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DEBRIS FLOW WHICH OCCURRED IN THE APUAN ALPS (TUSCANY – ITALY) DURING THE RAINFALL EVENT OF 19TH JUNE 1996

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Debris flows which occurred in the Apuan Alps (Tuscany - Italy) during the rainfall event of 19th June 1996

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INTRODUCTION

The Apuan Massif, which forms the northern limit of the Apennines, is characterized, from a climatic point of view, by a high rainfall rate; this is due to its geographic and morphological configuration. The rain gauges situated at higher levels record values above the mean 3000 mm per year.

The rainfall of 19th June 1996 was however of an exceptional magnitude, much greater than the events previously recorded in the area. The precipitation caused immense damage, destroying villages and infrastructures situated close to the valley bottoms of the area affected.

The devastating action of the phenomenon was aggravated by the enormous quantity of sediments, which were transported by the streams. The considerable mass of debris was not only due to the erosive processes which have affected some channels but, above all, it was due to the transportation of mobilized masses caused by the many debris- flow which have occurred on the two slopes of the Apuan chain (Versilia and Garfagnana).

The aim of this paper is to identify, by analysing the morphological and pluviometric data of the phenomenon, the limit equilibrium conditions which, when exceeded, trigger to the landslide typology under examination.

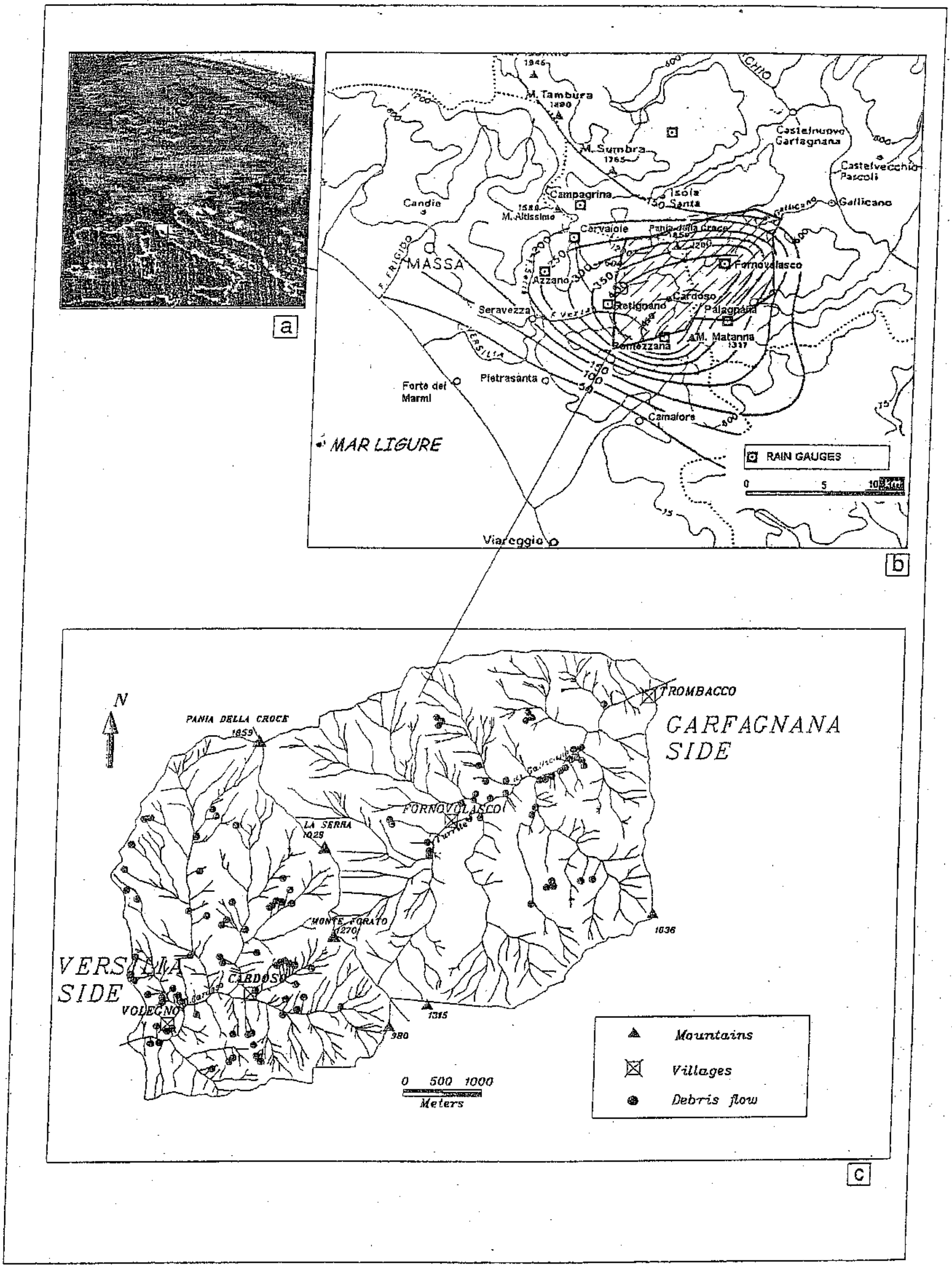

Fig. 1 – a) Meteosat image during the storm; b) The Apuan area with isohyet curves (total millimetres) (Rapetti, Rapetti - 1996) and rain gauges; c) debris flow locations in the study area (in the West side we can see the Versilia basin and in the East side the Garfagnana region)

THE STORM EVENT OF 19TH JUNE 1996

The considerable number of rain gauges allows us to obtain a good detailed reconstruction of the rainfall event, which affected an area of about 700 km^2 in all. On about 60 km^2 of this area, exceptional precipitation values were recorded , as shown in Fig.1, where the meteosat image, the Apuan area and the studied basins are reported.

The phenomenon is characterized by a high space-time variability: in fact the precipitation begins on the Versilia side, here it reaches its maximum intensity at around 6 a.m..

In the early afternoon, after a relative pause of a few hours, the phenomenon starts up again in an intense and widespread manner, and in this case the maximum intensity values occur in the high basin of Turrite di Gallicano River (in the East side Garfagnana Fig.1).

The rainfall event is well showed by cumulative rain curves of the three gauges of Pomezzana, Retignano and Fornovolasco (Fig.2), the closest ones to the studied area.

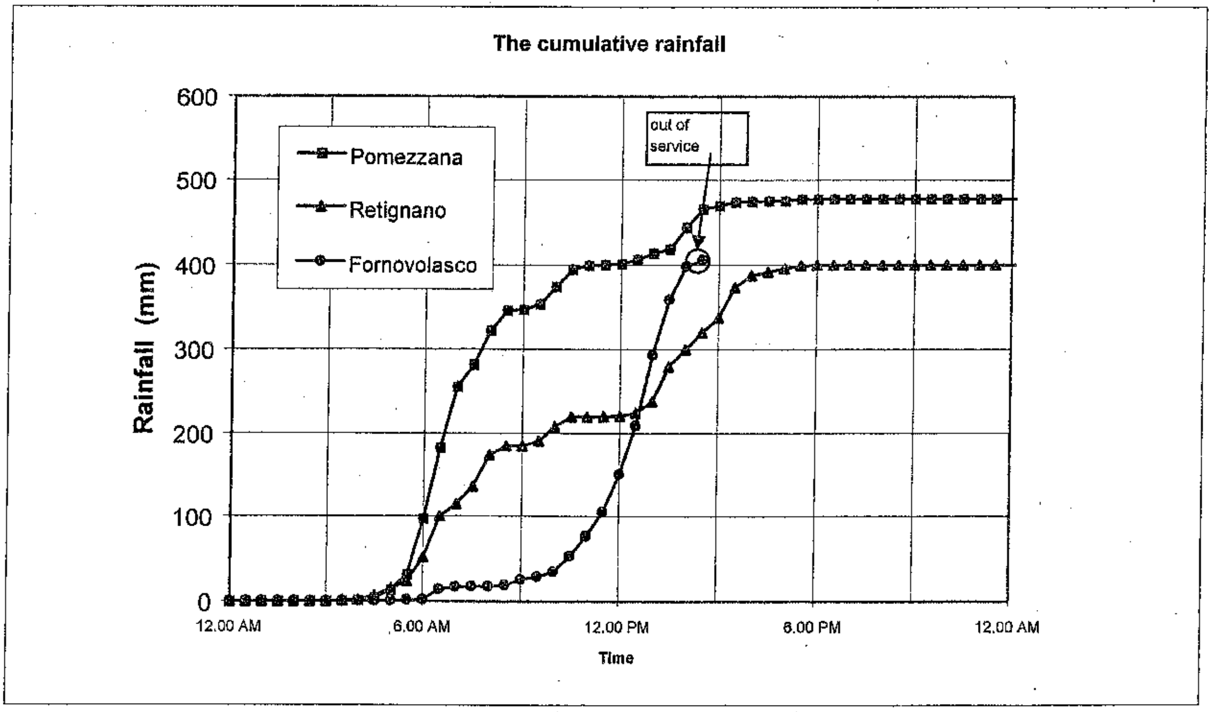

Fig 2 – Cumulative rainfall curves.

- The first rains, initially of moderate intensity, occur at 4 a.m. on 19th June, they already involve almost the entire area later to be affected by the heavy storm.
- The first "extreme" data are recorded at 6.00 a.m.: Pomezzana gauge (in the West side Versilia basin Fig.1) records the maximum precipitations of the morning, with a total of almost 300 mm of rain from 6.00 a.m. to 9.00 a.m. Retignano gauge (situated about 5 km

NW) records a homogeneous and substantially synchronized rainfall graph, albeit with lower intensity peaks.

- 10.00 a.m.: after a brief pause the rain starts up again and is highly intense throughout the entire area; the Garfagnana side also begins to report significant rainfall.
- 12.00 a.m.: the widespread rain continues and once again becomes of an exceptional intensity in the area of the high basin of Turrite di Gallicano river. Before breaking down, from 11.30 a.m. to 2.00 p.m. Fornovolasco gauge records about 300 mm of rain, with peaks of 84 mm in half an hour.
- The rain stops at 6.00 p.m.

The absolute exceptionality of the event is confirmed by the statistical analysis carried out with the rainfall time series. The usual statistical analysis carried out using Gumbel's model of probability shows very high return times for the phenomenon, within the limits deriving from the reduced range of the time series drawn up. Fig. 3 shows the comparison between the data on the event and the climatic probability curve, corresponding to return times of 100 and 1000 years obtained for the Fornovolasco gauge. From this comparison we can deduce that the most critical rain duration is between three and six hours.

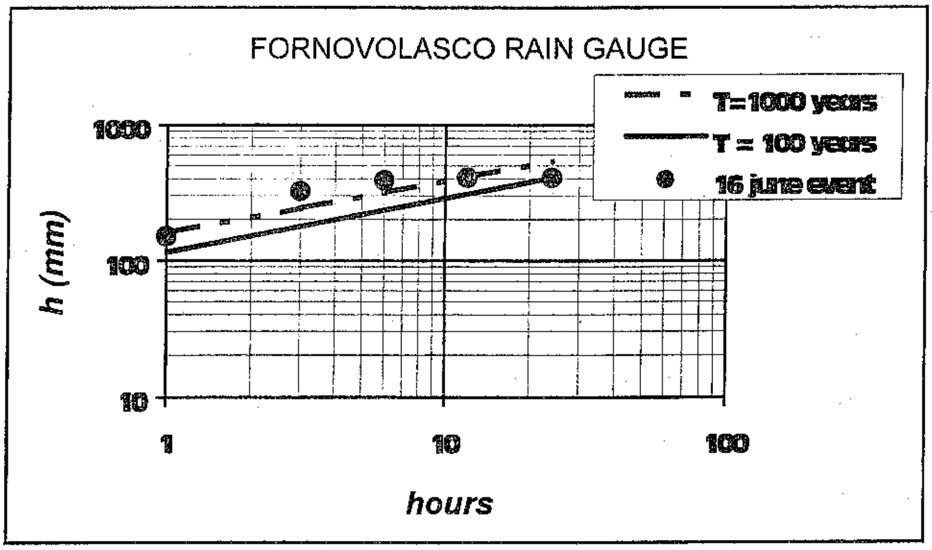

Fig 3 – Fornovolasco rain gauge – statistic relations

The event cannot be attributed to the scenarios which normally cause important rainfall in the area. We had not the transit of a defined front (the precipitation developed and ended in an area with a relatively high and levelled pressure), and we had not a convective phenomena triggered off by radiant heating (the phenomenon was triggered off at 4.00 a.m.).

Recent studies (Castelli et al. - 1998) identify the meteorological cause of the phenomenon with the presence of a "pocket" of hot air, which triggers off convective phenomena of extraordinary intensity when it goes over the Apuan chain. The meteosat satellite image shows how the development of the phenomenon was clearly limited to the Versilia-Apuan area (cf. FIG. 1).

GEOMORPHOLOGICAL OUTLINE OF THE STUDY AREA

This paper analyzes the debris-flows which occurred in one area of the Apuan Alps which was particularly affected by the heavy storm. It covers an area of 26.8 km^2 composed of the upper portion of the catchment basins of the Cardoso and Turrite di Gallicano streams. The boundary between the two basins is formed by the ridge of the Apuan Chain with NW-SE direction; the coastline lies about 35 km from the ridge.

The area has a very complex geology, given the presence of several superimposed tectonostratigraphic units. The deeper seated units have been affected by metamorphic phenomena.

The lithological variability and the geologic structure as well, condition the morphology of the area. The higher peaks which have very steep slopes, are formed by marls. At lower heights, the Pseudomacigno and Verrucano Formations predominate. These are composed of alternations of micaceous sandstones and schists, and of mica-schists and phyllites with quartzites, respectively. The slopes are often asymmetric for structural reasons (the highest gradients can be observed in correspondence with the slopes with layering dipping into the slope).

The landslide susceptibility of the area depends, albeit indirectly, on the geology, in that the latter has conditioned the morphology. Almost all the landslides (about 90%) affected thin debris layers covering a bedrock with dipslope bedding attitude (Pseudomacigno and Verrucano formations) and with prevalent gradients ranging from 30° to 40°. The scree is practically absent in high gradients, while gradients of less than 25° imply morphological conditions which are incompatible with instability phenomena due to debris flow.

THE DEBRIS FLOW PHENOMENA DURING THE PRECIPITATION OF 19TH JUNE 1996

In the study area a census was taken of 157 landslides in total (102 in the western side and 55 in the eastern side), with a mean density on the territory of 6 landslides per square kilometre (Gambetta Vianna, 1998; Leonasi, 1998)). The landslide distribution on the study area is shown in fig. 1. The main geometric characteristics of the landslides were measured (Gambetta Vianna, 1998; Leonasi, 1998).

In most cases in the scarp of the landslide the bedrock is exposed. Nearly all the landslide areas are covered with high trunk coppice (chestnut). The tree trunks were obviously transported towards the valley with the debris mass.

The landslides occurred mainly in concave areas or small impluvia. Many landslides occurred immediately downstream from roads and paths which facilitated the concentration of the runoff waters in the points in which the landslide were triggered off.

The phenomena can be classified as soil-slip debris-flows. The movements started with the slipping of the soil and the debris on the substratum, the fluidified mass then descended rapidly, flowing along depressed areas and mobilizing more material along the sides of the channel. The thickness of the materials involved in the movement varies from 0.5 to 2.0 metres and the predominant ones is 1.5m. This range of thickness is according with other experimental data (Ellen and Flemming,1987; Guadagno,1991, Iotti and Simoni,1997). Figure 4 shows frequency of landslides according to four thickness classes (41 cases).

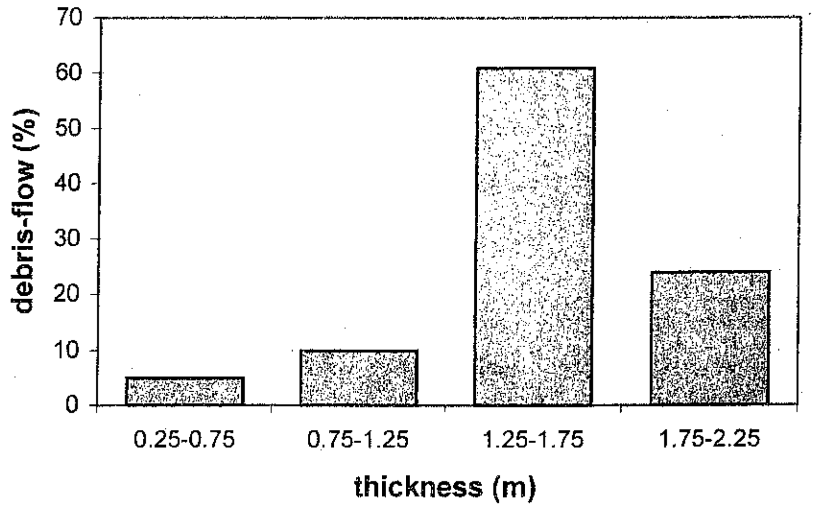

Fig 4 – Debris flow thickness distribution

Among the characteristic data collected in these studies, the classification of the landslides on the basis of the slope of the terrain is of particular interest.

The slope classes adopted correspond to those indicated by Campbell. (1975), as the phenomena described in the above study were found to have considerable similarities.

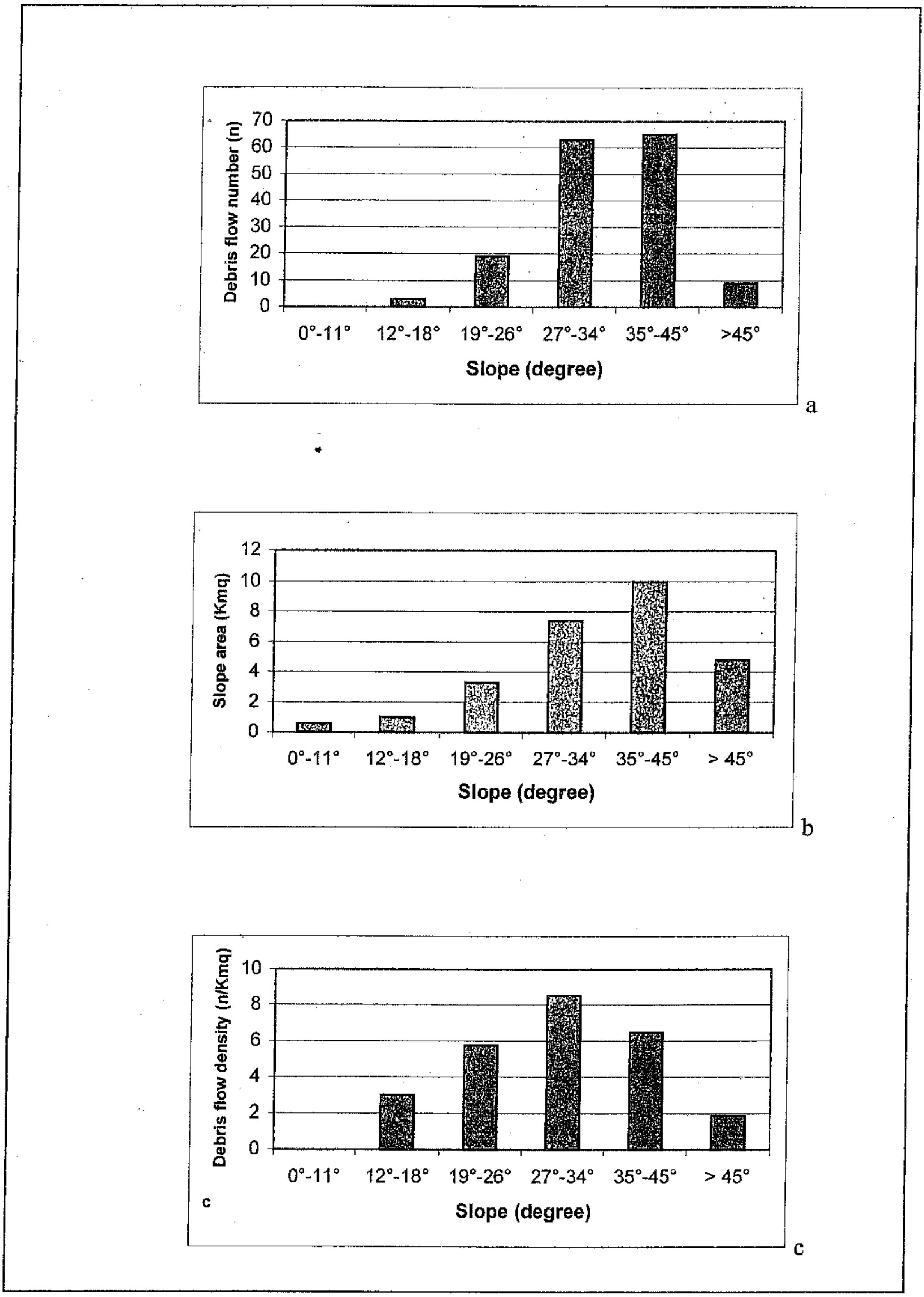

Fig.5 - Debris flow related to slopes

In order to characterize the slope influence on the phenomena, we took into consideration the variability of the morphology of the territory, connecting the landslide frequency with the respective frequency of the relative slope class.

The overall result is given in figure 5 ,where the number of landslides (fig.a), the areal extensions (fig.b) and the number of landslides per km^2 (fig.c) are represented for each slope class.

From this elaboration the distribution curve takes on a symmetric configuration and the greater vulnerability of the slopes included in the class 27° - 34° , with respect to the other classes considered, is clear. This result is in accord with previous works on debris flows (Guadagno., 1991), and also on the other types of landslides (Carrara et al. 1977).

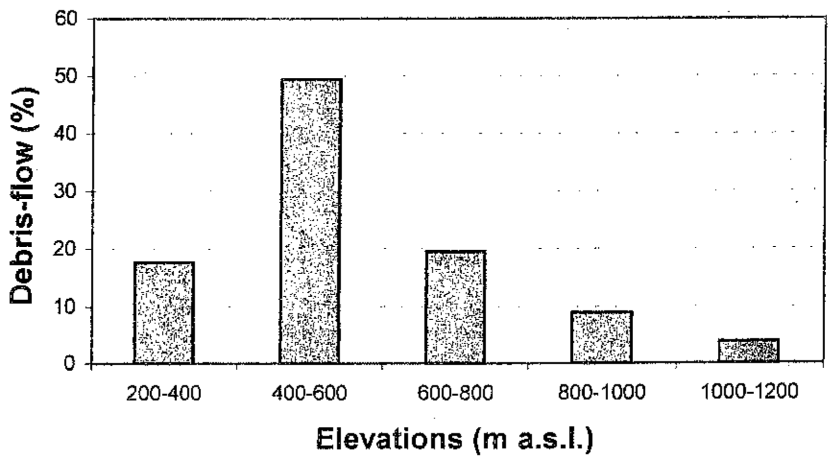

Fig.6 - Debris flow related to elevations

Finally figure 6 shows the number of landslides for each altitude range: we have the highest landslide frequency in the 400 - 600 metre range, and not at the greater heights as we would expect if the landslides were only conditioned by rainfall intensity. The greater landslide susceptibility of the 400 - 600 metre elevation range can be attributed to the correspondence between this class and the slope of the terrain which is the most critical. Furthermore we can take it that, in the higher altimetric ranges, the rainfall did not have markedly higher intensities than the ones measured at lower heights.

GEOTECHNICAL PARAMETRIZATION

Samples were taken from a total of 27 landslides in order to characterize the materials affected by debris-flows from a geotechnical point of view. The material was taken from the detachment niche at the intermediate height between the topographic surface and the slip surface. Grain size analyses using sieving and sedimentation apparatus were carried out on all the samples, as well as consistency limit tests.

The experimental results are shown in figures 7 and 8. The diagrams highlight the fact that the material is composed mainly of sand and gravel with low percentages of silt, while the clayey fraction is low. The fraction passing through the 40 a.s.t.m. sieve has liquidity limits ranging from 35 to 52 and plasticity index values which vary from 10 to 19. The material has grain size similar to other debris flow which moved rapidly (Ellen and Fleming, 1987).

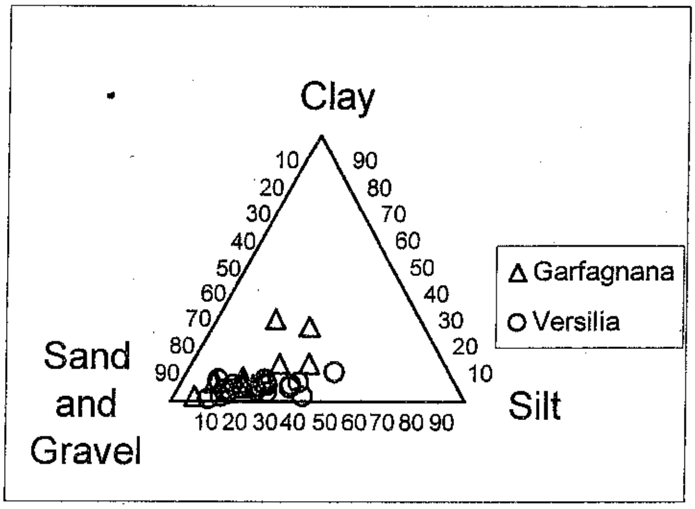

Fig 7 - Grain size (triangular diagram)

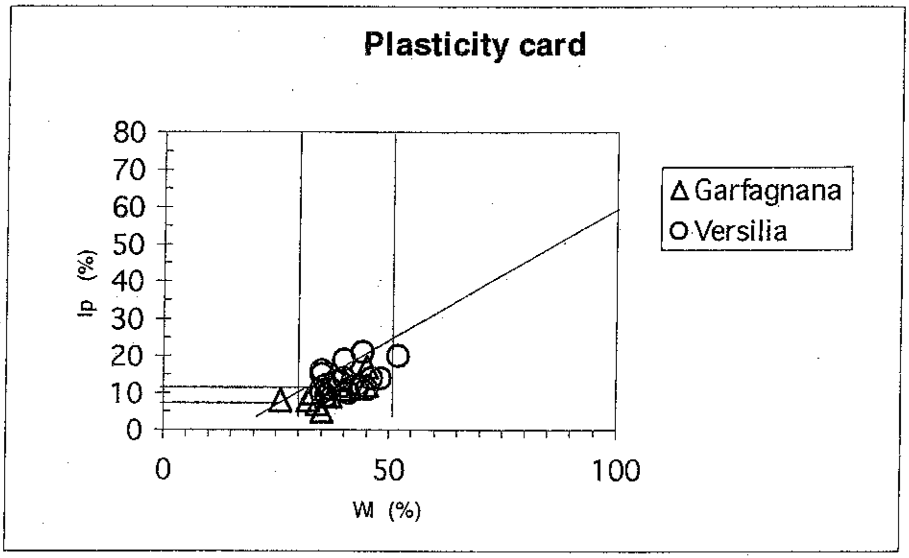

Fig 8 – Plasticity chart

Consolidated and drained direct shear tests with Casagrande equipment were carried out on 9 samples deriving from different landslides. The specimens were reconstituted with material which passed through the 40 A.S.T.M. sieve and they underwent CD testing after saturation.

With the exception of one sample with a finer granulometry, which gave an angle of internal friction $\varphi'= 26°$, the others gave friction angles ranging from 31° to 39°, with a mean value $\varphi'_m = 37°$ and a coesion c'= 0 kPa.

STABILITY ANALYSIS

Given the triggering mechanism of the movement and the geometric conditions, the stability analysis was carried out using the "indefinite slope method", (assuming the thickness to be constant and disregarding the side effects).

In order to estimate the in situ cohesion , due to various causes such as capillarity phenomena and effects induced by plant roots, back analysis was carried out calculating , for a fixed φ' value, the cohesion values corresponding to the safety coefficient Fs=1 for slope gradients varying from 20° to 50° for different depth of water table. The figure 9 shows the curves calculated adopting a friction angle $\varphi' = 37°$, unit weight $\gamma = 19$ KN/m^3 and $\gamma_s = 21$ KN/m^3 , respectively over and down the water table, and thickness of the landslide material h=1.50 m.

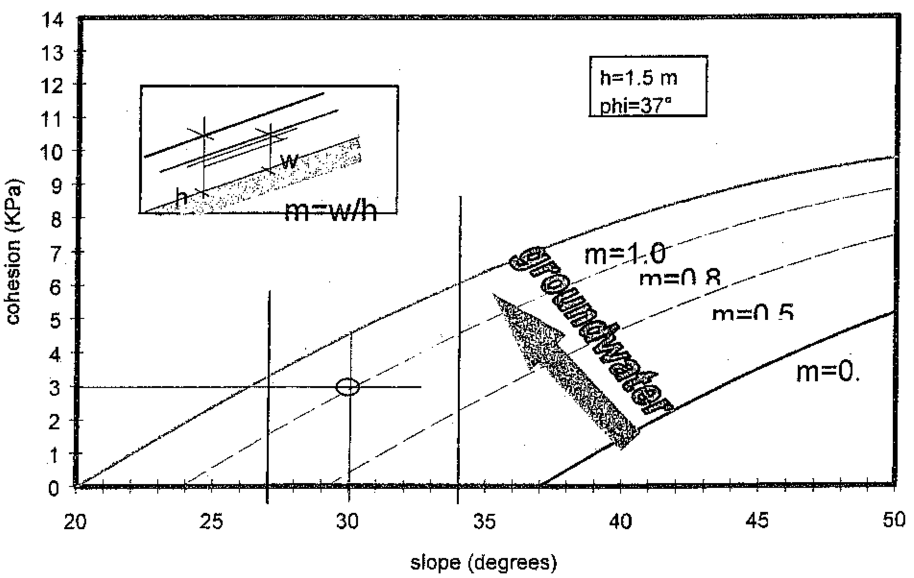

Fig 9 – Back analysis: stability for different water table heights

The area between the curves indicates the cohesion values which must be attributed to the soil in order to obtain the limit equilibrium condition in the various rising stages of the piezometric surface.

If we assume a low cohesion c'=3 kN/mq, for roots and structural soil effects, we obtain limit equilibrium conditions in a wide range of slope gradient values (26°- 44°); for the two limit values of this range, instability occurs for the conditions corresponding to no water table and water table at ground level, whereas for intermediate values the instability varies depending on the ratio between the height of the water table and the landslide thickness.

Stability analysis on a landslide with mean characteristics, i.e. slope gradient β=30° and thickness h=1.50 m, and adopting the geotechnical parameters indicated above (cohesion c'=3 kPa, internal friction angle φ'=37°, unit weight γ = 19 KN/m^3) give a safety coefficient Fs= 1.55; the Fs value drops to 1 when the ratio between the water table height and the landslide thickness rises to m=0.80.

THE RAINFALL EFFECTS

As described in the previous paragraph, the instability is due to the rise of the water table above the critical heights. The high permeability of the soils affected, as well as the precipitation history (there was practically no rain in the days prior to the event), allow us to assume that there was no hight water table conditions (prima falda) at the beginning of the phenomenon.

The critical height h_c, which corresponds to the water table height necessary to bring the safety coefficient below 1, depends on the thickness of the landslide material. In the geometric conditions indicated in the previous paragraph, corresponding to the most frequent cases, h_c= m x h = 1.20m.

Assuming a porosity n = 0.30, supposing a pre-existent saturation degree S= 20%, we obtain a unit water volume in critical equilibrium condition of

$$Vw=0.29\ m^3/m^2=290\ mm$$

which corresponds to about 70% of the mean precipitation which affected the study area (over 400 mm everywhere).

The same value definitely appears high if compared with the usual infiltration coefficient of deep forested slopes. We must however take into consideration the fact that the mobilized materials have grain sizes which give medium-high permeability and that the phenomena occurred in particular morphological conditions which favoured the local feeding of the aquifer.

CONCLUSIONS

- The event analyzed is definitely exceptional, given the fact that it has a statistical recurrence of several hundred years for the rainfall durations of 1 -3 - 6 hours.
- This rainfall rate has proved to have a destabilizing effect for debris-flow type phenomena with thicknesses of less than 2 metres, whereas it has not triggered off other slides phenomena.
- The model studied gave results which are compatible with the phenomena surveyed.

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