Available online at www.sciencedirect.com
SCIENCE DIRECT®
Tectonophysics 418 (2006) 255–276
TECTONOPHYSICS
www.elsevier.com/locate/tecto

The recent fault scarps of the Western Alps (France): Tectonic surface ruptures or gravitational sackung scarps? A combined mapping, geomorphic, levelling, and ^{10}Be dating approach

Jean-Claude Hippolyte [a,*], Gilles Brocard [b], Marc Tardy [a], Gérard Nicoud [c], Didier Bourlès [d], Régis Braucher [d], Gilles Ménard [a], Blaise Souffaché [e]

[a] *Laboratoire de Géodynamique des Chaînes Alpines (LGCA) UMR CNRS 5025, Université de Savoie—Campus Scientifique Technolac, 73376 Le Bourget du Lac Cedex, France*
[b] *Department of Geology and Geophysics, University of Minnesota, 310 Pillsbury Drive SE, Minneapolis, MN 55455-0129, USA*
[c] *Laboratoire Environnements, Dynamiques et Territoires de la Montagne (EDYTEM) UMR CNRS 5204, Université de Savoie—73376 Le Bourget du Lac Cedex, France*
[d] *Centre Européen de Recherche et d'Enseignement des Géosciences de l'Environnement (CEREGE) UMR CNRS 6635, Université Aix-Marseille 3, 13545 Aix en Provence, France*
[e] *Sisyphe, UMR CNRS 7619, Université Paris 6, 4 place Jussieu, 75252 PARIS CEDEX 05, France*

Received 30 June 2005; received in revised form 17 February 2006; accepted 24 February 2006
Available online 19 April 2006

Abstract

In the Western Alps, some recent scarps were previously interpreted as surface ruptures of tectonic reverse and normal faults that agree with microseismicity and GPS measurements. Our analysis shows that in fact there are hundreds of recent scarps, up to 30 m high and 2.1 km long, with only pure normal motions. They share the same characteristics as typical sackung scarps. The scarps are mainly uphill facing, parallel to the ridge crests and the contour lines. They are relatively short (less than 2.1 km) with respect to tectonic fault ruptures, and organized in swarms. They cut screes and relict rock glaciers with a slow (commonly 1 mm/year) average slip rate. In the Aiguilles Grives massif these sackung scarps clearly express the gravitational toppling of sub-vertical bedding planes in hard rocks. In contrast, the Belledonne Outer Crystalline Massif exhibits scarps that stem from the gravitational reactivation of conjugate tectonic faults. The recent faults extend to about 1600 m beneath the Rognier ridge crest, but are always above the valley floor. The main scarp swarm is 9.2 km long and constitutes the largest sackung ever described in the Western Alps. ^{10}Be dating of a scarp and offset surfaces shows that >4 m slip may have occurred rapidly (in less than 3800 years) sometimes between the end of the glaciation and 8800 ± 1900 years ago. This dating, together with the location of some faults far from the deep glacial valleys, suggests that sagging might have been triggered by strong earthquakes during a post-glacial period of probably enhanced seismicity. The Belledonne and Synclinal Median faults (just beneath the Rognier sackung) could have been the sources of this seismicity.

Keywords: Western Alps; Quaternary fault scarps; Active tectonics; Gravitational movements; Sackung; Geomorphology; Deep-seated gravitational deformation

* Corresponding author. Tel.: +33 4 79 75 87 15; fax: +33 4 79 75 87 77.
E-mail addresses: jean-claude.hippolyte@univ-savoie.fr (J.-C. Hippolyte), broca002@umn.edu (G. Brocard), marc.tardy@univ-savoie.fr (M. Tardy), nicoud@univ-savoie.fr (G. Nicoud), bourles@cerege.fr (D. Bourlès), braucher@cerege.fr (R. Braucher), gilles.menard@univ-savoie.fr (G. Ménard), Blaise.Souffache@ccr.jussieu.fr (B. Souffaché).

0040-1951/$ - see front matter
doi:10.1016/j.tecto.2006.02.009

1. Introduction

In Savoy (northern French Alps), recent fault scarps have been reported in two main areas (Fig. 1): first at Les Arcs ski resort within the inner alpine units (Goguel, 1969), and second at the Rognier Mountain in the Belledonne Outer Crystalline Massif (Bordet, 1970). These scarps were interpreted as resulting from movements on active faults. The faults of Les Arcs area would be normal (Goguel, 1969). The faults of the Rognier Mountain would be reverse (Bordet, 1970), but these were only observed from helicopter. In a critical reappraisal of potential active faults in SE France, Sebrier et al. (1997) proposed that these scarps could result either from gravitational processes such as sackung or from faults triggered by post glacial rebound. However, in a more recent synthesis Baize et al. (2002) concluded that the faults of the Rognier Mountain are active sinistral strike-slip faults, with a minor normal component.

Recent advances in seismotectonics (Sue et al., 1999) and permanent GPS networks (Calais et al., 2002) indicate that the present deformation in the inner alpine units is mainly extensional (Fig. 1). This new vision of the active deformation of the Alps seems in agreement with the description of normal faults scarps in the inner Alps (Goguel, 1969), and reverse or strike-slip faults scarps in the outer Alps (Bordet, 1970; Baize et al., 2002). This agreement motivated a detailed study of these two sites. We present the results of our geomorphic analysis, and a mapping of the recent scarps with the discovery of many unknown scarps in particular in the Belledonne massif. This analysis is completed by a dating of one of these recent fault scarps using the ^{10}Be time exposition method. Our multidisciplinary approach allows us to discriminate between the different interpretations: active tectonic faults, sackung faults, or combined processes.

2. Possible origins of a recent fault scarp in the Alps

In mountainous areas, recent fault scarps can be created either by coseismic tectonic surface rupture, by

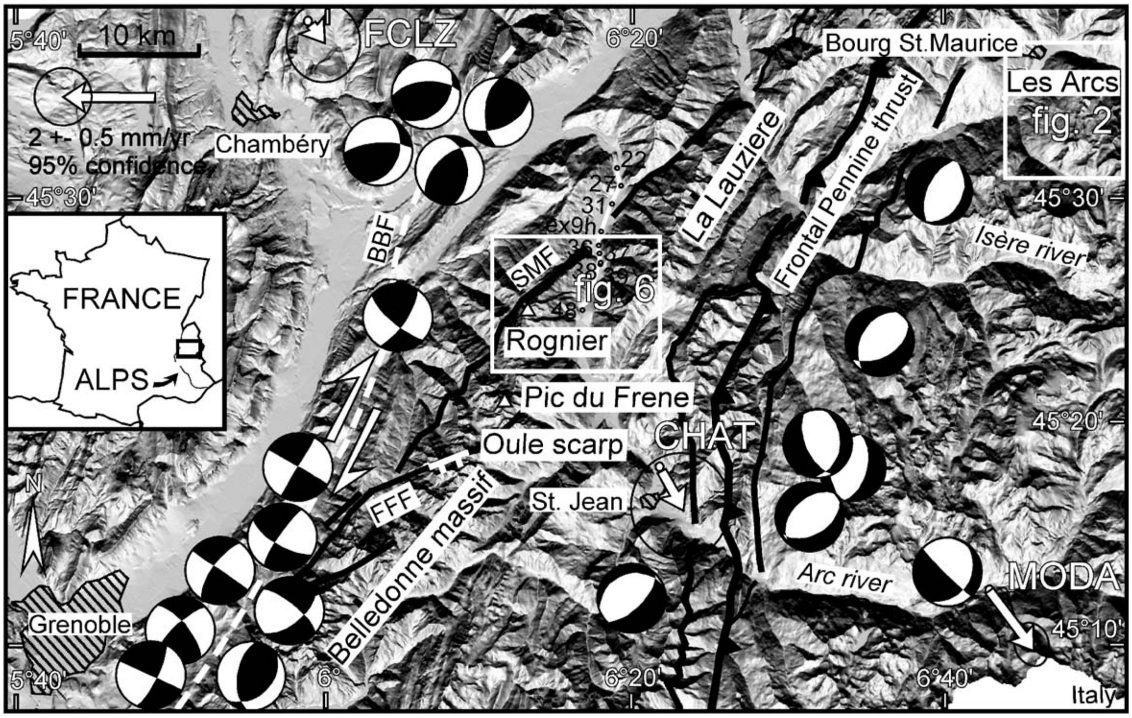

Fig. 1. Active deformation of the Alps of Savoy. GPS vectors relative to stable Europe from continuous measurements at sites located in the Feclaz Mountain (FCLZ), or in the Maurienne Valley (CHAT and MODA) are from Calais et al. (2002). Focal planes mechanisms (magnitudes 1.2 to 2.5) are from Thouvenot et al. (2003) for the Belledonne Border fault, and from Sue et al. (1999) for the inner Alps. BBF: Belledonne Border fault. SMF: Synclinal Median fault. FFF: Fond de France fault. Numbers refers to the levelling benchmarks of the profile of Fig. 12. The outer alpine units (west of the frontal Pennine thrust) are characterized by strike-slip reverse deformation while the inner alpine units (east of the frontal Pennine thrust) are characterized by NW–SE extension and a SE-directed GPS vector at Modane. Recent scarps of reverse faults have been reported at the Rognier (Bordet, 1970) in the outer compressional area. Recent scarps of normal faults have been reported at the Les Arcs (Goguel, 1969) in the inner extensional area.

post-glacial isostatic rebound, or by deep-seated gravitational spreading (sackung). The latter scarps are often large gravitational faults that can be mistaken with tectonic faults (cf. McCalpin, 2003). It is even possible that fault scarps result from a combination of these tectonic and gravitational forces. If a recent fault-scarp results only from tectonics, its kinematics must be consistent with the present regional crustal deformation.

2.1. Current tectonic deformation of the Western Alps

The present deformation of the Western Alps was often considered as the continuation of Miocene to early Pliocene crustal shortening. More recent analyses have revealed widespread extension in the inner part of the Western Alps. Effectively, focal plane mechanisms show the coexistence of radial extension characterizing the inner zones, and transpression characterizing the front of the belt and the innermost area (Fig. 1) (Frechet and Pavoni, 1979; Pavoni, 1980; Eva et al., 1997; Sue et al., 1999). Continuous GPS measurements over 2.5 to 5 years agree with the seismotectonic analyses and show clear southeast-directed velocities (1.7 ± 0.4 mm/year) at Modane (inner Alps, Fig. 1) relative to stable Europe and to the GPS station of Saint Jean des Vignes in the Alpine foreland (Calais et al., 2002).

Interpretation of fission tracks ages (Seward and Mancktelow, 1994) and fault plane striations (Sue and Tricart, 2002) suggest that extension mainly occurs as the normal reactivation of the Frontal Pennine thrust (Fig. 1), the external boundary of the inner Alps. However, no clear geomorphologic observations along the trace of this fault or of neighboring faults have been found to confirm its activity.

The recent fault scarps reported by Goguel (1969) and Bordet (1970) are located successively just east and west of the Frontal Pennine thrust (Fig. 1). Thefore they could represent a surface expression of the active tectonic deformation. But, in this area of high relief they could also represent large-scale slow gravitational deformation known in the geological literature as “sackung” (Zischinsky, 1966, 1969; Radbruch-Hall et al., 1976; Savage and Varnes, 1987; McCalpin and Irvine, 1995).

2.2. “Sackung” deep-seated gravitational deformation

The term “sackung”, from the German word for sagging, was first proposed for the surface manifestation of deep-seated rock creep in foliated bedrocks of the Alps (Zischinsky, 1966, 1969). In the definition example, deep-seated rock creep had produced a ridge-top trench by gradual settlement of a slablike mass into an adjacent valley (Zischinsky, 1969).

This term is now used to refer to the process of deep-seated sagging (e.g. Savage and Varnes, 1987). Following McCalpin and Irvine (1995), the term “sackung” denotes slope sagging, gravitational spreading of bedrock ridges or deep-seated gravitational slope deformation revealed by linear features in mountainous landscapes; the plurial forms “sackungs” or “sackungen” refer to the surface forms created by that process. Thus, “sackungen” is a collection of linear geomorphic features of gravitational spreading origin including: anomalous ridge-top troughs, double-crested ridges (Paschinger, 1928), uphill-facing scarps (also termed antislope scarps, Radbruch-Hall et al., 1977), downhill-facing scarps, tension fissures, closed depressions, and linear troughs.

The main difference between sackung and a classical landslide is the long term slowness of the process (average spreading rates across multiple sackung scarps of 0.4–5 mm/year, Varnes et al., 1990) together with the large size of the structure in movement, the often deep-seated character of the failure, the absence of well-defined boundaries, and the typical absence of compressional morphology at the toe (McCalpin and Irvine, 1995).

“Sackugen” are found dominantly along ridge crests in massive competent rocks (granite, quartzites, volcanic rocks, etc.). They trend roughly parallel to slope contours and to the strike of bedding, foliation, joints or faults. They can create giant steps in the landscape. Typical sackung scarps range in length from 15 to 500 m and in height from 1 to 10 m (Bovis, 1982; Bovis and Evans, 1995; McCalpin and Irvine, 1995; Jarman and Ballantyne, 2002; Schwab and Kirk, 2002).

While sackung faults have long been recognized in most of the Alps and in the Carpathians (Jahn, 1964; Zischinsky, 1966; Nemcok, 1972; Mahr, 1977; Forcella, 1984; Onida et al., 2001), in the French Alps only a few cases of slow slope deformations are studied. The most famous are the Séchilienne “landslides” near Grenoble (Fig. 1; Vengeon, 1998), and La Clapière “landslide” to the north of Nice (Follacci, 1987; Vibert, 1987; Guglielmi et al., 2000; Bigot-Cormier et al., 2005) with little linear features in the same valley interpreted as “sackungen” (Flageollet, 1989).

2.3. Post glacial rebound deformation

In former glaciated regions, recent fault scarps can also be created by crustal movements occurring during deglaciation in response to crustal unloading. Holocene

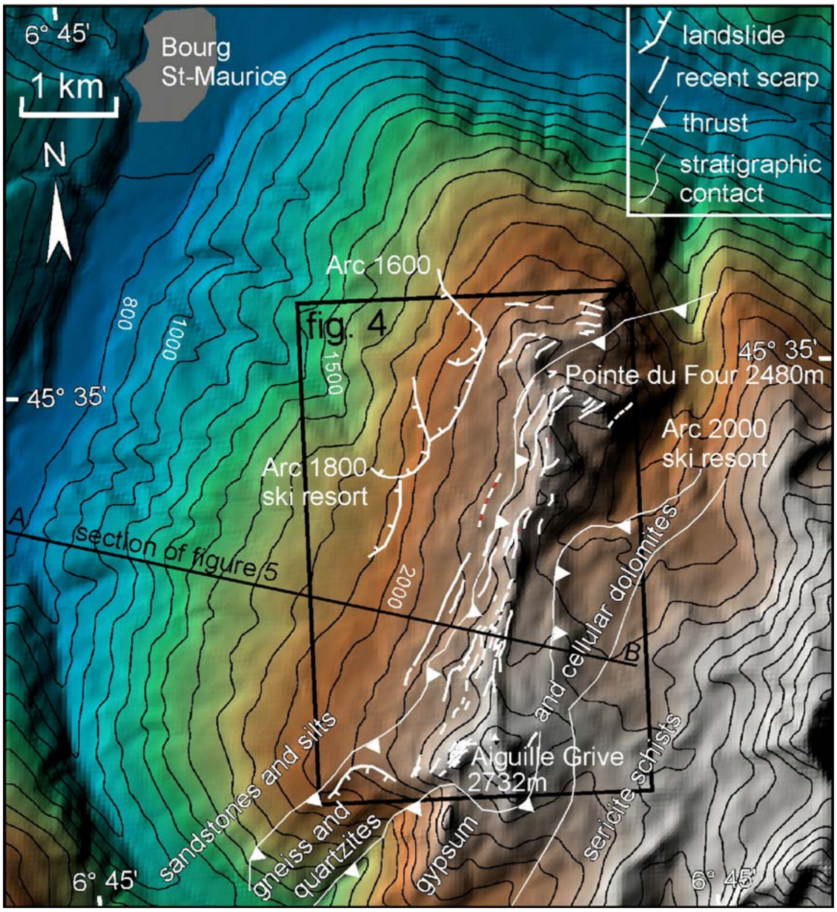

Fig. 2. The recent scarps of Les Arcs ski area (Aiguille Grive massif) and their relationships with the elevation contour lines and the geology. Most of the scarps are dipping uphill (cf. the detailed mapping of Fig. 4).

shoreline dislocations or post-glacial fault scarps interpreted as resulting from post-glacial isostatic rebound have been reported in particular in Sweden (e.g. the Parvie fault with a vertical displacement of 10 m and a rupture length of 150 km, Muir-Wood, 1989), in Ireland (Mohr, 1986; Knight, 1999) and in the Scottish Highlands (Fenton, 1991; Firth and Stewart, 2000; Stewart et al., 2000; Jarman and Ballantyne, 2002). In compressional settings, deglaciation induces a rapid decrease in the vertical stress and facilitates reactivation of palaeotectonic faults (Muir-Wood, 1989). When fault scarps are parallel to the major glaciated valleys and their kinematics indicates that the valley is uplifted with respect to the surrounding mountains ranges, the scarps were supposed to form by differential glacial isostatic rebound between the valley sides and its floor that supported a larger thickness of ice sheet (Mollard, 1977; Ego et al., 1996; Sébrier et al., 1997). Finally, it is probable that owing to changes in stress, earthquake activity was increased during the post-glacial rebound (Gregersen and Basham, 1989; Stewart et al., 2000).

3. Recent scarps in the Aiguille Grive massif

3.1. Previous works

In 1969, Goguel discovered recent fault scarps offsetting mountain slopes in the Aiguille Grive massif, which is now also the domain of Les Arcs ski resort, near Bourg St Maurice (Figs. 1 and 2). The Aiguille Grive massif is situated in the inner (Briançonnais) Alpine units, close to the Frontal Pennine thrust (Fig. 1). It is composed of sub-vertical N–20–E trending formations including from west to east (Fig. 2): a thick sequence of Late Carboniferous “houiller” silts and sandstones, a thrust contact dipping to the East, a local thin gneiss slice, thick white Permo-Triassic quartzites that form the crest of the ridge and which are overthrusted on the east-facing slope by Triassic cellular dolomites, gypsum and micaschists (Antoine et al., 1993).

Goguel (1969) observed and mapped several linear scarp morphologies in the Carboniferous and Permo-Triassic formations in a 700 m wide and 3.5 km long strip (Fig. 2). This author pointed out that some of these linear structures deflect gullies (Fig. 3a) and cut and offset “block accumulations”, which are in fact inactive rock glaciers (Fig. 4). He concluded from the vertical offset of 4–5 m of mountain slopes and from the

Fig. 3. Views of recent scarps (indicated by white arrows) at Les Arcs ski resort. Refer to Fig. 4 for the location of the photos and the scarps numbered. Numbers in meters indicate scarp height or trough width (V_t = vertical throw). We classify these scarps in: antislope fault scarps (figures a, e, g and h), ridgetop troughs (figures c, d and i), and tensional antislope scarps (figures f and j) depending on their mechanism and their location on the ridge (cf. text and Fig. 5). Note on photo b that the antislope fault scarps # 6 is recent because it cuts the screes, and still separates inactive lower screes partly covered by grass, from active upper screes without vegetation.

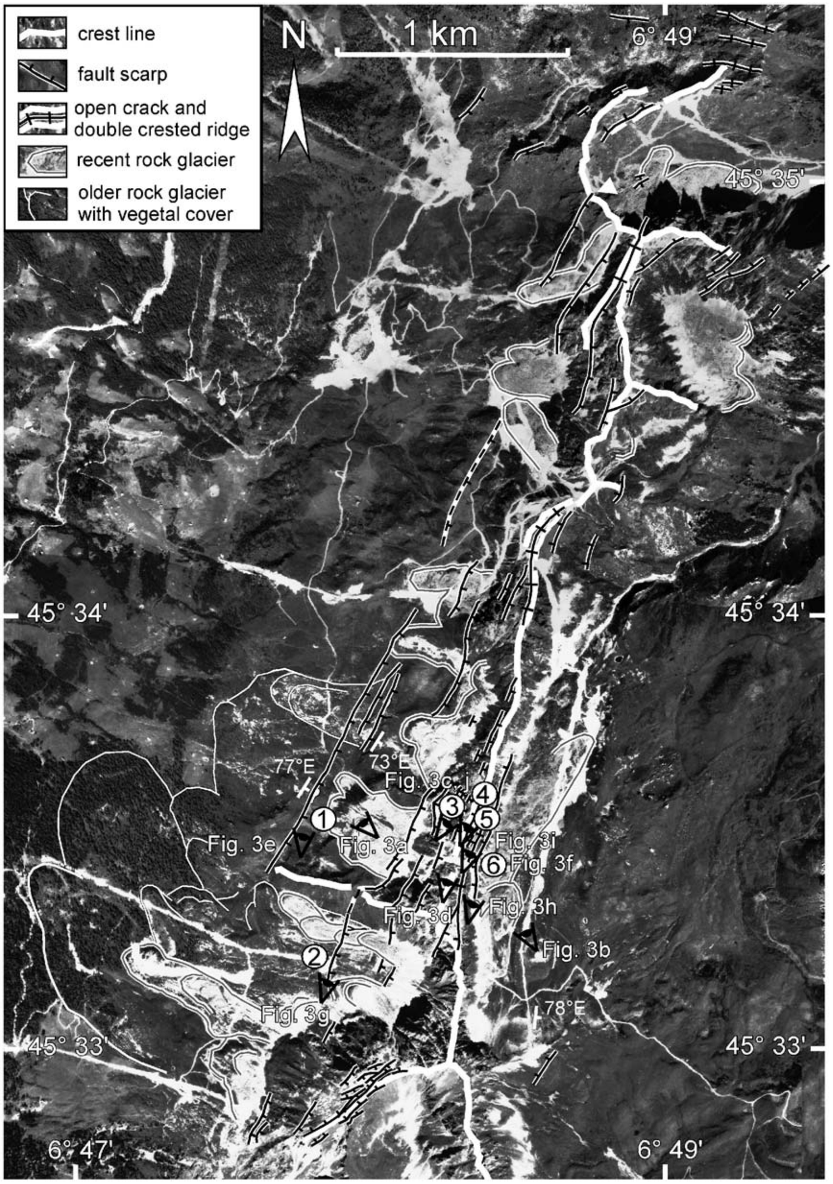

Fig. 4. Map of the recent scarps and rock glaciers of Les Arcs ski resort on orthorectified aerial photos.

apparent east-dip of the lower fault scarp (Fig. 3a) that these morphologies denote the activity of recent, possibly seismic, extensional faults. Note that if they were active faults, their NNE-trend would agree with the present SE-extension of the inner Alps.

3.2. Morphology and origin of the scarps

To check Goguel's conclusions, we have mapped the linear recent morphological scarps of the Aiguille Grive massif, on orthorectified aerial photos (Fig. 4). We found about 80 recent scarp for 5.3 km along the ridge crest (Figs. 2 and 4). Moreover, we made geomorphologic observations that suggest a new interpretation for the origin of these scarps.

We could distinguish two main types of scarp morphologies: (1) uphill-facing fault scarps, that apparently offset the mountain slopes in a normal sense (Fig. 3a, b, e, g); and (2) open cracks or troughs (Fig. 3c, i, j).

The uphill facing scarps are present on the two slopes of the range (Fig. 4). They trend and dip parallel to the bedding planes of the Carboniferous sandstone and the Permo-Triassic quartzites (Figs. 2 and 4) suggesting that they could represent bedding plane faults. Their movement is always pure dip-slip. Except a few downhill facing scarps in the southern extremity of the massif (Fig. 4) the recent scarps are always facing uphill. This geometry and the symmetry with the crest of the ridge suggest a gravitational control. This hypothesis is confirmed by the observation that their trend follows the trend of the crest and turns with it, varying in orientation from NNE–SSW to E–W (Fig. 4).

We found the open cracks in the Triassic quartzites (Figs. 2 and 3c, i). The voids have opened along the bedding planes (Fig. 3c). The main open cracks are found on the top of the ridge (Fig. 4). Where they are present, the top of the ridge is singularly characterized by a double crest (Fig. 3d) with an intervening depression filled with blocks (Fig. 3i). The excavations made for the ski area allow the observation that one open crack does not extend in depth at more than 30 m (Fig. 3j) suggesting that these structures are relatively superficial. As for the uphill facing scarps, the trend of the open troughs follows the ridge crest varying in orientation from NNE–SSW to E–W (Fig. 4) thus suggesting a gravitational control. These voids along the crest denote ridge lateral spreading. Moreover, the double-crest is a classical sackung landform. These observations support the model of gravitational origin for these structures.

Goguel (1969) had also noted the presence of some open cracks but he interpreted them as not having the same origin as the fault scarps. On the contrary we believe that the faults and the open voids are genetically related, for three reasons: 1) they are all fresh recent linear morphologies of probable similar age, 2) they all strike parallel to the crest line, 3) some uphill facing scarps, located in the slope between the ridge top trough and the lower normal faults, are mainly of tensional origin. Effectively, in contrast to the deepness of their trough, they show only slight elevation changes across (Fig. 3f). They result from downslope tilt of a joint plane and erosion as proposed by Jahn (1964) (Fig. 5).

This new type of uphill facing scarp allows us to classify the sackung scarps of the Les Arcs area in three categories according to their morphology and mechanism: the ridgetop troughs, the tensional antislope scarps, and the antislope fault scarps (Fig. 5). This classification takes into account the principal models proposed for the formation of uphill-facing scarps by deformation of sub-vertical structures (Jahn, 1964; Zischinsky, 1966; Tabor, 1971; Bovis, 1982).

The mechanism for the formation of these sackung scarps depends on their position in the slope (Fig. 5). The ridge top trough results from the toppling of sub-vertical layers of quartzites toward the two sides of the

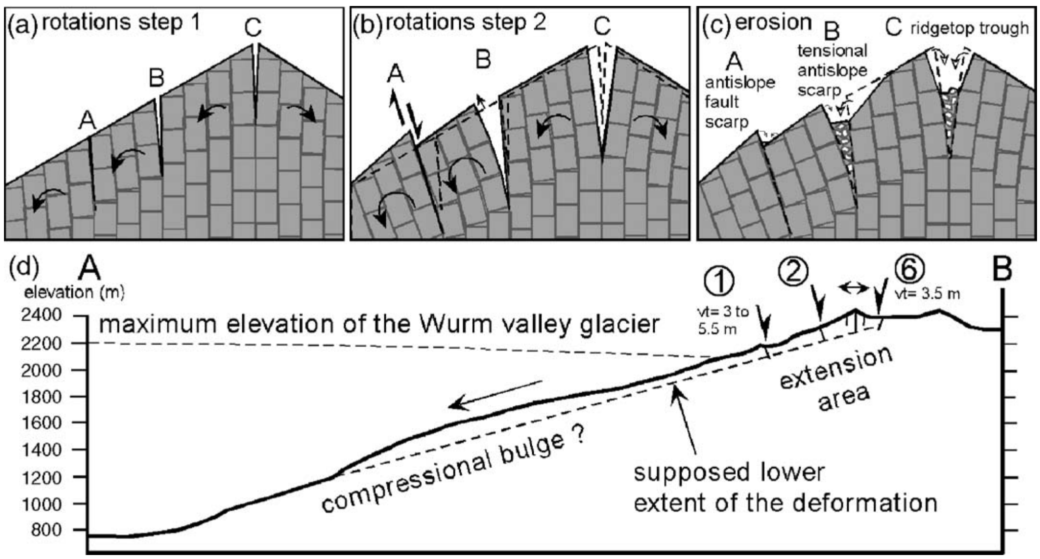

Fig. 5. Model of gravitational sackung deformation at Les Arcs. (a, b, c) Three-step model for the formation of the three types of sackung scarps (A, B, C) observed in the Aiguille Grive massif. The scarps results from rotation of rock layers and erosion. The ridge top troughs (C) are tension cracks resulting from toppling of vertical rock layers in the two slope directions. The tensional antislope scarp (B) are tension crack in the slope. Their morphology results from toppling of the rock layers and erosion mainly of the steep uphill face (Jahn, 1964). The antislope fault scarp (A) result from bookshelf bedding plane slip. (d) Topographic cross-section of the Aiguille Grive massif (location in Fig. 2) with location of the deformation. The valley glacier was settled on the Carboniferous silts and sandstones. We propose that when it retreated the silts crept gravitationally toward the valley. The convex profile of the toe of the range could represent the resulting compressional bulge, while extension is well expressed in the vertical bedding of the quartzites present along the crest.

range. Note that in this case, rocks have fallen from the two walls to partially fill the trough and its present width (more than 8 m at site 4, Figs. 4 and 3i) is not representative only of its opening but depends also on the bulk of fallen rocks and on the deepness of the voids.

The tensional antislope scarps (Fig. 5) are also tension voids but located in the slope, and therefore result from only one direction of bedding plane toppling. As noted by Jahn (1964), their apparent vertical offset results from the tilt of the bedding planes and erosion of the uphill face. The antislope fault scarps are the lowermost sackung structures of Les Arcs, and do not show any opening (Fig. 3a, g). Considering that these scarps follow the bedding planes (Fig. 4) we propose that they result from downslope flexuring of the rock layers with bookshelf-type rotations ("flexural toppling" of Goodman and Bray, 1976; model A in Fig. 5). In contrast to the other scarps, the rotation of the footwall block is equal with those of the hanging-wall block.

Note that the open cracks are restricted to the upper part of the ridge which is made of thick quartzite layers and characterized by steep-slopes, while the fault scarps are located in a lower-dipping slope made of sandstones with silt interlayers that facilitate layer flexuring. We conclude from this distribution and mechanism, that the recent scarps observed in Les Arcs area clearly result from slope gravitational deformation.

3.3. Hypotheses on the age of the scarps

The morphology of the scarps suggests that they are very recent. That the troughs are not yet filled shows that deformation was more rapid than sedimentation (Fig. 3b and e). In Fig. 4 we can see on the western side of the ridge, boulder accumulations with a lobate structure. They are talus-derived rock glaciers resulting from the cryogenic movement of the screes during a Lateglacial cold period. In Fig. 4 we can distinguish two types of glacier deposits corresponding to two successive cold events. The rock glaciers poorly vegetated are the most recent. In the western Alps, recent rock glaciers are generally attributed to the Dryas or subboreal episodes (e.g. Kaiser, 1987). In dark in Fig. 4, we can see vegetated rock glaciers that extend farther downslope. They lie at places that were covered by ice at Würm (Antoine et al., 1993), and they are locally overlain or remobilized by the recent rock glaciers. These older rock glaciers have a possible Oldest Dryas age.

The westernmost fault scarp (#1 in Figs. 3a and 4) downdrops the uphill side of the slope and an old rock glacier by 3 to 5.5 m. On its northern tip this scarp has stopped a recent rock glacier (white in Fig. 4). It suggests that the fault moved after the Oldest Dryas, and before or during the Younger Dryas.

To the south another scarp (#2 in Fig. 4) cuts both old and recent rock glaciers. This fault downthrows the uphill side of the vegetated rock glacier by 5 m (Fig. 3g). Locally, this older rock glacier is overlain by a tongue-shaped recent rock glacier (north of #2 in Fig. 4). The fault downthrows its surface by only 2 m (Fig. 3g). These relationships reveal the progressive movements of fault #2: 3 m between the end of activity of the old rock glacier and the end of activity of the recent rock glacier (about 3.7 ky if both of Dryassic age); 2 m since the end of activity of the recent rock glacier (about 11.5 ky). This evolution suggests a deceleration of the fault activity with time.

A similar pattern of high initial rates of deformation followed by slower rates was evidenced in Colorado by trench studies of a sackung trough, and ^{14}C dating (McCalpin and Irvine, 1995). This pattern is consistent with a gravitational deformation starting rapidly after the removal of lateral support from the sides during deglaciation, and slowing down with time.

3.4. Proposed model

We conclude that the recent deformation at Les Arcs is not of tectonic origin because it is not located on a main fault but is controlled by the morphology and trend of the ridge. Moreover the structures observed are typical of sackung deformation which is generally well expressed in hard brittle rocks. To explain the formation of superficial scarps we propose a mechanism that has already been proposed for sackung in vertical bedding: the flexural toppling mode (Goodman and Bray, 1976; Bovis, 1982). Downslope tilt of the vertical stratigraphic layers (toppling) together with erosion of the uphill face of a tension crack, contribute to antislope scarp developments (Fig. 5a, b, c). As observed at Les Arcs, this model implies opposite senses on the two slopes of the ridge, and that opening component increases toward the top which results in ridgetop splitting. The rotation of blocks leads to the deceleration of the movement with time. Two favorable conditions existed at Les Arcs for this ridge spreading: a high relative relief since the retreat of the valley glacier, and sub-vertical bedding planes parallel to the valley.

Gravitational downslope flexuring of subvertical layers was already proposed for the deformation of steeply dipping valley sides that have undergone glacial debuttressing (e.g. Bovis, 1982). In apparent contradiction with this model the sackung scarps of Les Arcs are confined to the upper 300 m of the range. However, as could be observed in other sackungen (e.g. Radbruch-Hall et al., 1976) the 1400 m of lower debuttressed slope, made of carboniferous silts which deform more plastically, shows a large bulging (Fig. 5d). We propose that this bulging of the valley flank has accommodated the uphill extensional deformation. Following glacial debuttressing the mountain slope is creeping, with diffuse extensional and compressional deformation in the carboniferous silts and sandstones, and brittle extensional deformation (sackungen) in the more competent rocks of the crest.

4. Recent scarps in the Belledonne massif

4.1. Previous works

In 1970, Bordet published photos taken from a helicopter showing four recent fault scarps located on the eastern flank of the Pointe de Rognier Mountain in the north of the Belledonne external crystalline massif (Figs. 6, 7 and 8). Bordet interpreted the fault scarps as active reverse faults that cannot result from gravitational movements. But these faults scarps were only interpreted from aerial views. In 1995, Ghafiri mentioned seven fault scarps in the granite massif on both sides of the mountain crest. In 2002, Baize et al. described Bordet's faults as rectilinear scarps trending NE–SW with only decimetric vertical offsets. They interpreted them, and two other faults in the Combe de Lachat area (Fig. 7), as active sinistral strike-slip faults. This discrepancy in the observations motivated field studies and a detailed mapping of the recent fault scarps of northern Belledonne.

4.2. Morphology of the scarps

In 2002 we examined the fault scarps photographed by Bordet (1970) (Fig. 8). The most visible fault scarp on Bordet's photos (#2 in Fig. 7) lies in the middle of the

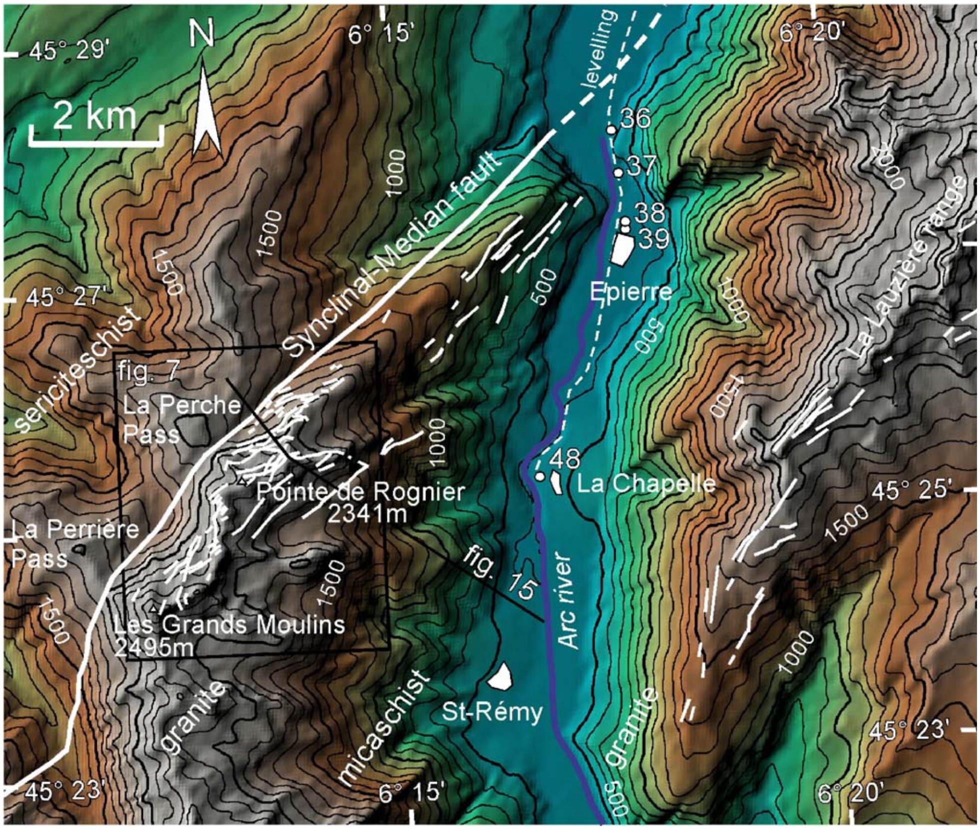

Fig. 6. The recent scarps of the Rognier area (Belledonne massif, Fig. 1) and their relationships with the topographic contours lines (elevation in meters) and the geology. All the scarps dip uphill. Numbers 36–39, 48, refer to the levelling benchmarks of the profile of Fig. 12. Note that apparently the Rognier scarps do not cross the floor of the Arc Valley along the Synclinal Median Fault. East of the Arc River, recent scarps are present, not along the Synclinal Median Fault, but in front of the Rognier slope along the Lauzière crest line, supporting a glacial debuttressing gravitational origin of the scarps.

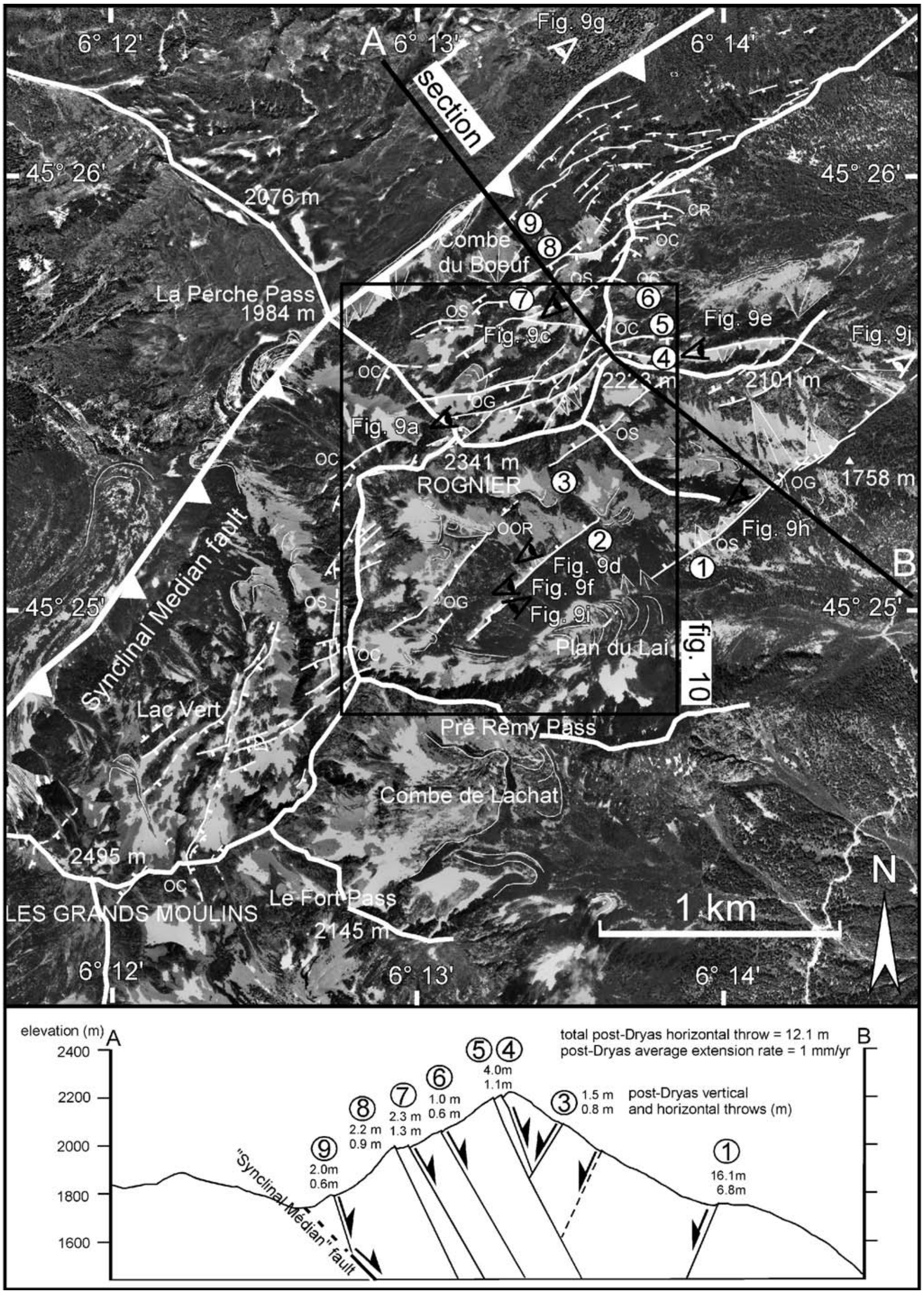

Fig. 7. Map of the fault scarps, rock glaciers and screes of the Rognier Mountain on orthorectified aerial photos (location in Fig. 6) and cross-section of the recent faults (dip measured in the field) in the main deformation area. OC: offset of crest. OS: offset of scree; OG: offset of a rock glacier. OOR: offset of the outer ridge of a rock glacier. CR: open cracks.

slope at about 1950 m elevation and is 870 m long (Fig. 9d). It is clearly a normal fault with no lateral component: the fault plane is locally well preserved and dips uphill; the V-shape intersection of the fault with the topography confirms that it dips to the west (Fig. 9f). We could observe striations on the fault plane but these are quartz lineations indicating a reverse movement that occurred in deeper setting during the Alpine

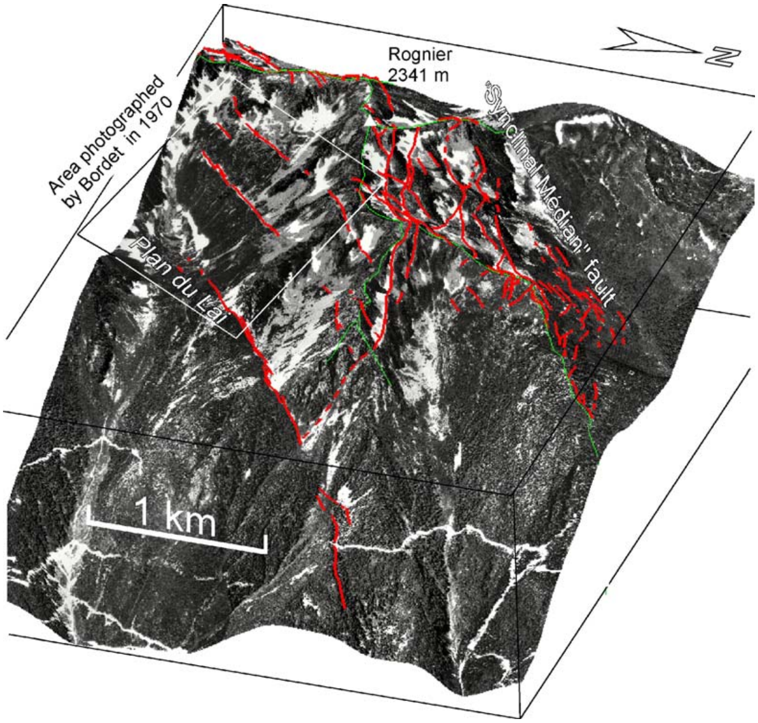

Fig. 8. 3D-view of the Rognier Mountain recent scarps (looking to the WSW). The recent fault scarps (red lines) are well visible because they stop debris and border the screes. Ridge crests are shown by green lines. Uphill facing scarps are present on the two sides of the Rognier range, and in particular in the fractured granites along the Synclinal Median fault. On the northern extremity of the range one can observe the connection between the antithetic recent faults.

deformation, and predates the recent normal throw. We measured the scarp heights using double frequency GPS. The maximum measurable vertical offset is 5.3 m (Fig. 9f). The recent activity of this fault is clearly demonstrated by the offset of the slope and the still visible free face of the antislope scarp despite the ongoing filling of the depression by rockfall debris (Fig. 9d).

The two upper fault scarps photographed by Bordet (1970) belong to the six fault segments lying between 2110 and 2200 m elevation revealing the activity of a 1550 m long fault (#3 in Fig. 7). This fault is also a normal fault with a few meters vertical throw that stops the active screes. Moreover, in its southern part, it cuts and offsets the lobate structure of a rock glacier, and in its central part it cuts only the external lobe of another rock glacier (Fig. 7, #3), attesting for a probable syn- to post-Dryassic age.

The lowest fault scarp, and the least visible on Bordet photos, is however the most spectacular (Figs. 8 and 9h, j). It runs between 1730 and 1820 m elevation and is at least 1330 m long (Fig. 7 #1). It is composed of several parallel or en echelon segments (Figs. 7 and 9j). The fault plane is locally well preserved showing that it is also an uphill dipping normal fault. This is furthermore confirmed by its V-shaped intersection with the topography (Fig. 9h). Its total vertical throw is more than 30 m near the point 1758 m (Figs. 9h and 7). This large scarp not only stops the falling blocks, but also offsets screes and an inactive rock glacier of probable Younger Dryas age (Fig. 9j). The normal fault uplifted the frontal lower part of the rock glacier by 16 ± 1 m after this cold period, while the total offset of the slope is about 30 m. It allows an evaluation of the vertical throw rate of 1.4 ± 0.1 mm/year and an estimate age of about twice the age of the rock glacier, that is 20–23 ky, for the beginning of the deformation if this rate is supposed constant through time.

Farther south, in the Combe du Lachat area, we did not find evidences for recent strike-slip movements as asserted by Baize et al. (2002). This area is in fact covered with rock glaciers with their own deformation (Fig. 7). However, the fault scarps discovered by Bordet (1970) are not the only ones, as noted by Ghafiri (1995). We found more than 80 other recent faults scarps, with vertical offset commonly between 1 m and 18 m, on both sides of the Rognier Ridge ((Figs. 7, 8 and 10)). Most of the scarps are confined on the western slope of

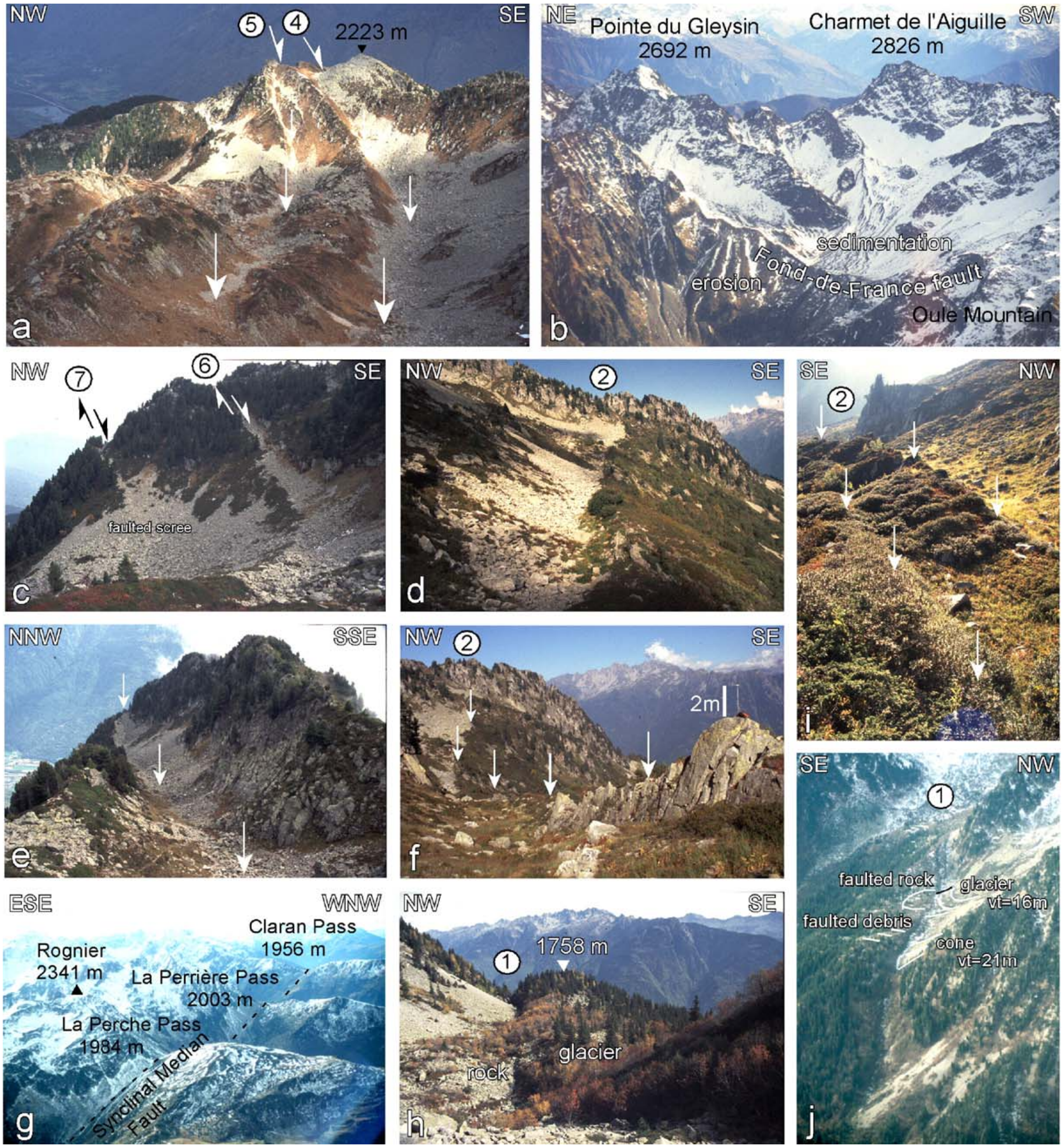

Fig. 9. Views of recent fault scarps in nothern Belledonne. Refer to Fig. 7 for the location of the photos and the scarps numbered. a) Triple-crested ridge; b) the Oule uphill facing scarp, corresponding to the reactivation for at least 2.1 km of the Font-de-France fault (location in Fig. 1); c) “saw teeth” morphology of a secondary crest line cut by two normal faults; d) fault scarp first noted by Bordet (1970); e) tensional antislope scarp; f) measurement by GPS of a 5.3 m high normal fault scarp; g) trace of the Synclinal Median fault; h) normal fault scarp offsetting by 16 m a rock glacier; i) two parallel normal fault scarps; j) aerial view of fault #1.

the Rognier Mountain between the crest and the Synclinal median fault (Figs. 7 and 8). The entire scarp swarm extends 9.2 km long from the Grands Moulins to Epierre (Fig. 6). The two largest scarps of the northwestern slope are found along the crest just north of the Rognier Mountain and have 12 m and 18 m vertical offsets (Fig. 7 #4 and 5). They cut the summit 2223 m (Fig. 9a) and result in three crest lines and two intervening deep depressions along the main crest (Fig. 10) similar to what we saw in the Arcs case study. On the northwestern slope the main faults cut secondary crests resulting in “saw teeth” morphologies (Fig. 9c).

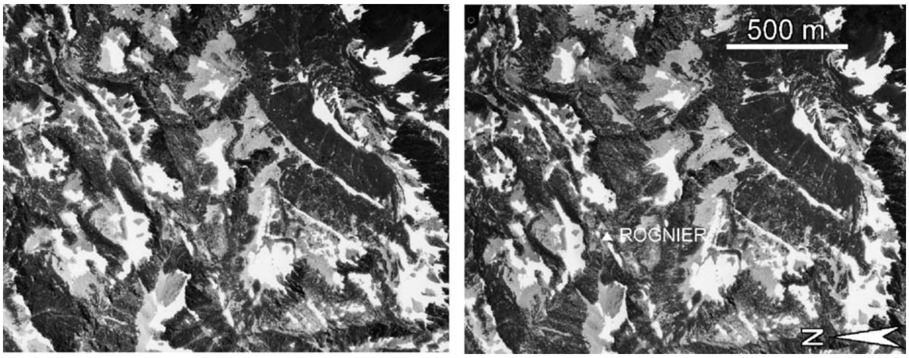

Fig. 10. Stereoscopic aerial view of the Rognier main scarp area (location in Fig. 7). Scarps are visible as strait lines between dark vegetated slope and light unvegetated screes or snow areas in troughs.

In all these areas the scarps have the same characteristics: 1) They are exclusively anti-slope scarps. 2) They retain or offset screes and rock glaciers. 3) They show only pure normal movements with 1 m to 30 m vertical offsets. 4) They trend around NE–SW which is the trend of most of the faults in the Belledonne massif. 5) They are inherited faults. Quartz lineations on the fault planes indicate strike-slip dextral to reverse movements that occurred in deeper conditions probably during the Tertiary compression.

4.3. Origin of the scarps

Several observations could suggest that these scarps result from active tectonics: (1) the recent movements occurred on tectonic faults. (2) The fault scarps make a 1,6 km wide strip along the Synclinal Median fault which is one of the main faults of the Belledone Massif. (3) The deformation is extensional and the trend of extension is NW–SE, as expected for tectonic deformation according to GPS and seismotectonic data (Fig. 1). (4) The estimated slip rates are under 2 mm/year as can be expected for tectonic deformation. (5) The deformation area is long (9.2 km) as typical for tectonic deformation. (6) The scarp height to length ratios (between 0.006 and 0.02) fits that of seismic normal faults (between 0.004 and 0.07 for cumulated slips; Manighetti et al., 2001).

To test this tectonic hypothesis we looked for recent deformation in the Belledonne Massif far from the Arc Valley (Fig. 1). To the south we found a few antislope recent scarps in two areas: the Pic du Frene and the Oule Mountain (Fig. 1). The Oule scarp is the larger. It is visible for 2.1 km along the ENE-trending Fond de France fault (Fig. 9b), and with other parallel scarps (Fig. 11) it makes a 3.5 km long deformation area that disappears to the south into an area of rapid erosion and is hidden to the north by the glaciers. But there is no continuity between these recent deformation areas in the Belledonne Massif. Even if the Rognier, Pic du Frene and Oule scarp areas appear to form a dextral en-echelon pattern (Fig. 1), this arrangement could be coincidental. To the north of the Pointe de Rognier, we found antislope scarps that follow the Rognier range and descend to within 300 m of the valley floor (Fig. 6), but do not cut the valley floor, and are not present east of the Arc River along the Synclinal Median fault as would be expected if the movements were of tectonic origin.

Finally, many observations suggest that the formation of the scarps of the Belledonne massif was mainly driven by gravity: (1) Except for the Oule, the largest scarps are not present along the main tectonic fault, but along the ridge crests (2) Like sackung scarps, the fault scarps are sub-parallel to the slope contours and to the crest (Fig. 6). (3) The scarps that follow the crest change in strike with it. This dependence results in the reactivation of several trends of inherited fractures where the crest curves sharply (Fig. 7). (4) The fault movements are dip-slip without any strike-slip component. (5) Close to the crest some scarps show only slight elevation changes across them and result mainly from opening, indicating crest lateral spreading (Fig. 9e). (6) The scarps are relatively short (less than 2.1 km) compared with tectonic fault ruptures which are commonly several kilometers to tens of kilometers long (Wells and Coppersmith, 1994; Sébrier et al., 1997; McCalpin, 2003). (7) Between the Grand Moulins and Epierre, there is not a single fault as often observed for seismogenic faults, but a swarm of multiple parallel scarps characteristic of sackungen (McCalpin, 2003) (Fig. 7). (8) The normal movements always indicate a subsidence of the mountain top, where the lithostatic stress is the higher. (9) We have typical sackung structures like antislope scarps, multiple-crested ridges,

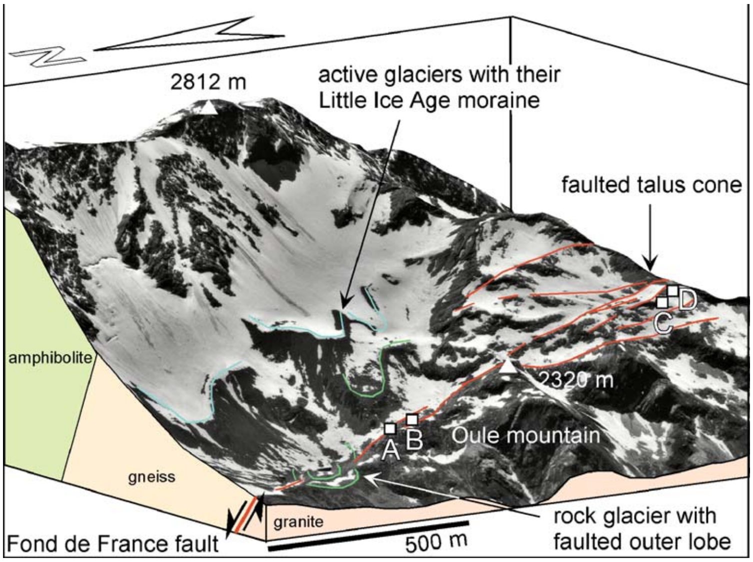

Fig. 11. 3D-view of recent fault scarps of the Oule mountain (looking to the ESE). The main fault scarp is along the Fond de France fault (Fig. 1). It spreads into several fault scarps at its southern and northern extremity. Like at the Rognier sackung, all the fault scarps dip uphill and a symmetry with the crest is observed north of this view confirming a gravitational control. A, B, C and D indicate the sites of sampling for ^{10}Be dating (Table 1).

and ridge top troughs. Therefore we conclude that most of the fault scarps we have studied in the Belledonne massif are sackung (gravitational) scarps.

However, the fact that most of the scarps are of gravitational origin does not prove that there is no active fault beneath them. The Synclinal Median fault is marked in the topography by three passes each with an apparent triangular faceted spur of several tens of meters high in micaschists (Figs. 6 and 9g). But, we found recent scarps cutting the rock glacier only at La Perrière Pass. They are two parallel scarps with vertical throw of less than 5 m in rock glaciers blocks and we could not distinguish between their tectonic or gravitational origin. We also checked levelling data from the Arc valley. Repeated levelling surveys between 1890 and 1969 indicate a subsidence of up to 10 cm of the Arc valley south of the benchmark 36 (Figs. 6 and 12). This constant 1 mm/year rate suggests that part of the crustal extension measured by GPS (Fig. 1; Calais et al., 2002) could be localized on the Synclinal median fault zone. However, even if the Synclinal median fault was active, its deformation rate could be independent from the rate of gravitational spreading.

4.4. Hypotheses on the rates of deformation, and ^{10}Be dating

Determination of deformation rates can help to better understand the origin of the scarps. As the Rognier recent scarps often cut Dryassic rock glaciers, we could evaluate the deformation rate along a section in the main deformation area (Fig. 7). With a total post-Dryassic horizontal throw of 12.1 ± 2 m (Fig. 7), we obtain an average post-Dryassic extension rate of 1 ± 0.2 mm/year. This value is lower than the SE-extensional rate revealed by the GPS station of Modane (1.7 ± 0.4 mm/year; Calais et al., 2002) (Fig. 1), but is also in the range of the deformation rates of sackungen (0.4–5 mm/year, Varnes et al., 1990; McCalpin and Irvine, 1995) and does not allow discrimination between tectonics and gravitational origins.

We also chose to date the Oule fault scarp located 11 km south of the Rognier sackung, on the Font-de-France fault (Fig. 1), because in this area we have locally a single scarp that records all the gravitational deformation (Fig. 9b). Its gravitational origin is attested by the presence near its northern extremity of two other uphill facing scarps along the ridge crest. Like the fault #3 in Fig. 7, the Oule fault scarp cuts the external lobe of a rock glacier but not its inner lobe (Fig. 11), suggesting an activity during the Dryas. Constraining surface fault displacements can be achieved by cosmogenic isotopes, either by dating deformed geomorphologic features (Bellier et al., 1999; Siame et al., 2001), or by dating the scarps themselves (e.g. Benedetti et al., 2002). We have combined both methods. We have sampled at site A (Fig. 11) where the footwall scarp surface looks well preserved over 13 m (Fig. 13a) and where the hanging-wall bedrock outcrops close to the scarp. Quartz was sampled at 0.5 m, 3.0 m and 7.0 m above the base of the

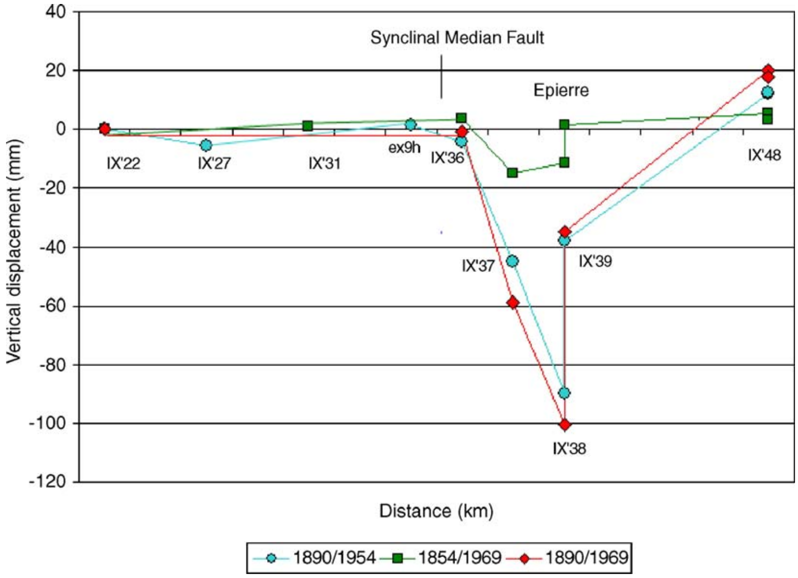

Fig. 12. Compared levelling along the Arc valley across the Synclinal Median fault (location in Figs. 1 and 6). Compared levelling for the periods 1890–1954, and 1954–1969 reveals a steady state subsidence around Epierre starting near the synclinal median fault (bench 36, Fig. 6).

scarp (Fig. 13a). The samples were expected to yield equal to increasing exposure ages with increasing height since the upper samples were exhumed before the lower ones. However, because the scarp could have started to form under a local glacier, we also assessed the end of glacial shielding by sampling glacially eroded surfaces (sites A and C, Figs. 11 and 13d).

We also sampled a probably faulted moraine (B in Fig. 11) and a scree cut by the fault scarp (site D in Fig. 13). In site B, we sampled quart-veins in an amphibolite 4×4×3 m block. In site D, we sampled a quartz vein on a 50 cm long block, downslope of a 4 m high scarp (Fig. 13c).

The 250–500 μm quartz grains isolation, dissolution, ^{9}Be spike addition and alkaline Be precipitations were

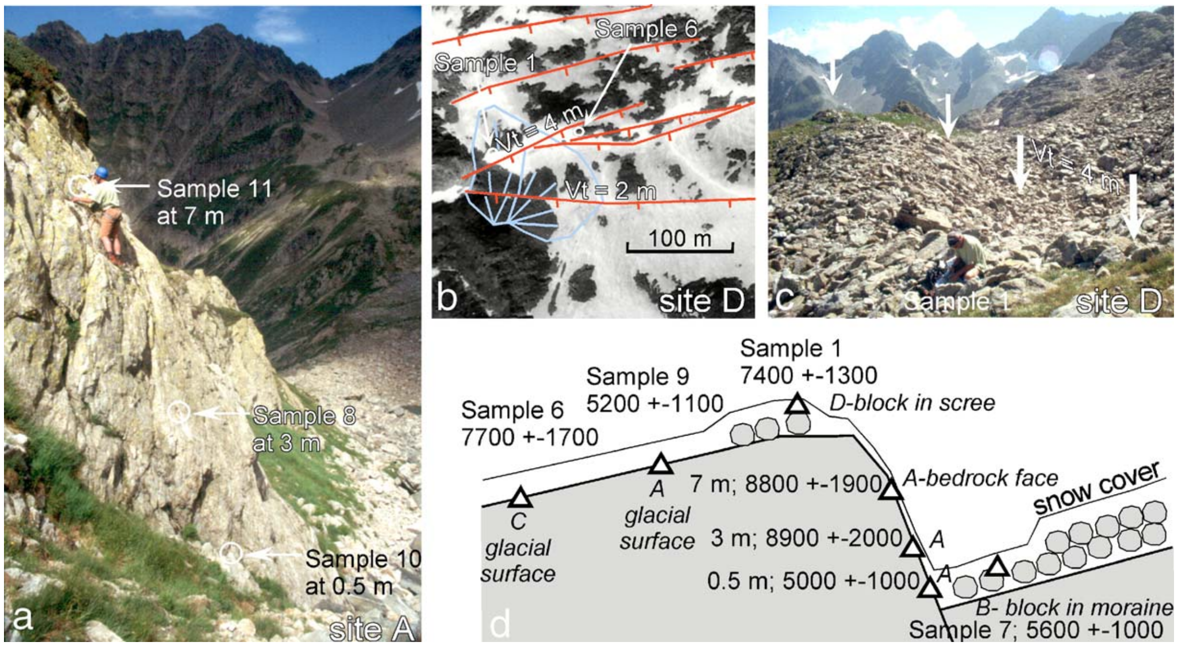

Fig. 13. Location of the samples dated by ^{10}Be concentrations. See location of sites in Fig. 11. a) View of the main fault scarp at site A. b) Aerial photo of the faulted debris cones of site D. c) View of the faulted debris cone and location of sample 1. d) Schematic composite cross-section illustrating the relative location of all the dated samples. V_t = vertical throw.

Table 1
^{10}Be concentration along the Oule scarp

Site	Sample	Type	^{10}Be/^{9}Be 10^{-13}	[^{10}Be] 10^{5} atom g^{-1}	Elevation (m)	Scaling factor	Shielding factor	Apparent age (ky)
A	10	Fault scarp 0.5 m	1.862	1.3±0.2	2170	5.781	0.7013	5.0±1.0
	08	Fault scarp 3.0 m	1.160	2.2±0.4	2170	5.781	0.7013	8.9±2.0
	11	Fault scarp 7.0 m	1.425	2.2±0.4	2180	5.825	0.7013	8.8±1.9
	09	Glacial surface	1.345	1.8±0.3	2180	5.825	0.9661	5.2±1.1
B	07	Block in moraine	2.383	2.0±0.2	2230	6.049	0.9583	5.6±1.0
C	06	Glacial surface	3.711	3.2±0.5	2390	6.818	0.9726	7.7±1.7
D	01	Block in scree	4.320	3.0±0.2	2390	6.818	0.9610	7.4±1.3

^{10}Be uncertainties (1σ) include a 3% contribution conservatively estimated from observed standard variations during the runs, a 1σ statistical error in the number of ^{10}Be events counted, uncertainty on the blank correction (associated ^{10}Be/^{9}Be blank ratio was $(1.0\pm0.3\times10^{-14})$, and a 15% uncertainty on in situ ^{10}Be production rate within quartz (Gosse and Klein, 1996). ^{10}Be/^{9}Be ratios were calibrated against the National Institute of Standards and Technology (NIST) standard reference material SRM 4325 (^{10}Be/^{9}Be$_{NIST}=26.8\pm1.4\times10^{-12}$). Latitude (45,311°) and altitudes scaling factors are from Dunai (2000).

performed following the protocol of Bourlès et al. (1989). ^{10}Be measurements were performed on the Tandetron Accelerator Mass Spectrometry (AMS) facility at Gif-sur-Yvette, France. We have thus normalized our ^{10}Be concentrations by a factor of 0.875 to make them comparable to those based on ICN standards. For the apparent age calculation, we use the high-latitude sea level production rate of (Kubik et al., 1998) obtained on a landslide that occurred in the Alps during the Dryas (5.42±0.3 atom g^{-1} year^{-1}), recalibrated according to Dunai (2000) scaling (Niedermann, 2002). Topographic shielding is evaluated following Heidbreder et al. (1971) and Dunne et al. (1999).

Results are given in Table 1 and Fig. 13. Note that the ages of the offset surface (samples 6, 9 and 1) are younger that the oldest age of the bedrock fault scarp (Fig. 13). This could mean that the 13 m high scarp at site A was partly formed before the erosion at sites C and A, and the formation of scree at site D. In fact, given the location, both the effects of snow cover and erosion rate have to be considered (Fig. 14). The common effect of snow shielding and erosion is an underestimation of the exposure ages. Snow covers this area most of the year. Spatially, the snow cover is very uneven. The snow shielding thus varies from one sample site to another and might alter the pattern of relative ages. Temporally too, the snow thickness has varied unpredictably.

Erosion rates are not easier to assess. Values between 0.5 and 1.0 cm ky^{-1} (zone a, Fig 14) are common in mountains of similar elevations and latitudes (e.g. Nishiizumi et al., 1993; Bierman et al., 1995; Bierman et al., 1999; Phillips et al., 1997), but higher rates, up to 5–10 cm ky^{-1} (zone b, Fig. 14), have been measured (Nishiizumi et al., 1993).

We can however assess the minimum age for the scarp formation at sites A and D, and the highest possible exposure age along the scarp. The moraines located at the front of the retreating local glaciers (Fig. 11) were emplaced during the Little Ice Age (100–500 years BP). For the Oule scarp to be reached by these glaciers requires a 170±15 m lowering of the local Equilibrium Line Altitude (ELA). At 80 km to the NE, the ELA depletion reached 290 m during the Younger Dryas (~11.5–12.8 ky cal BP; Alley, 2000), and 400 m during the Gschnitz cold stadial (~14,8 ky cal BP; Maisch, 1981; Dorth-Monachon, 1986; Coutterand and Nicoud, 2005). Thus, during the Younger Dryas, unified glaciers likely covered the place of the present-day scarp. And during the Gschnitz, the entire scarp was covered with ice. All the dated sites were uncovered

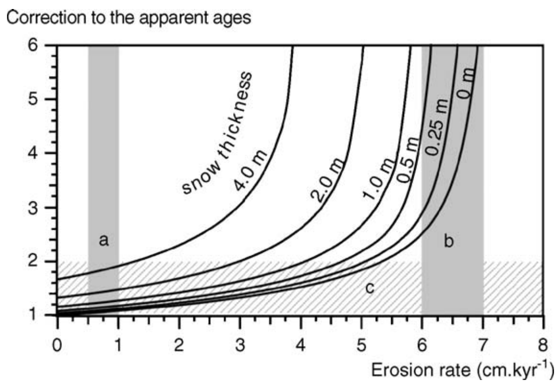

Fig. 14. Effects of snow cover and erosion rates on exposure ages. Exposure age changes as a function of the erosion rate, expressed as a correction factor to the apparent exposure ages (as in Table 1), for different values of snow cover (in meters). Snow shielding is calculated with an annual cover of 8 months, a snow density of 0.25 g cm^{-3}, and a rock density of 2.70 g cm^{-3}. a,b,c shaded zones: a: lower values of crystalline rock erosion rate obtained using cosmonuclides in mid-latitude mountains; b: highest values reported; c: probable range of snow shielding and erosion rate cumulated effects within the studied area.

between the Gschnitz and the end of the Younger Dryas, that is, between 15 and 12 ky. The absolute ages of the late activity are thus bracketed by the apparent ages (5–10 ky) and the deglaciation ages (12–15 ky). In addition, the exposure ages of the glacial surface (site C) and offset scree (site D) are undistinguishable. They indicate that part of this late scarp formation was coeval to the retreat of the local glaciers or closely followed it. Despite large correction factors, the dating of the Oule thus confirms the recent age of the deformation.

The relative pattern of apparent ages can give an upper value for the rate of tectonic denudation of the scarp if we admit that the entire scarp at site A post-dates the local glaciers (Fig. 13). Indeed, although the glacial surfaces yield ages undistinguishable from the upper scarp ages, no glacial striae are discernable on the scarp at this location. If post-glacial, the ages suggest a rapid (0 year ± 1900) fault slip from point 7 m to point 3 m; a slower movement from point 3 m to point 0.5 m, and a much slower movement from point 0.5 to the base of the scarp (if we suppose that the movement has not stopped). We conclude that much of the movement occurred beginning Holocene, and has slowed down.

4.5. Proposed model

The distribution of the recent scarps of the Belledonne massif and the probable rapid denudation of the Oule scarp support the gravitational interpretation. The most debated feature of a sackung is its extent at depth. In the case of the Rognier sackung, a view from the north (Fig. 8) shows how the antithetic faults connect on each other for constructing the section at depth (Fig. 15). We conclude that deformation may extend down to 600 m elevation, 1600 m beneath the Rognier crest. But conversely we observe that no fault was reactivated to the east of fault #1, with a connection deeper than the valley floor (Fig. 15), thus supporting again a gravitational model.

The most frequently proposed triggering mechanism for a sackung is the glacial debuttressing (e.g. Radbruch-Hall, 1978; Agliardi et al., 2001; Onida et al., 2001). We propose also as main mechanism for the Rognier sackung, glacial erosion followed by the retreat of the Würm valley glacier and the debuttressing of the valley slopes. During the Würm, the Arc glacier incised the valley but maintained a pressure on the valley sides (Fig. 15a). When the glacier has withdrawn, the valley flanks decompressed and deformed (Fig. 15b). This horizontal movement inducing a probable steepening of the ridge slope was accommodated by the sagging of the Rognier crest along the pre-existing Synclinal Median fault zone.

Note in Figs. 1 and 6 that the sackung occurred in a portion of the Synclinal Median fracture zone where the crest is close to the main fault and trends parallel to it. Moreover, sagging occurred only along the crest segment close to the deep glacier valley (Fig. 6), in agreement with a debuttressing model. The Rognier "sackungen" disappear toward the south, where the glacial valley gets away from and perpendicular to the fault zone. This model implies decompression of the two valley sides and formation of sackung scarps where a crest is parallel to the faults and to the glacial valley. We effectively found sackung scarps on the eastern side of the Arc River along the crest of the Lauziere massif (Fig. 6). As in the Rognier Mountain, the sackung scarps, that in places offset rock glaciers, follow the crest parallel to the valley and change of strike with it.

Finally, the huge Rognier sackung is the consequence of several favorable conditions: (1) presence of the fracture zone of a large alpine fault, (2) presence of a deeply incised glacier valley, (3) the crest line and the valley trend subparallel to the fractures.

5. Discussion

That, for these two sites, we found scarp swarms that do not cross the valley but follow the ridge crest, clearly shows that the deformation is related to the topography and according to our cross section is superficial (above the valley floor). Therefore, for these examples, we can rule out as main mechanism crustal deformation, like post-glacial rebound or plate tectonics, and conclude for a gravitational origin (sackung). Glacio-isotasy may have formed recent scarps in the Alps, like it has done in the northern Europe. Such origin was proposed for the post glacial scarps found on 88 km along the Rhine–Rhone line in the Swiss Alps (Eckardt et al., 1983; Geiger et al., 1986). But this origin is improbable for the scarps swarms we have mapped, which are less than 10 km long and are strictly linked to a single ridge.

Furthermore, that deformation at the Rognier site is related to glacial debuttressing is supported by the presence of another sackung in the Lauzière range on the opposite side of the valley (Fig. 6). But from the conclusion that these valley slopes deform under gravity, arise the question of the stability of the GPS sites. In fact, the GPS sites CHAT and MODA are located on the valley floor (Fig. 1) and we conclude that the similarity in trend between the gravitational spreading of the range and the crustal extension deduced from geophysics (Fig. 1) is coincidental and mainly

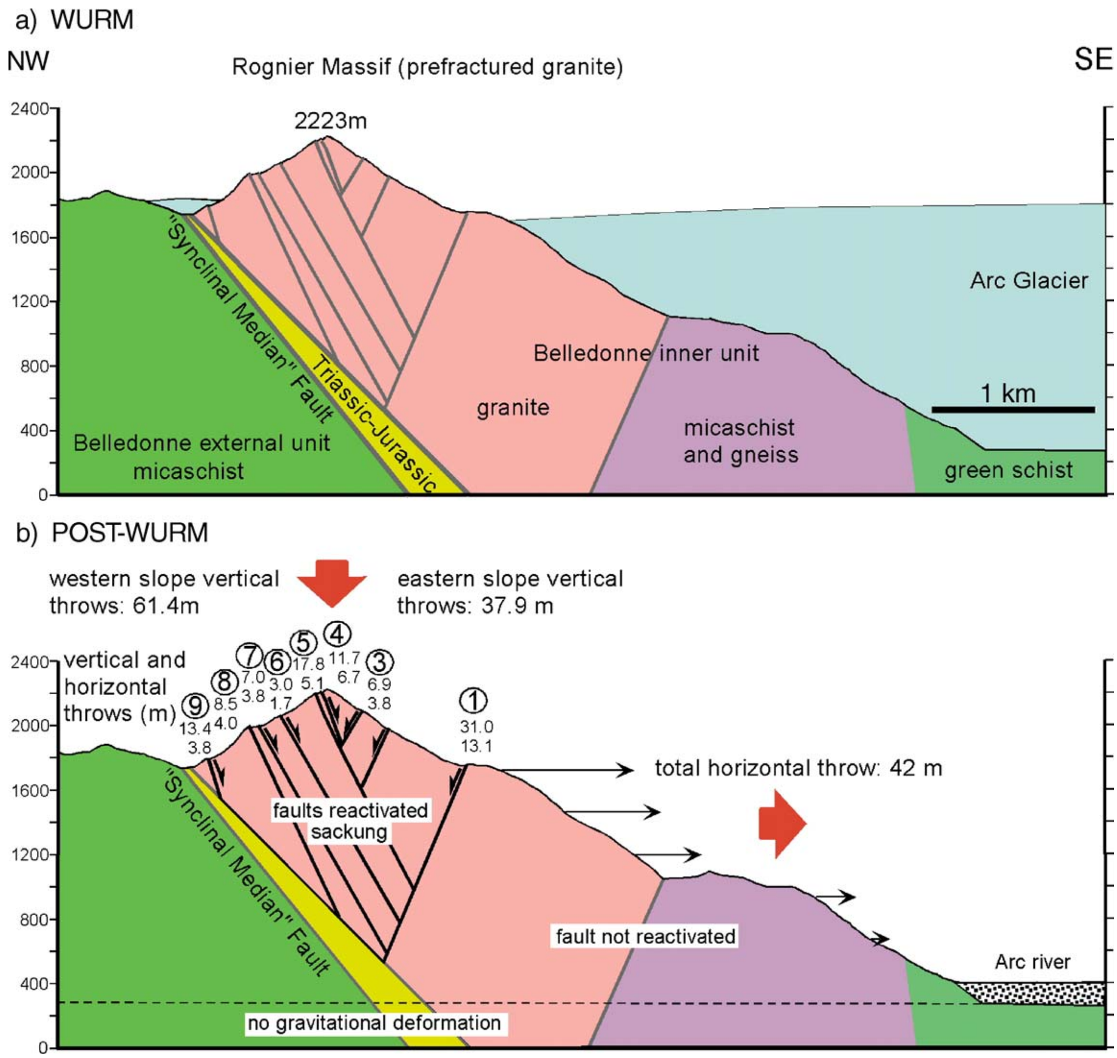

Fig. 15. Model of gravitational sackung deformation at the Rognier (cf. Figs. 6 and 7 for location of the section). Same fault numbers as in Fig. 7 with measured total morphologic vertical throw and corresponding horizontal throw. Note that all the faults activated during the post-glacial period are conjugate faults that connect above the valley floor supporting a superficial gravitational model. In a first step the Würm glacier erodes the valley and supports the slope, reaching the elevation of 1700 m. In a second step, the glacial debuttressing results in an increase of the slope compensated by a horizontal extension in the granite massif through reactivation of conjugate faults of the Synclinal Median fracture zone.

results from the structural heritage. Therefore, their deformation rates are independent.

6. Contribution to the understanding of the origin of sackungen

6.1. Typology of sackungen

In the French Alps, the Sechilienne and La Clapiere "landslides" were the only previously studied sackungen (Vibert, 1987; Flageollet, 1989; Vengeon, 1998; Bigot-Cormier et al., 2005), but they differ from the ones described here because they are localized at the base of the mountain slopes. At the Arcs and at the Rognier we found larger "sackungen" characterized by their main scarps along the crest. This type of gravitational structure could be called "ridgetop sackungen". They correspond to the gravitational collapse of an entire ridge and involve the two slopes.

Even if the scarps morphologies are similar, the examples of Les Arcs and the Belledonne correspond to two different mechanisms of sackung. Varnes et al. (1989) distinguish three types of sackung: a) spreading of rigid rocks overlying soft rocks (Radbruch-Hall et al., 1976; Radbruch-Hall, 1978); b) sagging and bending of foliated rocks (Zischinsky, 1969) and c) differential displacement in hard but fractured crystalline igneous rocks. The example of Les Arcs corresponds to the second category, and the Rognier to the third. At the Arcs, vertical bedding planes topple under gravity generating frequent open voids. In the Belledonne faulted granites, deformation mainly consists of fault slips.

6.2. Subsurface geometry

The subsurface geometry of the sackungen is often debated. One hypothesis is that a "sackungen" is the shallow surface manifestation of toppling and flexural slip along discontinuities (flexural toppling of Bovis, 1982) that do not penetrate into a mountain mass to any great depth (Jahn, 1964). This model is in agreement with our observations at Les Arcs. But note that at the Sechilienne sackung, which results from toppling of fractured micaschists, repeated levelling in a 240 m long gallery shows that flexural toppling can occur at more than 100 m depth (Vengeon, 1998). Another hypothesis is that the scarps may pass downward, at depth of several hundred meters, into a zone of rock creep (Zischinsky, 1969; Mahr, 1977; Radbruch-Hall, 1978; Morton and Sadler, 1989; Jarman and Ballantyne, 2002). In the Rognier sackung the extremity of the range allows the exceptional observation of fault connections supporting the idea that fault movements may occur at several hundred meter depth. It allows us to propose a new mechanism for the deep sagging in hard crystalline rocks: antithetic fault movements (Fig. 15).

6.3. Origin of sagging

Sackung can result from three different factors that are usually combined and can be influenced by other factors like groundwater fluctuations: a) gravity forces that produce slow deformation in a rock mass. These gravity forces act especially where valley-side slopes have been steepened by glacial erosion (Tabor, 1971; Radbruch-Hall, 1978; Bovis, 1990). b) Stress release due to glacial debuttressing resulting in progressive development of joints (e.g. Jarman and Ballantyne, 2002). Strain recovery against late Pleistocene glacial loading is suggested by slowing deformation rates deduced from trench observations along a scarp in Colorado (McCalpin and Irvine, 1995); c) earthquake shaking that induces lateral spreading and differential settlement of rock masses (Beck, 1968; Radbruch-Hall et al., 1976; Mahr, 1977; Solonenko, 1977; Dramis and Sorriso-Valvo, 1983; Morton et al., 1989; Ponti and Wells, 1991; McCalpin and Hart, 2003).

It is clear that in the Arc and Rognier cases, the glacial erosion and debuttressing are the main factor for the slope instability. Whether slip on sackung faults result from slow creep or sudden vertical displacement is often a matter of debate (McCalpin, 1997; Onida et al., 2001; McCalpin, 2003). At Les Arcs, the flexural slip mechanism suggests that deformation occurred by creeping, but at the Rognier, where movements occurred on conjugate fault planes submitted to large lithostatic stresses, fault slip can have been triggered by strong earthquake shaking.

Ridge crest collapses were observed after several severe earthquake shaking in California (Hart et al., 1990; Ponti and Wells, 1991; Harp and Jibson, 1996) and in Alaska near Gillet Pass, where a sackung curvilinear uphill-facing scarp formed during the 3 Nov. 2000 M 7.9 Denali earthquake (cf. McCalpin's web site, http://www.geohaz.com). Along the San Andreas fault, McCalpin and Hart (2001) trenched several sackung ridgetop depressions and showed that they recorded repeated ground ruptures of 1m or more of vertical displacement probably generated during large earthquake shaking.

The Oule scarp is probably a sackung activated by earthquake shaking. Effectively it is unlikely to result from gravity alone because it is located far from the deep glacier valleys (Fig. 1). Moreover the main failure plane dips toward the mountain range which does not permit a simple gravitational sliding. Furthermore, our cosmic-ray exposure dating shows that fault slip may have been rapid (Fig. 13). In agreement with this earthquake shaking hypothesis, seismic lines of the Bourget Lake at Chambéry (Fig. 1) revealed a large sublacustrine landslide that may have been also triggered by earthquake shaking around 11 ky ago (Chapron et al., 1996; Van Rensbergen et al., 1999).

Therefore, even if the Rognier sackung was clearly favored by glacial debuttressing, it is possible that the fault movements were triggered by similar postglacial earthquake shakings. Earthquakes may have been generated on the Belledonne fault (Thouvenot et al., 2003), the Frontal Pennine fault, or the Synclinal-Median fault, just beneath the sackung.

7. Conclusions

The recent fault scarps Les Arcs and Rognier result primarily from gravitational deformation. All the mapped scarps show similar morphologies typical of gravitational sackungen, with the development of multiple-crested ridges, ridge-top troughs, closed depressions, and tension cracks along the crest, while dominantly uphill-facing scarps with pure dip-slip movements are produced on the slopes. The latter trend parallel to slope contour lines, bedding planes and fault strikes. Like most sackungen described in the literature, they develop in massive competent rocks, where deformation is brittle and well visible. They belong to the "ridge-top sackungen" type, with scarp swarms occurring close to the ridge crests. The gravity control is further evidenced on both sites by the symmetry of the fault movements across the crest (opposite senses of movement on opposite slopes). Sackung scarps at Les Arcs and the Oule exhibit a deceleration of the movements with time. The two studied sites differ by the depth and mechanism of the deformation. The Aiguille Grive "sackungen" results from shallow (a few tens of meters) flexural toppling of sub-vertical stratigraphic layers. The Rognier "sackungen" results from reactivation of conjugate faults in granite possibly down to a depth of 1600 m beneath the summit, but still above the valley floor. This limit in depth of deformation, rule out a differential glacial isostatic rebound interpretation, but support a gravitational interpretation. The ^{10}Be dating of a fault scarp confirms its recent age, and suggests that its movement was triggered by large earthquakes. Our dating and those of a sublacustrine landslide (Chapron et al., 1996) tend to support the idea that post-glacial rebound was a period of enhanced seismicity. Other factors such as granite pre-fracturing, glacier incision, lowering of friction due to either water pore pressure or ice within faults, and glacial debuttressing, can have contributed to form at the Rognier the largest sackung ever described in the western Alps.

Acknowledgements

This work was supported by the University of Savoy and the "Conseil Général de l'Isère" through the "Pôle Grenoblois d'Etudes et de Recherche pour la Prevention des Risques Naturels". We are grateful to C. Crouzet, M. Meghraoui and J. McCalpin for constructive discussions during this work.

References

Agliardi, F., Crosta, G., Zanchi, A., 2001. Structural constraints on deep-seated slope deformation kinematics. Engineering Geology 59, 83–102.

Alley, R.B., 2000. The Younger Dryas cold interval as viewed from central Greenland. Quaternary Science Reviews 19, 213–226.

Antoine, P., Barféty, J.C., Vivier, G., Gros, Y., Fudral, S., Landry, P., Fabre, J., 1993. Carte géol. France (1/50 000), feuille Bourg-Saint-Maurice (727)—Orléans: BRGM. Notice explicative par P. Antoine et al. (1992), 110 p.

Baize, S., Cushing, M., Lemeille, F, Granier, T, Grellet, B., Carbon, D., Combes, P., Hibsch, C., 2002. Inventaire des indices de rupture affectant le Quaternaire en relation avec les grandes structures connues en France métropolitaine et dans les régions limitrophes. Mémoire de la société géologique de France, Mémoire H.S., no 175, 142p, 1pl. H.T.

Beck, A.C., 1968. Gravity faulting as a mechanism of topographic adjustment. New Zealand Journal of Geology and Geophysics 11 (1), 191–199.

Bellier, O., Bourlès, D.L., Beaudouin, T., Braucher, R., 1999. Cosmic ray exposure (CRE) dating in wet tropical domain: late Quaternary fan emplacements in central Sulawesi (Indonesia). Terra Nova 11, 174–180.

Benedetti, L., Finkel, R., Papanastassiou, D., King, G.C.P., Armijo, R., Ryerson, F., Farber, D., Flerit, F., 2002. Post-glacial slip history of the Sparta fault (Greece) determined by ^{36}Cl cosmogenic dating: evidence for non-periodic earthquakes. Geophysical Research Letters 29 (8–10), doi:10.1029/2001GL014510.

Bierman, P., Gillespie, A., Caffee, M., Elmore, D., 1995. Estimating erosion rates and exposure ages with ^{36}Cl produced by neutron activation. Geochimica et Cosmochimica Acta 59 (18), 3779–3798.

Bierman, P.R., Marsella, K.A., Patterson, C., Davis, P.T., Caffe, M., 1999. Mid-Pleistocene cosmogenic minimum-age limits for pre-Wisconsinan glacial surfaces in southwestern Minnesota and southern Baffin Island: a multiple nuclide approach. Geomorphology 27, 25–39.

Bigot-Cormier, F., Braucher, R., Bourlès, D.L., Guglielmi, Y., Dubar, M., Stéphan, J.-F., 2005. Chronological constraints on processes leading to large active landslides. Earth and Planetary Science Letters 235, 141–150.

Bordet, P., 1970. Les failles vivantes du massif des Grands Moulins (Massif cristallin externe de Belledonne). Géologie Alpine 46, 43–47.

Bourlès, D., Raisbeck, G.M., Yiou, F., 1989. ^{10}Be and ^{9}Be in marine sediments and their potential for dating. Geochimica et Cosmochimica Acta 53, 443–452.

Bovis, M.J., 1982. Uphill-facing (antislope) scarps in the Coast Mountains, southwest British Columbia. Geological Society of America Bulletin 93, 804–812.

Bovis, M., 1990. Rock-slope deformation at Affliction Creek, southern Coast Mountains, British Columbia. Canadian Journal of Earth Sciences 27, 243–254.

Bovis, M.J., Evans, S.G., 1995. Rock slope movements along the Mount Currie "fault scarp", southern Coast Mountains, British Columbia. Canadian Journal of Earth Sciences 32, 2015–2020.

Calais, E., Nocquet, J.M., Jouanne, F., Tardy, M., 2002. Current strain regime in the Western Alps from continuous Global Positioning System measurements, 1996–2001. Geology 30 (7), 651–654.

Chapron, E., Van Rensbergen, P., Beck, C., De Batist, M., Paillet, A., 1996. Lacustrine sedimentary records of brutal events in Lake Le Bourget (northwestern Alps–southern Jura). Quaternaire 7, 155–168.

Coutterand, S., Nicoud, G., 2005. The Arve glacier retreat stages between Cluses bolt and Chamonix ombilic during the Tardiglaciaire (Arve valley Haute Savoie). Quaternaire 16 (2), 85–94.

Dorth-Monachon, C. 1986. Contribution à l'étude de la morphologie glaciaire de la vallée de l'Arve (Haute-Savoie, France). Essai de reconstitution paléogéographique, Unpublished PhD University of Lausanne.

Dramis, F., Sorriso-Valvo, M., 1983. Two cases of earthquake-triggered gravitational spreading in Algeria and Italy. Rendiconti della Società Geologica Italiana 6, 7–10.

Dunai, T.J., 2000. Scaling factors for production rates of in situ produced cosmogenic nuclides: a critical reevaluation. Earth and Planetary Science Letters 176, 157–169.

Dunne, J., Elmore, D., Muzikar, P., 1999. Scaling factors for the rates of production of cosmogenic nuclides for geometric shielding and attenuation at depth on sloped surfaces. Geomorphology 27, 3–11.

Eckardt, P., Funk, H., Labhart, T., 1983. Postglaziale Krustenbewegungena an der Rhein–Rhone–Linie. Vermessung, Photogrammetrie, Kulturtechnik 2, 43–56.

Ego, F., Sébrier, M., Carey-Gailhardis, E., Beate, B., 1996. Do the Billlecocha normal faults (Ecuador) reveal extension due to lithospheric body forces in the northern Andes? Tectonophysics 265, 255–273.

Eva, E., Solarino, S., Eva, C., Neri, G., 1997. Stress tensor orientation derived from fault plane solution in the southwestern Alps. Journal of Geophysical Research 102, 8171–8185.

Fenton, C.H. 1991. Neotectonics and palaeoseismicity in North West Scotland. Unpublished PhD Thesis, University of Glasgow, 403 pp.

Firth, C.R., Stewart, I., 2000. Postglacial tectonics of the Scottish glacio-isostatic uplift centre. Quaternary Science Reviews 19, 1469–1493.

Flageollet, J.C., 1989. Les Mouvements de Terrain et leur Prévention. Masson, Paris. 221 pp.

Follacci, J.P., 1987. Les mouvements du versant de la Clapière à Saint Etienne de Tinée (Alpes Maritimes). Bulletin de Liaison des Ponts et Chaussées 150/151, 39–54.

Forcella, F., 1984. Brevi note sulla tettonica gravitativa di versante nelle Alpi centrali. Bollettino della Societa Geologica Italiana 103, 689–696.

Fréchet, J., Pavoni, N., 1979. Etude de la sismicité de la zone briançonnaise entre Pelvoux et Argentera (Alpes occidentales) à l'aide d'un réseau de stations portables. Eclogae Geologicae Helveticae 72, 763–779.

Geiger, A., Kahle, H.G., Gubler, E., 1986. Recent crustal movements in the Alpine–Mediterranean region analysed in the Swiss Alps. Tectonophysics 130, 289–298.

Ghafiri, A.1995. Paléosismicité de failles actives en contexte de sismicité modérée: application à l'évaluation de l'aléa sismique dans le Sud-Est de la France. Unpublished PhD Thesis, University Paris-Sud Orsay, France.

Goguel, J., 1969. Des failles d'âge quaternaire à Peisey (Savoie). Bulletin de la Société Géologique de France 7 (XI), 35–39.

Goodman, R.E., Bray, J.W., 1976. Toppling of rock slopes. Proc. Special Conf. on Rock Engineering and Foundations and Slopes, vol. 2. ASCE, Boulder, Colorado, pp. 201–234.

Gosse, J.C, Klein, J., 1996. Production rate of in situ cosmogenic ^{10}Be in quartz at high altitude and mid-latitude. Radiocarbon 38 (1), 154–155.

Gregersen, S., Basham, P.W., 1989. Summary, conclusions and recommendations. In: Gregersen, S., Basham, P.W. (Eds.), Earthquakes at North Atlantic Passive Margins: Neotectonics and Postglacial Rebound. Kluwer, Dordrecht, pp. 1–6.

Guglielmi, C., Bertrand, F., Compagon, J.P., Follacci, J., Mudry, 2000. Acquisition of water chemistry in a mobile fissured basement massif: its role in the hydrogeological knowledge of the La Clapiere landslide (Mercantour massif Southern Alps, France). Journal of Hydrology 229, 138–148.

Harp, E.L., Jibson, R.W., 1994. Landslides triggered by the 1994 Northridge California earthquake. Bulletin of the Seismological Society of America 86, S319–S332.

Hart, E.L., Bryant, W.A., Wills, C.J., Treiman, J.A., 1990. The search for fault rupture and significance of ridgetop fissures, Santa Cruz Mountains, California. In: McNutt, S.R., Sydnor, R.H. (Eds.), The Loma Prieta earthquake of 17 October 1989. California Division of Mines and Geology Special Publication, vol. 104, pp. 83–94.

Heidbreder, E., Pinkau, K., Reppin, C., Schöenfelder, V., 1971. Measurement of the distribution in energy and angle of high-energy neutrons in the lower atmosphere. Journal of Geophysical Research 76, 2905–2916.

Jahn, A., 1964. Slope morphological features resulting from gravitation. Zeitschrift für Geomorphologie. Supplement 5, 59–72.

Jarman, D., Ballantyne, C., 2002. Beinn Fhada, Kintail: An examples of large-scale paraglacial rock slope deformation. Scottish Geographical Journal 118, 59–68.

Kaiser, B., 1987. Les versants de vanoise: enjeux traditionnels de fonctionnement morphoclimatique. Unpublished PhD Thesis, University Paris 7, 2 vol., 1088 p.

Knight, J., 1999. Geological evidence for neotectonic activity during deglaciation of the southern Sperrin Mountains, Northern Ireland. Journal of Quaternary Science 14, 45–57.

Kubik, P.W., Ivy-Ochs, S., Masarik, J., Frank, M., Schlüchter, C., 1998. ^{10}Be and ^{26}Al production rates deduced from an instantaneous event within the dendro-calibration curve, the landslide of Köfels, Ötz Valley, Austria. Earth and Planetary Science Letters 161, 231–241.

Mahr, T., 1977. Deep-reaching gravitational deformations of high mountain slopes. Bulletin of the International Association of Engineering Geology 16, 121–127.

Maisch, M., 1981. Glazialmorphologische und Gletschergeschichte Untersuchungen im Gebiet zwischen Landwasser-und Albulatal. Physische Geographie 3 (Zürich).

Manighetti, I., King, G.C.P., Gaudemer, Y., 2001. Slip accumulation and lateral propagation of active normal faults in Afar. Journal of Geophysical Research 106, 13667–13696.

McCalpin, J.P., 1997. Ridgetop splitting and spreading: nonseismic sackungen or earthquake-induced secondary deformation? Assoc. of Engineering Geologists, Program with Abstracts, 40th Annual Meeting, Sept. 30–Oct. 4, 1997, Portland, OR, pp. 126–127.

McCalpin, J.P., 2003. Criteria for determining the seismic significance of sackungen and other scarplike landforms in mountainous regions. In: Hart, E.W. (Ed.), Ridge-Top Spreading in California: California Geological Survey, Open-File Report, 1CD-ROM.

McCalpin, J.P., Irvine, J.R., 1995. Sackungen at Aspen Highlands Ski Area, Pitkin County, Colorado. Environmental and Engineering Geoscience 1, 277–290.

McCalpin, J.P., Hart, E.W., 2001. Holocene displacement history of ridgetop depressions (sackungen) in the San Gabriel Mountains, southern California. Geological Society of America, Abstract with Programs 33 (3), A30.

McCalpin, J.P., Hart, E.W., 2003. Ridge-top spreading features and relationship to earthquakes, San Gabriel Mountains Region, Southern California: Part A. Distribution and description of ridge-top depressions (sackungen): Part B. Paleoseismic investigations of ridge-top depressions. In: Hart, E.W. (Ed.), Ridge-Top Spreading in California: California Geological Survey, Open-File Report, 1 CD-ROM.

Mohr, P., 1986. Possible Late Pleistocene faulting in Iar (West) Connacht Ireland. Geological Magazine 123, 545–552.

Mollard, J.D., 1977. Regional landslide types in Canada. Geological Society of America, Reviews in Engineering Geology 3, 29–56.

Morton, D.M., Sadler, P.M., 1989. The failings of the Pelona Schist; landslides and sackungen in the Lone Pine Canyon and

Wrightwood areas of the eastern San Gabriel Mountains, southern California. In: Sadler, P.M., Morton, D.M. (Eds.), Landslides in a Semi-arid Environment, with Emphasis on the inland Valleys of Southern California. Inland Geological Society, Publication, vol. 2, pp. 301–322.

Morton, D.M., Campbell, R.H., Jibson, R.W., Wesson, R.L., Nicholson, C., 1989. Ground fractures and landslides produced by the North Palm Springs, California, earthquake of July 8, 1986. In: Sadler, P.M., Morton, D.M. (Eds.), Landslides in a Semi-arid Environment, with Emphasis on the inland Valleys of Southern California. Inland Geological Society, Publication, vol. 2, pp. 183–196.

Muir-Wood, R., 1989. Extraordinary deglaciation reverse faulting in northern Fennoscandia. In: Gregersen, S., Basham, P. (Eds.), Earthquakes at North-Atlantic Passive Margins: Neotectonics and Post-glacial Rebound. Kluwer Academic Publishers, Dordrecht, pp. 141–173.

Nemcok, A., 1972. Gravitational slope deformation in high mountains. Proc. 24th Int. Geol. Cong., Montréal, Sect., vol. 13, pp. 132–141.

Niedermann, S., 2002. Cosmic-ray-produced noble gases in terrestrial rocks: Dating tools for surface processes. In: Porcelli, D., Ballentine, C.J., Wieler, R. (Eds.), Noble Gases in Geochemistry and Cosmochemistry. Reviews in Mineralogy and Geochemistry, vol. 47, pp. 731–784.

Nishiizumi, K., Kohl, C.P., Arnold, J.R., Klein, J., Fink, D., Middleton, R., Lal, D., 1993. Role of in situ cosmogenic nuclides ^{10}Be and ^{26}Al in the study of diverse geomorphic processes. Earth Surface Processes and Landforms 18, 407–425.

Onida, M., Galadini, F., Forcella, F., 2001. Application of paleoseismological techniques to the study of Late Pleistocene–Holocene deep-seated gravitational movements at the Mortirolo Pass (central Alps Italy). Geologie en Mijnbouw 80, 209–227.

Paschinger, V., 1928. Untersuchungen uber Doppelgrate. Z. Geomorph. N.F., vol. 3, pp. 204–236.

Pavoni, N., 1980. Comparison of focal mechanisms of earthquakes and faulting in the Helvetic zone of central Valais (Swiss Alps). Eclogae Geologicae Helvetiae 73, 551–558.

Phillips, F.M., Zreda, M.G., Gosse, J.C., Klein, J., Evenson, E.B., Hall, R.D., Chadwick, O.A., Sharma, P., 1997. Cosmogenic ^{36}Cl and ^{10}Be ages of Quaternary glacial and fluvial deposits of the Wind River Range, Wyoming. Geological Society of America Bulletin 109 (1), 1462–1463.

Ponti, D.J., Wells, R.E., 1991. Off-fault ground ruptures in the Santa Cruz Mountains, California—ridge-top spreading versus tectonic extension during the 1989 Loma Prieta earthquake. Bulletin of the Seismological Society of America 81, 1480–1510.

Radbruch-Hall, D., 1978. Gravitational creep of rock masses on slopes. Chap. 17. In: Voight, B. (Ed.), Rockslides and Avalanches Natural Phenomenal. Developement in Geotechnical Engineering, vol. 14. Elsevier Sc. Publ. Co, Amsterdam, pp. 608–657.

Radbruch-Hall, D., Varnes, D.J., Savage, W.Z., 1976. Gravitational spreading of steep-sided ridges ("Sackung") in Western United States. Bulletin of the International Association of Engineering Geology 14, 23–35.

Radbruch-Hall, D., Varnes, D.J., Colton, R.B., 1977. Gravitational spreading of steep-sided ridges ("Sackung") in Colorado. Journal of Research of the U.S. Geological Survey 5 (3), 359–363.

Savage, W.Z., Varnes, D.J., 1987. Mechanics of gravitational spreading of steep-sided ridges (Sackung). Bulletin of the International Association of Engineering Geology 35, 31–36.

Schwab, J.W., Kirk, W. 2002. Sackungen on a Forested Slope, Kitnayakwa River. In: Forest Sciences, Extension Note # 47, British Columbia Ministry of Forest Prince Ruper Forest Region (ed). www.for.gov.bc.ca/prupert/Research 1–6.

Sébrier, S., Ghafiri, A., Bles, J.L., 1997. Paleoseismicity in France: fault trench studies in a region of moderate seismicity. Journal of Geodynamics 24 (1–4), 207–217.

Seward, D., Mancktelow, N.S., 1994. Neogene kinematics of the central and western Alps: Evidence from fission-track dating. Geology 22, 803–806.

Siame, L., Braucher, R., Bourlès, D.L., Bellier, O., Sébrier, M., 2001. Datation des surfaces géomorphologiques repères par le ^{10}Be produits In-Situ: implications tectoniques et climatiques. Bulletin de la Société Géologique de France 172 (2), 223–236.

Solonenko, V.P., 1977. Landslides and collapses in seismic zones and their prediction. Bulletin of the International Association of Engineering Geology 15, 4–8.

Stewart, I.S., Sauber, J., Rose, J., 2000. Glacio-seismotectonics: ice sheets, crustal deformation and seismicity. Quaternary Science Reviews 19, 1367–1389.

Sue, C., Tricart, P., 2002. Late-Alpine syncollisional extension in the core of the Western Alps inferred from faulting analysis. Journal of the Geological Society (London) 159, 61–70.

Sue, C., Thouvenot, F., Fréchet, J., Tricart, P., 1999. Earthquake analysis reveals widespread extension in the core of the Western Alps. Journal of Geophysical Research 104, 25611–25622.

Tabor, R.W., 1971. Origin of ridge-top depressions by large scale creep in the Olimpic Mountains. Geological Society of America Bulletin 82, 1811–1822.

Thouvenot, F., Fréchet, J., Jenatton, L., Gamond, J.F., 2003. The Belledonne Border Fault: identification of an active seismic strike-slip fault in the western Alps. Geophysical Journal International 155, 174–192.

Van Rensbergen, P., De Batist, M., Beck, Ch., Chapron, E., 1999. High-resolution seismic stratigraphy of glacial to interglacial fill of a deep glacigenic lake: Lake Le Bourget Northwestern Alps, France. Sedimentary Geology 128, 99–129.

Varnes, D.J., Radbruch-Hall, D., Savage, W.Z., 1989. Topographic and structural conditions in area of gravitational spreadig of ridges in the western united states. U.S. Geol. Surv. Prof. Paper, vol. 1496, pp. 1–28. Washington.

Varnes, D.J., Radbruch-Hall, D.H., Varnes, K.L., Smith, W.K., Savage, W.Z., 1990. Measurement of ridge-spreading movements (sackungen) at Bald Eagle Mountain, Lake County, Colorado, 1975–1989. U.S. Geological Survey Open-File Report 90-543, p. 13.

Vengeon, J.M. 1998. Déformation et rupture des versants en terrain métamorphique anisotrope. Apport de l'étude des Ruines de Séchilienne. Unpublished PhD Thesis, Joseph Fourier University, Grenoble 1, 186 p.

Vibert, C. 1987. Les glissements de Lax le Roustit et Saint Etienne de Tinée. Unpublished PhD Thesis, Ecole des Mines de Paris, 206 pp.

Wells, D.L., Coppersmith, K.J., 1994. New empirical relationships among magnitude, rupture length, rupture width, rupture area, and surface displacement. Bulletin of the Seismological Society of America 84, 974–1002.

Zischinsky, U., 1966. On the deformation of high slopes. Proc. 1st Conf. Int. Soc. Rock Mech., Lisbon, Sect., vol. 2, pp. 179–185.

Zischinsky, U., 1969. Über Sackungen. Rock Mechanics 1, 30–52.