

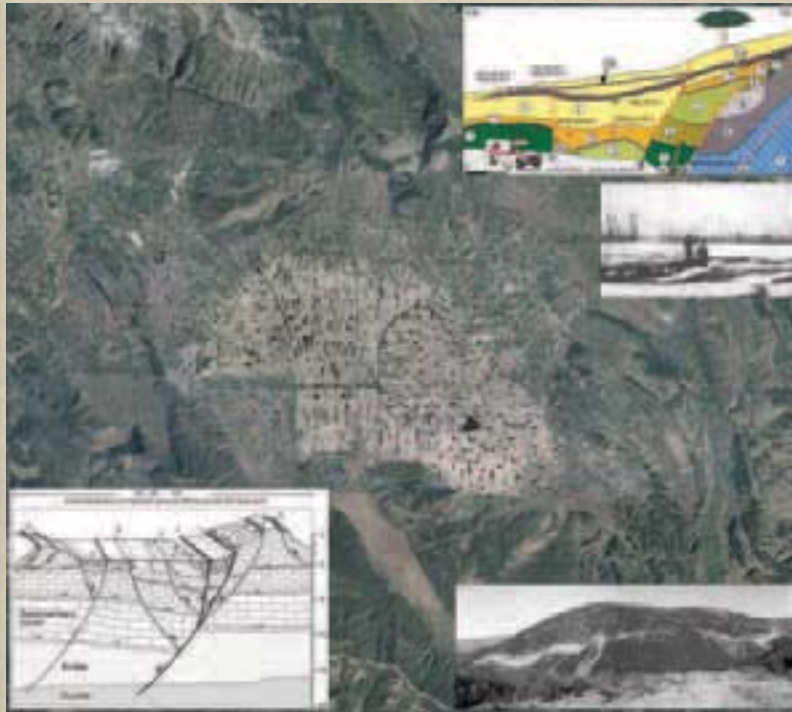
Volume n° 6 - from P55 to PW06



Field Trip Guide Book - P65

**32nd INTERNATIONAL
GEOLOGICAL CONGRESS**

**BASIN AND RANGE IN THE
CENTRAL AND SOUTHERN
APENNINES**



Leader: A.M. Blumetti

*Associate Leaders: A.M. Michetti, F. Dramis,
L. Guerrieri, B. Gentili, E. Tondi*

Florence - Italy
August 20-28, 2004

Post-Congress

P65

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Front Cover:

*Aerial photograph of the Fucino Basin composed with
images of historical, Holocene and Quaternary evidences
of surface faulting*

Leader: A.M. Blumetti

Associate Leaders: A.M. Michetti, F. Dramis, L. Guerrieri, B. Gentili, E. Tondi

Introduction

Blumetti, Guerrieri, Michetti & Serva

Main focus

This field trip will review the geological evidence of post-Miocene continental rifting in the Central and Southern Apennines, and in particular, the recent behaviour (Late Pleistocene to Holocene) of the system of active capable normal faults, as showed by the extensive amount of stratigraphic, geomorphic and paleoseismic data gathered in the last decade.

Special emphasis will be given to i) Quaternary, and especially Holocene, tectonics and surface processes interactions; ii) integrated observations of the recent landscape evolution; iii) the related understanding of natural hazards (including ground motion, ground rupture, and large landslides) and risk mitigation strategies.

Thus, the main focus of the field trip is the relation between coseismic ground effects of historical earthquakes, and paleoseismic events along capable faults, and the typical features of the associated fault-generated mountain fronts and extensional basins. With this aim, the trip will look at the evidence of large gravity slope deformations too. In fact these features are often the result of landscape evolution influenced by surface faulting earthquakes.

Itinerary

The trip will start in Umbria, in the area affected by a recent seismic sequence (26.09.1997, Mw=5.6 and 6.0, Colfiorito earthquake), that produced huge damage to historical buildings and monuments (i.e. San Francesco Monastery in Assisi, stop 1.1). The ground effects (surface faulting, landslides) of the 1997 seismic event will be observed in the context of the Colfiorito basin evolution and related fault-generated mountain fronts (stop 1.2). The last part of the day will be dedicated to the observation of block sliding phenomena and sackungen on the SW slope of Monte Fema (Sibillini Mts., stop 1.3).

On the Adriatic Sea, in the Ancona landslide (stop 2.1) we will speculate about the relationships between the surface effects of coseismic fracturing and the very deep-seated gravitational processes. Then, we will visit Montelparo village (stop 2.2), which has been affected by a large translational landslide that reactivated during the 1703 Norcia earthquake. After lunch, the field trip will move to the south, to Abruzzo, to analyze the Quaternary evolution of fault slopes

bordering the L'Aquila basin and Upper Aterno valley and the implications, in terms of active tectonics and seismic hazard, deriving from paleoseismological analyses along Upper Pleistocene to Holocene fault escarpments (stop 2.3).

The third day will be located in the inner sector of the Central Apennines, between Latium and Abruzzi. A first stop will be focussed on the role of the Fiamignano fault and the associated large-scale gravity slope deformations in the landscape evolution of the Salto river valley (stop 3.1). Then, most of the time will be dedicated to the Fucino basin (stop 3.2), that is one of the larger intermountain basins in the Apennines, characterized by high rates of deformation along the bordering normal faults. Long term geological effects will be integrated, with the observation of surface faulting features related to modern (i.e. 13.01.1915, M=7.0 Fucino earthquake), historical, and paleoseismic events. The last part of the field trip (days 4 and 5) will develop in Campania, in the epicentral area of the 23.11.1980, M=6.9, Irpinia earthquake. A large variety of coseismic ground effects (surface faulting, gravity slope deformations, and tectonic-karstic basins) will be examined in detail. In fact, these features are the typical seismic landforms in this sector of the Southern Apennines.

Field map references

In this guidebook we will refer to the following geological maps of the Carta Geologica d'Italia at the 1:100,000 scale:

- Day 1: Sheets n°: 123 "Assisi", 131 "Foligno" and 132 "Norcia";
- Day 2: 118 "Ancona" and 139 "L'Aquila";
- Day 3: 145 "Avezzano" and 146 "Sulmona";
- Day 4: 186 "S. Angelo dei Lombardi" and 187 "Melfi".
- Day 5: 174 "Ariano Irpino"

Geological setting of the Apennines

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 (Malinverno and Ryan, 1986; Doglioni et al., 1996; D'Agostino

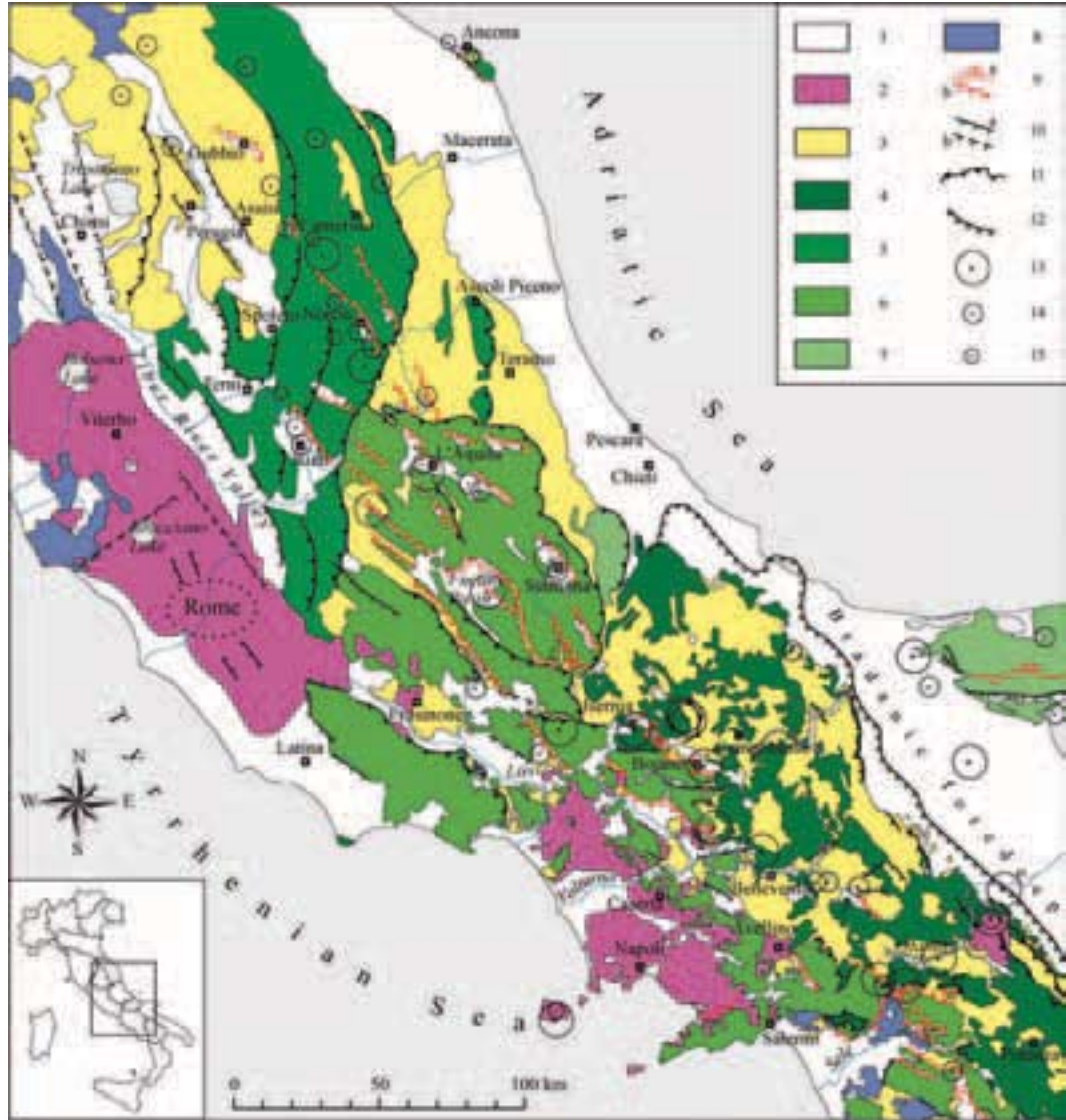


Figure 1 - Schematic geologic map of the Central Apennines (mainly after Mostardini and Merlini, 1986; Bonardi et al., 1988; Bigi et al., 1990, Michetti et al., 2000b). Legend: 1) Marine and continental sedimentary deposits along the Tyrrhenian margin, inside the tectonic depressions of the Apennine and in the Bradanic foredeep (Upper Pliocene - Quaternary); 2) Volcanic deposits related to the activity of the Tyrrhenian margin (Upper Pliocene - Quaternary); 3) Synorogenic deposits of external foredeep and piggy - back environments (Tortonian - Middle Pliocene); 4) Umbria-Marche pelagic and transitional limestones and marls (Mesozoic-Cenozoic); 5) Carbonate-siliceous-marly deposits of the Molise and Lagonegro units (Upper Triassic - Miocene); 6) Carbonate sequences of the Apennine platform structural unit (Upper Triassic - Miocene); 7) Carbonate platform sequences of the Apulian Foreland (Upper Jurassic - Upper Miocene); 8) Slightly metamorphosed and metamorphosed basinal deposits and piggy - back deposits of the Internal units (Jurassic - Lower Miocene); 9) Normal fault, capable of producing strong surface faulting earthquakes ($M_e > 6.0$), a) strike-slip, b) normal; 10) Normal fault a) outcropping, b) buried; 11) Main overthrust; 12) Boundary of the allochthonous Apennine units; 13) Location of earthquakes with macroseismic equivalent magnitude $M > 6.5$; 14) Location of earthquakes with macroseismic equivalent magnitude $6.0 < M_e < 6.5$; 15) Location of earthquakes with macroseismic equivalent magnitude $M_e 5.5 < M_e < 6.0$ (after Boschi et al., 1995; Camassi & Stucchi, 1997).

et al., 2001).

The Apennines are a NW-SE-trending Neogene and Quaternary fold and thrust belt (Figure 1; Mostardini and Merlini, 1986; Patacca and Scandone, 1989; Doglioni et al., 1996). Since the Late Pliocene, following the opening of the Tyrrhenian Sea (Cinque et al., 1991; Patacca et al., 1990), extensional tectonics progressively shifting to the east, have determined a number of deep tectonic basins, hosting mainly marine deposits and volcanics on the Tyrrhenian side. In the inner sectors of the Apennines, normal-fault-bounded intermountain depressions have developed, typically as northwesterly elongated half and full graben, up to several tens of kilometres long, bounded by steep limestone faults-generated-mountain front, and hosting a thick Quaternary continental sedimentation (e.g. Terni, Rieti, Norcia L'Aquila, Fucino, Sulmona, Isernia, and Bojano basins). The master faults dip prevalently southwest. However NE-dipping master faults are also present, for instance those controlling the evolutions of the Leonessa, Bojano, and Irpinia areas.

Basin and Range in the Apennines

Studies on active tectonics and paleoseismicity confirm that the present-day tectonic setting of the Apennines is guided by a system of Quaternary normal faults, which determine a still immature basin-and-range morphology (Michetti et al., 2000a; Serva et al., 2002), and are responsible for frequent moderate-to-strong earthquakes, with typical shallow crustal hypocentral depths (Amato et al., 1997).

Earthquake magnitude of the strongest events is in the order of M 7, with recurrence intervals ranging from a few hundreds, to a few thousands of years. However, it is important to note that continuing Quaternary to present compression and folding is well documented on the Adriatic side of the Central Apennines (e.g. Patacca et al. 1992), about 20 km east of the belt of extensional basins, associated with low to moderate seismicity.

This landscape is comparable to the immature topography typical of the Basin and Range Province of the Western United States, in terms of earthquake surface faulting, mountain building, slopes evolution, and gravity deformation. Similarly, the Apennine intermountain basins show a great variety of landscape features which appear to be directly connected with the magnitude of the associated "typical" earthquake (the seismic event due to the typical movement along the active faults bordering the basin). Detailed

geological studies on these basins suggest that the growth of normal fault segments is controlled by repeated earthquakes, causing surface faulting.

Ground effect analyses produced by moderate to large normal faulting earthquakes identify a lower magnitude threshold for the occurrence of surface faulting, that is, between 5.5 and 6.0, in good agreement with worldwide relations between surface faulting features and earthquake magnitude (Wells & Coppersmith, 1994; Mohammadioun and Serva, 2001).

The recurrence interval of surface faulting events ranges from a few hundred to a few thousand years, and normal fault slip rates are on the order of 0.1 to 1.0 mm/yr (Galadini & Galli, 2000; Michetti et al., 1996; Michetti et al., 1997; Pantosti et al., 1993). Short term (Holocene) and long term slip rates (Quaternary), relative to the capable faults mentioned, are in good agreement, although changes due to fault growth and interaction during the Quaternary may in any case be hypothesized (Roberts et al., 2002; 2004; Roberts and Michetti, 2004). According to Slemmons & de Polo (1986), slip rate data can also be related to the "typical earthquake" magnitude associated with a capable fault.

Moreover a comparison of the earthquake surface rupture (trace and length of the coseismic fault scarps) and the geomorphology (size of the Quaternary tectonic basin), corroborates the interpretation of the intermountain basins as the result of repeated strong earthquakes over a geological time interval.

In Figure 2 we propose a model illustrating, for two magnitude thresholds, the amount of surface faulting and the nature and distribution of coseismic effects (primary and secondary), generally associated with the typical earthquakes. For magnitudes $M = 6.0$ (A) and $M = 7.0$ (B), the typical rupture length should range approximately between a few km, to tens of km. The coseismic displacement should range from a few cm (A) to tens of cm, up to one meter (B).

The model also accounts for the secondary ground effects, such as landslides, sackungen, and liquefaction phenomena, which are associated with the seismic events.

Local factors can mask the shape and the size of the basin. These factors are essentially i) the rate of erosional and sedimentary processes, controlled by lithology, climate and shape of the drainage network; and ii) pre-existent geological structures, such as thrusts and folds related to orogenic compressive stresses.

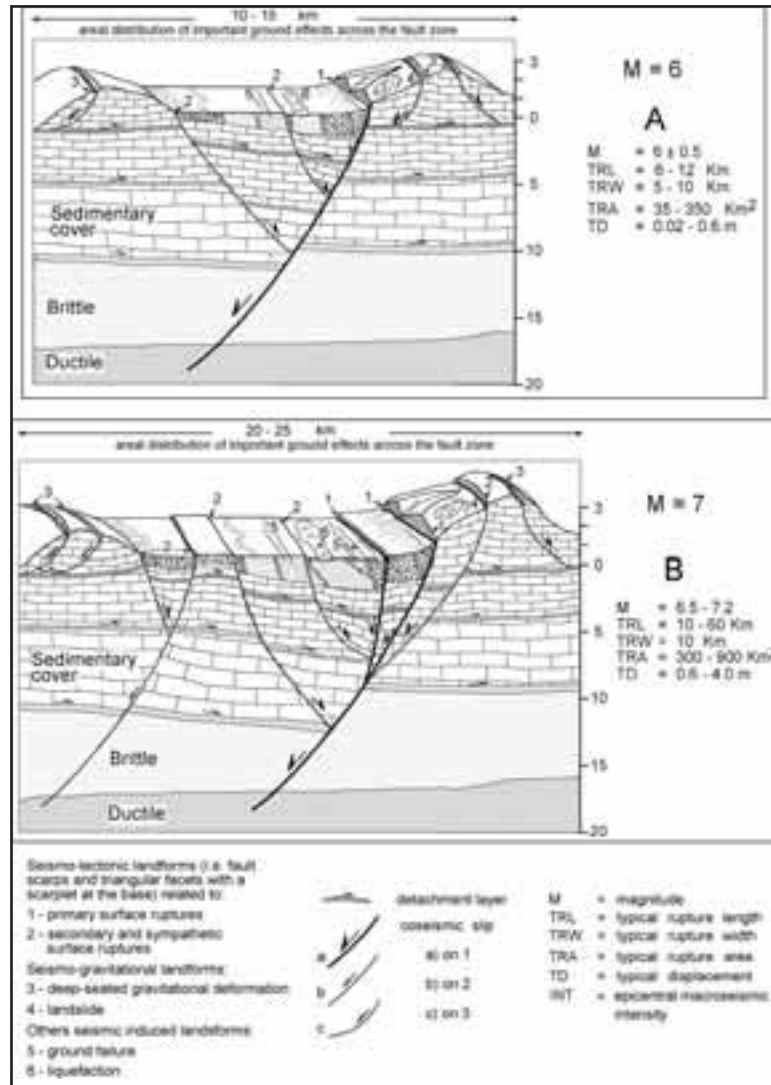


Figure 2 - Schematic block-diagram of two Quaternary intermountain basins associated with $M = 6$ (A) and $M = 7$ (B) characteristic earthquakes: the picture illustrates the typical tectonic and gravitational-tectonic coseismic effects. On the right, the characteristic values of surface-faulting parameters (rupture length, rupture width, rupture area, vertical displacement) are reported.

Moreover, it is important to outline that this model does not represent older intermountain basins, such as Pliocene ones, located in the western part of the Apennines (close to the Tyrrhenian coast). In fact these basins, although formed in the same way, are at present characterized by i) smoother landforms; ii)

lower seismicity; iii) and thinner crust, as evidenced by high heat flow and volcanic phenomena.

Field itinerary

DAY 1

Stop 1.1:

Assisi, Basilica inferiore. Damage due to the Colfiorito, 1997, Central Italy earthquake.

Blumetti A.M. and Marsan P.

Introduction

This stop will illustrate the damage which occurred in the Holy Monastery of Assisi during the 1997 Colfiorito seismic sequence.

During the years 1994-1998, the National Seismic Survey ("Servizio Sismico Nazionale", or SSN in the following text), has developed a static and dynamic control system, concentrated on a sector of the Holy Monastery in Assisi. The aim of this operation was to study the behavior of structures during small and moderate size earthquakes, since an important increase in

damage was observed connected with even small-scale seismic events. The risk of serious damage after large-scale earthquakes, although characterized by comparatively far epicentral distances, was also taken into consideration.

To this end, a dynamic control system was implemented, formed by Kinematics SSR-1 accelerometers, furnished with FBA-23 and FBA-II devices, and a 0.10 g full scale, in order to reveal the largest number of seismic events. A Kinematics SSA-2 accelerometer with a 1g full scale was also installed to reveal events that eventually would be able to saturate the SSR-1. The events that occurred

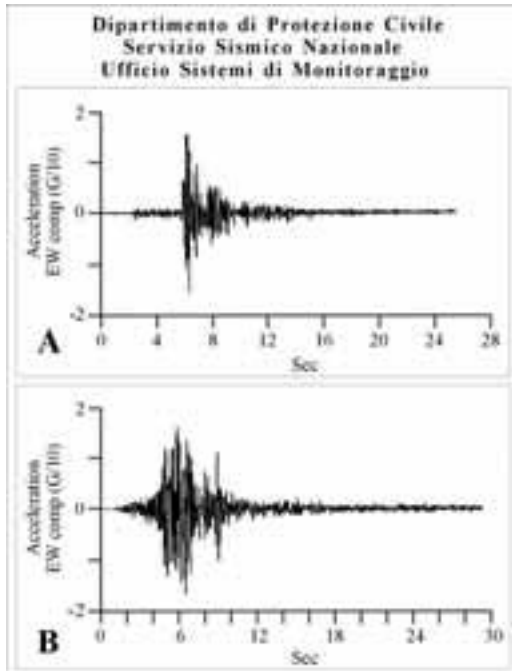


Figure 3 - Accelerograms of the two earthquakes, registered in the Assisi Holy Monastery, which occurred in the Colfiorito area, on September 26, 1997, A) at 2.33 a.m. and B) at 11.33 a.m. (local time).



Figure 4 - Firemen working among the ruins after the fall of the ceiling upon the major altar.

on 26 September 1997, and those that followed, have unfortunately proved in the most pessimistic terms all previous hypotheses, and have caused

extensive damage on the entire plan of the structures constituting the Holy Monastery.

Damage account

On September 26, 1997, at 00:33 and 09:40 (GMT) two moderate earthquakes ($M_w = 5.7$ and $M_w = 6.0$) struck several historical towns and monuments in the Umbria - Marche region of central Italy, causing the deaths of 12 persons, injuries to 140 people, and leaving about 80,000 people homeless.

After the event occurred at 2.30 a.m. (local time), a SSN technical team reached the Holy Monastery, on the request of the Civil Protection Department, to check the importance of the damage that occurred due to a $M_w 5.7$ seismic event that occurred at an epicentral distance of about 20 km, and to look at the records that had been obtained through the monitoring network installed on the western part of the Monastery. Both the 9/26 two main events records are reported in Figure 3. While a technician was going towards the acquiring stations for the discharge of data, another two were reaching the inner part of the San Francesco Upper Basilica, where some technicians were already executing the necessary investigations on the spot, and where some cracks on the vaults were noticed for having produced some crumbling of the plaster from the frescoes. The second event (local time 11.40 a.m., $M_w = 6.0$), had a smaller epicentral distance (about 15 km).

This second earthquake occurred while the technicians were working in the Upper Basilica's inner area, in the presence of the friars themselves. The dramatic effects, that reached all over the world through television images, unfortunately caused four fatalities and some injured people (Figure 4), as a consequence of the fall

of two sectors of the ceiling, located on the main access, and on the "altare maggiore" (major altar) (Fig 5).

The original roof, with its wooden truss, had been replaced by arches in bricks which hung over beams in



Figure 5 - The fallen sail upon the major altar.

concrete), some ten years before. Nevertheless, there is no evidence to suggest that this replacing operation of the roof was connected with the sudden sinking that involved the elements located symmetrically on the extremes of the basilica's longer side. Another element of the basilica that became famous through the damage that occurred, is the southern tympanum, whose stability was compromised by the rapid succession of seismic events that followed, although these were not as strong as the main shaking that occurred on 9/26. Had the tympanum collapsed completely, it would have involved the damage of particularly valuable frescoes, such as the "Cristo del Cimabue", located in the Upper Basilica just lying below (point A in Figure 6). However, the emergency counter measures, realised a few hours

before the 10/14/1997 shaking, prevented this from occurring. The bell tower was damaged too, in the cell where the bells hang, after the main event occurred on 9/26, and the key vault face to south was cracked, while the shaking that followed, which continued up to 10/14, damaged in part the central key vault and the north key vault, but in a more serious way. Possible falls have been avoided thanks to the efficacy both of the chaining of principle elements, and the south-eastern comer-edge pillar covering. Few structural elements haven't suffered heavy damage. In fact, towards the western part, we can note that both the southern side, in the upper cells, and in "the dining-room" below, just near the vaults (point B in Figure 6), and the north side where the museum was located, have been damaged. The inside colonnade, near the cistern at the back of the Basilica, held out thanks to the chaining surrounding it. Going westwards, we find the relatively younger Papal Palace in the Holy Monastery structure, just near the S. Geronzio courtyard. Here, the damaging level is even linked to the serious renewing of the pre-existing cracks, that became much wider, and to the fact that the northern side of the Papal Hall roof collapsed because of the corresponding partial collapse of the tympanum. The partial collapse of the tympanum also strongly deformed some bedroom roofs. As described, some pre-existing cracks became much wider, and even the Papal Hall floor was damaged. Again, in the Papal Hall, in order to have a static control of possible deformations and movements, in addition to the dynamic sensors, automatic devices to measure further deformation of pre-existing



Figure 6 - Southern view of the Assisi Holy Monastery.

cracks, and inclinometers to check verticality, had been installed too. The strain gauges were able to measure deformations on the main crack, which corresponds to the S. Geronzio courtyard, up to the first seismic event which occurred on 9/4, and were able to show a widening of 0.05 cm. This crack, after the shaking occurred on 9/26/1997 at 2.30 a.m., suddenly widened to around 1 cm, to later go off the scale – that is, more than 2 cm – with the 11.40 a.m. main event.

The serious level of damage that affected the Holy Monastery at Assisi as a consequence of the seismic events, was caused mainly by the unexpected strength of the principle shaking event that took place at 11.40 a.m. on 9/26, but even by all the other events that took place in the days that followed, which were associated with a rate of magnitude corresponding to around 5. Another important aspect to be considered, is that the most important damage in Assisi was concentrated in the Holy Monastery and in the Upper Basilica. This fact can be attributed in part to a local situation where the Holy Monastery's foundation rests on calcareous rock on the northern side, while on the southern one, it is built also on sloped deposits and artificial filling terrain. This situation is well illustrated in the large square behind the Basilica, where half of the square, the valley side, shows large cracks caused by the sudden settlement that affected the very poorly-coherent filling materials on which this side is built.

Stop 1.2:

The Colfiorito Seismic Zone

Tondi E. & Michetti A.M.

The fault array affecting the mountainous area extending east and south of the Colfiorito village is

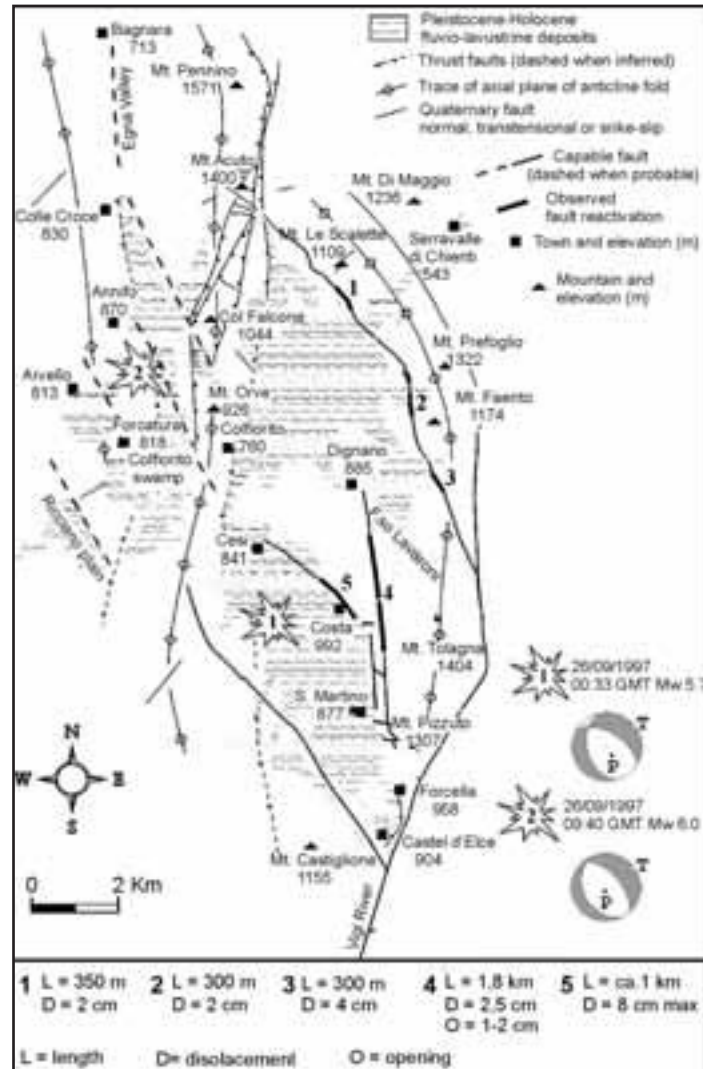


Figure 7 - Fault data, focal mechanism solutions, and ground deformation within the mesoseismic area of the Colfiorito earthquake sequence (data from Amato et al., 1998b; Cello et al., 1997, 1998b).

made up of several faults which cut through bedrock units and Upper Pleistocene deposits (Figure 7).

Faults belonging to this array display a general NNW-SSE trend, and their recent activity has been studied by several authors, who also discuss their potential for coseismic ground displacement (Cello et al., 1997; Tondi et al., 1997, and references therein). Associated with this array is also one of the best studied Quaternary Apennine intramontane basins: the Colfiorito basin, which is characterized

by an array of nested tectonic depressions, filled with lacustrine and alluvial deposits (Figure 7). Mammalian remains in the lake sediments suggest that the Colfiorito basin developed since, at least, the end of the Early Pleistocene (Ficcarelli and Silvestrini, 1991; Figure 8).

area (some 15 km south of Colfiorito), and a roughly 45 km deep earthquake ($M_s = 5$) which occurred 25 km north of Colfiorito. Seismological data of the 1997 mainshocks suggest almost pure normal faulting on northwest-southeast trending planes, dipping 40° - 50° to the southwest (Figure 7; Amato et al., 1998).

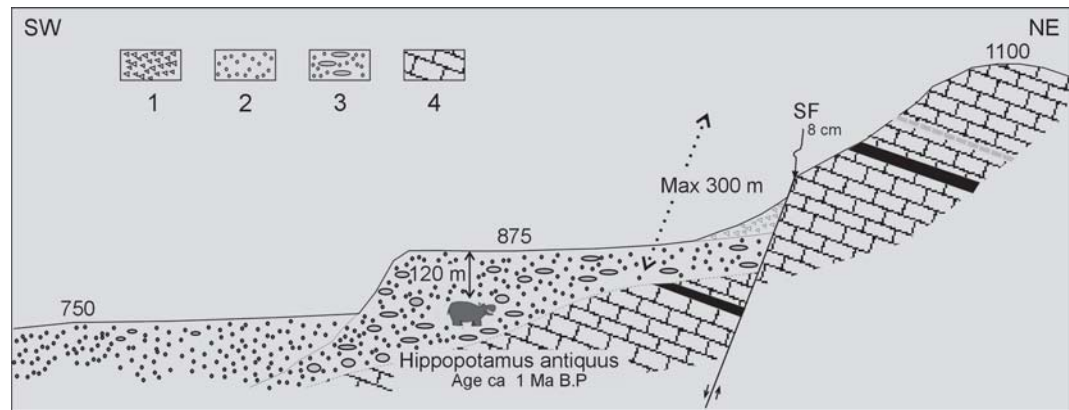


Figure 8 - Schematic cross-section of the Colfiorito basin stratigraphic and tectonic relations, as exposed in the Collecorti area. The offset is lower than 250 - 300 m in the Mesozoic limestone bedrock; therefore, the Quaternary slip-rate must be lower than 0.3 mm/yr. Legend: 1. Holocene slope deposits; 2. Late Pleistocene to Holocene lake and marsh deposits; 3. Lower to Mid Pleistocene fluvial and lake deposits; 4. Mesozoic pelagic limestone sequence; SF. Surface faulting along the Costa fault, produced by the Sept. 26, 1997 earthquakes.

Analysis of kinematic data on fault planes in both bedrock units and continental deposits (striae and shear fibers on slickenside surfaces), indicates a quite consistent orientation of the slip vector from differently oriented fault surfaces, leading to an almost pure, left-lateral, strike-slip motion along the north-south trending planes, and to transtensional slip (with an increasing sinistral component of motion), along the NW-SE to NNW-SSE oriented ones.

On September 26, 1997, at 00:33 and 09.40 (GMT) two moderate earthquakes ($M_w = 5.7$ and $M_w = 6.0$) had their epicenters in the Colfiorito area. The epicenter (Figure 7) of the first shock was located midway between Cesi and Costa, whereas the second shock occurred south of Annifo (about 6 km to the NNW of the first event). Both these events occurred within the Colfiorito basin. A major foreshock ($M_s = 4.8$) was also recorded in the Cesi area by the local seismic network of the University of Camerino on September 4, whereas many aftershocks ($M_s < 4.7$) affected the whole epicentral area over the months that followed. The two major earthquakes are part of a seismic sequence which also include a mainshock recorded on October 14 ($M_w = 5.7$) in the Sellano

The 1997 seismic crisis can be related to other historical and paleoseismological events of similar size which have affected the Umbria-Marche region over the last millennium (Boschi et al., 1997; Vittori et al., 2000). According to Italian seismic catalogues, the strongest earthquake that has ever occurred with an epicentral area close to that of the 1997 events, occurred in 1279 ($I = X$ MCS).

Most of the fault segments mapped as capable faults (as according to the IAEA, 1991) within the epicentral area of the September 26, 1997 earthquakes (Cello et al., 1997; Tondi et al., 1997), typically mark the interface between bedrock and slope deposits occurring at the base of the range fronts bordering the Colfiorito basin. Observed surface ruptures due to the 1997 seismic sequence were mapped along three main faults (refer to Figure 7; Cello et al., 1998b; Cello et al., 2000; Vittori et al., 2000): that is, 1) the Colfiorito border fault, 2) the Cesi - Costa fault, and 3), the Dignano - Forcella fault.

Stop 1.2.1:

The Colfiorito border fault

The Colfiorito border fault (Figure 9 and 10) is

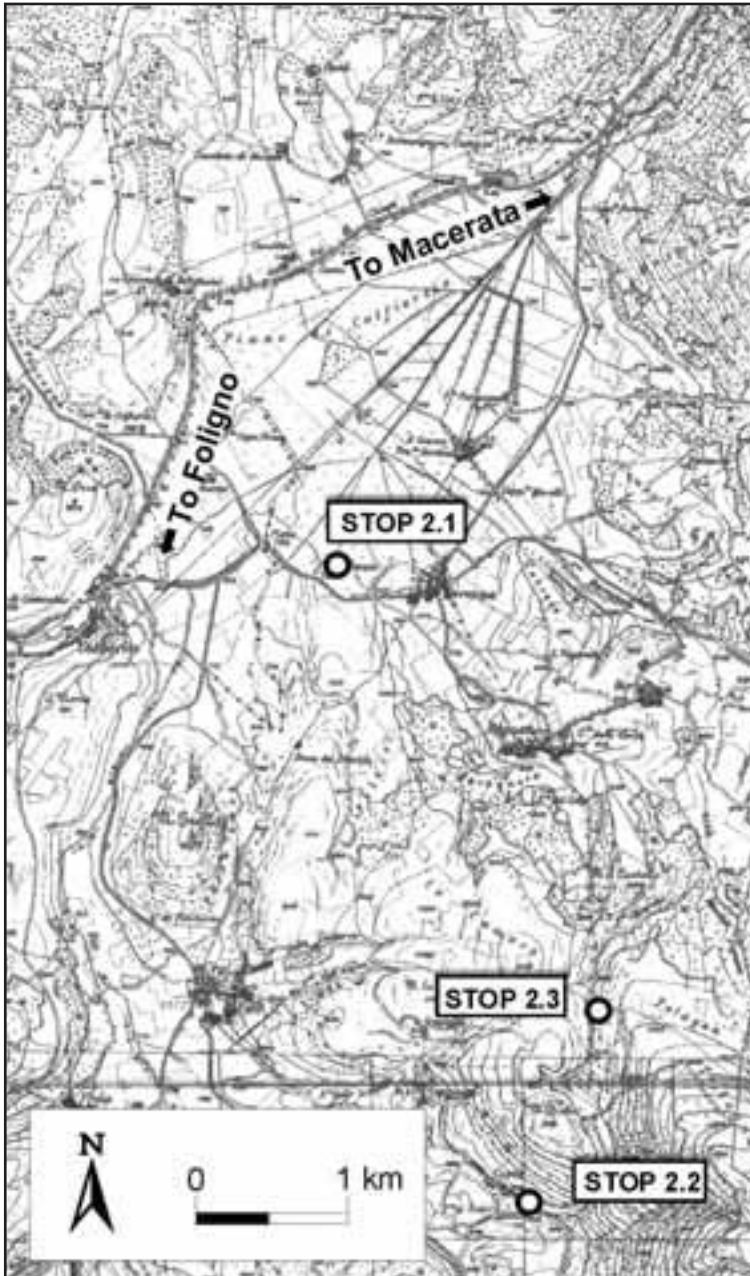


Figure 9 - Location of the stops in the Colfiorito basin

exposed at the surface for a length of about 7 km, and is characterized by differently-oriented segments (refer to Figure 7). At Mt. Le Scalette, the fault trends from N130° to N150°, and is characterized by a 2 to 5 m high slickenside-bedrock fault scarp. The stratigraphic and geomorphological offset

measured across the Colfiorito border fault is 150 to 200 m (Figure 11). The fault displaces Middle Pleistocene and recent lake sediments, damming the drainage of the Colfiorito basin into the Chienti River valley (Centamore et al., 1978).

Surface fault reactivation associated with the 1997 seismic sequence, produced free-faces 2 to 4 cm high over a length of ca. 550 m (Figure 12).

Stop 1.2.2:

The Cesi-Costa fault

The Cesi-Costa fault (Figs. 7 and 9) trends roughly N130°. It is characterized by the occurrence of triangular facets, minor perched valleys, and fresh-looking slickensides, dipping 50° to 70° SW. The stratigraphic offset measured across the Cesi-Costa fault is in the range of 100 to 150 m (Figure 11). In the proximity of the Costa village (Figure 13), fault reactivation, following the 00:33 GMT mainshock of September 26, is recorded by a newly-generated, continuous, 7-8 cm high, free-face (Figure 14). The estimated total length of the ruptured segment along the Cesi-Costa fault is about 1 km (Figure 12).

Stop 1.2.3:

The Dignano-Forcella fault

The Dignano-Forcella fault (Figs. 7 and 9) trends from N 160° to roughly N-S, and is characterized by limestone slickensides dipping 60°-70° SW. The slickenside-bedrock scarp is 1 to 1.5 m high, and a less than 20 m thick slope deposit sequence, including different generations of well-bedded periglacial



Figure 10 - The Colfiorito border fault. View of the Colfiorito basin, and of the mountain front which was generated by the border fault; the arrow indicates the Chienti River valley.

breccia, is juxtaposed against the fault plane. The maximum stratigraphic offset across fault 2) is about 100 m (Figure 11); locally, old cemented breccia is also offset. Fault reactivation following the 1997 seismic sequence along the Dignano-Forcella fault is characterized by the occurrence of a continuous, 2.5 - 3.0 cm high, free-face (Figure 15) exposed at Fosso Lavaroni over a length of about 200 m (Figure 12).

The 1997 earthquake sequence that occurred in central Italy offered a unique opportunity to study the distribution of coseismic surface faulting effects related to moderate-sized seismic events. Fault reactivations associated with this seismic sequence have been surveyed at several sites along mapped capable faults within the Colfiorito basin area. Coseismic surface displacements occur along pre-existing faults, and these faults are responsible for

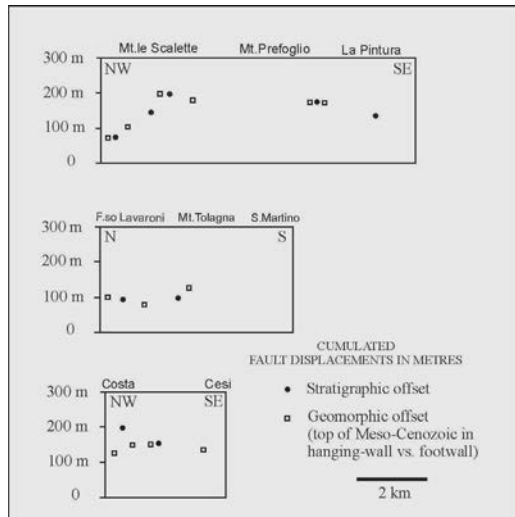


Figure 11 - Comparison of stratigraphic and geomorphological displacements associated with the capable faults of the Colfiorito area. Sections are parallel to the fault traces shown in Figure 12 (Cello et al., 2000).

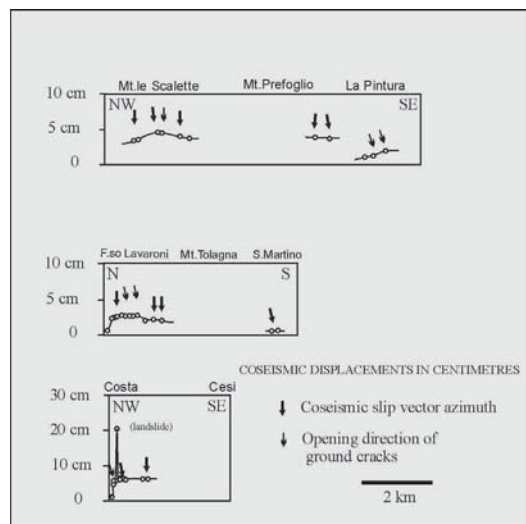


Figure 12 - Coseismic displacements and coseismic slip-vector azimuth recorded along the capable faults of the Colfiorito basin area after the two mainshocks of the 26 September, 1997, earthquake sequence (Cello et al., 2000).



Figure 13 - Costa village: Falling houses during the main shock ($M_w=6.0$) of the Colfiorito earthquake. 9:40 GMT, 26th September 1997.

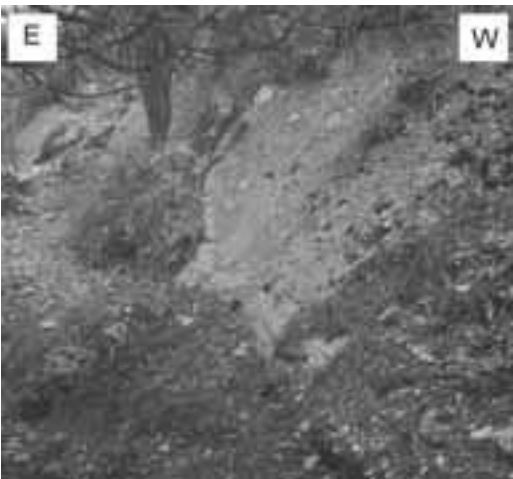


Figure 14 - Slickenside, with basal free-face, along the Costa bedrock fault scarp; photo taken on September 27. Fig. 15 - Dignano - Forcella fault. Site: Fosso Lavaroni. View of the bedrock fault scarp at Fosso Lavaroni. Note that free-face height is remarkably uniform (2.5 cm).



Figure 15 - Dignano - Forcella fault. Site: Fosso Lavaroni. View of the bedrock fault scarp at Fosso Lavaroni. Note that free-face height is remarkably uniform (2.5 cm).

the recent evolution of the area, and for the growth of fresh limestone scarps and slickensides characterized by geomorphic features which are unequivocally related to paleoseismic surface faulting. The observed offsets (a few centimeters), are remarkably constant over tens or hundreds of meters, and fit quite well with the empirical relations between fault displacements and lengths derived by Wells and Coppersmith (1994).

Stop 1.3: Deep-seated gravitational slope deformations in the Umbria-Marche Apennines: Mt. Fema.

C. Bisci, B. Gentili & G. Pambianchi

Deep-seated gravitational deformations are particularly frequent in the axial portion of the Umbria-Marche Apennines belt, where the tectonic uplift was more intense, giving rise to steep and long slopes. Among them, sackungs (rock flows) and lateral spreads are more frequently found.

One well-known example of this type of gravitational movement is the one affecting Mt. Fema (Dramis et al., 1994; 1995), in the Natural Reserve of Torricchio.

Mt. Fema is a moderate relief (1573 m a.s.l.), located to the NW of the ancient village of Visso, close to the boundary between the Marche and Abruzzi Regions. It is elongated in a ca. NNW-SSE direction, and is made up of stratified, marly limestone with intercalated, thin marly levels (Scaglia rosata formation). From a structural point of view, it represents an east-verging overthrust, bordered to the west by minor normal faults (Figure 16).

Its western slope, remodelled on a fault scarp, shows strata more or less regularly dipping to the west (i.e. in the same direction of the slope); stratigraphic and tectonic conditions are therefore favourable to mass movement activation. Along its mid and high portion, wide open cracks are present (Figure 17).

They are the most visible testimonies of a deep-seated deformation ("sackung", or rock flow, Mahr & Nemčok, 1977; Bisci et al., 1996)

which involved the whole calcareous slope. Oral witnesses indicate the reactivation of some of these fractures, during the 1979 earthquake (VIII MCS), whose epicenter was located only a few ten kilometers away. Moreover, counterslope steps, scarps and wide trenches, are also present (Figure 18).

The visit will start from the top of Mt. Fema (not normally accessible to private visitors, since the road is blocked by a bar at the base of the slope, at the

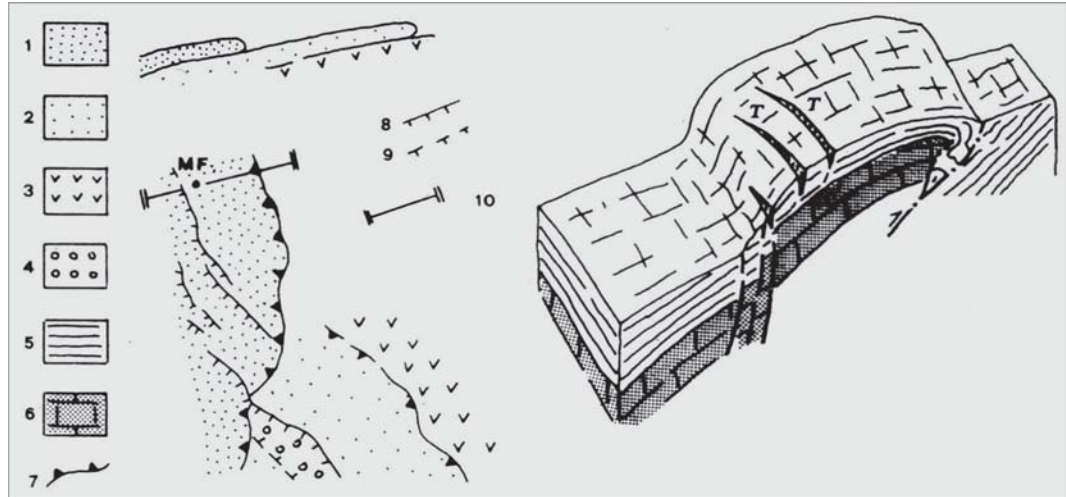


Figure 16 - Geological and geomorphological sketch of the Mt. Fema area (from Dramis et al., 1994). Legend: 1) upper overthrust units; 2) lower overthrust units; 3) foredeep units; 4) continental deposits; 5) stratified limestone, marly limestone and marl; 6) massive limestone; 7) thrust; 8) main normal fault; 9) normal fault; 10) location of the cross section; T - trench; MF - Mt. Fema.

boundary of the Natural Reserve of Torricchio).

A short walk (ca. 20-30 minutes) will lead to the area where the open cracks are present, and most of the features are clearly visible.



Figure 17 - Open crack along the western slope of Mt. Fema (from Dramis et al., 1995).



Figure 18 - Counterslope and trench along the western slope of Mt. Fema.

DAY 2

Stop 2.1:

The Ancona Landslide

Dramis F., Gentili B. and Pambianchi G.

Along the Adriatic coast, trenches parallel with the coastline, locally bordered by fractures and steps lowering seawards, have been found in wave-cut cliffs, presently inactive and separated by the sea through narrow coastal belts. These landforms are frequent on the north-eastern slopes of compressive structures with an Adriatic vergence, made up of lower-middle Pleistocene clayey-sandy-conglomeratic terrains (Cantalamessa et al, 1987). These structures are still active, as testified by the hypocentral mechanisms of earthquakes which have recently affected the area (Gasparini et al, 1985; Riguzzi et al, 1989).

Figure 19 - Aerial view of the December 1982 Ancona landslide.

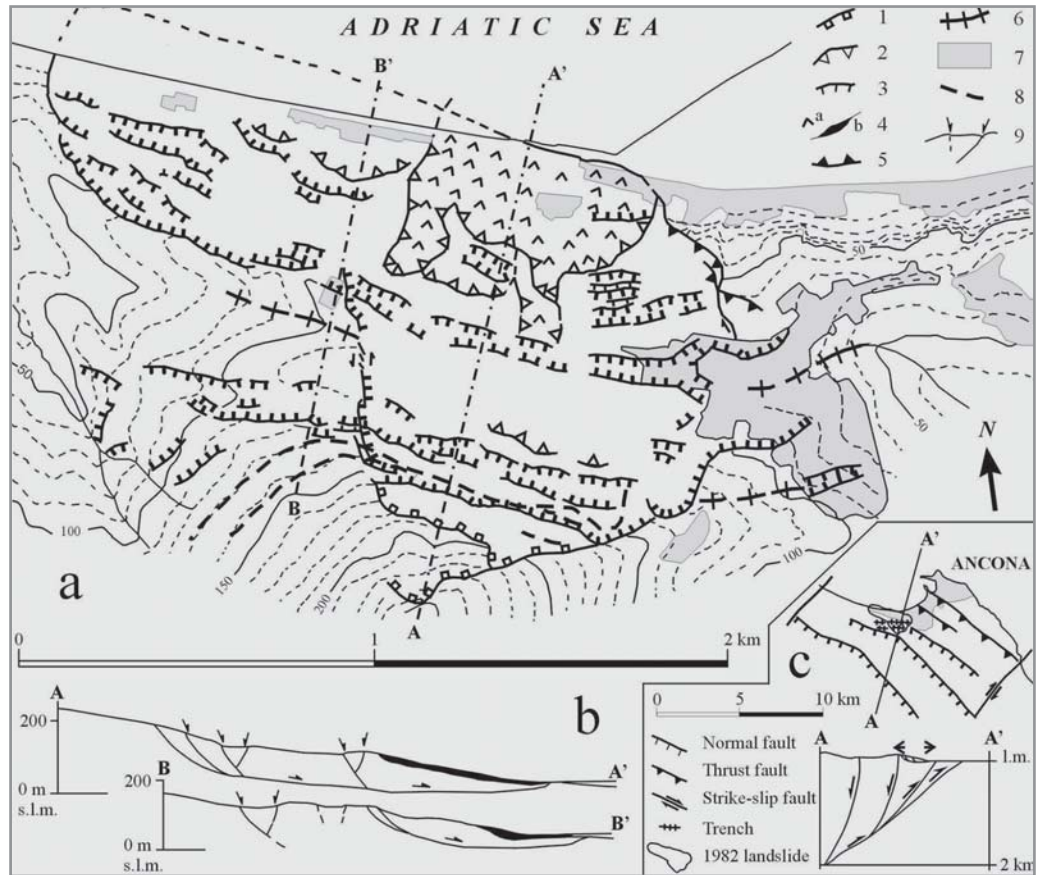


The origin of the above landforms is to be linked mainly to deep-seated gravitational deformations (tectonic-gravitational spreading), induced by active compressional tectonics (Dramis and Sorriso-Valvo, 1994). Within this framework, large scale rotational-translational landslides and listric faults, lowering towards the Adriatic Sea, are produced (Coltorti et al, 1984).

One of the most representative examples of the above-mentioned phenomena, is the deep-seated



Figure 20 - A 16th century post-house, showing the cumulative effects of recurrent landslide movements on the Montagnolo slope.



gravitational slope deformation which involved the north-facing slope of Montagnolo Hill, in the western outskirts of Ancona (Figure 19). Here, on the evening of the 13th December, 1982, after a period of heavy rain, a huge landslide took place, over an area of more than 3.4 km², from about 170 m a.s.l. to the Adriatic coast (Coltorti et al, 1984).

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 fatalities.

The slope hit by the landslide has had a long history of gravitational movements (Bracci, 1773; Segrè, 1920; Figure 20). In 1858, it was the site of a landslide even larger than the recent one (De Bosis, 1859). More shallow mass movements, still large in an absolute sense, have occurred in the landslide area. Of these, the Barducci mudflow, is well known for its continuous activity, and the resulting damage to the coastal road and railway (Segrè, 1920).

From a stratigraphic point of view, the lithotypes outcropping on the landslide affected slope are the following:

- 1) Lower-to-Middle Pliocene deposits (grey-blue marly clays, 20-40 cm thick, alternating with grey or grey-black compact sands up to 60 cm thick);
- 2) Pleistocene deposits, consisting of five transgressive-regressive cycles of pelitic-arenaceous units, with a total thickness of about 20 m.

The area has been uplifted, starting at the end of the 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 events of December 1982, shows a characteristic landslide morphology, with trenches, scarps, steps, undrained depressions, and reverse slopes (Figure 21). Moreover, aerial photo analysis shows that several deep open fractures and

flexural scarps were produced after 1956 but before 1979 in the area of the main detachment zone of the 1982 landslide. Most likely, similar surface displacements 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 was characterized by an amount of precipitation not particularly relevant from a hydrological point of view. Therefore, a fundamental role for the generation of the 1982 landslide was played by the coseismic opening of the numerous tectonic fractures.

A rugged foot-slope zone extends towards the sea; the steepening of the foot-slope seems to be accounted for by sea erosion, which was still active at the end of the 18th century (Bracci, 1773). Subsequently, the general advance of the coast line, caused by widespread deforestation in the inner mountain areas, and, more recently, by the harbour embankment and the along-shore protective measures, has built up a narrow belt of earth which separates the foot-slope from the sea. This belt is presently densely inhabited (apart from the sectors affected by the landslide, where all the previous buildings were destroyed), and crossed by two highways and a railway.

Stop 2.2:

The Montelparo Landslide

Angeli M.G., Pontoni F. and Dramis F.

The medieval village of Montelparo is located in a hilly



Figure 22 - Frontal view of the landslide area

area of the Marche Region, to the east of the Apennine chain. It is affected by a large translational landslide that develops from 580 m a.s.l. to 340 m a.s.l., with a length varying between 700 and 1100 m, and an average width of 600 m (Figure 22).

The landslide body is part of a monocline dipping 10°-12° NE, made of well-stratified sandstones overlying a

◀ **Figure 21 - Geomorphological sketch: (a) of the Ancona landslide slope; (b) Landslide sections; (c) Tectonic pattern. Key to symbols: 1, old non-reactivated main scarp; 2, shallow landslide scarp; 3, scarp linked to deep-seated deformation; 4, landslide body; 5, shearing surface; 6, track of trench; 7, intensely urbanized area; 8, fractures formed between 1956 and 1979, most likely generated during the 1972 earthquake sequence (after Cotecchia, 1997); 9, shearing surface (inferred when hatched).**

deep-seated clayey bedrock. A system of normal faults, and the erosion operated by the Fosso di S. Andrea stream, flowing at the toe of the slope, have favored the detachment of the sandstone slab (up to 65-70 m thick) – which slid on a clayey, sloping slip surface – and the formation of a well-marked depression, filled



Figure 23 - Graben structure (the horizontal arrow indicates the moving part of the village)

with debris material on top of the hill (Figure 23). The lowermost part of the sliding slab, much thinner than the upper one, is characterized by progressive disruption into blocks, divided by cracks (up to 3-5 m deep, and 2-3 m wide).

In other words, the Montelparo hill is split into two parts: the uphill part that is stable, whereas the downhill part is slowly moving as a whole (the buildings do not show any major tilting fracturing). In the intermediate area, a process of continuous settlement occurs (Angeli, 1981). This last area is known in geotechnical literature as a “graben”, on the base of the classification done by Skempton and Hutchinson (1969). On the northern flank of the hill, the upper limit of the depression is

exposed, showing the sub-vertical plane cut into the bedrock in contact with the filling debris.

Damage induced by the landslide in the built-up area have been documented since the XVII century. An important reactivation of the movement in coincidence with the high intensity earthquake which struck the area in 1703, is reported in the historical literature (Pastori, 1781). Major damage to buildings occurred only within this slowly expanding depression and along its margins.

Levelling surveys carried out from May 1977 to March 1979 established that along the trench, vertical displacements of 20-25 cm took place. The average speed varied from 0.3 cm/month for the first 16 months, to ca. 2.5 cm/month for the remaining 6 months. In addition, precision topographic surveys, carried out in 1980, showed horizontal displacements of up to 2-3 cm (Angeli et al, 1996). By comparing a cadastral map dating back to 1935, and a new map derived from aerial photographs taken in 1970, it was revealed that, in 35 years, the trench had widened by about 3 m, at an average rate of 8 cm/year.

In the short term, the critical hydraulic conditions are similar to the ones occurring in rockslides, where a triangular water pressure diagram operates inside the graben area (on the subvertical face of the moving mass), and a rectangular one acts on the sloping clayey slip surface. An indirect confirmation of this mechanism was provided by the acceleration of the movement in coincidence with a rainfall critical event occurring in December 1999, when significant piezometric peaks were also recorded. According to the mechanism invoked, any increase in water level makes the water thrusts (on the rear of the landslide body and at its base) increase exponentially. Hence the importance of maintaining the water levels permanently low, well

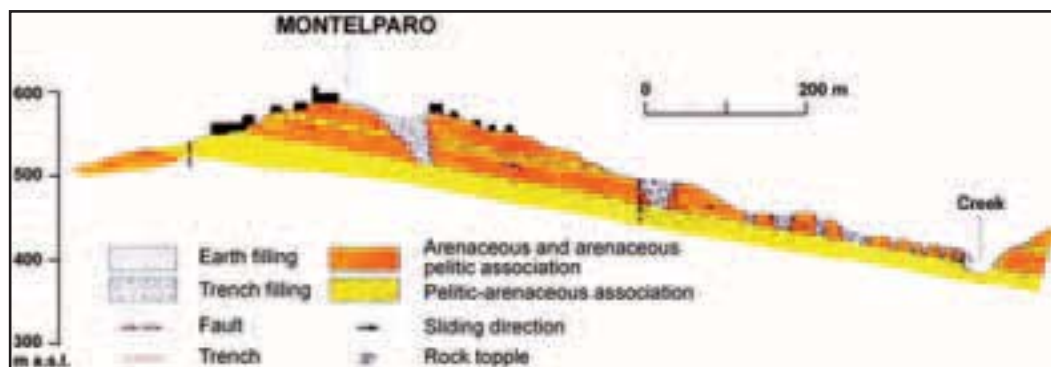


Figure 24 - Section along the Montelparo slide, showing the monocline setting of the bedrock made of stratified sandstone with clayey interlayers.

below the critical values.

The main control works (Angeli and Pontoni, 2000; Angeli et al., 2002) belong to the category of deep drainage, useful to increase the shear strength on the slip surface (Figure 24).

Stop 2.3:

Seismic landscape analysis in the L' Aquila Basin and Upper Aterno Valley

Blumetti A. M.

Stop 2.3.1:

A 24 Roma-L'Aquila, service area "Aterno": geological overview of the L'Aquila Basin and Mount Pettino fault escarpment

The L'Aquila Basin and the Upper Aterno Valley are two adjoining tectonic depressions (Figs. 25 and 26), of the Central Apennines, located between the Gran Sasso and the Monti d'Ocre-Velino-Sirente stratigraphic-structural Units (Accordi et al., 1988). The NE edges of these basins are bordered respectively by the Mount Pettino and Mount Marine faults, which are part of a NW-SE-trending segmented fault system with Late Quaternary activity (Figure 25; Demangeot, 1965; Bosi, 1975; Blumetti, 1995; Bagnaia et al., 1996; Roberts and Michetti, 2004; Roberts et al., 2004).

These two faults are arranged with an en échelon geometry, and are separated by a roughly 3 km wide right lateral step (Figure 26). They are at the base of steep fault escarpments (Figs. 27 and 28), where a scarp identifies the point where the fault outcrops (Figure 29 and 30). This scarp may be interpreted as a scarplet, i.e. arising from coseismic reactivation of a capable fault.

Both the faults are morphologically evidenced by a very wide, cataclastic belt. This is affected by accelerated erosion, giving rise to calanque-type reliefs (Figs. 27, 28, 29, and 30).

Mount Pettino and Mount Marine escarpments show all the characteristics of fault-generated mountain fronts (Wallace, 1978; Blumetti et al., 1993): their lower portion is very steep, and shows wonderful trapezoidal and triangular facets, separated by typical wine-glass-valleys.

Moreover, both the faults cut the piedmont belt which is made up of slope-waste and alluvial-fans-deposits, Late Pleistocene-Holocene in age, giving rise to fault scarps up to 10 meters high. One of these scarps, crossing an alluvial fan located at the base of

the Pettino fault, has been studied through detailed topographic surveys (*3 in Figure 26 and circle in Figure 27), with both GPS (Blumetti et al., 1997) and teodolite (Giuliani et al., 1998) methods. Both methods evidenced a fault scarp with a vertical offset of about 3m in Late Pleistocene deposits. The age of these deposits is not well constrained; a possible age of between 30,000 and 10,000 years BP would lead to a slip-rate of 0.1-0.3 mm/yr.

Other detailed topographic surveys carried out along the Mount Marine fault escarpments, on fault scarps dislocating for 8-10 meters slope-waste deposits dated back to $31,710 \pm 760$ and $23,330 \pm 300$ years BP, led to a slip-rate of 0.25 and 0.43 mm/yr (Figure 30; Galadini and Galli, 2000). As regards the Mount Marine fault, we will discuss in detail the dislocation of Late Pleistocene deposits in the next stop.

The Monte Marine fault (and, less evidently, the Mount Pettino fault escarpments), abruptly interrupts a fairly flat top surface, which is a remnant of a "low

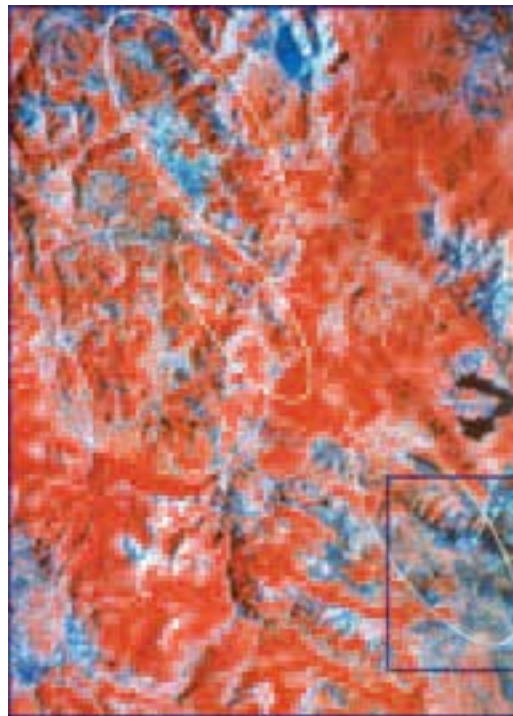


Figure 25 - Landsat satellite image of the fault system extending from the Norcia basin to the L'Aquila basin. The ellipsis shows the areas affected by the three main shocks of the January-February 1703 Central Italy seismic sequence, and the faults activated. In the inset box, the L' Aquila Basin and the Upper Aterno Valley.

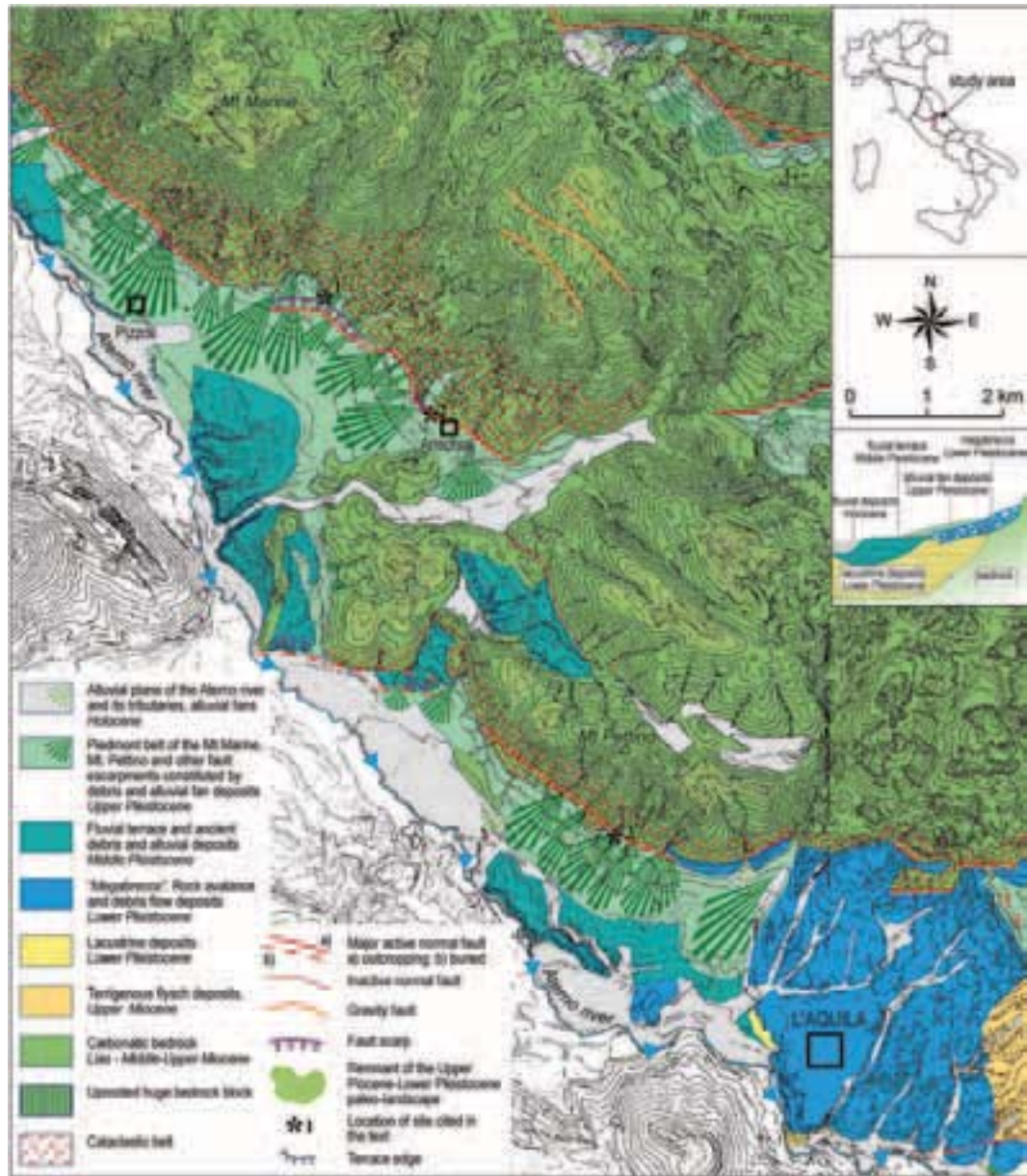


Figure 26 - Geological map of the left Upper Aterno Valley and L'Aquila basin and surrounding reliefs, and schematic stratigraphic correlations of Quaternary deposits.

energy relief landscape” that has been described here and in other areas of the Central Apennines. This type of landscape is the result of a Late Pliocene to Early Pleistocene evolution under arid-semi-arid climatic conditions, during a tectonic phase characterized by a very low rate of activity (Demangeot, 1965). Entrenched in this paleo-landscape, (and, as regards

the more ancient, down-thrown by the basin-border-faults), there are quaternary flights of terraces. In Figure 26, only the ones to the left of the Aterno are represented.

The more ancient deposits outcrop in the city of L'Aquila (Via Mausonia Unit; Blumetti et al., 2002).



Figure 27 - Panoramic view of the Mount Pettino fault escarpment. Circle indicates the location of the faulted fan, where detailed topographic measurements were carried out.



Figure 28 - Panoramic view of the Mount Marine fault escarpment.



Figure 29 - Fault scarp et the base of the Mount Pettino fault escarpment.

Based on lithostratigraphic and geomorphological data, they may be correlated with the “Madonna della Strada” complex, outcropping in the western part of the L’Aquila Basin and dated back to Lower

Figure 30 - Detail of the fault scarp et the base of the Mount Marine fault escarpment. Note the faulted debris deposits.



Pleistocene, thanks to the finding of a complete individual of *Mammuthus (Archidiskodon) meridionalis vestinus* (ascribed to the Farneta Faunal Unit of Late Villafranchian age; Esu et al., 1992).

Lithologically, the Via Mausonia Unit is made up of thin (a few cm-thick) beds of sandy silts, locally clayey, and sands, with frequent lignite lenses, of a lacustrine environment (Blumetti et al., 2002).

Above these sediments, just in the historical center of L'Aquila, paleo-landslide deposits are encountered, called "Megabrecce" (Demangeot, 1965). In their upper portion, these deposits grade into high-energy, alluvial-fan deposits (Blumetti et al., 2002). These deposits, as regards the lower part, should correlate with the Aielli and Poggio Poponesco Breccias, that we will encounter in our next stops (3.1 and 3.2). These ancient deposits were probably related to a basin quite different from the present, being guided by different tectonic elements.

The "Megabrecce" Unit shows a top surface that is a reworked depositional surface. This is down-through towards the east by an antithetical fault of the Middle Aterno Valley system, so that it is at the top of a fault escarpment partially visible in the right border of Figure 26 (Blumetti et al., 2002). Towards the Aterno River, on the other hand, it hangs above the valley floor for an height of about 50 m. In the L'Aquila basin, a single terraced unit is entrenched in this surface. This is an alluvial terrace, which is suspended above the Aterno River from a height of about 20 m and it is constituted of calcareous gravels with channeled structures and cross-bedded sand lenses. The sedimentary structures indicate that the sedimentation environment is fluvial, with braided channels and paleocurrents which had a direction similar to the present course of the Aterno River. A leached paleosol, resting in part on volcanic material, caps the terrace. The soil profile suggests that it must have evolved over a relatively long period (probably interglacial). The age of the terrace is estimated to be the upper part of the Middle Pleistocene (Blumetti et al., 1996; Blumetti et al., 2002). In the Upper Aterno Valley, the Middle Pleistocene terrace is particularly evident at the confluence of the Aterno River with a large stream that flows in the transfer zone between the Mount Marine and Mount Pettino faults. As mentioned before, these two faults are arranged with an échelon geometry, and are separated by an about 3 km-wide, right-lateral step.

As is common in this situation (Jackson and Leeder, 1994), the geometry of the two fault-segments causes

the transfer zone to be occupied by a large stream. Exploiting the natural slope created by the en échelon step-over, this flows from the footwall of the Mount Pettino fault into the hanging wall of the Mount Marine fault.

This situation has probably lasted from the end of the Middle Pleistocene, because, as already said, a fluvial terrace about 25 meter high is very evident in this position.

Underneath this terrace, sediments characterized by a reverse polarity, interpreted as deposited during the reverse Matuyama Chron, in Lower Pleistocene, have been signaled (Messina et al., 2001), but they have not been reported in Figure 26 for a matter of scale.

An evident wind gap, and the presence of hanging and faulted Middle Pleistocene alluvial deposits upon and along the Mount Pettino fault escarpment, indicate that, in the past, in previous stages of the fault's growth (Cowie and Sholz, 1992), a stream, with catchment behind and parallel to the Mount Pettino ridge, flowed into a paleo-Aterno Valley, possibly in the ancient transfer zone between a "paleo Mount Marine fault" and a "paleo Mount Pettino fault". The "paleo Mount Marine fault" could correspond to a fault located just on the SE prosecution of the present fault, marked in Figure 26 by a thin line. In this hypothesis, this fault would not be active any more.

Up on the Middle Pleistocene terrace, the Upper Pleistocene slope-waste and alluvial fan deposits which constitute the Mount Marine and Mount Pettino piedmont belts, are found (Figure 26).

Stop 2.3.2:

Mount Marine fault escarpment and the Colle Site.

The present activity of the Mount Pettino and Mount Marine fault system is testified by seismic catalogs, recording in the area of L'Aquila some of the highest-intensity events in central Italy (1349, 1461, and 1703, I = X MCS; 1762, I = IX MCS, Gruppo di Lavoro, CPTI, 1999).

The 3 km wide step-over that actually separates the two faults, is too short to constitute a barrier to the propagation of faulting during a strong earthquake, so that the two faults can be simultaneously reactivated, for a total length of about 20 km. This length, applying the equations linking the surface rupture length with the magnitude of an earthquake (Wells and Coppersmith, 1994; Mohammadioun and Serva, 2001), leads to Mw = 6.5.

On the other end, paleoseismological studies have

indicated the occurrence of Late Pleistocene to Holocene paleo-earthquakes with a magnitude of about 6.9, on the Monte Marine fault (Blumetti, 1995; Moro et al., 2002).

In particular, this zone was the epicentral area of a main shock of the January-February 1703 seismic sequence in central Italy, one of the strongest in the whole history of Italian seismicity. It was characterized by the progressive NNW to SSE propagation of three main shocks (January 14th, I=X; January 16th, I= VIII, and February 2nd, I= IX, 1703) along the axis of the Apennines Chain, and parallel to the major extensional tectonic lines (Figure 25). This propagation is presumed to have occurred along a segmented fault system (Blumetti, 1995).

Historical investigations and field surveys have shown that during the seismic sequence, surface effects occurred mostly in a NW trending belt (Figure 25), along tectonic structures that show geological and geomorphological evidence of activity. These surface effects were mainly tectonic, both primary and subordinate. Many of the tectonic ruptures were also partly gravity driven (Blumetti, 1995).

Primary and subordinate tectonic effects have been recognized along the master fault at the bottom of the Mount Marine fault escarpment, in a belt between the villages of Pizzoli and Arischia. Here, during the 2nd February, 1703 earthquake, two small craters ejecting sulfurous water opened up (Grimaldi, 1703; Uria de LLanos, 1703; Parozzani 1887). This suggests the occurrence of liquefaction, a secondary effect probably related to primary surface faulting. In fact, in this area Upper Pleistocene-Holocene alluvial-colluvial deposits are deformed, close to the bedrock-alluvium contact, to form subsidiary fault scarps a few meter high, and parallel to the main slope (Blumetti, 1995; Bagnaia et al., 1996).

Paleoseismological analysis carried out in a site close to the Arischia village revealed the occurrence of repeated surface faulting events during the Upper Pleistocene to Holocene. In particular one earthquake, with a magnitude of about 6.9, occurred a short time after 29,690 years BP.

In this same deformed belt, close to Colle, a wide exposure still shows faults which dislocate an alluvial-colluvial sequence, which is some meters high, (Figure 31) and Late Pleistocene-Holocene in age (it contains a layer dated back to 27,000 years B.P.). The analysis of this section show that all the layers involved in the deformation are dislocated for an height that is higher than the one of the outcrop.

For example, layer 6 (dated back to 27,000 years B.P.), is up-thrown by more than 3 meters by the western fault, and down-thrown by more than 6 meters by the eastern fault. It is impossible to estimate how many events caused this dislocation, whereas it is possible calculate a slip rate on both the faults, from the time of the developing of the dated soil, of 0.33 mm/y. The last event involve a recent colluvium that is dislocated by about 70 cm. This throw, applying the equations linking the maximum or the average surface displacement with the moment magnitude of an earthquake (Wells and Coppersmith, 1994), leads to about $M_w = 6.7$.

This event can probably be related to the 2 February 1703 earthquake.

Recent trenching across one of the above-mentioned subsidiary fault scarps, has indicated the occurrences of five displacement events within the last 15,000 years, with the most recent probably related to the 2 February 1703 earthquake, as indicated by the displacement of historical colluvial deposits (Moro et al., 2002).

Some features of the relief might be interpreted as the result of the combined action of seismic activity and gravity. Many subordinate tectonic rupture events leaded a landscape with a low energy relief (that lasted until the Early Pleistocene) to be progressively uplifted and dismembered in a horst and graben structure.

The top of the Mount Marine ridge is fairly flat as a whole, but in detail it is composed of small horst-and-graben-like forms. At the top of this ridge, on the mountain sector between the village of Arischia and the "Piano di Rotigliano" (Figure 26), "colossal" 1703 ground effects have been described, occurring in the territory of Colle, on and around a hill called "Colle del Grillo" (Uria de LLanos, 1703). These ground effects have been localized by interviewing the inhabitants of the surrounding villages, who have indicated Colle del Grillo, a hill not named on the current official topographic maps. Here, tectonic effects were probably amplified by gravity, in a kind of gigantic, deep gravitational deformation, which led to the lowering of the Colle del Grillo hill, and led to substantial modifications in the slightly undulating morphology of the top of this ridge.

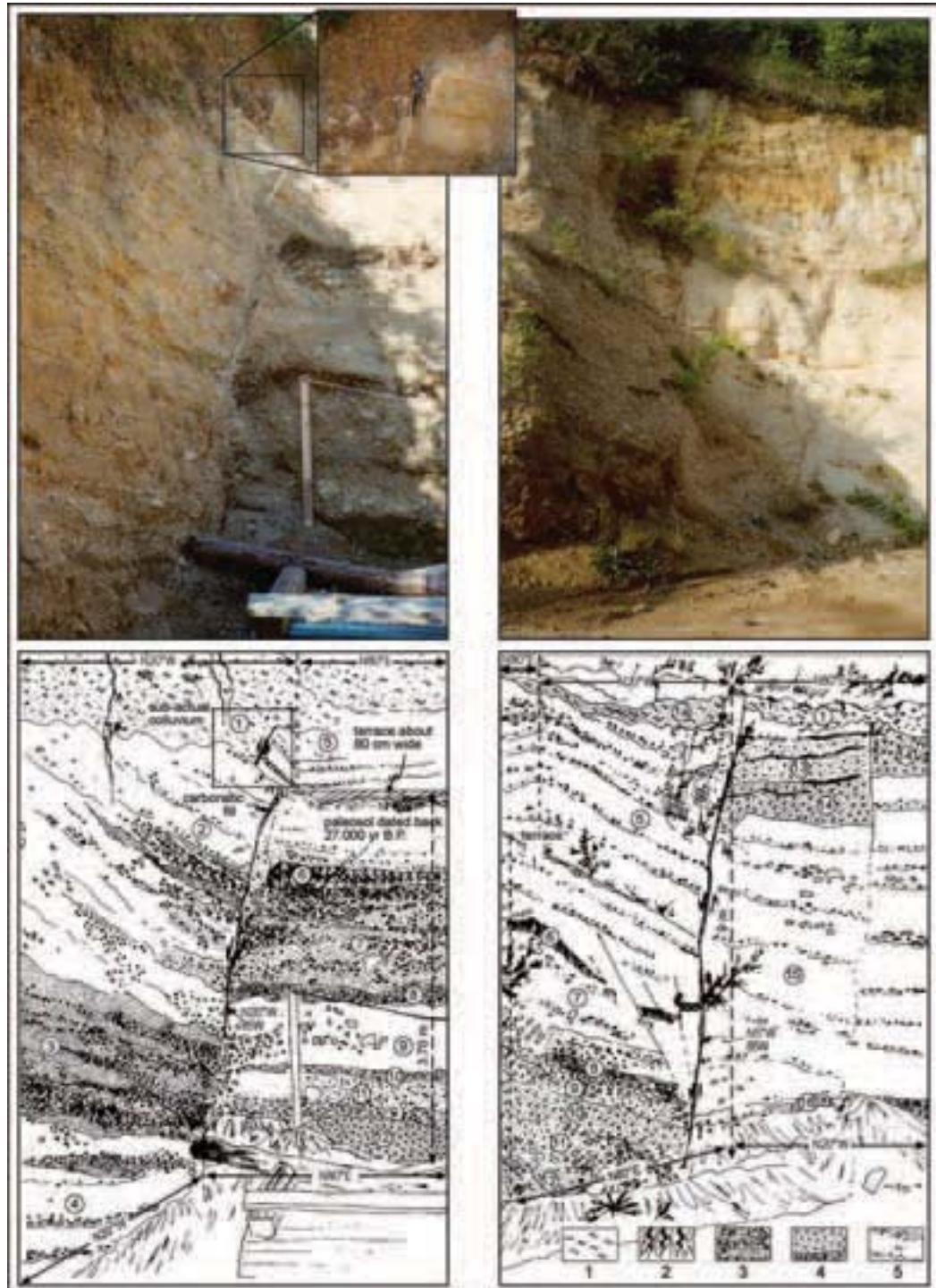


Figure 31 - Two stratigraphic sections drawn from pictures taken in the cut of Colle. The two sections are 8 meters apart. Legend: 1) Alteration; 2) Paleosol; 3) Sand and fine gravel; 4) Volcanic material; 5) Gravel.

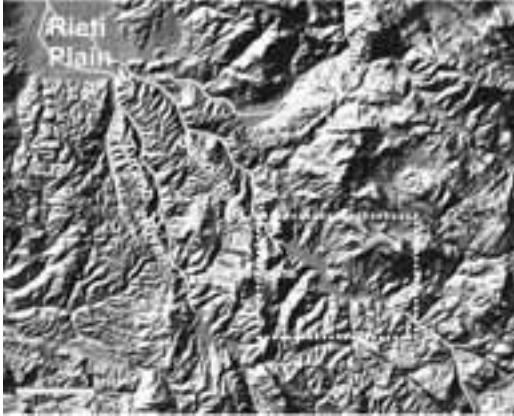


Figure 32 - The Salto river drains a wide sector located between the Sabini and Cicolani mountains, before flowing into the Velino river, near Rieti.

DAY 3

Stop 3.1:

Holocene surface faulting along the Fiamignano fault, and associated large scale gravity slope deformation (Salto river valley, Rieti)

Guerrieri L. and Silvestri S.

This stop is focused on the role of active tectonics and associated large-scale gravity slope deformations in the landscape evolution of the Salto River valley (Rieti).

The Salto River drains a wide sector from SE to NW in the Rieti Plain, which is located between the Sabini Mts. (Mt. Navegna, 1508 m), and the Cicolani Mts. (Mt. Nuria 1888 m). Since 1940, the Salto River fluvial processes are controlled by a 90 m high dam located at Balze S. Lucia, which has created a narrow and deep artificial lake.

Stop 3.1.1:

SS 578 Rieti-Torano, exit “Gamagna”: geological overview and active processes in the Salto River valley

The Salto River valley is crossed by a first-order thrust system (“linea Olevano-Antrodoco”, *Auct.*), trending about N-S, which contacts Meso-Cenozoic pelagic calcareous and marls (serie Umbro-Marchigiana), with Cretaceous neritic limestones (serie Laziale-Abruzzese), and Neogene turbiditic sequences (see Bigi & Costa Pisani, 2003 and bibliography herein). The older continental deposits, still preserved on the northeastern sector of the Salto valley (Sabbie di Piage and Breccie di Poggio Poponesco, Pliocene,

Bertini & Bosi, 1976), record an ancient intermountain basin (Salto Basin), filled with sands, conglomerates, and breccias. At that time, this area had become part of the Rieti Basin drainage network (Villafranchian, *Auct.*) as indicated by relict erosional surfaces. Related continental deposits have been eroded in this area, but are preserved near Grotti, some km to the NE (*braided* and *alluvial fan* facies, according to Barberi & Cavinato, 1993).

In the Middle Pleistocene, the regional uplift and the tectonic lowering of the Rieti Basin triggered a fast deepening of the Salto valley, almost to the current valley floor elevations. Analogously with the contiguous valleys (i.e. the Velino valley; Carrara et al., 1993), since the Upper Pleistocene the erosional and sedimentary processes have been controlled basically by the growing and downcutting of travertine barriers strictly connected to the climatic conditions.

The resulting landscape is characterized by a high energy relief that promotes instability on the slopes (surface landslides in terrigenous deposits, rock falls in vertical slopes). Moreover, on the northeastern slope, it is possible to recognize deformations that indicate deep-seated gravitational movements at different evolutionary stage (see the following stops). The development of these deformations is controlled (*sensu* Dramis, 1982) by topographic factors (high energy of the relief), and lithological factors (limestones topographically above sandstones and clays). Moreover, their catastrophic evolution is controlled by the Fiamignano fault, that is capable of producing surface faulting during strong earthquakes.

Stop 3.1.2:

Poggio Poponesco (Fiamignano): Holocene tectonic activity along the Fiamignano fault, and associated large-scale gravitational movements

The Fiamignano fault is a NW-SE-trending normal fault that downthrows to the SW with a maximum of at least 2200 m. The polyphasic evolution of this fault, during the compressive Neogene stage and Plio-Pleistocene extensional regime, has had different interpretations (i.e. Bosi, 1976; Capotorti & Mariotti, 1992; Bosi et al., 1994; Morewood and Roberts, 2000; Bigi & Costa Pisani, 2002).

Concerning the recent activity, a bedrock fault scarp (Figure 34A) interrupts the slope profile continuity for more than 10 km, between Staffoli (to the NW) and Brusciano (to the SE). In the Central Apennines, this is a primary feature typically associated with the



Figure 33 - Aerial view of the middle sector of the Salto river valley, at present filled by an artificial lake (ortophoto by TerraItaly, 2000). White dots indicate the location of stops.

major Holocene normal faults. More evidence of recent activity for this fault is clearly visible in the Poggio Poponesco area, where i) an Upper Quaternary erosional channel and Holocene colluvial deposits are cut by the Fiamignano fault (Figure 34B and 34C), ii) Upper Pleistocene slope deposits are tilted, and at present

dip in the opposite direction (Figure 34C). Trenches and counterscarps, visible in different sites along the fault (Poggio Poponesco, Castiglioni, and S.Vittoria, Figure 35 and 36), have been interpreted as the surface expression of large-scale gravity slope deformations associated with the Fiamignano fault.

Stop 3.1.3:

Borgo San Pietro (Petrella Salto): catastrophic evolution of large scale gravity slope movements

The Borgo San Pietro village is founded on chaotic deposits reworked by areal erosional processes during the Last Glacial Maximum (Figure 36). Above the Colle della Sponga village, it is possible to observe a

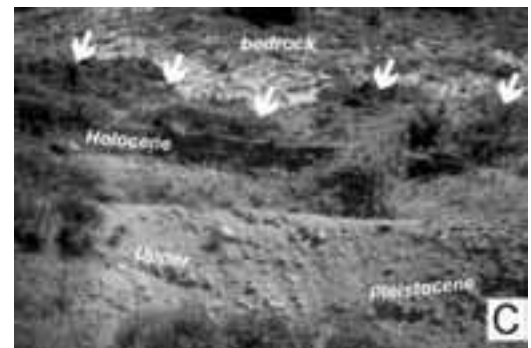


Figure 34 - Evidence of active tectonics along the Fiamignano fault in the area of Poggio Poponesco (see text).

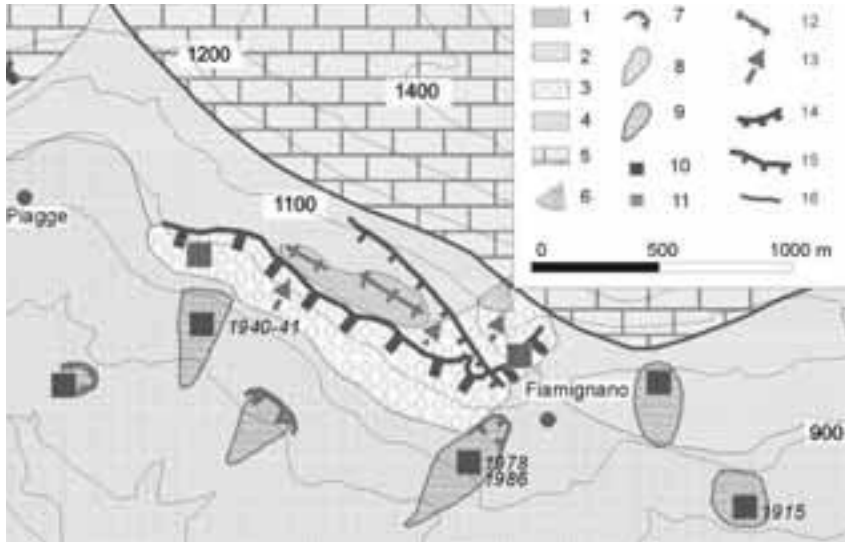


Figure 35 - Geological and geomorphological sketch in the area of Poggio Poponesco (from Guerrieri et al., 2001). Legend: 1) Eluvial and colluvial deposits; 2) Recent talus; 3) Old cemented breccias; 4) Clays, sand,s and sandstones; 5) Limestones and calcarenites; 6) Alluvial fan; 7) Landslide main scarp; 8) Landslide body; 9) Landslide body; 10) Sliding historical events; 11) Rock fall historical event; 12) Trench; 13) Counterscarp; 14) Rock fall scarp; 15) Fault scarp; 16) Fiamignano fault



Figure 36 - Geological and geomorphological sketch in the area between Borgo S. Pietro and Colle della Sponga (from Guerrieri et al., 2001). Legend: 1) Eluvial and colluvial deposits; 2) Old cemented breccias; 3) Clays, sand,s and sandstones; 4) Limestones and calcarenites; 5) Chaotic deposits reworked by areal erosional processes; 6) Landslide main scarp; 7) Landslide body; 8) Slope affected by deep-seated gravitational movements; 9) Trench; 10) Counterscarp; 11) Erosional channel; 12) Erosional surface; 13) Fiamignano fault

large concave detachment area along the carbonatic ridge (Figure 37). These are stratigraphic and geomorphologic evidences of a collapsed huge block (paleolandslide).

It is possible to speculate that it was the last stage in the evolution of large-scale gravity slope deformations, as described before. This catastrophic evolution requires, in our opinion, triggering factors such as strong earthquakes, together with a maximum energy of relief. Thus, the risk of catastrophic collapse is higher when the Salto River valley floor is at low-stand stages, fostered by climates cooler and drier than the present ones.



Figure 37 - Aerial photo, showing a large detachment area above the Colle della Sponga village and the landslide body near Borgo San Pietro. These features have been interpreted as evidence of an ancient catastrophic collapse

Figure 38 - Tectonic map, showing the system of capable normal faults in the Abruzzi Apennines; the distribution of historical seismicity from 1000 to 1900 A.D. is also represented. Note the absence of earthquakes in the Fucino area before the 1915 event, and compare with paleoseismic data in Table 1.



Stop 3.2:
Paleoseismology and Quaternary evolution of the Fucino Basin
Blumetti A.M. and Michetti A.M.

Stop 3.2.1:
Salviano Mt.- Overview of the history and Quaternary geology of the Fucino Basin

The Fucino Basin (Figs. 38 and 39) is the largest tectonic basin of the Abruzzi region. It lies in the middle of the central Apennines, surrounded by mountain ranges higher than 2000 m (Mt. Velino, 2486 m a.s.l.), which are shaped essentially into Meso-Cenozoic carbonate shelf sediments. The basin does not represent a concentration of the hydrographic network or of the important rivers. On the contrary, around it the upper course of the main rivers flows NW (Salto), S then SW (Liri), S then E (Sangro; Figure 38).

The central part of the basin, a plain between about 650 and 700 m, which is hydrologically closed, was occupied during the Late Glacial and Holocene by the third largest lake in Italy (ca 150 km²). In the second century A.D. the Roman Emperor Claudius initiated the drainage of Lake Fucino. This was accomplished through the excavation of a 6-km long tunnel, mostly carved in the Mesozoic limestone, one of the most remarkable engineering projects in Roman history. The last drainage of this area was performed at the end of the last century by Alessandro Torlonia. In the year 1875 A.D., Lake Fucino disappeared completely

from the map.

The Fucino basin is a typical intermountain normal-fault-bounded structure within the Apennines, segmented, normal fault system, which extends from southern Tuscany south, to the Calabrian Arc, and represents one of the most seismically active provinces of the Mediterranean region (Michetti et al., 2000; D'Agostino et al., 2001). Major normal faults, representing the nearest segments of this system, are the Mt. Magnola-Mt. Velino fault to the NW ("e" in Figure 38; Morewood and Roberts, 2000), and the Sangro Valley fault to the SE ("i" in Figure 38). Seismic reflection profiles (Mostardini & Merlini, 1988; Cavinato & alii, 2002), indicate that the Fucino structure is a half-graben, controlled by the master fault along the NE border of the basin, i.e. the Celano-Gioia dei Marsi normal fault (Beneo, 1939; "a" in Figure 38), and parallel subsidiary faults (e.g. the Parasano-Cerchio and Aielli-Giovenco faults; "b" and "c" in Figure 38). Within this style of faulting, tectonic inversion (Quaternary normal slip on pre-existing reverse faults) is very well documented – for instance,

Trench site	Locality	Strike	Dip	Holocene Vertical Offset	Holocene Horizontal Offset	Vertical Slip-Rate (mm/yr) and time-window	Recurrence Time (yr) and time-window	Total Events	Paleoearthquake Ages	Coseismic Vertical Slip
1	Strada Statale Marsicana	NW-SE	SW	> 5 m	none	0.4-0.5 (last 20000 yr)	4500-5000 (20000)	5	1915 AD; 6500-3700 BP; 7300-6100 BP; 13600-12300 BP; 19100-18500 BP	ev1: 0.50 m
2	Colle delle Cerese	N50W	SW	> 3 m	none	0.35-0.40 (last 7000 yr)	1000-1800	3	1915 AD; 1100-1500 AD; 7200-6540 BP	ev1: 0.70 m
3	Colle delle Cerese (Cave)	NW-SE	SW	10 - 15 m	none	0.35-0.40 (last 4200 yr)	1400-2100	3	1915 AD; 150-1349 AD; 3760 BP; 150 AD	ev1: 0.70 m
4	Molimi di Venere	NW-SE	SW	> 3 m	none	0.35-0.40 (last 10000 yr)	1200-1500 (2000)	4	1915 AD; 1300-1500 AD; 7120-5340 BP; 10400-7120 BP	ev1: 0.70 m
5	Casali D'Aschi	NW-SE	SW	5 m	none	0.35-0.40 (last 20000 yr)	800-1000 (3000); 3300-5500 (33000)	7	1915 AD; 1200-1400 AD; 2783 BP-1300 AD; 4700-2800 BP; 10400-7120 BP; 20000-10000BP; 32520-20000 BP	ev4: 2 m
6	Trasacco (Fosso 41)	N22W	WSW	> 3 m	none	0.3-0.4 (last 7000 yr)	1500-1800 (2000)	2	1915 AD; 1000-1349 AD; 3700-3500 BP; 7120-5000 BP; 10790-7120 BP	0.5 m; 0.6 m
7	Trasacco (Strada 37)	NW-SE	SW	> 3 m	none	(s.a.a.)	1800-2000 (7000)	5	1915 AD; 1000-1349 AD; 3700-3500 BP; 7120-5000 BP; 10790-7120 BP	ev1: 0.50 m
8	Trasacco (Strada 38)	N60W	SW	> 3 m	none	0.3-0.4 (last 10800 yr)	1600-1800 (10800)	7	1915 AD; 1000-1349 AD; 3700-3500 BP; 7120-5000; 10790-7120 BP (2 events); > 10.790 BP	evs 1,2,3,4; 0.55 m; evs5,6,7; 0.15 m
9	Trasacco (Strada 10)	NW-SE	SW	> 3 m	none	0.3-0.4 (last 12000 yr)	1800-2000 (12000)	8	1915 AD; 1000-1349 AD; 3700-3500 BP; 7000-5000 BP; 10790-7120 BP (3 events); > 12000 BP	ev1: 0.1 m
10	Luco dei Marsi (Strada 42)	NW-SE	NE	> 3 m	none	0.8 (last 1500 yr)	1000-1800 (1500)	2	1915 AD; 500-1500 AD	0.1 m; 0.15 m
11	Luco dei Marsi (Strada 45)	NW-SE	NE	> 3 m	none	(s.a.a.)	(s.a.a.)	2	1915 AD; 500-1500 AD	0.1 m; 0.15 m
12	San Benedetto dei Marsi	NW-SE	SW	> 3 m	none	1.0-1.6 (last 2000 yr)	500-800 (2000)	3	1915 AD; 885-1349 AD; 550-885 AD	ca. 0.5 m; ca. 0.5 m; > 1 m
13	Piano di Pezza	N120E	SW	3.5 m	none	0.7-1.2 (Holocene)	800-3300 (5000)	2	860-1300 AD; 1900 BC	2-3.4 m; 1.2-2.5 m
14	Vado di Pezza	N135E	SW	12-16 m	none	(s.a.a.)	(s.a.a.)	3	860-1300 AD; 1900 BC; 3300-5000 BC	2-3.4 m; 1.2-2.5 m; 2.0 m
15	Campo Porcato	N165E	W	6.5-11 m	5.5-8.5 m	(s.a.a.)	(s.a.a.)	2	860-1300 AD; 1900 BC	2-3.4 m; 1.2-2.5 m

Table 1: Synopsis of paleoseismological analyses in the Fucino basin and nearby areas (site numbers as in Fig. 38; data from Michetti et al., 2000b, and references therein).

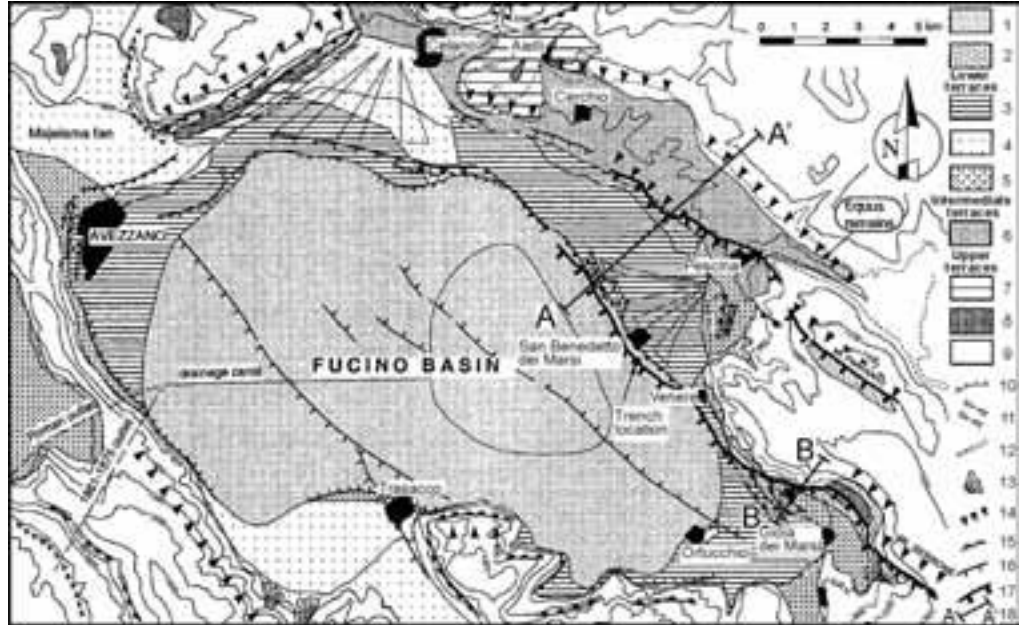


Figure 39 - Geological map of the Fucino Basin, displaying the surface faulting associated with the January 13, 1925, M7 earthquake. The location of the Equus cf. Altidens site is also shown. Legend: 1) Historical lake; 2) talus deposits (late Glacial and Holocene); 3) alluvial deposits (Holocene); 4) alluvial fan deposits (late Glacial); 5) fluvio-lacustrine deposits (late Glacial); 6) fluvio-lacustrine deposits (Middle Pleistocene); 7) fluvio-lacustrine deposits (Upper Pliocene?-Lower Pleistocene); 8) breccias (Upper Pliocene-Middle Pleistocene); 9) sedimentary bedrock (Meso-Cenozoic); 10) fluvio-lacustrine terrace edge; 11) trough; 12) V-shaped valley; 13) alluvial fan; 14) fault scarp; 15) fault scarp within the lower terraces; 16) Holocene normal fault; 17) Holocene normal fault reactivated during the 1915 earthquake; 18) Cross-section trace.

along the SW range front of the Velino Massif (e.g., Nijman, 1971; Raffy, 1979). Normal faulting appears to have been ongoing during the whole Quaternary and is still very active today. This is documented by (a) displacement and tilting of lacustrine formations, and slope deposit sequences, (b) progressive offset of young sediments, revealed by trench investigations of Holocene normal fault scarps, and (c) observation of coseismic and paleoseismic surface faulting events (Jan. 13, 1915, M 7 Avezzano earthquake and related paleoseismic studies; Oddone, 1915; Serva & alii, 1988; Galadini & alii, 1995; Michetti & alii, 1996; Galadini & alii, 1997; Figs. 39 and Table 1). These data show that the two major normal fault segments (Celano-Gioia dei Marsi and M. Magnola-M. Velino faults), are characterized by Middle Pleistocene to present slip-rates of 0.5 to 2.0 mm/yr, and total Quaternary offset in the order of 2000 m (Cavinato & alii, 2002; Roberts et al., 2002).

The range fronts bounding the Fucino basin are fault escarpments. The whole geomorphologic setting of

the basin shows a clear tectonic control. In particular, the Quaternary activity of the master normal fault zone at the NE border generated several flights of lacustrine terraces. Over the Quaternary, these were progressively uplifted, tilted and faulted, and younger terraces repeatedly developed in the down-thrown block. Therefore, sedimentation is mostly influenced by tectonics. In Figure 39, the Quaternary terraces are grouped into three major orders, namely "upper", "intermediate" and "lower terraces", separated by prominent fault scarps (Figure 40).

The "upper terraces" include two main terrace surfaces of Upper Pliocene (?) -Lower Pleistocene age. The highest one culminates at 1050 m a.s.l., and represents the top of the more ancient lacustrine cycle, as demonstrated also by wave-cut-terraces in the bedrock. The second one is faulted and reworked by a depositional surface of the Alto di Cacchia unit, culminating at 950 m a.s.l. (Figure 39; Blumetti et alii, 1993; Bosi et alii, 1995; Bosi et al., 2003). Two intersecting NW-SE and SE-NW trending normal

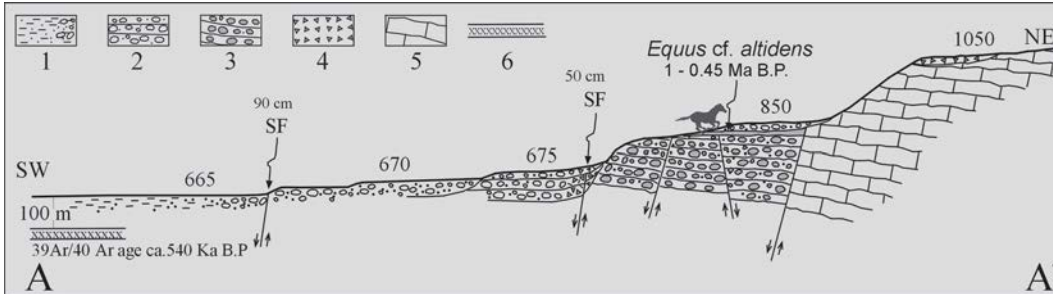


Figure 40 - Synthetic geological profile across the Quaternary terraces at the NE border of the Fucino Basin (location in Figure 39). The “intermediate” and “lower terraces” deposits are shown, whereas the “upper terrace” is here represented by an erosional surface with only a thin layer of deposits on it. Legend: 1) Holocene deposits of the Fucino Lake; 2) Upper Pleistocene to Holocene alluvial fan deposits; 3). Middle Pleistocene fluvial and lake deposits; 4). Pliocene?- Lower Pleistocene? breccias; 5) Mesozoic pelagic limestone sequence; 6) Dated tephra layer; SF). Surface faulting that occurred during the January 13, 1915 earthquakes.

faults border the “upper terraces” and generate fault scarps up to 100 meters high (Figure 39; Raffy, 1981-1982; Blumetti et alii, 1993). The “intermediate terraces” include two main Middle Pleistocene terrace surfaces. The higher one is divided

The “lower terraces” constitute the part of the basin at elevations ranging from 660 m to ca. 720 m a.s.l., where Upper Pleistocene to Holocene lacustrine and fluvio-glacial deposits crop out. These are both depositional and erosional surfaces (Raffy, 1981-

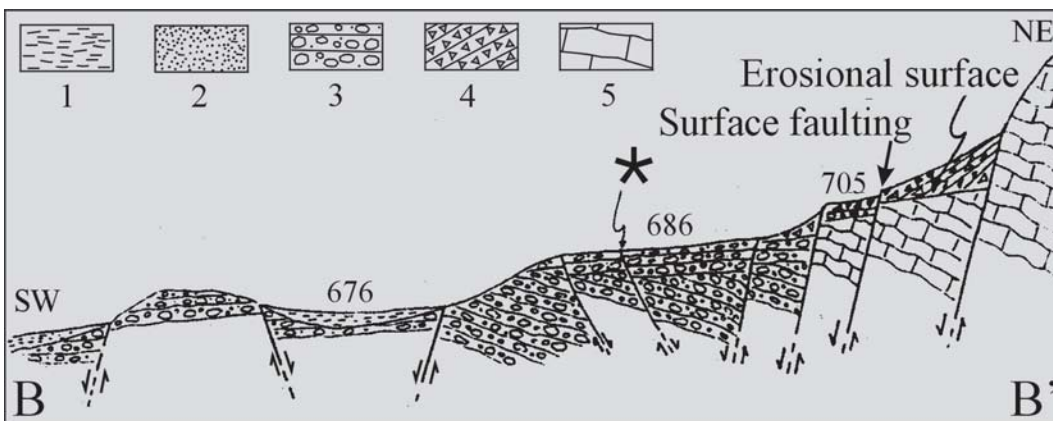


Figure 41 - Cross section in the “lower terraces” near Gioia dei Marsi (see location in Figure 39). Legend: 1) talus deposits (late Glacial to Holocene); 2) Lacustrine silt and clay (Holocene); 3) Fan delta gravel (late Glacial); 4) limestone bedrock.

by Bosi & alii (1995) into three orders “developed at base levels not very different from each other” (Bosi et alii, 1995; Messina, 1997). We will consider them as a single terrace, culminating at 850-870 m a.s.l. (Figure 40). Slightly entrenched in this terrace, there is a second surface that, in the Giovenco River valley and in the surroundings of the village of Cerchio, is about 800-830 m a.s.l. Both these terraces are limited to the SW by a major fault scarp up to 100 m high (Figure 40).

1982; Blumetti et alii, 1993; Giraudi, 1988). The central part of the basin between 649 and 660 m a.s.l. is the bottom of the historic lake. To date the above-mentioned lacustrine terrace, there are the following chronological constraints. The latest Pleistocene to Holocene evolution is very well defined by a wealth of archaeological, radiocarbon, and tephrochronological data (Radmilli, 1981; Narcisi, 1993). The works by Giraudi (1988) and Frezzotti & Giraudi (1992; 1995) provide an extensive review of

these data, building up a detailed paleoenvironmental reconstruction for this time interval. During this period, the maximum high stand of the lake is dated at ca. 20 to 18 ka B.P. It produced a prominent wave-cut terrace, well-preserved at several sites along the basin margins (Figure 41; Raffy, 1970).

Information on the 20 to 40 ka B.P. time interval mostly comes from radiocarbon dating of the Majelama fan stratigraphy. This shows (1) a fluvial sedimentary environment since ca. 30 ka B.P., (2) the formation of a thick paleosoil, developed from volcanic parent materials at 33 to 31 ka B.P. In the center of the basin, volcanic horizons originating from the Alban Hills district at ca. 40 to 50 ka B.P. are found 10 to 15 m below the ground surface (Narcisi, 1994). Pollen data show that in the same area the lake sediments deposited during the Eemian period are at ca. 60 to 65 m of depth (Magri & Follieri, 1989).

Chronological data for the period before the Late Pleistocene are very poor. The only available dating is a $^{39}\text{Ar}/^{40}\text{Ar}$ age of ca. 540 ka B.P. from tephra found at a depth of 100 m in the center of the basin (Figure 40; Follieri et alii, 1991).

New stratigraphic analyses within the “intermediate terraces”, recently led to the discovery of the first

species characterizes Italian fauna from the end of the Early Pleistocene, and is no longer documented during the Late-Middle Pleistocene. In particular, the biochronological distribution of this equid spans between the Pirro and Fontana Ranuccio Faunal Units (c.a. 1 Ma to 0.45 Ma; latest Villafranchian to latest Galerian in terms of Mammal Age). Therefore, the find of “*Equus cf. altidens*” seems to confirm the middle Pleistocene age of the Cerchio-Collarme-Pescina sequence (“intermediate terraces” in Figure 39).

Regarding the chronology of the “upper terraces”, two different interpretations can be found in the literature. According to Bosi et al. (1995, 2003), the Aielli formation is Pliocene in age, based on regional stratigraphic correlations. According to Raffy (1979), the Aielli formation is Late-Lower Pleistocene in age, based on the amount of volcanic minerals in the Aielli lake deposits, and on regional geomorphology.

Stop 3.2.2:

The Mt. Serrone fault escarpment

The Filippone Hotel is located at the base of the Mt. Serrone normal fault scarp (Figure 42). We have the opportunity here to have a close-up view of a landscape feature that is nearly identical to the Magnola, Velino, and Tre Monti fault scarps observed



Figure 42 - Panoramic view of the south-western flank of the Mt. Serrone fault escarpment. Halfway up the slope, the Holocene fault scarp, that was reactivated during the January 13, 1915 earthquakes, can be seen.

Middle Pleistocene mammal remains in the Fucino basin. This is a fragmentary maxillary bone with the teeth of an equid, found in September 1996, at Ponte della Mandra site (830 m a.s.l.; Figure 39), in a complex alluvial sequence of the Giovenco valley. The paleontological determination ascribes the fossil remains to “*Equus cf. altidens*”. This

during the first Stop. From the swimming pool of the hotel, it is possible to see one of the spectacular fault planes belonging to the Celano - Gioia dei Marsi fault. This bedrock fault scarp was reactivated during the 1915 earthquake, according to eye witnesses. Therefore, the geomorphic characteristics of the Mt. Serrone fault scarp can be used as a model for

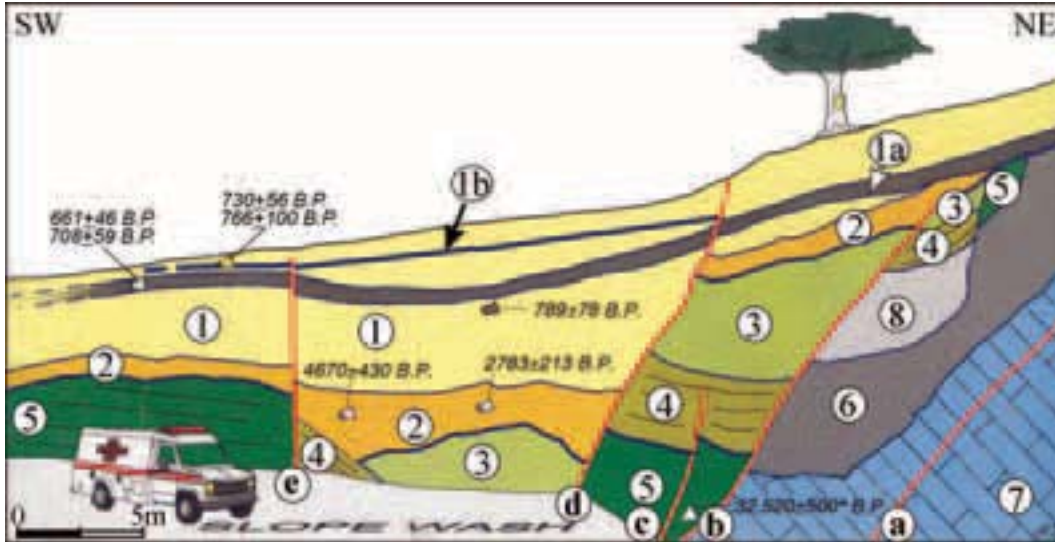


Figure 43 - Geological section of the Casali d'Aschi outcrop. Legend: 1), 2), 3), 4), 5), 6) Different levels composing a succession of colluvial and debris deposits (late Glacial to Holocene); 7) limestone bedrock; 8) not outcropping part; a), b), c), d), e) faults, (after Galadini et al., 1995).

understanding the evolution of similar landforms throughout the Apennines, which appears to be controlled by the repeated occurrence of strong recent earthquakes.

Route: We go back (North) along the Marsicana Road until we reach the Casali d'Aschi cross (8 min - 4 km); we turn right (East), and enter the Casali d'Aschi village (4 min - 1 km), where we will stop in a quarry at the base of the mountain slope.

Stop 3.2.3:

Displaced deposits of Casali d'Aschi and overview on the paleoseismology of the Fucino Basin

The quarry exposures allow detailed observations of the 1915 earthquake fault zone at the contact between the limestone bedrock and the slope waste deposits (Figure 43). The slope deposit stratigraphy has been very well studied in this area, providing reliable chronological constraints for dating the last movements of the fault. Thanks to the finding of Neolithic and Middle Bronze Age pottery fragments in level 2, and of other pottery fragments of the Middle Ages period in levels 1 and 1a, (all dated through the thermoluminescence method), four late Holocene surface faulting events have been detected at this site by Galadini et al. (1995). The last event is most likely due to the 1915 earthquake, while other two events appear to have occurred during the Middle Ages, in agreement with the San Benedetto trenching

site (Michetti et al., 1996).

In the following section we will synthesize the data that have emerged from the many paleoseismic analyses carried out in and around the Fucino Basin. This will lead to some considerations on the growth of the Fucino tectonic structure by repeated strong earthquakes.

Along the eastern border of the Fucino basin, and its NW extension in the Ovindoli and Piano di Pezza area (Salvi and Nardi, 1995; Pantosti et al., 1996), the Holocene paleoseismology and deformation rates have been investigated at several sites along the trace of the Celano - Gioia de' Marsi (sites 2, 3, 4, 5, and 12 in Figure 38; Michetti et al., 1996; Galadini et al., 1995; Galadini et al., 1997b), and the Parasano - Cerchio faults (site 1 in Figure 38; Galadini et al., 1995; Galadini et al., 1997b) and Ovindoli - Pezza faults (sites 13, 14, and 15 in Figure 38; Pantosti et al., 1996), as described in Table 1.

It is possible to view these paleoseismological results in terms of the variation in deformation rates and earthquake recurrence along a single tectonic structure. For instance, the extension rates vary with distance from the center of the Fucino basin. In fact, as already pointed out, most of the eastern part of the basin is bounded by two parallel normal faults, the Parasano - Cerchio fault (or Marsicana Fault of Galadini et al., 1997b), and the Celano - Gioia de' Marsi fault, both reactivated during

the 1915 earthquake (Oddone, 1915; Serva et al., 1988; Galadini et al., 1995), and showing Holocene displacement. If extension rates observed along these two traces, using the San Benedetto trenches (1.0 to 1.6 mm/yr; site 12 in Figure 38; Michetti et al., 1996) and the Marsicana trenches (0.4 to 0.5 mm/yr; site 1 in Figure 38; Galadini et al., 1997b), are summed up, the cumulative value near the center of the segment is 1.4 to 2.1 mm/yr for a 45° dipping fault. This is significantly higher than the value at the NW termination of the Fucino structure in the Ovindoli - Piano di Pezza area, where a maximum extension rate of 1.0 to 1.2 mm/yr can be derived for a 45° dipping fault plane. The data in Table 1 also indicate that the variation in extension rates is probably due to a higher frequency of earthquakes per unit time at the center of the fault compared to its NW termination.

It is important to note that the NW lateral termination of the Fucino tectonic structure occurs within a high mountain area where Mesozoic carbonates outcrop in the hangingwalls of the Quaternary faults. We interpret this area as a transverse bedrock ridge. The Quaternary throws on the faults are less than a few hundred meters in this location (Nijman, 1971; Salvi and Nardi, 1995). In contrast, the center of the Fucino basin is marked by the juxtaposition of Mesozoic carbonates and Neogene - Quaternary sediments across the faults. As already pointed out, Quaternary fault throws in this location are in the order of 1 - 2 kilometers. Therefore, the pattern of coseismic Holocene deformation, as recorded at the trench sites indicated above, is in good agreement with the Quaternary geological and geomorphic setting of the Fucino structure and its NW termination in the Ovindoli - Piano di Pezza area. This strongly suggest that the growth of the Fucino extensional structure can be interpreted as the cumulative effect of many earthquake rupture sequences throughout the Quaternary.

Other capable faults have also been ruptured by Holocene earthquakes in the Fucino basin. Galadini et al. (1997b) have trenched 2 other faults in addition to the Celano - Gioia and Parasano - Cerchio Faults and, assuming that the faults dip at 45°, the implied rates of horizontal extension are of ca. 0.4 - 0.5 mm/yr across the Trasacco and Luco de' Marsi faults. Therefore, the overall extension rate in the Fucino tectonic structure during the Holocene might be in the order of 3 to 3.5 mm/yr. We propose the following evolutionary model for the Fucino Basin.

The location of the southern and western margins

of the ancient lake which occupied the Fucino basin before the latest glacial is unknown. High continental terraces are stranded in the footwall of the Celano-Gioia dei Marsi normal fault, at elevations up to 1050 m a.s.l.. Most likely lacustrine sediments of the same age are buried below the modern deposits in the hangingwall of this master fault. Seismic reflection data from Cavinato et al. (2002), clearly show that continental deposits are several hundreds of meters thick toward the San Benedetto Fault. On the southern and western borders of the Fucino Basin, we can observe only the main younger terrace (at c.a. 720 m a.s.l.), which forms a narrow banquette at the foot of the mountain slopes. Since all the available data indicate that the Fucino basin was an endoreic, closed depression over the whole Quaternary, it is very difficult to believe that erosional processes could have obliterated any trace of the previous terraces. We can conclude that the Fucino Basin extended progressively to the west and to the south, following the continuing Quaternary hangingwall subsidence of the Celano-Gioia dei Marsi normal fault segment. The most spectacular evidence of this process was the geomorphic changes observed during the Jan. 13, 1915, Avezzano earthquake. Therefore, (a) the sequence of lakes that occupied this depression, (b) their size, and (c), the related landforms (fault scarps, flights of terraces) and deposits, all appear to be mostly controlled by active extensional tectonics capable of producing strong seismic events.

DAY 4

Stop 4.1:

1980 Southern Italy earthquake: surface faulting, ground fracturing, landslides and lateral spreading phenomena

Blumetti A.M. & Michetti A.M.

The Irpinia-Basilicata earthquake of the 23 November, 1980, was one of the largest normal-faulting events in the Apennines chain ($M_s = 6.9$ NEIS, and seismic moment $M_0 = 26 \times 10^{18}$ Nm, Westaway, 1993; epicentral intensity $I^0 = IX-X$ MCS, Postpischl, 1985). The main shock took place at latitude 40.724°, ± 1.4 km, and longitude 15.373°, ± 1.4 km, and its nucleation point was at 10-12 km of depth. Three different fault segments ruptured during the main shock, as deduced by the 0, 20, and 40 seconds subevents seen on the seismograms.

The earthquake heavily damaged more than 800 localities, mainly located in the Campania and Basilicata regions, killing ca. 3,000 people

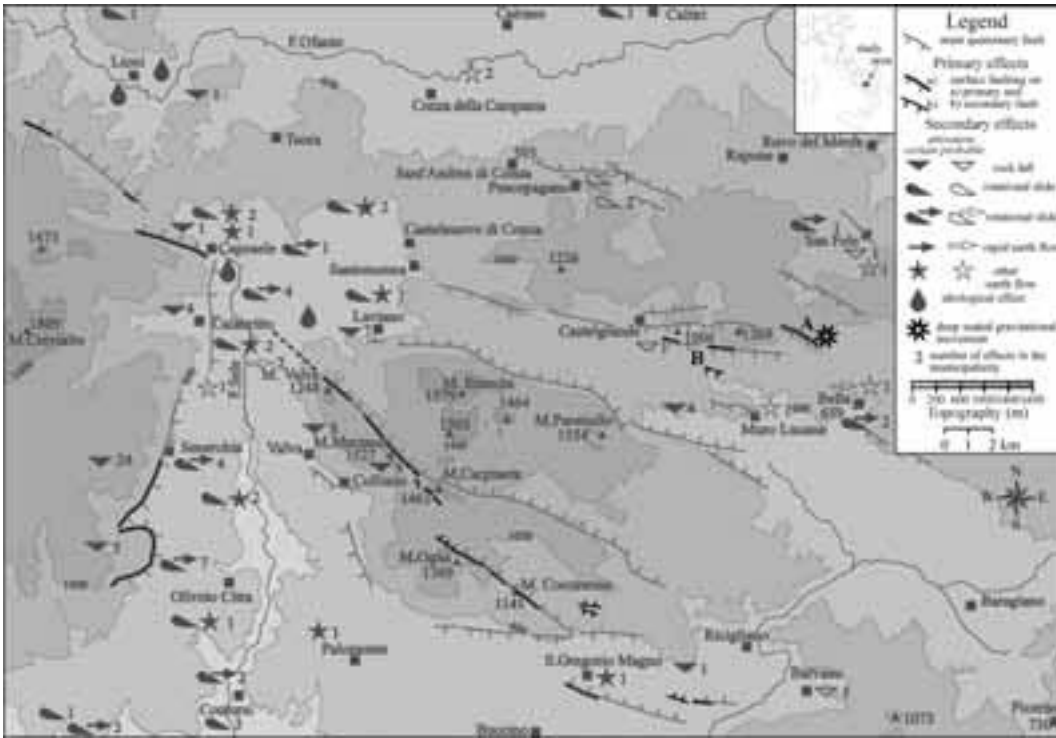


Figure 44 - Epicentral area of the November 23, 1980, event: locations of ruptured fault segments and earthquake-induced ground effects A and B, indicate locations for surface faulting as described in the text.

(Postpischl, 1985).

The high magnitude, and the local geomorphological setting, determined widespread effects on the physical environment. Many surface fractures were mostly located within the VIII isoseismal, specially focused

in the epicentral area. More than 200 earth slides, distributed over an area ca 20,000 square km wide (Esposito et al., 1998), and at least ten liquefaction events, were reported either in the near and the far field (Galli and Ferrelì, 1995). Many hydrologic



Figure 45 - Costa Pannicaro fault-generated range front. Arrows indicate the location of the fractures mapped by Carmignani et al. (1981).



Figure 46 - Costa Monticello site. a) panoramic view of the sackung; b) details of the trench in the Mesozoic limestone bedrock; c) free-face due to primary surface faulting at the base of the bedrock fault scarp near Costa Monticello. Mr. Ferracane (left) witnessed the event and provided invaluable information.

anomalies in the large carbonatic aquifers of this sector of the Apennines were observed before, during, and after the shock. Almost 70 springs with mean discharge rates higher than 500 liters/second showed a strong discharge and there were other anomalies, mainly located in the high Sele River valley and the Matese area (Esposito et al., 1999).

Soon after the earthquake, several groups mapped the coseismic geological effects which covered a large part, but not the whole epicentral area. Some of them didn't believe that any of those effects could have been of a tectonic nature, and ascribed all of them either to pure gravitational sliding or to debris compaction (Carmignani et al, 1981). Figure 44

Other Italian researchers interpreted as tectonic the roughly 7 km long, till 50 cm down-thrown

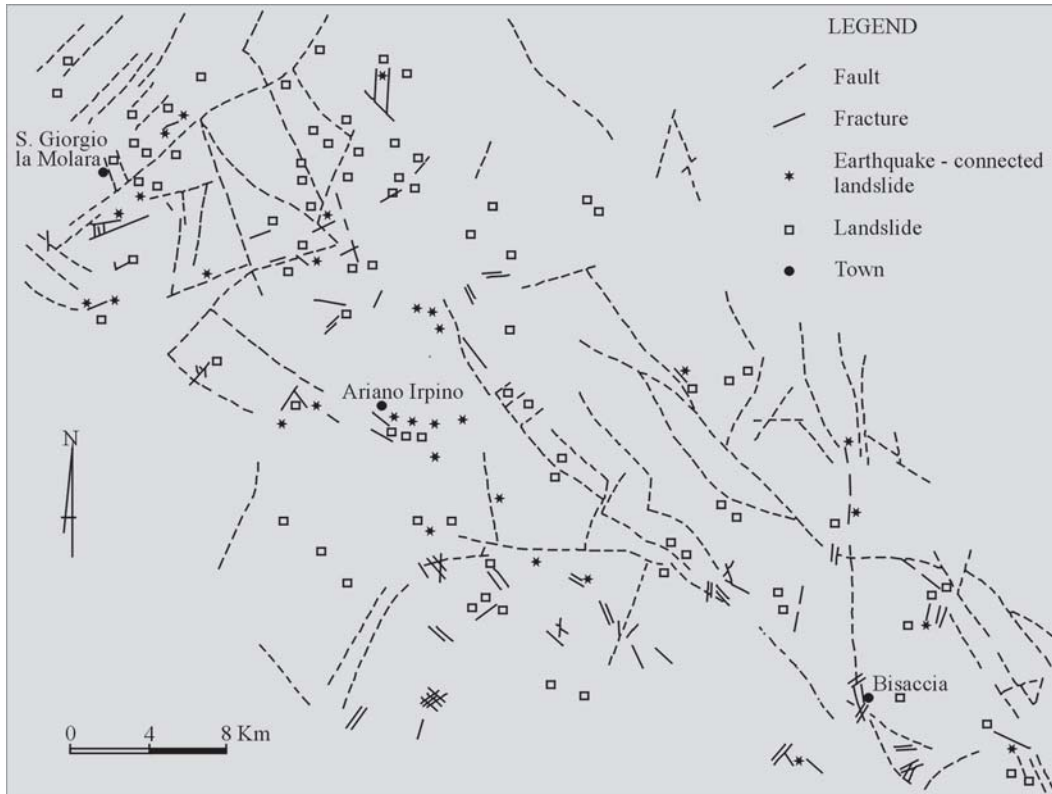


Figure 47 – Earthquake-induced ground fracture in the Ariano Irpino area.

reactivation that occurred along the western border fault of the Sele Valley; the about 2 km long fracturing that occurred around Piano di Pecore, on the Marzano ridge (Cinque et al., 1981); and the roughly same length fracturing that occurred in Pantano San Gregorio Magno (Bollettinari & Panizza, 1981). This idea was totally rejected by most parts of the Italian scientific community.

In 1984, Westaway and Jackson recognized as a surface faulting a part of the rupture that occurred along the Marzano ridge, and since then, the Italian scientific community started to accept the idea that surface faulting could also occur in the Apennines. In successive papers (Westaway and Jackson, 1987; Pantosti and Valensise, 1990), two segments, with total lengths of about 40 - 48 km, were reconstructed, the first running along the base of the northern border of the Picentini mountains, and the second cutting the Marzano ridge and reaching, in several steps, Pantano San Gregorio Magno. These surface faulting effects, striking 300-315°, prevailing dipping to the northeast and with a vertical offset, ranging 40 to 100

cm (Westaway, 1993; Pantosti & Valensise, 1993), on the whole, are considered to be primary tectonic effects related to the 0 and 18s sub events. More difficult was the identification of the fault responsible for the 40 sec event, notwithstanding the comparable magnitudes (Mw 6.2-6.5, 6.4 and 6.3 for the 0, 20 and 40 sec respectively; Westaway, 1993).

Stop 4.1.1:

The Bella slide and the antithetic surface faulting from the 1980 Irpinia-Lucania earthquake

Recently the geological effects produced by the 23 November 1980 Irpinia - Lucania earthquake have been revised and reinterpreted (Blumetti et al., 2002). The comprehensive revision of the literature on the geological phenomena induced by the November 23, 1980, earthquake, allowed us to draw the map in Figure 44, including its surface faulting, soil fracturing, landslides, and deep-seated gravitational slope deformations.

As for surface faulting effects, other than published reports (including Cinque et al., 1981; Bollettinari &



Figure 48 - Earthquake-induced ground fracture cutting a house

Panizza, 1981; Carmignani et al., 1981; Westaway & Jackson, 1984; Pantosti & Valensise, 1990) were reconsidered the original field maps drawn immediately after the earthquake by Carmignani et al. (1981). Field inspection, coupled with new eyewitness reports, lead to a substantial reinterpretation of the available information, indicating that surface faulting and fracturing in the epicentral area of the Irpinia earthquake was much more important and significant in terms of tectonic and paleoseismic interpretation, than had been reported before.

In particular, a set of ground fractures and fault scarps, mapped by Carmignani et al. (1981) at the Costa Pannicaro site (see arrows in Figure 45), between Muro Lucano and Castelgrande, each of them a few hundred meters long, clearly represents evidence for surface faulting, being associated with a down-to-the-SW vertical displacement of 10 to 20 cm for a total length of ca. 4 km (location B in Figure 44). At the nearby Costa Monticello site (location A in Figure 44), eyewitnesses described the coseismic reactivation of (a) a sackung, with the formation of a trench in the limestone bedrock of the local mountain

slope, ca. 200 m long and up to 2 m wide and 4 m deep (Figs. 46A and 46B), and (b), a SW-dipping limestone

bedrock fault scarp at the base of the local mountain range front, with the formation of a ca. 20 to 30 cm free-face (Figure 46C) for a length of some kilometers. The entire surface effects described above are located along a set of parallel faults which belong to the same tectonic structure (Figs. 44 and 45).

Assuming the whole structure broke during the 1980 earthquake, the end-to-end rupture length would be of ca. 8 km. Considering that the Costa Pannicaro and Costa Monticello fault traces lie very close to the surface projection of “the likely 40-s fault plane” interpreted by Bernard & Zollo (1989; see their Figure 10), based on levelling and seismological data, this surface rupture might have been associated with the sub-event that occurred at 40 seconds along a SW-dipping fault ($M_s = 6.3$, Westaway, 1993).

As regards the deep-seated gravitational deformation, we stress that such huge gravity-driven phenomena are very often earthquake-induced. As much as the surface faulting effects, with which they are typically closely associated, their repetition at each strong event leaves a strong mark on the landscape of the Apennines.

DAY 5

Stop 5.1:

Gravitational phenomena triggered by the 1980 Southern Italy Earthquake

Dramis F., Gentili B. and Pambianchi G.

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 has been possible to survey directly, with more scientific methods, the surface effects of strong earthquakes, also comparing them with instrument records of the shock. In this way, it has been possible to understand (Radbruch-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 to the characteristics of the shock, particularly with Arias intensity (Arias, 1970). It has also been outlined that many earthquake-induced mass movements are



Figure 49 - The high escarpment which divides the old town from the modern one (out of the picture, on the right side). Secondary escarpments are visible within the built-up area. All of them can be interpreted as landslide scarps.

also connected with other seismic ground effects (such as fracturing and faulting).

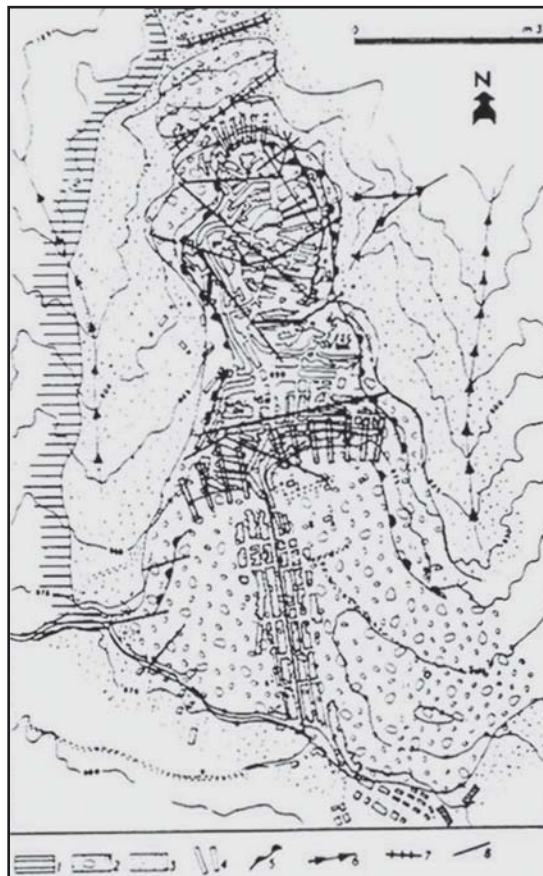
As far as lithological features are considered, the importance of identifying “engineering geological formations”, has to be stressed (Cotecchia, 1978; Canuti et al., 1988); research to this end is presently being carried out throughout the Italian territory.

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 the saturation of terrain, variations of piezometric level, etc.).

Figure 50 - Simplified geomorphological map of the Bisaccia area. Legend: 1. Varicolored Clays; 2. conglomerates; 3. debris; 4. main landslide escarpment; 5. edge of the conglomerate platform scarp retreating through mass movements; 6. stream erosion; 7. trench; 8. minor landslide scarps and fractures reactivated by the November 1980 earthquake (after Crescenti et al., 1984, modified).



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). These kind of landslides (Tinsley et al., 1985), often involve deep-seated beds too, disturbing very large areas far away from the epicenter.

Cases of Large-Scale Landslides Induced by the 1980 Earthquake in Southern Italy

The most outstanding phenomena triggered by the 1980 earthquake were mass movements of different types (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; Dramis and Blumetti, in press), at least partially determined by the quite high relief of the area, the poor geotechnical characteristics of most of the outcropping rocks, and the high water content of the terrains (due to heavy rainfall in the days preceding the seismic event). These movements often appeared to be connected with ground fractures which were widespread, both isolated or joined in groups (Figs. 49 and 50), up to several kilometers long, also relatively far from the epicentral area of the earthquake (Cantalamessa et al., 1981; Genevois and Prestininzi, 1981; Dramis et al., 1982; Bisci and Dramis, 1993).

The 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.).

Stop 5.1.1:

The Bisaccia landslide

An important mass movement triggered by the 1980 southern Italian earthquake was the deep-seated sliding (Crescenti et al., 1984), that involved most of the town of Bisaccia (Figure 49), located quite far away from the epicenter (the earthquake here reached only a VII MCS intensity). This town is built up over a terrace-like platform made up of a 50 m thick Pliocene-Quaternary, polygenic, conglomerate overlying strongly-disturbed allocthonous clays (Varicolored Clays - Argille Varicolori formation) of Miocene age.

A 40 m high escarpment divides the historical town into two parts: the historical center (in the lower sector), and the modern sector (in the upper one)

(Figs. 49 and 50). At its foot a 20 m deep trench, partly overlying the clayey substratum, is present. The filling materials are quite recent, as testified by the finding of masonry fragments near the bottom by exploration boreholes.

Immediately downslope of this depression, there is a 13th century tower, deeply emplaced in the conglomerate bedrock, and strongly tilted upslope: On the clayey slopes bordering the conglomerate platform, counterslopes and depressions are frequent.

The geomorphological framework of the town area can be interpreted as that of a deep-seated multiple rotational slide, within a mass involved in a slow lateral-spreading process. It caused the fragmentation of the conglomerate platform into blocks more or less turned counterslope, as well as that of the high escarpment and the trench at its base (Crescenti et al., 1984).

All along the historical center, coseismic ground fractures and scarplets, mostly corresponding to the margins of the conglomerate blocks, have been recognized. They experienced recurrent reactivation on the occasion of past seismic events, as clearly testified by detailed maps of surface effects carried out by the local municipality immediately after the strong earthquakes of 1930 and 1962.

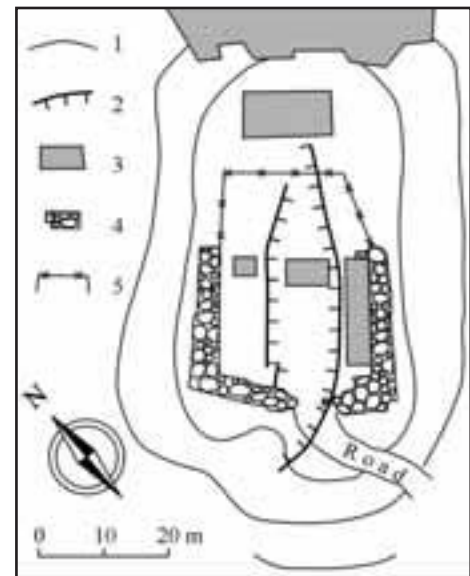


Figure 51 - Sketch of the graben-like depression:
1 - approximate contour line; 2 - trench scarp;
3 - building; 4 - castle ruins; 5 - fence
(after Dramis and Sorriso-Valvo, 1983).



Figure 52 - Scarplet bordering to the south-west of the graben-like depression within the Trevico lateral-spread.

Damage to the built-up area was generally not extreme, even if widely diffused. In fact, many artifacts were simply tilted together with the underlying conglomerate blocks, without suffering a complete destruction. The most relevant disruptive effects occurred in connection with the earthquake reactivated ground fractures and scarplets, and all along the edges of the conglomerate platform, where a number of buildings collapsed, as a consequence of local mass movements (mostly falls and slumps).

Stop 5.1.2:

The lateral spreading of Trevico

Quite different is the deep-seated lateral spreading (Dramis and Sorriso-Valvo, 1983; Carton et al., 1987) which affected the village of Trevico, located at 1089 m a.s.l., on top of a hill made of 200 m thick Pliocene-Quaternary conglomerate with sandy levels overlying Upper Miocene sandy clays.

This phenomenon, typical of high relief hills modeled in solid bedrock (such as limestone, sandstones,

Figure 53 - The observatory entrance stairs cut by the Figure 54 scarplet.





Figure 54 - A photograph likely taken after the 1930 earthquake, where it is possible to observe the same fracture observed in 1980 cutting the stairs of the observatory building.

conglomerates etc.) (Dramis et al., 1995), consisted of the deepening of a small graben-like depression (which is some 50 m long, about 15 m wide, and 2 m deep), on top of the hill where a military meteorological observatory is located (Figure 51). Immediately after the earthquake, superficial ruptures (scarplets up to some dm high), were observed on both sides of the depression (Dramis and Sorriso-Valvo, 1983). One of these cut into the observatory bordering wall and the entrance stairs of a building (Figs. 52 and 53).

As also hypothesized for other similar phenomena, the gravitational deformation could be the effect of seismically-induced oriented accelerations, even if other mechanisms (probably connected with the presence of water) contributed to the genesis of this deep-seated gravitational deformation.

As reported by local inhabitants, also this movement experienced in the past, had recurrent reactivation during

the previous seismic events (Dramis and Sorriso Valvo, 1983). A striking evidence of the past occurrence of the same ground effects is provided by an old photograph (Figure 54), most likely taken after the 1930 earthquake, where it is possible to observe the same fracture observed in 1980, cutting the stairs of the observatory building.

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