Appendix A

Submarine Failures and Associated Tsunamis, Norway
- Literature Review

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Introduction

Submarine failures and associated tsunamis have been documented in Norway for more than 125 years. Unlike subaerial slope failures, however, the precise shapes and volumes of subaqueous failures has been, until recently, poorly known. As well, documentation of the waves created and the distance where they were observed from the site of generation is generally poorly documented, even for relatively recent events.

This review presents accounts of various Norwegian subaqueous landslide-tsunami events, with a more in-depth description of the better documented cases. Some events have been described in great detail in the literature, particularly where recent multibeam bathymetry and sub-bottom profiling have been used in focussed investigations. Most older studies relied on eye witness accounts and limited echosounding, sub-bottom profiling and geotechnical information.

Table 1 presents a summary of the better known submarine failures with associated waves in coastal Norway. Jørstad (1968) undertook one of the first compilations of slope failures and waves in Norway, including subaqueous failures; much of his paper focusses on subaerial failures entering coastal waters and the ensuing tsunamis. The other references cited for Table 1 contain more detailed summaries of specific cases. The ten events highlighted in the table are summarized in greater detail below. There were several significant submarine slope failures and tsunamis in the Trondheim area between 1888 and 1990. These are better studied than most in Norway and will be considered together in this report.

1. Nidelva River Delta Front – “W” Failure 1888

The Trondheimsfjorden sediments near the Nidelva River consist of up to 125 m of glaciomarine clays, fjord basin muds, and deltaic silty sands and gravelly sands. The marine limit lies at about 175 m.a.s.l.; following deglaciation, the mouth of the Nidelva River (Figure 1) moved progressively northward and deltaic materials advanced over the underlying glaciomarine and fjord basin muds.

L’Heureux et al. (2010a, b) describe several events which took place between about 1888 and 1990 in the Trondheim Harbour area. One of these (their Landslide “W”) (Figure 1) was interpreted to be about 100 years old and “… could have been triggered in association with the flow of sediments following the 1888 event”. It took place on an average slope of about 5-6°, had a volume of 1.3 x 10^6 m^3 and occurred as a result of toe erosion.
Table 1. Subaqueous marine Norwegian failures and associated waves (Terzaghi, 1956; Jørstad, 1968; Aarseth et al., 1989; Longva et al., 2003; L’Heureux et al., 2007, 2010a, b; Bøe et al., 2003).

<table>
<thead>
<tr>
<th>Date</th>
<th>Location</th>
<th>Type</th>
<th>Volume (1000 m$^3$)</th>
<th>Max. Distance Wave Observed (km)</th>
<th>Wave Height (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10/2/1883</td>
<td>Ormestad, Sandar</td>
<td>Clay slide</td>
<td>~30 ?</td>
<td>?</td>
<td>?</td>
</tr>
<tr>
<td>?23/4/1888</td>
<td>Nidelva River delta front, Trondheim (&quot;W&quot;)</td>
<td>Translational slide</td>
<td>1,300</td>
<td>1</td>
<td>4-5</td>
</tr>
<tr>
<td>23/4/1888</td>
<td>Trondheim Harbour &quot;Brattora&quot;</td>
<td>Flow slide</td>
<td>~ 3,000</td>
<td>?</td>
<td>?</td>
</tr>
<tr>
<td>2/5/1930</td>
<td>Orkdalsfjord</td>
<td>Flow Slide</td>
<td>60,000</td>
<td>20</td>
<td>Large</td>
</tr>
<tr>
<td>31/8/1940</td>
<td>Finnvika, Vefsn</td>
<td>Clay slide, Flow slide?</td>
<td>?</td>
<td>5-6</td>
<td>&gt;5</td>
</tr>
<tr>
<td>8/10/1950</td>
<td>Ilsvika, Trondheim</td>
<td>Flow slide</td>
<td>2,000-3,000</td>
<td>?</td>
<td>?</td>
</tr>
<tr>
<td>7/5/1959</td>
<td>Sokkelvik, Nordreisa</td>
<td>Clay slide</td>
<td>3,000 – 4,000</td>
<td>8</td>
<td>4</td>
</tr>
<tr>
<td>18/6/1962</td>
<td>Stjernoy, Øksfjord</td>
<td>Clay slide</td>
<td>?</td>
<td>?</td>
<td>?</td>
</tr>
<tr>
<td>21/2/1964</td>
<td>Follerø, Surnadalen</td>
<td>Clay slide</td>
<td>30 ?</td>
<td>1-2</td>
<td>?</td>
</tr>
<tr>
<td>8/5/1966</td>
<td>Oslofjord</td>
<td>?</td>
<td>?</td>
<td>3-6</td>
<td>?</td>
</tr>
<tr>
<td>?/9/1967</td>
<td>Nordfjord</td>
<td>Flow slide?</td>
<td>15,000</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>25/4/1990</td>
<td>Nidelv River Delta front, Trondheim</td>
<td>Flow slide</td>
<td>5,000 – 6,000</td>
<td>?</td>
<td>?</td>
</tr>
<tr>
<td>20/6/1996</td>
<td>Finneidfjord</td>
<td>Clay slide</td>
<td>1,000</td>
<td>?</td>
<td>?</td>
</tr>
</tbody>
</table>
Figure 1. Location of four submarine failures off the Nidelva Rover near Trondheim discussed in this report – “W”, “1888”, “1950” and “1990”. Contour intervals are 5 m. (L’Heureux et al., 2010a).

The “W” slide is characterized by a slide scar at a depth of about 80-160 m near the head of the Nidelva Channel (Figure 1); the scarp is 5 to 10 m high and 1,100 m long. The slope at the base of the scarp is 5-6° but increases to 26° in Nidelva Channel. Blocks and landslide debris occur immediately below the scarp.
Figure 2. Shaded relief of the “W” failure area with high-resolution seismic profile (L’Heureux et al., 2010a).
Figure 3. Physical properties of Core GS08 located just upslope from the head scarp of the “W” failure (see Figure 1 for location.) (L’Heureux et al., 2010a).
Figure 4. Results from three triaxial tests on laminated clay-rich beds from core GS08 (L’Heureux et al., 2010b).

The failure plane is a distinct seismic reflector which can be traced upslope from the failure scarp. This plane is a 1.5 m-thick laminated clay which was sampled in coring (Figure 3, 4). The failure plane is overlain by an acoustically transparent unit. Based on sediment thicknesses in the slide scar and measured sedimentation rates, the failure event took place about 100 years ago; it is conjectured that it was probably related to the more significant failure events of April 23, 1888, the Brattøra slides.

L’Heureux et al. (2010a, b) suggest that sediment gravity flows through the Nidelva Channel in the April 23, 1888 Brattøra event might have undercut the toe of the slope below the “W” slope leading to the failure. The authors infer, based on slide dynamics and tsunami simulations, that the “W” failure was responsible for the wave which struck Trondheim on April 23, 1888. The wave was witnessed in the fjord between the Munkholmen Island and Ilsvika (Figure 1).

Fluid escape features are common throughout Trondheimsfjorden with pockmarks identified along the flanks of the Nidelva Channel. These features are typically 10-15 m in disameter and less than one metre deep. (Note that similar probable fluid escape features have been identified in Douglas Channel, British Columbia off Bish Cove and Emsley Cove and southwest of Coste Island.) These unfavourable artesian conditions were interpreted by L’Heureux et al. (2010a) as one of the important pre-conditioning factors leading to the failures near Trondheim. Figure 5 shows the Factor of Safety for the four failures plotted against shear strength ratio.
2. April 23 1888 Landslide – Brattøra Slide

The most devastating failure-tsunami in the Trondheim area took place on April 23, 1888 though as Andresen and Bjerrum (1967) show, there had been several other previous failures along the coastline in 1919, 1944 and 1950; the last is described below. The April 23, 1888 event occurred at low tide and created a wave which swept over a 5-7 m high embankment along the shore. A portion of the railway station at Brattøra, which had been built on fill 8 years earlier, failed into the fjord along with a jetty (Andresen and Bjerrum, 1967) (Figure 6; L’Heureux et al., 2010b). Meteorological conditions were abnormal in that total precipitation and snow accumulation were more than twice normal values in the winter of 1888; intense melting and rainfall may have contributed to the failure by increasing pore pressures in coastal and nearshore sediments (L’Heureux et al., 2010a).

**Figure 5.** Factor of Safety for four submarine failures in the Trondheimsfjorden plotted against the undrained shear strength ratio (L’Heureux et al., 2010a). The “1888 Landslide” is also referred to as the Brattøra slide.
Eyewitnesses stated that the failure started out in the fjord and retrogressed towards the shoreline (Andresen and Bjerrum, 1967) (Figure 7). It was concluded that a large wave formed midway between Munkholmen Island and the coast (Figure 7); it should be pointed out that this location would appear to correspond to what L’Heureux et al. (2010a, b) refer to as Slide “W” and would agree with the failure sequence which they postulate. The solid “stone jetty” rested on a layer of sand dipping offshore at 8-15°. A smaller slide occurred several hours after the first slide and was located to the east where it destroyed a jetty more than 30 m long (“Pir landslide”; L’Heureux et al., 2010b; Figure 8). Boreholes later confirmed that the failures took place in silty sand (Andresen and Bjerrum, 1967).
Figure 7. Interpretation of the April 23, 1888 failure in Trondheim harbour (Andresen and Bjerrum, 1967). Note the location of the initial wave (“starting wave”) which appears to correspond approximately to L’Heureux et al.’s (2010a) slide “W” location.

Recent investigations (L’Heureux et al., 2010a, b) have provided greater insights into the events of April 23, 1888, with some significant reinterpretations of Andresen and Bjerrum’s (1967) early work. Multibeam bathymetric data show that the failure was in two adjacent areas separated by two parallel ridges (Figure 1, 8). The two headwall scarps are up to 10 m high and extend 600 m. The slide failure plane dips at about 6°. The stratigraphy of the intact sediment lying between the two failure zones shows an upper laminated unit underlain by a very distinct reflector over a very transparent unit; failure is inferred to have occurred within this transparent unit (Figure 9). The authors conclude that much of the landslide mass disintegrated following failure, producing a flow slide which was funnelled down the Nidelva Channel. Some slide debris remains locally within the slide scars.
As noted above, failure is believed to have taken place within the thin, laminated clay-rich beds (Figure 9). Stability analyses show that the slopes for all of the failures were only marginally stable. Previous investigations (Andresen and Bjerrum, 1967) had concluded that the failure took place in unconsolidated fine silty sand and was a flow slide.
The results of L’Heureux et al.’s (2010b) slope stability analyses are presented here; the reader is referred to the original paper for details on the methods used. Table 2 summarizes the physical properties that L’Heureux et al. (2010b) present. For undrained conditions, the Brattøra slope was found to be stable at or less than a 15° slope. For such conditions, the most critical failure surface passes through the top of the weak clayey bed (Figure 10). If there is a strength reduction of 2 (i.e., $S_t = 2$), the slope becomes metastable (Factor of Safety = 1.0) if greater than 9°. When using a sensitivity of 5 (as found in nearby cores) all slopes were found to result in factors of safety below unity.

In the case of the “W” slide, under undrained conditions, failure occurs when the Nidelva Channel slope exceeds 30°, as would occur with channel sidewall undercutting. Initial
failure begins at the top of the weak, clay-rich bed where undrained shear strength is least. Figure 11 summarizes the slope stability analyses for these two slides.

Figure 10. Simplified model used for slope stability analysis with SLOPE/W for: (a) the Brattøra landslide; and, (b) the W-landslide.
Table 2. Physical and strength parameters used during the stability analyses at the W- and Brattøra sites

<table>
<thead>
<tr>
<th>Layer</th>
<th>Description</th>
<th>Thickness (m)</th>
<th>$\gamma$ (kN/m$^3$)</th>
<th>$\phi$ ('$^\circ$)</th>
<th>$c'$ (kPa)</th>
<th>$S_{u}$ at the top of layer (kPa)</th>
<th>$\Delta S_{u}$ (kPa/m)</th>
<th>Data from</th>
</tr>
</thead>
<tbody>
<tr>
<td>W-landslide</td>
<td>Unit 1: Silty sand</td>
<td>5</td>
<td>20</td>
<td>35</td>
<td>0</td>
<td>-</td>
<td>-</td>
<td>Core GS08</td>
</tr>
<tr>
<td></td>
<td>Unit 2: Weak clayey bed</td>
<td>2</td>
<td>19</td>
<td>33$^a$</td>
<td>6$^b$</td>
<td>8,0</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Unit 3: Clayey silt</td>
<td>$&gt;$10</td>
<td>19</td>
<td>35</td>
<td>2</td>
<td>21,0</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>Brattøra</td>
<td>Unit 1: Coarse sand and gravel</td>
<td>10</td>
<td>22,0</td>
<td>37</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>CPTU-1 and CPTU-2</td>
</tr>
<tr>
<td></td>
<td>Unit 2: Loose silty sand</td>
<td>6-15</td>
<td>20</td>
<td>35</td>
<td>0</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Unit 3: Weak clayey bed</td>
<td>2</td>
<td>19</td>
<td>33$^a$</td>
<td>6$^b$</td>
<td>13,5</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Unit 4: Loose silty sand</td>
<td>$&gt;$10</td>
<td>20</td>
<td>35</td>
<td>0</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
</tbody>
</table>

* Mohr-Coulomb drained strength parameters obtained from triaxial tests on clayey samples from core GS08 (see Fig. 6c)

Figure 11. Results from slope stability analysis for the Brattøra and W- slopes. Results from undrained analyses are presented with circles while the results from drained analyses are shown with squares. For the Brattøra site, the effect of changing the pore pressure ratio ($r_u$) on the factor of safety is shown. Similarly, the effect of long-term strength reduction (due to sensitivity; $S_t$) in the weak layer is shown.
The sequence of events postulated by L’Heureux et al. (2010a, b) began with a relatively small failure near Brattøra which rapidly evolved into a flow slide (“1” in Figure 12) which was constrained by the Nidelva Channel as it flowed downslope; the Nidelva Channel had not been an active channel, linked with a fluvial source, for the past 2-3,000 years. This flow slide undercut the channel sidewall resulting in the “W” failure (“2” in
Figure 12). This failure was a translational slide which developed rapidly and which led to the generation of the observed tsunami. Failure “3” (Figure 12) resulted from the tsunami inundation; failure “4” occurred several hours after the initial event.

The tsunami created by the “W” slide (Figure 13) was modelled, assuming that the failure was a Bingham fluid, using the BING model (Imran et al., 2001). The yield strength was set at 50 Pa based on results in L’Heureux et al. (2009). Maximum slide velocity from the model was 30 m s\(^{-1}\) and was reached after downslope movement of 1400 m. Significant simplifications were made in the tsunami modelling which otherwise used a conventional approach. Tsunami inundation was computed using the ComMIT/MOST) model.

Figure 13. Results from tsunami simulation showing: (a) surface elevation after 2 min; (b) after 5 min; and, (c) maximum surface elevation during the total simulation time.

Maximum offshore wave heights were about 1-2 m near Brattøra and 0-0.5 m opposite Brattøra at Rørvika (Figure 14).

Maximum inundation heights were up to 4-5 m a.s.l., in good agreement with observations of eyewitnesses who estimated run-up heights of 5-7 m. It appears that the nearshore Brattøra slide was too small to have produced a tsunami; the wave was caused
by the “W” failure, confirming eyewitness accounts that the wave came from out in the bay (L’Heureux et al., 2010b).

**Figure 14.** Time series of tsunami wave amplitudes computed in various ways at Brattøra (left) and Rørvika (right). LSW- linear hydrostatic; disp – linear dispersive; Bouss – Boussinesq (both dispersion and non-linearity).

### 3. 1950 Nidelva Landslide

The 1950 failure event was related to land reclamation along the waterfront at Ilsvika which had started in 1949 (Figure 1). By October 1950, elevations in this area had reached 12 m and offshore failure began (L’Heureux et al., 2010a). Failure occurred along 150 m of the recently filled coastline. Multibeam bathymetric data show that the failure is 600 m wide with a 20 m-high head scarp located close to shore; total volume is estimated to have been 2-3 x 10⁶ m³. The scarps around the failure diminish offshore to less than 1 m. The failure plane has a slope of about 6°. Circular pockmarks (fluid escape features) occur within the slide area (Figure 15); the largest pockmark is 15 m in diameter with a depth of 1.5 m. As with the previous failures, the 1950 event appears to have occurred on a clay horizon.
Figure 15. The 1950 failure near Trondheim showing the head scarps and pockmarks within the slide area. Contour interval is 10 m. (L’Heureux et al., 2010a)

Cutting through the failure areas is a channel (Ilsvika Channel) which originates at the coast off a small creek; this channel is 40-50 m wide, 3-5 m deep and joins the Nidelva Channel about 1 km offshore. The authors conjectured that the channel is active and is developed by gravity-driven undercurrents (turbidity currents) during stream floods.

4. 1990 Nidelva Landslide

The most recent event near Trondheim occurred on April 25, 1990 near the mouth of the Nidelva River (Figure 1). Initial failure occurred at 1621 when cables on the seafloor and in a road embankment were ruptured. Twenty minutes later part of the coastline east of the initial failure failed into the fjord. The failure can be subdivided into four zones (Figure 16). The headscarp has a maximum height of 28 m along the shoreline. The slope of the failure plane within Area 1 is 2-3° and shows slickensided features. This failure plane is exposed within Area 1 and corresponds to a thick transparent unit seen on seismic profiles. Coring shows that this unit is a thick, over-consolidated clay horizon.

Areas 2, 3 and 4 have hummocky surfaces and compressional ridges. The weak horizon in Area 1, on which failure is inferred to have occurred, is absent in Areas 2, 3 and 4 and failure appears to have occurred here on a “secondary slip plane” (Figure 16).
Following initiation of failure, Area 1 sediments appear to have disintegrated and were channelized through a gully north of the 1990 slide scar (Figure 1). Areas 2, 3 and 4 failed retrogressively after the removal of Area 1 and slid on a secondary slip plane located at a shallower position stratigraphically.

Figure 16. The 1990 failure area near Trondheim showing the 4 separate units and inferred failure planes.
Summary of Nidelva (Trondheim) Failures

L’Heureux et al. (2010a, b) summarize the pre-slide conditions which led to failure. These include the geologic conditions, inherited through the post-glacial development of the region, such as sea level fall, changing loci of sedimentation, superposition of deltaic sedimentary units on marine muds, interstratification of low-permeability, sensitive clays with coarser deltaic sediments and the presence of significant artesian pressures in offshore areas.

L’Heureux et al. (2010a) conclude that the failures resulted from translational shear within the laminated, sensitive clay-rich beds. The remoulding of this basal clay-rich horizon, combined with its low permeability, are thought to have resulted in water entrapment beneath the landslide masses leading to long run-out distances and hydroplaning. These flows were highly erosive downstream, leading in some instances to secondary failures due to undercut slopes (e.g., the “W” slide).

Many of the failure triggers were interpreted to be anthropogenic in nature, including deposition of fill, vibration related to construction activities and the detonation of explosives.

Figure 17 summarizes in a schematic fashion the pre-failure and failure conditions of these events. It points to an evolution of the failures from initial block sliding to debris flows and flow slides through to turbidity currents. Retrogressive failures occurred in the 1888, 1950 and 1990 events. As retrogression progresses landward, it can ultimately lead to liquefaction of loose delta front sands and silts; eventually retrogression ceased along the coastline due to higher undrained shear strengths in clayey beds and the presence of coarse littoral and manmade materials (L’Heureux et al., 2010b).
Figure 17. A. Schematic representation of the pre-failure and failure conditions of submarine slides in the Trondheim Bay area. B. Stages of failure from block slides through debris flows to turbidity currents. (L’Heureux et al., 2010a).

The following are the conclusions of L’Heureux et al. (2010a) regarding these submarine failures near Trondheim:

- Clay-rich (event) beds with contrasting sedimentological and physical properties are found on land and in the bay of Trondheim in delta and prodelta deposits. These beds are shown to be softer and more sensitive than the surrounding sediments.

- Pre-landslide conditions result from a complex interaction of variables including loading of the clay-rich beds by deltaic progradation during the Holocene, unfavourable groundwater conditions (i.e. artesian pore-pressures) and locally steep seafloor gradients from erosion and/or sediment accumulation.

- Detailed morphological observations of slide scars combined with high resolution seismic data suggest that gravitational slope failure initiates by translational
movement based in the softer and more sensitive clay rich beds. These observations are confirmed by limit equilibrium slope stability back-analyses.

- The probability of large-scale failure may be significantly greater where steep local gradients (>10°) are present and where recent construction activity causes localized loading and disturbance of the soft and sensitive clay-rich beds. The landslides show disintegrative and retrogressive behaviour and may affect large areas in the fjord and on land.

- Finally, results presented here illustrate the importance of on- and off-shore relationships for understanding shoreline mass wasting processes and warns against slope stability evaluation simply based on cores from land and low resolution bathymetric data in the fjord. A proper shoreline slope stability assessment requires detailed morphological analyses combined with geological understanding of the sediments and their physical properties both on- and off-shore.”

5. 1996 Finneidfjord Landslide

About midnight on June 20, 1996 subaqueous slope failures occurred about 50-70 m offshore creating waves and releasing bubbles (Figure 18). The failure retrogressed landwards and 25-30 minutes later consumed part of the adjacent highway, carrying away 250 m of shoreline killing one driver, before engulfing a house and killing 3 occupants. Failure activity continued for about one hour. Shoreline geotechnical investigations prior to the failures showed that there were about 10 m of soft sensitive clays, with horizons of quick clay and silt, overlain by up to 5 m of sand (Longva et al., 2003).

The undrained shear strength of the clayey sediments from these onshore investigations ranged from 10-50 kPa (Best et al., 2003), being between 25-40 kPa where they were less sensitive and 10-20 kPa in sensitive zones. Sensitivity varied from 5 to 35, though quick clay sensitivities of 40 to 60 were encountered in an area south of the slide. Leaching of these glaciomarine clays was thorough with measured salinities less than 1 PSU.
There were several stages in the failure sequence as shown in Figure 19. The initial detachment surface appears to have been a well-defined stratum with high-amplitude seismic reflectors. This initial failure led to subsequent retrogressive quick clay failures involving about $1 \times 10^6$ m$^3$ of sediment. This bed occurs over the entire area and is between 1 and 9 m thick. Although it has not been sampled, it is believed (Best et al., 2003) that the high-amplitude reflectors originate from gas trapped in porous sand, contributing to excess pore pressures. These authors focus on the possible ways by which gas in these sediments could have led to the observed failures. They state: “… the sensitive clays on the seabed at this location are not uniform … and sometimes contain ‘fabric’ (the non-uniform distribution of different particle sizes within the deposit, for example: varving; silt pockets within the clay; silt and sand particles along fissure planes). Fabric is extremely important for drainage of clays … since silt and sand sized layers and laminations provide a fast preferential drainage path for excess pore pressures …” The presence of gas bubbles, as was seen geophysically, can reduce the overall permeability by blocking the larger pores which carry most of the flow; excess pore pressures would thus be expected. Best et al. (2003) also suggest that gas bubbles can significantly weaken clay horizons, particularly when large (much larger than the mean particle size) bubbles form. Based on laboratory tests, they state that reductions in undrained shear strength occur associated with “bubble flooding”, at higher back pressures, and, thus, might be expected to be more significant at greater water depths.

According to Longva et al. (2003) “ … all available data suggests that excess pore pressure was the fundamental cause for triggering the initial slide”. Elevated pore pressures were the result of long periods of heavy rain and a high groundwater table prior
to the failures. In addition to natural sources, there was some evidence that leakage of water mains may have contributed to the elevated pore pressures. Other possible anthropogenic triggers included ground tremors from heavy traffic on rough roads and explosive detonations related to tunnel construction nearby. Geotechnical evaluations have shown that the beach slope materials had low factors of safety (≈ 1.1) prior to failure.

Figure 19. Slope failure sequence at Finneidfjord, Norway, June 20, 1996. (Longva et al., 2003)
6. Follafjord January 9, 1952

Follafjord is located about 250 km north of Trondheim. On January 9, 1952 a sequence of subaqueous failures was observed with associated waves (Terzaghi, 1956) and is shown in Figure 20. From Terzaghi (1956) the events were as follows:

“The dredge D shown at the upper left-hand corner of Figure [20] was tied by three anchors, #1 to #3, to the sea bottom and by two cables, #4 and #5, to the shore. On January 9, 1952, anchor chain #3 and cable #5 snapped on account of a local sand slide. The chain had a length of about 150 feet. The sea was calm. A few minutes later a wave with a height of four or five feet swept past the dredge. It came from the head of the fiord. After the sea had calmed down, about 14 minutes after the failure of anchor chain #3, anchor chain #1 and cable #4 snapped and anchor #2 dragged the dredge over a distance of about 1000 feet from its original location to the place D’ indicated in Figure [20]. The velocity at which the moving anchor pulled the dredge was estimated to be between eight and ten miles per hour.”

Figure 20. Map of Follafjord and the sequence of events of January 9, 1952 (Terzaghi, 1956). The slide scarp is shown by the hachured line.
The pier at the head of the fjord and the ferry boat landing pier (Figure 20) were both damaged by the sand slides.

Terzaghi (1956) concluded that the sequence of events began with a shallow sand liquefaction failure at the original site of the dredge. This liquefaction spread towards the head of the fjord; with distance the depth to which the sand underwent liquefaction increased. At the head of the fjord a major failure took place with liquefaction at great depth; this failure resulted in the observed tsunami. He concluded that following this large fjord-head failure, liquefaction spread again down-fjord, involving a much greater volume of sand and resulting in a significant failure at the site of the dredge. This failure moved the dredge and anchor #2 northeastward into the centre of the fjord; anchor #2 became so deeply buried that it could not be recovered. Terzaghi (1956) estimated the propagation velocity of the liquefaction “front” was between “… 4 and 6 miles per hour”.

Although he presents the alternative, Terzaghi (1956) discounts the possibility that the second large failure at the site of the dredge could have been caused by undercutting of the fjord sidewall slope by the passage of the channelized sandy flow slide/turbidity current emanating from the head of the fjord. This writer believes, based on observations in many other fjords, that the alternative explanation of erosion by channelized turbidity currents or flow slides is, in fact, quite probable and more likely than the scenario presented by Terzaghi of a rapidly back-and-forth migrating liquefaction “front”.

7. Orkdalsfjord May 2, 1930

Both Terzaghi (1956) and Andresen and Bjerrum (1967) discuss the major failure events in Orkdalsfjord, located about 30 km southwest of Trondheim, (Figure 21, 22) and Jørstad (1968) provides a brief summary with some limited details on the waves produced. Bøe et al. (2003) provide recent multibeam echosounding data for the area (Figure 23).

The failures took place in the early morning at the time of an exceptionally low tide (Andresen and Bjerrum,1967). The following observations are from these authors:

“(1) At Storaunet about 1000 to 1500 m$^3$ of a minor filling which had been recently placed along the shore disappeared at 7:48 a.m. in a slide that stretched 500 m along the edge of the fjord; it was preceded by the formation of a small wave some distance out in the fjord.

(2) A few minutes later a 600- to 700-m-long slide at Orkanger, approximately 2 km away, destroyed some piers and harbour constructions.
(3) On the opposite side of the fjord a slide took place at Furenstranden at 7:55 to 8:00 a.m.

(4) Three kilometers from Storaunet the Ofstad-Sandloken telephone cable crossed the fjord at a water depth of a maximum of 300 to 350 m. At 7:55 a.m. the cable was broken, indicating that the slide had propagated from Storaunet with a speed of about 25 km/hour.

(5) Approximately 18 km further along the fjord, the Vorpenesset-Stadsbygden cable crossed at a water depth of 500 m. Failure of this cable at 9:40 a.m. corresponded to an average speed of propagation of 10 km/hour since the occurrence of the initial slide."

The total volume of material involved in the three slides was estimated at $25 \times 10^6$ m$^3$ (Andresen and Bjerrum, 1967) though Karlsrud and By (1982) revised that to $60 \times 10^6$ m$^3$. Boreholes near slides A and B (Figure 22) revealed thick deposits of “very loose and soft nonplastic silt, having a water content of about 33 percent, corresponding to a porosity of about 49 percent” (Andresen and Bjerrum, 1967). They also state that it is “most likely that the water in the soil beneath the bottom of the fjord was under artesian pressure produced by high pore pressures in the fissures of the bedrock”.

Recent multibeam bathymetric data revealed in detail the slide scars of the 1930 event and showed areas of debris flow and slide deposits with large blocks (Figure 23; Bøe et al., 2003).

There appears to be little information on the waves except that Jørstad (1968) states that “… a large flood wave came which caused considerable damage along the shore” and that the wave was witnessed 30 km away.

Figure 21. Location map showing the location of Orkdalsfjord (left) in relation to Trondheim and the Nidelva River delta failure area (adapted from Bøe et al., 2003).
Figure 22. Map of Orkdalsfjord showing the locations of the three slides and the cables broken during the events of May 2, 1930 (Terzaghi, 1956).
Figure 23. Shaded-relief multibeam echosounding data from Orkdalsfjord (Bøe et al., 2003). Note that the scarp shapes and positions are somewhat different from those presented by Terzaghi (1956) and Andresen and Bjerrum (1967).
8. Nordfjord September 1967

Aarseth et al. (1989) provide a short account of the September 1967 failure and tsunami at Nordfjord located about 250 km southwest of Trondheim (Figure 24). The failure apparently occurred in relatively deep water (213 m) and produced three waves up to 1 m high 10 km away. A telephone cable was impacted (“stretched”) in this and in a previous failure and had to be lengthened both times. Comparison of hydrographic survey data collected in May 1967 and those obtained in April 1983 showed an increase in water depths of 55 m (213 to 268 m); Aarseth et al. (1989) estimate that the failure volume was $15 \times 10^6$ m$^3$.

A longitudinal high-resolution seismic profile shows that a 25 m step in the bathymetry moved 300 m to the east as a result of the failure (Figures 24-26).

![Figure 24. Bathymetry of Nordfjord showing (inset box) a comparison between soundings obtained in 1967 and 1983 and the eastward movement of the 25-m step at a depth of about 220 m. (Aarseth et al., 1989).](image)
Figure 25. Longitudinal “boomer” record along Nordfjord showing the “step” in the seafloor and the disrupted sediments to the west and stratified sediments to the east. Note that the vertical scale is in milliseconds two-way-travel time. (Aarseth et al., 1989).

Figure 26. Interpretation of a longitudinal “boomer” record showing the edge of the stratified sediments, the original position of the failed mass and the disrupted seafloor to the west. Note that the vertical scale is in milliseconds two-way-travel time (Aarseth et al., 1989).

Very little information exists about this subaqueous slope failure in Oslofjord, south of Hurum (Figure 27), and the associated tsunami event other than that reported by Jørstad (1968) which is presented here:

“On the calm waters of the fjords [sic] two waves suddenly arose. The first “foaming and dirty” wave was estimated to 3 m height, the second “shiny” wave to 6 m. The waves were observed both from the land and from the fjord. At one locality on Hurum traces of waves was [sic] seen 48 m from the shore and up to a height of 4 m above sea level. A wreck of a 60 foot boat was moved 350 m outwards and on the shore blocks up to 2 cu m in size were moved.”

Figure 27. Northern Oslofjord showing the location of the Hurum Peninsula. The precise location of the failure was not provided by Jørstad (1968), the only readily available source which refers to this event. The peninsula is approximately 10 km wide.
On a very low tide an estimated $3-4 \times 10^6 \text{ m}^3$ failed along the coastline near Sokkelvik (Figures 28, 29) (Jørstad, 1968). Nine lives were lost in the slide and 8 more individuals were caught in the slide but survived. There is some uncertainty as to whether the slide retrogressed from farther out in the fjord towards shore or originated at the coast. Fishing nets, placed at 100 m water depth 2.5 km from the slide area, were swept 6 km down the fjord. Four-metre high waves were produced and were detected 8 km from the source. There were three waves consisting of a smaller wave followed by two larger ones; there was an estimated 2 minute lag time between the first wave and the third wave (suggesting to this writer that there were several discrete tsunamigenic failures involved).

Figure 28. Location map of Norway showing the position of Sokkelvik.
Conclusions

Subaqueous slope failures and locally generated tsunamis have been frequent occurrences and have resulted in significant damage and loss of life. Tsunamigenic failures have apparently occurred from the coastline to water depths of more than 200 m. The following is a summary of some of the common observations related to these failures:

(1) Many events occurred at a time of very low tide.

(2) Exceptionally high precipitation (or rapid snow melt) accompanied some of the failures. These conditions led to significant artesian pressure at depth contributing to failure.

(3) Failures occurred typically in materials ranging from loose silty sands and sands, to quick clays. They were often found to be associated with sensitive clay strata.
(4) Styles of initial failure ranged from debris flows to flow slides. In some cases there was a clear evolution of failure style from debris flows through to turbidity currents.

(5) Over-steepening of slopes by toe erosion from flow slides or turbidity currents led to secondary failures and tsunamis in some instances.

(6) High pore pressures and artesian conditions were manifested in pock marks in deep water; near Trondheim they appear to have been associated with large failures. Similar pock mark features have been observed in parts of Douglas Channel, British Columbia.

(7) Many failures had an anthropogenic trigger such as emplacement of fill in coastal areas, construction vibrations (e.g., explosive detonations, heavy truck movements, tunneling) or highway traffic on rough roads.

(8) Failure volumes as small as $1.3 \times 10^6$ m$^3$ were capable of creating tsunamis with run up heights of at least 4-5 m. Volumes of $3-4 \times 10^6$ m$^3$ were able to create deep water waves of up to 4 m in height.
References


Appendix B

Bornhold, B, 2010, Coastal and Ocean Resources Inc.,
“Initial Results, Kitimat Arm Geophysical Survey”, Report to AMEC, 18 pp.

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Initial Results

Kitimat Arm Geophysical Survey

September 2010

Report Submitted
To
AMEC Earth and Environmental
Prince George, BC

by

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September 30, 2010
**Introduction**

Kitimat Arm witnessed two locally generated tsunamis in October 1974 and April 1975 which reached amplitudes of 5-10 m in much of the northern part of the fjord. These waves were created by underwater slope failures originating in the northeasternmost part of the inlet (1974), near the eastern edge of the Kitimat River delta, and in Moon Bay (1975) in muddy cohesive sidewall sediments.

Previous work by AMEC, WorleyParsons and Coastal and Ocean Resources engineers and geoscientists had identified subaqueous polygons around the inlet based on their seabed morphology and their likelihood of exhibiting significant thicknesses on steep slopes (Figure 1).

The intent of the September, 2010 geophysical program, carried out with Terra Remote Sensing Inc., a geophysical survey company, was to acquire high-resolution seismic profile data, especially in the polygons which were ranked highest in terms of tsunami hazard. The focus of the surveys was on the slopes of the basins; most survey lines were surveyed from as shallow as possible to the base of the slopes where basinal sediments could be seen acoustically. One hundred and forty-six seismic lines were surveyed ranging from a few hundred metres to several kilometres in length; most lines were shore-normal with less frequent shore-parallel tie lines. All of the polygons from 11 through 25 were surveyed with the exception of Polygon 4 (which was too shallow) and most of Polygon 3 where the coarse sediments on the modern Kitimat River delta resulted in virtually no acoustic penetration and therefore no means of assessing sediment thicknesses. Thus, all high priority polygons were surveyed in detail during the program.

Initially a reconnaissance suite of shore-normal lines was surveyed at a spacing of 200 m. Following this, areas of concern were were infilled to create a shore-normal line spacing of 100 m and shore-parallel lines were added, at depths appropriate to the geology observed in the shore-normal data, to create a dense grid. As well, shore-normal survey lines at a 100-m spacing were acquired off the area of the proposed NGP Terminal.

*The intent of this preliminary report is to provide an overview of the conditions as seen in the field during data collection and after a cursory study of the hard copy seismic profiler monitor records. More in depth analysis of the data and creation of sediment thickness maps requires analysis of the digital data collected during the survey work.*
Figure 1. Map showing locations of subaqueous polygons evaluated during sub-bottom profiling study
Instrumentation

The survey work was carried out using the 7.5-m jet boat “Doppler” owned by Terra Remote Sensing Inc. (Figure 2). Navigation was by GPS, yielding horizontal positional accuracy of about 2-3 m.

A 200 kHz narrow beam Knudsen echosounder was used for bathymetry throughout the survey with the data recorded digitally and as analogue sounding rolls.

A “boomer” seismic profiler with a 3-m hydrophone array was used to assess sediment thicknesses along the slopes. The catamaran and hydrophone array were towed off the stern on the port side of the vessel. Data were recorded digitally for later analysis with paper monitor records produced during the field program.

A sound velocity profile was conducted during the survey work to verify the velocity of sound in sea water to a depth of 50 m.
**Initial Results**

The seismic system was well able to penetrate presumed muddy sidewall sediments to bedrock (or possibly hard glacial materials in some cases) where they existed. In some areas, such as on the Kitimat River delta front and the subaqueous portions of some of the smaller sidewall fan deltas (e.g., Wathl Creek, Wathlsto Creek) dense sandy and gravelly sediments prevented acoustic penetration. Geophysical surveys on steep slopes, such as those in Kitimat Arm, is particularly difficult at times due to side echoes (returns from within the acoustic “cone” at the seafloor which are not directly beneath the vessel); thus, reflectors from, typically, shallower parts of the slope can appear within the sub-bottom record and be problematic in terms of interpretation. Biogenic gas in sediments (in bubble phase) can mask reflectors beneath it; while gas was not commonly observed, it did occur in one part of Polygon 13 in steeply dipping fine-grained sediments.

The polygons which exhibited the thickest and most extensive fine-grained sediments on steep fjord sidewalls were 13 and 22. Polygon 13 lies immediately north of, and overlaps with, the proposed NGP Terminal site. Polygon 22 lies on the east side of the inlet between the Kitamaat Small Craft Harbour and the north side of the Wathlsto Creek fan delta. Part of the north portion of Polygon 11 exhibited a zone of steeply dipping fine sediments. Polygon 20, earlier interpreted as the sidewall remnants of the October 1974 failure, showed significant remnants of fine sediments on very steep slopes.

The following are brief summaries of the results from Polygons 11 through 25 surveyed during this program. Sediment thicknesses were determined assuming an acoustic velocity for these soft muddy materials of 1500 ms⁻¹ (approximate speed of sound in seawater). Actual velocities may be somewhat higher (perhaps up to 1600 ms⁻¹) which would increase estimates by about 6%.

**Polygon 11**

Fine sediments occur to thicknesses of about 10-15 m in this polygon, mainly along the northern part (Figures 3 and 4). The sediments form an uneven drape over highly irregular bedrock. Slopes in this polygon average 18° between 90 and 110 m where significant sediment accumulations occur.
Figure 3. Shore-normal sub-bottom profile in Polygon 11 showing fine-grained sediment accumulation over bedrock.
**Figure 4.** Contour-parallel sub-bottom profile at about 100 m water depth showing sediment thicknesses of 10-15 m over irregular bedrock.

**Polygon 12.**

Slopes within Polygon 12 are very steep and no significant thicknesses of sediment were observed overlying bedrock.

**Polygon 13.**

Muddy sidewall sediments in Polygon 13 can reach more than 35 m in thickness and exhibit slopes of between 5° and 18° (Figures 5 and 6). In the northern part of the polygon in depths greater than about 80 m, slopes can exceed 25°; in these areas, sediments are generally less than 10 m in thickness.

There are no evident pre-existing sidewall failures within this polygon based on both the multibeam bathymetric data and the sub-bottom profiles.
**Figure 5.** Thick (up to 35 m) muddy sediments on the steep sidewall slope of Polygon 13.
Biogenic gas is evident in the central part of the polygon in water depths of between about 40 and 50 m. It is evident acoustically between 10 and 15 m below the seabed (Figure 7).
Figure 7. Shore-parallel seismic profile through Polygon 13 showing the thick transparent sediment sequence and the presence of biogenic gas.

Polygon 14

While there are no pre-existing slope failures in Polygon 13, at the southernmost end of Polygon 14, just north of the boundary with Polygon 13, there is a significant failure area and presumed debris flow accumulation at the base of the slope (Figures 8 and 9). The area of this failure zone is about 64,000 m².

In general there is little (< 5 m) sediment on most of the very steep slopes within Polygon 14 except at the southern end near the north end of Polygon 13. Slopes are up to 45⁰. The only accumulations are at the toe of the slope.
**Figure 8.** Shaded relief of a slope failure at the southern end of Polygon 14. The failure length is about 400 m and is about 160 m wide.

**Figure 9.** Seismic profile near the southern end of Polygon 14 showing the debris flow accumulation at the base of the very steep slope (see Figure 8).
Polygons 15 and 16

Sediments on the slopes within these polygon range from 15 m (20 m water depth) to more than 30 m (115 m water depth) near the site of the 1975 slope failure off Moon Bay (Figure 10). This polygon includes the source area of the 1975 event which created a major local tsunami; sediments within this trough created during the event are highly reflective, indicative of denser and/or coarser sediments (Figure 11). The trough is approximately 15-20 m deep. The north side of the trough (Polygon 16) is an intact zone which was apparently unaffected by the 1975 failure.

![Figure 10.](image)

*Figure 10.* Thick sediments on the slope in Polygon 15 south of the 1975 slope failure off Moon Bay.
Sediments south of the trough show thicknesses of more than 30 m (Figure 11). There remain 15-20 m of sidewall sediments beneath the trough floor.

The higher reflectivity of sediments on the intact north side of the trough (Polygon 16), resulting from its closer proximity to the Kitimat River delta front, prevents full acoustic penetration to bedrock; a minimum of 10 m of sediment can be seen in this zone. It is probable that sediments in Polygon 17 are a similar thickness to that seen in Polygon 15 (~ 30 m).

Figure 11. Trough created during the 1975 slope failure off Moon Bay. Note the more reflective sediments in the trough and on the north side of the trough and the thick sediments south of the trough.
Polygons 17 and 18

Polygon 17 is the north-trending branch trough believed to have been created during the 1975 slope failure event. Interestingly, however, the trough, unlike that off Moon Bay, is filled with about 2-3 m of fine-grained sediments overlying presumably coarse delta front sediments; the latter prevent acoustic penetration deeper with the sequence.

Polygon 18 is a morphologically smoother zone on the Kitimat Delta front (off the Eurocan facilities) and is characterized by highly reflective surficial sediments (presumably coarse delta front materials) preventing acoustic penetration more deeply within the sequence. No fine sediments were seen here except a thin (< 1-2 m) layer at the base of the slope.

Polygons 19 and 20

Polygon 19 lies at the eastern edge of the modern Kitimat River delta and is the principal locus of flow and sediment delivery to the deeper delta front. Within this polygon was the October 1974 failure which led to a significant local tsunami.

Sediments are generally more reflective than farther south along the fjord sidewalls rendering interpretation more uncertain. It appears that on the east side of the prominent trough which emanates from the principal distributary on the delta there remains up to about 20 m of sediment on slopes which can commonly reach 12° (Figure 12).
**Figure 12.** Sidewall slope in Polygon 19 along the east side of the Kitimat River delta. Note the thick sequence of sediments above the incised trough (left of image).

Sediments to the west of the trough on the delta front are very reflective (coarse) preventing acoustic penetration into the sequence.
Polygon 21

This polygon constitutes the delta front of Wathl Creek. As expected, no fine surficial sediments were encountered on this delta.

Polygon 22

Polygon 22 extends from the southern edge of the Wathl Creek fan delta to the north edge of the Wathlsto Creek fan delta. Between the latitude of the Kitamaat Small Craft Harbour to the northern side of the Wathlsto Creek fan delta there is a very extensive accumulation of fine grained sediments along the slope which can reach 22 m in thickness (Figure 13). These sediments occur primarily between about 80 m and the base of the slope, though shallower accumulations of thinner (5-10 m) materials also occur. Slopes below 80 m can commonly lie between 10 and 24°. Local steeper slopes exist within this polygon. No evidence of past failures was noted in either the multibeam bathymetric data or the sub-bottom profiles.

It is noteworthy that, unlike the Wathl Creek fan delta, significant thicknesses of transparent sediments have accumulated on the northern third of the Wathlsto subaerial delta. It is presumed that this side of the delta has been inactive for some time with the locus of coarse sediment deposition having been most recently on the southern and western sides.
Figure 13. Thick sequence of transparent sediments over bedrock within Polygon 22.

**Polygon 23**

This polygon includes the Wathlsto Creek fan delta. The southern two-thirds of the delta front consists of highly reflective coarse sediments which could not be penetrated acoustically.


Polygons 24 and 25

These polygons were characterized by very steep slopes with little or no accumulation of fine sediments.
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Kitimat Harbour Geophysical Investigation Report

Dear Shane

The report “Initial Results - Kitimat Arm Geophysical Survey, September, 2010” was followed by a more focussed and in-depth review of the data. This subsequent review substantiated the findings in the preliminary report.

In view of this, there was no need to revise and re-submit an updated summary of the geophysical results and the initial report stands. Other specific topical maps and interpretations were incorporated in other parts of the overall regional study.

Regards

Brian D. Bornhold, Ph.D., P.Geo.  
Vice President