Evolutionary behaviour of the Tessina landslide

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Summary

The Tessina Landslide, a complex movement which has been active for over the past 50 years in the Alpago area of Belluno (Italy), was triggered in October 1960. As it is typical of Tertiary Flysch formations, its evolution has been characterized by periodical medium-large roto-translational slides in the source area evolving into earthflows, often jeopardizing the safety of the lower valley. Both the landslide and earthflows have been extensively monitored and studied over the past few decades. The main aim is a better understanding of their evolution mechanisms and the selection of appropriate mitigation strategies to reduce risk to the valley below.

This paper presents data recorded by two real-time monitoring systems that have recently been installed in the area. The first system measures groundwater pressure and the displacements in a lateral section of the landslide that plays a crucial role in the stability of the entire area. The second, a new photogrammetric-based system which daily photographs the upper basin affected by the roto-translational movements, will permit us to evaluate the mechanisms triggering the earthflows.

The data collected until now at the site have made it possible to better understand the stability of the various sections as well as the processes involved in earth flow formation. This information will make it possible to evaluate the effectiveness of future mitigation measures. Finally, a new black-box model that will predict the mobility of the lateral section of the landslide has been developed, and its performance has been compared to that of a viscous model commonly used to analyse slow-moving landslides.

Keyword: earth-flow, creeping landslide, Flysch formation, monitoring, digital image.

1. Introduction

Complex landslides in Flysch-type formations are commonly found in Italy, both in the Alpine and Apennine regions, as well as in other countries in the Mediterranean basin (e.g. Borgatti et al., 2006; Van Asch et al., 2007; Arbanas et al., 2013; Corominas, 2015; Di Maio et al., 2013; Jamei, 2015). Their management is often difficult due to the high cost linked to stabilization interventions, which, in many cases, fail to ensure long-lasting solutions, and the distress of the people living in the proximity of the landslide [Picarelli, 2011].

Decisions by the local authorities should in any case be based on extensive knowledge and understanding of the landslide’s evolution and accurate, long-term monitoring of the instability processes taking place. Significant improvements in monitoring devices and interpretation methods have, over the last few decades, contributed to providing reliable data. In this case, new technologies such as radar interferometry (e.g. Pieraccini et al., 2003), TDR sensing for humidity detection (e.g. Bittelli et al., 2012), high frequency data recording and remote acquisition and transmission (e.g. Qiao et al., 2013) deserve mention.

The aim of this paper is to analyse the behaviour of a complex landslide that has been active in an Alpine region in North-Eastern Italy for more than 50 years and, in particular, to present the preliminary results produced by two monitoring systems that have been recently installed there to permit experts to study the landslide long-term evolution. In consideration of the fact that inside the area coexist sections characterized by various types of kinematics linked to different movement mechanisms, two systems based on different approaches have been developed and applied.

The first system aims to monitor a lateral sector of the landslide, referred to as the “Pian de Cice”, whose slow-moving pattern in the past increases during rainy periods, generally in the spring and autumn, but is not apparently linked to local groundwater variations, possibly as a result of non-continuous measurements of groundwater oscillations in piezometers.

A continuous monitoring system, consisting of a fixed inclinometer probe coupled with a pore pressure transducer, was installed inside a borehole to investigate any correlation that might exist between the rate of displacement and oscillations in the groundwater level. The data collected over approximately an eighteen-month period led to more insights into

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the hydro-geological seepage response in that sector, including the effects of an exceptional rainfall leading to a partial collapse of the slope.

The second monitoring system was stationed to substitute the old, less efficient topographic surveying station that had been installed in 1994 and required costly maintenance. It was our intention to develop a low-cost system guaranteeing a continuous recording of displacements in the large depletion basin. Two photo-cameras, which take digital images every day at noon, were thus installed at two positions 12 m away from one other. An automated procedure for photogrammetric digital image processing and the topographic reconstruction of the basin surface at different time points has outlined the evolution of the depletion basin over time. In addition, an algorithm to estimate the superficial displacements of various sections of the landslide by superimposing two images and measuring the difference between the ground configurations that they uncover is presently under development. Although the procedure is not yet ready to be used, a time-sequence of digital images could provide other important details concerning the landslide’s evolution. It would be possible, for example, to identify the fastest moving sections of the landslide, information that would be useful to optimize photogrammetric interpretation as well as to estimate snow permanence time or the time span between the beginning of a rainfall and the development of springs, etc.

Following a brief description of the landslide’s history, the paper presents the preliminary results that have been produced by these new monitoring systems. These data have confirmed the findings of past investigations focusing on the kinematics of the depletion basin and have clearly demonstrated the influence of groundwater oscillations on the displacements of Pian De Cice. Finally, a new black-box model to predict the movements of the sector was developed and calibrated using the ground pore water pressure data recorded; the movement pattern predicted by the new model was compared with the prediction of a viscous dynamic model usually used to analyse the behaviour of slow-moving landslides.

2. The history of the Tessina landslide

The Tessina landslide is located on the southern slope of Mont Teverone in the Alpago valley situated in the North-Eastern Italian Alps near Belluno (Fig. 1). The landslide began in October 1960 following the collapse of a small area on the western border of the present depletion basin, probably due to the construction of a new forest track.

At present, the landslide consists of some massive rota-translational slides occurring in its upper part and periodically evolving into earthflows developing through the steep channel in its lower zone [Pasuto et al., 1992; Angeli et al., 1994, 2000; Mantovani et al., 2000a, b].

The landslide develops within 1220 m and 625 m a.s.l with a total longitudinal extension of nearly 3 km and a maximum width of about 500 m in its upper portion. It involves the Belluno Flysch Formation (Middle Eocene) and Flysch Formation (Lower Eocene) consisting of alternating marly argillaceous and calcarenite layers with a total thickness of approximately 1000-1200 m. This formation constitutes the impermeable bedrock of the entire sliding area and crops out at the foot of Mont Teverone which is made up of Fadalto limestone (Upper Cretaceous) arranged in a sub-vertical orientation (Fig. 2). Large
and thick portions of calcareous moraines (deposits of Piave Wurmian glacier) and detrital fans cover the Flysch bedrock, which crops where the sliding mass moved away the covering materials (Fig. 3).

During the 1960s there were several cases of reactivation involving a total volume of about $5 \times 10^6$ m$^3$ of material and causing the Tessina valley to be progressively filled with displaced material. These movements seriously endangered the village of Funes, situated on a steep ridge, originally very high above the riverbed, but now at nearly the same level as the earth flow surface [Dall’Olio et al., 1987].

In April 1992, a rotational slide with a 20 - 30 m deep failure surface caused the detachment of an approximate volume of $10^6$ m$^3$ from an area on the left side of the Tessina stream. The movements initially caused the formation of a 15 m high scarp, a 100 m displacement of an unstable mass, the destruction of drainage systems set up a few years earlier, and they continued with a steady intensity until June 1992, causing the mobilization of another $3 \times 10^5$ m$^2$ (a total volume of about $2 \times 10^6$ m$^3$). The highly fractured and dismembered material from the collapse area channelled along the riverbed where, due to continuous remoulding processes and rising water content, it became increasingly more fluid, thus generating small earth flows converging into the main one. The rapidly increasing level of risk to the valley below induced the Civil Protection Agency to order the temporary evacuation of inhabitants of Funes and Lamosano.

After that event, some temporary protection measures for the inhabitants were adopted and geological surveys and geotechnical investigations were considered indispensable to predict the landslide’s behaviour [Angeli et al., 1994]. Since the landslide was of a considerably large magnitude and its stabilization certainly expensive, the Italian Ministry for Civil Protection preferred to keep its evolution under control with a permanently monitoring of its activity and the installation of an early-warning system.

The only remedial measure taken after 1992 was a 1200 m long tunnel, which was excavated inside Mont Teverone, at the rear of the main upper scarp, in order to drain the area and to reduce the instability of that side of the slope. Even if the tunnel drains large quantities of water – and, in fact, now the municipality utilizes the tunnel discharge for electrical power production - the land continues to slide [Cola et al., 2009]. Over the years, many other collapses at a frequency of about 3 years have led to the formation of earth flows, similar in evolution to the 1992 event but involving smaller volumes of earth. The risk level
has never again reached that verified in 1992, and it
has never again been necessary to evacuate the two
villages.

Since 1992, the geomorphological risk has been
analyzed, geological and geophysical data has been
collected, and the landslide has been efficiently
monitored [Silvano and Pasuto, 1991; Pasuto and
Silvano, 1995]. In addition, the data on the Tessina
landslide has been utilized as a benchmark case
for evaluating new monitoring systems and validating
advanced numerical models. A new photogram-
metry technique [Avolio et al., 2000; Mantovani et
al., 2000; Hervas et al., 2003] and a ground-based
SAR interferometry [Tarchi et al., 2002] were used
to control the geomorphological evolution of the
landslide for two brief periods. In addition, a cellular
automata simulation model [Avolio et al., 2000]
and a run-out model based on the SPH integration
method [Cola et al., 2008] were later used to pre-
dict the propagation of earth flows that could be
caused by collapses of the upper basin. Finally, Marco-
ato et al. [2009] analysed the long-term displace-
ments of the Pian de Cice sector using a finite dif-
ference method and an elasto-visco-plastic constitut-
ive model.

The source area is presently approximately 0.52
km² wide (Figs. 1 and 4) and the total displaced vol-
ume is equal to about 7·10⁶ m³. Geomorphological
evidence indicates that there is a collateral landslide
affecting the eastern slope of the depletion zone, i.e.
the Pian de Cice. The stability of that sector of the
slide is considered crucial for the future evolution of
the entire area since its collapse could lead to a rapid
enlargement of the source area.

3. The monitoring system installed in 1992

The first monitoring system, which was installed
in 1992 [Angeli et al., 1994] and later improved, con-
sisted of an array of sensors and measuring instru-
ments, including piezometers and inclinometers set
up in medium-deep boreholes, multiple-base wire
extensometers, and a topographic system with an au-
tomatic landmark detector for measuring the sur-
face movements of 30 benchmarks (Figs. 1 and 4).
Two alarm units, installed above the villages of Funes
and Lamosano, and three video cameras recording
slide movements in the areas considered most crit-
ical, i.e. the upper accumulation basin and the area
as upstream of Funes and Lamosano, completed the
system.

Data from the peripheral stations were transmit-
ted to the control centre, which is situated in Lamos-
ano’s city hall. This early-warning system was part of
the Civil Protection plan, which included the order
to evacuate the population from residential areas in
the event of impending danger.

The benchmarks were prevalently located in the
depletion zone of the landslide because its control
was crucially important to the early warning system.
In fact, the area was constantly active and character-
ized by three main scarps (Fig. 4): in the past, large
mass movements in this area always induced global
landslide reactivation. Figure 5 shows the pattern of
displacements over time observed during the 1997-
1999 period for some benchmarks located in the up-
per basin and close to the main scarps. The pattern
is typical of the evolution observed in the upper ba-
sin: the displacement rate varies from season to sea-
on, having higher values in the spring and autumn.
The mean annual rate remains approximately con-
stant or shows a slight increase during some years
(see the curves referring to benchmarks n. 9 and n.
304). Moreover, abrupt increases due to rapid force-
ful accelerations, as displayed by benchmark n. 106,
have been observed in those sectors involved by the collapse.

The topographic survey revealed that the most active sector, characterized by displacements exceeding 1 m/year, was located within the main scarp where the material slides intermittently towards the upper accumulation zone (Fig. 4). The highest velocities (up to several decimetres per hour) were recorded in correspondence to the superficial earth flows within this sector. The measurement campaign carried out using the ground-based SAR apparatus during an extreme event that occurred between the 25th of September and the 13th of October 2000 [TARCHI et al., 2002] confirmed these data. The interferogram sequences showed a complex displacement field with high movement concentrations in two sectors with elongated shapes located beneath the landslide’s main scarps. The highest peak rates of a few centimetres per hour were recorded in the western sector (near the landslide’s right flank), but the movements progressively expanded upwards in both sectors during heavy rains and these displacement patterns clearly appeared to be related to the propagation and retrogression of the superficial earth flows caused by the rainfall.

The interferometric method proved to be a powerful tool uncovering differences in the displacements measured using the SAR apparatus and registered by the topographic station amounting to less than ±3mm. Nevertheless, the high cost of the instrument and its complex elaboration method led the researchers to search more economic solutions suitable for the long-term monitoring.

A ground-based SAR interferometry is not yet in operation at the Tessina landslide. State-of-the-art technology based on stereophotogrammetry applied to digital imaging interpretation (which in the near future will replace interferometry) is currently under development. Although the new system will lower the monitoring system’s precision slightly, it is more economic and will certainly provide an effective early-warning system for the Tessina landslide.

4. The monitoring system at Pian de Cice

4.1. Past monitoring activities

The unstable Pian de Cice sector lies between 1100 m and 900 m a.s.l. on the eastern slope of the Tessina valley (Fig. 4). The three benchmarks (307, 308, and 309), that have proved to be resistant to landslide movement in this area, are occasionally monitored by conventional topographic survey systems.

Despite the fact that monitoring was occasionally interrupted, the data recorded from 1997 until now have shown that these three benchmarks are moving continuously at a rate between 2.8 and 10 cm/year (Figs. 6a, b) with a seasonal pattern similar to that observed in the depletion basin (compare curves of Figs. 6 and 5). The accumulated movements are relatively extensive: for instance, the benchmark 309

Fig. 5 – The displacement over time of some topographic benchmarks in the upper basin in the 04/03/1997 - 30/09/1999 period.

Fig. 6 – The displacement recorded at some topographic benchmarks at Pian de Cice in the 04/03/1997 - 30/09/1999 (a) and 26/04/2008 - 07/07/2008 (b) periods.
(B309), located in a central position in the unstable block, accumulated a displacement of about 15 cm between 1997 and 1999.

As the stability of that sector plays a crucial role for the stability of the entire slope, with critical implications for the safety of the local inhabitants, an additional geological and geotechnical investigation campaign was carried out in 2007 to better define the local geological status and its kinetic evolution as a function of meteorological conditions [MARCATO et al., 2009]. The investigation included geophysical and geomorphological surveys coupled with inclinometric and piezometric measurements.

Two Energy-Recovery Linac (ERL) tomographic sections and core samples collected from the boreholes made it possible for us to reconstruct a litho-stratigraphic sequence of the unstable block composed of Flysch Formation covered by moraine deposits of variable thicknesses (Fig. 7).

The displacement measurements carried out in the past at inclinometers I1 and I2 clearly showed a slip surface located at depths of 26 m and 10 m, respectively, along with an annual cumulative displacement of 4.5 cm. No failure surface was detected in borehole I3. The movement rate recorded along the slip surface was the same as that measured by the topographic station (benchmark 309) with a sliding direction of 230° with respect to north. Unfortunately, these inclinometers are no longer accessible due to the excessive accumulate deformations.

According to the results of the inclinometer, the slip surface seems to be located approximately along the contact between the loosened Flysch layer and the moraine deposits. The unstable volume is approximately $6 \times 10^5$ m$^3$ extending nearly 300 m in the central longitudinal section with a maximum width of about 150 m.

In order to evaluate the relationship between the rainfall, the ground water table (GWT) variations, and the displacements, the data collected at the nearest meteorological station of Roncadin (about 500 m distant from benchmark 309) and the GWT elevation continuously monitored by piezometer P1 were analysed.

The 10 year observation series (1997-2007) showed very slight fluctuations (0-0.5 m) in the GWT which remained almost constant at the contact between the moraine deposits and the Flysch bedrock: this suggests that there was a rapid groundwater outflow at the P1 location (Fig. 4) and pore water pressure dissipation in the moraine deposit. MARCATO et al. [2009] concluded that only extremely unfavourable meteorological conditions...
logical conditions could induce instability in this zone, despite the fact that no reliable correlation between GWT and displacements could be established.

4.2 New monitoring devices

The difficulty in establishing a relationship between the GWT level and the displacements made necessary to modify the monitoring system layout by installing new devices in a more advanced position. In June 2012, an inclinometer guide (namely the I1bis) was driven down to 30 m close to the existing inclinometer I1 (Fig. 4). After 3 months, when the vertical displacement profile clearly showed there was a shear band at the interface between the moraine and altered Flysch (Fig. 8), two in-place 1-meter long inclinometer probes (IPI1 and IPI2) and a pore water pressure transducer (PPT1) were set in the borehole, respectively, at depths of 25.5-26.5, 27.0-28.0, and 28.65 m.

The IPI2 probe unfortunately failed immediately because of electrical problems and was unable to provide any information. The other inclinometer probe regularly recorded data every hour from 29/09/2012 to 5/3/2014 when it reached an upper measurement limit because of a sudden landslide acceleration.

Converting the tilting angle into displacements, Figure 9 shows the pattern of horizontal displacement over time according to probe IPI1. It has to note that the main direction of this displacement is 260°N according to the dip direction of the slope.

Between 29/09/2012 and 5/3/2014, the total displacement measured using the IPI1 was approximately 35 cm (rotation of 19.3°), with the greatest movement taking place in January and February 2014. In view of the fact that the vertical profile of the I1bis inclinometer showed a displacement localization in a thickness of 2.2.5 m (see profile of Figure 8) with the sliding soil mass above behaving essentially as a quasi-rigid block, the displacement at the slope surface can be calculated by multiplying the 35 cm displacement by the shear band thickness. Consequently, the total superficial cumulative displacement at the ground surface close to I1bis was approximately 70-85 cm.

To check the validity of this calculation of the local deformation process, some data from benchmark B309 were used. Between 21/5/2013 and 13/6/2013 (i.e. 21 days), the B309 moved 32.5 mm and over the same time interval, the 1-m long inclinometer probe IPI1 tilted 14.8 mm/m (0.85°). Dividing 32.5 mm by 14.8 mm/m, produces 2.2 m, which falls in the range of the shearing zone thickness (2.0-2.5 m) above outlined.

For comparison purposes, figure 9 also outlines the 1-hour rainfall height from the nearby meteorological station and the pressure head $h = u/\gamma_w$ recorded at PPT1.

It has to note that the upper 10 m of the inclinometer casing is cemented while the lower part is in contact with the soil. For this, we can assume that the casing behaves like an open standpipe and the pore pressure measurement at PPT1 can be considered a linear function of the height of the entire
water column inside the standpipe. Consequently, the pressure head \( h \) can be assumed equal to the GWT elevation above the transducer and, being \( z_t = 28.65 \text{m} \) the depth of PPT1 from ground surface, the GWT depth \( D \) can easily be obtained by \( D = z_t - h = 28.65 - h \).

Despite the fact that the moraine cover lying above the Flysch is highly permeable, as the data reported in the follow indicate, a partially undrained response of Flysch in shearing condition cannot be completely excluded: in that case, the relationship between the pore water pressure at PPT1 and the GWT level in the standpipe would not be unique. For this reason, in the follow analysis we preferred to use the pressure head measurement of PPT1 without converting the data into the GWT depth.

The cumulative rainfall, whose rate rarely exceeded 15 mm/h, was about 3750 mm over the entire period and was unevenly distributed throughout the different seasons/years. In 2013, the total accumulated rainfall reached 2145 mm with a higher meteoric precipitation in spring (March – May) and autumn (October – November). The beginning of 2014 was an exceptionally rainy period: in fact, although it rarely exceeds 200 mm in other years, in January and February the total rainfall was 844 mm. It is important to note that several important snowfalls also occurred in the 2013-2014 winter.

The pressure head values at IP11 were strongly correlated to the rainfall pattern, varying between 10.5 and 7.8 m, a wide interval when compared with the GWT variations observed in P1 in 1997-2007.

With the exception of the dry summer season, the pressure head in IP11 rose rapidly as soon as it began to rain (there is an average delay of 5 hours between the rain peak and the maximum \( h \) value recorded in PPT1) and gradually fell during the days following the rainfall. In summer, the trend of the pressure head seemed relatively stable and not influenced by the meteorological contribution. In fact, even if some precipitation episodes also occurred in July and August 2013, the pressure head continued to fall gradually at a constant rate, approaching a long-term minimum low value of 7.8 m. At the beginning of 2014, the exceptional amount of rain maintained the pressure head above 8.6 m for two months and above 9.6 m for 4 days (from 31/1 to 3/2).

Examining only the data collected during 2012-2013, the displacement rate increased or decreased according to \( h \) variations, showing higher values during the spring period, thus confirming what already observed with the benchmark topographic monitoring (Figs. 6a and b). The displacement rate was about 2-5 times greater than that observed in 1997-99 and 2007-08 at benchmark B309 (5.5 and 12 cm/y respectively), clearly confirming that the Piano de Cice block was accelerating.

The high rate reached in March 2014 may have been due to a partial slope collapse, as a new scarp became evident at the rear of position I2 and a general lowering of about 15 cm is now evident in the area. In the second half of March 2014, however, the sliding movement strongly decelerated to a negligible displacement rate consistent with the natural de-watering of the slope.

4.2.1 Filtering Displacement Measurements

A reliable displacement rate value is necessary if we are to correctly analyse the relationship between the displacement rate, the pressure head at PPT1, and the rainfall. Since the first quantity is calculated as the ratio between the displacement increment and the time interval, its pattern is strongly dependent, at very slow displacement rates, on the size of the time interval selected. The shorter the interval adopted for calculating the displacement gradient, the larger the effect of individual measurement error on the displacement data and the larger the oscillations of the displacement rate over time. On the contrary, the larger the time interval, the smoother the pattern of the displacement rate, but the relationship between the pressure head fluctuation and the landslide acceleration may be less evident.

To evaluate the best time interval to interpret the monitored data in terms of the displacement rate, three different time step values were tentatively considered, namely 1, 5, and 24 hours. Figure 10 outlines the displacement mean rates calculated according to these intervals over a total period of 6 days. The raw 1h mean rates were excluded from the graph due to excessive and meaningless oscillations including even unrealistic negative values. The more stable 24h mean rate plotted on the graph tends to
hide relevant information on the landslide response, while the 5h interval seems to be the most suitable value accounting for the accuracy of the monitoring system. The 5h rate should be more sensitive to the water pressure variations, but, as clearly shown in figure 10, it still presents relatively large variations that regularly occur, probably due to the unavoidable noise of the measuring device due to the daily intensity excursion of the solar panel power supply. Using the Fourier analysis, we identified two 1-day and 12h cyclic signal components which we eliminated by filtering the data.

Figure 10 shows the 1h mean rate calculated on filtered data superimposed onto the 5h and 24h rates determined on raw data. It is evident that data filtering leads to much smaller but still unrealistic oscillations, probably due to the level of accuracy of the monitoring instruments. To eliminate these residual oscillations, the filtered 1h rate is further averaged using a central-moving mean method over 5h: the new data are plotted in figure 10, for the sake of comparison, and they will be considered the reference for the landslide behaviour analysis presented in the following part of the paper.

4.2.2 The displacement rate in relation to pore water pressure and amount of rain

Here, we examine the monitored data and measurements in four particularly significant periods to analyse the relationships between the amount of precipitation, the groundwater pressure, and the displacement rates. The selected periods are the following:

1) 24th November - 17th December 2012;
2) 21st October - 26th November 2012;
3) 28th February – 15th April 2013

Figures 11a-d depict the data on the amount of precipitation, the pressure head and the displace-
ment rate, the latter determined as the 5h-mean value on filtered data described in section 4.2.1, for all the selected periods.

In the first period, i.e. 24th November - 17th December 2012 (Fig. 11a), a rainfall event, occurring after 15 dry days, was exceptionally relevant (157 mm of total rainfall in 56 hours with a maximum intensity = 18 mm/h). This was considered a particularly important finding that would permit us to evaluate the effect of a single rainfall event on the evolution of the Pian de Cace block.

As already explained above, the rainfall caused the pressure head to rise rapidly. About 3 hours after the maximum rainfall intensity, it reached the temporary maximum peak of 10.5 m; after the rainfall ended, it remained almost stable at approximately 10.0 m for about 1.5 days and then gradually fell.

This particular pattern of the pressure head recorded at PPT1 may have been linked to the occurrence of two different seepage phenomena:

a) During the rainfall event, we can suppose that the water rapidly infiltrates vertically towards the permeable moraine materials. Since, in the upper 10 m long portion, the annular space between the inclinometer casing and the ground is filled with impermeable clay and the water takes about 3 hours to complete its vertical path, the seepage velocity results equal to 9.3·10^-4 m/s ±10^-5 m/s which is a proper value for a vertical seepage in sandy-gravel soil;

b) After the rain ended, the groundwater flowed parallel to the top of the Flysch bedrock, probably along less permeable paths, thus requiring a longer time to discharge the large volume of water accumulated in the uphill slope.

From the time when the rainfall began, the displacement rate gradually increased for about 3 days, both as a function of the rainfall persistence as well as of the time the pressure head remained above 10.0 m. The maximum acceleration reached was equal to 2.25·10^-3 cm/h² = 0.11 mm/d². After the rainfall stopped, the pressure head began to fall and the acceleration gradually decreased, with a noticeable maximum rate occurring 82h after the rainfall peak.

In the second period, i.e. between 21th October - 26th November 2012 (Fig. 11b), four subsequent rainfall events of medium-high intensity occurred within a short interval of time (the peak-to-peak time distances are 97, 98 and 156 hours respectively), and could thus be considered a single rainfall episode.

The total precipitation during the four events was 82 mm in 36 hours, 80 mm in 24 hours, 78 mm in 40 hours, and 149 mm in 45 hours; the maximum intensity varied between 10 and 13 mm/h. The maximum pressure head at PPT1 was reached during the 3rd event, even if it did not correspond to the maximum intensity, probably because the rain start-
ship between rainfall and the hydrological behavior of a slope. The irregular morphology of the deep layers causes, moreover, a non-homogeneous strain distribution and strain rate within the sliding mass causing local excess ground pore pressure development during movements and its dissipation during at-rest conditions. The macro-scale effect of these phenomena is a variable hydro-mechanical response inside the soil mass, but also variable over time at the same position.

In landslides characterized by large displacements along well-defined sliding surfaces in cohesive materials, the available shear strength is the residual one [Skempton, 1985], which eventually increases slightly during quiescent periods [Angel et al., 1994; Stark and Hussain, 2010].

Clay soils exhibit creeping behaviour; the deformation in the soil under any constant effective stress depends, consequently, on time. An ideal creep behaviour is usually divided into three types of responses occurring in sequence: the primary creep, occurring immediately after stress application in which the flow rate progressively decreases until it is nullified; the secondary creep in which the flow occurs at a constant rate; the tertiary creep in which the rate of acceleration sharply increases leading to failure. Depending on the stress level in relation to the final resistance, the soil may respond with two or all three components of the sequence.

According to Ter Stepanian [1963], Suklje [1969] and Yen [1969], creep can also occur due to shear stress lying within the range between the peak and the residual shear strengths depending on the history of the landslide. Following a detailed examination of the most useful time-dependent constitutive models, Vulliet and Hutter [1988] suggested that creeping strains occur at every shear stress level, i.e. both for shear stress that is lower or greater than the residual strength. Even if these investigators did not distinguish between primary, secondary, or tertiary creeps, it is reasonable to assume that the tertiary creep takes place if the shear stress is greater than shear resistance.

On the basis of laboratory creep tests performed in residual stress conditions, Briat et al. [2011] demonstrated that secondary and tertiary creeps occur when the mobilized shear strength along the sliding surface overcomes the residual resistance. The delay of failure depends upon the difference between the mobilized shear strength and the residual strength, the difference being the factor influencing the acceleration of an unstable mass. Briat et al. defined the Residual-state Creep Stress Ratio (RCSR) as the ratio between the mobilized shear stress and the residual shear strength and they observed that, in secondary and tertiary creeps, the time of failure depends on the RCSR value: when RCSR moves from 1.0 to 1.02, the time of failure decreases from more than 10⁸s to 100s. Similar results were described by Di Maio et al. [2013].

All together, these studies suggest that achievement of a complete failure is possible only if/when the secondary and tertiary creep phases last long enough to allow high displacement rates: if they do not, the landslide returns to more stable conditions. Observing the displacement pattern of Pian de Gice (see for example, the one presented in figure 11a), this sector appears to be in an accelerating phase, but the creep phase is generally too short to cause the slope to collapse and, consequently, as the pressure head at PPT1 falls, the displacement rate decreases.

Despite the fact that it is difficult to analyse a creeping slope, many attempts based on simple approaches to model the displacement pattern as a consequence of a hydrological response have been described in the literature. The majority of the models proposed are linked to the hypothesis of a friction-viscous soil behavior and to a representation of a very long, wide, thick sliding mass with the GWT and the sliding surface parallel to the ground slope. For a slope with a shear resistance described in accordance with the Mohr-Coulomb criteria, i.e. depending on both cohesion and friction, the momentum balance can be defined as:

\[ \tau - \tau_v = ma \]  

(1)

where:
- \( \tau \) is the mobilized shear stress = \( \gamma b \) \( \tan \phi \)
- \( \tau_v \) is the shear strength of soil along the sliding surface = \( c' + (\gamma h) \cos \alpha \tan \phi' \)
- \( \gamma \) is the unit weight of the soil
- \( \alpha \) is the surface slope angle
- \( h \) is the piezometric level on the sliding surface
- \( c' \) and \( \phi' \) are the Coulomb parameters of resistance
- \( m \) is the mass of a soil column with a base of 1m
- \( a \) is the gradient rate = \( dv/dt \)

While several models have been proposed to consider the viscous soil response, Bingham's law and the equation proposed by Vulliet and Hutter [1988] are probably the most commonly used. The Bingham model states the proportionality between shear stress and velocity as:

\[ v = z_b \tau_v \eta = \tau_v / \eta_d \]  

(2)

in which \( \eta_d = \eta / z_b \) is the coefficient of dynamic viscosity, defined as the ratio between viscosity \( \eta \) and shear band thickness \( z_b \).
The Vulliet-Hutter equation introduces a power relation between the same quantities as:

\[ v = \frac{\tau_v}{\eta_o} \]  

(3)

which requires two material constants \( \eta_o \) and \( b \), i.e. the intrinsic viscosity and a power exponent.

Introducing one of the above viscous models, for example equation (3), in the momentum balance equation, we have:

\[ \gamma \sin \alpha - \gamma \cos \alpha \tan \phi' \cdot \eta_v \cos \alpha \tan \phi - \eta_o \beta^b \cdot \gamma \ln g \]  

(4)

The second term of equation (1) is generally negligible (e.g. with regard to Pian de Cice the mean acceleration during a rate increase is approximately \( 10^{14} \cdot 10^{15} \text{ m/s}^2 \) and the term \( m \omega^2 \gamma \ln g \) is found to be equal to \( 10^{13} \cdot 10^{15} \text{kPa} \). The momentum balance equation may be consequently expressed as:

\[ v = \eta_o (\gamma \sin \alpha - \gamma \cos \alpha \tan \phi' \cdot \eta_v \cos \alpha \tan \phi - \eta_o \beta^b \cdot \gamma \ln g) \]  

(5)

Since in the second term of equation (5) the only variable quantity is \( h \), this equation expresses a direct relationship between the velocity and the piezometric level at the sliding surface that can be calibrated on the basis of laboratory test results or in-situ measurements.

It should be noted that the calibration could be carried out by assigning reasonable values to parameters such as \( c' \), \( \phi' \), \( \alpha \), \( \eta \) and \( z_b \) or \( \eta_o \) and \( b \), but may be also obtained without necessarily relating them to real in-situ conditions, especially when some of these parameters are not easily determinable.

These models successfully describe the behaviour of some landslides, but there are also cases in which they are unable to adequately reproduce the landslide’s movement. By implementing the Bingham or the Vulliet-Hutter equations within a “dynamic” model, Corominas et al. [2005] were able to successfully simulate the displacement patterns of the large Vallcebre landslide (Barcelona, Spain) in relation to the groundwater oscillations. The excellent database on the shear strength of soils involved in the sliding process and the assumption that the shear resistance was equal to the residual value enabled the authors to calibrate relationships between the velocities and water pore pressures in some points obtaining almost constant viscous parameters. The predictable errors in the two viscous models were of the same order of magnitude.

The key to success in the Vallcebre landslide model was the strong correlation between the GWT level, showing maximum excursions of 4-8 m, with velocity. If this kind of correlation does not exist, the viscous model cannot be applied in such a simple way and other hypotheses must be utilized. In the case of the Alverà mudflow (Cortina d’Ampezzo, Italy), Angeli et al. [1996] used the Bingham model, but in order to attain a satisfactory simulation of its slow-moving behaviour they introduced two piezometric threshold values, the lower one corresponding to a complete cessation of landslide motion and the upper one apparently linked to a re-activation of its movement. This means that as the GWT rises, the landslide is reactivated only when the upper threshold is exceeded, but during the falling phase, it ceased only when the GWT falls below the lower threshold value. It is to be noted that the difference between the two thresholds was 40 cm corresponding, for the Alverà landslide, to a shear resistance modification of approximately 1 kPa. The authors, who considered this difference as the gain strength occurring during a rest period, did not provide any information about the relationship between the GWT position and the velocity, but it is probable that, in contrast to the Vallcebre landslide case, the relationship was not unique.

Another example concerns the La Valette landslide [Van Asch et al., 2007]. In this case, the authors demonstrated there was no objective relationship between the flow rate and the GWT level neither during the rising nor the falling limbs of the GWT, meaning that a unique set of viscous parameters is unable to describe the time dependent behaviour of the landslide. In particular, Van Asch et al. observed that, for a given GWT level, the velocity and the acceleration are higher in the rising than in the falling phases: they ascribed this behaviour to the undrained response of the soil which induces increasing pore water pressure during displacement.

Before we go on to analyse the relationship between the displacement rate and the pore water pressure at the Pian de Cice sector, some preliminary comments are necessary.

The first concerns the rapid rise in the PPT1 pressure head observed during the rainfall event (see Fig.11a). As previously discussed, the groundwater pressure reached its minimum level 3-6 hours after the maximum rainfall intensity, but the change in velocity during that short interval was negligible. Moreover, during every intense rainfall event, the pore pressure presented two subsequent oscillations. The first, occurring during the rainfall, was characterized by a very high pressure variation rate that permitted a maximum pressure head of 8.2-8.4 m to be reached; the second, starting after the rainfall ended and characterized by a smaller pressure change rate, led to a maximum pressure head of approximately 7.75 m to be reached.

This particular behavior could be tentatively explained in view of the fact that the first rapid water level variation in the standpipe piezometer was main-
ly influenced by local infiltration inside the stand-pipe during groundwater percolation. As a consequence, this sudden pressure rise was not directly related to a real transient groundwater seepage into the slope mass and it did not affect the displacement rate. This anomalous pressure response, which must be taken into consideration, certainly affects data interpretation.

After carefully analyzing the maximum variation of the pore water pressure vs. time in the subsequent rising phase, we applied an automatic filter that removed changes in the pressure head values that were greater than 18 cm/h, smoothing the time pattern of this quantity, as shown in figure 12a. It is important to remember that the correction is applied to time intervals that do not exceed 10 hours, thus leading to a more regular pattern in the pressure head’s ascent.

The second comment concerns the selection of the most suitable interpretation model. Equation (5) should strictly apply to planar translational slips but, in view of its simple mathematical form, many researchers have even applied it to non-planar slope conditions. Of course, the discrepancy between measured data and model predictions observed in some cases (i.e. van Asch et al., 2007) may also be due to the assumption that there was a planar slip surface, which is not in accordance with the real in-situ hydro-morphological conditions.

As the ground water condition at Pian de Cice is being monitored by only one position, it is difficult to assess since the groundwater seepage is parallel to the slope surface. Moreover, the schematic longitudinal section depicted in figure 7 fails to support the hypothesis of a planar slip. Despite these conflicting findings, we have proposed a black-box model to describe the relationship between the pressure head and the displacement rate which is outlined below.

The graph in figure 13 outlines in a double logarithmic scale the displacement rate $\nu$ vs. the pressure head $h$ determined during the rising and falling phases of the GWT. The data are, of course, only a selection of those available in the database, specifically chosen to give an idea of how the system responds in different conditions.

In a double logarithmic plot, data lying on a line segment satisfy a relationship such as the one given by equation (5). Data lying on a unique line would mean a landslide behaviour according to the Vulliet-Hutter model, thus confirming the applicability of a viscous material model to the $\nu-h$ relationship. In our case, the data do not fit this condition, probably...
as a result of the discrepancy between the landslide morphology and the corresponding hydrological behaviour and the planar failure surface assumption, as explained above.

Some singular patterns can, nevertheless, be found in figure 13. It is evident that the data points seem to fit linear segments with various orientations on the v-h plane. We can, in particular, pinpoint three specific orientations, labelled in the figure as lines of type a, b and c.

The data stretch from one segment to another with continuity: for instance, following the falling phase in the 23/11/2013 – 23/12/2013 period, the data align on three segments parallel to lines of type a, b and c.

The transition from one segment to the subsequent one occurs when the pressure head falls below two specific thresholds: threshold F3, pinpointing the passage from line c to line b, and threshold F2, referring to the passage from line b to line a. For the rising phase (i.e., data of the 10/10/2013-14/10/2013 interval), the measurement points are located on two segments separated by the threshold F1: if \( h \) is smaller than F1, the segment is oriented such as line a, otherwise the segment is parallel to line b.

It is important to note that the segment orientation is fixed while the segments change position according to the previous acceleration history: in fact, the data referring to the rising phases corresponding to various events are located on different parallel segments. Moreover, in contrast to what was demonstrated by VAN Asch et al. [2007], the velocity in the GWT falling phase is generally higher than that corresponding to the rising one. We are convinced that this response is due to the deep morphology of Flysch, which forms an isolated depression in the upper part of the slope (see Fig. 7) where the groundwater probably accumulates during a rainfall, then slowly drains towards the lower part. This would justify the amount of time that the pressure head needs to return to the smallest values after a rainfall ends and the existence of a period characterized by a contemporaneous increase in the displacement rate and a fall in the pressure head.

A careful analysis of the data indicates that if the pressure head has not previously exceeded F3 \((h>F3)\) in the falling phases, the velocity retraces the same type-b segment followed in the rising phase. Otherwise, it follows a type-c segment until it reaches the F3 threshold, with a real acceleration of the landslide. In fact, when the pressure head falls below the F3 threshold, the velocity follows a new type-b segment, parallel to the one followed in the rising phase but located above it, with a hysteretic response. Moreover, if the pressure head falls to a point overcoming threshold F2, once again the rate moves on a type-a segment located parallel and above the type-a segment followed during the rising phase.

In order to reproduce this very complex behaviour, we propose an algorithm based on the following three fundamental \(v-h\) relationships:

1) \textit{type-a line.} \( v_a = \eta_a h^{b_a} \) \hfill (6)
2) \textit{type-b line.} \( v_b = \eta_b h^{b_b} \) \hfill (7)
3) \textit{type-c line.} \( v_c = \eta_c h^{b_c} \) \hfill (8)

with exponents \( b_a, b_b \) and \( b_c \) obtained by means of data calibration, while values \( \eta_a, \eta_b, \eta_c \) must be assigned in relation to the current pressure head \((h_i)\) and of the pressure head’s rising or falling condition. Figure 14 outlines the algorithm flow-chart: it is designed assuming that point \((h=0, v_0)\) is the starting condition, the initial in-situ condition measured. For the following time steps \((t_{i+1})\), the algorithm choses the segment on which the point falls on the basis of the value of pressure head \(h_{i+1}\) at current time and of the previous acceleration sequence, and it consequently calculates the flow rate with the relations:

1) \textit{condition a:} \( v_{i+1} = \eta_a h_{i+1}^{b_a} \) \hfill (9)
2) \textit{condition b:} \( v_{i+1} = 10^{\log v_i + b_b \log h_{i+1}/h_i} \) \hfill (10)
3) \textit{condition c:} \( v_{i+1} = 10^{\log v_i + b_c \log h_{i+1}/h_i} \) \hfill (11)
The black-box model has seven parameters: exponents $b_a$, $b_b$, and $b_c$, thresholds $F_1$, $F_2$ and $F_3$ and the initial displacement rate $v_0$. A good description of the velocity pattern was found with the following values:

- $b_a = 20.7$, $b_b = 4.6$ and $b_c = -4.6$;
- $F_1 = 8.35$ m, $F_2 = 8.60$ m and $F_3 = 9.53$ m;

being $v_0$ the value of in-situ measured $v$ at the beginning of the simulation.

In figure 12$^c$ the pattern of velocity determined by the proposed model is compared to the in-situ measurement. The correspondence seemed to be very good, since the model accurately reproduces the maximum $v$ values of the main rain events and takes into consideration the $v$ increase in the de-watering phases. It predicts, moreover, the gradual increasing velocity pattern that was observed during the overall as well as the final acceleration periods, even if the displacement rate during the latter, i.e. approaching a partial collapse, did not exactly correspond to the measurements.

Some occasional acceleration events is not predicted well. The change in velocity that occurred at the end of January 2013, probably due to the snow cover load not accompanied by a rise in the pressure head, is not taken into consideration by the model. Similarly, the model does not correctly reproduce the trend during the snowmelt period that followed, when the in-situ pore water pressure increased without any acceleration in displacement. This induces us to conclude that the model is able to describe the $v$-$h$ relation during rainfall events but not the effects caused by the presence of snow.

Figure 12$^b$ shows a very satisfactory correspondence also between the total displacements measured with IP11 and the model.

Finally, in order to further examine the differences between this model and the classical viscous one, we selected a series of short intervals in which the quantities $v$ and $h$ are almost constant and we calculated their mean values during each interval. The mean data are then outlined in a double logarithmic scale (Fig. 15), subdivided into 5 groups:

1. the data corresponding to rising pressure for events occurring after November 2012;
2. the intervals on very long falling limbs occurring after November 2012;
3. the data in an intermediate situation, i.e. in the middle between the initial rising and final falling phases occurring after November 2012;
4. the data referring to the snowmelt period that occurred in spring 2013;
5. the data referring to the initial 2 months during which measurements were taken (before November 2012).

This subdivision was used with the intent of examining if the relationship $v$-$h$ is influenced by significant factors, namely the snowmelt effect or if the
rate was excessively low during the period immediately after the inclinometer was installed because the adherence between the casing and ground was imperfect.

It is clear that a unique regression line could not interpolate all the data and only two series of data (those referring to the rising and falling GWT) seem to be well represented by the following equations:

fit for the rising phase:
\[ v_a = 1.1 \cdot 10^{16} h^{-0.84} \quad (r^2=0.856) \quad (12) \]

fit for the falling phase:
\[ v_a = 4.3 \cdot 10^{29} h^{-21.56} \quad (r^2=0.930) \quad (13) \]

The displacement rate predicted via equation (12) and the relative cumulative displacements are, respectively, outlined in figure 12b and 12c and compared with the experimental data and the prediction of the new model proposed. It is evident that the rates predictable with equation (12) are within the range of the measured data, but they vary within two limit values that remains constant during all the monitored period or consequently to multiple rainy events, since equations (12) and (13) are based on an objective relationship between \( v \) and \( h \). The velocity is thus slightly overestimated in the first period and greatly underestimated in the last one. It is, moreover, evident that it is nearly impossible to reproduce the rising and the falling limbs at the same time.

On the other hand, equation (13) greatly overestimates the velocity during the entire period monitored: the relative curve is not drawn in figure 12c because it falls outside of the range for long intervals. Consequently, the total displacement calculated using equation (13) results about 5 times greater than the measured one and a model based on this equation is absolutely inadequate for reproducing the Pian de Cice time behaviour.

5. Current evolutional kinematics in the depletion basin

To keep the landslide evolution in the upper basin under control, a new image-based landslide monitoring system, constituted by two digital reflex 18 MPixel cameras, was installed near the original topographic station. Each camera captures one digital image of the depletion area at noon every day. Even if the final aim of this system is to develop a stereo-photogrammetric method to evaluate superficial displacements in the area, the image sequences collected until now provide a suitable database to analyse the kinematic evolution and the mechanisms triggering the earthflow.

In order to confirm the effectiveness of digital imaging in monitoring and interpreting the landslide evolution, a sample of images taken on 20th and 29th May 2013 is outlined here. It is important to point out that in the previous 85 days (from the 6th March to 29th May 2013) about 60 rainy days occurred with an exceptional cumulative rainfall of 670 mm. The images, moreover, confirmed that snow covered the entire area at the end of February. When the warmer season set in, snowmelt was followed by water infiltration and a GWT rise, as already shown in figure 11. Following these particular meteorological conditions, the shallowest soil in the depletion basin was well saturated, highly collapsible, and erodible.

Figure 16 compares the images taken by one of the digital cameras on the 20th and 29th May 2013. Displacement of the ground surface is noticeable even visually. Moreover, even if the moving area is
wide and far from the camera’s viewpoint, the images’ high resolution provides a small scale well-defined view. The high resolution also made it possible to use digital image processing techniques to detect superficial displacements as well as their temporal sequence.

Figure 17 shows, for example, enlargements of images in figure 16, included here to highlight the movement in the upper sectors on the left hand side of the depletion basin. Figure 18 is a processed image obtained by applying image processing methods to the images in figure 17 in order to identify the zones that are moving. A Gaussian blur filter is first applied, the image is then normalized and the absolute difference in the images is computed. In this way, the most active areas are brightly coloured, as shown by figure 18. Yellow dashed lines also highlight the differences between the images.

Finally, some well-defined objects like trees and boulders were tracked using normalized cross-correlation and their paths pointed out in figure 18 by means of red vectors which also help to indicate the direction in which the areas are moving.

From a wider perspective, it is worth noting that the steepest slopes were subject to intense superficial erosion, especially in the area where the Flysch outcrops. In addition, many shallow and relatively small mass movements mobilized the layer of loosened Flysch overlapped by moraine deposits. This is particularly evident in figure 16 at the lateral flanks, where material slides transversally on the planes formed by the harder calcareous layers of Flysch which seem to be stable.

The images in figure 17 show some springs emerging at the base of moraine deposits a few hours after the rainfall began, thus confirming the significant drainage properties of the moraine lying above the less pervious Flysch which acts as a base for groundwater seepage.

An analysis of the image sequence also highlights the formation process of earth flows draining into the intermediate channel. Accumulating temporarily at the toe of steep slopes, material is pushed towards the channel (this is clearly evident in the low part of figures 17 and 18).

Due to this driving action, the material constituting the more advanced portion of the mud flow that accumulated progressively increases the front part of the slope up to reach instable conditions. The material accumulated at the channel head is visible at the centre of figure 16a, but a part of this material failed on May 28th leaving a large depression, as shown in figure 16b. In this case, the collapse mobi-
lized a small volume of soil without causing any risk to the valley.

It is important to note that this long image sequence confirms the results previously obtained through benchmark monitoring and ground-based SAR interferometry, thus permitting a subdivision of the upper basin into several portions characterized by different displacement mechanisms.

As was observed for the fixed inclinometer at Pian de Cice, the images allowed us to highlight the movements involving moraine deposits and the layer of loosened Flysch with a thickness varying from less than 1 m to a maximum of about 25 m. The displacements occur almost immediately after the onset of rainfall, but some sectors require 2 or 3 days to show a visually detectable displacement rate.

Conclusion

The Tessina landslide is a complex landslide, which has been active for more than 50 years in the Northern Italian Alps. Its evolution is typical of Flysch formations and characterized by medium-large roto-translational slides, periodically occurring in the source area, and evolving into earth flows. Both the upper landslide and the earth flows have been extensively monitored over the last few decades by means of various types of instruments in the attempt to enhance our knowledge concerning the processes involved in earth flow formation in view of adopting effective mitigation strategies.

Two new monitoring systems have been installed in the area during the last three years and the data collected until now has led to important advances in our understanding of landslide processes.

One system, constituted by a pair of fixed digital cameras taking images at selected intervals, was installed to keep the paths of detached slope scarp evolving into earth flows under control. This system also provides images of the entire landslide area and has uncovered various important factors (i.e. snow cover extension, which remains over time) that can partially explain the movements that have been observed. It was found that superficial ground movements, extremely sensitive to rainfall, usually involve the upper layers constituted by highly permeable sandy-gravel moraine overlapping the Flysch sedimentary rock. Improved digital imaging interpretation that will be available in the near future will convert stereo-photographic imaging into true measurements of the superficial displacements of the entire depletion basin.

The other system, which continuously measures groundwater pressure and deep localized displacement on the southern lateral flank (i.e. Pian de Cice), which seems to prone to an incipient collapse, has provided interesting data that has been presented and discussed here.

The data collected until now from the lateral flank is, however, particularly difficult to interpret since it does not appear to be any clear relationship between three fundamental variables: rainfall, groundwater pressure variation, and the displacement rate. This may be due to various factors, among which the variability of groundwater seepage during the different phases of rainfall events. The main outcome is that the lateral landslide flank is continuously moving at a variable rate that reaches a minimum value of 2 cm/year in the dry season, but can increase 10-100 times as a consequence of rainfall.

The data have been tentatively interpreted with the aid of a black-box model, and the results have been compared with those obtained using the Vulliet-Hutter equation for viscous soil behaviour. Three significant threshold values for ground water level were selected, separating different hydro-mechanical responses occurring as the ground pore pressure rises and falls during and after a rainfall period. The approach is far from simple, but, excluding the influence of a snow load, it seems to be able to describe the asynchronous response of the displacement rate to the groundwater pressure rise/decrease and the gradual acceleration of the sliding mass observed during repetitive or extended rainfall events.

References


Bettelli M., Valentino R., Salvatorelli F., Rossi Pisa P. (2012) – Monitoring soil-water and displacement con-


Evoluzione della frana del Tessina

Sommario

La frana del Tessina è una frana complessa attiva da più di 50 anni nella regione dell’Alpago (Belluno, Italia). La sua evoluzione, tipica dei movimenti franosi che si verificano nelle formazioni Terziare di Flysch, è caratterizzata dalla formazione di frane roto-traslazionali medio-grandi, che si verificano con cadenza elevata e si trasformano in colate di fango e terra che mettono a rischio le valli sottostanti. Nelle ultime decadi, la frana è stata intensamente studiata e monitorata allo scopo di meglio comprendere i suoi meccanismi evolutivi e adottare le più idonee strategie di mitigazione del rischio.

L’articolo presenta alcuni dati raccolti mediante due diversi sistemi di monitoraggio continuo, recentemente installati nell’area. Il primo sistema misura la pressione dell’acqua interstiziale e gli spostamenti in un settore laterale avente un ruolo cruciale per la stabilità dell’intera area; il secondo è un nuovo sistema fotografico, che cattura immagini digitali del bacino superiore con cadenza giornaliera, allo scopo di valutare i meccanismi di formazione delle colate nella zona interessata dai movimenti roto-traslazionali.

Le osservazioni raccolte fino ad ora permettono di capire come cambino le condizioni di stabilità delle diverse porzioni della frana, oltre a fornire importanti informazioni sul processo di formazione delle colate, informazioni necessarie per valutare la fattibilità e l’efficacia di possibili lavori di mitigazione da realizzare in futuro. Infine, grazie ai dati raccolti nel settore laterale è stato proposto e calibrato un nuovo modello “a scatola chiusa” per la predizione dei movimenti del settore monitorato: i suoi risultati sono confrontati con quelli di un modello viscoso molto spesso utilizzato per l’analisi della mobilità delle frane lente.