Tectonic vs. gravitational morphostructures in the central Eastern Alps (Italy): Constraints on the recent evolution of the mountain range

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A R T I C L E   I N F O

Article history:
Received 7 March 2008
Received in revised form 22 December 2008
Accepted 22 February 2009
Available online 28 February 2009

Keywords:
Faulting
Active tectonics
Deep-seated gravitational slope deformation
Relief
Numerical modelling
Central Eastern Alps

A B S T R A C T

Deep-seated gravitational slope deformations (DSGSDs) influence landscape development in tectonically active mountain ranges. Nevertheless, the relationships among tectonics, DSGSDs, and topography are poorly known. In this paper, the distribution of DSGSDs and their relationships with tectonic structures and active processes, surface processes, and topography were investigated at different scales. Over 100 DSGSDs were mapped in a 5000 km² sector of the central Eastern Alps between the Valtellina, Engadine and Venosta valleys. Detailed lineament mapping was carried out by photo-interpretation in a smaller area (about 750 km²) including the upper Valtellina and Val Venosta. Fault populations were also analysed in the field and their mechanisms unravelled, allowing to identify different structural stages, the youngest being consistent with the regional pattern of the ongoing crustal deformation. Finally, four DSGSD examples have been investigated in detail by geological and 2D geomechanical modelling. DSGSDs affect more than 10% of the study area, and mainly cluster in areas where anisotropic fractured rock mass and high local relief occur. Their onset and development is subjected to a strong passive control by mesosopic and major tectonic features, including regional nappe boundaries as well as NW–SE, N–S and NE–SW trending recent brittle structures. The kinematic consistency between these structures and the pattern of seismicity suggests that active tectonics may force DSGSDs, although field evidence and numerical models indicate slope debuttressing related to deglaciation as a primary triggering mechanism.

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1. Introduction

The occurrence of dynamic feedbacks between tectonic and surface processes has been emphasized by a number of authors (Molnar and England, 1990; Burbank and Anderson, 2001). Surface erosion and sediment transport by glacial, fluvial and gravitational processes modify the topography and redistribute large masses over significant distances. This may result in changes of the style and rate of tectonic deformation (Burbank and Anderson, 2001). On the other hand, the rate and type of erosion are dominated by climate, rock type, fracturing, and topography. The latter is strongly controlled by vertical tectonic uplift related to flexure, isostasy, thrusting and crustal thickening processes (Molnar and England, 1990). Thus, tectonics drives surface processes, which in turn could force tectonic processes evolving a positive feedback. The relationships between tectonic processes and surface erosion by fluvial, glacial, and catastrophic landslide processes have been studied by a number of investigators (Densmore et al., 1998; Burbank and Anderson, 2001; Brocklehurst and Whipple, 2007; Korup et al., 2007). On the other hand, an important role in landscape development in tectonically active mountain ranges is played by large-scale, deep-seated gravitational slope deformations (DSGSD).

DSGSDs are large mass movements involving high-relief slopes from their toe up to or beyond the ridge. They can involve nearly all rock types, and are characterised by poorly defined lateral boundaries, large volumes (often > 0.5 km³) and thickness, and relatively low present-day displacement rates (few to tens mm/yr; Varnes et al., 1990; Ambrosi and Crosta, 2006). Further diagnostic features of DSGSD are the development of persistent linear morpho-structural features (Agliardi et al., 2001). Scars, trenches, grabens, double or multiple ridges are the more frequently described tensional features in the upper slopes, whereas counterscarps (i.e. antitensional scars) are mainly observed in the middle part. Bulging, buckling folds or highly fractured rock masses are compressional features occurring at the slope toe. Large catastrophic slope instabilities (rockslides and rock-avalanches) also often occur in DSGSD areas.

In general, DSGSDs appear closely related to specific geological and structural features (bedding, foliations, joints, faults, etc.) and morpho-topographic factors, and frequently occur in tectonically active areas (McCalpin, 1999; Crosta and Zanchi, 2000; Agliardi et al., 2009). Phenomenological and mechanical descriptions of the processes leading to DSGSD formation have been proposed by several authors, interpreting features observed in the field as representative of specific mechanisms including: postglacial debuttressing of

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oversteepened slopes and associated groundwater changes (Agliardi et al., 2001; Ballantyne, 2002), topographic stresses in high-relief slopes (Radbruch-Hall, 1978; Varnes et al., 1989), effects of tectonic stresses (Miller and Dunne, 1996), earthquake ground shaking (Radbruch-Hall, 1978; McCalpin, 1999), and fluvial erosion of the slope toe (Crosta and Zanchi, 2000).

A major pitfall in the analysis of the relationships between tectonic and DSGSD-related features is their morphological convergence. Widespread morpho-structural evidence of postglacial surface ruptures in the Alps (Forcella et al., 1982; Crosta and Zanchi, 2000; Sue and Tricart, 2002; Persaud and Piffner, 2004) could be related to recent or active tectonic processes, to the gravitational deformation of large rock slopes, or to a combination of them (McCalpin, 1999; Agliardi et al., 2009). Gravitational morpho-structures have been often interpreted as the morphological expression of active faults, possibly biasing the overall interpretation of the tectonic evolution of some areas (Forcella and Orombelli, 1984; Bovis and Evans, 1995; Hippolyte et al., 2006). Misinterpreting morpho-structural evidence could also influence the assessment of the seismic potential of active faults (McCalpin, 1999), the reconstruction of the glacial/paraglacial history, as well as the evaluation of postglacial sediment budgets in large alpine catchments. On the other hand, tectonic controls on DSGSD can be difficult to ascertain, especially in tectonically active areas with low magnitude seismicity (M<5; Slejko et al., 1989).

This study aims at mapping and characterising tectonic and gravitational morpho-structures at different scales (i.e. regional to single slope scale), in order to investigate the distribution of DSGSDs and their relationships with tectonic structures, active tectonic processes, surface processes, and topography. Aerial photo-interpretation over 5000 km² between Valtellina, the Engadine valley and Val Venosta (central Eastern Alps, Fig. 1) allowed mapping of more than 100 DSGSDs. Detailed morpho-structural lineament mapping was carried out by photo-interpretation and field surveys in smaller areas (about 750 km²) including Upper Valtellina and Upper Val Venosta, where mesoscopic fault populations have also been analysed. These data were compared to seismological, geodetic and in situ stress data, to discuss the relationships among structural settings, active faulting, and DSGSD. Finally, four DSGSD have been analysed in detail by geomechanical numerical modelling, in order to evaluate the influence of tectonic and surface processes on the development of these phenomena.

2. Geological setting and active tectonics

2.1. Geological setting and evolution

The study area is part of the Austroalpine nappe stack, exposed between the Engadine and the Periadriatic Lineament (Fig. 1). The pre-Alpine and Alpine tectono-metamorphic evolution of these units has been described by several authors (Ratschbacher, 1986; Thöni and Hoinkes, 1987; Ratschbacher et al., 1989; Martin et al., 1991; Thöni, 1999; Söllva et al., 2005).

The Oetztal and the underlying Sesvenna-Campo nappes, which form most of the area, mainly consist of poly-metamorphic basement including gneiss and schists, with subordinate orthogneiss, amphibolite,
and marble, separated by poorly metamorphic Permo-Mesozoic sedimentary rocks. Both units show well-preserved amphibolite-facies associations overprinted by greenschist-facies assemblages. The main difference between the two units is the occurrence of Variscan eclogitic relics in the Oeztal nappe. A high-pressure Eo-Alpine metamorphic event was also recognized north of Merano in the Texel Massif, which is tectonically interposed between the two main units. The Campo nappe, consisting of phyllites in the region around Bormio, lacks these peculiar assemblages and shows more preserved Variscan associations. The sedimentary cover of the Sesvenna-Campo nappe (i.e. Ortler and S-charl units) mainly consists of Upper Palaeozoic terrigenous metasediments, passing to thick Triassic carbonate successions exposed north of the Zebru Line.

Major regional structures occurring in the area are the Schlinig Line (Schmid and Haas, 1989; Froitzheim et al., 1997) and the Zebrù Line (Conti et al., 1994). The former gently dips to the E–SE along the Upper Val Venosta, stacking the Oeztal nappe above the Sesvenna-Campo nappe. Schmid and Haas (1989) interpreted this structure as an intra-basement shear zone (Fig. 1), associated with the Early Cretaceous west-directed tectonic transport of the Oetztal Nappe. The same structure was reactivated during the Late Cretaceous as a shallow normal fault post-dating nappe stacking (Froitzheim et al., 1997). The Zebrù Line occurs north of Bormio and steeply deeps to the North, separating the Campo nappe from the overlying Ortler sedimentary cover. Several minor units (e.g. the Grosina and Chavalatsch nappes) are also included within the Austroalpine stack. Thick mylonitic and cataclastic layers occur along the main tectonic contacts.

Following nappe emplacement, exhumation of HP units, and at least two stages of post-nappe folding, N–S shortening synchronous with lateral continental extrusion along strike-slip and normal faults occurred since Late Oligocene (i.e. Periadriatic, DAV, Brenner Lineament) marking the recent evolution of the belt. E–W to ESE–WNW dextral motion was associated with back-folding and related back-thrusting along the North Giudicarie and Merano–Mules NE–SW trending branch of the Periadriatic Lineament, now interpreted as an original restraining bend of this fault, leading to the formation of ductile to brittle shear zones between 30 and 20 Ma ago (Martin et al., 1991; Prosser, 2000; Mancktelow et al., 2001; Müller et al., 2001; Viola et al., 2001, 2003). Exhumation of the Penninic units now exposed in the Tauern window occurred at the same time along the N–S trending Brenner normal fault (Selverstone, 1988).

During the Middle Miocene, back-thrusting in the Southern Alps was coeval with the northward propagation of the North Giudicarie fault along the Passiria valley (north of Merano), as a NNE–SSW sinistral strike-slip fault. The Engadine Line, a sinistral oblique-slip fault bounding the Engadine window to the east, also became active at this time (Schmid and Froitzheim, 1993) causing the exhumation of the Penninic units in the Engadine window. Both lines still show seismic activity (Caporali et al., 2005).

2.2. Recent and active tectonics

Recent and active tectonics was recognised in the central Eastern Alps by several authors, based on seismological data (Sljeko et al., 1989; Deichmann and Baer, 1990; Braunmüller et al., 2002; Caporali et al., 2005); GPS permanent network and precise levelling data (Kahle et al., 1997; D’Agostino et al., 2005; Tesauro et al., 2006), as well as by photogeological analyses (Forcella et al., 1982), mesoscopic structural
analyses of fault populations and in situ stress determinations (Zanchi et al., 1995; Persaud and Pffner, 2004; Reinecker et al., 2005; Agliardi et al., 2009).

The region is characterised by upper crustal microseismicity, increasingly more frequent toward the internal sector of the belt and particularly clustered in the Engadine, Valtellina and Venosta areas (Slejko et al., 1989; Fah et al., 2003; ECOS, ETHZ-Red Puma, Harvard CMT sources; Fig. 2). In spite of the large uncertainty affecting the localisation of small ($M_b<2$) and shallow events, epicentres cluster in the Graubunden between Chur and the Engadine, in the Upper Valtellina and Mustair valley, and along the western flank of the Upper Val Venosta (Fig. 2). Earthquakes within the internal Alps are restricted to the upper 15 km of the crust, contrary to the northern foreland where earthquakes occur throughout the crust down to the Moho at 30 km (Deichmann and Baer, 1990). In the study area, the Earthquake Catalogue of Switzerland (Fah et al., 2003) reports about 200 earthquakes with moment magnitude ($M_w$)>2 in the 1975–2008 period, with maximum values reaching 4.9. Two earthquakes with $M_w>4$ occurred on 29–31 December 1999 north of Bormio (maximum $M_w=4.9$; Braunmiller et al., 2002). Moment tensor determinations, based on full-waveform inversion of the main shocks, indicate normal faulting mechanisms with nodal planes striking NNW–SSE (Braunmiller et al., 2002). Another notable event was recorded on 17 July 2001 near Merano ($M=4.8$; Caporali et al., 2005). For this event, focal planes suggest strike-slip motion along a NE–SW left-lateral fault parallel to the North Giudicarie Line. Recent tectonic activity along NNE–SSW sinistral and NW to WNW–ESE dextral strike-slip faults had already been suggested by Forcella et al. (1982) based on photogeological investigations.

We tested the entire set of available fault planes solutions (Fig. 3a) through the P and T dihedra method (Angelier, 1984), to evaluate seismic data in terms of stress regime. The compatibility areas for $\sigma_1$ and $\sigma_3$ and their centroids (Fig. 3b), obtained by using 12 couples of nodal planes, suggest a horizontal N–S trending $\sigma_1$ axis and a horizontal E–W trending $\sigma_3$ axis. These results agree with crustal deformation measurement by GPS permanent networks (D’Agostino et al., 2005; Tesauro et al., 2006). These show a rotation of the compressional geodetic strain axes to N–S in the Eastern Alps, possibly as a result of the counter-clockwise rotation of the Adria microplate (D’Agostino et al., 2005), causing a compression consistent with the
present stress regime in the central Eastern Alps (Müller et al., 1992; Mariucci and Müller, 2003; Reinecker et al., 2005).

Available geodetic levelling data (Kahle et al., 1997) suggest that the area between Chur and the Upper Valtellina and Venosta valley is presently undergoing a surface uplift up to 1.2–1.4 mm/yr (Fig. 2). This rate is consistent with the high values of topographic relief (>1500–2000 m) of the area, which are typical of landscapes developed under tectonic forcing (Kühni and Pfiffner, 2001; Montgomery and Brandon, 2002). Persaud and Pfiffner (2004) related the uplift rate in the area to ongoing tectonic processes, although other researchers proposed that it could be related to isostatic reaction to crustal overthickening and erosion (Schlunegger and Hinderer, 2001), or Last Glacial Maximum (LGM) glacier removal (Gudmundsson, 1994).

3. Spatial distribution of DSGSD

DSGSDs were mapped at the regional scale to compare their distribution to the local topographic, geological and tectonic settings. Mapping was based on the analysis of aerial and satellite images at different scales. This allowed to recognise and map 105 DSGSDs (Fig. 4), each one extending between 0.3 km² and 34 km² (Fig. 5a), whereas the total area involved in DSGSD is about 460 km² (9.2% of the study area). Three main DSGSD clusters were recognised north of Tirano (Valtellina), around Bormio (Upper Valtellina) and between the Mustair valley and the Upper Val Venosta. The first cluster includes the Padrio-Varadega DSGSD, i.e. the largest observed in the Alps, covering an area of about 34 km² (Ambrosi and Crosta, 2006). The Bormio cluster includes about 25 DSGSDs aligned WNW–ESE south of the Zêbrü Line. The Mustair-Venosta cluster consists of about 30 DSGSDs occurring near two major structural features, i.e. the Schling and the Gallo Faults. Clustering of DSGSDs suggests that their development is constrained by some combinations of local conditions. Thus, we analysed the distribution of DSGSD versus lithology, local relief, active seismicity, and the present-day rate of uplift.

The distribution of DSGSDs with respect to lithology was investigated by intersecting the DSGSD map with a reclassified version of the 1:500,000 Geological Map of Switzerland (Swisstopo, 2007; Fig. 4a). Rock types were grouped according to their typical geomechanical behaviour (i.e. average rock mass properties and anisotropy; Bieniawski, 1989; Hoek and Brown, 1997) into: granitoid, metabasite, orthogneiss, metapelite (mainly phyllite), paragneiss, marble, carbonate and terrigenous (mainly sandstone and marl) sedimentary rock. Quaternary deposits, mapped only when occurring in large accumulations (e.g. alluvial valley fill, thick glacial cover), were considered separately (Fig. 5b). Notwithstanding the uncertainties due to map scale, about 68% of the mapped DSGSD occur in metapelite or paragneiss, with a DSGSD density up to 15% (Fig. 5b). Sandstone and marl are also significantly affected by DSGSD (density up to 13%), although their exposure is here limited. Carbonates and granitoids are poorly affected by DSGSD, whereas metabasite and marble are involved only when interlayered in metapelite and paragneiss.

The distribution of local relief in DSGSD areas was analysed using the SRTM digital elevation model (cell size ~90 m). Local relief (Fig. 4b) was calculated as the difference between the minimum and maximum elevation within circular windows with a radius of 5 km for each grid cell for the studied area. The size of the counting circle accounts for the wavelength of major valley spacing and allows...
comparison to other studies (Kühni and Pfiffner, 2001; Montgomery and Brandon, 2002).

In the study area, DSGSDs are associated with average local relief frequently exceeding 1500–2000 m (Fig. 5c). The minimum local relief detected in DSGSD areas also exceeds 1000 m (Fig. 5c).

The distribution of DSGSDs was also compared to the pattern of the present-day rate of rock uplift (Kahle et al., 1997). Although the study area is located at the boundary of their dataset, available data suggest that the entire area is undergoing a high uplift rate (>1.2 mm/yr), with a maximum north of Bormio (about 1.4 mm/yr), where recent earthquakes also cluster (Fig. 2). In the study area, the absolute frequency of DSGSD shows a positive correlation to uplift rate (Fig. 5h).

4. Detailed mapping of morpho-structural features

4.1. Mapping and classification

A detailed photogeological survey was carried out over about 750 km², including the Upper Valtellina (600 km²) and Val Venosta (150 km²). We mapped both tectonic lineaments and morpho-structural evidence of gravitational processes related to more than 30 DSGSDs. We used black and white (1:33,000 scale, 1954; 1:20,000 scale, 1983) and colour aerial photos (1:20,000 scale, 1980–82), and colour orthophotos (2000). Lineaments have been classified as either tectonic or gravitational according to a variety of morphological criteria (McCalpin, 1999; Persaud and Pfiffner, 2004) including: persistence, plan shape, continuity across major ridges or valleys, relationships with major tectonic structures (e.g. nappe boundaries, thrusts, faults), association to other tectonic structures or gravitational features, kinematics inferred by different indicators (e.g. relations with glacial, periglacial and alluvial landforms and deposits), and amount of displacement (when possible).

Tectonic lineaments (i.e. master fractures, faults) are generally persistent and rectilinear, with low associated topographic relief. These features strongly condition the drainage pattern and the localization of minor rock slope instabilities (e.g. rock falls, block slides). Recent tectonic features cross major topographic features as ridges or watershed divides. Postglacial faults appear as solid outcrop zones, crossing glacial landforms or deposits, and can even occur across slopes.

Gravitational morpho-structures are related to DSGSD phenomena (Figs. 4, 6–8), and include double-crested ridges, scars, counterscarps, gravitational half-grabens with associated block tilting, and open trenches.
These structures often show a trend similar to that of tectonic features, but are commonly less continuous, arcuate, and occur as swarms of multiple associated scarps. These can have individual offsets up to several tens of meters, and are usually confined in specific slope sectors showing evidence of dislocation with respect to neighbouring slopes.

4.2. Tectonic vs. gravitational morpho-structures

4.2.1. Upper Val Venosta: the Resia Pass area

The photogeological survey along the Upper Val Venosta covers the area from Glorenza to the Resia Pass (Fig. 7). Over 1500 lineaments were mapped, including fractures as well as scarps, counterscarps and trenches related to DSGSD.

The main fracture system trends NE–SW, i.e. the same trend as the NE–SW nodal planes of seismic events (Fig. 3). Several valleys also show this trend (e.g. Mustair, Zerzer, Roja, Planol, and Plawenn valleys; see Fig. 7). These fractures cross nappe boundaries (e.g. Schlínig fault) and bound the NW margin of the Permo-Mesozoic massif of Cima di Termine (Vallunga Line). Nevertheless, the main morphological features (e.g. the Upper Val Venosta) follow N–S and NNW–SSE, trends, which are less evident from the analysis of fracture lineaments (Fig. 7). N–S and NNW–SSE fractures occur along both valley flanks, along the double ridge connecting Cima 10 and Cima 11, where a possibly active fault has been detected, east of Curon Venosta and S. Valentino alla Muta up to Plan dei Morti (Fig. 7). Here N–S trending normal faults were recognised. NW–SE fractures are also common in the area, especially around Malles Venosta and in the upper Schlínig valley. E–W fractures locally occur, i.e. west of Lago della Muta and south of Cima Pian del Lago.

Postglacial gravitational reactivations of NE–SW, N–S, NNW–SSE, and E–W fractures are widespread, in particular on the western flank of the Upper Val Venosta (Fig. 7). Double and multiple ridges, scarps
Fig. 8. Comparison between (a) tectonic lineaments and (b) DSGSD morpho-structures along the upper part of Valtellina south of Bormio (area 3 in Fig. 1). Same symbols as in Fig. 7.
and counterscarps related to DSGSDs are particularly frequent along the Cima 10–11 ridge, where they trend N–S, and along the Mt. Watles ridge, where NE–SW trending morpho-structures occur. The Cima 11 double-crested ridge suggests a downthrow of about 100 m (Fig. 6a). NW–SE trending DSGSD morpho-structures occur in the Schling valley along the Mt. Watles and Mt. Rodes ridge. Less frequent E–W to ENE–WSW trending morpho-structures occur above Laudes and along the northern slope of the Zerzer valley (Fig. 7). DSGSDs are not as significant along the eastern flank of the Upper Val Venosta, where the Permo-Triassic sedimentary cover crops out, and where lineaments are generally less persistent.

4.2.2. Upper Valtellina

The investigated sector of the Upper Valtellina (Fig. 8) is bounded to the north by the Italian–Swiss border, to the east by the Gavia Pass area, to the south by the upper Val Grosina and to the west by the Foscagno Pass area. Near Bormio, the valley is cut more than 2000 m deep among mountain peaks up to 3000 m a.s.l., consisting of Late Palaeozoic Sondalo Gabbro, Bormio Phyllites, and Grosina Gneiss units. We mapped over 4000 lineaments and prepared two separate maps for structural features and DSGSD-related gravitational morpho-structures (Fig. 8a and b).

Trends of mapped lineaments (Fig. 8a) are close to those recognised in nearby areas. Frequent and persistent NW–SE fractures occur south of Bormio, whereas NE–SW fractures generally occur more discontinuously. N–S trending fracture lineaments are evident along the Upper Valtellina as along the Upper Val Venosta. E–W trending faults south of Mt. Zandila could be related to the Insubric Line (Figs. 1 and 4), the most important alpine structure occurring a few kilometres to the south. The relationships between NW–SE trending systems and the N–S and NNW–SSE ones may suggest that they are synchronous and related to dextral transtension. NW–SE fractures correspond to dextral strike-slip fault, whereas N–S and NNW–SSE faults are pure normal and normal/oblique faults forming extensional stepovers and duplex structures. These are

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**Fig. 9.** (a) Mesoscopic structural analysis of fault populations related to the oldest stage recognized in the study area. Great circles are faults, slickenside lineations are dots with arrows indicating the sense of motion. Convergent arrows correspond to the horizontal direction of the \( \sigma_1 \) axis; divergent arrows to the horizontal direction of the \( \sigma_3 \) axis. Principal stress eigenvectors are represented as stars with five (\( \sigma_1 \)), four (\( \sigma_2 \)) and three branches (\( \sigma_3 \)). Outline stars represent the solution obtained with the R4DT inversion function (Angelier, 1984), filled stars the solution for INVD (Angelier, 1990). (b) Mesoscopic structural analysis of fault populations related to the youngest stage recognized in the study area. Same symbols as in panel (a).
evident especially between the Foscagno Pass and Mt. Trela (NW sector of Fig. 8a).

In the Bormio area, major gravitational morpho-structures related to DSGSD (Fig. 8a and b) occur along the Valfurva, Valdidentro and Val Viola (Fig. 6e), where they mainly follow NE-SW fracture trends. DSGSDs occur in metamorphic rocks (i.e. phyllite, paragneiss) and are generally characterised by large and complex trenches and aligned downhill facing scarps. Along the high-relief slopes of Upper Valtellina, the occurrence of N-S fractures is frequently associated with DSGSD. NW-SE trending structures also occur, such as a graben-like system of trenches and uphill-facing counterscarps that cross the ridges east of Corno di San Colombano (3050 m a.s.l.). Deformed Holocene rock glaciers and slope deposits suggest recent motions along these structures (e.g. at Ruinon DSGSD in Valfurva, Figs. 6d and 8b; Agliardi et al., 2001). In other cases, rock glaciers lack their source areas (e.g. Migiondo at Sondalo, south of Morignone, Fig. 8b; Guglielmin and Orombelli, 2003; Crosta and Frattini, 2004).

The frequency and length of the mapped linear tectonic features (9692) and DSGSD morpho-structures (6511) were statistically analysed and compared (rose diagrams in Fig. 8). The trends of tectonic and gravitational morpho-structures are very close. In both cases, the more important sets trend NW-SE (N125°–130°) and NE-SW (N35°–45°), whereas the N-S and E-W sets are less significant. The orientations of gravitational features are more scattered and conditioned by topography (i.e. valley trend, drainage pattern).

5. Mesoscopic analysis of fault populations in the central Alps

A detailed field study of several hundred of recent faults was undertaken at 43 survey points in the study area (Fig. 9). Most of the field work was carried out in Val Venosta, whereas in Upper Valtellina very detailed observations were performed in an underground tunnel, where in situ stress was measured by hydrofracturing (Canetta et al., 1992). Paleostress analysis of striated fault surfaces was performed to determine the principal stress directions through iterative (R4DT) and direct inversion (INVD) methods (Angelier, 1984, 1990). Numerical solutions were accepted for values of the average angles between computed maximum shear stress on the fault plane and observed
striations lower than 10°–12°. Larger deviations were only considered in the case of separation of complex fault assemblages. The geometry and relative chronology of slickensides, as well as the geometrical consistency among sets of conjugate faults allowed us to recognise two main brittle tectonic stages.

5.1. NW–SE compression stage

The oldest fault assemblage found in Val Venosta consists of conjugate sets of E–W trending dextral and NNE–SSW to N–S trending sinistral strike-slip faults. Along the Merano–Mules segment of the Periadriatic Line, in the Passiria valley, these faults are often associated with NW–dipping NE–SW reverse dip-slip faults and previous mylonitic shear zones displaying a similar geometry and kinematics (Viola et al., 2001). Stress-tensor determinations indicate a horizontal NW–SE $\sigma_1$ axis associated with a NE–SW $\sigma_3$ horizontal axis. NW–SE normal dip-slip faults have been found in association with strike-slip faults. They can be related to local permutation of the maximum and intermediate stress axes (Fig. 9a). This stage can be related to dextral motion and associated SE-vergent back-thrusting which occurred along a restraining bend of the Insubric Line (North Giudicarie Line and Meran-Mules Line) during the Miocene (Prosser, 2000; Viola et al., 2001).

5.2. N–S compression stage and related E–W extension

Conjugate sets of NNW–SSE trending dextral and NNE–SSW trending sinistral strike-slip faults were recognised at most of the studied sites, associated with N–S trending normal dip-slip faults (Fig. 9b). NNE–SSW sinistral strike-slip faults were also found in association with NNE–SSW normal and normal-sinistral faults (e.g.: Tubre valley). Stress-tensor determinations relative to this stage suggest an E–W horizontal $\sigma_3$ axis, associated either with a horizontal $\sigma_1$ axis (i.e. when strike-slip faults occur), or with a vertical $\sigma_1$ axis (i.e. when N–S trending normal faults dominate).

Both mesoscopic observations and mapped lineaments suggest that the Upper Val Venosta developed along a N–S sinistral strike-slip fault, subsequently reactivated as a normal fault. This may be related to the Glorenza fault, which deeply dissects the mylonitic rocks related to Mesozoic thrusting and normal faulting along the Schling Line. NW–SE and NE–SW conjugated strike-slip faults related to a N–S compression have also been found at several sites indicating the importance of recent tectonic deformation. Mesoscopic data suggest that N–S trending normal-oblique faults are the most recent ones, post-dating motions along the conjugated sets of NNW–SSE dextral and NNE–SSW sinistral strike-slip faults. This tectonic regime, consistent with the present-day seismicity (Fig. 3) and with the left-lateral motion along major tectonic structures (i.e. North Giudicarie Fault, Passiria Fault and Engadine Line, trending between NE–SW and NNE–SSW), has probably been active at least since the Late Miocene (Müller et al., 2001).

Fault analyses were also carried out into an exploratory tunnel at the Premadio power station, near Bormio (Canetta et al., 1992), at a depth of about 200 m within the Bormio Phyllites. The tunnel crosses a N–S trending normal fault, representing the northern branch of an important fault system occurring along the axis of Upper Valtellina (Berra et al., 2002). Here a complex polyphase assemblage has been recognised through the analysis of cross-cutting relationships (Fig. 10a). Two main stages have been distinguished, closely matching the observations made in the nearby Venosta area. The oldest one consists of a NW–SE compression causing the activation of conjugate sets of E–W dextral strike-slip faults and NNE–SSW to N–S striking sinistral faults. For this stage, stress-tensor determinations indicate a horizontal NW–SE $\sigma_1$ stress axis associated with a NE–SW $\sigma_3$ horizontal axis (Fig. 10b). Measured NW–SE normal dip-slip faults can be
6. DSGSD modelling: selected case studies

The mechanisms of DSGSD development were studied in detail and investigated through 2D stress-strain numerical modelling at four sites (Fig. 4). These were selected according to their morpho-structural evidence, interaction with man-made structures, and availability of geomechanical or monitoring data.

6.1. Mt. Watles DSGSD (Upper Val Venosta)

Mt. Watles is located on the western flank of the Upper Val Venosta (“A” in Figs. 11a, 11a), and is delimited by the Adige river to the east, the Schlüng valley to the southwest, and Zerzer valley to the northwest. Mt. Watles slope has a local relief of about 1500 m with average slopes of 20°–30°. The DSGSD (Agliardi et al., 2009) affects an area of about 10 km² within paragneiss with thick orthogneiss and amphibolite layers of the Oetztal nappe (Figs. 1 and 11). This is juxtaposed to the sedimentary cover of the Sesvenna–Campo nappe along the Schlüng fault (Froitzheim et al., 1997), which crosses the western flank of Mt. Watles.

Mt. Watles is crossed by scarps and counterscarps with single offsets up to some tens of metres (Fig. 6c). A link between recent brittle structures (Fig. 7) and gravitational morpho-structures is evident (Agliardi et al., 2009). These consist of huge NE trending ridge-top trenches, scarps and counterscarps, sometimes filled with lacustrine and peat deposits (Fig. 11a). In the upper part of the slope, asymmetrical half graben-like systems of scarps and counterscarps re-activate NE–SW recent fractures containing cataclastic bands and fault gouge (Agliardi et al., 2009). In the middle and lower slope sectors, rockslide scars and accumulations suggest a partial collapse. According to Agliardi et al. (2009), differences in rock rheology below and above the Schlueng Fault could have localised deformations in the metamorphic of the hanging wall, belonging to the Oetztal nappe.

Radiocarbon dating of peat deposits formed along major morpho-structures (Agliardi et al., 2009) demonstrates that slope deformation started during the Lateglacial and continued during the Holocene in some slope sectors. Field evidence includes glacial and periglacial landforms and deposits cut or obliterated by gravitational morpho-structures in the Mt. Watles area and surroundings (e.g. Cima 10, Cima 11, Cima Termine).

6.2. Ruinon DSGSD (Upper Valtellina)

The Ruinon DSGSD (“B” in Figs. 4, 11b) occurs in the Upper Valtellina, east of Bormio along the middle Valfurva, where both valley flanks are affected by large DSGSDs. The area consists of phyllites of the Campo nappe, including isoclinally folded marble and prasinite layers (Agliardi et al., 2001). The DSGSD developed over an area of about 5 km² along a pre-existing deep-seated surface dipping to NW, and reactivated recent NW–ESE and N–S fractures. Displaced marble layers and observed offset along the main DSGSD headscarp suggest a downthrow of about 150 m (Fig. 6d). According to Agliardi et al. (2001), the geometry and displacement pattern of the DSGSD are constrained by NW trending master fractures, continuously crossing most of the northern flank of the Valfurva. Along these fractures other gravitational deformations occur (e.g. Pietra Rossa; Fig. 8b). These fractures, consistent with recent dextral strike-slip faults, have been related to neotectonic activity by Forcella and Orombelli (1984). On the contrary, their recent reactivation is suggested by normal dip-slip gravitational shear failures crossing the Holocene rock glacier of the Cavalleray valley (active between 10 and 6 ka, Fig. 11b). The evolution of NW trending gravitational scarps and half-grabens led to the progressive failure of the lower slope sector during the Holocene (Agliardi et al., 2001). This is indicated by large landslide accumulations and by the occurrence of an active, 30°10 m² complex rockslide at the DSGSD toe (between 1400 m and 2150 m a.s.l.), presently causing major concern (Crosta and Agliardi, 2003). Measurements by the Permanent Scatterers technique (Colesanti et al., 2004) suggest that the entire DSGSD and the confining Corna Rossa area are lowering at a vertical rate of 7–24 mm/yr.

6.3. Val Viola–Bosco del Conte DSGSD (Upper Valtellina)

The Bosco del Conte DSGSD (“C” in Fig. 4) occurs on the southeastern flank of the Val Viola (Figs. 8 and 12), an ENE trending valley linking the Italian–Swiss border to the Upper Valtellina. The most important structural feature in the area is the Mt. Verva–Corno delle Pecore–Dossos Peneglia thrust, stacking the Grosina Unit above the Campo Unit (Corradini et al., 1973). The thrust dips to the SSE with an average dip of 35°–45°, and is associated with a break in the hillslope gradient. Both valley flanks show widespread gravitational morpho-structures including scarps, counterscarps and trenches, mainly reactivating pre-existing NE–SW fractures. N–S, NW–SE and NW–ESE fracture sets are also important. Both valley flanks are also characterised by large landslide accumulations and DSGSDs.

The Bosco del Conte DSGSD extends on about 3 km² between 2515 m a.s.l. and 1550 m a.s.l. (Fig. 6e). The NW slope sector is cut by sharp NE–SW scarps and counterscarps up to 1 km long and up to 50 m high. The N and E slopes down to the confluence of the Cardone and Viola valleys are less steep, without large morpho-structures but with widespread evidence of past slope failures. Along the N slope sector, a large landslide probably up to 100 m thick, interrupts the DSGSD morpho-structures to the east. A smaller complex landslide, 20 to 30 m thick, occurs on the western side of the slope. Counterscarps are usually filled by the abundant blocky debris masking the bedrock. In the upper part of the dome-shaped slope, some scarps formed along dark phyllitic and cataclastic layers. The upslope development of the DSGSD seems to be constrained by the exposure zone of the orthogneiss–phyllite thrust. Some of the mapped morpho-structures cross-cut glacial deposits at 1950 m a.s.l. along the eastern limit of the DSGSD. Deposits related to nivation are also cut by some morpho-structures in the Mt. Verva area, whereas large landslides along the slopes postdate most of the DSGSD movements.

Fig. 11. Geomorphological and morpho-structural maps of the Mt. Watles (a) and Ruinon (b) DSGSD areas (location: “A” and “B” in Fig. 4). A simplified geological sketch of the Mt. Watles area is portrayed in the inset: units 1 (orthogneiss), 2 (amphibolite) and 3 (paragneiss) belong to the Oetztal nappe; units 4 (Triassic carbonates), 5 (Permian-Triassic sandstones) and 6 (paragneiss) belong to the Sesvenna-Campo nappe (modified after Agliardi et al., 2001, and Agliardi et al., 2009).
Fig. 12. Morphostructural analysis of the Val Viola Bormina area and Bosco del Conte DSGSD (location: “C” in Fig. 4). a) Fracture lineaments from photogeological analysis. b) Geomorphological map with the main structures, morpho-structures and landslides.
6.4. Cima di Mandriole DSGSD (Peio)

The Cima di Mandriole DSGSD (“D” in Figs. 4, 13) occurs in the upper Noce valley (Trentino–Alto Adige), few kilometres south-east of the Ruinon area. It extends over about 6 km², involving the Campo nappe metamorphics north of the Peio Line (Fig. 1), a NE trending Cretaceous normal fault separating the Campo from the Tonale nappe (Martin et al., 1991). Rocks include isoclinally folded paragneiss associated with mylonitic orthogneiss, crossed by ENE–WSW and NW–SE master joints and normal to oblique dip-slip faults.

The DSGSD involves the entire slope down to the 53 m high Pian Palù dam (Fig. 13). Heavily fractured rock masses on the left bank of the dam site were observed during its construction (Desio, 1961). Thus, the dam was built using large concrete blocks with friction joints, able to stand large deformations. Cataclastic shear zones with loose coarse gouge were encountered during borehole drilling and tunnel excavations, indicating discrete failure surfaces related to gravitational movements (Fig. 13). These follow a slightly curved trend and terminate in alluvial deposits, which were found below fractured rock at the slope toe (Desio, 1961). In the upper slope, steep head scarps with downthrow exceeding 100 m occur (Fig. 6e), reactivating a NE trending fault. Above 2300 m a.s.l., the slope is crossed by swarms of rectilinear counterscarps up to 200 m long, cross-cutting glacial erosional surfaces and rock glacier deposits. Below 2300 m a.s.l., scarps and counterscarps form gravitational half-grabens. These represent the head of a series of large concentric landslides with morphological evidence increasing downslope. The statistical analysis of tectonic and gravitational lineaments revealed that most tectonic fractures trend NE–SW, whereas E–W and N–S ones are few but persistent. NE–SW fractures clearly constrain the geometry of the DSGSD, showing a trend very close to the slope direction.

Geological and site investigation data (Crosta et al., 2000) suggest that the DSGSD moved over a basal shear zone over 200 m deep, causing the formation of an active wedge in the upper slope. In the

Fig. 13. Geomorphological and morpho-structural map of the Cima di Mandriole DSGSD at Peio. (location: “D” in Fig. 4). Contour interval: 50 m.
lower part of the slope, morpho-structures were activated as large landslides with multiple shear surfaces up to 100 m deep, recognised in boreholes as low core recovery and catastrophic layers. Cross-cutting relations between morpho-structures and glacial/periglacial deposits suggest a Lateglacial activation and further Holocene activity in the upper slope. At the slope toe, displacement monitoring revealed ongoing activity with average rates of 5–6 mm/yr.

6.5. Numerical modelling

2D stress-strain numerical modelling of the analysed DSGSDs was carried out using the explicit finite difference code FLAC (Itasca, 2005), to test different onset mechanisms and constrain the role of inferred controlling factors. Finite difference grids were generated, fitted to reconstructed pre-DSGSD slope profiles and constrained by displacement boundary conditions. Based on test model runs and previous experiences in modelling large rock slope failures (Agliardi et al., 2001; Ambrosi and Crosta, 2006), we assumed a Mohr–Coulomb elasto-plastic yield criterion and used an ubiquitous-joint model (Itasca, 2005) to simulate the anisotropy due to pervasive foliation or closely spaced joints. Major structural features (e.g. nappe boundaries, regional faults, master joints) were also included, either as regions with different geomechanical properties (e.g. Mt. Watles), or as discrete interfaces crossing the finite difference grids (e.g. Cima di Mandriole). Geomechanical properties of rock masses and ubiquitous joints (Table 1) were obtained through standard rock mass characterisation based on field and laboratory data (Bieniawski, 1989; Hoek and Brown, 1997).

All the studied DSGSDs occur in valleys heavily glaciated during the LGM. Thus, stress conditions corresponding to the glacial and pre-glacial evolution of the studied slopes were simulated by sequential modelling including ice loading at LGM glacier stage, subsequent deglaciation and related processes (e.g. changes in groundwater conditions). Assumed LGM ice surface elevations (Fig. 14) were based on literature (Florinseth and Schlüchter, 2000) and on the distribution of mapped glacial landforms and deposits, with total ice thickness ranging between 1200 m and 2000 m for the studied cases. In situ stress was also applied as initial condition when available (e.g. seismotectonic and hydrofracturing data). Moreover, simulations neglecting the effects of glacial loading and unloading were carried out for comparison. Model parameters were calibrated by comparing simulation results (i.e. cumulative displacements, type of failure, distribution of plasticity indicators) to geomorphological and morpho-structural constraints. Surface and deep displacement monitoring data both along the slope and the dam were also considered for the Cima di Mandriole DSGSD.

Numerical models simulating deglaciation provided a good account for the major morpho-structural features recognised and mapped at surface, thus confirming their gravitational origin (Fig. 14). Model results suggest that tensile fracturing parallel to topography, shearing through the slope and slippage along joints are strongly controlled by glacial unloading/rebound, whereas these features were not simulated by “ice free” models. In our simulations, plastic deformation close to the highest ridge initiate soon after the beginning of deglaciation, whereas deep shear zones of differing thicknesses generally develop later from the slope toe toward the crest. The development of tensile fracture zones along the slopes is controlled by the slope profile and the orientation of the ubiquitous joints. Model results also suggest that natural rock slopes subjected to DSGSD remain in a steady-state condition of dynamic equilibrium with accelerations for changes in controlling factors (e.g. groundwater table position). This corresponds to creeping conditions along an already formed failure surface.

According to numerical models, master joints and faults are major constraints on DSGSD onset and geometry especially in gentle slopes, thus confirming the strong passive structural control suggested by field investigations. The orientation of planes of weakness (i.e. foliation or closely spaced joints) also significantly controls the development of DSGSD. In fact, no large deformations developed in models when ubiquitous joints were not included. Where ubiquitous-joint orientation is favourable to slope instability, failure requires little internal deformation (e.g. Watles). Although brittle deformation is associated to single morpho-structures, at depth the rock mass experiences plastic deformation along subcircular or composite shear zones. These are often associated with anisotropic shear zones emerging at different elevations, bounding active wedges or half-graben structures (e.g. Cima di Mandriole, Ruinon, Bosco del Conte). This geometry was also observed by other authors through different modelling approaches (e.g. analytical solutions: Savage and Smith, 1986; physical models: Chemenda et al., 2005).

7. Discussion

Our multidisciplinary analyses considered the relationships among tectonic features, active tectonic processes, surface processes, topography, and the development of DSGSD.

We showed that the occurrence of major structural features (i.e. fractures, master joints) is essential for explaining the distribution and general features of DSGSD as well as for modelling their onset. This is confirmed by numerical models, unable to simulate the observed DSGSD deformation patterns if these structural and geological elements are neglected (e.g. Watles, Cima di Mandriole DSGSD). In the central Eastern Alps DSGSD develop in anisotropic, moderately strong fractured rock masses, mainly schistose metamorphic rocks cut by recent NW–SE, N–S, and NE–SW recent fracture systems. DSGSD occurs preferentially along high relief (>1000 m), structurally-controlled valley sides overdeepened by glacial erosion. Main trends of DSGSD morpho-structures always closely fit the orientation of recent brittle structures, thus confirming the significant structural control on their development pointed out by several authors (Radicchi-Hall, 1978; Agliardi et al., 2001; Hippolyte et al., 2006). Tectonic structures passively control the kinematic degrees of freedom in rock masses and influence rock mass strength, thus constraining DSGSD localisation, geometry, and morpho-structural pattern (Agliardi et al., 2001). Morpho-structures trending close to slope direction are generally more developed, suggesting a primary control of topographic stress on the activation of existing tectonic features. This is evident in upper Valtellina, where DSGSD features mainly trend N–S and NW–SE, as well as in the upper Val Venosta, where NE–SW and N–S trends dominate.

Table 1

Summary of average values of the geomechanical properties adopted in the numerical modelling of DSGSD case studies.

<table>
<thead>
<tr>
<th>DSGSD</th>
<th>Rock mass</th>
<th>Ubiquitous joints</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mass density</td>
<td>Bulk modulus</td>
</tr>
<tr>
<td></td>
<td>kN/m³</td>
<td>GPa</td>
</tr>
<tr>
<td>Mt. Watles (A)</td>
<td>27</td>
<td>3</td>
</tr>
<tr>
<td>Ruinon (B)</td>
<td>27</td>
<td>10</td>
</tr>
<tr>
<td>Bosco del Conte</td>
<td>27</td>
<td>7</td>
</tr>
<tr>
<td>Cima di Mandriole (D)</td>
<td>27</td>
<td>10</td>
</tr>
</tbody>
</table>
The consistency between fault kinematics inferred by mesoscopic analysis and fault plane solutions of earthquakes suggests that N–S trending normal dip-slip faults and conjugate NE–SW sinistral and NW–SE dextral strike-slip faults could be active in the central Eastern Alps. Forcella and Orombelli (1984) suggested that WNW–ESE scarps and trenches, related to DSGSD in Valfurva (Bormio DSGSD cluster; Figs. 8 and 11), could be surface expressions of neotectonic faults. On the other hand, Agliardi et al. (2009) suggested that the occurrence of fault systems kinematically related to the active stress field in the central Eastern Alps (Müller et al., 1992; Mariucci and Müller, 2003) may contribute to explain the widespread distribution of DSGSD in the Resia Pass area (Mustair-Venosta DSGSD cluster). We confirm a widespread gravitational reactivation of possibly active brittle structures over a larger area. This could suggest possible active controls by ongoing tectonic processes on DSGSD onset, i.e. by active fault slip, fault zone weakening, in situ stress or seismic shaking. The high values of present-day rock uplift (>1 mm/yr) and local relief (often exceeding 1500–2000 m) associated with DSGSD also suggest a possible tectonic forcing of DSGSD development. According to Montgomery and Brandon (2002), these relief values are typically

Fig. 14. Sketches comparing the results of field morpho-structural study and numerical modelling for the case-study areas. Left: interpretative cross sections with the main tectonic and morpho-structural features. Right: model results in terms of computed maximum shear strain increment, outlining strain localisation and shear band development in slopes subjected to postglacial unloading. Local LGM ice surface elevations after Florineth and Schlüchter (2000).
found in tectonically active areas, where they are associated with landscape evolution by large landsliding processes. Nevertheless, the distinction between tectonically active structures and fractures reactivated by DSGSD is not obvious. In most cases gravity acts at faster long-term average strain rates than neotectonic faulting, masking the occurrence of active fractures moving at low displacement rates (0.1 mm/yr of N–S shortening in the Alps; Westaway, 1992). Radbruch-Hall (1978) introduced a range for the measured rates of large-scale mass rock creep (from 2 cm/yr to 20 cm/d). Varnes et al. (1999) and Bovis (1982) report displacement rates ranging respectively between 1.4 and 5.0 mm/yr, and between 0.14 and 0.75 mm/yr. Colesanti et al. (2004) and Ambrosi and Crosta (2006) report vertical displacement rates ranging between 7 and 30 mm/yr for DSGSD areas in the Central Italian Alps. These rates are much higher than for any observed active fault in the area. This seems to confirm that many scars and countercarps in alpine environments must be considered the result of gravitational phenomena rather than of neotectonics and seismic activity.

In the central Eastern Alps, displacement rates of DSGSD are also much higher than present-day rock uplift rates measured by precise levelling networks (i.e. up to 1.4 mm/yr in the study area; Kahle et al., 1997). DSGSD rates also greatly exceed erosion rates for the Alps estimated by Hinderer (2001), with average values since the Late al., 1997). DSGSD rates also greatly exceed erosion rates for the Alps. Continuing changes in elevation along large slope sectors related to DSGSD can result in a significant focused lowering of topography over a time scale of $10^3$–$10^4$ years. At the aforementioned displacement rates, and assuming steady-state DSGSD, this could locally sum up to 100–300 m since the Lateglacial. Since DSGSDs are widespread (e.g. about 10% of the study area), they could significantly influence erosion rates and sediment yield at catchment scale in areas with high relief (also a proxy of erosion; Ahnert, 1970), especially when glacial and fluvioglacial erosion/deposition processes are exhausted. DSGSD control on erosion may include: exhuming bedrock by reactivation of tectonic faults (Hippolyte et al., 2006), damming or modifying drainage networks both on slopes and valley floors, supplying sediments to stream erosion at different timescale (e.g. from catastrophic slope failure to slow toe bulging).

It has been shown that large DSGSDs occur in areas that underwent glaciation and glaciation, as well as in unglaciated areas (Crosta and Zanchi, 2000). Nevertheless, most DSGSDs occur in glacial valleys and that processes related to deglaciation (e.g. slope unloading, ground-water level fluctuations and water pressure increase because of permafrost conditions, ice melting and fracture unloading) are required for large displacements to occur in numerical models. This agree with previous findings (Brückl, 2001; Agliardi et al., 2001; Ballantyne, 2002). According to the available data, maximum ice surface elevation during LGM in the studied areas was ranging between 2500 and 3000 m a.s.l. (Florineth and Schlüchter, 2000), thus leading to a significant debuttressing of slopes during deglaciation. Postglacial DSGSD triggering or evolution is also suggested by morpho-structures cross-cutting glacial and periglacial deposits, and confirmed by radiocarbon dating of peat deposits sealing Mt. Watles DSGSD features (Agliardi et al., 2009).

Meigs and Sauber (2000) presented the case of a very large instability with features similar to those of DSGSD. This developed in a 13 years period following rapid glacial retreat, with net down-dip displacement at the base of the landslide of about 300–400 m. This example suggests that the displacement rate can be higher just after deglaciation. Radiocarbon dating (Agliardi et al., 2009) also demonstrates that sudden activation and largest movements occurred just after LGM collapse (18–16 ka; Lambeck et al., 2002; Ivy-Ochs et al., 2004; Kelly, 2006). Nevertheless, available monitoring data (Colesanti et al., 2004) suggest that after an initial accelerating stage, slopes movements keep on at slow rates. This is in agreement with the results of Cruden and Hu (1993) and Prager et al. (2008), which suggested that the period of paraglacial slope adjustment (Ballantyne, 2002) could span up to thousand years. Finally, the occurrence of continuing slope movements over long time periods could also suggest a sort of ongoing positive feedback between slope instability, uplift, and valley deepening. In fact, high uplift rates enhance valley deepening, which in turn causes stress concentration at the slope toe, resulting in an increased slope activity and further fracturing.

8. Conclusions

DSGSDs are a common feature in the postglacial morphological evolution of the central Eastern Alps, involving more than 10% of the present surface and thus huge rock volumes. Their extent suggests a complex interplay with relief. Thus, significant controls on erosion processes at different timescale can be ascribed to DSGSD, which can remain active for long time periods at low displacement rates. Regional clustering of DSGSD suggests that, in general, they are more frequent when some combination of local factors occurs, including strongly anisotropic rock masses, local relief > 1000 m, and rates of rock uplift > 1 mm/yr. This also suggests possible links between enhanced relief production forced by tectonics (Montgomery and Brandon, 2002) and the onset of DSGSD.

Morpho-structures related to DSGSD in the area strictly follow the trends of recent NW–SE and N–S fractures in Valtellina and by N–S to NE–SW ones in the Upper Val Venosta. This confirms the strong passive structural control on DSGSD already outlined by some authors. The kinematic consistency between these structures and the pattern of seismicity also suggests that possibly active brittle structures are subjected to widespread gravitational reactivation. This possibly suggests active tectonic forcing of DSGSD, also supported by the recorded values of uplift and local relief.

Finally, numerical modelling suggests that deglaciation following LGM collapse and related paraglacial processes are the major triggers of DSGSD. The onset of these phenomena needs some pre-existing structural features (e.g. pervasive foliation, “Alpine” low-angle mylonitic to cataclastic fault zones, recent steep to vertical fractures) and geological conditions (e.g. occurrence of paraderivatives), especially when related to active fault systems.

Acknowledgements

The authors are indebted to C. Ambrosi, D. Bassanelli, and M. Villa for assistance in the collection and analysis of data, A. Azzoni and A. Frassoni of ISMES, and ENEL Hydro S.p.A. and Provincia Autonoma di Trento for providing field testing and monitoring data. The paper benefited from the careful reviews of M. Jaboyedoff and O. Korup. The research has been partially funded by a FIRB (“Multi-disciplinary approach for large landslides hazard assessment”) and a MIUR project (“Catastrophic slope failures: characterisation, monitoring and modelling for hazard assessment”).

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