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Mechanics of a tectonized soil slope: influence of boundary conditions and rainfalls

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ABSTRACT

The Vadoncello landslide, mobilized in December 1993 and still active, is the reactivation of a landslide which took place, within the highly tectonized soils of a turbiditic formation, after the 1980 Irpinia earthquake (Southern Italy), when it was dragged by the movements of a larger landslide body at the toe of the slope, the Serra dell'Acquara landslide. The Vadoncello landslide has been studied by means of the results of comprehensive investigations and monitoring carried out within a EEC funded research project, as well as by means of successive data. Consequently the complex geological model of the slope has been defined, in which the chaotic successions of soil and rock strata are grouped into soil complexes, the location of different landslide bodies is identified within the slope (the 1993-'95 Vadoncello landslide, the 1980 Vadoncello landslide and the 1980 Serra dell'Acquara landslide bodies) and two hydrogeological complexes are recognized. The soil mechanical properties are shown to be very poor, the deep soils being close to gross yield and therefore prone to large plastic straining due to even limited loading changes. The soil behaviour is consequently an important factor to the slope instability. The study of the soil displacements, both at the surface and at depth, shows that the landslide is composite, being formed of a shallow rotational slide at the top of the slope, a shallow earthflow downslope and an underlying mechanism of slow and long-lasting irrecoverable movements, which are also monitored on the Serra dell'Acquara landslide body, at the toe of the Vadoncello slope. These slow movements are considered to be consequent to the plastic flow of the weak clayey soils in the slopes, which may be activated by seasonal rainfall effects, by the frequent low-medium intensity seismic events occurring in the area, and also by the morphological changes

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resulting from the slow movements themselves. Rainfall intensity is not found to be the single direct cause of the shallow landslide reactivation, which instead is proposed to be due to the combination of the effects of the low return-period rainfalls in winter 1993 and the mechanism of slow movements active both at depth in the slope and at its toe.

Keywords: Deformation, landslide, movement, rainfall, scaly clay, structure, tectonics.

INTRODUCTION

The risk assessment of landslides and slope movements takes priority over adequate town-planning and road-construction in tectonically active chain areas, where slopes are formed of highly tectonized soils and rocks. In such areas, disturbed soil masses are crossed by both shear surfaces and discontinuities of tectonic origin together with those resulting from past landsliding (A.G.I. 1985). The large-scale features (ranging from tens of metres to kilometres) of these soil and rock masses are referred to as “complex mega-structures” by D’Elia *et al.* (1998) and the corresponding geological formations are defined as “structurally complex formations” (Esu 1977).

The definition of a geotechnical model representing a slope in structurally complex formations and of use in its stability analysis can be quite difficult. The present paper discusses the case study of the Vadoncello slope, which exemplifies these difficulties since it is located in the Sele River Valley in the Southern Apennines (Italy), a chain area made up of an overthrust nappe system enplaced since Upper Oligocene-Lower Miocene (Figure 1) and where slow tectonic movements still occur. The geological formations lying in this area are tectonized and often allochtonous, and the clayey units are often formed of intensely fissured clay shales and scaly clays. In the latter clays, fissures and shear planes form an intricate network, dividing the material into millimetric hard scales (Picarelli 1998), so that its structure is defined as “complex meso-structure” at the scale of the laboratory sample (Esu 1977; D’Elia *et al.* 1998). Due to these meso-structural features, the strength properties of the scaly clays are very poor (Guerriero *et al.* 1995; Picarelli & Olivares 1998; Fearon 1998).

In engineering terms the behaviour of a slope is the output of a boundary value problem of equilibrium for soils subjected to loading conditions which vary with time due to processes of either natural or human origin, e.g. rainfalls, earthquakes, excavations, surcharge loading. Plastic straining and irrecoverable movements in the slope, consequent to even temporary

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changes in the loading conditions, may be slow and progressive, since long time can be necessary for the effective stresses to change at all points in the slope from an initial to a final stable equilibrium. This process of deformation takes place within the well-established "progressive failure mechanism" (Bishop 1967; Potts *et al.* 1997), which is generally considered to only take place in the "pre-failure" stage of the initial slope failure (Vaunat *et al.* 1994; D'Elia *et al. 1998*), although it should rather be considered a mechanism generally applying to slopes experiencing significant deformations and so also to the reactivation of movements on slopes already failed in the past. In fact the reactivation of ancient landslides may follow slow and progressive plastic straining, with either the formation of new slip surfaces or the slipping of the soil mass along a pre-existing slip surface occurring only at the end. The landslide discussed in this paper, the Vadoncello landslide, is a reactivation and enlargement of a previous failure process and is shown to be connected to slow and progressive irrecoverable deformations active at depth and at the boundary of the slope, which are likely to be the response of very weak scaly clays and fissured marls to loading changes effect of rainfalls, seisms and changes in morphology.

An important part of the data used in the study resulted from a comprehensive investigation and monitoring of the slope carried out from December 1994 to the end of 1996 within a research project financed by the European Community (EEC Report 1996). These and other data are examined and discussed in the paper to reconstruct the geological section and the mechanics of the slope. The monitoring of the boundary conditions of the landslide has proved to be particularly important to interpret the mechanism of deformation active in the slope and the causes of instability. Also, the poor mechanical properties of the soils in the slope are considered to be a key factor of the slope instability, therefore they are discussed in some detail.

GEOLOGICAL SETTING OF THE VADONCELLO LANDSLIDE

The Vadoncello landslide was reactivated in December 1993 and lies South-East of the town of Senerchia (Figure 2), in the upper Sele River Valley (Figure 3). This is a tectonically active valley formed in Pliocene-Pleistocene (Caiazzo *et al.* 1992), bordered by carbonatic massifs. The last important seism in the valley occurred on 23 November 1980, with epicentre in Laviano (Figure 3), magnitude M_s=6.9, acceleration 0.33g at Sturno and duration of about 70 seconds (Cotecchia 1986). Between 1993 and 1996 the area has been hit by several minor

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earthquakes, the most important of which was on 3 April 1996 (M_L=4.3) (EEC Report 1996), about 30 Km from Senerchia, when an acceleration of 0.04g was monitored on the Vadoncello slope by means of an accelerometer (Del Gaudio *et al.* 1997).

The soils forming the slopes South-East of Senerchia (Figure 2) are part of the allochtonous Variegated Clay Formation (Ogniben 1969). They consist of grey-green scaly clays, grey-yellow marly limestones, grey marls, sandstones and varicoloured scaly clays, all of which are interbedded chaotically. The slopes are bordered by karst calcareous rocks (the Picentini Mountains) to the West, which host the most important aquifer.

The town of Senerchia lies on a debris slab of 100 m maximum thickness (Cotecchia 1986), which floats above the Variegated Clay Formation (Figure 2). The debris is formed of carbonatic breccias within a silty-sandy matrix, sometimes interbedding volcanic silts. On the whole the slab is of high permeability, being the second important acquifer of the area. The clayey soils of the Variegated Clay Formation form a very low permeability boundary for both of the upstream acquifers, and several springs occur at the outcropping contact between Variegated Clay soils, the Picentini Mountains (hundreds of litres per second flow) and the debris slab.

On the slopes South-East of Senerchia several landforms are indicative of past landsliding. In particular, the Vadoncello landslide is bordered by two important landslide bodies: the Serra dell’Acquara landslide at the toe and the Cimitero landslide at the top (Figure 2). The Serra dell’Acquara landslide is about 2.5 km long, 500 m wide and maximum 33 m deep, and was mobilized by the 1980 earthquake (Cotecchia *et al.* 1986). Due to this event, translational displacements started at the top of the slope (475-490 m a.s.l.) fifteen hours after the first shock, followed by the development of a mudslide. Ten days later, 20 metres uplift was recorded at the toe of the landslide and, about one month after the earthquake, the landslide toe moved further downslope. The 1980 Serra dell’Acquara landslide has been assumed to be the reactivation of a pre-existing landslide by Maugeri *et al.* (1982) and Cotecchia *et al.* (1986), although aerial photos taken in 1955 show some landslide activity only between 274 m and 199 m a.s.l. (Budetta 1983).

The upper part of the Serra dell’Acquara landslide is bordered to the left by a fault edging the debris slab. This is crossed by the rear scarp of the Cimitero landslide (Figure 2), that is a relict rotational landslide body (Cotecchia *et al.* 1986) deep-seated in the clayey units of the Variegated Clay Formation. This landslide was not reactivated by the 1980 earthquake, when

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landsliding took place on the left flank of the Serra dell'Acquara landslide only in the lower part of the Vadoncello slope (the 1980 Vadoncello subsidiary landslide), where clayey soils outcrop.

Since the end of 1980 till 1993 no significant displacements have been observed on all these slopes, except for some slow retrogression of the rear scarp of the 1980 Vadoncello landslide. On 29 December 1993 a sudden acceleration of ground movements took place on the Vadoncello slope (Figure 2), giving rise to a rotational slide at the top of the slope (Figure 4a) and, downslope, to a landslide process which will be referred to as earthflow in the following, although its deformation mode is between that of mudslides (Hutchinson 1988) and earthflows (Cruden & Varnes 1996), since strong deformations occur within the landslide body along with discrete sliding (Bromhead 1992). Ground movements have been active on the slope since December 1993, although the landslide activity has recently decreased. Figure 4(b) shows both the channel and the accumulation zone of the earthflow in summer 1994.

INVESTIGATION AND MONITORING

During the EEC research, numerous continuous coring boreholes were drilled on the slope roject (EEC Report 1996), both inside and outside the landslide body (Figure 5), of depth varying between 21.5 and 57.3 metres (Table 1). The boreholes were equipped with either piezometers or inclinometers (Figure 5; Table 1) and the continuous corings were examined (Wasowski 1995) in order to define the geological section of the slope. Since March 1995 the displacements at ground level have been monitored by means of both topographic and GPS (Polemio & Trizzino 1999) surveying, and those at depth by means of inclinomenters. In addition, in 1995 an aereophotogrammetric survey was carried out (Figure 5). The pore water pressures have been monitored by means of standpipes, Casagrande piezometers and electric piezometers (Table 1), the latter providing a quasi-continuous monitoring (Polemio 1996). Monitoring of temperature and rainfalls has been provided by both a station located close to the slope (Figure 2) and one located in the town. An accelerometer located on the slope has measured the local seismic accelerations. The monitoring data (EEC Report 1996), together with the interpretation of the soil profiles and the results of laboratory testing, have provided the observational database for the understanding of the mechanics of the slope.

LITHOLOGICAL SECTION OF THE VADONCELLO SLOPE

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The schematic lithological section of the Vadoncello slope, together with the changes in slope profile since the end of 1980, is shown in Figure 6. This is the result of in situ geological surveys, a study of the soil profiles, and an analysis of both index test data from Fearon (1998) and new data. The variation with depth of soil and rock type, as well as of the degree of remoulding, has been generally found to be frequent and chaotic (Wasowski 1995), the soil type often changing even within ten centimetres. However, a thorough study of the soil profiles has lead to identify defined alternations and/or successions of strata, which occur in the different parts of the slope and have been grouped into different “soil complexes” as shown in Figure 6.

Complex G represents the earthflow debris resulting from landsliding on the slope since 1993 till March 1995, when it was 550 m long, 140 m wide and about 10 m deep. Within this complex three main landslide debris are interbedded with discontinuous topsoil. The deepest debris is the most clayey, principally formed of dark-grey to yellow-grey scaly clays, whereas the most superficial debris is the most silty and sandy. The degree of remoulding of the soils is maximum in the top 5-6 metres of complex G, although remoulding is still significant at larger depths and neither a bedrock nor a shear band can be distinguished at the base of this complex.

The lower part of this complex overlies the 1980 Vadoncello landslide, defined as complex H. Further down-slope this overlies the 1980 Serra dell’Acquara landslide body, which is defined as complex I. Complex H is principally formed of grey scaly clays, sometimes silty, including marl and sandstone clasts. Clays in complex I are more sandy and silty at the top and at depth they are often interbedded with marl and calcareous clasts. On the whole complex I is fairly ununiform and several blocks of breccias float within it.

The soils which have not been subject to dislocation until January 1995 have been grouped into complexes A to F and will be referred to as “in situ” soils in the following. Complex A outcrops up-slope the rear scarp, is formed of calcareous mega-breccias and finer detritus and represents the South-East margin of the debris slab in Figure 2. Complexes B to F are part of the Variegated Clay Formation. Complex B is principally formed of scaly marly clays, light brown to grey in colour, sometimes interbedding fractured marls and marly limestone. Clays of complex C are also scaly, but less marly and far more plastic than complex B. They are of a darker brown-grey colour and may interbed marl and siltstone strata. The scaly clays of complex D are darker, but less plastic than complex C and also interbed marls and calcareous marls. Both complexes B and C outcrop behind the rear scarp, whereas complex D crosses this scarp (Figure 2).

Complex E outcrops just below the rear scarp and is also present at depth further down-slope. It is made up of grey-green clays and light green marly clays, marls, marly limestones and silty sandstones. Finally there is complex F, which forms the central part of the slope and also underlies complex I at the toe. This is formed of grey to dark grey scaly clays, interbedding marly strata and thin (about centimetres) siltstone layers.

The set-up of the strata in complexes B to F is chaotic, with the lapideous strata being fractured and discontinuous. On the whole the complexes seem to be verging up-slope. According to the classification of complex structures proposed by Esu (1977), these soils have structures A_2, B_2 and B_3. Soils in complexes B to F can be as disturbed as the soils which occur below 6-7 m depth within complexes G to I.

PHYSICAL, MINERALOGICAL AND MECHANICAL PROPERTIES OF THE SOILS

The index properties of scaly clay samples from the different complexes in Figure 6 are plotted in the Casagrande plasticity chart shown in Figure 7. Fearon (1998) and Fearon & Coop (1999) discuss the inadequacy of index testing to interpret the nature of the scaly clays in the slope, since the disaggregation of the clay particles forming the scales is not necessarily complete with standard remoulding. Therefore both the liquid limit and the plasticity index of these clays may be underestimated by standard index testing and the data in Figure 7 should only be accounted for a comparison between the clayey soils in the slope rather than to compare the nature of these clays with that of other non-scaly clays. Nonetheless the data show that most of the clays in the slope are of high plasticity. No significant difference exists between the data for the in situ complexes B to D and the landslide complexes G to I. Complex C is the most plastic, followed by complexes D and F.

The liquidity indexes (LI) and the void ratios of clayey soil samples are plotted against depth below ground level in Figures 8 and 9 respectively, where the samples from the landslide debris (complexes G, H and I) are distinguished from those from behind the rear scarp (complexes B, C and D). Both the soil void ratio and liquidity index follow similar trends with depth. This suggests that the differences in void ratio are primarily effect of differences in the soil meso-structures, and only secondarily effect of their differences in index properties. The in-situ liquidity indexes (complexes B, C and D) do not decrease with

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depth, as would be the case for soils consolidated only due to surcharge loading, probably due to the original disturbed nature of the in-situ soil meso-structures. However, the in-situ soil samples have lower liquidity indexes than the landslide debris soils; thus landsliding seems to have further opened and disturbed the original in-situ soil structure.

Mechanical testing has been carried out on several soil samples, three of which will be examined in the following (samples P5(1), I6(1) and I6(3), from 16.7 to 28.5 m depth; Figure 6) in order to exemplify the behaviour of the in-situ soils and show that this is a critical factor of the instability of the slope. In particular, the soils tested are scaly clays of high plasticity (CH), with clay fraction in the range 53-61%, and belong to complexes B and C (index properties in Table 2). Their calcite contents, measured by X-ray diffraction, indicate that sample P5(1) is a slightly marly clay (12%), whereas both samples I6(1) and I6(3) are clayey marls (31-35 %). The mineralogical composition of the three samples is fairly similar, the clay minerals being 51% on average, followed by calcite (26%), quartz (15%), feldspars (2%) and traces of pyrite and siderite. The clay minerals are interstratified illite-smectite (79%, illite content higher than the smectite content), kaolinite (15%), illite (4%) and chlorite (2%).

Results of high-pressure oedometer tests (σ_{vmax}'=18 MPa) on both samples I6(1) and I6(3) after reconstitution in the laboratory at an initial water content about 1.5 the liquid limit (Burland 1990), are shown in Figure 10. These define the one-dimensional normal compression lines of the reconstituted clays, ICLs (Burland 1990). The figure also shows results of restrained-swelling high-pressure oedometer compression tests (σ_{vmax}' = 18.7 MPa) on specimens from the undisturbed samples, whereas Figure 11 shows results of oedometer tests in which the soil was first unloaded to low pressures before reloading. The different void ratios of specimens from the same sample, e.g. I6(3) and I6(3)b, or P5(1) and P5(1)b (Figures 10 and 11) are indicative of the dishomogeneity of the soil in the in-situ samples.

The compression indexes of the reconstituted specimens, C_c* (measured between 100 and 1000 kPa vertical pressure) are equal to 0.44 and 0.52 for samples I6(1) and I6(3) respectively. Despite the fact that the measured plasticity indexes are not representative of the nature of the clays tested, both these C_c* values are close to the corresponding values resulting from the relation between C_c* and the plasticity index PI (Schofield and Wroth 1968):

$$C_c^* = G_s\ PI/2 \qquad (1)$$

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being C_c* from equation (1) equal to 0.45 for sample I6(1) and to 0.56 for sample I6(3). Thus the index properties of the scaly clays determined with a remoulding procedure similar to that used during conventional soil reconstitution (Burland 1990), appear to be related to the "intrinsic properties" of the clays measured by testing the reconstituted, as is generally the case for the reconstituted non-scaly clays. For both the reconstituted samples in Figure 10, C_c* reduces to about 0.2 at $\sigma_v' = 18$ MPa. Microstructural analyses have been carried out by means of the scanning electron microscope (SEM) on specimens of both reconstituted clay and natural undisturbed clay. SEM pictures of the fabric of the reconstituted clay and of the natural clay are shown in Figures 12 (a) and (b) respectively. Scales, formed of oriented particles, are seen in the undisturbed fabric. However, relicts of the original scales and particle aggregates occur also within the matrix fabric of the reconstituted. These observations confirm that conventional remoulding does not disaggregate completely the particles forming the scales.

The compression curves of the natural clay specimens plot to the left of the ICLs in Figure 10, as already observed with other scaly clays and clay shales from the Apennines (Guerriero *et al.* 1995; Picarelli & Olivares 1998), although scaly clay behaviour has been seldom compared at high pressures with that of the same clay reconstituted. This comparison instead proves to be indicative of the mechanical effects of the natural scaly meso-structure. This gives rise to particularly high re-compression indexes and low void ratios, the latter of which correspond to a stretch of ICL characterized by quite high stiffness and relating to high vertical effective stresses. Consequently the state of the natural sample is to the left of the ICL and joins it only after compression to very high pressures (Figure 10), in contrast to what generally observed for non-scaly sensitive clays, which generally have higher void ratios and stiff re-compression curves crossing the ICL pre-gross yield (Burland 1990; Cotecchia & Chandler 1997; Cotecchia & Chandler 1999). "Gross yield" is here referred to the soil state at which a significant reduction in stiffness and increase in compression index occur in connection to a change in the soil hardening properties (just yield for Burland 1990; secondary yield for Leroueil & Vaughan 1990).

The "gross yield" states of the natural scaly clays are located to the left of the ICL (Figure 10). For each of the specimens the gross yield pressure σ_y' has been identified within the stretch of compression curve of maximum curvature (Casagrande construction). Otherwise,

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gross yield pressure values, consistent with those in Figure 10, have been identified in the plot of the soil one-dimensional stiffness, M ($=d\sigma_v'/d\varepsilon_v$), against the vertical effective stress σ_v' (Figure 13), where the gross yield pressure corresponds to the vertical effective stress just before the minimum stiffness M (Janbu & Senneset 1979). Figure 13 shows that the natural scaly clay is only slightly stiffer than the reconstituted one at any vertical pressure, whereas sensitive natural clays generally have pre-gross yield stiffness far higher and post-gross yield stiffness lower than the corresponding (same vertical effective stress) stiffness of the reconstituted soil. This is because sensitive clays experience significant structural degradation at gross yield (Cotecchia & Chandler 1998), whereas the scaly clays are already fairly disturbed pre-gross yield and gross yield does not give rise to significant structural damage. For the same reason, the gross yield points on the one-dimensional compression curves of the scaly clays are not as well defined as those of sensitive clays.

Figure 11 shows that the undisturbed scaly clay samples exhibit particularly wide swelling-reloading loops, with quite high swelling indexes C_s (0.05-0.15). The C_s values are only slightly lower than the compression index (C_c) values (0.15-0.2), this being probably due to both the high proportion of swelling minerals in the clay and the very poor bonding existing between the scales, which, as such, does not restrain swelling. The swelling minerals would render the swelling capacity of these soils even higher if the clay particles were not bonded into scales. The high swelling capacity of the soils in the slope is likely to cause significant deformations during soil re-wetting due to either rainfall infiltration or the rising of the water table.

Due to the significant hysteresis in the unloading-reloading cycles, during reloading the gross yield states appear to shift upwards and further to the left of the ICL (Figure 11). This is even more apparent in the M-σ_v' plot in Figure 14, where the minimum values of stiffness M in the unload-reload oedometer tests are seen to occur at much lower σ_y' values than for the corresponding loading tests from the swelling pressure (Figure 13). However, the recompression curves in Figure 11 appear to join the same normal compression lines defined in Figure 10, which therefore are state boundary envelopes for the corresponding soil samples in one-dimensional compression. If these state boundary envelopes, together with the corresponding gross yield states, lie on the soil state boundary surface (SBS) in the specific volume ν – deviatoric stress q - medium effective stress p′ space, the data in Figure 10 imply that the natural scaly clays have smaller state boundary surface and lower strength than the corresponding reconstituted clays. Guerriero *et al.* (1995) and Picarelli & Olivares (1998) have found the peak strength envelopes of two overconsolidated natural scaly clays to be smaller than those of the corresponding reconstituted clays. This difference in strength envelopes is consistent with the gross yield behaviour of the soils shown in Figure 10.

Drained direct shear tests have been carried out on specimens from sample I6(3), which were consolidated to vertical pressures from 98 kPa to 980 kPa (Figure 15). Both the specimens consolidated to 980 kPa and 784 kPa vertical effective stress strain harden (increase in friction angle: arctang $\tau/\sigma_v'=\phi'$) and contract up to large strains; ϕ' reduces towards the residual value ϕ_r' (=5-6°; Fearon 1998) only at very large strains. Thus these specimens appear to exhibit a “wet behaviour” (Schofield & Wroth 1968) and reach a “pseudo-critical state”, chracterized by both zero strength and volumetric strain increments and ϕ_{CS}'=12°-15°, after about 2-3 mm displacement. Figure 16 shows that the consolidation states of these two specimens lie close to the normal compression line of sample I6(3) in the e-logσ_v' plane, as should be for a wet behaviour. On the contrary, the specimens sheared at lower vertical effective stresses (σ_v'=588, 392 and 98 kPa) dilate and strain soften significantly since low shear strains, according to a “dry behaviour”, and plot further to the left of the normal compression line of sample I6(3) (white dots in Figure 16). Thus it seems possible to distinguish, in the e-logσ_v' plane, a domain of consolidation states close to the normal compression line of the natural soil where the soil behaviour is wet, and a domain further to the left, where the behaviour is dry, in accordance with critical state soil mechanics. Triaxial test results on similar samples are presented by Fearon & Coop (in prep.). More testing is currently under way in order to assess the extent to which the critical state framework can interpret the behaviour of these soils.

The maximum friction angles exhibited by the I6(3) specimens vary from the very low ϕ' values at the pseudo-critical state (ϕ_{CS}'=12°-15°) to ϕ_p'=24°, exhibited at peak by the most dilative specimen (σ_v'=98kPa). These low ϕ' values confirm the poor mechanical properties of these soils, which are weaker in the natural than in the reconstituted state. The soil mechanical properties do not improve with depth. In addition, whereas the shallow samples consolidated to the in-situ stresses exhibit some dilation and peak strength, with ϕ_p' in the range 23°-30° (Fearon 1998), the deep samples exhibit a wet behaviour during shear after consolidation to their in-situ stresses, with maximum friction angles ϕ_{CS}' in the range 12°-17° (Fearon 1998).

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Therefore the deep soils in the slope appear to be close to the SBS, so that even small changes in the loading conditions can cause them to gross-yield, giving rise to large plastic strains and irrecoverable movements. Consequently, the poor mechanical properties of the in-situ soils in the Vadoncello slope appear to be a critical inherent factor to the slope instability and to the long-lasting progression of slow movements, which will be shown to take place at depth in the slope and downslope its toe.

GROUND MOVEMENTS AND SLOPE DEFORMATION MECHANISM

Since 1980 a slow retrogression of the rear scarp has taken place at the top of the Vadoncello slope (about 445 m a.s.l.). In December 1993 a rotational slide caused significant loss of material between 420 and 450 m a.s.l., causing the immediate failure of a road (Figure 4a) and an earthflow was mobilized downslope, forming complex G. Falls, rotational slipping at the rear scarp and earthflowing have been active since December 1993, with a significant acceleration of the ground movements in 1995, when a house sitting on the top of the slope (Figure 6) was damaged.

The data from topographic surveying during 1995 and 1996 (data from EEC Report 1996) lead to identify three main areas on the slope within each of which the horizontal displacement rates were quite uniform. These areas (A, B and C) are shown in Figure 17 with the average displacement rates and the maximum and minimum displacements measured during four time intervals. The displacement rates in area A relate to the retrogressive slipping at the top of the slope (source area), whereas area B corresponds to the earthflow channel and area C covers the earthflow accumulation zone and part of the Serra dell’Acquara landslide deposit. The highest displacement rates in area B occurred in the second period of observation, between 18 May and 23 June 1995 (> 30 m/month) and were still very high (20 m/month) in the third period, between 23 June and 14 October, when the displacement rates were maximum in both areas A and C. Therefore the average activity of the whole landslide appears to have been highest, in 1995, between the end of June and end of October. The vertical displacements at the toe of the slope are shown in Figure 18 and indicate an initial bulging in May-June 1995, followed by settlements in the period of maximum landslide activity.

The displacement directions show that the part of area C overlapping the Serra dell’Acquara landslide has moved South-East, following the main direction of the Serra

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dell'Acquara landslide body, in contrast to the North-South direction of all the other displacements on the Vadoncello slope. In addition also the Serra dell'Acquara landslide, outside area C, is seen to move, since surveying points such as T4 (Figure 17) moved at rates between 0.003 and 0.05 m/month in the monitoring period. These observations demonstrate that the Serra dell'Acquara landslide has been moving slowly in its own direction in 1995, influencing the toe of the Vadoncello slope and probably the overall activity of the Vadoncello landslide, as discussed later.

The displacements logged at depth in the slope by means of inclinometers I1, at the toe, and I4, at the top of the slope respectively (Figures 5 and 6), are shown in Figure 19. The inclinometer data for I4 (Figure 19a), behind the rear scarp, show the largest displacements to occur in the top 9 metres, although smaller displacements were logged also down to 15 m depth; this inclinometer was sheared at both 15 m and 9 m depth on 9 May 1995. The deeper displacements (between 9 and 15 m depth) indicate that either the retrogressive failure mechanism at the top of the slope was deeper than 9 m or it lies above a slower progressive mechanism of deformation active at depth. The occurrence of deep deformations in the slope is confirmed by the I1 inclinometer readings, Figure 19(b). These readings show that on 22 May 1995, before the stage of maximum activity of the whole landslide, significant displacements were taking place down to 16-17 m depth below the toe of the earthflow body G, and that smaller displacements were still active down to about 23-24 m depth. Inclinometer I1 was sheared on 26 August 1995 at 16-17 m depth, within complex I (Serra dell'Acquara landslide body, Figure 6). The I1 inclinometer readings have been drawn in Figure 20 in relation to the logging day, together with the average displacement rates monitored at the surface in the toe area. The plot shows that the deep and the superficial displacements at the toe of the slope accelerated at the same time in May-June 1995 and that the consequent maximum activity of the toe of the slope was related to deep movements and occurred at the same time as the maximum activity of the rest of the landslide. All these observations suggest that the activity of the Vadoncello landslide is related to displacements active at depth, which are far slower than the displacements within the shallow landslide bodies: the rotational slide and the earthflow. These deep displacements are found to take place also within complex I and are therefore connected to the slow displacements of the Serra dell'Acquara landslide monitored at the surface. They may be considered to influence the stability of the shallow soils on the Vadoncello slope and are probably one of the triggering factors of the fast shallow landsliding. The inclinometers located in the central part of the slope (I2 and I3, 17.5 and 22 m deep respectively; Table 1), logged no significant displacements until the occurrence of sudden shallow failures (4-5 m depth), probably because they were floating before being sheared at shallow depths.

The mechanism of deformation active at depth in the Vadoncello slope and on the Serra dell'Acquara landslide can be classified as a "long-term mechanism of slow, plastic deformation and irrecoverable movements". The deep soils involved in this deformation process within the Vadoncello slope, have been shown to have low strength and in situ states close to the state boundary surface, conditions which make them prone to large plastic flow due to even small changes in loading. Thus the mechanism of irrecoverable movements is possibly consequent to even limited changes in loading result of rainfall infiltration, seismic events and changes in morphology, the latter of which in turn arise from the deep-seated deformations. These deformations are so slow as not to cause alarm of sudden instability for the Serra dell'Acquara slope, which is quite flat, but they may be critical to the stability of the steeper Vadoncello slope.

RAINFALLS AND HYDROGEOLOGY OF THE SLOPE

The statistical study of the cumulative daily rainfalls, ranging from 1 to 180 days, has been carried out to characterize the rainfall effect on four landslides which have been mobilized from 1980 to 1996 in the area around Senerchia (Polemio 1997), including the Vadoncello landslide. This study has shown that the return period of the cumulative rainfalls occurring just before landslide activation has generally been much less than 10 years. Therefore landsliding has not been the direct consequence of cumulative rainfall in the four cases examined. The cumulative rainfalls before the 1993 December Vadoncello landslide are shown in Figure 21. These were of particularly low return period (< 2 years); as such, they could not be the only cause of the sudden reactivation of the Vadoncello landslide.

The piezometric regime of the Vadoncello slope has been deduced from the readings of the piezometers set up in the slope (Figure 5). The readings of the Casagrande and the electric piezometers set up at small distance down the same borehole (see Table 1) have been generally consistent between each other. Due to the very large displacements taking place in 1995 at shallow depths, the working life of most Casagrande piezometers was very short (e.g. few weeks for piezometers in P2 and P3), whereas the working life of the electric piezometers

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has been longer due to the possibility for the electrical cables to slip in the bending PVC tubes (Polemio 1996). Also, since the electric piezometers were set to log readings every 0.5 to 6 times per hour, they could record rapid changes in piezometric level following rainfalls.

The piezometric data logged at larger depths (Table 2) define a deep hydrogeological complex with piezometric levels between 12.8 and 15 m below ground level. The readings of the electric piezometers in this hydrogeological complex have never been affected by short duration-high intensity rainfall events during the monitoring period. On the contrary, the ground water flow taking place at shallower depths, above 7-9 m depth below ground level, is often not steady and defines a hydrogeological complex located within the most disturbed earthflow soils, which also have the highest void ratios (Figure 9) and coefficient of permeability. The flow in this hydrogeological complex is affected by rainfall infiltration, as shown, for example, by the readings of the electric piezometer in borehole P2 at 7.3 m depth, which are plotted in Figure 22 against time along with the corresponding rainfalls. Also shown in the figure are the corresponding readings of the deeper electric piezometer in the same borehole (15.3 m depth). It is seen that short and heavy rainfalls in March 1995 triggered changes in piezometric level (from 7.3 to 5.9 m below ground level, pore water pressure change about 15 kPa) in the shallower hydrogeological complex in less than 15 hours each time, but not in the area surrounding the piezometer in the deeper hydrogeological complex. The pressure changes in the top complex were dissipated in less than 20 hours each time. Thus, the pore water pressures in the soils below complex G appear not to exhibit immediate response to rainfalls which, on the contrary, cause quite immediate changes in pore water pressure within complex G. This observation suggests that the superficial displacements in 1995-96 were related, at least in part, to high intensity - short duration rainfalls. The deep-seated displacements seem not to have been directly related to these rainfalls, although the change in soil weight caused by rainfall infiltration in the shallow soils and possible water infiltration in fissures and fractures at depth may have influenced the loading conditions of the deep soils in the long term and contributed to the deep displacements. No doubt the presence of steady positive pore water pressures at depth has to be considered an inherent cause of instability for the Vadoncello slope, since it reduces the deep soil strength.

CONCLUSIONS

The understanding of the set-up and behaviour of the soil masses within the Vadoncello slope has required a comprehensive analysis of investigation and monitoring data. The soils have been found to be disturbed also at large depth and to have very poor mechanical properties. Oedometer tests have shown the gross yield state of the clayey materials to lie to the left of the ICL; therefore the natural soils in the slope have a smaller state boundary surface than the corresponding reconstituted soils. The consequent low values of gross yield pressure and shear strength of the natural soils make them prone to large plastic flow, so that they may be at gross yield even at large depths and be subject to plastic straining due to even small changes n the loading conditions.

The Vadoncello landslide may be defined as a composite landslide according to Cruden & Varnes (1996), taking place as superimposition of different mechanisms of deformation and failure: a retrogressive slipping mechanism at the top of the slope, an earthflow downslope and an underlying deep-seated mechanism of slow irrecoverable movements. The first two failure mechanisms are rather shallow and fast (the first about 15-20 m deep, the second about 10 m deep). The mechanism of slow irrecoverable deformation involves soils at depth in the Vadoncello slope and soils in the Serra dell’Acquara slope. The occurrence of these slow displacements has been revealed by both inclinometer readings and topographic surveying, carried out both inside and outside the Vadoncello landslide. In particular, the monitoring outside the Vadoncello landslide has proved to be important, e.g. surveying point T4 (Figure 5) on the Serra dell’Acquara landslide.

The reactivation of the shallow failure processes on the slope in December 1993 seems not to have been caused solely by the rainfalls preceding the event, both because these were of particularly low return period and because electric piezometer readings have shown that the groundwater pressures in the slope are not immediately affected by rainfalls except for the pore water pressures in the landslide debris consequent to the 1993 event. Rainfalls may have had direct and immediate effects on the shallow movements after 1993, but not either on the deep movements progressing in the long term or on the overall activity of the landslide in June-October 1995, when significant accelerations of the movements took place also at depth. Rainfalls probably cause changes in loading for the deep soils in the long term, e.g. seasonal loading pulses, due to the change in weight of the top soils consequent to rainfall infiltration and possibly due to the infiltration of water into deep fissures and fractures. As such rainfalls may contribute to keep active the long-lasting mechanism of deformation active at depth, which may be also activated by other environmental factors, such as seismic loading, frequently monitored in the area, and the changes in slope geometry consequent to the deep movements themselves. Thus it is suggested that rainfalls, seismic loading and the poor strength properties of the soils are causes of the mechanism of irrecoverable deformations active in the Vadoncello slope at depth and in the Serra dell'Acquara slope. This mechanism influences the stability of the Vadoncello slope and may be considered one important factor of the slope failure in 1993, although this factor probably then added to the direct effects of other climatic and seismic factors, e.g. the normal intensity rainfalls preceding the 1993 December event.

Modelling the failure and deformation processes taking place in the slope requires numerical applications with models implementing the constitutive behaviour of the scaly clays. Results of a numerical modelling of the Vadoncello slope have been reported by Cotecchia *et al.* (2000) and appear to confirm the interpretation of the failure processes proposed here. Further research into the constitutive modelling of scaly clays is under way.

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Figure Captions

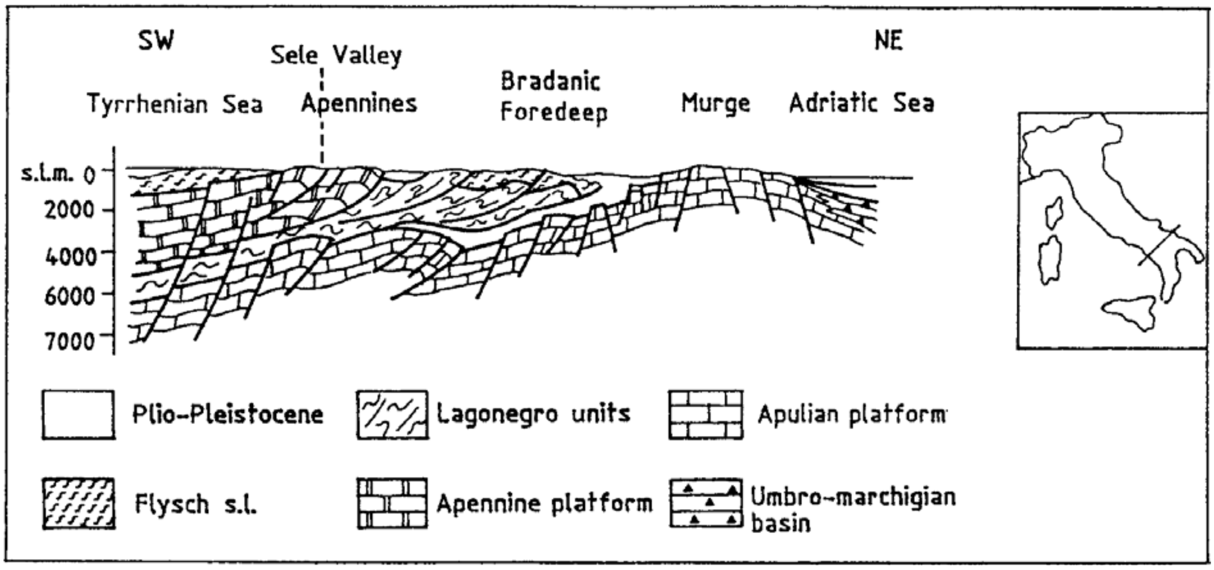

Fig. 1. Schematic geological section through Southern Apennines (after Sella et al. 1988).

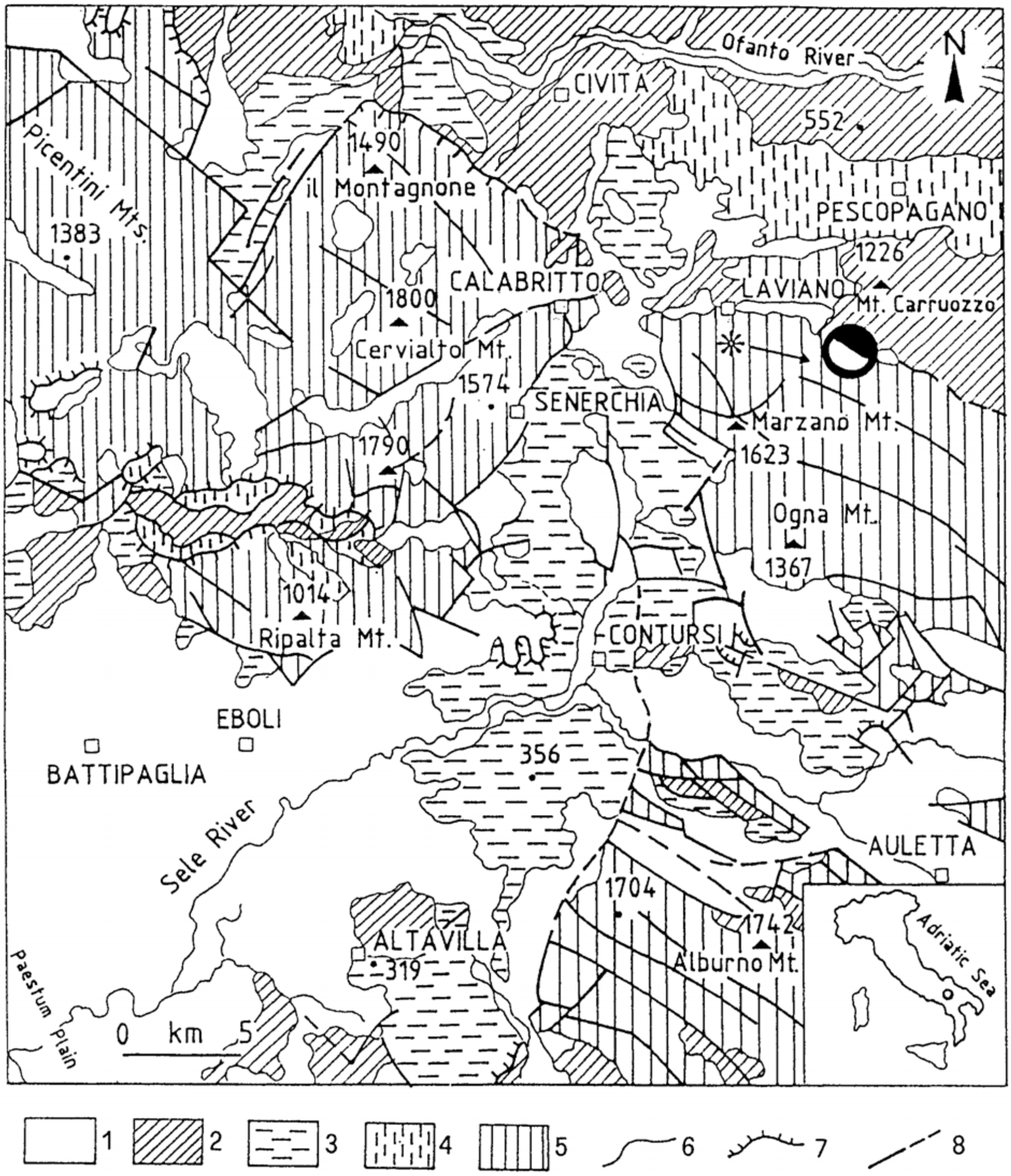

Fig. 2. Geological map of the area (after Cotecchia et al. 1986, modified). Keys: 1, Vadoncello landslide (a) and Serra dell'Acquara landslide (b); 2, talus debris, eluvial and colluvial deposits; 3, debris slab (a) and mega-breccia deposit (b); 4. Variegated Clay Formation (Paleogene-Upper Cretaceous): varicoloured clays with calcareous-marly blocks or strata (a), interbedding limestones, marls and clays (b); 5, Picentini unit (Cretaceous-Upper Triassic): limestones and dolomitic limestones; 6. stratigraphic contact; 7. fault; 8, strata attitude: a) 0-10°, b) 10-45°, c) > 45°; 9, landslide crown; 10, crown of the 1980 Vadoncello slide; 11, reactivated toe of the 1980 Serra dell'Acquara landslide; 12, principal springs.

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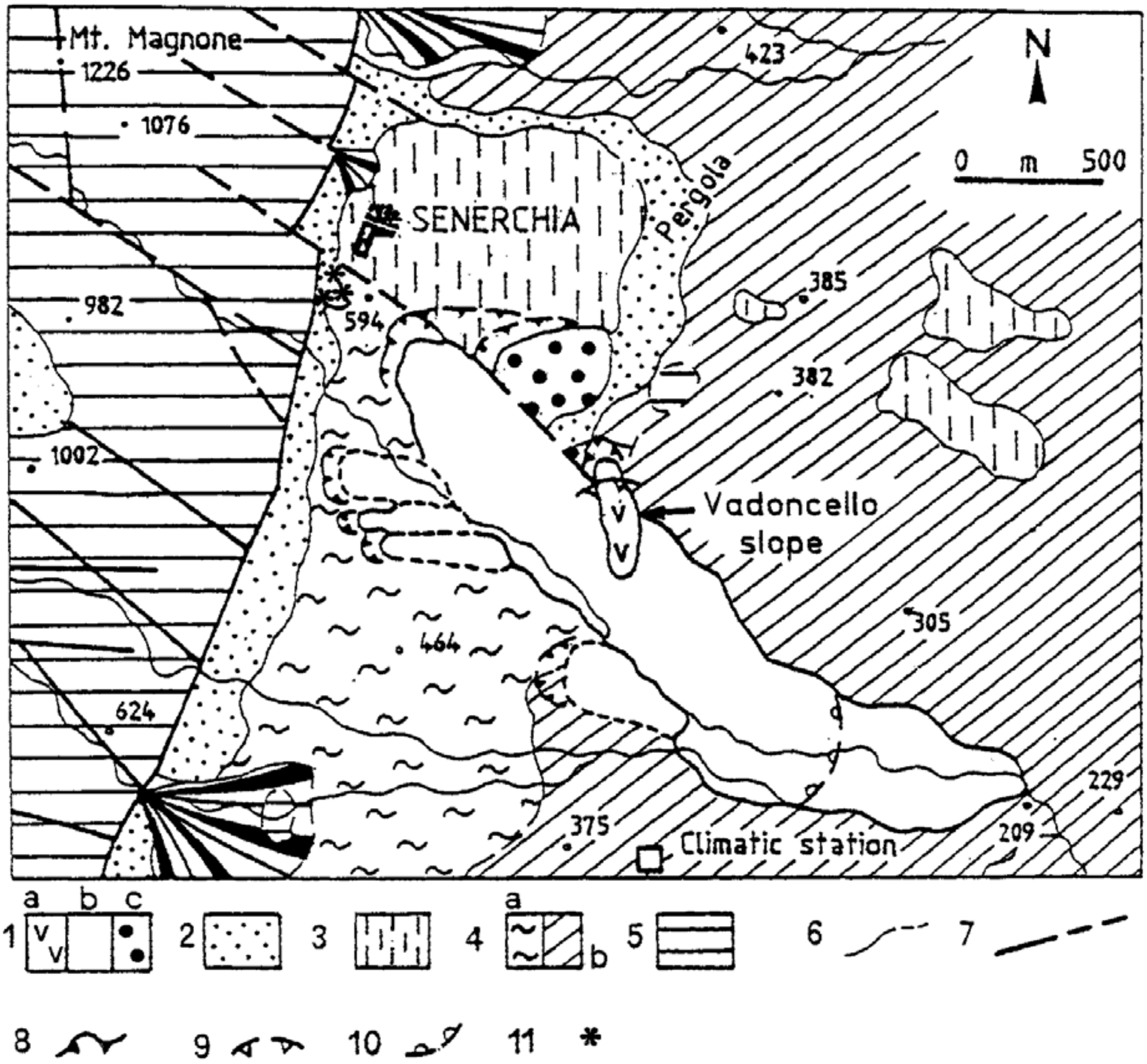

Fig. 3. Geological map of the upper Sele River Valley (after Bonardi *et al*. 1988). Keys: 1, Quaternary sedimentary deposits; 2, Foredeep deposits (Calvello unit; Lower Pleistocene-Lower Pliocene); 3, Neogenic lithological units; 4, Sicilide Complex (Lower Miocene-Cretaceous); 5, Lagonegro II unit (Oligocene-Lower Triassic); 6 Carbonatic platform units (Eocene-Upper Triassic); 7, stratigraphic contact; 8, tectonic contact; 9, faults.

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Fig. 4. The Vadoncello landslide in summer 1994: rear scarp (a); channel and accumulation zone (b).

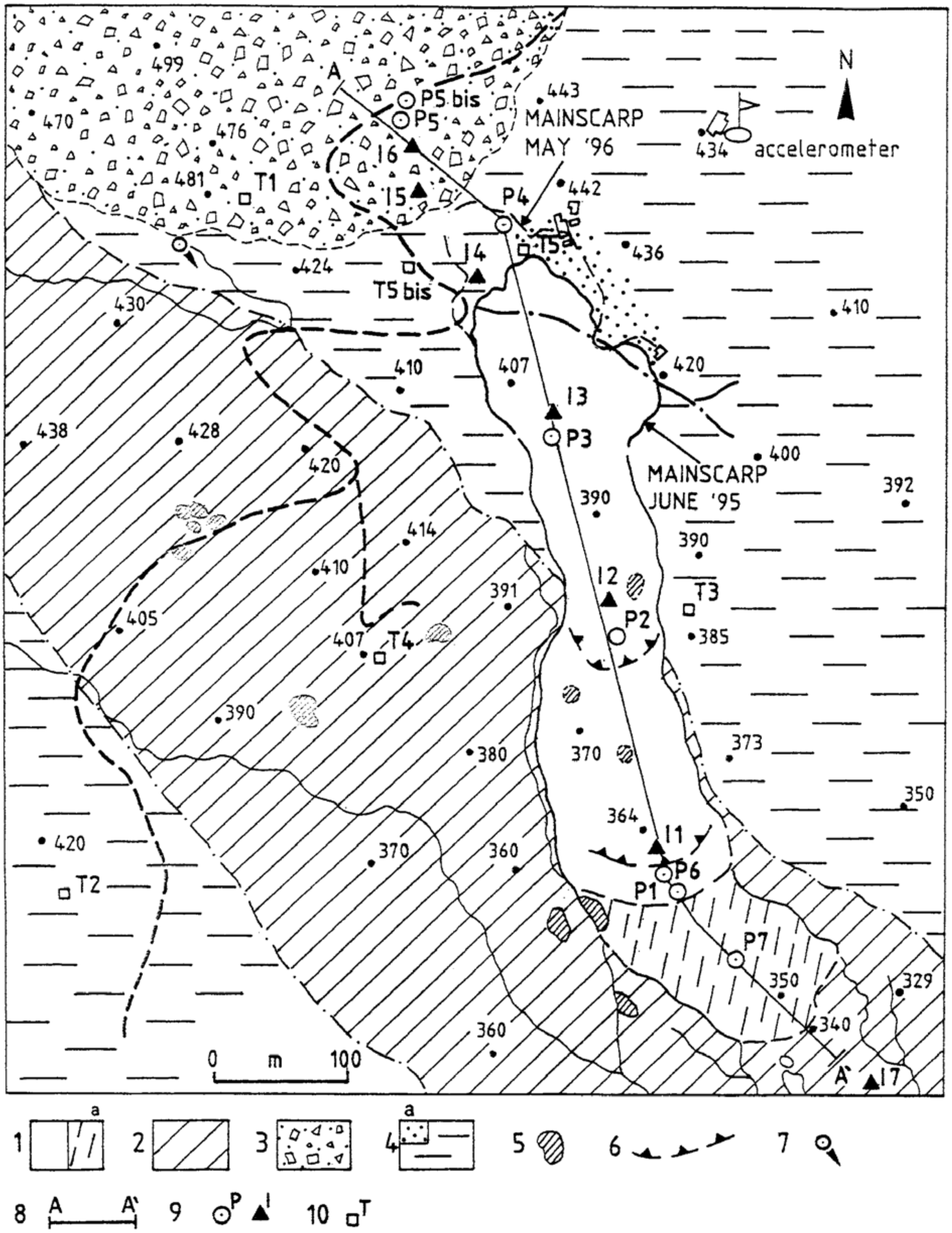

Fig. 5. Geomorphological map of the Serra dell'Acquara-Vadoncello slopes and monitoring system (after the EEC Project Report 1996). Keys: 1, 1993-'95 Vadoncello landslide, overlying the Serra dell'Acquara landslide in area (a); 2, Serra dell'Acquara landslide; 3, debris slab; 4, Variegated Clay Formation and floating calcareous debris (a); 5, stagnation waters; 6, internal toes; 7, strata attitude; 8,

stratigraphic contact; 9, spring; 10, line of section in Fig. 6; 11, piezometers (P) and inclinometers (I); 12, topographic control stations outside the Vadoncello landslide.

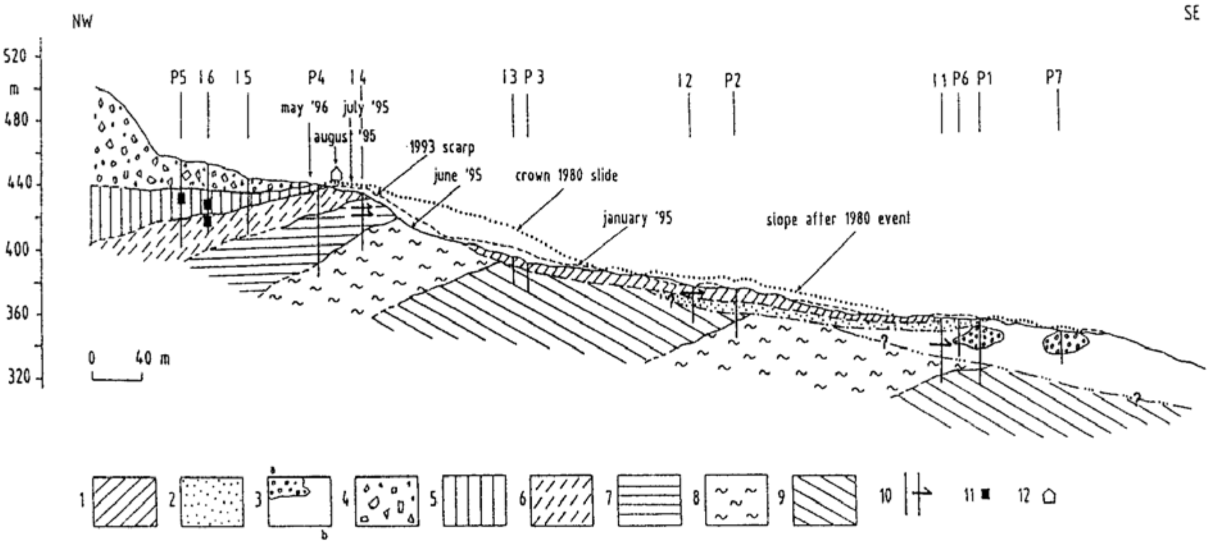

Fig. 6. Cross section (line in Fig.5) of the Vadoncello slope. Keys: 1, Complex G; 2, Complex H; 3, Complex I with calcareous blocks floating in it (a); 4, Complex A; 5, Complex B; 6, Complex C; 7, Complex D; 8, Complex E; 9, Complex F; 10, piezometers (P1-P7) and inclinometers (I1-I6) with depth of shear; 11, undisturbed samples discussed in the paper; 12, house.

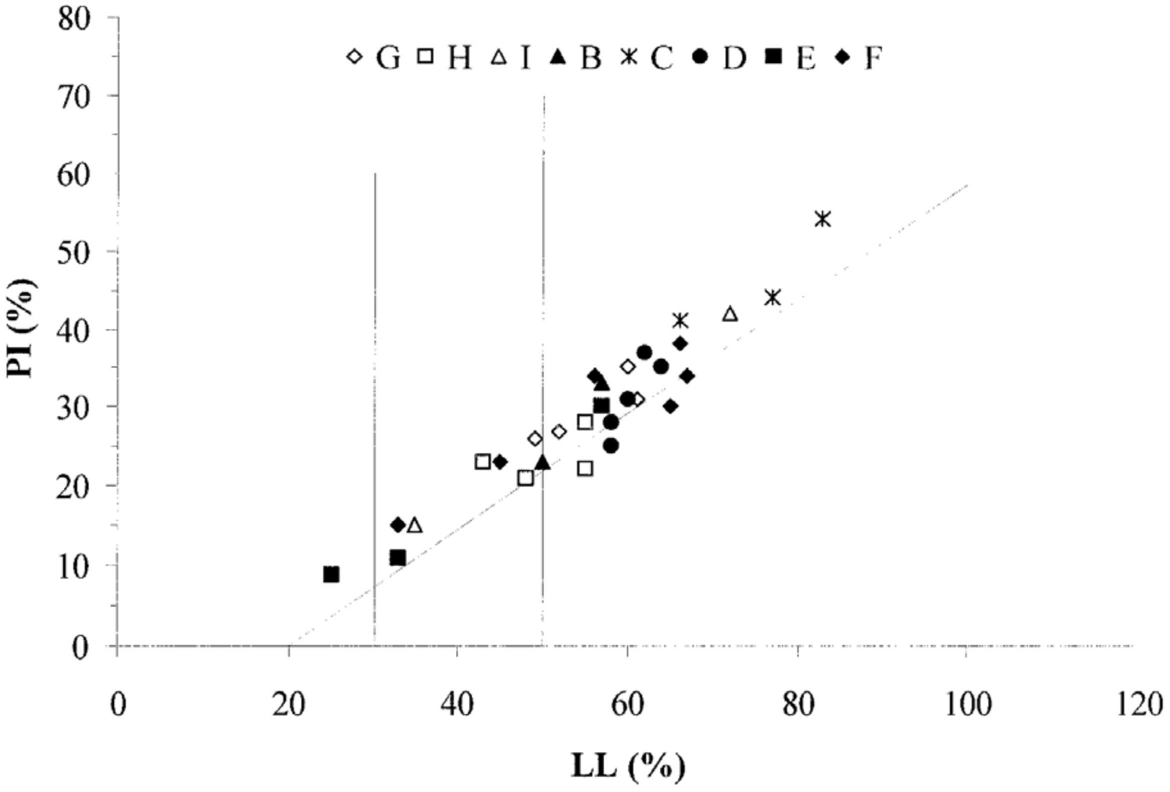

Fig. 7. Plasticity chart for "in situ" and landslide samples.

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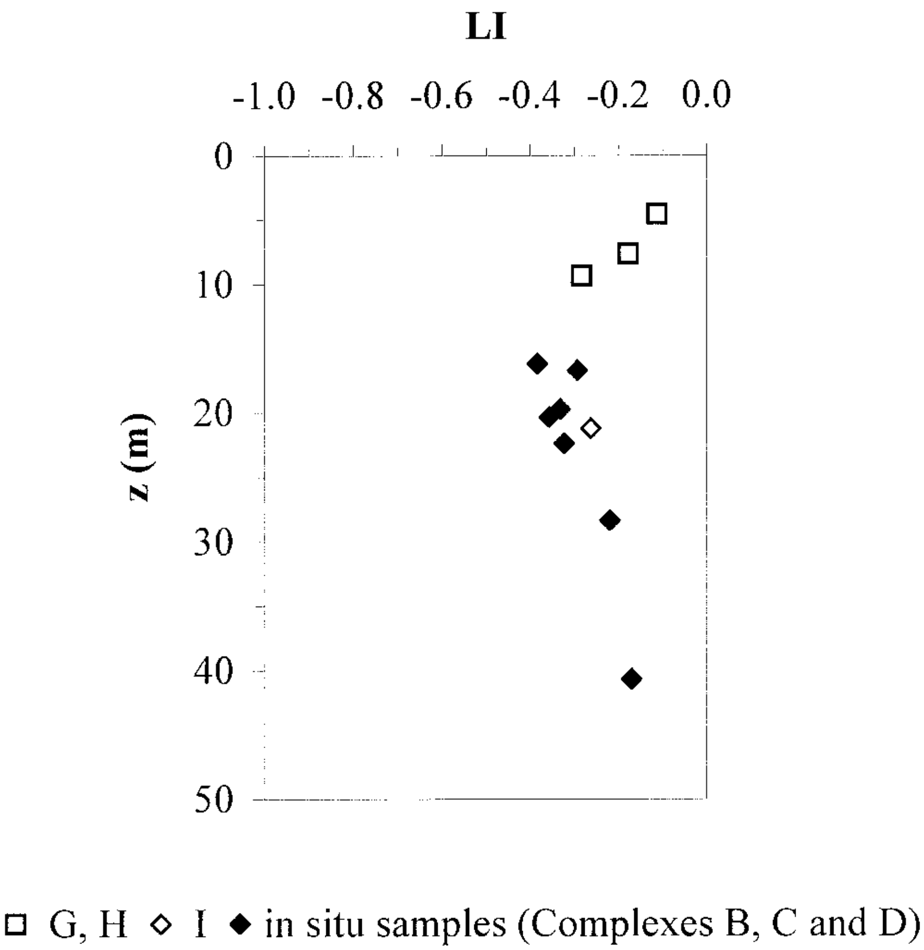

Fig. 8. Liquidity indexes of “in situ” and landslide samples.

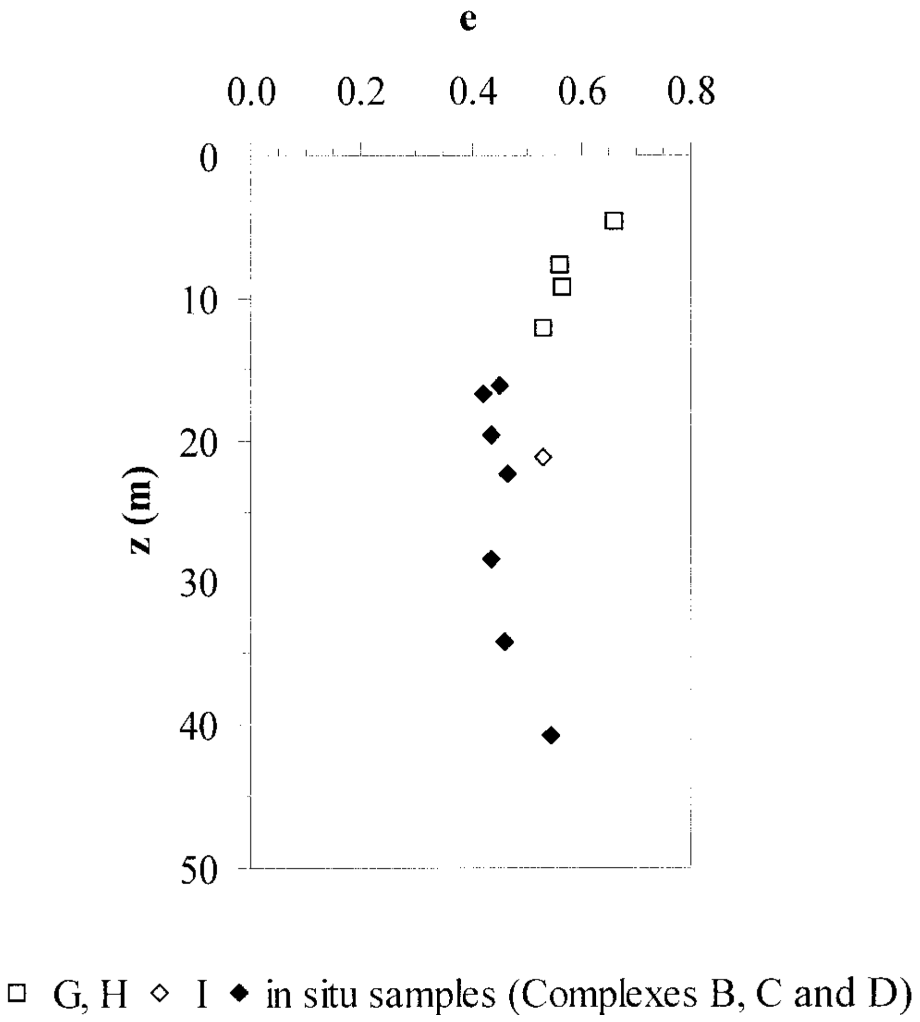

Fig. 9. Void ratios of "in situ" and landslide samples.

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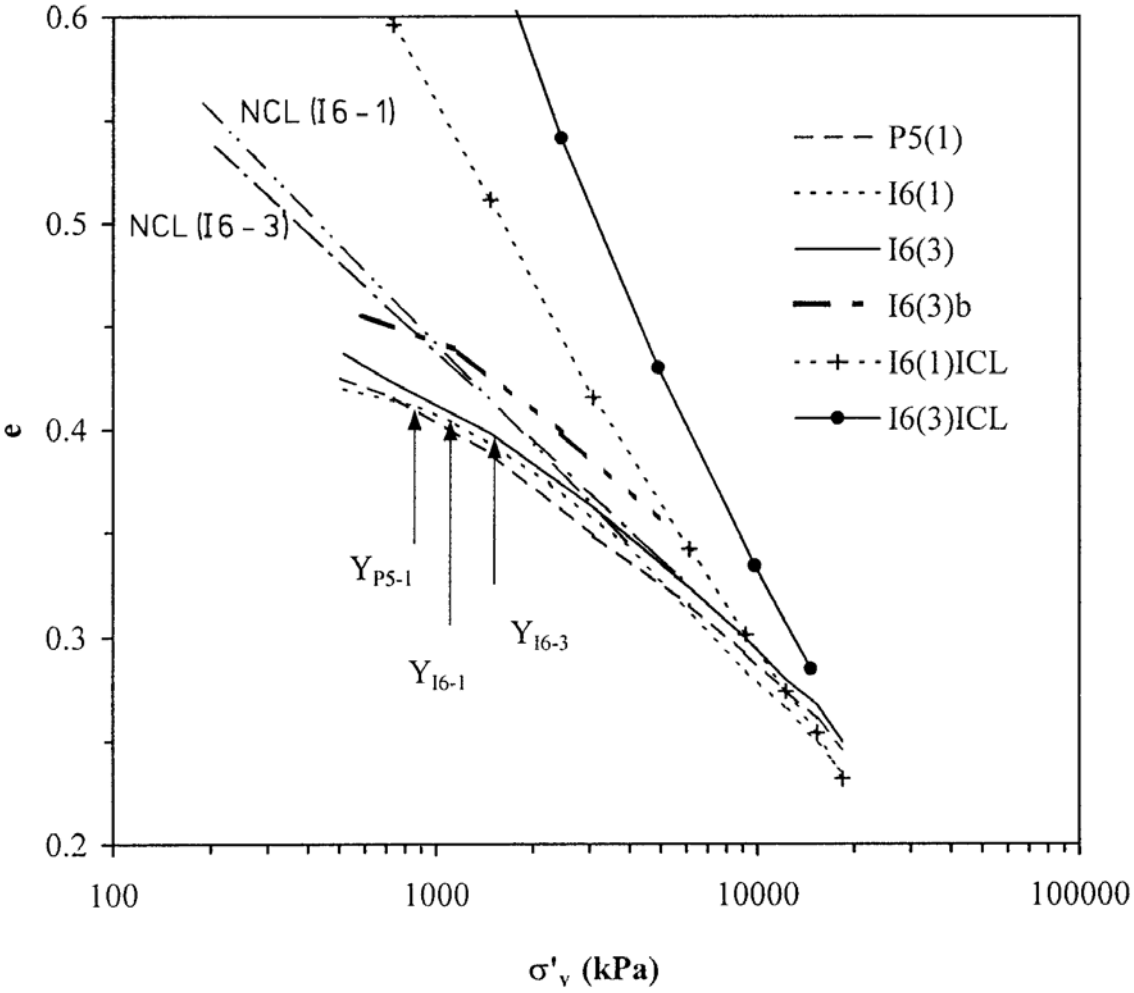

Fig.10. Natural and reconstituted samples [I6(1): 16.7 m; I6(3): 28.35 m; P5(1): 19.7 m]: oedometer test results.

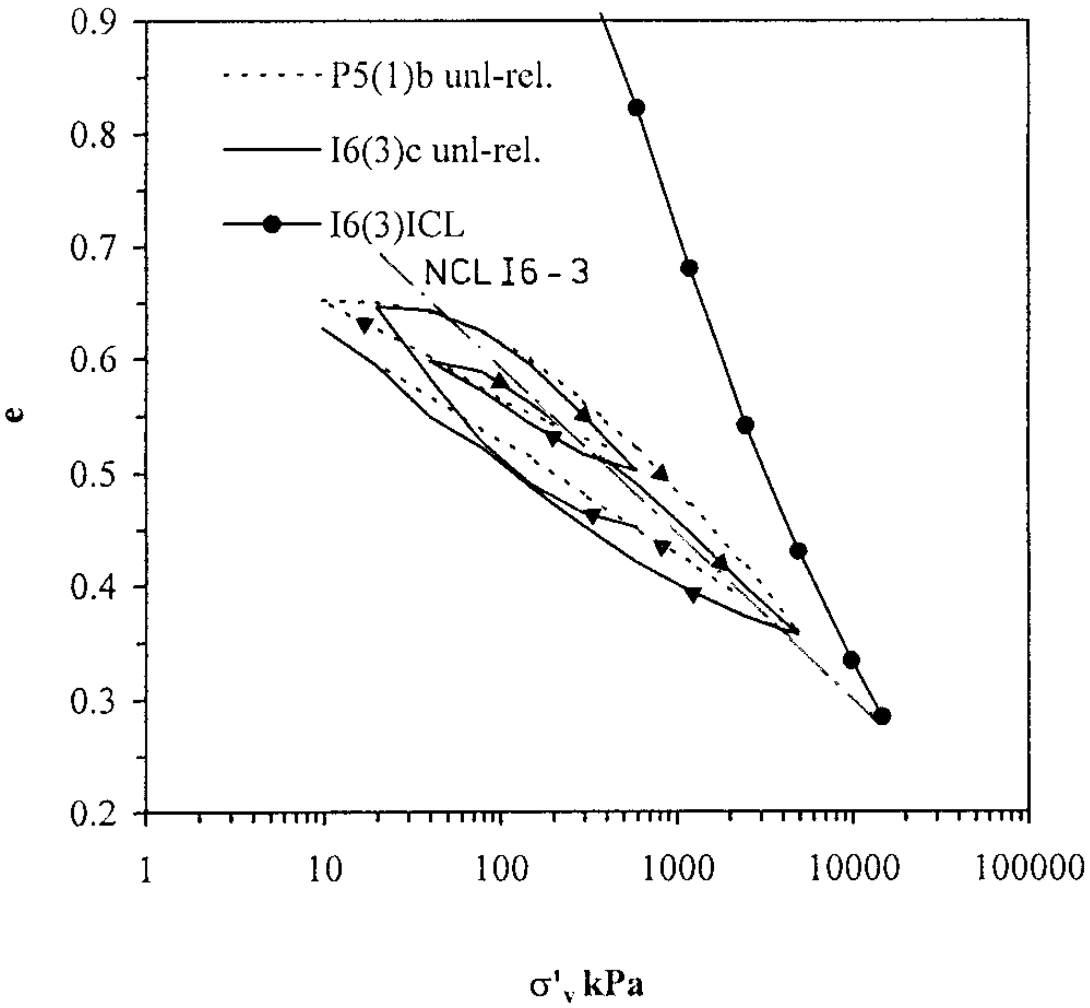

Fig.11. Unload-reload oedometer test results.

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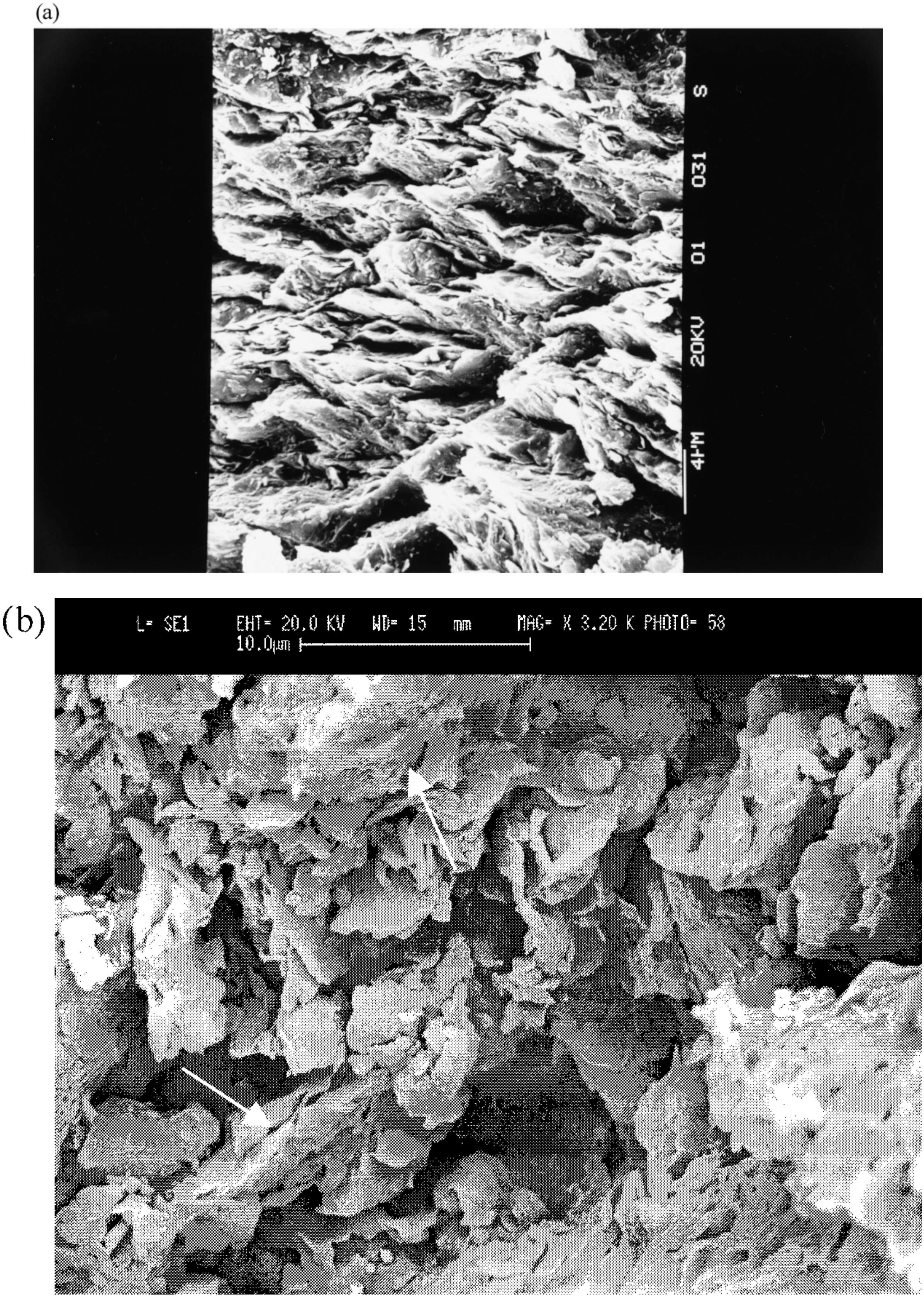

Fig. 12. SEM pictures of fabric on vertical fractures: (a) natural sample I6(1) and (b) reconstituted sample.

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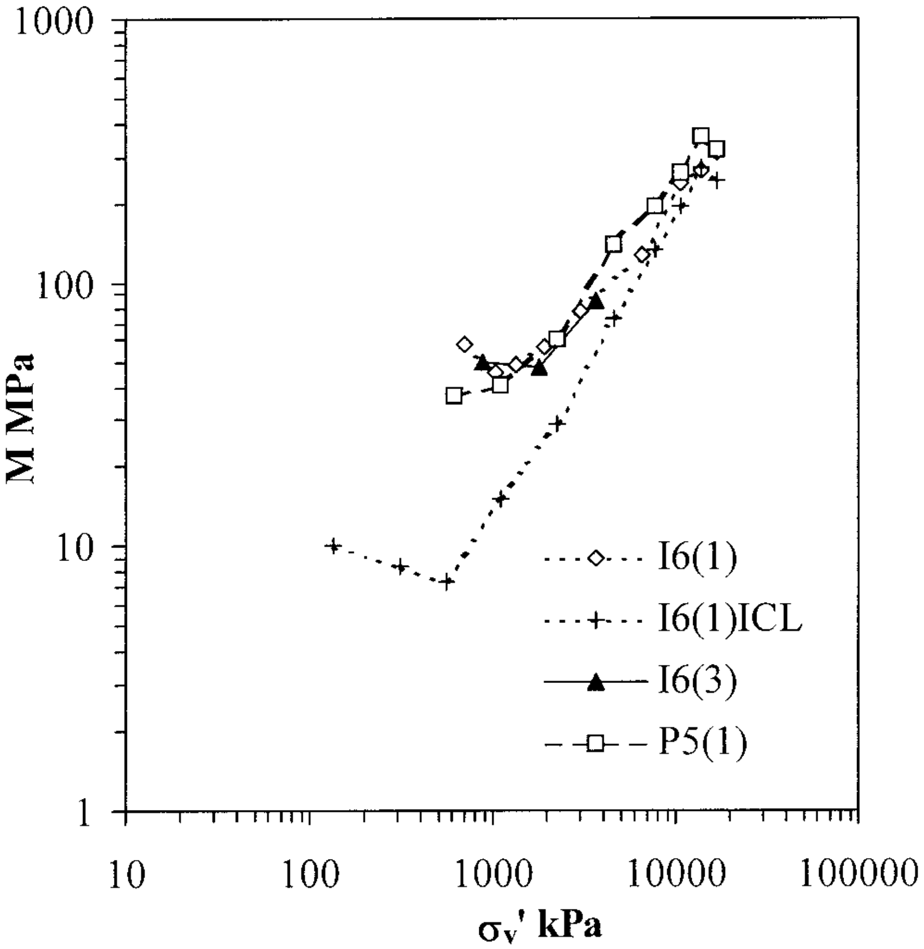

Fig.13. Oedometric stiffnesses of natural and reconstituted samples in Figure 10.

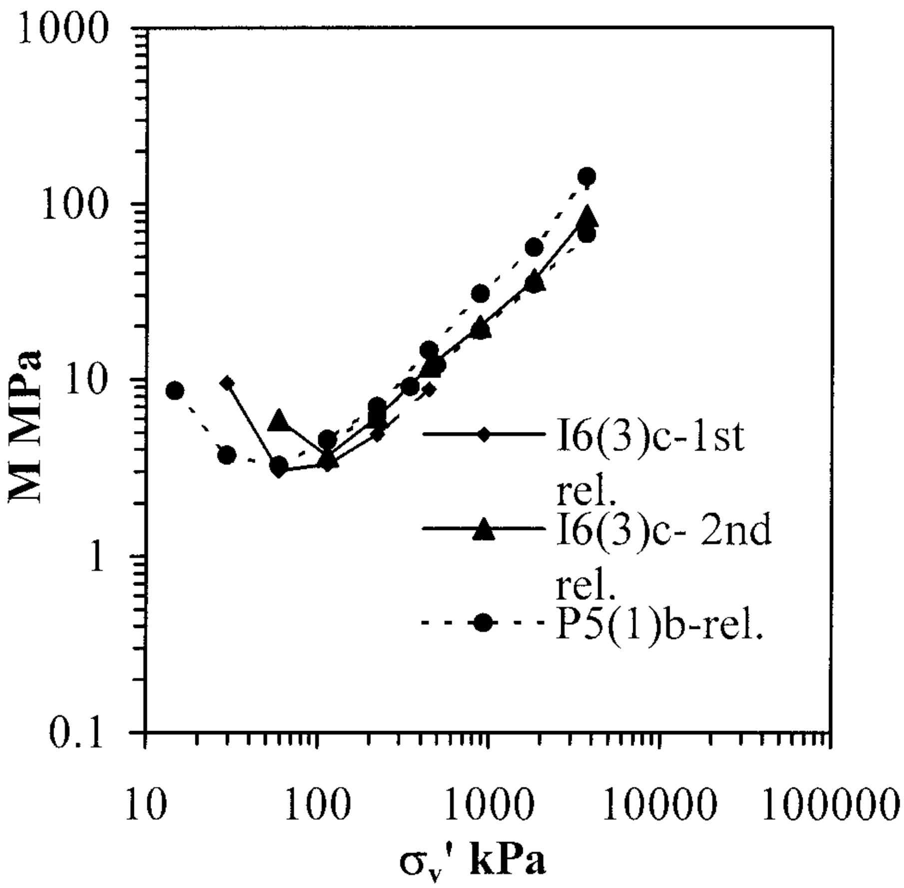

Fig.14. Oedometric stiffnesses of natural samples during reloading after swelling.

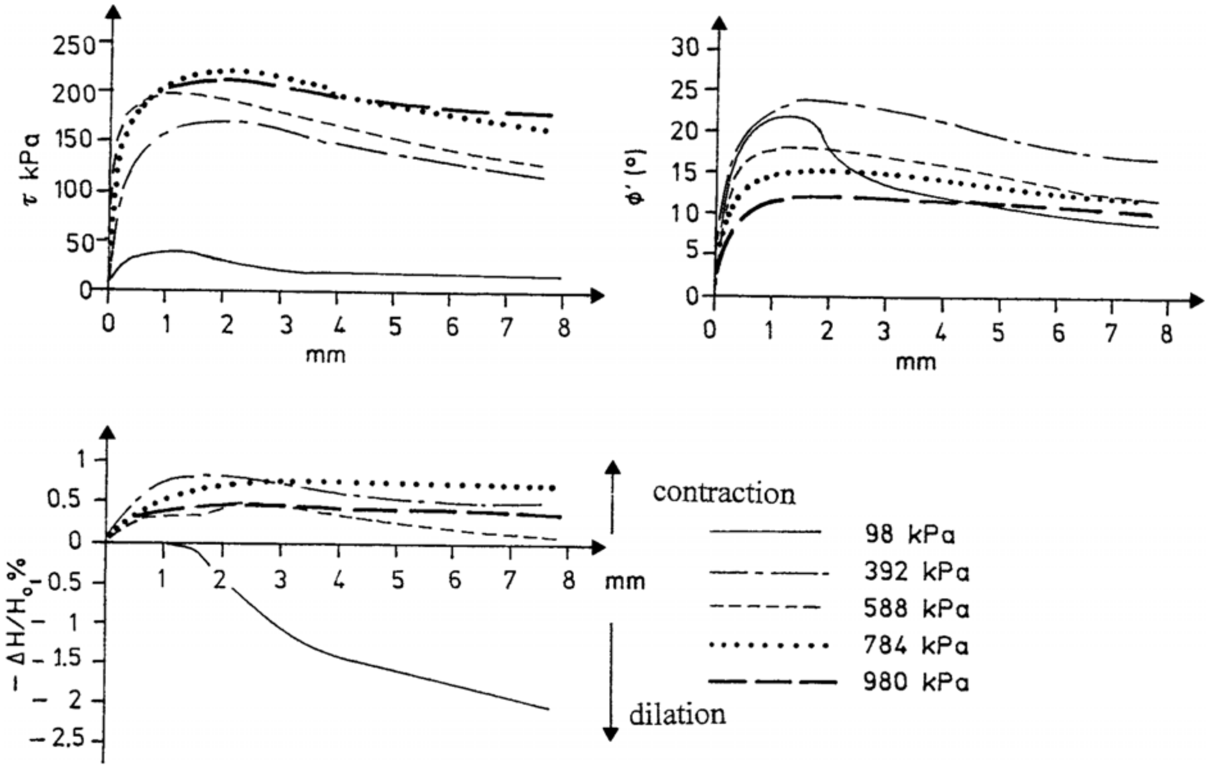

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Fig.15. Results of direct shear tests on sample I6(3).

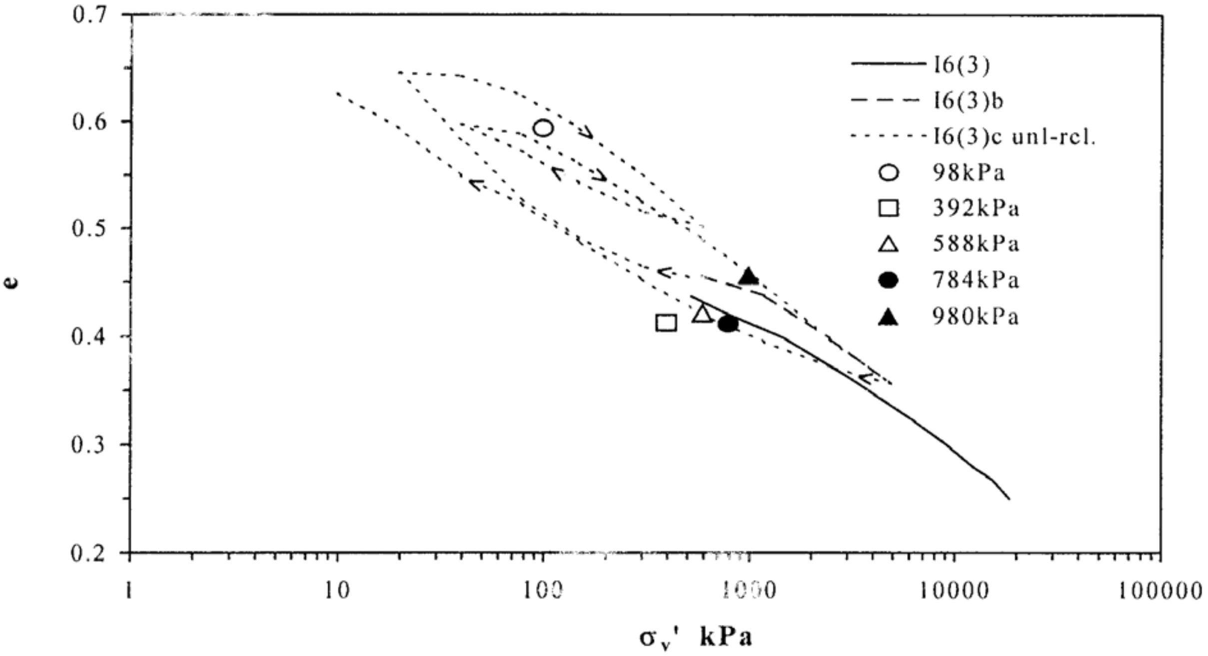

Fig.16. The consolidation states of direct shear test specimens, I6(3), with the compression curves from both unload-reload and restrained swelling high-pressure oedometer tests.

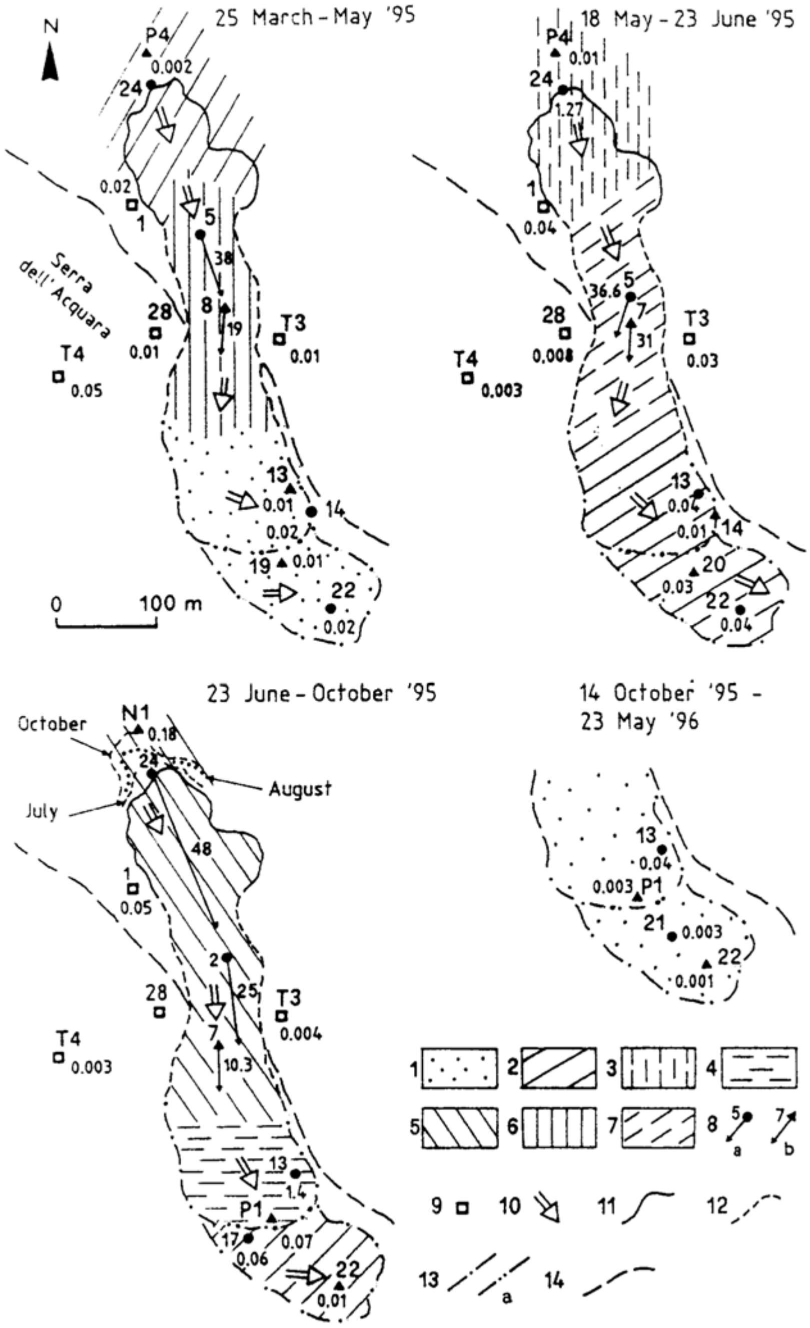

Fig. 17. Horizontal displacement rates (m/month). Keys: 1, 0÷0.02; 2, 0.02÷0.2; 3, 0.2÷0.4; 4, 0.4÷10; 5, 10÷20; 6, 20÷30; 7, > 30; 8, maximum (a) and minimum (b) rates; 9, topographic control stations outside the Vadoncello slope; 10, main displacement directions; 11, area A; 12, area B; 13, area C (a, earthflow accumulation zone; b, part of Serra dell'Acquara deposit); 14, side-border of the Serra dell'Acquara landslide body.

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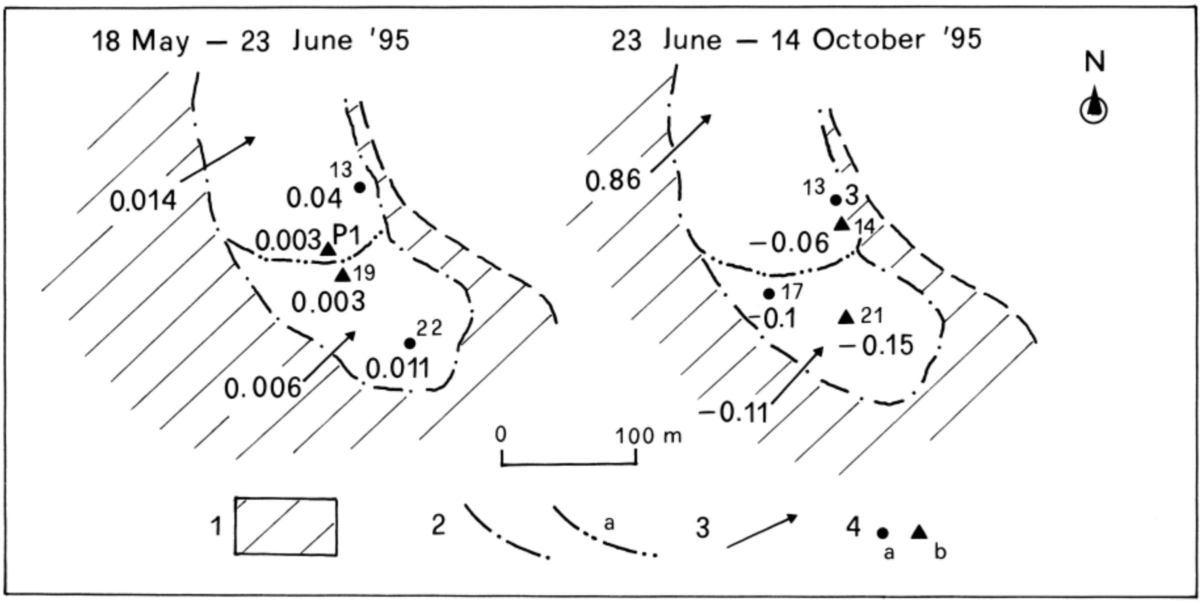

Fig. 18. Vertical displacements (m) within the lower part of the Vadoncello slope. Keys: 1, Serra dell’Acquara landslide; 2, area C (a, earthflow accumulation zone; b, part of Serra dell’Acquara deposit); 3, average vertical displacements; 4, maximum (a) and minimum (b) displacements.

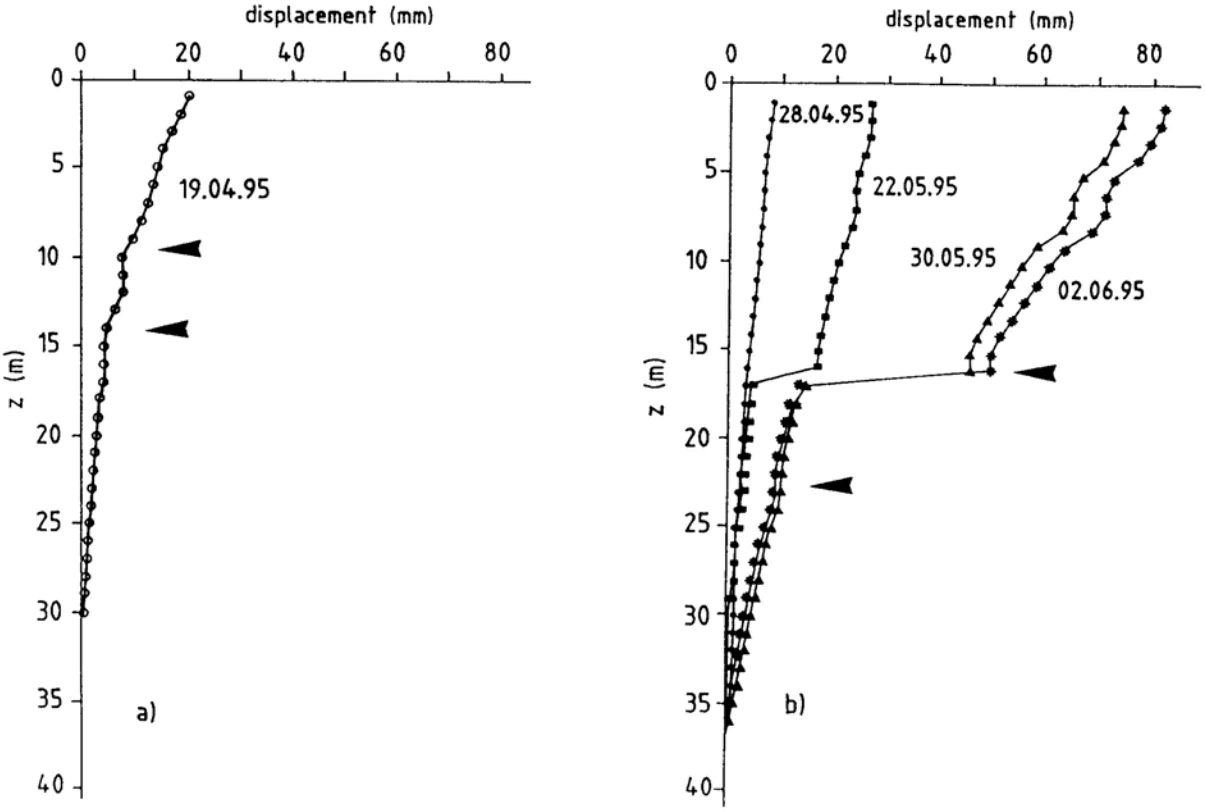

Fig. 19. Inclinometer readings: I4 (a) and I1 (b).

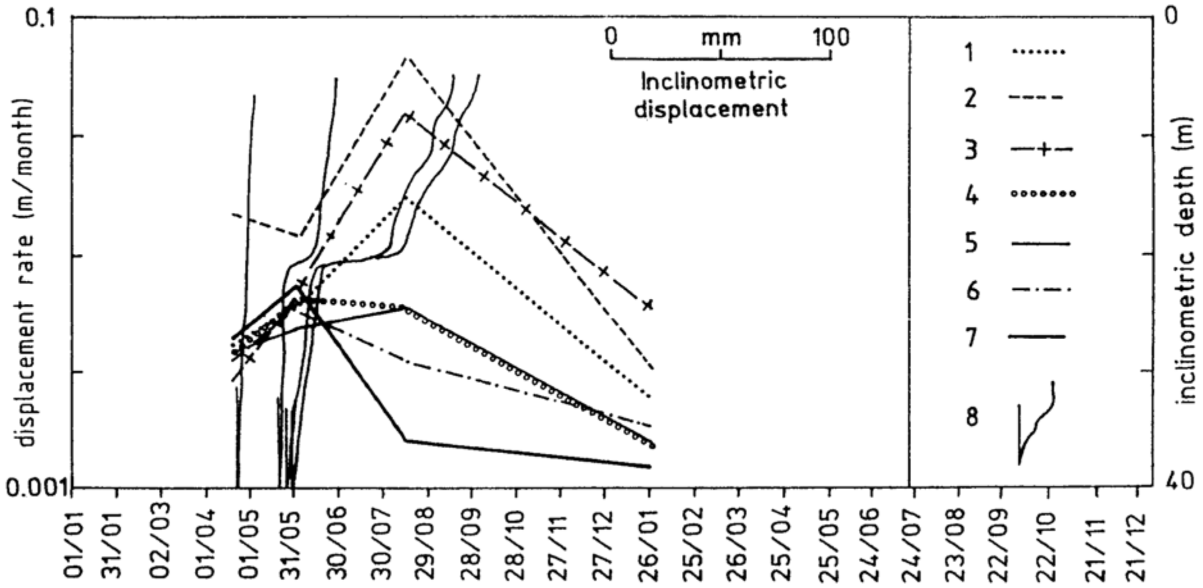

Fig. 20. Horizontal superficial and deep displacements in area C of the Vadoncello slope. Keys: stations of topographic network 1, P1; 2, D15; 3, D13; 4, D19; 5, D20; 6, D21; 7, D22; 8, I1 inclinometer readings.

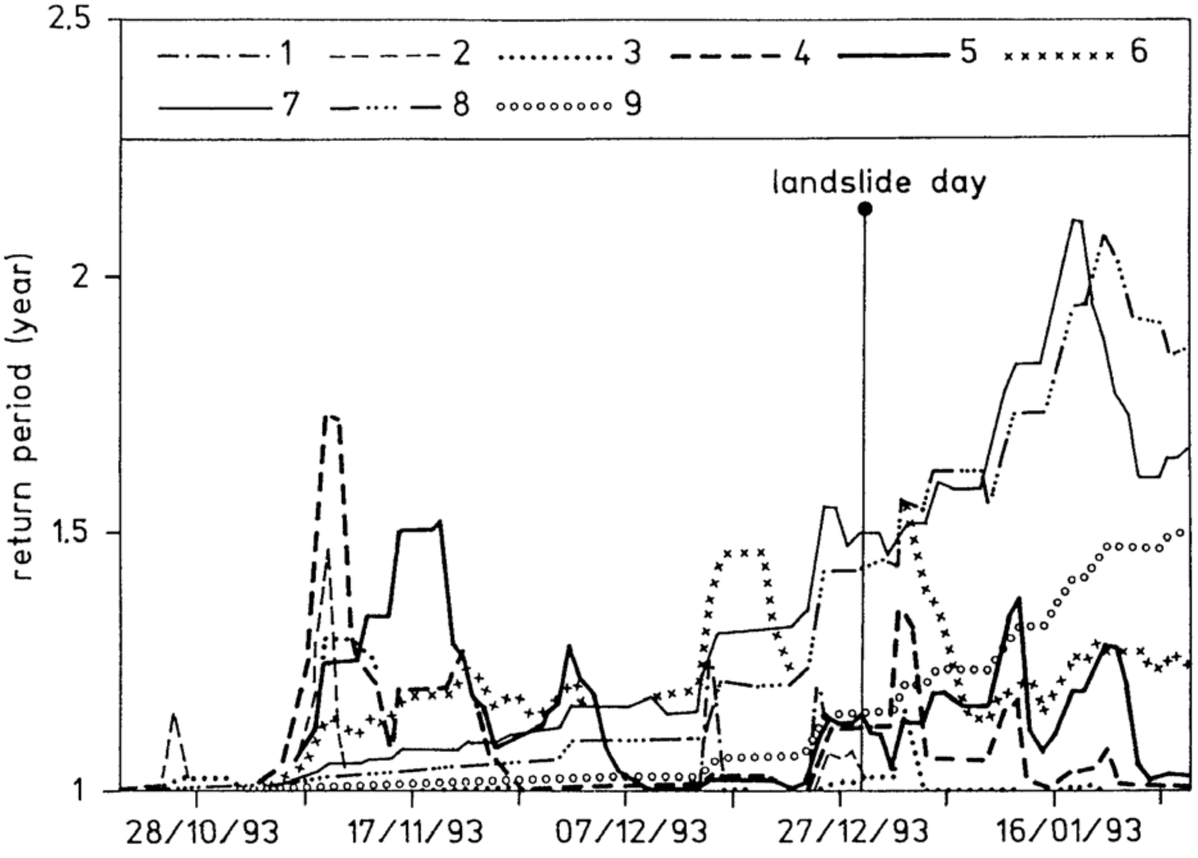

Fig. 21. Return periods of cumulative daily rainfalls: 1) 1, 2) 5, 3) 10, 4) 20, 5) 30, 6) 60, 7) 90, 8) 120, 9) 180 days of cumulative rainfall. The return period is calculated using the generalised extreme value distribution function.

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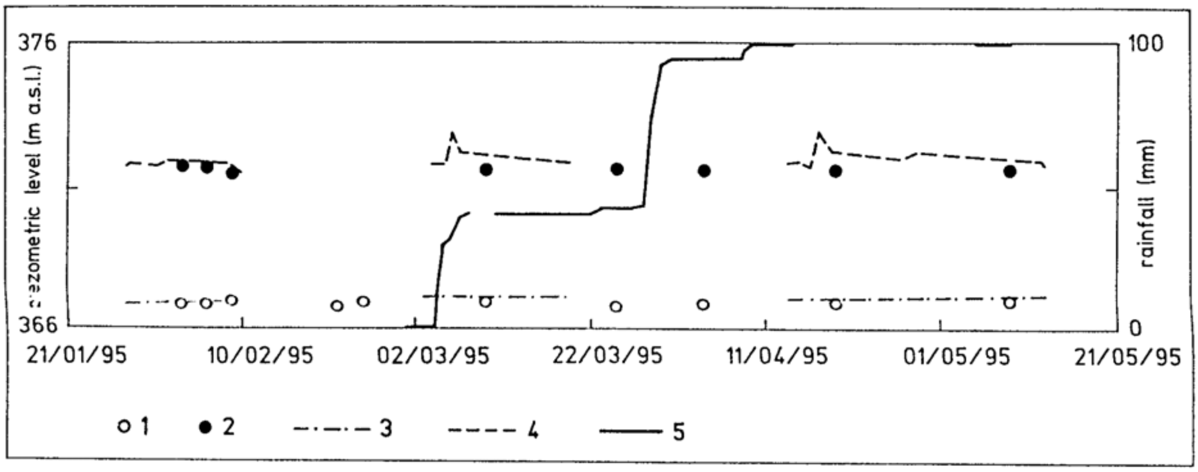

Fig. 22. Piezometric levels logged by means of piezometers in borehole P2 and rainfalls. Keys: 1, Casagrande cell CC1 (15.5 m depth); 2, Casagrande cell CC2 (7.8 m depth); 3, Electric cell EC1 (15.3 m depth); 4, Electric cell EC2 (7.3 m depth); 5, cumulative rain.

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TABLE 1. Summary of the borehole instrumentation installed during EEC Research Project

Borehole	Depth (m)	Inclinometer casing (m)	Piezometer type and depth (m)
P5bis	52.0		O.P.
P5	52.7	--	O.P.
I6	40.2	40.0	--
I5	36.0	30.0	--
P4	57.3	--	C.C. - 12.0 C.C. - 18.8 E.C. - 28.3 C.C. - 34.0 E.C. - 46.0
I4	35.0	33.0	--
I3	22.0	20.6	--
P3	21.5	--	E.C. - 8.3 C.C. - 8.5 E.C. - 15.5 C.C. - 15.8
I2	23.0	17.5	--
P2	32.0	--	E.C. - 7.3 C.C. - 7.8 E.C. - 15.3 C.C. - 15.5
I1	40.1	36.3	--
P6	23.0	--	E.C. - 14.0 E.C. - 20.0
P1	39.6	--	C.C. - 17.5 C.C. - 21.2
P7	22.0	--	O.P.
I7	33.0	30.0	

O.P. = open pipe, C.C. = Casagrande cell, E.C. = elettric piezometer

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TABLE 2. Summary of the laboratory data

Sample	z	w	LL	LP	PI	LI	S	γ	γ_d	e
	(m)	%	%	%	%		%	KN/m^3	KN/m^3	
I2/D1	4.20		61	30	31					
P3/I1	4.58	22.0	52	25	27	-0.111		19.8	16.23	0.659
I2/D2	6.00		48	27	21					
I4/D1	7.10		58	30	28					
P2/I1	7.65	18.4	49	23	26	-0.177	97	20.4	17.23	0.563
I4/I1	7.90		58	33	25					
P3/D1	8.80		60	25	35					
I2/D3	8.90		43	20	23					
I2/I1	9.35	19.1	55	27	28	-0.282	94	20.5	17.21	0.564
I2/D4	11.35		55	33	22					
I2/I5	12.20	19.4						20.8	17.42	0.546
I2/I5	12.20	18.7					96	20.9	17.61	0.529
P3/I3	12.40	23.8	66	28	38	-0.111	100	20.2	16.32	0.650
I2/D5	13.15		67	33	34					
I3/D3	15.70		56	22	34					
I4/I2	16.19	17.1	60	29	31	-0.384	92	21.7	18.53	0.453
I2/D6	16.60		65	35	30					
I6/I1	16.70	18.4	57	24	33	-0.170				
I6/I1	16.70	14.4				-0.292	93.7	21.66	18.94	0.422
I1/D1	19.50		35	20	15					
P5/I1	19.70	16.0	50	27	23	-0.478	100	21.76	18.76	0.436
P5/I1	19.70	19.4				-0.331	96.3	20.38	17.07	0.577
I5/I2	20.35	17.3	77	33	44	-0.357				
I1/I1	21.20	19.0	72	30	42	-0.262	96	20.9	17.56	0.533
P4/I3	22.40	17.7	64	29	35	-0.323	100	21.6	18.35	0.467
I1/D2	24.95		57	27	30					
I6/I3	28.35	16.0	66	25	41	-0.220	100	21.72	18.72	0.438
I1/D3	28.65		25	16	9					
I1/D4	31.30		33	22	11					
I1/D5	33.50		45	22	23					
I6/I4	34.25	16.8						21.5	18.41	0.463
I1/D6	38.15		33	18	15					
P4/I5	40.65	18.7	62	25	37	-0.170	98	20.7	17.44	0.544
P5/I3	52.35		83	29	54					