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Examples of multiple rock-slope collapses from Köfels (Ötz valley, Austria) and western Norway

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Abstract

This paper describes the effects of large massive rock-slope failures on subsequent slope stability. Three examples of large rock-slope failures from the Austrian Alps and Norway demonstrate that failure increases the probability of further collapses. At Köfels, Austria, a Holocene rock-slope failure several km³ in size filled the Ötz valley. The morphology of the deposits indicates that at least one subsequent failure occurred along the head scarp of the first failure, most likely a slide of similar size. Debris of the second landslide slid over the older deposits, forming the famous Köfels frictionite. At least three rock-slope failures, all of them in excess of 10⁶ m³, occurred from the same mountainside within the last 6000–8000 years at Tafjord in western Norway. The most recent of these failures in 1934, triggered a destructive tsunami. Five large failures with volumes >50,000 m³ occurred at Ramnefjell, Norway within 50 years; two of them caused considerable damage and a large number of death due to the formation of destructive tsunamis. Two-dimensional finite element models of rock-slope stability before and after the Köfels and Tafjord landslides show that massive rock-slope failures produce: (a) irregular slopes, parts of which are as steep or steeper than the slope before failure and which represent new zones of instability; and (b) zones of weakness related to slow slope deformation and related cracking.

Keywords: Rock avalanche; Landslides; Recurrence; Numerical stability modelling

1. Introduction

Rock avalanches are catastrophic high velocity, long-runout landslides defined by Heim (1932) as sturzstroms with a volume in excess of 10⁶ m³. They move at velocities of up to several 100 km/h (Plafker and Ericksen, 1978) and travel further than can be accounted for by the internal angle of friction of loose rock debris (32°; Hsü, 1975). Rock avalanches impacting a lake or the sea may trigger destructive tsunamis, with displacement waves up to several tens to hundreds of meters high (Hendron and Patton, 1985; Grimstad and Nesdal, 1990). Although rock avalanches are relatively rare in space and time (Abele, 1974) compared to other types of landslides (e.g. rock falls, debris flows), they are extremely destructive and thus have a high hazard potential (Plafker and Ericksen, 1978; Evans, 2002).

Hence, it is important to characterize and understand mountain fronts where rock avalanches can occur and to understand their recurrence times.

Some rock avalanches occur in clusters (Abele, 1974; Eisbacher, 1979; Blikra *et al.*, 2006) along slopes, that are prone to failure due to their tectonic and structural history or lithology (Hermanns and Strecker, 1999; Strecker and Marrett, 1999; Hermanns *et al.*, 2006). Repeated failure of the same slope has been documented from different places in the world (Plafker and Ericksen, 1978; Grimstad and Nesdal, 1990; Weichert *et al.*, 1994; Hermanns and Strecker, 1999; McSaveney, 2002; Blikra *et al.*, 2006), with recurrence times ranging from a few hours (Weichert *et al.*, 1994), to a few years (Plafker and Ericksen, 1978; Grimstad and Nesdal, 1990), to several thousands to tens of thousands of years (Hermanns *et al.*, 2001, 2004).

Here, we will present examples of recurrent rock avalanches at Köfels in the Ötz valley of the Austrian Alps and Tafjord and Loen in western Norway (Fig. 1) to make the point that repeated rock-slope failures along the same valley wall are common

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phenomena. Five rock avalanches occurred at Loen within 50 years of the last century. There has been one event at Tafjord in the 20th century, but prehistoric events can be inferred from studies of marine deposits. The Kfels landslide deposit is prehistoric and has been interpreted to be the result of a single mountain-front collapse (Heuberger, 1966; Preuss, 1971; Masch et al., 1985). Reinterpretation of the deposit and the break-away zone, however, indicates multiple events and considerable slope deformation since the last failure. The new interpretation is in better accordance with previously published seismic and sedimentological data and with numerical modelling of rock-avalanche movements. We test our hypothesis that large catastrophic landslides can reduce, rather than increase, slope stability through numerical modelling of rock-slope failures at Kfels and Tafjord.

2. Settings

2.1. Kfels landslides

Pichler (1863) first described the morphology of the tz valley between Umhausen and Lngenfeld. He pointed out that the Umhausener basin is separated from the Lngenfelder basin by an area of weak gneiss, which is cut by tz River along the 400-m-deep Maurach gorge. The gorge separates a terrace at Kfels from Tauferberg and a plateau at Niederthai on the opposite valley slope (Fig. 2). He further described the Kfels terrace being amphitheatre-shaped. This description is in accordance with the morphologic interpretation of the terrace as a landslide deposit (Erismann et al., 1977).

Kfels became famous when pumiceous glass, found within a boulder deposit, was attributed to volcanic activity (Pichler, 1863) (Fig. 2). This hypothesis persisted into the 20th century, when the deposit was attributed to maar volcanism (Penck, 1925). Due to the absence of any other evidence for volcanism in the tz valley, Lichtenegger (1929, 1930) interpreted the boulder deposit as postglacial ‘‘Bergsturz’’ (mountain collapse) and the glass as older rock that formed in the Tertiary. Almost a decade later, Suess (1937) and Stutzer (1937) suggested that a meteorite impact caused the collapse of the Fundus ridge and produced the boulder deposit and glass. This interpretation was supported later by geochemical data, indicating that the glass is of similar composition to some meteorites (Kurat and Richter, 1968, 1972) and by quartz in pumice and gneissic rocks from Kfels indicating shock metamorphism (Surenian, 1988). Leroux and Doukhan

Figure 1. Location of study sites in Austria and Norway.

(1993) and Lyons et al. (1993) argue, however, that these data indicate heating rather than shock. The meteorite impact hypothesis persists in the scientific literature (Tollmann and Tollmann, 1994).

Preuss (1971, 1974) first proposed a landslide origin for the Kfels deposit and frictional melting as the mechanism of glass formation. Erismann et al. (1977) estimated the kinetic energy of the displaced rock mass and demonstrated that the heat necessary to melt the rock could have been produced by the landslide itself. They named the pumice ‘‘frictionite’’. The frictionite does not occur at the base of the landslide deposit but in dikes several decimetre to meter thick at the top of the landslide deposit. Glass has also been found within the landslide deposit in erosional cuts in the Maurach gorge, formed by tz river (Masch et al., 1985) (Figs. 2 and 3). This glass has formed by partial to complete melting of the landslide debris, as indicated by electron microscopy studies, and has been termed hyalomylonite (Masch et al., 1985). Hyalomylonite differs from the frictionite in its thickness (1 mm to 3 cm

Figure 3. Topographic map of the failed mountain front at Köfels and related deposits, also showing the location of different glass findings (after Masch et al., 1985). Dashed line is eastern limit of secondary deposits in the southern sector; dotted line delimits a tentative interpretation of deposits of secondary failure in the northern sector. Cross-section A–A' (Fig. 9) was used for numerical slope stability analyses. Arrow indicates location of Nösslachkapelle (see text).

thick) and absence of porosity. Similar hyalomylonites are also associated with the Langtang landslide in the Himalayas (Masch et al., 1985). The exceptional occurrence of rock melts within a landslide was attributed by Abele (1974) to the exceptional volume of the Köfels deposit ($>2 \text{ km}^3$), being the largest landslide in crystalline rock in the Alps.

Erismann et al. (1977) introduced a two-stage model of landsliding, that explains the generation of glass on a secondary sliding surface (Figs. 2 and 3). They suggest that the secondary sliding surface formed when the bulk of the landslide was stopped by its impact on the eastern slope of Ötz valley and an upper portion split off and travelled uphill an additional 2 km.

Recent studies include geophysical investigations to precisely determine the landslide volume and prelandslide topography (Brückl et al., 2001), and computer modelling of the run-out (Sørensen and Bauer, 2003). Both studies provided insights, that cannot be explained by the model of Erismann et al. (1977). Their model cannot explain the formation of

Figure 2. Sequential series of one-failure hypothesis of unstable rock mass on the slope (A) emplacement (B) and subsequent incision (C) of the Köfels rockslide (after Erismann et al., 1977). Arrows indicate occurrences of frictionite along secondary sliding surface and hyalomylonite in the Maurach gorge.

a secondary sliding plane within the rock mass, along which material from behind overruns the front of the slide. The frontal part of the landslide will continue to move ahead due to its lower friction (Sørensen and Bauer, 2003). Brückl et al. (2001) found three seismic reflectors in the Maurach gorge, which they interpreted to be (1) intact bedrock, (2) prelandslide sediments of the Maurach gorge, and (3) a secondary sliding plane. The uppermost reflector is 950–1000 m a.s.l. located more than 100 m below the hyalomyonite outcrops.

In addition, the Köfels rock-avalanche deposits have been used to calibrate production rates of the cosmogenic isotopes ^{10}Be and ^{26}Al (Kubik et al., 1998), in part because they have been independently radiocarbon dated. Pieces of wood recovered from a reconnaissance gallery in the Tauferberg gave a conventional ^{14}C age of 8710 ± 150 years BP (Heuberger, 1966), and an AMS ^{14}C age of 8705 ± 55 years BP (Ivy-Ochs et al., 1998). In contrast, ^{10}Be ages from the landslide deposit on both sides of the Maurach gorge and from the primary sliding plane at Fundus ridge differ by several thousand years. Ivy-Ochs et al. (1998) attribute this difference to post-landslide erosion.

2.1.1. New morphologic observations

The landslide deposit west of Maurach gorge is

nearly flat with a few depressions and ridges, that are tens of meters deep or high and tend west–southwest parallel to the direction of movement of the landslide (Fig. 4). Farther west is an 80-m step in the terrace tending south–southeast parallel to the Fundus ridge and southwest directly southwest of Köfels. The step has a lobate form and sharply terminates the west–southwest tending depression and ridges.

Tauferberg, on the east side of Maurach gorge, has a hummocky surface with no preferential orientation of features. At this site, the largest blocks occur within the landslide body (Heuberger, 1966; Sørensen and Bauer, 2003). Reworked river gravels and till overlie the landslide deposit at Tauferberg (Ampferer, 1939; Sørensen and Bauer, 2003), suggesting that these sediments were eroded from the Ötz valley and its riverbed by the landslide.

The crown of the landslide is amphitheatre-shaped and steepens from south to north. The steepest slopes ($>50^\circ$) are at Scharltle (Figs. 2 and 4). Sliding surfaces with friction polish are preserved in this area (Suess, 1937; Ivy-Ochs et al., 1998). This part of Fundus ridge coincides with the change in the orientation of the break away and includes two north-striking depressions adjacent to a 100-m-high wall (Fig. 5). The distal depression is straight, more than 10-m-deep, ca. 250 m long and intersects the

Figure 4. Head scarp of the Köfels mountain collapses and lobate secondary deposits (triangles) cutting off depressions tending WSW–ENE (arrows), detail of air photo D 9701 of the flight Waldstands-aufnahmen 1954, Bundesamt für Eich- u. Vermessungswesen, Wien, Austria.

Figure 5. Detail of Fig. 4, showing N–S striking depressions (triangles) indicative of postfailure deformation of Fundus ridge.

breakaway wall (Fig. 5). The curved proximal depression is about 500 m long, a few meters deep, and no more than 130 m distant from the breakaway wall. It links the south–southeast and south–southwest trending parts of the breakaway wall (Fig. 5). The two depressions are evidence of ongoing deformation of Fundus ridge.

2.2. Tafjord rock avalanches

Tafjord is one of the innermost inlets of Storfjorden in western Norway (Figs. 1 and 6). On April 7, 1934, $1.5 \times 10^6 \text{ m}^3$ of rock fell from the wall of the fjord (Fig. 6), mobilised the talus cone below, and entered the sea (Kaldhol and Kolderup, 1936; Bugge, 1937). The tsunami produced by the landslide reached a maximum height of 62 m above sea

Figure 6. The Tafjord and Loen localities in western Norway. (A) The Tafjord site with the slide scar and profile used in numerical modelling (see Fig. 12). The frame shows the details of the rock-avalanche deposits in Fig. 7; (B) The Loen site with the slide-source area and the rock-avalanche deposits.

level. Geological mapping reveals many slide scars on the slopes above the fjord, and about 10 large rock-avalanche deposits have been detected at the bottom of the fjord through bathymetrical and seismic investigations (Blikra et al., 2006).

2.2.1. Rock-avalanche deposits

Swath bathymetry reveals the morphology of the

Tafjord rock-avalanche deposits. Features include boulder cones, concentric ridges bordering the rock-avalanche debris, and secondary mass flows extending out the impact sites (Blikra et al., 2006). A boulder cone and large secondary mass flows occur on the seafloor below the 1934 slide scar (Fig. 7).

Reflection seismic data demonstrate that the rock-avalanche deposits are the product of several

Figure 7. (A) Shaded-relief image of the seafloor at the site of the 1934 Tafjord slide. Dashed line is seismic profile presented in panel B. Position I marks outer limit of the deposits of the 1934 slide. Positions II and III mark two older rock avalanches; (B) seismic reflection profile showing the relationship between rock-avalanche deposits and fine-grained marine sediments.

events, with debris lobes of different ages lying on top of one another. At least three individual rock-avalanche deposits have been identified below the 1934 slide scar, and the 1934 deposit is the smallest (Fig. 7). *Blikra et al. (2006)*, interpreted these deposits to be younger than 8000 years.

2.2.2. Source area of the rock avalanches

Bedrock at Taffjord is quartzo-feldspathic mica gneiss of the Risberget Nappe (*Tveten et al., 1998*). The foliation dips moderately to steeply to the south. The 1934 slide plane follows the foliation and is a steep slope (60°) below a sub-vertical rockface extending from sea level to ca. 800 m a.s.l. (Figs. 6 and 8A). Unconsolidated breccias have been found in biotite-rich layers, weak layers that probably play a fundamental role for the stability conditions (*Braathen et al., 2004*). Before the failure in 1934, a large block, 150 m high and 200 m wide (1.5 million m^3), was separated from stable rock by fractures (Fig. 8A). Fracture width increased from a few decimeters to more than 1 m between 1900 and 1933 (*Bugge, 1937; Kaldhol and Kolderup, 1936; Bjerrum and Jørstad, 1968*) and the widening of the crack accelerated further until the failure in 1934.

2.3. Loen rock avalanches

Two disastrous rock avalanches occurred at Loen

in western Norway in 1905 and 1936 (Figs. 1 and 6, Table 1). Large volumes of rock fell from the mountain Ramnefjell and impacted Loenvatnet lake, triggering tsunamis that killed 61 persons in 1905 and 73 persons in 1936 (Fig. 8B).

The source rocks are gneiss with zones of amphibolite. The subvertical mountain slope is penetrated by a set of fractures extending into the mountain side, and zones of amphibolite seem to operate as weak planes below these vertical joints (*Bjerrum and Jørstad, 1968*).

On January 15, 1905, a block from a height of about 500 m, approximately 50,000 m^3 in volume, impacted the talus below. It mobilised the slope and increased its volume to 350,000 m^3 . This debris struck the lake, causing a tsunami with a maximum run-up of 41 m. Several rock falls occurred in the following months, and on September 20, part of Ramnefjell collapsed generating a powerful tsunami. In August 1936, a number of smaller rock falls were observed at Ramnefjell, and on September the 13, a block about 1 million m^3 in volume released from the upper part of Ramnefjell 800 m above the lake (event 3). The block fragmented and plunged into the lake causing a tsunami that achieved a maximum run-up of 74 m. This event was followed by a series of small rock falls. A crack developed behind the 1936 rupture surface, but a monitoring program did not detect any changes in the slope from 1936 to

Figure 8. Slide scars of the Taffjord and Loen rock avalanches. (A) Langhammaren at Taffjord. The slide scar extends to about 800 m above the sea level; (B) Ramnefjell at Loen. The upper part of the scarp is about 900 m above the lake level.

Table 1

Historical rock-slope failures and related tsunamis from Loen

Date	Volume of collapsed rock (m ³)	Volume of entrainment (m ³)	Tsunami height (m)
15/01/1905	50,000	300,000	40.5
20/09/1905	15,000	50,000	>15
13/09/1936	1,000,000		74.2
21/09/1936	100,000		ca. 40
06/10/1936	?		?
11/11/1936	>1,000,000		>74
22/06/1950	1,000,000		ca. 15

Data from Grimstad and Nesdal (1990).

1946. In 1947, a new part of the rock at the top of the slope failed, and a new fracture was discovered parallel to the fracture that led to the failure in 1936. Several hundred thousand m³ of rock failed along this fracture on June 22, 1950. The lake was almost full of rock-avalanche deposits from the previous events, thus only a small tsunami occurred. This last failure opened a deep-seated fracture, which is a potential sliding surface for a new series of failures (Bjerrum and Jørstad, 1968).

3. Geomechanical modelling

Geomechanical modelling has become an important tool for better understanding of slope stability and rock-avalanche run-out (e.g. Crosta et al., 2003; Bhasin and Kaynia, 2004; Eberhardt et al., 2004). Here we model stress changes and deformation prior to and after failure to determine if the removal of large rock masses can contribute to further destabilisation of the slope. We use two-dimensional finite element models to estimate stresses within the slopes that failed at Köfels and Tafjorden (Figs. 9–12).

3.1. Model input data

The finite element analysis for Köfels is based on the partial differential equation toolbox of MATLAB (The Mathworks Inc.). Model input data come from Brückl (2001) and Brückl et al. (2001), and have been estimated for landslide deposits using the relationship between porosity and bulk modulus of Han et al. (1986) (Table 2). Reconstructed topography of the slope, with 100 m contours, prior to failure is taken from Brückl et al. (2001). Present day topography comes from 1:25,000 scale maps with

Figure 9. Finite element mesh of the failed mountain sector in Ötz valley. The line A–A* according to Fig. 2 points out the profile used for the model. The black area is the root of the mountain; the dark grey area is bedrock that did not fail, the grey area at the top of the mountain is the failed rock mass; light grey zone is a former bedrock area, now occupied by debris; and the brightest grey area is rock debris from mountain collapse. The black box is highlighted in Fig. 10.

20-m contours. The profile location (see Fig. 2) was selected to include in the analysis the section of Fundus ridge that most obviously deformed after failure (Fig. 5). The density of triangular elements was increased in regions with assumed large stress gradients, as suggested by Brückl (2001).

The initial stress state for Fundus ridge was estimated from the original topography and gravitational effects. This equilibrium stress state was transferred to the topography (to preserve the original shape of the slope). Poisson's ratio and only influence of gravity yield the expected horizontal-to-vertical stress ratio of 0.33 in the lower, undisturbed region of the model.

The ratio between acting maximum shear stresses $[(\sigma_{\max} - \sigma_{\min})/2]$ and resisting shear stresses in each triangular element was calculated to define critical zones in terms of the Mohr–Coulomb criterion. The resisting shear stress is the highest shear stress a material can withstand under an applied load without failure $[c + \tan \phi^* (\phi_{\max} + \phi_{\min})/2]$ based on Mohr–Coulombs theory. If the ratio exceeds 1, the model suggests material failure. If the ratio is <1 , the material is stable. The higher the ratio, the more likely the material is to fail (Fig. 10).

In situ rock stress measurements were made during the construction of the 5360 m long tunnel (the Heggura Road Tunnel) in 1980–1982 through Precambrian gneiss under the slide area in Tafjord (Broch and Sørheim, 1984). These data, along with laboratory testing results and the Hoek–Braun failure criterion (Hoek et al., 2002) provided input data for analysis of the Heggura rock slope (Panthi and Nilsen, 2005), Table 2. The analysis performed uses the Phase² 2-D finite element model (Rocscience, 1998–2001) shown in Fig. 11.

3.2. Results

3.2.1. Distribution of critical zones at Köfels

The spatial distribution of critical zones at Köfels is similar to the distribution of maximum shear stress after deglaciation, as modelled by Brückl (2001). Our results reveal that, prior to landsliding, the steep rock slope bordering the valley floor is a natural stress concentrator. Likewise most critical zones in the postlandslide landscape occur at the transition between the rock slope and the landslide deposit. Both the slope angle and the ratio between acting and resisting shear stresses are lower after the

Table 2
Parameters used for model

	Köfels	Tafjord
Model	2-D finite elements (plain strain)	2-D finite elements (plain strain)
Meshing	8000 3-noded-triangles	1878 3-noded-triangles
Constitutive relation	Isotropic linear elastic	Isotropic linear elastic
Material properties		
Young's modulus (E)	50 GPa (intact rock) ^a 38 GPa (landslide debris) ^b	5 GPa (rock mass) 2 GPa (landslide debris)
Poisson's ratio (m)	0.25 (intact rock) ^c 0.27 (landslide debris) ^d	0.21 (intact rock) 0.23 (landslide debris)
Density (q)	2600 kg/m ³ (intact rock) ^a 2300 kg/m ³ (landslide debris) ^b	2750 kg/m ³ (intact rock) 2150 kg/m ³ (landslide debris)
Critical zones calculations	Mohr–Coulomb criterion	Hoek–Brown criterion
Angle of internal friction (ø)	22.50 ^a	
Cohesion (c)	4.1 MPa ^a	
Uniaxial compression strength (rc)		67 MPa (intact rock) 7.5 MPa (landslide debris)
Hoek–Brown parameter (m)		2.53 (intact rock) 0.30 (landslide debris)
Hoek–Brown parameter (s)		0.036 (intact rock) 0.000045 (landslide debris)
Roller boundaries	See Figs. 9 and 11	
Left vertical	3000 m a.s.l. (top of highest ridge)	270 m b.s.l. (fjord bottom)
Right vertical	1000 m a.s.l. (valley bottom)	1000 m a.s.l. (top of highest ridge)
Horizontal	4000 m b.s.l.	700 m b.s.l.
Horizontal/vertical stress ratio K	0.33 (only gravitational)	0.49 (from in situ measurements)

Data from Tafjord after Panthi and Nilsen (2005).

^a After (Brückl, 2001).

^b Derived from (Brückl, 2001).

^c After (Brückl et al., 2001).

^d Derived after (Brückl et al., 2001).

landslide suggesting that the slope is more stable.

3.2.2. Displacements in Tafford

The unloading caused by the 1934 major slide has caused displacements of considerable dimensions, indicating maximum total and horizontal displacements due to unloading of the rock mass of about 21 cm and 6 cm, respectively (Fig. 12). The modelled total displacements vary from 5 cm at the base to 21 cm in the middle part of the slope. It decreases again towards the top of the slope, at the edge of the plateau, where it is 3–4 cm. The deformation is thus especially prominent in the middle to lower part of the former slide scar.

4. Discussion

The giant slope collapse at Köfels has been the subject of controversy for 150 years, and several contrasting hypotheses for its cause have been proposed (volcanism, meteorite impact and glacial oversteepening of the slope). *Erismann et al. (1977)* explain both the failure and glass formation most simply, arguing that the glass was formed on secondary sliding surfaces when the upper part of the landslide mass slid across the lower part. The presence of river gravel on top of Tauferberg (*Ampferer, 1939; Sørensen and Bauer, 2003*), however, argues against this hypothesis. If the upper block detached from the lower part of the landslide when the lower

Figure 10. Finite-element model of critical zones shown as ratio between acting maximum shear stresses [$(\sigma_{\max} - \sigma_{\min}) / 2$] and resisting shear stresses in terms of the Mohr–Coulomb criterion in the Fundus ridge, Ötz valley, before failure (A) and at present (B).

Figure 11. Heggura slope topography and FEM-mesh representing situation (A) before and (B) after the 1934 slide. Profile location shown in Fig. 6.

part struck the east slope of the Ötz valley, the gravels could not have been eroded from the riverbed and transported to the top of Tauferberg. Also, modelling of the run-out of the Köfels landslide cannot explain this landslide behaviour (Sørensen and Bauer, 2003). Run-out models of other rock avalanches, likewise, cannot explain a secondary sliding plane (Crosta et al., 2003).

Ötz River has cut a deep gorge into the landslide body, but the original surface is preserved on top of Tauferberg and at the southern margin of the landslide deposit. In the latter area, a lobe of landslide debris cuts west–northwest striking depressions and ridges. A similar lobe exists directly southwest of Köfels. We conclude that these deposits are not the result of a single event, but rather two or more fail-

Figure 12. (A) horizontal and (B) total displacements (in m) along the Heggura slope after the 1934 slide.

ures of Fundus ridge. Due to erosion of the landslide deposits, it is not possible to reconstruct the history of the events in detail. A tentative interpretation, however, is given in Fig. 2. The first event involved failure of a rock mass of the order of 109 m^3 along the southern part of Fundus ridge. The second event, or sequence of events, involved volumes of 107 m^3 in the southern sector and $108\text{--}109 \text{ m}^3$ in the northern sector. This model explains the preserved morphologic features on top of the deposit as well as the river gravels on top of the Tauferberg. According to the model the gravels were scraped from valley during the first event. It also accounts for the glass as melted rock at the base of the basal sliding plane of the second landslide. The young surface exposure age (Ivy-Ochs et al., 1998) of the polished break away surface below the Schartle is also in accord with the model. This surface had already been interpreted by Suess (1937) as a sliding plane formed by a landslide subsequent to the initial collapse of Fundus ridge. Finally, a multiple landslide model also explains the stratigraphic sequence of a borehole at Nösslachkapelle, which indicates two generations of impoundment of the Längenfelder basin (Fig. 2, Ampferer, 1939). In Tafjord and Loen in western Norway, there is no doubt about the multiple nature of the rock-slope failures.

There are numerous examples of multiple failures of the same rock slope from around the world. Plafker and Erickson (1978), for example, found a lobe of rockavalanche deposits below the 1970 Huascarán avalanche deposit (Cordillera Blanca, Peru), and interpreted it to be the deposit of a prehistoric failure. Mathews and McTaggart (1978) found prehistoric rock-avalanche deposit outside the deposit of the 1965 Hope rockslide in southwestern British Columbia, Canada. Hermanns et al. (2001, 2004) investigated multiple rock avalanches in arid northwestern Argentina. Lobes geomorphology and cosmogenic nuclide ages were used to define different generation of landslides separated by several thousands to tens of thousands of years. Two rock avalanches at Mount Cook in the Southern Alps of New Zealand were separated by 118 years and two rock avalanches at nearby Mt. Flechter were separated by only 88 days (McSaveney, 2002). The Randa rockslides in 1991 in the Swiss Alps, involving $30 \times 106 \text{ m}^3$ of bedrock, occurred only 3 weeks apart (Eberhardt et al., 2002). Four large multi-million cubic meter rockslides occurred between 1862 and 1979 at Mount Tsao-Ling in Taiwan. Three of them happened on the same slope (Hung et al., 2002).

From the rock mechanical point of view unstable mountain slopes remain critical as long as the slope

gradient is not reduced by landsliding. Rock-slope failures often produce slopes with segments that are as steep or steeper than the original slope. At Köfels the entire postlandslide slope is less steep than the slope before collapse, but an unfavourable ratio of driving to resisting stresses remains at the base of the steep section. The postfailure stress state in the steepest part of the slope is nearly the same as the stress state in the prefailure slope.

Only a small part of the slope at Tafjord has failed. The postfailure slope has, as at Köfels, steeper and gentler inclined sectors. However, at Tafjord the steeper sector is more inclined than the prefailure slope. Here, areas of high shear or tensile loads are more pronounced than at Köfels, resulting in displacements of decimeters. This response likely has great significance for the stability of the post-failure slope because of the development of new, opening tensional joints. The tensional joints may increase the probability of new failures. The importance of the opening of tensional joints is highlighted by the series of rock-slope failures at Ramnefjell in Loen during a 50-year period.

Finite element models of the Köfels and Tafjord slopes indicate that they remain critical or near-critical, and the spreading within Fundus ridge continued after the initial collapse. We postulate that slope failure may occur at both sites in the future.

A possible explanation for repeated slope failures lies in the redistribution of stress over time. Diffusive, continuous mass redistribution occurs as a result of erosion, resulting in stress relaxation. In such cases, the deviatoric components of the stress tensor decrease over time and hence the overall stress state becomes less critical. In contrast, after a sudden, catastrophic redistribution of mass during a landslide, a more critical stress state is produced than in the first instance, even if the final slope has the same geometry, because there has been no time for stress relaxation. This means that slopes that once collapsed are more likely to fail again, compared to similar slopes that have not experienced such a sudden large mass redistribution.

5. Conclusion

We have presented examples of landslides from the Austrian Alps (Köfels) and western Norway (Loen and Tafjord) with repeated, non-random rock-slope failure along the same mountain slopes. Numerous rock avalanches occurred at Loen between 1905 and 1950 and both historical and prehistoric rock avalanches can be inferred at Tafjord from marine deposits. The Köfels deposits had been previ-

ously interpreted as the deposit of a single mountain front collapse, but new geomorphologic observations indicate more than one event. The new multiple event interpretation better explains published geophysical and sedimentological data, numerical modelling of rock-avalanche movements, as well as surface-exposure ages.

Numerical modelling has been used to analyse stresses and displacements resulting from rock-slope failures at Kfels and Tafjord. Two-dimensional finite element models of rock-slope stability prior to and following the Kfels and Tafjord failures indicate that the sudden release of large rock masses causes: (a) irregular slopes where parts are as steep or steeper than the slopes before failure, resulting in new zones of instability; and (b) zones of weakness that cause tensional cracks to open, further reducing slope stability.

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References

- Abele, G., 1974. Bergstrze in den Alpen. Wissenschaftliche Alpenvereinshefte, vol. 25. Ausschsse des Deutschen und sterreichischen Alpenvereins, Mnchen, pp. 1–231.
- Ampferer, O., 1939. ber die geologischen Deutungen und Bausondierungen des Maurachriegels im tztal. *Geologie und Bauwesen*, 25–43.
- Bhasin, R., Kaynia, A.M., 2004. Static and dynamic simulation of a 700-m high rock slope in western Norway. *Engineering Geology* 71, 213–226.
- Bjerrum, L., Jrstad, F., 1968. Stability of rock slopes in Norway. Publication-Norwegian Geotechnical Institute 79, 1–11.
- Blikra, L.H., Longva, O., Braathen, A., Anda, E., Dehls, J., Stalsberg, K., 2006. Rock-slope failures in Norwegian fjord areas: examples, spatial distribution and temporal pattern. In: Evans, S.G., Scarawcia Mugnozza, G., Strom, A.L., Hermanns, R.L. (Eds.), *Massive Rock Slope Failure: New Models for Hazard Assessment*, NATO Science Series, IV: Earth and Environmental Sciences, vol. 49. Springer, Berlin, pp. 489–511.
- Braathen, A., Blikra, L.H., Berg, S.S., Karlsen, F., 2004. Rock-slope failures of Norway; type, geometry, deformation mechanisms and stability. *Norwegian Journal of Geology* 84, 67–88.
- Broch, E., Srheim, S., 1984. Experience from the planning, construction and supporting of a road tunnel subjected to heavy rockbursting. *Rock Mechanics and Rock Engineering* 17, 15–35.
- Brckl, E., 2001. Cause-effect models of large landslides. *Natural Hazards* 23, 291–314.
- Brckl, E., Brckl, J., Heuberger, H., 2001. Present structure and prefailure topography of the giant rockslide of Kfels. *Zeitschrift fr Gletscherkunde und Glazialgeologie* 37, 49–79.
- Bugge, A., 1937. Fjellskred fra topografisk og geologisk synspunkt. *Norsk Geografisk Tidsskrift* 6, 342–360.
- Crosta, G.B., Imposimato, S., Roddeman, D.G., 2003. Numerical modelling of large landslides stability and runout. *Natural Hazards and Earth System Sciences* 3, 523–528.
- Eberhardt, E., Willenberg, H., Stead, D., Coggan, J.S., 2002. An integrated numerical analysis approach applied to the Randa rockslide. In: Rybr, J., Stemberk, J., Wagner, P. (Eds.), *Landslides*. Balkema, Lisse, pp. 355–362.
- Eberhardt, E., Stead, D., Coggan, J.S., 2004. Numerical analysis of initiation and progressive failure in natural rock slopes - the 1991 Randa rockslides. *Rock Mechanics and Mining Sciences* 41, 69–87.
- Eisbacher, G.H., 1979. Cliff collapse and rock avalanches (sturzstroms). *Canadian Geotechnical Journal* 16, 309–334.
- Erismann, T., Heuberger, H., Preuss, E., 1977. Der Bimsstein von Kfels (Tirol), ein Bergsturz-(Frikzionit). *Tschermak's Mineralogische und Petrographische Mitteilungen* 24, 67–119.
- Evans, S.G., 2002. Single-event landslides resulting from massive rock slope failure: characterising their frequency and impact on society. In: Evans, S.G., Martino, S. (Eds.), *NATO Advanced Research Workshop "Massive rock Slope Failure: New Models for Hazard Assessment"*. Celano, Italy, pp. 32–37.
- Grimstad, E., Nesdal, S., 1990. The Loen rockslides - a historical review. In: Barton, M., Stephansson, W. (Eds.), *Rock Joints*. Balkema, Rotterdam, pp. 1–6.
- Han, D.H., Nur, A., Morgan, D., 1986. Effects of porosity and clay content on wave velocities in sandstones. *Geophysics* 51, 2093–2107.
- Heim, A., 1932. Bergsturz und Menschenleben. *Beiblatt zur Vierteljahrsschrift der Naturforschenden Gesellschaft in Zrich* 77, 1–217.
- Hendron Jr., A.J., Patton, F.D., 1985. The Vaiont Slide, A Geotechnical Analysis Based on New Geologic Observations of the Failure Surface. U.S. Army Corps of Engineers, Waterways Experiment Station, Vicksburg, MS.
- Hermanns, R.L., Strecker, M.R., 1999. Structural and lithological controls on large Quaternary rock ava-

- lanches (sturzstroms) in arid northwestern Argentina. *Geological Society of America Bulletin* 111, 934–948.
- Hermanns, R.L., Niedermann, S., Villanueva Garcia, A., Sosa Gomez, J., Strecker, M.R., 2001. Neotectonics and catastrophic failure of mountain fronts in the southern intra-Andean Puna Plateau, Argentina. *Geology* 29, 619–623.
- Hermanns, R.L., Niedermann, S., Ivy-Ochs, S., Kubik, P.W., 2004. Rock avalanching into a landslide-dammed lake causing multiple dam failure in Las Conchas valley (NW Argentina)—evidence from surface exposure dating and stratigraphic analyses. *Landslides* 1, 113–122.
- Hermanns, R.L., Niedermann, S., Villanueva Garcia, A., Schellenberger, A., 2006. Rock avalanching in the NW Argentine Andes as a result of complex interactions of lithologic, structural and topographic boundary conditions, climate change and active tectonics. In: Evans, S.G., Scarascia Mugnozza, G., Strom, A.L., Hermanns, R.L. (Eds.), *Massive Rock Slope Failure: New Models for Hazard Assessment*, NATO Science Series, IV: Earth and Environmental Sciences, vol. 49. Springer, Berlin, pp. 513–537.
- Heuberger, H., 1966. Gletschergeschichtliche Untersuchungen in den Zentralalpen zwischen Sellrain-und Ötztal. *Wissenschaftliche Alpenvereinshefte*, 20.
- Hoek, E., Torres, C.C., Corkum, B., 2002. Hoek–Brown failure criterion. *Proceedings, North American Rock Mechanics Society Meeting*, Toronto: ON., pp. 267–273.
- Hsü, K.J., 1975. Catastrophic debris streams (sturzstroms) generated by rockfalls. *Geological Society of America Bulletin* 86, 129–140.
- Hung, J.-J., Lee, C.-T., Lin, M.-L., 2002. Tsao-Ling rockslides, Taiwan. In: Evans, S.G., DeGraff, J.V. (Eds.), *Catastrophic Landslides: Effects, Occurrence and Mechanism*, Reviews in Engineering Geology, vol. 15. Geological Society of America, pp. 91–115.
- Ivy-Ochs, S., Heuberger, H., Kubik, P.W., Kerschner, H., Bonani, G., Frank, M., Schlüchter, C., 1998. The age of the Köfels event. Relative, ^{14}C and cosmogenic isotope dating of an early Holocene landslide in the central Alps (Tyrol, Austria). *Zeitschrift für Gletscherkunde und Glazialgeologie* 34, 57–70.
- Kaldhol, H., Kolderup, N.-H., 1936. Skredet i Tafjord 7. April 1934. *Bergens Museums Årbok* 1936 Naturvitenskapelig Rekke 11, 1–15.
- Kubik, P.W., Ivy-Ochs, S., Masarik, J., Frank, M., Schlüchter, C., 1998. ^{10}Be and ^{26}Al production rates deduced from an instantaneous event within the dendro-calibration curve, the landslide of Köfels, Ötztal Valley, Austria. *Earth and Planetary Science Letters* 161, 231–241.
- Kurat, G., Richter, W., 1968. Ein Alkalifeldspat-glas im Impaktit von Köfels/Tirol. *Naturwissenschaften* 55, 490.
- Kurat, G., Richter, W., 1972. Impaktite von Köfels. *Tschermak's Mineralogische und Petrographische Mitteilungen* 17, 23–45.
- Leroux, H., Doukhan, J.-C., 1993. Dynamic deformation of quartz in the landslide of Köfels, Austria. *European Journal of Mineralogy* 5, 893–902.
- Lichtenecker, N., 1929. Bergsturz und Bimssteingang von Köfels im Ötztal (Tirol). *Geographischer Jahresbericht aus Österreich* 14 u. 15, 228–254.
- Lichtenecker, N., 1930. Zur Deutung des Bimssteinvorkommens von Köfels im Ötztal. *Verhandlungen der Geologischen Bundesanstalt (Wien)*, 173–179.
- Lyons, J.B., Officer, C.B., Borelia, P.E., Lahdonsky, R., 1993. Planar lamellar substructures in quartz. *Earth and Planetary Science Letters* 119, 431–440.
- Masch, L., Wenk, H.R., Preuss, E., 1985. Electron microscopy study of hyalomylonites-evidence for frictional melting in landslides. *Tectonophysics* 115, 131–160.
- Mathews, W.H., McTaggart, K.C., 1978. Hope rockslides, British Columbia, Canada. In: Voight, B. (Ed.), *Rockslides and Avalanches*. Elsevier, Amsterdam, pp. 259–275.
- McSaveney, M.J., 2002. Recent rockfalls and rock avalanches in Mount Cook National Park, New Zealand. In: Evans, S.G., DeGraff, J.V. (Eds.), *Catastrophic Landslides: Effects, Occurrence and Mechanism*, Reviews in Engineering Geology, vol. 15. Geological Society of America, pp. 35–70.
- Panthi, K.K., Nilsen, B., 2005. Numerical analysis of stresses and displacements for the Tafjord Slide, Norway. *Bulletin of Engineering Geology and the Environment*. doi:10.1007/s10064-005-0009-y.
- Penck, A., 1925. Der postglaziale Vulkan von Köfels im Ötztale. *Sitzungsberichte der Preussischen Akademie der Wissenschaften* 11, 218–225.
- Pichler, A., 1863. Zur geognosie Tirols. II. Die vulkanischen Reste von Köfels. *Jahrbuch der Kaiserlich Königlich Reichsanstalt*, Wien, 591–593.
- Plafker, G., Ericksen, G.E., 1978. Nevados Huascaran avalanches, Peru. In: Voight, B. (Ed.), *Rockslides and Avalanches*. Elsevier, Amsterdam, pp. 277–314.
- Preuss, E., 1971. Über den Bimsstein von Köfels/Tirol. *Fortschritte der Mineralogie* 49, 70.
- Preuss, E., 1974. Der Bimsstein von Köfels im Ötztal/Tirol, die Reibungsschmelze eines Bergsturzes. *Jahrbuch des Vereins zum Schutze der Alpenpflanzen und-Tiere* 39, 85–95.
- Rockscience Inc., 1998–2001. PHASE2: 2D Finite Element Pro-gram for Calculating Stresses and Estimating Support Around Underground Excavation. Rockscience Inc., Toronto, ON, Canada.
- Sørensen, S.-A., Bauer, B., 2003. On the dynamics of the Köfels sturzstrom. *Geomorphology* 54, 11–19.
- Strecker, M.R., Marrett, R.A., 1999. Kinematic evolution of fault ramps and role in development of landslides and lakes in northwestern Argentine Andes. *Geology* 27, 307–310.
- Stutzer, O., 1937. Die Talweitung von Köfels im Ötztal (Tirol) als Meteorkrater. *Zeitschrift der Deutschen Geologischen Gesellschaft* 88, 523–525.
- Suess, F.E., 1937. Der Meteorkrater von Köfels bei Umhausen im Ötztale, Tirol. *Neues Jahrbuch für Mineralogie, Geologie und Paläontologie, Abhandlungen* 72, 98–155.
- Surenian, R., 1988. Scanning electron microprobe study

- of shock features in pumice and gneiss from Koefels (Tyrol, Austria). *Geologisch Paläontologische Mitteilungen* 15, 135–143.
- The Mathworks Inc., Partial Differential Equation Toolbox, Natick, MA, USA, www.mathworks.com.
- Tollmann, E.K., Tollmann, A., 1994. The youngest big impact on Earth deduced from geological and historical evidence. *Terra Nova* 6, 209–217.
- Tveten, E., Lutro, O., Thorsnes, T., 1998. Bedrock map, Ålesund, scale 1:250,000. Geological Survey of Norway, Trondheim.
- Weichert, D., Horner, R.B., Evans, S.G., 1994. Seismic signatures of landslides; the 1990 Brenda Mine collapse and the 1965 Hope rockslides. *Bulletin of the Seismological Society of America* 84, 1523–1532.