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A GEOMORPHOLOGICAL RECONNAISSANCE OF STRUCTURALLY-CONTROLLED LANDSLIDES IN THE DOLOMITES

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INTRODUCTION

The mountainous region of northeast Italy is known as 'The Dolomites', the name arising from the dominance of dolomitic limestones among the geological formations that characterise the region. The mountains are the product of a complex tectonic history which, in the region that includes Longarone and Cortina d'Ampezzo, is thought to have included uplift of perhaps 3-5 km (Doglioni, 1987). Present uplift rates are high, possibly as much as 6-7 mm y^{-1} (Dr F Podda, University of Trieste, pers. comm. 2012). The present mountain slopes, that provide the potential for mass movements, were largely shaped by glacial and subsequent fluvial erosion processes. The persistence of the (north-facing) Marmolada glacier down to ~2850 m altitude and the terminal moraines from the Piave Glacier south of the Venetian Pre-Alps near Vittorio Veneto are indicative of significant Quaternary glaciation extending into this region. The summit of the Marmolada mountain, at 3342 m, is the highest point in the region. Maximum summit elevations decrease eastwards from here, across the region of interest to this paper, being around 2500 m adjacent to the Piave Valley and <2000 m around the Tagliamento Valley east of Ampezzo. Local (valley-scale) relief is typically in the range 1200-1600 m. Mass movements are unsurprisingly common in such a tectonically active mountainous region, their types and characteristics being strongly related to the geology, particularly with respect to structural controls.

Within about 100 km (by road) from Longarone and Vaiont, many landslides of many different types and characteristics can be readily identified and accessed. In general, the different types can be clearly associated with their respective topographic and geological contexts. Those associated with 'soft rocks' are found at the southern edge of the Dolomites region and within the northern half of the region, where the respective lithologies generally underlie more gentle slopes below prominent limestone outcrops. Small shallow slides and extensive landslides, including the Tambre and Tessina earthflows, occur in the Eocene flysch in the Alpago Basin at the eastern end of the Belluno Trough. The mid-slopes of the Alpago Basin are also characterised by the forms of deep-seated rotational and/or compound slides in the flysch. Further northeast, the soft mudrocks such as those of the San Cassiano Formation have given rise to many mass movements including the Corvara earthflow and the shallow landslides in and around Cortina d'Ampezzo, as well as shallow rotational slides and deep-seated rotational or compound slides involving failure within the mudrocks resulting in major displacements of the overlying limestones. The upper valley slopes immediately west of the Gardena Pass are characterised by very large tension cracks and subsided limestone blocks that can be categorised as 'cambering' failures, again due to failure of the weaker underlying San Cassiano Formation.

Landslides founded entirely in the limestones/dolostones have occurred as far south as the ancient Piave

Valley at Fadalto and Nove, where very large rockfalls or rockslides formed landslide dams that have been exploited for hydro-power electricity generation (Coppola & Bromhead, 2008). The Fadalto landslide involved failure across or through the bedding of the limestone, and the Nove landslide can be assumed to be similar in this respect although it is so old that the original source area cannot be reliably identified. Further north, a moderately large and very distinct deep-seated rotational landslide on the western side of the Pelmo massif (summit elevation 3168 m) has also developed across the bedding, which further highlights the role of joints and fractures through the rock mass associated with the regional tectonics in promoting large-scale instability. The other landslides that are not causally associated with the bedding of the limestones are the debris flows. Although small debris flows are common throughout the Dolomites, two distinctive concentrations occur along the eastern slope of the Boite Valley southeast of Cortina and below the 2000-2200 m high ridge immediately north of the same town. Most of these are appear to originate from deeply dissected fracture or bedding zones where structural weaknesses have allowed water penetration and thus promoted localised preferential freeze-thaw weathering.

The primary focus of this paper is a group of landform-defining rockslides, some of which may be better described as 'block slides', that appear to have been primarily controlled by the bedding within, or perhaps adjacent to, the major dolomitic lithological units. Most of the examples involve units of the Dolomia Principale of Upper Triassic age. We make a distinction at this point between 'bedding-plane failures' (also known as 'dip-slope failures') and 'bedding-controlled failures'. In a dip-slope failure, the landslide occurs due to failure along one or more parallel planar or near-planar bedding planes and the failed bedding surface dominates the post-failure morphology of the slope. This type of landslide is more common in hard rocks where the structural weakness provided by the bedding planes contrasts strongly with the intact rock mass strength. Bedding-controlled landslides are commonly found in argillaceous rocks, particularly stiff plastic overconsolidated clays (e.g. Barton, 1984; Cooper *et alii*, 1998). The basal part of the deep-seated rotational or compound slip surface in most failures of these materials is planar, being controlled by a 'weak layer' arising from a (brief) interruption to an otherwise continuous and generally uniform sedimentary accumulation, i.e. a bedding plane (e.g. Bromhead *et alii*, 2002; Dixon & Bromhead, 2002). The low strength of the 'weak layer' gives rise to the instability and then controls the depth of the failure, but does not dominate the morphology or internal geometry of the landslide.

CONTROLS ON DIP-SLOPE FAILURES

The occurrence and consequences (geomorphological and possibly socio-economic) of dip-slope failures can be strongly related to the geometric relationship(s) between the dip and strike of the bedding planes and the pre-failure topographic surface. The configuration of the bedding planes and any other structural features results from the tectonic history of a location, having developed over geological timescales. The pre-failure topographic surface, on the other hand, results from the combined effects of various geomorphological processes acting on the slope and/or adjoining landform units (such as the river channel at the foot of the slope) over typically much shorter timescales, of the order of perhaps 10^3-10^5 years rather than e.g. 10^6-10^8 years.

Cruden (2000) classified these geometric relationships based on observations of rockslides in the Canadian Rockies (Tab. 1). As is usually the case, reality often exhibits more complex arrangements than any simple classification scheme can represent. Hence, if

anaclinal slope	bedding dips in the direction opposite to the slope	
cataclinal slope	bedding dips in the same direction as the slope	
	overdip slope	slope is steeper than the dip
	dip slope	slope is parallel with the dip
	underdip slope	dip is steeper than the slope

Tab. 1 - Classification of slope–bedding relationships (after Cruden, 2000)

Landslide	Length (m)	Height (m)	Mean slope (°)	Upper slope (°)
Alleghe	1500	880	30	31
Borta	2070	1040	27	32
Pineda	2220	1130	27	25
Vaiont (East)	1450	800	29	35
Cinque Torri	3000	~700	18	~30 at the head

Tab. 2 - Characteristics of the study landslides the river at the foot of a dip slope incises rapidly to create a narrow gorge, then the lower part of the slope becomes an overdip slope and the likelihood of instability is significantly increased. Furthermore, although this framework is designed for the intepretation of slopes underlain by planar bedding, it does provide a basis for assessing the stability status of slopes formed on anticlinal or synclinal structures. For example, the Vaiont landslide occurred on a dominantly dip (or underdip) slope but instability was promoted by the overdip slope at the toe, where the extremely steep sides of the river gorge allowed the critical bedding plane(s) to daylight.

DIP-SLOPE LANDSLIDES IN THE DOLOMITES

We have examined the main features of four distinct examples of large dip-slope landslides and what we think may be a very large dip-slope or bedding-controlled landslide, in the southeastern Dolomites region. These landslides and the coordinates of their head scars are as follows:

Alleghe	046°24'19"N, 011°59'44"E
Borta	046°22'04"N, 012°48'17"E
Pineda	046°16'54"N, 012°21'04"E
Vaiont (East)	046°15'15"N, 012°21'04"E
Cinque Torri	046°30'15"N, 012°02'00"E

Summary characteristics of these landslides are shown in Table 2. The lengths and heights (vertical differences in elevation between the head and the toe) are derived from various topographic map sources, with the consequence that the derived slope gradients may be ±1–2° of the stated value. This level of approximation does not affect the interpretations or the key message of this paper.

Fig. 1 - The Alleghe landslide in 2011, seen from above Alleghe village

ALLEGHE LANDSLIDE

The major rockslide at Alleghe occurred in January 1771 and was followed later that year by another, smaller one from the same slope (Eisbacher & Clague, 1984; Di Biasio *et alii*, 2000; Coppola & Bromhead, 2008). Figure 1 shows the planar form of the (combined) landslide scar, the failure surface being defined by a small number of closely spaced bedding planes in the limestone (Fig. 3). The displaced mass comprised a thick wedge of rock that slipped from a valley side slope that was previously convex in profile and plan. The northern edge of the failure surface marks the line where the critical bedding plane(s) daylighted, whereas the southern margin comprises a high lateral scar defined by high angle joints (Fig. 2).

The original failure formed a landslide dam that was subsequently engineered to become effectively permanent. Before the failure, the river was 30-40 m below the present lake level. If the landslide scar is projected downslope at the same gradient, it appears to coincide with the approximate position of the river at that elevation. Since the post-failure scar is planar,

Fig. 2 - The upper part of the scar showing the bedding control and the lateral joint-controlled margin

Fig. 3 - Part of the failure surface (seen from below the landslide) showing it to be controlled by very few separate bedding planes

it follows that the original valley slope adjacent to the river was probably steeper than the bedding, i.e. that instability was promoted by the lower slope being an overdip slope. However, this geomorphological context of a somewhat overdeepened river valley also reduced the impacts of the landslide. The dam was formed because the narrow lower valley prevented runout of the debris, and the permeable nature of the debris enabled it to persist without catastrophic failure and flood (Coppola & Bromhead, 2008). The village was, however, hit by a devastating wave from the second slide (Eisbacher & Clague, 1984).

BORTA LANDSLIDE

The rockslide that destroyed the village of Borta, on the Tagliamento River near Caprizi, occurred in August 1692. It is similar to the Alleghe slide in many respects, but although the affected slope was significantly longer and higher than at Alleghe (Tab. 2), the failed mass was much smaller with an estimated volume of around 30,000 m^3 (Coppola & Bromhead, 2008). Inspection of aerial imagery (e.g. GoogleEarth) suggests that the failure involved the upper part of a main valley slope that had been dissected by first-order tributary streams. The extent of the steeper upper slope scar can be clearly seen in Fig. 4, where the bedding planes in the dolostone that define the failure surface are clearly visible (Fig. 5). Figures 4 and 5 both show the lateral scar on the eastern side of the failure, controlled by joints.

Several point of similarity with the Alleghe landslide can be identified. These include the bedding plane shear surface with the same gradient, the joint-controlled lateral scar on one side, the triangular form of the main scar (width increasing downslope) and the formation of a landslide dam. Finally, the geomorphological context suggests a similar geometric predisposition to instability, although the evidence is less clear. It seems likely that the incising tributary streams on the main valley slope created overdip slopes along the upslope sides of the streams. As such, they also defined the downslope extent of the failure and thus limited the scale and persistence (only a few weeks) of the dam.

PINEDA LANDSLIDE

The age of the Pineda rockslide is uncertain, but its scale is readily apparent (Tab. 2, Fig. 6). The ridge between the present Vaiont Lake and the Mesazzo Stream has been identified as a deposit of the Pineda slide (e.g. Paronuzzi & Bolla, 2012), which strongly suggests that a large landslide dam existed for some period of time following the landslide. Unlike the other landslides, the upper scar is less steep than the overall slope but nevertheless it appears that failure of bedding planes in the Soverzene Formation defined the event (Fig. 6b), resulting in a triangular sub-planar scar in the landscape (Fig. 6) similar to the landslides described above.

The Pineda slide is characterised by a lower mean gradient of the bedding plane failure surface compared with the other landslides, although the form of this upper slope unit is slightly convex (Fig. 6b) and a dip of 30° is shown for the lower part of this unit by Paronuzzi & Bolla (2012). Clearly the geological control over this slope failure is more complex than at the landslides discussed previously, but it seems reasonable to suggest that, as a dip slope, the upper slope unit was susceptible to the development of instability, and that this instability was ultimately brought about

Fig. 4 - The Borta landslide in 2011, seen from upstream

Fig. 5 - Part of the failure surface showing the dolostone bedding planes

by glacial oversteepening of the lower part of the valley slope. Figure 6a appears to show the landslide scar extending some distance below the upper slope unit. This landslide may therefore have begun with failure of the underlying 'soft' rocks (the marly Scaglia Rossa (CANCELLI *et alii*, 1991) and the Marne di Erto which is transitional from the Scaglia Rossa to the Flysch), and retrogressed upslope.

The lithology of the upper bedding failure surface of the Pineda landslide and its general structure, i.e. the folded front of a laterally extensive over-thrust, are the same as for the cliffs above Casso village that overlook the Vaiont dam. There have been several potentially damaging rockfalls/rockslides from these cliffs in recent years, requiring significant expenditure on mitigation and stabilisation measures. It seems ironic that a landslide hazard assessment of the valley before 1963 would probably highlight the risks associated with these outcrops of the Soverzene Formation, based on the evidence of Pineda and the Casso deposits, but entirely fail to recognise any realistic hazard and risk from the northern slopes of Mt Toc or indeed from the reservoir, some 200 m lower than the town.

VAIONT LANDSLIDE

This research led us to re-examine some features of the Vaiont landslide in the context of other major rockslides in the region. For this reason, attention is focused here on the East side of the Vaiont slide, which has often been largely overlooked in the past because of the closer proximity of the West side to the dam.

If considered as a separate landslide, Vaiont (East) is similar in many respects to the landslides described above. The scar is much wider than those of the other landslides but (i) its pre-failure profile characteristics are similar (Table 2), (ii) it formed a landslide dam (Fig. 7), (iii) it has a joint-controlled lateral scar at its (eastern) lateral margin, and (iv) the visible failure surface has a roughly triangular (actually more of an arch) plan form. The latter feature is easily missed on casual inspection of aerial imagery because of a large forested unit at the extreme southeast corner of the landslide that was displaced but remained on the failure surface. This displaced block clearly resulted from retrogressive development of the landslide. Such retrogression often accounts for the upward-narrowing plan forms of landslide scars, as observed in all of the cases in this paper.

The view of the landslide in Fig. 7 is somewhat misleading in that it does not show how the Vaiont Gorge turned slightly northwards, away from Mt. Toc, where the landslide dam now begins. In any case, cross-sections indicate that the East side failure surface was probably mostly planar. Furthermore, the sides of the gorge were much steeper than the typically 30-40° dip that defines the failure surface and thus the toeslope of the entire landslide was an overdip slope.

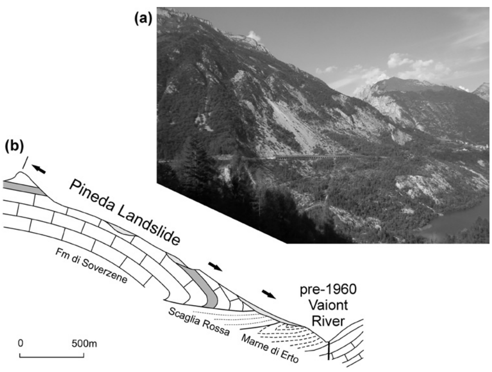

Fig. 6 - The Pineda landslide: (a) seen from the Vaiont landslide in 2012; (b) geology according to BESIO & SEMENZA (RIVA et alii, 1990)

CINQUE TORRI LANDSLIDE

The small rocky outcrop near Cortina d'Ampezzo known as the Cinque Torri ('Five Towers') (Fig. 8) has been demonstrated to be the product of mass movements involving failures of the weak argillaceous rocks under the load exerted by the Dolomia Principale and suggested to be a small part of a much larger-scale landslide system involving bedding-controlled failure(s) (VIERO *et alii*, 2010). The argillaceous rocks (the Travenanzes and Heiligkreuz Formations) are 150-180 m thick and underlain by the Dolomia Cassiana, with the entire sequence having a moderate dip of 30° towards N10°E. VIERO *et alii*

Fig. 7 - The Vaiont landslide in 2011, seen from the east

Fig. 8 - The Cinque Torri and Averau, with Pelmo (3168 m) in the distance.The white line indicates the hypothesised failure surface (broken line = concealed at depth). Snow highlights the topographic details

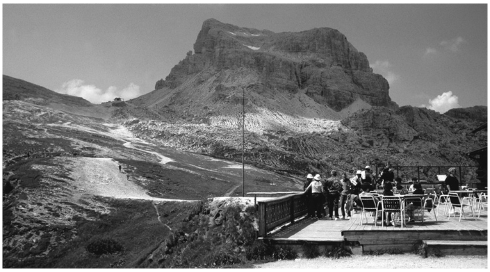

Fig. 9 - View towards Averau from the Cinque Torri. The planar slope on the left is the bedding-plane failure surface highlighted in Fig. 8

(2010) show everything to the right of our Fig. 8 as a 'reference stable area'.

In our view, the geomorphology of the site indicates a landslide system possibly extending over several square kilometres. The influence of the 30° dip of the bedding can be seen in many parts of the landscape in Fig. 8. The left side of Averau appears to be the head of a large translational dip-slope landslide (Fig. 9) that probably extends northeastwards for up to 4 km down to the Falzarego River with the upper surface of the Dolomia Cassiana as the failure surface. This is consistent with VIERO *et alii*'s (2010) suggestion of such a landslide although we think it extends another kilometre further upslope. However, it is unclear what structural forms or failure surface geometry in the lower half of the slope give rise the low overall gradient of the failed slope (Tab. 2).

DISCUSSION

We have examined five major landslides in the southeast Dolomites region, four of which are dip-slope failures and the fifth may also be of the same type although details of the lower part of its slope are not known.The dip angles of the bedding plane failure surfaces ('Upper slope' in Tab. 2) are very similar for all of the landslides and the dip-slope failures are characterised by almost identical mean pre-failure slope gradients.

Detailed investigations of the Vaiont landslide over many years have confirmed the presence of thin clay layers between limestone beds within the Fonzaso Formation, which is the stratigraphic unit that forms the visible failure surface. The Pineda landslide involved a lower dip angle than the other cases, which is lower than the normal range of friction angles between unweathered limestone surfaces (SELBY, 1993), and is thus consistent with a failure strongly controlled by the strength of clay layers. However, in deformed strata like this (Fig. 6), flexural slip between beds is likely to have reduced the available shear strength - in this case, of the clay - to its residual value, with the cohesion required for stability probably having been provided by rock-bridges (STURZENEGGER & STEAD, 2012).

At the Borta landslide, only the upper part of the slope failed, but both here and at Alleghe the bedding appears to be almost entirely planar. The dip angles lie just inside the range of friction angles for unweathered limestone surfaces. This can be interpreted as indicating that failure in these cases was controlled either by clay layers and rock bridges or simply by the bedding contacts without any clay but possibly weathered due to rainwater penetration directly into the bedding at the top of the slopes.

The Cinque Torri landslide is perhaps the easiest to explain. In this case, the controlling shear strength would have been that of the basal layer of the dominantly argillaceous Heiligkreuz Formation. The observed ~30° dip angle in the upper part of the slope is steep enough to account for the occurrence of the landslide, but its apparently limited development may be associated with the much lower overall slope gradient.

The four dip-slope landslides are typical of others described from around the world. The largest known terrestrial landslide on Earth, Saidmerrah in the Zagros Mountains of southwest Iran, is a dip-slope landslide involving failure of planar bedding surfaces in Cretaceous limestone, although the dip angle is rather steeper than in many other cases. The role of rock bridges as contributors to the available shear strength in folded limestone sequences was demonstrated recently using the Palliser Rockslide, one of many such dip-slope landslides in the limestones of the Canadian Rockies (STURZENEGGER & STEAD, 2012). Interestingly, the Frank Rockslide in Alberta was for many years thought to have been primarily bedding-controlled, although it has now been shown that the structural control was dominated by the joint sets through the massive limestone (e.g. PEDRAZZINI *et alii*, 2012).

Two further geomorphological issues became apparent as a result of our study of the landslides in the Dolomites. Firstly, the formation of a landslide dam depends not so much on the absolute volume of the failed mass and the duration of the movement (i.e. rate of accumulation of the deposit), but more importantly on how these two characteristics relate to the scale of the valley and the discharge in the river at the time of failure. Three of the landslides are of moderate scale but all involved rapid accumulation of their debris within narrow and steep-sided valleys. At Vaiont, the glacial-scale valley trough was effectively dammed as the fluvial gorge was entirely infilled.

Secondly, although formation of the landslide dams will have supplied some sediment to the downstream river channels, in the Dolomites it appears that the more significant sediment impact has been for the dams to interrupt the downstream transfer of very high (glacio-) fluvial sediment loads. Very rapid accumulation of sediments upstream of landslide dams is best demonstrated by reference to infilling of the lake formed by the Alleghe landslide, which had extended as far as Caprile (3 km upstream of the present residual lake) within weeks of the event. Such rapid sedimentation is also readily apparent in the residual lake upstream of the Vaiont landslide and in the river above the Borta slide. Finally, although not relating to a dip-slope failure, mention must be made of the infilling of the Piave Valley following the landslide and dam formation at Fadalto. The net accumulation of sediment arising from this event can be traced as far upstream as Castelmezzano and provided the major source of debris that the flood wave from the Vaiont landslide was able to entrain and which consequently further increased the intensity of the destruction at Longarone (BROMHEAD *et alii*, 1996).

CONCLUSIONS

Two key messages arise from this brief survey of the large structurally controlled landslides of the south-eastern Dolomites. Firstly, failure along the bedding is as likely in this region as it is in any other mountainous region dominated by strong sedimentary rocks, often carbonates, and subjected to significant tectonic deformation followed by vigorous glacial and/or fluvial erosion. Secondly, we have shown that the eastern half of the Vaiont landslide appears to be entirely typical of the other dip-slope landslides in the study area. We make no assessment of the influence (or not) of the western half on the occurrence of the failure of the eastern half, but we do emphasise the fact that there should be nothing inherently surprising about the east side failure. The detailed mechanisms of the overall landslide may still be very uncertain, but the underlying general context and characteristics are clear. The critical implication is that any assessment of the risks of major landslides that may arise from any future infrastructure developments in mountain regions, in the Dolomites or indeed elsewhere in the world, should include the Vaiont example as a typical case and not exclude it as being 'unexplained'.

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