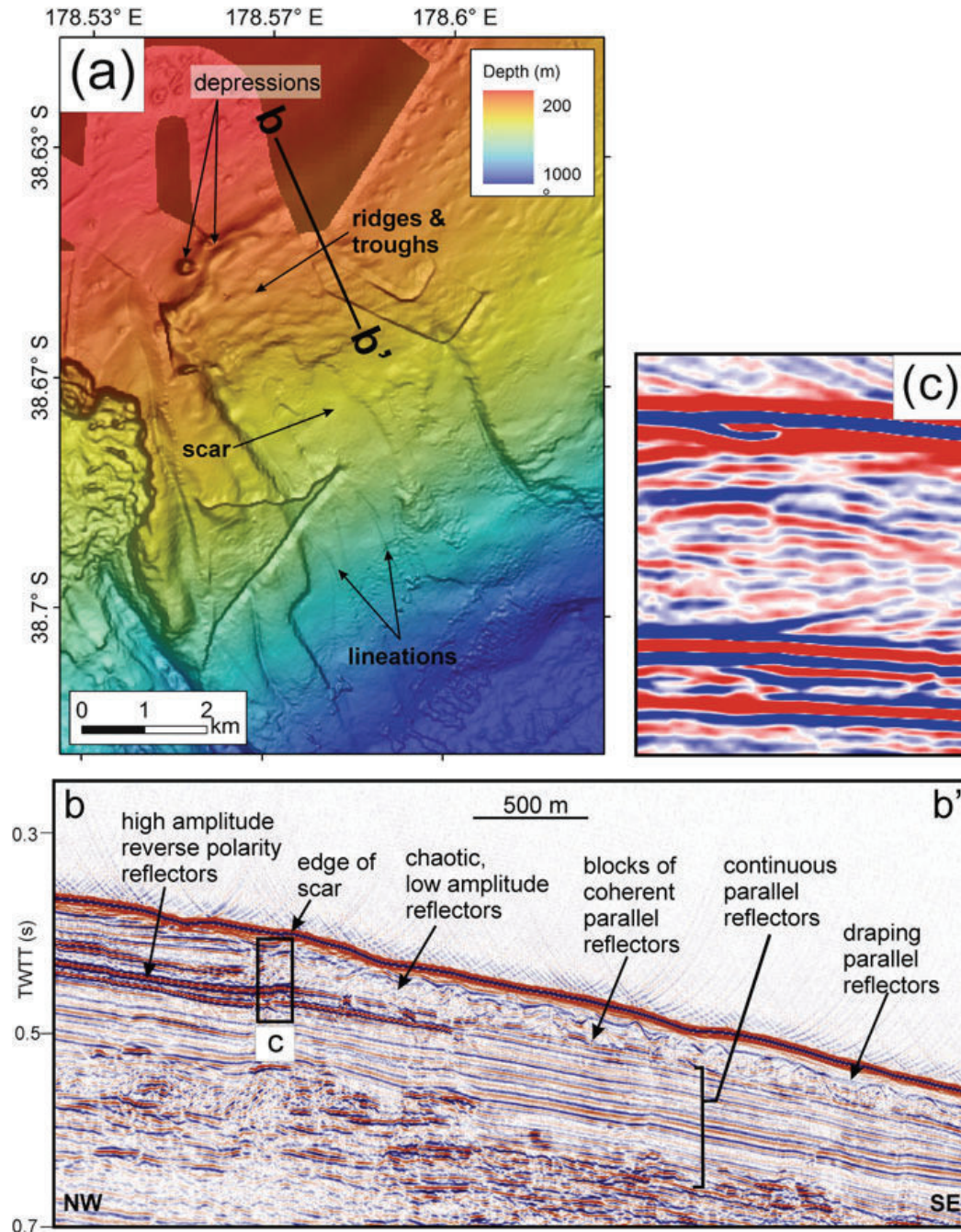
### 98 **42.4.1 Morphology**

99 The continental slope within the study area has an average slope gradient of  $5.5^\circ$   
100 towards SSE. The morphology is dominated by an elongated scar with a length of  
101 8 km, width of 4 km, and 60 m depth (Fig. 42.2a). The downslope limit of the scar  
102 coincides with the regional base of the continental slope where it is contiguous with  
103 the Tuaheni sedimentary basin, at 975 m depth. Its headwall is located in the upper  
104 continental slope, at a depth of 250 m. Smaller scars, sharing a similar morphology  
105 and distal limit, are located 1 km to the north-east of the elongated scar. The seafloor  
106 morphology across the upper section of the scar predominantly consists of a sub-  
107 dued, repetitive pattern of ridges and troughs oriented parallel to the isobaths. In the  
108 downslope section of the scar, the morphology is smoother and intersected by  
109 lineations that are up to 3 km long, 5 m deep, and oriented perpendicular to the  
110 isobaths. These lineations and the western boundary of the scar are intersected by a  
111 4.5 km long and 20 m high SW-NE oriented escarpment. Circular depressions that  
112 are up to 200 m wide and 30 m deep are located at the headwall of the scar.

### 113 **42.4.2 Sub-seafloor Architecture**

114 The seismic expression of the sub-seafloor in the study area comprises a sequence  
115 of continuous, parallel, gently-dipping seismic reflectors that is at least 150 m thick  
116 in places (assuming a seismic P wave velocity of  $1600 \text{ m s}^{-1}$  for depth conversion).  
117 Two reflectors within this sequence are characterised by high amplitude and reverse  
118 polarity, and are recorded at an average depth of 55 m below the seafloor  
119 (Fig. 42.2b). These high amplitude reflectors occur across the upper continental  
120 slope, but are either significantly weaker or entirely absent where the elongated scar  
121 with the ridge and trough morphology is present. Here, the upper part of the seismic  
122 sequence is generally characterised by a unit of irregular, chaotic, low amplitude





**Fig. 42.2** (a) Bathymetric map of study area draped on a slope gradient map and showing principal morphologic elements of the scar. (b) Seismic reflection profile P3106 across the scar. (c) Enlarged section of profile P3106 showing reverse polarity of high amplitude reflector

reflectors, although triangular blocks of coherent, parallel, downslope-dipping 123  
 reflectors are visible. This unit has a variable thickness, with a maximum of 124  
 40 m. The base of this chaotic unit is a planar reflector that connects with the 125  
 high amplitude reflector further upslope. The chaotic unit is also covered by a 126  
 draping unit of parallel reflectors, which is characterised by irregular thickness and 127  
 reaches a maximum thickness of 35 m. 128

## 129 **42.5 Discussion and Conclusions**

### 130 **42.5.1 Nature of Mass Movement**

131 The downslope-dipping, parallel seismic reflectors across the study area are  
132 interpreted as layers in a stratified sediment package (Mountjoy et al. 2009). We  
133 infer that the material in this package is similar to that which failed in the adjacent  
134 Tuaheni Landslide Complex. This consisted of muddy sedimentary deposits, which  
135 accumulated during periods of eustatic sea level lowering, overlain by a Holocene  
136 hemipelagic succession (Carter and Manighetti 2006; Paquet et al. 2009). The ridge  
137 and trough morphology, and the chaotic seismic sequence with isolated blocks of  
138 coherent reflectors, are signature characteristics of submarine spreading (Micallef  
139 et al. 2007). We therefore interpret the upper section of the elongated scar  
140 documented across the study area as evidence of a submarine spreading event that  
141 comprised thin, extensional deformation of the lowstand units, occurring along  
142 stratigraphic surfaces, and which was later draped by Holocene sedimentation.  
143 This mode of failure corresponds to model 2 proposed by Micallef et al. (2007) for  
144 the Storegga Slide, where a thin slab ruptures under tension into a series of coherent  
145 blocks that translate and tilt downslope along a quasi-planar failure plane. The  
146 downslope section of the scar has undergone a higher degree of sediment evacuation,  
147 likely a result of translational sliding or more plastic deformation. The lineations  
148 may correspond to furrows eroded by debris flows into the failure surface.

### 149 **42.5.2 Role of Shallow Gas**

150 Limit equilibrium modelling by Micallef et al. (2007) showed how spreading can be  
151 pre-conditioned or triggered by three processes – loss of support, increase in total  
152 weight upslope, and an increase in pore pressure. Loss of support is a potential  
153 trigger of spreading in the region because of sediment evacuation in the downslope  
154 section of the scar. We exclude increase in total weight upslope as a potential cause  
155 because there are no indications of loading of sediment from a slope failure in the  
156 seismic data. An increase in pore pressure is also a likely cause of spreading in the  
157 study area. We interpret the high amplitude and reverse polarity reflectors in  
158 Fig. 42.2b as the top of an accumulation of gas within the sediments. The circular  
159 depressions, which we interpret as pockmarks, provide additional evidence of  
160 sub-seafloor overpressure. We are not able to determine whether the gas has bio-  
161 genic or thermogenic origin. In bubble phase, gas is known to markedly increase the  
162 pore pressure, which decreases the effective stress of the seafloor sediment, creating  
163 weak layers that are prone to failure (Crutchley et al. 2010; Field 1990). In our study  
164 area this effect is enhanced by the low permeability of the fine-grained material that  
165 failed. The absence or low quantities of shallow gas in the elongated scar, and the  
166 correspondence of the depth of failure with that of the shallow gas, indicate that the

latter has played a key role in establishing the failure surface. Additionally, dynamic changes in pore pressure may have been triggered by two factors. The first is a drop in sea level during the Last Glacial Maximum, with an associated reduction in effective stress as gas came out of solution due to lower hydrostatic pressures. This could explain why the absence of gas is more noticeable in the deeper part of the slope, where the reduction of the hydrostatic pressure would have been more pronounced. The second factor is seismic loading. The active fault most proximal to the study area is the Ariel Bank Fault (Fig. 42.1b), which is inferred to have a late Quaternary displacement rate in the range of 3.0–6.5 mm year<sup>-1</sup> (Barnes et al. 2002). Peak ground acceleration, estimated from probabilistic seismic hazard modelling of regional earthquake sources, is in the order of 0.3–0.4 g with a return time of 475 years (Stirling et al. 2002). Some moderately large magnitude historic earthquakes have also occurred in the vicinity (e.g. the 1931 M7.8 Napier earthquake). The escarpment crossing the western boundary of the scar and the lineations may also be interpreted as the location of a recently-active fault.

To evaluate the potential failure mechanisms and perform stability analyses that take into account the role of gas charging and seismic loading, there is the need to acquire long sediment cores and carry out in situ geotechnical measurements from the study area.

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