NGU Report 2001.108

Hazard evaluation of rock avalanches; the Baraldsnes - Oterøya area

NGU
Norges geologiske undersøkelse
Geological Survey of Norway

Geological Survey of Norway
N-7441 Trondheim, Norway
Tel.: 47 73 90 40 00
Telefax 47 73 92 16 20

REPORT

Report no.: 2001.108	ISSN 0800-3416	Grading: Open

Title: Hazard evaluation of rock avalanches; the Baraldsnes - Oterøya area	
Authors: Blikra, Lars H.[1], Braathen, Alvar[1] & Skurtveit, Elin[2] ([1] NGU [2] NGI)	Client: Hydro Technology and Projects
County: Møre og Romsdal	Commune:
Map-sheet name (M=1:250.000) Ålesund	Map-sheet no. and -name (M=1:50.000) Brattvåg (1220 III); Vestnes (1220 II)
Deposit name and grid-reference:	Number of pages: 33 Price (NOK): Kr. 260,- Map enclosures:

Fieldwork carried out: September 2001	Date of report: 15.02 2002	Project no.: 2962.01	Person responsible: Øystein Nordgulen

Summary:

The report provides an evaluation of the hazard related to Skår/Baraldsnes area in accordance with a contract with Hydro Technology and Projects. Geological and historical data on rock-avalanche events demonstrate that the high-risk areas are concentrated to the inner part of deep fjords of western Norway. The studied locations, Baraldsneset, Oppstadhornet and Hellenakken, are situated outside the high-risk areas. Major avalanches cannot be excluded from this region, but the probability is much lower. There are no indications of larger-scale instability structures on Skårahornet near Baraldsneset, and the probability for large rock avalanches from this area are considered very low and at least lower than a yearly probability of 10^{-4}. Debris flows triggered along Kloppelva at the western slope of Skårahornet can reach the fjord in the western part of the considered area and a hazard zone with an estimated yearly probability of 10^{-4} is present.

Several areas exhibiting collapsed bedrock characterize the 700-m-high slope towards Oppstadhornet. There, distinct large blocks show intense fracturing and overall down-slope sliding both along the foliation and on cross-fractures. Surface disturbances at two locations indicate recent movement; these failures are younger than ca. 11500 years. The volume of the entire failure is estimated to be c. 20 mill m^3 (scenario 1), including a 2-3 mill m^3 large steep eastern part (scenario 3) that is characterized by fractures and fissures showing evidence of movements. A potential more limited slide from the upper parts of the collapsed area (scenario 2) may influence a mass of 5-7 mill. m^3. This part constitutes the unstable top layer modelled by NGI. Estimates of potential run-out distances indicate that a large rock avalanche involving the entire area may go 0,7 – 1,2 km into the fjord (scenario 1), whereas the uppermost top of the slide may reach 0,5-0,7 km (scenario 2) and the steep eastern part 0,5 – 0,9 km (scenario 3) into the fjord basin.

Probability estimates are difficult due to lack of measurements of potential movements and inaccurate earthquake predictions. It cannot be excluded that a large-size rock avalanche could occur from the Oppstadhornet area, but with an annual probability of less than 10^{-4} (senario 1). However, a future slide will most probably occur in one of the upper sliding surfaces in areas of higher gradients (scenario 2), with an estimated annual probability in the order of 10^{-4}. The estimated annual probability for scenario 3 is also in the order of 10^{-4}. A test of ongoing movements can be conducted through Differential Radar Interferometry (dInSAR). In long term, it is recommended that the entire unstable area is monitored with respect to potential movements.

Along a steep mountain east of Baraldsneset, at Hellenakken, there are open fractures indicative of relatively recent movement and therefore instability. An estimated volume of 2.5-5 mill m^3 may be involved in the unstable mountain slope. It cannot be excluded that a failure can occur from Hellenakken with an annual probability in the order of 10^{-4}. It is concluded that most of the volume will be deposited onshore, but a volume of less than 0,5 mill m^3 may enter the fjord. It cannot be excluded that a larger portion of the slide may reach the fjord (1-2 mill m^3); the annual probability for this is considered to be low (less than 10^{-4}).

Keywords: Risk	Hazard	Rock avalanche
Slide	Bedrock failures	Creep
Debrisflow	Earthquake	Stability

INNHOLD

1. Introduction

In accordance with a contract with Hydro Technology and Projects (No. NHT-B44-5120825-00), this report presents the results of a geological risk evaluation related to the Skår/Baraldsnes area. The project is based on the description given in the technical note of 15.08.2001 from the Geological Survey of Norway (NGU), Norwegian Geotechnical Institute (NGI) and the County Council of Møre & Romsdal.

The report is based on work related to the following main areas:

- Regional distribution of large rock avalanches and possible unstable rock slopes in Møre & Romsdal
- Skår/Baraldsneset
- Oterøya/Oppstadhornet
- Hellenakken

A hazard evaluation for large rock avalanches from the steep Skåregga Mountain near Baraldsneset is performed on the basis of geological studies. The probability that a large rock avalanche from the unstable part of Oppstadhornet, on the other side of the fjord, can reach the fjord is evaluated, and the volume and possibly run-out distance of a potential event is estimated. Possible unstable features on the steep mountain Hellenakken, west of Baraldsnes, are evaluated in more general terms.

The fieldwork was carried out by Lars Harald Blikra (NGU), Alvar Braathen (NGU), Elin Skrutveit (NGI) and Einar Anda (The County Council of Møre og Romsdal). Peter Robinson (NGU) supplied data on the bedrock geology and has together with Einar Anda reviewed the report. This report is prepared in close interaction with NGI and their separate reports (NGI Reports 20001472-2 and 20001472-3).

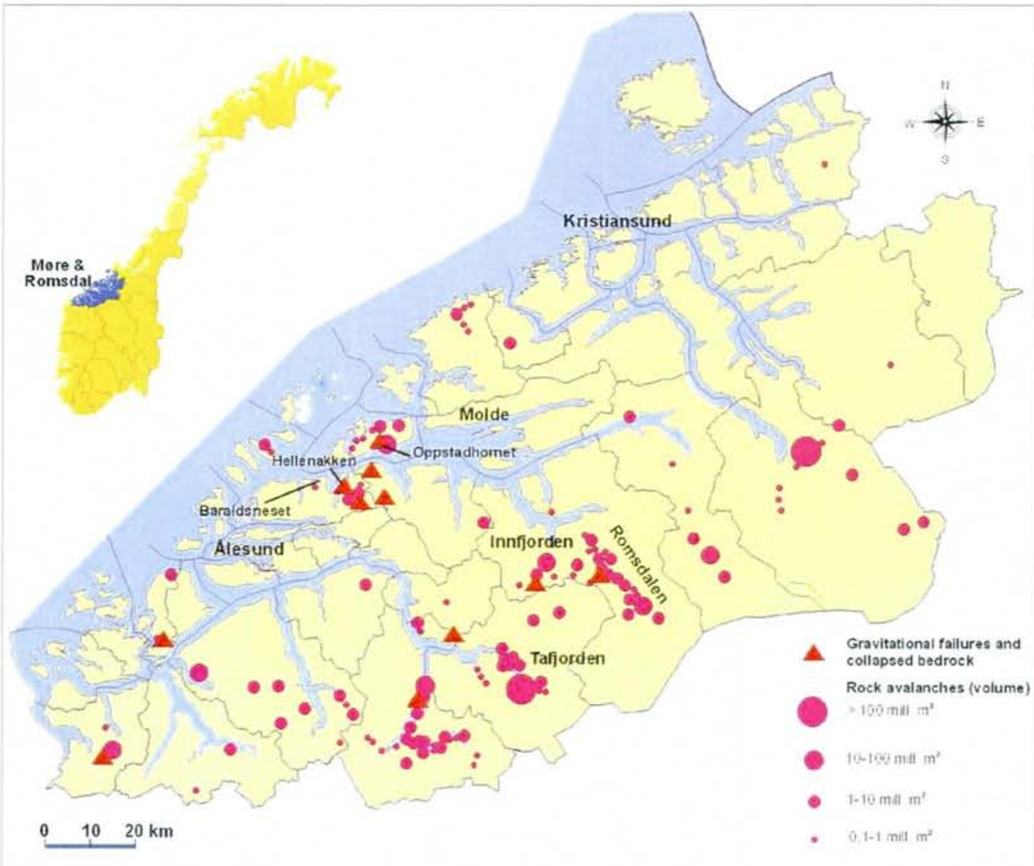


Figure 1. Distribution of rock avalanches and potential unstable rock slopes in Møre og Romsdal.

2. Regional distribution of rock avalanches and unstable rock slopes

The Geological Survey of Norway, together with the County Council of Møre & Romsdal, and with financial support also from Statens naturskadefond, has performed regional mapping and, at some sites, detailed geological studies related to rock-avalanche hazard (Blikra & Anda, 1997; Blikra *et al.* 1999). This project started in 1996 and has included both studies of past rock avalanches and detailed studies of potential source areas for such events. Tsunami waves generated by large rock avalanches have caused major disasters in Norway. Many of these events are restricted to relatively limited zones in the counties Møre & Romsdal and Sogn & Fjordane. The Tjelle slide in 1756 is closest to the present investigated area; an event that caused major flood waves noticeable more than 40 km from the site. Geological studies confirm that several zones have been affected by large-scale rock avalanches throughout the postglacial period (Blikra *et al.*, 1999). Furthermore, a significant number of rock-avalanche events have occurred within these areas/zones (Fig. 1), and many of them are from the latest part of the postglacial period (the last 5000 years). Major tsunami waves related to rock avalanches in general represent a higher risk than other types of avalanches in the region, due to their spatial coverage, and high impact on urban or densely populated areas.

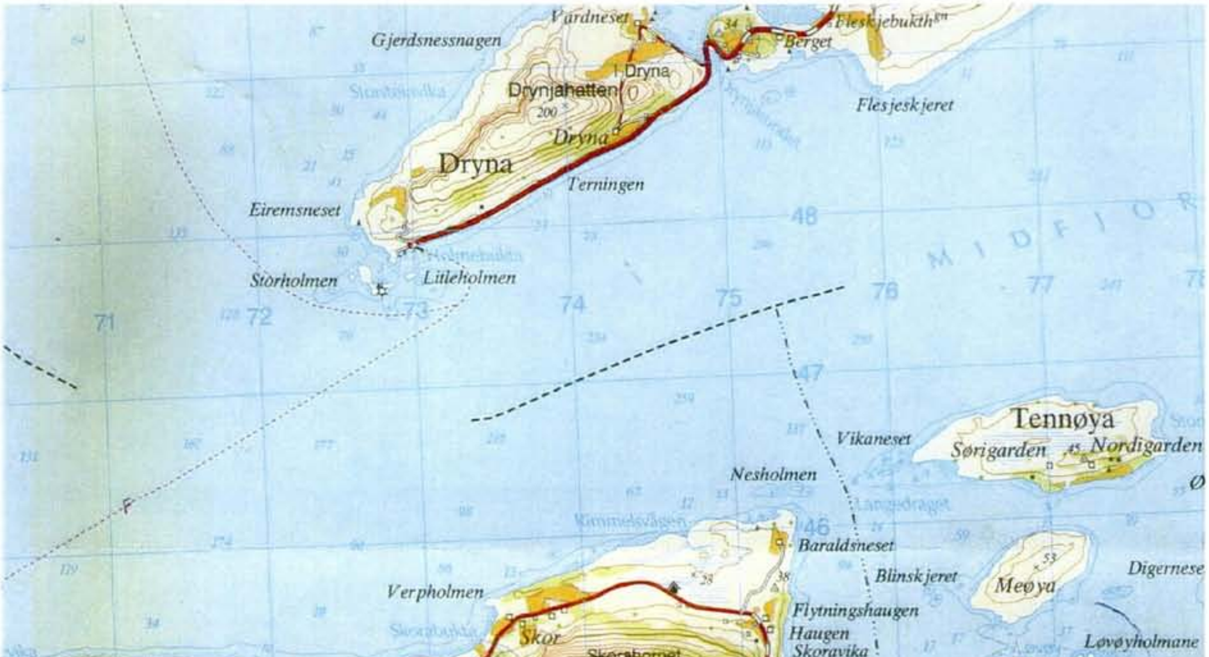


Figure 2. Topographic map locating the Baraldsneset area (map sheet 1220 III). **Each square of the map is 1 X 1 km in size.**

3. Skår/Baraldsneset

The area in question is located on a peninsula between Midtfjorden and Vatnefjorden, named Baraldsneset (Fig. 2), approximately 30 km WSW of Molde. It consists of a fairly flat lowland with exposed bedrock along the coast, which is bound in the south by a north-facing, 400 m high mountain-side up towards the Skårahornet.

Figure 3. View from the NE towards the Baraldsneset peninsula and the mountain Skårahornet.

3.1 Skårahornet

In the lowland, thin zones of quartzite, amphibolite, garnet-mica schist, granitic augen gneiss, and meta-gabbro are the main lithologies (Robinson, 1995; Tveten *et al.*, 1998). The mountain-side and mountain is made up of biotite gneiss and amphibolite. All rocks are well foliated, with an overall E-W trend and sub-vertical dip.
The mountain-side is steep, although with some variation. Parts are sub-vertical, but more commonly there is a stair-step like geometry above a fairly steep (~40-50°) basal slope. Most of the mountain-side is covered with dense vegetation. In the west, there are Quaternary deposits that, in addition to rock falls, constitute another hazard.

On the mountain above Baraldsneset, there are no signs of significant instability. At places, fractures reactivate the steep foliation, and some cross-fractures divide the rock into large blocks (Figs. 3, 4 and 5). Fracture frequencies[1] (ff) are low, generally in the range of 1-2 fractures/metre. One minor fracture zone, oriented nearly perpendicular to the slope, was identified. Here, ff reached 3-4. Importantly, on the mountain and especially marginal to the steep slope, no observed blocks appear to be unstable. The only observations of instability

[1] Fracture frequency is recorded as the number of fractures that intersect a metre-long line-segment.

were found along the upper part of the slope. There, cm-wide, open crevasses along the upper and inner margin of blocks (Fig. 5) indicate movement by rotation. A fall-out or toppling process generally causes collapse of single blocks.

Figure 4. The Skårahornet mountain viewed from the ENE.

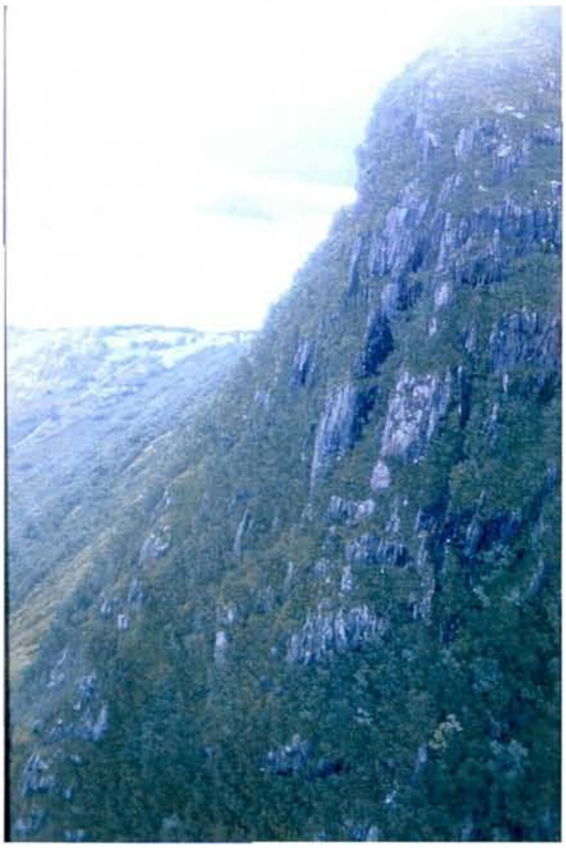


Figure 5. The Skårahornet mountainside viewed from the west. Note the step-like geometry caused by the foliation and block fall-out from the rock-face.

3.2 Debris-flows at Kloppelva

The small valley along Kloppelva, south of Skår, are characterized by a relatively thick cover of glacial till. Kloppelva marks a deep incision (debrisflow channel) into this till cover, and debrisflow tracts also occur further east (see Fig. 6). Debrisflows along Kloppelva have developed a large fan extending from 100 m above sea level to the shoreline. Some of the debrisflow events that have occurred are relatively large with up to 2 m high lateral levees and individual boulders more than 2 m in diameter. There are also distinct slide scars in the glacial till further east (Fig. 7), and two well-defined debrisflow channels run down against Skår (Fig. 6). Blocks triggered from the western slope of Skårahornet may trigger debrisflows in the glacial-till and such events may reach the depression just south of Einargarden.

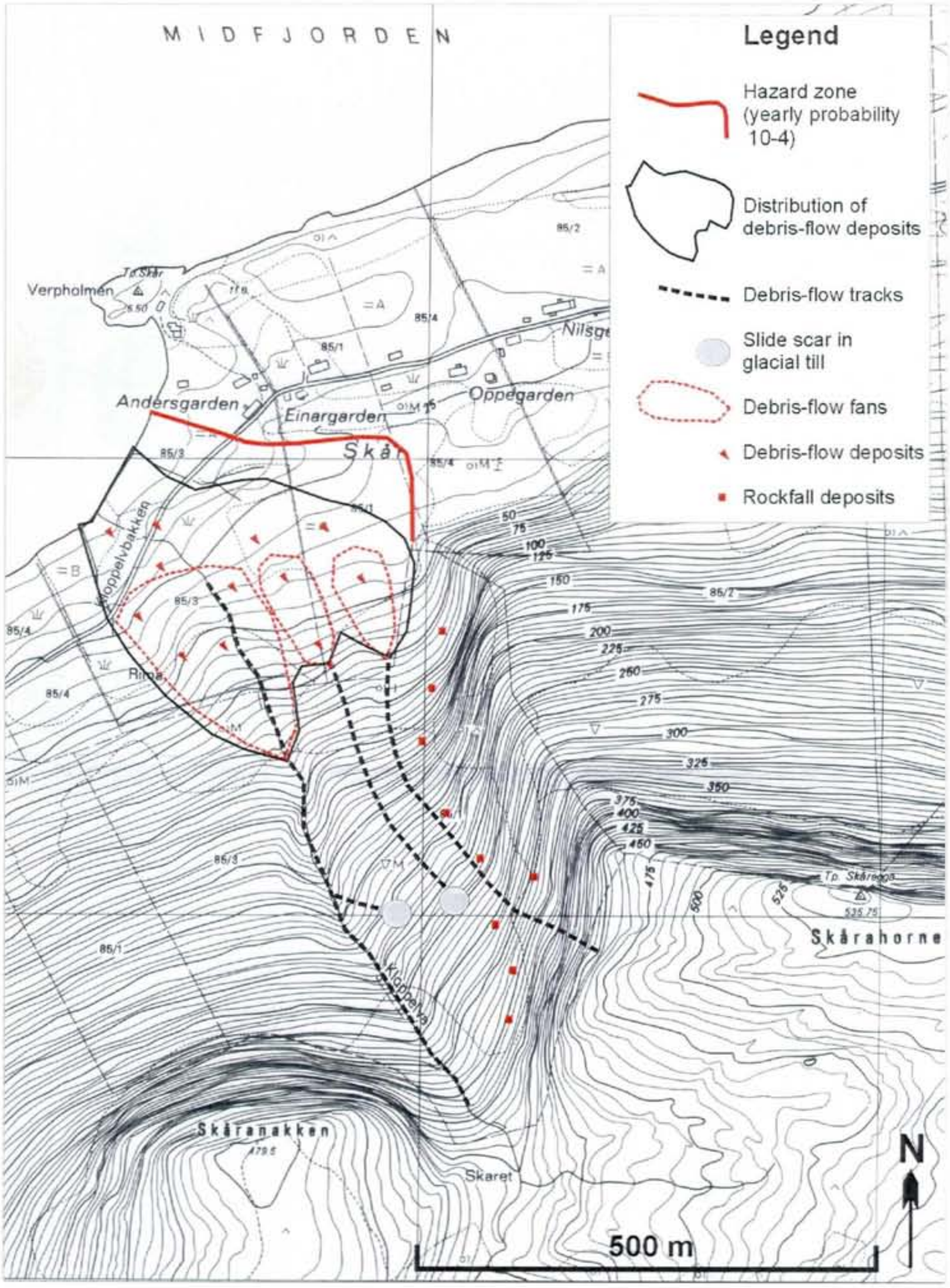


Figure 6. Map from the Koppelva area showing features formed by debrisflow activity and with the estimated hazard zone of an annual probability of 10^{-4}.

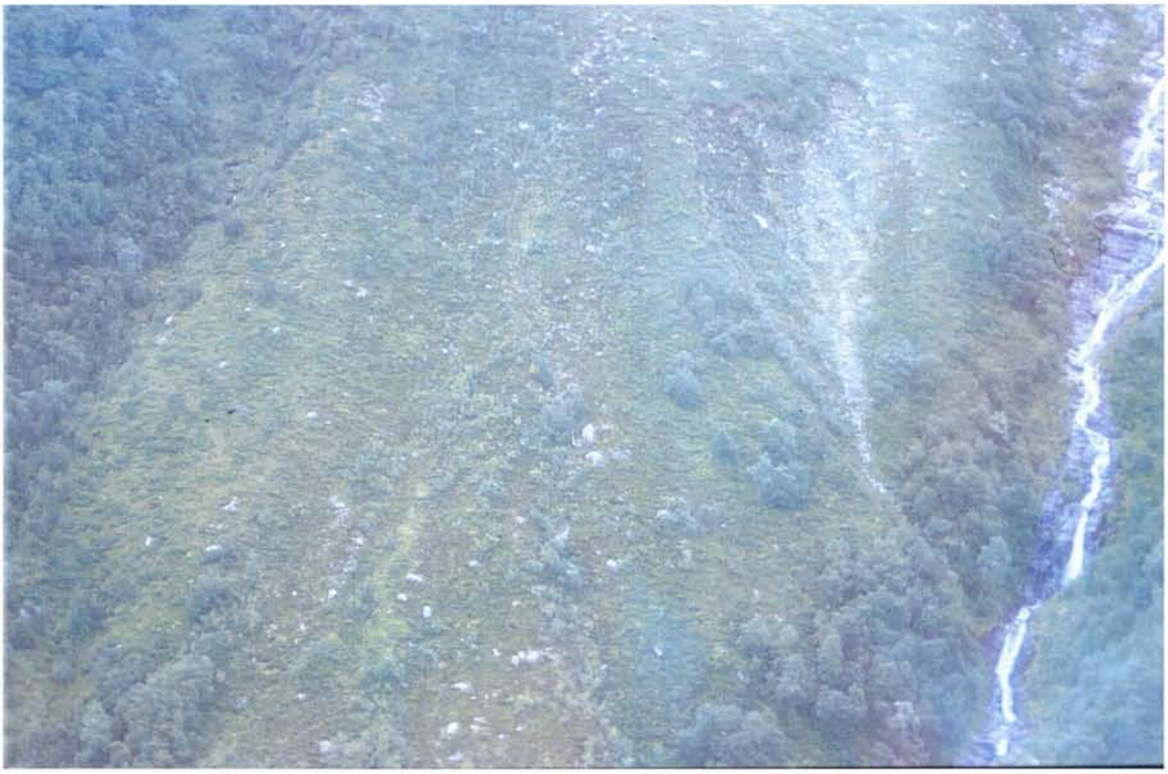


Figure 7. Photograph from Koppelva showing a slide scar from debris flow. The small valley is filled with a relatively thick glacial-till cover

4. Oterøya/Oppstadhornet

An unstable mountain-side has previously been reported (Robinson *et al.*, 1997) along the southern slope of the island Oterøya, approximately 15 km west of Molde (Figs. 8 and 9). Within a several km wide and up to 700 m high area, actually stretching from near-shore to the peak of Oppstadhornet, open crevasses and clefts reflecting block sliding, are found (Fig. 10). The slope of the soil-covered mountain slope varies from a moderate ~20-30° angle in the lower and western part, to bare and steep cliff-faces in the upper and especially eastern part. Under the steeper portions, talus and block fields make up the deposits. Further down, the area is vegetated with dense forest, mostly hiding geological features.

Rocks of the area include granitoid gneiss that hosts two zones of schists, and subordinate meta-gabbro. All rocks are well foliated, typically with a steep to moderate dip to the south, i.e. towards the fjord.

4.1 Geological description

The area can be divided into three parts: (A) a large slide-block field, (B) the steep mountain face to the east, and (C) clefts in the lower, forested part.

Slide-block field

The slide-block field is approximately 600 m wide and 1 km long/high, and is characterised by significant internal faulting/sliding marginal to larger collapse structures (Figs. 10 and 11). The field is bound by three key structures. In the west, a NW-SE striking and steeply NE dipping fracture-zone, reactivating a Devonian (?) fault, is clearly visible. The fracture-zone forms a transfer structure that has seen highly down-oblique sliding, and that forms the western termination of the collapsed field. Above the field, the main, upward-bounding structure is clearly displayed as an escarpment (Figs. 12 and 13). This structure is

superimposed on the steeply SE-dipping foliation as well as a narrow zone of Devonian (?), carbonate-cemented fault breccia. The upward bounding structure acted (or acts) as a master fault during collapse, as is easily seen by the large escarpment, steep and up to 20 m high. Rocks of the hangingwall are highly shattered, whereas the footwall section reveals moderate ff (3-4) that promptly diminishes away from the escarpment. In the east, the termination of the collapse field is identified as a steep NNW-SSE striking and steeply SW dipping structure, which has acted as a transfer fault.
Internally, the slide-block field is segmented into large blocks by two sets of steep fractures; NW-SE and NE-SW striking. The latter set is sub-parallel to the master fault and the foliation in the rocks. The fault sets are different in that they show down-S transport on the NW-SE fractures and down SE transport on the NE-SW fractures, respectively. Many of the blocks in the field have rotated: surface blocks rotate out from the slope, i.e. clockwise, in addition to the general down sliding, whereas the overall pattern of the collapse field is that of rotation toward the slope (anticlockwise) above a basal detachment.

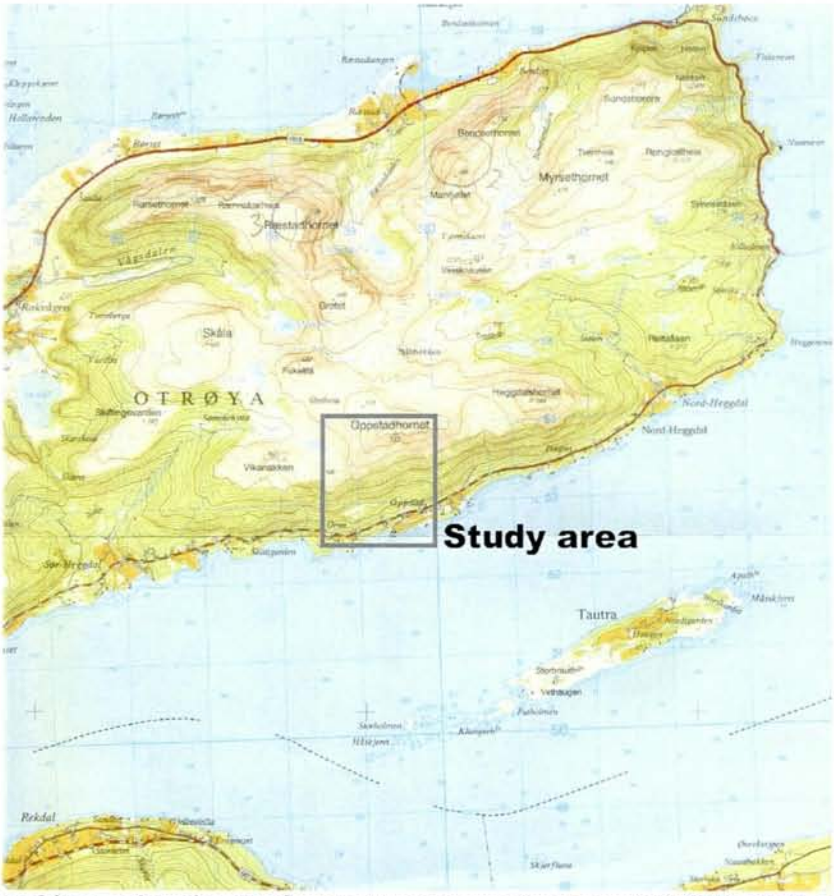


Figure 8. Topographic map locating the Oppstadhornet area (map sheet 1220 II). Each square of the map is 1 X 1 km in size.

Four blocks with different deformation style have been identified within the slide-block field (see blocks of Figs. 9 and 10). A southwestern block (block 1) is moderately shattered. Internally, this block shows indications for minor down-SE sliding (< 1 m) on SW-NE striking fractures; however, the main structure is the bounding NW-SE cross-fracture zone to the NE, which reveals down-S movement. The next block to the NE, block 2, is deformed in a similar fashion, albeit deformation is more intense. There has been moderate (< 2 m) down-SE movement on SW-NE striking fractures, and some NE-SW striking cross-fractures reveal down-S and sinistral separation. Some also appear as open fissures, consistent with sub-

horizontal movements. Northeastward, the block margin is shown as a zone of NW-SE cross-fractures, clearly visible on photographs (Figs. 10 and 11), which shows down-S movement.
Block 3 constitutes the largest feature of the slide-block field. Internally, it is segmented by two NE-SW striking normal faults with down-SE separation in the range of 3-7 m, as revealed by escarpments (Fig. 14). The upper normal fault links up with a NW-SE transfer structure, seen as a several-metre wide open cleft clearly visible on photographs (Fig. 15).
Block 4, found below blocks 1, 2 and 3, is different from the other blocks in that it is dominated by roughly shaped, side-stepping and open NW-SE striking fissures that partly link up with foliation-parallel fractures (Fig. 16). Its lower part forms a highly fractured bulge with a steep front, which may delineate the tip-line of a slope-parallel detachment within the slide-block field (see cross-section, Fig. 9b).

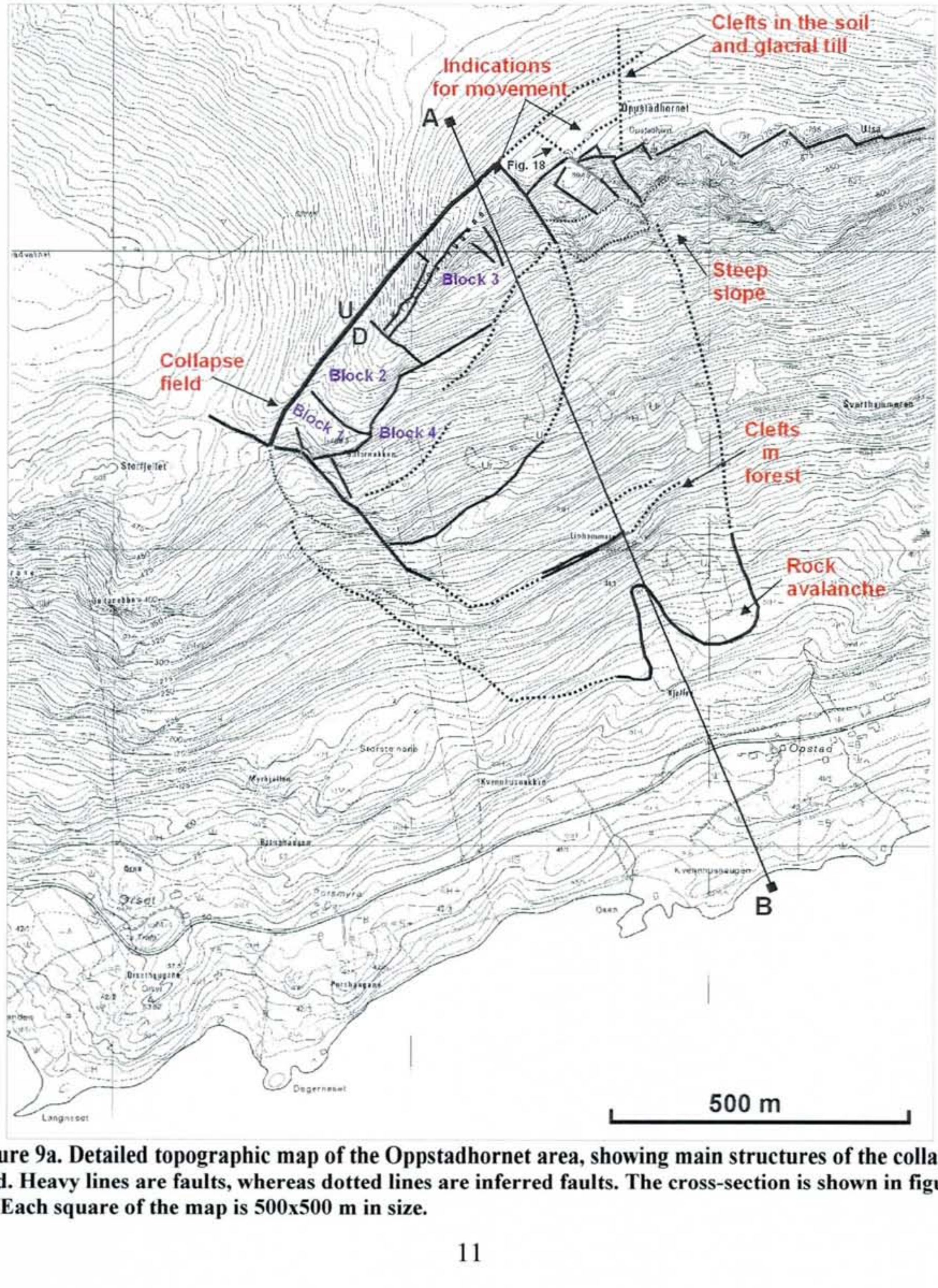
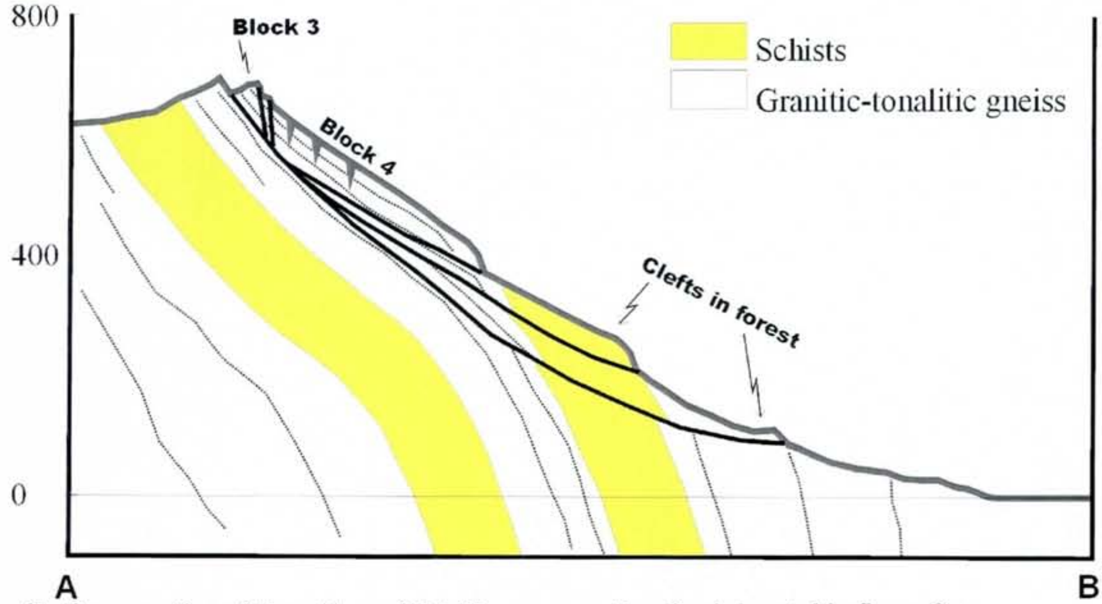


Figure 9a. Detailed topographic map of the Oppstadhornet area, showing main structures of the collapse field. Heavy lines are faults, whereas dotted lines are inferred faults. The cross-section is shown in figure 9b. Each square of the map is 500x500 m in size.



Figur 9b. Cross-section of the collapse field. The cross-section line is located in figure 9a.

Steep mountain face to the east

The steep slope to the east is characterised by fractures and fissures that reactivate the steeply SSE dipping foliation, and that generate moderately sized blocks. In addition, several mega-blocks are identified (Fig. 9a). They are bound by similar fractures in the back, and some have slid 10-30 m down without being significantly internally shattered (Fig. 17).
On the plateau around Oppstadhornet, there are several 1-2 m deep clefts in the soil and moraine. These relatively small structures indicate that fractures similar to those in and around the slide-block field are partly reactivated in this area as well (Fig. 18).

Lower part of the slide-block field

In the lowland, near the foot of the mountain-side, at least two clefts have been observed. They are found approximately 10-m above small cliffs, and are characterised by blocks that have sunk 2 to 5-m, i.e. they appear as small grabens parallel to the cliff-face. From a hazard point of view, they are insignificant, however, they indicate that the entire slope down to sea level have experienced or experiences movements. One can speculate that a basal detachment for the overlying slide-block field actually emerges in this area, near sea level (see cross-section, Fig. 9b).

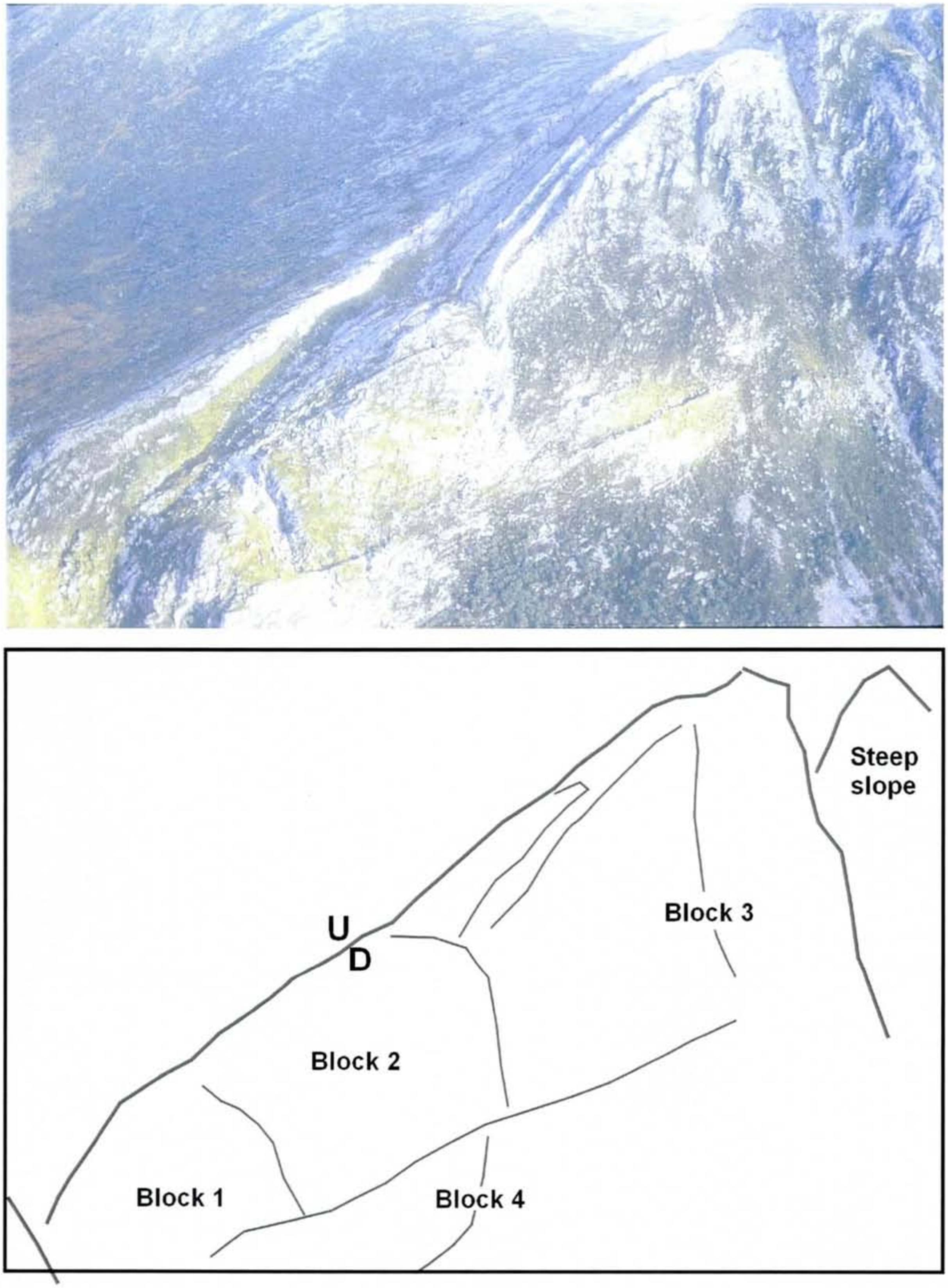


Figure 10. View from the south of the mountain-side below Oppstadhornet. Main features of the collapse field are displayed in the line drawing.

Figure 11. View from the north, showing the master fault in the back of the collapse field, and a major, wide-open cross-crevasse.

Figure 12. View from the southwest of the collapse field of Oppstadhornet. Note the bounding master fault and both strike-parallel and cross-fractures that subdivide the area into several blocks.

Figure 13. Close-up photograph of the master fault in the back of the collapse field. The rock-face, or escarpment, is 10-15 m high. The view is to the north.

Figure 14. Internal foliation-parallel faults in block 3 of figure 10. View is down-hill to the southwest.

Figure 15. Open crevasse that link up as a cross-fracture to one of the foliation-parallel faults of block 3. Note person for scale. View is to the southeast.

Figure 16. Internal fracturing of block 4 of figure 10. View is down-slope to the south.

Figure 17. Photograph towards southwest down-slope the steep mountain side near Oppstadhornet. Note the large block in the centre, which has slid down without breaking apart.

Figure 18. Small clefts at the ridge near Oppstadhornet. Such clefts reflect opening of crevasses/fractures in rocks below the topsoil. View is towards Oppstadhornet, from the WSW.

4.2 Age of deformation and possibly present movements

The slide feature is formed after the deglaciation of the region since it shows no sign of glacial reworking. The deglaciation occurred ca. 14.000-15.000 years ago. The slide area does not shows indications of being affected by processes related to permafrost conditions. The latest phase with permafrost conditions in the area was during the cold Younger Dryas period, ca. 13.000-11.5000 years ago. It is thus concluded that the entire slide area was formed after ca. 11.500 years ago, and that the failure is not connected to unstable conditions shortly after the deglaciation.
At two locations, there are indications for recent movements (Fig. 9a). At the foot of the master fault escarpment, near its NE termination, a depression or a collapsed part in the lower part of a the rockfall cone indicate modern movement. Collapsed soil and sediments into clefts or crevasses on the plateau near Oppstadhornet may also indicate quite recent movement. However, measurement of possible movement is needed in order to confirm or rule out present instability.

4.3 Volume estimates

Volume estimates are based on; (i) the area affected by failures, and (ii) a thickness of between 50 and 75 m in the upper part, and (iii) ca. 25 m in the lower part of the slide. The volume of the entire possibly unstable mountain-side is then estimated to be 10-20 mill m^3 (maximum 20 mill m^3, hereafter named scenario 1 of section 7).
A potential slide of a more limited upper part of the main slide feature is named scenario 2 and may influence a mass of 5-7 mill m^3. This part constitutes the unstable top layer modelled by NGI (Bhasin & Løset, 2002).
The maximum estimate of 20 mill m^3 also includes the steep eastern part that is characterized by fractures and fissures showing evidence of movements. This part, which is situated at higher altitudes (705-730 m above sea level), is estimated to involve a volume of 2-3 mill m^3 (hereafter named scenario 3).

5. Hellenakken

Possibly open crevasses have been observed by an air-photo reconnaissance study on Hellenakken (400-450 above sea level), and by observations from the helicopter (Fig. 19). The bedrock on Hellenakken consists of biotite gneiss with zones of amphibolite and with a subhorizontal foliation. The open fracture on the Hellenakken mountain ridge can be observed for about 500 m in length and is situated 35 to 75 m from the mountain edge to the west (Figs. 20 and 21). With an estimated thickness of the possibly unstable zone of between 100 to 200 m, this gives an estimated total volume of 2,5 to 5 mill m^3. No detailed mapping of this structure has been performed.

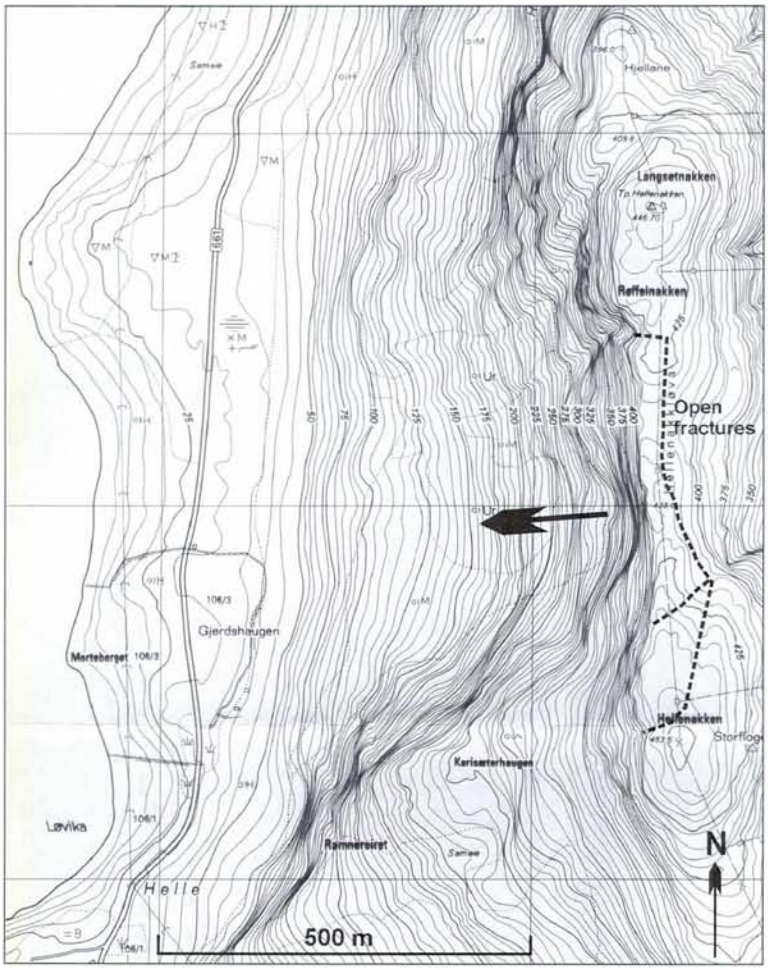


Figure 19. Map of the Hellenakken area with the localization of open fractures observed from aerial photographs and from the helicopter.

Figure 20. View from the south of the crevasses found along the ridge of Hellenakken. Oppstadhornet can be seen in the background.

Figure 21. Close-up of photograph of figure 20, showing the cleft(s) to be several metres deep.

6. Possible triggering mechanisms for rock avalanches

Triggering mechanisms for large rock avalanches in the region is poorly understood, but extreme rainfall building up high water pressure in fractures, and high-amplitude earthquakes are probably the most important factors. Increased movement seen as widening of clefts, have been observed for quite long periods before rock-avalanche events (e.g. Tjelle, Tafjord and Loen slide). This suggests that gravitational creep can be active for long periods, and the failure itself does not necessarily need to be controlled by earthquake or rainfall events.

6.1 Creep

Creep is a relatively slow process that in many cases seems to occur prior to rock-avalanches, as hinted at above. This is well illustrated by a disastrous avalanche in Italy (Erismann & Abele 2001), which caused collapse of a hydropower dam and thereby flooding of a populated valley. During filling of the dam, creep was recorded along one of the slopes of the dam margin. With increasing water level, i.e. increasing pore pressure in the rock, creep accelerated. Because of this, the filling of the dam was stopped, resulting in diminishing creep that gradually was reduced to zero. At this stage it was concluded that the unstable area had stabilised, and filling of the dam was sustained. Some hours later, the entire unstable area avalanched into the dam. This suggests that, in this case, initial creep resulted in strain softening of the basal shear layer, hence reducing the critical basal shear strength.

In order to trigger creep for an unstable area, the total forces acting on the basal sliding surface have to overcome the critical frictional resistance. Two common mechanisms for reduced friction, acting either isolated or in symphony, are; (I) increased pore fluid/water pressure along the shear surface, and (II) weathering of the shear surface. It is important to

keep in mind that these processes are time dependent; hence stability evaluations for a given time are temporally unreliable.
Case I: Increased water pressure reduces the normal stress on the shear plane, thereby reducing the shear strength. Creep in this context means that sliding starts at a critical value of pore pressure, and that the pressure immediately drops below the critical value when movement takes place. Hence, captured pore pressure is released during microscopic slip events, which appear repeatedly. If the pressure is preserved, or increased, rather than released, the slip event will develop and accelerate, and likely evolve into a full-scale rock avalanche, unless some mechanism blocks the development, for example an obstacle at the toe of the sliding rock body.
Case II: Weathering of the basal shear plane can be viewed along two lines: (a) Mechanical and/or chemical breakdown of microscopic obstacles on the surface which are stress localisers. Thereby, the smoothness of the plane is enhanced, commonly resulting in reduced friction. (b) Chemical alteration of fractured and damaged minerals at or near the shear surface that, for example, may result in formation of new clay minerals. Such minerals may lubricate the shear plane, and therefore reduce the shear strength.

In all cases, as soon as a critical shear value is reached, movement will accelerate unless other factors act in the opposite direction. Accelerating movement result in destruction of minerals or clasts at the basal shear surface, which then develop into a soft layer of crushed rock (gouge and breccia), lubricating and reducing the shear strength, i.e., a strain softening process. Development of a gouge zone, and especially if it becomes continuous, dramatically reduces the stability of a collapsing area.

Figure 22. Rotated, highly unstable blocks close to the toe of block 4.

Why is a creep process most likely for the Oppstadhornet area? Firstly, there are indications for total down-slope movement in the range of 25-50 m of the unstable field. If this displacement happened in one event, it most likely would have developed into a full-scale avalanche. Secondly, there are several indications for creep. One observation is the rotated blocks that indicate they have been squeezed out (Fig. 22); another is the localised collapse of the soil above fractures. The latter actually indicate that creep is ongoing. A test of slow creep versus punctual slip can be conducted through Differential Radar Interferometry (dInSAR), a method at present being explored at the NGU. Such data have been recorded since 1992 and, when combined, has a vertical resolution of ca. 0.5 mm. Our conclusions agree with that of Harbitz (2002). They argue that if the slide were triggered off in one event and then slid down several 10s of metres, it would evolve into a full-scale avalanche.

6.2 Earthquakes

A review of historical and recent data on seismically induced landslides shows that the full range of landslide types can be initiated by seismic activity, and 81% of all slope failures occurred in the region with mean horizontal peak ground acceleration (PGA) greater than 1.5 m/s^2 (Sitar & Kahazai, 2001). Seismic-hazard analysis for Norwegian onshore and offshore areas shows that Oppstadhornet is found between the 2.7 and 3.0 m/s^2 PGA contour lines for annual probability of 10^{-4} and between 1.0 and 1.2 m/s^2 PGA contour lines for annual probability of 10^{-3} (NORSAR & NGI 1998). A peak ground acceleration of 2.7-3.0 m/s^2 is comparable to an earthquake with magnitude Ms 6,5 within a distance of 20-30 km, whereas a PGA of 1.0-1.2 is comparable to magnitude around Ms 6 for the same distance. It is worth nothing that the predicted values of PGA contour lines is based on the historical earthquake database and have not been analysed with respect to geological structures.

The earthquake distribution of Southern Norway from 1750-1999 shows that Oppstadhornet is located 100-200 km away from the area with high earthquake activity found off and along the western coast of Norway (Fig. 23; Dehls *et al.* 2000). Earthquakes of magnitude 6 or higher have never been recorded in Norway.

Figure 23 shows earthquake data (NORSAR) plotted on the lineament map with a Landsat 7 satellite image base (NGU's structural database). Most earthquakes are found offshore northern West Norway, where the NE-SW striking Møre-Trøndelag fault complex intersects the N-S trending Bergen lineament zone. There, a limited number of earthquakes have been of magnitude 5 to 6. It is worth mentioning that only a few earthquakes have been recorded from the Molde region, which is situated along the SE margin of the Møre-Trøndelag fault complex. This major fault zone is likely dominated by creep (Pascal & Gabrielsen 2001), hence, it is an open question whether or not sufficient build-up of stress may occur along the zone. However, if the fault complex periodically locks, significant earthquake swarms can be expected. A recently detected neotectonic fault in Innfjorden 35 km SSE of Molde (Anda *et al.*, in prep.) is connected to a zone characterized by many large rock avalanches (Fig. 23). A possible continuation have been observed further south. The earthquake database does not reveal recent neotectonic activity on these structures.

The review made by Jibson (1996) on the use of landslides for palaeoseismic analysis shows that one normally needs an earthquake of about magnitude 6 to trigger rock avalanches. This author also concludes that there is a relationship between earthquake magnitude and the size of the area affected by earthquake-triggered landslides. If we use these data in the present case, it can be concluded that an earthquake of at least $M > 6,5$ on the neotectonic fault in Innfjorden is needed in order to affect the slide block on Oppstadhornet (45 km from the Oppstadhornet area, Fig. 23). However, the regional distribution of rock-avalanches shows that large earthquakes associated with the neotectonic fault did not trigger avalanches very far to the northwest in the direction of Oterøya (Fig. 1). Another possibility, large earthquakes from the offshore area, more than 100 km from Oppstadhornet, requires at least an earthquake of M > 7-7,5 to have an effect on the collapsed slide block on Oppstadhornet. If such large earthquakes (greater than magnitude M 7) have been common in the past, there should be geological evidences in the form of neotectonic faults both in the offshore and onshore areas, and possibly widespread rock-avalanches in steep terrain in the areas further to southwest from the Oppstadhornet area. This is not truly supported by the geological record.

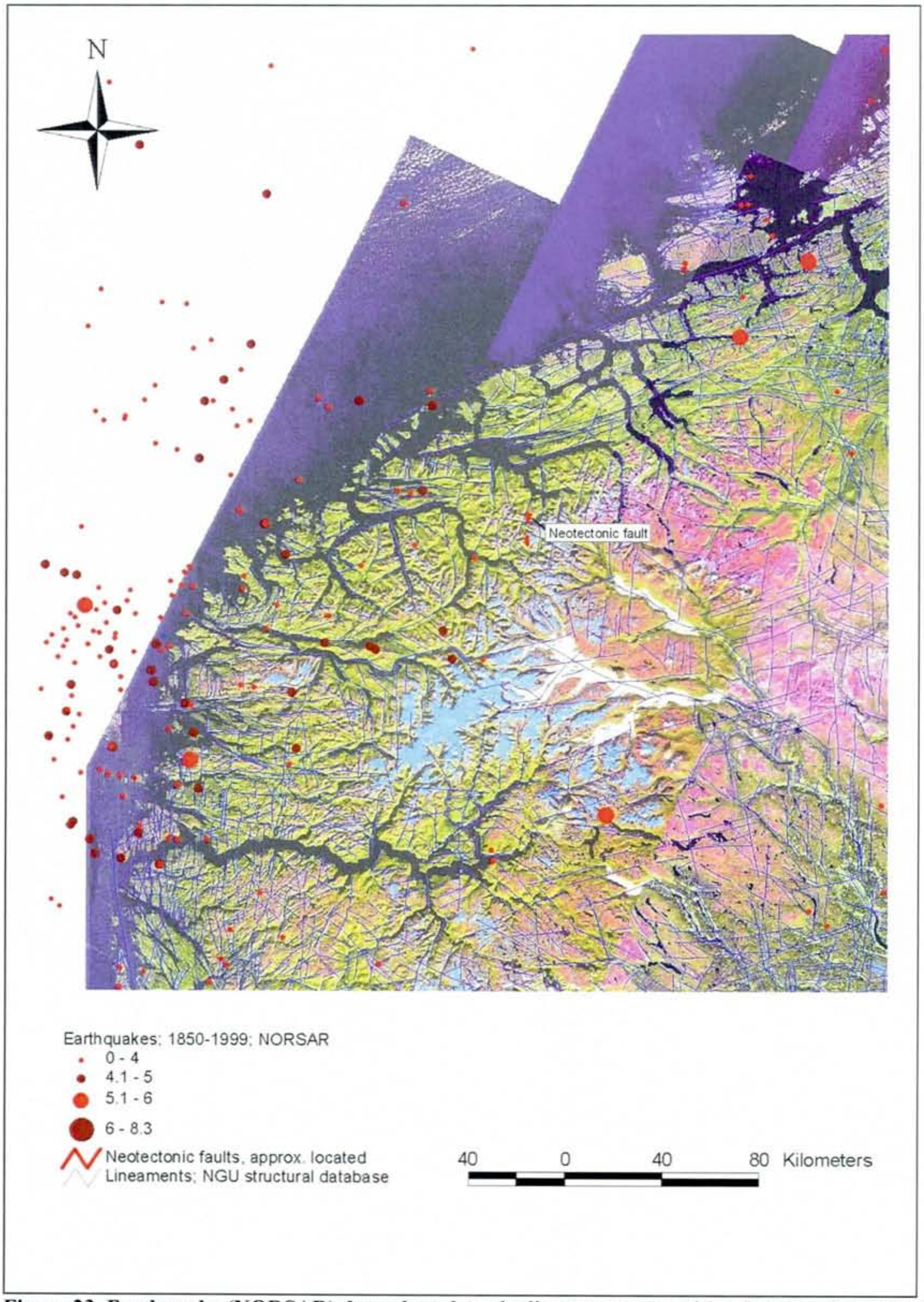


Figure 23. Earthquake (NORSAR) data plotted on the lineament map with a Landsat 7 satellite image base (NGU's structural database). The neotectonic fault is shown (Anda *et al.*, in prep), and a possibe continuation of the neotectonic fault is indicated further to the south.

Another complicating factor is the possibility of "seismic pumping" observed during historical earthquakes (Sibson *et al.*, 1975), which may influence the pore-pressure and thus the stability condition.

7. Run-out distance

Run-out predictions of large-scale rock avalanches are normally treated by use of the relationship between the volume of the sliding mass and the height of drop/length of run-out factor *(H/L)*. This relationship has been well established by several authors (McEwen, 1989; Nicoletti & Sorriso-Valvo, 1991; and in a recent review by Erismann & Abele, 2001). However, the international data have several weaknesses, mainly since there are a limited number of events with relatively small volumes. Only a few are less than 10 mill m^3, the smallest have a volume of 7 mill m^3. Furtermore, only a few events are characterized by small fall heights. Another factor which may play an important role is the type of bedrock involved in the slide.

Due to the reasons above, we have collected data from 25 geologically mapped rock-avalanche deposits in Norway, predominantly from the Western Gneiss Region (Table 1). These are plotted against the volume of the sliding mass and the H/L factor (Fig. 24). The individual events are divided into two groups based on the fall height; events found in the fjords are grouped in a separate category. The Norwegian examples plot above the regression line defined by the data of McEwen (1989). Another important factor exemplified in these data is that events with low fall heights and relatively small volumes plot very high in the diagram, also compared with other Norwegian examples in the gneiss region.

Table 1. Data from 24 geological mapped Norwegian rock avalanches. The data are collected for this report.

No.	Name	Reference	Year	Bedrock	Land/ sea	V (mill m^3)	L	H	H/L
1	Verkilsdalen, Rondane	Dawson *et al.* 1986		Schists and phyllite	L	15	1600	675	0,42
2	Tjelle, Langfjorden		1756	Gneiss	L & S	15	2000	750	0,375
3	Melkevoll, Olden	Nesje, in press		Gneiss	L	0,25-0,5	750	480	0,64
4	Rørsetura, Oterøya			Gneiss	L & S	2,5	>1100	650	<0,59
5	Gravem, Sunndalen	Blikra, 1994		Gneiss	L	0,3-0,5	1500	900	0,6
6	Sørdalen*			Gneiss	L	2,5-5	1500	675	0,45
7	Urdabøuri			Gneiss	L	16-23	1350	470	0,35
8.	Erdalen, Stryn	Nesje, 1984		Gneiss	L	8-12	1010	460	0,46
9	Hjelle, Stryn	Nesje, 1984	1980	Gneiss	L	0,5	575	730	1,27
10	Bjørkum, Lærdal	Blikra & Aa, 1996		Gneiss	L	0,15-0,3	550	400	0,73
11	Furuneset, Tafjorden			Gneiss	L & S	0,5-1	1500	900	0,6
12	Langhammaren, Tafjorden		1934	Gneiss	L & S	2-3	1500 (2000)*	850	0,57 (0,43)
13	Grande*, Geirangerfjorden			Gneiss	L & S	0,5-0,8	1250	1350	0,93
14	Hysket, Geirangerfjorden			Gneiss	L & S	1,5-3	1125	550	0,49
15	Stølaholmen, Fjørland			Gneisss	L	3-4	960	420	0,55
16	Berrføttene, Fjærlandsfjorden			Gneiss	L & S	50-100	4000	1000	0,25
17	Frykkjelen, Fjærlandsfjorden			Gneiss	L & S	2,5-3	2200	950	0,43
18	Kubergan 1, Tromsø	Blikra, 1994		Garnet-mica schists	L	8	700	375	0,54
19	Kubergan 2, Tromsø	Blikra, 1994		Garnet-mica schists	L	5	640	350	0,55
20	Nakkevatnet*, Lyngen			Garnet-mica schists	L & Lake	15-25	2350	900	0,38
21	Grøtlandsura, Salangen			Garnet-mica schists	L & S	6-12	1200	500	0,42
22	Skjersura*, Valldal			Gneiss	L	12-15	1750	1000	0,57
23	Hellaren, Grovfjorden			Garnet-mica schists	L & S?	100-150	4000	900	0,23
24	Gumpedalen*, Sørreisa			Marble	L	30-50	2200	720	0,33
25	Store Urdi, Jotunheimen*	Anda & Lømo, 1981		Gneiss	L	15	1400	400	0,29

* The H/L value is a maximum date due to a topographic impact against opposite slope

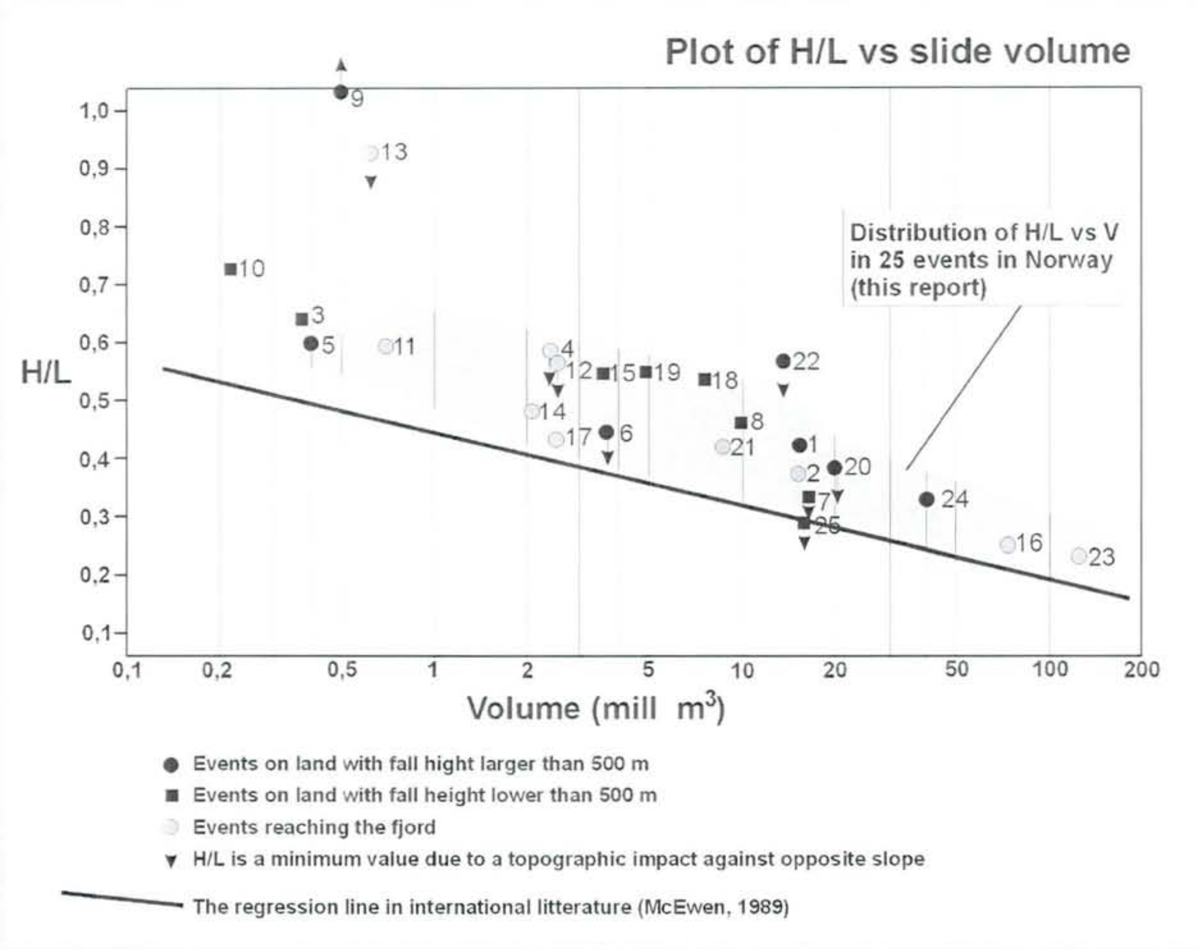


Figure 24. Plot of H/L (height of drop/length of run-out) vs. rockslide volume from the Norwegian data in Table 1.

8. Hazard evaluation

The distribution of rock-avalanche events clearly demonstrates that the frequency of events is highest in the inner fjord areas (Fig. 1). However, one concentration area is found around Oterøya and another area further to the southwest. A database containing historical avalanche events, including snow avalanches and debrisflows, is under construction at NGU in close collaboration with Astor Furseth (Nordal commune in Møre & Romsdal). These data (Figure 25) clearly show the high number of rock-avalanche events with related tsunamis in the inner fjord areas. All events with related tsunamis are found in inner fjord areas, except for the Tjelle event in Langfjorden.

From the data on geological mapped rock-avalanche events (Fig. 1) and the historical once (Fig. 25) it can be concluded that the high-risk areas are concentrated to the inner fjord areas of western Norway (Fig. 26). However, it cannot be excluded that large rock-avalanche events can occur outside this area, but the probability is significantly lower. It is clear that the concentrations of geologically mapped rock-avalanche events and gravitational fractures in the Oterøya area (Fig. 1), not are found present in the historical data (Fig. 25). This suggests that the distinct concentration of events in the Oterøya area is relatively old, and probably formed just after the deglaciation, or alternatively, is related to very rare events. The possibility that large earthquakes have played a role cannot be excluded. Actually, the geographic concentrations of rock avalanches hints at a common mechanisms, e.g. one or more earthquakes. The Oppstadhornet area is located about 45 km from the only mapped postglacial fault in southern Norway, but the distribution of rock-avalanche events suggest that possible earthquakes related to this fault did not affect the stability of rock slopes over such long distances. If large earthquakes in the offshore area, more than 100 km from Oppstadhornet, should affect the stability here, they most probably need to be in the order of M 7-7,5. Such large-size earthquakes should give prominent surface ruptures. Based on the missing geological evidences for this, it can be concluded that it is not probably that earthquakes in the order of 7-7,5 have occurred in near-shore western Norway in postglacial time.

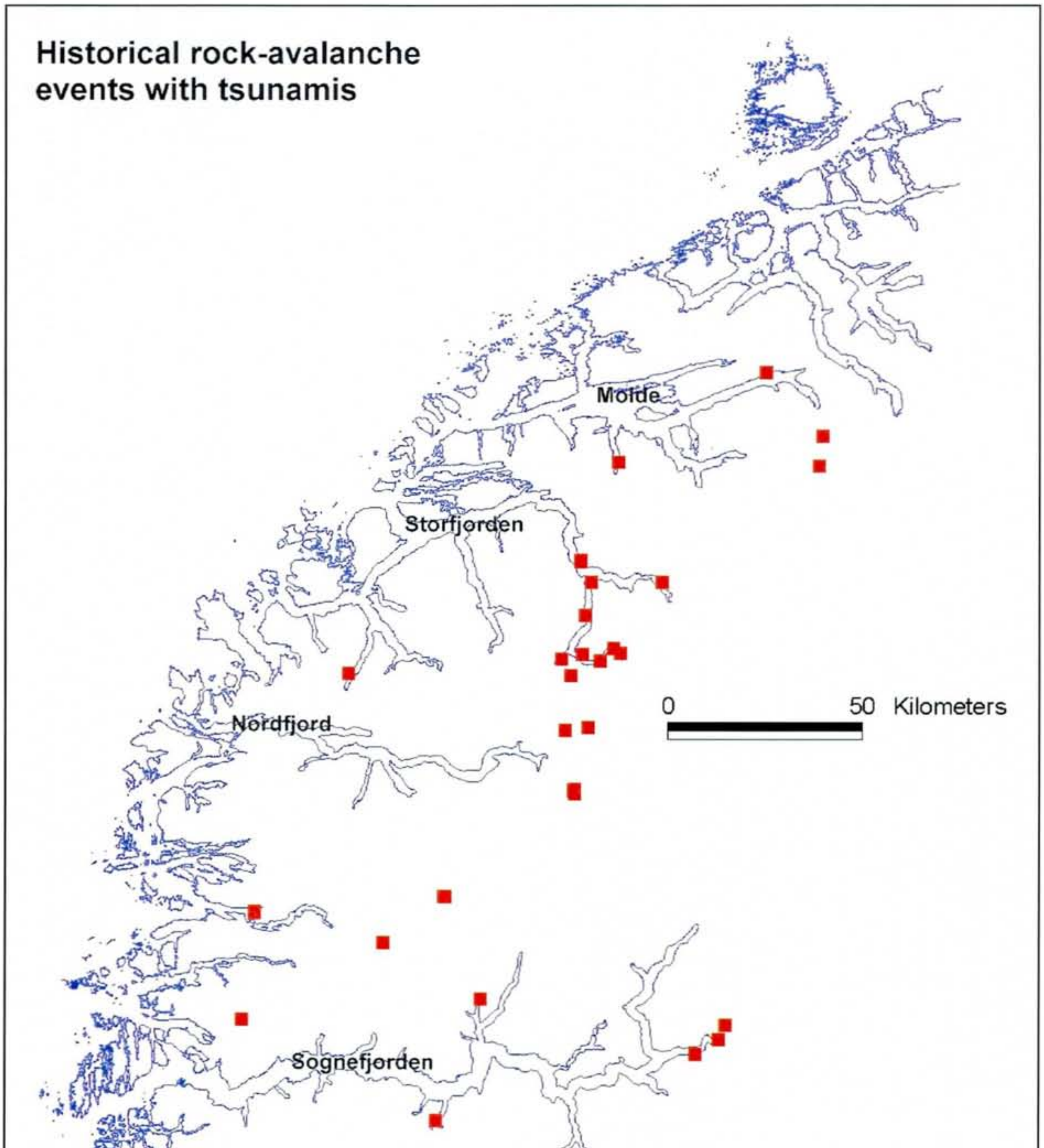


Figure 25. Historical rock-avalanche events with observed tsunamis in Møre & Romsdal and Sogn & Fjordane county extracted from the historical database (NGU/Astor Furseth).

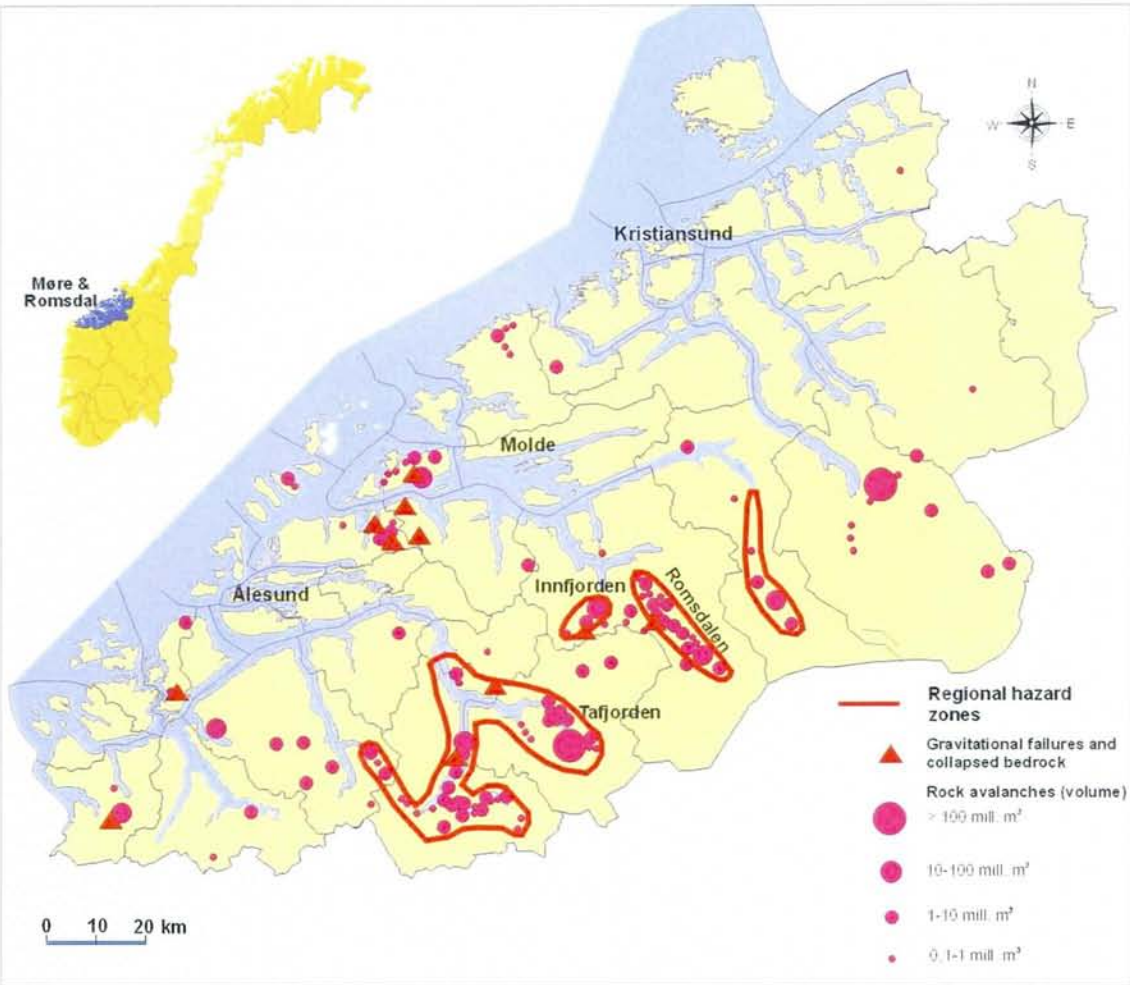


Figure 26. Regional hazard zones in Møre & Romsdal county. These zones are characterized by a high number of both historical events and geological mapped rock-avalanche deposits.

In the following the individual sites will be analysed in terms of run-out potential and probability estimates.

8.1 Skår – Baraldsneset

No major fractures with the potential to act as a slide plane were found, and the fractured zone observed at one site along the upper part of the slope would most likely collapse as single blocks.

In principle it can not be excluded that a large rock avalanche can be triggered from the steep face of Skårahornet, but the probability for large rock avalanches from this area is considered very low and at least lower than an annual probability of 10^{-4}. Debris flows triggered along Kloppelva can reach the fjord, and collapses of glacial till due to rockfall from the western slope of Skårahornet may reach the depression just south of Einargarden. A hazard zone with an estimated annual probability of 10^{-4} is shown in Fig. 6.

8.2 Oterøya – Oppstadhornet

8.2.1 Run-out estimates

Possible run-out distances have been estimated with basis in the Norwegian data shown in figure 24. Events witch are thought to be most comparable with the Oppstadhornet case are plotted in the diagram of figure 27. The estimates of possible run-out distances gave the following results (see Fig. 27):

Scenario 1 (entire potential slide area)
Volume: 20 mill m^3
H (height of drop): 830 m (650 m on land and 180 m in the fjord)
H/L: 0,34-0,43
Estimated L (length of runout): 1,9-2,4 km (0,7-1,2 km into the fjord).

Scenario 2 (topmost upper part)
Volume: 5-7 mill m^3
H (height of drop): 830 m (650 m on land and 180 m in the fjord)
H/L: 0,40-0,47
Estimated L (length of runout): 1.7-1,9 km (0,5-0,7 km into the fjord).

Scenario 3 (eastern steep part)
Volume: 2-3 mill m^3
H (height of drop): 900 m (720 m on land and 180 m in the fjord)
H/L: 0,43-0,52
Estimated L (length of runout): 1.7-2,1 km (0,5-0,9 km into the fjord).

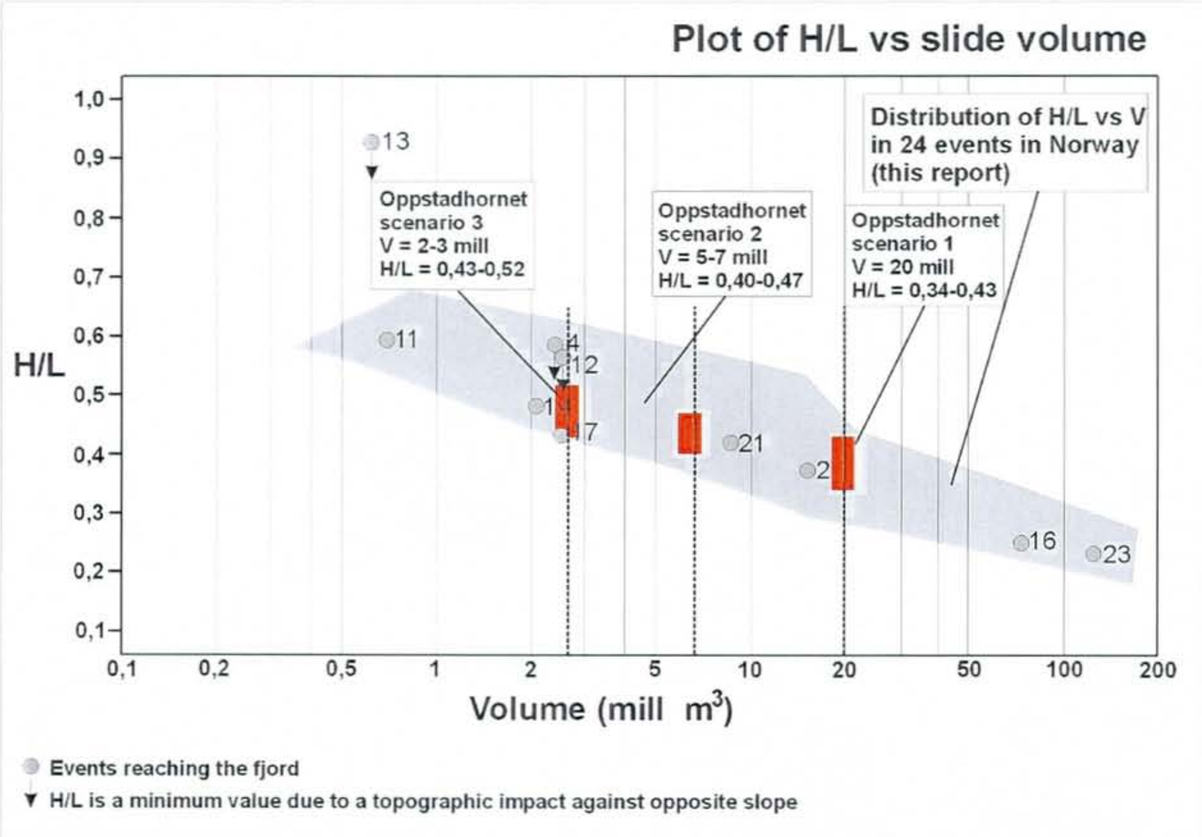


Figure. 27. Plot of H/L (height of drop/length of run-out) vs. rock-slide volume (Fig. 24) with the plot of possible slides from Oppstadhornet. The diagram also shows individual events most comparable with the situation at Oppstadhornet (rock avalanches entering the fjord).

Figure 28. Estimates of possible run-out zones for three scenarios of sliding from the Oppstadhornet area. Each square of the map is 1 X 1 km in size.

8.2.2 Probability estimates

In general, calculation of probability values is hampered by significant uncertainties, and it is especially difficult for events with low probabilities. In this case, one major drawback is the lack of measurements of present movements in the unstable area. Unfavourable conditions in terms of high water saturation and high-magnitude earthquakes can trigger parts of the mountainside at Oppstadhornet (see section 6 and Bhasin & Løset, 2002). The predictions of future earthquakes are far from accurate, but a major earthquake cannot be totally excluded over a period of 10.000 years (see chapter 6). With these uncertainties in mind we have estimated probability values for 3 different scenarios.

Scenario 1 (20 mill m^3)

The failure is younger than ca. 11.500 years and there are two observations that indicate that parts of the slope have moved recently (see chapter 4.2). Numerical modelling by NGI (Bhasin & Løset 2002) concludes that it is less likely that the whole rock mass of up to 20 million m^3 can become unstable. One key question, and uncertainty, is the risk for a large-magnitude earthquake, which can trigger this unstable mountain slope. With this in mind, the annual probability is predicted to be less than 10^{-4}.

Scenario 2 (5-7 mill m^3)
The slope gradient is gentler in the lower part of the slide block, and our interpretation indicates that there are several basal detachments or sliding zones forming a sole to the sliding mass. A future slide will most probably occur in one of the upper sliding surfaces in areas of higher gradients. Numerical simulation made by NGI (Bhasin & Løset 2002) also concludes that this part may become unstable due to either earthquake or reduction of the residual friction angle due to weathered and saturated surfaces. The annual probability is estimated to be in the order of 10^{-4}.

Scenario 3 (2-3 mill m^3)
The estimated probability for scenario 3 with a volume of 2-3 mill m^3 is thought to be higher due to the fact that this part have not yet slid down, and besides, it is situated in an area of higher slope gradients and with a higher fall height. An annual probability in the order of 10^{-4} has been estimated.

The entire unstable area should be monitored through a series of measurements of potential movements. More precise measurements of possible movements and dating of existing failures will strengthen the probability estimates. A test of ongoing movements can be conducted through Differential Radar Interferometry (dInSAR), a method presently being explored at the NGU.

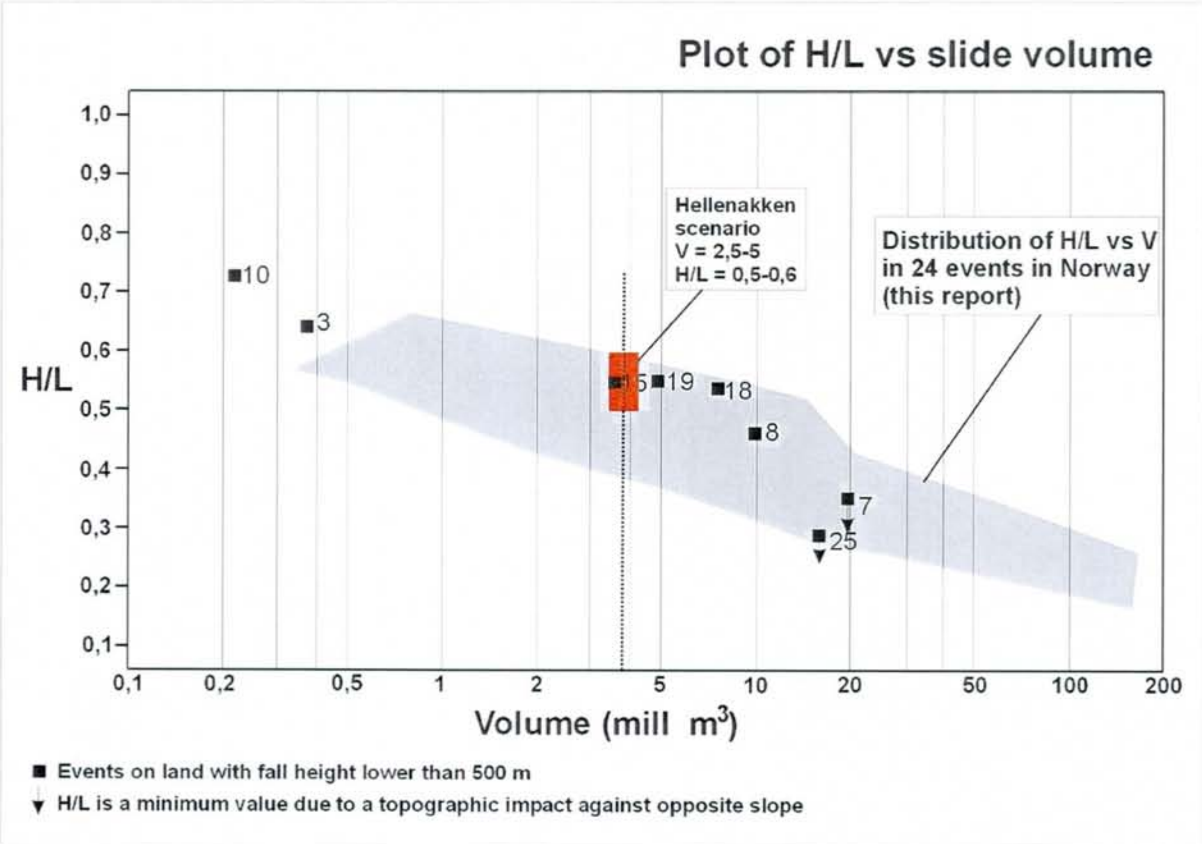


Figure. 29. Plot of H/L (height of drop/length of run-out) vs. rock-slide volume (Fig. 24) with the plot of a possible slide from Hellenakken. The plot also shows individual events most comparable with the situation at Hellenakken (small fall heights).

8.3 Hellenakken

No detailed studies have been conducted on this possibly unstable mountain ridge.

8.3.1 Run-out estimates

An estimate of possibly run-out distance has been done by using the H/L vs slide volume relationship. The Norwegian data presented in figure 24 show that events with small fall heights have a relatively low run-out distance (low H/L value). Rock-avalanche events characterized by fall heights lower than 500 m are presented in Fig. 29, and the H/L value for a volume of 2.5 to 5 mill m^3 from Hellenakken has been estimated to be 0,5-0,6. A height drop of 400 m (down to the shoreline) will give an estimated run-out distance about 700-800 m. The distance from the potential unstable area to the shore is 750 m and the estimates therefore indicate that a failure from the Hellenakken might reach the shoreline. However, most of the mass will be deposited onshore, and the slide volume reaching the fjord will be small (probably less than 0,5 mill m^3). It is thus concluded that a large tsunami caused by failure from the Hellenakken is not likely, but cannot be ruled out.

8.3.2 Probability estimates

Evaluation of the probability for an avalanche is difficult because no detailed studies have been performed, and no measurements of modern movements are available. It cannot be excluded that a large avalanche from the Hellenakken can occur with an annual probability of about 10^{-4}. However, probably most of the volume will be deposited onshore and the risk for a large tsunami is low. It is concluded that most of the volume will be deposited onshore, but a volume of less than 0,5 mill m^3 could reach the fjord. It can not be excluded that a larger portion of the slide may reach the fjord (1-2 mill m^3), but the probability for this is low (less than 10^{-4} pr. year).
The potentially unstable area should be monitored with measurements of modern movements.

9. Conclusions and suggestions for further investigations

Geological and historical data on rock-avalanche events demonstrate that the high-risk areas are concentrated to the inner fjord areas of western Norway. The studied areas at Baraldsneset, Oppstadhornet and Hellenakken are situated outside this high-risk zone. However, it can not be excluded that large rock-avalanche events can occur in this region as well, but the probability will be much lower.

There are no indications of larger-scale instability structures on Skårahornet near Baraldsneset, and the probability for large rock avalanches from this area are considered very low, and at lower than an annual probability of 10^{-4}. Debrisflows triggered along Kloppelva at the western slope of Skårahornet can reach the fjord in the western part of the considered area, and a hazard zone with an estimated annual probability of 10^{-4} is presented.

The Oppstadhornet area is a zone characterized by large rock avalanches and bedrock failures. Several areas of collapsed bedrock occur on the 700-m high southeastern slope of Oppstadhornet. Distinct large blocks show intense fracturing and overall down-slope sliding both along the foliation and on cross-fractures. Surface disturbances at two locations indicate recent movement; these failures are younger than ca. 11.500 years. The volume of the entire failure is estimated to be c. 20 mill m^3 (scenario 1), including a 2-3 mill m^3 large steep eastern part (scenario 3) that is characterized by fractures and fissures showing evidence of movements. A potential slide of a more limited upper part of the main slide feature is named

scenario 2 and may influence a mass of 5-7 mill m^3. This part constitutes the unstable top layer modelled by NGI (Bhasin & Løset, 2002).
Due to several weaknesses in the available international data on run-out of large rock avalanches, we have collected data from 25 mapped rock-avalanche deposits in Norway, predominantly from the Western Gneiss Region. These are plotted against the volume of the sliding mass and the H/L factor (height of drop/length of run-out). Estimates of potential run-out distances indicate that a large rock avalanche involving the entire area may reach 0,7 – 1,2 km into the fjord (scenario 1), whereas the uppermost top of the slide may reach 0,5-0,7 km (scenario 2) and the steep eastern part 0,5 – 0,9 km (scenario 3) into the fjord basin.

There are no measurements of present movements in the unstable area, making probability estimates difficult. Unfavourable conditions in terms of high water saturation and high-magnitude earthquakes can trigger parts of the mountainside at Oppstadhornet (see section 6 and Bhasin & Løset, 2002). It cannot be excluded that a large-size rock avalanche, with a volume up to 20 mill m^3, could occur from the Oppstadhornet area, but with an annual probability of less than 10^{-4} (scenario 1) For example, a relatively large-magnitude earthquake could trigger off this unstable mountain slope. However, a future slide will most probably occur in one of the upper sliding surfaces in areas of high gradients (see Bhasin & Løset 2002). The annual probability of a slide involving 5-7 mill m^3 is estimated to be in the order of 10^{-4} (scenario 2). The annual probability for a rock avalanche from the eastern part with a volume of 2-3 mill m^3 is also estimated to be in the order of 10^{-4}.

Along a steep mountain west of Baraldsneset, at Hellenakken, there are open fractures indicative of relatively recent movement and therefore instability. An estimated volume of 2.5-5 mill m^3 may be involved in the unstable mountain slope. No detailed studies have been conducted on this feature, but it cannot be excluded that a large failure could occur from Hellenakken with an annual probability of about 10^{-4}. It is concluded that most of the volume will be deposited onshore, but a volume of less than 0,5 mill m^3 may enter the fjord. It cannot be excluded that a larger portion of the slide may reach the fjord (1-2 mill m^3), but the probability for this is low (less than 10^{-4} pr. year).

A test on ongoing movements can be conducted through Differential Radar Interferometry (dInSAR), a method presently being explored at the NGU. SAR scenes may be used as a first approach to detect mm vertical movements over the last 10 years. In order to quantify the stability situation, monitoring stations should be installed, at least on Oppstadhornet. These may detect ongoing movements and unravel changing stability conditions. In addition to the possibility of predicting potential slide events, this would also make the probability estimates more accurate. The probability estimates are also hampered by limited data on actual triggering factors. There are only a few age determinations on observed prehistorical failures and large rock avalanches in Møre & Romsdal. A dating program focused on cosmogenic dating on sliding planes and boulders within rock-avalanche deposits would yield better data on the frequency and time distribution of such events. If they cluster around specific time periods, that would strengthen the idea that seismic triggering is important.

10. References

Anda, E. & Lømo, L. 1981: Utlavassdraget. Kvartærgeologiske og geomorfologiske undersøkelser. 10 års vernede vassdrag Rapport 4. Geologisk institutt, Universitetet i Bergen.

Anda, E., Blikra, L.H. & Braathen, A. In prep.: Berill postglacial fault in Romsdal, western Norway.

Bhasin, R. & Løset, F. 2002: Ormen Lange – Ilandføringsalternativer. Numerical Simulation of a Rock Slope at Oppstadhornet. *NGI Report* 20001472-03.

Blikra, 1994: Kvartærgeologisk kart Tromsø 1534 III. *Geological Survey of Norway.*

Blikra, L.H. & Anda, E. 1997: Large rock avalanches in Møre og Romsdal, western Norway. *NGU Bulletin* 433, 44-45.

Blikra, L.H., Anda, E. & Longva, O. 1999: Fjellskredprosjektet i Møre og Romsdal: status og planer. *NGU rapport* 99.120, 21 pp.

Blikra, L.H. & Aa, A.R. 1996: Skredfarekartlegging i Lærdal i samband med den nye stamveien Oslo-Bergen. *NGU Report* 96.055

Erismann, T.H. & Abele, G. 2001: *Dynamics of Rockslides and Rockfalls.* Springer, 316 pp.

Dawson, A,G., Matthews, J.A. & Shakesby, R.A. 1986: A catastrophic landslide (sturzstrom) in Verkilsdalen, Rondane national park, southern Norway. *Geografiska Annaler,* 68A, 77-87.

Dehls, J., Olesen, O., Bungum, H., Hicks, E.C., Lindholm, C.D. & Riis, F. 2000A: Neotectonic map: Norway and adjacent areas. *Geological Survey of Norway.*

Harbitz, K. 2002: Ormen Lange – Ilandføringsalternativer. Rock slide generated tsunamis – Run-up heights at Baraldsneset and Nyhamn. *NGI Report* 2001472-2.

Jibson, R.W. 1996: Using landsides for palaeoseismic analysis. In: J.P McCalpin (ed.) Palaeosismology. International geophysics series 62. Academic Press, San Diego, 397-438.

McEwen, A.S. 1989: Mobility of large rock avalanches. Evidence from Valles Marineris, Mars. *Geology 17,* 1111-1114.

Nesje, A. 2002: A rockfall avalanche in Oldedalen, inner Nordfjord, western Norway, dated by means of a sub-avalanche *Salix* sp. tree trunk. *Norsk Geologisk Tidsskrift,* in press.

Nesje, A. 1984: Kvartærgeologiske undersøkingar i Erdalen, Stryn, Sogn og Fjordane. Master thesis, University of Bergen.

Nicoletti, P.G. & Sorriso-Valvo, M. 1991: Geomorphic controls of the shape and mobility of rock avalanches. *Geological Society of America Bulletin* 103, 1365-1373.

NORSAR & NGI 1998: Development of a seismic zonation for Norway. Final report for Norwegian Council for Building Standardization (NBR), march 15, 162pp.

Pascal, C. & Gabrielsen, R. 2001: Numerical modelling of Cenozoic stress pattern in the mid-Norwegian margin and the northern North Sea. *Tectonics* 20, 585-599.

Robinson, P. 1995: Extension of Trollheimen tectono-stratigraphic sequence in deep synclines near Molde and Brattvåg, Western Gneiss Region, southern Norway. *Norsk Geologisk Tidsskrift 75,* 181-198.

Robinson, P., Tveten, E. & Blikra, L.H. 1997: A post-glacial bedrock failure at Oppstadhornet, Oterøya, Møre og Romsdal: a potential major rock avalanche. *NGU Bulletin* 433, 46-47.

Sibson, R.H., Moore, J.McM & Rankin, A.H. 1975: Seismic pumping – a hydrothermal fluid transport mechanism. *Journal Geological Society of London* 131, 653-659

Sitar, N. & Khazai, B. 2001: *Characteristics of seismically induced landslides in recent eartquakes.* In:M. Kühne, H.H. Einstein, E.Krauter, H. Klapperich, R. Pöttler (eds.). Int. Conf. On Landslides, Causes, Impacts and Countermeasures,. VGE, Davos, Switzerland.

Tveten, E., Lutro, O. & Thorsnes, T. 1998. Map sheet Ålesund M 1:250000. *Norges geologiske undersøkelse.*