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Formation and Preservation of Large Scale Toppling Related to Alpine Tectonic Structures - Eastern Alps

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KEYWORDS

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Abstract

Two textbook examples of rock-slope failure (RSF) by deep-seated toppling show the importance of pre-existing Alpine tectonic structures, e.g. strike-slip faults, for toppling formation, and document specific geomorphological and rock mechanical conditions at the toe and at the flanks as a prerequisite for the preservation of toppling structures.

The study area is dominated by micaschist and paragneiss and situated in Eastern Tyrol northwest of Lienz, close to a major NW-SE trending strike-slip fault. The upper slope is characterised by a saw-tooth morphology due to a series of up to 1.5 km long anti-slope scarps with an upslope trend towards lower dip (80° to 45°) and with up to 300 m wide graben structures in the topmost part. The pattern of these linear structures shows a striking similarity with that of the vertically oriented brittle structures (faults) linked to the major fault system. These features together with the hydrological regime (no surface flow, location of the springs) indicate a deep-seated gravitational slope deformation (DGSD). A bulging toe, typical for the RSF of sackung (sagging) as defined by Zischinsky (1966, 1969), is absent.

Kinematic analyses according to Goodman and Bray (1976) demonstrate toppling as the only plausible RSF mechanism due to interlayer-slip along this set of discontinuities (faults and joints). Flexural toppling is inferred to be the dominant mode of RSF, whereas the block-flexural mode is apparent in the topmost area where discontinuities have been rotated beyond the limit of flexural toppling. The observed rock disintegration with open joints can be described as the result of an initial collapse along the existing discontinuities.

Glacial over-steepening during the Last Glacial Maximum (LGM) followed by the loss of support in the early Lateglacial phase of ice-decay most probably caused the onset of the RSF. At present day, the mass-movements towards the main valley seem to have ceased due to a self-stabilizing effect of the flexural-toppling mode at the lower part of the DGSD. In addition (secondary) rock creep along the flanks, especially in the area of the heavily disintegrated graben structures towards the tributary valleys, may also have had a major impact on slope stability due to the reduction of load in the topmost part. This unloading at the flanks in combination with the orientation of the cross-joint system (re-activated foliation gently dipping into the slope), which prevent a downward slip, and the absence of fluvial erosion at the toe, caused the preservation of these rare toppling structures.

A transition from toppling to sackung (Poisel, 1998), the major RSF in crystalline rocks in Alpine valleys, did not occur in these cases. However, as a strictly structure-controlled process, toppling is regarded as a significant initial phase of RSF development. Thus, its impact on valley shaping and landscape evolution especially in the longitudinal valleys of the Alps appears to be underestimated.

Zwei Beispiele von Böschungsversagen zeigen die Bedeutung von prä-existenten, alpinen tektonischen Strukturen, insbesondere von Seitenverschiebung, für die Bildung tiefgreifender Kippung („toppling") auf. Anhand dieser Beispiele werden die spezifischen geomorphologischen und felsmechanischen Bedingungen am Böschungsfuß sowie an den oberen Hangbereichen dargestellt, die zur Erhaltung der Kippungsstrukturen geführt haben.

Das aus Glimmerschiefer und Paragneis aufgebaute Untersuchungsgebiet liegt in Osttirol, nordwestlich von Lienz, und befindet sich an einer bedeutenden NW-SE streichenden Seitenverschiebung. Der obere Hangabschnitt ist durch eine „Sägezahn"-Morphologie charakterisiert, mit einer Serie von hangparallelen und bis zu 1,5 km langen, in den Hang einfallenden Bewegungsflächen („antislope scarps"). Diese weisen einen hangaufwärts gerichteten Trend zu geringeren Einfallswinkeln (von 80° auf 45°) auf. Im obersten Bereich liegt eine bis zu 300 m breite Grabenstruktur vor. Das räumliche Verteilungsmuster dieser linearen Strukturen zeigt eine große Übereinstimmung mit jenem der vertikalen Sprödstrukturen (Störungen), welche parallel zur Hauptstörung verlaufen. Diese Böschungsmerkmale belegen zusammen mit den hydrologischen Charakteristika (kein Oberflächenabfluss, Verteilung der Quellen) eine tiefgreifende gravitative Hangdeformation. Ein vorgewölbter Böschungsfuß, typisch für Sackungen sensu Zischinsky (1966, 1969), ist nicht vorhanden.

Kinematische Analysen nach Goodman und Bray (1976) belegen „toppling" infolge der Bewegung entlang von bevorzugten Trennflächen (Störungen und Klüften) als den einzig plausiblen Mechanismus des Böschungsversagens. „Flexural toppling" (Biegekippen) ist als dominanter Mechanismus abzuleiten. Im obersten Bereich lassen deutlich über das Limit von „flexural toppling" rotierte Trennflächen „block-flexural toppling", eine Zwischenform von Biege- und Blockkippen, erkennen. Die erfasste Felsauflockerung mit offenen Klüften wird als Resultat eines initialen Kollapses an der Gesamtheit der vorhandenen Trennflächen erklärt.

Glaziale Übersteilung während des Würm-Hochglazials, gefolgt vom Verlust des stützenden Widerlagers während der Eiszerfallsphase, verursachte wahrscheinlich das Einsetzen der Hangbewegung im frühen Spätglazial. Die Massenbewegung Richtung Haupttal dürfte durch den Selbststabilisierungseffekt des im unteren Hangabschnitt erfolgten Biegekippens gegenwärtig inaktiv sein. (Sekundäres) Felskriechen an den Flanken in Richtung der Seitentäler, besonders im Bereich der aufgelockerten Grabenstrukturen, hat ebenso einen wichtigen Einfluss auf die Böschungsstabilität durch Entlastung des oberen Bereiches. Weiters lässt die flach in den Hang einfallende Schieferung, die als Querklüftung reaktiviert wurde, kein Gleiten hangabwärts zu. Diese Faktoren sind zusammen mit der fehlenden fluviatilen Erosion am Böschungsfuß maßgeblich für die Erhaltung dieser ausgedehnten Kippungs-Strukturen.

Ein Übergang von Kippung zur Sackung (Poisel, 1998), dem häufigsten Böschungsversagen in Kristallingesteinen, erfolgte in diesen Fällen nicht. „Toppling", als ein vorwiegend strukturell kontrollierter Prozess, ist als bedeutende Initialphase in der Entwicklung von Böschungsversagen zu betrachten. Die Auswirkung dieses Typs von Massenbewegung auf die Talformung und die Landschaftsentwicklung, im speziellen der alpinen Längstäler, erscheint demzufolge als unterschätzt.

1. Introduction

Our perception of the formation and shaping of Eastern Alpine landscapes by tectonic, fluvial, glacial and gravitational processes has made great progress in the last two decades (e.g. Frisch et al., 1998; van Husen, 2000). For example, the large longitudinal valleys following major fault systems, generated during the Miocene lateral extrusion of the Eastern Alps (Ratschbacher et al., 1991), and a rearranged drainage pattern, developed by fluvial incision from the Miocene onwards (Frisch et al., 1998), exemplify this complex relationship. Finally, valley shaping by Pleistocene glaciers resulted in the modern topography with steep slopes and overdeepened valleys.

In such a setting, one of the most prominent morphological

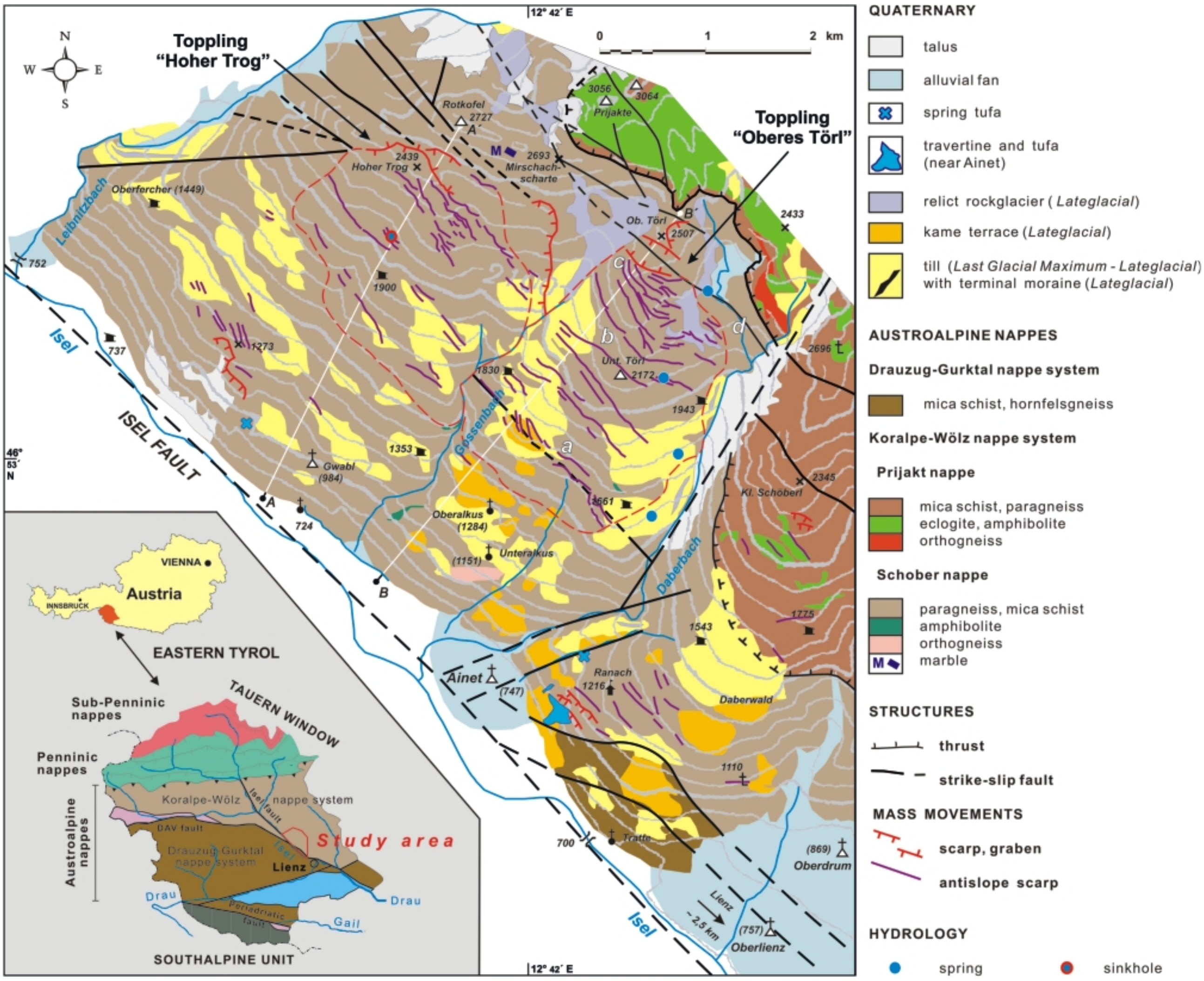

Figure 1: Location of the study area, general tectonic setting and geological sketch map of the southwestern part of the Schobergruppe with the slopes of Oberes Törl and Hoher Trog (compiled and simplified after Linner 1994, 1995, 2005, and Reitner, 2003). The location of the geological cross sections (A-A´and B-B') are indicated by white lines. The positions of letters a, b, c and d denote regions of the Oberes Törl slope, of which structural data are displayed separately in Fig. 4.

features are antislope scarps (synonymous with antithetic, obsequent, uphill-facing scarps or counterscarps, cf. Bovis, 1982) running parallel to the valley axis and to major faults. Their formation has been attributed either to neotectonic movements (Gutiérrez-Santolalla et al., 2005), glacio-isostatic rebound due to unloading following ice-decay after the Last Glacial Maximum (LGM), or to mass movements (Agliardi et al., 2009; Hippolyte et al., 2006). Ustaszewski et al. (2008) even argued that these scarps should be called composite faults as they regarded them to resemble all three processes. Moreover, different views on the specific gravitational mechanisms responsible for the formation of rock-slope failures (RSF) and deep-seated gravitational slope deformations (DGSD; Dramis and Sorriso-Valvo, 1994) exist, ranging from common toppling, i.e. flexural to block toppling (Freitas and Watters, 1973; Goodman and Bray, 1976; Bovis, 1982; Nichol et al., 2002), slope deformations assigned as sackung (e.g. Gutiérrez-Santolalla et al., 2005; Hippolyte et al., 2006; Agliardi et al., 2009) and various combination of these and other gravitational processes (see discussions in Bovis, 1982 and Poisel, 1998).

However, as most important lifelines of modern society run along such valleys the knowledge on the formation of such a phenomenon is of interest not only from a theoretical perspective, but may have a profound impact on engineering as well as on risk assessment, as toppling might develop under specific circumstances into catastrophic rock avalanches (Nichol et al., 2002).

Based on detailed geological mapping focusing on Quaternary morphology and sediments as well as on structures of the basement rocks we present two cases of large scale toppling from the Isel valley northwest of Lienz, located on the southwest slope of the Schobergruppe (Eastern Tyrol, Austria; Fig. 1). Morphology, structure and their relation to lithology and large scale tectonic structure will be used to characterise the kinematics of mountain slope deformations and to discuss their formation and contribution to Alpine landscape development. The occurrence of this rare but significant type of RSF will be discussed together with other examples in the Eastern Alps.

2. Geological and Geomorphological Setting

The Austroalpine nappes between the Tauern Window and the Periadriatic fault system (Schmid et al., 2004) represent the tectonic setting for the study area. These nappes are composed of the western part of the Drauzug-Gurktal nappe system in the Deferegger Alps and Lienz Dolomites and the Koralpe-Wölz nappe system in the Schobergruppe (Fig.1). There, the high-pressure metamorphic Prijakt nappe was thrust over the Schober nappe (Behrmann, 1990). The Prijakt nappe is characterised by eclogite, amphibolite, mica schist, paragneiss and orthogneiss, while the Schober nappe of this area exhibit somewhat monotonous paragneiss and micaschist (Pestal et al., 2009). The thrust between the two nappes is delineated by the occurrence of eclogites and marked by an obvious mylonitic texture, which extends several hundreds of meters in the hanging- and footwall. Therefore the foliation dips quite uniform to the E and SE at low to medium angles (Behrmann, 1990; Bücksteeg, 1999).

The brittle deformation south of the Tauern Window is controlled by the Periadriatic fault system, which mainly accommodated the Miocene lateral extrusion of the Eastern Alps relative to the Southalpine units (Mancktelow et al., 2001). Kinematically linked with dextral strike-slip movement along the Periadriatic fault prominent NW-SE-trending faults are found in the Isel valley NW of Lienz and in the Debant valley within the Schobergruppe (Pestal et al., 2005). Consequently steep dipping faults are common in the slope to the Isel valley, with morphological significant faults across Mirschachscharte and around Oberes Törl as well as a set of parallel faults SW of Rotkofel (Fig.1).

The location of the Isel valley along a fault has undergone fluvial incision, which is regarded to have started concurrent with the Early Miocene activity of faulting (Frisch et al., 1998). This general pattern of landscape evolution was significantly shaped by multiple Alpine glaciations. Nevertheless, any glacial sediment older than Upper Pleistocene is missing. A tentative trimline reconstruction in the Isel valley upstream of Lienz indicates an ice surface of the Isel glaciers in 2200 to 2300 m a.s.l. (Reitner, 2003) for the time of the climax of Würmian Pleniglacial (= Last Glacial Maximum – LGM) around ~ 24-21 kyr BP (Preusser, 2004). A patchy coverage by basal till (Fig. 1) and roche moutonnée are evidence of glacial action, however, the typical U-shape valley form is also absent in valley cross-sections, which have not been affected by mass movements. Data on the extent of overdeepening are only available from the basin of Lienz, which is located in the confluence of the former Isel and Drau glacier, with maximum bedrock depth in 520 m below present valley floor based on a gravity survey (Walach, 1993). It can be assumed, however, that the significant narrowing of the Isel valley from Lienz upstream to its closest point northwest of Ainet, with an only 300 m broad valley floor, is accompanied by a substantial decrease of glacial overdeepening.

The rapid collapse of the network of valley glaciers in the surrounding of Lienz basin during the early Lateglacial phase of ice-decay (~ 21-19 kyr BP) is well documented by kame terraces (ice-marginal sediments; Fig. 1). A complete deglaciation of the Isel valley in this segment can be inferred around 19 kyr BP in accordance with the available geochronological data of comparable valleys in the Eastern Alps (van Husen, 2000). A local glacier tongue occupied parts of the lower reaches of the tributary valley of Leibnitzbach and Daberbach during different stadials of the Oldest Dryas (Buchenauer, 1990), as indicated by terminal moraines (Fig.1). In the last phase of the Oldest Dryas until the Bölling Interstadial (14.7 kyr BP), when reforestation started as well as during the Younger Dryas (12.9 -11.7 kyr BP; references see Boch et al., 2005) the valley slopes above about 2100 - 2200 m a.s.l. were affected by discontinuous permafrost, as indicated by rock glacier deposits (= relict rock glaciers; Buchenauer, 1990; Fig. 1). The lower limit of present day permafrost on south exposed slopes occurs at 2700 m a.s.l. (Buchenauer, 1990), which is above the study area.

3. Field Evidence

3.1 Oberes Törl

The site is located on the southwestern flank of Oberes Törl between the creeks Daberbach in the SE and Gossenbach in the NW on a slope extending in elevation from 715 m a.s.l. at the valley floor of the river Isel to 2560 m a.s.l. (Figs. 1 and 2). The slope is predominantly formed by paragneiss and mica-schist. According to morphological features and to the rock mass properties the slope can be divided into two areas with different features.

The lower slope below the village Oberalkus (1284 m a.s.l.) with an average slope angle of 35° is characterised by glacially abraded bedrock with occasional cover by till or glaciofluvial sediments. With the exception of a restricted rockfall area at the steepest part of the toe N of Ainet, there are no indications for mass movements.

The upper slope shows a completely different morphology with saw-tooth-surface due to the presence of several slope parallel, NW-SE trending depressions up to 1.5 km long in length above Oberalkus and exceptionally between the locations Unteres Törl and Oberes Törl (Figs. 2 and 3a). These depressions are characterised by ridges and steep upslope-facing surfaces dipping at an average of 35° (15°-50°) compared to the 25° average slope gradient (Fig. 3b). These surfaces are referred to as antislope scarps (=uphill-facing scarps; Bovis 1982, Bovis and Evans, 1996), where the difference between the depression and the crest of the ridge, ranges between 1 and 40 m. Most of them are developed within bedrock (Figs. 3b and 3c). Some smaller features close to the Gossenbach creek (see Fig. 1, in 1800 m a.s.l. N of Oberalkus) show a quite accentuated, “fresh” morphology despite a thin cover by glacial to glaciofluvial sediments. The most apparent examples occur above 2000 m a.s.l. (Fig. 3b), which can be traced almost across nearly the entire slope, indicating a cut across the N-S trending edge of the slope with an altitudinal difference of at least 100 m between the highest and the lowest part of its trace. In addition the intersection line of those antislope scarps with the form of the slope indicates steep to vertical scarps SW of Unteres Törl, whereas the apparent dip of those features into the slope decreases in the uphill direction successively to about 45° (Figs. 2, 3d and Fig. 4). This trend is accompanied by a closer spacing of the antislope scarps as well.

The topmost part of the slope is characterised by an almost 300 m broad graben, limited by a normal scarp at the summit of Oberes Törl and an antislope scarp towards the southwest (Figs. 1 and 2). The typical graben structure is evident especially on the northwestern flank, where the graben structure is accompanied by a series of NW-SE trending antislope scarps (Fig. 3d).

Among the brittle structures, a single dominant vertical fault striking parallel to the Isel fault can be unambiguously traced across the location Mirschachscharte to the NW-SE trending graben structure of Oberes Törl (Fig. 1). Considering connected faults, of which the main traces are shown on the map, it is likely that numerous subordinate faults developed close to the fault Mirschachscharte - Oberes Törl. In addition, along the rocky flank of the Leibnitzbach creek furrows indicate a set of parallel NW-SE-trending faults, which presumably continue into the slope between Oberes and Unteres Törl, where a dense set of antislope scarps with the same strike and a dip of > 45° occurs (Figs. 1 and 3d). Moreover, roadcuts show antislope scarps exposing slickensides with subhorizontal lineations.

The most apparent group of discontinuity planes within the outcrops consist of vertical to steeply (90°-80°) dipping joints with a similar NW-SE trend as the Isel fault and parallel structures (s. above, Fig. 4). Due to joint planes of > 1 m^2 and spacings of 1-3 m this discontinuity appears persistent at least in the scale of common outcrops (Fig. 3e). Subordinate sets of joints strike N-S and NNE-SSW and thus parallel to the valley

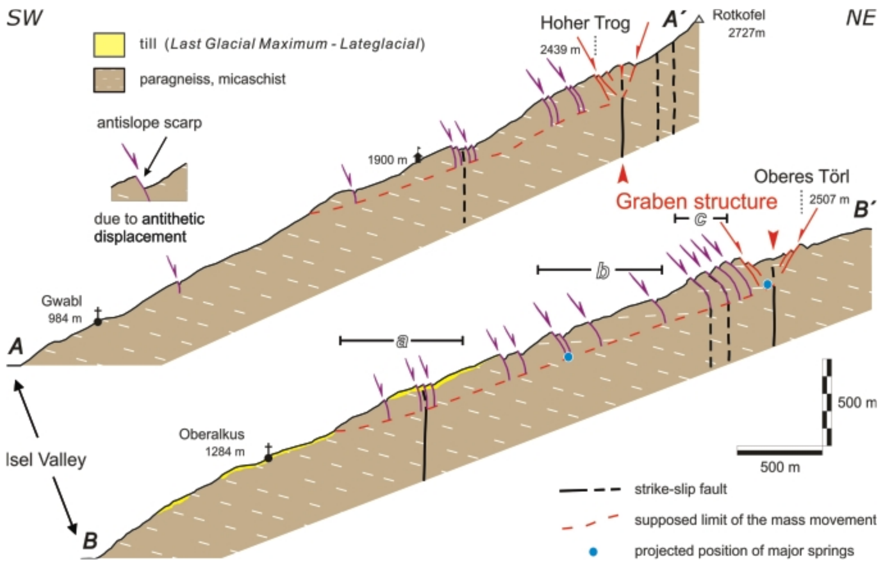

Figure 2: Geological cross sections of the slopes Oberes Törl and Hoher Trog (for location see Fig. 1). Faults are indicated by black lines. White dashed lines denote the apparent orientation of foliation. The estimated depth of movement and collapse (rock disintegration along pre-existing joints) based on the springs occurring in the lateral position (projected into the slope body) is shown by the red dashed line. The regions of the structural data displayed in Fig. 4 are indicated by a, b and c.

of the Daberbach (rose diagram in Fig. 4).

An opening along the foliation, dipping in general between 20° and 45° towards east to southeast, occurs by comparison only subsidiary. However, the foliation shows with the onset of antislope scarps above Oberalkus a remarkable change of orientation in upslope direction compared to the regional trend (a-d in Fig. 4) towards an apparent horizontal position in the cross-section (Fig. 2). In the uppermost areas, where the antislope scarps dip 45° (Fig. 3d), an even gentle dip in S to SW direction (Fig. 2) can be observed (c in Fig.4). Such a downward decreasing rotation of the structural elements is also evident in the dip of the NW-SE striking joints, showing a similar trend as the antislope scarps (a-c in Fig.4).

Based on these observations the rock mass between Oberalkus and the graben structure at Oberes Törl shows a disintegration mostly along the joints and thus a collapse in the sense of Davies et al. (2006) due to dilation and hardly any sign of chemical weathering beyond oxidation along the joints (Fig. 3f). Only the eastward flank of the graben at Oberes Törl (Fig. 3g) indicates a higher amount of rock disintegration with a gradual transition from jointed bedrock to, firstly, an agglomeration of rock blocks which still show indications of structural arrangement and to, secondly, a blockfield. The latter passes into a relict rock glacier (Reitner, 2003), which occurs at the eastward limit of the slope (Fig. 1). Such a succession in rock quality and morphology, typical for the toe of sagging (Zischinsky, 1966) as well as subordinate N-S to NE-SW trending extension structures cross-cutting the main structural elements are regarded to resemble a minor SE-directed creeping mass-movement.

The hydrological situation of the slope between the creeks Daberbach and Gossenbach is characterised by the absence of springs on most of the slope above Oberalkus (Fig. 1). Although most of the linear depressions are dry, several are characterised by boggy ground or the temporary pounding of water at least in their deepest parts. One group of springs occurs on the western flank of the Daberbach in nearly all cases in prolongation of the linear depression with the most prominent example in the southeastward prolongation of the graben structure at Oberes Törl (Fig. 1). The second minor group is located close to the lower limit of the disintegrated part above Oberalkus, but without any evidence of a defined spring line.

The peculiarity of the hydrological system of the slope is also obvious by comparing the creeks Gossenbach and Daberbach. The latter has a rather steep and extensive alluvial fan, which reflects erosion of glacial sediments and talus fan material in its lower and middle part. In contrast, the Gossenbach has in relation to its drainage area, which overwhelmingly consists of the slopes of Oberes Törl and Hoher Trog (see below) a hypotrophic alluvial fan.

3.2 Hoher Trog

This slope spans the southwestern flank of Rotkofel (2727 m a.s.l.) down to the Isel valley (Figs. 1, 2 and 3a). The lateral limits are defined by the creek Gossenbach and the rocky flank to the Leibnitzbach. The lithology and structural setting are identical to that of Oberes Törl.

According to the morphological features and the characteristics of the rock mass this slope may be divided into three segments.

The lowermost part, which is only present in the area NW of the Gwabl village, characterised by an up to 35° slope gradient with a small headwall and talus cone below that reflecting the steepness. The middle slope segment with a mean slope angle of 25° extends from the valley floor south of Gwabl village as a more than 1 km band NW towards the locality Oberfercher. A glacially shaped morphology with small roche moutonnée is typical for this part. Antislope scarps are apparent as a group just above the head scarp area of the lowermost slope segment. This is the only location within this slope segment where rock disintegration with open joints is evident.

The upper slope segment reaching into the region of Hoher Trog (2439 m a.s.l.) is marked by the onset of NW-SE trending antislope scarp sets from 1500 m a.s.l. in the east and 1800 m a.s.l. in the west resulting in a sawtooth-surface (Figs. 3a and 3h). The average slope gradient is 25° with some steeper sections with a gradient of 30°. The dimension of the individual antislope scarps varies from length of 100 m to around 1 km and exhibit vertical differences between the depression and the crest of the antislope scarp in the range of 1 to 10 m. Similar to the Oberes Törl the upslope-facing planes show as well a slope of 25-50°. The spacing between ridges reduces significantly towards Hoher Trog where above 2100 m a.s.l. groups of antislope scarps take on a bundle-like appearance.

A high-elevated trough (translation of Hoher Trog), is located SW of Rotkofel (Figs. 1 and 3a). It is a complex graben structure which in the western part is made up by a single E-W trending linear depression. The eastern prolongation consists of a bundle of NW-SE trending antislope scarps (Fig. 3h). With this development the graben structure gets an up to 200 m broad infill, which morphologically represents an irregular rupture surface. Contrary to the example of Oberes Törl, a defined normal scarp delineating the upper limit of this graben structure is absent. This may be due to superficial mass wastage from the southern flank of the Rotkofel, which has an average slope of 35°. Scarps with downward creeping mass movements are present at the flank of Rotkofel only to a minor extent, as indicated in the eastern part south of Mirschachscharte (Fig. 1) There, the zig-zag run of the antislope scarps (Fig. 3h) below may reflect the lateral pressure of this marginal and still active creeping mass movement. It is noted that indications are absent that the whole flank of the Rotkofel down to ~ 2500 m a.s.l. may represent a scarp.

The intersection line of the antislope scarps and linear depression with topography denotes that the case of Hoher Trog is the morphological expression of a > 150 m deep-reaching rupture of the slope (Figs. 1 and 2). From this morphological relationship, as in the case of Oberes Törl, it can be concluded that the antislope scarps in the lower part represent more

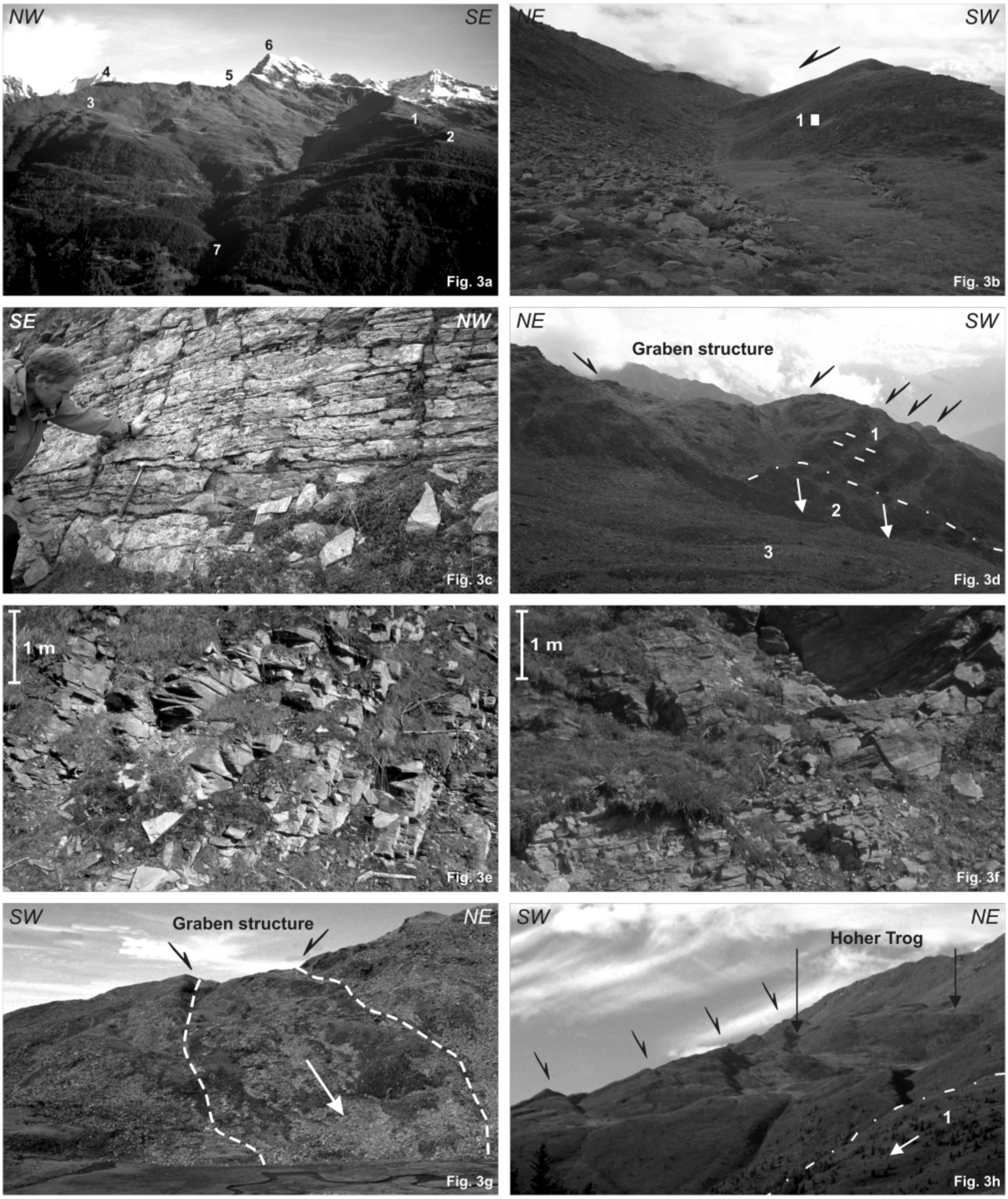

Figure 3: Fig. 3a. View of the slopes of Oberes Törl and Hoher Trog with multiple linear elements (antislope scarps) from the opposite Isel valley flank. (1) Oberes Törl, (2) Unteres Törl, (3) Hoher Trog, (4) Rotkofel, (5) Mirschachscharte, (6) Prijakt, (7) Gossenbach creek. Fig. 3b. Typical antislope scarp developed in mica schist/paragneiss at the slope of Oberes Törl in 2340 m a.s.l. White bar (1) indicates the size of a person located at the outcrop shown in Fig. 3c. Fig. 3c. Detail of Fig. 3b. View of the NE dipping fault/joint plane with foliation gently dipping towards south and thus intersecting the fault/joint plane at a low angle. Vertical traces denote NE-SW striking joints. Fig. 3 d. Graben structure at Oberes Törl followed by a series of antislope scarps as indicated by the arrows (view towards SE). White lines (1) indicate the daylighting foliation gently dipping towards south (structural data of the part below this graben are shown in stereo plot c of Fig. 4). Dashed line and white arrows (2) denote secondary mass movement detaching along NE-SW striking joints towards tributary valley of Gossenbach with relict rock glacier of Lateglacial age (3). Fig. 3e. Paragneiss with a closely spaced vertical NW-SE striking joint set and foliation dipping gently into the slope. Fig. 3f. Open joints within the rock mass of Oberes Törl indicate collapse, i.e. dilation, along pre-existing joints. Fig. 3g: View towards NW onto the eastern flank of the graben structure of Oberes Törl (dashed line und white arrow, see also Fig. 3d) with rock mass creeping towards the valley of Daberbach creek. Fig. 3h. View towards NW onto the graben structure of Hoher Trog and anti-slope scarps (black arrows) showing a zig-zag run of the crest line due to clockwise rotation at their SE flank caused by the creeping mass movement S of Mirschachscharte (1; white arrow indicates direction of movement and bulging toe).

or less vertical planes, whereas the upper ones give evidence to surfaces dipping rather steeply into the slope (Fig. 4).

The morphological features show striking similarities to the large-scale tectonic structures, i.e. brittle faults at the northwestern flank the Rotkofel with respect to trend and distribution. On the one hand the NW-SE running antislope scarps do not only show the same strike as the vertical Isel parallel faults but also occur in many cases directly in the prolongation of mapped faults. On the other hand the apparent younger E-W trending brittle fault finds its morphological continuation in the western part of the graben of Hoher Trog.

The dominant joints are those striking parallel to the Isel fault and dip more or less vertically (Fig. 4). Their spacing ranges between 1-3 m. The foliation gently dipping SE to NE is the second apparent structural element within the scale of outcrops (Fig. 4). A minor joint set trending WSW-ENE is subordinate. The disintegration of the rock mass with open joints is evident along the tracks of the upper slope segment and especially also above the timberline near Hoher Trog.

The upper slope does not show any superficial drainage in the major part. There are only two springs and an ephemeral water flow in the upper drainage area of the Gossenbach between 1800 and 2100 m a.s.l., all of them located in the downward continuation of the above described linear morphological elements. Two ponds, which are downward limited by antislope scarps, are present as well. Interestingly, one of them with a position in 2070 m a.s.l. just in the slope line of the Hoher Trog, has a superficial inflow but no outflow and acts, thus, as a sinkhole (Fig. 1).

4. Kinematic Tests

Due to the structural setting of both slopes with a set of discontinuities (especially faults) which have a vertical orientation or a steep dip into the slope in combination with a slope-parallel trend, common toppling (Goodman and Bray, 1976) seems to be a reasonable slope failure mechanism (Fig. 4). In order to test such assumption, simple kinematic analysis was performed in a first attempt based on Goodman and Bray (1976) using the basic condition for flexural toppling:

$$\Theta > \phi + 90 - \Psi \qquad (1)$$

where Θ is the slope angle, ϕ is the angle of friction and Ψ is the dip angle of the controlling discontinuity. In a first assumption the angle of friction, which actually is not known, was set zero and modern slope angles of ~ 25° were taken.

The results (Fig. 4) show that the overwhelming part of the dominant NW-SE trending joint and fault system falls into the field, where flexible toppling is kinematically possible. The same is true for the poles of most antislope scarps, whose dip was deduced from their intersection with topography. It is noted that the antislope scarps located at the upper part of Oberes Törl with a dip < 65° are beyond the flexural toppling field. The same is true for all foliation data.

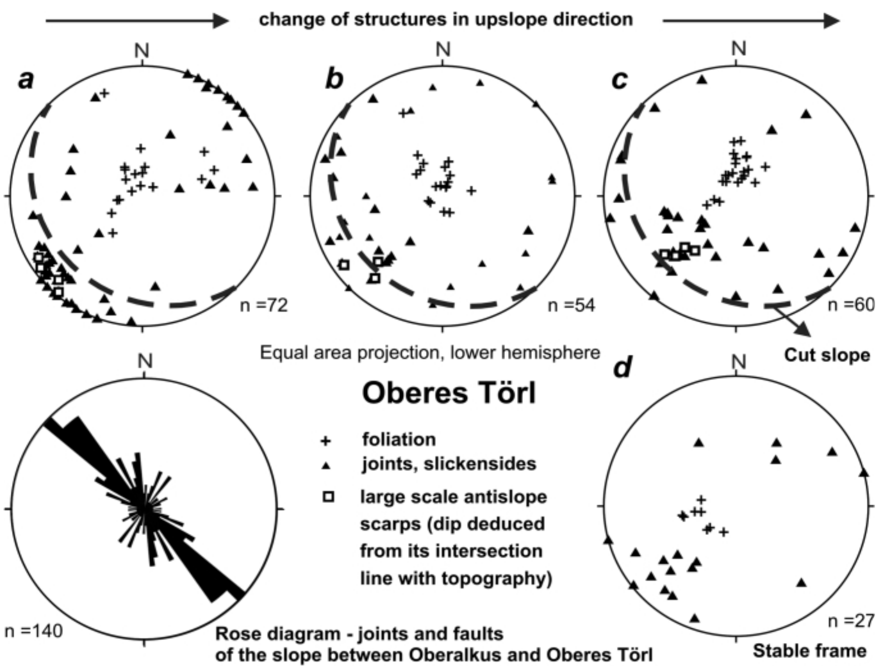

Figure 4: Slope of Oberes Törl: Stereo plots of joints, foliation and antislope scarps from the slope segments above Oberalkus (a; approximately 1400 -1900 m a.s.l.), around Unteres Törl (b; approximately 2000-2200 m a.s.l.) and from the uppermost antislope scarps adjoining to the graben structure (c) exposed at the NW flank (see Figs. 1, 2 and 3d). Great circles denote the orientation of the slope with an average gradient of 25°. The plots a-c show an upslope change of these structural elements i.e. concurrent decrease of dip angle of antislope scarps and NE dipping joint system. The orientation of foliation shows a shift from E-SE to S-SW. Downward decreasing rotation of structural elements is also evident by comparison to data from an area 1 km SE of the graben (stereo plot d - “stable frame”; location indicated in Fig. 1) with a subglacially shaped surface and no indication of mass movements. The rose diagram of joints and slickensides displays the dominance of a NW-SE striking joint system parallel to the Isel fault over those with a N-S and NNE-SSW trend.

5. Discussion

5.1 Mass Movement Structures and Slope Failure Mechanism

Already the morphological characteristics of the upper segments of both slopes with a graben structure and a sawtooth-surface, which differ substantially from that of the subglacially smoothed lower slopes, may only be explained by gravitational mass movements. It is highlighted that there is no major head scarp and no defined bulging toe indicated by the morphology (Fig. 2 and 3a). In addition no significant downslope change in the properties of the rock mass, in the sense of a downward increase of disintegration as described for typical sackung (Zischinsky, 1966) can be observed. The rocks of the areas with sawtooth-surface

surface exhibit more or less the same orientation of the tectonic structures as in the areas with no indication of mass movements, but with open joints.

A comparison of the spatial distribution of parallel trending NW-SE vertical faults with that of the graben structures and antislope scarps (Fig. 1) show that the slope deformation structures are evidently linked to areas with persistent brittle deformation. It even appears that the bundles of antislope scarps, for example at Oberes Törl, reflect in much more detail the intensive brittle segmentation of this area, than it was able to perceive it in the stable frame.

Faults and connected joints internally rotated in sense of the pivot point located within the body by up to 45° as shown for Oberes Törl (Figs. 2, 3d, 4 and 5) and to a lesser amount of approximately 10-20° at Hoher Trog (Figs. 1 and 5) suggest in combination with the observed disintegration large-scale toppling as mechanism for these DSGDs.

Simple kinematic tests according to Goodman and Bray (1976) support this first assumption as the dominating brittle tectonic structures (faults, joints) trending NW-SE, parallel to the slope, enable flexural toppling (Figs. 4 and 5). Thus, this type of RSF seems to be a reasonable mechanism for the formation of all described morphological and structural features of Hoher Trog and most of Oberes Törl. If we consider the limited amount of displacement along the antislope scarps than no major change in the general slope gradient between the pre-failure and modern topography may be expected. A tentative reconstruction of the former morphology for the slope of Oberes Törl indicates nearly the same slope above Oberalkus as the presence of steeply dipping antislope scarps in this part of the slope excludes major bulging due to kinematic reasons. In the area between 2200 and 2400 m a.s.l. where the largest amount of rotation of antislope scarps occurred, a slightly lower gradient of the pre-failure topography appears likely. Consequently, the most obvious morphological change occurred due to subsidence in the graben structure.

Interlayer slip as a precondition for this failure type (Goodman and Bray, 1976) focussed along pre-existing brittle faults. Their persistent nature allowed such a concentration of strain in the course of the RSF. A low angle of friction along the fault planes is assumed due to the presence of cataclastites with slickensides. Anyhow, any speculation of the rock properties is complicated by the reduction of normal stress due to possibly high hydrostatic pressure along the faults, a fact that should be considered at least for the initiation of the movement.

However, with this given topography and the simplifying assumption of $\phi = 0$ flexural toppling should have stopped and thus stabilized itself, when Ψ (dip to discontinuities) had fallen below approximately 65° due to tilting in the course of slope failure. Nevertheless, the field evidence from at least the uppermost part of Oberes Törl (Fig. 4) with rotated fault and joint planes dipping approximately 45 ° into the slope requires an additional mechanism of slope deformation.

The discontinuity planes of the rock masses between the main antislope scarps had to accommodate the toppling movement. The most obvious candidate for this process are the closely spaced joint systems parallel to the NW-SE trending faults, with a more or less vertical orientation (Fig. 3e). Like in the case of the major brittle faults this discontinuity planes enabled interlayer slip according to the criteria of Goodman and Bray (1976; Fig. 4), even if they are not as persistently developed regarding their lateral and vertical extent compared to large scale brittle structures. In addition, the angle of friction along these joint planes is assumed to be higher compared to that of the fault planes, due to higher roughness. The foliation with an apparent gentle dip into the slope as indicated in the cross sections (Fig. 2) is reactivated in general as a cross-joint-system in relation to the dominant fault and joint system. Its original position does not enable either sliding or toppling. Nevertheless, the initial toppling of the rock columns between the main antislope scarps led to a rotation of the foliation, which in the extreme resulted in a more or less horizontal position. Such a rotation of the cross-joint system is regarded to have facilitated the overall bending at small scale.

A different effect is proposed for the most tilted area in the uppermost region of Oberes Törl, at the downslope limits of the graben structure (Fig. 3d), where an even apparent gentle downslope dip of the foliation planes with an angle less than the slope gradient is observable (Figs. 3b, 3c and 4). In this case the combination of the incremental increase of rotation in the upslope direction due to flexural toppling of the slope below and the related overturning of the foliation (cross-joints) may have enabled a rock-failure mode which is between the flexural type and block toppling (Goodman and Bray, 1976; Nichol et al., 2002). In latter mode the cross-joints provide release surfaces for the rotation of columns, which than rotate forward driven by their own weight and not by a principal stress parallel to the slope as assumed for flexural toppling (Nichol et al., 2002). Block-flexural toppling characterised by pseudo-continuous flexure of long columns through accumulated motions along numerous cross joints (Goodman and Bray, 1976) describes this part of the slope best. Limited sliding along overturned cross joints may have occurred in the small scale but did not result in large scale scarps with synthetic displacement in the scale of mapping. Such a model consistently explains the up to 45° dipping antislope scarps, which are kinematical impossible according to the pure flexural mode. Thus, the slope failure mode of Oberes Törl as well as of Hoher Trog is dominated by flexural toppling, whereas at least the uppermost part of Oberes Törl shows block-flexural toppling.

From the mapping results a minimum depth of these DGSDs of 100 m can be inferred based on the intersection of the antislope scarps with morphology. If we take the position of the springs in prolongation of some of the scarps into account, than the depth of disintegration may reach down to 200 m at least for Oberes Törl but most probably also for parts of Hoher Trog. According to the flexural toppling mechanism a decrease of accompanied displacement towards the depth is supposed similar to that of sackung as defined by Zischinsky (1966). The basal limit of disintegration within the cross-sections (Fig. 2) denotes only the supposed lower limit of toppling and no clearly defined rupture surface.

5.2 Onset of Movement and Conditions of Preservation

As far no geochronological data like exposure datings of antislope scarps (e.g. Hippolyte et al., 2006, 2009) are available, any consideration regarding the age of the mass movements has to rely on some assumption: Evidently the Isel glacier shaped the slope of the valley flanks up to its trimline, which is in the regional setting located between 2200 and 2300 m a.s.l. Subglacial erosion below this limit resulted in a relatively oversteepened pre-failure topography. The condition of a free-face was given with the Isel glacier collapse-like down-wasting during the Early Lateglacial phase of ice-decay (Reitner, 2007) which lasted approximately from 21 kyr to 19 kyr. We suppose that the toppling started concurrent with this loss of support, as on the hand the faults as the pre-existing zones of weakness failed due to their structural characteristic. A lack of vegetation and an abundance of meltwater, especially on this southward exposed flank, may have led to joints filling with water. Hence, the first opening of gaps in the course of the mass movements resulted in high hydrostatic pressure in the fault zones. These conditions should have reduced the internal friction substantially and facilitated interlayer slip. Rockglaciers dating from the Oldest Dryas (Buchenauer, 1990) were already formed in the disintegrated rock mass of Oberes Törl where their flow paths followed depressions generated by toppling (Fig. 1; Reitner, 2003). This interpretation of Lateglacial slope evolution is supported by the dating of a travertine E of Ainet (Fig. 1), which is located at the toe of small scale toppling slope close the study area. The U/Th-date of 13.5 kyr BP (Boch et al., 2005) shows that a deep-reaching solution transport, probably due to deep-seated mass movements, existed already at that time. Small antislope scarps cutting a rock-glacier deposit SE of Oberes Törl supposedly denote at least partial younger reactivations of the toppling RSF.

In total the RSF seems to be settled at present day due to the self-stabilising of the dominant flexural toppling mode according to the kinematic stability criterion of Goodman and Bray (1976), a situation which had been found in similar cases (Nichol et al., 2002).

Another pre-requisite for the relative stability of the two toppling slopes as well for the preservation of their structures lies in the location of their lower limits in the middle part of the overall slope. Thus, no slope-toe erosion by the Isel river has been possible. In addition secondary mass-movements at the toe, which develop in already disintegrated rock masses and which are common in such a position at sackung-slopes, are completely absent. The other major condition for the preservation of this peculiar saw-tooth morphology is the orientation of the foliation (cross-joint system) dipping into the slope which does not support sliding also in the disintegrated part. Even in the overturned part near Oberes Törl such a failure is unlikely due to the non-penetrative nature of the cross-joints. In addition, the creeping rock mass at the eastward flank of the graben structure at Oberes Törl (Fig. 3g) with a movement towards the tributary valley and thus more or less perpendicular to that of the main DGSD should have resulted in an unloading of the uppermost area (graben) and, hence, have had a stabilising effect on the toppling RSF. Like in a similar case (Reitner et al., 1993), this chronology of mass movement processes may reflect that of deglaciation as the tributary valley got ice-free during the Lateglacial after the main valley (Buchenauer, 1990).

Reactivations of these RSFs may occur in principal due to a weakening of the toe caused by secondary rock creep at the lower flanks (facilitated by a joints trending parallel to the tributary valleys) or directly in the frontal position. The potential

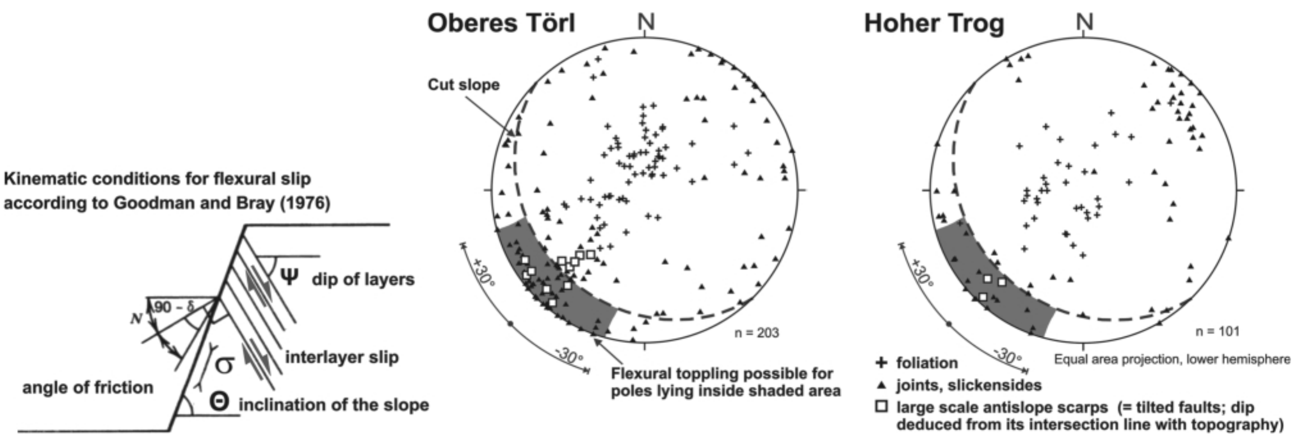

Figure 5: Stereo plots of structural data (poles of joints, foliation and antislope scarps) of the upper slope segments of Oberes Törl and Hoher Trog and kinematic analyses for flexural toppling according to Goodman and Bray (1976). The great circles display the cut of the slopes with an average gradient of 25° towards SW. As a simplification, the friction angle along joint surfaces is disregarded (see text). Flexural toppling due to interlayer slip is possible for vertical to NE-oriented discontinuities, whose poles fall into the shaded area. This is true for joints and antislope scarps parallel the Isel fault, which resemble tilted faults from the slopes Hoher Trog and the lower part of Oberes Törl (see also Fig. 4a). The formation of antislope scarps located at the upper part of Oberes Törl (see also Fig. 4b-c) with a dip less than 65° and, hence, with poles plotting outside, cannot be explained solely by flexural toppling.

effect of higher precipitation may be quite limited due to the joint aquifer caused by dilation in the course of toppling. In addition, the current situation makes a transition from toppling to sackung due to the stress at the toe and a progressive collapse as described by Poisel (1998) unlikely.

5.3 Comparison of Toppling Failures and Their Distribution in the Eastern Alps

Toppling failures are commonly described, but partly in different context. The definitions by Goodman and Bray (1976) include common toppling consisting of flexural, block and block-flexural type, and secondary toppling, e.g. slide-head-topple also referred as sliding-induced toppling (Bovis and Evans, 1996) as a consequence of a block slide. As a complication, transitions from toppling to a classical sackung in the sense of Zischinsky (1966, 1969) may occur, as the stress at the toe may lead to an increase in rock disintegration and finally to a plastic behaviour typical for the convex slope of sackung due to a complete collapse of the rock mass into joint-controlled particles. In an intermediate state graben structures and antislope scarps are accompanied with a bulging toe (Poisel, 1998). Additionally toppling-induced sliding (Bovis and Evans, 1996) may mask the initial failure process.

Examples of anaclinal slopes, characterised by discontinuities with a vertical orientation or a dip into the slope (Cruden, 1989) with a "pure" toppling failure similar to that of Hoher Trog and Oberes Törl are rarely described in the literature.

Examples from British Columbia (Bovis, 1982) and Scotland (Holmes and Jarvis, 1985) appear to be the most similar examples from outside the Alps. The closest analogues concerning structural setting as well as morphology are the examples from the Swiss Alps presented by Ustaszewski et al. (2008). There the toppling failure occurred as well along pre-existing vertical faults, which in the view of the authors were as well re-activated by postglacial differential uplift as a result of deglaciation. However, in the cases of Oberes Törl and Hoher Trog no indication has been found to support any postglacial rebound (Ustaszewski et al., 2008) as a relevant process in the formation of antislope scarps. Such a process should have had an imprint outside the mass movements as well, which has not been detected so far during our intensive mapping campaign. Unlike Ustaszewski et al. (2008) any endogenetic reason for the formation of antislope scarps in the sense of neotectonic re-activation is excluded in our examples as the brittle tectonic structures have been tilted by a force acting parallel to the slope surface.

More examples are known from the Eastern Alps, especially from the longitudinal valleys following major Miocene faults, e.g. Möll valley (Pirkl, 1972). There, brittle faults parallel to the Möll fault may have been as well important for toppling as indicated by subsurface investigations in the adjoining slopes with the same RSF (Clar and Demmer, 1979; Horninger and Weiss, 1980). Another prominent case of flexural toppling with interlayer-slip along the schistosity surfaces of phyllonites and micaschists of the Gailtal basement is described by Reitner et al. (1993) from the flank of the Gail valley, which follows the Periadriatic fault system.

Numerous antislope scarps accompanied by deep seated toppling failure are also evident in the steeply dipping Quarzphyllit complex in the Salzach valley close to the Salzach fault (Heinisch et al., 2003). Further toppling failures have been reported by van Husen (1995) and Fischer (1997) within Palaeozoic schists of the Carnic Alps, where the slopes are often characterised by antislope scarps (Schönlaub, 1997 and 2000).

The slopes of Oberes Törl and Hoher Trog are even outstanding if compared with the adjoining slopes along the Isel valley, which were mapped and analysed in the same detail. The adjoining slope in SE direction is intersected by the continuation of the Isel fault-system, but exhibits a general dip of foliation towards SW. There prominent mass movements are developed showing different stages of slope disintegration (Reitner, 2003), which probably started with toppling failure.

6. Conclusions

In the area Oberes Törl and Hoher Trog, the characteristic saw-tooth morphology of the slope, which is caused by a series of antislope scarps, the graben structure in the uppermost part and the hydrological regime (no surface flow, location of springs), all indicate a DGSD with a probable depth of rock disintegration of about 200 m. The structural control is indicated by the pattern of mass movement structures (antislope scarps and graben) trending parallel to brittle structures linked to the Isel fault system. Kinematic analyses according to Goodman and Bray (1976) prove large scale toppling as the only plausible RSF mechanism due to interlayer-slip along vertical discontinuities (faults and joints). Flexural toppling is inferred to be the dominant mode of RSF, whereas the block-flexural mode is supposedly more important in the topmost region, at least SW of Oberes Törl, where antislope scarps with the lowest dip and the most rotated pre-existing discontinuities (faults, joints and foliation) occur. The observed rock disintegration with open joints can be described as the result of an initial collapse (in the sense of Davies et al., 2006) along existing discontinuities.

In a probable scenario, glacial over-steepening during the LGM followed by the loss of support in the early Lateglacial phase of ice-decay caused the onset of the RSF. Plenty of meltwater, and no vegetation to balance the hydrology of the slope, resulted in high hydrostatic pressures along the discontinuities and thus high strain rates during this climatic period. At present day, the mass-movement seems to have ceased due to a self-stabilizing effect of the flexural-toppling mode at the lower part of the DGSD. In addition, (secondary) rock creep along the flanks, especially in the area of the heavily disintegrated graben structures, towards the tributary valleys may also have had a major impact on slope stability due to the reduction of load in the topmost part.

As transitions from toppling to sackung may occur (Poisel, 1998), our examples should highlight the important factors for preserving typical large scale toppling. Firstly, the orientation of the cross-joint system, in this case the foliation, prevents a downward slip due to kinematic reasons. In combination with missing fluvial erosion at the toe and unloading at the flanks caused by secondary slope creep, this seems to be the most plausible explanation for the conservation of large-scale toppling as a rare member of DGSD.

Our results emphasise the importance of large-scale toppling as a mainly structure-controlled process, and an initial stage of DGSD, which may be focussed in valleys following major fault zones, like the longitudinal valleys of the Eastern Alps. This special kind of RSF should be separated from other mass-movement types on geological maps, if possible, in order to anticipate other potentially more disastrous DGSDs or even catastrophic rock slope failures like rock falls (e.g. McAffee and Cruden, 1996) and rock avalanches (e.g. Nichol et al., 2002). In addition, in such a setting large scale toppling initiating further mass wasting has to be regarded as a major, possibly underestimated process in valley shaping and landscape evolution of Alpine regions.

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