Paleoseismicity of the continental margin of eastern Canada: Rare regional failures and associated turbidites in Orphan Basin

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ABSTRACT

The eastern Canadian continental margin is a typical glaciated passive margin where historic earthquakes have triggered submarine landslides. This study compares seismological estimates of earthquake recurrence with the geological record over the past 85 k.y. offshore of Newfoundland to assess the reliability of the geologic record. Heinrich layers in cores provide chronology at ~3–5 k.y. resolution in high-resolution seismic-reflection profiles across headscars and mass-transport deposits. Landslide-generated turbidites on the basin floor have distinctive petrology, sedimentology, and distribution, with ~1 k.y. chronologic resolution. Large slope failures occurred synchronously over margin lengths of 50–300 km. Since 85 ka, four failures have affected a >150-km-long sector of the slope and 18 failures were large enough to be recognized in seismic-reflection profiles and/or cores. The widespread failures were earthquake triggered; other mechanisms for triggering laterally extensive synchronous failure do not apply. A frequency-magnitude plot of length of failed slope was calibrated by the published relationship that an order-of-magnitude increase in failed slope length corresponds to two orders of magnitude of earthquake energy, together with the published estimate of $M_w = 8.0$ for the largest earthquake on the Canadian eastern continental margin. Mean recurrence interval of $M = 7$ earthquakes at any point on the margin is estimated at 25–30 k.y. from both seismological models and the sediment failure record. On such a margin with modest sedimentation rates (~0.3 m/k.y.) and low seismicity, sediment failures provide a robust estimator of past seismicity.

INTRODUCTION

The record of past seismicity can be inferred, under some circumstances, from a stratigraphic record that contains a proxy record of seismic activity (Adams, 1990). Submarine failures result from a range of preconditioning factors (Mosher et al., 2010a). On formerly glaciated passive continental margins, sediment failures in stratified proglacial and contourite drift sediments have been generally ascribed to seismic triggering, as in the case of the Storegga (offshore Norway; Kvalstad et al., 2005) and Grand Banks (offshore Newfound-land, Canada; Dowsee, 1948) slides. Smaller infrequent shallow failures that appear synchronous in multiple valley systems on the continental slope are also most likely the result of seismic triggers (Jenner et al., 2007). In situ geotechnical measurements confirm that most continental margin sediments on such formerly glaciated margins, except on slopes of more than a few degrees, are stable and would require seismic accelerations associated with a large earthquake (likely magnitude $M = 5$ or greater) to fail (e.g., Mosher et al., 1994; Nadim et al., 2005). However, failure may be triggered by erosional or tectonic steepening of slopes (e.g., Lamarche et al., 2008) or may occur as a result of loading failure on weak layers, particularly in areas of high sedimentation rate in active deltas (Clare et al., 2016). Factors that precondition slope sediments to be more liable to failure include high pore pressures due to high sedimentation rates, dissociation of gas hydrates by increased bottom temperature or sea-level fall, or leakage of deep fluids (Masson et al., 2006). However, several studies have concluded that severe earthquakes are the likely triggers for failure on slopes of less than a few degrees (e.g., Nadim et al., 2005; Locat et al., 2009; Goldfinger, 2011), especially in areas of modest or low sedimentation rates. Severe historical earthquakes (with peak ground acceleration >0.3 g) have generally resulted in shallow retrogressive failures affecting the upper few tens of meters of sediment (e.g., Papatheodorou and Ferentinos, 1997; Piper et al., 1999), but in cases where a buried weak layer is present, retrogression may affect a much thicker section (Kvalstad et al., 2005; Mosher and Piper, 2007). In areas of high sedimentation rates, such as many deltas, failure may be triggered by processes unrelated to earthquakes, particularly in shallow water where wave and tidal effects on pore pressure are important (Hampton et al., 1996; Flemings et al., 2008).

The eastern Canadian continental margin was the site of the $M = 7.2$ Grand Banks earthquake in A.D. 1929 and the $M = 7.4$ A.D. 1933 Baffin Bay earthquake. Both epicenters were located beneath the continental slope, within a 6000-km-long zone corresponding approximately to the continental slope zone of Mazzotti and Adams (2005). This estimated frequency is incorporated in the 2005 Building Code of Canada (Adams and Halchuk, 2003). The frequency of large submarine landslides...
through the Quaternary in the vicinity of the 1929 Grand Banks earthquake and landslide, where paleotectonic structures have probably concentrated earthquakes (Mazzotti, 2007), suggests that earthquakes with effects similar to those in 1929 have occurred approximately once every 50–120 k.y. (Piper et al., 2005).

The purpose of this study is to investigate the relationship between seismological models and the record of sediment failure on the formerly glaciated eastern Canadian continental margin. We use a particularly well-known area of the continental margin, in Orphan Basin off northeastern Newfoundland, where there is an excellent stratigraphic record of large submarine failures on the continental margin. The work is based on extensive high-resolution (sparker) seismic-reflection profiles (resolution in the order of 0.3–0.4 m), with ground truth provided by a suite of piston cores. We use previously published chronologic control back to 85 ka based on Heinrich event stratigraphy calibrated by radiocarbon dating (Tripsanas and Piper, 2008a) that is correlated to the reference section from Integrated Ocean Drilling Program (IODP) Site U1302 (Channell et al., 2012; Mao et al., 2018). This chronologic control is used to date observed slope failure surfaces and overlying mass-transport deposits in seismic-reflection profiles. Basin-floor turbidites are classified as to whether they were linked to failure events (Hampton, 1972; Tripsanas et al., 2008) or generated by ice-margin processes (Tripsanas and Piper, 2008b). Only the former are used to assess the record of slope failure. Limit equilibrium slope-stability analysis using the known geometry of the slopes and strength and density data from piston cores is reported for typical continental slope areas that have intermittently failed. We present arguments that the record of failures is a record of large-earthquake triggering. We compare this record with that elsewhere on the eastern Canadian margin. Finally, we compare the paleoseismic record with estimates of earthquake frequency based on historical records and strain rates, in order to draw conclusions about the reliability of the paleo-failure record for estimating paleoseismicity.

### GEOLOGICAL SETTING

**Morphology of the Continental Margin**

Orphan Basin is a bathymetric embayment in 2000–3000 m water depth on the northeastern Newfoundland continental margin. It is bounded to the west, east, and southeast by the Grand Banks, Orphan Knoll, and Flemish Cap, respectively (Fig. 1). At times when the Laurentide Ice Sheet extended onto the continental shelf, western Orphan Basin was supplied with sediment through an ice stream in Trinity Trough. Glacialic debris-flow deposits accumulated on the floor of the basin during glacial maxima, forming the Trinity trough-mouth fan (Fig. 2; Tripsanas and Piper, 2008b). Both at glacial maxima and during ice retreat, subglacial meltwater from ice streams in Trinity Trough and Notre Dame Trough to the north supplied fine-grained sediment plumes that were transported southeastward by the Labrador Current (Mao et al., 2018).

In contrast, glacial ice extended only across the inner Grand Banks at the last glacial maximum (Shaw et al., 2006).

The regional morphology of the continental slope is described by Mosher et al. (2007). Specific to Orphan Basin, four geomorphic areas are recognized on the continental slope (Fig. 2): the north slope, the slope off Trinity Trough, the south slope, and the slope off Sackville Spur. The north slope has gradients of 1.5°–2.5° between 400 and 600 m water depth. The slope off Trinity Trough is smooth and gently inclined (1°–1.7°). In contrast, the south slope has regional inclinations of 2°–4° between 300 and 1500 m water depth, and is highly dissected by multiple canyons, with relief of as much as 500 m and local slopes of as much as 20°. These canyons are organized into two systems, the Sheridan and Bonanza canyon systems. The slope off Sackville Spur is also steep (2°–3° from 1000 to 2500 m water depth) and is dissected by multiple shallow gullies with a relief of 30–120 m that are separated from each other by depositional overbank areas. Canyons and gullies on the Grand Banks margin pass downslope into basin-floor channels.

**Tectonic Setting and Modern Seismicity**

Orphan Basin formed initially during Early Cretaceous rifting of Iberia from North America, with crustal hyperextension accentuated by the southeastward translation of the Flemish Cap microplate (Sibuet et al., 2007). Oceanic spreading extended northwards to the Charlie Gibbs fracture zone by the Late Cretaceous (Aptian–Santonian; Keen et al., 2014).

The short instrumental historical record of seismicity (Fig. 1) shows a cluster of earthquakes near the 1929 Grand Banks hypocenter and beneath the Laurentian Fan. Otherwise there is a sparse record of small earthquakes on the southeastern Canadian continental margin with little relationship to major tectonic lineaments such as the Charlie Gibbs fracture zone in oceanic crust or along the Cumberland belt at the southern margin of Orphan Basin (Enachescu, 1987). The short length of the instrumental record means that it is a poor indication of regional seismic hazard (Mazzotti, 2007).

**Late Quaternary Sediments and Sedimentation Processes**

The slope and floor of Orphan Basin are characterized by multiple and widespread tongues of mass-transport deposits (MTDs) and turbidites. The MTDs include the glacigenic debris-flow deposits that extend from the slope off Trinity Trough to the western basin floor and are capped by channels cut by hyperpycnal subglacial meltwater discharge (Tripsanas and Piper, 2008b). The near-seabed deposits date from the last glacial maximum in the area, from 28 to 20 ka. Glacial meltwater also created hypopycnal muddy plumes that drifted southeastward in the Labrador Current, depositing red-brown muds at sedimentation rates of 0.3–0.5 m/k.y. between 28 and 15 ka on the south slope out to water depths of 2300 m (Tripsanas and Piper, 2008a). Episodic Late Pleisto-
cene and Holocene failures on the south slope transformed into turbidity currents that flowed through channels in the southern part of the basin, most recently in the Sheridan failure at ca. 7 ka (Tripsanas et al., 2008). Interbedded with all of these sediments are thin detrital carbonate–rich beds, termed Heinrich layers, which correlate with episodic discharge of glacial meltwater from the Hudson Strait ice stream (Rashid et al., 2003; Channell et al., 2012).

Regional Stratigraphic Control

Detrital carbonate-rich beds deposited during Heinrich events provide the basis for widespread lithostratigraphic correlation, augmented by a distinctive succession of red-brown muds deposited from meltwater plumes. The precise identification of particular Heinrich events younger than 40 ka is based on numerous previously reported calibrated radiocarbon dates on the planktonic foraminifera *Neogloboquadrina pachyderma* sinistral (Tripsanas and Piper, 2008a, 2008b). The greater bulk density of the detrital-carbonate beds interbedded with muddy terrigenous sediment commonly results in corresponding high-amplitude reflections in sparker seismic-reflection profiles (resolution in the order of 0.3–0.4 m) where sedimentation rates are high (Fig. 3). Age control on the older section is provided by the presence of North Atlantic ash zone II just below Heinrich event H5a in several cores and by the isotope stratigraphy of core MD95-2025 (Fig. 2) (Hiscott et al., 2001) and IODP Sites U1302 and U1303 (Channell et al., 2012). We use the chronology of Heinrich events summarized by Olsen et al. (2014). For the age of the top of the detrital-carbonate interval in the Heinrich 1 event, we use the age of 16.3 ka of Gil et al. (2015).

Seismic markers for the past 85 k.y. correlate regionally throughout Orphan Basin (Fig. 4), as documented by Tripsanas et al. (2007). Older markers that correlate to mid-Pleistocene marine isotope stages (Campbell, 2005) are generally too deep to be imaged in most sparker profiles.
High-resolution seismic-reflection data were acquired by the use of a Huntec DTS 480 J sparker system, towed at ~100 m below sea level (Mosher and Simpkin, 1999). A total of 85 cores were collected from Orphan Basin using the AGC Long Corer (Driscoll et al., 1989), which recovered cores up to 12 m long, and a smaller Benthos piston corer, which recovered cores <5 m in length. Trigger-weight gravity corers were used for the better recovery of the upper metre of sediment. High-resolution seismic-reflection data were acquired by the use of a Huntec DTS sparker system, towed at ~100 m below sea level. Power output was 480 Joules at 4 kV, fired at 1-1.5 s, and a 7.3 m-long, 24 element (AQ-16 cartridges), oil filled streamer towed behind the sparker system, towed around 100 m below sea level. Flows were directed at 0.03-0.05 m/s and recorded in digital format and analogue hard copy on an EPC 9800 thermal chart at a 0.25 s sweep.

Methods

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Supplementary Data: Methods

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Supplemental Data. Details on methods. Please visit https://doi.org/10.1130/GES2001.S1 or access the full-text article on www.gsapubs.org to view the Supplemental Data.
The floor of the basin (Fig. 7) is also characterized by widespread sub-parallel reflections. These are also interrupted by rare MTDs, some of which originate on the south slope and others that are glacigenic debris-flow deposits derived from Trinity Trough, described by Tripsanas and Piper (2008b). Channels on the basin floor are pathways for turbidity currents carrying sand and gravel and may also contain MTDs. A buried channel near the bedrock high southwest of Orphan Knoll (Fig. 8) contains a record of at least two MTDs.

Cores

Cores on the basin floor commonly contain very fine-grained sand and silt turbidites interbedded with muds. These appear temporally related to the rapid supply of glacial meltwater plumes to the basin, and their red-brown color indicates a source from Trinity Trough (Tripsanas and Piper, 2008a). Downflow from the glacialic debris-flow deposits on the Trinity trough-mouth fan are sandy turbidites that correlate with erosional surfaces on the fan interpreted as resulting from infrequent hyperpycnic flow of meltwater (Tripsanas and Piper, 2008b). These turbidites consist of normally graded sand beds capped by red-brown silts and muds (Fig. 9C). Although the mean grain size decreases upwards, such beds are well sorted and have a very small mud component except in the silt and mud cap (Figs. 10A, 10C).

A second type of turbidite is found in channels draining from the south slope and in a few cores elsewhere in the basin; it consists of massive gravel and sand (Fig. 9A, upper bed), or a bed of massive sand overlain by massive sandy gravel (Fig. 9A, lower bed) that in more distal settings is capped by a graded medium- to fine-grained sand at the top of the bed (Fig. 9A, lower bed). More proximal cores include gravelly diamicton (Fig. 9B). Proximal sands and gravels are poorly sorted and include a fine muddy “tail” on a grain-size distribution curve (Figs. 10B, 10D). Tripsanas et al. (2008) showed that such a turbidite was deposited above an erosional surface during a large regional failure on the south slope at ca. 7 ka and illustrated its proximal-to-distal evolution. Such turbidites were interpreted as “seismo-turbidites” initiated by submarine landsliding triggered by an earthquake (Tripsanas et al., 2008).

The age of turbidites in cores (right hand panels in Figs. 11, 12) has been estimated from their stratigraphic position relative to Heinrich layers (H1–H4), Holocene detrital-carbonate layers (NI, LA), local meltwater plume beds (R1–R7), and hyperpycnal erosional layers (E1–E3) previously dated by Tripsanas and Piper (2008a, 2008b). The “seismo-turbidites” of Tripsanas et al. (2008) are of value in precisely dating landslide events. The “hyperpycnites” of Tripsanas and Piper (2008b) are distinctly different petrologically and unlikely to be associated with landsliding.

Geotechnical Properties of Cores from the Continental Slope

Twenty-four (24) sediment cores taken on the slopes of Orphan Basin are used to characterize the index properties of the sediments to a maximum subbottom depth of 12 m. Core 2003033-19 (Fig. 2) was chosen as representa-
tive of slope sediments. The bulk density in hemipelagic sediment shows the expected increasing trend with depth (dashed line in Fig. 13A) and is higher in intervals rich in ice-rafted detritus and in carbonate-rich Heinrich layers. Water content is also dependent on lithology and decreases slightly down core. The vane shear strength values increase to 5 m downcore, but the trend becomes negative to the bottom of the core (Fig. 13C). The maximum past pressure ($P'$) values obtained from consolidation tests (Table 2) suggest that the sediments are normally consolidated to slightly overconsolidated (Fig. 13D).

Continuous undrained strength profiles were calculated for each core using the triaxial vane shear data. The $S_u/\sigma'$ ratio in the normal consolidation stress range (NSP) was estimated to be 0.48 from the triaxial data. The seismic character on Trinity trough-mouth fan, up-dip from the illustrated basin floor stratigraphy, distinguishes stratified intervals (dashed pattern) and mass-transport deposits (gray tone; e.g., Fig. 7), some of which are glaciogenic debris-flow (GDF) deposits. Sparker profiles are located in Figure 2.

**Figure 4. Correlation of deeper stratigraphy in core MD95-2025, including an oxygen isotope curve (Hiscott et al., 2001), to high-resolution sparker profiles in Orphan Basin, showing correlation to Heinrich (H) events and to marine isotope stages (MIS).** Depth scale on seismic profile based on an assumed acoustic velocity, $v$, of 1500 m s$^{-1}$. Interpreted seismic character on Trinity trough-mouth fan, up-dip from the illustrated basin floor stratigraphy, distinguishes stratified intervals (dashed pattern) and mass-transport deposits (gray tone; e.g., Fig. 7), some of which are glaciogenic debris-flow (GDF) deposits. Sparker profiles are located in Figure 2.

| Table 1. Summary of Stability Analysis Using Miniature Vane Shear Data from 18 Cores, Orphan Basin |
|-----------------|-----------------|-----------------|-----------------|
| Minimum         | 0.7             | 3.2             | 0               |
| Maximum         | 83.6            | 18.3            | 27.9            |
| Average         | 13              | 8.1             | 11.7            |

**Notes:** Factor of safety—the ratio of undrained shear strength of sediment to the shear stress on a slip surface, such that failure will occur when the value is <1. Critical slope—the slope at which the factor of safety is 1. Critical earthquake coefficient—the ratio of the seismic acceleration to the acceleration of gravity that yields a factor of safety = 1. Critical thickness—the depth of the slip surface at which the factor of safety = 1. For further explanation, see text section “Geotechnical Properties of Cores from the Continental Slope” and Supplemental Material (text footnote 1).
hesion ($c'$), pore pressure value ($A_f$), and friction angles ($\phi'$) used in the modified Mohr-Coulomb equation were obtained from the triaxial tests (Table 2). Multi-sensor core logger bulk density was used to calculate the effective overburden stress ($\sigma'_v$). In addition, $\sigma'_v$ values and miniature vane undrained shear strength measurements from the 24 cores were used to estimate $S_u/\sigma'_v$. The averaged $S_u/\sigma'_v$ from the 24 cores was 1.27 in the upper 2 m and 0.35 for sediments at depths of greater than 2 m. The strength profiles are plotted with the vane measurements from core 2003033-19 (Fig. 13C) and show good correlation to 5 m depth. Below 5 m, the vane data values are lower than those of the three profiles.

The normalized strength values (0.48) obtained from the triaxial tests (Table 2) are higher than the average $S_u/\sigma'_v$ (0.35) obtained from the miniature vane shear strength (Table 1; Fig. 13C). The undrained unconsolidated (UU) shear strength value matches the normalized strength parameter (NSP) profile (Fig. 13C). These various measured values of $S_u/\sigma'_v$ are consistent with estimates made elsewhere. On the continental slope beyond the eastern periphery of the 1929 Grand Banks landslide, $S_u/\sigma'_v$ from three triaxial tests ranges from 0.33 to 0.40 (Mosher et al., 2010b). Both east of the Grand Banks landslide and in Orphan Basin, vane shear measurements suggest that $S_u/\sigma'_v$ falls to 0.25 at some levels in cores and that the cores show normal consolidation. Karlsrud et al. (1985) suggested that a typical value of $S_u/\sigma'_v$, for lean normally consolidated Norwegian clay is 0.3–0.35, based on anisotropic consolidation triaxial tests. Lee et al. (2004) suggested that $S_u/\sigma'_v$, is generally 0.3 for normally consolidated sediments, whereas Skempton (1969) gave a range of 0.2–0.5 for normally consolidated clay.

The critical pseudo-static acceleration ($k_c$) to cause failure was calculated at various depths following Lee and Edwards (1986), assuming that $S_u/\sigma'_v$ is constant with depth and that the sediments are normally consolidated. This calculation was done using $S_u/\sigma'_v$ values of 0.35 (vane data) and 0.48 (triaxial data). The minimum earthquake coefficient ($k_x$) calculated for each core site was also included (Fig. 14B). This analysis suggests that on slopes of 6°, a seismic coefficient of >10% is required to cause failure in most of the sediment, but that some intervals sampled only by vane shear measurements might require a seismic coefficient of only 4%–5% on slopes of 1°. There was only one site of the 24 studied where the calculated static factor of safety was less than unity, and the slope angle for this site was 1°.

**Maximum Unfailed Slope Angle on the Continental Slope**

The apparent greatest slope angle beyond which sediments have failed was measured at 61 locations on canyon walls and steep continental slope from seismic-reflection profiles within ±30° of orthogonal to local strike. An example is shown in Figure 5, where undisturbed sediment above headscarps shows apparent slope angles of 1.7°, 1.2°, and 2.2°. The entire data set plotted as a histogram (Fig. 14A) shows a maximum angle of unfailed slope of ~3°, with higher than 3° sustained only in glacial till. Low critical-slope measurements may result from the apparent dip effect where profiles were not precisely orthogonal to the local strike.
Figure 6. High-resolution sparker dip profile from the south slope (Bonanza sector) of Orphan basin, showing seismic stratigraphy and multiple mass-transport deposits (MTDs). Profile is located in Figure 2. H1–H3—Heinrich event layers; MIS—marine isotope stage.
DISCUSSION

The Use of Turbidites to Recognize Failures

Previous work in Orphan Basin has suggested two principal mechanisms for the initiation of turbidity currents: hyperpycnal flow of subglacial meltwater producing "hyperpycnites" (Tripsanas and Piper, 2008a) and transformation of landslides on the continental slope likely triggered by earthquakes producing "seismo-turbidites", as exemplified by the Sheridan failure at ca. 7 ka (Tripsanas et al., 2008). Distinguishing these two types of turbidite is necessary if turbidites are to be used as indicators of paleoseismicity. Such a distinction is applicable, e.g., in fiord basins where turbidites have been used to interpret paleoseismicity (St-Onge et al., 2012). Therefore we discuss criteria based on sedimentological and petrographic character that may be applicable to other settings.

The stratigraphic distribution of large hyperpycnal flows is inferred from characteristic straight erosional channels extending downslope from the limit of till on the upper slope and interbedded with glacigenic debris-flow deposits (Tripsanas and Piper, 2008b; Piper et al., 2012). It is possible that some of the smaller beds that we previously classified as "hyperpycnal" were initiated at an ice margin by the plume settling process described by Hizzett et al. (2018) from the Squamish Delta (British Columbia, Canada). We therefore use the more general term “glacigenic turbidite” in this paper. Hyperpycnal flows and glacigenic turbidites originated only when the Newfoundland ice sheet extended to the shelf edge or upper slope. Seismic-reflection profiles show that in the last 100 k.y., glacigenic debris flows and associated hyperpycnal flow erosion occurred only during the last glacial maximum, from 15 to 28 ka (Tripsanas and Piper, 2008a). Turbidites occurring before or after this time interval likely resulted from the flow transformation of landslides, or possibly as storm-generated sediment gravity flows from the Grand Banks shelf.

In Orphan Basin, almost all sandy turbidites dating from the last glacial maximum share a similar petrography, dominated by iron oxide–stained quartz grains derived from the Carboniferous strata eroded by the Trinity Trough ice stream. These sands were described and discussed in detail by Tripsanas and Piper (2008b). Muds eroded from the same source also have a red-brown color that is imparted to the glacigenic turbidite beds. Such glacigenic turbidites occur as amalgamated sand deposits in channels and lobes and as thin-bedded (<5 cm thick) sand and mud turbidites in overbank settings (Tripsanas and Piper, 2008a). Conversely, seismo-turbidites have variable petrographic composition due to their widespread sourcing principally from the Grand Banks, as documented in detail for the 7 ka Sheridan failure by Tripsanas et al. (2008).

The presence of inverse grading at the base of a sand bed has been widely used as a criterion for hyperpycnites (Mulder et al., 2003), even though inverse grading can also develop in high-density turbidity-current gravel-sand deposits due to kinematic sieving in a grain flow (Lowe, 1982). Furthermore, not all hyperpycnal flows result in inverse grading (Lamb and Mohrig, 2009). In Orphan Basin, amalgamated channel-sand deposits interpreted as glacigenic
flows range from medium to fine grained and are very well sorted, and some beds are normally graded (e.g., Fig. 10A). Reverse grading has not been recognized (Tripsanas and Piper, 2008a).

On the other hand, thick proximal and channel-floor seismo-turbidite deposits have grain size ranging from gravel to fine-grained sand, are poorly sorted, and contain a considerable amount of mud (5%–35%; Fig. 10B). They are commonly accompanied by a basal zone consisting of diamicton (muddy gravel). This characteristic is attributed to the limited hydraulic sorting that was imposed on the wide grain-size range of the failed sediment masses. Thin-bedded overbank and distal turbidite deposits, of both glacigenic and seismo-turbidite origin, display similar grain-size characteristics (Figs. 10C, 10D), so that their distinction is based on their detrital petrography (color) and age relative...
to the occurrence of shelf-crossing glacial ice. Thus the origin of at least the thin-bedded turbidites can be quite elusive, and their usage as indicators of seismicity requires corroborating evidence.

**Timing and Distribution of Failures**

The stratigraphic position of headscarps and MTDs within sparker profiles allows their age to be estimated to within a few thousand years, depending on sedimentation rates (Fig. 11). Intermittent turbidites on the basin floor can generally be dated to better than a thousand years based on the precision of the lithostratigraphy in Orphan Basin. The spatial distribution of smaller and shallower headscars and MTDs is determined with confidence from seismic-reflection data. The along-slope extent of evidence of failures, including headscars, MTDs, and seismo-turbidites, is shown schematically in Figures 15 and 16.

For the youngest widespread slope failure in Orphan Basin at ca. 7 ka, Tripsanas et al. (2008) demonstrated from cores and sparker profiles a clear genetic relationship between retrogressive failure and its progressive transformation into a turbidity current, which caused channel erosion and deposited...
first a diamicton and then a gravel and sand turbidite. The 7 ka failure was recognized from both seismic-reflection and core data along a 280-km-long segment of the continental slope in the Bonanza, Sheridan, and Trinity sectors (Figs. 2, 15A; Tripsanas et al., 2008). A probable MTD within an older slide complex on northwest Flemish Cap is dated as mid- to late Holocene in core 201 1031-13 (Cameron et al., 2014), but dating is not sufficiently precise to be sure that it correlates with the main ca. 7 ka failure of Tripsanas et al. (2008). If it were correlative, the length of the failed segment would increase to 370 km.

By analogy with the 7 ka failure, a widespread failure on the south slope at ca. 25 ka (Figs. 5, 6) corresponds to a coarse sand turbidite with an erosional base (Fig. 9A, lower bed) on the basin floor, just below Heinrich event layer H2 (Fig. 11). A large slide on northwest Flemish Cap (Fig. 15C) is also just below H2 (Cameron et al., 2014). MTDs of apparently similar age, just below H2, are reported from the southeastern Grand Banks slope (Rashid et al., 2017) and southern Flemish Pass (Rudolph et al., 2018). The 25 ka failure extends over at least a 120 km length of the south slope. If the failures in Flemish Pass and the Grand Banks are correlative, the failure extends over 345 km. A 24 ka gravel-sand turbidite is present near the bedrock high southwest of Orphan Knoll, probably implying failure on the north slope, for a failure extent of ~400 km if all of these deposits represent a single event.

Other Late Pleistocene failures are of lesser extent. The ca. 19.5 ka failure recognized in seismic-reflection data from the Bonanza sector (Fig. 6) has a correlative gravel-sand turbidite in the Sackville sector, implying source failure over at least 45 km of margin (Fig. 15B). A 22.5 ka MTD in the Sheridan sector is of only local extent. Throughout the southern part of the basin below the entire south slope, there are one or two gravel-sand turbidites (upper bed in Fig. 9A) dated at 12–14 ka and an additional turbidite at ca. 15 ka in the Sackville sector (Fig. 11). No corresponding widespread failure on the south slope is recognized in sparker profiles, perhaps because there is insufficient overlying draping sediment cover to allow it to be discriminated from the very

Figure 10. Sequence of grain-size distributions through turbidite beds in Orphan Basin deposited from hyperpycnal flows or other glacial margin processes (glaciogenic turbidites) and those initiated by slumping (seismo-turbidites). Analyses are numbered and color coded from base to top.
Figure 11. Summary of distribution of mass-transport deposits (MTDs) in high-resolution sparker profiles and cores from Orphan Basin and the occurrence of gravel-sand turbidites and erosional surfaces in cores back to 40 ka. Age scale is based on Heinrich event layers (H1-H4), early Holocene detrital-carbonate layers (NI—Noble Inlet, LA—Lake Agassiz), local meltwater-plume beds (R1–R7), and hyperpycnal erosional layers (E1–E3) previously dated by Tripsanas and Piper (2008a, 2008b). Error bars are estimated on a case-by-case basis depending on precision of stratigraphic markers. IRD—ice-rafted detritus.
Figure 12. Summary of distribution of mass-transport deposits (MTDs) in high-resolution sparker profiles and cores from Orphan Basin and the occurrence of gravel-sand turbidites and erosional surfaces in cores from 40 to 80 ka. Age scale is based on Heinrich event layers (H4–H6) correlated to core MD95-2025 (Hiscott et al., 2001) and Integrated Ocean Drilling Program Sites U1302 and U1303 (Channell et al., 2012). Error bars are estimated on a case-by-case basis depending on precision of stratigraphic markers; older age limit of open error bars is unconstrained. MIS—marine isotope stage.
Figure 13. Downcore plots of geotechnical measurements from core 2003033-19, Orphan Basin. (A) Bulk density from multi-sensor core logger. Dashed line shows trend for low-carbonate muds. (B) Water content from discrete samples. (C) Shear strength from miniature shear vane and one undrained unconsolidated test; also shows predicted trends based on various assumptions discussed in text, using the Mohr-Coulomb equation, the normal consolidation stress range, and assuming the normalized strength ratio $S_u/\sigma'_v = 0.35$. (D) Shear stress as calculated from downcore data and measured from two consolidation tests. (E) Downcore spectrophotometer values, $L^*$—black to white; $a^*$—green to red. (F) Lithology log, legend across bottom; IRD—ice-rafted detritus. Blue bars across all panels show correlative Heinrich event layers H1–H3.
widespread 7 ka failure, but the turbidites could be the result of failure on steep canyon walls. The 24 ka gravel-sand turbidites and erosional surfaces in the Bonanza and Sackville sectors appear more extensive than the 25 ka turbidites, but 25 ka failures and MTDs are widespread in Flemish Pass (Fig. 15C; Cameron et al., 2014; Rashid et al., 2017). It is uncertain which event is represented by the failures and MTDs on the south slope. Gravel-sand turbidites of limited extent are also found at 29 ka (Fig. 11), but as with the 12–14 ka turbidites, no corresponding failure is recognized in seismic-reflection data in the Bonanza and Sackville sectors.

The record of failure and corresponding gravel-sand turbidites is less precisely dated prior to 35 ka because of the lower sedimentation rate and fewer cores penetrating so deep. Furthermore, seismic-reflection imaging of slope failures is less complete. A ca. 37 ka failure in the Bonanza (Fig. 6) and parts of the Sheridan sectors resulted in a gravel-sand turbidite (Fig. 11). A ca. 37 ka failure is also recognized on part of the north slope, but it is not certain that the two correlate (Fig. 15D). Failures of similar age were reported in central Flemish Pass (Huppertz and Piper, 2009) and on the slope southwest of Flemish Pass (Toews, 2003), both areas where large failures are infrequent. The conservative estimate of the extent of a single failure is 100 km on the south slope; if all the failure were synchronous, then the extent would be 520 km. A gravel-sand turbidite was found in the Sackville sector only at 31.5 ka; the slope failures recognized in seismic-reflection data between Heinrich event layers H3 and H4 in these two sectors might, therefore, represent two discrete events of even shorter failure length.

A 45 ka gravel-sand turbidite in the Bonanza sector has no recognized correlative failure on the south slope (Fig. 12). Failure at ca. 58 ka is recognized on the south slope (Fig. 7) and from a MTD adjacent to the bedrock high southwest of Orphan Knoll (Fig. 8) and corresponds to a gravel-sand turbidite in the

### Table 2. Summary of Consolidation and Triaxial Test Results from Core 2003033-19, Orphan Basin

| Core depth (m) | LL (%) | PI (%) | Soil classification | $C_v$ (kPa) | $P'_c$ (kPa) | OCR | $c'$ (kPa) | $q'$ (°) | $A_f$ | $S_u/\sigma_{u, v}$ |
|----------------|-------|-------|---------------------|-----------|-------------|-----|-----------|--------|--------|----------------|
| 4.6            | 20.1  | 10.8  | CL-ML               | 0.12      | 42.1–61.6   | 1.1–1.62 | 6.8       | 35.6   | 0.47  | 0.48           |
| 6.5            | 26.1  | 10.4  | CL                  | 0.2       | 52.2–90.0   | 0.87–1.5 | 0.1       | 36.1   | 0.25  | 0.48           |

Note: $S_u/\sigma_{u, v}$ is calculated using a correction of 0.80 to account for the effect of isotropic consolidation. LL—Liquid Limit; PI—Plasticity Index. Soil classification: CL—lean clay, ML—lean silt; $C_v$—Compressibility Index; $P'_c$—preconsolidation pressure; OCR—overconsolidation ratio; $c'$—effective cohesion; $q'$—effective friction angle; $A_f$—Skempton’s pore pressure parameter at failure; $S_u/\sigma_{u, v}$—normalized strength ratio. For further explanation, see text section “Geotechnical Properties of Cores from the Continental Slope” and Supplemental Material (text footnote 1).
Figure 15. Maps showing possible regional extent of the larger failures in and near Orphan Basin. Colored bars show areas of failed slope from this study; core lithologies provide additional control on the extent of failure. Ellipses define schematic along-slope extent of other failures from the literature, namely: 1—Cameron et al. (2014); 2—Huppertz and Piper (2009), with ages corrected after Rudolph et al. (2018); 3—Tøews and Piper (2002); 4—Rashid et al. (2017); 5—Rudolph et al. (2018); 6—Marshall et al. (2014). Bathymetric contours in meters, at 250 m intervals on the shelf and 1000 m intervals in deep water. BON.—Bonanza; MTD—mass-transport deposit.
Bonanza sector (Fig. 12). This failure extends along at least 220 km of the south slope, but whether this is the same failure on the bedrock high is uncertain (Fig. 15E). A widespread 70 ka failure on the south slope (Figs. 6, 7) is represented in cores by only a mud-silt overbank turbidite (Fig. 12); channel cores do not penetrate to this stratigraphic horizon. A deeper MTD adjacent to the bedrock high (Fig. 8) has an extrapolated age of ca. 85 ka (Fig. 15F). It is approximately the same age as a major failure on the eastern side of Orphan Knoll (Toews and Piper, 2002), 150 km to the east, and as a failure in central Flemish Pass (Campbell et al., 2002; Huppertz and Piper, 2009), 380 km to the south.

In order to evaluate whether the longer-distance correlations in Figure 15 are likely or not, three different interpretations of the failure record have been compiled. The “maximum” scenario assumes that any large failures that have an overlapping age estimate within error are the product of the same earthquake trigger. The “conservative” scenario assumes that failures of similar ages separated by areas with no evidence for failure represent separate triggers. The “balanced” scenario (Fig. 16) applies geological judgement based on data available in intervening areas to correlate some of the features that were separated in the conservative scenario.

A plot of length of failed slope against cumulative failure rate or its inverse, the recurrence period (Fig. 17), is used to determine whether the conserva-

Figure 16. Age distribution of widespread failures in Orphan Basin, also showing variation in sedimentation rate on the south slope [from core 2004024-024 of Tripsanas and Piper (2008a); >80 ka extrapolated from core MD95-2025] and variation in eustatic sea level [compiled from Peltier and Fairbanks [2006] and Chappell [2002]]; datum is present sea level. For discussion of uncertainties in failure length and definition of the “balanced” scenario see text. LA—Lake Agassiz event; NI—Noble Inlet event; H1–H7a—Heinrich event layers; MIS—marine isotope stage.

Figure 17. Cumulative frequency distribution for the failure length proxy for earthquake magnitude. Failure lengths follow the maximum, balanced, and conservative scenarios as discussed in the text. Red bar shows estimated recurrence interval of large failures around the epicenter of the 1929 Grand Banks earthquake (Piper et al., 2005). Gray dashed line is the best fit straight line for the balanced scenario data. Earthquake magnitude scale is speculative; its positioning is based on the estimate by Adams and Halchuk (2003) of Mw = 8.0 for the largest earthquake on the Canadian eastern continental margin, further calibrated by the finding of ten Brink et al. (2009) that an order-of-magnitude increase in failed slope length corresponds to a two-point increase in earthquake magnitude. For further discussion, see text.
tive, balanced, or maximum estimates of failed slope length are more probable. The length of failed slope is assessed from the distribution of correlative failures (either scarps or MTDs) along the continental slope. The 1929 Grand Banks failure occurred along 250 km of the slope (Mosher and Piper, 2007), and failures of similar size have an estimated recurrence interval of between 50 k.y. and 120 k.y. based on a 1 m.y. record on the continental slope (Piper et al., 2005). The plot of recurrence period using maximum estimates of failed slope length in Orphan Basin suggests a recurrence interval of only 25 k.y. for 200-km-long failed lengths of slope, the size of the 1929 Grand Banks failure; this value is inconsistent with the record inferred from seismic-reflection profiles and cores (Fig. 11). Use of the balanced and conservative estimates for large failures results in more failures being recognized, but overall the failures are of lesser extent. A scenario between the balanced and conservative estimates best predicts failure lengths consistent with the failure length and magnitude of the 1929 Grand Banks failure.

Correlation of Failures with Potential Forcing Factors

The widespread distribution of the 7 ka regional failure surfaces and MTDs in multiple slope canyon systems led Tripsanas et al. (2008) to argue that a large earthquake was the probable trigger for failure. Similar arguments can be made for most of the older failures and corresponding turbidites recognized in this study. Several other mechanisms have been proposed in the literature that might result in a widespread failure: (1) Lateral spreading deformation at a weak décollement horizon, commonly a buried MTD horizon, can lead to catastrophic failure (e.g., Mosher et al., 2004). Such décollement surfaces are limited by bathymetry and unlikely to occur in multiple canyon systems. Where found on the southeastern Canadian margin, they generally underlie successions >30 m thick (Piper, 2001), in contrast to the <20 m thickness of most failures on the south slope of Orphan Basin. (2) Increased pore pressure in shelf-edge deltas generally formed at times of lowered sea level (e.g., Locat et al., 2009) can be ruled out because the upper slope underlain by glacial till has not failed. (3) Bearing-capacity failure as a result of ice loading on the outer shelf (Mulder and Moran, 1995) can also be ruled out by the lack of failure in the upper slope. (4) Many authors have proposed a relationship between dissociation of gas hydrates and failure, with a temporal relationship during glacial cycles as a result of either falling sea level or rising bottom-water temperature (e.g., Maslin et al., 2004). A bottom simulating reflector is present on Sackville and Orphan Spur to the south and north of Orphan Basin respectively (Mosher, 2011), but none has been recognized within Orphan Basin. Furthermore, there is no evidence that sufficient excess pore pressures (i.e., >30% hydrostatic) can be generated in the upper 20 m of the seabed (e.g., Christian and Heffler, 1993; Strout and Tjelta, 2005) to cause failure. The dissociation of gas hydrates may, however, weaken slope sediments, and failure could result from a lower-magnitude earthquake (Kvalstad et al., 2005). In Orphan Basin, bottom waters have been cold (<3°C) even in the Holocene; pressure changes due to a fall in sea level (Fig. 16) are thus the dominant control on gas hydrate dissociation. (5) Almost simultaneous failure in multiple canyon systems could result from episodic release of subglacial meltwater from ice sheets grounded on the upper slope, leading to erosion and undercutting of canyon walls (Piper and Gould, 2004). There is no evidence for this process on the south slope from the timing of failure correlating with meltwater events (Fig. 16) or from slope morphology showing canyon-widening events. We conclude that none of the five mechanisms discussed above account for the large synchronous failures in Orphan Basin. A few smaller failures, such as the small MTD dated at 22.5 ka in part of the Sheridan sector, might have been triggered by any of a number of factors, including (1) bioturbation leading to oversteepening on a canyon wall or (2) higher pore pressures due to dissociation of gas hydrate following the prolonged fall in sea level. We consider, however, that earthquake triggering remains the most likely cause.

Although the total number of recognized failures is small, failures appear to have been more abundant during maximum ice extent (30–20 ka) and during deglaciation (20–12 ka) (Figs. 11, 16) compared with the 40–80 ka time span (Fig. 12). There is a similar observation along the Scotian margin (Mosher et al., 2004). This higher failure frequency is not considered to be solely an artifact of the ability to image and core deeper horizons. There are two processes that might be responsible for this variability in failure frequency: (1) Loading and unloading by glacial ice on parts of the outer continental shelf may have led to greater seismicity on reactivated basement structures, as argued by Wu (1998). In Fennoscandia, where ice kilometers in thickness overlay crystalline bedrock, an immediate deglacial burst of seismic activity is well documented (Arvidsson, 1996). However, offshore of Newfoundland, much thinner ice partially supported by marine buoyancy overlay shelf sediments, so loading and consequent seismicity would have been less extreme. (2) Generation of high excess pore pressure, as a result of high sedimentation rates during glacial maximum or from dissociation of gas hydrates during falling sea level prior to the last glacial maximum (Fig. 16) would result in smaller seismic accelerations being required to trigger a failure (Mosher et al., 1994). A 0.3 decrease in the magnitude of the earthquake required to create a failure of a particular size would double the apparent rate of earthquakes inferred from the failure record, using the relationships of ten Brink et al. (2009).

Estimating Seismic Magnitude from Failures

Whether continental slope sediment fails during an earthquake is a consequence of the gradient, the shear strength of the sediment, and the magnitude of seismic accelerations (Mosher et al., 1994). Given that the gradient is highly variable on a slope incised by canyons and that most failures that we have studied appear retrogressive, any continental slope area with a mean gradient of >2° will have some local gradients >6° that would be sufficient to initiate retrogressive failure (Piper et al., 1999). Such areas are indicated by yellow or red tones in Figure 2.
The maximum distance from an earthquake epicenter at which failure will take place was estimated by ten Brink et al. (2009) for various earthquake magnitudes and seafloor gradients and was found to vary substantially with estimates of $S_{f}/\sigma'$, for the sediments. For example, for a magnitude 7.2 earthquake, failure would take place on a 6° slope as much as 43 km from the epicenter for $S_{f}/\sigma' = 0.3$. The extent of failure would decrease with higher values of $S_{f}/\sigma'$, and increase with earthquake magnitude.

The $M_n = 7.2$ 1929 Grand Banks earthquake created widespread landsliding over a distance of ~250 km along the continental slope (Piper et al., 1999; Mosher and Piper, 2007). Like the 7 ka failure on the south slope of Orphan Basin, this event involved principally retrogressive failure. The two areas have similar variability in local gradients, including slopes >6° on which failure likely initiated (Piper et al., 1999). The central area of the Grand Banks landslide likely had higher pore pressures, indicated by the abundant shallow gas in sediments in that region (Piper et al., 1999).

Widespread sediment failure along a 100 km segment of the continental slope might therefore be interpreted in several ways. If the triggering earthquake were located beneath the slope, according to the calculations of ten Brink et al. (2009), it would have a magnitude of $M = 6.8$ for $S_{f}/\sigma' = 0.2$, $M = 7.5$ for $S_{f}/\sigma' = 0.3$, or $M = 6.4$ if calibrated against the 1929 Grand Banks earthquake. If the triggering earthquake were located beyond the failed area, e.g., on the continental shelf, then the magnitude estimate would be a minimal estimate. Thus the estimation of earthquake magnitude is difficult from sediment strength data because of the inherent uncertainty about the strength properties of the weakest layers and the location of the causative earthquake relative to the failed slope. Furthermore, variations in depth of earthquakes would influence sediment response.

In our study, we have recognized large failures occurring over a distance of at least 200 and probably 300 km along the margin (conservative and balanced scenarios respectively) and small failures affecting a margin length an order of magnitude shorter. Using the calculations of ten Brink et al. (2009), a two-point increase in earthquake magnitude (e.g., from 6.0 to 8.0) results in an approximately ten-fold increase in the length of the failed slope for an epicenter on the slope, assuming constant $S_{f}/\sigma'$. It is therefore probable that the observed range of failure lengths in Orphan Basin has recorded about a two-point range of magnitudes. The maximum earthquake magnitude for the eastern continental margin is estimated as $M_n = 8.0$ (Adams and Halchuk, 2003; corrected to moment magnitude by J. Adams, 2010, personal commun.), implying that earthquakes ranging in magnitude from near this maximum to magnitude ~6.0, giving a two-point range in magnitudes, are likely to be recorded by observed submarine landslides and turbidites. The scale of most-probable magnitude in Figure 17 has thus been positioned to cover this range of magnitudes and is consistent with the determined 7.2 magnitude of the 1929 Grand Banks failure, with an epicenter beneath the upper continental slope. Because not all epicenters are likely to lie beneath the continental slope, the estimate of the length of failed slope from an earthquake of known magnitude using Figure 17 represents a maximum length, or conversely the record from submarine landslides will tend to underestimate the frequency and magnitude of earthquakes along the continental margin. If smaller failures in Figure 17 were not triggered by earthquakes, but by some other process, it would not invalidate the interpretation of the larger more extensive failures for which an earthquake trigger is necessary to account for their distribution across multiple slope-drainage systems.

### Regional Occurrence of Large Failures and Comparison with the Instrumental Record

There is no evidence that seismicity in Orphan Basin is regionally lower than elsewhere on the southeastern Canadian margin. The south slope overlaps a major tectonic lineament, the Cumberland belt, and the north slope terminates at the Dover fault, aligned with the Charlie Gibbs fracture zone (Fig. 1; Enachescu, 1987). The recurrence interval of large failures in Orphan Basin is similar to that found elsewhere on the southern parts of the eastern continental margin, as summarized by Piper (2005). For example, on the central Scotian Slope, Jenner et al. (2007) recognized four small failures in the past 15 k.y., comparable with the number on the south slope of Orphan Basin over the same period (Fig. 16).

There is no evidence that smaller earthquakes might act to cause seismic strengthening of slope sediment (Sawyer and DeVore, 2015). Such seismic strengthening is apparent from cores on active margins as higher shear strengths, with $S_{f}/\sigma' >0.4$. Such strengthening is not apparent in cores from the south slope of Orphan Basin, or from many other areas of the eastern Canadian margin, where cores generally show normal consolidation (e.g., Mac-Killop et al., 2004; Mosher et al., 2010b). With the low rates of seismicity on the Canadian passive margin, submarine failure thus provides a reliable estimator of seismic hazard.

From seismological models, Mazzotti (2007) argued that random seismicity under a typical intraplate strain rate of $10^{-12}$ a$^{-1}$ would produce one $M = 7$ earthquake per thousand years over the 10$^6$ km$^2$ of the Canadian eastern continental margin zone, or a recurrence interval of 33 k.y. for the 3.3% (200 km) of the eastern continental margin likely to affect Orphan Basin. Our best estimate for $M = 7$ earthquakes on geological grounds is a recurrence interval of 25 k.y., implying seismicity only a little higher than that of the random seismological model. The seismological model and the submarine failure record are reasonably consistent in their estimates of seismic risk on the eastern Canadian margin, providing additional evidence that the large failures are earthquake triggered.

### CONCLUSIONS

1. With good coverage of high-resolution seismic-reflection profiles, chronologically tied to long cores, a correlation of submarine landslides is developed for Orphan Basin.
2. Two main types of turbidite beds are recognized. Those with massive gravel and sand, with gravelly diamicton in more proximal cores, are
correlated with major retrogressive slope failures. Those with abundant red-brown mud and iron-oxide-stained quartz sand were deposited from glaciogenic flows that were at least in part of hyperpycnal origin.

3. Although the mean $S'/\sigma'$ ratio of typical sediment in Orphan Basin is $\sim0.35$, implying considerable stability, $S'/\sigma'$ falls to as low as 0.25 in some weak layers. Slopes steeper than 3° show widespread failure, except where underlain by glacial till.

4. A frequency-magnitude plot of length of failed slope is calibrated by the finding of ten Brink et al. (2009) that a two-point increase in earthquake magnitude results in an approximately ten-fold increase in the length of the failed slope, and the estimate by Mazzotti and Adams (2005) of $M_w = 8.0$ for the largest earthquake on the Canadian eastern continental margin. This plot is consistent with the recurrence interval estimated for the $M_w = 7.2$ Grand Banks earthquake, provided that balanced or conservative estimates of failed slope length are used.

5. The frequency of earthquakes estimated from the record of large, generally retrogressive slope failures is similar to that estimated from seis-mological modeling by Mazzotti (2007). It suggests that the slope failure record on passive margins with low seismicity can be used as a proxy for paleoseismicity.

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