Considerations about the Mechanics of Slow Active Landslides in Clay

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Abstract. Slow active landslides in clay include slides, mudslides and spreads. Movement is induced by any change of effective stress and by creep; in the very long-time, some role may be played by a change in soil properties. Looking at geological phenomena causing movement, pore pressure fluctuations and erosion have a strong influence on shallow translational slides and mudslides, while creep or erosion and other geological phenomena of stress relief govern movement of deep seated slides and spreads. In several cases, excess pore pressures generated by changes of boundary conditions may play a significant role.

Keywords. Clay, slide, mudslide, spread, active slow slope movement

3.1 Introduction

Slow landslides are widespread in geomorphological contexts where stiff clays or clay shales crop out. According to old chronicles it can be argued that there are landslides which have been active from thousands of years.

The risk posed by slow slope movements is rather low, but management of landslides interacting with urban areas, infrastructures and lifelines raises peculiar problems since, even though the evacuation of people is not a pressing need, in the long-term movements can severely damage structures or interrupt the serviceability of lifelines, while slope stabilization or reinforcement of exposed elements is too expensive or non-effective. As a consequence, land management requires a complex and delicate cost-benefit approach in which assessment of future slope behavior and comparison among the effects of different remedial measures take a crucial role.

This paper concerns natural slopes in clay, and is mainly based on the state-of-the art report presented by L. Picarelli and C. Russo at the 9th International Symposium on Landslides (Rio de Janeiro 2004), but includes further data and considerations drawn from more recent papers and reports.

3.2 General Features of Slow Active Landslides

Slow active landslides involve a number of natural slopes much larger than what would be expected. The main types of slow active landslides are slides, mudslides and spreads. Movement is a result displacement along internal discontinuities (as the slip surface) and of internal strains. The landslide body is generally subjected to a constant driving force due to self-weight, thus movement is triggered by change of the resisting force caused by change of boundary conditions, and by viscous deformations. Furthermore, in the long-term, any change of soil properties due to weathering or to other processes of soil deterioration, can play some role. Finally, change of slope morphology caused by movement itself may restrain further movement. Therefore, the displacement rate depends on geometric features of the landslide body, on the rate of effective stress change and of soil deterioration and on viscous properties of both landslide body and discontinuities.

Usually, the expression "active landslide" refers to (those) soil bodies which have experienced a general failure (in the sense discussed by Urciuoli et al. 2007) and are still moving. Their behavior is represented by the last part of the schematic diagram reported in Fig. 3.1, which depicts a typical time-slope displacement relationship from the so-called pre-failure stage, i.e. before general slope failure, to the post-failure stage (Leroueil et al. 1996). Slow slope movements have much a longer life than rapid landslides: in some cases they last centuries or even thousands of years. Even though movement can appear permanent, i.e. characterised by a constant velocity, very often it displays a continuous change of the displacement rate, as a function of pore pressure fluctuations.

According to the classification proposed by Cruden and Varnes (1996), landslides can be classified slow as far as their velocity is less than 13 m month^{-1}, but for a velocity less than 16 mm yr^{-1}, they are categorised as extremely slow. However, here, the adjective slow will be used in a flexible sense, to indicate a long-lasting movement unable to provoke, in the short-time, any significant consequence on people and goods.

Many active landslides in clay (slides and mudslides) present a slip surface, generated by shear strain localization and soil rupture, over which the soil mass slides advancing on the slope. However, just after failure, mudslides display a peculiar flow-like style attaining a rapid to moderate velocity followed by a slow decline, while the style of movement progressively turns to slide (Picarelli 2001).

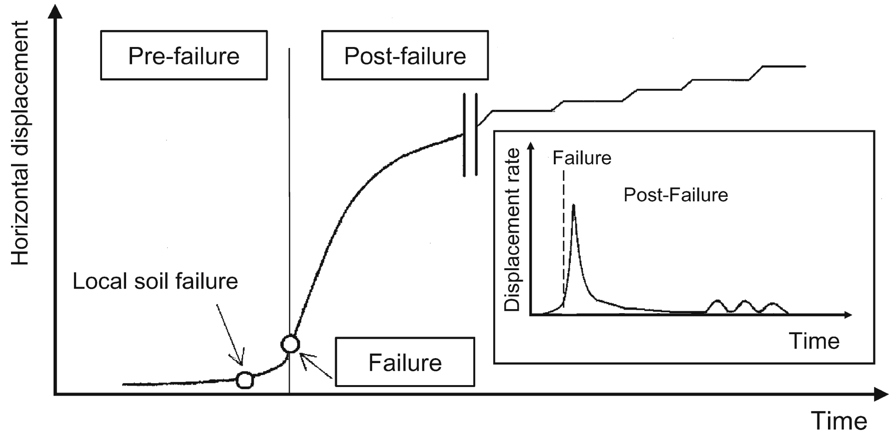


Fig. 3.1. Simplified scheme of slope movement from pre-failure to arrest (after Picarelli 2000)

In some cases people include in the category of landslides soil masses which did not experience any general slope failure, but are subjected to continuous movement due to perceivable internal strains. For instance, lateral spreads caused by valley rebound can be driven by deformation of even large soil masses, but not by slope failure. This implies that a persistent slip failure does not necessarily exist. Also movements caused by creep or by internal soil deformation provoked, for instance, by ongoing progressive failure or by internal stress redistribution, reveal a pre-failure more than a the post-failure stage. Moreover, these phenomena will not necessarily lead to general slope failure.

3.3 Considerations about the Mechanics of Active Slides and Mudslides

3.3.1 Active Slides

Slow active slides generally move over a shear zone located at the base of the landslide body. This zone is the result of a complex process of strain localisation occurring in the pre-failure stage, whose final effect is the formation of a shear surface, also called slip or sliding surface, internal to the shear zone.

Urciuoli (2002) analyses the formation of the shear zone in the case of infinite slope, showing that is accompanied by rotation of the principal stresses and formation of minor shears (Skempton 1967; Morgenstern and Tchalenko 1967). As the direction of principal stresses becomes consistent with the direction of the theoretical failure plane, i.e. parallel to the ground surface, a persistent slip surface eventually forms and slope failure takes place. This occurs simultaneously in all points of the failure plane because of uniformity of the state of stress. The thickness of the shear zone depends on the initial state of stress and on the shear strength of soil: for a critical value of the coefficient of earth pressure at rest, it is nil; for high values of OCR it is quite small; for small values, it is relatively large. In other cases formation of the shear zone is a more complicated process because of non uniformity of the state of stress. Generally it can be said that it starts forming in the most stressed part of the slope, propagating into the soil mass as a result of a mechanism of progressive failure which culminates in the general failure (Urciuoli and Picarelli 2004).

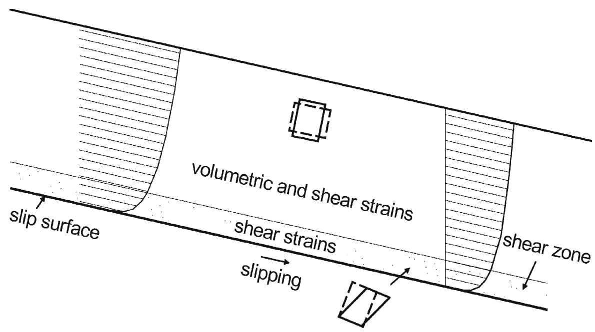
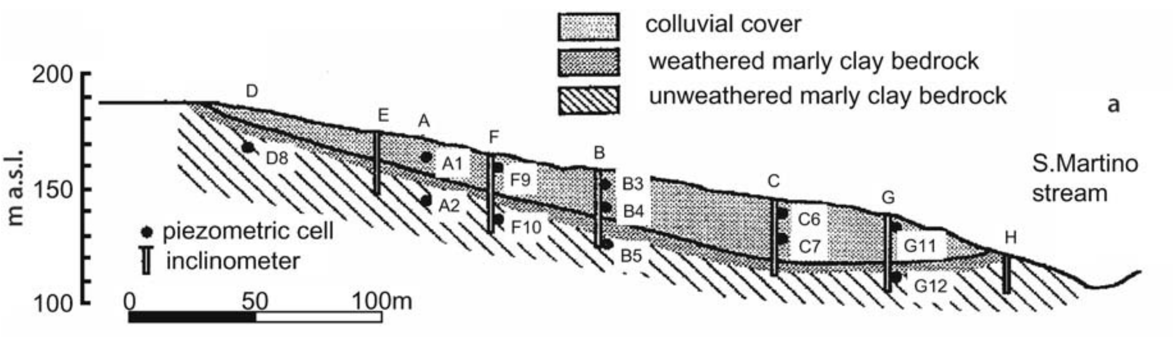


Fig. 3.2. The nature of movement of slides (from Picarelli and Russo 2004)

In the post-failure stage, the mobilized soil mass can eventually move over the slip surface. Even if immediately after failure the landslide may experience a strong acceleration because of brittle soil behavior, eventually it progressively slows down due to change of both slope morphology and Generalised Brittleness Index of soil (D'Elia et al. 1998), while the shear strength along the slip surface approaches the residual value. At residual, the resisting force along the slip surface depends only on the normal effective stress and, possibly, on viscous effects associated with the displacement rate. In the final stage, i.e. once the friction angle has attained a constant value, slope movement is governed by the cyclic balance between driving force and resisting force, being the result of sliding along the slip surface and of internal strains (Fig. 3.2). Slipping occurs every time the residual strength is mobilized along the shear surface, for instance when pore pressure trespasses a critical value. Internal strains are caused by changes of the effective stress field and by viscous phenomena, but in the very long-term additional strains can be provoked by change of stiffness due to soil deterioration. Concerning this point, Picarelli (2000) shows that the residual strength of clay shales may decrease with time because of changes of grain size due to progressive breaking of bonds linking aggregates of clay particles. In addition, accounting for data provided by Di Maio (1996) and by Di Maio and Onorati (2000), Picarelli et al. (2006) illustrate possible long-term effects of infiltration of fresh water in marine clay of high plasticity leading to a decrease of both peak (softening) and residual shear strength, but certainly, even soil stiffness experiences some decay. Some numerical analyses on the effects of soil deterioration on slope movements are reported by Yoshida (1990) through a manipulation of the Hoek and Brown (1980) strength criterion.

The behavior of translational slides (an example is reported in Fig. 3.3a) is governed by the component of the driving force in the direction of movement. Since such a component is essentially constant, the displacement rate depends on changes of the resisting force caused by changes of effective stress and on soil viscosity. Quite a similar framework held for compound slides, most of which are characterised by an almost rectilinear and subhorizontal lowermost part of the slip surface (Fig. 3.3b). However, in this last case the non uniformity of the state of stress may determine a complex interaction between the rear and the front part of the landslide body. The behavior of circular slides strongly depends on changing balance between the driving and resisting force due to modifications of the morphology caused by movement itself (D'Elia et al. 1998). As a consequence, circular slides generally run short distances and are active for quite a short time.

In clay deposits the groundwater table is located at rather a small depth from the ground surface, and fluctuates during the year as a consequence of seasonal water recharge and discharge. In uniform clay, the magnitude of pore water fluctuations decreases with depth because of the time required by pore pressure to reach equilibrium with boundary conditions, compared to the duration of the wet season. Kenney and Lau (1984) present excellent experimental data regarding a slope in Ontario, while Cavalera (1977) reports analytical solutions of this problem for sinusoidal changes with time of hydraulic conditions at the ground surface. This suggests that pore pressure fluctuation due to rainfall (or to ice melting) strongly governs the behavior of shallow translational slides, while its role is minor for deep-seated landslides. In contrast, the behavior of deep seated slides should be essentially governed by creep, i.e. by soil deformation occurring under an essentially constant state of stress, or by stress changes due to geological phenomena, as erosion. A well known slide driven by fluvial erosion is the huge La Frasse slide, which has an extension of 42 km^3. Monitoring dates back 170 years, since construction of the major cantonal road connecting the town of Aigle to the



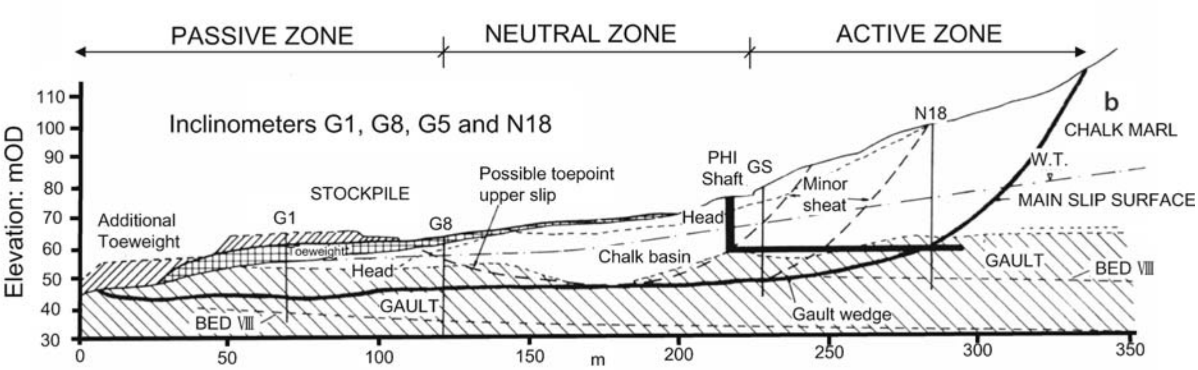


Fig. 3.3. Examples of slow active slides. **a** The essentially translational Fosso San Martino slide (from Bertini et al. 1986); **b** the compound Castle Hill slide (from Varley and Warren 1995)

Col des Mosses Pas, but the landslide is much older, as dating of wood fragments suggests (Tacher et al. 2005).

Literature reports numerous examples of slow slides. A number of these, as the La Frasse slide, are have been active for hundreds or even thousands of years, moving with an average displacement rate as low as a few centimeters per year or less. An example of long-lasting landslide is Castle Hill slide shown in Fig. 3.3b, which is believed to have formed about 10 000 years ago. Its displacement rate in the last tens of years, argued from deformation of a pipeline located within the landslide body, is about 1–2 mm yr^{-1}. The displacement rate of long-lasting slow movements is often assumed to be uniform with time, mainly because of a lack in continuous measurements, but once more data are available, some landslides show themselves to be intermittent (stick slip movements) being driven by pore pressures changes occurring over very short periods of time (Van Genuchten 1984). This naturally disproves the assumption of creep as cause of movement.

An example of creeping slide is reported by Picarelli and Simonelli (1991), who discuss deformation phenomena affecting the neighborhood of a small Italian town (Fig. 3.4), located on a gentle slope. The buildings rise on coarse calcareous debris covering a deposit of highly fissured sheared clay shales. At the time of investigations, the majority of dwellings and walls built in the last tens of years presented only some very small cracks, while only the oldest constructions (a monastery built some centuries ago, a building and a high retaining wall) displayed large cracks. Monitoring through some inclinometer purposely installed in the area revealed the presence of a slow active slide whose slip surface is located just at the top of clay shales. The displacement rate was quite constant (around 1 cm yr^{-1}) and uniform all over the area (Fig. 3.5). Readings carried out with Casagrande piezometers indicated that the groundwater table is located at the base of the landslide body, some decimeters above the top of clay, and that it experiences only very little annual fluctuations because of high permeability of the debris, preventing formation of an aquifer. Therefore the state of stress is practically constant and no significant future changes of this can be predicted. The small internal deformation of the landslide body depends on the high stiffness of debris; moreover, the absence of significant differential movements can explain the absence of severe damages on dwellings and walls built recently, while large cracks across the walls of the monastery and of the other old structures testify deformations cumulated over a much longer period of time.

Another example of slope movement presumably caused by creep is reported by Corominas et al. (2005), who describe the Vallcebre landslide in fine grained stiff deposits (Fig. 3.6a,b). Figure 3.7 reports the water levels measured

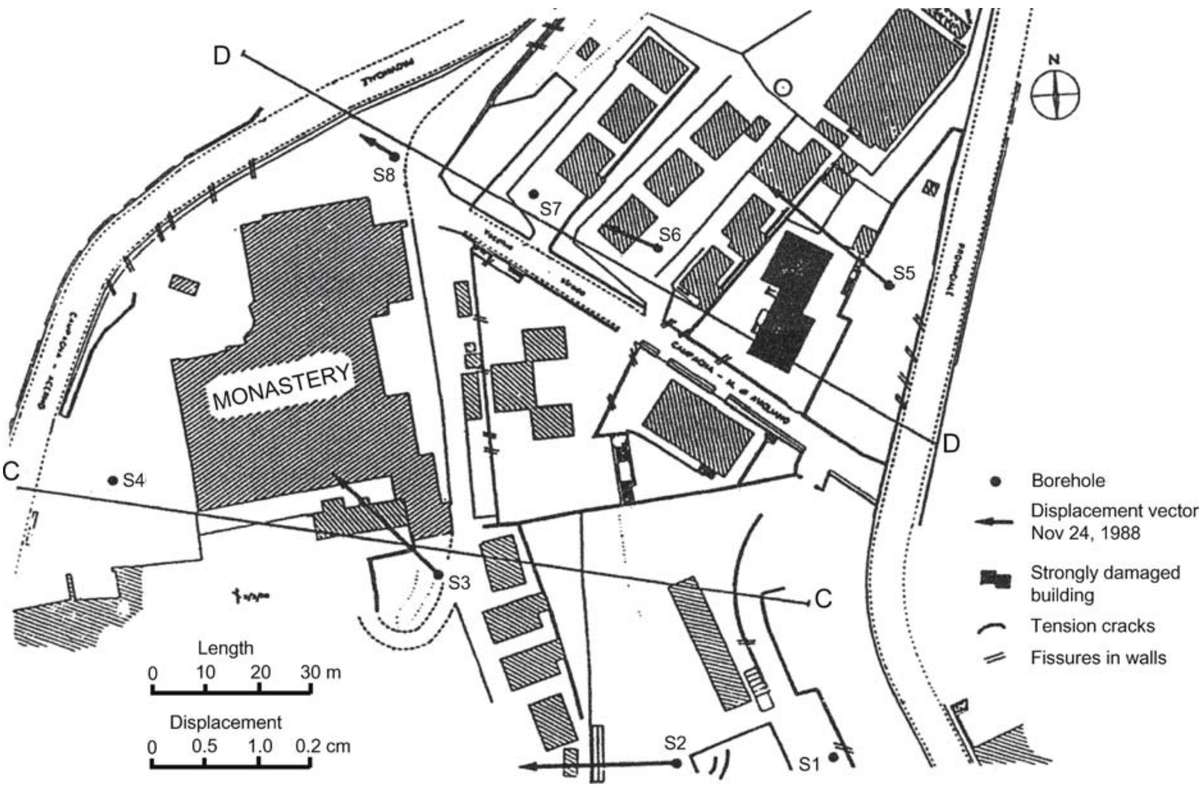


Fig. 3.4. A urban area experiencing slow movements and displacement vectors between 1987 and 1990 (after Picarelli and Simonelli 1991)

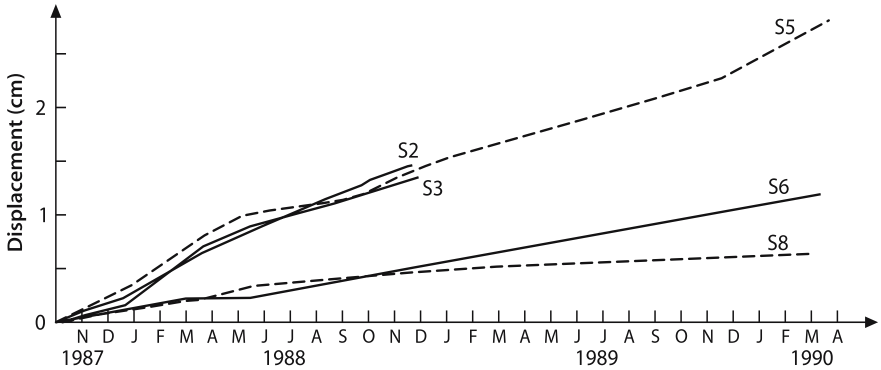


Fig. 3.5.
Cumulated displacements in the urban area of Fig. 3.4 (from Picarelli and Simonelli 1991)

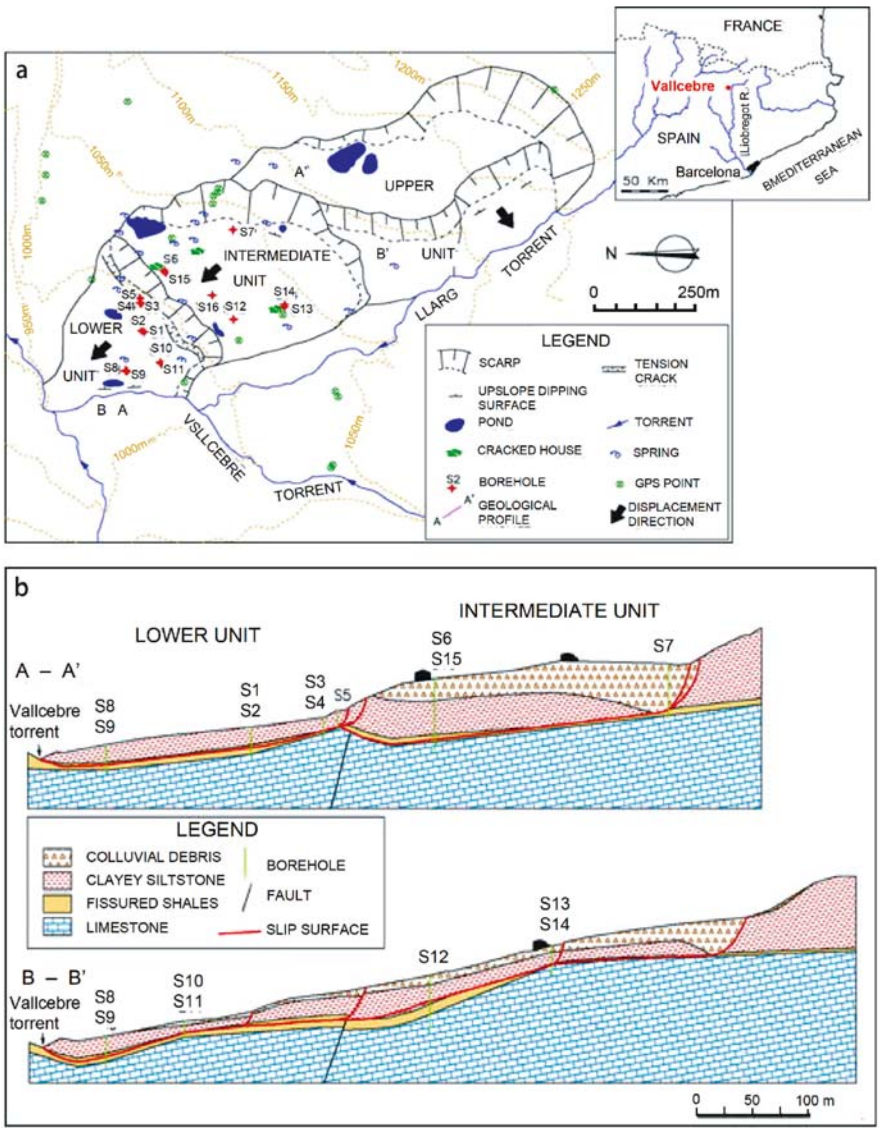


Fig. 3.6.
The Vallcebre slide (from Corominas et al. 2005): (**a**) plan, (**b**) longitudinal sections

with a standpipe piezometer and displacements at vertical S2 (Fig. 3.6a). It shows that in some periods of the year pore pressures remain constant, even though the landslide moves anyway. For instance, between April and June, 1997, the displacement rate was around 15 mm month^{-1}, pore pressure being constant, but a sudden increase of this, followed by a decrease to a value slightly higher than before, determined a doubling of velocity. The Authors

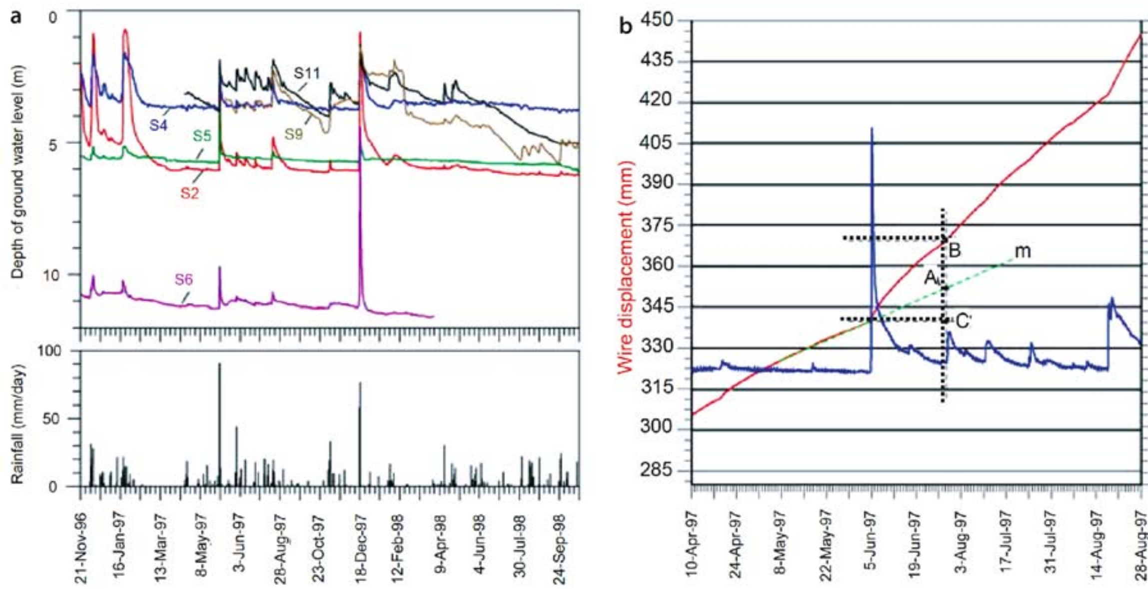
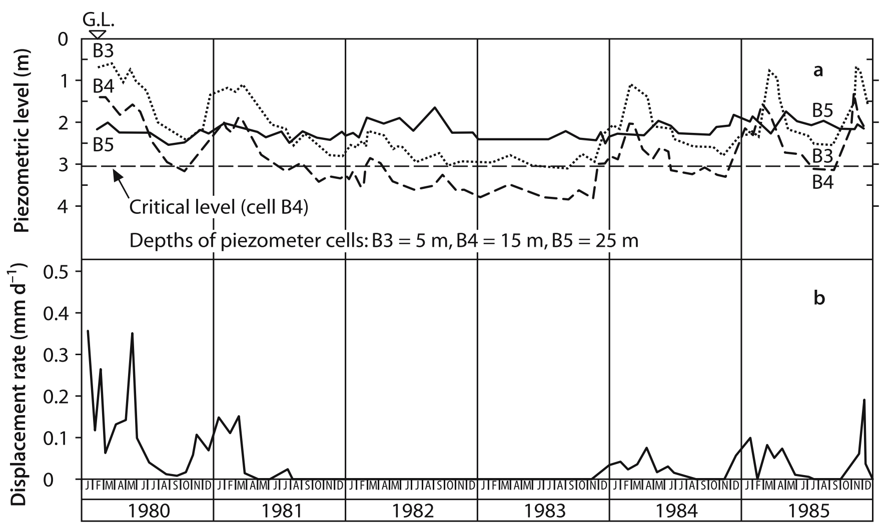


Fig. 3.7. Groundwater level, rainfall and displacements at vertical S2, Vallcebre landslide (from Corominas et al. 2005)

Fig. 3.8. Fosso San Martino slide. **a** Ground-water levels; **b** displacements of the ground surface (from Bertini et al. 1986)



assume that the increase of the displacement rate is contrasted by increase of the residual shear strength due to rate effects. This idea is shared by other Authors, who report similar considerations about the behavior of soils involved in slow movements (Vulliet 1986; Angeli et al. 1996; Vulliet and Hutter 1998).

A more complex viscous soil behavior is hypothesised by Bertini et al. (1986) for the Fosso San Martino slide, in fine grained colluvial soils (Fig. 3.3a). Figure 3.8 reports some results of monitoring covering a period of 6 years. Movement is directly correlated to rainfall and to consequent pore pressures changes, starting once the mobilized shear strength along the slip surface is about 95% of the residual value. When pore pressure decreases below a threshold value, as between summer, 1981, and fall, 1983, movement stops. In wetter periods, it re-starts following pore pressure rising. As remarked by other Authors (Cartier and Pouget 1988; Corominas et al. 2005), the displacement rate increases as the pore pressure increases, but the relationship between pore pressure and displacement rate is non-linear. In addition, Bertini et al. show that, for the same value of pore pressure, the velocity displayed during groundwater rising is higher than during lowering. Hence, two different relationships can be established between pore pressure and displacement rate; for the same reason, reactivation and arrest occur for different pore pressure thresholds (Fig. 3.9a). The Authors attribute such a behavior to the viscous behavior of soil along the slip surface, accounting for both primary and secondary creep. In fact, they argue that:

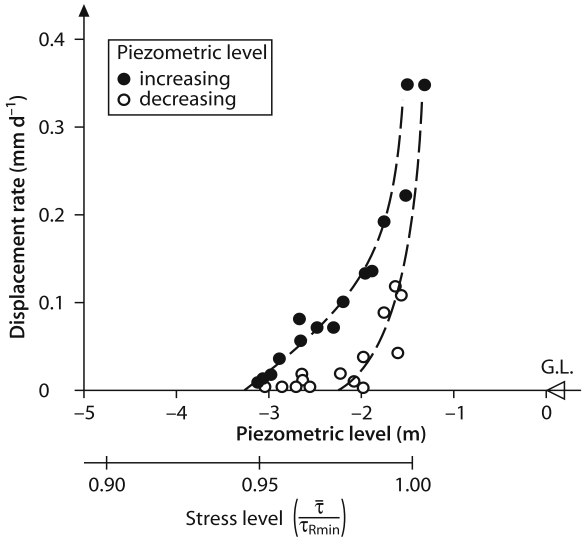


Fig. 3.9. Relationship between displacement rate and water level (from Bertini et al. 1986)

1. during the stage of pore pressure rising, the displacement rate is the result of two contrasting phenomena: increase of the rate of primary creep, which is associated with increasing stress level, and decrease of the rate of steady-state creep, which depends on time;
2. in the stage of pore pressure decrease, the two effects have the same negative sign and the resulting velocity is smaller.

The existence of different thresholds for increasing and decreasing pore pressure is stressed by other Authors too, but sometimes their observations are opposite, suggesting different interpretations. Moore and Brunsden (1996), as Bertini et al., find that the threshold pore pressure required to reactivate the Worbarrow Bay mudslide, Dorset, is lower than the one required to stop movement, but they explain this apparent inconsistency with a change in pore water chemistry during periods of rest. They also note that each reactivation requires a pore pressure larger than previous reactivations. Differently, in the case of the Alverà mudslide, Angeli et al. (1996) note the existence of two different thresholds corresponding to reactivation and arrest, but in this case the lowest one corresponds to arrest. They assume that the higher threshold required for reactivation is due to some increase of the residual strength occurring during the period of rest.

A non linear relationship between pore pressure and displacement rate has been found by Mandolini and Urciuoli too (1999) for the Miscano mudslide in softened clay shales (Fig. 3.10). According to Fig. 3.10b, movement starts when the groundwater table is located at a depth around 2 m: in this condition, the mobilized friction angle is about 13°. A full shear strength mobilization, corresponding to a high displacement rate (tending to infinite),

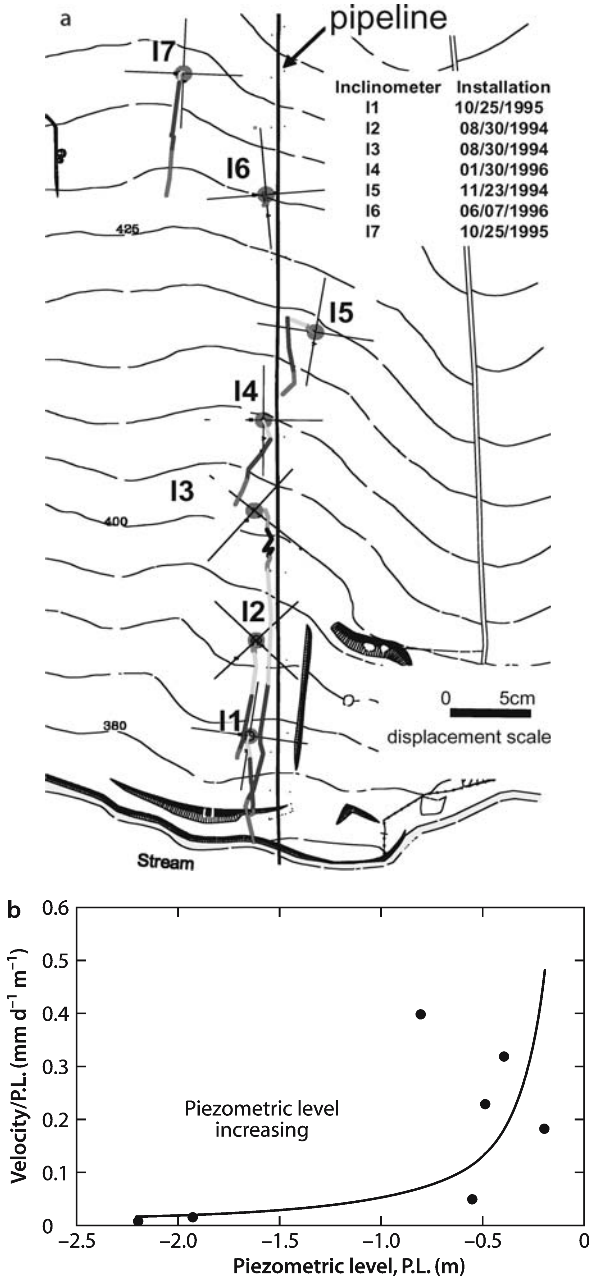


Fig. 3.10. Displacements (**a**) of the Miscano mudslide (from Picarelli et al. 1999), and relationship (**b**) between velocity divided by groundwater level and groundwater level (from Mandolini and Urciuoli 1999)

seems to occur once the groundwater table reaches the ground surface, i.e. for a mobilized friction angle of 18°. The ratio between the first stress level (creep threshold) and the second one (full shear strength mobilization) is then only 70%, which seems to be very low when compared to data provided by Bertini et al. (95%) and to laboratory experiences (Boucek and Pardo-Praga 1984). This raises some doubts about the real mechanisms of movement that will be discussed below.

Every assumption has a conceptual model behind. Referring to slopes, many researchers have the rigid-plastic constitutive law in their mind. According to this model, the landslide body should move along the slip surface as a block, with a velocity depending on the constitutive law of the slip surface.

Assuming that the residual friction angle is the one mobilized just at the beginning of movement (13° in the case of the Miscano mudslide), a conceptual model capable to justify the landslide behavior resembles the one proposed by Corominas et al., which assumes a rate-depending residual strength along the slip surface. Therefore, once a unbalanced force (difference between the driving and the resisting force due to basal friction) establishes, the soil mass tends to accelerate, but this is contrasted by growing resistance along the slip surface: the net effect depends on both the magnitude of the unbalanced force and the rate of shear strength increase. Naturally, the dependence of the residual strength on the displacement rate, thus the validity of proposed model, should be checked by laboratory tests. Unfortunately, available data do not help very much. Through tests on clay, Kenney (1967) and Skempton (1985) remark a negligible or little increase of the residual shear strength with the displacement rate. Furthermore, no influence is noticed by Hungr and Morgenstern (1983) for sand, while data collected by Picarelli and Urciuoli (1988) and by Tika et al. (1996) through laboratory tests on clay, are more problematic, suggesting that in some cases excess pore pressure (either negative or positive) can be induced by fast movement. This adds a complication in the model, which should account for drainage conditions. Furthermore, in this case viscosity is not necessary to explain the slope behavior, because increasing shear strength would essentially be a consequence of induced negative pore pressure.

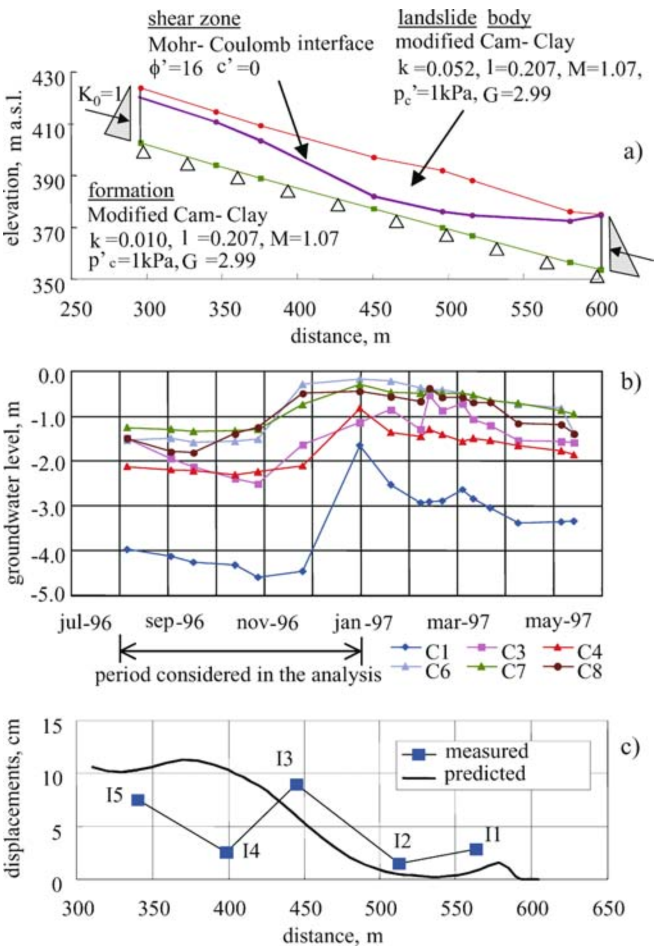


Fig. 3.11. Analysis of displacements of the Miscano mudslide. **a** Adopted model; **b** measured groundwater levels; **c** measured and calculated displacements (from Picarelli et al. 1999)

If we assume that the true residual friction angle is the one which is mobilized when the displacement rate attains an "infinite" value (18° in the case of Miscano mudslide, when the piezometric level reaches the ground surface), the model to be adopted for interpretation of the landslide behavior should include pre-failure creep (as assumed by Bertini et al.), which is able to justify movements occurring for pore pressures less than the critical value required to provoke a full shear strength mobilization. Therefore, once again, viscosity of the slip surface could play a fundamental role, but in this case, prior to mobilization of the shear strength, which does not necessarily depend on the rate of movement. In this last case, a full shear strength mobilization would lead to catastrophic consequences.

The two models of slope behavior could be easily unified, but both consider the soil mass as a block moving along a viscous interface. Reality is more complex since soil is deformable and both initial and induced state of stress are not uniform in the slope: as a consequence, internal strains may play a significant role on landslide behavior. This last point is stressed by Picarelli et al. (2004) who argue that in many cases the landslide body is not entirely mobilized by pore pressure fluctuations, which cause internal strains associated with slipping along only part of the sliding surface.

Figure 3.10a shows that the displacement field of the Miscano mudslide at a given time is not uniform along the slope, with decreasing values downward. This can be explained accounting for pore pressure changes in the wet season (Fig. 3.11a), which are faster upslope than downslope, causing a temporary mobilization of the uppermost part of the landslide only, while the local shear stress mobilized at the base of the lowermost remains lower than the shear strength. This causes a compression of the landslide body induced by the overstress caused by restraint imposed by the lowermost part of it. Such a phenomenon has been simulated by the FEM. The slip surface has been characterised with elastic-perfectly plastic elements along which is operative the residual shear strength, while the landslide body has been modeled with an elastic-plastic constitutive law (Cam-Clay) adopting the parameters reported in Fig. 3.11b. Slope displacements have been calculated for the same pore pressure increase which has been measured in site, as in Fig. 3.11a. The results of the analysis are comparable to reality (Fig. 3.11b): in particular, significant movement develop even though the safety factor of the lowermost part of the mudslide body remains higher than one. Therefore, this simple model confirms that movement can be essentially caused by internal strains more than by full mobilization of the landslide body.

The same mechanism can also explain the non linear increase of displacement with increasing pore pressure, discussed above (Figs. 3.9a and 3.10b): in fact, this result depends on the increase of the length of the part of the landslide body which is mobilized by pore pressure changes (Picarelli and Russo 2004). In addition, since pore pressures vary with time, movement too varies with time, so that soil viscosity is not necessary to explain the rate-depending behavior of slide.

Naturally, previous simplified interpretation of slope behavior does not exclude mobilization of a viscous component of displacement within the landslide body and along the slip surface, which can be easily incorporated in the model (Vulliet 1986).

Similar effects on slope behavior, i.e. displacement caused by internal deformation of the landslide body, can be provoked by uniform pore pressure increase, if movement is locally constrained by:

- variation of the geometry of the landslide body, because of increasing thickness and/or decreasing slope of the slip surface moving from upward to downward;
- local change of the width of the landslide (3D effect), as in mudslides which present a neck in between the alimentation zone and the main track;
- any change of the shear strength parameters along the slip surface.

These considerations offer alternative scenarios about slope behavior. In fact, at least in the case of long landslides, even very small strains, when integrated to the entire length of the landslide, could give rise to a significant component of displacement. As an extreme case, for a 100 m long landslide having a free upper boundary (the crown) and a fixed lower boundary (the toe), an average compressive strain of 10^{-2}% due to pore pressure rising, can trigger a displacement of the uppermost part of the landslide body of 1 cm, solely due to compression. It is worth noting that, in stiff clay, a strain equal to 10^{-2}% can be provoked by increase of pore pressure in the order of 1 kPa, i.e. by a rising of the water table in the order of 10 cm.

Described mechanisms of slope movement, which might be governed by constraints due to boundary conditions or non uniform geometric and mechanical parameters of soil, have been investigated by Russo (1997) and discussed by Picarelli and Russo (2004) using both an elastic-plastic and an elastic-viscous-plastic constitutive law of soil. In particular, the use of an elastic-viscous-plastic constitutive law allows to perform more sophisticated analyses. As an example, a slope model as the one used to interpret movement of the active Miscano mudslide, can account for the effects of relaxation occurring in any phase of pore pressures decrease (dry season), when the compressive state of stress stored in the landslide body in previous wet season (when pore pressure increases) progressively disappears as a consequence of stress relaxation. It is worth noting that the magnitude of deformation occurring in the wet season depends on viscous soil properties and on time span prior to the new pore pressure increase. Such a deformation history governs the magnitude of pore pressure (threshold) required to activate new movements, which decreases as the magnitude of relaxation increases.

Similar considerations could be made about opposite mechanisms inducing extensive strains in the slope.

All these remarks about the possible role of soil deformability on slope movement appear consistent with data provided by some Authors (Wilson 1969; Jappelli et al. 1977; Nakamura 1984; Pouget 1996).

3.3.2 Active Mudslides

The typical flow-like style of mudslides is revealed just at failure or immediately after failure, when the landslide body displays a high mobility, spreading along the slope and filling pre-existing tracks. In this stage the mobilized soil mass moves quite rapidly over the ground surface, eroding and incorporating the top soil and the upper part of the underlying deposit: as a consequence, it can experience some subsidence (Hutchinson 1970; Corominas 1995). The described mechanism of movement implies that the base of the mudslide body is subjected to high shear stresses which are presumably responsible for intense remoulding of quite a thick basal slice of soil. In stages following first post-failure movements, the soil mass decelerates assuming a slide style, but the occurrence of previous flow-like mechanisms can be still easily recognized from the assumed morphology of the landslide body. This typically displays an alimentation zone, a track and a fan-shaped accumulation zone. In this last stage of slope deformation prior to definitive arrest, a well defined persistent slip surface, which could not survive to movements occurring in the flow-like stage, presumably governs the mudslide behavior.

Mudslides are bounded by a main scarp that dominates a depletion area (Fig. 3.12). This alimentation zone feeds the landslide body with "fresh" soil masses (surges),

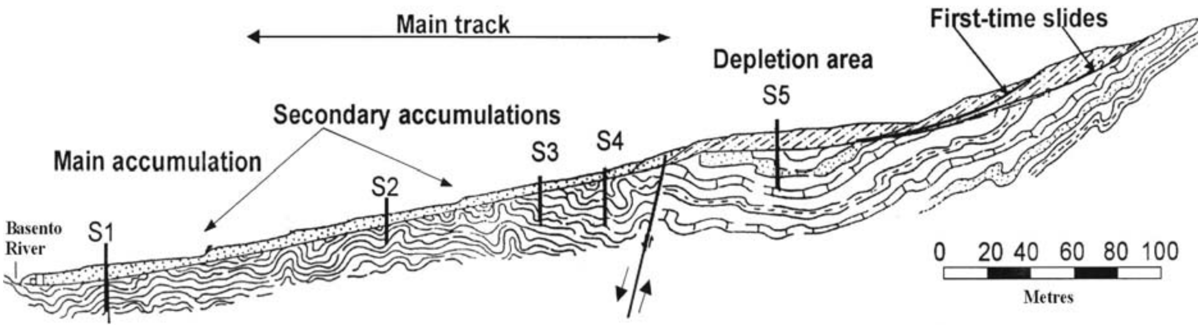


Fig. 3.12. The Brindisi di Montagna mudslide (from Cotecchia et al. 1984)

mobilized by local failures of the scarp, which are conveyed into the main track, causing periodical flow-like reactivations of movement. Further alimentation is provided by slips occurring along the flanks of the track. Surges form secondary accumulations along the track, but can reach the toe of the slope, thickening the accumulation zone. The presence of independent active or temporarily quiescent landslide bodies within the area occupied by the mudslide body is revealed by inclinometer profiles which show more shear zones along the same vertical.

To sum up, the main features of mudslides concern:

- their morphology, which reveals the mechanism of movement which characterises the initial post-failure flow-like phase: when channellized within a pre-existing track, long-term movements are restrained and governed by a complex 3D condition;
- the features of the soil mass, which is highly remoulded as a consequence of high deformations and softening experienced during first movements (Hutchinson 1988; Picarelli 1993);
- the thickness and fabric of the shear zone: while in the case of slides this is quite thin (generally a few centimeters) and displays a set of minor shears, in the case of mudslides it is rather thick (up to one meter) and fully remoulded (Picarelli et al. 2005): often the structure of the parent formation is completely obliterated.

Several Authors (Hutchinson and Bandhari 1971; Picarelli 1988;) argue that the initial flow-like style is a consequence of building-up of positive excess pore pressures. Pellegrino et al. (2004) discuss the main mechanisms which are supposed to be responsible for triggering of excess pore pressures. They include:

- rupture itself;
- seismic loading;
- accumulation of debris over a pre-existing mudslide body as a consequence of secondary failures or of rapid erosion along the main scarp or flanks of the track;
- loading caused by surges traveling over the mudslide body.

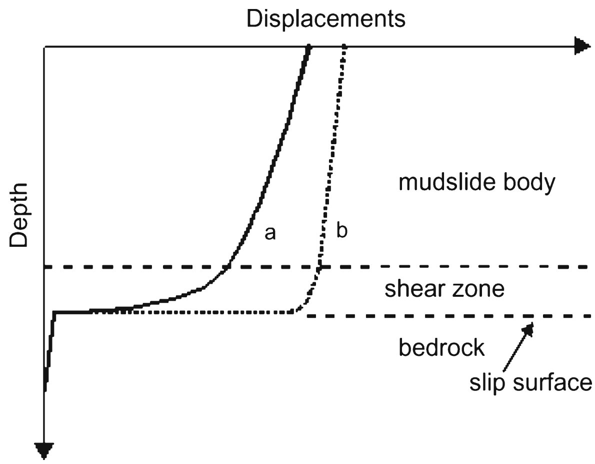


Fig. 3.13. Typical displacement profiles of a mudslide. **a** In the first stage of movement (flow-like style); **b** in the last stage of movement (slide style) (from Comegna et al. 2005)

Dissipation of excess pore pressures leads to a decrease of the rate of movement while the soil mass takes the features of a slide: in this stage movement localizes along a slip surface, while shear strains within the landslide body become smaller than in previous flow-like stage (Fig. 3.13). However, some data suggest that even this sliding phase can be characterised by development of moderate excess pore pressures because of internal stress changes.

Figure 3.14 shows pore pressures measured within the Masseria Marino mudslide at a depth of about 3 m. Figure 3.13a concerns a quiescent zone and Fig. 3.13b an active zone subjected to slow movement. While in the quiescent zone pore pressures display smooth seasonal fluctuations, quite in a good agreement with the results of numerical simulations carried out using pluviometer data and adopting the permeability of soil obtained by in situ and laboratory tests, in the most active zone pore pressures display rapid fluctuations which do not seem to follow a logic course (Comegna et al. 2004). The most reliable explanation is that any local soil deformation can trigger excess pore pressure: where deformation is by compression, induced pore pressure is positive.

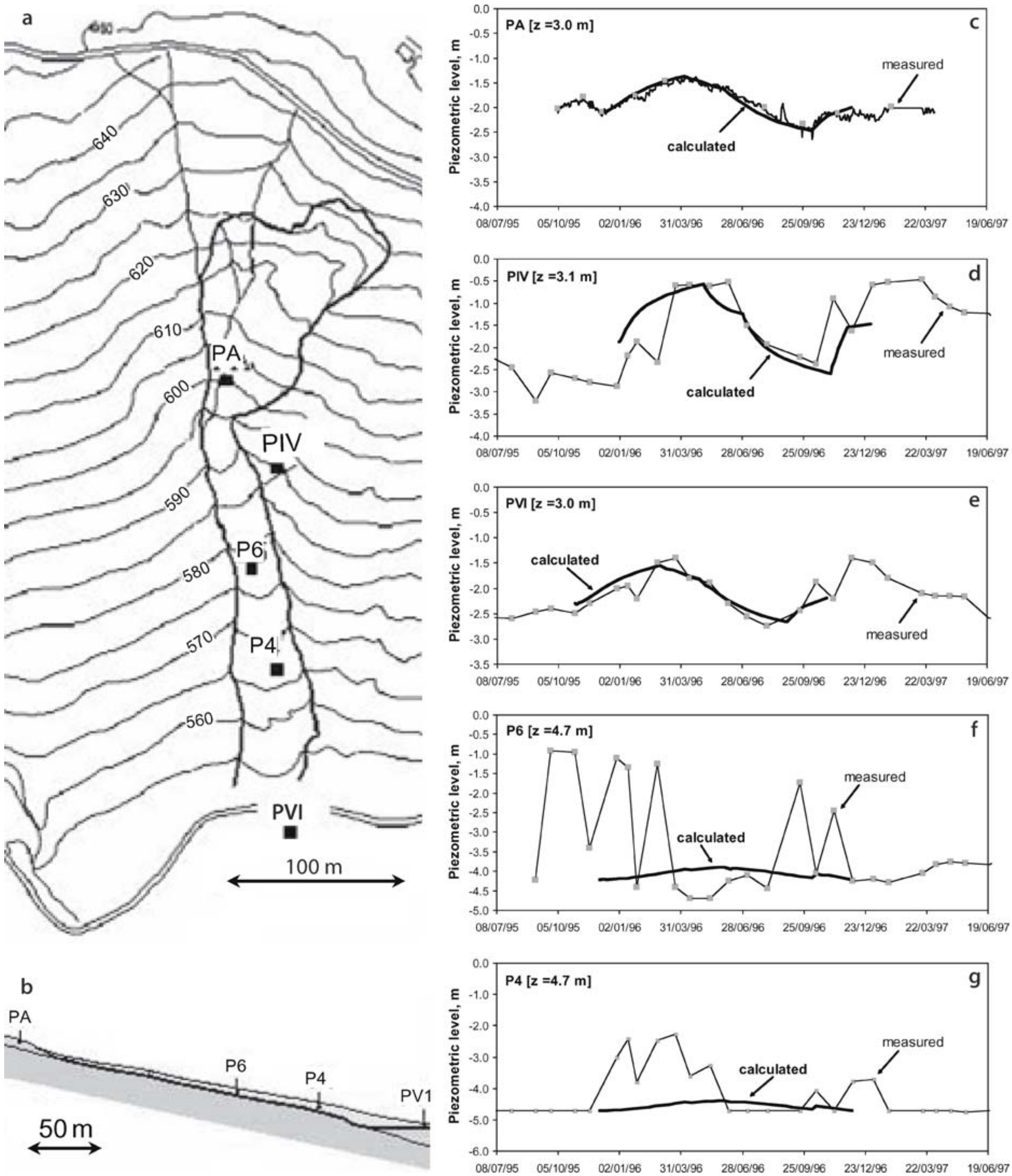


Fig. 3.14. Masseria Marino mudslide: location of the piezometers considered in the analysis (**a, b**); calculated and measured water levels in quiescent (**c, d, e**) and active zones (**f, g**) of the mudslide (after Comegna et al. 2005)

Such a mechanism has been checked through a simplified analysis. The mudslide body has been modeled with a non linear elastic constitutive law characterized by an isotropic yielding law ("Soft-Soil Model" present in the library of the code PLAXIS). In addition, has been considered a basal 1 m thick softer shear zone. The soil parameters adopted in the analysis have been obtained by the best fitting of laboratory tests (Comegna 2005): their values are reported in Table 3.1.

The sliding surface located at the base of the shear zone, has been simulated by interface elements along which is operative a residual friction angle of 13°. The analysis

Table 3.1.
Soil parameters used to simulate the behaviour of the Masseria Marino mudslide

	γ_{sat} (kN m^{-3})	λ	k	e_0	ν	K_0^{NC}	O.C.R.	c' (kPa)	ϕ' (°)	K (m s^{-1})
Mudslide	20	0.057	0.026	0.59	0.35	0.58	3	8	25	10^{-9}
Shear zone	20	0.103	0.038	0.68	0.35	0.61	1	4	23	10^{-9}

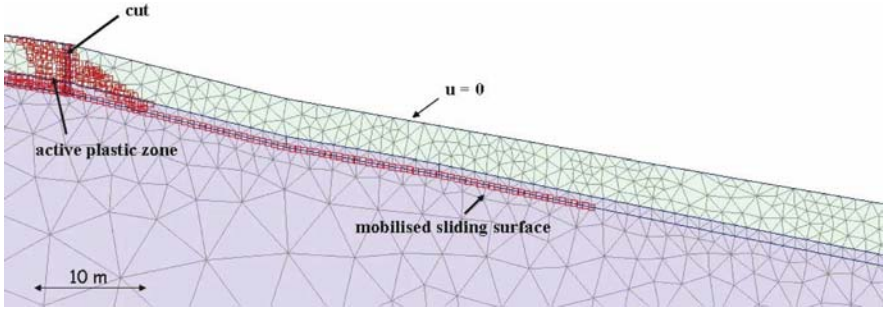


Fig. 3.15.
Plastic active zone formed as a consequence of pore pressure rising (from Comegna et al. 2007)

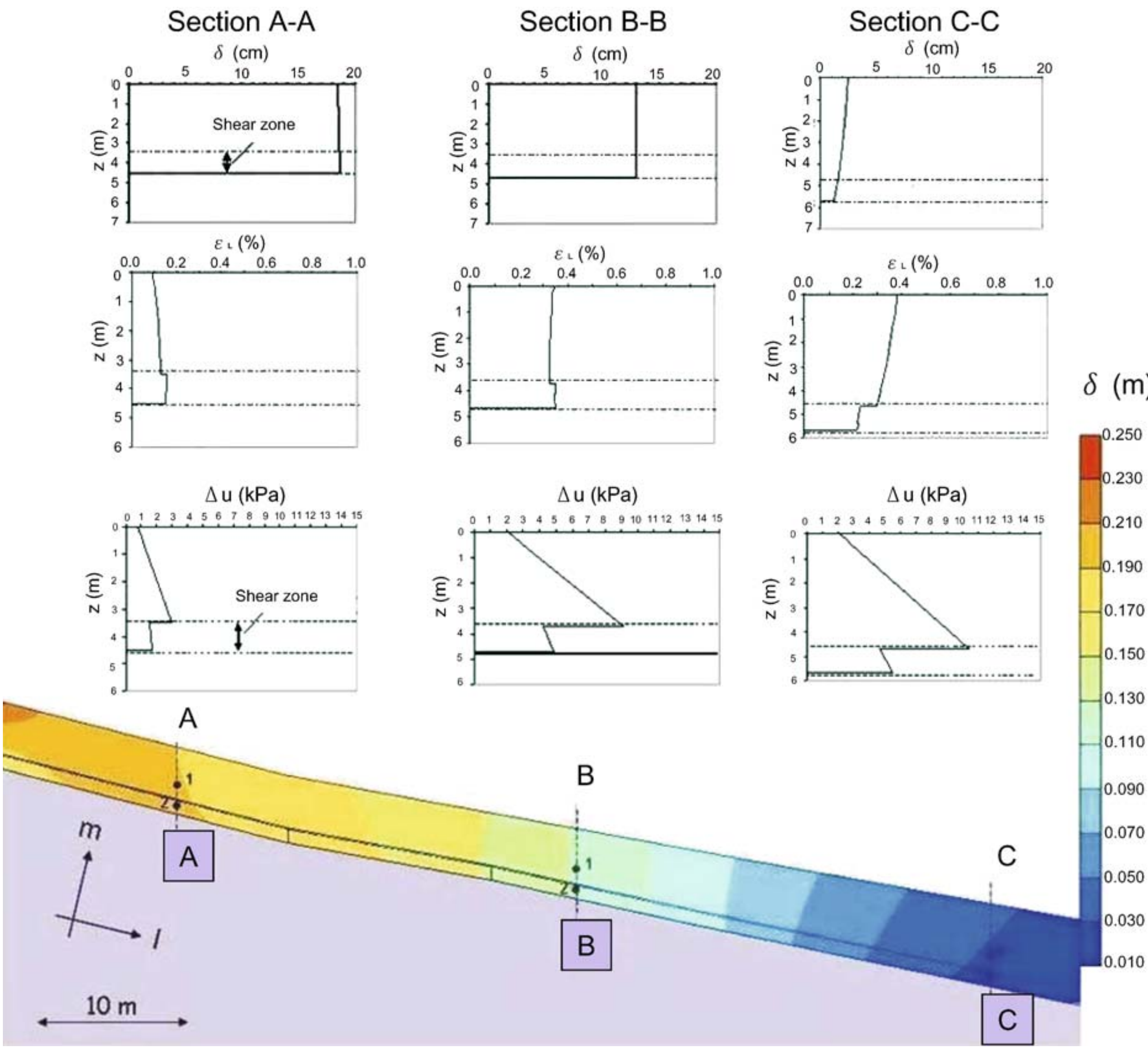


Fig. 3.16. Displacements, longitudinal strains and excess pore pressures at three sections of the mudslide (from Comegna et al. 2007)

simulates the pore pressure rising induced by rainfall, by imposing a water film along the ground surface (pore pressure equal to zero). During this phase, only part of the slip surface is mobilized, but the length of the mobilized part propagates downward as pore pressure increases. As a result, the active mudslide body slides downward, compressing the still stable part located ahead; at the same time, an active zone forms immediately upslope (Fig. 3.15). In order to simulate local soil rupture in the active zone, a vertical cut has then been imposed. Assuming that the deformation of the landslide body after cracking is fast enough to trigger excess pore pressures, the concerned stage of analysis has been performed assuming "short-term" (undrained) conditions. As a consequence of cracking, the mobilized part of the sliding surface further propagates downward while pore pressures rise due to

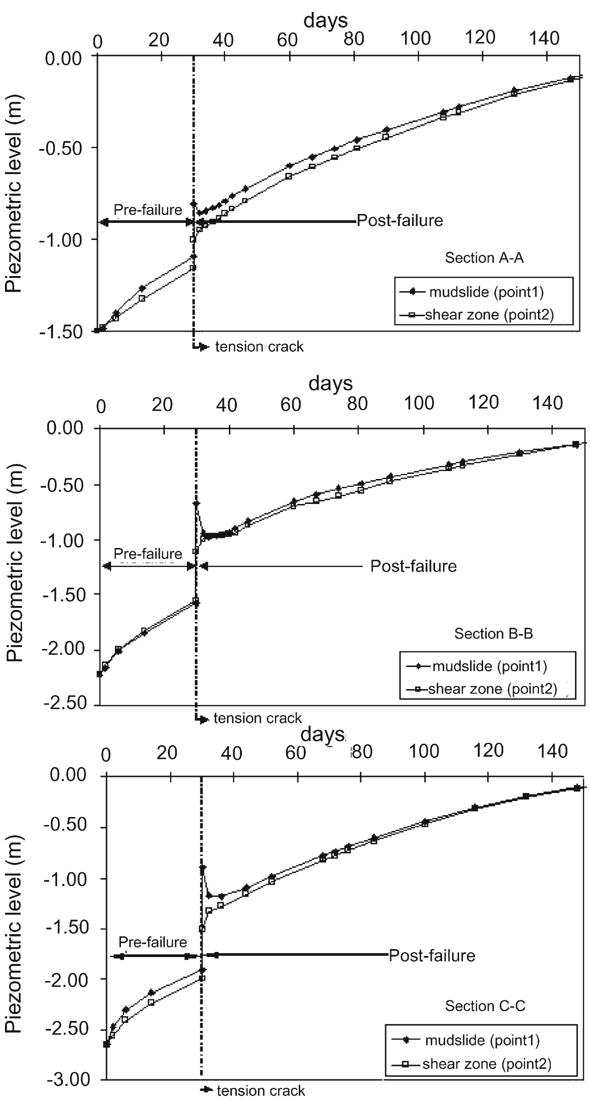


Fig. 3.17. Pore pressure evolution at two points within the mudslide body (from Comegna et al. 2007)

undrained compression. However, the longitudinal strain ε_L is not uniform because of the different stiffness of mudslide body and shear zone; this difference in stiffness is also responsible for lower excess pore pressure in the shear zone. Longitudinal strains, pore pressures and displacements at three different sections of the slope are reported in Fig. 3.16.

During the following stage of analysis, excess pore pressures are allowed to equalize. This phase is characterized by the overlap of two opposite phenomena: dissipation of excess pore pressure and continuing rising of water table due to infiltration. Figure 3.17 reports the evolution of pore pressure at two points, one located in the shear zone and the other one in the mudslide body, in three sections of the landslide. Because of the high gradient of the piezometer head closely around the two points, excess pore pressures rapidly dissipate: a decrease in the mudslide body is associated with increase in the shear zone. Such a result could explain the anomalous drops of pore pressures monitored in the Masseria Marino mudslide (Fig. 3.14b).

3.4 Consideration about the Mechanics of Active Lateral Spreads

Lateral spreads represent a special category of slow active landslides in clay. They are so slow to be recognized only through sophisticated monitoring or from continuing damage to old structures. According to literature, lateral spreads can be triggered by tectonic uplift, by removal of lateral support following glacial retreat or by river erosion (Radbruch-Hall et al. 1976).

An interesting example is movement which involves the Monte Verna hill (Fig. 3.18) where San Francis retired in the last years of his life (Canuti et al. 1990). The hill is constituted by jointed massive calcarenites and bedded sandstones which rest on a deposit of highly fissured sheared clay shales. It is subjected to lateral movements, essentially driven by squeezing out of clay shales. Movement causes opening of vertical joints in the rock slab.

A schematic description of the mechanisms which govern spreading induced by valley formation is shown in Fig. 3.19. This concerns the case of a rock slab resting on clay, as in Fig. 3.18. Erosion starts developing along fractures of the slab (stages 1 and 2). The stress relief caused by unloading provokes rising of the valley floor; as a consequence, the clay deforms forming an anticline (stages 3 and 4). Further deepening of the valley reaches the buried top of clay, which squeezes out bringing about lateral deformation of slopes. As a consequence of imposed drag forces, a shear zone forms at the contact between clay and rock bed, and the slab is subjected to fracturing as a consequence of tensile stresses imposed by shear (stage 5). Induced movement causes opening of

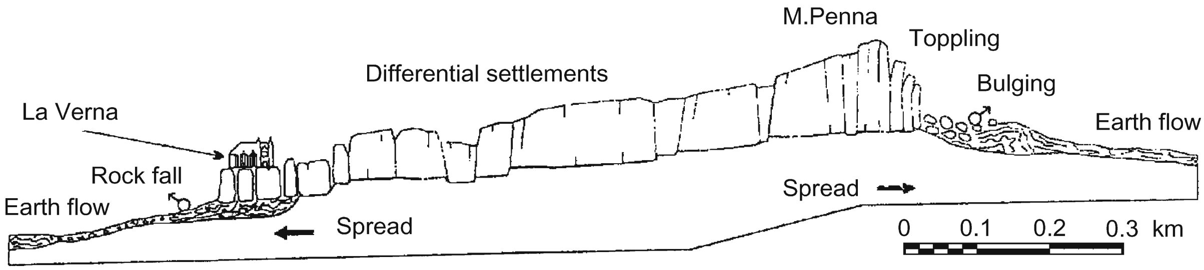


Fig. 3.18. Deformation phenomena of the Monte Verna mountain (from Canuti et al. 1990).

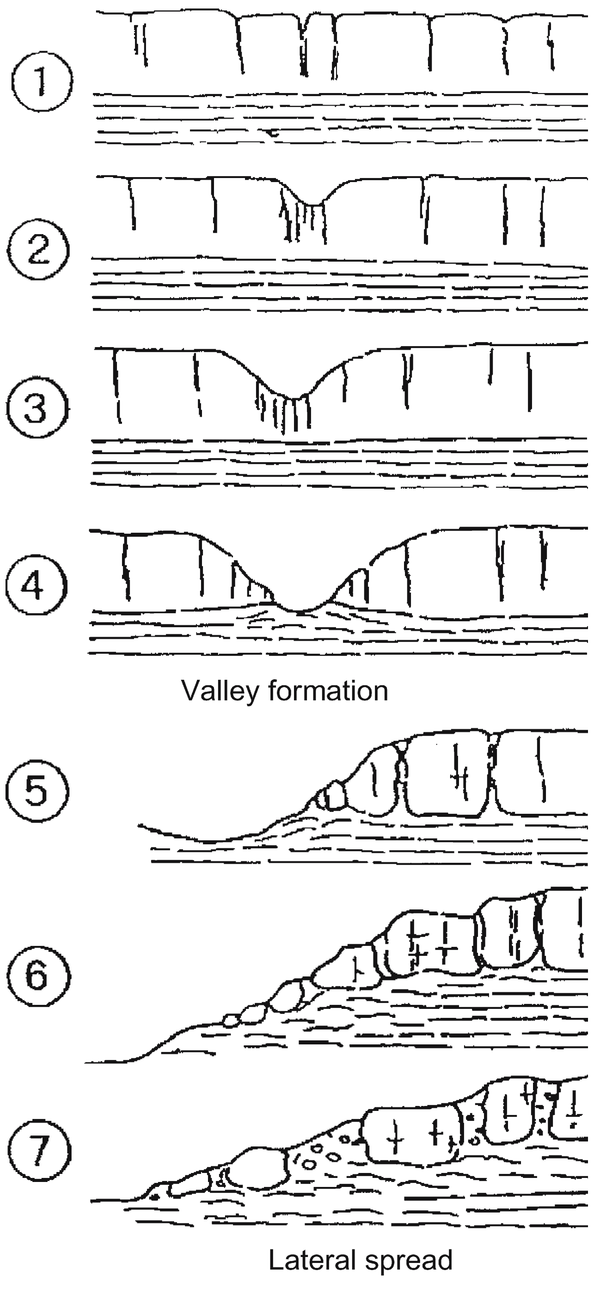


Fig. 3.19. Effects of river erosion on the deformation pattern of a rock slab resting on clay (from Pasek 1974).

vertical joints which can be filled by clay spread from the bottom. In addition, internal rock blocks subside, while lateral blocks tilt, forming a trench at the top of the slab. Further movements trigger landslides along the boundaries of the slab contributing to their complete split up (stages 6 and 7).

The geological phenomena which govern such a process, even when very fast in geological terms, are extremely slow in the human perception, thus consequent movements are extremely slow. However, in case of very thick clay deposits, unloading due to erosion can be slower than the time required for dissipation of induced negative excess pore pressures (Neuzil 1993), causing a delay of associated deformations. In seismically active areas, earthquakes can add their effects to erosion and creep, producing acceleration of deformation.

Other cases of spreads can be found in the Italian scientific literature (Crescenti et al. 1994; Fenelli et al. 1996; Chelli et al. 2006). Several of them concern towns located on the top of hills and mountains in the geologically active area of Apennines. A well known example is the Orvieto hill, in the Tiber valley (Lembo-Fazio et al. 1984; Tommasi et al. 2006), which is constituted by a tuff slab resting on stiff clay. A similar situation is described by Tommasi and Rotonda (1995) who discuss the deformations of the calcareous San Leo cliff, on top of which stands a Renaissance's Castle where the famous Cagliostro was hold in jail until his death.

An interesting case is the Bisaccia spread, in Southern Italy (Di Nocera et al. 1995). Bisaccia rises on a slightly cemented conglomerate slab with thickness exceeding 100 m, which rests on intensely fissured highly plastic clay shales. In the last 300 000 years, the area has been deeply eroded by downcutting along two parallel faults (Fig. 3.20), through a geological process very similar to the one described in Fig. 3.19. Concentrated erosion left an elongated hill in the middle. Presently, the floor of the two valleys is lower than the bed of the conglomerate slab over which stands the town. The slab is divided into large blocks separated by vertical cracks. Its boundaries are subjected to landslides involving both upper conglomerates and lower clay shales. Some movements involve both conglomerate blocks and clay at the foot (slides); others involve only clay and generally display a flow-like style (Picarelli et al. 2006). The top of the hill presents a series of morphological steps. In the past, these have been smoothed with man-made ground or bridged with stairs. In addition, buildings show fissures and cracks; old reparations reveal the occurring of a process of general deformation.

The area is shaken by earthquakes having a return time of some tens of years. In the last century, two strong earthquakes occurred in 1930 and in 1980 (the last one characterized by a Magnitude $M = 6.3$). Seismic events systematically cause cracks in the masonry of old buildings and in pavements. The main effects of the 1930 and 1980 earthquakes were a series of fractures on pavements. Most of the cracks generated by the two events coincide, and are located along natural morphological steps; in addition, buildings most severely damaged by the 1980 quake are located along such alignments (Fenelli 1986). This means that the damages were mostly caused by opening of fractures.

The area has been systematically investigated since 1981. The first campaigns included field and laboratory investigations which allowed to obtain a complete mechanical characterization of clay shales (Picarelli et al. 2002). In 1985 and 1989, thus after the Irpinia earthquake

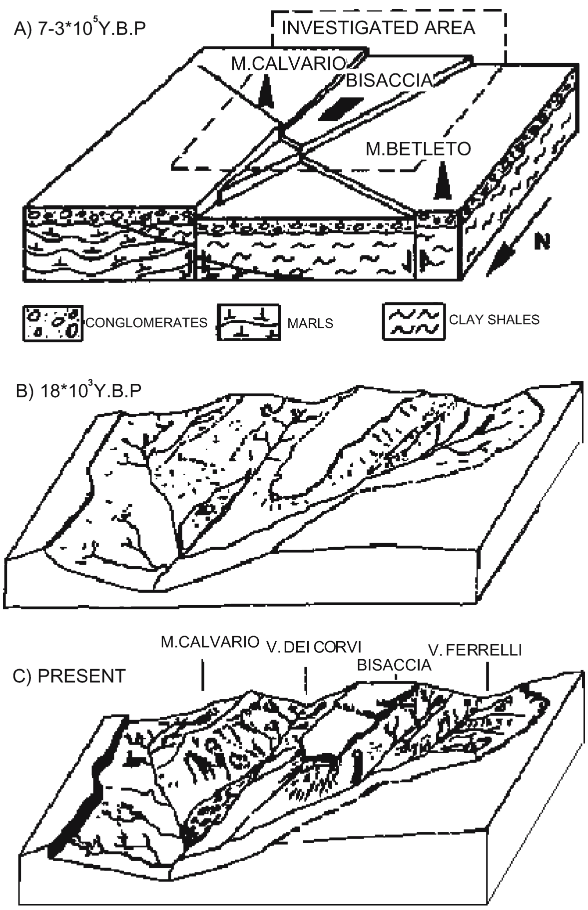


Fig. 3.20. The Bisaccia hill (from Fenelli and Picarelli 1990)

(1980), two verticals, one in the eastern valley and the other in the urban area, were instrumented with Casagrande and vibrating wire piezometers (Fenelli and Picarelli 1990; Di Nocera et al. 1995). Finally, on February, 1981, a number of benchmarks were installed in the town to monitor deformations induced by the quake. Monitoring covered more than seven years, until October, 1988.

Figure 3.21 reports pore pressures measured along the two instrumented verticals. Deficient values (locally negative, i.e. below the atmospheric pressure) have been measured below the bed of the eastern valley where unloading has its major effect. In contrast, pore pressures above their theoretical value have been measured just below the slab. However, accurate readings have shown a constant decrease in time.

These results have been interpreted accounting for the geological phenomena affecting the site. In fact, accounting for the thickness and low permeability of clay shales, erosion is quite fast in geological terms and is assumed to trigger negative excess pore pressures (Fenelli and Picarelli 1990). Di Nocera et al. (1995) show that swelling of clay can explain the geomorphological shape of the two valleys and the distribution of the water content in the subsoil; other consequences of erosion are cracking of the slab, squeezing out of clays and a shear zone recognized at the conglomerate-clay shale contact.

Formation of the hill has been simulated by FEM analyses (Picarelli and Urciuoli 1993; Di Nocera et al.

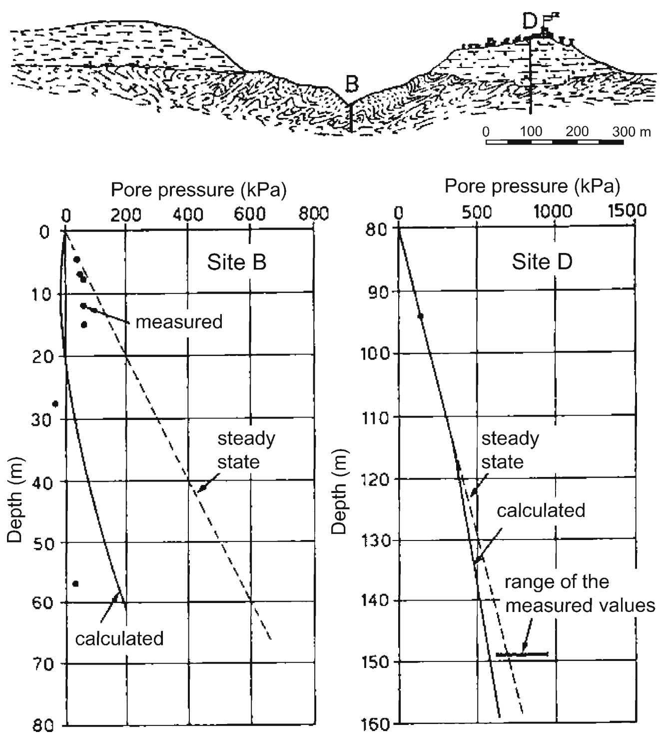


Fig. 3.21. Pore pressures measured in the Bisaccia area (from Fenelli and Picarelli 1990)

1995) assuming that erosion started 300 000 years ago, starting from an elevation corresponding to the present top of the hill. Figure 3.22 shows calculated vertical displacements of the valley floor, at the depth of its present elevation, and of the top of the hill. The figure shows that the valley floor is rising with a velocity of about 0.04 mm yr^{-1}, while the hill sinks in clay shales with a velocity of one order of magnitude less. Horizontal displacement at the contact between conglomerate and clay shale is shown in Fig. 3.23. Calculated movement is in the order of 0.2 mm per century; the differential displacement along the conglomerate-clay shale interface is responsible for formation of a shear zone. It is worth noting that the deformations calculated in the analysis not only depend on the assumed rate of erosion, but also on the rate of pore pressure equalization.

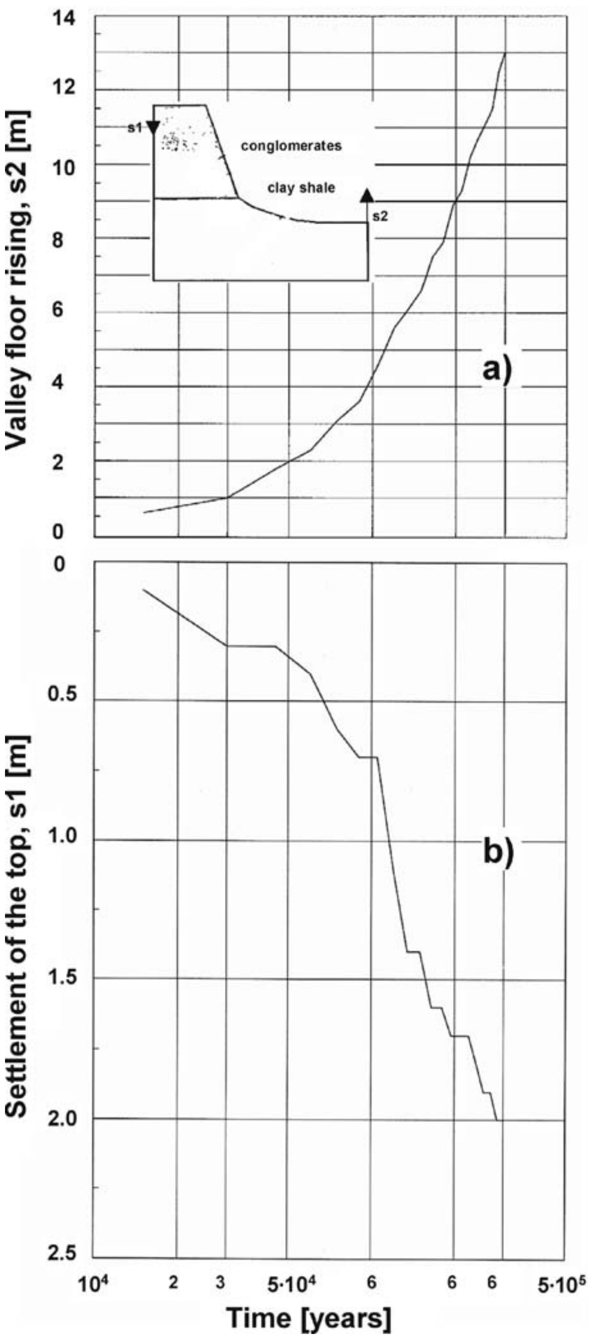


Fig. 3.22. Calculated vertical displacement of the valley floor (**a**) and (**b**) of the top of the hill (Picarelli and Russo 2004)

Previous data about the effects of the 1980 earthquake suggest that strong seismic events can accelerate movement of the hill. Figure 3.24 shows the settlements measured from February, 1981, to respectively February, 1982, and October, 1988, along a longitudinal section of the town (Fenelli et al. 1992). The location of major and minor fractures in the slab is represented by arrows. A profile of the hill along the same section is also shown in a magnified scale.

The figure shows that:

- until October, 1988, the entire urban area underwent a general subsidence with settlements increasing from South to North, where the hill is truncated by a steep high slope;
- settlements increased with time; eight years after the quake, the maximum value measured along this section was about 12 cm: the average settlement rate in the examined period was about 1 cm yr^{-1};
- the general subsidence of the hill is discontinuous, probably because of independent movements of conglomerate blocks; in particular, some blocks experience a vertical translation, others a rotation;
- the trend of vertical displacements is very similar to the morphological profile of the hill, that seems hence determined by cumulated deformations induced by erosion and by seismic events.

Despite the absence of further measurements after 1988 due to loss of benchmarks, the development of new

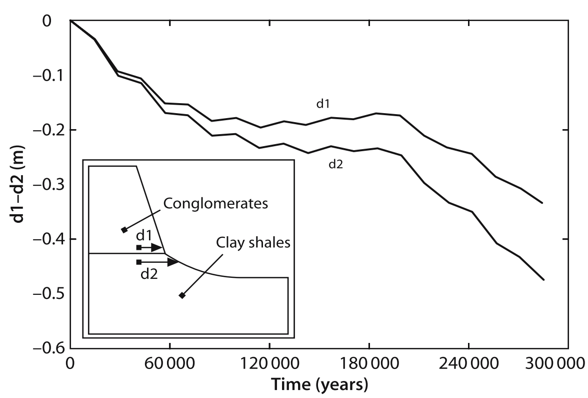


Fig. 3.23. Calculated horizontal displacement at the conglomerate-clay shale interface (Picarelli and Russo 2004)

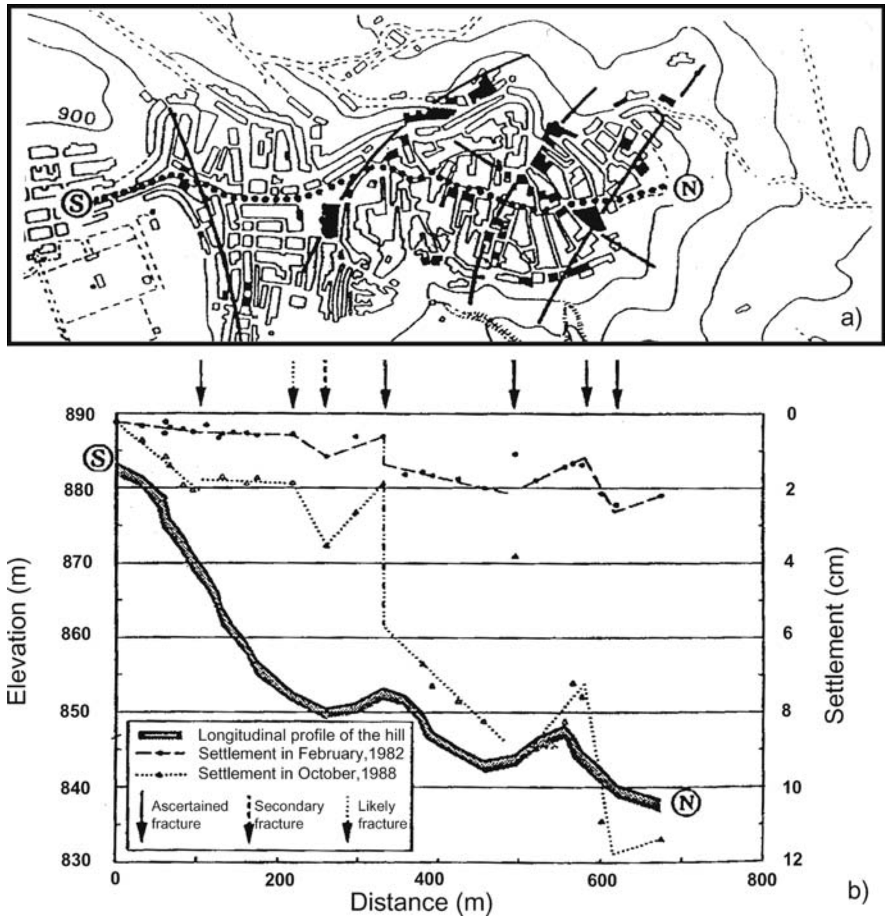


Fig. 3.24.
Vertical displacements measured after the 1980 earthquake and longitudinal profile of the hill (from Fenelli et al. 1992)

cracks in buildings demonstrate that movements were still continuing. Unfortunately, no data are available about horizontal displacements, but certainly they were not negligible.

More information are provided by numerical analyses (Olivares 1997; Lampitiello et al. 2001). The simulations have been carried out using seismogram data recorded in the town during the earthquake, deconvoluted to reproduce the seismic input motion at the top of bedrock. The analyses demonstrate that, in a large part of the clay shale deposit under the slab, the quake induced shear strains larger than the volumetric threshold, causing a local building up of positive excess pore pressures. The dissipation of these is responsible for delayed vertical displacements, as shown above.

Summing up, the hill is subjected to two different geological phenomena: erosion and earthquakes. Both contribute to slope movements. Erosion brings about a deepening of the valleys surrounding the hill. The superimposed effects of earthquakes cause a general subsidence of the slab, with sinking of the blocks into the clay shale deposit.

3.5 Conclusions

Typical slow active landslides in clay are translational slides and mudslides, but other types of movements, as spreads, also belong to the category of slow active landslides.

Active slides and mudslides advance along a pre-existing slip surface and are driven either by changes of boundary conditions, as those caused by rainfall or erosion, or by viscous deformations. The effects of pore pressure fluctuations prevail in the case of landslides of moderate thickness, while those of erosion and of creep can be significant in the case of deep-seated landslides. To understand the mechanics of slides and mudslides, the model of rigid-plastic body is not adequate. In fact, since any change of boundary condition provokes a non homogeneous variation of the stress field, internal deformation of the landslide body is a normal condition: since slow landslides present very small displacements, internal deformation is a significant component of movement. However, available data show that positive excess pore pressures self-generated by movement itself may play a not negligible role, at least in the case of mudslides; following movements is affected by processes of pore pressure equalization. The displacement rate is then a function of the rate of stress change, of the rate of excess pore pressure dissipation and of soil properties, possibly including a time-dependent component.

The mechanics of lateral spread is more complicated. In some cases, deformation induced by erosion and, possibly, by earthquakes, is prominent. However, once again, excess pore pressures can play a significant role, mainly when the movement involves a thick deposit of highly plastic clay: in such a case, excess pore pressures can be either negative, as a consequence of erosion, or positive due to seismic input. Pore pressure equalization (consolidation or swelling) may strongly affect soil movement.

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