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Italian Agency for Environment
Protection and Technical Services

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*Since October 6, 2002, the **Italian Agency for Environment Protection and Technical Services (APAT)** has been operative. This new institution incorporates the former Italian Agency for Environment Protection (ANPA) and the Technical Services of the Presidency of the Council of Ministers, (Geologic, Hydrographic and Mareographic Services).*

*According to the directives, and under the supervision of the Ministry for the Environment and Territory, **APAT** will keep on carrying out all the technical-scientific tasks for which it has been responsible so far, consisting in the protection, monitoring and control of the environment, soil and water protection, prevention of technological risks and nature conservation.*
*So far as its institutional duties are concerned, **APAT** will continue being a reference point in terms of collaboration, consultancy, assistance, service and support to the Public Administrations, as defined by relevant agreements.*

In the framework of a consolidated environmental network, it will be always responsability of the Agency, to work towards the integration of the Informative System into the SINA network, in which both the National Cartographic System and the Environmental Regional Informative Systems (SIRA) will have to be integrated.

***APAT**'s targets, priorities and resources will be defined by a triennial activity program, updated on a yearly basis, according to the directives coming from the Ministry for the Environment and Territory.*

***APAT**'s bodies consist in the General Director (assisted by a consultive Committee) and the Board of Revisors; the Agency is divided into Departments and interdepartmental Services. A new feature is represented by the institution of a Federal Council, headed by the General Director and composed of legal representatives of the Regional and Provincial Agencies for Environment Protection (the ARPAs and APPAs), and one representative of the State-Region Conference.*

*Within this new framework, the correctness, reliability and integrity of the data and technical surveys performed by the Agency's experts, which always characterized the institutional reports and documents previously produced by ANPA, will be improved by **APAT**, in such a way as to contribute to the transparency and authoritiveness of the environmental information.*

The General Director
Giorgio Cesari

Preface

In the framework of the APAT (the Italian Agency for Environment Protection and Technical Services) activities dealing with the prevention and reduction of hydro-geologic risks, between September 21 and 27, 2001, a field workshop, titled "Seismically Induced Ground Rupture and Large Scale Mass Movements" has been held.

The organization of the event was carried out in strict co-operation with the Department of Geological Sciences of "Roma Tre" University, the Department of Earth Sciences of Camerino University, the Department of Chemical, Physical and Mathematical Sciences of Insubria University of Como, IRPI-CNR in Cosenza, and with the patronage of INQUA (International Union for Quaternary Research).

People from universities, research institutions and professionals of Italy, U.S.A., Australia, Japan, Venezuela, France, Sweden, Belgium, Poland and Egypt have taken part in the event.

With the purpose of providing the convenors with guidance, a "Field Trip Guide Book" was produced. A significant improvement of this book has resulted in the present document, representing a useful tool for understanding the structural-geological framework of the central and southern Apennines, as well as the relationship between neotectonics, seismic features and gravitational phenomena which characterize this mountain belt.

This document is, therefore, intended to provide a contribution to the knowledge of Italy's physical environment which, due to its relatively young geologic and geomorphologic framework, is particularly subjected to natural phenomena such as earthquakes and landslides that are potentially catastrophic in terms of human consequences, unless an adequate awareness of their evolutive dynamic processes is raised.

The General Director
Giorgio Cesari

Itinerary of the field trip (September 21-27, 2001)

1. Geological and tectonic setting of the Appennines and Calabrian arc

N. D'Agostino[1], F. Dramis[1] and P. Molin[1]

1. Geological and tectonic setting of the Apennines and the Calabrian Arc

Most reviews of the late-Tertiary evolution of the Tyrrhenian-Apennines system emphasize the eastward migration during the Neogene of paired extensional (in the west) and compressional (in the east) belts, together with flexural subsidence of the Adriatic foredeep and volcanism, all of which are envisaged as responses to the `roll-back' of the subducting Adriatic-Ionian lithosphere (Fig. 1, Elter *et al.*, 1975; Royden, 1993; Serri *et al.*, 1993; Malinverno and Ryan 1996; Faccenna *et al.*, 1996; Jolivet *et al.*, 1998). Thus the progressively eastward-younging foredeep and syn-rift deposits in the Apennines record the coeval activity and migration of the paired compressional-extensional belts in the Neogene (Patacca *et al.*, 1990). However, during the Quaternary the flexural subsidence, compressional deformation and eastward retreat of the subduction hinge all decreased dramatically (Patacca *et al.*, 1990; Kruse and Royden, 1994; Cinque *et al.*, 1993) and the Apennines became dominated by crustal extension and vertical movements. Seismic reflection profiles in the Adriatic Sea show that the Mesozoic-Cenozoic sequence which is deformed by thrust anticlines is in turn overlain by prograding Quaternary deltaic sequences fed by streams draining the eastern flank of the Apennines, with little evidence of compressional deformation after the Early Pleistocene (Dondi *et al.*, 1985; Ori *et al.*, 1993; Argnani *et al.*, 1997). This Quaternary depositional pattern marks a dramatic change in subsidence rate and sediment supply from the Pliocene, during which up to 7,000 m of sediment accumulated in a flexural trough close to the thrust front (Bigi *et al.*, 1992). The Quaternary evolution thus involves the final infilling and extinction of the Mio-Pliocene Adriatic foredeep (Ori *et al.*, 1993) and a regional NE tilting of the whole Adriatic coastal belt of central Italy (Dufaure *et al.*, 1989; Dramis, 1992; Kruse and Royden, 1994).
Evidence of present-day thrusting is contained in weakly deformed and tilted Quaternary deposits (Bigi *et al.*, 1997) and moderate compressional seismicity on the NE side of the Northern Apennines (Frepoli and Amato, 1997) where intermediate seismicity down to a depth of 90 km may suggest that subduction is still active (Selvaggi and Amato 1992). In the area of the central Apennines that is currently undergoing extension, normal faulting has been active since the Upper Pliocene-Early Pleistocene (Patacca *et al.*, 1990; Bosi and Messina, 1991). Normal faults cut a bedrock sequence dominated by resistant Mesozoic limestones and

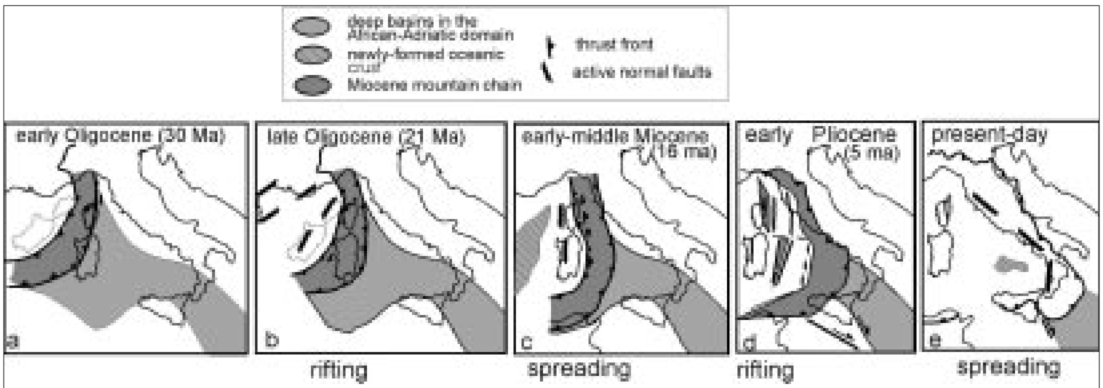

Fig. 1 – Simplified scheme of the evolution of the Tyrrhenian-Apennines system (modified from Faccenna *et al.*, 1999).

[1] Università degli Studi Roma Tre. Largo S. Leonardo Murialdo 1, 00146 Roma

less resistant Upper Miocene flysch that were previously deformed by NW-SE striking Neogene thrust faults (Parotto and Praturlon, 1975; Bigi *et al.*, 1992). This extension is responsible for Pleistocene intermontane basins that are partially filled with alluvial, fluvial and lacustrine deposits and coarse conglomerates or breccias (Cavinato *et al.*, 1993b; Cavinato and DeCelles, 1999).

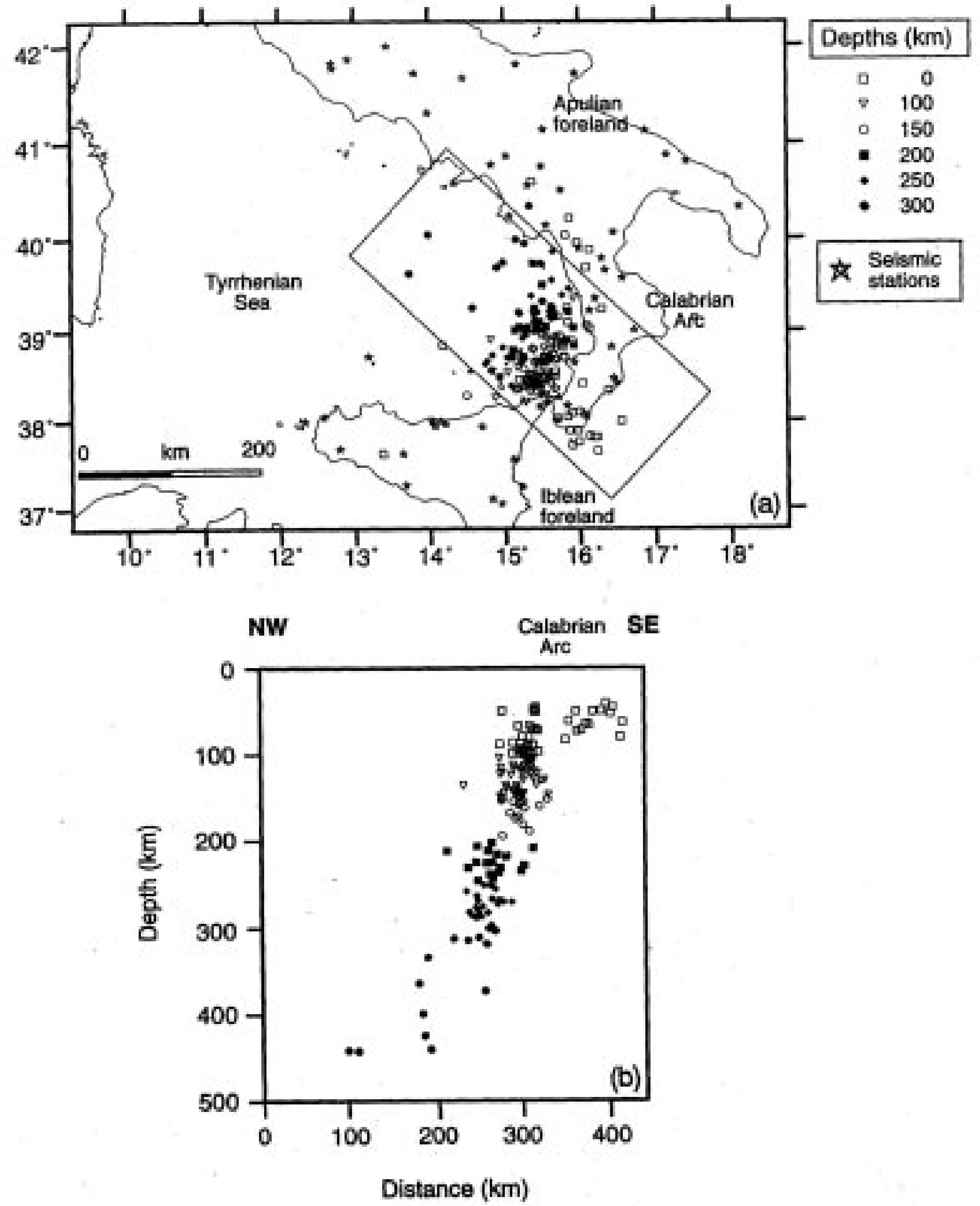

Fig. 2. (a) Hypocentral map of Southern Tyrrhenian seismicity from 1988 to 1993; (b) NW-SE vertical section showing the events located within the box in (a). From Selvaggi and Chiarabba (1995).

More to the south, in the Calabrian arc, the Apennines are a narrow, rugged peninsula, underlain by pre-Alpine plutonic and metamorphic rocks, continental and oceanic metamorphic rocks ("Complesso Ofiolitico", Auct.) overlain tectonically Mesozoic carbonates and flysches (Ogniben, 1973; Amodio-Morelli *et al.*, 1976; Dietrich, 1976; Lanzafame *et al.*, 1979; Critelli, 1990). Similarly to the rest of Italy, since the late lower Pleistocene, the Calabrian arc was affected by extensional tectonics and regional uplift (Vezzani, 1968; Tortorici, 1980, Tortorici, 1981; Ciaranfi *et al.*, 1983; Lanzafame and Tortorici, 1981; Boccaletti *et al.*, 1984; Colella *et al.*, 1987; Moretti, 1993; Tortorici *et al.*, 1995; Moretti and Guerra, 1997) and its present dynamics is well expressed by crustal and subcrustal seismicity (Gasparini *et al.*, 1982; Ghisetti and Vezzani, 1982; Cristofolini *et al.*, 1985; Tortorici *et al.*, 1995; Monaco and Tortorici, 2000). The distribution of epicenters and the relative focal mechanisms indicate the activity of extensional N-S fault systems mainly inland and strike-slip and compressional activity off-shore the Ionian coast (Amato *et al.*, 1997; Frepoli and Amato, 2000; Monaco and Tortorici, 2000). The location of intermediate and deep earthquakes (Fig. 2) allows to define the geometry of the southern Tyrrhenian subduction zone (Amato *et al.*, 1997). The seismicity distribution and tomographic images reveal a continuous 40-50 km thick slab that abruptly increases its W-NW dip from sub-horizontal in the Ionian Sea to a 70° dip in the Tyrrhenian (Gasparini *et al.*, 1982; Kiratzi, 1994; Bruno *et al.*, 1995; Selvaggi and Chiarabba, 1995; Lucente *et al.*, 1995; Frepoli *et al.*, 1996; Piromallo and Morelli, 1997).

2. Geological and geomorphological evidence of Quaternary regional uplift

Various geological and geomorphological data indicate that a widespread surface uplift occurred during the Quaternary (Demangeot, 1965; Ambrosetti *et al.*, 1982; Dufaure *et al.*, 1989; Dramis, 1992), when the tectonics of the Apennines was already dominated by crustal extension. Scattered outcrops of highly dissected marine Messinian—Lower Pliocene conglomerates, weakly deformed by later compressional structures and frequently made up of clasts (granites, metamorphic rocks) whose source area is hundreds of kilometers distant (Accordi and Carbone, 1988), are found in the Apennines at high elevations (>1,500m in the Gran Sasso range), indicating that profound changes in elevation and morphology occurred in the Quaternary. One of the most significant effects of the uplift is found in the Tiber river valley close to the Tyrrhenian coast (Fig. 3). Here an Early Pleistocene shoreline that is continuously exposed for almost 100km has been uplifted to an elevation of 200—400m (Ambrosetti *et al.*, 1987; Alfonsi *et al.*, 1991; Girotti and Piccardi, 1994). This ancient shoreline follows the long-axis of the Apennines and provides an important reference for the evaluation of vertical movements. It shows almost no short-wavelength deformation over its nearly continuous exposure, but smoothly decreases in elevation to the south, suggesting that it was raised by a large-wavelength regional uplift rather than by localized fault activity, which is apparently responsible only for minor local effects (Alfonsi *et al.*, 1991).
Paleontological analyses (Gliozzi and Mazzini, 1998) in the more western parts of the Rieti and Terni basin fills showed the existence of brackish marshes influenced by the contiguous Early Pleistocene Tyrrhenian Sea, showing that the basin floors were approximately at sea-level and that were significantly uplifted after the Early Pleistocene. Remnant Neogene-Pleistocene marine deposits found by Marinelli *et al.* (1993) in the Latium and Tuscany areas on the Tyrrhenian side of the Apennines increase in elevation to the NE, also supporting the suggestion of regional uplift. On the NE side of the Apennines the Adriatic foothills are characterized by a NE-dipping Pleistocene sequence made up of fine-grained marine deposits passing upward into sandy and conglomeratic deltaic deposits at the top of the

Quaternary sequence (Cantalamessa *et al.*, 1986; Ori *et al.*, 1993). From the geomorphological point of view the whole Adriatic coastal region of central Italy displays a homogeneous evolution in the Quaternary, characterized by a regional NE tilting that is clearly reflected in the NE-trending parallel drainage network (Demangeot, 1965; Dufaure, *et al.*, 1989; Dramis, 1992). Furthermore, transitional fluvial to marine Lower-Middle Pleistocene deposits at the top of the Quaternary sequence are uplifted up to 500 m some kilometers inland of the coastline (Fig. 3; Cantalamessa *et al.*, 1986; Bigi *et al.*, 1995), giving

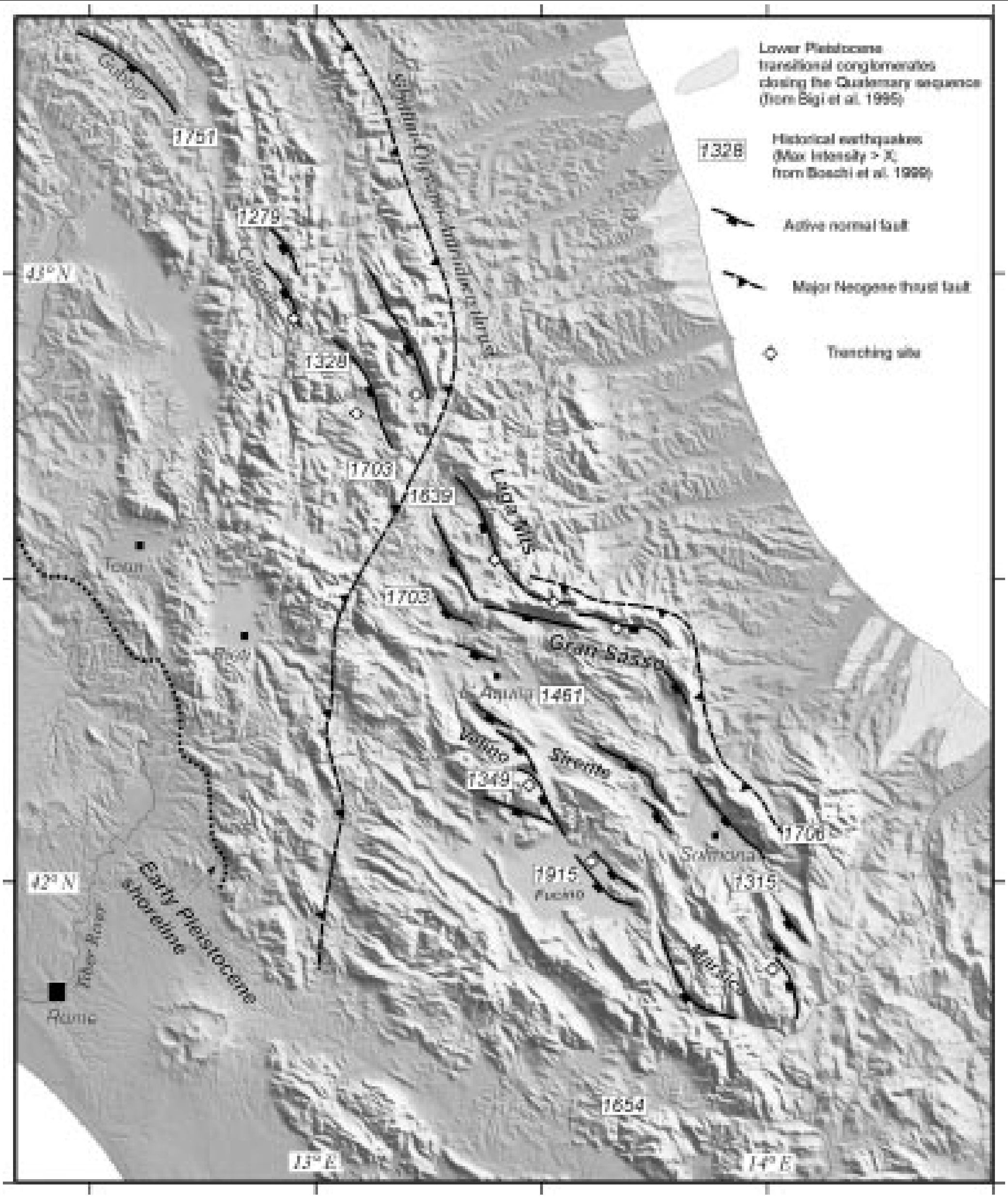

Fig. 3. Shaded relief map with faults active in the Late Pleistocene-Holocene from Galadini and Galli (2000), major Neogene thrust fronts and historical earthquakes. Soft shading on the Adriatic side of the Apennines represents Lower Pleistocene transitional conglomerates closing the Quaternary sequence. On the Tyrrhenian side the trace of the Early Pleistocene shoreline is also shown.

an indication of the amount of uplift on the Adriatic side. A precise evaluation of regional uplift rates in the Middle-Late Pleistocene is hampered by the relative scarcity of remnant shoreline deposits and uncertainties in their ages, especially on the Adriatic coast. Nevertheless, approximate long-term uplift rates have been derived by Bigi *et al.*, (1995) obtaining values ranging between 0.3 and 0.5 mm/yr over the last 1 Ma. Similar values (0.1-0.26 mm/yr) can be estimated for the long-term uplift on the Tyrrhenian side on the basis of the elevation (200—400 m) of the Early Pleistocene shoreline (1.5-2 Ma). These rates are likely to be higher along the central axis of the Apennines itself. The geological and geomorphological data summarized above suggest a general doming of the Central Apennines on a wavelength larger than 150—200 km during the Quaternary. A post-700kyr regional bulging of the Southern Apennines is envisaged by Bordoni and Valensise (1999), based on the elevations of marine deposits of `Tyrrhenian' age (corresponding to isotopic stage 5e of 125 ka). A similar Quaternary evolutionary pattern has been also described by Argnani *et al.* (1997) for the Northern Apennines, showing that the regional Pleistocene doming is a process that affected the whole Italian peninsula.
The effects of the regional uplift that affected the whole Italian peninsula in the Quaternary are also present in the Calabrian arc, where in previous subsiding basins the uplift produced a marine regression in the late lower Pleistocene (Tortorici, 1980; Colella *et al.*, 1987; Moretti, 1993). In middle-upper Pleistocene the interaction between the regional uplift and the glacio-eustatic variations generated sequences of both marine and fluvial terraces that allow to compute and uplift rate of 0.6-1.0 mm/yr (Tortorici, 1980; Lanzafame and Tortorici, 1981; Carobene and Damiani, 1985; Carobene *et al.*, 1986; Colella *et al.*, 1987; Colella, 1988a,b; Gliozzi, 1988; Carobene *et al.*, 1989; Carobene and Dai Prà, 1990; Palmentola *et al.*, 1990; Carobene and Ferrini, 1991; Moretti, 1993; Sorriso-Valvo and Sylvester, 1993; Westaway, 1993; Sorriso-Valvo *et al.*, 1996b; Cucci and Cinti, 1998; Mauz and Hassler, 2000; Molin *et al.*, 2001). Another morphological evidence of the broad uplift is a rolling upland of subdued local relief standing at the higher elevations of the present topography (van Genderen, 1970; Dramis *et al.*, 1990; Sorriso-Valvo and Sylvester, 1993; AA.VV., 1995; Gabriele *et al.*, 1997; La Pera and Sorriso-Valvo, 2000; Molin, 2001). This landform is a relic of an ancient landscape that developed before late lower Pleistocene, in conditions of base level stability, of very low tectonic activity or of compensation between surface processes and morphological effects of tectonics (van Genderen, 1970; Dramis *et al.*, 1990; Gabriele *et al.*, 1997; Molin, 2001).

3. Quaternary stratigraphy of the Apennines intermontane basins

The stratigraphy of the Quaternary intermontane basins in the central Apennines basins has been studied closely and synthesized in the studies by Cavinato *et al.* (1993b) and Bosi and Messina (1991) among others. Lithological and paleontological correlations throughout these basins highlight common regional tectonic and climatic controls in addition to the influence of the basin-bounding normal faults. The first deposits associated with the formation of the intermontane basins consist of coarse-grained breccias and conglomerates generally assigned to the Upper Pliocene - Early Pleistocene (Demangeot, 1965; Cavinato *et al.*, 1993b; Bosi and Messina, 1991; Bagnaia *et al.*, 1989). Low-relief surfaces, locally preserved above the major calcareous ridges (Demangeot, 1965; Bosi and Messina, 1991; Dramis, 1992) and occasionally corresponding to outcrops of these early breccia deposits, are significantly dissected and fragmented by erosion and tectonic deformation. A progressive younging in the age of the oldest deposits of the intermontane basins is observed from west to east, with Middle-Late Pliocene sediments at the bottom of the the westernmost basins (Terni and Rieti)

and Early Pleistocene sediments at the bottom of the more eastern ones (Sulmona and Colfiorito; see Fig. 3 for location).The Early Pleistocene is characterized by lacustrine environments in most of the intermontane depressions, recorded by widespread lake beds that are revealed within the incised basin fills (Michetti and Serva, 1990; Bosi and Messina, 1991; Cavinato *et al.*, 1993b) or have been drilled (GE.MI.NA., 1962). The Early-Middle Pleistocene lacustrine deposits are generally overlain by units that are transitional from lacustrine and low-gradient fluvial environments to coarser deposits representative of alluvial fans (Blumetti and Dramis, 1992; Miccadei *et al.*, 1998). This transition is frequently marked by erosion and incision of the lake beds so that the Middle Pleistocene deposits are often entrenched and unconformably overlie the fluvial-lacustrine units. In some cases depositional surfaces are preserved within the lacustrine and fluvial deposits while in others it is more difficult to estimate the amount of incision of the fluvial-lacustrine sequences. After the Middle Pleistocene, deposition of lacustrine sediments in the intermontane basins was drastically reduced and continued only in basins that maintained internal drainage. During the Pleistocene abundant pyroclastic materials derived from alkaline-potassic volcanic centers on the Tyrrhenian coast provide radiometrically datable tephra layers interbedded within the continental sequences (Cavinato *et al.*, 1993b; Miccadei *et al.*, 1998).
In summary, the intermontane basins generally record a history in which the Early-Middle Pleistocene continental fluvial-lacustrine environments were later incised and covered by alluvial fans. This succession is consistent with the capture of internally-draining, fault-bounded basins by the regressive headward erosion of major regional streams cutting down to a lower base level. The changes in facies and environment are unlikely to be climatically induced, since some closed lake basins, such as Fucino, have survived to modern times. Over the Quaternary as a whole, this evolution is also typical, with slightly different timing, of the intermontane basins of the Northern (Argnani *et al.*, 1997) and Southern Apennines (Capaldi *et al.*, 1988).

4. Distribution of active deformation in the Apennines and Calabrian arc

Earthquake focal mechanisms show that the central Apennines are undergoing NW-SE extension, with seismicity concentrated along the main topographic ridge (Anderson and Jackson, 1987; Amato *et al.*, 1997) on active normal faults that overprint earlier compressional structures (Vittori *et al.*, 1997; D'Agostino *et al.*, 1998; Fig. 4).
The total rate of extension across the Apennines is not yet well constrained. Seismic moment summations yield estimates between 0.9 and 3.5mm/yr (Jackson and McKenzie, 1988; Pondrelli *et al.*, 1995) but are likely to be imprecise as the seismicity is relatively low and dominated by a few large events. Analysis of the deformation revealed by the Italian first-order triangulation networks in the interval 1865-1963 yields estimates of 3mm/yr as an upper bound for the extension rate accommodated in the central Apennines (Hunstad and England, 1999). Both earthquake slip vectors and VLBI measurements suggest the motion of the Adriatic relative to Europe can be described by rotation about a pole in northern Italy (Anderson and Jackson, 1987; Ward, 1994), with the VLBI data predicting an extension rate in the Apennines of about 3mm/yr at the 43\degr N increasing southward to 6mm/yr at the latitude of Matera (Fig.5). First GPS estimates of active crustal extension (D'Agostino *et al.*, 2001) show that strain accumulation in the interval 1994-1999 is concentrated in a 40 km wide belt extending at a rate of 6±2mm/yr (1 σ). The low levels of internal deformation in the Adriatic Sea suggest that much of the extension in the Apennines is absorbed by shortening in the Dinarides (Anderson and Jackson, 1987) with some also accommodated in the external part of the Northern Apennines (Frepoli and Amato, 1997). The pattern of active deformation in the Apennines (Figs. 4 and 5) is revealed by the historical and instrumental seismicity and

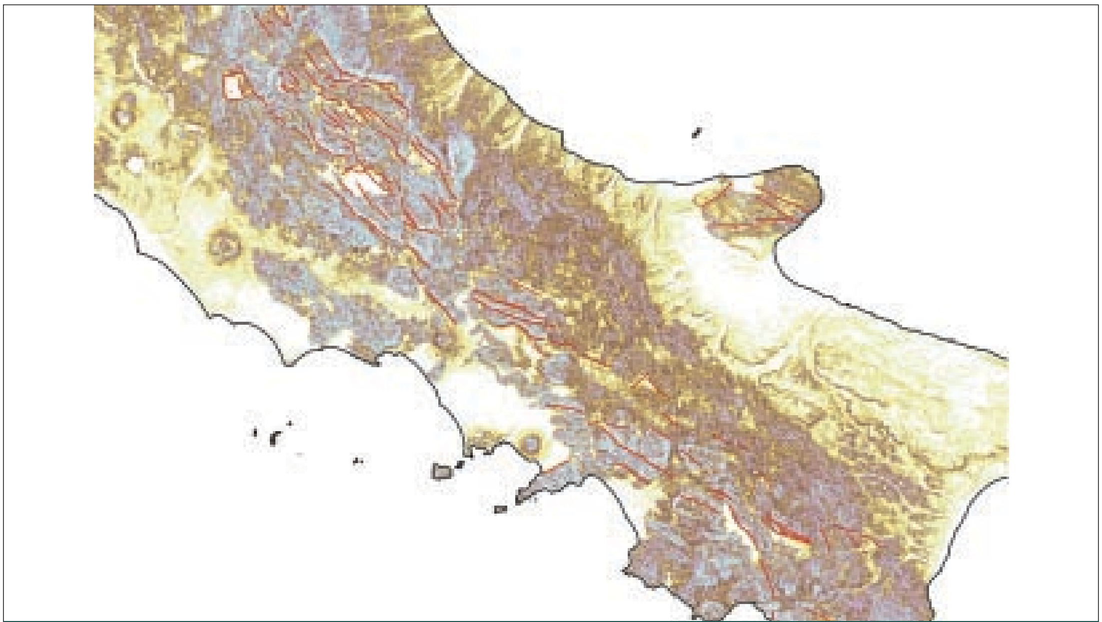

Fig. 4. Quaternary normal faults showing evidence of late Pleistocene to Holocene coseismic surface faulting associated with strong (M > 5.5) earthquakes in Central – Southern Italy (modified from GNDT, 2000; Michetti *et al.,* 2000).

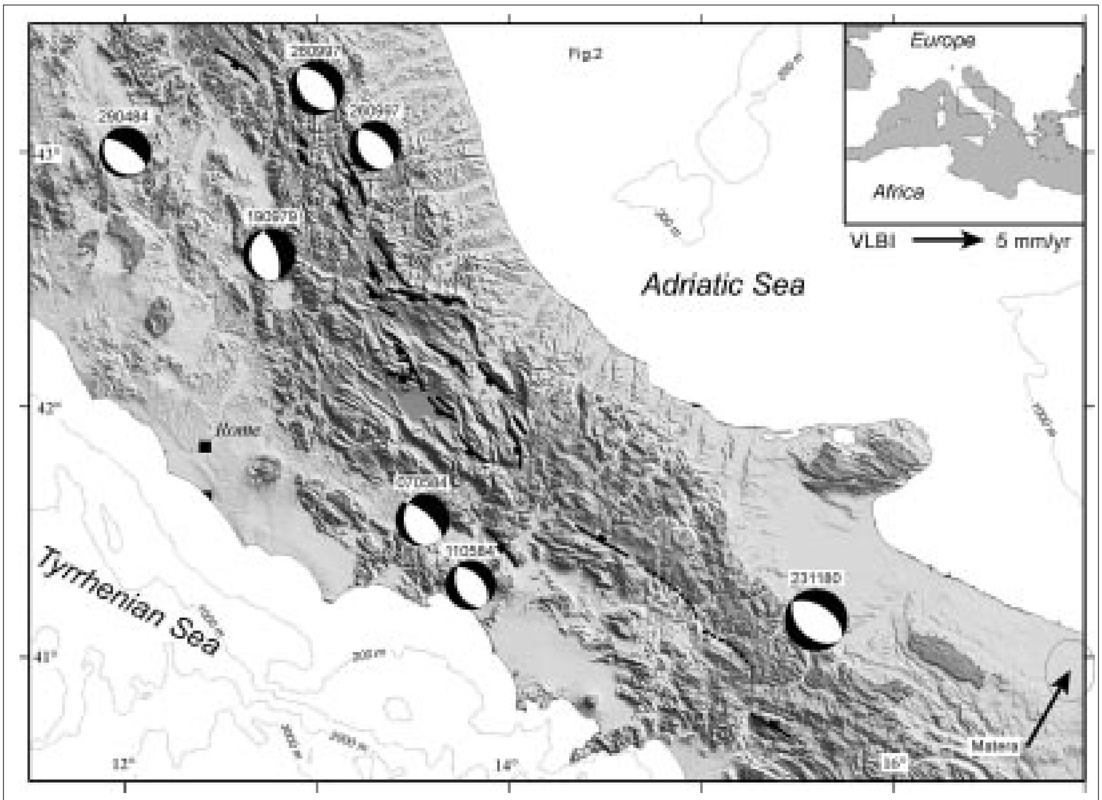

Fig. 5. Seismotectonic map of central Italy. Dark shading represents area above elevation 500 m. Focal mechanisms are from the Harvard CMT catalogue. Active faults (thick lines) are from Galadini and Galli (2000) and Valensise and Pantosti (2000).

by the distribution of faults that have been active in the Late Pleistocene and Holocene. The historical and instrumental seismicity is generally confined to a belt whose width varies along the length of the Italian peninsula. In the northern and southern Apennines the active belt is only 20—30km wide and may correspond to a single active fault system (Valensise and Pantosti, 2000). However, in the central Apennines (approximately between the latitudes of 41.5 N and 42.5 N) the largest earthquakes (Selvaggi, 1998) and the faults active in Late Pleistocene and Holocene time seem to be distributed over a broader width of at least 50 km and involve at least two major sub-parallel fault systems.

The increase in width of the actively extending belt in the central Apennines is associated also with an increase in average elevation and the across-strike width of the Apennine topographic belt itself. In this central area the eastern normal fault system bounds the Laga and Gran Sasso massifs (Blumetti *et al.*, 1993; Giraudi and Frezzotti, 1995) and the Sulmona basin (Vittori *et al.*, 1995), while the western one starts from the northern end of the l'Aquila basin, crosses the Velino-Sirente massif, bounds the Fucino basin, and continues southward through the Marsica region. The western fault system has produced several substantial earthquakes in the last thousand years (Fig. 2). The last shock (I_max=X MCS; Boschi *et al.*, 1999) of a sequence that occurred in 1703 is thought to have ruptured the fault system in the upper Aterno valley (Blumetti, 1995; Cello *et al.*, 1998a). Trenching studies along the Ovindoli fault in the Velino-Sirente massif have revealed a previously unknown earthquake that occurred between 860AD and 1,300AD (Pantosti *et al.*, 1996). The normal fault system bounding the Fucino basin is known to have ruptured in 1915 (M 6.9) when co-seismic surface faulting was described at the time by Oddone (1915) and later re-investigated by others (Serva et at., 1988; Michetti et al., 1996). In 1984 the southern Latium earthquakes (M 5.8 and 5.2) occurred in the Marsica region SE of the Fucino basin (Westaway *et al.*, 1989). The eastern fault system shows evidence of Late-Pleistocene to Holocene activity (Galadini and Galli, 2000) but can not be associated with any known historical earthquakes with the possible exception of the Aremogna fault which may have been activated in the 1349 earthquake (Valensise and Pantosti, 2000).

This apparent quiescence may suggest that the eastern system is now inactive and that extension is taken up only by the western fault system. Alternatively, it may indicate that seismicity is clustered episodically on to a single fault system with cycles whose time scale (perhaps 10^4—10^5years) is longer than the historical or paleoseismological catalogue: a suggestion that has also been inferred from geomorphological or historical data in Nevada, Turkey and Greece (Wallace, 1987; Ambraseys, 1989; Jackson and Leeder, 1994; Jackson, 1999) and which has some support from numerical models (Cowie, 1998). Further insights are provided by GPS measurements in the interval 1994-1999 (D'Agostino *et al.*, 2001) showing significant active strain accumulation across the western fault system suggesting also the possible existence of another undetected active fault system more to the SW. This summary of the active deformation in the central Apennines highlights two main points: (1) that the active extension is concentrated along the main topographic ridge of the Apennines, and (2) that the increase in width of the actively extending belt between 41.5 N and 42.5 N correlates with the higher elevation and increased width of the topographic belt.

The Quaternary tectonic activity of the Calabrian arc (Fig. 6) is mainly represented by N-S and NNE-SSW extensional fault systems that extend all along the inner flank of the arc(Tortorici *et al.*, 1995). According to the age of faulted rocks and to morphological featuresof the fault scarps, vertical slip rates of 0.5-1.2 mm/yr for the last 700 kyr have been computed (Monaco and Tortorici, 2000). This strong activity is confirmed by historical and instrumental data about earthquakes occurred since 1,000 AD (Postpischl, 1985; Boschi *et al.*, 1995, 1997). These data show a distribution of crustal seismicity of M 5, concentrated mostly along the above mentioned fault systems, although several

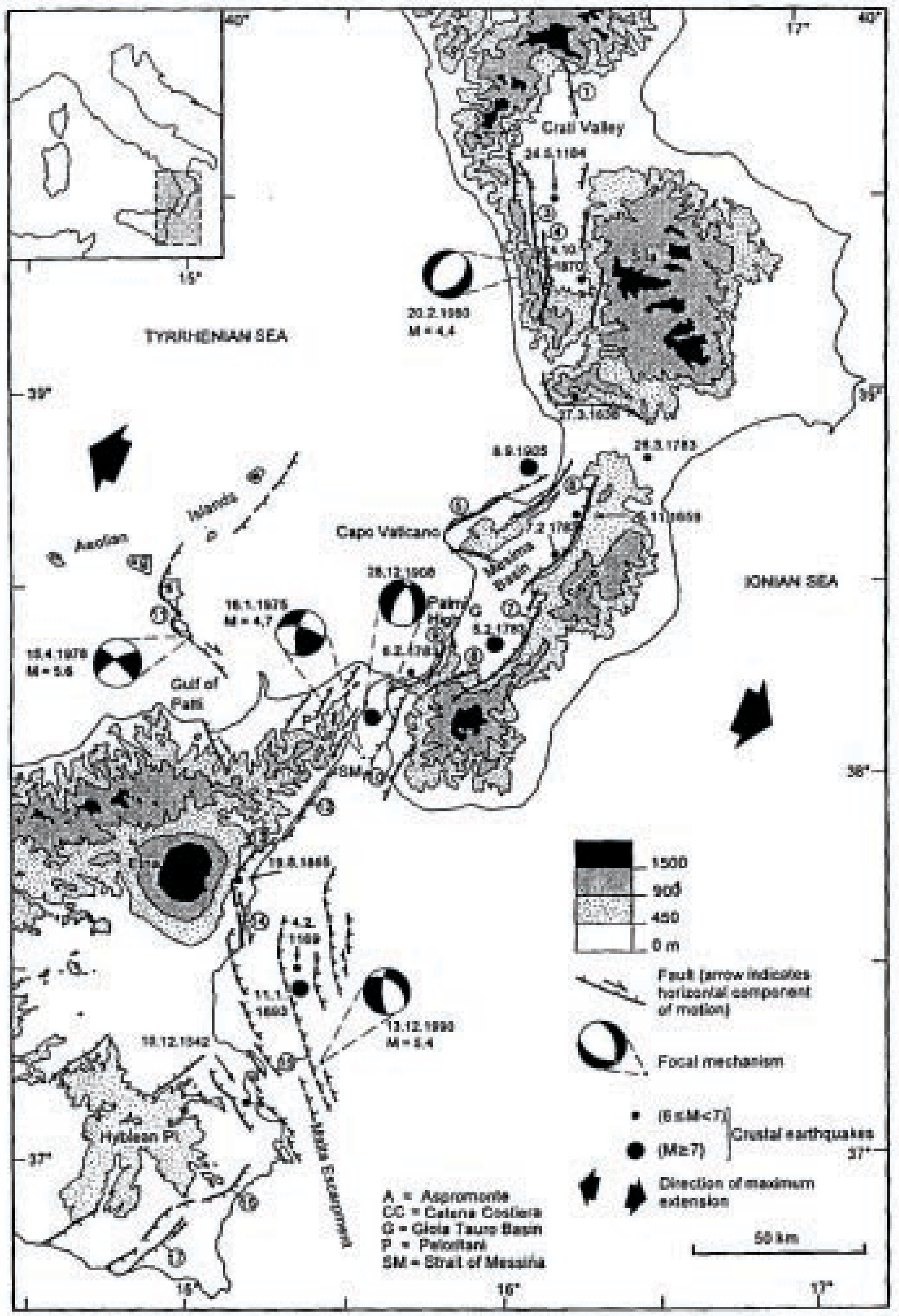

Fig. 6. Seismotectonic map of the Calabrian arc and eastern Sicily. Topography is from 1:100.000 topographic maps of Istituto Geografico Militare. Crustal seismicity (H < 35 km) (Postpischl, 1985; Boschi et al., 1995, 1997) and focal mechanisms of more recent events are from Cello et al. (1982), Gasparini et al. (1982), Anderson and Jackson (1987) and Amato et al. (1991). Numbers refer to fault segments: 1, Castrovillari; 2, Fognano Castello; 3, S. Marco-S. Fili; 4, Montalto-Rende; 5, Capo Vaticano; 6, Serre; 7, Cittanova; 8, S. Eufemia; 9, Scilla; 10, Reggio Calabria; 11, Lipari Vulcano; 12, Messina-Taormina; 13, Piedimonte; 14, S. Alfio-Acireale; 15, Western offshore; 16, Avola; 17, Rosolini-Pozzallo. (From Monaco and Tortorici, 2000).

events occurred more to the east inland and in the Ionian sea (Moretti *et al.*, 1990; Chiodo *et al.*, 1992; Monaco and Tortorici, 2000).

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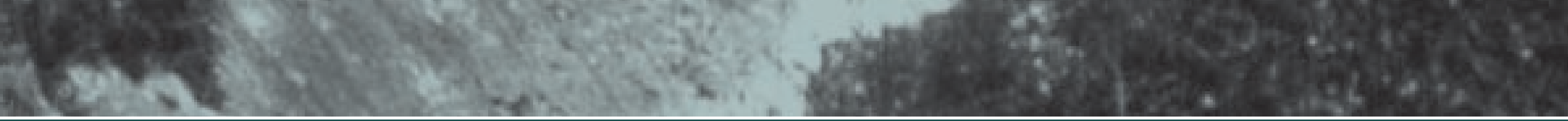

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2. Neotectonics and large-scale gravitational phenomena in the Umbria-Marche Apennines, Italy

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Introduction

The role played in the evolution of the relief by deep-seated gravitational deformations along slopes, has been pointed out by several authors who analysed their typology and distribution in different geological and geomorphological conditions (Jahn, 1964; Ter-Stepanian, 1966 and 1977; Zischinsky, 1969; Nemcok, 1972; Radbruch-Hall *et al.*, 1976; Mahr and Nemcok, 1977; Guerricchio and Melidoro, 1981; Sorriso-Valvo, 1984; Savage and Swolfs, 1986). These phenomena are particularly frequent in Italy, where they have been favoured by the noteworthy relief, the complex geologic structure, the tectonic and seismical activity, and the frequence of extreme rainfall characterising its territory (Dramis *et al.*, 1983; Sorriso-Valvo ed., 1984, 1987 and 1989). Some of them evolved into huge landslides of different type, or sometimes are connected to the latter by a cause/effect relationship.

Deep-seated gravitational deformations and their evolution into landslides

Deep-seated gravitational deformations and their related huge landslides are complex phenomena taking place through a wide variety of mechanisms whose genesis and evolution are controlled by several factors, among which structure, relief, and tectonic and seismic activity have a particular importance (Terzaghi, 1962; Radbruch-Hall *et al.*, 1976; Solonenko, 1977; Radbruch-Hall, 1978; Dramis *et al.*, 1983; Carton *et al.*, 1987; Savage and Varnes, 1987).
The most typical morphological elements of deep-seated gravitational deformations are the following (Radbruch-Hall *et al.*, 1976; Mahr and Nemcok, 1977; Radbruch-Hall, 1978; Dramis *et al.*, 1983; Pieruccini, 1988; Dramis and Sorriso-Valvo, 1994):

a) surface extension is generally ranging around at least 1 square km;
b) thickness of deformed mass ranges around several tens (or sometimes hundreds) metres;
c) displacements are reduced with respect to the volume of involved materials;
d) a continuous shear surface delimiting the deformed mass is lacking;
e) evolution is very slow, happening in "geologic" times with long periods of inactivity or extremely reduced activity, alternated with short and sudden activations, often as a consequence of earthquakes or extreme rainfall;
f) deformational mechanism is continuous creep, with accelerations and creep ruptures;
g) kinematics is almost always influenced by active or residual regional stress and high confined pressure;
h) topography has a reduced influence whilst structure plays a strong control on the location and typology of the deformation;
i) deformed slopes are often invoved in more surficial (secondary) landslides.

The most recurrent deformational typologies are sackungs, lateral spreadings and deep-seated block slides (Jahn, 1964; Zischinsky, 1966 and 1969; Varnes, 1978; Dramis and Sorriso-Valvo, 1994).
Sackungs are connected with rock flows affecting huge jointed or stratified rigid masses, which can be considered homogeneous as a whole. These, being loaded for extremely long

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periods by their own weight may slowly deformate, in analogy to what happens during tectonic folding. It is easy to understand that to activate these deformational phenomena, slopes have to be high enough to induce strong gravitational stress in the bedrock, and resistant enough to maintain their overall stability. Gravitational stress produces shear surfaces, often coinciding with existing tectonic discontinuities (joints, faults, thrust or backthrust planes) in the surficial portion of the slopes, whilst more in depth the high pressure induces plastic deformation, without a proper yelding surface (Mahr and Nemcok, 1977). The highermost portion of the deformed slope is affected by extensional stress which produces high angle shear planes, to which counterslope steps and graben-like trenches, are associated. The lowermost portion of the slope is affected by compressional stress determining bulging and, sometimes, low-angle shear planes (Nemcok, 1972; Mahr, 1977; Ter-Stepanian, 1977; Guerricchio and Melidoro, 1981; Savage and Swolfs, 1986; Savage and Varnes, 1987).

Surface effects on the upper part of the slopes are very similar to those produced by listric faults, making it difficult to recognise them properly in the field. On the other hand, on slopes (such as those produced by normal faults) disposed perpendicularly to the minimum stress, extensional and gravitational stresses being consistent can undergo feedback phenomena, mostly in coincidence with strong earthquakes (Dramis *et al.*, 1982).

Lateral spreads are characterised by horizontal extension of the relief, balanced by shear or tensile fractures and may happen in the following ways.

a) Bilateral spread: gravitational deformations produce a bilateral extension of the relief without generating any shear surface or basal flow zone. Double ridges locally found at the top of high and elongated reliefs modeled in rigid and "homogeneous" rocks have been attributed to these phenomena (Jahn, 1964). Another deep-seated gravitational deformation with bilateral extension of the relief but associated with both trench-like depressions at the summit and shear planes at the base, has been described by Beck (1968) and included by Varnes (1978) in his classification of mass movements.
b) Tectonic-gravitational spread: gravitational deformations affect morpho-structural reliefs produced by active thrusting. Spreading starts as a consequence of tectonics but then evolves because of gravity, predisposing conditions favourable to a successive activation of huge landslides (Dramis and Sorriso-Valvo, 1994).
c) Deep-seated block spread: deformations involve rigid and thick rocky masses overlying sub-horizontal less-competent layers (Cruden and Varnes, 1993). In these conditions, gravitational stresses can trigger plastic flows (sometimes combined with extrusion phenomena) in the underlying material, causing rupture into blocks of the rigid masses, often following existing discontinuities. Blocks can experience subsidence, limited traslation, rotation, tilting or downlifting, giving origin to counterslopes, steps and trenches in this case too.

Deep-seated block slides (Varnes, 1978; Agnesi *et al.*, 1987; Dramis *et al.*, 1987) are characterised by relatively small displacements of overall intact and thick rigid blocks overlying faintly sloping less competent layers. Their main surface effects are large scarps and trenches, in the upper and intermediate portion of the slope, and bulging associated with frequent shear planes with geometry typical for a reverse fault, in its lowermost portion.

Taking into account the characteristics of the above described phenomena, it is evident that only sackungs and lateral spreads, of type a (possibly less the Beck type) and b, fit well in the general definition of deep-seated gravitational deformation, whilst deep-seated block spreads and block slides should better be included among landslides, being characterised by displacements along more or less continuous yelding surfaces. Anyhow, taking into account the moderate entity of displacement with respect to dimension of involved masses and the extremely slow evolution with dominating creep phenomena, it has been decided to include the above said phenomena among

deep-seated gravitational movements, separating them from landslides s.s.
After a generally very long evolutionary phase, deep-seated gravitational phenomena can evolve in huge landslide movements of different types (Guerricchio and Melidoro, 1973; Radbruch-hall, 1978; Nemcok, 1982), often characterised by "scale" factors (Goguel, 1978). From this point of view, deep-seated gravitational deformations can be considered as preparing stages for huge gravitational collapses that, anyhow, not always complete their evolution (Hutchinson, 1988; Dramis and Sorriso-Valvo, 1994). The most striking phenomena are those connected with deep-seated block slides, whose evolution into translational slides can sometimes produce superimposition of terrains having dimensions comparable with those of tectonic phenomena (Nijman, 1971).

Geology and neotectonic evolution of the Umbria-Marche Apennines

The Umbria-Marche Apennines is a complex fold-and-overthrust arcuate belt with a NE vergence (Calamita and Deiana, 1988), created as a consequence of Neogene compressional tectonics which affected a thick sedimentary cover.
The latter includes at its base a thick (some 800 m) massive calcareous shelf complex (lower Lias), followed by a well stratified pelagic and emipelagic sequence (ca. 1400 m thick) made up of limestones, marly limestones, cherty limestones and marls (middle Lias - lower-middle Miocene). The sedimentation ends with more or less thick turbiditic terrains (Tortonian - lower Pliocene), with evaporitic and lagoon facies (Messinian) intercalated in its upper portion (Cantalamessa *et al.*, 1986).
The most striking surficial effects of compressional deformation are two main anticlinoric structures (the Umbria-Marche ridge s.s. to the West and the Marche ridge to the East), whose axes have an overall trend NW-SE to the North, and N-S up to NNE-SSW to the South. An important overthrust is that of the Sibillini Mts, which emplaced Mesozoic-Neogene terrains of the Apennine Ridge over the more external Tertiary deposits of the eastern area. Another overthrust (the Valnerina one) superposed terrains of the internal ridge over those of the Camerino basin, located in between the above said ridge and the easternmost anticlinoric structure. To the South, the Camerino synclinorium looses its identity and the two above said ridges merge in the Sibillini Mts. massif.
In the eastern sector of the area, in the peri-Adriatic belt, the compressional structures are unconformably covered by sands, conglomerates and pelites (lower Pliocene - lower Pleistocene), whose surficial setting follows a monocline gently dipping to the NE (Ori *et al.*, 1991). Over most of the area, the tectonic phase responsable for thrusting took place in the timespan middle Miocene - middle Pliocene, but it continued in more recent times along the Adriatic coast, where reverse faults and faint anticlinalic structures with eastern vergence (such as those of Porto San Giorgio, Polverigi and Senigallia) affecting Pleistocene deposits are present (Cantalamessa *et al.*, 1987). Listric normal faults lowering seawards the above sediments are sometimes associated to these structures.
Present-day activity of compression is still continuing in the easternmost part of the area, as testified by focal mechanisms of earthquakes (Gasparini *et al.*, 1985; Riguzzi *et al.*, 1989).
Starting from the upper Pliocene, the area experienced extensional tectonics, which, moving eastwards from West, was following compressional tectonics. In this way, normal faults, mostly trending in parallel to the Apennines and faulted blocks degrading towards the Tyrrenian Sea, have been generated. Extensional tectonics largely reactivated previous compressive discontinuities, producing noteworthy inversion phenomena, located at different depths (Nijman, 1971; Calamita and Pizzi, 199).
In the axial portion of the Apenninic belt, the extensional phase tooke place in continental conditions because of a general uplift started at the end of the compressive phase. In this

framework, tectonic activity largely influenced the evolution of the landscape, where fault scarps dividing the relief into blocks (mostly in the SW of the area) show morphological evidence (Centamore *et al.*, 1980).
Combined action of apenninic normal faults and transversal faults which played as transfer elements (Calamita and Pizzi, 1992), produced intramontane tectonic depressions bordered by showy fault scarps. Starting from upper Pliocene - lower Pleistocene (Ge.Mi.Na., 1963; Blumetti and Dramis, 1992), these depressions hosted lacustrine basins whose sedimentary sequences add up to some hundreds m. Present-day activity of extensional tectonics is testified by both tectonic discontinuities, displacing also the most recent deposits (Calamita *et al.*, 1982; Blumetti *et al.*, 1993), and focal mechanisms of the strong earthquakes frequently affecting the area (Gasparini *et al.*, 1985).
Starting from the end of the lower Pleistocene and up to the present, the whole Central Italy has been involved in a faster uplift, even though with different rates (Demangeot, 1965; Ambrosetti *et al.*, 1982; Dufaure *et al.*, 1989; Dramis, 1992). As a consequence, marine deposits of the peri-Adriatic Pliocene-Pleistocene Basin assumed their typical monocline structure, reaching altitudes exceeding 200 m a.s.l. close to the coastline and up to more than 1.000 m a.s.l. mountainwards, at Mt. Ascensione (Cantalamessa *et al.*, 1987). Also along the Tyrrhenian side, altitudes up to some hundred metres of Pleistocene marine deposits and shorelines (Ambrosetti *et al.*, 1987; Alfonsi *et al.*, 1991) testify for the uplift. Along the axis of the Apennines, the uplift was even stronger, as demonstrated by the altitude of planation surfaces and continental terraced deposits correlated with Pleistocene sea levels (Dufaure *et al.*, 1989; Dramis, 1992).
Along the western side of the Apenninic belt, the uplift interacted with extensional tectonics enhancing the activity of existing normal faults and producing new ones featured by very long and straight apenninic lay-outs, sub-vertical planes and moderate displacements.
The erosion deepening caused by the uplift produced widespread phenomena of morphoselection, which throwing into morphologic relief the calcareous anticlinoric structures (Ciccacci *et al.*, 1985). This is particularly evident along the eastern side of the belt, where erosional phenomena created differences in height up to more than 1000 metres between the overthrust fronts and the terrigenous deposits located eastwards. The deepening was less strong along the Tyrrhenian side, because of the progressive lowering to the West of planation and deposition surfaces caused by normal faults (Dufaure *et al.*, 1989; Dramis, 1992).
The uplift manifested itself with different amplitude moving from North to South, thus determining in the area the creation transversally to the chain of three main morpho-structural undulations whose elevation progressively lowers moving from the South northwards (Dramis *et al.*, 1991). This is clearly shown by interpretation and altitudinal analysis of planation surfaces, and by sedimentological-stratigraphic analysis of Pliocene-Pleistocene marine closure deposits. Also the bedrock structure of the Apenninic calcareous ridges, which as previously said has an overall axial trend to the North, perfectly fits with the above morphological setting.
Transversal morphostructures are delimited by anti-apenninic faults, whose activity seems to have continued, with moviments varying in time, starting at least from the Tortonian up to the present. These discontinuities strongly influenced sedimentation in Miocene-Pliocene-Pleistocene basins and, after the emersion, the evolution of drainage network (Boccaletti *et al.*, 1983; Ciccacci *et al.*, 1985).
Recent and present activity of these trasversal tectonic elements seems to vary both in space and time: in fact, they show to have acted as normal, trans-tensive and trans-pressive. This can be explained both in an extensional stress field SW-NE, and in a compressional stress field N-S, either alternating or superimposing to the former (Centamore *et al.*, 1980; Calamita and Pizzi, 1992).

Historical seismicity of the area

Archive reserch allowed to point out the occurrence in the area of about 60 intense (VIII - IX MCS) earthquakes from 1000 to 1984, with a strong clustering of events in the last 150 years (Baratta, 1901; C.N.R., 1985).
Historical investigations also evidenced more ancient strong earthquakes such as the 102 B.C. Valnerina event, the 217 B.C. event of Val Tiberina (occurred during the Lake Trasimeno battle between the Hannibal and the Roman armies), and the VIII-IX MCS events which struck Spoleto in 446 and 801 B.C.
The strongest earthquakes occurred in the Norcia - Cascia area in 1328, 1703, 1719, 1730, 1783, 1859, 1979. Among these, the strongest ones (both X MCS) were the 1328 event (which destroyed Norcia, Preci and Montesanto) and the 1703 event (which caused about 10.000 victims in a vast area between Norcia and L'Aquila).
Another area affected by strong earthquakes is the Assisi-Foligno-Spoleto basin. Here, in 1832, a IX-X MCS event caused severe damages to the towns of Bevagna, Cannara and Foligno (Anonimous, 1832; Rutili Gentili, 1832; Mercalli, 1883).
A particularly destructive seismic period in Central Italy was XVI century, when two of the strongest earthquakes of the last millennium occurred: the already mentioned 1703 event (Norcia - L'Aquila), the 1781 event (Cagli-Pesaro), and the 1799 one (Camerino) (Osservatorio Geofisico dI Macerata, 1982-Com. Mont. Fabriano).

Typology and distribution of deep-seated gravitational deformations in the Umbria-Marche area

Systematic geomorphologic analysis of the Umbria-Marche area allowed to recognise a noticeable amount of deep-seated gravitational deformations and large-scale landslides of different type (Dramis *et al.*, 1995). Altogether, some 500 phenomena have been individuated, with a frequency of 0.06 per square km, a mean extension around 2 square km and a maximum area of about 15 square km.
In the area, the main genetic factors of these phenomena are the following:

a) bedrock geology is characterised by high thickness of massive calcareous and arenaceous rocks overlying levels less resistant or thinly stratified with thin pelitic intercalations;
b) structure derives from compressional tectonics to which the presence of strong residual stress and shear zones, along which large-scale gravitational displacements can occur, are connected;
c) the area recently underwent a strong uplift, mostly along the axial belt of the Apennines, where the presence of resistant rocks allowed the creation of high and steep slopes bordering either deep transversal valleys or overthrust fronts;
d) extensional tectonics originated high fault scarps which sometimes brought to the surface potential deep sliding planes, whilst compressional tectonics, still active along the coast, is producing folding along whose eastern sides huge gravitational phenomena can take place;
e) seismicity is quite high, mostly along the chain axes and, subordinately, along the Adriatic coast (Postpischl ed., 1985).

First activation and successive evolution of the deep-seated gravitational phenomena observed in the area can be dated only by means of relative methods, lacking any good datable element. Particularly useful to this end are the relationships between slope elements deformed or involved in large-scale landslides and stratified debris deposited in periglacial environment (Coltorti and Dramis, 1988), terraced alluvial deposits, lacustrine and morainic deposits. These relationships allow to set the first activation of many phenomena before the last

Pleistocene cold phases (or, at least, contemporarly to them), and, in many cases, to recognise Holocene reactivations (Carraro *et al.*, 1979; Dramis *et al.*, 1988; Gentili and Pambianchi, 1993). Only rarely historical or recent reactivations of the phenomena were obsreved, in coincidence of either earthquakes (Coppola *et al.*, 1978; Blumetti *et al.*, 1990) or extreme precipitations in the terrigenous lithotypes of the Adriatic belt (Coltorti *et al.*, 1985).

Deep-seated graviational phenomena and large-scale landslides along Western (fault) slopes of the Apenninic ridges

Many cases of deep-seated gravitational deformations can be observed along the western sides of both the calcareous Apenninic belts and the massive arenaceous Laga Mts. Particularly important are the sackung and block slide phenomena found in the Sibillini Mts., along the large fault slopes bordering to the East the Cascia, Norcia and Castelluccio depressions, and the SW slope of Mt. Fema.

On the highermost portion of these slopes, counterslope steps, scarps and trenches are present, in addition to open fissures, from some ten centimetres up to a few metres wide, whose freshness is testified by lack of debris inside them

Threnches, scarps, steps and couterslopes, located at the top of slopes, together with huge landslides, partially covered by stratified debris, characterise the Mt. Alvagnano -Castel S. Maria structure (Fig.1) to the South (Blumetti *et al.*, 1990). Seismic and historical data testify for an evident reactivation of some landforms during the 1703 earthquake (X MCS), whilst less intensive surface effects were produced by the 1979 event.

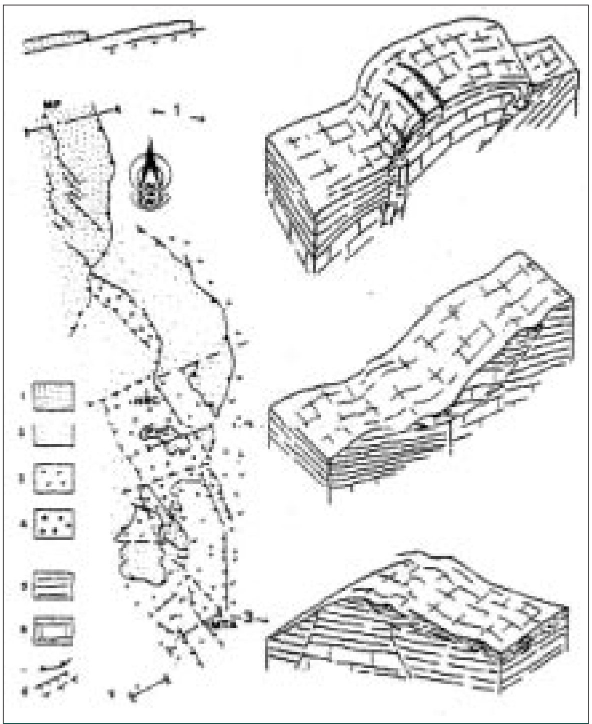

Fig. 1. Structural sketch of the Mount Fema-Mont Alvagnano area, and cross-section representing deep seated gravitational deformation and large scale landslides on its western fault slopes: 1)upper overthrusted units; 2) middle overthrusted units; 3) lower units; 4) deposits filling the depression; 5) limestones and marly limestones (Middle Lias-Miocene); 6) limestones (Lower Lias); 7) overthrust; 8) normal fault; 9) cross-section traces: MF – Mt. Fema; MTA – Mt. Alvagnano; NRC – Norcia; PVC – Poggio Valaccone.

The above seems to testify for a step-like evolution of deep-seated gravitational deformations in correspondance with major earthquakes. It also seems possibe to attribute the origin of the huge landslide accumulations found at the base of fault slopes (Blumetti *et al.*, 1993) to collapse of rocky masses previously affected by deep-seated gravitational deformations, probably as a consequence of past earthquakes.

Block spreads and deep-seated block slides, sometimes evolving into huge translational slides, found along the NW side of Mt. Gorzano (Monti della Laga), a little South of the Sibillini Mts., are connected with fragmentation of thick arenaceous bodies into blocks, due to normal faults (Dramis *et al.*, 1987).

Other gravitational phenomena, found along the large fault slopes bordering to the East the intermontane tectonic depressions, are huge translational slides, probably deriving from deep-seated block slides. Those developed as a consequence of exhumation of potential sliding surfaces, corresponding to previous compressional shear planes (Nijman, 1971), operated by the above said border faults. A showy example is possibly represented by the small relief of Poggio Valaccone, located at the base of a fault slope bordering to the East the Norcia depression (Fig.1). A similar but smaller phenomenon has been recognised along the eastern slope of the Castelluccio basin.

Deep-seated graviational deformations and large-scale landslides along the eastern slopes of ridges and overthrust front

Gravitational morphogenesis reaches its maximum development on the eastern slopes of calcareous ridges, along oversteepened sides of both folded structures and overthrust fronts. Along these slopes, steps, trenches, undulations and intense fracturing can be observed, together with large scale landslide scarps; at the foot of the slopes, wide and thick landslide scree deposits, sometimes coalescing, overlie marly-clayey levels of the most recent terrains of the Umbria-Marche Succession. Generally, those deposits are often covered by upper Pleistocene stratified slope-waste deposits, which sometimes are tilted counterslope; the two units are locally interfingered thus testifying for an ancient origin of the movement and for its successive reactivations (Coppola *et al.*, 1978; Gentili and Pambianchi, 1993). Main causes of these complex gravitational movements are to be found in the intense tectonic deformation of the bedrock, which has been folded and overthrusted, and in the presence of residual compressional stresses. Recent uplift with connected creation of erosional differences of height, favoured the activation of bilateral spreadings of the type described by Jahn (1964), which originated double ridges, lateral spreads and sackungs, mainly along slopes modelled on thick stratified calcareous formations; also deep-seated block slides, along either existing shear planes or zones, or marly-clayey stratigraphic levels, were activated which, because of strong altitude differences, easily evolved into large-scale gravitational collapses.

Deep-seated graviational deformations and large-scale landslides along the slopes of transversal valleys

Deep-seated gravitational deformations and large-scale landslides are present, even though with less frequency, also inside the chain, where deep incisions transversal to its axis caused by fast and intense uplift allowed the intersection at various heights of slopes with potential sliding surfaces, mostly represented by marly layers dividing more rigid calcareous beds. The overall bedding accordingly to the slope, due to the general northward dipping of the structures, favoured the activation of huge block spreads and deep-seated slides on southern valley-sides (Dramis *et al.*, 1988; Gentili and Pambianchi, 1993). Somewhere, sackung phenomena have been recognised too (Coppola *et al.*, 1978). One of the most representative examples of deep-seated gravitational deformation can be found along the portion of the Marche ridge in between the Fiastrone and Chienti Rivers, slightly South of Camerino (Dramis *et al.*, 1988; Gentili *et al.*, 1992). There, at the base of the slope, modeled in marly limestone, shearing planes and counterslope tilting of Holocene alluvial deposits can be observed. In its median portion, the

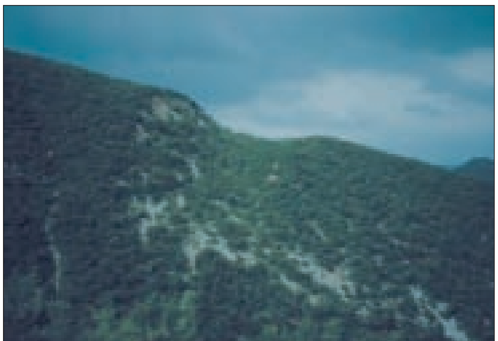

Fig. 2. The Valdiea trench

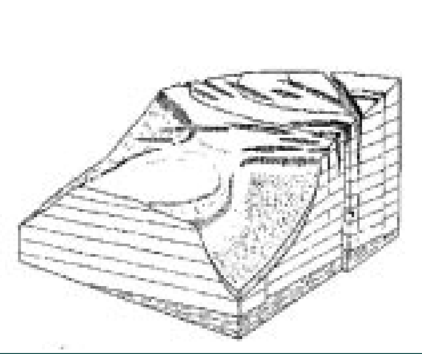

Fig. 3. The Mt. Frascare deep-seated gravitational slope deformation

slope showss a trench whose freshness suggests a very recent activation of the deformation (Fig. 2). This is further confirmed by oral tradition among local people, speaking of reactivation of the deformation as a consequence of the strong 1741arthquake (IX MCS). Slightly mountainward along the valley bottom, an alluvial terrace of the same order contains lacustrine silty and clayey levels with abundant vegetal remnants, whose age is comprised between 27,000 and 30,000 years b.p. (Damiani and Moretti, 1968). This demonstrates the step-like activation of the landslides during the deposition of upper Pleistocene alluvial deposits.

Sometimes, the connection with recently active transversal discontinuities is particularly evident. Particularly interesting for this are the deep-seated gravitational phenomena characterising the northern slope of Mt. Frascare , on the right side of the Fiastrone River (Fig. 3). In the uppermost part of a slope affected by a wide block spread, involving marly-calcareous levels overlying marls, steps and trenches trending SSW-NNE are present, following an en échelon system transversal with respect to the slope and fitting with a right transcurrent displacement of the N-S trending faults which cross the area.

Somewhere else, finally, relationships between gravitational deformation and tectonic displacement are not clear, phenomena being possibly induced only by high relief. Such is the case of the NW slope of Mt. Le Siere (North of Fabriano), modeled in thinly stratified marly limestones and affected at its summit by a showy sackung.

Cases of historical and recent reactivation of gravitational phenomena in the area

Historical research, mostly focussed on the Valnerina area, pointed out reports of earthquake triggered gravitational phenomena.

For example, from the 1328 earthquake collapses of "mountains" and "opening of "abysses" in the ground were reported (Villani, 1848; Baratta, 1901). Probably some particularly fresh morphological elements in the area (such as scarplets, trenches and landslide bodies) could be related to this event.

Baratta (1901) reported that during the 1703 earthquake, Mt. Alvagnano, close to Norcia, was strongly affected by surface deformation over a belt more than 1500 yards large and about 32 palms. Geomorphological features on top of the hilltop (such small fresh scarplets within a large trench), connected with the reactivation of a deep seated gravitational slope deformation, have been referred to that event (Blumetti *et al.*, 1990).

In connection with the 1703 earthquake, the same author reported about the opening of a long and wide ground fracture on Mt. Corvo, close to the Leonessa tectonic depression, and a deep hole on top of Mt. Amaro, in the sorroundings of Sigillo. Field geomorphological evidences seem to be consistent with the historical record.. In particular, the presence in the area of a deep depression with loose debris at the bottom supports the hypothesis that it was related to the collapse of a karstic cave in the Scaglia Rosata limestone. A similar depression, denominated “buca del terremoto” (earthquake hole), was produced in the Scaglia Rosata limestone may be observed close to Camerino, on top of Mt. Colleluce. Nothwistanding the reference to earthquake in the landform name, there is no historical record of any seismic event connected with the possible collapse.

Finally, three huge landslides were triggered in the area of Mt. Nerone - Piobbico during the catastrophic earthquake of 1781 (Bisci *et al.*, 1995).
Concerning the reactivation of deep-seated gravitational slope deformations, A showy example can be found along the portion of the Marche ridge in between the Fiastrone and Chienti Rivers, slightly South of Camerino (Dramis *et al.*, 1988; Gentili *et al.*, 1992; Gentili and Pambianchi, 1993). There, at the slope feet, two shearing planes can be observed within the lowermost member of the rigid marly-calcareous Scaglia Rosata, whose uppermost portion show a mountainward dip, tilting in the same direction the Holocene fan deposits too. In its median portion, the slope has a thench whose freshness suggests a recent activation of the deformation. This is further confirmed by oral tradition among local people, speaking of a reactivation of the deformation as a consequence of the 1741 earthquake (Dramis *et al.*, 1995) and, possibly, the 1799 earthquake. Slightly mountainward along the valley bottom, an alluvial terrace of the same order contains lacustrine silty and clayey levels with abundant vegetal remnants, whose age is comprised between 27.000 and 30.000 yaers B.P. (Damiani and Moretti, 1968). This demonstrated a step-like activation of the landslides during the deposition of upper Pleistocene alluvial deposits (Dramis *et al.*, 1988). About 500 m to the south (close to Valdiea), a huge mass of Scaglia Rosata limestone, related to a gravitational event, overlies the above mentioned alluvial materials.


Fig. 4. Fracture on the Mt. Fema south-western slope reactovated by 1979 Norcia earthquake

Another possible historical reactivation of deep-seated gravitational slope deformation is that of the 1866 earthquake (VIII MCS) wich struck the area of Spoleto. Concerning this event reported that on the eastern side of the Spoleto tectonic depression (between Spoleto and Campello) "the mountain was opened over one mile length". Field survey confirmed the presence in the area of trenches up to 2 km long, parallel to the slope and clearly connected with deep-seated gravitational deformation. The extreme freshness of these features seems to testify a possible reactivation in very recent times.
A showy reactivation of a sackung type gravitational deformation, connected to the 1979 earthquake (VIII MCS) was that observed by oral testimonies of local people. who reported about the opening of cracks and the reactivation of trench scarplets on the slopes of Mt. Fema (Dramis *et al.*, 1995; Fig. 4).
During the same earthquake, a number of ground fractures, more than 100 m long, were produced also along the northward continuation of the trench of Mt. Alvagnano, close to the town of Norcia (Blumetti *et al.*, 1990).

Deep-seated graviational deformations and large-scale landslides along the Adriatic Coast.

Along the Adriatic coast, trenches trending in parallel with the coastline and sometimes delimited by fractures, and steps lowering seawards, almost identical to fault scarplets, have been found in wave-cut cliffs, generally no more active and separed by the sea through small beaches. These landforms are frequent along NE slopes of compressional structures with an Adriatic vergence, made up of lower-middle Pleistocene clayey-arenaceous-conglomeratic terrains (Cantalamessa *et al.*, 1987), presently still active as testified by hypocentral depth and

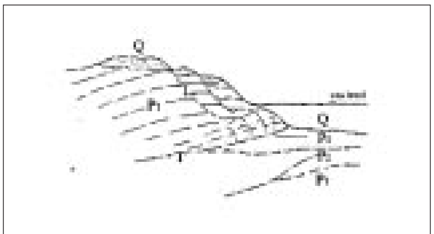

Fig. 5. Cross-setion of a deep-seated gravitational deformation along the Adriatic coast: P_1 - Lower Pliocene; P_2 - Middle Pliocebe;
Q - Quaternary; T - Thrust, L - Listric fault

focal mechanisms of earthquakes (Gasparini *et al.*, 1985; Riguzzi *et al.*, 1989). The origin of the above landforms is to be referred mainly to deep-seated gravitational deformations (tectonic-gravitational spreading), induced by the compressional tectonics which prosecutes in uplifting the structures (Dramis and Sorriso-Valvo, 1994). Within this framework, large scale rotational-traslational landslides and listric faults lowering toward the Adriatic Sea are also produced, whose morphological surface effects can be easily confused among them and with those caused by deep-seated gravitational deformations (Coltorti *et al.*, 1985; Guerricchio, 1988) (Fig.5).

Along slopes affected by deep-seated gravitational deformations and/or listric faults, large-scale landslides are present whose activity is step-like, with long dormant periods separated by brief movements. Those reactivations are connected with evolutionary steps of deeper phenomena and seem to be mainly triggered by extreme rainfall (Coltorti *et al.*, 1985).

One of the most representative examples of deep-seated gravitational slope deformation can be observed near the town of Ancona. On the evening of the 13th December 1982, after a period of particularly heavy rain, a huge landslide took place on the north-facing slope of Montagnolo Hill, in the western outskirts of Ancona, over an area of more than 3.4 km^2, from about 170 m a.s.l. to the Adriatic coast (Crescenti *et al.*, 1983; Coltorti *et al.*, 1984; Crescenti ed., 1986).

The phase of rapid deformation, which started without warning, lasted only a few hours and was followed by a longer period of settling. More than 280 buildings were damaged beyond repair and many of them collapsed completely. The Adriatic railway, along the coastline, was damaged over a distance of about 1.7 Km. Luckily, there were no victims.

The slope hit by the landslide has had a long history of gravitational movements (Bracci, 1773; Segrè, 1920). In 1858 it was the site of a landslide even larger than the recent one (De Bosis, 1859). Other smaller landslides, still large in an absolute sense, have occurred in the same area. Of these, the "Barducci" landslide, a flow type mass-movement, has been known for a long time, because its continual and intense activity produced visible damage to the coastal road and railway (Segrè, 1920).

From a stratigraphic point of view the lithotypes outcropping are the following:

1) Lower and Middle Pliocene deposits (grey-blue marly clays, 20-40 cm thick, alternated with grey or grey-black compact sands up to 60 cm thick).

2) Pleistocene deposits consisting of five transgressive-regressive cycles of pelitic-arenacoeus units with a total thickness of about 20 m.

The area has been uplifted starting by the end of Early Pleistocene. Coquinic panchina and sands at the top of the clayey beds are probably related to the early stages of the uplift. These deposits are found at the top of Montagnolo Hill (250 m a.s.l.) and at more than 350 m in the surrounding area.

From a geomorphological point of view, the study area displays an overall smoothed morphology, with moderate relief and gentle slopes. The observation of aerial photographs, taken before the event of December 1982, shows a characteristic landslide morphology with trenches, scarps, steps, undrained depressions and reverse slopes. Downslope, a rugged foot-slope zone extends towards the sea; the steepening of the foot-slope seems to be accounted for by sea erosion, before building of the harbour embankment and along-shore protective measures.

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Some remarks on the relations between the 1972 Ancona earthquake sequence and the Ancona landslide

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Most relevant issues under discussion in this stop

The visit to the Ancona landslide allow to clearly document the central role played by earthquake tectonic deformation and surface rupture in the evolution of large-scale landslides. The comprehensive study of this landslide performed in the last 20 years unequivocally shows that rainfall alone would not have been able to produce the catastrophic landslide event occurred on December 13, 1982. Therefore, this is an excellent case for a meaningful discussion of the relations between surface effects of coseismic faulting and very deep-seated gravitational processes (the following section is modified from Brunamonte, 2001).
The continuously increasing knowledge about the tectonic structures that are actively deforming the earth surface, defined as capable faults in IAEA (1991) and recently mapped in Italy by Vittori et al. (1997) and Michetti et al. (2000), provides new interpretative tools for the analysis of the relationships among the structural setting of the topographic relief and the landslide development. The role of the recent fragile tectonics has been essentially investigated up to now from the point of view of the assessment of the seismic potential of a geologic structure. However, active tectonic ground faulting and fracturing deeply influence the bedrock mechanic behavior, and may also represent a fundamental factor for the localization of sizeable gravitational phenomena. The study of large recent bedrock gravity movements, such as the Ancona 1982 landslide that we are visiting during this stop, demonstrated that the landslide evolution, especially the location of the main detachment zone, and the hydrogeological setting responsible for the (re)activation of the gravitational movements, are dominated by pre-existing faults having dimensions at the same scale of the affected slopes (Cotecchia, 1997; Brunamonte, 1997). In these conditions, faults showing young displacement, typically related to coseismic surface rupture phenomena, control the predisposition of the slope to fail. The occurrence of repeated, moderate to strong ($M \geq 5$) earthquakes plays a critical role in the slope geomorphic evolution, either through the periodic fracture remobilization and allowing a hydrogeological circulation at the scale of the whole slope even within very-low permeability terrains. However, where the lithology is mostly composed by clay, the effects of the seismicity are masked by the interaction with the surface processes, which very effectively model the landscape, and habitually cancel the morphological evidence of coseismic tectonic ruptures within few days to few weeks after the causative earthquake. Moreover, it should be noted that in most cases the determining cause that eventually triggers the landslide reactivation is the rise of the water table due to a period of heavy rainfall.

Seismological data and surface fractures during the 1972 earthquake

Figure 1 shows the focal mechanisms of several moderate to strong earthquakes in Central Italy. The solutions for the Ancona 1972 sequence (see also Table 1) clearly show a strike-slip, compressional style of faulting, in agreement with the Adriatic foredeep setting. The earthquake sequence in fact lasted for about two years, from 1972 to 1974. Typical focal depths in the Ancona area ranges between 4 – 8 km. The 1972 sequence is typical for the

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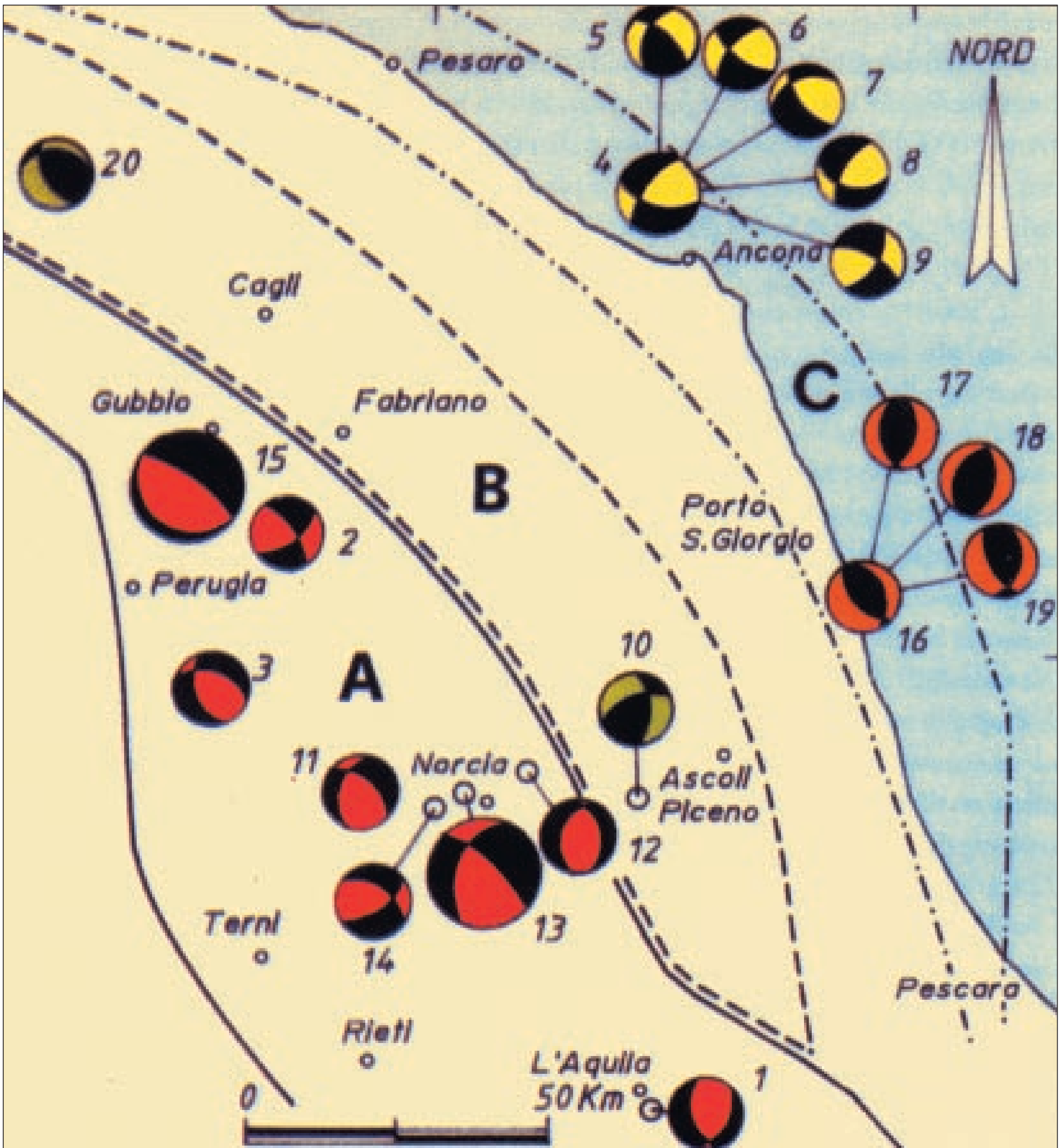

Figure 1. Fochal mechanisms for damaging seismic events in the period 1958-1987 in the Umbria Marche region (after Brozzetti and Lavecchia, 1994). Numbers 4 to 9 refer to the 1972 Ancona earthquake sequence. A) extensional tectonic setting of the Apennine mountain belt; B) transitional Apennine piedmont zone; C) compressional tectonic setting of the Adriatic foredeep.

Ancona region, similar events being well documented in the historical catalogues.
The examination of two airphoto coverages demonstrated that several deep open fractures and flexural scarps were produced after 1956 and before 1979 in the area of the main detachment zone of the 1982 landslide (Fig. 2). Most likely, similar surface displacement should be related to the 1972-74 earthquake sequence (Cotecchia, 1997). The rainfall period that occurred in Ancona 10 days before the catastrophic 1982 landslide has been characterized by an amount of precipitation not particularly relevant from the hydrologic point of view. Therefore, a fundamental role for the generation of the 1982 landslide has been played by the coseismic opening of the numerous tectonic fractures illustrated by the analysis

Table 1: Most relevant earthquakes of the Ancona, 1972, seismic sequence (after Catalogue of Strong Italian Earthquakes on the Web, ING-SGA, http://storing.ingrm.it/cft/index.htm)

date	time	lat	long	Io	Imax	sites	ref	epicentral zone
1972 01 25	20 25 11	43.62	13.35	6.0	7.0	24	102	Medio Adriatico
1972 02 04	02 42 53	43.58	13.30	7.5	8.0	75	102	Medio Adriatico
1972 02 04	09 19 04	43.58	13.28	7.5	7.5	56	102	Medio Adriatico
1972 02 05	01 27 00	43.60	13.50	6.0	7.0	2	102	Medio Adriatico
1972 02 05	07 08 42	43.65	13.33	7.0	7.0	6	102	Medio Adriatico
1972 02 05	15 14 48	43.60	13.50	6.0	7.0	3	102	Medio Adriatico
1972 02 06	01 34 14	43.60	13.50	7.0	7.0	1	102	Medio Adriatico
1972 06 14	18 55 46	43.58	13.42	8.0	8.0	17	102	Medio Adriatico

of pre- and post-earthquake airphoto coverages.

Figure 2. Main geomorphic features in the area of the Ancona landslide, including the location of fractures and flexural scarps most likely generated during the 1972 earthquake sequence (modified after Cotecchia, 1997).

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3. Camerino (MC) - Colfiorito (PG). 1997 Central Italy Earthquake: tectonic ground rupture along the Costa Fault (Colfiorito) and large scale gravitational phenomena near Sellano (PG)

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Introduction

The Colfiorito basin has been the epicentral area of a major seismic sequence in 1997. During the visit in the Colfiorito area we will have the opportunity to observe two sites where coseismic ground displacement has been documented, the Costa fault and the Sellano sites. The interpretation of the tectonic vs. gravitational nature of this ground displacement has generated quite a long debate. We will present the data, and the interpretation that is in our opinion better supported by the data. This presentation will be mostly based on the paper by Vittori et al. (2000). During the field trip we will stop at the Costa fault site and at the Sellano gravity fracture site only; the section including the relative descriptions are marked by an underlined title. However, this guidebook also illustrates several other sites in the Colfiorito area, which we will not have the opportunity to check in the field, but are relevant for understanding the issues discussed below.

Most relevant issues

Since September 1997, the Umbria - Marche region in Central Italy has been affected for several months by a seismic sequence, which has caused the loss of 12 lives and severe damage to ca. 300 localities, including many old historical towns, as Assisi, Camerino and Foligno (Fig. 1). A first shock on September 3 (at 22:07 GMT, Ml 4.5) was followed, on September 26, by two moderate earthquakes (Fig. 2A) with epicenter in the Colfiorito area (Mw = 5.7 at 00:33 GMT and Mw = 6.0 at 09:40 GMT), which produced the highest damage (Intensity IX-X MCS, Camassi et al., 1997), and by two other main shocks on October 3 (Mw = 5.4) and 14 (Mw = 5.7), the last one with epicenter more to the south, near Sellano (Fig. 1). Due to the vicinity of the main shocks, the macroseismic intensity rating, based on the MCS scale, depicts the cumulative effect of several events (Camassi *et al.*, 1997). All these events were characterized by shallow focus (4 to 9 km of depth) and nucleated along northwest-southeast-striking and southwest-dipping faults with dominant normal dip slip components, as indicated by a) the tectonic structure (e.g. profile in Fig. 1) and the geological evidence discussed in this paper, b) the focal mechanism solutions and the distribution of aftershocks (Boschi and Cocco, 1997; Morelli *et al.*, 1998; Cattaneo *et al.*, 2000) and c) the geodetic measurements, in particular GPS data and SAR interferometry (Fig. 2C; Salvi *et al.*, 2000).
Since the very morning of Sept. 26 a field survey was conducted in the epicentral area, aimed at describing the earthquake environmental effects (Figs. 2 and 3). Among these effects, which included typical features such as ground fractures, slides, hydrogeological anomalies, we observed, along the west-verging northwest-southeast-trending normal faults bordering the

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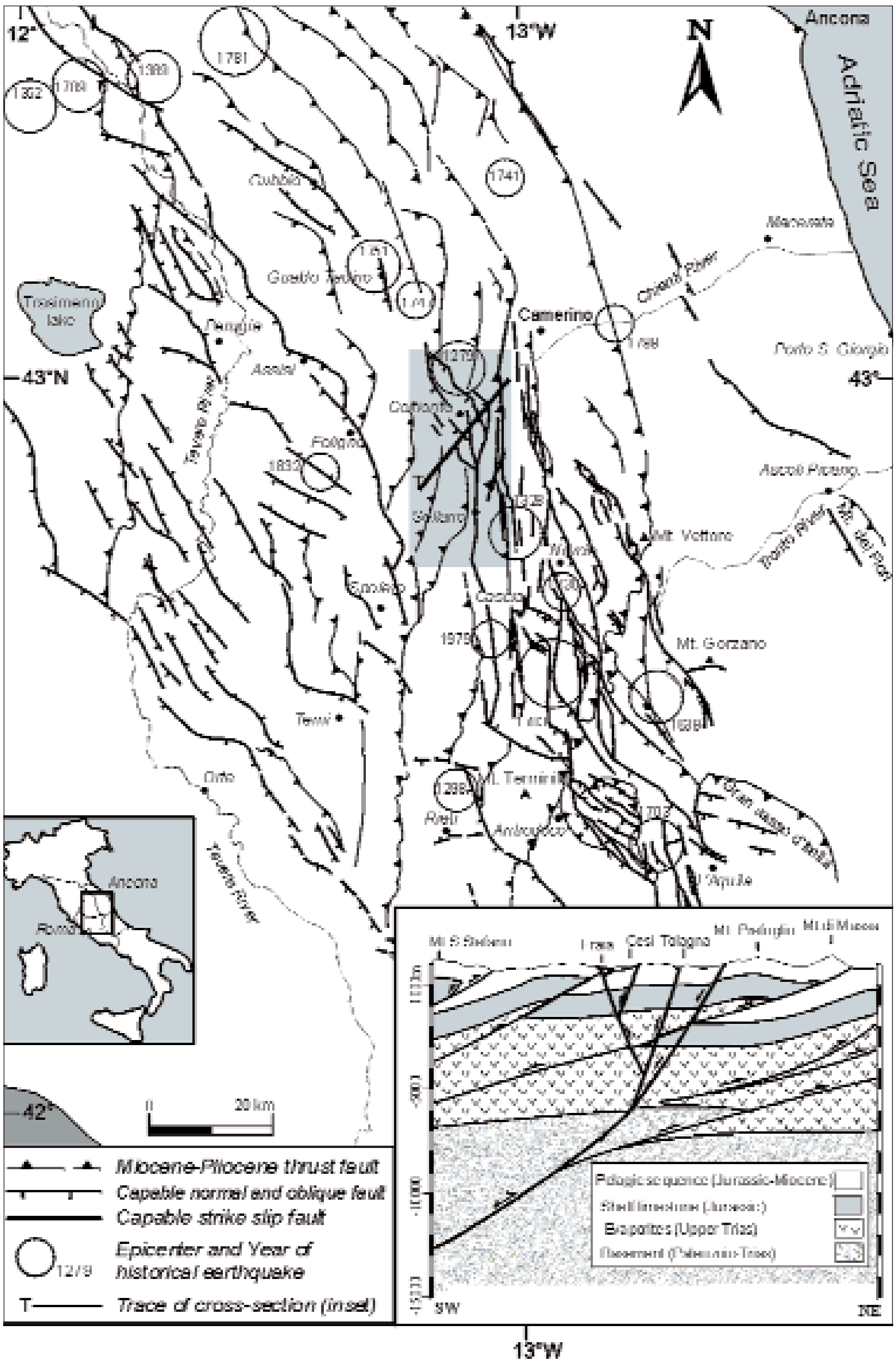


Fig. 1. Structural sketch of Central Italy (from Calamita and Pizzi, 1993; Vittori et al., 1995; Cello et al., 1997); circles mark epicenters of I ≥ IX MCS earthquakes (from Boschi et al., 1995; 1997; Camassi and Stucchi, 1998), shaded area shows location of figure 2. Inset is an interpretative cross-section of the Colfiorito area, based on surface geology and seismic reflection lines. (From Cello et al., 1998a, modified).

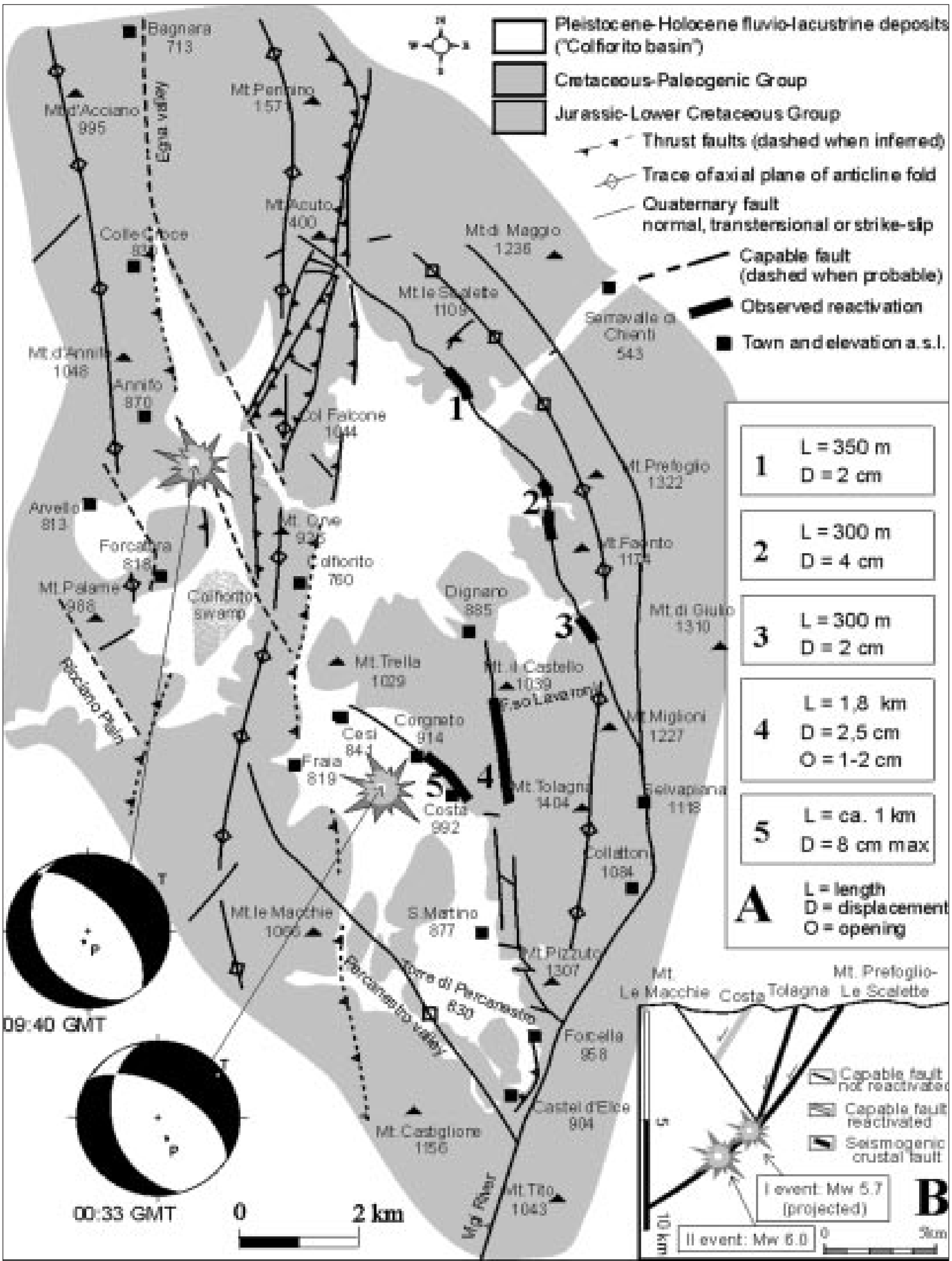


Fig. 2 Surface faulting during the September 26, 1987, earthquakes drawn over the map of capable faults of the Colfiorito basin (from Cello et al., 1998a, modified); A) observed vertical displacement, length, and opening of surface ruptures at sites 1 to 5 described in text. Epicenter locations are inferred from preliminary instrumental data and ground effects distribution (see Fig. 4); focal mechanisms are after Harvard CMT catalogue; B) Interpretative section showing the inferred prosecution at depth of the main capable and reactivated faults and the hypocenters of the 26 September 1997 earthquakes.

intermontane Colfiorito tectonic basin and its southern prolongation, systematic normal slips in the range of 2 to 8 centimeters inside discontinuous rupture zones with an overall length of ca. 12 kilometers.

One delicate question, which has arisen among Italian geologists and geophysicists (Boschi and Cocco, 1997; Basili *et al.*, 1998; Cello *et al.*, 1998a), is the interpretation of the displacement observed along sections of these fault scarps (slickensides) as true coseismic faulting or as shaking effects, without a direct link to the seismogenic slip. As a matter of fact, the level of magnitude of the main events, between 5.5 and 6.0 (Mw), is in the lower boundary of observed cases of surface faulting during recent crustal earthquakes worldwide (Wells and Coppersmith, 1994). However, subdued surface faulting for such moderate earthquakes might be more common than expected, being easily missed or misintepreted. Deciding if the ground effects were due to surface faulting or to simple shaking is not a trivial matter, since it has enormous impact, for example, on how to scale the coseismic offset seen in paleoseismic trenches or along the slickensides, in order to assess the magnitude of the causative event.

In fact, the essential questions for seismic hazard assessment (SHA) (e.g., dePolo and Slemmons, 1990) where geological studies (*paleoseismology*, e.g., Vittori *et al.*, 1991) prove fundamental are: 1) is the last event typical (*characteristic*) for the area? 2) if a larger event is foreseeable, how big it will be? 3) what are reliable slip-rate and recurrence interval for the damaging events? We therefore briefly analyze the distribution and characteristics of the environmental effects related to the main shocks of the sequence, as observed in the field (see Cello *et al.*, 1998a; and Esposito *et al.*, 1998, for their detailed description), illustrating their relevance for SHA. Subsequently, we will focus on the coseismic surface displacement occurred between Renaro and Mevale, near Sellano, a feature, that in our opinion can be interpreted as the reactivation of a large-scale gravitational movement. The comparison among the tectonic rupture at Costa and the gravitational rupture near Sellano will allow us to discuss in the field the relative morphogenic role, the relations, and the criteria for discriminating, between tectonic-earthquake and gravity-driven deformation features in the active crustal extension setting of the Apennines.

Geological and seismological framework

The Umbria-Marche region is part of the east-verging Neogene thrust and fold belt of Central Italy (Bally *et al.*, 1988; Patacca *et al.*, 1992; Calamita *et al.*, 1994). Middle-late Quaternary normal faults and related intermontane tectonic depressions, overprinted onto the thrust structure, are the preferential *loci* of strong to moderate earthquakes, as shown by historical and instrumental data (Fig. 1) and geological evidence (e.g. Cello *et al.*, 1997; Vittori *et al.*, 1997). In particular, this system of capable normal faults is

represented in the epicentral area of the 1997 earthquake sequence by 3 main normal fault segments: the central Colfiorito segment, the Norcia segment to the SE, and the Gubbio segment to the NW.

The Colfiorito basin (Fig. 2A), elongated in a NNW-SSE direction, is a typical actively expanding, fault-bounded, depression inside the Apennines; its flat valley floor is slightly above 800 m a.s.l. and is now artificially drained toward NE in the Chienti river valley, being the natural drainage hindered by the recent activity of the Colfiorito border fault. The surrounding mountains reach maximum elevations of ca. 1570 m a.s.l. (Mt. Pennino) and are mostly made of mesozoic limestones and marls. The valley fill, not exceeding 120 m, is made of alluvial and lacustrine sediments, whose deposition started about 1 million years ago, slightly before the Jaramillo palaeomagnetic event (Coltorti *et al.*, 1998). Several authors had already evidenced Late-Pleistocene to Holocene dip-to-oblique slip offsets along the main faults bordering the epicentral area, with maximum cumulative Quaternary

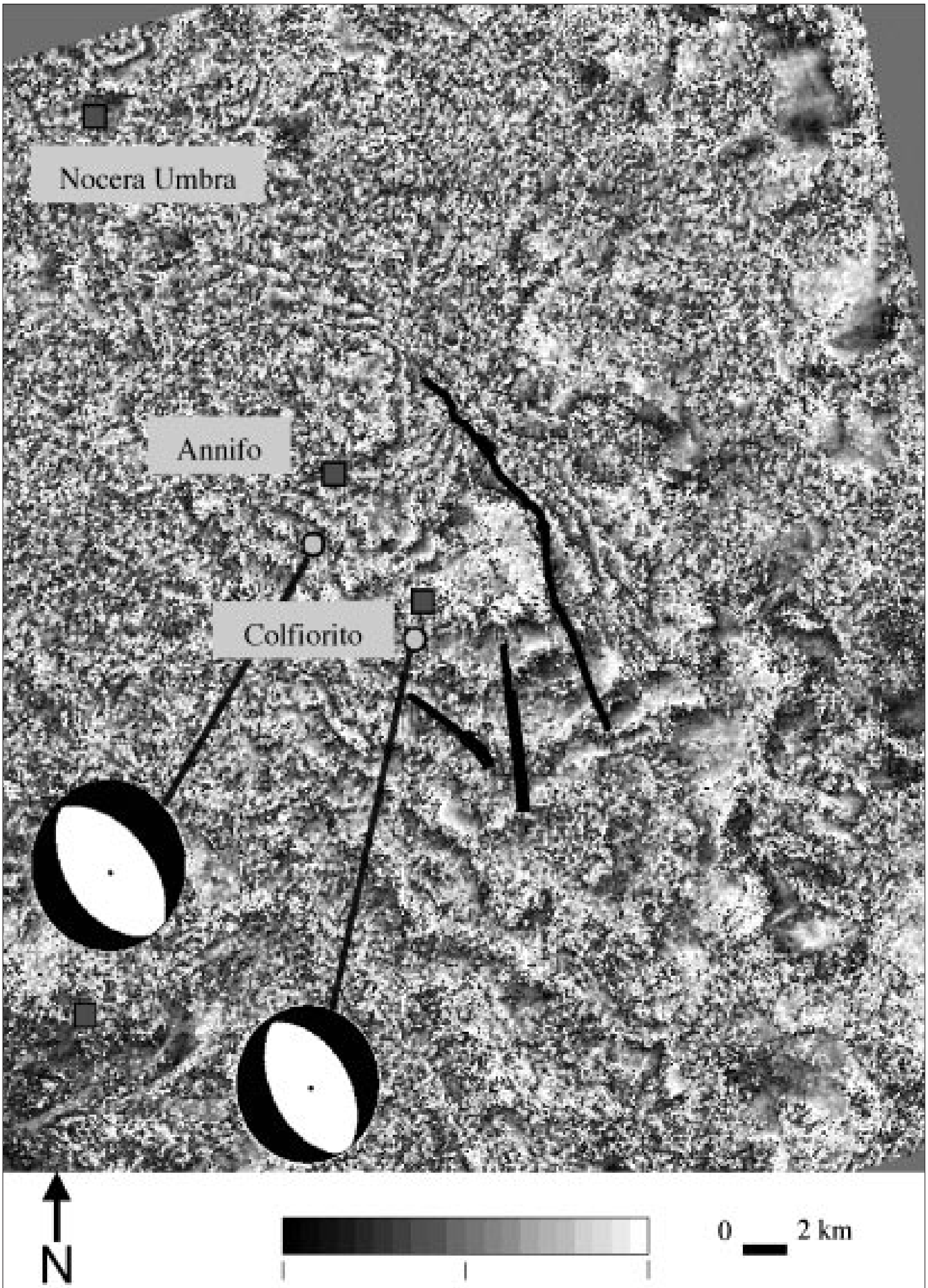


Fig. 2C. September 26, 1987, Colfiorito earthquakes Mw = 5.7 – 6.0: focal mechanisms, displacement field from radar interferometry (1 fringe = 28 mm; Salvi et al., 2000) and surface faulting (Cello *et al.*, 1998).

stratigraphic offsets in the order of 150 to 200 m (e.g., Centamore *et al.*, 1978; Calamita and Pizzi, 1993). In particular, Cello *et al.* (1997) and Tondi *et al.* (1997) had mapped in detail capable (*sensu* IAEA, 1991; i.e., the subset of active faults with the potential for surface rupture, commonly associated to moderate to strong crustal earthquakes) normal faults in the region including the Colfiorito basin, also emphasizing their potential for coseismic ground displacement. Furthermore, historical catalogues show that events of similar and even larger size repeatedly affected the Umbria-Marche region over the last millennium (Boschi *et al.*, 1995; 1997; Monachesi and Stucchi, 1997; Camassi and Stucchi, 1998).

The available instrumental data for recent moderate events in the Umbria Marche region show that typical hypocentral depths are in the order of 6 to 15 km (Haessler *et al.*, 1988; Deschamps *et al.*, 1992), and that the prevailing focal mechanism solutions suggest mostly normal faulting along roughly NW-SE-trending structures (Anderson and Jackson, 1987; Scarpa, 1992; Montone et al., 1997).

To allow a comparison of the ground effects of the 1997 sequence with historical documentation, we provide a short summary of the preceding events and their known effects on the environment. Table 1 and Fig. 1 show a selection of the most relevant seismic events included in the presently available historical catalogues and syntheses (Postpischl, 1985; Boschi *et al.*, 1995; Boschi *et al.*, 1997; Castelli, 1997; Camassi and Stucchi; 1998). These earthquakes were generally felt in a wide area of Central Italy including Rome, and occurred in long seismic crises sometimes preceded by foreshocks. The magnitudes given here for pre-instrumental events are those listed in the catalogues, based on correlation estimates between intensity and magnitude for instrumental data. Intensities are given in the MCS (Mercalli-Cancani-Sieberg) scale, particularly suitable for masonry constructions; see for example Reiter (1990) for a comparison with MM and MSK scales. A generalized map of the earthquake ground effects reported in the region for historical and modern events is presented in Fig. 3. A more detailed discussion of this topic can be found in Esposito *et al.* (2000).

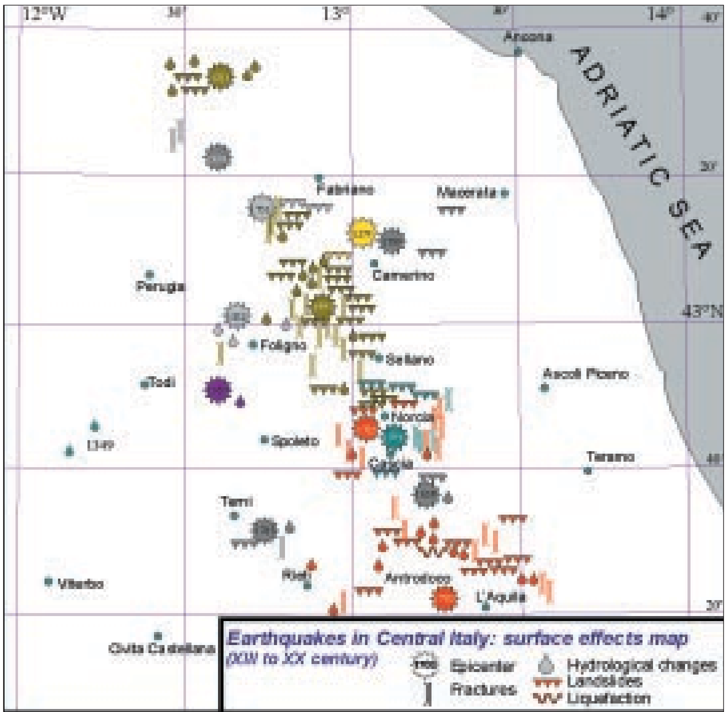


Fig. 3. Distribution of ground effects induced by historical earthquakes since XIII century.

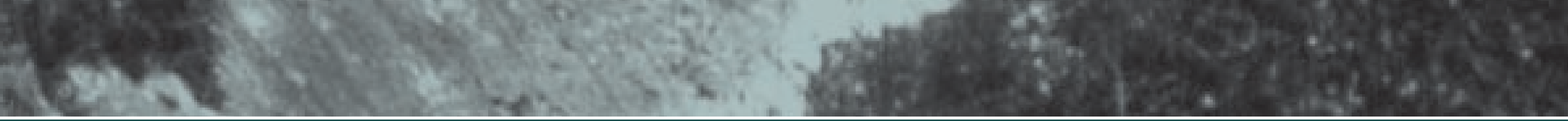


Table 1. Major historical and recent earthquakes in Umbria-Marche and surrounding region (I max ≥VII MCS) and inferred magnitudes as listed in NT4.1 (Camassi and Stucchi, 1998) and CFTI (Boschi *et al.*, 1995; Boschi *et al.*, 1997) catalogues.

Date	Epicenter	I max NT4.1	magnitude NT4.1	magnitude CFTI
268 b.C.	Picenum	-	-	-
100 b.C.	Picenum	8,5	-	5,8
99 b.C.	Norcia	9	-	5,6
1279 04 30	Camerino	10	6,7	6,6
1298 12 01	Reatino	10	6,4	6,3
1328 12 01	Norcia	10	6,7	6,2
1349 09	Valle del Salto	9.5	6,4	6,7
1352 12 25	Monterchi	9	6,2	5,7
1389 10 18	Bocca Serriola	9	6,2	6,0
1458 04 26	Città di Castello	9	6,2	5,4
1599 11 05	Cascia	8,5	5,9	-
1639 10 07	Amatrice	10	6,7	5,4
1703 01 14	Norcia	10	6,7	6,5
1703 02 02	L'Aquila	9	6,2	-
1719 06 27	Valnerina	7,5	5,2	-
1730 05 12	Norcia	9	5,9	6,4
1741 04 24	Fabrianese	9	6,2	6,4
1747 04 17	Fiuminata	9	6,2	5,7
1751 07 27	Gualdo Tadino	10	6,7	6,0
1781 06 03	Cagliese	10	6,4	6,0
1785 10 09	Piediluco	8	5,5	5,6
1789 09 30	Val Tiberina	9	5,9	5,4
1791 10 11	Scopoli	7,5	5,2	-
1799 07 28	Camerino	9,5	6,2	5,6
1832 01 13	Foligno	8,5	5,9	5,7
1838 02 14	Valnerina	8	5,5	-
1859 08 22	Norcia	8,5	5,9	-
1917 04 26	Monterchi	9,5	5,6	5,7
1930 10 30	Senigallia	8,5	6,0	5,9
1972 02 04	Ancona	8	4,5	5,6
1972 06 14	Medio Adriatico	8	4,3	5,4
1974 12 02	Monte Fema	7	4,7	-
1979 09 19	Norcia	8,5	5,9	5,8
1984 04 29	Gubbio	7	-	5,6
1987 07 03	Porto S.Giorgio	7	4,9	-
1997 05 12	Massa Martana	7	4,5	-
1997 09 26	Colfiorito	9,5	6,0	-

Ground effects of the 1997 umbria-marche earthquakes

Based on field investigation, we recognized ground effects in a ca. 700 Km^2 wide roughly elliptical area around the Colfiorito epicentral zone (Fig. 4) (Cello *et al.*, 1998a; Esposito *et al.*, 1998; Esposito *et al.*, 2000). The observed phenomena were classified as *primary effects* (surface faulting) and *secondary effects* (ground fractures, landslides, local ground settlements and hydrological changes). The former results from the propagation up to the

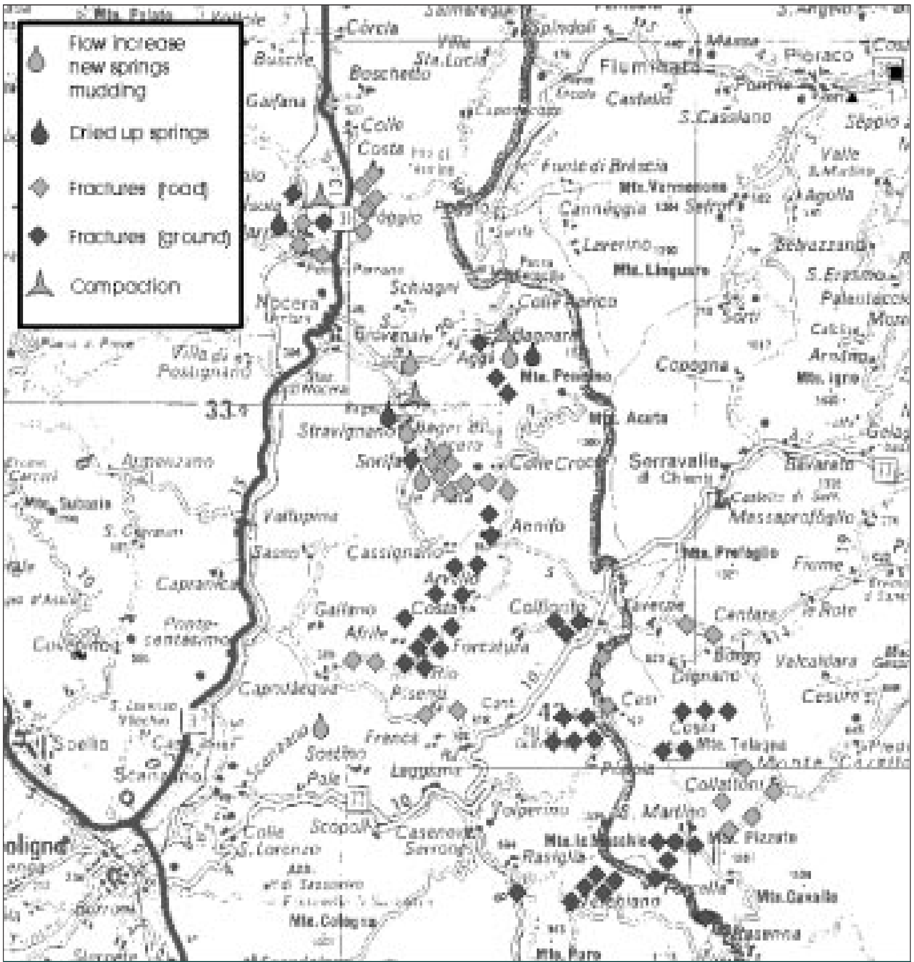


Fig. 4. Distribution of ground effects of the 1997 Umbria-Marche sequence (after Esposito *et al.*, 2000)

surface of the seismogenic slip at depth, giving direct evidence of the size and kinematics of the earthquake. Instead, the presence, activation threshold, spatial distribution, number and size of secondary effects depend on frequency content and duration of seismic shaking, which is a function of the local geology (stratigraphy, water saturation, morphology) and of the travel path of seismic waves, as well as of the source parameters.
In the following sections we describe the secondary effects first, because they have also served as indicators of the most likely location of primary faulting. However, the gravitational feature at Sellano is illustrated after the primary effects, to allow better comparison between surface faulting features and non-tectonic earthquake ruptures.

Secondary effects

Ground fractures, either in Quaternary colluvial, and fluvio-lacustrine deposits and in rock formations, were concentrated in a NNW-SSE elliptical area (Fig. 4) and mainly struck conformably with the reactivated faults (inset in Fig. 5A). Their highest concentration and dimensions well fit the macroseismic and instrumental epicentral area. About the typology of the phenomena (examples in Fig. 5A), the open fractures (tension cracks) were prevalent, from a few meters up to hundreds of meters long, generally from some millimeters to a few centimeters wide and with relative vertical displacements ranging from a few millimeters to 5 centimeters.

The most common landslides were represented by rock falls (ca. 60%), probably due to the relatively high vulnerability of the outcropping formations in the epicentral area, mostly consisting of densely fractured limestone and marly limestone, sometimes interbedded with clay and sandstones; rotational and translational slides also occurred (ca. 35%), with only a few cases of earth flows.

Most of the rock falls, consisting of some cubic meters of stones, occurred along the artificial scarps (road cuts) which border the main and secondary roads. Sliding slopes involving considerably larger volumes of material occurred at Sorifa (thousands of cubic meters), at Stravignano Bagni (hundreds of cubic meters), and in Val Nerina (blocks larger than 10 cubic meters).

Rotational sliding phenomena mainly occurred in debris deposits. Three interesting cases were observed respectively at Afrile, at middle slope of the valley right above the Acciano soil dam and on the down-slope side of the road Bagnara-Colle Croce. For these landslides were observed: arcuate crowns some hundreds of meters long, open fractures developed along the main scarps, about tens of centimeters wide and with vertical displacements varying from units to tens of centimeters (Fig. 5B). At Monte d'Annifo an already existing rotational slide was re-triggered, with a main scarp about 200 m long, and a total vertical displacement of about 40 cm. It must be noted that some of these slides continued their motion for days after the Sept. 26 quake, for example the Acciano slide.

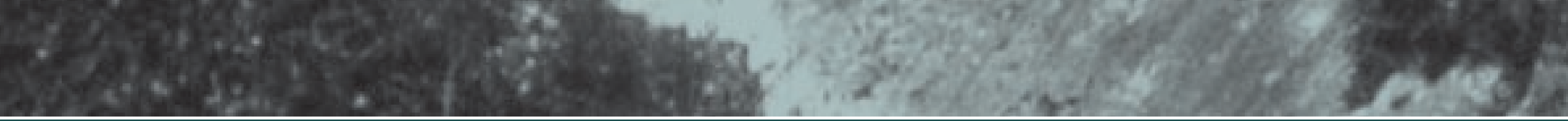
The landslide distribution indicates that the area of maximum density is consistent with the VIII-IX MCS damage level. Most of the landslides occurred within a distance of 10 km (ca. 60 % of the total), their number decreases progressively within a distance of 20-25 Km (Esposito *et al.*, 1998). It should be noted that, fortunately, the very dry conditions of the late summer season prevented the occurrence of many rotational slides, being most of the area very prone to this phenomenon in wet conditions (Guzzetti and Cardinali, 1989).

For the same reason, no liquefaction was observed, but only localized ground compaction and artificial fill settlements, mainly concentrated near Nocera Umbra. At Le Molina (Nocera Umbra), a road entering the town suffered noticeable settlements (some centimeters). This phenomenon was enhanced by the contact between the road and a little bridge, whose rigid structure (made of reinforced concrete) laid on a large wall of gravel caissons stiffer than the road subsoil.

A complex phenomenon was observed in an earth dam near Acciano, where significant settlements deformed the top rim of the dam. Some small deformations observed along one of the two slopes of the dam suggest that both compaction and localized slides probably occurred.

A major surface effect occurred on the hill of the Holy Monastery in Assisi. A wide area of the *Piazza Inferiore* (lower square) in front of the church suffered noticeable vertical displacements; this area, delimited by an arch-shaped fracture and by the border of the square towards the hill, appears as the main scarp of a landslide body. Nevertheless, no clear evidence of a sliding phenomenon was observed along the slope of the hill, which is partially retained by an ancient and huge masonry wall. Hence, once again, the interpretation of the phenomenon is rather difficult, being probably caused by the combination of different earthquake-induced mechanisms; in any case the settlement of the subsoil materials appears to play a significant role.

Several hydrological effects were mainly concentrated in the area around Nocera Umbra. They include flow increase, turbid water and drying up of existing springs, and even creation of new springs. On September 26, after the first event (00:33 GMT), water mudding affected the Topino River springs near Bagnara for several hours; after the second event (09:40 GMT)



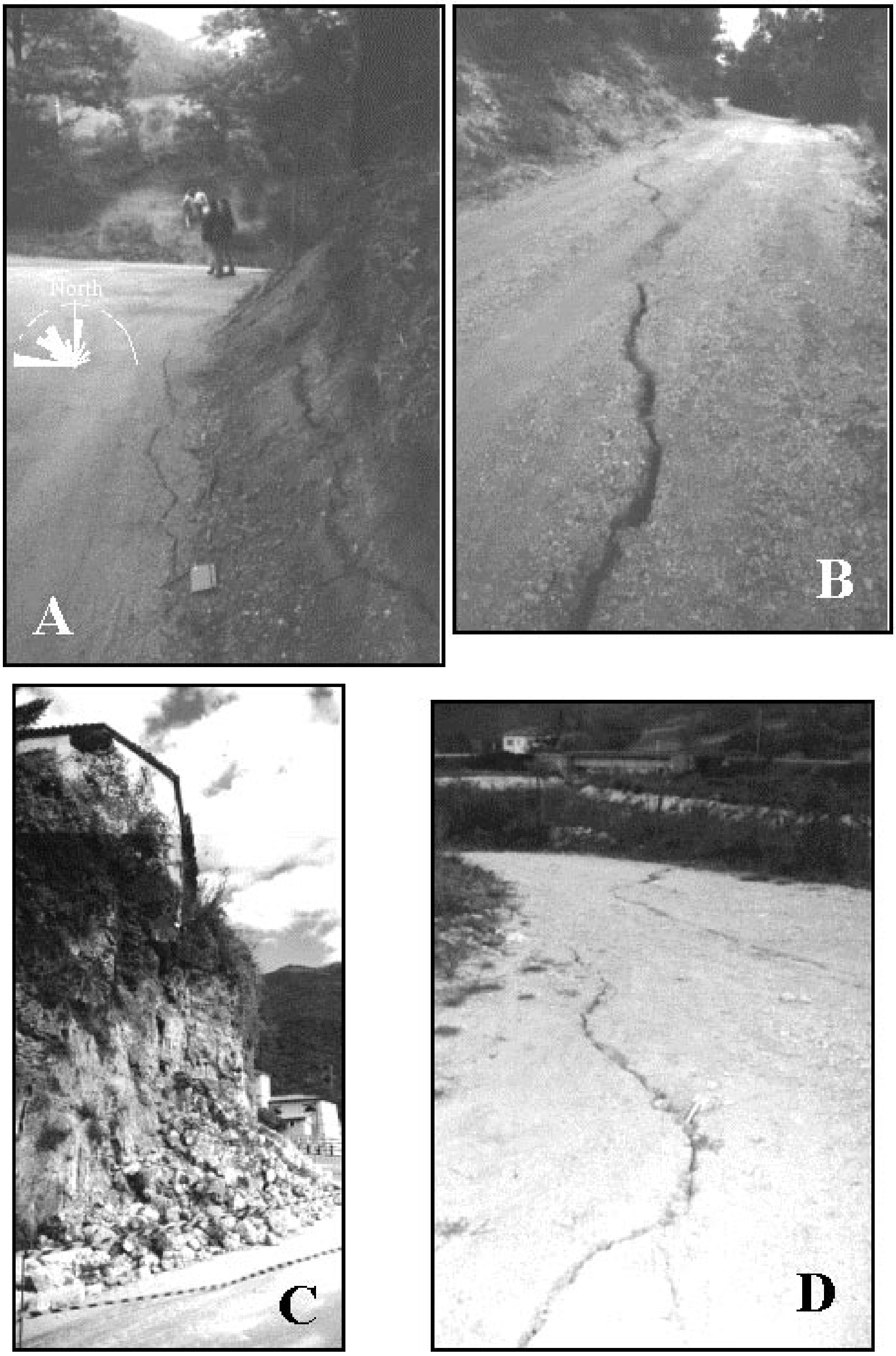


Fig. 5. Examples of ground effects of the Sept. 26, 1997, events (location in Fig. 4). A) ground crack that developed for several hundred meters near Col Pasquale; here local inhabitants noted a violent gas emission (in inset: windrose diagram of fracture orientations within the earthquakes area); B) Typical landslide crown at Afrile; C) rock fall near Stravignano; D) ground settlement in the Topino river bed near Bagni di Nocera.

Fig. 6. View of the Costa-Corgneto range front taken from SW; arrow marks the site of reactivated slickensides shown in Fig. 7 (site 5 in Fig. 2A).

the flow was interrupted for about two hours and then started again with mud water at an intensity lower than the regime flow (100 l/s). Analogous phenomena occurred at S. Giovenale water wells. Other significant hydrological effects were observed after the 9:40 event: new springs were active for a few hours in the area of Le Prata village; some springs temporarily dried up, e.g. the Angelica spring near Bagni, and a small spring that fed a fountain in the center of Isola; water mudding affected for several hours the Roveggiano stream at Capodacqua and the Montenero spring at Campodonico.
Finally, gas emissions and some variations in chemical parameters of spring waters were observed at different locations (Calcara *et al.*, 1997), also outside the epicentral area (e.g., Umbertide town, Martinelli and Albarello, 1997).
In general, it was noted a level of secondary effects in good accordance (Esposito *et al.*, submitted) with what expected for the corresponding levels of intensity based on the historical database (Serva, 1994).

Primary effects

Based on the preliminary appraisal of the distribution of damage and environmental effects, we checked all the already known capable faults located within the epicentral area (Colfiorito basin, Fig. 2A) in the hours and days following the main shocks, observing a subtle but systematic slip on some of them (see Cello *et al.*, 1998a, for a more detailed description). These faults, which border the range fronts, generally mark the interface between the Meso-Cenozoic calcareous or marly bedrock and the late Quaternary slope deposits, typically related to the latest Glacial age (20 to 15 ka B.P.) (Coltorti and Dramis, 1987). In general, the scarps carved in calcareous rock are well exposed and often display up to 2-3 meters high well preserved slickensides.
We mapped surface ruptures along several segments of the main faults (Fig. 2): a) Costa-Corgneto segment of the Cesi-Costa fault; b) Tolagna segment of the Dignano-Forcella fault; c) Mt. Le Scalette, Mt. Faento and La Pintura segments of the Colfiorito border fault. The Costa rupture occurred during the 00:33 shock, because the road cracks associated to it were observed before the second 09:40 event; for the other cited ruptures we cannot discriminate which was the causative event between the two. The coseismic motion was generally normal dip-slip, with only a minor left-lateral component near the Costa village. These faults have been re-surveyed several times during a period of more than one year to document possible additional slips, which have not occurred, and the morphological evolution of the free-faces.

Costa-Corgneto segment This fault segment (Fig. 2A, site 5) trends ca. N130 and shows a fresh-looking slickenside for most of its length, generally marking the contact of Mesozoic limestone with stabilized and cohesive slope scree, which hosts a dense wood (Fig. 6). Behind Costa, a coseismic slip affected the slickenside determining a continuous 7-8 cm high free-face over a length of 80 m, marked by a brown strip due to a veneer of soil coating the base of the fault plane (Fig. 7A-

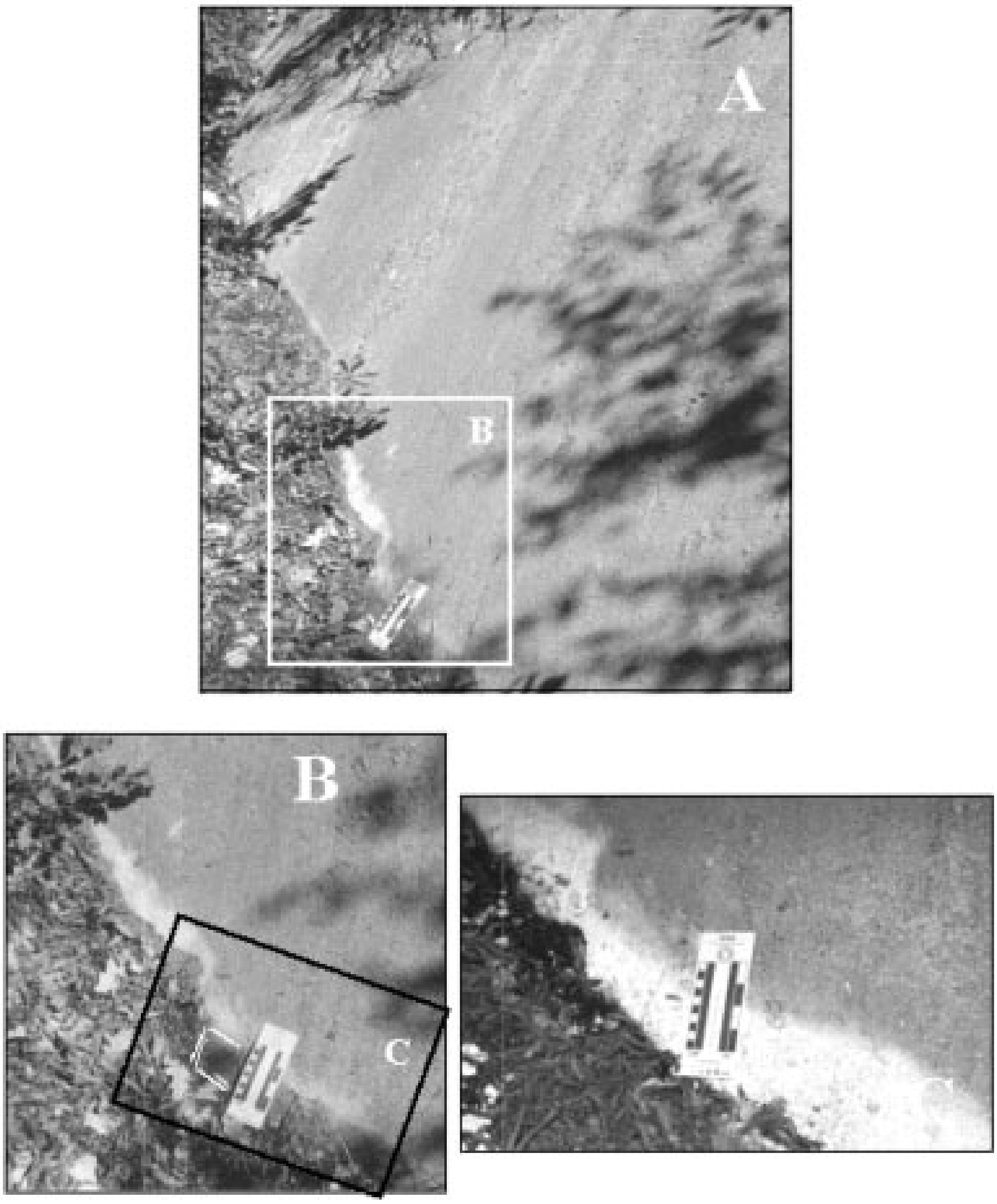


Fig. 7. A) Slickenside with basal 7-8 cm high free-face along the Costa bedrock fault scarp; B) Detail of the free-face, marked by a brown soil-coated strip. Photographs taken on September 27, 1997; C) Detail of the free face taken about one year later, on October 1998.

7B). One year later a pale strip had remained (Fig. 7C), while the soil had been washed out. A similar reactivation was observed on a 40 m long segment of the limestone scarp near Corgneto, where the brown strip marking the free-face was 4 cm high. Between Corgneto and Costa, and SE of Costa, the bedrock fault scarp showed only local evidence of tension gashes in the soil. In the SE termination of the Costa scarp free-face, where the bedrock is present in both sides of the slickenside, we observed small faults and fractures in the slope immediately below the fault scarp and along the dirt road connecting Costa to the Mt. Tolagna area. Very likely, the few centimeters of measured coseismic fault slip were, here and between Costa and Corgneto, spread over a network of minor hanging-wall fractures. Hence, we have concluded that the end-to-end rupture length along this segment is about 1 km.

Other relevant sites

Tolagna segment - At Fosso Lavaroni (Fig. 2A, site 4; Fig. 8) the bedrock scarp, N160 trending and SW dipping, 1 to 1.5 m high, juxtaposes slope deposits against Mesozoic limestone. Over a length of ca. 200 m the coseismic motion produced a continuous, 2.5 cm high free-face. At the SE termination of the scarp, it displaced of the same amount a dirt road. Here the fault plane is exposed in the road cut and the observed thickness of slope deposits is minimal. Moving south after the road, the bedrock fault scarp maintains the same height, but, being completely mantled by soil, it shows only diffuse ground cracks at its base over a distance of 1.6 km along the slope of Mt. Tolagna; hence, the cumulative end-to-end length of the ruptures associated with this fault segment is ca. 1.8 km.

Mt. Le Scalette, Mt. Faento and La Pintura segments - At Mt. Le Scalette (Fig. 2A, site 1) N130 to 150 trending, SW dipping, fresh slickensides outcrop along a 2 to 4 m high bedrock fault scarp. This segment records the maximum stratigraphic offset of the whole Colfiorito border fault, that is 150 to 200 m. At its SE termination, the fault displaces middle Pleistocene and recent lake sediments, damming the NE-ward drainage of the Colfiorito basin into the Chienti River valley (Centamore *et al.*, 1978). The coseismic offset in this area produced free-faces 2 to 4 cm high over a discontinuous length of ca. 350 m (Figs. 9 and 10). As well, along the slope of Mt. Faento (Fig. 2, site 2) we observed a 4 cm high basal free-face along a ca. 300 m long section of the slickenside in the Maiolica Fm. in contact with slope deposits.

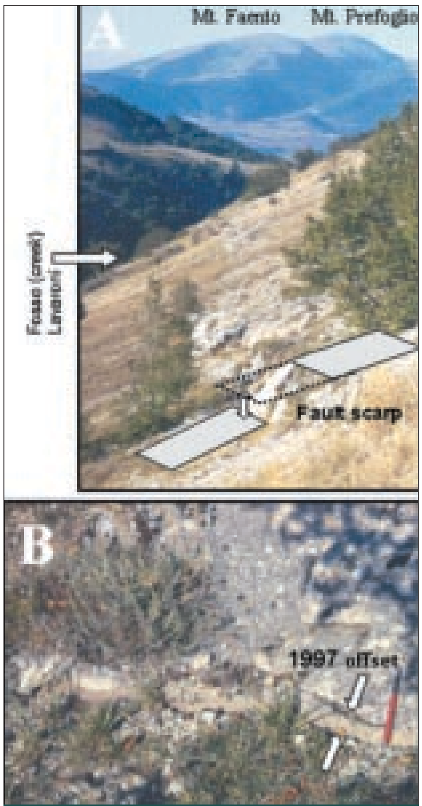


Fig. 8. A) View of the Tolagna bedrock fault scarp near Fosso Lavaroni (site 4 in Fig. 2A). The free-face height is remarkably uniform (2.5 cm) while slope deposits thicken from SE (right side of this image) to NW; B) detail of the free face, partially coated with soil. Photographs taken on September 27.

At La Pintura (Fig. 2, site 3) the coseismic rupture cut obliquely across the slope intersecting a winding dirt road at two sites. Along this segment the observed rupture length was ca. 300 m with a maximum displacement of 2 cm.

The mapped surface breaks provide a minimum value for rupture length assessment

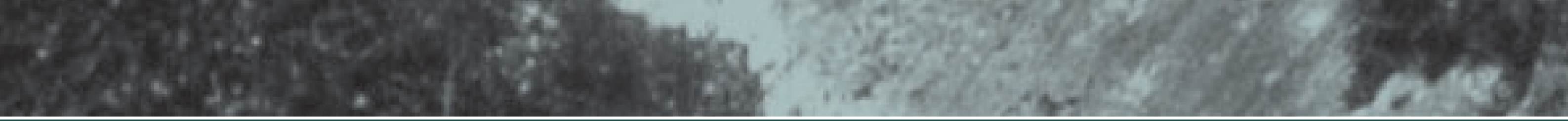


Fig. 9. A) View of the Mt. Le Scalette range front from SW; note the bedrock fault scarp with the reactivated slickensides; B) Detail of the free-face (4 cm high, brown soil-coated strip). Photographs taken on October 3.

along this fault. As noticed at La Pintura, the fault reactivation can be precisely mapped and the offset measured, only where reference features such as slickensides or road cuts occur along the fault trace. Likewise, ground breaks having only a few centimeters of slip are generally undetectable to map where the wood covers the fault zone. Therefore, Cello *et al.* (1998a) have estimated the minimum rupture length of the Colfiorito border fault as the end-to-end reactivation between Mt. Le Scalette and La Pintura, i.e. no less than ca. 6 km (Fig. 2).

Evidence for large-scale coseismic mass movements near Sellano

Either on September 26 and October 14 other surface ruptures occurred unrelated to known capable faults, for instance near Fraia, Sellano, Renaro, Mevale and Rasenna (Fig. 4). In particular, discontinuous superficial breaks occurred over a distance of about 1.7 km between Renaro and Mevale during the October 14 shock (Fig. 11). Near Renaro the fracture was up to 5 cm wide on October 14, and opened up to ca. 10 cm until October 29. The rupture partly run within the Scaglia Fm. parallel to bedding. Some authors (Basili *et al.*, 1998) attribute to the Renaro-Mevale ruptures a tectonic significance. Addressing the interested readers to them for details of their interpretation, we underline that the tectonic significance of these features is questionable in our judgment mainly because of a) the morphological and structural setting, for example the lack of a normal fault in the vicinity of the ruptured zone with any evident pre-existing morphological and stratigraphic offset; b) the displacement was not instantaneous, but continued for days after the quake; c) the displacement was essentially in the shape of a tension crack up to 10 cm large, without appreciable vertical or lateral components. In our opinion, this feature might be interpreted in several ways, all tectonically controlled but related to an indirect expression of the faulting at the hypocenter. Most likely, it was a gravity-related *sackung*-like phenomenon (e.g. Beck, 1968; Radbruch-Hall *et al.*, 1976), very common feature all along the Alps and the Apennines (e.g., Blumetti, 1995, and references therein). Less likely, it may also be considered a *subordinate rupture*, as defined in Clark *et al.* (1991) or a *fraglia* as defined in Serva (1995). This last term comes from *frana* (landslide) and

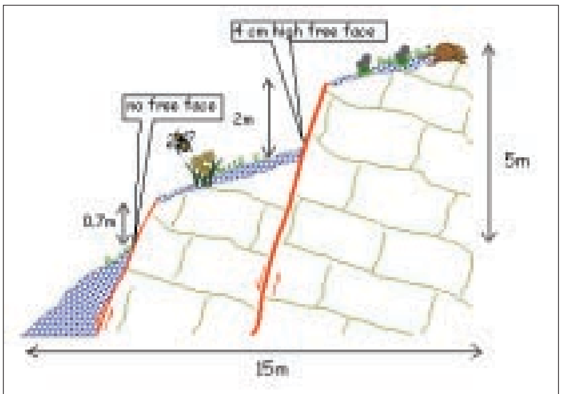


Fig. 10. Interpretative geological section across a 15 m long portion of the Mt. Le Scalette fault, where coseismic slip occurred within the limestone bedrock.

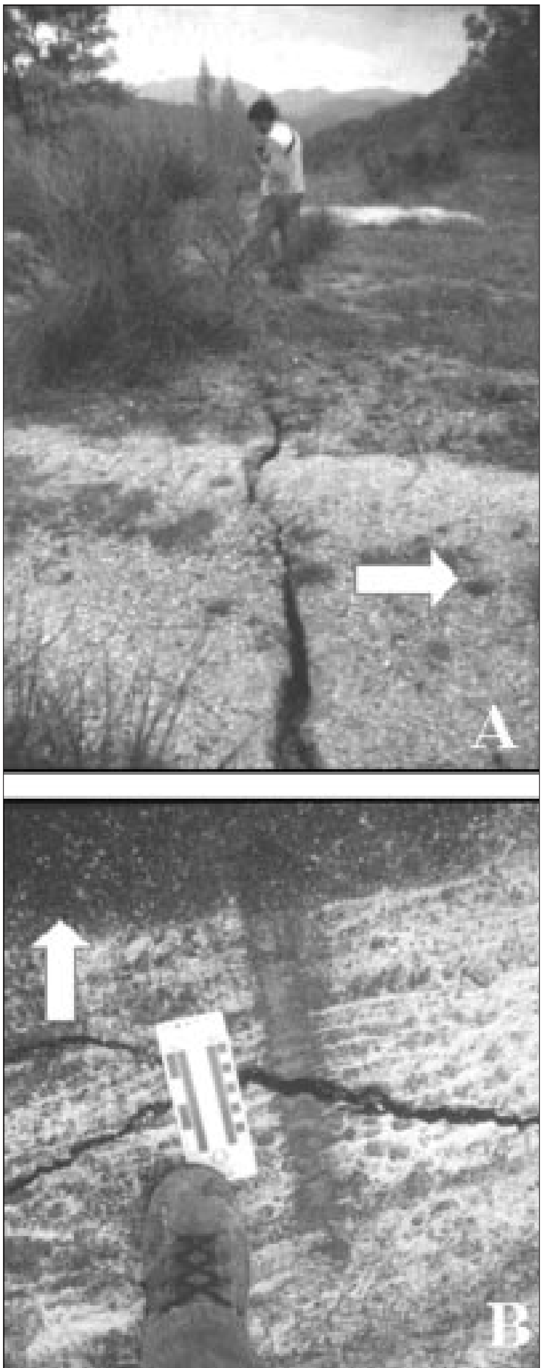


Fig. 11. A) Tension gash up to ca. 10 cm wide near Renaro (ca. 5 km south of Rasenna, Fig. 4) along a fracture zone extending NNW-SSE for ca. 1.7 km between Renaro and Mevale originated during the Oct. 14, 1997, earthquake; B) Some ten meters to the north, the same fracture crossed a paved road, which, repaired the day after the earthquake, showed that this fracture was still widening on Oct. 29, 1997.

faglia (fault) and has been proposed to describe a set of features, still poorly studied, resulting from a combination of these two phenomena, i.e., fault-controlled very deep landslides.

It is important to note that the Sellano, October 14 earthquake ruptured a fault section located at the SE termination of the Colfiorito normal fault segment. In the Apennines many remarkable, large-scale, gravity deformations occur at or near normal fault segment boundaries.

Discussion

The significance of the observed slip along faults bordering the Colfiorito basin has caused a great debate among Italian scientists, having been interpreted either as (1) the compaction of the debris deposit, (2) a gravity effect or (3) a true fault reactivation.

The reliability of hypotheses (1) and (2) can be tested based on soil mechanics principles, which allow to evaluate the behaviour of the debris deposit under the seismic excitations. As a case-study, we have analysed in some detail the fault reactivation near Costa. Here the free face was practically uniform (7 to 8 cm) along the contact between the bedrock and the recent deposits, even if the latter had thickness variable from less than one meter

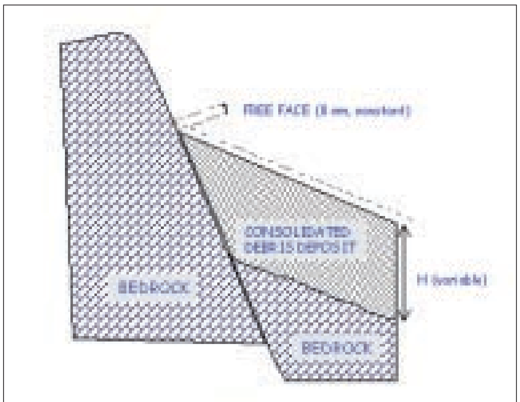


Fig. 12. Schematic section of the reactivated fault zone at Costa, used for modeling the possible causes of observed slip.

to tens of meters (Fig. 12). We have estimated the order of magnitude of the potential settlements induced by the September 26 first earthquake from the geotechnical viewpoint. First of all, it must be observed that, even if the dynamic excitation and the material stiffness are assumed constant along the fault, nevertheless the amount of settlement varies with the height of the debris deposit: in particular, on the basis of a simple model, the settlement is a quadratic function of the deposit thickness. Hence, varying the deposit height between units and tens of meters, the settlement should vary at least of two order of magnitude. A rough but effective evaluation of such settlements, adopting the most conservative values of debris stiffness (thousands of kg/cm^2) and seismic accelerations (ca. 0.5g constant in the debris deposit), gives maximum settlements of the order of a few millimeters, at least one order of magnitude lower than the measured one. In the calculation we have taken into account only the vertical component of the acceleration, being the horizontal components relatively small in the near field. In conclusion, the hypothesis (1) is not reliable.

About the hypothesis (2), it must be firstly underlined that, because of the stiffness of the debris material, the landslide mass should behave as a rigid body, without significant volumetric changes. As a consequence, noticeable variations of the shape of the deposit should be evident elsewhere (downward), consistently with the amount of measured settlements of the scarp; furthermore, a soil or debris slip would have occurred on a plane inclined less than 30°-35° (rest angle), for example along the contact with the bedrock underneath the debris deposit, thus displaying an appreciable horizontal component away from the slickenside. Instead, the relative movement occurred exactly along the contact surface between the slickenside and the debris deposit: neither tension gashes were generally observed, nor any small graben-like settlement (gravity faulting). The illustrated conclusions, based on physical considerations and on simple but effective analyses, allow to assess that the observed phenomenon is the experimental evidence of a rigid sliding movement between the fault bedrock and the debris formation, which perfectly agrees with the hypothesis (3) of the capable fault reactivation.

Taking into account the geologic and geomorphologic recent evolution of the reactivated bedrock fault scarps and the Quaternary setting of the Colfiorito basin, many other pieces of evidence sum up to indicate that the surface ruptures associated with the September 26, 1997, earthquakes in the Colfiorito basin have a tectonic significance. Coseismic surface slip occurred along pre-existing Quaternary faults responsible for the recent tectonic evolution of the area. Fresh limestone slickensides were reactivated, which are characterized by geomorphic features unequivocally related to paleoseismic surface faulting, such as a) the position within prominent bedrock fault scarps at the base of young tectonic slopes and b) the juxtaposition with thick, faulted latest Glacial to Holocene stratified slope-waste deposits. It is worth noting that such deposits must have suffered repeated seismic shakings during the Holocene (extrapolating the relatively short historical sample), which had certainly determined their settlement. The soil-coated strip observed just after the earthquake has become after about one year (during which the soil has been washed out) similar to the pale strip described for the 1915, Fucino, earthquake (Vittori *et al.*, 1991; Michetti *et al.*, 1996). The same feature has been described for recently activated limestone earthquake scarps worldwide, see for example Stewart and Hancock (1990) for the Egean region, which also discuss its genesis and evolution. Therefore, the Colfiorito surface ruptures confirm that the so-called *nastri di faglia* (Segre, 1950; Vittori *et al.*, 1991; Yeats *et al.*, 1997) are in general evidence of very recent paleoseismicity, although local erosional conditions, which a correct geomorphological analysis can easily identify, may enhance or even produce the same feature over limited (meters to tens of meters) distances.

Along the reactivated fault segments, surface offset is remarkably constant over tens to several hundreds of meters, while the thickness of slope deposits varies from a few centimeters to tens of meters. Comparable displacements have been observed at sites showing different morphological features, such as slope angle and position of the reactivated fault along the slope. Slope deposits were in dry conditions and generally stabilized by tree root penetration. Finally, we have monitored the free faces in the following days and months without detecting any additional slip, which should have been expected in case of gravity movements, due to the numerous events of the seismic sequence (some with M > 5) and the very humid environmental conditions (rainfalls and snow) that have characterised the subsequent months. As previously described, significant creep has instead widened in the following days the tension cracks opened after the Renaro-Mevale earthquake on October 14, which we are prone to interpret as a *sackung*-like movement. As well, post-seismic movements were observed in several slides in the epicentral area, all displaying features and morphological settings sensibly different from those of the reactivated fault scarps. As a matter of fact, the different long-term morphological evolution of deep-seated fault slips and superficial, landslide or sackung-like movements make it quite easy their discrimination.

Geodetic, satellite and SAR data have recognised a permanent ground deformation in the Colfiorito area (Salvi et al., 2000), which is in good agreement with a normal slip along the NW-SE-trending and SW-dipping basin border faults. As well, the aftershocks related to the Colfiorito sequence concentrate in the hangingwall of the reactivated faults. The fault parameters and seismic moments (0.4 and 1.0 $x10^{18}$ Nm) of the two main shocks, as derived from strong-motion data are compatible with the observed surface ruptures, taking into account the shallow hypocenters: 38 and 37 cm of slip along 6 and 12 km long and 6 and 7.5 km wide faults, respectively.

It is noteworthy that the application to the Colfiorito events of the magnitude vs. rupture length and magnitude vs. displacement relationships, derived from the empirical database for earthquake faulting worldwide (Wells and Coppersmith, 1994), suggests values in good agreement with what observed (4 to 8 cm maximum over lengths of several kilometers) (Fig. 13).

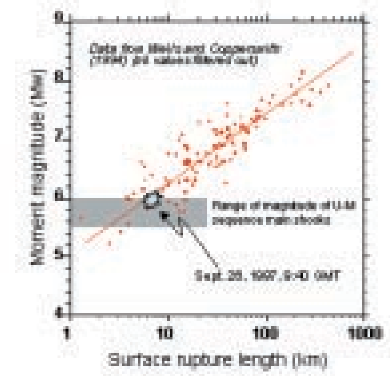


Fig. 13. Plot of Mw vs. surface rupture length from worldwide database (Wells and Coppersmith, 1994). The Colfiorito main shocks are in the lower zone of occurrence of surface faulting.

Some constraints on the seismic potential in the Colfiorito area

Following the Sept. 26 events we have observed up to 8 cm of coseismic slip. We know that at the end of the last Glacial epoch the mountain slopes were shaped as relatively regular surfaces because smoothed by a very effective climatic morphogenesis (Brancaccio *et al.*, 1979; Blumetti *et al.*; 1993; Giraudi, 1995, Giraudi and Frezzotti, 1997). Hence, we can assume the interval 18 to 14 ky (ending part of last Glacial to beginning of Holocene) as a reliable age estimate of mountain slopes in the Apennines; therefore, these slopes may be assumed as reference surfaces to measure the tectonic mobility of the capable faults of this region. The present displacement of these slopes across fault planes, estimated as 2-3 meters maximum (Cello *et al.*, 1997), is the cumulative effect of repeated coseismic slips during the Holocene. Such time span is that of major interest for SHA, although the repeat time of large earthquakes can be in the order of several thousands of years. On this basis, it is possible to estimate a Holocene slip-rate along dip of 0.1-0.2 mm/year (0.07-0.14 vertical and horizontal components for 45° dipping fault planes), without taking into account any possible strike-slip component.

Although the time span of historical information is relatively long, only the 1279 event appears to fit the last earthquake for epicenter location, intensity and distribution of effects. On this basis, we assume the 1279 event as the preceding event in the Colfiorito area, yielding an interval time of ca. 7 centuries. This would give ca. 20-25 such events since the end of the last Glacial epoch. The repetition of 25 Colfiorito-like events would determine the observed Holocene displacement with 8-12 cm of slip per event. Actually, the maximum observed slip was 8 cm (Costa segment), while 2 to 4 cm were observed on the main fault system. It is likely that such slip represents only a portion of the actual slip, being the "missing" part distributed in a way not directly visible on the ground, because of the small offset and the "ductile" characteristics of the hangingwall deposits and soil-vegetation cover. Such hypothesis might be supported by the ca. 20 cm of vertical slip measured by the geodetic network.

If we extrapolate such slips per event and return periods to a time range of one million years, which is the inception time of the basin (see Geological Background), we obtain a rough estimate of vertical offsets of ca. 100-150 m, in agreement with the stratigraphic and morphological offsets (150-200 m, Calamita and Pizzi, 1993). Hence, reasonably we may hypothesize that, being magnitude 5.5-6 the threshold for the occurrence of surface faulting, the local morphology can derive from the cumulative effect of repeating events of such magnitude, once the climatic forcing is filtered out.

Obviously, without additional constraints, it would be arbitrary to assume the last Colfiorito event as "typical". In fact, based only on the historical record, how can we rule out the possibility of previous events larger than it, may be alternating in time, and the occurrence of clustered activity? The September 26 rupture was distributed in two sub-events; if they ruptured at the same time, the size of the earthquake would have been larger, probably affecting also the amount of offset. Can this happen in the future? What seismotectonic significance should we attribute to the local geological and geomorphological parameters? A tentative answer may be given comparing the shape of the Colfiorito basin, which is made of a nested network of small depressions with limited thickness of the valley fill and is bounded by fault segments up to 15 km long, to other intermontane basins of the Apennines where much stronger earthquakes have occurred. A typical example is the Fucino basin, located in the Abruzzo region, ca. 130 km to the southeast, which hosted the Avezzano earthquake (magnitude ca. 7) in 1915. Here the geological slip-rate since the Middle Pleistocene is at least 0.4 mm/y, high enough to drown the secondary structures within the basin, and the length and stratigraphic offsets of the master faults (Michetti *et al.*, 1996) are sensibly larger than those in the Colfiorito basin. These differences can derive from the younger inception of the Colfiorito basin and its much lower slip rate. Another interesting fact is that the hypocentral depths of the Colfiorito events appear significantly shallower than that of the major seismic events of this century in Italy. Does this necessarily imply a much lower seismic potential, as also suggested by the historical seismicity? How thick is the seismogenic layer in the area and how this affects the fault segmentation? What is the potential of nearby basins, e.g. Norcia and Gubbio, which display characteristics somehow in the middle between Colfiorito and Fucino?

To answer these main questions for SHA, we need to determine, as precisely as allowed by geological, geomorphological and geophysical methods, a number of parameters on geometry and recent history of all capable faults in the area (Cello *et al.*, 1997; Michetti *et al.*, 1997; Tondi *et al.*, 1997; Vittori *et al.*, 1997; Azzaro *et al.*, 1998). At this end, we emphasize the need for paleoseismological analyses in the area and along other tectonic structures in the Apennines, showing similar historical seismicity levels and evidence of capability.

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Colfiorito - Rieti, Valle del Salto. Holocene surface faulting along the Fiamignano Fault and associated large scale slope deformation

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After more than ten years of research conducted by the Authors and others research groups (e.g. Bertini & Bosi, 1976; Raffy, 1983; Barberi & Cavinato, 1993; Michetti *et al.*, 1995, and references here in), a significant geological database is available for the region of Central Italy centered on the Rieti basin. To understand the pattern of recent gravitational deformations and capable earthquake faulting along the Fiamignano Fault, it is very important to frame the local setting into the Quaternary evolution of the surrounding areas. Figures 1 to 4 show the example of the changes in the drainage pattern of the Rieti basin and nearby areas in Central Italy over the Quaternary. These maps are based on mostly 1:10,000 scale field survey of the Quaternary deposits, interpretation of airphoto coverages with different scales, geomorphic analyses, paleoseismic and stratigraphic trench investigations soil surveys, geophysical prospecting, radiocarbon and Ar/Ar dating, and detailed analyses of several ad hoc drilled boreholes. The most significant result from these investigations is the understanding of the relative importance of tectonic and climatic processes in controlling the landscape evolution. In tectonically active provinces like Central Italy, only once typical earthquake ground effects, style of faulting, fault rupture length, displacement per event and slip-rates have been reasonably well constrained, it is possible to reconstruct the local geomorphic evolution. The growth of the Valle del Salto, Rieti, Terni and Leonessa intermountain basins and footwall mountain ranges is here related with the activity and interaction of the dominant normal faults in the area, which strongly influenced the drainage network (e.g., Michetti *et al.*, 1995; Mazzi, 1996; Guerrieri *et al.*, submitted; Brunamonte *et al.*, in prep.).

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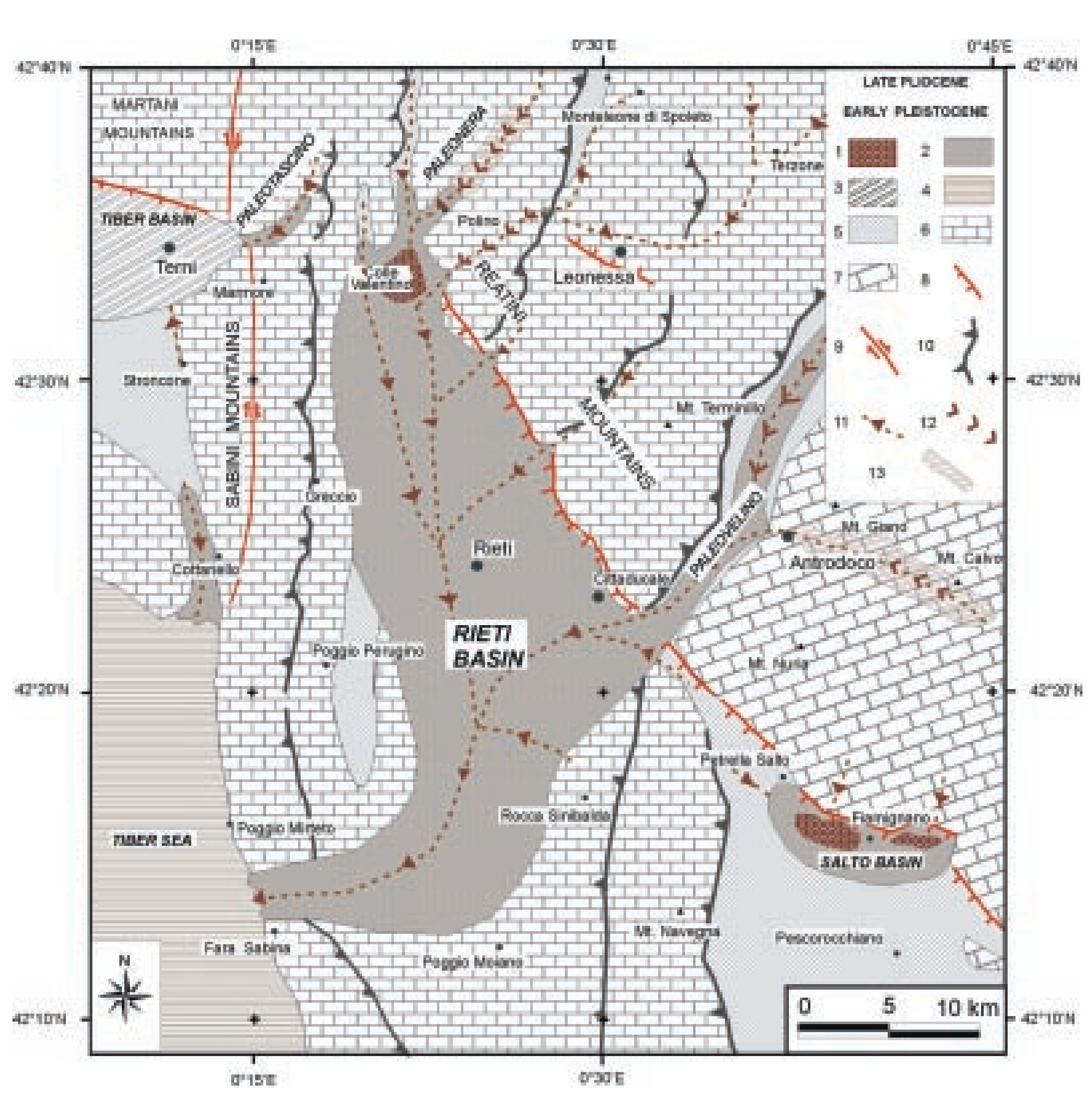

Fig. 1. Late Pliocene – Early Pleistocene: 1. Continental deposits of Rieti and Salto basins still preserved as remnants; 2. Hypotetical distribution of not preserved continental deposits; 3. Lacustrine deposits (Ponte Naja Unit, Tiberin Basin); 4. Tiber Sea sequence;
Bedrock: 5. Flysch deposits (Miocene); 6. Limestones and marls of Umbrian sequence (Jurassic-Miocene); 7. Limestones of Latium and Abruzzi sequence (Triassic-Miocene);
Structural elements: 8. Normal fault, active in Late Pliocene–Early Pleistocene time; 9. Strike-slip fault, active in Late Pliocene – Early Pleistocene; 10. Thrust;
Geomorphological features: 11. Rivers and streams: the triangle indicates the draining direction; 12. Wind-gap; 13. Erosional paleosurface.

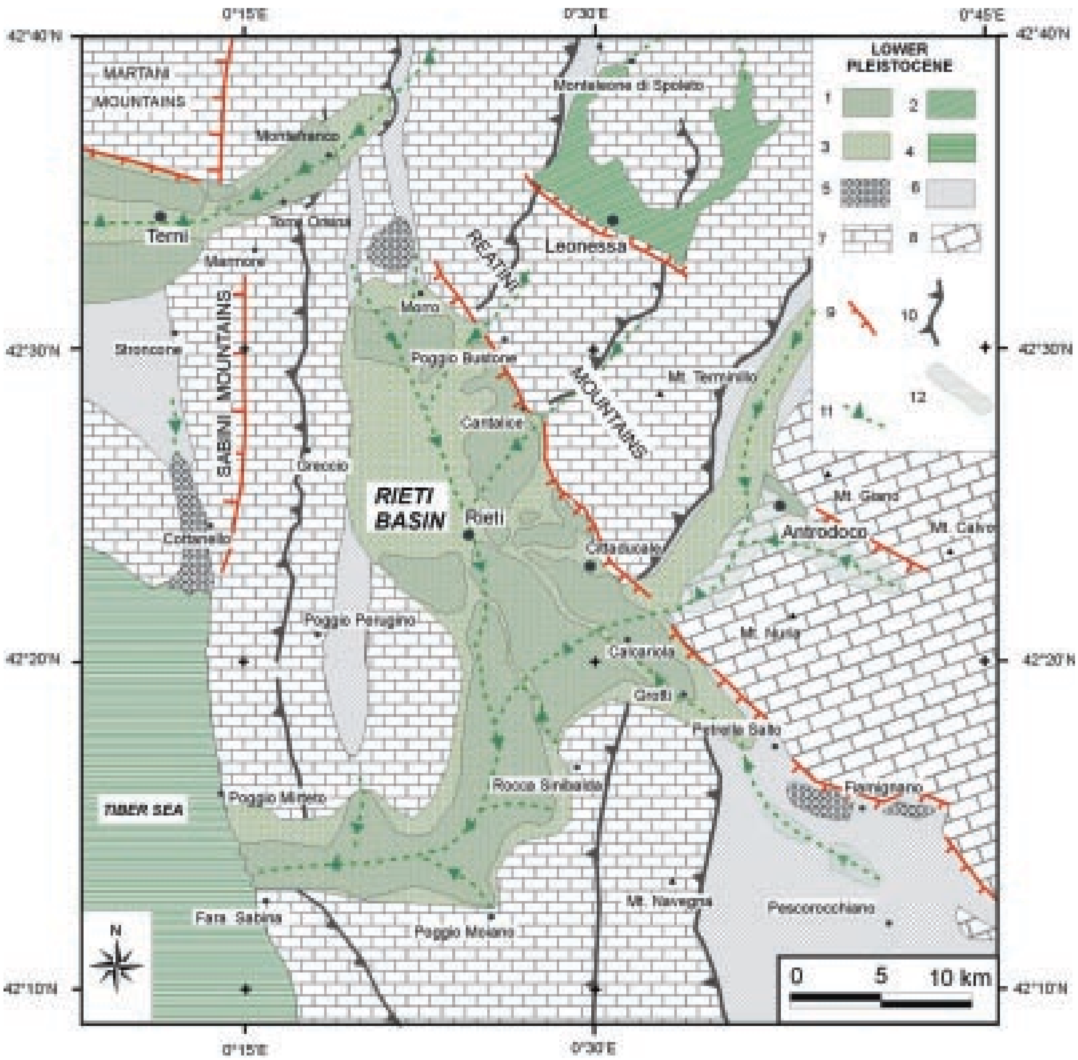

Fig. 2. Lower Pleistocene: 1. Fluvial deposits of Rieti and Terni basins still preserved as remnants; 2. Lacustrine and marsh deposits of Leonessa basin; 3. Hypotetical distribution of not preserved continental deposits; 4. Tiber Sea sequence.
Bedrock: 5. Older continental deposits (Late Pliocene – Early Pleistocene); 6. Flysch deposits (Miocene); 7. Limestones and marls of Umbrian sequence (Jurassic-Miocene); 8. Limestones of Latium and Abruzzi sequence (Triassic-Miocene).
Structural elements: 9. Normal fault, active in Lower Pleistocene time; 10. Thrust;
Geomorphological features: 11. Rivers and streams: the triangle indicates the draining direction; 12. Erosional paleosurface.

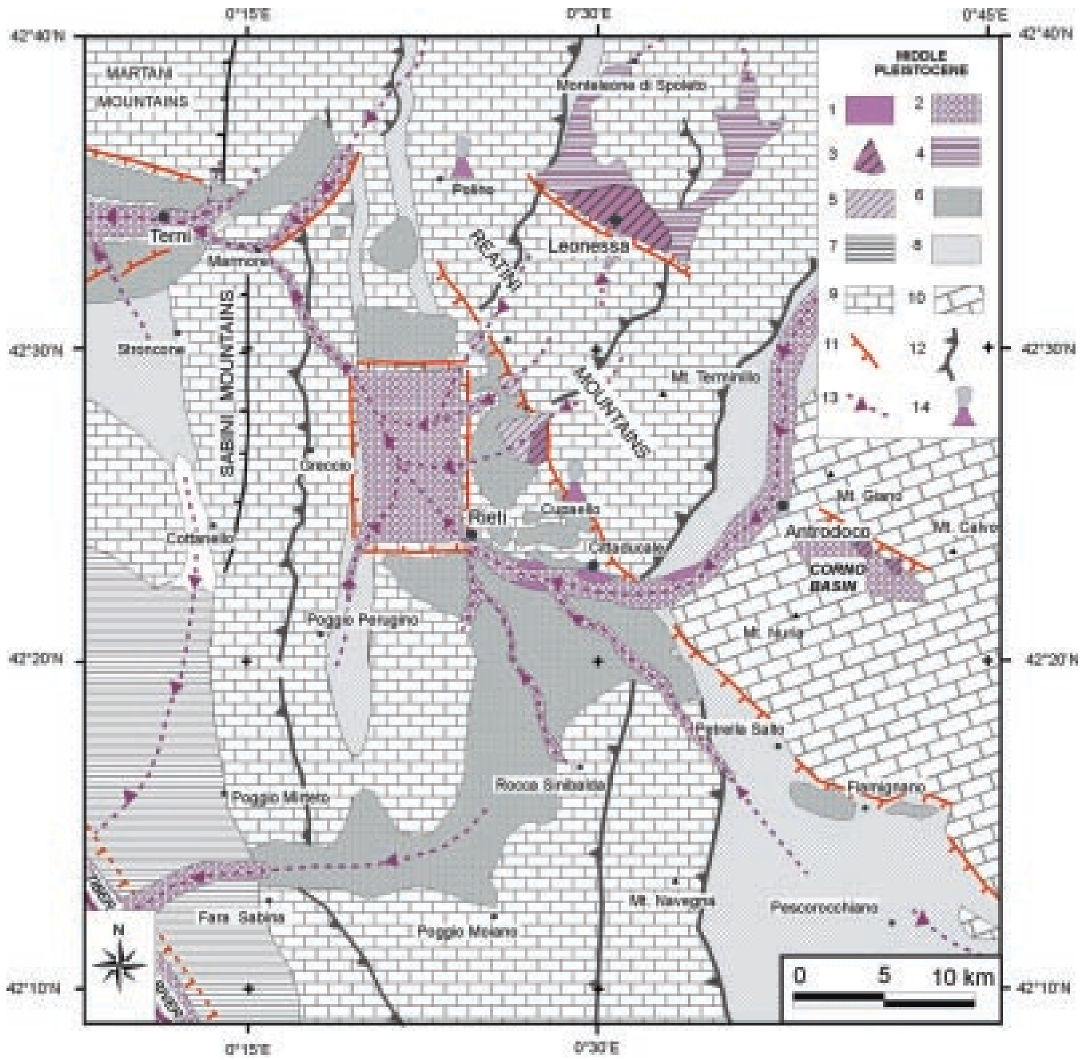

Fig. 3. Middle Pleistocene: 1.Travertines with related fluvial and lacustrine deposits; 2. Hypotetical distribution of not preserved fluvial deposits; 3. Alluvial fan deposits; 4. Lacustrine and marsh deposits of Leonessa basin; 5. Hypotetical distribution of not preserved alluvial fan deposits;
Bedrock: 6. Older continental deposits (Late Pliocene–Early Pleistocene); 7 Sands and clays (Tiberin Sea sequence, Pliocene–Early Pleistocene); 8. Flysch deposits (Miocene); 9. Limestones and marls of Umbrian sequence (Jurassic-Miocene) 10. Limestones of Latium and Abruzzi sequence (Triassic-Miocene);
Structural elements: 11. Normal fault, active in Middle Pleistocene time; 12.Thrust;
Geomorphological features: 13. Rivers and streams: the triangle indicates the draining direction; 14. Local volcanic eruption.

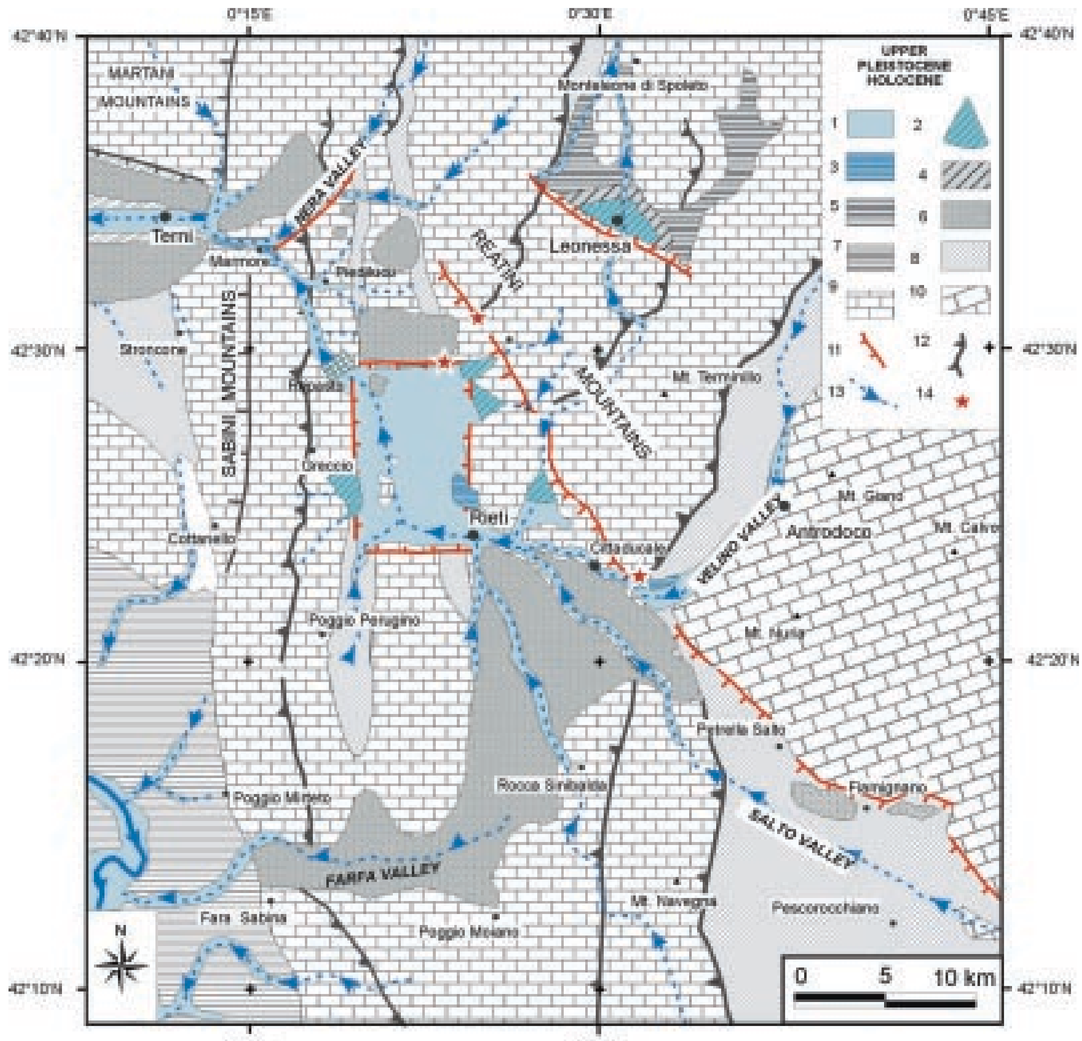

Fig. 4. Upper Pleistocene – Holocene: 1. Fluvial and lacustrine deposits; 2. Alluvial fan deposits; 3. Travertines with related fluvial and lacustrine deposits;
Bedrock: 4. Older alluvial fan deposits (Middle Pleistocene); 5. Lacustrine and marsh deposits of the Leonessa basin (Lower-Middle Pleistocene); 6. Older fluvial deposits (Late Pliocene-Lower Pleistocene); 7. Sands and clays (Tiber Sea sequence, Pliocene–Early Pleistocene); 8. Flysch deposits (Miocene); 9. Limestones and marls of Umbrian sequence (Jurassic-Miocene); 10. Limestones of Latium and Abruzzi sequence (Triassic-Miocene);
Structural elements: 11. Normal fault, active in Upper Pleistocene to Holocene time; 12. Thrust; 13. Rivers and streams: the triangle indicates the drainage direction; 14. Late Quaternary evidence of surface faulting and site of exploratory trenching for paleoseismological analyses.

However, in the long term the river drainage system have been able to overcome the tectonic control, due to average erosional and depositional rates much higher than the average slip-rates of the local normal faults. In mid-Pleistocene times, after 2 to 3 Ma since the inception of crustal extension in this sector of the Apennines, the fluvial network developed in one single, large catchment basin (the Nera River catchment basin; Fig. 2-4). Obviously, this history of the landscape depends on the peculiar features of the Central Apennines, in terms of Quaternary climatic conditions, rates of tectonic activity and paleoseismological relations. As illustrated in Figures 1 to 4, the Salto River Valley have been an independent, normal-fault controlled, sedimentary basin only in late Pliocene to early Pleistocene times. Subsequently, the drainage network evolution led to the opening of a direct connection with the Rieti basin. The geomorphic setting of the area is shown in Figure 5. The present slope height, strictly connected to the climatically controlled talweg elevation (Fig. 6; see Carrara et al., 1993; Calderoni et al., 1995; Calderini et al., 1998), promotes surficial instability mainly related to the lithological and structural features of the slopes and triggered by intense rainfall events or by earthquakes, as suggested also by historical documents.

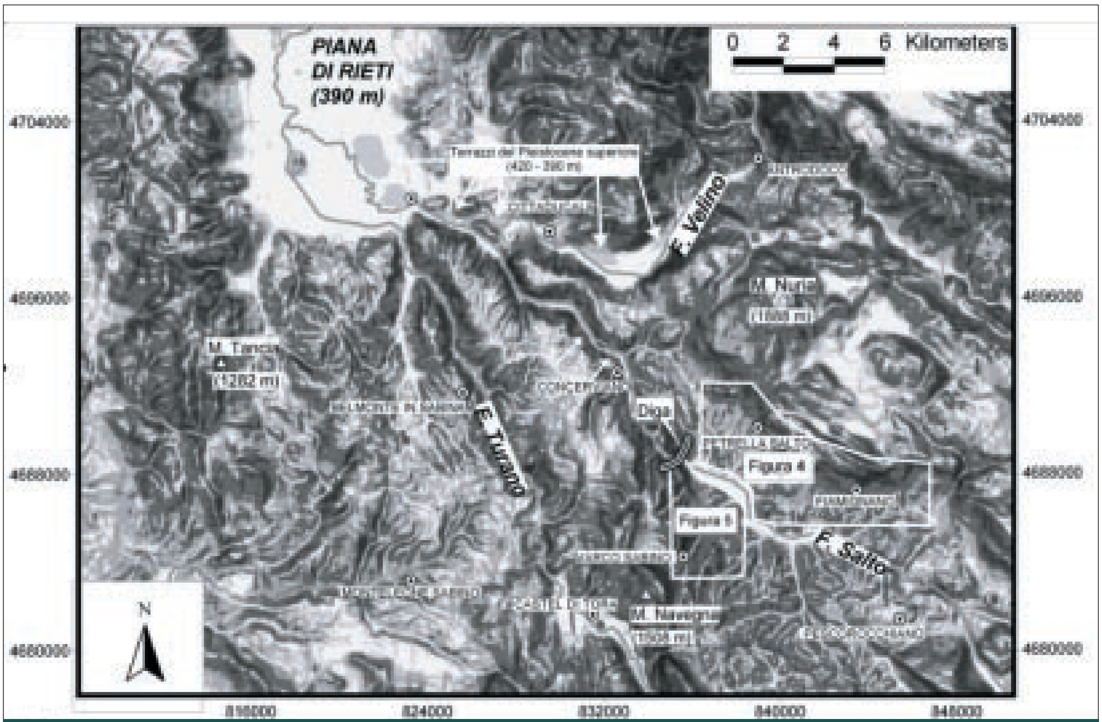

Fig. 5. Location of the studied area and relations with the Rieti area

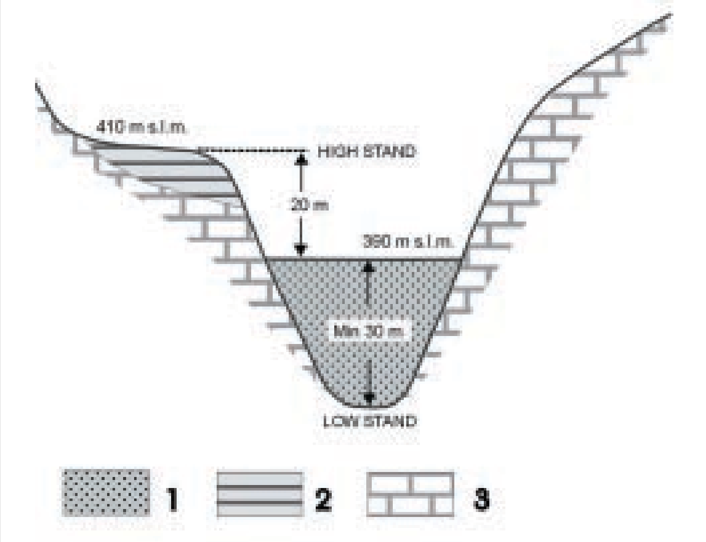

Fig. 6. Schematic section in the downstream sector of the Salto River dam showing the variation of the valley floor elevation due to recent climatic changes, including the Last Glacial Maximum Low Stand, due to erosion of the Marmore travertine threshold. Legend: 1) Recent alluvial deposits (Holocene); 2) Fluvial terrace (Upper Pleistocene); 3) Bedrock.

However in the long-term landscape evolution, a primary role was played by deep seated gravitational movements which sign the north western sector (Fiamignano and Petrella Salto) probably due to a greater slope height and also to the presence of a capable normal fault. Evidence for this will be reviewed in the field in the area between Petrella Salto and Fiamignano (Figs. 7 to 10).

Gravitational collapses, carved only in the pre-Holocene landscape (Colle della Sponga and S.Vittoria paleolandslides; Fig. 9) and not recorded in the historical data, is probably related to higher and steeper slopes during low-stand climatic conditions.

A sequence of flooding events is well known in the Rieti basin, downstream of the studied area, constrained in the Holocene stratigraphy and historical documents. These processes were probably influenced by anthropic deforestation since the Middle Age, and today are controlled by a dam.

The recent activity of the Fiamignano Fault is well documented at this site (Bosi, 1975; Mariotti and Capotorti, 1988). A ca. 8 km long, continuous, Holocene bedrock fault scarp characterizes the base of the mountain front between Petrella Salto and Brusciano (Fig. 10A). Here, a sequence of young, post-glacial small drainages carved in the bedrock are systematically offset across the fault scarp (Fig. 10B). The recent activity of this fault is also documented by the analysis of Late Glacial to Holocene slope deposits. At the Poggio Poponesco site, stratified Late Glacial debris are backtiled against the fault (Fig. 10C). Overlaying Holocene colluvial debris, including a soil horizon dated 4440 + 140 yr BP, are also offset by the fault (Fig. 10D). These observations strongly suggest that earthquake surface faulting along the Fiamignano Fault may influence the deep-seated gravity deformations of the NE slope of the Salto River Valley.

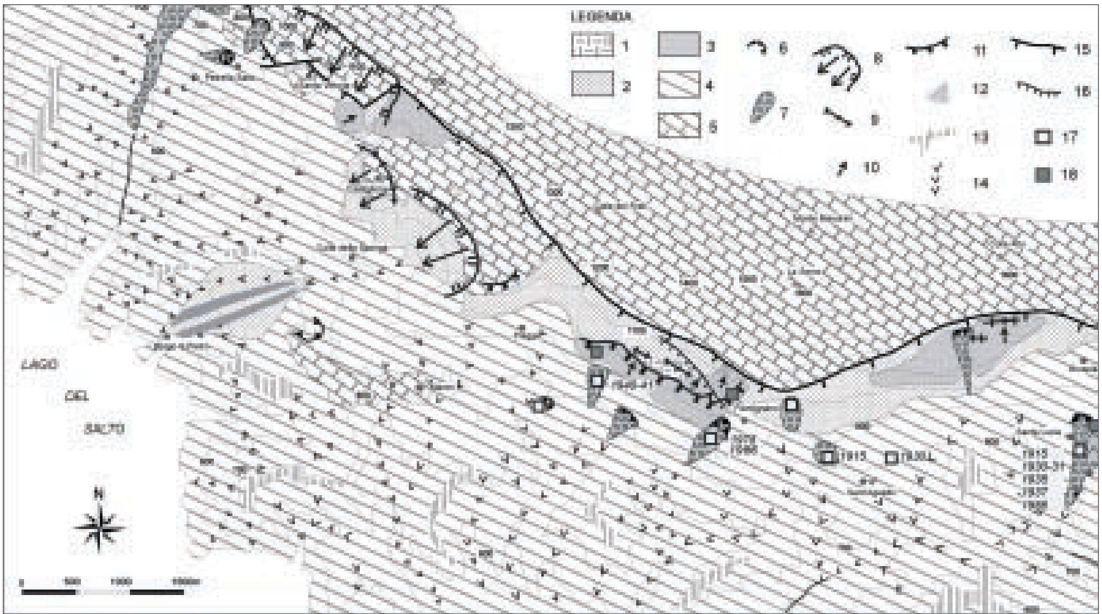

Fig. 7. Geological and geomorphological sketch in the area between the Petrella Salto and Fiamignano villages and the Salto Lake. Legend: 1) Eluvial and colluvial deposits; 2) Recent talus; 3) Old cemented breccias; 4) Clays, sands and sandstones; 5) Limestones and calcarenites; 6) Landslide main scarp; 7) Landslide body; 8) Slope affected by deep seated gravitational movements; 9) Trench; 10) Sector of a slope inclined in the opposite direction; 11) Rock fall scarp; 12) Alluvial fan; 13) Erosional surface; 14) Erosional channel; 15) Staffoli-Fiamignano fault; 16) Fault scarp; 17) Sliding historical events; 18) Rock fall historical events.

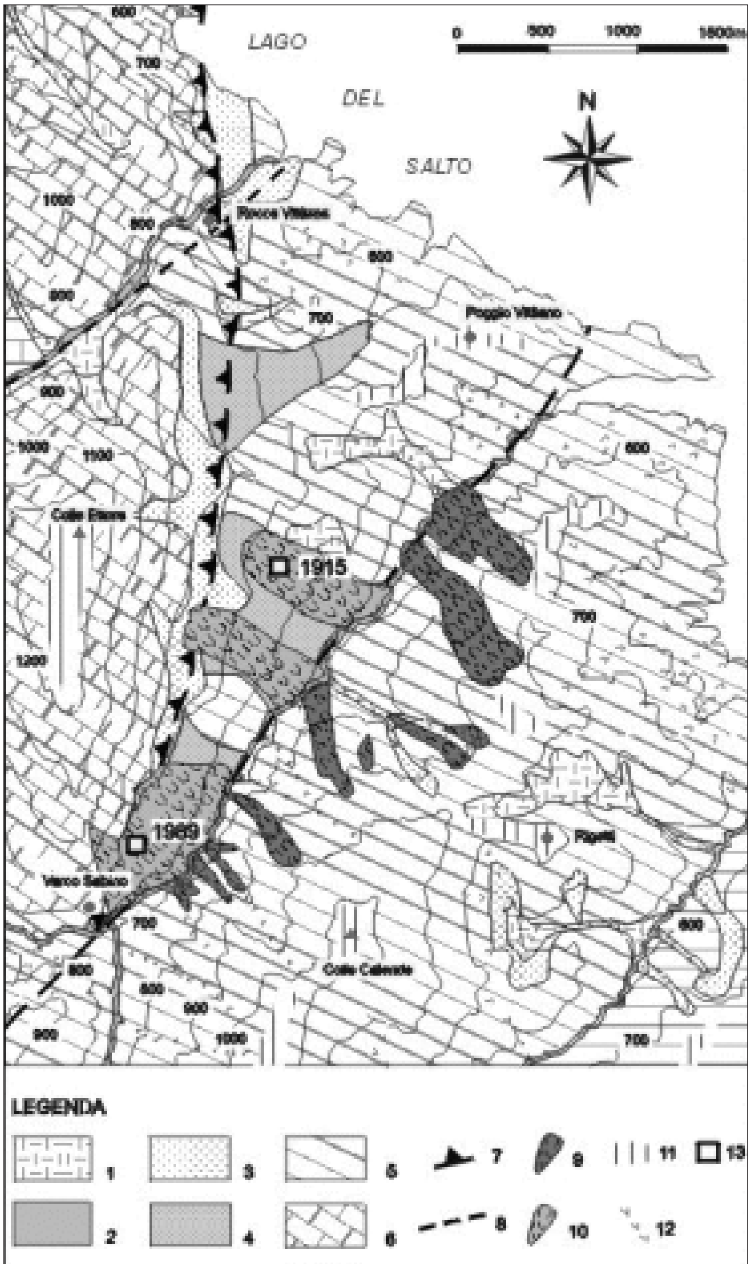

Fig. 8. Geological and geomorphological sketch in the area between the Varco Sabino village and the Salto Lake. Legend: 1) Eluvial and colluvial deposits; 2) Recent alluvial deposits; 3) Recent talus; 4) Old cemented talus; 5) Clays and sandstones; 6) Limestones and calcarenites; 7) Olevano-Antrodoco thrust line; 8) Normal fault; 9) Earth flow; 10) Complex landslide; 11) Paleosurface; 12) Erosional channel; 13) Historical landslide event.

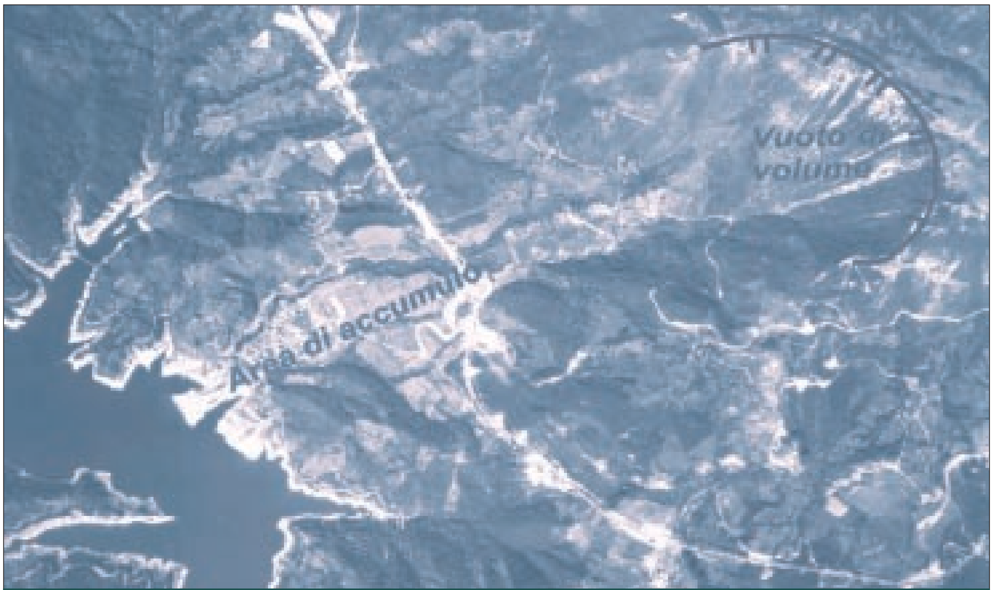

Fig. 9. Aerial photo showing the ancient landslide affecting the slope up to Colle della Sponga village. The landslide accumulation body is still preserved near Borgo San Pietro village, shaped by areal earth surface processes.

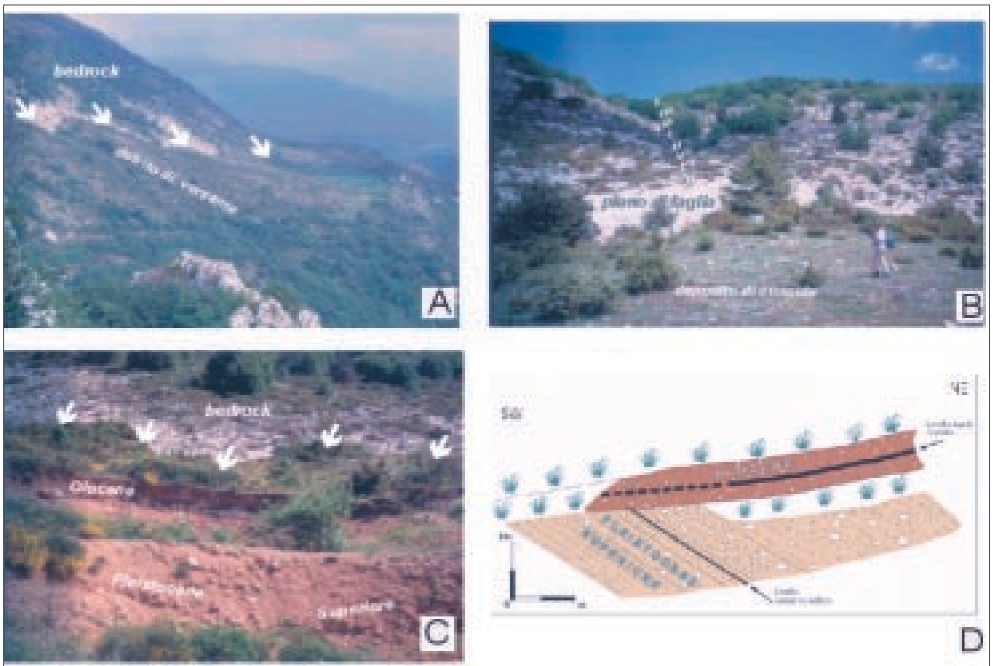

Fig. 10. Tectonic activity along the Staffoli-Fiamignano fault A) Panoramic view of Poggio Poponesco area: the outcropping rock fault scarp is evident; B) Detail of the previous picture showing the same fault displacing an Upper Pleistocene erosional channel; C) Slope deposits sequence studied in detail; D) Upper Pleistocene slope deposits are tilted and at present dip in the opposite direction. Also Holocene colluvial deposits are displaced by the Staffoli-Fiamignano fault.

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4. Gravitational phenomena triggered by the 1980 Southern Italy Earthquake

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Earthquake-triggered landslides in Italy

Several historic records and oral traditions exist of very large gravitational movements triggered by earthquakes in Italy (see, for example, Oddone, 1930; Cotecchia *et al.*, 1969; Govi, 1977; Dramis *et al*, 1982; Crescenti *et al.*, 1984).
In recent times it was possible to survey directly, with more scientific methods, surface effects of strong earthquakes (Panizza *et al.*, 1987), also comparing them with instrument records of the shock. In this way it was possible to understand (Radbruck-Hall and Varnes, 1976; Keefer, 1984) that the typology and dimension of triggered mass movements are strictly related to both litho-structural features of the site and characteristics of the shock, particularly with Arias intensity (Arias, 1970). It has also been outlined that many earthquake-induced mass movements are also connected with other seismic ground effects (such as fracturing and faulting).
As far as lithologic features are considered, the importance of identifying "engineering geological formations" has to be stressed (Cotecchia, 1978; Canuti *et al.*, 1988); research aiming to this end is presently being carried out throughout the Italian territory.
It has also been pointed out that earthquake-triggered mass movements may involve slopes already characterized by instability, but normally dormant or evolving at a very slow rate. Even though earthquakes can trigger phenomena of any kind and dimension (ranging from very small and shallow ground failures to huge landslides and deep-seated gravitational movements), a typical feature of earthquake related landslides (i.e. of phenomena which generally reactivate only as a consequence of strong seismic shocks) is their wide extension and elevated depth. These kind of mass movements, being activated only by extreme events (mainly strong earthquakes and, subordinately, intense rainfalls), typically show recurrent activity, alternating long steady periods with sudden reactivations.
Among earthquake-induced surface effects, lateral spreadings, causing progressive "graben like" sinking on hill tops, are reported (Solonenko, 1977; Dramis *et al.*, 1983).
Very important for the activation of landslides (and, of course, of earthquake-induced ones too) are also hydrogeological conditions (such as saturation of terrain, variations of piezometric level etc.). Particularly frequent on saturated sandy-silty sediments are liquefaction phenomena which can produce instability either directly (because of flow slides along saturated sandy-silty slopes) or indirectly (by allowing the mobilization of overlying terrains). This kind of landslides (Tinsley *et al.*, 1985) often involve deep-seated beds too, disturbing very large areas far away from the epicenter.

The southern Italy earthquake

The 1980 Earthquake in southern Italy The last highly energetic earthquake that affected the Italian peninsula happened in November 1980 (Deschamps and King, 1983; Westaway and Jackson, 1987) and involved a wide portion of the Campania and Basilicata regions, in southern Italy, even though it was perceived (III MCS) in a large part of the Italian territory. The main shock reached a magnitude of 6.8-6.9R and produced widespread and severe damage (for this event too the intensity was estimated around IX-X MCS), also causing about 4,000 casualties and many injuries.

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The source was a dip-slip normal fault with Apennine (NW-SE) strike and vergence towards the Tyrrhenian Sea (i.e. SW) and the focus was some 18 Km deep. The shock was, in many aspects, similar to those affecting more or less the same area in 1930 and 1962 (the latter reached intensity XMCS).

Coseismic surface faulting (Fig. 1) was recorded in the area (Cinque *et al.*, 1981; Bollettinari and Panizza, 1981; Pantosti and Valensise, 1990) as well as vertical movements (Cotecchia, 1982; Arca and Marchioni, 1983). More widespread were surface fractures (Fig. 2), mostly rectilinear and up to several kilometers long, that were created in a variety of lithologies (including loose material, such as alluvial deposits and inactive landslide bodies), both isolated and in parallel or joined groups, as a consequence of interference between seismic, tectonic and gravitational stress (Carmignani *et al.*, 1981; Dramis *et al.*, 1982). Many of these discontinuities opened as a reactivation of fractures created by past earthquakes (e.g. in 1930 and in 1962), as reported by several authors (Alfano, 1930; Oddone, 1930; Vari, 1930; Serva, 1981). Their opening frequently lead (both in 1980 and during past earthquakes) to gas emission (and, sometimes, ignition). Along some of them a sharp increase in Helium content of the soil has been recorded (Dramis *et al.*, 1982) that may indicate their depth (or at least their connection with deep-seated beds).


Fig. 1. 1980 Southern Italy Surface fault and historical eartquakes

However, the most outstanding phenomena triggered by the earthquake were mass movements of different type (Cantalamessa et al., 1981; Cherubini et al., 1981; Cotecchia, 1981, 1982; Genevois and Prestininzi, 1981; Agnesi et al., 1983; Crescenti et al., 1984; Bisci and Dramis, 1993), at least partially determined by the quite high relief of the area, the poor geotechnical characteristics of most of the outcropping rock and the high water content of terrains (due to heavy rainfall in the days preceding the seismic event). These movements often appear to be connected with the above described ground fractures too and quite frequently affected entire slopes. These gravitational phenomena mainly moved immediately after the earthquake and their activity lasted only for a short period; most of them represent the reactivation of landslides activated by past earthquakes.

Calcareous formations were locally mobilized, quite close to the epicenter (such as at Castelgrande, Nusco, Valva, Bella-Muro Lucano, Balvano-San Gregorio Magno etc.). More frequent and widespread were mass movements on Tertiary flysch (such as at Laurenzana, Sant'Angelo le Fratte, Teora, Oliveto Lucano etc.) and on Pliocene-Quaternary deposits (such as at Bisaccia, Avigliano, Tricarico, Accettura, Balvano, Lioni, etc.).

Particularly frequent among seismically-induced landslides were huge block slide phenomena that are generally bounded by ground fractures (Fig. 3) which in many cases continue also beyond the sliding body, sometimes reaching the opposite slope (Dramis *et al.*, 1982). These phenomena, which generally experienced a sudden activation and show a long body (up to 2-3 Km), often being associated with other gravitational phenomena (most of an earth flows

Fig. 2. Earthquake induced ground fracture cutting a house

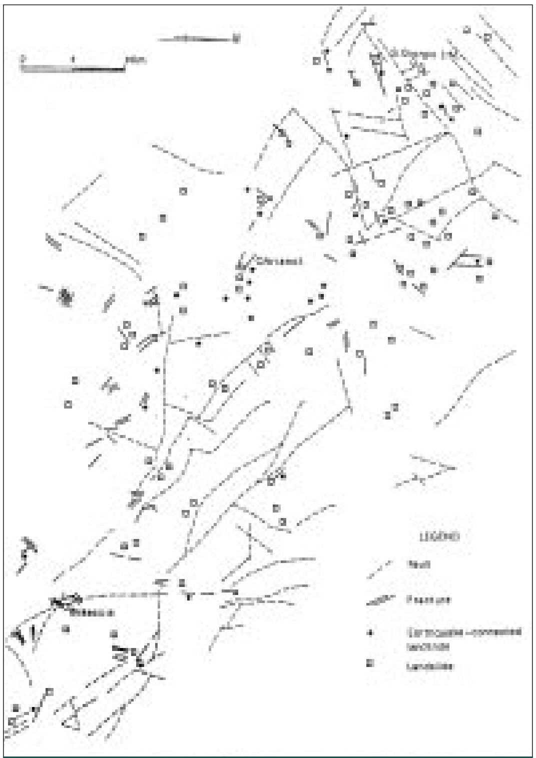

Fig. 3. Pattern of fractures and landslides induced by the 1980 Southern Italy earthquake between S. Giorgio la Molara and Bisaccia

and subordinately, mud flows and rotational slides). Block slides whose movement direction showed significant differences from the slope gradient were also recognized. Lateral spreads and deep-seated slope deformations were diffusely recorded too, as well as translational slides, earth flows and topples.

Cases of huge landslides induced by the 1980 earthquake in southern Italy

San Giorgio la Molara

Among the several landslides caused by the shock, particularly significant looks to be the huge phenomenon that occurred near the town of San Giorgio la Molara (not very close to the epicenter), that moved a few minutes after the earthquake, causing displacements of some tens of meters.
It was a more or less a global reactivation of a some 3 Km long dormant movement interesting pelitic-arenaceous terrains that also in the past (as in 1688, 1805 and 1930) reactivated only as a consequence of strong earthquakes (Dramis *et al.*, 1982).
In the neighboring area, along the Tammaro River (Fig. 4), several other phenomena of slope instability and deformation were recorded (Genevois and Prestininzi, 1982) on Miocene sediments (gray and variegated marls and argillites with calcarenitic beds), among which a big block slide (whose upper portion moved as a huge earth flow) and several slump-earth flows (in the upper part of the slopes), translational and rotational slides and earth flows (in the lower part). In the same area, many fractures (among which one can be followed for some 4 Km) and counterslope ponds were also recognized.

Fig. 4. Tammaro river bridge deformed by earthquake induced landsliding.

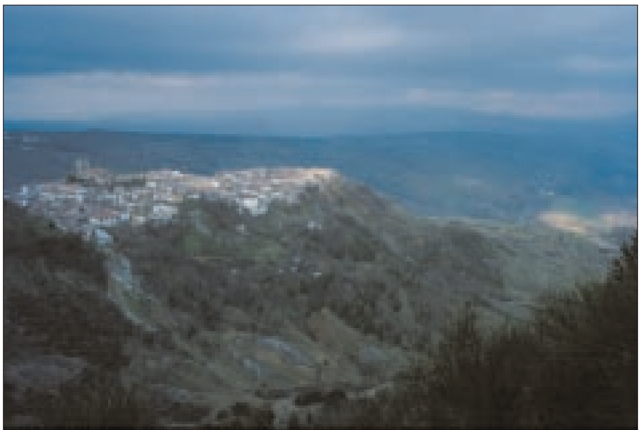

Fig. 5. View of Bisaccia cut by a gravitational scarp

Bisaccia

Another important mass movements was the multiple deep-seated sliding that involved most of the town of Bisaccia (Crescenti *et al.*, 1984), located quite far away from the epicenter (the earthquake here reached only a VII MCS intensity) and built up over a clastic formation of a Quaternary age (polygenic conglomerate with abundant sandy-clayey matrix) overlying strongly disturbed allochtonous clays (Miocene "Varicolori" clays) and divided by a some 40 m high escarpment (Figs. 5-6). Along the slopes, modeled in the clayey deposits, counterslopes and depressions, often fined by alluvial and colluvial material, are frequent. This complex phenomenon was also the reactivation of a dormant one that in the past had shown recurrent activations in correspondence of major earthquakes. Also most of the ground fractures recognized along the sliding body (including buildings) experienced in the past recurrent seismic reactivation, as testified by a detailed mapping of surface effects carried out by the municipality immediately after previous events (1930 and 1962).
Damage to buildings was not extreme because of the typology of the movement (many artefacts were simply rotated together with soil wedges but not destroyed). It has to be pointed out that in Italy several other examples exist of villages and towns built .over the body of dormant landslides that in the past have been shown to reactivate during earthquakes.

Trevico

Quite different is the mass movement recognized in Trevico (Carton *et al.*, 1987), typical of places where solid bedrock (such as limestone, sandstones etc.) outcrops and they are characterized by a high value of relief (mountainous areas). It is a deep-seated lateral spreading (Figs. 7-8-9) affecting conglomerates overlying clays and consisting in the deepening (up to some decimeters) of a small graben-like depression (some 50 m long, about 15 m wide ad 2 m deep) located on the top of an hill. Surficial ruptures look to be related to deep-seated shearing surfaces and the trench formed because the ridges spread aside; it was not directly generated by tectonic phenomena but is of a gravitational origin (Dramis and Sorriso-Valvo, 1983.
As also hypothesized for other similar phenomena, it is not probable that the movement could be due to oriented acceleration only; it seems more likely that other mechanisms (probably

connected with the presence of water) contributed to the genesis of this deep-seated gravitational deformation. Also this movement in the past experienced recurrent reactivation in connection with seismic activity. graben-like features like this one, and having the same step-like evolution, have also been observed elsewhere, such as in Algeria after the El Asnam earthquake (7.3 Ms) in 1980 (Dramis and Sorriso-Valvo, 1983).

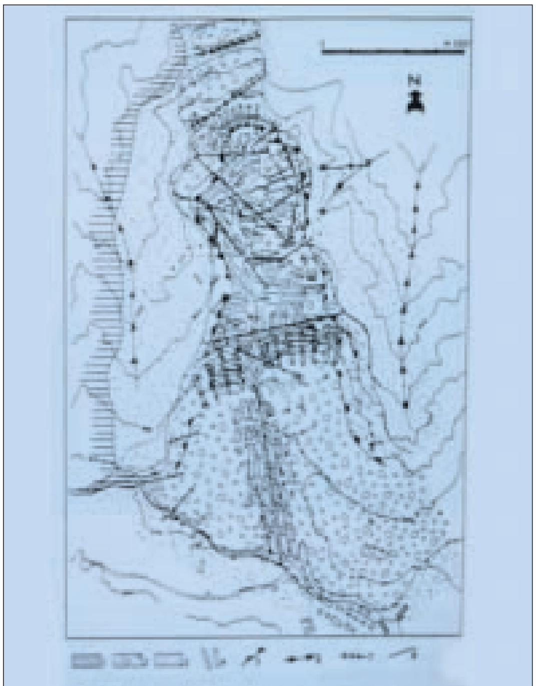

Fig. 6. Simplified geomorphic map of the Bisaccia area:
1 - "Varicolori" clays;
2 - conglomerates;
3 - debris;
4 - main landslide escarpment;
5 - erosional scarps retreating by mass movement;
6 - stream erosion;
7 - trench;
8 - fault or fracture (after Crescenti et al. 1984).

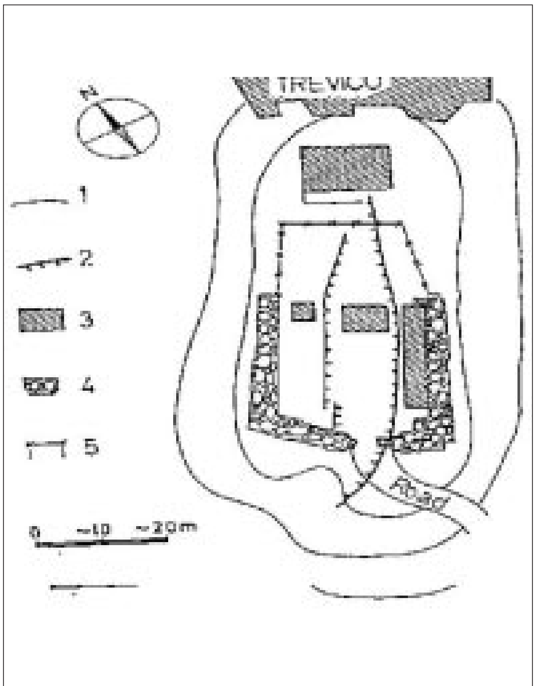

Fig. 7. Sketch of the graben–like depression:
1 - approximate contour line;
2 - trench scarp;
3 - building;
4 - castle ruins;
5 - fence (alter Dramis and Sorriso- Valvo, 1983).

Fig. 8. A ground fracture cutting the stairs of the meteo station at Trevico

Fig. 9. Scarplet along the border of the graben-like depression within the Trevico earthquake reactivated lateral-spreading

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Slope movements triggered by the 1980 Irpinia Earthquake

Mario Parise[1] & Janusz Wasowski[1]

Slope instability phenomena have long been a constant feature in the Southern Apennines of Italy, with earthquakes and meteoric events representing the main triggering factors. Landslides often have resulted in dreadful losses both in economic terms and in terms of human lives. In particular, history of Irpinia has always been accompanied by catastrophic seismic events that therefore played a significant role in contributing to determine the low socio-economical development of the region, ehich is essentially based on rural economy. The occurrence of these earthquakes, affecting an environment which was already characterised by widespread landsliding due to local geology and tectonic history, exhacerbated the predisposition of the territory to slope movements activity.
The outcropping lithologies in Irpinia can be grouped as follows:

1) Alluvial deposits (Pleistocene - Holocene). They are present along the courses of the Sele and Ofanto Rivers and their main tributaries.
2) Conglomerates, sands and grey-blue clays of Lower-Middle Pliocene. These rocks, representing regressive successions, crop out on both sides of the Ofanto River, and form the hills on which several towns are located.
3) Shales, marls, chert limestones, sandstones, varicoloured clays, and clayey-marly-arenaceous flysch of the Molise - Lucanian basin (Upper Cretaceous - Paleocene). The flysch successions constitute the majority of the outcropping rocks, and show prevailing clayey lithofacies. These materials, with marked heterogeneity and anisotropy induced by a long geological and tectonic history, are defined in the current geotechnical literature as "structurally complex formations" (Esu, 1977). In relation to their peculiar mechanical and geotechnical behaviour, they are very susceptible to macroscopic slope deformation and to mass movement in general.
4) Limestones, dolomitic limestones and dolomites (Trias - Cretaceous). Carbonate terrains form the Picentini Mountains and the Mt. Ogna - Mt. Marzano ridge.

Morphology of this portion of the Southern Apennines is strongly dependent on structural and neotectonic evolution of the study area: two different landscapes are recognizable, one typical of the carbonate domains, the other of the terrigenous materials. A carbonate landscape, with high energy relief, occurs along the eastern and western margins of the Sele valley, characterized by steep and sometimes subvertical slopes; it has a hydrographic network whose rectangular pattern testifies strong structural control. Water courses develop following the network of faults and fractures, and they are often entrenched as a consequence of the rapid Plio-Quaternary orogenic uplift, which, in the Picentini area, is on the order of several hundred meters (Capaldi *and others*, 1988). The areas where carbonate materials crop out are only marginally influenced by mass movements; these include mostly rockfalls and topples limited to the steep slopes bordering the calcareous massifs (Fig. 1).
The landscape carved in the terrigenous deposits is characterized by low-medium acclivity slopes, and by widespread creep and landslide phenomena. The hydrographic network shows a dendritic pattern only partly controlled by tectonic discontinuities. In this landscape it is possible to identify several orders of terraces and remnants of ancient erosional surfaces formed as a consequence of different phases of rivers downcutting.
Mass movements represent the main geomorphic process active in the area. In fact, landslides varying in age, dimensions and state of activity, are present on most slopes. Complex landslides, starting as rotational or translational slides in the upper portions and evolving to flow in the medium-lower sectors, are widespread. Main scarps of landslides are often located in

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the proximity of the contact between the flysch and carbonate deposits; the presence of multiple coalescing crowns testifies to the occurrence of several episodes of movement. The larger landslides are characterized by elongated bodies, which reach the base of the slopes; gravitational phenomena of small dimensions, but with more evident activity, are often superimposed on these, or develop at their margins.
The 23 November 1980 earthquake triggered numerous slope movements in the flysch terranes cropping out in the upper valley of the Sele River, and in the Ofanto River as well. The majority of these phenomena were reactivations of ancient or dormant mass movements; the spatial distribution of slope failures, which are mainly of the slide-flow type, appears to be strongly dependent on factors such as the presence of old landslide masses, the structural and hydrogeological setting, and the direction along which the horizontal components of the seismic ground motion were greatest.

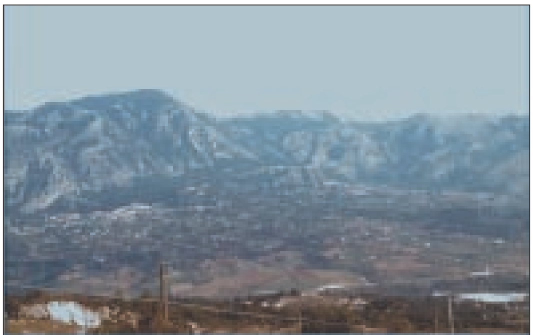

Fig. 1. Mt. Valva – Mt. Marzano – Mt. Ogna ridge, bounding the eastern side of the upper valley of the Sele River. From left to right, the towns of Valva, Collianello (at the top of the prominent rocky spur), and Colliano are visible. One of the rockfalls triggered by the 1980 earthquake is shown by the white scar and path in the carbonate slope uphill from Collianello. Note the very different landscape between the carbonate mountains and the valley carved in flyschioid (mostly clayey) materials.

The Calitri Landslide

The Calitri hill (Fig. 2) is characterized by a Pliocene regressive succession including clays and silty grey-blue clays, soft sandstones, sands and conglomerates. In the middle-lower portion of the hilly relief as well as in some areas of the town scaly clays outcrop. They consist of a chaotical sequence of dark grey clays with reddish and green patches and stone-inclusions composed of marly calcareous rocks and calcirudites (Red Flysch or Varicoloured Clays). The relationships occurring between the Varicoloured Clays and other formations are presumably of tectonic origin. However, the detection of their exact nature is by no means easy, due to the occurrence of countless landslides.
Within the described lithotypes, silty-marly clays interbedded with sandstones and Varicoloured Clays appear considerably fissured and deformed. This contrasts significantly with the relatively less disturbed structural setting of Pliocene grey-blue clays.
The area surrounding Calitri has been historically hit by landslides which are periodically reactivated by major meteoric and/or seismic events (HUTCHINSON & DEL PRETE, 1985; BUDETTA *and others*, 1990).
The widespread failures are closely related to the occurrence of clayey lithologies, poor geotechnical characteristics (chiefly Varicoloured Clays), countless tectonic discontinuities, highly-permeable materials (conglomerates and sandstones) overlying pelitic deposits (Pliocene clays and Varicoloured Clays), and markedly steep slopes. The most landslide-prone sections coincide with outcrops of Varicoloured Clays or their contact zones with the overlaying lithotypes.
Calitri is well known in the recent Italian engineering geology literature as the site of one of the largest deep-seated mass movements remobilized by the 1980 Irpinia seismic event (Fig. 2). The Calitri landslide (approximately 850 m long and up to 100 m deep), located at a distance of less than 20 km from the 1980 epicenter, has been extensively studied by HUTCHINSON & DEL PRETE (1985), and by CRESPELLANI *and others* (1996).
The reactivation of the Calitri landslide on November 23rd, 1980, destroyed or seriously dam-

Fig. 2. Calitri hill, from the Ofanto river valley. The 1980 Calitri landslide is still recognisable in the center of the photo.

aged over 100 houses and caused the death of 7 people. Within the body of the Calitri landslide, HUTCHINSON & DEL PRETE (1985) distinguished four main elements:

1) a major, deep-seated slide, with a volume on the order of 20×10^6 m^3, occupying the upper two-thirds of the valley slope, with its main scarp in the old town. It shows a part-rotational, part-translational character, and its depth has been estimated at about 100 meters;
2) associated secondary retrogressive slides around the rear scarp of the main slide;
3) shallow slides in the toe area of the main slide, which supply material in the head area of the fourth element:
4) shallow translational mudslides which form part of the colluvial apron extending down to the Ofanto River.

According to eyewitness accounts reported by HUTCHINSON & DEL PRETE (1985), cracks appeared along the main scarp about at the time of the main earthquake shock; slow downward movements of the ground then proceeded for several hour to produce the eventual, near-vertical displacements of 1 to 2 meters in that vicinity. Significant movements of the main slide and its secondary slides appear to have been essentially completed within about 24 hours.

At present, the 1980 Calitri landslide shows only local and fairly superficial movements. However, some damage to structures present within the affected area indicates much deeper deformations. Superficial phenomena are generally seasonal, and provoked by major spring-time meteoric events. Many other slope movements affect today the Calitri municipality: most of them are related to outcrops of Varicoloured Clays and to their difference in permeability with overlying materials. In many cases, active landslips are also related to anthropogenic activities on the slopes. In recent years, however, landsliding activity seems largely reduced and chiefly dependent on meteoric events (PARISE & WASOWSKI, 1998). As far as the great Calitri landslide is concerned, it is currently affected by superficial landsliding and erosional phenomena, which, given their location in the middle-lower section of the main body, might prove destabilizing in the long run.

Despite the typically slow evolution of mass movements and, therefore, corresponding low risks involved, the space distribution of failures conditions directly the town development and the construction of new road networks in the neighbouring areas.

The upper Sele river valley

The largest slope movements triggered by the 1980 Irpinia earthquake in the Sele valley are the Buoninventre landslide, near Caposele, and the Serra dell'Acquara landslide, near Senerchia.

The **"Buoninventre" landslide** (named after the main locality affected by the slide, in the municipality of Caposele) is one of the largest mass movements triggered by the 23rd November, 1980 Irpinia earthquake (BUDETTA, 1983; CARRARA *and others*, 1986; COTECCHIA *and others*, 1986). The site is located in the upper reaches of the Sele river, at a distance of less than 10 km from the epicenter of the earthquake: the area is geologically characterized by outcroppings of structurally complex formations severely affected by landsliding, whose main remobilizations appear to be related to meteoric and/or seismic events, or to anthropic actions.

The Buoninventre landslide is a complex slope movement, with length of about 3 km, made by multiple rotational and translational slides, whose deposits feed a wide flow-body. With an estimated volume of about 25 x 10^6 m^3 (COTECCHIA *and others*, 1986), the slide involves mainly old landslide materials, and, in addition, intensely folded and jointed rocks of turbidite origin (alternations of calcarenites, sandstones, marly limestones and clays). The chaotic setting of most of the outcropping lithologies, together with the high susceptibility to slope movements, derive from the complex geologic history of the area.

The availability of multi-year aerial photo coverage (1955, 1981, 1990 and 1995) in appropriate scale (ranging from about 1:30,000 to about 1:8,000) helped in assessing the morphological changes occurred in the last 40 years at Buoninventre site. Each stereopair of aerial photos was interpreted and a geomorphological map produced for each year. Hummocky topography, changes in the drainage surficial network, scarps, cracks, uphill facing areas, ponds were among the main features used to recognize and delimit areas with major degree of mass movements.

Comparison of the landslide activity maps in different years pointed out at Buoninventre site to an areal frequency of active landsliding which increased more than two and a half times following the 1980 seismic event. This increase is very similar to that registered in the Calitri area (PARISE AND WASOWSKI, 1999), which is not surprising considering that the epicentral distances of these two sites were comparably short (within 20 km).

Recent field obervations, including inclinometer measurements (PARISE & WASOWSKI, 1996), demonstrated that the main slide body at Buoninventre is stable. The earlier stabilization of the Buoninventre mass movement is most likely linked to the human activity in the area. In fact, in the mid-eighties, several large diameter drains were realized within the landslide body. In addition, the area affcted by the movement was shortly regained as an agricultural land. This meant a construction of an efficient network of simple surficial drainage works, and, importantly, their continuous maintenance.

Like in the Calitri case, the examination of the evolutionary trend of the Buoninventre landslide area reveals the sharp increase in mass movements activity related to the 1980 earthquake. Moreover, the overall landslide areal frequency (active and inactive movements) remained about constant in the years following 1980, which again indicated the persistency of the geomorphic changes caused by the seismic event.

The town of **Senerchia**, located on the west side of the Sele valley, is built on a Middle Quaternary detrital slab, lying on the boundary between the Mount Cervialto water-bearing carbonate horst and the Unità Sicilidi occupying the Sele graben. The geological and structural setting of the area, with the tectonic contact between the carbonate massif and the flysch deposits, where the latter plug the large aquifer in the limestones and dolomites, are the main factors of the widespread presence of landslides on the eastern slope of the Picentini Mountains. Significant surface faulting was observed along the Quaternary normal fault which border the mountain front (Figs.4 and 5). It is worth noting that the old village of Senerchia was located just across this normal fault.

Indeed, Senerchia was very badly hit by the 1980 earthquake. There was serious loss of life owing to the collapse of the whole southern part built on the variously faulted and dislocated detrital slab. In addition, the slab has been affected on the southern side by one of the largest mass movements reactivated by the 1980 earthquake, the Serra dell'Acquara mudslide (MAUGERI *and others*, 1982). Following the main shock on 23rd November 1980, the mudslide was gradually remobilized over a period of a couple of weeks. The subsequent shocks reactivated a 2,500 m long, and up to 500 m wide mudslide mobilizing a mass of about

28 x 10^6 m^3. The slip surface lies entirely in the pelitic-flysch sequence of Unità Sicilidi, at maximum depth of 33 m.
The phenomenon started from upstream and slowly spread downstream. In the initial phase the material slipped almost as a single mass. The main moving mass, on which local secondary movements of remoulded muddy material were superimposed, possibly came to exert pressure on the pre-existing zone of accumulation, causing an uplift of about 20 m along the line of contact. Remobilization of the zone of accumulation thus occurred about a month after the main shock. The weakness of the slope where the mass movement started derives from the structural setting, characterized by the presence of several tectonic discontinuities, and from the unfavourable hydrogeological situation which occurs upstream along the tectonic contact between the carbonate aquifer and the aquiclude formed by the flysch materials: the barrier springs that occur there pour out enormous quantities of water (200 l/s) into the unstable basin below.
Comparison of aerial photographs taken before and after the 1980 earthquake clearly reveal the reactivation of a landslide mass that had been dormant at least for the last 40 years, as confirmed by the age of several rural houses destroyed. However, it could well be that the slide had remained dormant even longer, so that probably it is not wrong to go back in time to the 1930 earthquake for its last movement preceeding the 1980 seismic event. What is certain, however, is that no appreciable remobilization occurred over the last forty years before 1980, even though several periods of very heavy rainfalls took place.

Fig. 3. Surface faulting along the bedrock fault scarp near Senerchia: reactivated fault plane affecting striated, poorly cemented, eboulis, and displacement of the roadway. Photo taken on 04.04.1981, at close distance along the road behind the village; courtesy of Prof. Albert Pissart.

Fig. 4. Surface faulting along the bedrock fault scarp near Senerchia: free-face at the base of the limestone fault plane. Photo taken on 04.04.1981, at close distance along the road behind the village; courtesy of Prof. Albert Pissart. Note the close similarity with the free-face produced during the 1997 Colfiorito earthquakes at the Costa bedrock fault scarp (see 3. Camerino (MC) - Colfiorito (PG), Fig. 7).

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5. Castrovillari (CS). Earthquake-induced ground ruptures and paleoseismology in the Mt. Pollino Area

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Introduction

The visit to the Mt. Pollino fault generated mountain front will be described using the paper by Michetti A.M., Ferreli L., Serva L. and Vittori E. (1997), *Geological evidence for strong historical earthquakes in an "aseismic" region: the Pollino case (southern Italy)*, Journal of Geodynamics, 24, 1-4, 67-86.

ABSTRACT

The Pollino Range is the southernmost segment of the Southern Apennines, at the boundary with the Calabrian Arc. While several strong earthquakes (magnitude 6.5 - 7.0) occurred in the nearby regions, apparently the Pollino area has known no historical evidence for seismic events of magnitude > 5. We carried out an airphoto interpretation and a field survey of the Pollino fault (the major Quaternary normal fault of the area) in order to geologically characterize the seismic potential of this structure. We dug two sets of trenches across fault scarps within the apex of latest Pleistocene to Holocene alluvial fans at Masseria Quercia Marina (MQM) and Grotta Carbone (GC) sites, in the central segment of the southern Pollino Range front. At both sites we identified two surface faulting events affecting the alluvial fan deposits and two overlaying colluvial units of historical age. The penultimate event produced a vertical offset of 80-90 cm at GC and 50-60 cm at MQM; while the last event produced a vertical offset of 40-50 cm at GC and few centimeters at MQM. Based on field observations and range front morphology, we hypothesize that the two historical earthquakes reactivated at least the entire length of the Masseria Marzano - Civita segment of the Pollino fault (rupture length about 18 km). For such events, the comparison with surface faulting earthquakes in the Apennines and abroad indicates a magnitude of 6.5 - 7.0. Therefore, the maximum potential earthquake and the seismic hazard of the Pollino area is significantly larger than that suggested by the available historical seismic catalogue.

Purpose and scope of the study

The Pollino Range is the southernmost segment of the Southern Apennines, at the boundary with the Calabrian Arc. Crustal extension has shaped these regions during the Quaternary with a system of tectonic basins (e.g., Bousquet, 1973; Ciaranfi *et al.*, 1983; Westaway *et al.*, 1989; Scandone *et al.*, 1992; Valensise *et al.*, 1994; Fig. 1) and is also active today as shown by seismological data (cf. Gasparini *et al.*, 1985). While several strong earthquakes (Intensity X MCS or greater, i.e. about magnitude 6.5 to 7.0; Postpischl, 1985; Boschi *et al.*, 1995; Camassi and Stucchi, 1996; Fig. 1) occurred nearby, seemingly the Pollino area has known no historical evidence for seismic events of Intensity greater than VII MCS (cf. Valensise and Guidoboni, 1995), i.e. about magnitude 5 (yet, for low Intensity events along the Pollino fault, see Magri and Molin, 1979).

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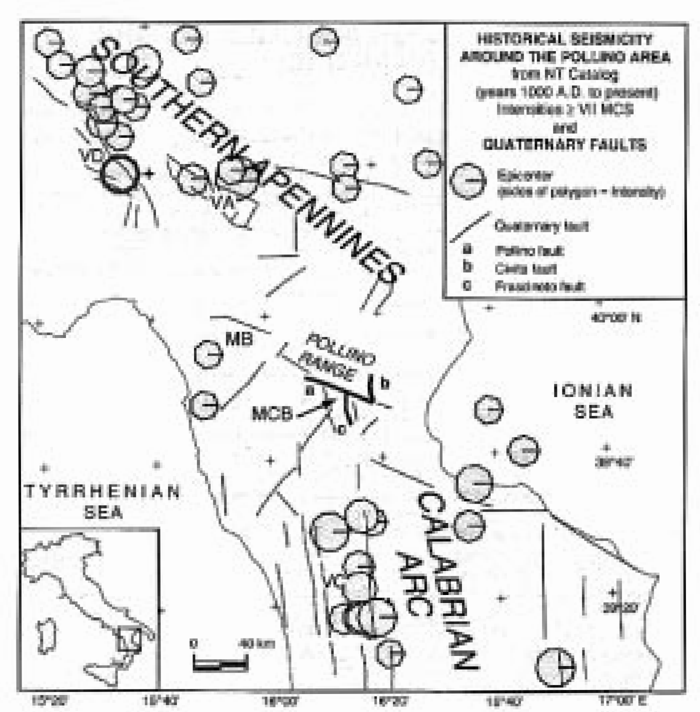

Fig. 1. Main Quaternary faults (from our own data and from Bousquet, 1973; Bosi, 1987) and historical seismicity (according to Postpischl, 1985, and Camassi and Stucchi, 1996; Intensity is given in the MCS, Mercalli-Cancani-Sieberg scale) in the area between the Apennines and the Calabrian Arc. Magnitude estimates provide values of 7 for I = XI and of 6.1 to 6.6 for I = X (Postpischl, 1985); note the lack of events with Intensity > VII near the Pollino Range, whereas faults marked "a", "b", and "c" do show evidence of recent surface faulting events. The Vallo di Diano (VD), Val d'Agri (VA), Mercure Basin (MB), Morano - Castrovillari Basin (MCB), and Valle del Crati (VC) are the main Quaternary extensional basins (e.g. Bousquet and Gueremy, 1968; 1969; Lippman-Provansal, 1985; Ascione et al., 1993; Russo and Schiattarella, 1993; Tortorici et al., 1995), corresponding (with the exception of MB and MCB) to the main historical seismogenic structures.

However, the Pollino fault, that is the major neotectonic structure of the area ("a" in Fig. 1; see also Figs 2 and 3), shows evidence of late Quaternary faulting with the same style (normal), and at least the same (but probably a larger) amount, than that observed in the neighboring seismogenic structures (e.g., Bousquet and Gueremy, 1969; Bosi, 1987; Russo and Schiattarella, 1993; Ghisetti *et al.*, 1994; Tortorici *et al.*, 1995; Fig. 1). It is reasonable to assume that in the seismotectonic environment of the Central-Southern Apennines and Calabrian Arc, major Quaternary normal faults do represent the surface expression of seismic sources capable of strong events (magnitude 6.5 to 7.0; e.g. Michetti, 1994; Vittori, 1994; Tortorici *et al.*, 1995); the lack of documented creep along either the Pollino fault and any other normal fault within the Central and Southern Apennines rules out the occurrence of significant aseismic slip.

Therefore, this apparent disagreement between geological and seismological data may have two explanations: (1) the seismic potential is higher than that suggested by the historical record and the last strong earthquake occurred in pre-historical times, since then the Pollino region being seismically quiescent; or (2) the last strong earthquake is historical but unrecog-

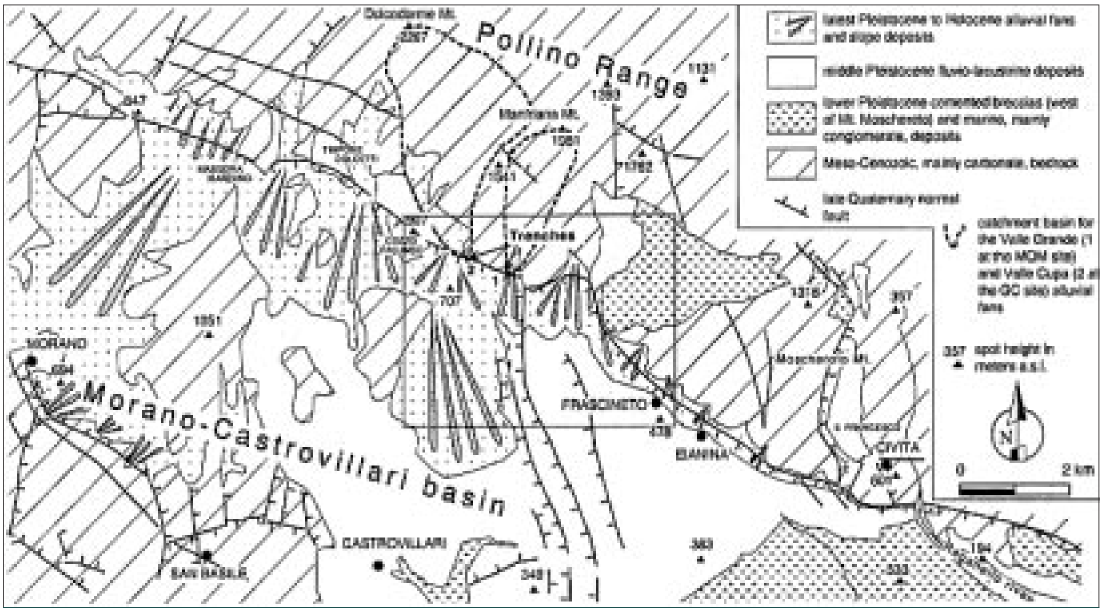

Fig. 2. Geologic map and late Quaternary faults of the Pollino region. Inset shows the location of Fig. 4

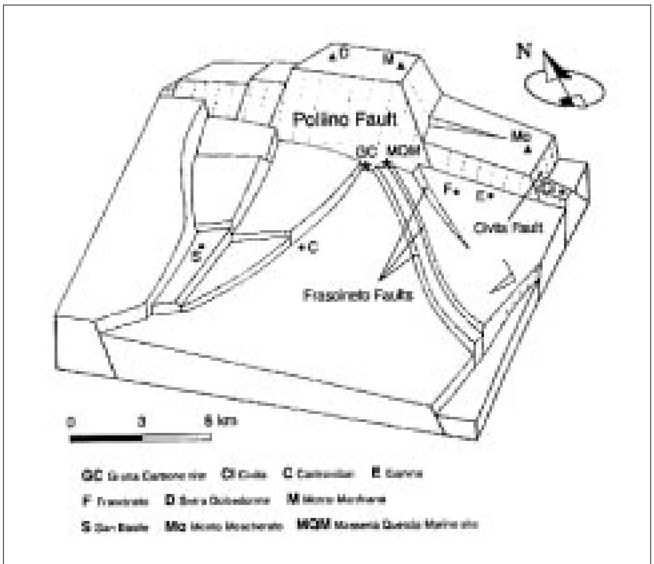

Fig: 3. Block diagram showing the structural relations in the Pollino region, near Castrovillari (after Bousquet, 1973, modified); dotted lines show the maximum dip on fault planes; the represented marker is the top surface of the lower Miocene. Post-lower Miocene vertical displacement on the Pollino fault reaches a maximum of ca. 3000 m, while post-lower Pleistocene vertical displacement reaches a maximum of ca. 400 m. Based on available geomorphic and structural data, since the beginning of the Quaternary extensional phase the N-S trending faults in the Castrovillari basin are interpreted as release faults (sensu Destro, 1995) confined within the hangingwall of a major active normal fault, the Pollino fault; likewise, cross faults developed in the Pollino fault footwall. In particular, recent activity is concentrated on the Civita and Frascineto faults.

nized, the catalogue is not complete, and the seismicity of the Pollino region is similar to that of the nearby areas. These hypotheses imply completely different scenarios for earthquake hazard characterization.

To address this issue, we firstly carried out field survey and airphoto interpretation along the range front generated by the Pollino fault, the related second-order faults (like the Frascineto cross faults, within the Pollino fault hangingwall, and the Civita fault, within the Pollino fault footwall; "b" and "c" in Fig. 1; Figs 2 and 3) and over the Morano-Castrovillari basin (Fig. 1), and then selected two sites for exploratory trenching across the Pollino fault. In this paper we summarize the results of our analyses in the Pollino region, and discuss the evidence for historical paleoseismicity. The initial findings from these investigations are described in Ferreli *et al.* (1994) and Vittori *et al.* (1995).

Table 1 – Radiocarbon samples

Samples	$d^{13}C$	Laboratory reported Radiocarbon Age[a], ^{14}C years B.P.	Calibrated Radiocarbon Age [b] (cal years at 2s)	Location [c] and Description
POL 2	-27,6	575 ± 210	1270-1615 A.D. (680-335 B.P.)	MQM1, top unit 1, wood
POL 4	-24,9	2470 ± 85	780-405 B.C. (2730-2355 B.P.)	MQM1, bottom unit 2, TOC [d] from colluvial soil
POL 6	-25,6	1030 ± 80	955-1150 A.D. (995-800 B.P.)	MQM1, top unit 2, TOC from colluvial soil
POL 7 *	-24,1	1451 ± 53	595-655 A.D. (1355-1295 B.P.)	MQM1, unit 2, charcoal
POL 8	-25,2	modern age		MQM1, unit 1 filling an open fissure, TOC from colluvial soil
POL 9	-24,4	1550 ± 75	425-610 A.D. (1525-1340 B.P.)	MQM2, unit 2, TOC from colluvial soil
POL 15	-24,5	2385 ± 75	750-390 B.C. (2700-2340 B.P.)	GC1, unit 2, TOC from colluvial soil
GC 1	-26.7	1355 ± 110	435-975 A.D. (1515-975 B.P.)	GC2, unit 2, TOC from colluvial matrix of debris
GC 2	-25.3	3705 ± 165	2565-1630 B.C. (4515-3580 B.P.)	GC2, unit 2, TOC from colluvial matrix of debris
GC 3	-25.1	5380 ± 130	4460-3950 B.C. (6410-5900 B.P.)	GC2, unit 2, TOC from colluvial soil
GC 4	-23.5	3240 ± 150	1880-1055 B.C. (3830-3005 B.P.)	GC2, unit 2, TOC from colluvial soil
GC 7	-23.2	4985 ± 90	3970-3635 B.C. (5920-5585 B.P.)	GC2, unit 2, TOC from colluvial soil
GC 8	-27.1	6095 ± 200	5435-4535B.C. (7385-6485 B.P.)	GC2, unit 2, TOC from colluvial soil
GC 11	-26.4	5075 ± 185	4330-3385 B.C. (6280-5335 B.P.)	GC2, unit 2.F3, TOC from colluvial soil
GC 13	-25.9	5840 ± 185	5215-4332 B.C. (7165-6280 B.P.)	GC2, unit 3.C, TOC from colluvial soil
POL 10	-25,0	19470 ± 395		colluvial soil within the alluvial fan deposits at Cava D'Atri, sampled at a depth of 6 m, TOC
POL 11	-25,6	14260 ± 200		colluvial soil within the alluvial fan deposits at Cava D'Atri, sampled at a depth of 3 m, TOC
POL 12	-25,6	25330 ± 1380		colluvial soil within the alluvial fan deposits at Cava A, sampled at a depth of 7 m, TOC
POL 14	-24,3	25915 ± 465		charcoal-rich colluvial soil within the alluvial fan deposits at Masseria Marzano, sampled at a depth of 5 m, TOC

[a] Radiocarbon analyses by GEOCHRON Laboratory, Cambridge, Massachusetts, USA.
[b] Age calibration calculated (cal) following Stuiver and Becker, 1993.
[c] Samples collected, for instance, at Masseria Quercia Marina site, trench 1, are marked as MQM1; Figs 2 and 4 show the map location of the sampling sites.
[d] TOC is for total organic carbon
* AMS dating

The Pollino fault

The Pollino region is part of the northwest-southeast trending Apenninic Chain built up as a thrust-and-fold belt essentially during the upper Miocene to lower Pleistocene (e.g. Royden *et al.*, 1987; Patacca *et al.*, 1992; Ghisetti *et al.*, 1994). Previous studies indicate that the Pollino fault firstly evolved in a compressional regional environment and was characterized by significant amount of both horizontal and vertical offset (Bousquet, 1973; Monaco, 1993; refer to Ghisetti *et al.*, 1994, for a critical summary of the pertinent literature).
At present, the Pollino fault is a WNW-trending structure (Fig. 2) characterized by an impressive range front with more than 1400 m of relief. The fault hangingwall hosts the Morano-Castrovillari basin, whose sedimentary filling reaches its maximum thickness (about 600 m of upper Pliocene to lower Pleistocene marine sequence plus about 300 m of middle Pleistocene to Holocene continental deposits; see Russo and Schiattarella, 1993, and references herein) against the fault plane. The N-S trending and antithetic faults in Figure 2, generated as second-order structures of the Pollino fault, also display clear evidence of late Quaternary activity. Our own data and previous structural and geomorphological studies (e.g. Bousquet and Gueremy, 1969; Bousquet, 1973; Russo and Schiattarella, 1993) clearly demonstrate that the

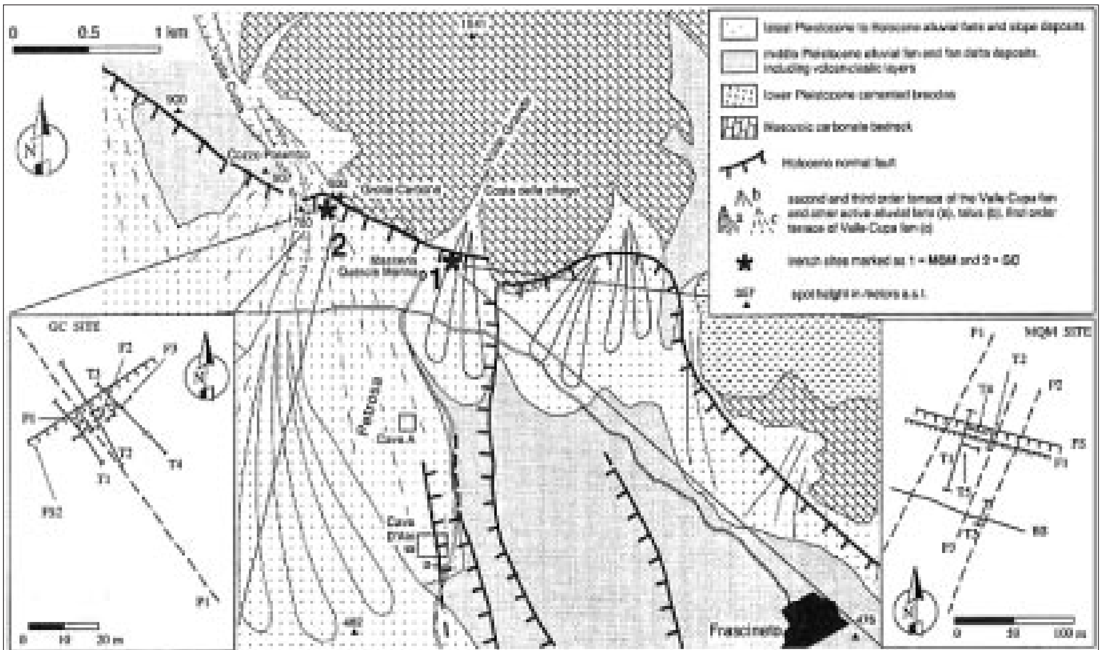

Fig. 4: Map of the trench sites area. Insets show detailed map of the trench sites: P) trace of the topographic profiles shown in Figs. 5 and 8; T) trench; F) fault; FS) fault scarp edge; BS) break in slope marking the boundary of the backtilted sector of the alluvial fan at MQM as illustrated in Fig. 5.

Pollino fault dominates the recent evolution of the region. Comparison of Figures 2 and 3 illustrates the hierarchy of active tectonic structures in the area. While part of the vertical offset is probably linked to pre-rift tectonic phases, inception of Quaternary extensional tectonics activated the Pollino fault as a major, purely normal fault. Consequently, the Morano-Castrovillari basin developed in the subsiding hangingwall of the Pollino fault, and second-order cross faults (Civita and Frascineto faults) formed. In particular, the Frascineto faults are typical *release faults* observed to form in the hangingwall of normal faults to accommodate extension along the strike of the dominant fault segment (Destro, 1995; Roberts, 1996). In the current stress field acting in the southern Apennines (e.g., Cello *et al.*, 1982) the above structural relations are the result of a NE- SW extension and of a strong regional uplift (cf. Bousquet, 1973; Ciaranfi *et al.*, 1983; Gasparini *et al.*, 1985). The nearby Mercure Basin (MB in Fig. 1; cf. Bousquet and Gueremy, 1968; Bousquet, 1973) shows a very similar setting, controlled by aspectacular WNW-trending active fault representing the extension of the Pollino structure; this demonstrates that the Pollino fault belongs to the main active and segmented normal fault system of the Southern Apennines (Fig. 1).

The nature of the Pollino fault is also corroborated by data from Ferreli *et al.* (1994), showing no evidence for significant strike-slip offsets in middle Pleistocene to Holocene landforms and deposits along the faults in Figure 2. The above data also allowed us to confirm the amount of late Quaternary normal displacement evaluated by Bousquet and Gueremy (1969) and Russo and Schiattarella (1993) along the Pollino fault and related minor faults (such as the Civita and Frascineto faults). Latest Pleistocene to Holocene displacement is apparent on the entire fault trace mapped in Figure 2. In particular, at Masseria Marzano fan deposits younger than 25,915 ± 465 ^{14}C years B.P. (sample POL 14, Table 1) are dragged and faulted, and similar relations exist near Timpone Dolcetti, as already illustrated by Bousquet and Gueremy (1969). Between Eianina and Civita, the range front base is oversteepened over a two km long section by a fresh limestone fault scarp showing associated features (non-karstified fault planes, thickening of Holocene talus against the fault planes, faulted Pleistocene breccias) typically observed along historical earthquake scarps in the Mediterranean region (e.g., during the 1915 Fucino earthquake, Serva et al., 1988; or the 1981 Corinth earthquakes, Jackson *et al.*, 1982; cf. Stewart and Hancock, 1994, and references herein).

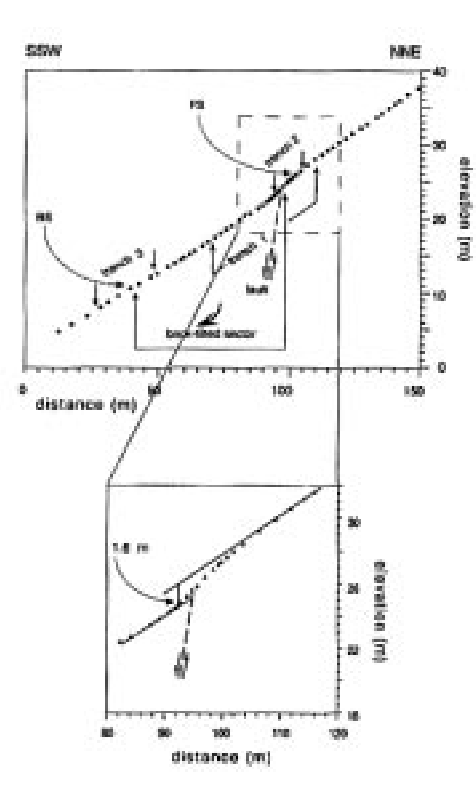

Fig. 5. Topographic profile at MQM (see location in Fig. 4); elevations relative to an arbitrary datum. Note the back-tilting of the alluvial fan deposits; no antithetic fault was found in trench 3, dug across its lower boundary (BS). The vertical displacement of 1.6 m across the main fault scarp (FS) is therefore greater than the net vertical tectonic offset at this site.

Ferreli et al. (1994) describe fault scarps affecting the apex of two adjacent young alluvial fans (Figs 2 and 4) at Masseria Quercia Marina (MQM) and Grotta Carbone (GC). Since it is reasonable to assume that the faulting history at these sites elucidate the most recent movements of at least the entire Masseria Marzano – Civita segment of the Pollino fault (about 18 km long), and this fault clearly controls the active tectonics of the whole area, we surveyed and trenched both the MQM and GC fault scarps in order to recognize evidence of the last surface faulting earthquakes.

Trench walls were mapped at 1:20 scale, and samples were collected for ^{14}C dating (Table 1) and soil analyses (P. Lorenzoni and M. Raglione, written comm., 1995). Since the stratigraphy of the trenches shows remarkable correspondences, we correlated the mapped units using the same unit numbers.

Masseria Quercia Marina site

The Pollino Range area, whose relief varies from more than 2200 m (Monte Pollino) to almost the sea level (Raganello River valley floor; Fig. 2), has a mountain-mediterranean climate characterized by 1000 - 1500 mm/yr of precipitation and 5 - 15° C of average temperature; at elevations higher than 1500 m the snow permanently covers the soil from December to March. The range front is essentially carved in carbonate rocks. This is the same geographic and lithologic setting of other intermountain tectonic depressions in the Apennines, like the Rieti basin and the Fucino basin in Central Italy (cf. Michetti *et al.*, 1995; Giraudi, 1995), where the latest Pleistocene to Holocene climatic history appears to be quite similar (e.g. Palmentola *et al.*, 1990). The latter features are relevant for interpreting the stratigraphy, morphology and faulting at MQM and GC sites.
The Valle Grande fan has a relatively small surface (Fig. 4) where active deposition still occurs during rare alluvial events; the present-day channel is not deeply entrenched. The deposition rate was much higher during the last glacial period, when the top of the Pollino range hosted significant glaciers (Palmentola et al., 1990) and the forest cover was completely removed from the slopes (cf. Watts, 1985). The scarp at MQM is very well preserved for most of its length within the fan surface, suggesting a very recent age of faulting. At the trench site the fault scarp is 1.6 - 1.8 m high, however, backtilting does amplify the net vertical displacement (Ferreli et al., 1994; see Fig. 5). Figure 4 shows the location of the trenches at MQM. The fol-

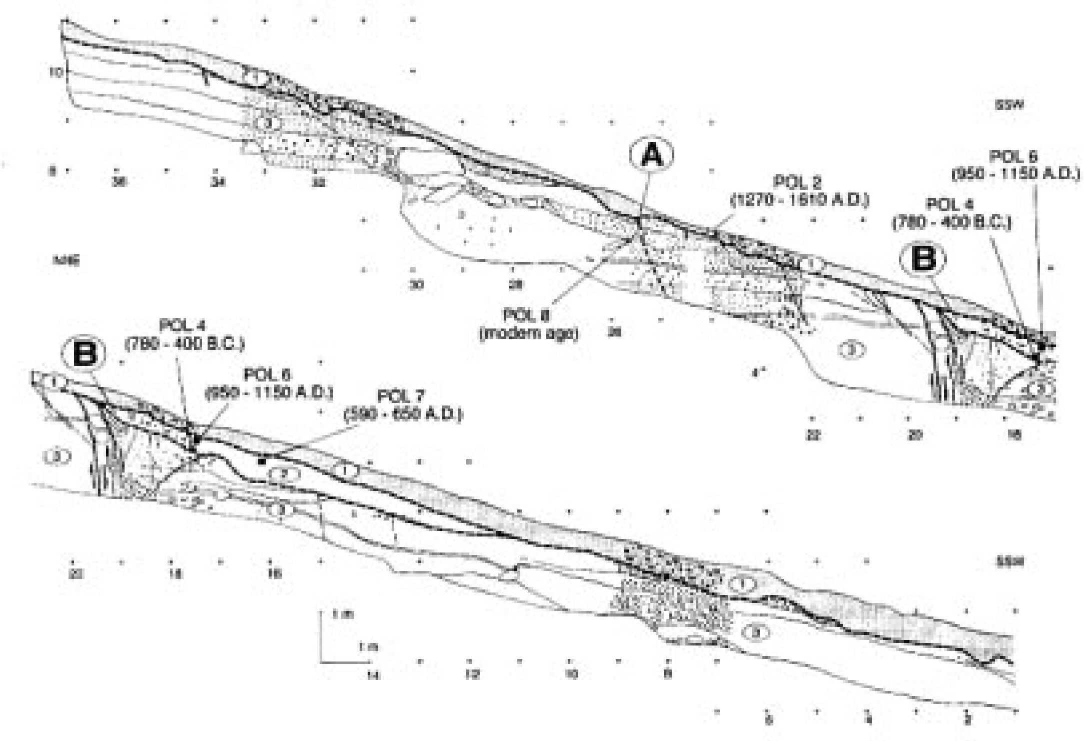

Fig. 6. MQM site, log of the eastern wall of trench 1 near the fault zone; the normal fault plane (marked by arrows) trends N108° and dips 70°S . Note location (black dots) and age (see Tab. 1) of samples (POL 2, etc.) collected for radiocarbon dating. Two main erosional surfaces (marked by bold dashed lines) separate the mapped depositional units (marked by encircled numbers): 1) upper colluvial soil, 10YR 3/2; 2) lower colluvial soil, 2,5YR 3/4; 3) bedded alluvial fan deposits made of carbonate clasts. A and B mark the evidence for the last and penultimate paleoseismic event.

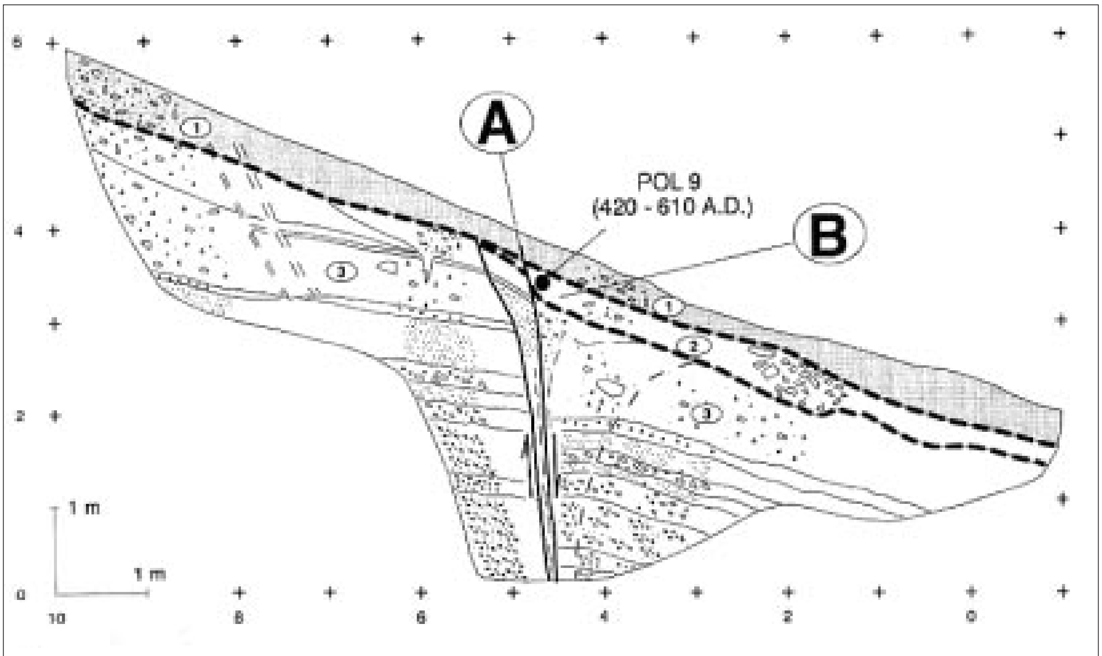

Fig. 7. MQM site, log of the eastern wall of trench 2; the fault plane has the same attitude as in trench 1. Symbols as in Fig. 6.

lowing discussion is based on data from trenches 1 and 2 (Figs 6 and 7) that illustrate evidence for past surface faulting events.

The fault plane visible in the trenches affects the fan deposits and two overlaying colluvial soils. The lower colluvial soil (unit 2 in Figs 6 and 7) thickens toward the fault plane, and is preserved only in a 8 m section of the fault hangingwall in trench 1 (Fig. 6). This indicates that when unit 2 was at the ground surface a movementof the fault, that we indicate as event B in the following (see Figs 6 and 7), produced rejuvenation of the scarp, and backtilting of the hangingwall. An erosional phase then removed unit 2 from the footwall and most of the hangingwall before deposition of unit 1. While we lack correlative units on both sides of the fault, the top of the alluvial sequence (top of unit 3 in Figs 6 and 7) indicates a displacement from event B of about 0.5 to 0.6 m; since the texture of unit 2 near the fault does not show evidence for the retreat of a big free face, this value can be seen as a reasonable approximation.

In trench 2 the fault also displaces a few centimeters of the upper colluvial soil (units 1.1 and 1.2), which also fills an open fracture in trench 1 (see "A" in Figs 6 and 7). On the whole, evidence for a fault movement younger than event B at this site is based on small effects, and we were at first very cautious about its interpretation. After finding a much more convincing evidence at GC, we now think that these effects may confidently be regarded as the fossil record of the last faulting event, i.e. event A.

Radiocarbon dating of charcoal from trench 1 (sample POL 7, see Table 1) shows that the deposition of unit 2 occurred in a time span comprised between 590 - 650 A.D. or later (the radiocarbon dates reported in the text and figures are the calibrated ages from Table 1 adjusted to the nearest decade), thus suggesting that the age of event B is younger than the VI century A.D. A similar timing is also corroborated by dating of total organic carbon (TOC) from unit 2 (POL 4, POL 6 and POL 9 in Figs 6 and 7), the corresponding ages being in agreement with stratigraphy and dating of wood from unit 1 (POL 2 in Fig. 6). This indicates that, despite the inherent uncertainty associated with the colluvial nature of this unit, the age of its organic matter correctly reflects the age of deposition at this site. Since unit 1 is the present-day soil, contamination from modern organic matter complicates accurate ^{14}C dating. However, dating of POL 2, while showing a wide age range due to the penetration of rootlets in the analyzed wood sample, suggests that the deposition of unit 1 originated before 1270 - 1610 A.D.; this is also the minimum age for event B. Pedological analyses indicates that the upper undisplaced section of unit 1 required at least 2 - 3 centuries to form (P. Lorenzoni and M. Raglione, writ. comm., 1995). To summarize, the time intervals comprised between the VI and XII century A.D. (for event B) and the XIII and XV century A.D. (for event A) represent our best estimate of earthquake dates at MQM.

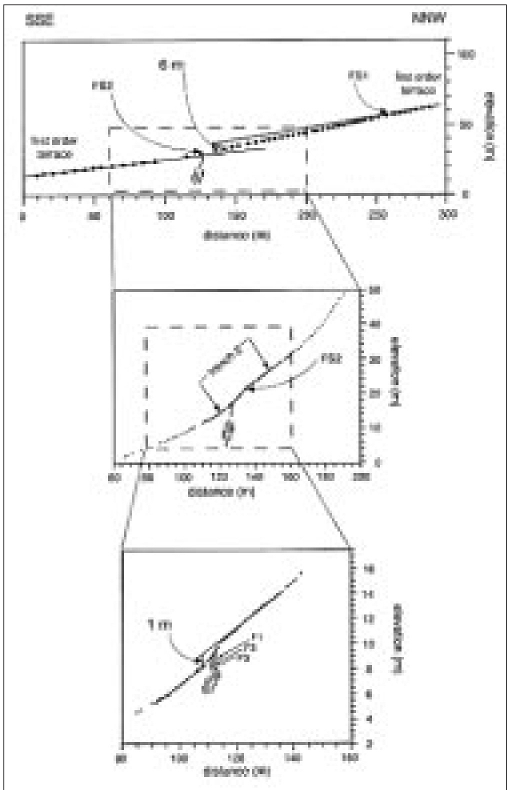

Fig. 8. Topographic profiles at GC (see location in Fig. 4); elevations relative to an arbitrary datum. Symbols as in Fig. 5. The upper profile includes both the main scarp (FS1) and the trenched scarplet at its base (FS2); the lower profiles were performed with increasing detail along the same trace.

Grotta Carbone site

The Valle Cupa fan has a catchment basin much wider than the Valle Grande fan (Figs 2 and 4), and this explains their different Pleistocene to Holocene evolution. While the Valle Grande fan has a regular surface with a feeble channel incision, the Valle Cupa fan displays three orders of recent terraces. Radiocarbon dating of colluvial soils (samples POL 10 and POL 11 from Cava D'Atri; and POL 12 from Cava A; ages in Tab. 1, locations in Fig. 4) beneath the first order surface in the piedmont belt strongly suggests that the corresponding phase of fan development started during the last Glacial stage and ended at about 14 kyr ago. For example, this is in very good agreement with the late glacial evolution of the Majelama fan in the Fucino area, in a similar geomorphic setting (e.g. Frezzotti and Giraudi, 1992); thus suggesting a common climatic control on this depositional phase.

The deposition rate during this phase at Cava A was higher than 0.6 mm/yr, and most likely higher than the corresponding vertical slip rate of the Pollino fault. This indicates that during this phase there was no fault scarp at GC, that is at the Valle Cupa fan apex. We assume that in the footwall of the Pollino fault the inception of fan apex entrenchment and, therefore, of displacement of the first order terrace surface, occurred at about 14 kyr B.P. Observations from trench exposures, however, show that at GC, that is in the Pollino fault hangingwall, the alluvial fan sedimentation continued until about 6 kyr ago (shortly after the deposition of unit 3.C and the dating of sample GC 13; Fig. 9); hence, downslope of GC the first order surface age increases with the distance from the range front fault. Topographic survey demonstrates that the first order surface offset at the main GC scarp (FS1 in Figs 4 and 8) is at least 6 m. Since we assume that the age of this surface in the Pollino fault footwall is about 14 kyr B.P., this yields a minimum vertical slip rate of 0.4 mm/year. This is consistent with the slip rate obtained from the GC trenches. The offset of the 7160 - 6280 yr old colluvial soil of unit 3.C (Figs 9 and 10; sample GC 13, Table 1) is greater than 2.5 m, yielding a minimum vertical slip rate of 0.35 - 0.4 mm/yr.

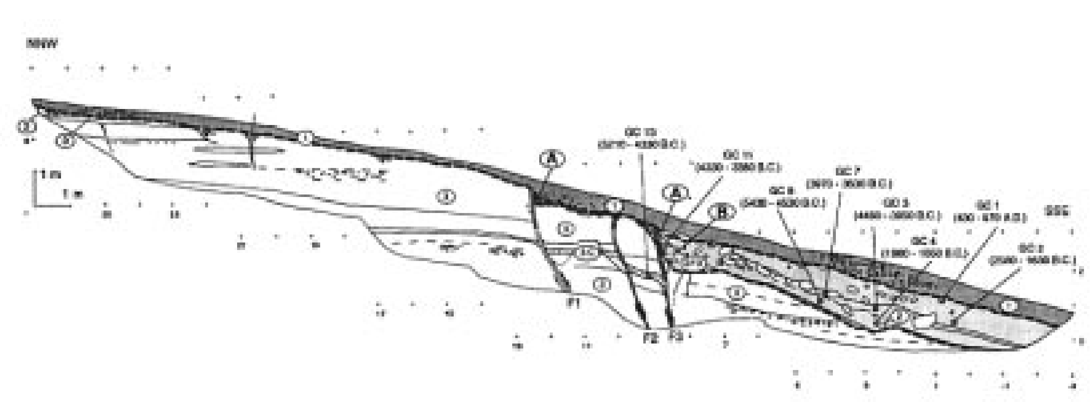

Fig. 9. GC site, log of the eastern wall of trench 2. The normal fault zone (fault planes F1, F2 and F3 are marked by arrows) has an average trend of N77° and dip of 70°S; however, the fault planes are not parallel to each other, as shown in Fig. 4. Note location (black dots) and age (see Tab. 1) of samples (GC 1, etc.) collected for radiocarbon dating. Two main erosional surfaces (marked by bold dashed lines) separate the mapped depositional units (marked by encircled numbers): 1) upper colluvial soil, 5YR between 2,5/2 and 3/3; 2) slope deposits including colluvial soils between 5YR 4/6 and 7,5YR 4/6, containing volcanic minerals, in the lower part, and carbonate debris in the upper part; 2.F3) colluvial wedge at the base of unit 2 near the fault plane F3; 3) bedded alluvial fan deposits made of carbonate clasts; 3.C) colluvial soil. A and B mark the evidence for the last and penultimate paleoseismic event.

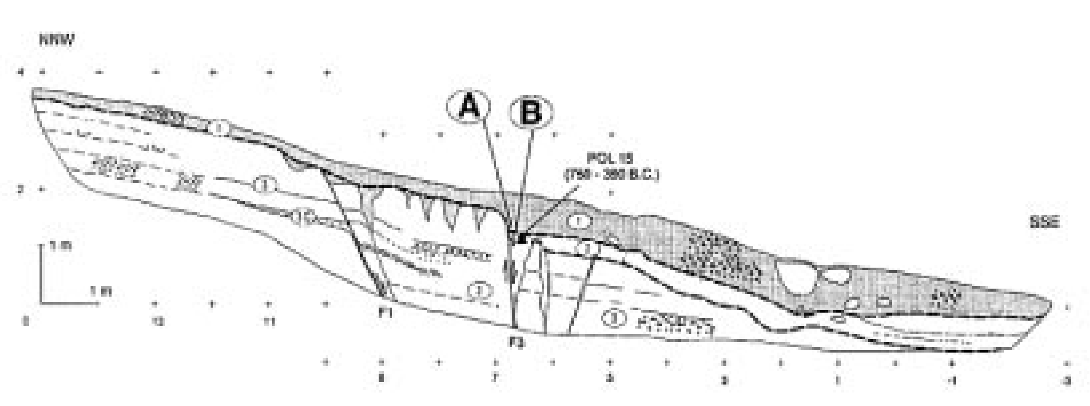

Fig. 10. GC site, log of the eastern wall of trench 1. Symbols as in Fig. 9.

A fresh scarplet located at the toe of the main scarp (FS1 in Fig. 8) marks the most recent fault movements at GC (FS2 in Figs 4 and 8). In the following we focus on observations from trench 1 and 2, dug across FS2, since these revealed the best record of past surface faulting earthquakes (Figs 9 and 10). The stratigraphy illustrates the final phase of the alluvial deposition, including a remarkable colluvial soil (unit 3.C), followed by two episodes of erosion and subsequent colluvial and slope deposition (Figs 9 and 10). The trenches exposed a normal fault zone composed of three listric discontinuity planes; only the fan deposits (unit 3) are clearly offset at each plane. The cumulative displacement of unit 3.C is significantly greater than that of the overlaying unit 2, evidence of faulting event(s) sealed by the first erosional phase.
Unit 2 is a slope deposit made of tephra-derived colluvial soils in the lower part and carbonate debris in the upper part. In trench 2 the lower part of unit 2 thickens close to the fault (Fig 9). The surface between unit 3 and unit 2 clearly indicates a degraded fault scarp free face, buried by the colluvial wedge of unit 2.F3. The height of this paleoscarp suggests that the causative event (event B, Fig. 9) produced a vertical displacement of about 0.8 - 0.9 m at F3; this is in good agreement with observations from trench 1 (Fig. 10). Moreover, in trench 2 at

F3 unit 2.F3 and unit 1 are affected by a younger faulting event (event A, Fig. 9) with a 0.1 m throw. Unit 1 is displaced by additional 0.2 - 0.3 m at F1. The cumulative vertical displacement for event A is therefore 0.3 to 0.4 m. This is consistent with the 0.4 - 0.5 m displacement estimated in trench 1 at F3 from the degraded fault scarp free face buried by unit 1 (Fig. 10).
Earthquake dates are not well constrained at GC because of 1) the lack of charcoal samples, and 2) the uncertainty involved in the interpretation of total organic matter dating from colluvial soils (see Table 1). Samples GC 3 and GC 4 show a chronostratigraphic reversal; sample GC 13, collected within the alluvial sequence (Fig. 9), has a ^{14}C age younger (see GC 8 in Fig. 9) or only slightly older (see GC 3, GC 7 and GC 11 in Fig. 9) than samples collected within the colluvial soils of unit 2. Furthermore, despite the similitude in the trench stratigraphy and pedologic features, most of the samples from unit 2 at GC appear to be significantly older than the corresponding samples at MQM. This different organic matter age and distribution in colluvial soils, that we interpret as correlative, may be explained by taking into account the different geomorphic setting at MQM and GC. The trench site at GC is near the first order terrace edge and 20 m above the present-day valley floor. The age of sample GC 13 clearly suggests that channel entrenchment took place shortly after 6 kyr B.P. The subsequent colluvial deposition at GC derives only from the nearby slopes, as also demonstrated by clast lithology and soil analyses (P. Lorenzoni and M. Raglione, writ. comm., 1995). In the Central Apennines, the major Holocene phase of soil development, forest expansion and slope stability occurred at 8 to 6 kyr B.P. (cf. Giraudi and Narcisi, 1995), about the same age range of samples GC 3, GC 8, GC 11, and GC 13 (Fig. 9). We therefore suggest that the organic matter produced during that phase was subsequently stored in the soils uphill from GC; this aged organic matter was then recycled at least twice in the trench stratigraphy, i.e. during the deposition of unit 3.C and during the deposition of unit 2. This reconstruction is consistent with typical features of colluvial deposition along Apennines carbonate mountain slopes (Giraudi, 1995; Michetti *et al.*, 1995; Frezzotti and Narcisi, 1996). For instance, a sharp climatically-controlled phase of soil erosion and re-deposition with stratigraphic inversion across a fault scarp has been documented by Lorenzoni *et al.* (1993) in the Rieti basin.
Based on the above considerations, we believe,that radiocarbon dates of unit 2 samples at GC record the age of the original organic matter, that is older than the age of the corresponding beds. Conversely, the internal consistency and the lack of reversals for ^{14}C ages at MQM strongly suggests that the active alluvial fan environment, where input of organic matter derives from periodical erosion of the whole catchment basin, allows satisfactory TOC dating. Therefore, to assess the age of faulting events at GC we follow the same chronological frame established at MQM, and interpret event A and event B as occurred in the same period at both sites.

Discussion and conclusion

Paleoseismic analysis at MQM and GC sites, located south of the Pollino Range, demonstrates that the Pollino fault ruptured during two historical surface faulting earthquakes. At both sites there is no evidence for significant lateral components of motion neither from displaced landforms nor from trench exposures, thus showing that the Pollino fault has had a mainly normal style of surface faulting. Our investigations and literature data strongly suggest that faults mapped in Figure 2 are presently acting as normal faults capable to produce surface displacement during strong earthquakes; indeed, Holocene reactivation has been proved for the Pollino, Frascineto and Civita faults (Ferreli *et al.*, 1994; Cinti *et al.* 1995a; in the latter, the Frascineto faults are referred to as "Castrovillari fault"). However, the structural and geomorphic setting definitely demonstrates that the master fault of the Pollino seismogenic structure is the Pollino fault (cf. Cinti *et al.*, 1995b).

Field observations and published data (Bousquet and Gueremy, 1969; Bousquet, 1973; Russo and Schiattarella, 1993; Ferreli *et al.* 1994) clearly show that the recognized events can be confidently associated with a rupture length of at least 18 km (Masseria Marzano - Civita segment, Fig. 2). Coseismic displacement from trench mapping is consistently larger at GC (0.8 - 0.9 m for event B, 0.4 - 0.5 m for event A) than at MQM (0.5 - 0.6 m for event B, less than 0.1 m for event A), in agreement with the suggested continuity of fault rupturing between the two sites. These features are coherent with contemporary surface faulting occurred in the Central and Southern Apennines (Serva *et al.*, 1988; Pantosti and Valensise, 1990) and corresponding paleoseismic evidence (Pantosti *et al.*, 1993; Michetti *et al.*, 1996), suggesting earthquake magnitudes ranging between 6.5 and 7.0 for both events A and B.

Although the ages from unit 2 at GC appear unsatisfactory, radiocarbon dating suggests that event B at MQM took place in the early Middle Ages, our best estimate being VI to XII century A.D. Radiocarbon dating and soil analyses also provide a basis for estimating the age of event A, our preferred range being XIII to XV century A.D. This is also in agreement with the very young faulted landforms. Main result of our study is, therefore, that two strong earthquakes are missing or unrecognized in the historical record of the Pollino region (cf. Valensise and Guidoboni, 1995, for a discussion of the available historical sources); accordingly, hypothesis (2) from the introduction appears to be the correct one. In other words, the present situation in the Pollino region might be similar to that of the Fucino basin before the 13.01.1915, magnitude 7, earthquake; i.e. the seismic catalogue until 1914 shows only subdued seismicity for this area, while paleoseismological studies (cf. Michetti *et al.*, 1996; Galadini *et al.*, this volume) found convincing evidence of strong historical events.

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North Calabria - 1783 and 1905 earthquake-triggered landslides

G. Chiodo[1], L. Merenda[1] and M. Sorriso-Valvo[1]

Most significant effects on the environment of several historical destructive earthquakes occurred in Calabria (Southern Italy) are briefly discussed.
As a preliminary result of a long-yterm study, a chronogical list of the Calabrian sites in which certain signs of sismo-induced landslides orderly have been reckonned, is produced. The data base so built also contains data on the seismic events as cause of the movement, on the geological, morphological and geothecnical features of the involved sites, and the information extracted by the historical sources. In the cases in which we have carried out the investigation on the ground, the database contains informations on typology of the movement according to the classification of Varnes (1978) and a synthetic evaluation of the damage.
Here we describe results derived from the study of three destructive earthquakes happened in Calabria in the 1638, 1783 e 1905 AD. The last two earthquakes will be illustrated, as data for 1638 event are poorly reliable. Two localities with seismo-induced landslides by this earthquakes will be visited during workshop, S. Lucido and Martirano Lombardo, north Calabria.

Itinerary of 2001 September 25 excursion.

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1 - The March 23, 1783 earthquake

The 28th March, 1783 earthquake was one of the most destructive earthquakes of Calabrian region in historical times.
This earthquake was the last of the five strongest earthquakes occurred in the period February-March 1783 (Tab. 1), that were part of a seismic period continued for about four years from 1783 to 1786. The total number of casualties, including those due to the epidemic waves and other aftershocks, reached 35.000 (Vivenzio 1783).

The first three earthquakes involved an area located in Southern Calabria (the Gioia Tauro plain) and Messina Straits. In the last strong earthquake of 28 March, the epicentral area migrated northwards in the Catanzaro area.

Tab.1: Table of the 1783 Earthquake of Calabria; I_0 = seismic intensity at the epicentre; I_{max} = maximum observed seismic intensity. *M* = derived magnitude. Data from CPTI Working Group, 1999

Major 1783 earthquake	I_0	I max	M_m
1783/02/05	XI	XI	7.1
1783/02/06	IX	X	5.8
1783/02/07	X-XI	X-XI	6.8
1783/03/01	VIII-IX	IX-X	6.0
1783/03/28	X	XI	6.6

The distribution of intensity of the 1783 28th March in the central part of Calabria is shown in Fig.1a. The geographic distribution of phenomena most probably related to earthquake-induced landslides is shown in Tab. 2 and Fig. 1b.

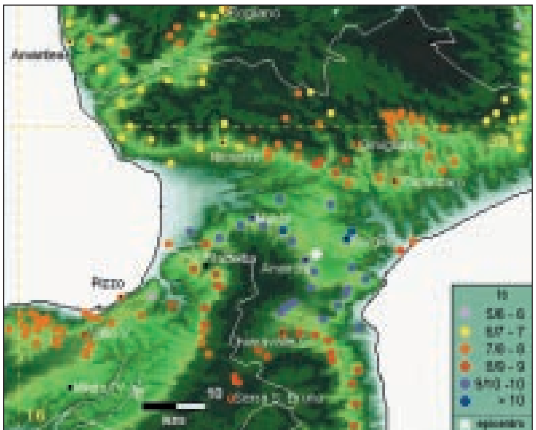

Fig. 1a. Distribution of intensity (March 28, 1783 earthquake). (Camassi & Stucchi, 2000)

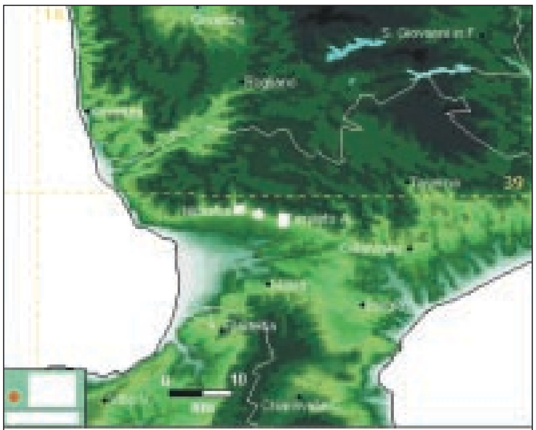

Fig. 1b. Landslides (*frane* in the legend) triggered by March 28, 1783 earthquake.

Table 2 - Sites with landslides and their distance from epicentre (March 28, 1783 earthquake)

Sites	Distance from epicentre	Sites	Distance from epicentre
Aiello Serra	4	Maida	12
Altilia (CS)	43	Marano Marchesato	53
Amato	15	Marcellinara	13
Borgia	2	**Martirano**	**36**
Caraffa Di Catanzaro	7	Pedace	53
		San Floro	4
Carolei	55	**San Lucido**	**70**
Cortale	7	San Mango	36
Girifalco	7	San Sostene	19
Laurignano	60	Squillace	3.5
Figline V.	49	Vallelonga	25

2 - The 1905 earthquake

The 8th September, 1905 earthquake was one of the most destructive earthquakes in XXth Century, in Italy. The cities and villages damaged by the earthquake were 326. Casualties totalled 600, and 300,000 people remained without cover.

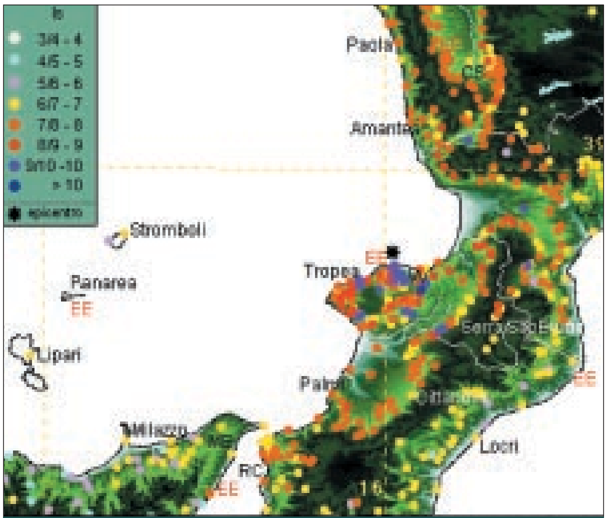

Fig.2a. Distribution of intensity (September 8, 1905 earthquake). In (Camassi & Stucchi,2000)

In the CPTI working group, 1999 report, the most recent publication on strong Italian earthquakes, it is reported an intensity of XI on the Mercalli scale, and a magnitude of 7.1 with the epicentre located offshore, in the S. Eufemia Gulf (Fig. 2a).

Table 3 – Sites with landslides and their distance from epicenter of 1905 earthquake.

Sites	Distance from epicentre	Sites	Distance from epicentre
Acri	90	Guardavalle	48
Aiello Calabro	40	Majerato (Angitola)	12
Amaroni	32	Marcellinara	40
Amato	35	**MARTIRANO**	**38**
Belmonte Calabro	42	Mileto	20
Briatico Vecchio	10	Parghelia	16
Caraffa di Catanzaro	38	Ricadi	24
Caulonia	57	Rosarno	32
Cessaniti	10	San Floro	39
Cirò	110	San Gregorio di Ippona	14
Cleto	38	San Leo Diruto	12
Conflenti	38	San Martino di Finita	80
Conidoni	10	Santa Sofia d'Epiro	88
Cortale	30	Seminara	50
Curinga	22	Stefanaconi	12
Dinami	28	Tiriolo	44
Feroleto Antico	34	Triparni	10
Filandari	16	Tropea	18
Gerocarne	24	Vallelonga	24
Gizzeria	26	Vena (VV)	12
Fitili	14	Zungri	16

The distribution of the 8th September of 1905 intensity is very irregular and the localisation of the epicentre from macroseismic data has been so far much debated.
In the sites near the Mt. Poro area, and in few more sites into and near the Savuto River Valley, the intensity reached higher value when compared with the nearby localities; in the same sites, contemporary reports indicate the presence of landslides occurred contemporaneously and shortly after the earthquake, but it is not clear what was the real space-time relationship between seismic event and instability phenomena. We could note that some phenomena are replicas of landslides occurred during earthquakes dating since 1600.
The distribution of intensity of the 1905, March 9th in the central part of Calabria is shown in Fig. 2a. The distribution of landslides most probably triggered by the earthquake is shown in Fig. 2b.

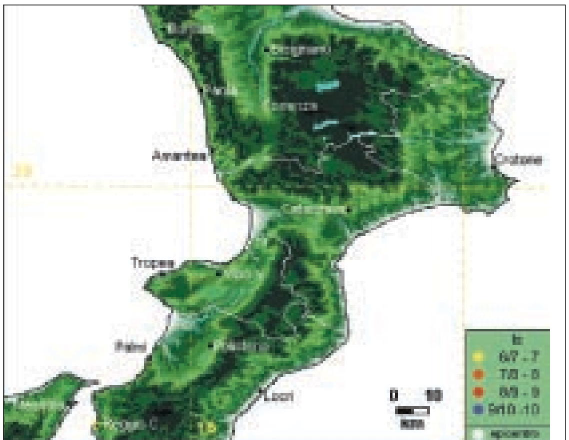

Fig. 2b. Landslides (*frane* in the legend) triggered by the 1905 earthquake. White lines are the province boundaries.

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The ancient large land-slide of Mount Martinella – San Lucido (CS), in the calabrian coastal chain, and its reactivation by the calabrian earthquakes of 1783

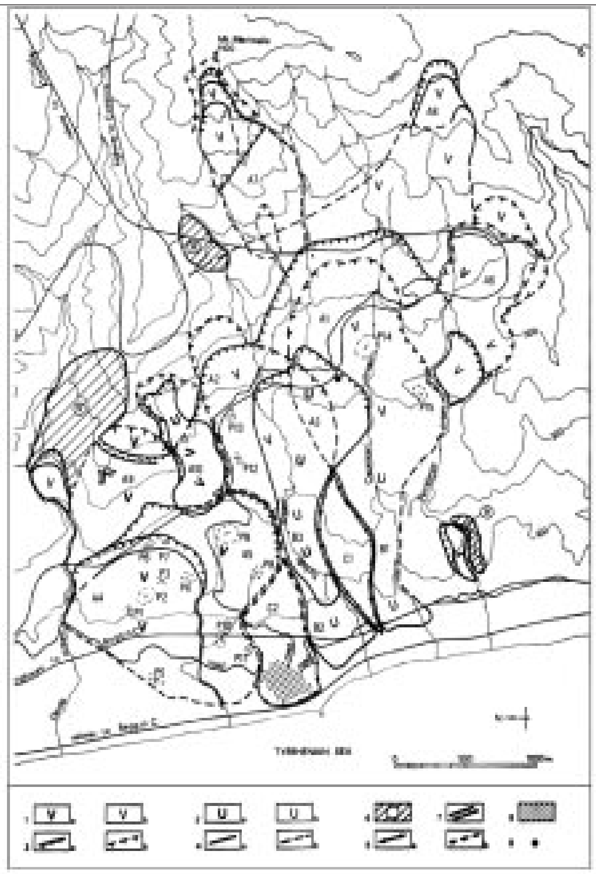
L. Merenda[1]

Studies of landslides distribution carried out many years ago (1980 – 1990) in the Calabrian "Catena Costiera" (Coastal Range), have shown that landslide phenomena of various type, age and degree of activity were widely diffused throughout the entire Range; often they are of quite considerable dimension.
A large landslide phenomenon, located on the western side-slope of the Tyrrhenian Coastal Range, developed from an elevation of about 1100 - 800 m a.s.l., over a length of 5 km and an average width of 2-3 km (with an area of about 10-15 km^2), in the area of Mt. Martinella – Falconara Albanese and San Lucido (province of Cosenza).
Evidence of numerous events of general and paroxysmal activation ("diastrofic movements") have been recognised. One of these reactivations is certainly correlated with the 1783 Calabrian earthquakes, occurred between February 5^{th} and March 28^{th} (Fig. 1). Currently, the slope movement is believed not to be exhausted; indeed, it still causes strong marine erosion, together with stability problems to important infrastructure facilities, such as the state road SS.108 and the railway, as well as to urbanised areas rather densely populated.
The Coastal Range is composed of calcareous-dolomitic nappes of the Apennine Chain, tectonically overlain by metamorphic rocks belonging to the nappes of the Calabrian Alpine Chain; in some places the former units crop out in tectonic windows.
Conspicuous and rapid uplift occurred in the Range, particularly during the Quaternary, through a wide distribution of faults and joints, producing the present "horst" configuration comprised between the "grabens" of the Crati valley and Tyrrhenian basin. Tortonian to Quaternary sedimentary sequences overlie the eastern and western flanks of the Range; these are also to be found at high altitude.
In the Middle-Lower Pleistocene, as a consequence of a left transcurrent fault (with a NNE direction) crossing the Falconara Albanese-San Lucido area, these units were affected by tectonic deformation of "pull-apart" type, which was probably re-activated by ancient earthquakes and by the 1783 tremors as "lateral spreading" and associated typology of more superficial "complex" landslides and flows.
Gneiss and underlying phyllites make up the basement of the landslide area. These are overlain by Miocene formations (conglomerate, sandstone and limestone) and Pleistocene continental conglomerate. Deep ruptures and dislocations are to be observed in these formations, also produced by mass movements.

Events of the year 1783

According to the chronicle of the times, from Autumn 1782 till the end of January 1783 "continual and intense" rainfalls in Calabria caused landslides and floods throughout the entire region (Sarcone, 1784).

[1] CNR-IRPI, Cosenza



- LANDSLIDE MAP

Legend: 1) *Slide* type lands.; 2) *Flow* type lands.; symbols "a" indicate moderately active phenomena, symbols "b" dormant phen.; 3) Scarps; 4) Debris limit; 5) Stabilized debris; symbols "a" indicate certain mapping, symbols "b" uncertain mapp.; 6) Landslide zone; 7) Widening fractures; 8) Town of S. Lucido; 9) Drillhole site.

Fig. 1. The S. Lucido landslide.

Since no quantitative data obviously exist, it is impossible to establish whether, and to what extent, the circumstances were all that exceptional. The existing evidence does however point to a flooding stage accompanied by a widespread triggering of slope movements.

While the stability of slopes was in these precarious conditions, the climax of one of the most devastating periods of seismic activity ever experienced in Calabria occurred: between February 5th and March 28th 1783 the five principal seismic shocks which hit the region ranged in intensity from IX to XI degrees MCS. These shocks, which primarily involved the area south of the 39th parallel, caused tens of thousands of victims and altered the morphology of the zone to such an extent that several hundreds of lakes were formed along the streams dammed by landslides (Cotecchia *et al.*, 1969; Iaccarino, 1978).

According to Sarcone (1784) ground movements started in the San Lucido area towards the end of December 1782, accelerated during January 1783 and finally became disastrous in February, after the first strong earthquake shock happened.

A committee from the "*Reale Accademia delle Scienze e delle Belle Lettere del Regno di Napoli*" (Royal Academy of Science and Arts of the Kingdom of Naples), which was led by Sarcone himself, sailed for Calabria in the following month of April.

The aim of the expedition was to:

i) check the extent of the damage;

ii) revise the maps of the region;

iii) examine the problem of how to drain the landslide lakes.

The Neapolitan scientists carried out a survey of the San Lucido area on April 12th and 15th 1783. They ascertained that the ground was still in movement and found what they describe in detail in their report as a "total disaster" (Sarcone, 1784). Draughtsmen employed by the committee prepared a very interesting map of the area which, albeit presenting some problems of interpretation, essentially corroborates our own survey (fig.1.b, North Calabria - 1783 and 1905 earthquake-trigged landslides).

It seems worth mentioning two other points which can be deduced from Sarcone's report:

i) the slope movements, in so far as they are described, appear to belong to the class of "*complex landslides*", as defined by Varnes (1978) in the upper part of landslide, followed by "*flows*" type downslopewards. This fully agrees with our data.

ii) accoprding to the description of the sliding materials, at depth it is of the same type as those encountered from 9m to 68m in our exploratory drillholes.

Unfortunately, the language used in the report is not always perfectly clear and some uncertainties arise. It is however clear that the committee members understood the inherent instability of the area. which they attributed to its lithology and geomorphic setting. They also appear to have understood the accelerating effect that an earthquake has on landslides.

Since then, two centuries have passed. Nowadays the area is rather densely populated: in addition to the town itself, it is dotted by many houses and small farms. Moreover, two highways, one important railway (two until few years ago), roads, aqueducts and power lines pass through or close to it. All this accounts for social and economic importance for this ancient and large landslide.

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Martirano (CZ). Landslide reactivated by the 1638 and 1905 earthquakes

Martirano Lombardo is a village settled in the river Savuto valley, north-west Calabria, on the top of hill with steep slopes.
The Geology of the site is rather complex: Upper Miocene sands and sandstones lye unconformably over allochtonous terranes made of Jurassic limestone overlying Paleozoic phyllites and granite belonging to the Alpine Units. Landsliding affets a great part of the hill (Fig. Next page).
This area suffered several historical earthquakes, that destroyed Martirano at least eigth times. The last three events occurred in 1638, 1783 and 1905 (see table bellow).
Landslides were reported in occasion of all three earthquakes, but it is certain which landslides moved only for 1638 and 1905 events.
After the 1905 event the village was split in two: part of people remained in the ancient village, part moved to the new Martirano Lombardo site. The new site is the one visible from the A3 freeway.
From the reports on the 1905 earthquake, it appears that an entire side of the village, named Verdesca, was completely destroyed by a large landslide (table below: "*...una frana fece precipitare in basso un colle...(oriente)...con tutte le abitazioni*" = "a landslide knocked down a hill....(east)...with all the houses,...". The same zone was involved by a landslide during the 1638 earthquake (*sdrucciolando in ruine* = rolling down in a mess).
This landslide is recognisable today. It is shown in Plate 1 (provided separately from this text), where the main geomorphologic features of the old Martirano zona are displayed.
In 1638, liquefaction phenomena occurred in the river Savuto alluvial deposits; these phenomena produced deep depressions and ground craks, from which sulphur-stinking steam was ejected.
Today, the old village is still suffering problems due to mass-movement.

HISTORICAL SEISMICALLY-TRIGGERED LANDSLIDES IN MARTIRANO (CALABRIA)

Date/ I Max	I Site	Transcription	Source
1638-03-27 XI	XI	*"Parte del monte della città si squarciò dall'un fianco, e* ***sdrucciolando in ruine*** *occluse il corso del fiume Bisanto, e dall'altro lato, dove declina in pianure, e fremendo vi scorre il fiume Acherone, hoggi volgarmente Savuto, s'aperse la terra in vaste voragini e in pozzi profondi; esalandone fuori le fetide nebbie di solfo, per oltraggiar il medesimo cielo, e palesar in Calabria un verace Acheronte. Pochi mesi da poi io mi condussi, per officio civile verso quel buon prelato, e per osservar di presenza gli effetti strani del terremoto….."*	D'Orsi (1640) Di Somma (1641)
1783-03-28 XI	VII	Il territorio si trova sconvolto dà movimenti……	Vivenzio (1784)
1905-09-08 X-XI	X	*"La montagna si squarciò in più punti, e ne scaturirono getti di acqua bollente…"* ***"Una frana fece precipitare in basso un colle*** *soprastante da questo lato (oriente) con tutte le abitazioni, sicchè passare era pericoloso tanto da sopra, quanto nel precipizio ch'era di sotto* *[…] mi aiutarono a passare per quella striscia larga 20 cm e per buoni 20 metri che s'era formata sul cedevole terreno, mentre sotto di noi s'apriva un abisso …"* ***"** *Infine quasi tutte le vittime (16 su 17 morti) furono nella frazione Verdesca […] fondate su una molassa miocenica molto franabile…"*	Rizzo (1907) Cotroneo (1905) Mercalli* (1906)

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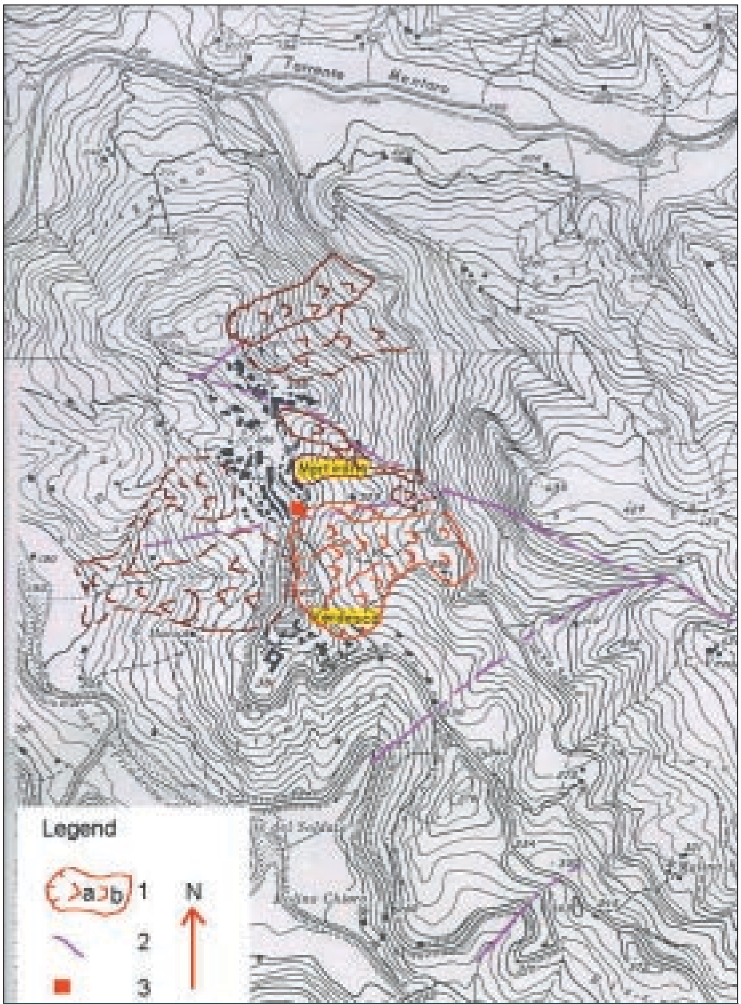

The Martirano area. The landslide in red is the one reported as triggered off by the 1638 and 1905 earthquakes.
Symbols of the Legend: 1 = landslide; a = earth/rock slide; b = earth flow; 3 = stop site.

The seismicity of Martirano area

Prof. I. Guerra[1]

The figures accompanying these notes (Fig. 1) represent at different scales two areas, both centred in Martirano, and the instrumental seismicity resulting from the catalogue of the Geophysical Laboratory at the Calabria University. Seismic foci have been located by using mainly the data acquired by the Regional Seismic Network of Calabria, managed by this laboratory, and those published by all the other scientific institutions operating in the area, first of all the Italian National Institute of Geophysics.
The full square identifies the Martirano's location, while the open circles correspond to the epicentres located since 1986 up to May 2001, selected on the basis of the precision of their focal parameters. Only earthquakes at depths less than 40 km have been retained. The magnitude of the shocks is generally very low. If one accepts the definition of micro-earthquakes as the shocks characterised by magnitude less than 3.0, the two maps are better called microseismicity maps.
Seismic events prior to 1986 have not been taken into account because the quality of the instrumental parameters in the years preceding the eighties in not finer than the corresponding locations by typical macroseismic methods.

The instrumental seismicity recorded in Calabria in the last years shows, particularly on the maps at small scale, several clusters of epicentres that allow to identify some areas characterised at the present by a more intense seismic activity. The correlation between tectonic structures responsible for the great historical earthquakes and the present-day instrumental microseismicity observed on a span shorter than twenty years is not an obvious matter. It results particularly questionable in Calabria, where the main seismogenic structures are quiescent since nearly a whole century, after the paroxysmal activity in the years 1783-1908. It is supported however by the observation that the areas with denser seismic clusters coincide with those where the amount of the released seismic energy more frequently in the last years has exceeded the perceptibility threshold.
With these limitations, the present-day micro-seismicity maps indicate that no cluster of the presently more intense seismic activity is located near Martirano. This site falls in a relative seismicity minimum, so that the hypothesis is strengthened that the effects of the earthquakes of March 27, 1638 and September 8, 1905 had their severity enhanced there by factors related to local geological-geomorphological conditions.

This statement is supported by the results of the analysis of the macroseismic filed carried out at the Geophysical Laboratory. The work involved the fitting of the macroseismic field resulting from contemporary historical chronicles to the well-known Blake's law on the attenuation of the seismic intensities. In spite of the simplicity of the numerical model and the contrasting complexity of the geological structures, the agreement between experimental data and theoretical model came out very strict. The intensity X of the MCS scale assigned to Martirano in this study is greater than the theoretical value corresponding to the best-fit model, so that this datum represents a positive anomaly of the macroseismic field. Of course, the anomaly could be much more pronounced repeating the same analytical procedure by using the intensity values assigned in the Catalogue of Strong Earthquakes in Italy by Boschi et al. (1995): these authors infact assign to Martirano the degree XI of macroseimic intensity.

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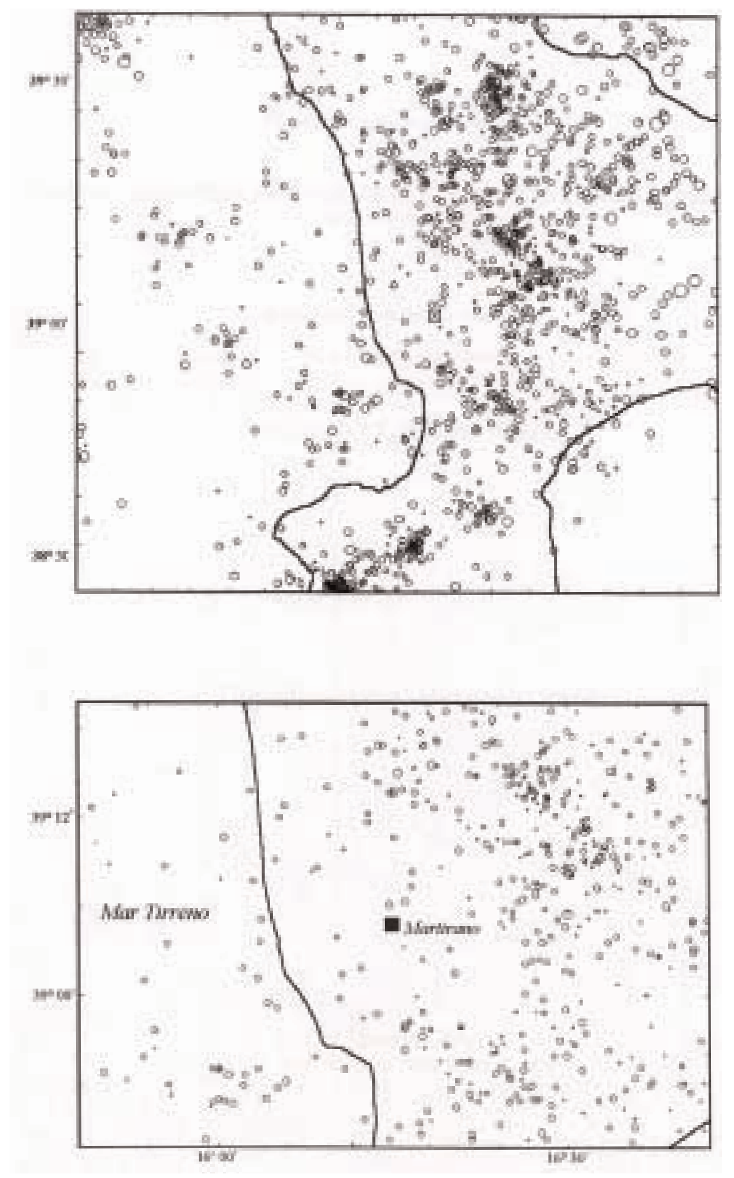

Fig. 1. Instrumental seismicity resulting from the catalogue of the geophisical Laboratory at the Calabria University

The same algorithms have been successively applied to the macroseismic observations of the 1905 earthquake. On the basis of the results obtained in this case, it is impossible to fit the data to the Blake's and other common mathematical models. It is noteworthy that the unusual space distribution of the macroseismic effects was already noticed by contemporary authors like Baratta and Mercalli. However, the strong positive anomaly of the intensity at Martirano in this case is evident in the same observational data. Again according to Boschi et al. (1995) intensities of the degree X or greater were observed in fifteen sites: fourteen are located on the Mt Poro promontory; the last is just Martirano, well away from all the others.

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