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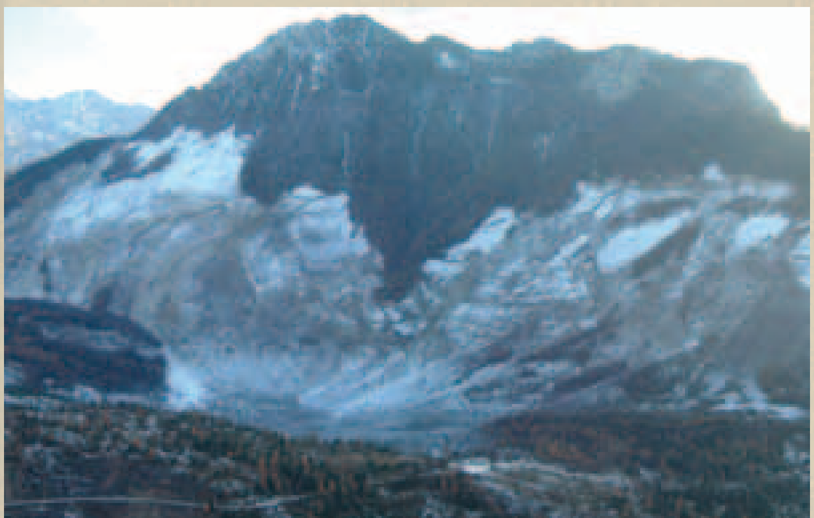
32nd INTERNATIONAL GEOLOGICAL CONGRESS

GEOMORPHOLOGY AND SLOPE INSTABILITY IN THE DOLOMITES (NORTHERN ITALY): FROM LATEGLACIAL TO RECENT GEOMORPHOLOGICAL EVIDENCE AND ENGINEERING GEOLOGICAL APPLICATIONS



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Post-Congress **P22**

Field Trip Guide Book - P22

Florence - Italy
August 20-28, 2004

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GEOMORPHOLOGY AND SLOPE INSTABILITY IN THE DOLOMITES (NORTHERN ITALY): FROM LATEGLACIAL TO RECENT GEOMORPHOLOGICAL EVIDENCE AND ENGINEERING GEOLOGICAL APPLICATIONS

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Front Cover:
Vajont landslide (photo A. Pasuto)

Leader: M. Soldati
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Introduction

Lisa Borgatti and Mauro Soldati

The superb landscape of the Dolomites is the result of their geology, which is different from the rest of the Alps and most of the mountains in the world. Here, pale dolomitic cliffs stand above gentle slopes often made up of dark volcanoclastic rock types.

The name “dolomite”, which applies to the mineral and to the rock, derives from Deodat de Dolomieu, the French chemist who formerly discovered it.

The Italian Dolomites are located in the south-eastern sector of the Alpine chain (Figures 1, 2). They are bounded to the north by the River Rienza, on the east by the River Piave, on the south by the River Brenta and the terminal section of the Torrent Fersina and on the west, for a limited extension, by the River Isarco (between Bressanone and Bolzano) and then by the River Adige. These boundaries correspond to the Pusteria Valley, Piave Valley, Val Sugana and Adige Valley, respectively. The main mountain groups are: Croda Rossa Group (3139 m a.s.l.), Dolomiti di Sesto (3152 m), Odle Group (3025 m), Tofàne Group (3243 m), Cristallo Group (3216 m), Antelào-Sorapìs Group (3263-3205 m), Catinaccio Group (3004 m), Sassolungo Group (3181 m), Sella Group (3151 m), Marmolada Group (3342 m), Dolomiti di Zoldo (Mt. Pelmo, 3168 m and Mt. Civetta, 3218 m) and Pale di San Martino Group (3193 m).

From the climatic standpoint the Dolomites are quite varied owing to their wide altitude range (1200–3200 m). The area has an Alpine climate, characterised by cold winters and mild summers. Precipitation patterns are typical of the Alpine climate, too: late springtime and summer are the most rainy periods, with a peak in July. Precipitation is intense and concentrated and occurs mostly during violent rainstorms. A secondary precipitation maximum occurs in November, following the prolonged autumn rains. Permanent snowfall in the valley bottom normally begins in December and lasts until April. On the highest peaks this period usually lasts a couple of months longer. The average snow thickness is 50 to 100 cm.

The annual average temperature is around 6°C. In summer, temperature ranges from more than 25°C during the day to 7°C during the night.

The field-trip is focused on geomorphology and engineering geology applied to slope instability in the Dolomites. Along the route to the Dolomites, vast landslide bodies occurring in geological and geomorphological contexts different from the dolomitic ones, will also be shown.

The main aim is to show significant cases of mass movements of various types, sizes and ages which

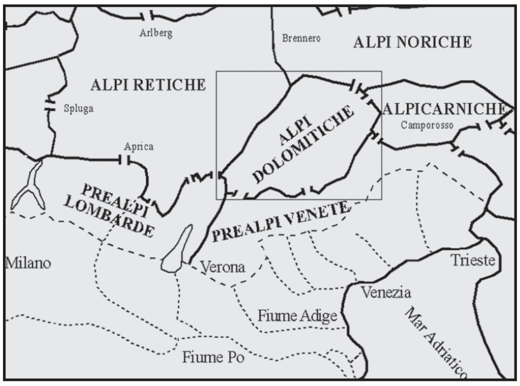

Figure 1 - Sectors of the Eastern Alpine system (after Bertoglio and De Simoni, 1979; modified).

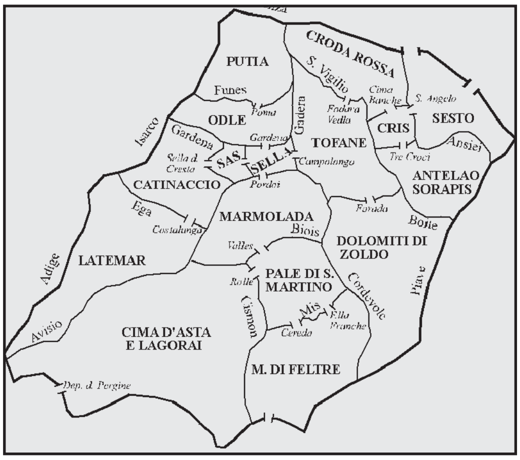

Figure 2 - The mountain groups of the Alpine Dolomites (after Bertoglio and De Simoni, 1979; modified).

have affected the dolomitic valleys since the retreat of the Last Glacial Maximum (LGM) glaciers, including the recent catastrophic Vajont landslide which occurred in 1963. Mass movements often interfere with transport infrastructures and developed areas where a large number of tourists are present during both winter and summer. In these conditions of high vulnerability, even mass movements of modest magnitude, like debris flows, could have severe and sometimes unacceptable consequences.

The secondary aims of the field trip are, on the one hand, to highlight the relationships between geological structures and landscape evolution and, on the other hand, to show the influence of Holocene climatic changes on slope instability processes.
The team leading the field trip has had considerable experience in co-ordinating national and international projects dealing with basic or applied geomorphology, with particular interest in slope instability processes.
This guide has not been arranged into single and specific stops as, at the time of printing, the number of participants was not yet known. Therefore, the Authors have preferred to give the text a more general layout, since more detailed information will be provided during the excursion. Furthermore, depending on the number of participants and means of transport used, some of the excursion routes may take across rough or wet mountain terrain. Considering also that weather conditions in the Alps may change rapidly, an umbrella, weather-resistant clothes and mountain-proof footwear could be necessary, as well as a professional field-geological outfit.

Field references
Topographic maps
Istituto Geografico Militare Italiano, Sheets at the 1:50,000 scale: 063 Belluno, 046 Longarone, 039 Cortina d'Ampezzo, 028 La Marmolada.
Geological maps
Servizio Geologico d'Italia (1996). Geological map of Italy at the 1:50,000 scale. Sheet 63 "Belluno".

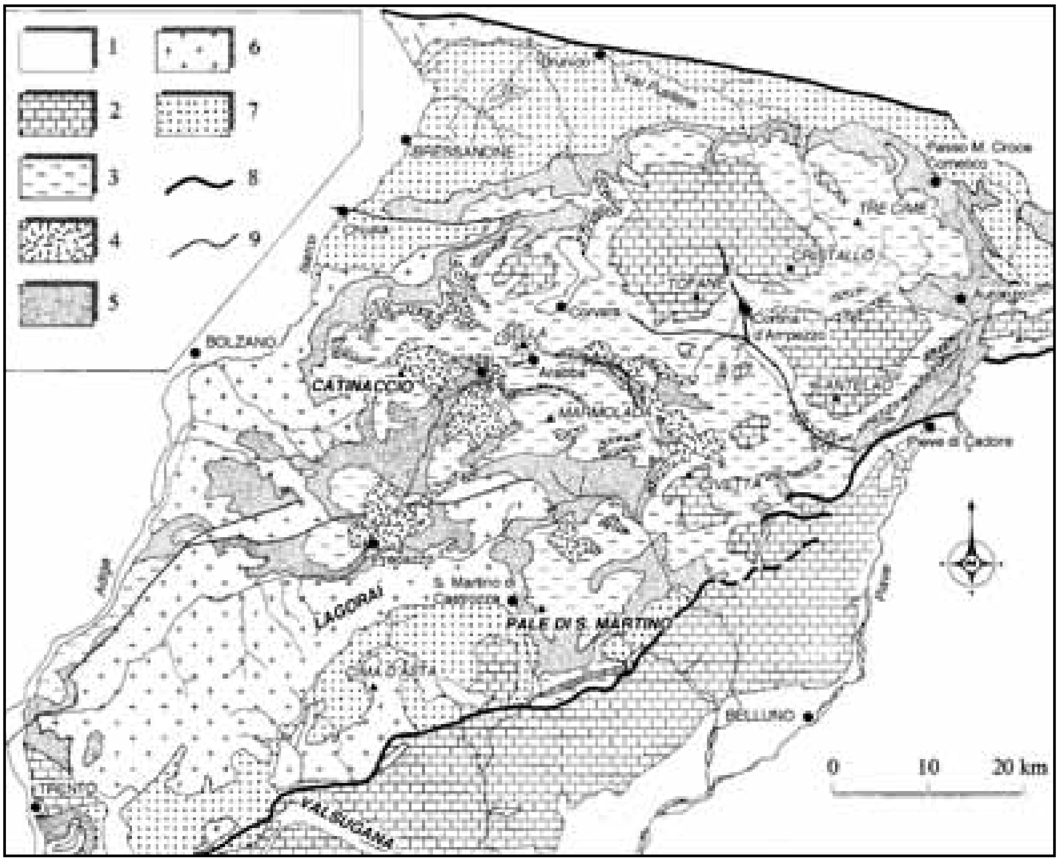

Figure 3 - Geological schematic map of the Dolomites. Legend: 1) Plio-Quaternary deposits; 2) dolomitic and carbonatic rocks, Norian-Cretaceous (Dolomia Pricipale, Dachstein Limestone, Calcari Grigi, Ammonitico Rosso); 3) dolomitic and clastic rocks, Carnian (Dolomia Cassiana, S. Cassiano Fm., Dürrenstein Dolomite, Raibl Fm.); 4) dolomitic, clastic and volcanic rocks, Ladinian (Sciliar Dolomite, Livinallongo Fm., volcano-clastic deposits); 5) clastic rocks, Permian-Lower Anisian (Arenarie di Val Gardena, Bellerophon Fm., Wengen Fm.); 6) volcanic rocks, Permian (rhyolites, rhyodacites and dacites); 7) Crystalline basement, Carboniferous (phyllites, micaschists and paragneisses); 8) main tectonic lineaments; 9) secondary tectonic lineaments (after Carton and Soldati, 1993).

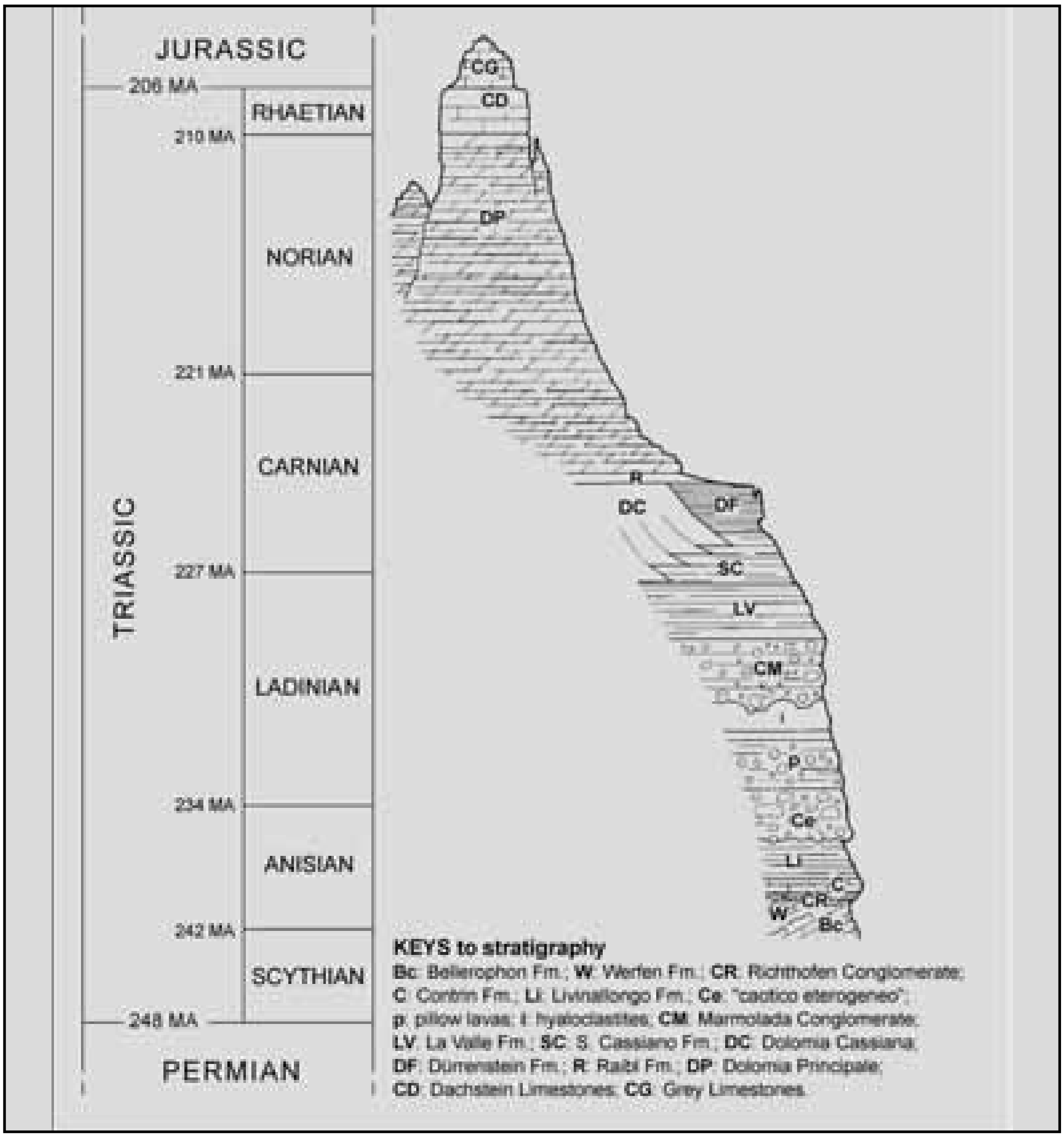

Figure 4 - Stratigraphic sequence outcropping in the Dolomites (after Soldati et al., 2004, modified).

Servizio Geologico d'Italia (1970). Sheet 11 "M. Marmolada". Geological map of Italy at a 1:100,000 scale, 2nd edition.
Servizio Geologico d'Italia (1977). Sheet 028 "La Marmolada". Geological map of Italy at a 1:50,000 scale.
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Geological features of the Dolomites

Chiara Siorpaes

From a geological point of view, the Dolomites belong to the Southern Alps, a segment of the African continental margin which was deformed from the Late Cretaceous to the Quaternary. Deformations had an early European vergence and an African vergence during the Tertiary. Two thrust belts are evident to the west and to the east of the Trento plateau (Figure 3).

The stratigraphic sequence consists of a Permo-Mesozoic sedimentary cover which was deposited in an ancient warm tropical sea and which unconformably overlies the Hercynian metamorphic basement (Figure 4). To the west and to the north they are bounded by the Neogene Periadriatic Lineament (Giudicarie Line and Pusteria Line, respectively) and to the south and to the south-east by the Valsugana and Pieve di Cadore Thrust: from a geological point of view these structural elements separated the Dolomites from the Alps s.s. to the north and from the Pre-Alps to the south.

The geological history of the area may be subdivided into three periods: first, the deposition of different sediments into a tropical see during the Permo-Triassic, Jurassic and Cretaceous, later the mountain chain building inside the Alps during the Tertiary and lastly, the modelling of the present landscape by weathering and erosion, glaciation, running water, rock falls, slides, and mass wasting in the last million years.

In the Early Permian the western part of the Dolomites was affected by volcanism and therefore it mostly consists of rhyolitic lavas (“Piattaforma Porfirica Atesina” or “Porfidi Quarziferi”); in the eastern part, only locally have these volcanic rocks erupted and actually crop out. The Piattaforma Porfirica lies on top of the Ponte Gardena Conglomerate which crops out discontinuously as does the underlying Hercynian metamorphic basement, which consists of grey and dark grey schists, phillites and paragneisses.

At the end of the Permian eruptions, the paleogeographic setting changed and sedimentation processes are recorded by the Val Gardena Sandstones and the Bellerophon Formation. The red sandstones (Val Gardena Sandstones) and the gypsum and black bituminous limestones (Bellerophon Fm.) indicated an evolution from alluvial plain to sabkha, and then to lagoon and shallow marine conditions. During the Triassic, a marine transgression westwards occupied a widespread region in south-central Europe and, in the Dolomites, sedimentation recorded shallow marine carbonate and terrigenous deposits mainly consisting of limestones, marly limestones, dolomites, gypsum, oolitic limestones, red sandstones and oolitic dolomites belonging to the Werfen Fm. (Scitian). White to light grey peritidal dolomite (Lower Serla Dolomite) stratigraphically overlie the Werfen Fm.

The Anisian sedimentation was mainly controlled by tectonics. Two zones with different subsidence can be identified: the Badioto-Gardenese High, corresponding to the western Dolomites, and the Cadore Basin, corresponding to the eastern Dolomites (De Zanche, 1990). Tectonics and relative sea level changes gave rise to complex paleogeography which is documented by units deposited in basin, lagoon, peritidal and continental environments. The units consist of terrigenous-carbonate sediments and carbonate shelves even during the Ladinian and Carnian. Massive Sciliar Dolomite and Cassian Dolomite, such as Catinaccio, Latemar, Sciliar (Schlern), Sassolungo (Langkofel), Odle (Geisler), Putia (Peitler), Pale di San Martino, Marmolada, and the lower part of the Sella, built up carbonate reefs, quite similar to the present-day Maldives or Bahamas, while the surrounding basins were filled by pelagic carbonate units belonging to the Buchenstein Group and to the Wengen Group p.p. Siliciclastic and volcanoclastic rocks together with ash tuffs and tuffites (“pietra verde”) rhyolitic and rhyodacitic in composition (Viel, 1979), are also present.

In the Upper Ladinian the whole region underwent a strong magmatic phase. In the western part of the Dolomites, basalt eruptions and comagmatic bodies such as granites, monzonites and syenites, can be recognised (De Zanche, 1990). In the eastern part, the Wengen Group includes volcanoclastic, siliciclastic, volcanic and subordinate marly rocks and pelagic limestones which progressively filled the basins (Viel, 1979; Pisa *et al.*, 1980). Post-volcanic evolution is again represented by a reef-basin model: carbonate platforms of “Cassian Dolomite” were built up by algae, corals and other marine invertebrates whereas the S. Cassiano Fm. (marls, clay, extrabasinal siliciclastic and resedimented calcarenites) was deposited in the gradually shallowing upward basins surrounding the platforms.

During the Upper Carnian, the Raibl Fm. (Upper Carnian) “may represent the reconquest of the Alpine domain by the German facies which had been previously exiled by Middle-Triassic tectonics” (Pisa *et al.*, 1980). It includes polychrome shales, aphanitic dolomites and limestones covering the entire region.

The deposition of this unit prepared the morphological conditions suitable for the formation of the peritidal dolomites belonging to the Dolomia Principale (Norian-Rhaetian?; De Zanche, 1990). The latter consists of white and light grey cyclic dolomites containing stromatolithic lithozones and fossils such as *Megalodon sp*. The famous mountains surrounding Cortina d'Ampezzo (Tofàne, Cristallo, Pomagagnon, Cinque Torri, Croda da Lago, Averau), the Tre Cime di Lavaredo (Drei Zinnen) in Sesto (Sexten), and the lower part of Pelmo and Civetta near Alleghe are made up of Dolomia Principale.
The Uppermost Triassic is represented by the Dachstein Limestone. It includes shallow water limestone containing large bivalves such as *Megalodon sp.*, *Dicerocardium sp.* and corals.
The Jurassic formations are composed of shallow-water well-bedded limestones (micritic, bioclastic and oolitic limestones) belonging to the Calcari Grigi (Lias) and by nodular red pelagic ammonite-rich limestones belonging to the Ammonitico Rosso (Dogger-Malm).
The Lower Cretaceous rocks are present only in the Altipiani Ampezzani area, Piz Boè and Gardenaccia mountain groups and they are made up of pelagic grey-blue marls, marly limestones and reddish to greyish silty limestones belonging to the Marne del Puez; the Ra Stua Flysch and the Antruilles Fm. (Upper Cretaceous) crop out only in the Altipiani Ampezzani area and they record turbiditic fine-grained sandstones, turbiditic bio-calcarenites and red marly limestones of deep-sea conditions.
In the Dolomites the Upper Oligocene - Lower Miocene Monte Parei Conglomerate is the only Tertiary unit. It crops out near the summit of the Col Bechei relief, between 2550 and 2580 m. The unit includes polygenic conglomerates and sandstones which unconformably overlie Jurassic limestones. They were probably deposited in shallow marine waters along an irregular coastline.
The Dolomites are located in the eastern Southern Alps and their tectonic history recorded not only the deformations related to Alpine compressional phases, but also shows many synsedimentary events occurring after the post-Variscan rift (Winterer and Bosellini, 1981).
During the Permian and Triassic the whole region underwent a rifting phase which is documented by paleogeographic and paleostructural elements such as the Atesina Platform and the Carnico-Bellunese Basin (Bosellini, 1965). The different thickness of the same units on the platform and the basin confirms this hypothesis (Bosellini and Doglioni, 1986). As described before, this complex Triassic paleogeography was partly controlled by extensional tectonics. Triassic compressive features are also documented in many parts of the Dolomites (Castellarin, 1982; Doglioni, 1985, 1987; Picotti and Prosser, 1987; Doglioni and Neri, 1988). The compressive structures have been interpreted as the results of a sinistral transpression, N70 trending (Doglioni, 1984).
During the Upper Triassic and Jurassic the Southern Alps became part of a southern passive continental margin when the Ligurian Ocean opened to the West. As a consequence, the thickness of the Jurassic sequence varies considerably from the western to the eastern Dolomites. In fact, in the Sella Massif the thickness of the Dolomia Principale is 300 metres and the Ammonitico Rosso unconformably overlies a Dachstein Limestone tilted block. On the other side, in the Altipiani Ampezzani area and in the Eastern Dolomites, the Dolomia Principale reaches 1000 metres and the Dachstein Limestone together with the Calcari Grigi is 600 metres thick. These remarks indicate the existence of a buried NNW-SSE striking growth fault which is probably located along the Badia Valley (Doglioni, 1987).
During the Tertiary (60 to 5 million years ago), the collision of the African and European continents displaced, crumpled and uplifted the marine sediments, thus giving rise to the Alpine chain. The Dolomites were deformed by two major compressional tectonic phases: the W-WSW-verging phase and the S-ESE-verging Neogene phase. In fact, the Upper Oligocene-Lower Miocene Monte Parei Conglomerate sutures the W-verging structures and is, in turn, deformed by the S-verging one.
During the W and WSW verging tectonics only the sedimentary cover of the Dolomites was involved and the main *décollement* zone is located within the gypsum facies of the Permian Bellerophon Fm. Moreover, important flats are also located at the top of the Dolomia Principale and within the Cretaceous marls. During this tectonic phase Upper Triassic to Cretaceous rocks were folded and thrust and some of the overthrusts are known as *Gipfelfaltungen* AUCTT. (summit overthrusts). The most characteristic structures are the Tofana III klippe, the Remeda Rossa-Piccola Croda Rossa klippe, the Col Bechei W-verging recumbent fold in the hanging-wall of a low angle thrust, the Sella Overthrusts (Doglioni, 1990).
The later Valsugana Neogene phase took place in S-

verging thrusts and folds involving the metamorphic basement with pop-up geometry. In the eastern part of the Dolomites the Valsugana Line was transferred along a sinistral transpressional zone into the Pieve di Cadore Line (Gianolla *et al.*, 1988). The major Neogene structural features are the Passo Falzarego and the Valparola Lines, the Antelào-Selva di Cadore Line and a set of wide and gentle ENE-WSW trending synclines (Gardenaccia, Sella, Pelmo and Sorapìs-Marmarole Synclines).

Not all the Dolomites reacted in the same way during the mountain-building process. The geometry of each structural phase is controlled by inherited stratigraphic and tectonic discontinuities, resulting in complicated polyphase structures which are the most important features controlling the Quaternary evolution of the Dolomitic region.

Geomorphological features of the Dolomites

Mauro Soldati and Alberto Carton

The spatial distribution of very different geological formations (Figures 3, 4), the intense tectonisation as well as the complex Quaternary evolution of the region determined a significant variety of landforms and surficial deposits (Carton and Soldati, 1993), which are described below (see Figure 5).

Structural landforms

The geomorphological features of the Dolomites are in most cases closely connected with geological structures. The tectonic events which have affected the region since the Cretaceous folding, faulting and overthrusting of thick piles of sediments, are still clearly legible in the landscape. As for faults, they have in several cases determined the direction of valleys and river cuts, as in the case of the Funès Valley, Tires Valley, upper Cismon Valley and S. Vigilio Valley. In addition, the presence of overthrusts has determined the formation of saddles, like those of the Valparola Pass and Falzarego Pass.

Furthermore, the alternation in space and time of different sedimentation environments together with recurrent magmatic activity, led to the present vertical and lateral contact between rocks having different mechanical behaviour, which has been so significant for the evolution of the landscape. A particular feature of the Dolomite landscape is given by morphoselection phenomena, which can be observed in the areas where Ladinian carbonate formations are cut through by volcanic dykes (Latemar and Costabella Group). Another typical example is given by the “cengia” of the Sella Group, where the pelitic rocks of the Raibl Fm. are interposed between Dolomia Cassiana at the bottom and Dolomia Principale at the top. However, all over the western Dolomites strong morphological contrasts can be observed on slopes where dolomitic rocks overlie clastic formations, like in the area of Cortina d’Ampezzo, where very steep-sided groups like Tofàne, Cristallo, Pomagagnon, Sorapìs and Antelào lie over the highly erodible marls and clays of the S. Cassiano Fm.

The location of some passes of the Dolomites is also linked to the presence of more easily erodible rocks; this is, for example, the case of the Gardena Pass, Sella Pass, Pordoi Pass and Campolongo Pass, which are shaped on clastic rocks weaker than the surrounding dolomites.

Glacial landforms

During the Quaternary, glaciers repeatedly (at least four times) occupied the Dolomite valleys and left very evident traces of their presence. However, at present there is no clear evidence of glacial deposits or erosional landforms older than the last glaciation and datable with certainty. Only limited outcrops are located in the area of Brunico (Klebesberg, 1956) and perhaps along the River Piave in the vicinity of Ponte nelle Alpi (Penck and Brückner, 1909). The oldest observable landforms are almost exclusively attributable to the Last Glacial Maximum (LGM). These are the forms that previous Authors attributed to the Würm glacial period. During this period, ice masses reached relevant heights in all the Dolomite valleys with only the highest peaks and plateaus remaining uncovered. The glaciers coming from the large Dolomite groups joined together and made up a network of intersecting glacial tongues. Some flowed over the present dolomitic passes, which functioned as transfluent saddles at the time: Sella Pass (2244 m), Pordoi Pass (2239 m), Gardena Pass (2121 m), Falzarego Pass (2105 m) and other minor passes such as Campolongo Pass (1875 m), Costalunga Pass (1745 m) and S. Pellegrino Pass (1918 m). The thickness of the ice in the Dolomite region reached over 1500 m in certain periods.

Glacial deposits, ascribable to the Last Glacial Maximum, are found only in the southern sectors of the Dolomites and not in the central or northern valleys. In the latter areas, the glacial surface was above the permanent snow-line and therefore completely contained within the alimentation area. In these same areas, only lodgement till and scattered material left during the general melting of the ice can be attributed

Figure 5 - Geomorphological schematic map of the Dolomites. Legend: 1) orography; 2) glaciers; 3) main glacial deposits; 4) movement direction of the Pleistocene glaciers; 5) rock glaciers; 6) ice- and snow-related phenomena; 7) main landslides; 8) areas intensely affected by karst phenomena; 9) troughs; 10) deeply eroded fluvial valleys; 11) talus cones and alluvial fans (after Carton and Soldati, 1993; modified).

to the peak of maximum advancement.

The most evident traces linked to glacial morphogenesis observable today, can be attributed to the subsequent melting phases, which took place intermittently and in a discontinuous manner, as defined in the classic model introduced by Penck and Brückner (1909) for the Alpine region, that is Lateglacial stadial phases. The reduction, followed by the complete disappearance of glacial masses, exposed various erosion landforms, among which, besides the well known U-shaped valleys, the most frequent are: steps, hanging valleys, sharp rocky crests and cirques. The freshness and permanence of some of these landforms is related to the type of bedrock. For this reason, their presence in the Dolomite area is not homogeneous, but follows the distribution of the Ladinian carbonate rocks. Traces of glacial erosion are better preserved in the latter, compared with volcanic rocks. Hanging valleys are particularly evident at the junction of small valleys with a main valley: the junction occurs through a step (outlet step). Significant examples of secondary valleys, hanging above a principal valley since they were formed by minor tongues, are present along the slopes of the Fassa Valley at the confluences with the Contrin Valley, Ciampac Valley and Duron Valley. Glacial cirques are scattered almost everywhere, especially near the valley heads. The cirques vary from semi-circular to elongated shapes and are isolated or arranged in groups or in a step-like morphology. They often show very steep ridges and offer suitable climatic conditions for the preservation and permanence of snow patches until late summer or even throughout the entire year. The flanked glacier cirques present in the Marmarole Group, the isolated and well formed Croda Rossa cirque and the Travignolo cirque, which is deeply carved in the north-western flank of the Pale di S. Martino Group, are mentioned here only as a few examples. Glacial deposits ascribable to the retreat phases of the LGM (Lateglacial AUCTT.) are quite frequent in the Dolomites (Castiglioni, 1964). They can be found within many valleys, mainly in sequences of frontal ridges. They normally present a rounded and not very prominent topographic profile, especially in comparison with the moraines deposited during the Little Ice Age. Stadial moraines enclose lakes Misurina (Cristallo Group), Carezza (Latemar Group), S. Pellegrino and Lagusel (Marmolada Group), d'Ayal in the vicinity of Cortina d'Ampezzo, Valparola (Tofàne Group), Calaita (Cima d'Asta-Lagorai Group) (Figure 5).

Other accumulations having an overall appearance similar to landslide deposits, but distributed over vast areas, should also be mentioned. They result from rock falls that occurred on the ancient glaciers thus becoming part of the moraines. Significant examples of this type are found below, in the vicinity of the main Dolomite passes.

Landforms related to recent or present-day glacier activity are not lacking, even if the existing glaciers in the Dolomites are small in size. Their limited development is mainly due to the Alps' decrease in mean altitude from west to east, with the resulting reduction of surfaces lying within the altimetric intervals with favourable climatic conditions for the formation and maintenance of glaciers. The Dolomites have a relatively low mean altitude for glaciation. Moreover, although there are numerous peaks in the Dolomites rising to above 3000 m, high relief energy is an unfavourable element for glaciation: the steepness of the walls greatly reduces the slope surface area lying within altimetric intervals of accumulation. Only a few glaciers are located on relatively upward-sloping surfaces suitable for receiving glaciers that extend over an entire slope.

The mountain groups which still have extremely small ice masses today are listed in the following, from west to east: Popera, Cristallo, Tofàne, Sorapìs, Marmarole, Antelào, Marmolada, Pelmo, Civetta and Pale di S. Martino. Although there were glacierets present at the beginning of the century, the Croda Rossa, Sassolungo and Sella groups have no glaciers at all today.

Periglacial landforms

Periglacial landforms are also very typical of the Dolomites. At these altitudes temperature plays an important rôle, both for its stability minus 0°C (with frozen ground for part of the year) and for its variability with passages above and below zero (frost-thaw cycles). Talus cones and scree slopes are a very common and sometimes spectacular feature of the region, linking at their base many mountain groups. These are due to frost shattering processes which have taken place since the LGM glaciers' retreat from the valleys. The extension, inclination and particle-size distribution depends very much on the lithology (i.e. physical and mechanical properties) of the source areas and on the characteristics of the tectonic jointing of the slope itself (Soldati, 1989). Other widespread periglacial landforms are protalus ramparts, which make up elongated ridges parallel to the slopes and are fed by debris falling from rock walls and sliding on snow patches.

Patterned grounds due to discontinuous frost action and/or to snow are locally observable at high altitudes on plateau areas, like on the Sella and Catinaccio groups. Gelifluction processes also give rise to evident morphological features (scars, lobes, small flows etc.) especially on slopes consisting of volcanoclastic and clayey formations. Several cases are observable in the high Cordevole Valley between Arabba and Pordoi Pass (Meneghel and Carton, 2002).

Avalanche cones are typical of the Dolomites, especially where structural discontinuities affect the slopes; this is the case of slopes characterised by dense jointing or by volcanic intrusions which have given rise to morphological discontinuities. Avalanches in some cases may threaten roads, villages, woods and crops, thus creating conditions of geomorphological risk.

In the Dolomites also rock glaciers are found, although they are not a typical landform of this mountain environment. They have the traditional shape of flows elevated over the surrounding terrain; the forms held to be active terminate at the front in a high, steep slope. They have an important rôle in climatic and paleoclimatic reconstructions, since they mark the lower limit of the discontinuous mountain permafrost. A comparison with other sectors of the Alpine chain reveals an almost complete absence of active and inactive forms; nearly all of them can, in fact, be considered relict forms. The altitudes of the frontal margins of the active rock glaciers decrease to 2225 m, whereas the inactive ones reach 1750 m. On the other hand, volcanic rocks (porphyries) appear to be quite suited for the formation of rock glaciers, as observable in the Lagorai Group (see Figure 5). The most spectacular form is found at S. Pellegrino Pass: it consists of a tongue with convoluted troughs which attains a length of some 1200 m.

Gravity-induced landforms

Gravity-induced landforms are widespread in the Dolomites and mainly consist of mass movements occurring from the Lateglacial to date (Soldati *et al.*, 2004). The frequency and magnitude of gravitational phenomena is proved to be very high in the last Post-glacial period when slopes no longer sustained by ice masses, were affected by many large scale landslides. Panizza (1973) showed how these landslides seem to be concentrated downstream of the confluence of glaciated valleys where "glaciopressure" might have been more intense, like downstream of the confluence of the T. Costeana and T. Boite valleys (south of Cortina d'Ampezzo), where large landslide accumulations are still visible today. Several mass movements have been recently dated in the area of Cortina d'Ampezzo and in the Upper Badia Valley (Alta Val Badia), where surficial deposits mostly consist of landslides. The analysis of the dated landslides has allowed correlations to be outlined between increases in landslide activity and climatic changes in the study areas. The first phase of marked slope instability is observed in the Preboreal and Boreal (about 11,500 to 8500 cal BP) and includes both large translational rock slides, which affected the dolomite slopes following the withdrawal of the LGM glaciers, and complex movements (rotational slides and flows) which affected the underlying pelitic formations probably favoured by high groundwater levels due to an increase of precipitation and/or permafrost melting. A second concentration of landslide events occurred during the Sub-boreal (about 5800 to 2000 cal BP), when slope processes mainly ascribable to rotational slides and/or flows took place in both the study areas. In the light of collected data, the events dated may be considered reactivations of older events linked to the phase of precipitation increase which has been documented in several European regions during this mid-Holocene period.

The recurrence in time of this landslide activity was certainly influenced also by other non-climatic factors, in particular geological-structural ones. First of all the spatial distribution of geological formations (Figure 3) with different mechanical characteristics must be taken into account. In particular, the occurrence of landslides is high where rigid and resistant rocks (dolomites, limestones etc.) showing brittle behaviour, overlie weaker rocks characterised by ductile behaviour (marls, clays etc.). This is typical of the area of Cortina d'Ampezzo and of the valleys around the Sella Group (Figure 5). Furthermore, also where the effects of tectonics were most intense, in correspondence with faults or overthrusts, mass movements have been favoured.

Gravitational landforms are also connected with the existence of deep-seated gravitational slope deformations, recently recognised in the Dolomites and particularly in the area of Cortina d'Ampezzo (Pasuto *et al.*, 1994; 1997). With respect to morphological evidence, they are generally characterised by the presence of trenches, gulls and uphill-facing scarps in the upper parts of the slopes and bulges in the lower parts (e.g. Tofàne, Lastoni di Formin and Faloria groups). It must be emphasised that deep-seated deformations may be the initial stage of large-scale mass movements whenever the deformation belt develops into a sliding plane. However it has been observed that the presence of these processes favours or induces "collateral movements" (rock falls, slides and flows) in the surrounding areas.

Karst landforms

Karst phenomena in the Dolomites are strictly related to structural and morphological conditions (Mietto and Sauro, 1989). Karst phenomena basically take place in the higher parts of the valleys and are mainly located in the high carbonate massifs (1200 to 3000 m), which rest on formations with little or no tendency at all to undergo karst processes. Among the numerous carbonate formations of the Dolomites, only the Marmolada Limestone, Sciliar Dolomite, Dolomia Principale and Calcari Grigi are particularly suited for karst processes. The formation containing gypsum of Permian and Triassic age (Bellerophon Fm.), outcropping on several slopes and in some valley floors has also undergone karst processes.

From a morphological standpoint, karst phenomena are prevalent on plateaus, in cirque floors or glacial hanging valley floors, on level summit areas, scarps and ridges. The most typical karst landforms in the Dolomites are large glacio-karstic depressions, blind valleys, fluvio-karstic dry valleys, dolines, glacio-karstic karren, landslide scree heaps with karren and slope karren. Undoubtedly, glacio-karstic depressions are the largest and most characteristic forms (length: 500 to 1500 m; depth: 20 to 60 m). The most typical examples of glaciokarstic depressions are found at the Lago Grande di Fùsses, Lago Remeda Rossa di Fosses, Tondi di Sorapìs, Lago Nero on Monte Popera, at Alpe dei Piani, north of Monte Paterno, in the Pale di San Martino near Forck di Sopra, Buse Alte, Sponde Alte, Buse di Coll'Alto, Riviera Manna and near Piani Eterni, in the Vette and the Feltrine Dolomites. Perennial or temporary lakes often exist inside the depressions.

Dolines are numerous: in the Alpe di Fosses their diameters range from 50 to 150 m. Very often, the margins of these forms have been modified by peri-glacial processes that are particularly effective in these environments. The karren are also very frequent. They occupy wide extensions, mainly on slightly inclined slopes, in highland plains and the floors of the glacio-karstic depressions. They are often complex forms affected by the structural features of the rocks involved, the modalities of glacial erosion and also by karst morphogenesis.

Field itinerary

DAY 1

From Venice to Fadalto
Mauro Marchetti

After leaving the city of Venice and its characteristic lagoon, we proceed to the north along the A27 motorway, across the alluvial deposits of the Venetian watercourses. We cross a mildly SE inclined slope, run through by numerous streams fed also by groundwater. At the surface these alluvial deposits are made up of silty-clayey sediments. In the Holocene the area nearest to the coastline has undergone a rather complex evolution, starting from the Flandrian transgression – which caused a retreat of several kilometres of the coastline – followed by slow progradation up to the present situation. This trend was partly counterbalanced by natural (due to sediment compaction) and artificial (following land reclamation activities and considerable drawdown of the water table). Eastward of the city of Treviso, the sediments grow coarser, with particle size distribution ranging from sand to gravel. These deposits make up the surface portion of the large alluvial fan of present-day River Piave. The main morphological features are large fan-shaped fluvial ridges, with the apex located at the river's outlet into the plain. The mainly gravelly grain size distribution in the fan's proximal portion, implies considerable water percolation through these deposits. This groundwater eventually emerges downstream, beyond the fan's lower boundary. The River Piave is a significant example of overimposed hydrography, since it cut its course through the Montello reliefs, between Conegliano and Montebelluna, with a rate equal to the that of the chain's uplift. The living Montello anticline is, in fact, characterised by fluvial terraces which were uplifted and arched on the west side, whereas the River Piave underwent a diversion to the east, up to its present position. Therefore, the course of the River Piave was changeable during the Quaternary. The uplift of the Pre-Alpine chain created an obstacle in the path of the river's downflow, forcing it in two directions: the eastward direction across the S. Croce Valley and Lapisina Valley (which we will visit during our excursion) has now been abandoned, whereas the westward direction, across the Quero narrows, which in the past was occupied with continuity by its tributary River Cordevole, is still active.

After leaving Conegliano to the west, we proceed towards the Cansiglio reliefs, travelling across the moraine hills of Vittorio Veneto, already recognised as such by Penck and Brückner (1909). This moraine amphitheatre bears witness to the outlet of a glacial tongue into the plain through the Vittorio Veneto narrows. Once across this anticline, we reach the S. Croce Valley, where evidence of glacial and fluvial modelling is still observable.

Stop 1.1:

Fadalto Landslide
Giovanni Battista Pellegrini and Nicola Surian

Introduction

The Fadalto landslide is located in the upper part of the Lapisina Valley, a transverse valley separating the Belluno Pre-Alps from the Cansiglio Plateau

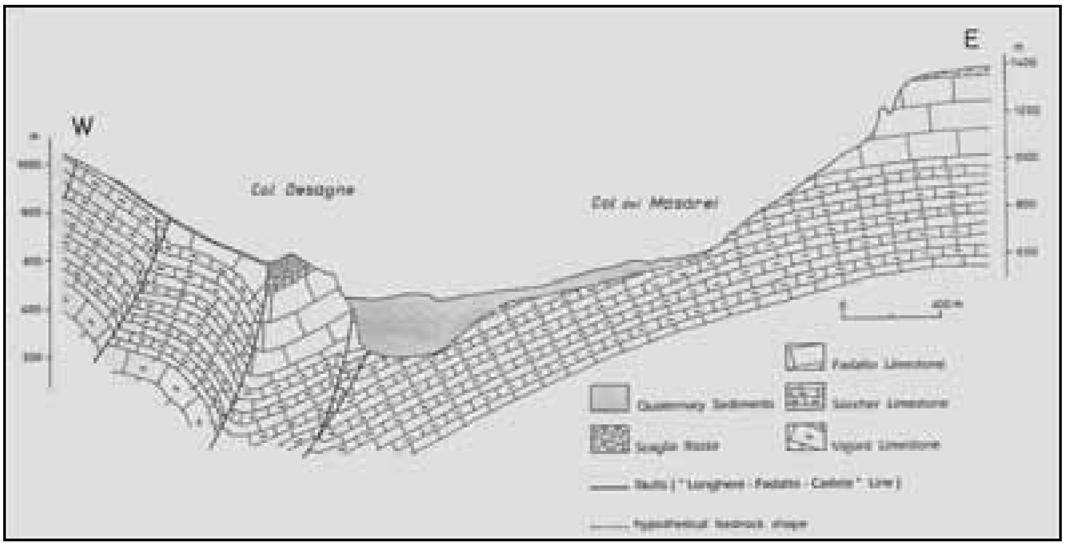

Figure 6 - Geological section across the Lapisina Valley (after Pellegrini and Surian, 1996).

(Venetian Pre-Alps). Since the retreat of the Würmian glaciers, 14,000-13,500 years BP in the nearby Vallone Bellunese (Surian and Pellegrini, 2000), the River Piave has flowed only through the Vallone Bellunese, abandoning its course through the S. Croce – Lapisina Valley. Even if glacial over-deepening could have taken place, the data available on bedrock morphology suggest that the Fadalto landslide could be the main cause of valley damming, the formation of the Santa Croce Lake and the subsequent change in the course of the River Piave. Nowadays, the Lapisina valley-bottom is filled with debris from several large landslides and the Fadalto Col is a wind gap. Along the valley there are three lakes, following the axis of the valley: Lake Morto at 274 m, Lake Restello at 177 m and Lake Negrisiola at 160 m. These natural lakes have been transformed into hydroelectric basins.

The Fadalto landslide is a complex process involving several types of movement (rock slide, debris slide etc.) which has been reactivated on several occasions since the Late Glacial (Pellegrini and Surian, 1996; Pellegrini *et al.*, 2004). The landslide affects the eastern slope of the Lapisina Valley, which is mostly built up of Fadalto Limestone, a bioclastic calcarenite and calcirudite with strata dipping towards the valley bottom. The main scarp is about 7 km long with vertical walls reaching a height of 400 m, whereas the accumulation zone is about 1.6 km^2. The highest point of the scarp is at 1400 m in altitude whilst the lowest point of the accumulation is at 305 m.

Geological setting

The tectonic structures had a main role in the Neogenic deformation of this sector of the Venetian Pre-Alps. The layout of the Lapisina Valley has in fact been conditioned by the presence of a structural depression existing between the “ramp anticline” of Col Visentin, which borders the Belluno Prealps to the east, and the Cansiglio massif. The complex syncline of the valley bottom, dislocated by a system of inverse and transcurrent faults connected with the line “Longhere-Fadalto-Cadola” (Figure 6), has meant that the Lapisina Valley has developed as a transverse valley in the Pre-Alpine chain.

The eastern slope of the Lapisina Valley is mostly made up of Fadalto Limestone (bioclastic calcarenite and calcirudite stratified in beds). The strata, with an average N 300° direction, dip towards the valley bottom with average inclinations of around 25°, while the boundary with the underlying Soccher Limestone (finely stratified micritic limestone, with thin argillaceous interstrata in the upper part) is located at about 1000 m (Figure 6). No faults were found in this slope of the valley. However, here a series of joint planes is visible at a mesostructural scale in the lower part of the slope, while at a larger scale they look like deep, vertical, N-S oriented cracks which isolate portions of the wall occasionally subject to collapse. The joint planes are all approximately perpendicular to those of the bedding. The most frequent ones show a N 120° direction; also those around N 160° and N 270° are statistically significant, while there is a smaller number of planes with N 200° direction. These joint systems intersect the bedding planes, subdividing the rock into blocks ranging in volume from a few cm^3 to hundreds of m^3. Gravity acts together with a component of extension oriented towards the valley bottom and parallel to the bedding and opens wide the joints intersecting the layers, thus exposing the slope to new failures.

Geomorphological analysis of the Fadalto landslide

In the past many Authors argued about the nature of the deposits that shut off the upper part of the Lapisina Valley and on the role of the landslide in the formation of Lake Santa Croce. Some suggested a glacial origin for these deposits (e.g., Dal Piaz, 1912; Venzo, 1939), whereas others considered the accumulation as the result of a landslide (e.g., Brückner, 1909; Panizza, 1973; Eisbacher and Clague, 1984). More recently, the landslide was studied by Pellegrini and Surian (1996) and Pellegrini *et al*. (in press). These Authors have mapped the landslide in detail and, by dating several mass movements, have also tried to investigate the specific causes of this complex gravitational process.

Various hypotheses have been formulated as to the relationship between the landslide and the formation of Lake Santa Croce. For instance, Brückner (1909) maintains that the landslide debris body accumulated on top of a rocky threshold forming the southern bank of the lake, and which had been formerly created by intense glacial over-excavation affecting the Tertiary rocks of the Alpago syncline.

Some boreholes – extending as far as the bedrock and carried out at La Secca, along the earth dam that bounds the northern shore of Lake Santa Croce – have provided the only reliable information. The presence of lacustrine deposits on top of the glacial deposits means that after the retreat of the Würmian glacier a stream could not have flowed towards the Lapisina Valley, therefore confirming the presence of an obstruction in the upper part of the valley, due to glacial and/or landslide deposits (Pellegrini and Zambrano, 1979).

The Fadalto landslide includes the large scarp of Mt. Costa and the area where the landslide accumulation is distributed, some 1.6 km^3. This area is made up of a series of hills which separate the northern part of the Fadalto valley from Lake Santa Croce. The accumulation attains a height of 587 m at Col dei Masarei, and 537 m at Col Brustolade, in the central part of the Fadalto Col (475 m). The debris constitutes the southern banks of Lake Santa Croce, whose average level is 380 m. Downstream, next to Lake Morto, the lowest elevation of the debris is 305 m. Considering the highest and the lowest altitudes of the accumulation, the landslide debris occupies an altimetric belt of about 280 m.

At first sight there seems to be a random distribution of hills and depressions, but an attentive analysis reveals the presence of distinct morphological sectors corresponding to different evolutive processes concerning the landslide accumulation. By means of geomorphological investigations, various landslide bodies were identified in the accumulation zone. These landslides may be grouped into three landslide sub-units: Col Brustolade – Col delle Vi, Col dei Masarei and Sassoi (Figure 7).

Over several years, large volumes of sediments have been excavated in the accumulation zone of the Fadalto landslide. At present quarrying activity is particularly intense on the eastern side of Col Brustolade – Col delle Vi landslide sub-unit, and this has provided us with the opportunity to observe new sections within the landslide deposits and to collect samples suitable for radiocarbon and dendrochronological dating.

Our knowledge of the chronology of the Fadalto landslide improved significantly thanks to the identification and dating of several mass movements occurring during the Holocene and after the main rock slide event that dammed the Lapisina valley in the Late Glacial (Pellegrini *et al*., 2004). One landslide (a small rotational slide within the accumulation zone at Col delle Vi) dates back to the late Atlantic (5375 ± 95 years BP), while the others (a rock avalanche on Col Brustolade – Col delle Vi sub-unit and a rotational slide on Col dei Masarei sub-unit) took place in historical times (within the last 14 centuries). This chronology has allowed the specific causes of these landslides to be investigated; although establishing a relationship between landslides and climate (or earthquakes) has turned out to be a difficult task. This was mainly due to the inaccuracy of radiocarbon dating especially when dealing with historical times. For this reason an attempt was made to improve radiocarbon dating by means of dendrochronology analyses. The results do not rule out the possibility that some mass movements in the Fadalto landslide took place during spells of a colder and more disturbed climate (5350-4900 years BP and AD 1300-1450) or at a transition between climatic periods with different characteristics (such as Atlantic and Sub-boreal). On the other hand, according to dendrochronological analysis, no relationship between the historical rock avalanche and climate (or earthquakes) can be inferred. Therefore, this rock avalanche is likely to have been triggered by a single meteorological event during a warm climatic spell rather than during a particular climatic phase.

Hazard implications

As for landslide hazard, it must be stressed that during the Holocene this landslide underwent several reactivations, which involved both the main scarp on the eastern slope of the valley and the accumulation zone. Besides, it was recognised that the landslide can be reactivated not only through rock slide, rotational slide and debris flow, but also through rock avalanche, a complex landslide not uncommon in the southern Alps. Since the geometry of the main scarp has not significantly changed through time, it can be argued that movements such as rock avalanche and rock fall might still take place in the future. On the other hand in the accumulation zone, rotational slides are more likely to happen.

From Fadalto to Longarone

Alessandro Pasuto

After leaving the Fadalto saddle (488 m) which divides the Livenza basin from the Piave basin, we go down as far as the shores of Lake Santa Croce, whose reservoir collects most of the water from the Alpago valley and is exploited for generating hydroelectric power.

Subsequently, we go along the relict Piave Valley and, after La Secca, forms ascribable to slope movements affecting these ridges in the very early Post-glacial can be observed.

After going through the village of Ponte nelle Alpi, we eventually reach a bridge over a deep, narrow gorge cut by the River Piave when it was diverted by the damming of the Lapisina valley. Following the river's right bank, we go upstream along the no. 51 state road to the north. Here the course of the River Piave is anastomosing, although its riverbed has narrowed considerably owing to the presence of infrastructures. Due to morphological conditions, there are almost no human settlements on the opposite slope. Small streams running through deep valleys can here be seen, some of which have been exploited for hydroelectric purposes. Just before arriving in Longarone, the confluence of the Torrent Maè with the River Piave can be observed. The basin of this secondary watercourse originates on the slopes of Mts. Pelmo and Civetta. On the opposite slope we can observe the deep gorge of the Torrent Vajont.

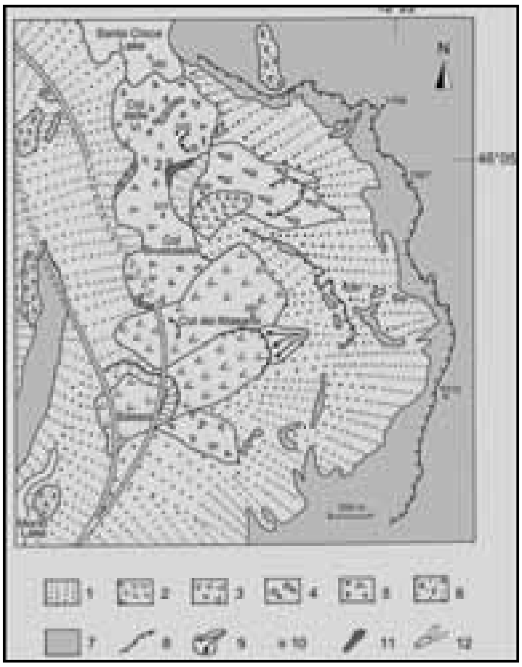

Figure 7 - Geomorphological sketch map of the Fadalto landslide. Legend: 1) slope debris; 2) earth flow; 3) rotational slide; 4) rock avalanche; 5) rock slide; 6) glacial deposit; 7) bedrock; 8) landslide scarp; 9) debris flow; 10) location of the exposures; 11) area where rock avalanche deposits were exposed; 12) motorway (after Pellegrini et al., 2004).

DAY 2

From Longarone to the Tessina landslide

Alessandro Pasuto

Proceeding southwards on a recently built stretch of motorway along the Piave Valley, one reaches the Tessina landslide. At La Secca the road turns to the left along the earth dam along the northern boundary of Lake Santa Croce, and eventually arrives in the village of Puos d'Alpago.

The Alpago area is an asymmetrical brachysyncline with a curved axis turning from a SW-NE to an E-W direction. Stratification is markedly inclined towards N and NE (60-70°), whilst it decreases considerably to the east and south (10-20°). The outcropping rocks belong to the Jurassic, Cretaceous and Tertiary formations, whereas the Quaternary deposits are mainly

made up of glacial and gravitational sediments.
Owing to the extensive presence of Tertiary flysch rock types, the Alpago basin is characterised by a considerable propension to disarray processes. Indeed, along the road leading from Puos to Chies d'Alpago it is possible to observe numerous landslide bodies with steep slopes and watercourses with over-filled riverbeds owing to the presence of large debris covers.
Once in the centre of Chies, it is possible to observe the first important case of a landslide affecting part of the village. The displacements caused by the landslide are witnessed by several cracks visible in the walls of many buildings. This slope movement is ascribed to a slow rotational-translational slide affecting material up to 30-35 m thick.
In correspondence with the village of Chies d'Alpago it is also possible to observe the foot of the earth flow which occupies the Tessina valley which is part of the landslide described at the next stop.

Stop 2.1:
The Tessina landslide
Alessandro Pasuto and Fabrizio Tagliavini

Introduction
The Tessina landslide is located on the south slope of Mt. Teverone, in the eastern part of the Alpago valley in the Province of Belluno (north-eastern Italy). It was triggered on 30th October 1960 by a rotational slide involving some 1,000,000 m^3 of material following intense rainfall (398.7 mm just in the month of October). In the following four years, always after considerable precipitation, three reactivations took place, affecting the same amount of material. They led to the formation of an earth flow which eventually came to a standstill 600 m from the village of Lamosano.
In 1966, although all the Belluno territory was struck by one of the most disastrous floods, the landslide did not record considerable reactivations and up to 1987 its activity was extremely reduced.
In August 1990 an important reactivation involving the upper part of the landslide's eastern sector occurred. Following this event, an earth flow which maintained its motion for another 200 m, was triggered in the lower accumulation area; this episode stopped after ten days.
Until 1991 the Tessina landslide had always resumed its movement by means of progressive displacements which had never caused particular risk situations for the villages of Funes and Lamosano.
In December 1991, perhaps in concomitance with a light seismic shock, significant movements occurred in the eastern sector of the source area, whereas in the following April a 40,000 m^2 wide area suddenly collapsed, giving rise to a flow which at the end of the month went over the Funes-San Martino road. In May a new reactivation affected the same sector and the flow went as far as the embankment built near the hamlet of Tarcogna. The maximum landslide advancement, which was followed by a decrease of the movement, was recorded on 15th June 1992.
In July and August 1993 new reactivations took place, which mobilised some 20,000 m^3 of material.
After these events, the landslide showed an overall longitudinal extension exceeding 2.5 km and a maximum width of about 500 m.
In 1995 the displaced material reached the village of Lamosano for the first time, causing a serious risk situation. This reactivation started on 21st May with the onset of a flow which in the following ten days moved 300 m downstream. Once again, the detachment affected the eastern sector of the source area.
Finally, in September 1998, without any apparent triggering cause, the whole upper accumulation resumed its movement, thus displacing all the main scarp and giving rise to a flow which in a few days went as far as the village of Funes.

Geological setting and morphological evolution
From the geological standpoint, the slope portion affected by the landslide belongs to the northern flank of the asymmetrical brachy-syncline of the Alpago catchment basin; this structure, which developed in Tertiary formations, shows a slightly curved trend of its axis (Mantovani *et al.*, 1976). In its northernmost part, this syncline is also affected by some important NNW-SSE arranged tectonic alignments. In the Alpago valley, formations ascribable to a chronostratigraphic sequence spanning from the Jurassic to the Quaternary crop out (Figure 8). Only the Mesozoic and pre-Oligocene Cainozoic formations are important as far as the Tessina landslide is concerned, whereas the subsequent ones (i.e., from the Oligocene and Miocene) are only marginally affected by the flow's lowermost portion. The calcareous, marly-calcareous and cherty-calcareous formations (Upper Cretaceous-Paleocene), are mainly organised in layers of variable thickness and are permeable due to fissuring.
The material involved in the slope movement belongs to the Flysch dell'Alpago (Lower Eocene). It is made

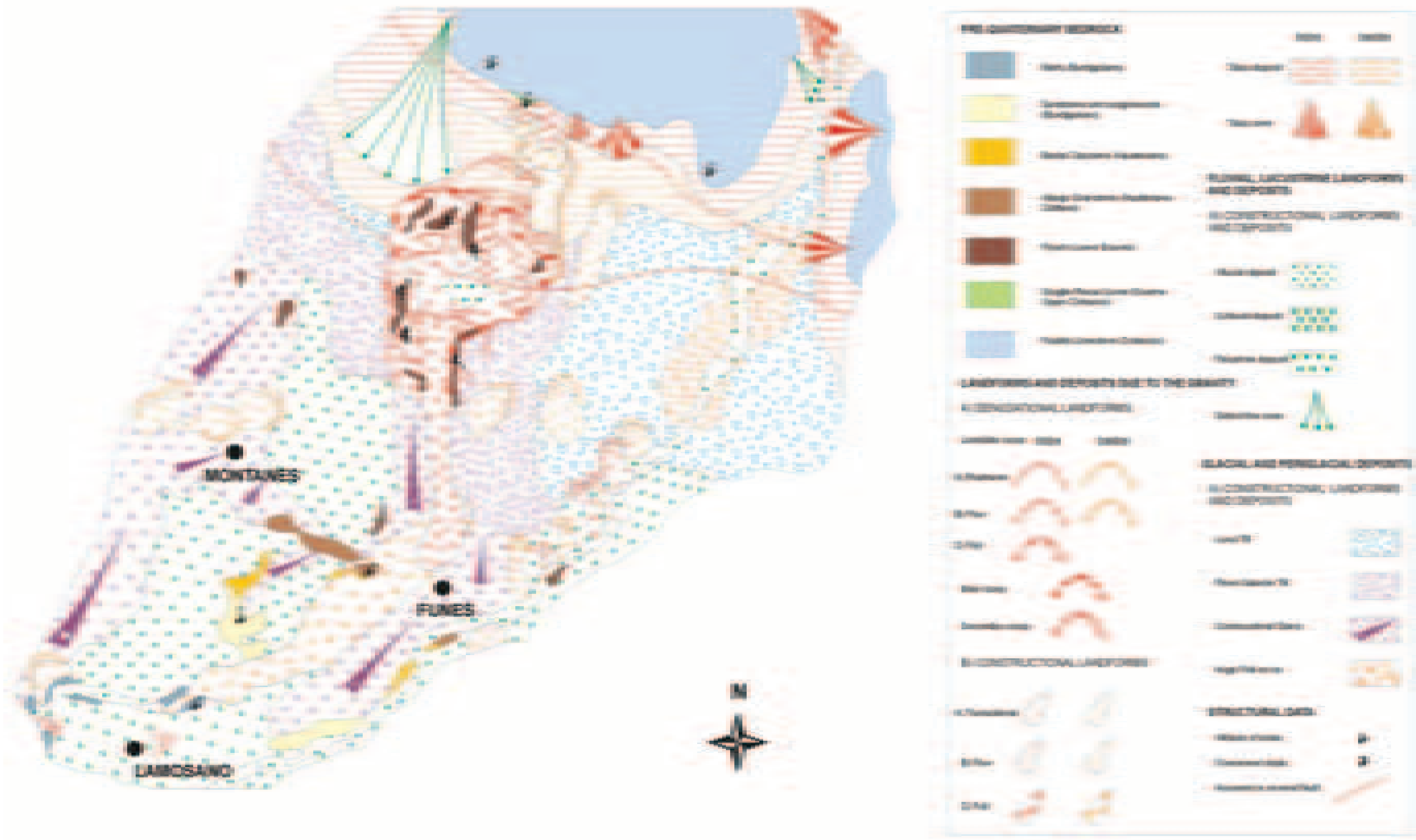

Figure 8 - Geomorphological sketch map of the Tessina landslide area.

up of a rhythmic alternance of marlstones, clay shales and calcarenite layers up to 1 m thick.
The overall thickness of the formation is 1000 to 1200 m; it makes up the substratum of the whole landslide area. This formation is particularly permeable owing to the high degree of fissures mainly resulting from tectonic stresses. Its attitude is generally dip-upstream, with overturned strata in correspondence with the main scarp and the upper landslide portion. More downhill the attitude is south-dipping, dip-downstream with low angle.
The Quaternary deposits are Pleistocene and Holocene in age and crop out with continuity from the foot of the Mt. Teverone slopes, covering the underlying formations with variable thicknesses. Apart from vast detritic covers found at the foot of Mt. Teverone, moraine deposits from the Piave glacier and local glaciers are found. The latter, in the form of glacis, are arranged on both sides of the valley.
From a topographic viewpoint, the Tessina landslide stretches from 1200 m to 640 m. The landslide is classified as complex movement including a rotational and translational slide in its upper part and a flow in its mid-lower part.
Before 1960 the Tessina valley was deeply cut by erosion and flanked by steep slopes. During the 1960s several reactivations, involving about 5 million m^3 of material, occurred, causing the filling of the Tessina valley with displaced material 30 to 50 m thick. These movements seriously endangered the village of Funes, which is situated on a steep ridge originally quite high above the riverbed, but now nearly at the same level as the mud flow. The collapsed sector of the April 1992 event occupies a 40,000 m^2 wide area, on the left hand-side of the Tessina stream, with an approximate volume of 1 million m^3. The movement corresponds to a rotational slide with a 20 to 30 m deep failure surface, affecting also the flysch bedrock. Initially it caused the formation of a 15 m high scarp and a 100 m displacement downstream, with consequent disarrangement of all the unstable mass and destruction of the drainage systems set up some years earlier.
The movement in this area continued, with an intensity of about several tens of centimetres per day, up to June, causing the mobilisation of another 30,000 m^2, with a total volume of about 2 million m^3 of displaced material.
The material from this area, which is intensely fractured and dismembered, was channelised along the riverbed where, owing to its continuous remoulding and increase of water content, it became more and more fluidified, thus giving rise to small earth flows converging into the main flow body. After these events the inhabitants of Funes and Lamosano were evacuated.
The landslide's evolution is still in progress and has led to a constant widening of the source area, which from the original 300,000 m^2 is now about 500,000 m^2. The total displaceable volume has been assessed as about 7 million m^3.
The velocity of the movement seems to vary considerably according to the episodes and sections considered. The maximum velocities attained were recorded uphill of Lamosano in May 1992, with displacements of 70 to 100 m/day along the main flow. In the same period, in proximity of Funes, velocities of about 25 to 30 m/day were recorded.

Instrumentation

The particular evolution and typological features of the slide, the extension of the area involved, the extent of the mud flow movement and the intense dislocation of the slide made it necessary to plan a series of innovative solutions for the installation of an automatic alert and monitoring system that could guarantee a sufficient level of safety for the population. The system consists of an arrangement of sensors and measuring instruments which transmit their signals to a central station by means of peripheral receiver units. The station is installed in a building in Lamosano and receives information about changes and movements of the mud flow and of the upper section of the slide. It processes the data, thus giving a visual layout of the situation on an appropriate peripheral unit installed at the Belluno Fire Station and sending out an alarm signal to suitably located stations should a situation of danger occur.
Two multiple-base wire extensometer units, measuring 280 m and 390 m, were installed in the upper section of the landslide in order to keep a constant check on the movement at the surface. These units, fitted with an appropriate scaler system, consist of a series of 12 measuring purpose-built apparatuses, capable of detecting movements as small as one millimetre.
On the upper landslide section an electronic distance measurement system (EDM) with an automatic landmark detector for measuring surface movements was also installed. Every 6 hours it monitors a network of more than 30 bench marks, assessing their displacements with respect to previous readings. The data are then recorded on a computer and sent to the

control centre, situated inside the Town Hall of Chies d'Alpago in Lamosano.
On the body of the mud flow two alarm units, one comprising three directional bars and an ultrasonic echometer, the other with two directional bars and an echometer were installed upstream of the villages of Funes and Lamosano.
Three videocameras were also installed for the purpose of recording and monitoring the slide movement in the areas considered as the most critical, that is, the upper accumulation zone and the two areas uphill of Funes and Lamosano.
The control centre receives data from the peripheral stations to which the sensors are connected. It also defines possible situations of danger and sends data and signals via modem to an alarm station located at the Belluno Fire Station.
In order to follow the developments of this process the station is integrated with an image acquisition and recording system which allows direct visual observation of the upper landslide section.
The peripheral units receive and perform preliminary processing of the data from the instruments (i.e. echometer, directional bars, extensometer) and verify their effectiveness.
In the event of a critical development of the situation, various alert levels can be determined (i.e. pre-alarm, normal alarm, serious alarm), with an indication of the peripheral units and sensors directly involved. It is also possible to have access to the data in real time mode, by connecting up each peripheral device and checking the instrumentation, as well as providing access to data concerning daily, weekly or monthly trends of each sensor by means of special programmes.
In recent years an innovative technique, based on radar interferometry and implemented using ground-based instrumentation, has been applied for monitoring the Tessina landslide. This technique has allowed multi-temporal surface deformation maps of the entire depletion zone of the landslide to be derived with a high resolution and accuracy. The application of the ground-based SAR interferometry to the Tessina landslide has been implemented by using portable field instrumentation that can be installed in a very short time (few hours). Derived data have shown the evolution of ground movements over almost the entire depletion zone of the landslide. Displacement rates up to about 1 m/day have been assessed with a millimetric accuracy and a pixel resolution of approximately 2x2 m.
Through the continuous monitoring of slope movements, the proposed technique has allowed the implementation of real-time maps showing the deformation field of those landslide sectors characterised by a good radar reflectivity and coherence (i.e. scarcely vegetated areas). These deformation maps are a tool for the interpretation of landslide kinematics and short term-evolution, providing, at the same time, data for an accurate analysis of temporal and spatial displacement fields.

From the Tessina landslide to Vajont

Alessandro Pasuto

The route towards the Vajont landslide is the same as the one followed during the transfer trip in the opposite direction. Having arrived in Longarone, we turn to the right across the River Piave going up the left-hand side of the valley. In correspondence with the first tunnel, we follow the valley of the Torrent Vajont going towards the administrative border between the regions of Veneto and Friuli-Venezia Giulia. This road was purposely built for the construction of the dam. Before the 1950s there was only the steep and narrow road going up the Vajont valley on the opposite slope. After passing some narrow tunnels dug directly into the rock, we arrive in proximity of the dam whose reservoir caused the terrible disaster of 1963.

Stop 2.2:

The Vajont landslide

Emiliano Oddone, Alessandro Pasuto and Fabrizio Tagliavini

Introduction

The Vajont reservoir in Italy's eastern Alps is located along the lower reaches of the River Vajont, close to its confluence with the R. Piave. Some 100 km NNW of Venice, the reservoir was created artificially in 1960 after the T. Vajont was dammed as part of regional expansion in hydroelectric-power generation. The dam is 261.60 m high and 190 m across the top, and at the time of construction was the highest and one of the most advanced double-arched dams in the world. After the reservoir level had been raised and lowered several times, the southern margin of Mt. Toc became eventually destabilised and, after nearly three years of intermittent creeping, it catastrophically collapsed at 22:39 hours on 9th October 1963 (see cover photo). Within 30-40 seconds, some 270 million m^3 of rock crashed into the reservoir, expelling a wave of water about 100-150 m high over the dam. The vast slope movement, ascribable to a rock slide, started to move along a surface corresponding to one or more dip-downstream clayey interbeds having the same angle as the slope, which characterise the Fonzaso Fm. (see below).

This slide displaced a 250-300 m thick mass of rock for 300 to 400 m horizontally involving a volume of 270 million m^3. It reached a velocity of about 20 to 30 m/s before running up and stopping against the opposite slope of the Vajont valley. The slide filled the valley and the reservoir in a few tens of seconds causing a wave of water propagating both upstream and downstream. This wave reached a maximum elevation of 935 m (235 m above the reservoir level). It swept across the dam and the Vajont gorge and eventually fell onto the Piave valley floor, where it destroyed the town of Longarone and neighbouring villages, claiming 1909 lives.

Geological studies for the construction of a 200 m high dam had started in the late 1920s. After World War II, the project was modified and the construction of the dam, which attained the height of 261.60 m, was eventually completed in 1959. The water volume of the reservoir reached 168 million m^3.

Mt. Toc and the southern slope of the Vajont valley were more prone to instability than expected, since the mountain's outer flanks consisted of an ancient landslide deposit and not of bedrock, as initially inferred. Giudici and Semenza (1960) first realised that Mt. Toc was covered by an ancient landslide deposit, citing a mylonitic belt and the remains of a slide with an abnormal attitude on the opposite northern side of the valley, isolated by the gorge cut by the T. Vajont. They also identified another massive ancient landslide, La Pineda landslide, upstream of the Vajont slide, which had previously dammed the T. Vajont with the formation of an impoundment.

Geological and geomorphological setting

The report by Carloni and Mazzanti (1964) should be considered as the starting reference for the geomorphological assessment of the Vajont area. These Authors suggested that during the last deglaciation the lower reaches of the T. Vajont were completely truncated by a north-east-directed landslide. The river subsequently cut a new course through the easily erodable materials of the landslide deposit, producing the deep and narrow Vajont gorge that today has decreased to a vertical drop of 240 m and from a width of 150 m to 20 m. The current E-W trend of the valley follows that of its pre-glacial predecessor, determined in turn by a regional tectonic trend. The topographic features of the area changed enormously following the 1963 slide, especially with the infilling of the gorge with loose landslide material consisting of moraine deposits, ancient alluvial deposits (fluvial and torrential), sometimes cemented, and slope debris, all Quaternary in age.

In the Vajont valley and adjacent mountain group geological formations, range in age from the Upper Triassic (Dolomia Principale) to the Middle Eocene (Flysch) (Figure 9). During the Lower Lias the area was part of a basin in which cherty micrites with pelagic fauna accumulated. Interbeds of calcarenites and oolitic or bioclastic calcirudites are often found; they were driven into the basin by gravity currents from marine shelves around the basin margins. Indeed, most of the characteristic differences between formations reflect variations in the material coming from marine shelves.

Bedrock

Flysch (Eocene; >200 m thick). A succession of turbiditic arenites intercalated with mudstones. The arenitic fraction is made up of calcarenites passing to grey or yellow litharenites; the mudstones are formed by marls and grey marly clays.

Erto Marls (Paleocene; 100-150 m thick). This informal unit represents the transition between the Scaglia Rossa and the Flysch and is formed by marls and subordinate grey marly limestones, intensely bioturbated, containing rare thin layers of calcarenites and litharenites.

Scaglia Rossa (Upper Cretaceous; ~300 m thick). A monotonous succession of marls and red marly limestones, completely lacking gravity resediments is the typical facies of Scaglia. This formation levelled the pre-existing topography because it covers evenly the whole area.

Soccher Limestone (Lower-Upper Cretaceous; 150 m thick). An alternation of microcrystalline limestones, calcarenites and bioclastic-intraclastic calcirudites coming from the Friuli shelf. The fine component is made up of decimetric layers of micrites, marly micrites and marls, that can be grey, red or greenish with nodules and beds of chert. There are numerous intraformational discordances interpreted as slump scars. The coarser components of the slump material can be further subdivided into bioclastic calcarenites and conglomerates-and-breccias that are sometimes associated with stratigraphic unconformities. The breccias and conglomerates are characterised by great lateral continuity and are visible in the mass which slid from Mt. Toc.

Ammonitico Rosso (Kimmeridgian Titonian; 5-15 m thick). Nodular reddish and grey micrites with ammonites, occurring as massive units or as layers of more than 1 m thick that differ only in colour from the classic facies outcropping in the Veneto area.

Fonzaso Fm. (Oxfordian; 10-40 m thick) The complex succession of levels between the Vajont Limestone and the Scaglia Rossa have been attributed to the Soccher Limestone. The Fonzaso Fm. is formed by finely graded biocalcarenites and chert-rich, brown micritic limestone in 5-20 cm thick beds with parallel and oblique *laminae*. Interbedded layers of green claystone (5-10 cm thick) are of particular interest because of their role in the movement of the Vajont landslide.

Vajont Limestone (Dogger; 450 m thick). Hazelnut oolitic calcarenites, massive or layered in thick beds, intercalated with decimetric layers of brown basinal micrites and intraformational breccias (clasts formed from the erosion of the micrites). The latter are more frequent in the upper part of the formation. Nodules and beds of brown chert can be locally present. Sedimentary structures are represented by graded beds with parallel or oblique *laminae*.

Landslide morphological features

The Vajont landslide deposits can be divided into three kinematic subunits, following the order of events occurring during the 1963 collapse (Figure 9).

Subunit A describes the northernmost part of the landslide which, now infilling the valley, was originally at the toe of the unstable slope and represents the material that displaced the water of the Vajont reservoir. It consists of a single mass of layers strongly deformed during collapse. All prefailure vegetation was removed. However, it has been possible to map the principal surface fractures and prefailure surface material which survived the return water wave after collapse (ancient collapse and glacial deposits). These features are interspersed with zones of fragment accumulation – the fragments being poorly-sorted mixtures ranging in size from gravel to blocks – produced during deposition by the return wave. Depressions are also visible along its western margin, above the dam.

Subunit B is an intermediate unit between the present failure scarp (at a higher elevation) and subunit A below. Prefailure vegetation is visible, trees showing classic landslide deformation marks along their trunks. There is no evidence of significant interaction between this unit and waves produced by the collapse and this is consistent with a second-stage collapse, shortly after the displacement of subunit A.

Subunit C is the portion of the landslide deposit directly against the dam, south of state road no. 251 (Val Cellina-Val di Zoldo) and north of subunit A. The surface is lower than that of subunit A and consists mainly of sand and gravel. The loose surface material completely covers the originally displaced slope (which underneath may be continuous with the adjacent subunit A) and appears to have been derived from fragments washed back onto the landslide by the return wave immediately after collapse.

Future perspectives

At present the areas affected by the 1963 slide are subject to numerous environmental and town planning restraints. Recently, in-depth investigations have been carried out in order to define the residual geological risk and facilitate the upgrading of the territory. These investigations, though, have identified a rather serious risk situation for the hamlet of Casso, owing to a rock fall accumulation located upstream of some houses in the upper part of the village. As for the slide deposit of 1963, no particular stability problems have been identified, whereas the surfaces of rupture are still affected by circumscribed debris slides which, nevertheless do not cause risk situations considering the inaccessibility of these areas.

Numerous initiatives have been enterprised in order to assess this territory and contribute to keeping the memory of the catastrophic disaster alive. Among these, are the creation of a permanent laboratory for the study of hydrogeological hazard and the spreading of information concerning the increased perception of these kinds of risk. This laboratory is jointly run by an Italian-Japanese team promoted by the Italian Foreign Ministry within the framework of the 7th Executive Programme for Scientific and Technological Co-operation between Italy and Japan.

At present, other initiatives concern the creation of the so-called “multi-centre museum”, consisting of historical-naturalistic paths, permanent exhibitions, video-cassettes and other educational and/or popular multimedia material about the sites where the catastrophe occurred, in order to keep alive the historical memory of what happened – especially among younger generations – and increase awareness of natural risks.

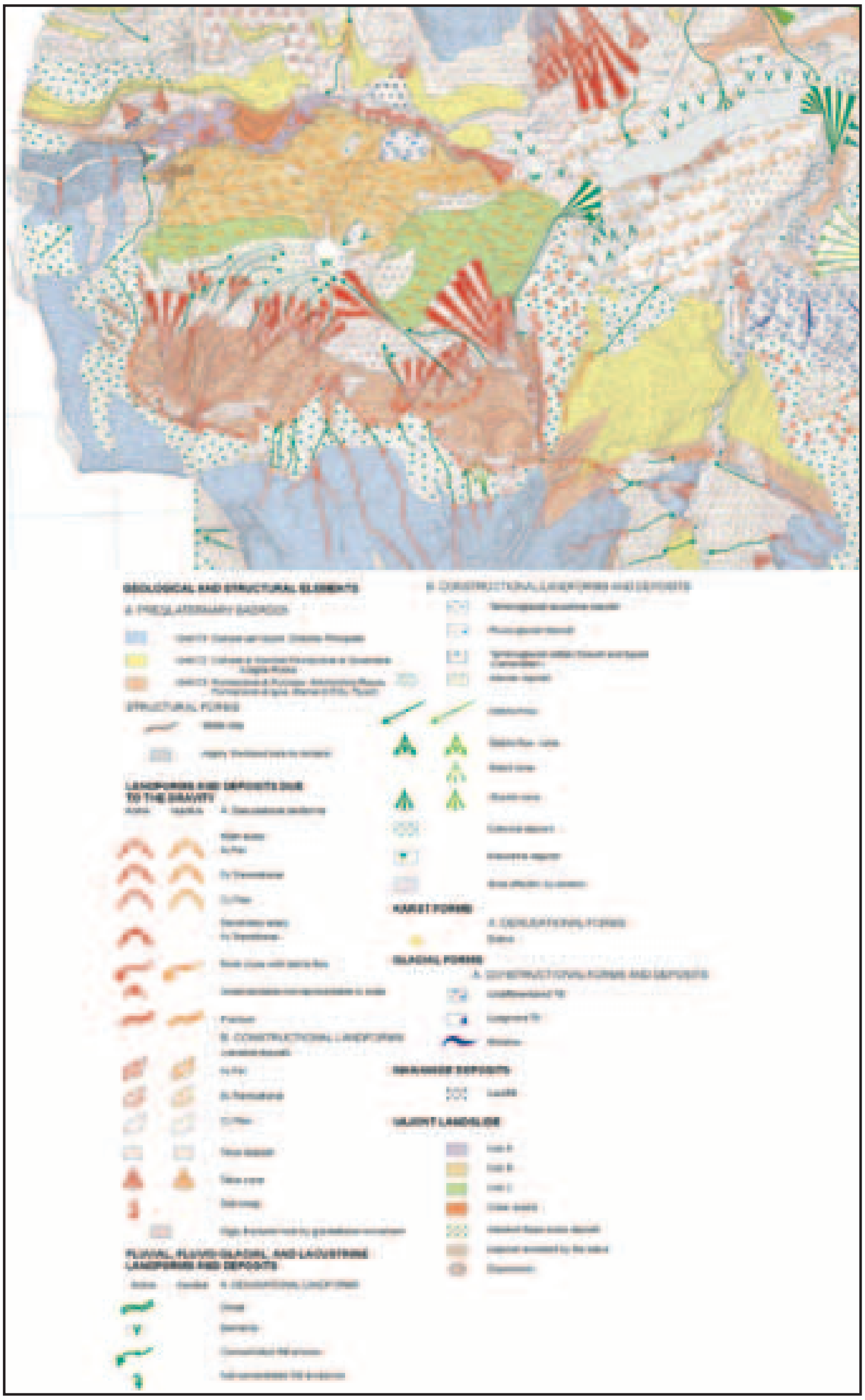

Figure 9 - Geomorphological sketch map of the Vajont landslide area

From the Vajont landslide to Cortina d'Ampezzo

Anna Galuppo, Alessandro Pasuto and Enrico Schiavon

Starting from the Vajont dam, we go down the same road that leads to Longarone, then follow the state road to the north running parallel to the R. Piave. Upstream of the village of Castellavazzo, the valley becomes rather narrow, with steep rocky slopes rising above a flat and overfilled valley floor. A new diversion of state road no. 251, which has been recently constructed, further reduces the already narrow river-bed, thus increasing a serious hydraulic risk situation. The valley of the R. Piave becomes wider at its confluence with the right-hand side tributary Torrent Boite. This portion of the Dolomites where the village of Perarolo di Cadore is situated, shows a rather complex situation from both the structural and hydrogeological disarray point of view. This area is, in fact, subject to numerous disarrangement processes, some of which directly affect inhabited areas, inducing high risk situations for the population.

S. Andrea landslide, Perarolo di Cadore

The bedrock corresponds to the most recent terms of the Triassic sequence (Dolomia Principale, Raibl and Dürrenstein formations). The large S. Andrea landslide, just upstream of Perarolo, affects both state road no. 51 and the railway. The active portion of the landslides has an approximate volume of some 4-500,000 m^3 and is constantly monitored by the Veneto Region. Observations from aerial photographs and field surveys have allowed a vast relict landslide involving gypsum and dolomite formations to be identified. Uphill of the accumulation of the active landslide a dormant earth flow affecting the red clays of the Raibl Fm. is found. Its accumulation material, which is substantially impervious, conveys surface water towards the active landslide.

Having arrived at the village of Tai di Cadore we turn left, going up the valley of the Torrent Boite. After a few kilometres, we reach the village of Valle di Cadore, which has been subject to instability processes since the disastrous flood of November 1966. After this village, the valley becomes gradually wider, assuming on its left flank the typical "U" profile resulting from the intense glaciation of the LGM. This valley is characterised by the presence of a W-verging overthrust front, which conditions its marked asymmetry. Vertical rock faces of limestones and dolostones with subhorizontal bedding characterise the valley's left flank, whereas at the foot of the slopes thick talus fan covers are found. On the contrary, on the right flank the strata bedding is dip-downstream and the big step-like features are due to tectonics.
Owing to this geological setting, the left slope, which is more densely inhabited, is subject to several debris flows which seriously threaten some villages such as Cancia and Acquabona.

Debris flows at Cancia

Debris flows have repeatedly struck the village of Cancia, affecting the western slope of Mt. Antelào and mobilising vast volumes of material. The dolomitic rocks cropping out along the slope are intensely jointed and partially covered by a thick debris deposit. These processes have been described in chronicles since the 14th century. In 1868, during the most destructive event ever recorded, a debris flow of over 100,000 m^3 buried the centre of Cancia claiming 11 lives. Starting from the 1950s, all the area has undergone intense development. In particular, a tourist village and a new area of Cancia have been built, whereas the manifold canal was artificially prolonged downstream.
On 7th August 1996, following very intense, concentrated, though not exceptional, precipitation, a debris flow was triggered which struck Cancia after travelling about 2800 m (Panizza *et al.*, 1998). The area affected by the latest flows is some 70,000 m^2, whereas the area which has been involved in more ancient slope movements may be estimated as about 200,000 m^2.
This process tends to reoccur with time, often creating high risk situations for the inhabitants of the area. In addition, it can be assumed that interventions made in the territory have probably had a negative influence on the transport and accumulation patterns of the landslide material. In particular, the construction of check dams and bank reinforcements have constrained the debris flow into a narrower channel, thus increasing its velocity and energy. Since it is practically impossible to stop possible new flows along the defile, remedial measures have been directed to mitigate the consequences in terms of risk. For this purpose, a flow control and containment weir has been constructed at the confluence of the two defiles at an altitude of 1300 m circa.

Acquabona debris flow

Mass wasting processes occuring at Acquabona (see Berti *et al*., 1999) are typical examples of debris flows in the Dolomites.

The debris flow channel is deeply cut mostly into the scree and ancient debris flow deposits. Boulders are present just at the bottom of the rock cliffs, but they do not constitute an effective boulder dam. Partly vegetated debris flow deposits (lateral levees and lobes) are visible on the middle-lower portion of the slope. Approximately 500 m uphill the deposition area (about 1200 m), where the slope locally decreases, the channel is partially clogged by a debris dam which may cause avulsion of the flowing mass. The terminal deposits are confined by a debris basin built for road safety purposes.

The initiation area of debris flow is located in the upper channel, between 1550 m and 1650 m. Debris flows are likely to be started by the “fire hose effect”, in which a rapid inflow of water mobilises low density and lightly cemented materials in the channel bed, where another possible initiation area is to be individuated at the natural rocky dam.

A fully automatic and remote-controlled monitoring system has been installed at Acquabona by Padova University. It consists of three on-site monitoring stations and an off-site master collection station located 1.3 km from Acquabona creek. Each on-site station is provided with a small data logger and a two-way radio. Data are transmitted to the master station, and then via modem to Padova University. The transmission time interval is five minutes in “pre-event” mode. When a threshold value is exceeded, the “event” mode performs data transmission at faster rate. The measuring instruments utilised are geophones, anemometers, rain gauges, superficial and deep pressure transducers and video systems.

After leaving the Acquabona area, we arrive in Cortina d’Ampezzo, where the valley is even wider owing to the presence of particularly erodable Triassic rock types. Naturally, the presence of these rocks affects also the stability conditions of all the area. Indeed, the valley around Cortina is particularly prone to disarray processes, with the presence of numerous landslides of highly diverse types and dimensions.

DAY 3

Stop 3.1:

Landslides in the Cortina d’Ampezzo area

Mauro Soldati

Introduction

Landslide investigations in the area of Cortina d’Ampezzo (Province of Belluno) (Figure 10) have been carried out on a multi-disciplinary basis since the beginning of the 1990s mainly within European research projects, involving primarily experts in geology, geomorphology and engineering geology. This has enabled the researchers not only to define a general picture of landslide causes, occurrence and evolution in the study area, but also to assess landslide hazard and therefore plan and install monitoring systems on the most landslide-prone slopes so as to forecast future movements

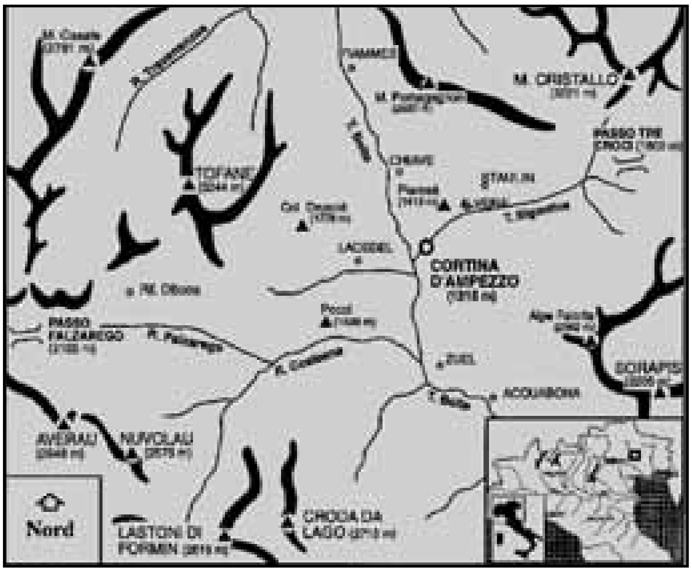

Figure 10 - Physiographic setting of the Cortina d’Ampezzo area.

Landslide causes

The geomorphological evolution and emplacement of superficial deposits, mainly made up of landslide deposits, are ascribable to the Upper Pleistocene and Holocene, in particular to the post-Würmian period. In fact, during the last Alpine glaciation the study area, apart from the highest peaks, was covered by a considerable ice thickness up to 1000 m. Of the vast glacial deposits which occupied the study area after the glaciers' withdrawal, only a few remnants were found, probably ascribable to Lateglacial stadial phases, which have been preserved thanks to the presence of rocky rises which hindered their erosion. On the contrary, no trace of deposits from the LGM phase has ever been found. This is mainly due to the numerous, vast mass movements occurring during the withdrawal of glacial masses. During this period the area was affected by several slope movements which in most cases involved dolomite rock types; their accumulations could have been reworked afterwards by slide and flow processes affecting the S. Cassiano Fm.

At present, the climate of the area can be defined as Alpine type, from cold to temperate, with variably cold winters and mild summers. The pluviometric characteristics of the area were deduced from data recorded at two meteorological stations: Cortina d'Ampezzo (1275 m) and the Codivilla-Putti Institute (1350 m). The data cover a time span from 1920 to 1990 with a yearly average of about 1108 mm. Late springtime and summer are the most rainy periods, with a peak in July. In this case, precipitation occurs mostly during violent rainstorms, with intense, concentrated events.

The area of Cortina d'Ampezzo appears to have always been prone to slope instability processes for various reasons. First of all, the structural conditions of the valley should be taken into account. The stratigraphic succession, in fact, features an alternation of dolomite rocks showing brittle mechanical behaviour and rocks with ductile mechanical behaviour. This situation has favoured the development of mass movements and deep-seated gravitational slope deformations; the latter, which have been widely recognised in the area, may have favoured or induced the occurrence of several landslides (Soldati and Pasuto, 1991). Furthermore, the incidence of tectonics is also significant. In fact, the dolomites were affected by intense jointing in correspondence with the principal faults, thus creating discontinuities that became potential sliding surfaces and preferential seepage zones where the water could reach and moisten the underlying marly and clayey formations.

Particularly rainy periods occurred during the years 1965 and 1966, when many Italian regions suffered disastrous floods. On that occasion the Cortina d'Ampezzo area was affected by numerous mass movements which mostly consisted of reactivations of older landslides.

As regards earthquake-induced slope instability, the only confirmed case of correlation is the Cinque Torri earth flow which occurred nearby in concomitance with the quake which took place on the 15th of September 1976 in the Friuli-Venezia Giulia Region (Zardini, 1979). In general, no relationship was found, mainly because the time scale of landslide occurrence and of the seismic series recorded are not comparable, since the latter refer only to historical time.

Description of the main landslides

Over thirty landslides of different type, size and age, some of which still active, have been identified in the whole area (Panizza, 1990; Panizza *et al.*, 1996; Pasuto *et al.*, 1997; Soldati, 1999; Soldati *et al.*, 2004) and many of these were dated by means of radiometric methods (for landslide location and dating information see Figure 11 and Table 1). The landslide characteristics are generally linked to both lithological and structural conditions; rock falls occurred in the higher parts of the slopes, where the dolomitic rocks outcrop, whereas slides and flows took place in the medium and lower sectors of the slopes, where marly and clayey formations are present. Extensive movements affecting the slopes from the top to the bottom have been ascribed to those of the complex type. The types of slope instability also seem to have changed through time. From the data gathered so far through ^{14}C dating, it seems clear that the Lateglacial and early Post-glacial period witnessed the majority of the largest mass movements in the area of Cortina d'Ampezzo (Soldati *et al.*, 2004). Huge rotational block slides occurred, involving both clayey and dolomite materials; their occurrence is still clearly witnessed by geomorphological evidence. Actually, the slope evolution of the area has always been characterised by the presence of recurrent landslides, mainly flows and slides involving the S. Cassiano Fm., generally of minor dimensions and higher frequency than those mentioned above, which have had a shorter persistence in time. For this reason only the most recent ones, as well as those still active today, are recognisable. Reactivations of older landslide deposits, which generally involved relatively small thicknesses (up to 15-20 m circa) of material in earth flows, may also be ascribed to this group.

The main landslides in the area of Cortina d'Ampezzo are described below (Figure 11 and Table 1) (see Pasuto *et al.*, 1997).

Col Drusciè landslide

This is a complex landslide whose body corresponds to the oldest sector of a wider accumulation. The finding of a tree fragment at a depth of about 15 m has allowed radiometric dating to be carried out, which produced a minimal age of 9000±150 years (Zardini *et al.*, 1984). The landslide body is made up of huge boulders of Dolomia Principale arranged in a chaotic manner, mixed with polygenic detritus.

Pierosà landslide

The landslide body is made up of boulders, up to some tens of cubic metres large, of Dolomia Cassiana and Dolomia Principale mixed with a clayey-silty matrix resulting from the weathered rock types of the S. Cassiano Fm. It may be inferred that this deposit is the relict remnant of a large complex landslide that detached from the crest of Mt. Pomagagnon. Part of this accumulation material was subsequently mobilised by flow processes. The finding of a tree trunk at a depth of about 4-5 m in Grava di Sotto, has allowed this event to be dated to 10850±80 years BP. Precision topographic measurement carried out in the past years has shown that the Pierosà ridge is moving to the south at a displacement rate of some 0.5 cm/year.

Zuel landslide

The Zuel landslide body is part of a vast complex movement that was triggered in the early Post-glacial on the slope of the Faloria massif, stretching from the Colle di S. Rocco as far as the present area of Pezziè.

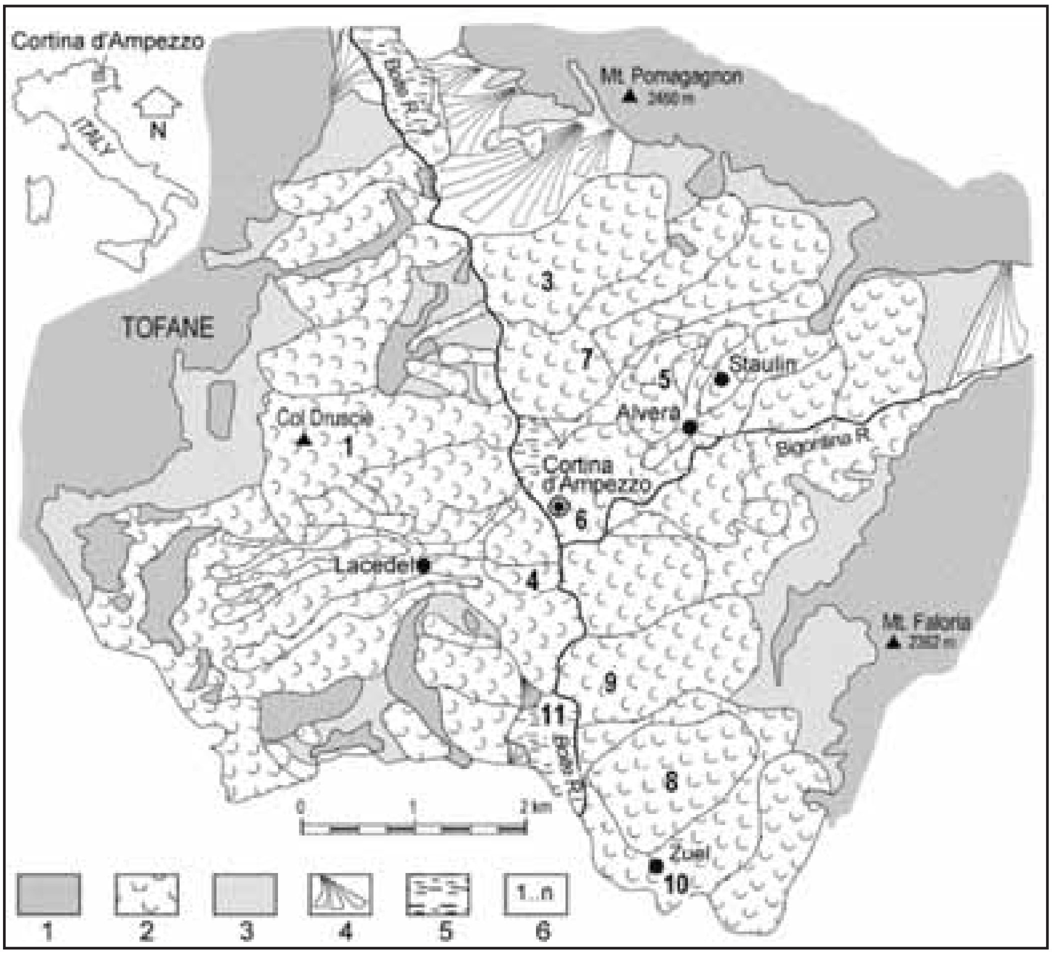

Figure 11 - Geomorphological sketch of the Cortina d'Ampezzo area (heights are given in metres). Legend: 1) bedrock; 2) landslide; 3) scree slope; 4) debris flow and/or alluvial fan; 5) lacustrine deposit; 6) landslide numbering according to Table 1 (after Soldati et al., 2004).

The landslide crown is quite visible, and consists of a regression of the rock walls between 2055 m and 2115 m. To the south a portion of slope, which is still in precarious equilibrium conditions, is confined by the scarp. Displacements, which have prevalently affected the Dolomia Principale Fm., have been favoured by the structural conditions of the slope, which is characterised by a dense discontinuity network.

The dating of two tree fragments found in a trench near the fork between state road no. 51 and the local Pezziè road has produced a conventional age of 9440±105 and 9215±105 years BP, respectively. These finds were also reworked by the subsequent Pezziè landslide. On the basis of geomorphological observations, it is assumed that the two dates refer to the same event.

Cortina d'Ampezzo landslide

This slide has involved the eastern sector of the Pierosà body in slow flows which have given rise to a fan-shaped deposit on which the town of Cortina d'Ampezzo is situated. Associated lacustrine deposits resulting from the damming of the Torrent Boite have been found in the Ice Palace area and palustrine deposits in correspondence with the Post Office square.

The materials making up the landslide body are prevalently composed of grey clays and silty clays incorporating large blocks of Dolomia Principale and frequent fragments of organic material. These deposits are now stabilised and correspond only partially to the "Begontina landslide" as mapped by Panizza and Zardini (1986). The most ancient landslide event so far detected, corresponding to an age of 8710±70 years BP, has been identified by means of a peat sample found at a depth of 23 m in a borehole near the centre of Staulin. A later reactivation was dated at 4350±60 years BP thanks to a wood specimen found south of Verocai (Cianderies). Further displacements go back to 3315±140, 2560±80 and 2465±125 years BP, as witnessed by the dating given by organic material recovered at various depths from boreholes near Staulin. Another radiometric dating carried out on a tree trunk found at a depth of 7 m in the centre of Cortina has allowed an event from 1460±30 years BP to be dated.

At present, two portions of the Cortina d'Ampezzo landslide are active in correspondence with the Staulin and Alverà hamlets (Figure 12). They are made up of two earth flows stretching from 1700 and 1500 m as far down as 1300 m, respectively, affecting surfaces of some 40 and 10 hectares, respectively. In 1935, both villages were listed among the settlements to be transferred.

Subsurface investigations have allowed two failure surfaces to be identified at depths of about 25 m and 8 m in the Staulin landslide, and 23 m and 5 m in the Alverà landslide (Angeli *et al.*, 1996; Gasparetto *et al.*, 1996).

ID	Landslide name	Landslide type	Sample code	Sample type	Site of collection	Depth (m)	Conventional Age (^{14}C yr BP)	Calendar Age 2σ (Cal. yr BP)
1	Col Drusciè	translational rock slide	—	tree trunk	excavation	14.50	[illegible]	[illegible]
2	Cadin	earth flow	GX-16315	tree trunk	excavation	3.50	[illegible]	[illegible]
3	Chiave	earth flow	B-74002	tree trunk	excavation	3.85	4520 +/- 60	[illegible]
4	Lacedel	earth flow	GX-17086	wood	borehole (S1)	42.20	[illegible]	12,277 - 11,201
4	Lacedel	earth flow	GX-17085	wood	borehole (S2)	22.20	[illegible]	[illegible]
5	Pierosà	translational rock slide	B-63381	wood	excavation	3.50	[illegible]	[illegible]
6	Cortina	earth flow	B-83624	peat	borehole	23.00	8710 +/- 70	[illegible]
6	Cortina	earth flow	B-63238	wood	excavation	9.00	4350 +/- 60	[illegible]
6	Cortina (Alverà)	earth flow	B-63340	wood	excavation	3.50	[illegible]	[illegible]
6	Cortina (Alverà)	earth flow	B-83623	wood	excavation	2.00	2560 +/- 80	[illegible]
6	Cortina (Alverà)	earth flow	B-3751	tree trunk	excavation	7.00	1460 +/- 30	[illegible]
6	Cortina-Staulin	earth flow	GX-17089	wood	borehole (S3)	24.00	3315 +/- 140	[illegible]
6	Cortina-Staulin	earth flow	GX-17088	wood	borehole (S3)	3.00	2465 +/- 125	[illegible]
7	Chiamulera	earth flow	[illegible]	tree trunk	excavation	4.50	[illegible]	[illegible]
8	Pezziè	earth flow	B-83622	wood	excavation	6.00	[illegible]	[illegible]
8	Pezziè	earth flow	B-83621	wood	excavation	2.50	[illegible]	[illegible]
8	Pezziè	earth flow	B-74003	tree trunk	excavation	3.50	[illegible]	[illegible]
9	La Riva	earth flow	GX-17080	wood	excavation	2.50	[illegible]	[illegible]
9	La Riva	earth flow	B-63342	wood	excavation	3.00	[illegible]	[illegible]
10	Zuel	translational rock slide	GX-17087	wood	excavation	3.50	9440 +/- 105	[illegible]
10	Zuel	translational rock slide	GX-17088	wood	excavation	3.50	9215 +/- 105	[illegible]
11	Campo / Zuel	sedimentation in dam lake	GX-17090	wood	excavation	4.00	[illegible]	[illegible]

Table 1 - Radiocarbon dating in the area of Cortina d'Ampezzo; data calibrated with Radiocarbon Calib. Program 4.1 by Stuiver and Reimer (1993), data set by Stuiver et al. (1998) (after Soldati et al., 2004, modified).

Lacedel landslide

This is made up of the deposits of some earth flow bodies characterised by two separate source areas which have contributed to the feeding of two large fan-shaped accumulation zones. The materials are made up of silty clays incorporating broken fragments of sandstones and calcarenites which have affected state road no. 48 and some houses, the latter only marginally. Along the Roncatto torrent the movement is clearly active (Rio Roncatto landslide). The movement assumes higher velocity in correspondence with the fan-shaped accumulation zone.

Boreholes have identified a main surface of rupture at a depth of some 10 m and a secondary surface 6 m below the ground surface. In the past years, displacements have attained velocities of about 2 m per year, causing serious damage to transport infrastructures and buildings. Dating carried out on tree findings recovered from a depth of about 22 m have given an age of about 9270±105 years BP.

Debris flows at Fiames and Acquabona

The debris flows occurring during heavy summer rainstorms can mobilise considerable amounts of material. In the Cortina d'Ampezzo area, large alluvial fans can be noticed. They were mostly originated by this type of process south of Fiames and Acquabona (Pasuto and Soldati, 2004; Berti *et al.*, 1999). These materials are of dolomitic origin with particle size distribution ranging from blocks to silt. Nearly all of them come from narrow defiles arranged on tectonic structures which can convey large amounts of detritus. These debris flows often interfere with transport infrastructures and developed areas and take place especially during summer when all the Dolomite region is visited by a large number of tourists. In these conditions of high vulnerability, even mass movements of modest magnitude could have severe consequences. As a matter of fact, on September 4th 1997, following a heavy rainstorm, the northern sector of the valley of Cortina d'Ampezzo was affected by a series of debris flows, two of which had serious consequences. One debris flow destroyed a road bridge and killed a police officer. A second flow, which affected the talus fans at the foot of Mt. Pomagagnon near the hamlet of Fiames, blocked state road no. 51 "Alemagna" and, after sparing some houses, barred the course of the Torrent Boite forming an impoundment, that still exists.

Active landslides

There are three main active landslides the Alverà, Staulin and Rio Roncatto earth flows (mudslides according to British usage) which affect clayey material derived from the weathering of the S. Cassiano Fm. The movements at Alverà and Staulin should be considered as partial remobilisations of the Cortina d'Ampezzo landslide, as mentioned above, while the Rio Roncatto landslide is a partial reactivation of the Lacedel landslide (Figure 12). These phenomena show different rates of displacement, although they evolve with similar mechanisms. The Rio Roncatto landslide shows displacements of about 2 m/year in the vicinity of Lacedel; the Alverà landslide of almost 20 cm/year and the Staulin landslide, which is almost dormant at present, of a few centimetres per year.

Historical records refer above all to landslides which have affected the villages of Alverà and Staulin: major events occurred in 1879, 1882, 1924, 1927, 1935, 1942 and 1951; in 1935, both these localities were included in the list of towns to be relocated at the expense of the Italian Government. As for Lacedel, which consists almost entirely of farming land, only a few records regarding periodic maintenance works on the roads were found. Remedial works, consisting of superficial drainage, retaining walls and soil mass remodelling, have been carried out on these landslides in the last few years with the aim of mitigating landslide effects on constructions.

Because of their potential risk for human activities, these slope movements have been monitored by means of automatic systems since 1989 (Angeli *et al.*, 1996; Gasparetto *et al.*, 1996). A few boreholes have been equipped with inclinometric tubes, Casagrande piezometers and wire extensometers. A climatic station, consisting of a tipping bucket rain gauge, an air thermometer and an ultrasonic snow gauge (giving the snow cover depth), has also been connected to the recording system. The sensors are linked to four peripheral units which acquire data by using instructions sent via radio from the central unit located in Cortina d'Ampezzo. The system has also been connected with a modem which allows real-time data transmission.

As regards the Staulin landslide, a well defined sliding surface was found at 19.5 m depth (Figure 13). The movement rate recorded is of 15.6 mm/year.

As for the Alverà landslide, inclinometric measurements revealed two sliding surfaces at a depth of about 5 m and 23 m (Figure 13). The movement rate recorded is 182.5 mm/year.

Concerning the Rio Roncatto landslide a principal sliding surface has been recognised at a depth of 10.5

Figure 12 - Location of the active mass movements: 1) Alverà, 2) Staulin and 3) Lacedel (after Soldati, 1999).

m (Figure 13). The average rate of movement recorded is 2.55 m per year for the recorded period. Another sliding surface has been found at a depth of 7.5 m. The life of the inclinometric tubes has been quite short due to significant displacement. Nevertheless, the data subsequently obtained from the wire extensometer has shown continuous deformation without a significant variation in velocity. This has been confirmed by repeated topographic surveys (Angeli *et al.*, 1996).

A good correlation between groundwater levels and displacements has been obtained, especially for the Alverà landslide to which a visco-plastic rheological model, capable of simulating the movement rate on the basis of recorded groundwater levels, has been applied. The model has been appropriately calibrated

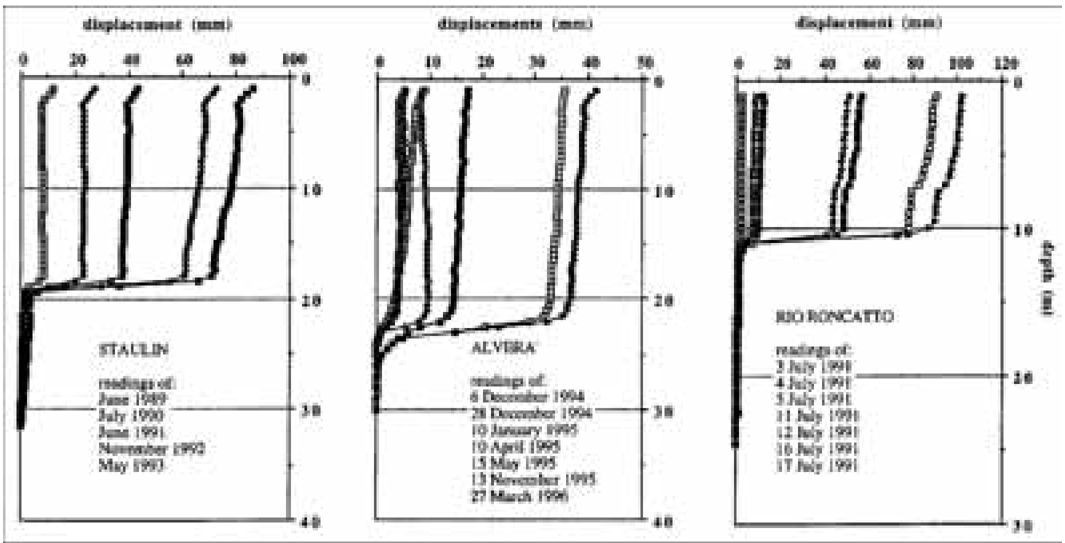

Figure 13 - Inclinometric curves of the Staulin, Alverà and Rio Roncatto landslides (after Soldati, 1999).

using the large amount of data collected, and it therefore gives an explanation of the relationship between groundwater level and the rate of the landslide movements. The validity of the model has been verified on the basis of a four year record of displacements and groundwater levels.

From Cortina d'Ampezzo to the Falzarego Pass

Mauro Soldati

After leaving the centre of Cortina d'Ampezzo, following state road no. 48, we go along the western slope of the valley and cross the track zone of the Lacedel landslide. Evident displacements and steps on the road in the vicinity of the hamlet of Lacedel mark this route. After some 500 m the excursion continues across a wood which contains huge boulders witnessing rock falls from the overlying dolomitic Pocol relief (Dolomia Cassiana).

Having reached the Belvedere of Pocol (1539 m), we can have an almost complete view of the Cortina d'Ampezzo area; so this is an ideal location for the observation of landslides and other geomorphological features in the valley.

From Pocol we move westwards towards the Falzarego Pass. After about 3 km, on the left-hand-side of the road, the landslide that was triggered by the Friuli Venezia Giulia earthquake of 15th September 1976 can be clearly observed. This is an earth flow affecting the slope underlying the spectacular dolomitic relief of Cinque Torri (2361 m). The movement affected the reddish clays and marls of the Raibl Fm., that are clearly visible in the landslide crown area (Zardini, 1979).

On the right-hand-side of the road, going up towards the Falzarego Pass, the imposing cliffs of the Tofàne mountain group surrounded by scree slopes can be observed. After a while, they are interrupted by avalanche cones and debris flows, while on the right-hand side there is a good view of the dolomitic peaks Averau (2647 m) and Nuvolau (2574 m) as well as the rock cliffs of Cinque Torri.

Stop 3.2:

The Lagazuoi Refuge

Alessandro Ghinoi

As we go up by cable car from the Falzarego Pass (2105 m) toward the Lagazuoi Refuge (2752 m) at the top of Mt. Lagazuoi, we cross a moraine ridge, active scree-slope deposits and a vertical dolomitic wall. Half way up the wall, it is possible to notice a "cengia" (a typical step-like morphology) chosen by the Italian Army in World War I as strategic location for a mine attack. In fact, the former front-line running between the Italian and the Austrian-Hungarian Armies is close by. For three years (from 1915 until 1918), there was a disastrous position-war that led to no significant advances for either of the two opposing armies (von Lichem, 1993). This part of Mt. Lagazuoi was conquered by the Italian Army while the westernmost part of it was under Austrian control. Both armies built a series of tunnels within the mountain with the aim of getting closer to the enemy and gaining positions. Usually at night, dynamite explosions could be heard from both sides: the louder the sound, the closer the enemy was. A few years ago, those tunnels were re-opened by a peaceful joint team of Italian, Austrian and German soldiers and they form an open-air museum of World War I that can be visited if one is equipped with helmets, torches and mountain boots.

Once at the top, there is a spectacular panoramic view showing all the beauty of the Dolomite scenery. Mt. Lagazuoi itself is a structural plateau slightly inclined toward NNW. It used to host a glacier that sculpted its top leaving clear signs of its erosional and depositional activity represented by *roches moutonnées* and till deposits.

To the west the dark shape of Col di Lana relief (2464 m) can be seen, which is composed of volcanoclastic sandstones (deposited by turbiditic currents during the Upper Ladinian) of La Valle Fm. Its NE bowl-shaped side used to contain a glacier whose deposits consist of frontal moraines and till, visible at its foot. Col di Lana, too, was the setting of some of the bloodiest battles between the Italians and Austrians, so bloody that it was renamed Col di Sangue (Bloody Hill; some 20,000 dead by the end of the war). An Italian mine destroyed the natural profile of the mountain crest, creating a pronounced concave shape and sweeping away 10,000 tons of rock (17th April 1916) (Dibona, 2000).

Behind Col di Lana, the glacier of Mt. Marmolada descends, from 3343 m, facing north, with a rather steep inclination. The mountain is formed by grey massive limestones (both detrital and organic) of the Lower Ladinian named "Marmolada Limestones". The current regional climatic ELA (Equilibrium Line Altitude) is between 2800 m and 2900 m, while the topographic ELA for this glacier has been calculated around 2700 m, using the 2:1-isoline method (Masini, 1998). During World War I the glacier hosted a network of tunnels and rooms where the Austrian-Hungarian Army had one of its most important headquarters at the front. The temperature inside the glacier was much more favourable than the outside one and the snow-avalanche risk was avoided. This under-ice protection system was called "the ice town" (Bobbio and Illing, 1997). Nowadays the glacier surface is used for skiing and for different events: one of the most curious being a golf tournament (!) in 1997.

If we look towards the west we encounter the bowl-shaped morphology of Mt. Settsass (2571 m). The clino-stratification of the dolomitic rocks mainly composing it (Dolomia Cassiana, Carnian) can be observed, with a dip-direction and inclination N30°. Of interest is also the so-called Piccolo Settsass (Small Settsass), which lies beneath the major peak looking south. Its presence seems to be ascribable to a sort of aborted coral reef, covered by deep-sea sediments coeval with the S. Cassiano Fm. Environmental conditions then became favourable for the development of a wider and thicker reef on top of deep-sea scarp sediments, which form the main body of Settsass (Panizza, 1991). On top of Mt. Settsass is a cap of thinly stratified dolomites of the Dürrenstein Fm. (Upper Carnian), formed above and within a tidal flat.

Behind Mt. Settsass, the Sella Group (3151 m) rises. Its base is formed by the south-verging progradation of a Carnian coral reef, with mega-breccia clinoforms shifting upward to massive dolomites. The depositional sequence continues with a toplap of the Dürrenstein Fm., followed by layers of marls, multi-colour clays, limestones and dolomitic sandstones of the Raibl Fm. (Upper Carnian). The top of the mountain is formed by Dolomia Principale (Upper Carnian-Norian), followed by a tiny summit of Calcari di Dachstein (Rhaetian) and nodular limestones of the Ammonitico Rosso Fm. (Malm-Dogger) at Piz Boé (3152 m). The two uppermost formations are separated by a well developed iron-manganese crust (Bosellini *et al.*, 1996).

The flat top of the Sella Group favoured the development of an ice cap that has mainly sculpted the eastern side of the mountain. A well-developed glacial cirque is visible on that side, with high vertical surrounding walls and a frontal moraine.

Looking NW, the Puez-Gardenaccia Group (2918 m) shows a depositional sequence similar to the Sella Group's. More frequent, at the top, are Jurassic and Cretaceous formations (Ammonitico Rosso and Marne del Puez, respectively), testifying a progressive and rapid uplift and emersion of the dolomitic area. One of the highest peaks of the group is Mt. Sassongher (2665 m) overhanging the village of Corvara in Badia; a source of rock fall events in the past it is likely to cause others in the future. Significant landslide events have recurred on the eastern side of the Puez-Gardenaccia Group, sometimes creating natural dams which favoured the development of lakes (Corsini *et al.*, 2000).

Looking north we face the Conturines Group (3077 m). This is a thick sequence of Dolomia Principale cut by a 45° south-verging thrust (Neogene) and with a summit of scattered Calcari Grigi di Fanes (Lower–Medium Lias). The slightly NE-inclined flat top was covered by an ice cap whose outlet lobes reached both the Valparola and the Boite Valleys (Sauro and Meneghel, 1995). Still well preserved are some frontal and lateral moraines of the Lateglacial whose relative dating allowed the reconstruction of that period's geomorphological evolution (Masini, 1998). In 1987, a man from Val Badia discovered the fossilised remains of a species of bear called *Ursus speleus*. These remains were found inside a hidden cave called "the hall of skulls", inside the mountain. This species is thought to have become extinct around 25,000 years ago, probably because of unfavourable climatic conditions and to hunting by man (Dibona, 2000).

One of the most famous mountain groups is visible if we look SE: the Cinque Torri (Five Towers, 2361 m). These are completely formed by Dolomia Principale, overlying the plastic terrains of the Raibl Fm. Their spectacular pillars are a clear example of lateral spreading that evolved into block slides, affecting a former plateau isolated by erosive processes active during the last glacial period. The pillars are now affected by degradation phenomena (mainly frost-shattering), giving rise to rock falls and topples. The overload caused by the weight of the pillars causes bulges to develop within the plastic material of the Raibl Fm. (Soldati and Pasuto, 1991).

Closing the panoramic circle there is, towards the east, a mountain group called Tofâne composed of three peaks: Tofana di Rozes (3225 m), Tofana di Mezzo (3243 m), and Tofana di Dentro (3237 m). All three are good examples of the Triassic stratigraphic sequence of the area, with Dolomia Principale as the dominant formation and only a small cap of Jurassic Calcari Grigi. As for Mt. Settsass and Mt. Conturines, the dip-direction of the strata is towards N and, as for Mt. Lagazuoi, the same south-verging thrust system has partly superimposed the group on top of the southern platforms. An important tectonic line with a SE-NW direction separates Tofana di Rozes and the other two Tofàne, while another line runs on the western sides of the Tofana di Mezzo and Tofana di Dentro. Of interest, from the geomorphological point of view, are the characteristic "cengie" on the southern slope of Tofana di Rozes, resulting from selective erosion driven by the alternation of rock types displaying diverse mechanical behaviour.

Figure 14 - Physiographic sketch map of Valparola; bedrock is depicted in grey

Stop 3.3:
Valparola
Alessandro Ghinoi

Leaving the Falzarego Pass for the Valparola Pass the road goes through a long, flat valley-bottom between Mt. Lagazuoi and Mt. Sass de Stria (2477 m). Both sides of the valley are formed by scree-slope deposits, debris cones cut by debris flows, protalus ramparts and rock fall deposits which have covered much of the till left since the end of the Lateglacial. Part of the blocks probably crumbled down after the frequent explosions of World War I, especially from the Cengia Martini. The road runs parallel to the Falzarego-Valparola south-verging thrust complex, which is thought to be composed of at least two parallel thrust planes: the Falzarego Line and the Valparola Line. The thrust complex has overlapped two distinct Carnian carbonate shelves, squeezing a succession of Carnian basinal facies in between (Bosellini *et al.*, 1982).

Before reaching the Valparola Pass, on the left side, there is a former Austrian bunker which has been recently refurbished and opened as a World War I museum. Opposite the bunker is a fossil rock glacier, better visible through aerial photographs, whose tongue crosses the road ending up a few hundred metres beneath it. Right at the pass, an example of a shallow lake formed by glacial erosion is visible (Figure 14); the lake's vegetation is typical of high-mountain lakes, with hydrophyte and hygrophyte plants. Near the lake some Mesolithic sites have been discovered, probably dating back to 10,000 - 7500 years BP (Panizza, 1991).

A set of parallel normal faults (NW-SE direction) have led to a complicated relationship between the formations outcropping NE of the pass, putting together, for instance, the Raibl Fm. (Upper Carnian) and the Dolomia Cassiana (Carnian). These faults are structures that formed during a Mesozoic extension phase and were reactivated as strike-slip faults during the Alpine phases.

From the pass as far as the northern foot of Mt. Settsass (Pre da Ciamena), three small parallel crests crop out from the red clayey terrains of the Raibl Fm., creating an interesting triple step-like morphology. These crests are made of the arenaceous layers of the uppermost part of the Raibl Fm. and, in this area, they show a clear cross-stratification.

The easternmost part of Mt. Settsass has been rounded by a glacier whose scouring effects are especially visible on the northern slope. Right underneath the highest peak (2571 m) facing N, two small glacial cirques now host debris cones covered by snow fields which sometimes last for an entire year.

The discovery of till deposits, lateral and frontal moraines (most of them widely covered by vegetation, therefore uneasy to detect) have led to a relative dating of the Lateglacial dynamic phases of the glacier bodies hosted within the valley. The oldest phase may date back to the Sciliar stadial. Testifying this phase are seven front moraines near the locality L'Armentarola (1615 m) and a lateral moraine, at the foot of Mt. Les Pizades, whose upper end is at 2025 m. At that time, it can be assumed that a considerable portion of the Valparola was still covered by an almost continuous ice body which may have left just a few peaks uncovered like, for instance, the Piz Ciampai (2295 m), near the pass.

Getting closer to the Gschnitz stadial, the almost continuous glacier body might have showed extensive fragmentation: the front moraine at Casere Eisenofner (1749 m) and another front moraine at Prei Glira (1996 m) belong to this period. The latter is completely covered by dense forest, its front is very high and steep and formed by huge dolomitic blocks. A small slope glacier on the northern side of Piz Ciampai belongs to the same period.

Before the onset of the Daun Stadium, one glacier tongue was hosted by the Vallone Pudres, a narrow N-S oriented valley at the westernmost part of Mt. Settsass, while at the foot of the highest part of the mountain (2571 m) two glacierettes were located. The latter were probably the last glacial bodies to remain also during the Daun stadial, thanks to their N orientation and to protection against direct solar radiation given by the surrounding vertical walls.

After the end of the Lateglacial, on the slopes left free by the glacier bodies, debris cones and scree-slope deposits began to accumulate, owing to the mechanical action of the frost/thaw cycles affecting the tectonically fractured dolomitic walls. During this phase some rock fall deposits (some of which of gigantic proportions) can also be placed. These rock falls resulted from a lack of support of the vertical slopes, previously provided by glacier bodies and to the erosional action that these bodies had carried out, widening and deepening tectonic discontinuities.

The rock glacier at Valparola Pass may have been active during this period: it has been demonstrated, in fact, how permafrost was active in the Dolomites starting from 10,500 years BP.

As in adjacent areas (Soldati *et al.*, 2004), with the onset of the Holocene, a more humid and less severe climate is likely to have favoured landslide activity within the plastic terrains of the S. Cassiano and La Valle formations. Currently we have datings (in this valley) for only two landslide events which allow us to place them within the Upper Sub-boreal.

Currently, debris flows are the most important geomorphological agent of the valley. Generated by debris deposits accumulating within the major fractures of the dolomitic massifs, they cut deeply through the apexes of the debris cones, leaving their deposition lobes at their bottom. The most impressive example lies beneath the Conturines Group, at the northern end of the valley. Its deposits have frequently dammed the Torrent Saré, but nowadays they have become a profitable source of income for a private mining company producing cement for the building industry. Active, but smaller debris flows descending from Mt. Lagazuoi affected the old path of the State Road and are approaching its new path, lower down the valley.

Even if they are not of great importance among active geomorphological agents, a few examples of earth slides-earth flows have to be mentioned, especially near springs at the permeability boundary between dolomitic formations and the plastic terrains of the Raibl and S. Cassiano formations.

Concerning the presence of man in Valparola, during the Middle Ages the valley was an extremely important stretch of the so called Strada della Vena (Road of the Vein) that was used for the transit of iron, cast iron and steel from the mines of Fursil (Province of

Belluno) to the Tyrolean markets. The valley has always been rich in water and wood therefore, along the sides of the Col dai Furns Torrent, a blast-furnace was built for smelting iron.
The soldiers of the Austrian-Hungarian Army took advantage of the valley's morphology to create one of the most impenetrable front lines of World War I. Their headquarters were on a debris deposit that filled a trench formed by a lateral spread (Plan de Lot, western side of Mt. Lagazuoi). Some tunnels were dug inside the mountain as a shelter from avalanches and for storing ammunition. The deep fractures within Mt. Settsass were perfect natural trenches and the strategic position of the mountain top made it a crucial observation point for anticipating the moves of the Italian adversary coming from Falzarego Pass.

From Valparola to Andraz Castle

Alessandro Ghinoi

After going back to the Falzarego Pass, we follow the winding road down to the River Cordevole valley. On our right side (looking west) we can see a N-S crest that starts from the southern tip of Mt. Settsass and reaches Mt. Col di Lana. On our left side (looking east) we can see a similar crest, with the same direction, that links Mt. Croda Negra (2518 m) to Mt. Pore (2405 m). Although covered by thick vegetation, the narrow valley is characterised by till deposits left by at least three different glacier tongues: one coming from the Valparola Pass, another from the Falzarego Pass and yet another from the southern faces of Mt. Settsass. The gigantic rock on top of which the Castle of Andraz (next stop) stands is an erratic boulder transported from Mt. Settsass (Dibona, 2000).

Stop 3.4:

Andraz Castle

Mario Panizza

The area where the Andraz Castle is built is made up of moraine deposits, left by two small glaciers that here were flowing together, one coming from the southern slope of Lagazuoi, the other from the Falzarego and Valparola passes. This deposit has a flat shape and makes up the right lateral moraine, whose frontal moraine can be found downslope, just after the confluence of Falzarego and Valparola torrents. Huge dolomite blocks can be observed; on top of one of these blocks the Andraz Castle is built (Figure 15). These blocks clearly came from the dolomitic cliffs of Lagazuoi and Sasso di Stria, and not from the overhanging Col di Lana, where volcanic rocks crop out.

Figure 15 - Andraz Castle (after Loss et al., 1998).

This castle is likely to have been built on the same site as a Roman *mansio*, by the Puchenstein family around 1000 AD. It is therefore coeval with the nearby ruins of the Rocca Pietore and Avoscan Castles. Around 1200 AD Hartwige von Puchenstein sold the castle to the Bressanone Bishop Prince Conrad von Rodenegg, who in 1221 appointed his two nephews Frederich and Arnold von Rodenegg-Schöneck as feudatories. In 1331 the Schönecks sold the rights of the castle and in 1350 it returned to the Bressanone Bishop Prince, who gave it as fief to the noble Stück family. When the male descent became extinct in 1461, the Bishop Prince installed the Captain of a small legion there. In 1803 the castle became the property of the Austrian Government. In 1853 it was sold to a local man, who, in order to recuperate the money spent, destroyed the whole roof and sold off the timber and other materials that made up this stronghold. It is said that the family of the new owner used old documents and parchments found in the castle to fuel their fires for a whole winter. In 1986 the castle was derelict, but the Veneto Region, which now owned the stronghold, decided to start complex restoration works which have produced good results, notwithstanding the decision not to restore the original wooden roof, but to substitute it with an iron framework and glass panels.

From Andraz to Arabba and the Upper Cordevole Valley

Alessandro Pasuto

Starting from Andraz, we travel along the left side of the upper Cordevole Valley. The road runs high above the valley floor near the slopes of Col di Lana, which, as previously mentioned, witnessed fierce and bloody fighting during World War I. Eventually, the village of Livinallongo del Col di Lana is reached. At a narrow hairpin bend, just before the village, we cross a small bridge over Rio Chiesa. This stream is dry for most of the year, but in the past it has caused serious debris flows and overfilling of the underlying square used as a parking lot. After Livinallongo, the road continues along the hillside, across the Permian-Triassic sequence typical of the Dolomite environment, as far as Arabba, at the foot of the Sella Group relief. This mountain ridge, mainly made up of Norian-Carnian dolostones, has a roughly circular shape and is surrounded by four mountain passes which link the Cordevole Valley with the Fassa, Gardena and Badia Valleys, respectively. Once in Arabba, we turn right and proceed uphill towards the Campolongo Pass. Later, after passing the administrative border between Veneto Region and Alto Adige/South Tyrol, we go down to the Badia Valley and, eventually, to Corvara.

Corvara in Badia and Upper Badia Valley

Alessandro Corsini

Introduction

The area of Corvara in Badia is located in the central part of the Dolomites (Eastern Italian Alps), in the Autonomous Province of Bolzano (South Tyrol) and is situated in the southernmost sector of the Alta Badia district (Upper Badia Valley). Access to the area of Corvara is possible through the Gardena Pass (2121 m), the Campolongo Pass (1679 m) to the south-west and the narrow gorge of the Gadera stream to the north, between the villages of Corvara in Badia and La Villa.

The mountain ridges culminate in the dolomite peaks of Puez-Gardenaccia (3025 m) and Sella (3151 m) in the western sector and in the marly pyroclastic peaks of Piz la Villa (2078 m), Col Alto (1980 m) and Pralongià (2571 m) in the eastern sector (Figure 16). The main watercourses (T. Pissadù, T. Rutorto and, after their confluence, T. Gadera), flow across the valley floors, which range in altitude between over 2000 m near the passes and about 1500 m near the northern boundary of the study area. The slopes are usually subvertical in the upper part of the ridge (in correspondence with the dolomite peaks) and moderately inclined in the mid-lower part.

Figure 16 - Location of the Corvara in Badia area (after Corsini et al., 2001).

Geological setting

The area of Corvara in Badia is illustrated by Sheet no. 11 "Monte Marmolada" of the Italian Geological Map at a 1:100,000 scale (Servizio Geologico d'Italia, 1970) and Sheet no. 028 "La Marmolada" of the Italian Geological Map at a 1:50,000 scale (Servizio Geologico d'Italia, 1977). Formations cropping out in the area have been illustrated in the geological setting of the Dolomites and are exhaustively described in literature (Leonardi, 1967; Brondi *et al.*, 1977; De Zanche *et al.*, 1993; Bosellini, 1996): they make up the sedimentary sequence of the Dolomites in the Upper Permian-Norian interval (Servizio Geologico d'Italia, 1977). The mountain groups of Puez-Gardenaccia and Sella are made up of massif or stratified dolostones from the Upper-Middle Triassic (Dolomia Cassiana, Dürrenstein Fm. and Dolomia Principale) and, subordinately, by marls and pelites from the Upper Triassic (Raibl Fm.) (Figure 17). The Pralongià ridge and the upper part of the slopes located beneath the Sella and Gardenaccia dolomite cliffs are made up of lithological complexes, prevalently with ductile mechanical behaviour and made up of alternating sequences of Upper-Middle Triassic calcarenites, marls and pelites (S. Cassiano Fm.) and Middle Triassic pyroclastic arenites and siltites (La Valle Fm.). The crest of the Piz La Villa - Col Alto ridge is made up of Middle Triassic arenites, pyroclastic conglomerates and lavas (Fernazza Fm.; Pillow Lavas), which are affected by a dense network of

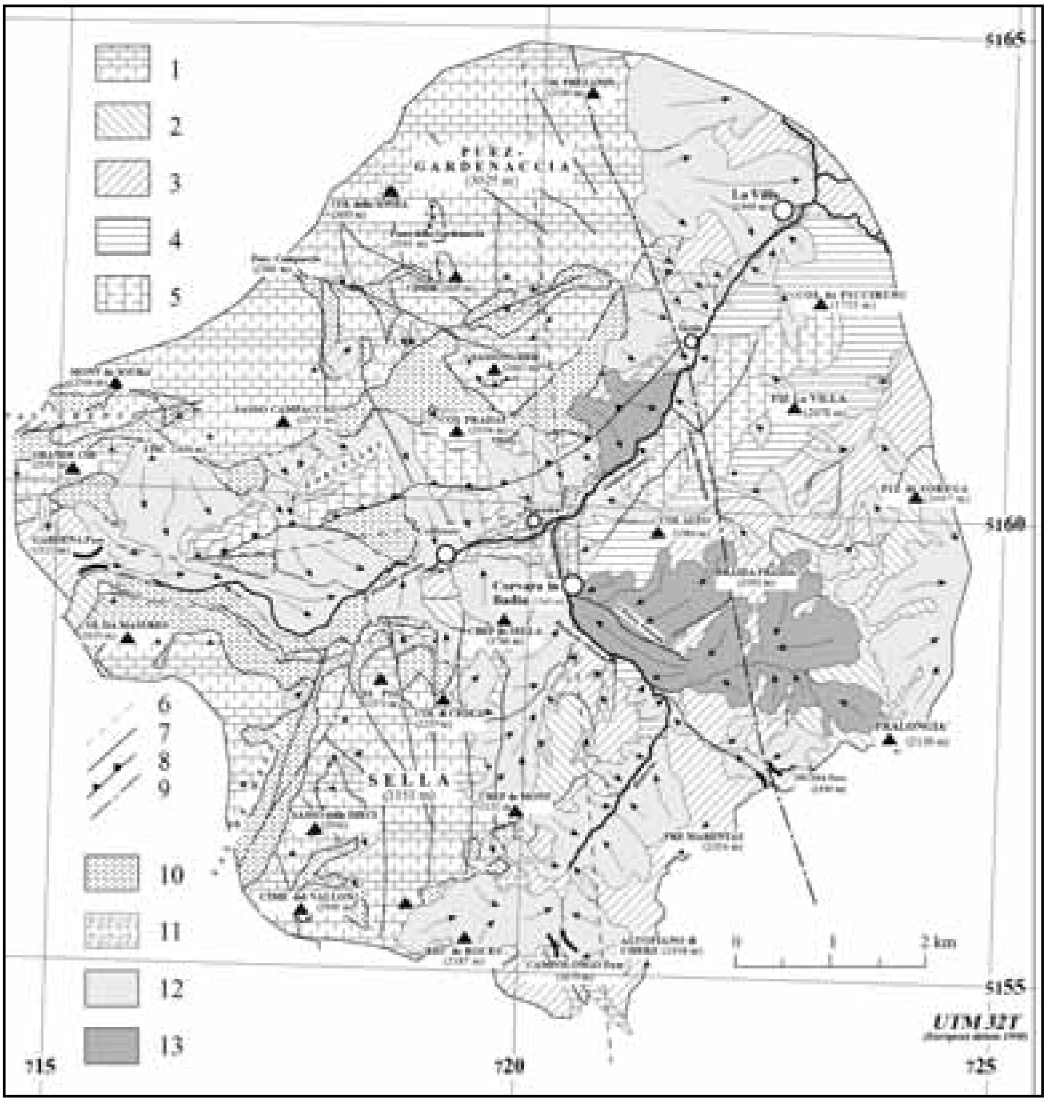

Figure 17 - Geological schematic map of the Corvara in Badia area. Legend: 1) dolomites and marls (Middle and Upper Triassic); 2) alternances of calcarenites, marls and claystones (Middle and Upper Triassic); 3) alternances of sandstones and siltstones (Middle Triassic); 4) volcanoclastic arenites and conglomerates (Middle Triassic); 5) limestones, sandstones and marls (Upper Permian-Lower Triassic); 6) palaeotectonic line (Upper Permian-Lower Triassic); 7) meso-neoalpine fault (Tertiary); 8) meso-neoalpine overthrust (Tertiary); 9) neotectonic fault (Plio-Quaternary); 10) scree slopes, scree fans (Lateglacial-Holocene); 11) alluvial or lacustrine deposit (Lateglacial-Holocene); 12) landslide (Lateglacial-Holocene); 13) Corvara and Col Maladat landslides (after Corsini et al., 2001).

joints with substantially brittle mechanical behaviour. In the north-western sector of the Piz La Villa - Col Alto ridge, and in the lower portion of the northern and eastern slopes of the Puez-Gardenaccia ridge, rock types displaying different kinds of mechanical behaviour crop out; they are mainly Upper Permian-Lower Triassic limestones, sandstones and marls (Bellerophon Fm.; Werfen Fm.; Richthofen Conglomerate; Contrin Fm.; Livinallongo Fm.).

Tectonic setting

The tectonic setting of the area results from three main deformational phases: Upper Permian basin subsidence and fracturing, Upper Oligocene-Lower

Miocene neoalpine and mesoalpine compressive phases and Middle Miocene-Upper Miocene south-alpine compressive phases (Bosellini, 1965; Leonardi, 1967; Doglioni, 1987; Doglioni and Bosellini, 1987; Doglioni, 1990; Castellarin *et al.*, 1992). Some of the structures derived from these ancient deformational phases were reactivated during the Pliocene and the Quaternary. This has been demonstrated for the N-S fault which cuts through the Sella Group along the Val de Mesdì. This fault is considered active between 5.2-0.7 Ma BP by Carton *et al.* (1980) and Zanferrari *et al.* (1982) as well as by Panizza *et al.* (1978), Slejko *et al.* (1989) and Castaldini and Panizza (1991). Also the NNW-SSE strike-slip fault, which runs on the Col Alto – Pralongià – Passo Incisa lineament, has been shown to have been reactivated on the basis of several congruous geomorphological elements (among which also the deep-seated failures affecting the Col Alto – Pralongià slope, among which the Corvara landslide). It is thought to have been active, with a transcurrent dextral movement during the Plio-Quaternary (Panizza *et al.*, 1978; Carton *et al.*, 1980; Zanferrari *et al.*, 1982; Slejko *et al.*, 1989; Castaldini and Panizza, 1991, Corsini and Panizza, 2003).

Geomorphological setting

On the whole, the geomorphological setting of the Corvara in Badia area is strictly related to the slopes' geological structure and reflects the different climatic conditions of the Quaternary (Corsini *et al.*, 1997). From a morpho-structural viewpoint, selective erosion processes leading to the exhumation of previously buried geological structures, have played a major role starting from the Miocene-Pliocene, when the main valleys of the area started to develop, with the modelling of the Gardena Pass and Pralongià. At present the slopes are practically vertical at the margin of the dolomite reliefs, highly inclined in correspondence with the Lower-Middle Triassic outcrops of limestones and pyroclastic arenites and mildly inclined and undulated in correspondence with the north-western slope of the Col Alto – Pralongià relief, where the Late Triassic sedimentary rocks crop out. As regards the landforms and deposits linked to the Würm Pleniglacial and Lateglacial periods, worthy of note are the Pleniglacial moraine ridges near Pralongià and the Gardena Pass, as well as the cirques, suspended valleys, glacial deposits and moraine arcs which witness the presence and progressive disappearance of valley glaciers which had developed during the Lateglacial and Holocene. In Pralongià shale clasts are found trapped inside Pleniglacial deposits or inside landslide bodies. These confirm the Austrian provenance of the Last Glacial Maximum glaciers, which stretched south as far as an altitude of 2300 m, as already hypothesised by Penck and Brückner (1909), B. Castiglioni (1936), Sacco (1939) and G.B. Castiglioni (1964). On the other hand, periglacial features such as the talus and alluvial fans bordering the base of the Sella and Puez-Gardenaccia Dolomite slopes are also evident. In the area of Corvara in Badia the landforms and deposits due to slope movements are particularly important. Landslides of various types and dimensions affect practically every sector of this area (Figure 17); some of these, in particular those which involve argillaceous materials, are at present active.

Landslide causes

The situations favouring landslide processes in the area of Corvara in Badia are extremely varied, mainly depending on the different rock types involved and stratigraphic and tectonic relations between outcropping formations. The development of numerous slope movements in correspondence with the margins of the Sella and Puez-Gardenaccia groups appears to be linked to the overlapping of jointed competent rocks, such as dolostones, on materials showing substantially ductile behaviour, such as the S. Cassiano and La Valle formations. In this case, various types of movement combine and take place on different stretches of the same slope, affecting materials with different mechanical characteristics, thus giving origin to vast landslides with a complex and composite movement style. Other complex landslides started because of structural weakness situations, or the presence of pelitic rock types which give rise to rotational slides accompanied by earth flows.

Slope instability

Many landslides have been identified and mapped in recent years in Corvara in Badia, and in the Upper Badia Valley in general.

In order to reconstruct Holocene gravitational slope dynamics, great importance is assumed by: i) rotational slides–earth flows, involving alternating sequences of Upper-Middle Triassic pyroclastic arenites, siltites, calcarenites, marls and pelites along the north-western slope of the Col Alto – Pralongià ridge; ii) rotational slides affecting Upper Permian-Lower Triassic limestones, sandstones and marls cropping out along the eastern slope of the Puez-Gardenaccia Group. Out of 40 radiocarbon age datings, some post-glacial phases of activity were determined in 13 (for

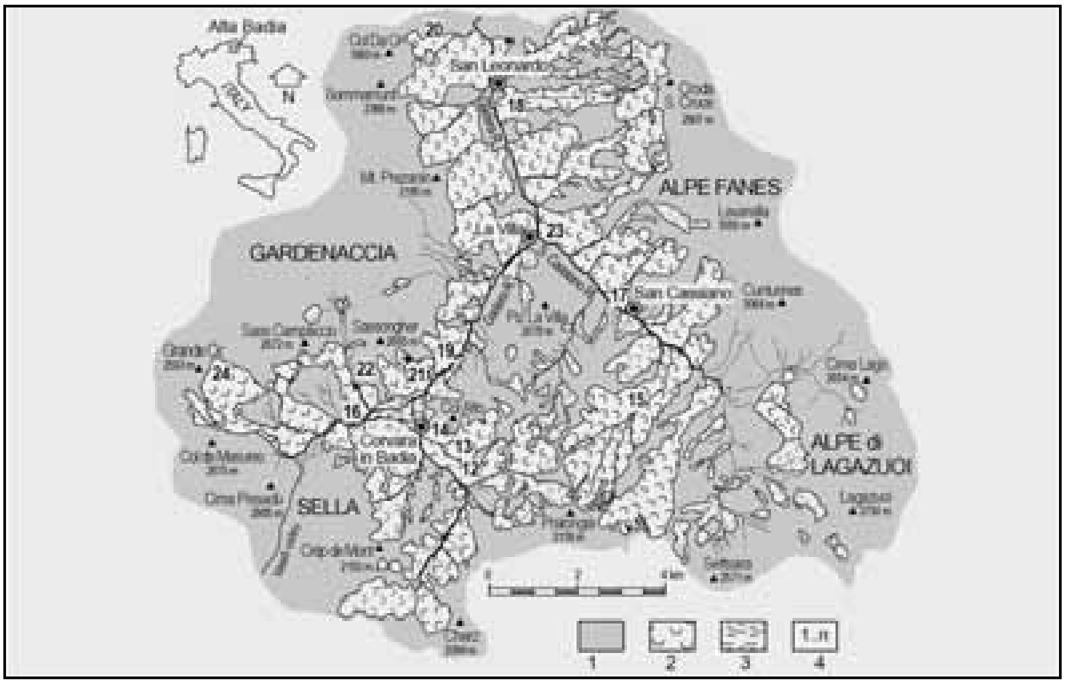

Figure 18 - Geomorphological sketch of Alta Badia area (heights are given in metres). Legend: 1) bedrock; 2) landslide; 3) lacustrine deposits; 4) landslide numbering according to Table 2 (after Soldati et al., 2004).

landslide location and dating information see Figure 18 and Table 2).

As regards only the Corvara in Badia area, more than 80% of the territory is covered with landslide deposits, ascribable to many different landslide types. Nevertheless, considering also socio-economic aspects and risk implications, the main landslides in this area are: the Corvara landslide unit, the Colfosco landslide and the Passo Gardena landslide. These mass movements are described in the following sections.

Stop 3.5:

Corvara landslide

Alessandro Corsini

The Corvara landslide affects an area of more than 2.5 km^2 located immediately uphill of the village of Corvara in Badia (Figure 19). It can be classified as an active, slow moving, deep-seated rotational earth slide – earth flow, which has an estimated overall volume of more than 300 million m^3 (Corsini *et al*., 1999). Present day movements of the Corvara landslide, ranging from 0.1 to more than 1 m/year at the surface and at depth, cause, on a yearly basis, damage to power lines, cable cars and, in particular, to state road no. 244, which connects Corvara to the Campolongo Pass and to the neighbouring tourist centre of Arabba. Moreover, over the past 30 years, many of the buildings in the village of Corvara have been erected in the proximity of the landslide toe, and could be affected by any advance of this gravitational process or by flash floods originating from the collapse of any ephemeral landslide dam that could be formed on the riverbeds of the streams flanking the landslide toe. Since the specific and total risk arising from this situation is potentially significant, both in relation to the extent and magnitude of the landslide and the socio-economic vulnerability and value of the elements at risk, monitoring has been carried out in close collaboration with provincial and local agencies within the framework of European and National research projects.

The morphology of the Corvara landslide, shows distinct source (S), track (T) and accumulation (A) areas (Figure 19). The source itself is made up of four morphological sectors (S1, S2, S3, S4) separated by rock ridges. The main morphometric characteristics are listed in Table 3. It is known from several radiocarbon dates that the landslide has existed since at least

ID	Landslide name	Landslide type	Sample code	Sample type	Site of collection	Depth (m)	Conventional Age (^{14}C yr BP)	Calendar Age 2σ (Cal. yr BP)
12	Corvara	rotational rock slide - earth flow	B-112032	wood	borehole (C1)	25.70	8820 +/- 50	10,152 - 9632
12	Corvara	rotational rock slide - earth flow	B-112033	wood	borehole (C1)	26.40	8560 +/- 90	9709 - 9334
12	Corvara	rotational rock slide - earth flow	B-112031	organic sediment	borehole (C3)	22.70	7920 +/- 70	9009 - 8543
12	Corvara	rotational rock slide - earth flow	B-154704	wood	borehole (C6)	69.70	8020 +/- 60	9030 - 8650
12	Corvara	earth flow	Ki-9233	wood	borehole (C6)	47.50	5543 +/- 72	6471 - 6199
12	Corvara	earth flow	Ki-9230	wood	borehole (C6)	19.10	4616 +/- 64	5575 - 5052
12	Corvara	earth flow	B-112029	wood	borehole (C2)	7.50	4260 +/- 70	5025 - 4575
12	Corvara	earth flow	B-112030	wood	borehole (C2)	20.00	4260 +/- 70	5025 - 4575
12	Corvara	earth flow	Ki-9234	tree trunk	erosion scarp	8.00	3888 +/- 64	4513 - 4094
12	Corvara	earth flow	B-105976	tree trunk	erosion scarp	6.00	3830 +/- 60	4417 - 3999
12	Corvara	earth flow	B-105977	tree trunk	erosion scarp	4.50	2860 +/- 60	3207 - 2792
12	Corvara	earth flow	B-93975	tree trunk	erosion scarp	5.00	2490 +/- 60	2750 - 2352
12	Corvara	earth flow	B-112034	wood	borehole (C4)	37.40	2260 +/- 50	2351 - 2129
13	Arlara	rotational rock slide - earth flow	B-105975	tree trunk	erosion scarp	3.5	6870 +/- 50	7789 - 7592
14	Col Alto	mud flow	B-93976	tree trunk	excavation	2.00	2350 +/- 60	2708 - 2183
15	Rio Pocol	earth flow	B-128367	tree trunk	erosion scarp	5.00	950 +/- 50	954 - 736
16	Colfosco	rock fall / avalanche	B-112024	tree trunk	excavation	3.50	4420 +/- 70	5030 - 4978
16	Colfosco	sedimentation in dam lake	Ki-9232	organic sediment	borehole (S4)	15.10	7088 +/- 72	8106 - 7751
16	Colfosco	sedimentation in dam lake	B-154701	wood	borehole (S3)	12.80	6810 +/- 60	7700 - 7560
17	San Cassiano	earth flow	B-128365	tree trunk	excavation	3.50	2260 +/- 60	2869 - 2715
17	San Cassiano	earth flow	B-128366	organic sediment	excavation	7.40	2610 +/- 60	2845 - 2496
ID	Landslide name	Landslide type	Sample code	Sample type	Site of collection	Depth (m)	Conventional Age (^{14}C yr BP)	Calendar Age 2σ (Cal. yr BP)
18	San Leonardo	earth flow	B-128368	wood	borehole (SL2)	11.00	4890 +/- 70	5839 - 5475
19	Col Maladat	rotational rock slide (begin of dam lake sedimentation)	B-112026	wood	borehole (B5)	20.20	9080 +/- 70	10,401 – 10,154
19	Col Maladat	sedimentation in dam lake	B-112028	wood	borehole (B6)	20.24	8810 +/- 70	10,173 – 9557
19	Col Maladat	sedimentation in dam lake	B-112025	wood	excavation	5.64	7740 +/- 80	8696 – 8386
19	Col Maladat	sedimentation in dam lake	B-112027	wood	borehole (B6)	4.40	6460 +/- 90	7563 – 7213
19	Col Maladat	sedimentation in dam lake	B-112023	wood and peat	excavation	4.20	3550 +/- 70	4075 – 3640
19	Col Maladat	sedimentation in dam lake	B-112022	wood	excavation	1.43	3210 +/- 60	3626 – 3272
19	Col Maladat	sedimentation in dam lake	HD-19408	wood	excavation	0.48	1249 +/- 22	1263 1090
19	Col Maladat	debris flow	B-154705	organic sediment	excavation	3.50	1320 +/- 60	1320 1100
19	Col Maladat	debris flow	Ki-9229	chalk	excavation	5.00	1281 +/- 64	1306 1058
19	Col Maladat	debris flow	Ki-9231	tree trunk	excavation	6.00	2537 +/- 64	2771 2359
20	Col da Oi	earth flow	Ki-7757	tree trunk	erosion scarp	2.00	3050 +/- 50	3363 – 3079
21	Greif-Pescosta	mud flow	Ki-9226	tree trunk	excavation	3.50	1490 +/- 64	1525 – 1288
21	Greif-Pescosta	mud flow	B-154703	peat	excavation	3.30	810 +/- 60	900 – 660
21	Greif-Pescosta	mud flow	B-154702	peat	excavation	2.80	340 +/- 60	510 – 290
21	Greif-Pescosta	mud flow	Ki-9227	organic sediment	excavation	2.50	378 +/- 64	529 – 298
22	Stella Alpina	debris flow	B-154706	organic sediment	excavation	5.00	1950 +/- 60	2030 – 1800
23	Sottocianin	rotational rock slide - earth flow	Ki-9228	tree trunk	excavation	2.50	8143 +/- 72	9398 8814
24	Passo Gardena	translational rock slide	R99-1258	surface exposure dating by cosmogenic ^{36}Cl AMS measurement in dolomite rock from the main scarp			from about 11,800 to 8500 yr BP with erosion rates from 20 to 10 mm/ka respectively	

Table 2 - Radiocarbon dating in Alta Badia; data calibrated with Calib 4.1 by Stuiver and Reimer (1993), data set by Stuiver et al. (1998) (after Soldati et al., 2004; modified).

about 10,000 cal BP, and that it underwent a second major phase of morphological development from about 5000 to 2500 cal BP (Corsini, 2000; Corsini *et al*., 2001). The landslide is active at present, with movement rates ranging from about 0.01 to 2 m/year. The distribution of activity is retrogressive at the crown and at the flanks of the source areas, slightly enlarging at the sides of the accumulation area, slightly advancing at the turn of the landslide into the main valley and potentially advancing at the toe.

From a geomechanical viewpoint, the bedrock can be defined as a weak rock (Bieniawski, 1989), that has a B1-type lithological and structural complexity (A.G.I., 1985). It is made up of a more or less regular alternation between sub-metre strata of hard rocks (marls, arenites and/or marly limestone with an uniaxial compressive strength between 50 and 25 MPa) and sub-metre to metre packets of thin strata of weak rocks (claystone or clayshales with uniaxial compressive strength <25 MPa) (Corsini, 2000). The ratio between the hard rock component and the weak rock component is variable, but is generally higher in the La Valle Fm. (average hard/weak ~ 1.5 to 2) than in the S. Cassiano Fm. (average hard/weak ~ 1). On

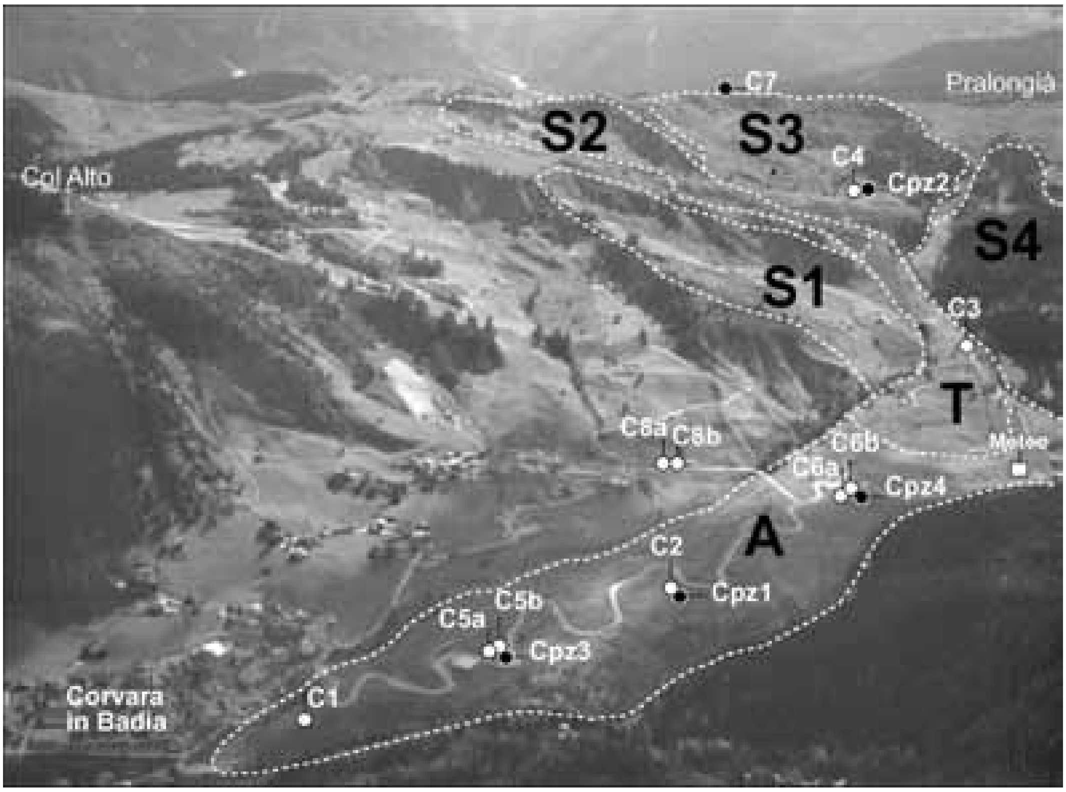

Fig 19 - View of the Corvara landslide, illustrating the location of the instrumented boreholes. Legend: A) accumulation area; T) track zone; S1, S2, S3, S4) source areas (after Corsini et al., in press; modified).

Characteristic	*Value*
Surface	~ 2.5 km²
Depth	~ 50 to 100 m
Volume	>300 millions m³
Max altitude at crown	2083 m
Min altitude at foot	1543 m
Difference of altitude	540 m
Max length	~ 3500 m
Max width	~ 1300 m
Length/width ratio	~ 2.7
Average slope gradient	~ 10%

Table 3 - Morphometric characteristics of the Corvara landslide (after Corsini et al., in press).

the unstable slope, the stratigraphic boundary between the two formations is gradual and is related to the progressive substitution of dark-coloured volcanic arenites (typical of La Valle Fm.) with yellowish marly limestone (typical of the S. Cassiano Fm.). Other major persistent discontinuities, besides stratification, affect these rock masses with spacing ranging from a few centimetres to decametres. Jointing and faulting is related to tectonic and neotectonic compressive stresses that have played an important role in determining slope instability in this particular location (Corsini and Panizza, 2003). The resulting mechanical behaviour of the bedrock can be defined as brittle near the surface, where confining loads are low and joints tend to split the mass; brittle-plastic at a depth of some tens of metres, where sliding can occur on clayey strata together with rupture of the hard strata; viscous-plastic at greater depths, where rock creep can take place on a longer term basis. Boreholes drilled down to the bedrock starting from axial points in the source, track and accumulation zone of the landslide, have shown that landslide material reaches a maximum thickness of about 50 to 100 m. Similar estimates of the depth of the bedrock have been obtained by some electric resistivity profiles carried out along traces linking or crossing boreholes (Corsini and Arndt, 2003) and refraction/reflection seismic survey.

Landslide material consists of clayey silts or silty clay, with mixed gravels and blocks made up of arenites, marly limestone and dolostone. Geotechnical parameters are summarised in Table 4. From borehole stratigraphy and radiocarbon dating on wood remnants found in cores, a system of four main layers seems to exist, at least on the track and accumulation zones of the landslide: an uppermost layer of landslide material that is relatively loose and inconsistent, a lower layer of landslide material that is more compacted and consistent, a cap of highly weathered bedrock and then the lowermost undisturbed bedrock. The two layers of landslide material have been correlated to different phases of development of the phenomenon, with the lower one being the result of a series of landslide events dating back to the early Holocene and the upper one representing landslide events dating prevalently from 5000 cal BP to 2500 cal BP and to the present (Corsini *et al*., 2001). In source area S3 only the most recent and loose layer of landslide material is present, and this seems to indicate that a retrogression of the landslide scarp of more than 500 m has taken place from about 3000 cal BP.

Parameter	*Value range*
Unit weight	15 ÷ 22 kN/m^3
Clay Fraction	30 ÷ 50 %
Plasticity Index	20 ÷ 30 %
Natural water content	20 ÷ 45 %
Undrained cohesion	30 ÷ 100 kPa
Effective friction angle	20 ÷ 30 °
Effective cohesion	10 ÷ 35 kPa
Residual friction angle	15 ÷ 20 °

Table 4 - Geotechnical characteristics of the landslide material (after Corsini et al., in press).

The monitoring network for the Corvara landslide has been designed to accommodate devices capable of defining and recording movement rates at the surface and underground, as well as groundwater levels and rainfall (Figure 19). A meteorological station, consisting of a rain gauge for rainfall recording, an air temperature thermometer and an echo-meter for snow depth measurement, was installed in the middle of the landslide in October 1999. As regards underground movements and the water table, devices such as inclinometers, down-hole wire extensometers, large diameter TDR cables and electric piezometers (connected to data loggers) have been located in boreholes on different points of the Corvara landslide. For surface movements, a network of 47 benchmarks for measurement with RTK differential GPS technique has been installed on the whole Col Alto – Pralongià slope.

Results obtained through monitoring with inclinometers and TDR cables have shown that multiple shear zones overlap at various depths within the landslide mass and that the maximum depth of the shear zones on each individual borehole ranges from 20 to about 50 m (Figure 20). The thickness of these shear zones is generally in the order of 1 to 2 m, and it can be appreciated that in between two shear zones, the displaced material slides as a whole without significant evidence of flow-like deformation.

GPS measurements from September 2001 to September 2002 have shown that horizontal movements have ranged from a few cm to more than one metre, while vertical movements have ranged from a few centimetres to about ten centimetres (Figure 20). It can be noticed that the maximum displacement rate occurs in the uppermost part of the accumulation lobe and in the track zone (from 0.20 to 1.20 m/year).

Piezometers have been used to prove a working hypothesis that a system of overlapping confined aquifers exists in the landslide body as the result of horizons of coarser materials occurring in the clayey mass as shown by borehole cores, For this reason, one type of measurement has regarded the

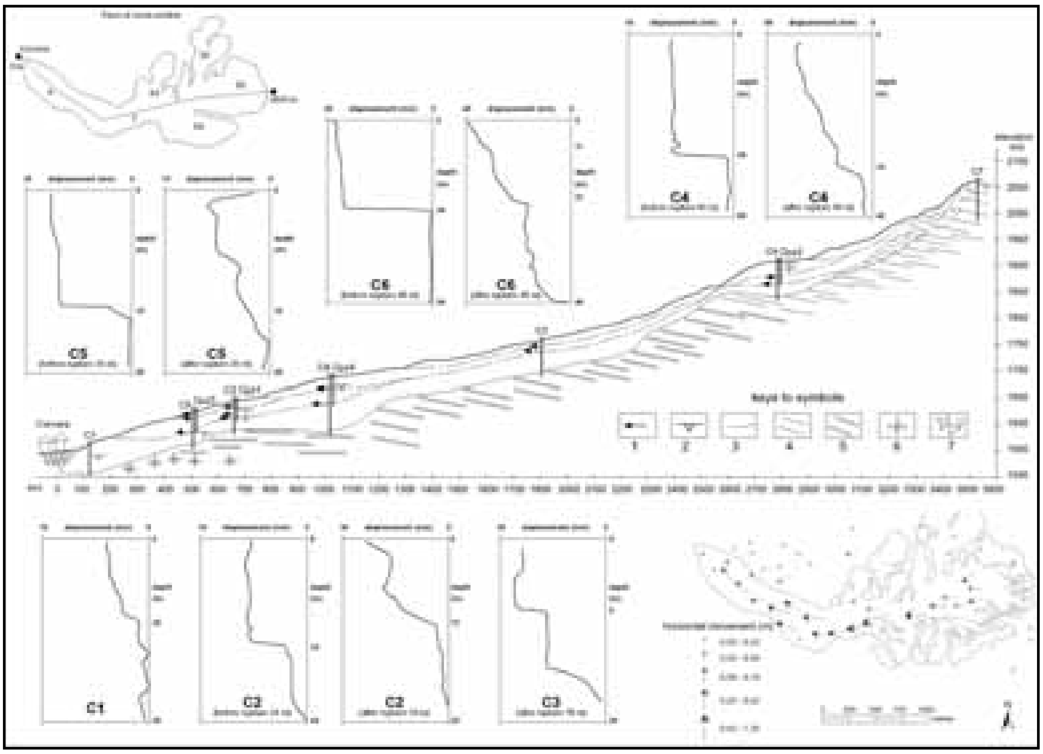

Figure 20 - Cross section based on boreholes, monitoring data and geophysical investigation. Legend: 1) Active shear surface, revealed by inclinometres or TDR; 2) Groundwater level, measured by piezometres; 3) Interpreted active shear surface; 4) S. Cassiano Fm. (turbidites); 5) La Valle Fm. (turbidites); 6) La Valle Fm. (massive sandstones); 7) Alluvial deposits. Horizontal Movements recorded by GPS monitoring from September 2001 to September 2002 (after Corsini et al., in press; modified).

water table generated by the contribution of all the aquifers intercepted by the boreholes crossing the landslide mass. This type of measurement has been carried out with electric transponders installed in unsealed piezometers C1, C2 and C4. Another type of measurement regards the pressure of water around main shear surfaces, that on the basis of borehole data seems to occur in correspondence with coarser horizons, which probably act as confined aquifers. To measure this type of pressure, electric transponders have been installed in open pipe piezometers fissured only for a tract including the main shear surface and sealed above and below.

Groundwater levels recorded form 2000 to 2002 in unsealed inclinometer C4 are plotted in Figure 21, together with daily rainfall recorded in the meteorological station specifically installed in the Corvara landslide and displacement data from the 29 m depth sliding surface. It must be remembered that inclinometer C4 revealed a major sliding surface at about 40 m in depth from 1997 to 1999, which led to rupture of the casing in August 1999. The sliding surface at 29 m in depth shown in Figure 21, is thus a secondary one that, however, has been detectable from 1997 to date, thus allowing a comparison between groundwater level and movement rates in this location. Figure 21 shows that since the high standing water table conditions of summer 2000, displacement rate along the 29 m deep sliding surface has had an increase only in the period following the high precipitation of summer 2001; since then the landslide has been practically dormant. This leads to the conclusion that while the deeper sliding surface has probably moved continuously before and after the water table rise, shallower sliding surfaces are more influenced also by smaller water table fluctuations related to precipitation patterns.

DAY 4

From Corvara to the Gardena Pass
Alessandro Corsini

Mesdì Valley and its extinct glacier

The N-S directed Val Mesdì, which cuts the Sella Group almost in half has developed in correspondence with a normal fault ascribable to the Neogene (Tertiary Alpine compressive phases, Leonardi, 1967; Doglioni and Bosellini, 1987; Castellarin *et al.*, 1992; Doglioni, 1990; Bosellini, 1996). This is the principal fault among a sub-parallel set of normal faults affecting the whole eastern portion of the Sella Group that, as a consequence, has assumed a marked and characteristic step-like morphology. The Val Mesdì fault line has been considered as having been active between 5.2-0.7 Ma BP by Carton *et al*. (1980) and Zanferrari *et al*. (1982) as well as by Panizza *et al*. (1978), Slejko *et al*. (1989) and Castaldini and Panizza (1991).

During the Holocene glaciers progressively withdrew into high mountain cirques and, at times – such as probably during the climatic optimum (hypsithermal) – they completely disappeared. The Little Ice Age of the 19th century caused the formation of new small glaciers in many narrow valleys and cirques that progressively

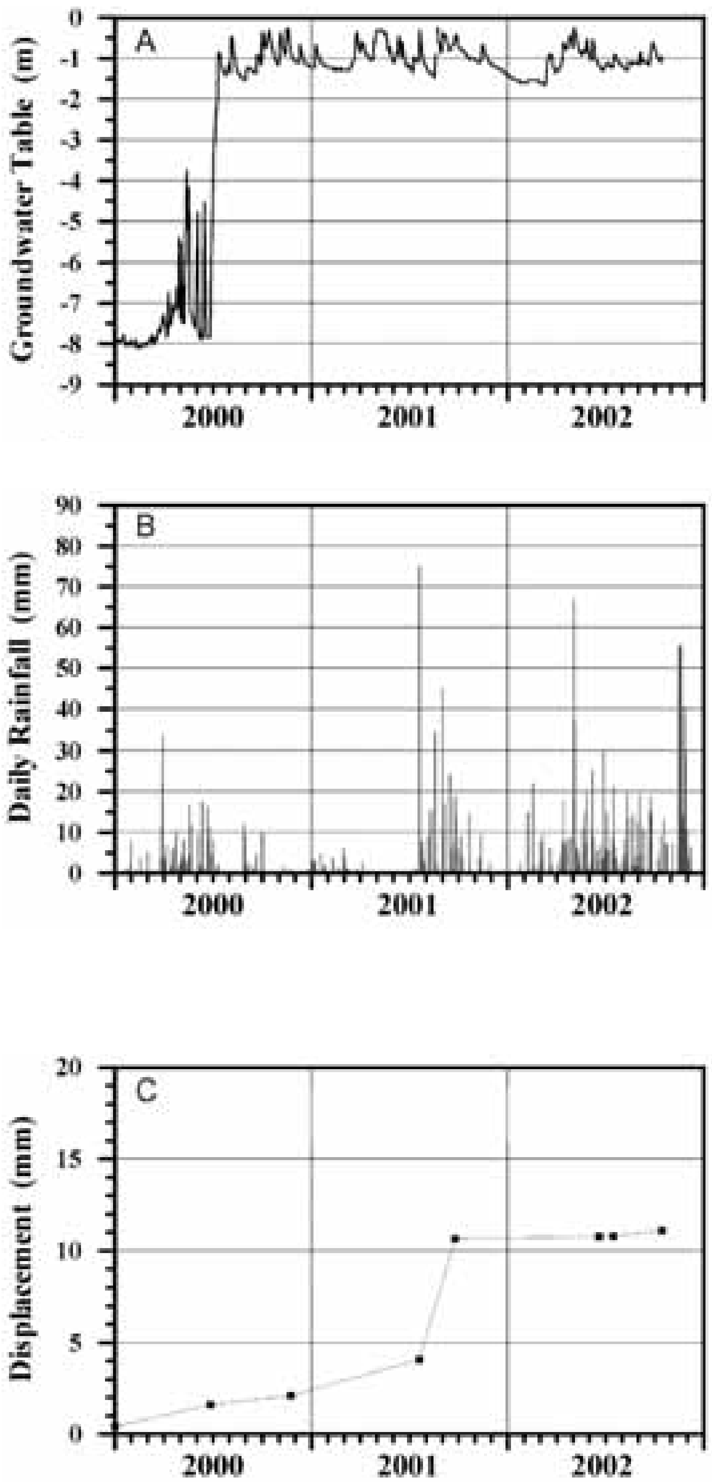

Figure 21 - Relationships between groundwater table fluctuations in C4 (A), precipitation records from 2000 to 2002 from the meteorological station installed in the Corvara landslide (B) and displacements (C) (after Corsini et al., in press).

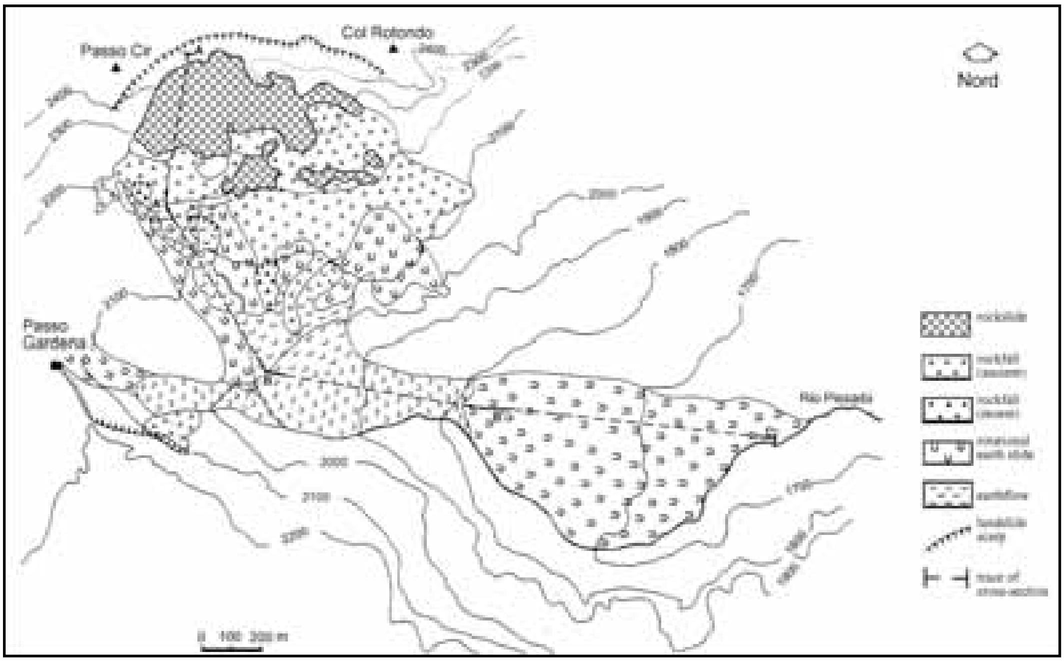

Figure 22 - Sketch map of the Passo Gardena landslide.

melted in the 20th century. The last known glacier in Val Badia was the "Dlacia de Val Mesdì", located at the head of Val Mesdì. Marinelli (1910) indicates a residual extension of the glacier at that time of about 0.14 km^2.

Colfosco landslide

The village of Colfosco is among the oldest settlements in Alta Badia, and still comprises some quite well preserved rural buildings of the 16th century. The village lies on a vast accumulation of large dolomite blocks sealed by a silty matrix, whose origin is probably ascribable to a rock fall – rock avalanche event. This event is believed to have taken place about 5000 years BP (Corsini *et al*., 2000). This date results from a tree trunk, with a diameter exceeding 1 m, which was found buried at a depth of 7 m among large dolomite boulders, which make up the landslide accumulation zone. However, other rock-avalanche events occurred even earlier, as suggested by the age of about 7000 cal BP obtained from organic sediments (typical of dam lake deposition) found in borehole cores drilled upstream of the landslide. The detachment area can be identified in the sub-vertical rock walls of Sass di Campaccio peak, within the Puez-Gardenaccia Group, from where other quite large rock falls have detached even in recent times. In this zone, three fault systems cross each other, causing the rock to be divided into large prisms subject to spreading and toppling. The Colfosco landslide, and especially its apical portion, has also been built up and reshaped by vast debris flows running down from the Stella Alpina valley. Local inhabitants report a debris flow event in 1918 that reached the village and destroyed a few buildings located in proximity of the debris flow channel.

Stop 4.1:
Passo Gardena landslide
Alessandro Corsini

The Passo Gardena landslide is a complex-composite-style mass movement extending over an area of about 1.4 km^2. It starts off as a rock block slide affecting dolomite rock types (making up clean rocky and spiky outcrops) – that in turn have fallen into coarse boulders – it evolves into a rotational slide affecting weak clayey rocks of the S. Cassiano and La Valle formations, and then becomes an earth slide – earth flow of some million m^3 of clay-rich material (Figure 22).

A sketched cross section showing this sequence of movements is depicted in Figure 23. Boreholes and a geophysical survey have shown that the thickness of the earth slide – earth flow portion ranges from

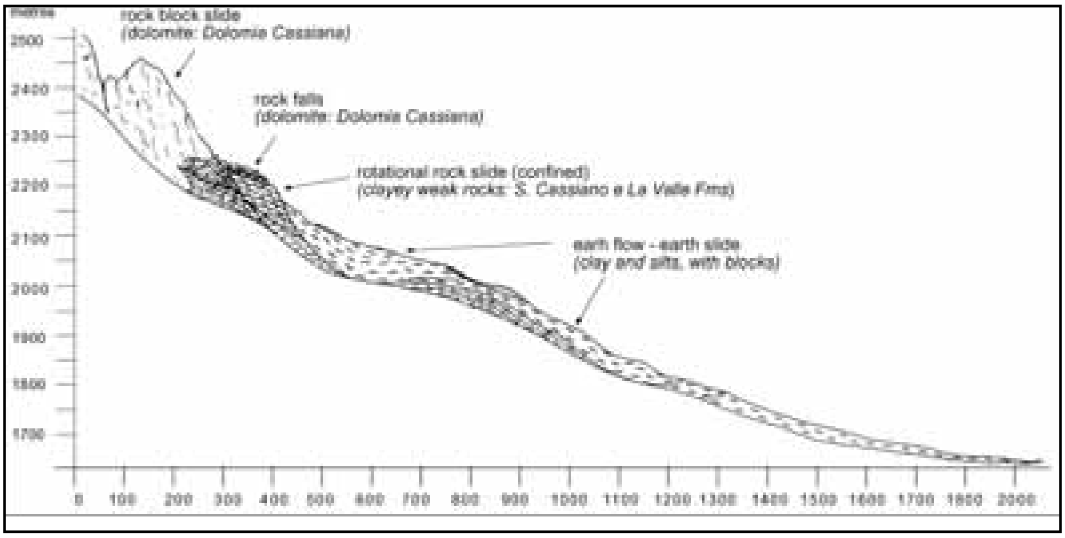

Figure 23 - Sketch section of the Passo Gardena landslide.

45 to 70 m. As regards present day movements, the uppermost rock block slide does not show evidence of ongoing activity, whereas large sectors of the flow-like processes are active, damaging a strategic roadway connecting the Alta Badia Valley with the adjacent valley.

Recent monitoring data from inclinometers have shown that movements in the earth flow portion take place on sliding surfaces located at about 20 m in depth. Downslope, movements are negligible, so at present the landslide is to be considered active-suspended or dormant. Seismic data have also shown that underneath the toe of the landslide, a 30 to 50 m thick level of glacial or fluvial-glacial gravel exists; this level probably acts as a natural drainage system, preventing the steeper portions of the earth flows from being active with higher displacement rates.

Synthesis of the relationships between landslide occurrence and climate changes in the Dolomites since the Lateglacial

Alessandro Corsini, Alessandro Pasuto and Mauro Soldati

Several investigations on landslide recurrence and climatic implications have been described in literature for various European regions among which, in particular, Austria, Great Britain, Italy, Norway, Poland, Spain and Switzerland. For a wider review aiming at illustrating the state of the art of these topics, with particular attention paid to Europe, see Borgatti *et al*. (2001) and Borgatti and Soldati (2002).

For the Dolomites, an effort has been made to assess which of the numerous radiocarbon dates collected in the study areas of Cortina d'Ampezzo (see Table 1) and Corvara in Badia (see Table 2) are related to events of a type or magnitude that might be indicators of Holocene climatic changes (Soldati *et al*., 2004). The degree of paleoclimatic significance of each of the dated landslides has been assessed on the basis of the following criteria: geomorphological context in which the landslide has taken place; type of landslide; material involved in the landslide; size/magnitude of the landslide; type of dated material; depth at which the dated material has been found (either from excavations or drillings); borehole stratigraphy when dated material is collected from drill holes.

The first phase of marked slope instability (Figure 24), is observed in the Preboreal and Boreal (about 11,500 to 8500 cal BP) and includes, on the one hand, large translational rock slides, which affected the Dolomite slopes following the withdrawal of the Lateglacial glaciers and the consequent decompression of slopes and, on the other hand, complex movements (rotational slides and flows) which affected the underlying pelitic formations and were probably favoured by high groundwater levels resulting from an increase of precipitation and/or permafrost meltdown. A second concentration of landslide events is found during the Sub-boreal (about 5800 to 2000 cal BP), when slope processes, mainly rotational slides and/or flows, took place in both the study areas. These slides may be

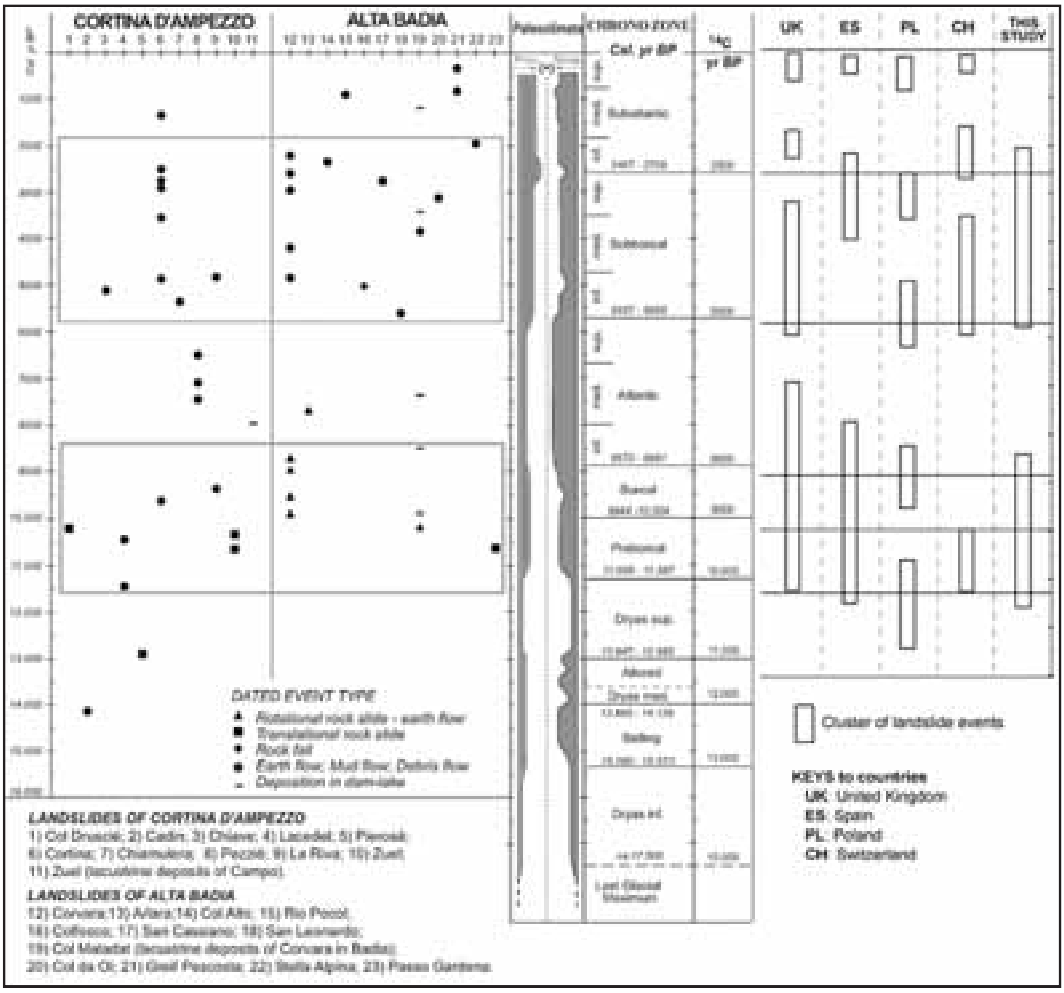

Figure 24 - Temporal distribution of "climatically significant" landslide events in the Dolomites and in Europe (after Soldati et al., 2004; modified).

considered as reactivations of older events linked to the phase of precipitation increase, which has been documented in several European regions during the mid-Holocene period. On the other hand, during the Little Ice Age, the scarce number of landslides dated in the study areas does not enable an increased frequency of landslides to be detected.

The recurrence in time of landslide activity since the Lateglacial was certainly also influenced by non-climatic factors. Among these, the influence of human activity, which is proved to have been crucial for other Alpine areas at least since the mid-Holocene (e.g. Provansal, 1995; Dapples *et al.*, 2002), seems to be almost irrelevant in the study areas where only seasonal or isolated settlements were present before the Middle Ages. Nevertheless, the present state of the investigations does not allow the assessment of the degree of influence of this and other possible non-climatic causes, which might have "disturbed" the climatic signal identifiable in the landslides studied.

In any case, one should consider that only large events have such a persistence in the landscape as to be identified after hundreds or thousands of years. Therefore, only a portion of the slope movements taking place in the past can contribute to the assessment of the relationships between recurrence of instability processes and climatic changes. Furthermore, since direct dating of mass movements by means of radiocarbon strictly depends on the presence of organic remains in landslide deposits, only the events which affected stable vegetation-covered areas – or landslide bodies which had previously been dormant long enough for new vegetation to grow – can be dated. It is therefore particularly difficult to obtain data ascribable to slope instability for the periods closest to glaciation, such as, for example, the Lateglacial, considering the lack or scantiness of tree cover in that epoch.

Notwithstanding this, in the light of the results so far obtained, many landslide events previously described might be considered as indicators of climatic changes. For this purpose, the comparison with bibliographic data on other European mountain areas seems to be significant (Figure 24). In fact, concentrations of landslides during periods similar to those of the areas studied were reported in various European regions, such as the United Kingdom, the Iberian Peninsula, the Alps and Eastern Europe.

In the light of these remarks, some analogies in the temporal distribution of landslides in various European regions and in the Italian Dolomites seem quite evident. Indeed, all the cases previously described can be fixed, more or less precisely, in the two periods of greatest activity identified. In Europe, the Preboreal period coincides with the oldest period of frequent postglacial slope movements, which has been identified by various Authors, whereas a second period coincides with an increase of precipitation reported in literature (Figure 24). However, the hypothesis of a direct influence of this climate change on landslides, partially confirmed by the European case studies above mentioned, still needs further verification. In fact, this trend of landslide recurrence, which inevitably shows differences among the European regions considered, is influenced not only by geographic, and therefore climatic, diversity but also by the amount and climatic significance of the data available. Moreover, non-climatic factors, in particular human influence, need to be analysed in detail, also in collaboration with other experts (i.e. archaeologists), and through comparison with other sites.

From Gardena Pass to Verona

Mauro Marchetti

After the Gardena Pass, we proceed along Gardena Valley towards the River Isarco Valley. Travelling across Gardena Valley, the complete stratigraphic sequence is exposed: from the Dolomia Principale to the Permian volcanites cropping out near Ortisei (see stratigraphic column, Figure 4).

Once in the River Isarco Valley, we take the A22 Brenner motorway and, just south of Bolzano, follow the main valley where the R. Adige flows. This wide Alpine valley stretches in a NNE-SSW direction from Bolzano to its outlet into the Po Plain, following an important fault and tectonic lineament system called the "Sistema Giudicariense". This system divides the Southern Alps of the Lombardy basin from the Venetian basin. In its terminal stretch, this valley is parallel to Lake Garda, from which it is divided by the ridge of Mt. Baldo.

The Adige Valley is characterised by a flat bottom due to fluvioglacial and fluvial sedimentation principally occurring from the LGM to the Lower Holocene. In the northern part, a Permian large porphyric platform marginally outcrops. Whereas, the geological formations cropping out in the lower Adige Valley and in the basins of its tributaries are Mesozoic and Cainozoic in age and the most representative terrains are calcareous and dolomitic formations (Dolomia Principale Fm., Noriglio Grey Limestone Fm., San Vigilio Oolitic Limestone Fm.).

The valley originated during an intense erosional phase that can be dated to the Late Miocene (Messinian), when the Mediterranean basin was isolated from the Atlantic Ocean (Bini *et al.*, 1977; Finckh, 1978). The Messinian fluvial erosion of the River Adige took place in a pre-existing valley, which was tectonically controlled by structures belonging to the Giudicarie System. During the Pleistocene glacial phases the valley was affected by the action of glaciers, with consequent glacial remodelling. During the LGM the glacier thickness reached 1700 m near Trento and 1500 m a little to the south. Other extremely active processes acting on the slopes and valley floor are linked to karst dissolution, fluvial dynamics of the River Adige and anthropogenetic activity.

Near the road junction of Rovereto Sud with the A22 motorway, the vast, impressive complex of "Lavini di Marco" is visible. This is an extensive landslide accumulation with various minor gravitational deposits. The landslide has long been famous, mainly because of a reference made to it by Dante Alighieri in his "Divine Comedy". The main landslide occupies a surface of about 6 km^2, with a maximum length of about 5 km, a maximum width of 1800 m and a difference in altitude of 1030 m (Orombelli and Sauro, 1988). The zone of detachment coincides with bedding planes and affects a modest thickness of the calcareous sequence (up to some tens of metres). Within the zone of accumulation it is possible to observe two different lobes. The major lobe shows a fan-like expansion on the valley floor for a distance of over 1 km; it is mostly covered by spontaneous vegetation and is characterised by several depressions on the flat, cultivated portion of the plain. The lesser lobe lies on the bulging part of the slope and is made up of large rock blocks and is mostly lacking in vegetation cover. These two lobes could either have been generated at different times or be the outcome of a single event during which a resistant rocky ridge located in the failure surface (Costa Stenda) could have separated the displaced material into two. On the basis of historical evidence, the landslide was ascribed to the early Middle Ages by several Authors (Sacco, 1940; Fuganti, 1969; Eisbacher and Clague, 1984). Nevertheless, this attribution is not sufficiently documented.

Southwards, near the road junction of Affi with the A22 motorway, the trip will proceed along the Rivoli Veronese glacial amphitheatre. It consists of a moraine ridge smaller than the Garda system but is a well preserved system of frontal moraine ridges deposited by the glaciers descending along the Adige Valley, where thick loess covers, red soils, and fluvioglacial terraces have also been preserved. Considering the state of conservation of the moraines and their stratigraphic nature, Cremaschi (1990) dated the whole amphitheatre to the Upper Pleistocene. Outside the Upper Pleistocene main moraine ridge, Middle Pleistocene eroded moraines and highly eroded and terraced glacial deposits, pushed against the north reliefs of Caprino Veronese, are also found. They are all ascribed to the Early Pleistocene.

The excursion continues downstream, towards the Po Plain, and will be concluded in Verona.

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Back Cover:
field trip itinerary

FIELD TRIP MAP

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