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FOURTH INTERNATIONAL CONFERENCE ON GEOMORPHOLOGY - Italy 1997

Guide for the excursion

GEOMORPHOLOGY AND QUATERNARY EVOLUTION OF CENTRAL ITALY

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INTRODUCTION

The excursion will start from Ancona and will continue in the inner portion of the central-southern Marche up to the end of the second day. The two following days will be devoted to the northeastern (Campo Imperatore - Gran Sasso) and southwestern (Fucino basin) sectors of the Abruzzi Apennines. During the last day the Sabatini Mts. area (Southern Latium) will be illustrated. During the excursion, some large areas, on the whole homogeneous from a physiographic point of view (fig. 1), will be crossed:
- the low-hills of the periadriatic belt;
- the medium-high to high hills of the central-southern Marche;
- the mountainous massifs and ridges of the Apennines, hosting several tectonic depressions;
- the volcanic complex of the Sabatini Mts.

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For the above areas, some of the main geomorphological features will be illustrated, such as large-scale gravitational movements, morphotectonic evolution of both the periadritic belt and the intra-Apenninic depressions, hypogean karstic phenomena, stratified slope-waste deposits and their relation with terraced alluvial deposits, Holocene fluvial dynamics, glacial and periglacial morphogenesis in the Apennines, and volcanic reliefs.

GEOMORPHOLOGICAL FEATURES OF CENTRAL-SOUTHERN MARCHE

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BEDROCK GEOLOGY

The geological evolution of the region started from the Tortonian, as a consequence of compressional tectonics associated to an arc-trench system migrating from the Tyrrhenian to the Adriatic Sea (CENTAMORE & DEIANA, 1986). Along the Apennines, compressional tectonics reached its climax during the upper Messinian, resulting in the first emersions, an ended in the Lower Pliocene. More recently, it continued (even though with limited effects) in the eastern portion of the region, where during the Lower Pliocene - Pleistocene a thick (up to 2300 m) marine succession was deposited, lying unconformably on the older terrains (folded and eroded) and following a gently dipping monocline. Compressional tectonics is still active along the Adriatic coast (DRAMIS & SORRISO-VALVO,

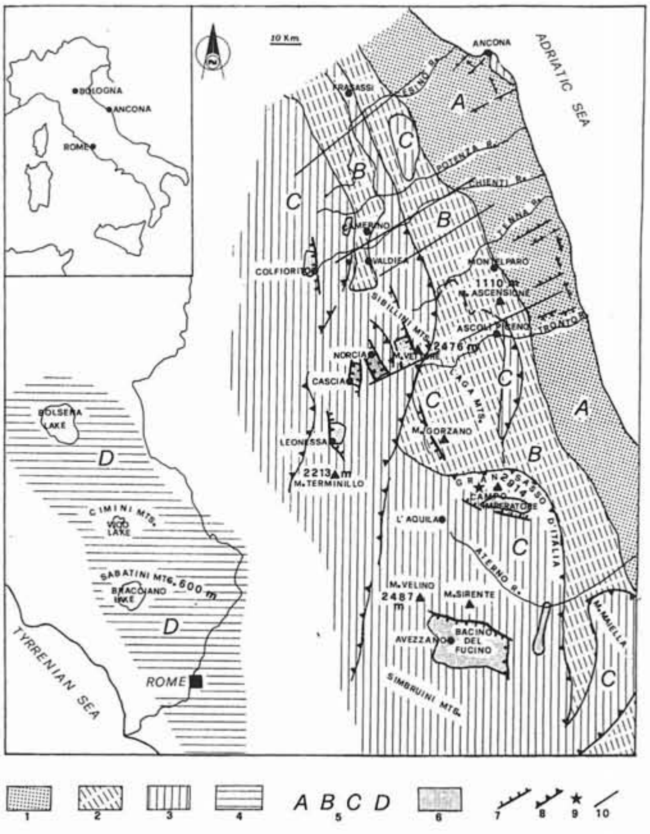

FIG. 1 - Simplified litho-structural and morphological setting of Central Italy. Eastern area with mainly clayey (1) and torbiditic sediments (2); Apennine chain mainly calcareous (3); western area with chiefly terrigenous and volcanic deposits (4). 5): periadriatic belt (A), medium-high and high hills belt (B); Apenninic belt (C); volcanic complex (D); 6) tectonic depression; 7) normal faults; 8) overthrusts; 9) the Calderone glacier; 10) main antiapenninic faults.

1994), as demonstrated by the focal mechanisms of earthquakes recorded offshore (RIGUZZI & *alii*, 1989).

The compression was followed by an extensional phase characterized by normal faults, lowering towards the Tyrrhenian Sea and reactivating at different depths compressional shear planes (PIZZI, 1992; CALAMITA & *alii*, 1994; BIGI & *alii*, 1996). This second phase was associated with a generalized uplift. Starting from the Lower Pleistocene, the latter strongly increased, involving also the outermost periadriatic belt (DUFAURE & *alii*, 1989; DRAMIS, 1992). This phenomenon, connected with a vaster uplift responsible for the formation of the whole Apenninic belt (DEMANGEOT, 1965; DUFAURE & *alii*, 1989; DRAMIS, 1992), gave rise to the overall monoclinal structure of the Pliocene-Pleistocene sediments (reaching 1100 m a.s.l. at Mt. Ascensione). Also the genesis of the main intra-Apenninic depression, whose older filling sediments are referred to the Lower Pleistocene (BLUMETTI & DRAMIS, 1993) is connected to the extensional tectonics. The breaching of these depressions and their integration in the hydrographic network, both generally referable to the Middle Pleistocene, is a consequence of headward fluvial erosion triggered by the uplift (DRAMIS, 1992; BLUMETTI & DRAMIS, 1993).

MAIN PHYSIOGRAPHIC AND GEOMORPHOLOGICAL FEATURES

The periadriatic belt

This area is characterized by low hills, transversally crossed by wide fluvial valleys and delimited to the East by wave-cut cliffs, either active or bordered by narrow beaches. There, Pliocene-Pleistocene marine sediments, made of by alternations of pelitic, pelitic-arenaceous, arenaceous-pelitic and arenaceous layers passing to the top to sandy conglomeratic deposits (Sicilian-Crotonian), crop out. On the whole, the layers dip gently towards the sea, even though locally (close to the shoreline) folded structures (coastal structures) have been found. These deposits are often crossed by both Apenninic and antiapenninic faults; moving inland, the closure deposits are progressively more uplifted, up to some hundred meters (500 m a.s.l. at Ripatransone, some 8 km from the shoreline).

Almost everywhere, slopes are covered by eluvial colluvial silty-clayey deposits which are particularly thick along the gentlest slopes and at least partially derive from wind-carried material (FARABOLLINI, 1995).

Among active morphogenetic processes, the most conspicuous are gravitational phenomena, such as flows (abundant along moderately steep slopes), falls and large slides (typical of wave-cut cliffs). The huge Ancona landslide, activated on 13th December 1982 (COLTORTI & *alii*, 1984), certainly represents the most representative example of the latter, set up on a deep-seated slope deformation (DRAMIS & SORRISO VALVO, 1994).

The medium-high to high hills belt

In this portion of the Marche, altitudes generally range between 200 and 700 m a.s.l., even though they may reach over 1000 m (1109 m at Mt. Ascensione), with a relief of some hundred meters (up to 500 m and more).

Reliefs are arranged according to alignments trending parallel with the Apennines and are modelled into the alternations of the Pliocene and Messinian (Laga Formation) turbiditic sediments (pelitic, pelitic-arenaceous, arenaceous-pelitic, arenaceous and arenaceous-calcarenitic-conglomeratic associations). The structure is rather complex for the oldest terrains, affected by folds having different radius, faults and overthrusts; on the other hand, Pliocene deposits (generally in discordance on the previous units) follow a regular monocline with a dip generally ranging around 15°-20° and progressively diminishing eastwards, closer to the coast. Also the layer direction varies from ENE to NE. At the surface, the structure therefore shows the overall configuration of a double flexure with upward concavity, whose main and secondary axes trend parallel to the Apennines and EW, respectively (INVERNIZZI, 1992; DEIANA & PIALLI, 1994). In the area, low throw normal faults together with two sets of joints connected with the double flexure (N70-N150; N20-N100) are present; a third set of fractures trending ca. NS and EW, generally affecting the stiffer layers, has been attributed to gravitational phenomena (GENTILI & *alii*, 1995).

The area is deeply cut by a hydrographic network formed of almost symmetrical major valleys, crossing transversally to the Apennines and markedly asymmetric minor ones. This gives rise to a typical overall morphostructural settlement with cuestas (GENTILI & *alii*, 1995) and widespread badlands along the steeper slopes facing between South and West (DRAMIS & *alii*, 1982; BISCI & *alii*, 1992).

The portion of the Apennines included here is made up by a complex system of folds, faults, overthrusts and, subordinately, backthrusts, both outcropping and buried. Folds constitute the most conspicuous structural features, even though the main structural elements are the overthrusts (fig. 2). In fact, the folds caused the overlying of calcareous and marly-calcareous Mesozoic units on the Messinian turbidites, thus giving origin to the arcuate belts of the Umbria-Marche-Sabina-Marsica Apennines and of the Gran Sasso d'Italia, respectively verging to the east and to the north (fig. 1).

The Apenninic belt

Besides some minor belts, two main belts (trending ca NW-SE and reaching altitudes ranging between ca. 1000 m and more than 2000 m a.s.l.) are present: the Umbria-Marchean and the Marchean ones. To the South, they merge in the Sibillini Mts. massif, reaching 2476 m a.s.l. (Mt. Vettore). More to the South, there follow the Laga Mts, reaching comparable altitudes (Mt. Gorzano, 2458 m a.s.l.), and the Gran Sasso d'Italia Massif, trending ca. WNW-ESE and made up by reliefs often exceeding 2500 m a.s.l. (including the highest peak in the Apennines with 2912 m a.s.l.). Relief is eveywhere high, often exceeding 1000 m (1400 m along the northern slope of the Gran Sasso), and slope angles are generally rather high.

The bedrock is generally constituted by stratified or massive limestone, marly limestone and cherty limestone, alternating with calcareous marls and marls (upper Trias - middle Miocene), and by sandstone in strata and banks separated by thin pelitic layers (Laga Formation, Messinian). The bedrock follows anticline structures whose settlement is complicated by overthrust planes trending towards the Adriatic, normal faults trending parallel to the Apennines and transversal faults, often transcurrent.

The bedrock often crops out or is covered by thin layers of superficial deposits, except for some, thick but not large, travertine outcrops. Along calcareous slopes stratified slope-waste deposits are frequent, regularizing large areas (COLTORTI & DRAMIS, 1995). The presence of brittle rocks (limestone and sandstone) overlying plastic layers (marl or clayey marl), tectonic discontinuities (faults, overthrusts and backthrusts), highly shattered layers of brittle rocks (sometimes strongly karstified), high relief (with higher values along the gorges transversally cutting the structures) and elevated seismicity, all together favored the genesis of deep-seated gravitational slope deformation (DRAMIS & *alii*, 1995). About 500 of these phenomena, characterized by an average extension of ca. 2 km^2 (up to 20 km^2) have been recognized, with a frequency of 0.06 phenomena per km^2 and a density of 0.12 (FARABOLLINI & *alii*, 1995).

In the described sector, several tectonic depressions are present (fig. 1), filled with thick lacustrine successions, having ages generally ranging between the Lower Pleistocene (Upper Pliocene?) and the Middle Pleistocene, followed by more recent continental (slope, fluvial) deposits. One of the above depressions is the Colfiorito basin, spreading for some 26 km^2 at an altitude of 750-880 m a.s.l. and bordered by reliefs ranging approximatively from 900 to 1200 m a.s.l.

The filling deposits, whose overall thickness exceeds 100 m in the central portion of the depressions, are constituted by alternations of silty-clayey and gravelly-sandy layers, with horizontal bedding. Along the borders, they pass to slope debris and alluvial fan deposits. Ancient landslide bodies (mostly regularized by slope debris) and rare deep-seated gravitational deformations are also present on the bordering slopes.

The hydrographic systems

The main rivers of central Italy reaching the Adriatic Sea originate from the highest areas of the Umbria-Marche Apennines (fig. 1). They cross the Marche region flowing WSW-ENE with a generally not very long course (only the River Tronto exceeds 100 km in length). Four main orders of alluvial deposits can be observed inside the fluvial valleys. Close to the coast, those terraces are entrenched in the Sicilian-Crotonian marine deposits. They are generally

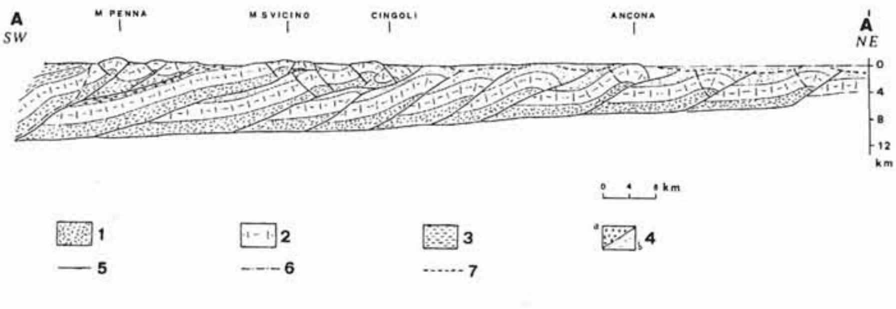

FIG. 2 - Cross section of the Umbro-Marchean Apennines (modified from CENTAMORE & DEIANA, 1986). 1: Trias evaporites and sedimentary basement; 2: carbonate sequence; 3: Miocene sequence in the Umbrian pedeapennine, 4: Miocene sequence of the terrigenous basin inside the Apennines (a); Neogene-Quaternary sequence of the eastern sector (b); 5: overthrust; 6: normal fault and associated faults; 7: lower-middle Pliocene transgression.

gravelly and have been produced by the interaction of cold and arid climatic phases with the general deepening of linear erosion following the tectonic uplift of the area. Terraced deposits connected with eustatic variations of sea level are present along the periadriatic belt, close to the coast (DRAMIS & CANTALAMESSA, 1993).

Some of the Marchean rivers (Esino, Potenza, Chienti and Tenna) will be visited during the excursion; there, relationships between terraced alluvial deposits and stratified slope-waste deposits, phenomena of fluvial erosion and deposition, and anomalies of the hydrographic network in the time range Middle Pleistocene - Holocene will be illustrated. Moreover, various levels of hypogean karst in the calcareous bedrock will be shown, correlated with the evolution of the hydrographic network (Frasassi Caves).

Recent uplift and fluvial cutting in the Adriatic piedmont of the Marche Apennines

The overall topographic setting of the Adriatic piedmont is primarly the result of selective erosion wich affected to a different degree the limestone Mesozoic structures, the Miocene sandstones and the conglomeratic, sandy clayey Plio-Pleistocene sediments. Such conditions are hardly favorable for the preservation of continental forms and formations. However, a precise cartographic study shows a considerable number of Quaternary levels which can be used to quantify the recent uplift of the Central Apennines.

On the external margin of the ridge, between Mt. Amandola - Mt. Banditello (Sibillini Mts., west of the town of Amandola), four piedmont benches are found at different elevation from about 1300 m to less than 900 m a.s.l. The two higher ones have been preserved from erosion because of the resistance of the limestone (Scaglia Rosata) bedrock (fig. 3). All the river terraces in the area are visibly entrenched in the lowermost piedmont level (RASSE, 1993-94).

Altitude measurements and geomorphological observations allow us to assume that the Sibillini Mts have undergone an uplift of more than 1000 m since the lower Pleistocene. Eastern tilting of the piedmont caused a consequent river system flow east, complicated by diversion and piracy phenomena, such as at Pian di Pieca (Chienti River basin).

ENVIRONMENTAL GEOMORPHOLOGY

Deep-seated gravitational slope deformations in the Apennine range

Along fault scarps and steep erosional slopes, deep-seated gravitational deformations and large scale landslides (up to several square kilometers wide) may be observed. These phenomena are very common in the axial part of the Apennines (fig. 1, C), which was affected by strong uplift and extensional tectonics during Late Quaternary. The predisposing factors were identified in the recent tectonic uplift of the area, which determined high relief values, and

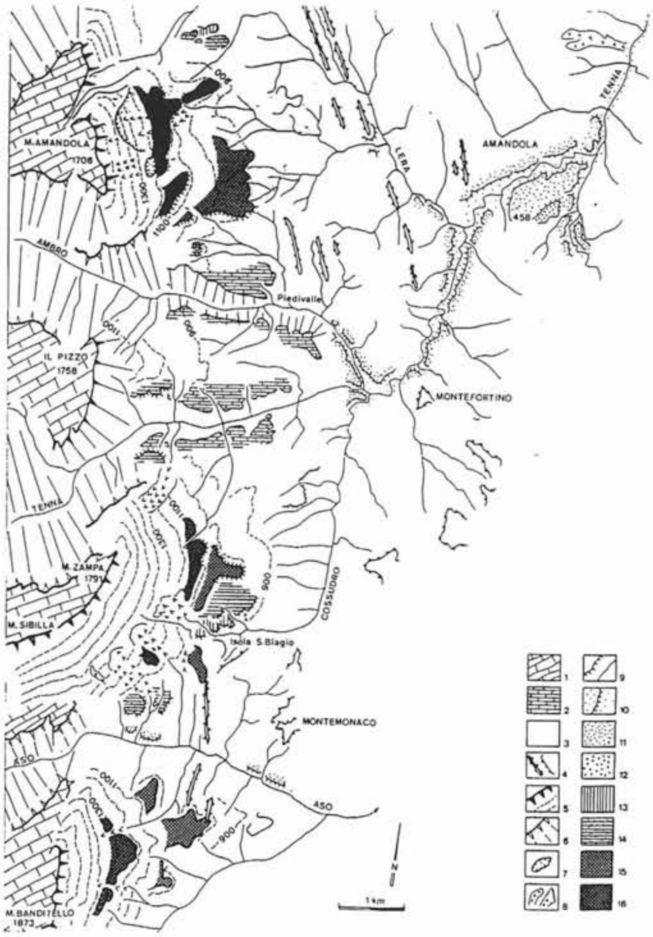

FIG. 3 - 1) Sibillini Mts. calcareous units; 2) structural surfaces; 3) cainozoic marly formations; 4) ridges on the turbiditic sediments; 5) overthrusts; 6) fluvial denudational slopes; 7) dolines; 8) badlands; 9) terrace edges; 10), 11), 12) alluvial terraces; 13), 14), 15), 16) planation surfaces along slopes (after RASSE, 1993-94).

in the intense fracturing of the bedrock (connected either with the shaping of the periadriatic monocline, or with gravitational deformation), as well as in the presence of rigid rocky bodies (sandstones and calcarenites) overlying more plastic marly-clayey levels. Among the stratigraphic discontinuities, the pre-transgressional erosional surface plays a leading role, since the main sliding surfaces coincide with it (GENTILI & *alii*, 1995).

Many cases of deep-seated gravitational deformations and large-scale landslides may be observed along the western sides of the limestone ridges. Important deep-seated gravitational phenomena found in the Sibillini Mts., along the large fault slopes bordering the east of the Cascia, Norcia, Castelluccio and Colfiorito depressions, and on the eastern slopes of calcareous ridges, along the eastern oversteepened sides of anticline folds and overthrust fronts (fig. 2). Along these slopes, steps, trenches, undulations and fractures may be observed, together with large-scale landslide scarps and wide scree deposits at the footslopes.

These materials are often covered by Upper Pleistocene stratified slope-waste deposits, which sometimes are tilted counterslope. The two units are locally interfingered thus testifying for an ancient origin of the movement and for its subsequent reactivations (FARABOLLINI & *alii*, 1995). The main causes of these gravitational deformations may be the following:

- recent uplift which induced deep valley incision;
- intense tectonic deformation of the bedrock;
- the presence of residual compressional stresses.

The most common types of deep-seated slope deformation on slopes modelled on thick stratified calcareous formations are bilateral spreadings (JAHN, 1964; DRAMIS & SORRISO-VALVO, 1994), which originated double ridges, lateral spreads and sackungs, mainly. Also deep-seated block slides (DRAMIS & *alii*, 1987), sometimes evolving into large-scale gravitational collapses, were activated.

Deep-seated gravitational deformations and large-scale landslides are also present inside the chain. Deep incisions transverse to the chain axis allowed the intersection, at various heights, of slopes with potential sliding surfaces, such as marly layers included in the limestone sequence beds. Due to the general northward dip of the beds, huge deep-seated block slides were activated on the southern sides of the valleys (GENTILI & PAMBIANCHI, 1994; DRAMIS & *alii*, 1995).

One of the most representative examples of deep-seated gravitational deformation is that between the Fiastrone and Chienti Rivers, slightly south of Camerino (GENTILI & *alii*, 1992; DRAMIS & *alii*, 1995). There, at the base of a marly limestone slope, shear planes and upslope tilting of Holocene alluvial deposits may be observed. In its middle portion, the slope shows a trench whose freshness suggests a very recent age for the deformation (fig. 4). This is further confirmed by oral tradition among local inhabitants reporting about the reactivation of the trench during the strong 1799 earthquake (IX MCS, PERGALANI & *alii*, 1985). Slightly mountainward along the valley bottom, lacustrine deposits including plant remnants, whose age is dated between 27,000 and 30,000 BP (DAMIANI & MORETTI, 1969) may indicate a previous activation of the gravitational movement during Upper Pleistocene.

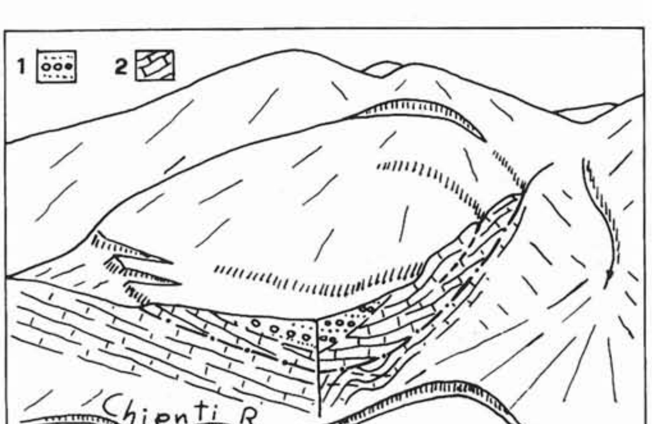

FIG. 4 - Block-diagram of the Valdiea landslide. 1 - Pleistocene-Holocene deposits; 2 - marly limestones.

The Ancona landslide

Along the Adriatic coast (fig. 1, A), trenches trending parallel to the coastline and steps lowering seawards have been found in wave-cut cliffs, generally no longer active and separated by the sea through small beaches. The origin of the above landforms is to be referred mainly to deep-seated gravitational deformations induced by active compressional tectonics (DRAMIS & SORRISO-VALVO, 1994). Within this framework large-scale landslides were also produced.

One of the most representative examples of deep-seated gravitational slope deformation can be observed near the town of Ancona (fig. 5). On the evening of the 13th December 1982, after a period of particularly heavy rain, a huge landslide took place on the north-facing slope of Montagnolo Hill, in the western outskirts of Ancona, over an area of more than 3.4 km^2, from about 170 m a.s.l. to the Adriatic coast (CRESCENTI & *alii*, 1983; COLTORTI & *alii*, 1984).

The phase of rapid deformation, which started without warning, lasted only a few hours and was followed by a

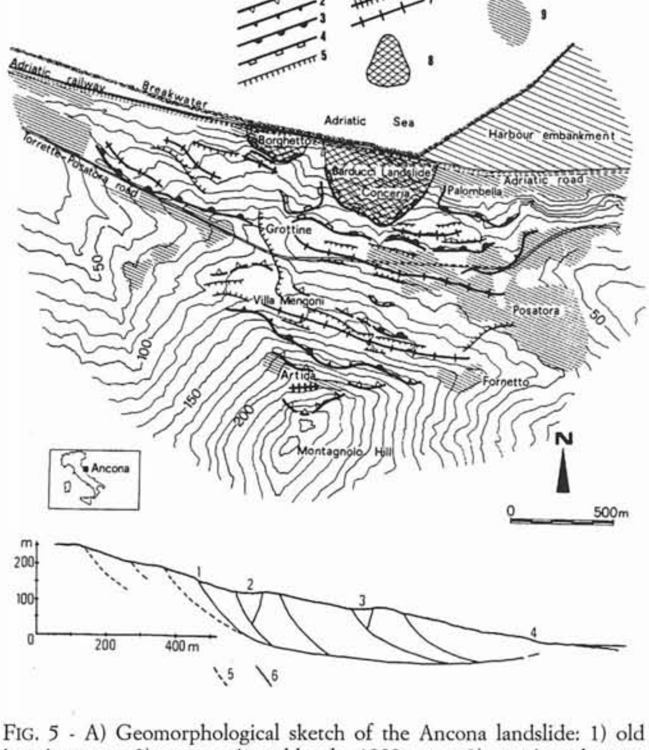

FIG. 5 - A) Geomorphological sketch of the Ancona landslide: 1) old inactive scarp; 2) scarp activated by the 1982 event; 3) reactivated scarp; 4) scarp active before the event; 5) minor scarp and step; 6) inactive trench; 7) reactivated trench; 8) main superficial landslide; 9) urban areas. B) Topographic profile of the Montagnolo slope showing hypothetical deep shear surfaces: a) main scarp; b) upper trench; c) lower trench; d) foot slope; e) inactive shear surface; f) active shear surface (after COLTORTI & *alii*, 1984).

longer period of settling. More than 280 buildings were damaged beyond repair and many of them collapsed completely. The Adriatic railway, along the coastline, was damaged over a distance of about 1.7 km. Luckily, there were no victims.

The slope hit by the landslide has had a long history of gravitational movements (BRACCI, 1773; SEGRÈ, 1920) and in 1858 it was the site of a landslide even larger than the present one (DE BOSIS, 1859). Other smaller landslides, still large as regard to the commonest ones, have occurred in the same area. Of these, the «Barducci» landslide, a flow type mass-movement, has been known for a long time, because its continual and intense activity produced visible damage to the coastal road and railway (SEGRÈ, 1920). From a stratigraphic point of view, the outcropping lithotypes are the following:

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

The tectonic setting of the area fits the general framework of the northern Apennines including folds with axes trending NW-SE and faults striking NW-SE and NE-SW. Three main tectonic phases have been recognized in the area (CRESCENTI & *alii*, 1983):

a) a folding phase dated middle Pliocene-Early Pleistocene;
b) a block-faulting phase during the Early Pleistocene which caused the subsidence of the area and the deposition of marly-silty clays;
c) an uplift phase, starting by 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 event of December 1982, shows a characteristic landslide morphology with trenches, scarps, steps, undrained depressions and reverse slopes. Downslope, a rugged foot-slope zone extends towards the sea; the steepening of the foot-slope seems to be accounted for by sea erosion, before building of the harbour embankment and along-shore protective measures.

Landslides and badlands in the Marchean high hills

Landslides and badlands have a basic importance in the modelling of the high hills of central-southern Marche (fig. 1, B; fig. 6) (BISCI & *alii*, 1992; DRAMIS & *alii*, 1992; FARABOLLINI & *alii*, 1992). Landslides are clearly influenced by the Pliocene-Quaternary network of faults and fractures, which also influenced the settlement and development of both the hydrographic network and the badlands (at the level of meso-landforms), as well as of the main structural scarps (FARABOLLINI & *alii*, 1992; GENTILI & *alii*, 1995).

In the valleys cut into gray-bluish clays, the slopes where the layers dip mountainward (i.e. those exposed between the West and the South) are intensely eroded by badlands (DRAMIS & *alii*, 1982), whilst the others are affected by translational slides and flows, sometimes of rather large dimensions.

Also the morphodynamic evolution of lithoid reliefs mainly happens by means of mass movements, such as falls and topplings, along the scarps where strata dip inside the slope, and (even though subordinately) translational slides (VARNES, 1978) where strata (constituted by rigid lithoid slabs of thick slope debris overlying clayey levels) dip outwards.

The Montelparo landslide

The ancient village of Montelparo is located on top of a hill modelled into an approx. 25-m-thick mainly arenaceous slab overlying pelitic-arenaceous terrains. The structure follows a monocline dipping NE with an angle of about 10°-12°. The eastern side of the hill is affected by a huge translational slide which starts at about 580 m a.s.l. and reaches down to the valley bottom (about 340 m a.s.l.), with an overall length of about 1100 m and a width ranging from 500 to 700 m. It involves more than half of the historic center, which has been displaced as a whole (buildings do not show any symptom of tilting or fracturing). Major damages occurred along the limits of the large trench located at the top of the landslide. There, it is possible to see a transversal protection wall built about three years ago and already crossed by cracks caused by differential downlift; moreover, fractures and steps in constructions and roads are diffusely present all along the sides of the trench. Along a cut in the left flank of the hill, it is possible to see the mountainward limit of the trench and, to the West, the sub-vertical plane cut into the intact bedrock in contact with the chaotic debris filling the depression.

A 46.5 m-deep borehole recently drilled at some 300 m from the main scarp, revealed the presence of slickensides within clays at a depth of about 37-38 m, testifying for a multiple shearing surface. Very likely this slip surface under the village may be located at a depth ranging between 65 m (the depth corresponding to a layering limit between arenaceous layers and the underlying clayey strata) and 100 m. Surface information makes it possible to infer the possibility of a rectilinear crossing of the clayey strata. Due to the erosion operated by the river at the toe of the slope, the lowermost part of the landslide is characterized by progressive disruption of the sliding body in blocks, divided by cracks up to 3-5 m deep and 2-3 m wide.

Precision levelling surveys carried out from May 1977 to March 1979 established that along the trench vertical displacements of 20-25 cm had taken 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. The comparison of a cadastral map dating back to 1935 and a new

FIG. 6 - Panoramic view of the Mt. Ascensione area. Glacis deposits are present at the top of badlands.

map, derived from aerial photographs dated 1970, determined that in 35 years the trench had widened by about 3 m, with an average velocity of 8 cm/year. In addition, precision topographic surveys carried out in 1980 gave horizontal displacements of up to 2-3 cm (ANGELI & *alii*, 1996).

Damage induced by the landslide, which severely influenced the development of the village causing a noteworthy reduction of its size, are documented starting from the XVII century. The first testimony of activity is in a report written in 1650. A booklet dated 1781 reports an important reactivation which took place during the strong displacement happening in coincidence with the strong 1703 earthquake.

Among the main predisposing factors for the movement, the structural setting of the area is of prevailing importance; the presence of transversal tectonic discontinuities, which disconnected the presently sliding body from the mountainward area, also seems to be very important.

The strong downcutting of the Fosso di S. Andrea stream is maybe the most important triggering factor. Anyhow, the ruinous effects of the 1703 earthquake suggest that, besides strong rainfalls, also seismic activity had an important role in the reactivations (and maybe in the genesis) of the phenomenon.

STRATIFIED SLOPE-WASTE DEPOSITS AND FLUVIAL DEPOSITS IN THE UMBRO-MARCHEAN APENNINES.

Stratified slope-waste deposits

Stratified slope-waste deposits are widespread in Central Italy. They are particularly frequent on rocky slopes (especially those composed by micritic and marly limestone) and are evidence of the cold climatic conditions which affected the area during Middle-Upper Pleistocene (COLTORTI & *alii*, 1983; COLTORTI & DRAMIS, 1987).

Detailed morpho-stratigraphic analysis and ^{14}C dating of alluvial and interfingered slope-waste deposits in a mountain area of the Umbria-Marche Apennines (upper Esino River basin) allow the chronological sequence of debris depositional phases during Late Pleistocene to be outlined (COLTORTI & DRAMIS, 1995). A few dates were also obtained from the Ponte di Crispiero sequence, in the contiguous Potenza River basin (fig. 6), which may be considered as a type section for the Late Pleistocene of Central Italy, Adriatic side (CHIESA & *alii*, 1990). The stratified debris was deposited after 50-60,000 years BP, in the Stadials of middle and upper Pleniglacial. During the Interstadials, most of the slopes of the basin were colonized by steppe vegetation and the production of stratified debris was probably restricted to the highest elevations. Increased rates of debris deposition occurred during the upper Pleniglacial.

The Ponte di Crispiero sequence

The Ponte di Crispiero sequence (fig. 7) has been studied in detail for its stratigraphical, pedological and palaeoethnological peculiarities (CHIESA & *alii*, 1990); approximately 8 m thick, it is lying on a straath terrace modelled on the «Scaglia Rosata» Formation (Eocene), at an elevation of 310 m a.s.l.. Laterally, this terrace is correlated with alluvial gravels from the final middle Pleistocene. The sequence covers the time span from the Last Interglacial to the end of the last glaciation. From the top, 4 main units have been recognized:

Unit 1: present topsoil;

Unit 2: stratified slope-waste deposits and intercalated chernozems;

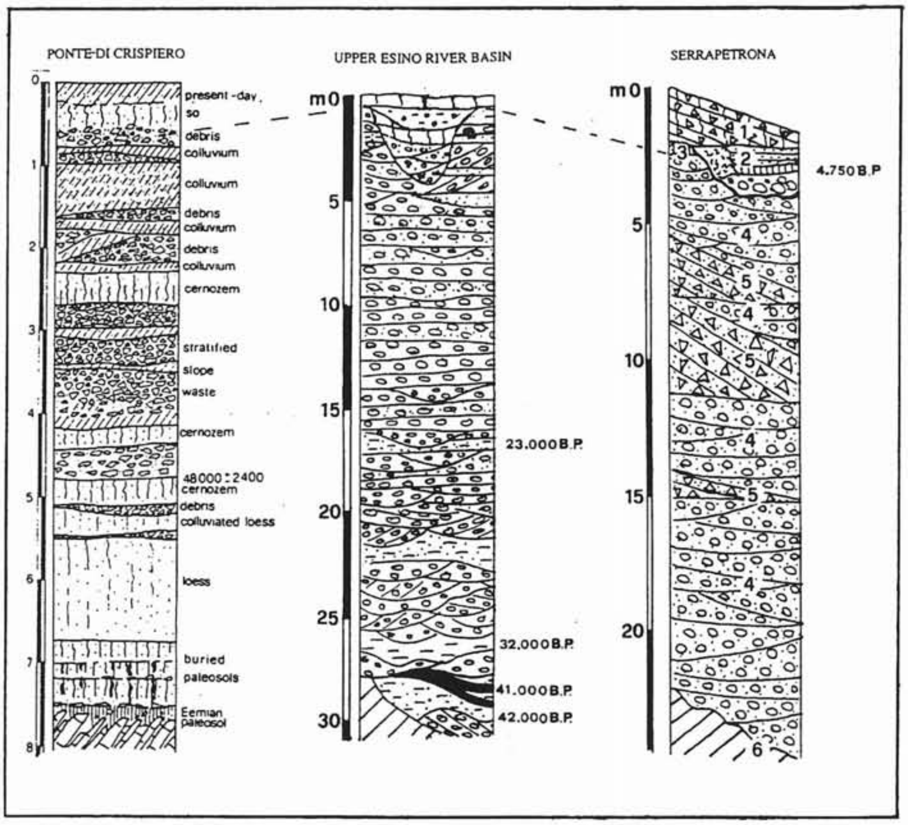

FIG. 7 - Correlation of sedimentary sequences in the Central Marche (modified from FARABOLLINI, 1995, and COLTORTI, 1996).

Unit 3: loess;
Unit 4: buried paleosoils.

Within the last unit, between IXB22ca and XB23t horizons, 7.5 m from the top, a rich assemblage of Musterian artifacts was discovered which, from a typological point of view, was attributed to the beginning of the Last Glaciation. Artifacts typologically attributed to the upper Palaeolithic were discovered at the surface of the sequence and within the IIC horizon. These are surely younger than 30,000 B.P.

Two ^{14}C datings on organic matter contained in the chernozems of the IVA1 horizon gave ages older than 34,200 and 48,000 B.P. (GX 14029 and GX 13957). Better results were obtained with a U/Th datation on calcareous concretions from VC1. The age of this layer, around 48,100 ± 2,400 B.P. allows us to establish that the underlying loess deposits belong to the lower Pleniglacial and confirms an attribution for the buried paleosoils to the time-span from the end of the Last Interglacial to the Early Glacial. The stratified slope-waste deposits were deposited during the middle-upper Pleniglacial and Late Glacial while the interlayered chernozems represent the Interstadials of this period.

The buried paleosoils reveal that the local slopes were affected by areal erosion after the end of the Last Interglacial, which strongly truncated the Bt lower horizon. However, sparse forested areas survived for a large part of the Early Glacial. In fact, the Interglacial soil is characterized by a coarse fraction represented exclusively by flint, a clay-rich matrix and a strong rubefaction. On the other hand, the overlying paleosoils show a progressive decrease in clay and organic matter and an increase in calcium carbonate. This last occurrence is associated to aeolian sedimentation which indicates the progressive onset of arid conditions. The loess deposition indicates the most arid period probably corresponding with the lower Pleniglacial. Only after this period did the local slopes begin to be deeply affected by cryoclastic conditions which continued till the end of the last glaciation.

The Serrapetrona sequence

Along the Marchean valleys, inside the Upper Pleistocene alluvial deposits, frequent interfingering with slope debris can be observed (COLTORTI & *alii*, 1983; CHIESA & *alii*, 1990; COLTORTI & *alii*, 1991; COLTORTI & DRAMIS, 1995). In the Serrapetrona area, in particular, inside the upper valley of a left-hand tributary of the River Chienti, it is possible to find some layers of stratified slope-waste deposits intefingered at different stratigraphic heights with alluvional sequence. The whole sequence (fig. 7), some 35 m high, is made up of alluvial gravels (4) intercalated with detrital sediments (5) constituted by flattened, subangular clasts with scarce silty matrix (clast supported). The latter material is constituted by bodies having a conspicuous inclined parallel stratification, more frequent in the central portion of the suc-cession.

The alluvial sediments (generally matrix supported) are made up of sub-rounded, flattened clasts, fairly compact, even though coarser layers are present.

The gravelly bodies are generally massive or show a slight low-angle cross-stratification. The accumulation of these bodies is given by the overlying of gravelly bars deposited by a multiple-channel river overloaded with debris. On the other hand, the stratified slope-waste deposits interfingering the previously mentioned sediments are closely connected with slope dynamics of the last pleniglacial. Their formation, connected with intense and repeated cryonival phenomena (very effective on the calcareous, strongly fractured rocks cropping out in the area), was activated by both slope wash and superficial mass movements (solifluction s.l., debris flow) along bare slopes (COLTORTI & DRAMIS, 1995).

At the top of the sequence, subangular clastic sediments (1), immersed in a brownish clayey-silty matrix (matrix supported), are present, with an overall thickness of 4 m. They bury a brownish calcic paleosoil (3), whose evolution seems to be connected with the evolution (during the Holocene) of a dense forestal cover, and partially cover a gravelly channel (2). The presence of resedimented historic ceramic material within the latter sediments suggest some connections with anthropic slope clearing. At the base of the channel (2), at a depth of ca. 2 m and for a thickness of ca. 1.5 m, coarse gravelly lenses (Gm of MIALL, 1985) are present, constituted by sub-rounded and sometimes embricated calcareous clasts immersed in a scarce silty-sandy matrix (partially open work). These lenses are alternated with thin layers (a few cm thick) of phytohermal travertine and stromatolitic crusts, testifying for a temperate-humid climate characterized by the presence of a widespread vegetal cover (CILLA & *alii*, 1994). The presence of a ca. 30-40 cm thick layer with a clayey texture and skeleton constituted by subangular clasts, containing not reworked ceramic fragments, allows the sequence to be referred to the early Holocene. The channel is filled by one meter of phytoclastic sand with planar stratification, laterally passing to phytohermal travertine. The analysis of the ceramic fragments allows them to be dated to the Eneolitic (V *millennium* B.P.), thus allowing a rate of 8 mm/year to be referred for Holocene downcutting (FARABOLLINI, 1995).

Travertine deposits

After a general downcutting of the Late Pleistocene deposits, slightly before 8260 ± 100 yr B.P. travertine deposition started in the mountain sectors causing the formation of dams and related small lacustrine basins or swamps. This process followed until the first half of the IV millennium B.P. During the travertine deposition, the surrounding area was covered by warm-temperate forests, which completely stabilized the slope. The travertine deposition ceased in the fluvial valleys as in other sites of the Umbria-Marche Apennines, as well in several areas of central-western Europe. The strong reduction of travertine formation has been connected to the consequences of human impact on the natural forest cover (CILLA & *alii*, 1994; COLTORTI, 1997) as well as to climatic variations at the regional scale (CALDERONI & *alii*, 1996).

Fluvial morphogenesis during Late Pleistocene - Holocene

The periadriatic valleys of Central Italy were affected by a general aggradation under a braided river system during most of the middle and upper Pleniglacial (CALDERONI & *alii*, 1991). Anastomosing rivers characterized by a fine sedimentation with peat layers, were established during the interstadial phases (Denekamp-Arcy, Hengelo, Kesselt, Tursac and Pre-Bolling). The general Pleistocene aggradation seems to have ended during the Late Glacial even though further dates are still needed to point to a more precise framework of the events (fig. 7).

A section at Smorlesi quarry, in the mid Potenza river valley, showed a sequence made up of mostly fine sediments (alternating silty-clayey layers, alluvial soils and cross-bedded sandy layers). Also ceramic fragments were recognized in the middle-upper part of the sequence inside sandy sediments. A fireplace located at about 11 m from the top of the sequence gave an age of 7210 ± 90 yr B.P. (Rome-508). Approximately 1 km downvalley, close to Fontenoce, a channel has been localized at the top of the same sedimentary unit, containing two superimposed Eneolitic settlement layers, separated by overflood silts. The dating of these layers gave ages of 4680 ± 100 and 4700 ± 100 yr B.P.

Sequences up to 7-8 m thick, showing the same stratigraphic and morphochronological features, have been recognized in the mid Tenna River valley. At the San Gualtiero quarries (fig. 8), a charcoal layer from the base of an alluvial sequence mostly made up of fine sediments gave an age of 7620 ± 80 yr B.P. Moreover, Bronze Age ceramic fragments and charcoal dated 3570 ± 70 yr B.P. were discovered at the base of a more than 10-m-deep channel, cutting the whole sequence and the underlying upper Pleistocene alluvial deposits. These ages testify a progressive deepening of the alluvial plain made by a single and sinuous channel up to 20 m high between Late Glacial and 8000 yr B.P.; processes of slow aggradation followed.

After the IV millennium B.P., the deposition of fine sediments ended and a rapid deepening of the thalweg was recorded. These modifications could be attributed to cli-

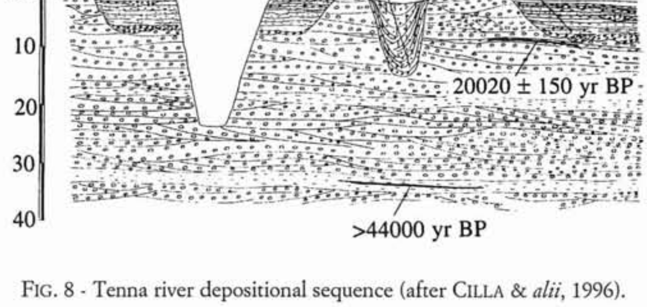

FIG. 8 - Tenna river depositional sequence (after CILLA & *alii*, 1996).

matic changes as well as to the result of human impact following the general occupation of the slopes for agriculture and farming purposes (COLTORTI & DAL RI, 1985; CALDERONI & *alii*, 1989; CILLA & *alii*, 1996)

THE KARSTIC COMPLEX OF FRASASSI

The Frasassi Gorge, located in the northern sector of the M.Valmontagnana anticline, is the site of an impressive karstic system in Central Italy (COLTORTI & GALDENZI, 1982; BOCCHINI & COLTORTI, 1990). It is mainly developed in correspondence with the «Calcare Massiccio» Fm. (Lower Lias) which crops out in the core of the anticline and is bordered by Jurassic synsedimentary faults, on the northeastern side. More than 80 caves have been mapped: many of them are interconnected, and the major one, the Fiume-Vento karstic complex, is over 12 km long. The caves are arranged on karst floors located at progressive elevations on the valley bottom. At least 7 superimposed floors have been described in the Fiume-Vento complex, from the thalweg (200 m a.s.l.) to the entrance of the Grotta Grande del Vento (400 m a.s.l.) but other minor floors are present at higher elevations. They record the presence of periods of stability (floor developement) alternating with periods of rapid river downcutting (CATTUTO, 1976; BOCCHINI & COLTORTI, 1990; TADDEUCCI & *alii*, 1995). The uppermost floors, to which several entrances are connected (Caves of Mezzogiorno, Paradiso, Grottone, Infinito, Inferno, etc.), are located along E-W or N-S faults and fractures. They suggest that the first karstification processes were connected to the reactivation of Jurassic tectonic lines which occurred during the uplifting movements and affected the area since the end of lower Pleistocene. Afterward, the lower floors developed in connection with NE-SW strike-slip faults through which sulphide ground waters rise up. A faster development of phreatic tubes and galleries along a NE-SW fault captured the rest of the karst systems and created the huge Fiume-Vento Complex.

MT. ASCENSIONE GLACIS

The neighborhood of Mt. Ascensione is characterized by the presence of a widespread detritic cover, presently reduced to small remnants suspended along the divides, made up of polygenetic and etherometric pebbly-sandy deposits, deriving from the erosion of the conglomeratic-sandy-pelitic bedrock (fig. 6). This deposit, belonging to a widespread accumulation glacis (CASTIGLIONI, 1935; DEMANGEOT, 1965; DRAMIS & *alii*, 1982; FARABOLLINI & *alii*, 1992), is present below Mt. Ascensione, at altitudes ranging between 850 m and 650 m a.s.l., with minor extension along the western slope.

The thickness varies from a few cm up to more than 40 m, with the maximum close to the mountain; the contact with the bedrock sometimes is in paraconformity, but more frequently onlapping, overlying both the Middle and Upper Pliocene, and Messinian terrains of the post-evaporitic member of the Laga Formation (CANTALAMESSA & *alii*, 1986).

The deposit is made up of alternating sandy and sandy-conglomeratic layers, from open-work to matrix-supported and/or clast-supported, having a chiefly lenticular geometry and variable thickness, with a dip from low to medium (10°-15°). The cementation is generally medium to low, even though sometimes the deposits are well cemented. At different stratigraphic heights, many brown paleosoils have been found, having a thickness ranging from 20 to 50 cm. Inside these paleosoils, clasts ranging from 2 to 50 mm have been found as well as coal elements, presently being dated. At the top of the sequence, thick deposits are present, made up of fine and very fine massive sand, sometimes well cemented, yellowish, with rare calcareous or cherty pebbles. Clast and matrix dimensions show noteworthy variations, both vertically and transversally; vertical variations are connected with the presence of graded layers, with typical coarsening upward or fining upward sequences. Moving from the proximal to the distal portions of the relief, a progressive increase of the sandy (medium to fine) matrix as well as an increase of sphericity and roundness of the pebbles is observed.

In the whole area, the dominating depositional mechanism is connected with processes of areal erosion and mass transportation (mass flow, 1995); the associated facies are mainly Gms, more rarely St, Sp and Sh (MIALL, 1985).

In the more proximal areas and along the western slope of Mt. Ascensione, layers due to gravitational morphogenesis (debris flow, solifluction s.l. and fall), constituted by clasts and blocks up to some meters large are intercalated in the massive levels. In the most distal portions, thin pelitic layers, probably connected to filling of abandoned channels, and intercalated gravelly lenses with sandy laminae have been found.

Along the western side of Mt. Ascensione, an outcrop shows in its middle-high portion layers with dominating mass transportation (debris flow and sheet-flood), whilst in the lower portion conspicuous channellization with erosional surfaces below gravelly channels (CH of MIALL, 1985) are present; the latter can be attributed to braided channel systems, with facies mainly Gm and subordinately Gms, Gp, St, Sp, Fm. Paleocurrents show an extremely variable directional pattern, ranging from N30° to N120° to the eastern side of Mt. Ascensione, from N330° to N270° to the western side and from N250° to N120° to the south.

On the basis of both geomorphologic-pedologic considerations and the analysis of stratigraphic sections, it is possible to date the glacis to the Upper Pleistocene, which regularized a more ancient topography (DRAMIS & *alii*, 1982; FARABOLLINI & *alii*, 1992).

THE TECTONIC DEPRESSION OF COLFIORITO

A thick sequence of alluvial and lacustrine deposits is present inside the remnants of wide paleovalleys in the

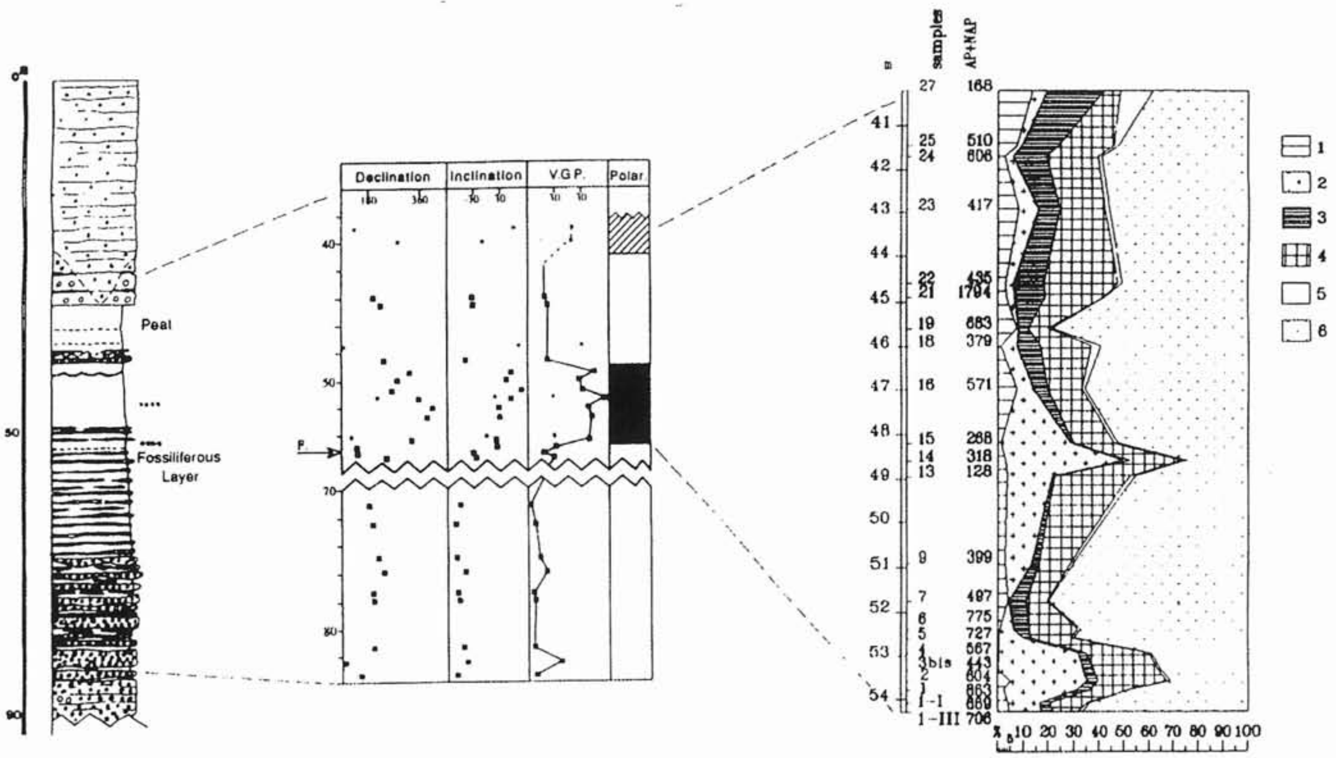

FIG. 9 - Main stratigraphic, paleomagnetic and palynological elements of the Colfiorito sequence. The lower part of the sequence, down to the fossiliferous level (F), is negative magnetized. A normal polarity is present upward followed by another earth magnetic field up to the top of the clayey sequence. The two upper samples are represented with little squares because their direction is uncertain. On the right side, a synthethic diagram with the main pollinic association is shown: 1) deciduous elements; 2) *Tsuga* and *Cedrus*; 3) *Abies* and *Picea* with *Betula* and *Fagus*; 4) *Pinus* with several *Pinaceae* grains; 5) other arboreous plants; 6) non-arboreous plants (after COLTORTI, 1996).

Colfiorito basin (Cesi) and at Colle Curti (Nera Basin) in the eastern extension of the same basin (fig. 1). The sequence (fig. 9), which is altogether about 100 m thick, is almost completely exposed at Colle Curti while only its middle-upper part may be observed at Cesi. From the bottom to the top, it is constituted by three units: 1) subangular and subrounded coarse gravels with intercalated thin clayey-silty and sandy layers as well as weakly developed alluvial soils (ca. 40 m thick); 2) massive or weakly laminated clays (ca. 20 m thick); 3) gravels (40 m thick) whose calcareous fraction is preserved only at the base of the sequence while, in the uppermost part (more than 20 m thick), only residual flint gravels are preserved. This unit constitutes the remnant of a thick alfisol whose original thickness was much greater than nowadays.

At Colle Curti, the intermediate unit contains a floristic association much more impoverished than the Late Pliocene and Early Pleistocene ones. Much of the typical elements of the humid subtropical forest as *Taxodium, Nyssa, Sciadopitys, Sequoia* and *Cathaya* are missing. *Tsuga* and *Carya* are still present although quite rare. The floral association suggests the presence of a pre-Cromerian pollen unit which can be correlated to the Bavelian flora.

The palinological record points to a landscape strongly dominated by an open vegetation represented principally by *Asteraceae, Cyperaceae* and *Poaceae* pollen grains, both at Cesi and Colle Curti. The repeated episodes characterized by such a vegetation typical of strong arid conditions is correlatable with the occurrence of glacial phases. No significant expansion of mesophilous forms with *Tsuga* and *Cedrus* at Colle Curti has been recorded. The absence of fresh Interglacial phases is the striking feature of both the palynological diagrams. A lack of documentation of Interglacials in the sedimentary record cannot be denied.

At the base of the intermediate unit, at Colle Curti, a rich mammal fauna has been excavated (FICCARELLI & MAZZA, 1990; FICCARELLI & *alii*, 1990; FICCARELLI & SILVESTRINI, 1991). The fossil findings are represented by *Hippopotamus antiquus, Stephanorhinus* cf. *hundsheimensis,* «*Elephas*» sp., *Pseudodama* ex group *nestii* advanced form, *Megaceroides verticornis, Canis arnensis, Canis* (*Xenocyon*) ex group *falconeri, Ursus* sp., an undetermined *hyaenid* and *Microtus* (*Allophaiomys*) form with more derived features than M.(A.) *ruffoi* and M.(A.) *pliocaenicus*. The presence of *Megaceroides verticornis* and *Microtus* (*Allophaiomys*) indicates a very early Galerian Mammal Age (upper part of the Early Biharian) (TORRE & *alii*, 1992).

At Cesi, only the upper part of the intermediate unit is preserved at the excavated site. The taxa hitherto recognised from Cesi are *Elephas* sp., an advanced representative of *Stephanorhinus hundsheimensis*, a *caballine equid, Hippopotamus* sp., *Megaceroides solilhacus, Cervus elaphus, Da-*

ma clactoniana, Bison schoetensacki and *Homotherium* sp and represent a faunal assemblage of the middle-late Galerian.

The Colle Curti sequence in the Geomagnetic Polarity Time Scale embraces a large part of the Matuyama Chron although the fauna corresponds to the beginning of the Jaramillo Normal Polarity Event (Chron C1r.1n). The Cesi fossiliferous is located a few meters above the M/B boundary (780 ka). In the middle part of the upper weathered unit a pyroclastic layer has been dated by Ar/Ar method and given an age of 437 ka. Therefore, these deposits record the climatic and tectonic records since the lower-middle Pleistocene.

THE CALDERONE GLACIER
(M. D'Orefice, M. Pecci & C. Smiraglia)

The Calderone Glacier, a small glacier (less than 5 ha) located in a small deep northward valley, between the three main peaks of Corno Grande d'Italia (2912 m a.s.l.) represents the southernmost glacier in Europe (42°28'15"N) after the complete melting of the Picado de Veleta Glacier in the Sierra Nevada (MESSERLI, 1967). Scientists disputed a long time about the glacial nature of the Calderone (DE MARCHI, 1573; DELFICO, 1794; MARINELLI & RICCI, 1916; KLEBESBERG, 1930; SUTER, 1939). It was included in the Italian glacier inventories in 1925 (PORRO, 1925; COMITATO GLACIOLOGICO ITALIANO, 1962) and, recently, also in the World Glacier Inventory (IHAS-UNEP-UNESCO, 1993).

From the geomorphological point of view, the Calderone may be divided into three sectors (D'OREFICE & *alii*, 1996), differently dipping and shaped. The lower sector, more easily visitable, is generated by a big and visible front-lateral moraine, rising about 40 meters over the bottom of the cirque and leaning against the eastern peak of Corno Grande. The moraine may be sub-divided into three zones: the inner, still active with ice-core visible in advanced summer-time, the middle, already deposed, but lacking in vegetation, and the outer, older and stabilized, with traces of phanerogamic vegetation. This morainic deposit may be generically linked to «Little Ice Age» phases. Upward, an intermittent lake (Sofia lake), generated by snow melting during early summer, may be related to a regressive phase of the glacier. In the same sector, on the hydrographic right side, many big rock fall blocks are laid down on the glacier surface. On the steep debris slope against the rock walls, several slide phenomena give rise to arched structures dipping downslope. These structures are characterized by an asymmetric profile with the upper face (ice cored) generally steeper than the lower one where a circular hollow is present, probably linked to melt waters. The superficial run off is collected by a little torrent which cuts the ice core for few decimeters. Also «ice tables» with 20-30 centimeter «pedestals» are easily observed at the end of summer (SMIRAGLIA & VEGGETTI, 1992).

The central sector of the glacier is covered by a discontinuous mantle of debris which shows small hollows and hills, typical of «black glaciers». The ice core is visible along the steepest cones and deposits, separated by meandering «bédières» which discharge water into circular ice-holes (SMIRAGLIA & VEGGETTI, 1992). On the lateral sides some transversal and longitudinal ice-trenches and crevasses may develop. The rock walls show typical «roches moutonées» features and glacial striae. The upper sector of the glacier is characterized by a thick debris cover, continuously fed by debris fall from the rock walls.

From the hydrological point of view, a modelling on Gis has been carried out in order to evaluate the water resources in terms of volumes of snow melted versus glacier covered area in 1994 and 1995 (summer season). Preliminary results show a reduction in 1995 of 25,000 cubic meters (365 mm in water equivalent) of deposited snow (D'OREFICE & *alii*, 1996).

ENVIRONMENTAL CONDITIONS AND PLEISTOCENE GLACIATION OF CAMPO IMPERATORE, GRAN SASSO
(C. Bisci, F. Dramis, B. Gentili, E. Jaurand & A. Kotarba)

Campo Imperatore is a wide NW-SE trending tectonic depression located in the inner section of the Gran Sasso Massif. Its western portion described below, is more than 13 km long and ca. 4 km wide, with elevations ranging from 2494 m a.s.l.. at is northernmost end (Mt. Aquila) to about 1480 m a.s.l. at its lowest point.

The present day climate of Campo Imperatore is characterized by 1120 mm of average annual rainfall, with minimum values in winter and summer (as reported by the records of a meteorological station located very close to Albergo Campo Imperatore, at 2138 m a.s.l.). The mean annual temperature is 4.7°C with an annual amplitude of monthly averages of 16.2°C (DEMANGEOT, 1965). The 0°C mean annual temperature is located around 2900 m. a.s.l., close to the highest peak, of Gran Sasso Massif-Corno Grande, 2919 m a.s.l.).

The outcropping bedrock is made up of the following units (GHISETTI & VEZZANI, 1986; GHISETTI & *alii*, 1990): massive dolomite (lower Lias), bioclastic grainstone (Dogger-Lower Cretaceous), marl, shale and cherty mudstones (Upper Lias-Titonian), bioclastic mudstone with cherty levels (Upper Jurassic-Lower Cretaceous), pelagic mudstone, marly mudstone and grainstone (Upper Cretaceous-Middle Eocene), packstone and grainstone with marly mudstone (Eocene-Oligocene), and laminated marl and marly mudstone (Lower-Middle Miocene).

The above terrain follows a monocline gently dipping to the northeast, even through the overall structure is the result of the tectonic superimposition of several units (following a NE trend), which took place during the Upper Miocene-Lower Pliocene compressional phase (PAROTTO & PRATURLON, 1975, GHISETTI & VEZZANI, 1990, BIGI & *alii*, 1991). During the Quaternary, the area was subject to extensional tectonics (CARRARO & GIARDINO, 1992, GIRAUDI, 1994) which produced systems of normal faults mostly trending ENE - WSW and bordering to the south the Gran Sasso Massif. This tectonics phase, which has continued up to the present, caused a progressive lowering of the central portion of the basin with respect to the surrounding ridges.

On the slopes, present day tectonic activity is indicated by the presence of showy triangular facets, fresh scarplets and intense badland-like erosional phenomena, the last of which affect extremely shattered dolomites. Locally, these faults cut Late Quaternary deposits (stratified slope-waste deposits and alluvial fans), too. Also the strong earthquakes recorded in the surroundings of Campo Imperatore (up to 10 MSC in the L'Aquila basin, located about 10-15 km to the west) clearly testify to present day tectonic activity (POSTPISCHL, 1985, SOCIETÀ GEOLOGICA ITALIANA, 1989). This extensional phase has gone together with tectonic uplift, whose intensity increased considerably by the end of Lower Pleistocene, causing relief enhancements of up to 1000 - 1500 m. (DEMANGEOT, 1965; AMBROSETTI & *alii*, 1982; DUFAURE & *alii*, 1989; DRAMIS, 1992).

The Campo Imperatore depression is partially filled by Quaternary deposits that are primarily of glacial, glaciofluvial, alluvial, colluvial and lacustrine origin. The thickness of these deposits is particularly high, as testified by a deep borehole drilled in the uppermost portion of the valley, which went through some 200 m of continental sediments (CO.GE FAR., 1979).

The study area represents a very typical example of glaciated mountain landscape. Both erosional and depositional landforms are present, showing evidence of both Alpine glacial morphology (at altitudes above ca. 1800 m.) and stagnant ice deglaciation pattern, characteristic of the flat-floored topography of piedmont glaciers of Central Asia or Canada at lower altitudes.

The effect of two main Pleistocene glaciations can be recognized in the area. These two glacial generations have been distinguished using local names: «Piano Racollo» for the older glaciation and «Coppe di Santo Stefano» for the younger one (fig. 10).

The Piano Racollo glaciation is the oldest and most extensive glacial phase recognized in the Campo Imperatore basin. It affected nearly the whole study area, from Mt. Aquila to the base of Mt. Bolza and Mt. Mutri. The maximum extent of glaciation was reconstructed on the basis of the distribution of erratic boulders and small moraine remnants. The erratics are deposited on the glacially polished limestone hills of the Maniere valley. In Piano Racollo, only small patches of moraines, often strongly remodeled, are still recognizable within the outwash plain, as well as narrow shelves of glaciofluvial sediments formed as ice-contact deposits. The Piano Racollo glaciation also produced glaciolacustrine sediments. Glacial deposits present in Piano Racollo are correlated to the last cold stage of Riss, like a few further remains in the Apennines (FEDERICI, 1980).

The Coppe di Santo Stefano glaciation is characterized by large and well-defined terminal (and, subordinately, lateral) moraines showing distinctive morphology. These glacial features differ significantly from those of the Piano Racollo glaciation, mostly because of their freshness. The maximum extent of the glacier is clearly marked by sharp morphological contrast between the hummocky morainic system and the almost flat outwash plain. Landforms of glacial deposition are the most conspiciuous features in the main valley depression at altitudes ranging around 1640-1590 m a.s.l. There, a stagnant ice deglaciation pattern dominates on form of a ca. 2.5 km-wide lobate moraine system, rising 25-30 m above the outwash plain. Supraglacial sediments deposited during the more recent glaciation of the Coppe di Santo Stefano underwent differential melting. This resulted in spectacular topographic irregulari-

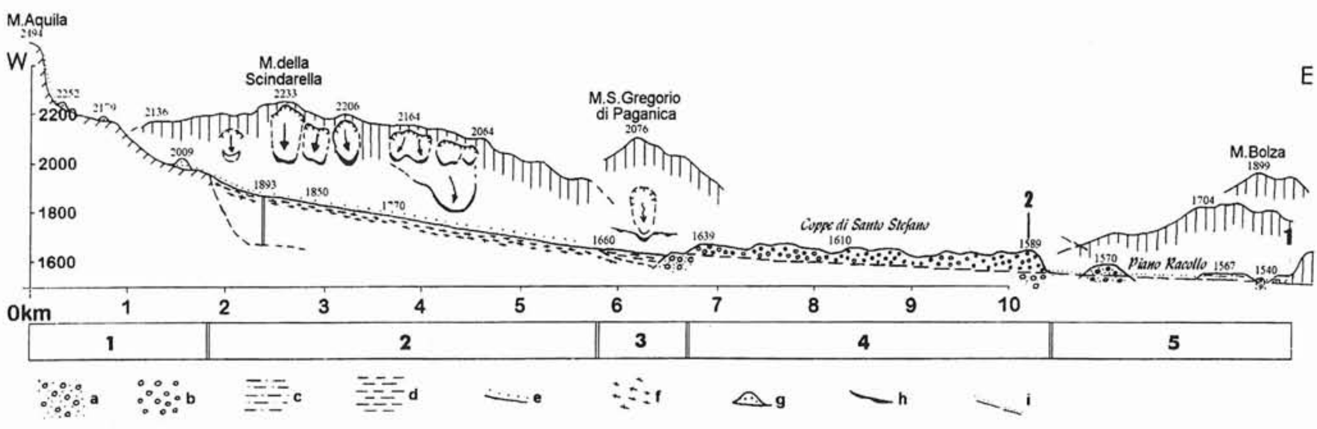

FIG. 10 - Schematic cross section along the main valley of Campo Imperatore plain, showing the principal geomorphologic units (numbers 1 to 5 in the bar below the section indicate the approximate extension of each unit; for their description refer to the text). The numbers 1 and 2 above the sketch, respectively, indicate the maximum extension of the Coppe di Santo Stefano and the Piano Racollo Glaciation. Legend: a) moraine (Piano Racollo glaciation); b) moraine (Coppe di Santo Stefano glaciation); c) esker (Coppe di Santo Stefano glaciation); d) fluviolacustrine deposits (Coppe di Santo Stefano glaciation); e) glacial erratics on bedrock topography (Coppe di Santo Stefano glaciation); f) well cemented conglomerate (breccia), eroded by glaciers; g) recessional moraine (Late Glacial); h) lateral moraine (Coppe di Santo Stefano glaciation); i) outwash plain (Coppe di Santo Stefano glaciation).

ties. Such morphology is similar to the so-called «karst topography on stagnant glaciers» described by CLAYTON (1964) and CLAYTON & MORAN (1974) in Canada. Within the widespread outwash plain, formed by glaciofluvial waters in front of the terminal moraines, paleochannels showing a braided river pattern are still well preserved in topography. The Coppe di Santo Stefano glacier was ca. 10 km long and its surface area was 21.5 km^2, *i.e.* approximately as big as the Argentiere glacier in the Mont-Blanc Massif. Behind this terminal moraine system, distinct recessional moraines are also well preserved at various altitudes. At least four main deglaciation stages are recognizable in the upper portion of the Campo Imperatore depression (*i.e.* around 2250, 2180, 2000 and 1740 m a.s.l.).

The Coppe di Santo Stefano glaciation culminated before 13,000 years B.P., since deposits correlated to the «tufo giallo napoletano» volcanic episode rest on the glacial drift (FREZZOTTI & NARCISI, 1989; CARRARO & GIARDINO, 1992). The glacial phase responsible for the deposition could be correlated to the Wurm III stage (Pleniglacial), which occurred in the Apennines from 21,000 to 18,000 years B.P. (FEDERICI, 1979). Periglacial processes were partialy simultaneous with the last glaciation and strongly affected the area after the ice retreat. Stratified slope-waste deposits (*grézès litées*) were produced within mountain slopes built up mainly of tectonically fragmented calcareous and dolomitic rocks. After the retreat of the last ice body, also the upper portion of the main valley was strongly affected by periglacial morphogenesis. A relict rock-glacier (GHISETTI & *alii*, 1990; DRAMIS & KOTARBA, 1994) is still fairly well preserved at an altitude of about 1900 m a.s.l.

Coalescing alluvial fans and composite debris flow cones are the most spectacular landforms at Campo Imperatore. They cover large areas between the steep, south-facing slopes and the valley bottom filled with drift deposits, thus fossilizing partly glacial depositional landforms. Alluvial fans were formed during Late Pleistocene, possibly during the second glacial phase, as testified by ^{14}C dating of paleosoils included within an alluvial fan located in the eastern portion of Campo Imperatore (FREZZOTTI & GIRAUDI, 1990). The more modern geomorphological studies by GIRAUDI (1994) show that alluvial fans have been developed in five phases, the last one being related to present day alluviation in the area.

FUCINO BASIN

(A.M. Blumetti, M. Coltorti, L. Ferreli, A.M. Michetti, C. Petronio, J. Raffy & R. Sardella)

INTRODUCTION

The Apennine Chain was shaped as a fold and thrust belt mostly during Upper Miocene to Early Pliocene. On the Tyrrhenian side of this compressional structure, since the Late Pliocene (?) - Early Quaternary («Villafranchian»)

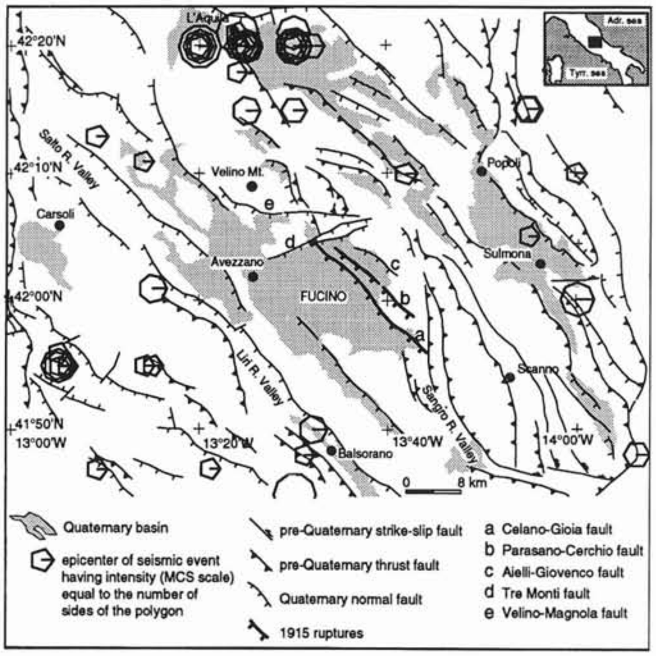

FIG. 11 - Tectonic map of the Abruzzi Apennines and historical seismicity from 1000 to 1900 A.D. Note the absence of historical seismicity in the Fucino area before 1901 and compare with paleoseismic data in Fig. 13.

normal-fault-bounded intermountain depressions, including the Terni, Rieti, and L'Aquila basins, evolved. Crustal extension in this area persisted during the whole Quaternary and is active today (fig. 11). However, it is important to note that continuing Quaternary to present folding is well documented in the Adriatic side of the central Apennines (e.g. PATACCA & *alii*, 1992), a few tens of kilometers east of the extensional basins.

The Fucino Basin (fig. 12) is the largest tectonic basin of the Abruzzi region. It lies in the middle of the central Apennines surrounded by mountain ranges more than 2000 meters in altitude (Mt. Velino, 2486 m a.s.l.), which are shaped essentially into Mesozoic-Cenozoic carbonate shelf sediments. The basin does not concentrate the hydrographic network or 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; fig. 11).

The central part of the basin, a plain between about 650 and 700 meters which is hydrologically closed, during the Late Glacial and Holocene, was occupied by the third largest lake in Italy (*ca* 150 km^2). In the 2nd century A.D. EMPEROR CLAUDIUS prompted the drainage of Fucino Lake. 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. Fucino Lake disappeared from maps.

THE LEADING ROLE OF TECTONICS

The Fucino basin is a typical intermountain normal-fault-bounded structure within the Apennines segmented normal fault system, which extends from Tuscany south to the Calabrian Arc and represents one of the more seismically active provinces of the Mediterranean region. Major normal faults representing nearest segments of this system are the Velino-Magnola fault to the NW («e» in fig. 11) and the Sangro Valley fault to the SE. Seismic reflection profiles (MOSTARDINI & MERLINI, 1988; CAVINATO & *alii*, 1993) 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 normal fault (BENEO, 1939; «a» in fig. 11) and parallel subsidiary faults (e.g. the Parasano-Cerchio and the Aielli-Giovenco faults; «b» and «c» in fig. 11). Within this style of faulting, tectonic inversion (Quaternary normal slip on pre-existing reverse faults) is very well documented, for instance along the SW range front of the Velino Massif (e.g., NIJMAN, 1971; RAFFY, 1979). Normal faulting appears to have been progressive during the whole Quaternary and is very active today. Evidence for this is documented by a) displacement and tilting of lacustrine formations and slope deposit sequences, b) trench investigations on Holocene normal fault scarps, and c) observation of coseismic and paleoseismic surface faulting events (Jan. 13, 1915, M7 Avezzano earthquake and related paleoseismic studies; ODDONE, 1915; SERVA & *alii*, 1988; GALADINI & *alii*, 1995; MICHETTI & *alii*, 1996; GALADINI & *alii*, in press; Figs. 12 and 13). These data show that the two major normal fault segments (Celano-Gioia and Velino-Magnola faults) are characterized by Middle Pleistocene to present slip-rates of 0.5 to 1 mm/yr, and total Quaternary offset in the order of 1000 m.

QUATERNARY LANDFORMS AND DEPOSITS

The range fronts bounding the Fucino basin are fault scarps. 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 downthrown block. Therefore, sedimentation is mostly influenced by tectonics. In fig. 12, the Quaternary terraces are grouped into three major orders, namely «upper», «intermediate» and «lower terraces», separated by prominent fault scarps.

The «upper terraces» include two main terrace surfaces of Late Pliocene (?)-Early 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 the faulted and reworked depositional surface of the Alto di Cacchia unit culminating at 950 m a.s.l. (fig. 12; BLUMETTI & *alii*, 1993; BOSI & *alii*, 1995). Two intersecting NW-SE

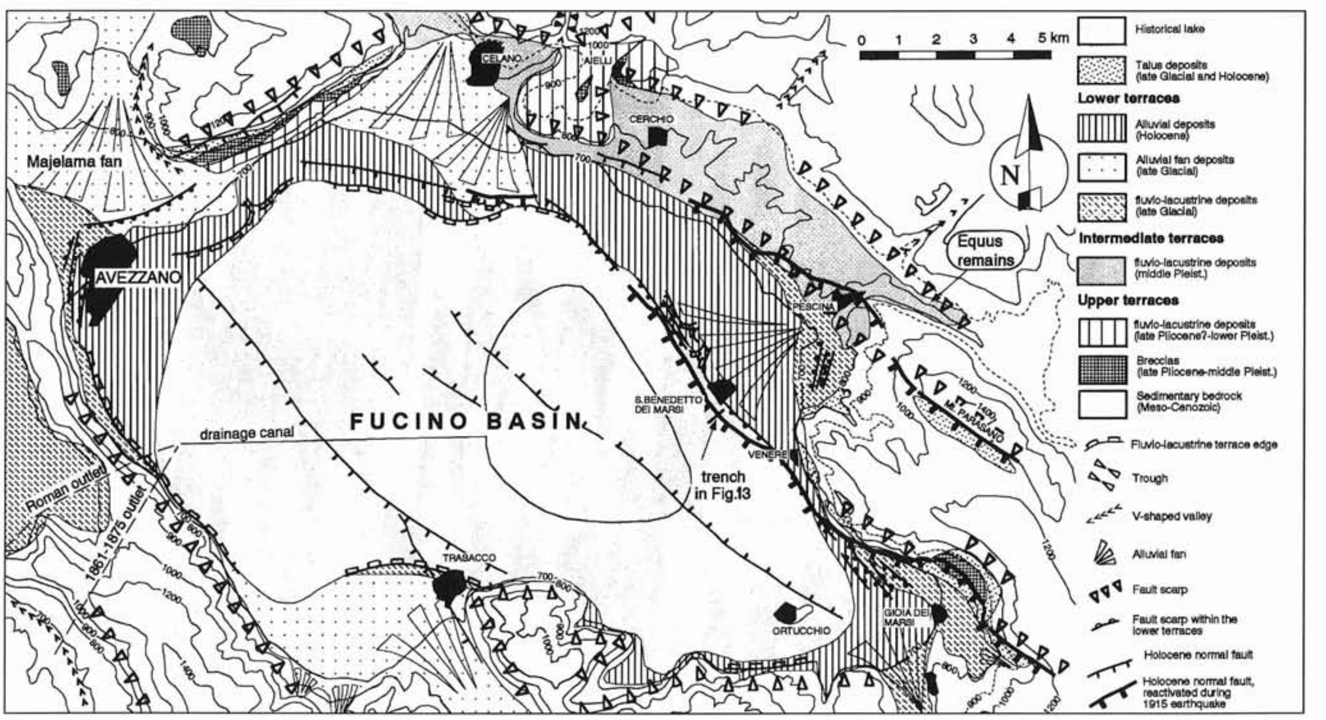

FIG. 12 - Geologic map of the Fucino basin, showing the surface faulting associated with the January 13, 1915, earthquake. Location of Fig.13 and of the *Equus* cf. *altidens* site is also shown.

and SE-NW trending normal faults border the «upper terraces», and generate fault scarps up to 100 meters high (fig. 12; RAFFY, 1981-1982; BLUMETTI & *alii*, 1993).

The «intermediate terraces» include two main Middle Pleistocene terrace surfaces. The higher one is divided by BOSI & *alii* (1995) into three orders «developed at base levels not very different from each other» (BOSI & *alii*, 1995; MESSINA, 1997). We will consider them as a single terrace culminating at 850-870 m a. s. l. Slightly entrenched in this terrace there is a second surface, that in the Giovenco River valley and in the surrounding of the village of Cerchio is at 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.

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, 1981-1982; BLUMETTI & *alii*, 1993; GIRAUDI, 1988).

The central part of the basin between 649 and 660 meters is the bottom of the historic lake.

CHRONOLOGICAL CONSTRAINTS

The latest Pleistocene to Holocene evolution is very well defined by a wealth of archaeological, radiocarbon and tephrachronological data. 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.

Just before the Roman period, the Fucino lake level appears to have been very low. Archaeological data show that a middle Holocene high stand developed at ca. 5.5 ka B.P., while the lake was at a minimum level at 9 to 8 ka B.P. A tephra coming from the eruption in the Mt. Etna volcano area dated ca. 14.5 ka B.P. locally buried lake shore gravel deposits (RADMILLI, 1981; NARCISI, 1993). The gravels contain upper Paleolithic artifacts of «pre-Bertoniana» facies and are related with the maximum level reached by the lake in recent times, dated at ca. 20 to 18 ka B.P. (fig. 13). This maximum high stand produced a prominent wave-cut terrace, well preserved at several sites along the basin margins (RAFFY, 1970). In the meantime, fluvio-glacial sedimentation took place on the Majelama fan, which originated from the Velino Massif (fig. 12). Due to tectonic tilting and faulting, it is very difficult to precisely define the absolute elevation of these past lake levels. However, in a general way it is possible to state that lake level fluctuations are climatically controlled, with high stands closely related with more arid cold climatic stages.

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 paleosol developed from volcanic parent materials at 33 to 31 ka B.P., and 3) a fluvio-glacial sedimentary environment in the interval pre-39.5 to ca. 33 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 39Ar/40Ar age of ca. 540 ka B.P. from a tephra found at a depth of 100 m in the center of the basin (FOLLIERI & *alii*, 1991). With regard to this, it is worth describing in some detail and for the first time the preliminary results of new stratigraphic analyses within the «intermediate terraces», which recently led to the discovery of the first middle Pleistocene mammal remains in the Fucino basin. This is a fragmentary maxillary bone with teeth of an equid found, in September 1996, at Ponte della Mandra site (830m a.s.l.; fig. 2), in a complex alluvial sequence of the Giovenco valley. The sequence have at the base an interglacial reddish soil developed on alluvial gravel, and continue with other soil less evolved, separated by decimetric alluvial levels. Upon this soil sequence there is another alluvial unit in braided facies, that culminate with a depositional surface at about 850m a.s.l. The whole sequence indicate the progressive climatic change from warm humid condition toward cold arid condition. The remains are a clast of one of the decimetric alluvial levels mentioned above. Like other nearby clasts of the same dimension (10 to 20 cm), the edges of the fossil sample are sharp, thus suggesting minimal reworking. The paleontological determination ascribes the fossil remain to *Equus* cf. *altidens*. The dental arc is anatomically connected and consists of P4/, M1/, M2/ and part of M3/; P4/ and M2/ are better preserved. These teeth are damaged in several points, and are embodied in a strongly cemented reddish calcarenitic matrix including also bone fragments and carbonate clasts 0.1 to a few cm in size. The structure and dimensions of the dental arc allow us to rule

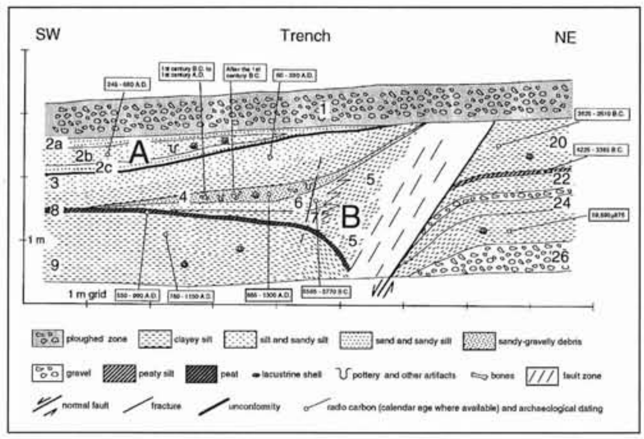

FIG. 13 - Log of the NW wall of the San Benedetto trench (see location in fig. 12). «A» and «B» indicate evidence for pre-1915 historical surface faulting events, as described in MICHETTI & *alii* (1996); numbers mark the stratigraphic units also described in this paper. The paleoseismological analysis clearly demonstrates that the historical seismic catalogue for the Fucino area (fig. 11) is not complete.

out any relation with caballoid horses. Likewise, the affinity with *Equus hydruntinus*, which is commonly found in Italy from the end of the middle Pleistocene to the end of the late Pleistocene, should be excluded based on a) the robustness of the teeth, b) the overall greater dimensions and c) the different morphostructural characters. The thickness of the tooth enamel, the flat interstiloid surface, and the apparently low protocone index allow us to attribute the Fucino equid to *Equus* cf. *altidens*. This species characterizes the 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 (ca. 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-Collarmele-Pescina sequence («intermediate terraces» in fig. 12).

Regarding the chronology of the «upper terraces», two different interpretations can be found in the literature. According to BOSI & *alii* (1995), the Aielli formation is Pliocene in age based on regional stratigraphic correlations. According to RAFFY (1979), the Aielli formation is late early Pleistocene in age, based on the amount of volcanic minerals in the Aielli lake deposits and regional geomorphology. New radiometric, paleontological and tephrachronologic dating is in progress to better constrain the timing of the early to middle Pleistocene evolution of this basin.

The southern and western margins of the ancient lake which occupied the Fucino basin before the latest glacial are not known. High continental terraces are stranded in the footwall of the Celano-Gioia normal fault at elevations up to 1050 m. Most likely lacustrine sediments of the same age are buried below the modern deposits in the hangingwall of this master fault. Boreholes indicate at least 200 m of lacustrine deposits without reaching the substratum, while seismic reflection and geoelectric survey strongly suggest that continental deposits are several hundreds of meters thick in the center of the basin. On the southern and western borders of the Fucino basin we observe only the main younger terrace (at ca. 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 think 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 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 an active extensional tectonics capable of producing strong seismic events.

THE SABATINI VOLCANIC COMPLEX
(S. Ciccacci, D. De Rita & P. Fredi)

The Sabatini volcanic district is placed in northern-central Latium about 20 km to the north of Rome and together with the other Latium volcanic districts, is part of the so-called «Roman Province», an alkali-potassic volcanic province developed inside a tectonically lowered area between the Tyrrhenian coast and the Apennine chain mainly during the Pleistocene.

The geodynamic cause of the volcanism is strictly connected to the post-Miocene rifting of the Tyrrhenian basin and to the post-orogenic extension in the Apennine chain. This extensional tectonism determined a NW-SE trending horst and graben structure, intersected by transversal faults. The alkali-potassic volcanism developed along these faults (fig. 14).

The Sabatini volcanic district developed in a NW-SE trending graben bordered to the west by the Mounts of Tolfa and to the east by the Mt. Soratte-Mts. Cornicolani horst. This main structure is interrupted in the central sector by the Baccano-Cesano structural high, which is interrupted by NE-SW trending faults and is interpreted to be the culmiation of a Miocene thrust sheet structure. The graben and the horst cut an allochthonous Cretaceous-Oli-

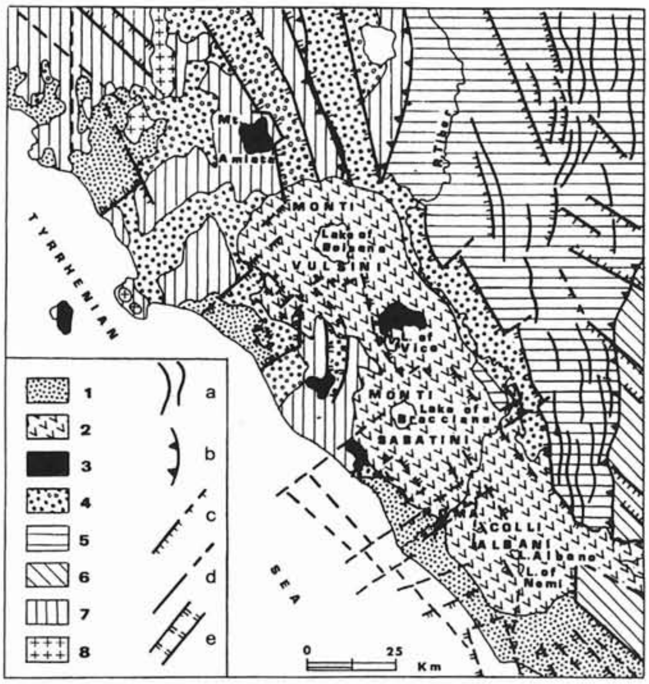

FIG. 14 - Geostructural sketch showing the tectonic arrangement in which the volcanism of Latium developed. 1) Present and recent alluvium; 2) k-alkaline volcanites; 3) plutonites and acid volcanites; 4) sands and clays neoautochthonous unit of Plio-Pleistocene; 5) Umbria-Marche sedimentary units (Trias-Upper Miocene); 6) Latium-Abruzzi carbonatic units (Trias-Upper Miocene); 7) allochthonous complexes (Sicilidi and Liguridi flysches) and Tuscan nappe ; 8) Tuscan autochthonous and paraautochthonous; a) fold; b) thrust c) normal fault; d) vertical fault; e) horst-graben structure (from CAPUTO & *alii*, 1989).

gocene flysch sequence (1000 m thick), which covers a Meso-Cenozoic sequence of pelagic limestones 1500 m thick. Below the volcanic products, Plio-Pleistocene marine sediments (clays, sandy clays and conglomerates) infill the graben for more than 500 m, while on the Baccano-Cesano horst they are either absent or very thin (FUNICIELLO & *alii*, 1976; DE RITA & *alii*, 1983)

The volcanism, with an areal type activity and numerous eruptive centers widespread over an area of about 1500 km^2, caused the emplacement of a huge amount of products prevalently made of pyroclastic flows, hydromagmatic products, lava sheets, and pyroclastic fall products (fig. 15).

The main volcanic edifices are located in the eastern sector of the Sabatini district, at the east of the lacustrine basin of Bracciano. The oldest volcanic activity began at the margins of the outcropping sedimentary structures and progressively migrated to the central sector. The oldest activity started in the western sector with rhyolitic and rhyodacitic products dated around 2.5 Ma. In the eastern sector the first products, dated approximately 0.6 Ma, belong to the Morlupo-Castelnuovo di Porto center at the western margins of the Mt. Soratte-Mts.Cornicolani horst. This center produced trachytic lavas and pyroclastics followed by hydromagmatic surge activity and ended with pyroclastic flow. After Morlupo, the main stage of activity was associated with the Sacrofano center. During the activity of Sacrofano, concurrent volcanism occurred in the other areas of the Sabatini volcanic district. A widespread pyroclastic and lavic activity from several small scoria cones occurred in the northern sector. Hydromagmatic or phreatic explosions produced some coalescent centers along a NNW-SSE trending fracture. Thick lavas and pyroclastic flows that rose along faults caused the collapse of the Bracciano depression. At the eastern border of the lake, small centers had an activity as very small strato volcanoes with limited hydromagmatic activity; moreover, huge pyroclastic flows crop out in the eastern part of the Sabatini volcanic district. The Baccano-Sacrofano eruptive center is one of the major volcanic structures of the Sabatini district.

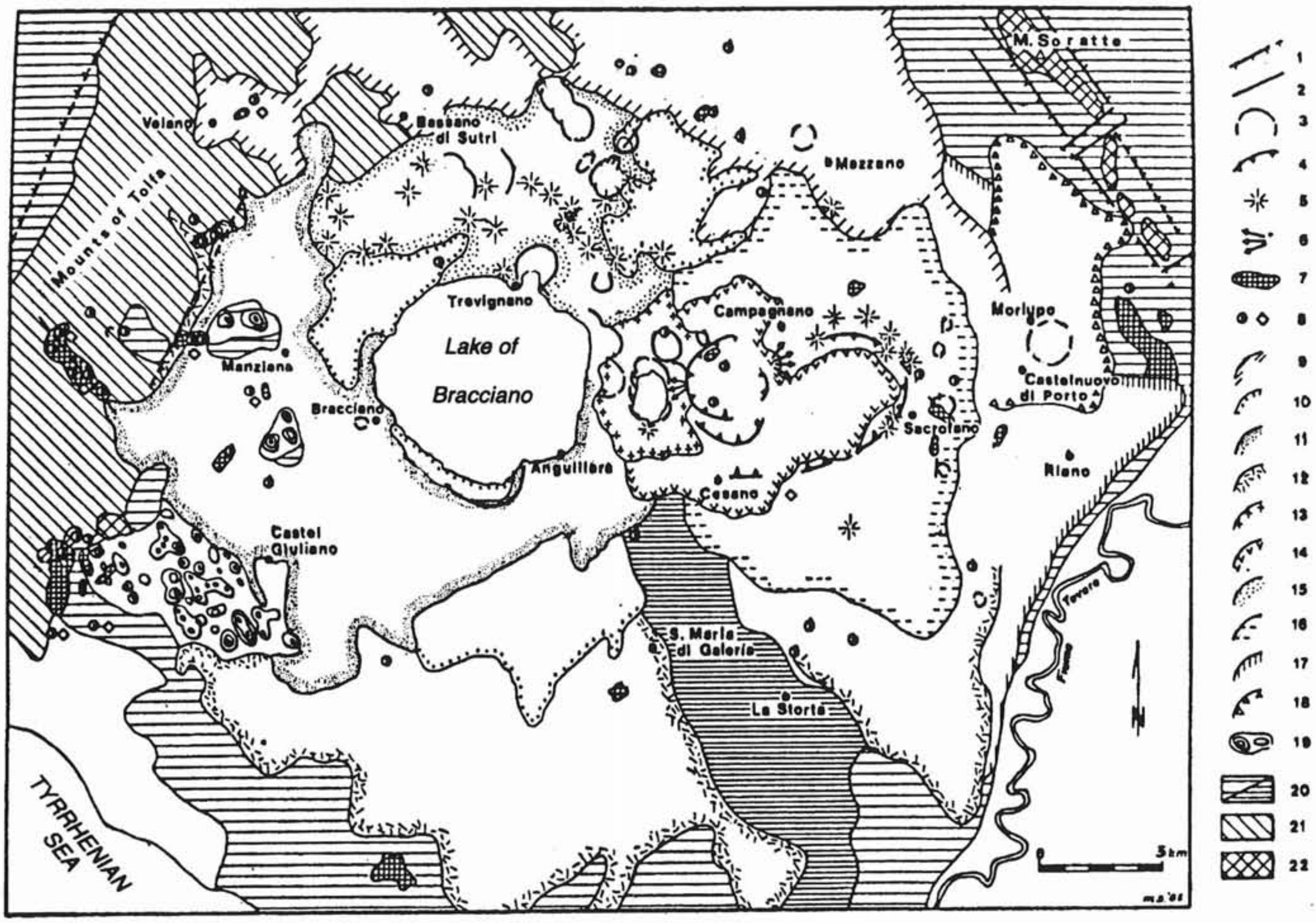

FIG. 15 - Geological map of the Sabatini Mts. volcanic district.: 1) main regional fault; 2) fractures; 3) crater rim; 4) caldera rim; 5) scoria cone; 6) tuff cone; 7) hydrothermally altered areas; 8) thermal springs and gas emission; 9) «Red tuff with black scoria « ignimbrite limits from Vico volcano (0.18 Ma); 10) limits of unsutured lava unit; 11) limits of the Bracciano pyroclastic flow unit; 12) limits of the «Red tuff with black scoria « from Sabatini volcanic district (0.4 Ma); 13) limits of the hydromagmatic products of Martignano craters; 14) limits of the hydromagmatic products of Baccano center; 15) limit of phreatomagmatic and magmatic activity of the northern sector; 16) upper pyroclastic flow from Sacrofano volcano; 17) lower pyroclastic flow from Sacrofano volcano; 18) limits of the explosive activity from Morlupo volcano; 19) acid volcanic products of the Cerite-Tolfa distícts; 20) Plio-Pleistocene clayey sediments; 21) Tolfa allocthonous units; 22) calcareous units of Mt. Soratte- Mts. Cornicolani structure.

This volcano produced the greatest volume of eruptive products. Pyroclastic flows dominated during its activity but scoria cones and lava flows also occurred. The type of activity was related to structural and hydrologic settings that controlled magma/water interaction. The Sacrofano volcano lies above a carbonate platform (Cesano structural high). These limestones, now buried under volcanic deposits, form the reservoir for the regional aquifer. Fracturing related to the distensive tectonics could rapidly mobilize the supply of water leading to phreatic or hydromagrnatic activity. The history of thc Baccano-Sacrofano center can be subdivided into three stages:

- construction of a pyroclastic edifice by predominantly Strombolian-Vulcanian activity (0.5 - 0.4 Ma),
- collapse of the Sacrofano caldera (0.3 Ma),
- development of the Baccano explosive center inside the old Sacrofano caldera and collapse of the Baccano caldera (0.3 - 0.008 Ma).

Hydromagmatic volcanism occurred at the final stage of the Sacrofano activity; it also represents a recurrent kind of activity at Baccano (DE RITA & *alii*, 1983).

The morphological characteristics of the Sabatini Mts. are directly related to the volcanic and tectonic phenomena that affected the area until recent periods and in particular are related to the broad extension of the volcanic blanket, mainly alkaline-potassium pyroclastics and lava flows.

Among the general features of the Sabatini area there is marked contrast between the composite aspects of the central zone, where positive and negative volcanic forms are present, and the landscape which is even and slightly inclined away from the central area.

The positive volcanic forms, made up of scoria cones and of other small strato-volcanoes, are responsible for the greatest elevations in the northern part, where at Monte Rocca Romana, to the north of Bracciano Lake, the maximum elevation (612 m a.s.l.) is reached.

Among the negative forms, the wide subcircular depression, occupied by Bracciano Lake, dominates. The other subcircular or ellipsoidal depressions of Martignano (the site of a lake), of Baccano and of Sacrofano are aligned with Bracciano Lake towards the east. But while these easternmost depressions display a typical - even if complex - feature of volcanic centers with a predominantly explosive activity, the former (Bracciano) does not have such a typology. This agrees with the geologic and volcanologic studies which attribute an essentially tectonic, or at most volcano-tectonic, origin to the depression (DE RITA & *alii*, 1996)

The morphology of the peripheral zone of the Sabatini Mts. is characteristic of the Latium volcanic area; the repeated pyroclastic flows caused the formation of surfaces slightly inclined from the eruptive centers outwards, and on these surfaces the erosive action of the exogenous agents has produced its modelling effects. Among these, the most diffuse and evident are due to channelled running waters that often cut up the volcanic plateau into large shelves separated by incisions and very deep valleys, forming a drainage network that varies in density from zone to zone, in connection with the typology of the outcrops and the local slopes.

As a consequence of the morphological aspects of the volcanic area, the surface drainage shows a generally centrifugal pattern, with a centripetal pattern inside the volcanic and volcano-tectonic depressions of the central sector. However, a more careful observation reveals that many anomalies exist within these patterns. The azimuthal analysis of stream channels, carried out to evaluate the possible tectonic control exerted on the drainage network (fig. 16 -

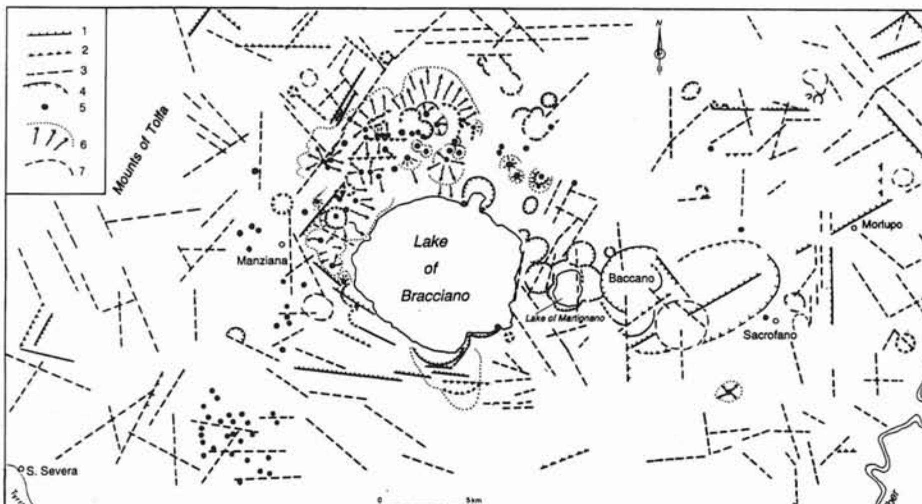

FIG. 16 - Map of volcanic forms and main tectonic directions inferred on a morphological basis: 1) fault; 2) supposed fault; 3) fracture; 4) volcanic depression; 5) volcanic relief; 6) slopes and morphological limit of accumulation round the emission centers; 7) depression of uncertain origin. At the bottom, the rose-diagrams show the orientation of stream channnels (from CAPUTO & *alii*, 1989).

bottom), revealed that the latter is preferentially emplaced in N-S, NE-SW, NW-SE and E-W directions, which are likely to be tectonically controlled (CAPUTO & *alii*, 1989).

The existence of these hypothetical tectonic directions was strengthened by results of the analysis of the morphological evidence of tectonics (BIASINI & *alii*, 1993). The study evidences that the tectonic directrices - drawn from the alignments of significant morphological features and from the location and shapes of the volcanic landforms - are preferentially oriented in NW-SE, NE-SW and N-S direction (fig. 16).

During the field excursion, only the easternmost sector of the Sabatini area will be examined in detail; this area stretches between the Tiber lower valley to the east and the Bracciano lake to the west. The aim is to analyse some important volcanic landforms of the region and the influence of tectonics on its geomorphological evolution.

First of all, the area of Morlupo-Castelnovo di Porto will be considered; this area is located on the eastern edge of the graben, in which the whole Sabatini volcanism evolved, and is the place where the Morlupo eruptive center developed about 0.6 Ma ago. This center, the oldest of the eastern sector of the volcanic district, is not discernible in the field, but its pyroclastics cover most of the area, even if some products of the westernmost and younger Sacrofano center are present. Morphodynamic processes are particularly marked all over the area; the morphogenetic action is mainly due to the channelled running waters; they cut the volcanic cover very deeply, which often makes the Plio-Pleistocene clayey-sandy lithologies crop out (fig. 17). Morphologic and morphometric characters of the drainage network show that many anomalies exist in the spatial arrangement and in the flow direction of fluvial channels, the latter being often inconsistent with regional and local slopes. Moreover, morphometric parameters show generally anomalous values, which testify for a poor organization of the drainage networks. Furthermore, the areal distribution of drainage density and of the relief amplitude values, calculated per unit of area, shows frequent irregularities that cannot be explained by lithological conditions alone.

Geomorphological studies evidenced that the main part of the morphological situation clearly depend on re-

FIG. 17 - Morphological map of Morlupo - Castelnuovo di Porto area: 1) structural subhorizontal surfaces; 2) main watershed; 3) secondary watershed; 4) drainage network; 5) river bed erosion deepening; 6) V-shaped small valley; 7) trough-floored small valley; 8) flat-floored small valley; 9) fluvial erosion scarp and polygenetic scarp influenced by structure; 10) fault scarp; 11) fault line scarp; 12) saddle corresponding to buried river beds; 13) alignment of mineral springs; 14) caldera rim. (from CICCACCI & *alii*, 1989).

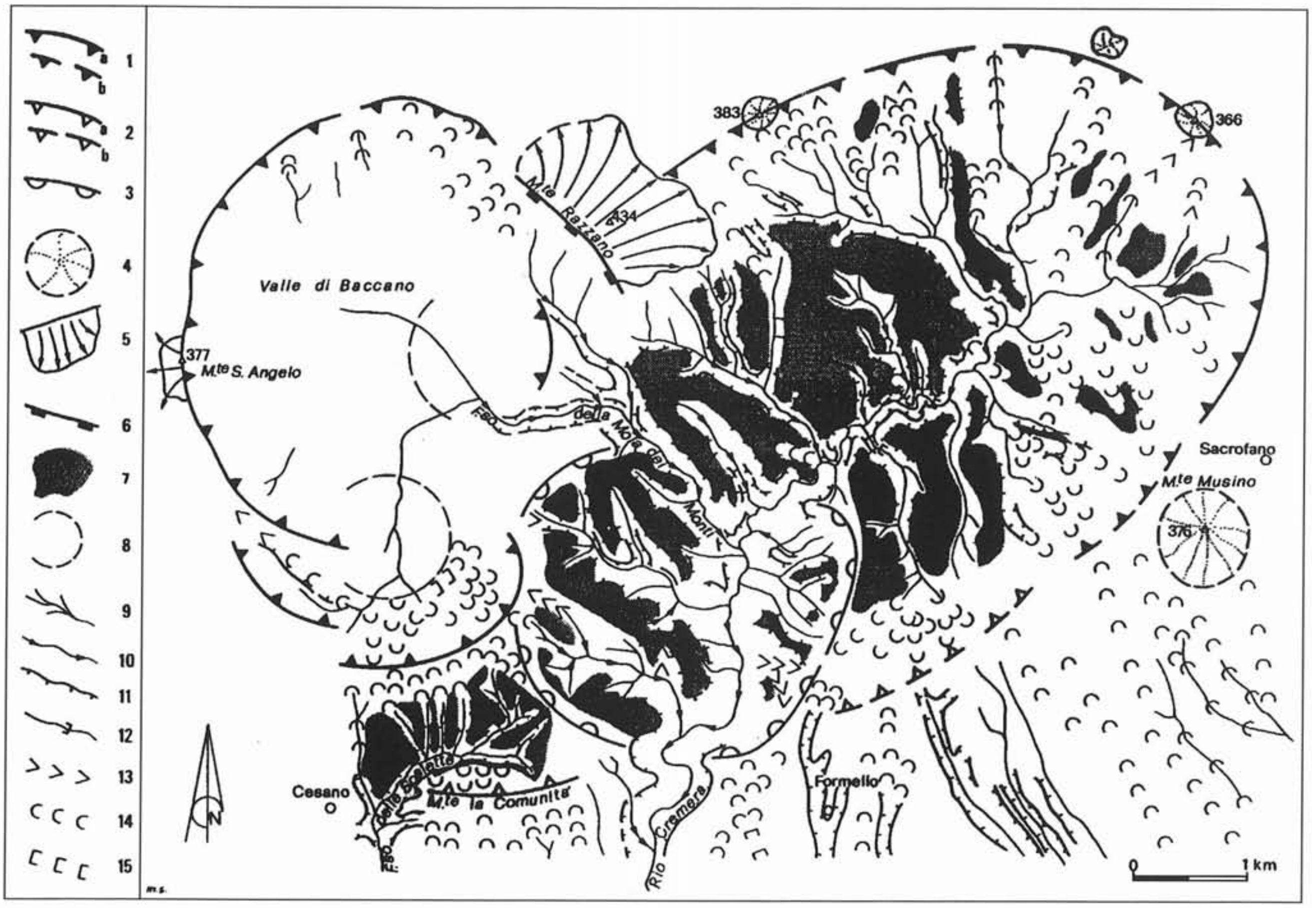

FIG . 18 - Morphological map of the Sacrofano-Baccano area: 1) Caldera rim, a) continuous, b) discontinuous; 2) Caldera buried rim, a) evident, b) hypothetical; 3) Sub circular depression edge of uncertain origin ; 4) Scoria cone; 5) Tuff cone; 6) Fault scarp; 7) Structural suborizontal surface; 8) Buried depression edge corresponding to thick volcanic breccias; 9) Drainage network; 10) River bed erosion deepening; 11) Fluvial erosion scarp; 12) Knick along stream channel; 13) V shaped small valley; 14) Trough-floored small valley; 15) Flat-floored small valley (From CICCACCI & *alii,* 1988).

cent tectonic directrices trending NW-SE, NE-SW and N-S (CICCACCI & *alii*, 1988). A little more to the west, the same tectonic alignments are easily recognizable in the Sacrofano-Baccano zone, where some of the most important volcanic edifices of the Sabatini area are present (fig. 17).

Analyses carried out allow us to establish on the whole the role that tectonic alignments played in defining the morphological evolution of the eastern Sabatini area (fig. 18). The outstanding alignment of NE-SW directed morphotectonic evidence, located to the north of the town of Sacrofano, was considered as corresponding to a tectonic dislocation of regional importance. It is along this direction that the Sabatini volcanism migrated with time and that the Sacrofano central edifice developed, giving rise in a late phase to the large homonymous caldera with the major axis trending NE-SW (fig. 19).

The NW-SE tectonic direction played a primary role both in the volcanic and morphological evolution of the area; it is this very direction which led the progressive sinking of the graben where the Latium volcanism developed. Later on, the same direction conditioned the collapse of the Baccano caldera and the evolution of its drainage network

Very important, finally, is the role of the N-S tectonic direction. Morphological studies showed that this direction strictly controlled the most recent morphological evolution of some important valleys. Moreover the azimuthal analysis of drainage network evidences that the N-S direction markedly prevails in the stream channels of 1st and 2nd order, which are likely to have been emplaced later than the other stream orders. For this reason, it is possible to suppose that the N-S tectonic directrices had been active in very recent time after the most important volcanic events. Recent geological and volcanological works (FACCENNA, 1993; DE RITA & *alii*, 1996) confirm this hypothesis: they point out, in fact, that the Central Latium area has been affected by a N-S strike-slip tectonism almost from 0.17 Ma ago to Late Pleistocene time. Moreover, this tectonic phase controlled the final hydromagmatic activity of the Sabatini volcanic distict, in the area between the Baccano caldera and the western side of Lake of Bracciano.

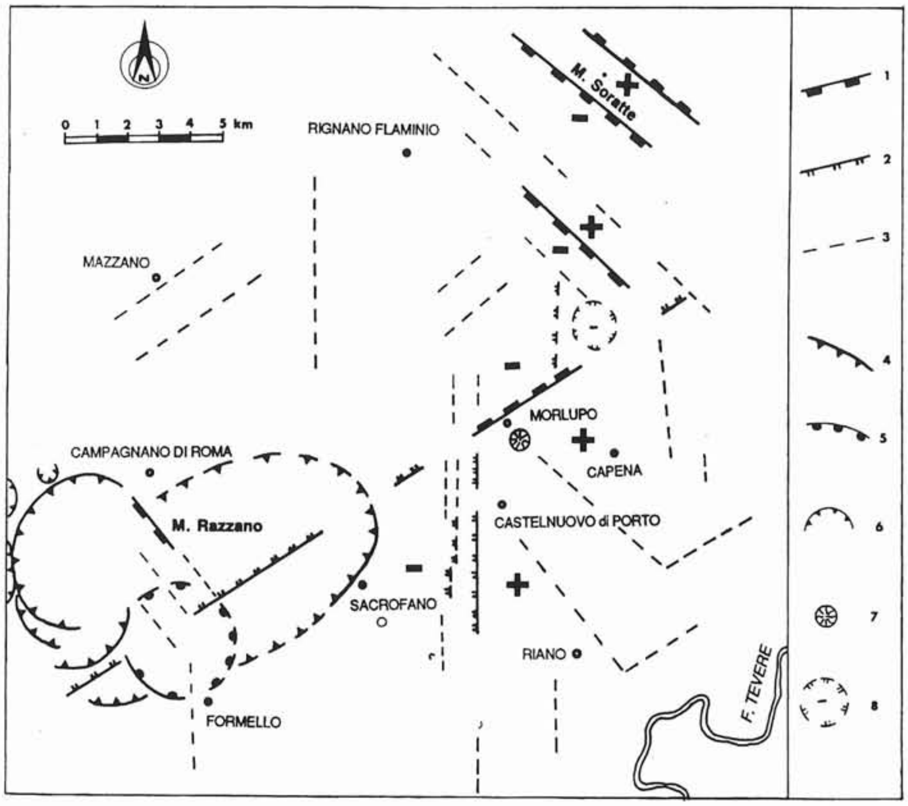

FIG. 19 - Morphotectonic sketch of the eastern Sabatini area.: 1) Fault; 2) Hypothetical fault; 3) Fracture; 4) Caldera rim; 5) Depression of hypothetical volcano-tectonic origin; 6) Crater rim; 7) Trachytic lavas ; 8) Depression of uncertain origin (From CICCACCI & *alii,* 1989).

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