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FIELD TRIP GUIDEBOOK

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CONTENTS

1. Geological Background for the evolution of the Apennines during the Neogene and the Quaternary (Coltorti M. & Pieruccini P.)

The Italian peninsula is made of an arcuate shaped mountain chain (the Apennines, FIG1), that reaches 2,912 m a.s.l. in the Gran Sasso Mt., with a mean altitude of about 1,500 m a.s.l. The Apennines form the watershed between the Tyrrhenian and the Adriatic drainage basins, even though the divide often does not match the maximum altitudes (MAZZANTI & TREVISAN, 1978).

The Apennines constitute a Neogene thrust-fold belt (FIG.2A) developed in a non-metamorphic regime, although low grade metamorphic and granite rocks are, in places, involved in the thrust sheets. The thrust fronts have complex imbricate geometry with a direction of transport mainly to the NE in the northern Apennines turning to the NS and NNE in the Calabrian area and EW in Sicily. The units involved in the tectogenesis belong to sedimentary sequences (FIG.2B), many kilometres thick, ranging from Palaeozoic to Lower Pleistocene in age. The onset of the "compressional" events is commonly explained as the consequence of the progressive migration to the east of the thrust fronts, today located in the Adriatic and Ionic sea (ELTER et al., 1975; BALLY et al., 1986; BARBERI et al., 1994; CARMIGNANI et al., 1994; BARCHI et al., 1998). This process is supposed to start during the Burdigalian with the opening of the Tyrrhenian Sea. The foreland basins, created during the eastward migration of the compression and containing different sedimentary sequences, were split by several overthrusts. In this classic model, the migration of the compressional front to the east is soon followed by the activation of extensional faults to the west.

In the north-western area, low angle E-dipping normal faults led to the creation of the "Serie Ridotta" with, in places, "tectonic omission" of overthrusted structural units (DECANDIA et al., 1998). The timing of the emplacement of the different thrust units is usually determined taking into account the age of the youngest formation involved. Their last activity is dated to the Early Pliocene in the northern Apennines (FIG. 2B), to the Middle-Upper Pliocene in the eastern areas (Gran Sasso and Maiella mountain fronts: CALAMITA et al., 1991; BIGI et al., 1991; 1996) and up to the Lower-Middle Pleistocene in the Southern Apennines (PATACCA ET AL., 1991). As previously mentioned, in the north-western side, most authors (ELTER et al., 1975; CALAMITA & DEIANA, 1988; 1995; BALLY et al., 1986; MARTINI & SAGRI, 1993; BARCHI et al., 1998; DECANDIA et al., 1998 and reference therein) suggest that the thrust fronts migrated to the west, since the Tortonian, together with the onset of extensional faults and related grabens on the back of the chain. However, many authors (BOCCALETTI et al., 1995; 1999; 2002 and reference therein; COLTORTI & PIERUCCINI, 1997a; 1997b; CALAMITA et al., 1999; BERNINI & PAPANI, 2002) evidenced that the tectonic basins in the western side of Italy have a compressional setting and interpreted them as “piggy back basins” on which extensional tectonics are superimposed only since the end of Early Pleistocene. “Compressional structures” were still active up to the Late Pliocene in the front of the chain. CALAMITA et al. (1999) and BERNINI & PAPANI (2002) ascribed these "compressive" features to the activity of east-dipping low angle normal faults that involve also the Paleozoic basement ('K Horizon' of DECANDIA et al., 1998; 'Chiana fault' of BARCHI et

al., 1998; 'Alto Tiberina fault' of BONCIO et al., 1998). The onset of high-angle extensional tectonics on the western side of the Apennine chain, with locally originating graben or semi-graben basins filled with continental deposits, displace the thrust planes and date back only to the Early Pleistocene (COLTORTI & PIERUCCINI, 1997a; 1997b; CALAMITA et al., 1999). However, many NW-SE trending normal faults re-activated older structures through processes of tectonic inversion (DECANDIA & TAVARNELLI, 1991; CALAMITA et al, 1994; BIGI et al., 1996).

Thrust fronts are also present further to the east, in the Periadriatic Basin, as well as in the Po Valley and the Adriatic sea where their timing is constrained by the Zanclean and the late Early Pliocene transgressional SURFACES (ORI ET AL., 1986; CALAMITA ET AL., 1991; ARGNANI & GAMBERI, 1996). Since the end of the Early Pliocene, strong uplift movements affected the axis of the chain whereas in the Periadriatic Basin they started at the end of the Pliocene, and today Late Pliocene coastal deposits can be found up to 1100 m a.s.l. (NANNI et al., 1986). The Adriatic sea is affected by subsidence. The present-day coast line is almost coincident with the limit between these two areas.

The area crossed during the Field Trip is made up of the Umbro-Marchean Structural Unit (FiIG. 2A-B) that constitutes the southern part of the Northern Apennines (FIG.1). It is a thrust-fold belt made of rocks belonging to the "autochtonous" Umbro-Marchean sedimentary sequence, although local thrusts affect the sequence (CENTAMORE & DEIANA, 1986). The older rocks (Trias to Oligocene in age) are mainly carbonatic whereas the younger units (Miocene) are terrigenous deposited both in the Foredeep (Marnoso-Arenacea Formation; Laga and Cellino Formation) or in Satellite (piggy back) basins generated inside the frontal thrusts of the Umbro-Marchean Domain (RICCI LUCCHI, 1986; 1987). In the northernmost sector, the allocthonous Ligurian and Epiligurian Units unconformably overlie all the previously mentioned units (FIG.2B). The Ligurian Units are mainly made of chaotic Cretaceous shaly-clayey, calcareous and marly rocks unconformably overlaid by the Epiligurian Units deposited in piggy back basins (Val Marecchia Gravity Flow, DE FEYTER, 1991). Their emplacement, in the frontal part of the chain, is dated up to the Early Pliocene because rocks of this age are buried by the thrusted units (CONTI, 1991; DE FEYTER, 1991). A very rapid denudational processes occurred since their emplacement because AFT and ZFT (apatite and zircon fission tracks) suggest that, in places, between 4 and 5 kilometres of rock, supposed to belong mostly to these allochtonous units were eroded since 5-4 Ma ago (ABBATE et al., 1999; ZATTIN et al., 2000; BALESTRIERI et al., 2003; BALESTRIERI et al., 2003). A denudation of ca 2000 m in the same time span has been recently pointed out also in the Central Apennine (CALAMITA et al., 2003).

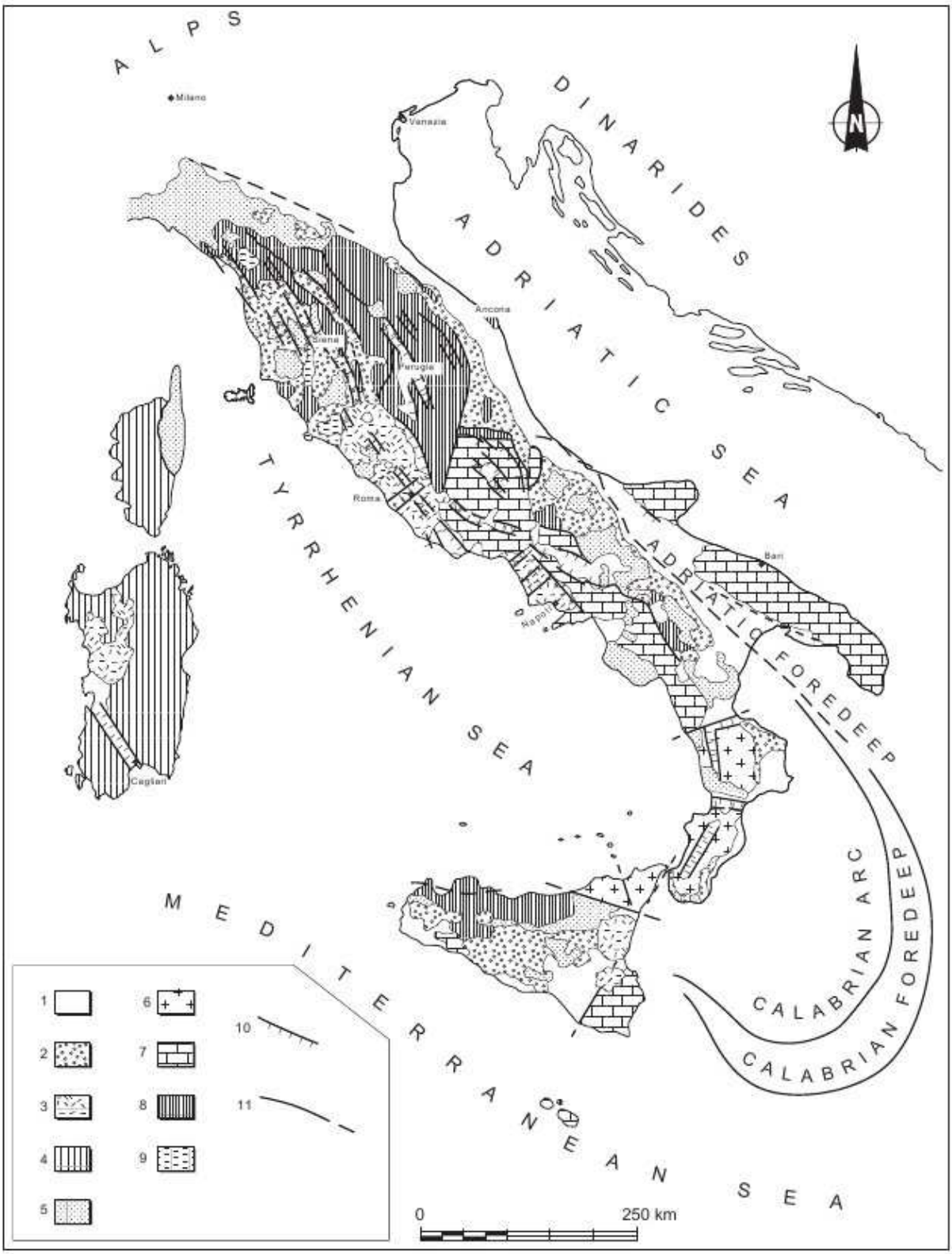

Fig. 1 – Structural scheme of the Italian Peninsula
1-Quaternary continental and marine deposits; 2-Pliocene continental and marine deposits; 3-Quaternary volcanic deposits; 4-Sardo-Corsican Massif; 5-Ligurian, Molisan and Sicilian Units; 6-Calabrian Units; 7-Latium-Abruzzi, Apulian, Iblean and Campanian-Lucanian Units; 8-Tuscan, Umbro-Marchean, Molisan, Lagonegrese, Imerese and Sicano Units; 9-Tuscan Autochton and Parautochton; 10-Normal faults; 11-Main thrust fronts.

2. Geomorphological background to the Apennines evolution during the Neogene and the Quaternary (Coltorti M. & Pieruccini P.)

The landscape of the area is mainly characterised by NW-SE trending mountain ridges (from west to east: the Amelia-M.Peglia Ridge, the Martani Ridge, the Umbro-Marchean Ridge, the Marchean Ridge, FIG 2A-C) separated by hilly morphologies with transversal and longitudinal valley systems. The ridges roughly correspond to the limestone terrains bounded to the east by thrust fronts whereas the hilly areas are modelled on the Miocene terrigenous sediments although minor and shorter limestone ridges are present. The maximum altitudes are reached in the southern sector where most of these ridges converge to generate the Sibillini Mts. The almost regular repetition of ridges and basins is interrupted in the northernmost sector where the mountain morphology (i.e. Carpegna Mt, Sasso Simone) is irregular due to the presence of the chaotic Ligurian terrains of the Val Marecchia Gravity Flow. The front of the Apennines, that coincide with the main overthrust front, runs along the so-called Antrodoco-Urbino Line, that delimits to the east of the Sibillini Mts. To the east of this line, in the Periadratic basin, there are other overthrusted units, partially outcropping from the Plio-Pleistocene sediments of the Periadriatic Basin (Cingoli Ridge, Monti della Cesana, Mt. Conero , etc.) but their elevation is always very limited. The hilly landscape of the Periadriatic Basin is mostly modelled on the Pliocene terrigenous deposits that reach altitudes of 1100 m a.s.l. at the Ascensione Mts. The mean elevation of this area (Periadriatic Basin) gradually decreases and in the easternmost sector that is cut by wide alluvial plains. A series of Early to Middle Pleistocene transgressional-regressional cycles from a coastal to outer neritic environment, are recorded in the area closer to the coastline (NANNI et al., 1986). The uppermost units are represented by a series of coalescent alluvial fans interlayered with coastal deposits. They generate a large terrace whose inner part, originally alimented from the mountain sector has been completely eroded.

In the Umbria and Marche Region four orders of alluvial terraces are well preserved in the valleys. The older terraces (Ist, IInd and IIIrd order) were deposited during the main cold phases of the Middle-Upper Pleistocene and dissected during the Interglacials (COLTORTI et al., 1991) whereas the dynamics of the Holocene terrace (IVth order) are mostly the result of human activities (COLTORTI, 1981; 1997; GENTILI & PAMBIANCHI, 1987). They are convergent terraces due to the interactions between the uplift of the chain and the Quaternary climatic changes (ELMI et al., 1987).

3. The Planation surface (PS) (Coltorti M. & Pieruccini P.)

Almost all the Apennine mountain summits are characterised by flat erosional surfaces (Planation Surface, PS; FIG. 3). Similar morphologies have been described across the Apennines from north to south (DEMANGEOT,1965; DESPLANQUES, 1969; GUEREMY, 1972; BERNINI et al., 1977; BARTOLINI,1980; SESTINI,1981; DRAMIS et al, 1991; CINQUE , 1992; AMATO et al., 1992; BOSI et al. 1996; COLTORTI & PIERUCCINI, 1997b). All the Authors agree that these surfaces are the remnants of low energy landscapes, modelled close to sea level, before the uplift and mountain building of the

Apennine chain in a period of relative tectonic stasis and predominance of denudation processes.

The main debate about the Apennine planation surface (PS) regards: A) what is the morphogenetic process responsible for this modelling; B) the number of events involved; c) the chronology.

According to COLTORTI & PIERUCCINI, (1999, 2002) the Apennine chain was planated close to sea level at the end of the Early Pliocene, and the PS is mainly the result of processes of marine erosion. The PS is correlated to the late Early Pliocene transgressional surface that in the Peri-Adriatic Basin cuts the Early Pliocene marine deposits and is buried under younger marine sediments (SIGNORINI, 1948; CANTALAMESSA et al., 1986; CALAMITA et al., 1991; 1995; ORI et al., 1986; ARGNANI & GAMBERI, 1996); to the south this unconformity and has also occasionally been observed in the Chain sector were it cuts Early Pliocene deposits (Formazione di Monte Coppe; Formazione di Rigopiano; VEZZANI & GHISETTI, 1998; BIGI et al., 1996). This transgressional event has been recognised in many parts of the world and dated between 3,7 and 3,2 Ma (PICKARD et al. 1988; CRONIN & DOWSETT, 1994; BERGREEN et al.,1995a,1995b).

The PS gently dips toward the Adriatic Basin since the northern Tyrrhenian was not yet opened, as suggested by the composition of late Early Pliocene coastal and continental clastic deposits made, among the others, by metamorphic coarse-grained pebbles (COLTORTI & PIERUCCINI, 2002 and ref. therein) indicating the western provenance of the watercourses. After its modelling, the PS underwent to positive (in the chain sector) and negative (i.e. Tiber and Peri-Adriatic Basins) tilting and folding tectonic deformations (CALAMITA et al., 1999; COLTORTI & PIERUCCINI, 1999; 2002). In the areas affected by negative movements it was buried under younger sediments. During the Pliocene and the Quaternary the PS was also affected by extensional faults, mainly NW-SE trending on the western side of the chain and by NE-SW trending transtensional faults along most of the chain (COLTORTI et al., 1996). NE-SW trending transtensional faults are also found above all in the southern sector of the Northern Apennines although in the study area these faults are absent.

The PS was later dissected by valleys filled with Pleistocene alluvial terraces. It constitutes a marker for the evaluation of the uplift after its modelling. However, in the peri-Tyrrhenian and peri-Adriatic areas, close to the coastline, there are younger trasgressive events of minor extent that modelled planation surfaces (and unconformities). In fact, in places, they also cut Early Pleistocene deposits (COLTORTI & PIERUCCINI, 2002).

In conclusion, the main assumptions about the PS are:

1. the PS was a plain of marine erosion modelled during the global transgression at the end of the Early Pliocene (ca. 3,7-3,2 Ma);
2. in the central sector of the Apennine chain there is a single PS.

The PS can be used as a morpho-stratigraphical marker that indicates the starting-point for the final emersion of the Apennine. The PS is also a valuable key-tool for neotectonic studies allowing the distinction between the pre-planation and post-planation structures (OLLIER, 1999). It is the starting-point to evaluate the uplift and the erosional rates after the late Early Pliocene.

Figure 3 - The planation surface across the Apennine: a view from the ESE

THE WESTERN SIDE OF THE PENINSULA: FROM PLIOCENE SYNFORM BASINS TO PLEISTOCENE EXTENSIONAL BASINS.

4. The Tiber Basin (Coltorti M. & Pieruccini P.)

The Tiber basin extends from Terni, to the south, to Borgo S. Sepolcro, to the north, where it almost joins the Upper Valdarno Basin (FIG.2A) (LOTTI, 1917; 1926; ALBANI, 1962; GE.MI.NA, 1963; CONTI & GIROTTI, 1977; CATTUTO et al., 1979; AMBROSETTI et al., 1978; 1987; BOCCALETTI et al., 1986; BARBERI et al., 1995; MARTINI & SAGRI, 1993). South of Perugia the Martani Ridge subdivides the basin into two portions (West Tiber Basin-WTB and East Tiber Basin-ETB) delimited to the west by the Narni-Amelia and M.Peglia Ridge, and to the east by the Umbro-Marchean Ridge (FIG.4). The WTB and ETB are filled with fluvial deposits up to 500 m thick; the older sediments are late Early Pliocene in age and constitute the older terrestrial evidence and the marker for the beginning of the Mountain Building of the Umbro-Marchean Apennines. It must be noticed that coheval basins with similar fluvial sequences are present in the Northern Apennines (Barga-Castelnuovo Garfagnana, Aulla-Olivola, Valdarno Basins, etc.) whereas both on the Tyrrhenian (Volterra, Valdarno, Siena-Radicofani Basins, etc.) and the Adriatic sides (Periadriatic area), marine deposition was still ongoing.

The Narni-Amelia, Peglia and Martani ridges, are bounded to the east by NW-SE trending overthrusts (CALAMITA et al., 1994). The western structural units overthrust the eastern one as well as the Marnoso-Arenacea Formation once lying on their top (DE FEYTER, 1991; DAMIANI et al., 1995). Further to the north, the Peglia Ridge Unit also outcrops as a tectonic window below the Cervarola Unit (DAMIANI et al., 1993). In the south-eastern part of the Martani ridge the thrust turns gradually into a high-angle transpressive-strike slip fault (Battiferro-Cottanello fault, CALAMITA & PIERANTONI, 1995). The structuring of this sector is dated to the late Serravallian (DECANDIA & GIANNINI, 1977) and/or to the Tortonian (DEIANA & PIALLI, 1994). However, severe re-activation of the Narni-Amelia and Martani overthrusts are referred to the Upper Messinian-Early Pliocene (CALAMITA et al. 1999; COLTORTI & PIERUCCINI, 1997b). Low angle NE-dipping normal faults are present in this area, north of Perugia, with more than 5000 m of displacement (MINELLI & MENICHETTI, 1990; BROZZETTI, 1995). They are related to an E-dipping detachment plane with a staircase geometry recognised in seismic profiles below the Apennine ridge where they reach a depth of 12-13 Km ("alto-Tiberina" fault, BARCHI et al., 1996; BONCIO et al., 1995). These authors suggest that this fault is still active and responsible for the recent seismic activity (BONCIO et al., 1995) although COLTORTI & PIERUCCINI (1997a; 1997b) pointed out that they were active previous to the late Early Pliocene. In fact the summits of the ridges are characterised by the Planation Surface (PS) (COLTORTI, 1996; COLTORTI & PIERUCCINI, 1997b) that indicates as the emplacement of the different structural units took place before the late Early Pliocene. Late Early Pliocene deposits also seal the Battiferro-Cottanello fault

at Montebibico as well as in the Spoleto basin (COLTORTI et al., 1997; COLTORTI & PIERUCCINI, 1997a).

The western slope of the Martani Ridge and the Umbria-marche Ridge are fault escarpments related to the presence of high-angle faults with trastensional to normal geometries (Martana and Spoleto-Foligno Faults).

Figure 5 – The Planation Surface on top of the Narni-Amelia Ridge

4.1 The West Tiber Basin (WTB) (Pieruccini P.)

The WTB sedimentary sequence (FIG.6), described by BASILICI (Fosso Bianco Fm,1997), is made of late Early-Middle Pliocene mainly fine-grained sediments with lignites deposited in a complex environment comprising lakes, swamps, deltas, alluvial fans and ribbon channels. BASILICI (1997) interpreted the depositional environment mostly as a wide and quite deep lacustrine basin with coarser grained sediments deposited by turbidity currents on wave dominated or wetland coastlines with rare alluvial fan or delta bodies. The age is based on the findings of mammal fauna (Triversa Unit, TORRE et al., 1992), palynological and magnetostratigraphical evidence (AMBROSETTI et al., 1995b; BASILICI, 1997).

At Todi, alluvial fan deposits with western provenance (Ponte Naia Fm, BASILICI 1995; ABBAZZI et al., 1997) unconformably overlie the older deposits and are dated to the Late Pliocene due to the presence of mammal fauna belonging to the Montopoli Unit (TORRE et al., 1992).

The Early Pleistocene sequence lies unconformable on the older deposits and is characterised by coarser-grained facies typical of wandering to meandering rivers, although the malacofaunas revealed hypo-haline conditions related to the vicinity of the coastline located to the west of the Narni-Amelia-Peglia Ridges (GIROTTI and PICCARDI, 1994). Its age is constrained on malacofauna and mammalofauna remains (S.Maria di Ciciliano Fm, AMBROSETTI et al., 1995b).

The total outcropping thickness of the Plio-Pleistocene sequence is about 550 m, although AMBROSETTI et al. (1993) assumes a thickness up to 2000 m in the southernmost sector based on gravimetric data.

There is debate about the origin of the WTB. Classically, it has been interpreted as a graben (AMBROSETTI et al., 1978; 1987; CATTUTO et al., 1979; BASILICI, 1997) or semi-graben (BARCHI et al., 1991). However, the Pliocene sediments do not show lateral facies changes close to the border of the basin and the existing normal faults displace the whole sequence suggesting that their activity is younger. Moreover, the Pliocene deposits rest in onlap along most of the western side of the Basin where they lie over a major unconformity that cuts the older thrust planes. The stratigraphical evidence as well as structural studies provided evidence for a compressional origin and evolution of the Pliocene Basin as a synform or satellite (piggy-back) basin (FIG.7) (BOCCALETTI et al., 1995; BONINI, 1998; COLTORTI & PIERUCCINI, 1997). The former authors firstly related this evidence to the presence of alternated periods of compressional/extensional regimes but in more recent times they pointed out that the whole evolution occurred in a compressional regime. In fact, the Pliocene sediments are affected by a a series of progressive unconformities. CALAMITA et al., (1999) suggested that the superficial folding is the result of a reactivation of a deeper detachment along the east-dipping low angle normal faults.

A main ca. NNW-SSE trending fault (Martana fault) is present along the eastern border of the Basin where it forms a fault escarpment more than 30 km long. This fault line has been considered as the master fault bordering the Plio-Pleistocene Basin (AMBROSETTI et al., 1987; MARTINI & SAGRI 1993; BARCHI et al., 1991). However, recent studies (PIERUCCINI & PIZZI, in prep.) demonstrated that this fault (with transtensional characteristics) was activated during the Early Pleistocene cutting the previous synform basin deposits. Coarse-grained debris rich facies are present close to the fault inside the Early Pleistocene sediments that are deformed up to a near vertical dip next to the fault plane. The end of the main activity of the fault is marked by the presence of Middle Pleistocene carbonate rich sediments (Acquasparta Fm, AMBROSETTI et al. 1995) all along the fault escarpment (FIG.8) that unconformably overlie the older units and are displaced at most by a few tens of metres. These sediments are mainly made of travertines (phytoclastic travertines, fissure ridge and spring travertines GOLUBIC et al., 1993; HANCOCK et al., 1999), carbonates muds, as well as carbonate-rich alluvial deposits.

The activity of the Martana Fault is related also to the E-W normal fault which is responsible for the sinking of the Terni Basin and the consequent divertion of the Tiber River at Todi. After this moment the deposition along the eastern border of the WTB is mainly characterised by debris-slope and alluvial fans sediments (Middle Pleistocene-Holocene) and the activity of the Martana Fault almost ceased as indicated also by the present-day low seismicity of the area (PIERUCCINI & PIZZI, in prep.).

In conclusion, the WTB evolved as a synform basin during the Pliocene when the accommodation space for the sedimentation was provided by large-scale folding due to the activity of compressional structures possibly as surface expression of a low-angle extensional regime (CALAMITA et al., 1999) or in a compressional regime (BOCCALETTI et al., 1996; BONINI, 1998). The Planation Surface, present on top of the western and eastern carbonatic Ridges as well as on top of the marly-arenaceous relief, was deformed and the late Early Pliocene WTB continental sediments constitute the first evidence of the uplift of the Apennine chain. The sediments are mostly fine grained deposited

inside a complex environment characterised by the presence of lakes and swamps. The Late Pliocene, scarcely represented, is characterised by an increasing coarse fraction and the deposition of alluvial fans. This phase may be related to the onset of cooler conditions (OIS 100, Gelasian stage RUDDIMAN & RAYMO,1988) that interacted with the ongoing vertical uplift. The drainage directions are not well established although the presence of Early to Middle Pliocene marine sediments to the west of the Narni-Amelia-Peglia Ridge suggest a direction of drainage inside the WTB to the south. However, a low-energy landscape lasted up to Early Pleistocene times as revealed by the hypo-haline conditions of the co-heval WTB sediments and by the presence of lithofagus holes few tens of metres below the relief summit to the west of the Narni-Amelia-Peglia Ridge (GIROTTI and PICCARDI, 1994). The onset of stronger vertical uplift movements and high-angle transtensional fault activity (Martana Fault) created, after the Early Pleistocene, steeper slopes and contributed to the onset of the present day drainage network. Although a steep fault escarpment is present, the sedimentation is driven by climate in terms of repeating cycles of sedimentation and pedogenesys. In fact, at Massa Martana the late Middle Pleistocene alluvial fan is weathered on top by the Eemian pedo-sequence and buried under the late Pleistocene alluvial fan (FIG. 9 COLTORTI & PIERUCCINI, 2004).

Figure 8 – The Martana fault escarpment from the south.

4.1.1 STOP 1

THE DUNAROBBA FOSSIL FOREST

From the http://www.forestafossile.it

ABSTRACT

The Dunarobba Fossil Forest (Terni, Umbria, Central Italy): lithostratigraphic, sedimentologic, palynologic, dendrochronologic and paleomalacologic characteristics.

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The Dunarobba Fossil Forest (DFF) is a singular paleontological case: namely, it is a perfectly preserved Pliocene forest formed of in situ fossil trees, more than 50 in number and up to 8 m high. The DFF is located near the village of Dunarobba (near Terni, Umbria, in Central Italy), and is enclosed in the deposits of the Tiber Basin, which is an extensional basin crossing Umbria from north to south. Continental deposits, from Pliocene to Holocene in age, fill this basin. Four lithostratigraphic units have been recognized in the studied area between the towns of Marsciano and Terni: -"Fosso Bianco" formation (FBf, Middle-Late Pliocene), which was deposited in a complex lacustrine system; - "Ponte Naja" formation (PNf, Late Pliocene), formed of alluvial fan deposits; - "Santa Maria di Ciciliano" formation (SMCf, Early Pleistocene), formed of fluvial alluvial deposits; - "Acquasparta" formation (Af, Early Pleistocene), which is a deposit settled into small isolated lacustrine carbonate basins. The DFF is enclosed in the "Fosso Bianco" formation, in deposits attributed to a coastal lacustrine wetland Tree trunks are buried by muds and subordinate lignite and sands (facies C1 subassociation, lithofacies a-e). The base of the trunks rests on dark buish grey clayey silts (facies c), interpreted as little evolved and hydromorphic paleosols. Clayey silts with silty-sandy non-

continuous undulated laminae, sometimes displaying cross-laminations (facies b) and clayey silts with planar, parallel and continuous laminae (facies a) make up deposits yielded by weak wave motion (or distal deposits of a delta system) and small ponds on a wetland lacustrine coastline, respectively; lignite (facies d) evidences swamp organic deposits, whereas uncommon sand lenses (facies e) can be interpreted as due to wave motion (or as deposits of a delta body). The DFF depositional environment was an area subjected to floodings, where the groundwater level was near, or above, the depositional plane for many months in the year. Slow and continuous sedimentation of clastic materials and a high subsidence caused trunks to be buried still during their life. Near the DFF other lacustrine coastal deposits outcrop, which however do not contain in situ fossil trunks (facies C2 subassociation). This depositional succession is characterized by lenticular sandy strata showing planar parallel or undulated laminations (facies f), alternating with muds (facies g); these deposits correspond to sequences formed during stormy wave motion and fair weather, respectively. A clayey sandy silty hydromorphic paleosol, with lignite beds on top (facies h) overlies these deposits. Laminated calcareous sediments, containing freshwater fauna and flora, (facies i) close the outcropping succession; these sediments formed in a small lacustrine basin isolated from clastic inputs. A different wave activity on the lake shores probably controlled the sedimentation of the two facies subassociations.

Where morphological barriers or wetland-vegetation hindered the wave action, facies C1 subassociation could be deposited, whereas facies C2 subassociation was deposited where shoreline was directly subjected to wave motion.

Palynological aspects - A palynostratigraphic sequence representing the upper 350 cm of the sediments surrounding one of the fossil trunks (trunk no. 49) was studied. The trunk is still in growing position and leans — like all the other trunks — some 10° towards the NE According to Biondi & Brugiapaglia (1991), all the trunks so far examined appear to belong to one species only, identified as Taxodioxylon gypsaceum (Goppert) Krausel, which became extinct during the Pliocene and had anatomical features similar to the present-day Sequoia sempervirens (Lamb.) Endl. The pollen analyses show a qualitatively rich flora with a good proportion of tertiary species that are no longer present in indigenous Italian flora, with a predominance of trees and shrubs (AP) with respect to herbaceous plants (NAP). Forest species most frequently represented in the pollen diagram are: Sequoia-type, Taxodium-type and Pinus subgenus Haploxylon, followed, in decreasing order of frequency, by: Alnus, Larix, Abies, Picea, Pinus subgenus Diploxylon, Zelkova, Tsuga, Cycadaceae (Stengeria-type), Carya, Sciadopitys, Betula, Salix, Quercus, and Castanea; there are also sporadic findings of pollen from Carpinus, Cedrus, Celtis, Ephedra, Eucommia, Fagus, Ilex, Juniperus, Ligustrum, Maclura, Nyssa, Ostrya, Pterocarya, Sambucus, and Tilia. We believe that the pollen classified as Sequoia-type should be identified as Taxodioxylon gypsaceum. Although the studies so far carried out suggest that the Dunarobba Forest would consist of a single species, on the contrary pollen analyses show that the forest in that area was mixed. The arboreal taxa found in pollen analyses can be divided into three groups, according to the time of their disappearance from Italy. The 1st group comprises the tertiary species which disappeared in the late Pliocene and/or Early Pleistocene times, i. e., Sequoia-type, Taxodium-type, Sciadopitys, Strangeria-type, Eucommia, Nyssa, Celtis; and, among the NAP,

Tillandsia-type; the 2nd group includes Cedrus, Tsuga, Pinus subgenus Haploxylon, Carya, and Pterocarya which became extinct during the Pleistocene; the 3rd group includes all the other species still growing in Italy today. The fact that Zeikova, Castanea, and Juglans have to be regarded as indigenous species, is discussed. On the basis of the pollen diagram, the time climate had to be warmer and more humid than at present, if the in situ occurrence of plants like the ancient Sequoia and the abundance of spores of Pteridophyta together with Tillandsia-type pollens, an epiphyte living in subtropical environments, are considered. Edaphic conditions in which Sequoia grew were those of an environment subjected to continual alluvial phases, as confirmed by the costant and abundant occurrence of Cyperaceae. Finally, the finding of pollen from Alnus and Salix. together with Potamogetonaceae and Alismataceae, is evidence of slow-flowing waters. Moreover, the constant finding of Taxodium-type pollen is further evidence for a coastal marshland close to the Dunarobba area. The Taxodioxylon forest had a luxuriant undergrowth, with abundance of Pteridophyta, mainly Lycopodium-type which is replaced by Osmunda-type in the uppermost layers. This change is probably related to the occurrence of different edaphic conditions. Finally, the Sciadopitys curve is discussed, in order to outline possible climatic conclusions from it. The dendrochronological research carried out on Taxodioxylon gypsaceum is a further support to palynological results. On the basis of the pollen results the sediments of the Dunarobba Fossil Forest showing an evident Taxodiaceae fades are attributed to the Pliocene. A more precise attribution is at this time impossible, because the sediment thickness so far studied is very limited.

Dendrochronological aspects - A dendrochronological analysis was carried out on a trunk of Taxodioxylon gypsaceum (Gippert) Krausel a homoxil wood sample put at disposal by the Umbria Region Archaeological Suprintendence. The wood was in a good state of conservation owing to the impermeability of the embedding clay. Ring widths were measured with Dendroware Aniol; mean value standard deviation, autocorrelation coefficient and mean sensitivity were calculated. Abrupt growth changes were also calculated and a spectral analysis was performed to obtain indications on past climate lengths of period. An image analysis system was used to investigate anatomical features. Preliminary tracheidograms were elaborated, and a floating curve of 565 years was built. The first period of the tree life is characterized by narrow rings, then by large ring widths for about 60 years. This pattern may be explained by the social status of the tree within the forest. Subsequently, a light biological trend is shown. In regard to the dendrochronological statistical parameters the mean sensitivity value was calculated 100 yr intervals: the value is in fact high for the total period of 565 rings (0.332) but it is not constant over the whole period. The recorded increase might indicate a worsening in the external ecological conditions; however it is difficult to date environmental modifications in a particular tree life period. As a matter of fact if abrupt growth changes are take into account, we see that these are slight, recorded for short periods and present over the whole length of the curve. Thus we think that any environmental modification must have occurred step by step. The autocorrelation coefficient values support the observations made on mean sensitivity. Important results arise also from the analysis of the lengths of period. These were calculated for the total 565 years of the curve and at 100 year intervals such as the dendrochronological statistical parameters. Second

and 3rd order frequences are the most common. The distribution in the real curve is at the right of the theoretical distribution indicating a climate tending to the oceanic one. From the interval analysis a decrease of the 3rd order frequencies versus an increase in the 2nd order ones, can be seen. In particular, uniannual variations reach the 62% frequency in the last 65 years suggesting a notable degree of environmental variability and a trend towards a continental climate. Beside frequency other high and medium fluctuations seem to be distinguishable. Spectral analysis allows to isolate 5-6 and 11 year fluctuations whereas 22 year fluctuation does not appear. The anatomical analysis provides information on short period. Tracheidograms indicate a species having a regular growth after an initial pause such as occurs in present conifers at high altitudes. The number of cells of latewood is 1/5 -1/3 of the total number of ring cells Three variables were taken into account: tracheid lumen diameter; total tracheid diameter; and wall tickness. Tracheidograms show a remarkable variability in cell diameter from the 10th to the 15th element. This suggests a greater sensitivity of the species to the climatic conditions during the early phases of cambial activity. The tracheid lumen diameter shows a greater variability from the 20th to the 24th element with a maximum in the 23rd element. The cell wall thickness shows the greatest variability in the region of the 19th and 23rd element. This may be due to the different typology of annual rings, in particular to the more or less abrupt trasition from large lumen and thin walled cells to small lumen and thick walled cells. Finally, an abrupt decrease in cell dimension just before latewood is recorded. Perhaps this is to be attributed to scarce raining. Although if the study was carried out on one sample only some preliminary conclusions can be drawn: - xylematic typology seems to indicate a seasonal fluctuation; - ring boundaries are 'in fact visible and both earlywood and latewood are well distinguishable; - in the case of intra-annual variations, different modes of transition from earlywood to latewood among rings are present and in most of them, an abrupt transition from earlywood to latewood is recorded evidencing a sudden change of an external factor (temperature?) which led to a vegetative stasis. In rings where the transition from earlywood to latewood is gradual (slight decrease in the cell lumen and slight increase in cell wall thickness) the impact of seasonal changes on tree life is probably less abrupt. Trecheidograms typology well illustrates seasonal differences. Although the floating curve cannot be dated, it can be assumed that the climate controlling the Dunarobba Fossil Forest tended to the oceanic type even if modifications toward a more continental climate may be hypothesised.

Malacological aspects - The DFF deposits contain a rich continental malacofauna which can be distinguished into two different assemblages: one - dominated by terrestrial pulmonates - is found in the clays encrusting the trunks (lithofacies c of hydromorphic paleosols) and indicates a forest environment with a swampy to very hygrophilous substratum; the occurence of molluscs typical of slowly moving water in sediments near the trunks marks the passage from lithofacies c to lithofacies d. The other assemblage - dominated by aquatic prosobranchiopods - characterizes the sediments of lithofacies b, a, g and i and is indicative of a submerged environment with moving-slowly moving or stagnant water. Both assemblages suggest climatic conditions warmer than the present ones. The assemblage with prevalent terrestrial pulmonates is chronostratigraphically attributed to Pliocene, because of its close resemblance to the molluscan faunas from the Pliocene

deposits of Piedmont (Triversa F.U.), France and Germany. The second assemblage neither proves nor excludes a Pliocene age. It comprises an endemic group of species recognized in other two Umbrian formations: Fosso Bianco (FBf) and Ponte Naja formation (PNf) at the Toppetti quarry near Todi, both attributed to Pliocene (FBf is Middle Pliocene in age). It can thus be concluded that the pulmonates assemblage points to a Pliocene age of deposits, while the prosobranchiopods assemblage spreads from Middle Pliocene to Early Pleistocene.

In conclusion, stratigraphic, palynological, and paleomalacologic data indicate that the Dunarobba Fossil Forest can be attributed to a time span in the Pliocene with climatic conditions which were warmer and more oceanic than the present ones. The dendrochronological study of a Taxodioxylon gypsaceum sample indicates a change towards worse and more continental climatic conditions; this observation is confirmed by palynological evidence, in particular by a decrease of the Sciadopithys pollen frequence.

4.1.2 Biochronology of Plio-Pleistocene mammal assemblages of Umbria (Central Italy) (Argenti P. * & Kotsakis T. **)

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Several Plio-Pleistocene mammal assemblages of Umbria have been collected in the "classic" deposits of the Tiberino Basin. Other fossil sites are located in the Tavernelle/Pietrafitta Basin, in the Valle Umbra deposits and in many karst fissures (ARGENTI, 2003; with bibliography).

Present day knowledge about terrestrial mammal faunas of the Plio-Pleistocene of Umbria does not allow attempting a detailed biochronology for the whole region. The chronological distribution of the mammalian fossiliferous layers is clear in the central and southern part of Umbria, because of the abundance of the fossils. In the northern part of the region there are instead many gaps due to the scarcity of fossiliferous beds. The fossil mammal remains found in Umbria are not uniformly distributed within the Plio-Pleistocene deposits, they are in fact concentrated in Late Pliocene and Early Pleistocene ones. Not all analysed localities are referred to a F.U. because in many sites only isolated fossils were discovered or the fossils come from ancient collections, which are impossible to review; the localities are referred to Mammal Ages (GLIOZZI et al., 1997) . Some mammal assemblages are kept in cave deposits.

The only mammal assemblage found in Umbria referred to Early Villafranchian (Middle Pliocene), was collected in the lignite mine of Santa Croce and Morgnano (Spoleto). This can be tentatively referred to Triversa F.U. for the presence of *Mammut borsoni*, *Anancus arvernensis* and *Tapirus arvernensis*. This assemblage is characteristic of the Early and Middle Pliocene (Ruscinian and Early Villafranchian). The beginning of Triversa F.U. should be around 3.2 Ma as suggested by paleomagnetic analysis at Fornace R.D.B. (Villafranca d'Asti, Piedmont). Some *Anancus arvernensis* remains were found near Montoro (Narni) and referred to Pliocene age.

Middle Villafranchian (Late Pliocene) mammal assemblages in Umbria are very rare. The assemblage of Cava Toppetti (Todi) is referred to Costa S.

Giacomo F.U. The evolutionary degree of the murid *Apodemus dominans* of Cava Toppetti is identical to the Rivoli Veronese (Veneto) population.

Most fossil mammal assemblages of Umbria belong to Late Villafranchian (Latest Pliocene and Early Pleistocene). The assemblage collected in Pantalla (Todi), characterised by several carnivore remains (*Panthera gombaszoegenesis*, *Lynx* cf. *L. issiodorensis*, *Canis etruscus*), is assigned to Olivola F.U. Torre Picchio (Montecastrilli) mammal assemblage is referred to Olivola F.U. (or between Olivola F.U. and Tasso F.U.) on the basis of stratigraphical correlations with marine strata, despite the presence of *Mimomys medasensis*, known from Late Pliocene of Ibero-occitanic palaeobioprovince. To Tasso F.U. are referred some assemblages such as those found at Villa Spinola, Ponte S. Giovanni and Piscille (Perugia), which come from the same levels. Tasso F.U. is characterised by several changes in the composition of the faunal assemblages with the appearance of typical Quaternary elements such as *Mimomys savini*, *Lycaon falconeri* (cfr. MARTÍNEZ-NAVARRO & ROOK, 2003), *Hippopotamus*, *Praeovibos* and *Leptobos vallisarni*. The fossils from S. Faustino (Massa Martana), and Vigna Nuova (Panicale) belong to Tasso F.U. too. Some mammal remains from Fighille clays (Citerna) are referred to the Late Villafranchian (Tasso F.U.) on the basis of mollusc assemblage (*Gastrocopta* (*Vertigopsis*) *dehmi*, *Multidentula helenae* and *Parmacella* (*Parmacella*) cf. *P.* (*P.*) sp.1. – cfr. CIANGHEROTTI & ESU, 2000). The Farneta F.U. is characterised by a faunal renewal: the cervids of *Megaceroides* clade (*M. obscurus*) and the voles *Allophaiomys chalinei* and *A.* cf. *A. ruffoi* appear. The mammal assemblages from Pietrafitta (Panicale) and Ellera (Perugia) are referred to Farneta F.U. for the presence of *Megaceroides obscurus*, the voles of the genus *Allophaiomys* and *Mimomys pusillus*. To the Late Villafranchian are also assigned many fossils collected from several localities between Deruta and Todi. The rhino skeleton of Capitone (Narni) (classified as an advanced form of *Stephanorhinus etruscus*) is assigned also to the Late Villafranchian. The Parrano-Frattaguida fauna (Parrano) is quite heterogeneous and its biochronological position is unknown; it could be referred to the mid- Late Villafranchian: in this site the sabre-toothed cat *Megantereon cultridens* and *Stephanorhinus etruscus* are present. The mammal remains from Fontignano (Panicale) are referred to the upper part of Late Villafranchian, for the presence of an advanced form of *Hippopotamus antiquus*.

The Galerian (latest Early Pleistocene and early Middle Pleistocene) mammal assemblages of Umbria are relatively scanty. The very rich small mammal assemblages found in the M. Peglia cave (San Venanzo), characterised by the presence of *Allophaiomys nutiensis* and *Allophaiomys burgondiae*, have been ascribed to Colle Curti F.U. The mammal remains (lower assemblage) found in the sands of Promano (Città di Castello) are also referred to the earliest part of the Early Galerian, probably to Colle Curti F.U., mainly for the presence of *Megaceroides verticornis* and of an advanced form of *Mammuthus meridionalis*. To Early Galerian must be probably assigned *Elephas (Palaeoloxodon) antiquus* from Fighille (Citerna), collected in sands overlaying the Fighille clays Another assemblage found in Promano sands (intermediate assemblage), has been referred to Middle Galerian, for the presence of *Stephanorhinus hemitoechus*. The fossil remains found near Marsciano are referred to the early Galerian for the presence of a cervid classified as *Megaceroides* cf. *M. verticornis*.

A marked faunal renewal occurs at the beginning of the Aurelian, with the extinction of some Galerian forms such as the archaic megacerine cervids, the archaic elaphine deer and the arrival of *Canis lupus* and *Ursus spelaeus*. During late Aurelian, *Capra ibex*, *Coelodonta antiquitatis*, *Mammuthus primigenius*, *Marmota marmota* spread over Italy and several taxa of large herbivores and carnivores progressively disappeared in correspondence with the successive phases of climatic deterioration. Aurelian assemblages (late Middle Pleistocene and Late Pleistocene) are not so common in Umbria. The mammal assemblage from M. Cucco cave (Costacciaro), characterised by the presence of *Coelodonta antiquitatis* and the abundance of *Capra ibex*, is referred to Late Aurelian. To the same period have been assigned some faunal assemblages collected in some caves whilst a few assemblages (e.g.: Triponzo) belong to Holocene.

Time scale	Geochronology		Mammal ages		Faunal Units	Umbria localities
0.0	**Holocene**					Triponzo
0.1	**Pleistocene**	Late	Aurelian	Late		M. Cucco
0.2				Early	Vitinia F.U.	
0.3					Torre in Pietra F.U.	
0.4			↕			
0.5				Late	Fontana Ranuccio F.U.	
0.6		Middle			Isernia F.U.	
0.7				Middle	Ponte Galeria F.U.	
0.8			Galerian			
0.9					Slivia F.U.	
1.0				Early		
1.1					Colle Curti F.U.	M. Peglia, Promano
1.2			↕			
1.3					Pirro F.U.	
1.4						
1.5		Early			Farneta F.U.	Ellera, Pietrafitta.
1.6				Late		
1.7						Colle S. Andrea, S. Faustino,
1.8					Tasso F.U.	Vigna Nuova, Villa Spinola
1.9	**Pliocene**				Olivola F.U	Pantalla, ?Torre Picchio
2.0						
2.1					Costa S. Giacomo F.U.	Cava Toppetti
2.2		Late	Villafranchian			
2.3				Middle	St. Vallier	
2.4					Montopoli F.U.	
2.5						
2.6						
2.7						
2.8						
2.9						
3.0						
3.1						
3.2		Middle		Early	Triversa F.U.	S. Croce di Spoleto
3.3						

4.2 The East-Tiber Basin (ETB) (Coltorti M. & Pieruccini P.)

The ETB is located between the Martani Ridge to the west and the Umbro-Marchean Apennine Ridge to the east (FIG. 4) (COLTORTI & PIERUCCINI, 1997a). Both ridges are characterised by the presence of well preserved remnants of the Planation Surface (PS) on top. In the Martani ridge the PS elevation decreases to the north where it affects also the marly-arenaceous rocks (Marnoso Arenacea Fm). Moreover, the PS dips to the east where it is buried under the Pliocene continental sequence.

The Pliocene sediments are found mostly to the west of the present day valley bottom, from the Spoleto area to the south up to Bevagna to the north over an extension of about 35 km. The sequence (FIG.10) is similar to the WTB, with a lower portion made of fine-grained floodplain and lacustrine sediments with lignites (Bastardo-Bevagna Unit, late Early-Middle Pliocene) and an upper part made of gravelly-sandy alluvial fan deposits (Montefalco Unit, Late Pliocene-Early Pleistocene?). Details on the facies and stratigraphy will be given later.

The Pliocene sequence shows onlap geometries over the pre-Pliocene bedrock and no normal faults are present along the western and the eastern sides. The accommodation space was created in a synform basin bounded to the east by the Martani Ridge thrust whereas the eastern borders of the Pliocene Basin are uncertain since they have been cut by Pleistocene high angle normal faults that generated an asimmetric graben with a master fault located in the eastern side (FIG. 7). However, close to Spoleto (Collicelli), onlap relationships of the Pliocene sediments over pre-Pliocene rocks has been observed. In the Spoleto area, lignites were exploited in the last two centuries, and mining reports indicate the base of the Pliocene sediments below the present-day sea-level indicating a strong deformation as also revealed by the sediments on the western side where they dip up to 30° to the east.

In the southern sector the Bastardo-Bevagna Unit is poorly preserved since it is unconformably overlied by a sequence up to 200 m thick of alluvial fan gravels and sands (Montefalco Unit) forming a bajada extending for about 15 km with provenance from the Martani Ridge (FIG.11). In the northern sector the Bastardo-Bevagna Unit is better represented and is unconformably overlaid by the alluvial fan gravels and sands of the Montefalco Unit, with provenance from ENE. The presence of thick unconformable gravelly-sandy alluvial fan deposits indicates that a higher relief was already present to the east of the study area.

After the deposition of these sediments the whole sequence has been affected by the activation of high-angle normal faults, the Pliocene sequence was dissected and the new accommodation space for the sedimentation is inside the graben of the Valle Umbra. In the case of Montefalco the alluvial sediments are uplifted up to 200 m above the present day valley bottom.

From the Middle Pleistocene the sedimentation is mainly driven by the interactions between climatic changes and vertical uplift. Inside the valleys there are three main terraced units (COLTORTI & PIERUCCINI, 1997a), with few exceptions (DECANDIA, 1982; CENCETTI, 1988; GREGORI, 1988; CATTUTO et al., 1992;) consistent with the present-day drainage network. The older Unit is affected by normal faults with displacements of a few tens of

metres whereas the late Middle Pleistocene and Late Pleistocene-Holocene units do not show important offsets.

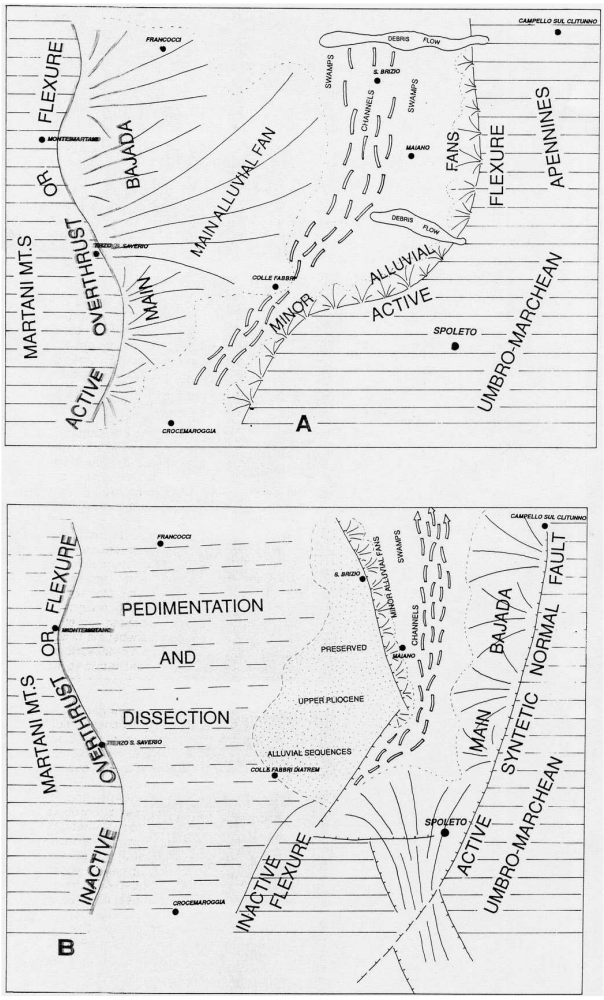

Figure 11 – Late Pliocene-Quaternary evolution of the East Tiber Basin in the Spoleto sector. A- Late Pliocene (?). A bajada of alluvial fans originated from the west was deposited in a downwarping basin. B – (Early Pleistocene?-Middle Pleistocene) Extensional faults displaced the Pliocene sequence and a graben basin was created. (from: Coltorti & Pieruccini, 1997a)

4.2.1 STOP2
The Bastardo-Ponte di Ferro area

The curious name of this village (Bastardo) is due to the fact that until the XIX century there was only a Travel Inn along the Flaminia road (one of the oldest roman roads) that crossed the Martani Ridge from Massa Martana to Foligno. At that time the beginning of mining works for the exploitation of the lignite led to the growth of a mining village around the main mines and the related furnaces. The people working in the mines came from different place around Italy and for that reason the place assumed the name of Bastardo, due to the lack of a precise, local origin. The name persisted also after the end of the exploitation and the inhabitants of Bastardo are called “Bastardesi” (something like Bastardians). They are very sensitive to this and do not like very much jokes on the name of their village.

In 1996, new quarrying works for lignite exploitation opened a large section (FIG. 12) with a spectacular view of the uppermost part of the late Early-Middle Pliocene ETB sequence (Bastardo-Bevagna Unit) and of the unconformable Late Pliocene-Early Pleistocene Montefalco Unit (FIG. 10).

The sequence is located in the westernmost sector of the ETB, next to the ridge that separates the present-day outcrop of the ETB from the WTB sediments. The bedrock of the ridge is the Marnoso arenacea Fm. (FIG. 7) close to the overthrust of the Martani Ridge. The top of the ridge is characterised by the presence of the PS and further to the west, close to the watershed, small patches of Pliocene sediments are still preserved.

The late Early-Middle Pliocene sequence (Bastardo-Bevagna Unit) is slightly deformed with a dip of about 5° to the E and shows clear onlap relationships with the bedrock.

The base of the log corresponds to the top of a lignite layer about 2 m thick made of compacted tree trunks, wood fragments, vegetal tissues, seeds and leaves.

The ca. 10 metres of sediments on top of the lignite layer are made of laterally continous and planar layers with a thickness ranging from 0,3 to 3 m of mostly clayey sediments. Two main lithofacies are recogniseable:

a) grayish-bluish massive clays (Fm) with concoidal fractures, rich in coal and vegetal floated fragments and molluscs, interbedded with very rare and very thin (1-2 mm) sandy silty laminae, sometimes ripple-laminated; locally, bioturbation and oxidised hard grounds are present.

b) dark gray laminites (Fl), made of organic-rich clays and ripple-laminated fine sands and silts, with many coal fragments; also in this case the thickness of the laminae ranges from 1 to 2 mm.

The upper part of the clayey sequence is abruptly truncated by an unconformity made of a channel belt filled with sands and subordinated gravels (Montefalco Unit).

The channels geometry changes according to their position. In fact the older channels in the central part of the picture, are cut directly on the cohesive clayey sequence and are deeper, narrower and have steeper sides (5th order bounding surfaces MIALL, 1995). The channel filling is made of fine well rounded pebbly lags and trough crossbedded poorly sorted sands rich in muddy matrix (St). The younger channels (4th order bounding surfaces) are wider and have much less steep sides. Their filling is predominantly made of mainly planar crosbedded coarse to very coarse sands with granules and pebbles and fine to medium well rounded gravels (Sp, Gp). The channels are elongated in ca. E-W direction whereas the paleo-flows measured on pebble embrications, long axies and sets indicates a provenance from E-NE.

During the deposition of the Bastardo-Bevagna Unit the environment was characterised by a low energy flood plain, with lakes and swamps where the lignite accumulated. The presence of hardgrounds, bioturbations and the lignite suggests the lakes and swamps were shallow. In this sector, further to the west, proximal facies made of gravels and laminated sands possibly deposited at the lakes margins have been also observed. However, in other sectors of the Basin (see stop 4) the same sequence is also affected by in-channel deposition and lateral accretion deposits. Moreover, the presence of small remnants of similar deposits close to the top of the ridge separating the ETB from the WTB suggests the late Early-Middle Pliocene Basin extended across the present-day ridge, characterised by a very wide low energy fluvial environment with a muddy floodplain and a shifting channel system. The warm climatic conditions matched with the very low relief energy enhanced the deposition of long fine-grained sequences.

An abrupt change of the geomorphic conditions is clearly indicated by the depositon of the Montefalco Unit with the onset of confined channel belts. The older channels indicate single-channels fluvial systems with rapid sedimentation and a mainly vertical accretion. Later, in this section, the lateral accretion features are predominant and the architectural elements (LA, SB) and facies associations are typical of sandy bed meandering rivers, possibly the outermost part of the alluvial fan complex of Montefalco (see stop 3).

The unconformity is a 5th order bounding surface related to climatic events (the onset of cooler conditions) as well as to tectonic events (the ongoing uplift of the Apennines and the increased relief energy) that together led to the deposition of coarser grained sediments with well defined flow directions.

4.2.2 STOP3
The Montefalco area

The historical town of Montefalco is built on top of a hill at 472 m a.s.l., about 270 m above the present-day valley bottom. The relief is made of a mainly gravelly-sandy sequence (Montefalco Unit) ca. 150 m thick, unconformably overlying the Bastardo-Bevagna Unit. The sediments are tilted up to 15° to the WSW and are characterised by the vertical stacking of gravelly, sandy and silty bodies with frequent lateral changes due to channel geometries. Processes of selective erosion created characteristic stepped slopes that help in the recognition of the lateral continuity of the gravelly bodies, since their exposure is limited due to vegetation and human induced modifications.

The town and the city walls are founded on the uppermost body of conglomerates made of subrounded to rounded poorly sorted gravels with a sandy matrix. Locally the gravels are massive, very poorly sorted, angular to subangular and matrix supported as result of highly concentrated flows (debris flows).

We have two stops here where we can observe the typical facies associations of the Montefalco Unit.

The main features are represented by channels with different geometries, from deep and steep to wide and shallow with channel floor scours (FIG.13). They are typically cut (FIG. 14) in Overbank fines sediments (FF) made of sands, silts and clays, massive to laminated (Fm, Fl, Sh, Sr), with molluscs and occasional paleosoils (paleoentisoils to paleoinceptisoils) or in Sandy bedforms sediments (SB) made of trough to planar crossbedded sands (St, Sp). Usually the thickness of the FF and SB deposits is limited but, depending upon the position of the main channels, it can reach 10 m. The channels are infilled by gravelly bodies several metres thick forming Gravel bars and bedforms (GB). The basal scours of the channels show typical lags made of coarse-grained gravels, often embricated, and the channel filling is dominated by trough cross bedded gravels (Gt), subangular to rounded with a sandy matrix, associated to sudden and rapid sedimentation at the first stages of the channel formation. Upward the number and the dimensions of the gravel bars increases and the main bedforms are planar crossbedded gravels (Gp) indicating the diverson of the flows as response to avulsion processes. Small sandy lenses (Sh, St) and thin muddy levels are also present.

In other small outcrops, lobes of Sediment Gravity flows (SG) made of unsorted, angular to subangular matrix supported gravels (Gmm) have been observed.

The overall characterestics of these deposits are consistent with deposition in a deep, gravel bed river (Model 3 of MIALL, 1996) in the inner/mid part of an alluvial fan with several channel-fill, bar progradation, flood events and abandonement paleosoil formation cycles. The paleoflow directions observed on the embricated clasts as well as on the sets indicate a provenance from ENE.

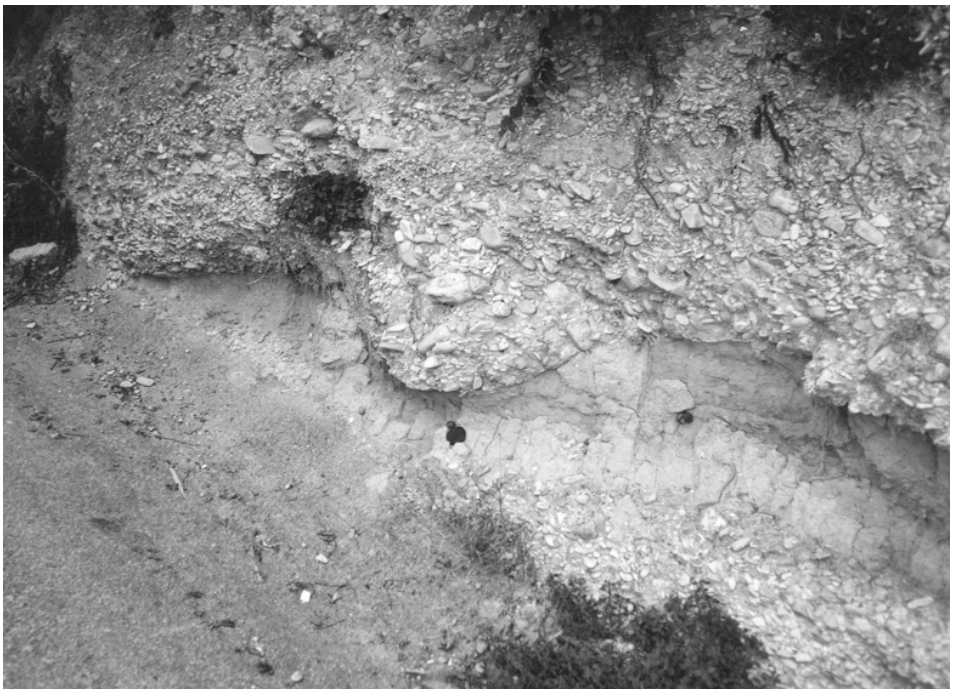

Figure 13- Channel floor scours with coarse-grained gravelly lags.

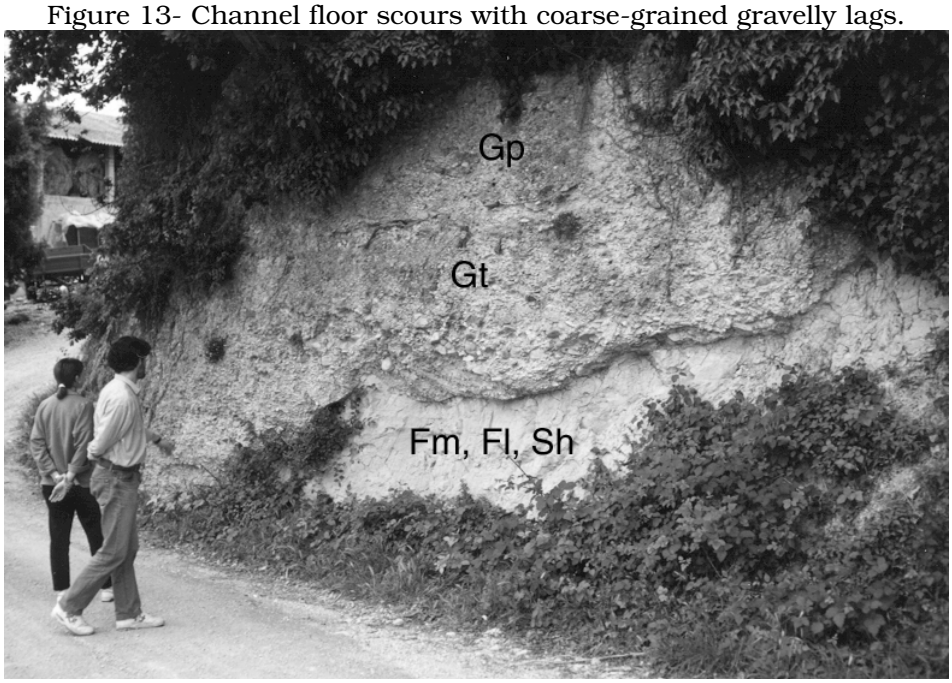

Figure 14 - Gravel bedforms filling a channel cut in the Overbank fines sediments.

4.2.3 STOP 4

The Bevagna area (with the contribution of Roberto Bizzarri, University of Perugia)

About 40 m of the sequence (FIG.15) belonging to the basal part of the ETB sequence has been observed close to the historical town of Bevagna, in a quarry exploited for sands and clays for a brick factory (Fornaci Briziarelli).

The vertical and lateral continuity of the sequence is disturbed by the presence of many low- and high-angle normal faults with offsets of a few metres, staircase geometry and main dip slip movements although horizontal components are common. Inverse faults are less represented. Due to the variability of the fault directions and their geometries it is possible that they are also related to local gravitational features.

Three main facies association can be recognized.

Facies Association	Lithofacies	Architectural elements	Interpretation
A	Sp Sr Sl St Sh (Fm)	LA SB CS FF CH	Wandering to meandering sandy bed river.
B	Sh Sr Sl (Fm)	CS CR CH	Crevasse channels and splays
C	Fm Fl Fr (Sr Sh)	FF	Distal floodplain, small lakes, swamps.

The Facies association A is characterised by the presence of wide channels filled with pebbly lags and laterally accreting planar crossbedded fine to coarse sands (FIG.16). Frequently muddy laminae drape the surface of the sets. Upwards the channel fillings are characterised by ripple-laminated sands interbedded with thin muddy and clayey massive to finely laminated beds. Locally, convolute bedding and water escape structures are common. This Facies Association occurs at the base of the log and indicates a river system dominated by lateral and vertical accretion (wandering to meandering?), migrating in a low-energy floodplain where after the channel avulsion muddy and clayey deposition was predominant.

The Facies association B is dominated by ribbon sand bodies and slightly erosional channels filled with thin beds of fine to medium sands interbedded with clays and silts. The sandy beds often show erosive bases. Dune and ripple laminations are common as well as convolute bedding due to load or water-escape. Thin organic-rich laminae, molluscs and root tracks also occur. This facies association has been interpreted as typical of crevasse channels and splays entering the distal part of a floodplain characterised by muddy deposition in small lakes and swamps. The main channel system was diverted in other sectors of the alluvial plain.

The Facies Association C is made of greyish to yellowish clays and silts massive to laminated (FIG.17). Whitish and blackish laminae (1-2 mm thick) typically alternating whereas vegetal tissues (leaves, seeds etc..), molluscs and root tracks are common. The silty-clayey thicker laminae often show a ripple lamination, both symmetric or climbing as well as occasional sandy laminae.

The deposition took place within a floodplain characterised by occasionally emerging shallow lakes and swamps.

The log shows an overall fining-upward trend, ending close to the top with the appearance of the first lignite beds. The observed lignite bed is about 1 m thick and is made of strongly compressed tree trunks, wood fragments and vegetal tissues with a clayey matrix and many molluscs. The fining upward trend is due to the avulsion in other sector of the alluvial plain of the main channel system, present at the base of the log (Facies association A). After the avulsion the channel evolution is mainly crevassing with floodplain muds and lignites deposited inside small ponds and lakes. Vertical accretion is therefore dominant. Here the deformation of the layers (up to 25° to the NW) is greater than the dip observed in the corresponding log of Stop 2 due to the combined effect of folding and faulting along the main high-angle normal faults that bound the graben of the ETB (FIG.7).

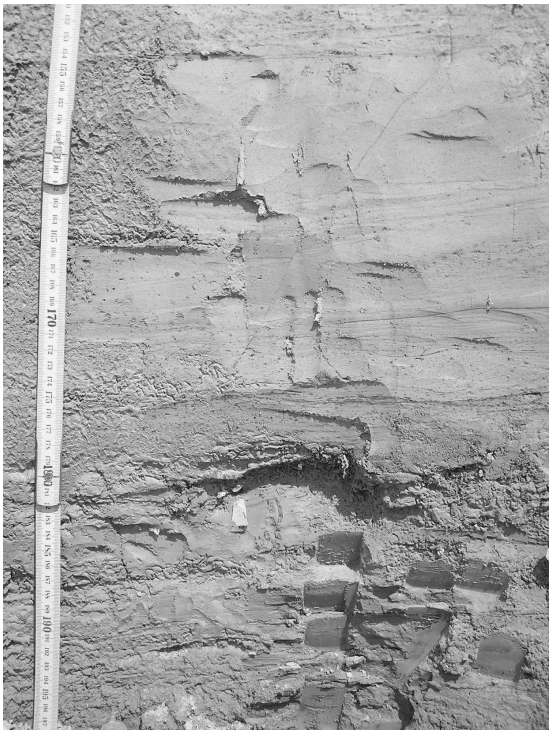

Figure 16 – Sandy laminated bedforms of the Facies Association A

Figure 17 – Organic-rich clayey laminae (Facie Association C)

4.3 Major steps in the evolution of the Tiber Basin (FIG 18)

LATE EARLY-MIDDLE PLIOCENE (Bastardo-Bevagna Unit)

After the PS modelling, (COLTORTI & PIERUCCINI, 1999) compressional tectonic events, that have been associated to the activity of low-angle east-dipping normal faults (CALAMITA et al., 1999) or to a compressional regime (BONINI 1998; BOCCALETTI et al., 1999), created a wide synform Basin, that is the accomodation space for the onset of continental deposition. The low energy landscape and the warm climatic conditions enhanced the deposition of moslty fine-grained sediments inside an alluvial system characterised by a wide floodplain with small lakes, swamps and few migrating channels. In this phase it is difficult to establish the paleoflow directions and Coltorti (in prep.) suggested a western provenance of the watercourses although AMBROSETTI et al. (1987) and BARBERI et al., (1993) based on the presence of co-heval marine deposits to the west of the Peglia-Narni Ridge (Argille di Fabro Fm. and Sabbie a Flabellipecten Fm.,) suggested a drainage parallel to the main structures (NNW-SSE). We suggest that in this phase the ETB and WTB formed a unique Basin, only later divided into two branches by the uplift and the ongoing compressional tectonics. In terms of Sequence stratigraphy (VAIL et al., 1977; MIALL, 1990) the accommodation space for the deposition of the late Early-Middle Pliocene sequence corresponds to a third-order cycle (MIALL, 1995) related to regional uplift and compressional tectonic features.

LATE PLIOCENE-EARLY PLEISTOCENE (?) (Montefalco Unit)

During this phase the onset of cooler conditions (OIS 100, RUDDIMAN & RAYMO, 1988) and the ongoing uplift of the Apennines favoured the deposition of coarser-grained sediments in alluvial fan bajadas indicating the presence of well defined slopes. In the northern sector of the ETB the alluvial fans have flow directions from the ENE (Umbro-Marchean Apennines Ridge) whereas in the southern sector (Spoleto) they originated from the west, that is the Martani Ridge. The Late Pliocene-Early Pleistocene sequence, mainly related to climatic changes, corresponds to a fourth or fifth order cycle

EARLY PLEISTOCENE

In the WTB the fluvial sedimentation is sandy-gravelly dominated whereas marine conditions still exist to the west of the Peglia Ridge. However, high-angle transtensional/normal faults (Martana fault) start their activity, revealed by strong deformations and local lateral facies variations of the Early Pleistocene sediments close to the fault escarpment.

MIDDLE PLEISTOCENE – UP TO THE PRESENT-DAY

The Tiber river is diverted and the southernmost sector of the WTB is no longer affected by fluvial sedimentation except for small stream systems close to the eastern slope. Along the fault escarpment, widespread travertine deposition took place related to climatic conditions and to the presence of hot springs. In the ETB the entire Pliocene sequence is affected by normal faulting and the present-day Valle Umbra graben was created. The deposition took place inside the graben by means of alluvial fan and fluvial deposits forming three main orders of alluvial terraces. The normal faults do not severely displace the late Middle Pleistocene-Holocene deposits although a low seismicity still affect the area (BONCIO et al., 1998). The Middle Pleistocene-Holocene deposits of the graben basin (ETB) and those associated to the normal fault escarpment (WTB) are fourth or fifth order cycles. In fact they are related to local phenomena without a regional importance. Their deposition is

tectonically driven in terms of accomodation space in a continously uplifting mountain chain and by climatic changes in terms of sedimentation and pedogenesis (Glacial-Interglacial cycles) and facies distribution (COLTORTI & PIERUCCINI, 2004).

5. The Gualdotadino Basin: fluvial evidence for a catastrophic event during the Late Glacial (Coltorti M.)

The Gualdo Tadino - Fossato di Vico Basin, is a tectonic depression located between Nocera and Sigillo. It is located at the transition between the "Umbro-Marchean Ridge" to the east, and the "Umbrian Pede-Appenine" to the west (CENTAMORE et al., 1978; 1979; BALLY et al., 1986; DEIANA & PIALLI, 1994, FIG. 2A). The "Umbro-Marchigiana Ridge" is made up of the classical limestone units, from the Jurassic to the Oligocene in age; The "Umbrian PedeAppenine" is made of the terrigenous sediments of the Marly-Arenaceous Formation, from the Oligocene to the Miocene in age, representing a pelitic-arenaceous association deposited on a raised basin margin ("basin plain edge", CENTAMORE et al., 1979). A series of overthrusts have affected the limestone and the terrigenous units. However, nowadays, an extensional fault marks the limit between the two sectors; it is a continuation of the important fault located further south that delineates the Umbra Valley (Spoleto-Foligno-Assisi, COLTORTI & PIERUCCINI, 1997a).

Some previous publications have underlined the peculiarity of the Quaternary deposits in the area. Detailed information is provided by the 1:50.000 map of "Fabriano" (CENTAMORE et al., 1979) and by BOSI et al. (1987). The first map highlighted the presence of alluvial deposits in the valley bottom and recent debris deposits on local slope. At higher elevation, two distinct alluvial terraced units, raised 3-10 and 10-20 m respectively from the fluvial plain, have been pointed out. Further to the south, there are also a fluvio-lacustrine formation, raised ca. 40 m above the valley bottom. BOSI et al. (1987) recognised various terraced units and, in addition, identified and mapped two detrital formations, the most recent named S. Facondino-Casella and the oldest S. Pellegrino - L'Ospedale. These formations crop out on both the eastern and western sides of the basin, and are composed of calcareous deposits coming from the Umbro-Marchean ridge. In FIG. 19, modified after BOSI et al., (1987), the terraced fluvial deposits of the Middle Pleistocene have been mapped together whilst the Late Pleistocene deposits have been subdivided in alluvial and debris deposits. The alluvial deposits, cropping out on the western side of the depression, have been considered slope deposits in the previous works; they are found at high elevation (up to 495 m) and are strongly deformed over a distance of ca 10 km.

In order to better understand the origin, chronology and deformative processes of these deposits, a study was carried out in an area delimited to the north by an EW line passing north of Fossato di Vico (up to the River Chiascio). To the south, the area, is delimited by a line passing south of Gualdotadino (up to the Rasina stream). These streams and their tributary cut widespread continental deposits of various grainsize, creating a series of terraces at elevations ranging from 590-360 m s.l.m. Between Col Bassano and Cerqueto, on the western side, strongly the deformed alluvial deposits crop out but, in the central part of the valley, no alluvial terraced units are present. In the northern and southern sectors alluvial terraces are preserved in places. In fact, it is possible to distinguish major differences also between the deposits of the latter sectors. In the north the alluvial deposits are dissected and form a series of terraces, the highest marked by the watershed (FIG.20). The sediments are sub-rounded to sub-angular gravels. These constitute the remains of extensive alluvial fans created at the feet of the Umbro-Marchean

Ridge, as can be inferred by the almost exclusively calcareous composition of the clasts. These terraces are weathered by a relict alfisoil, that it is usually severely trunkated, characterised by a strongly rubified clayey horizon (Bts), with a thickness greater than 1-2 m in places, and with only siliceous clast preservation. This is a classic soil profile associated to the Last Interglacial (COLTORTI, 1996; COLTORTI & PIERUCCINI, 2004), allowing the underlying alluvial units to be attributed to the MIS 6. The successive alluvial terraces are preserved inside valleys cutting the older alluvial deposits and the marly-aranaceous bedrock. This indicates that the incision, that began during the Last Interglacial, affected the bedrock. Of particular importance for the recent evolution of the area is the fluvial capture of the "Rigo Stream". Previously this important stream passed north of Colbassano, but during the final part of the Late Pleistocene, it was deviated with a visibly sharp bend to the south onto the plain of Fossato di Vico. In the Holocene the Late Pleistocene units were incised and terraced although the alluvial deposits inside this unit have limited width and thickness.

In the southern sector, south of the "Feo Stream", two terraced units are also present but the bedrock does not crop out. The older terrace is represented by a reduced remnant south west of Gualdotadino (Colle di Sopra). The more recent terrace, on which the town is built, forms an extensive gently flattened alluvial fan, which from the feet of the calcareous slope reach the western side of the basin. The deposits are angular and sub-angular gravels, consisting of mainly calcareous clasts, locally interrupted by entisoil with an A1C profile, few centimetres thick. The datings of the organic horizons of the soil allow us to establish their evolution during the end of the Late Pleistocene (see Stop 5). The mean grain-size decreases towards the distal zone of the alluvial fan, where fine to very fine gravels and coarse calcareous sand forms up to decimetric layers. The outcropping Gh and Sh structures are interlayered and, in some cases, lie inside very shallow channels (Ch; MIALL, 1985; 1996) created by sheet-flood dynamics. The sedimentary architectural environment is typical of a distal alluvial fan, or a well supplied braid-plain. Sediment was supplied from reworking of stratified slope-waste deposits, formed by crionival processes affecting the valley sides (COLTORTI & DRAMIS, 1988; 1995). To the south of the Feo Stream, the summit of the terraced alluvial fan lies at 525 m near the calcareous valley side, and at 450 m as it approaches the western Marly-Arenaceous valley side.

As previously mentioned, in the area between Corraduccio and Cerqueto (between the Rigo and Feo Stream), the outcropping deposits have been considered to be slope debris (CENTAMORE et al., 1978; BOSI et al., 1987). The latest authors distinguished two units separated by an erosive unconformity. The older unit outcrops along the whole valley side reaching an elevation of 495 m at S. Pellegrino and it is affected by metric to decimetric folding (FIG.20). This unit also outcrops on the valley bottom, showing the same deformation. A detailed sedimentological analysis shows that the sediments are similar to those forming the distal part of the Gualdotadino alluvial fan, and are therefore poorly reworked alluvial deposits. Gravitational processes are responsible for the observed folding, suggesting the presence of a sliding surface at their base. However, the bedrock has never been observed involved in the deformations. Inside a quarry near Cerqueto, it was possible to see the progressive deformation of alluvial deposits. The dating at ca. 20 ka of

a thin soil confirmed that they were formed contemporaneously with the undeformed deposits of the distal part of the Gualdotadino alluvial fan.

On the western valley side, close to S. Pellegrino, a more recent undeformed sequence of sediment cutting the older unit has been observed. It rarely exceed few metres in thickness and, in some places, it contains some arenaceous clasts. These sediments cover the entire slope and therefore indicate that, at the end of its deposition the valley bottom was closer to the present day one. The sedimentological characteristics of the deposits, similar to that of the distal alluvial fan, indicate slope-wash dynamics on bare slopes and therefore cold climatic conditions, still in the Late Pleistocene.

Although our investigations shows that sheet-flooding processes formed the deformed deposits, the events that led to the present day situation are not completely understood. It is unlikely that they were deposited at higher elevations than present, because there is a very short time frame available to explain the very rapid creation of the valley, as indicated by the second generation of sediments. This part of the valley was created very quickly during the end of the Late Glacial. The easier explanation is that the area was affected by deep gravitational deformation, the valley itself representing a huge trench. The main detachment surface would be in correspondence with the fault running along the calcareous valley-side between Fossato di Vico and Gualdotadino, which extends for a total of 10km. The sliding surface would become listric at depth, and it is possible that it reactivates, in opposite direction, old overthrust planes using a mechanism analogous to that of tectonic inversion. In S. Pellegrino, the maximum elevation at which deposits outcrop corresponds to an area of deposition, preceding the catastrophic activation of the deep-seated gravitational movements. The deformation of the deposits occurred after or contemporaneously with the tilting of the western side of the trench. This hypothesis is partly confirmed by the presence, along the eastern valley-side (the detachment escarpment), of a limited thickness of stratified slope-waste deposits (rarely thicker than 10 m). These deposits reach a thicknesses of over 20 metres in Gualdotadino (near the Hospital), inside a secondary valley, as well as along the eastern slopes located to the south of the area affected by the deformation. The limited thickness would be associated to the limited time span after the activation of the event, available for the deposition of crionival deposits.

An alternative explanation could be pointed out if we take into consideration the similarities of the S.Pellegrino deposits with fluvio-glacial and pro-glacial deposits observed further to the east, in the Marche region (Castelluccio di Norcia, COLTORTI & FARABOLLINI, 1995) and to the south, in the Abruzzi region (SERVIZIO GEOLOGICO D'ITALIA, in press) at higher elevation (over 1000 m). If the Gualdotadino valley had hosted a small valley glacier, fed from a large cirque that is located on the eastern slope, the fluvial deposits would have represented proglacial and supraglacial deposits. The rapid melting of the glacier would have explained the collapse of the deposits and their deformations. However, the southern exposure of the valley, the low elevation of the deposits, the fact that glacial deposits has never been discovered, and overall, the absence of similar deposits along the same valley as well as in the nearby areas (also at higher elevation) allow us to discard this possible explanation.

5.1 STOP 5
The Gualdotadino alluvial fan

Some sections (Fig.21), outcropping near the supermarket Lidl in Gualdotadino, allow the sedimentological characteristics of the inner part of the alluvial fan to be observed. The deposition of gravels, which can in this sector reach a depth of about 15 m, is interrupted by the evolution of soils with profile A1C. The deposits are made of horizontally and cross-bedded angular to sub-angular gravels, sometimes filling shallow channels (Gh, Ch). Occasional lenses and sandy/silty layers are present (Sh, Fh). This facies association is typical of a proximal to mid part of an alluvial fan architectural model. Three soils were dated within the 200 m long outcropping sequence. The first two soils have been dated in the more proximal sector (Lidl East, FIG 21) and gave ages of 27770 ± 300 yr BP (Beta 170673) and > 28460 yr BP (Beta 170675) respectively. The third dating came from a samples located in the more distal part of the sequence (Lidl East), and gave an age of >27190 yr BP (Beta 170676). This shows that during the sequence belong to the Late Glacial and the sedimentation, on this part of the alluvial fan, was interrupted for a brief period.

5.2 STOP 6
The Cerqueto deposits

Various sections from the same alluvial fan were identified near Cerqueto, allowing the analysis of the sedimentological characteristics from the more distal sector of the alluvial fan, before the Torrente Rasina dissects the aranaceous ridge. These are fine angular to subangular gravels, alternating with sandy and silty layers, giving Gh and Sh facies, sometimes deposited inside shallow channels. North of Cerqueto, in the central part of the outcropping sequence the datings of a buried soil, gave an age of 20.850 ± 120 yr BP (Beta 194852) (FIG. 22). The depositional environment is typical of an ephemeral distal braidplain (Mod. 11, MIALL, 1996) or sheet-flood fluvial plain subject to highly flashy discharge (Mod. 12). Some deformation is present in the section and, near the bedrock, the strata dip at 15° to the NW. The contact with the bedrock is a sliding surface, secondary and listric to the main one that should be located along the western escarpment. Analogous deformation in deposits with similar sedimentalogical characteristics is observed in various locations including the centre of the plain. Analogous deposits outcrop for about 10 km along the western slope of the basin, up to an elevation of 495 m a.s.l. at S. Pellegrino with respect to the valley bottom. Cerqueto marks the most northerly area in which the deformed deposits are recognised. They appear in a transitional area with non-deformed deposits outcropping south of T. Feo, on the distal part of the alluvial fan.

At C. Pennoni (FIG. 21) an analogous facies association was observed, characterised by alternating decimetric to centimetric fine and very fine angular gravels with coarse-grained calcareous sandy layers (Gh and Sh). The thickness of the outcropping sequence is greater than 10 m. Dating of molluscs in the sandy layers, from the central part of the sequence, gives a age of 20180 ± 80 BP (Beta 170678). In the vicinity, mollusc shells coming from the upper part of the sequence, just below the actual soil gave an age of 7500 ± 40 BP (Beta 170677). It is probable that the Holocene age of the molluscs is due to

alteration processes affecting the shell, given that in Italy similar depositional dynamics are not observed during this time interval.

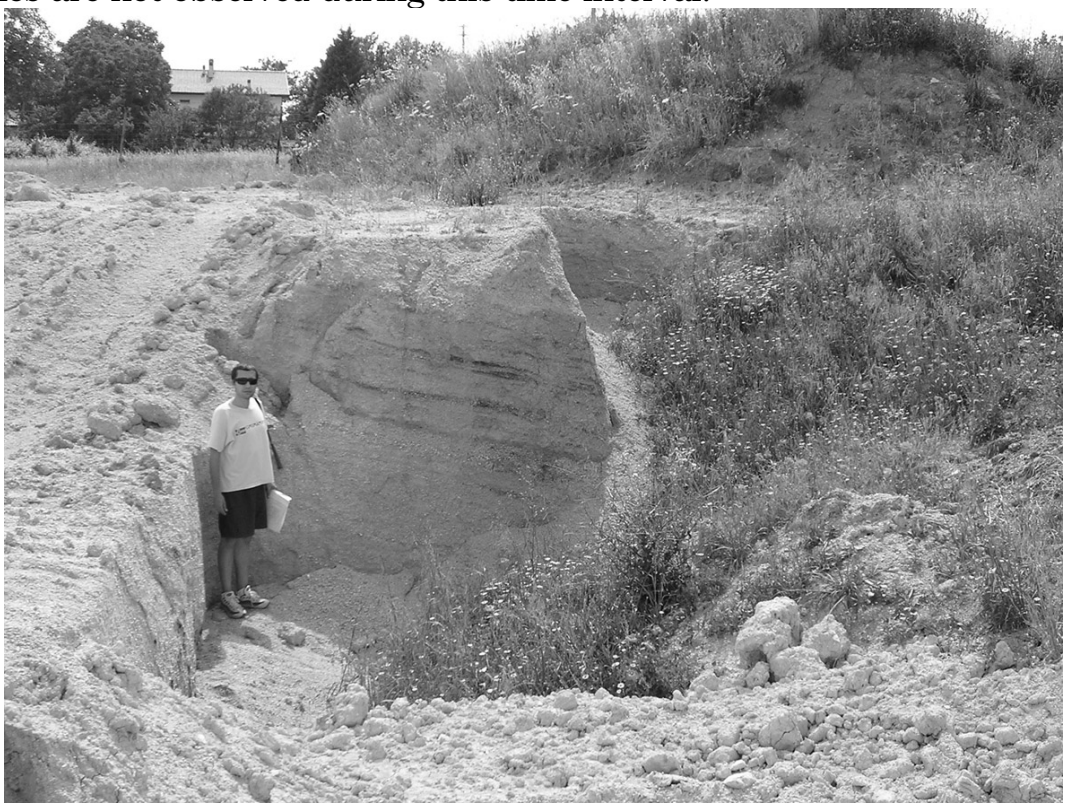

Figure 22 – The tilted alluvial fan deposits to the east of Querceto

From this Stop it is also possible to observe the fault plane which separates the Umbro-Marchean Ridge from the "Umbrian Pedappennine", that, in the sector between Fossato di Vico and Gualdotadino, may have been affected by deep gravitational movement (FIG.19, 23).

Figure 23 – The fault escarpment on the eastern side of the Gualdotadino Basin.

6. The Colfiorito Basin

The Colfiorito basin and nearby areas (FIG 24, 25), contains one of the best dated late Early and early Middle Pleistocene sequence of the Northern Apennine. The central part of the basin belong to the Chienti River drainage system, draining to the Adriatic side but, to the west, there are some closed depressions that constitute the transition with the Tyrrhenian areas. The Percanestro stream, located to the south of Colfiorito, once was a tributary of the Chienti River and nowadays it is tributary to the Nera-Tiber. It cuts one of these closed depressions evidencing a sequence over 90 metres thick. The sequence can be roughly divided into 3 parts where gravel deposition in a

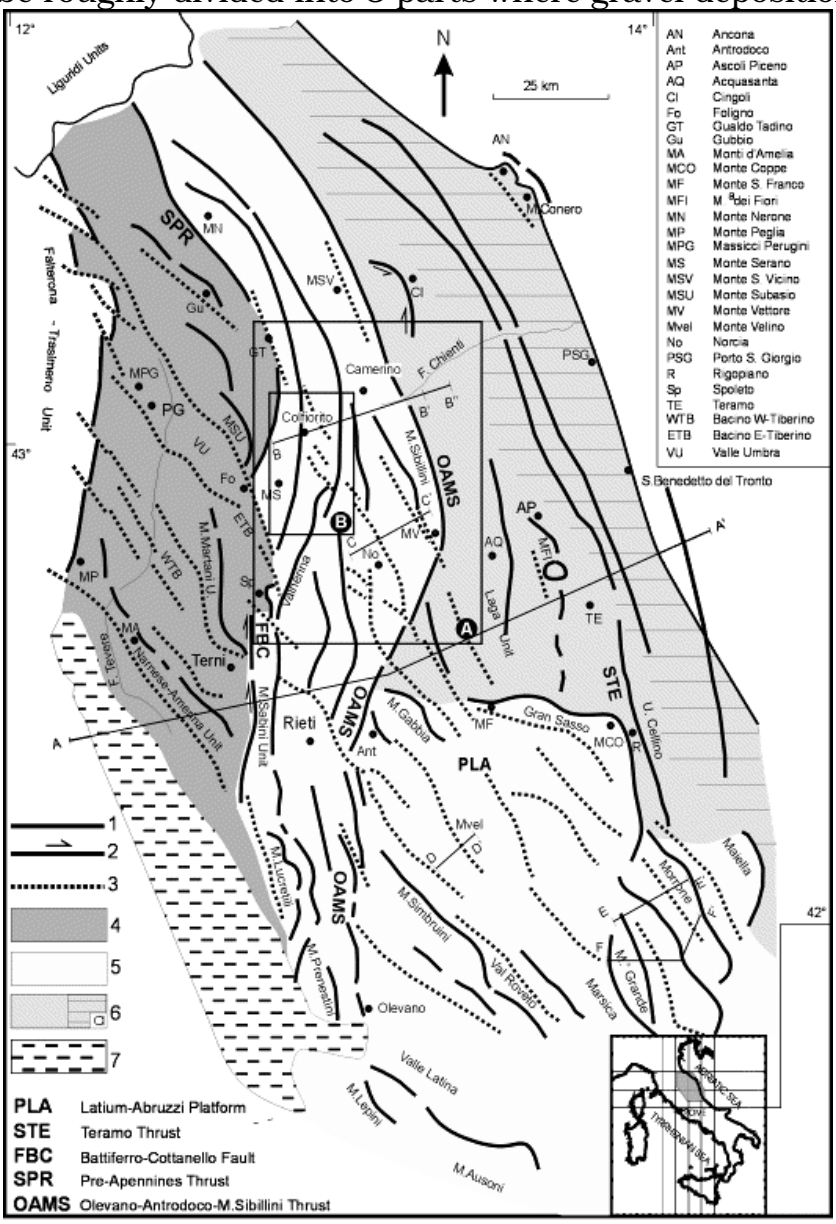

Figure 24 - Regional tectonic map of the Umbro-Marchean–Abruzzi fold and thrust belt (Central Apennines: Italy). (1) Thrust; (2) strike–slip and/or transpressive fault; (3) normal fault; (4) Western Umbria and Preapennines; (5) Umbro-Marchean–Abruzzi Ridge; (6) Pedeapennines, (a) peri-Adriatic area characterized by vast outcrops of Middle Pliocene–Lower Pleistocene transgressive deposits; (7) Quaternary volcanoclastic deposits.

braidplain environment dominates in the lower and upper parts, separated by a phase of clay deposition in a lacustrine environment.

The Colle Curti sequence (FIG. 26), is constituted, at the base, by subangular and angular coarse gravel bars (Gm, MIALL, 1996) and rare small channel fills (Gt). These sediments are interrupted upwards, by few decimeters thick layers of massive silts and clays (Fm) sometimes weathered into thin reddish vertisols. They are the result of the deposition in a braidplain environment (Model 2 of MIALL, 1996) with coarse material becoming progressively less abundant along with the establishment of very shallow lakes. 45 mt above the base of the outcropping sequence, inside medium size gravels interbedded in clays, a mammal assemblage has been discovered. The sequence continues with 14 meters of clays with few thin gravel lenses which bear witness of long lasting lacustrine conditions. It is followed by over 30 meters of gravel bars (Gm) with rare channel fills (Gt) interbedded with sandy layers (Sh and St) and with silts, clays and peat (Fm and C), at places weathered to blackish vertisols. They represent again the establishment of a gravelly braidplain alternating with periods of shallow swamps formation. At Voltellina village, pyroclastic sediments with sanidine crystals have been discovered dispersed in a black vertisol. Being contained in finer sediments these crystals probably represent a thin tephra layer. The age of the sample is ca. 427 ka.

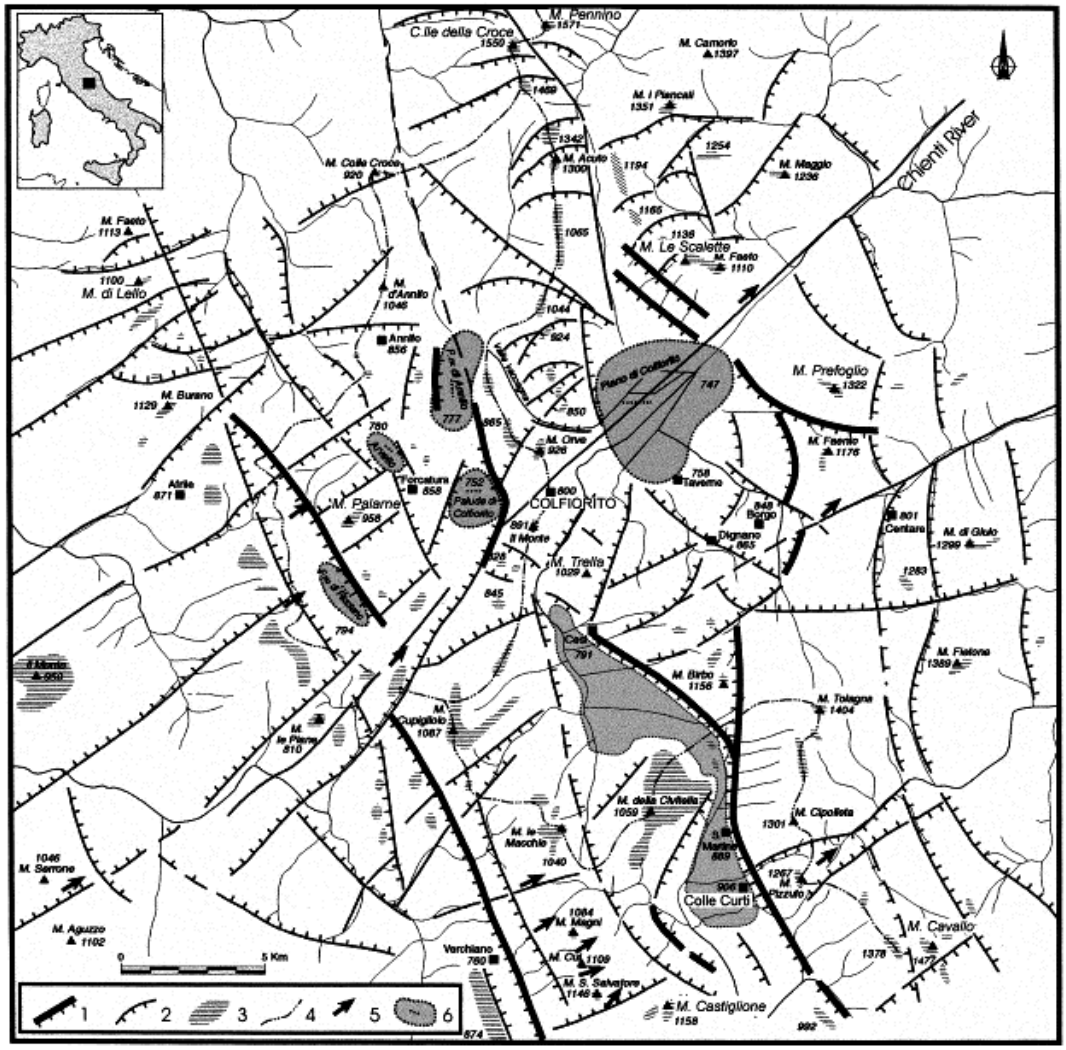

Figure 25 - Geomorphological map of the Colfiorito area. (1) Lineaments which correspond to the Quaternary normal faults, (2) lineaments (barbed lines show the downthrow-block as suggested by the displacement of the planation surface remnants), (3) planation surface remnants, (4) watershed divide, (5) trend of paleo-drainage, (6) intramontane basin, filled by Quaternay fluvio-lacustrine sediments.

The part of this sequence above the thick clay layer is weathered into a reddish soil with a Bts horizon similar to those presently developed in tropical warm environments. The pedogenetic process affected gravels mainly composed by limestone elements with a scarce flint component. These processes determined the complete leaching of the limestone gravels and the concentration of the flint. The present day percentage of flint in the gravels of the lower part of the sequence rarely exceeds 10-20 % suggesting a great reduction in thickness (ca 100 meters !!).

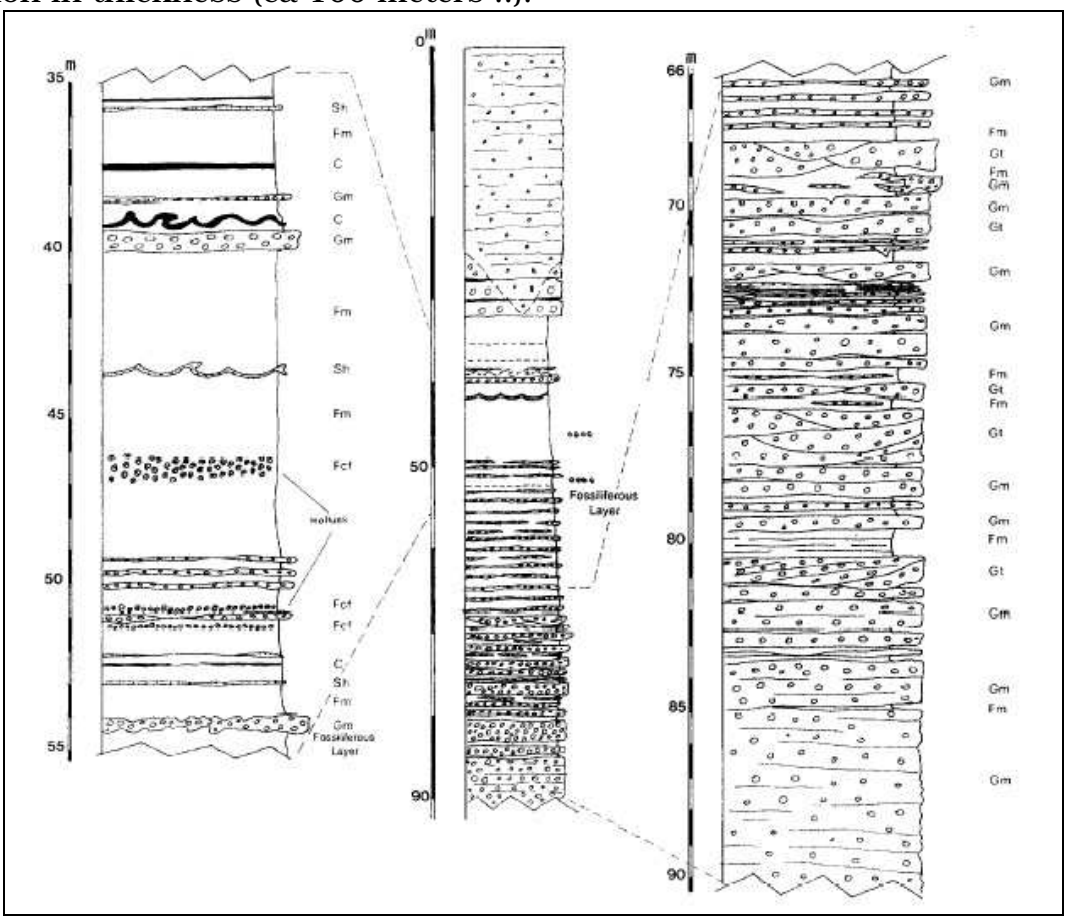

Figure 26 – Stratigraphy of the Colle Curti deposits with details of the outcropping and unweathered part of the sequence.

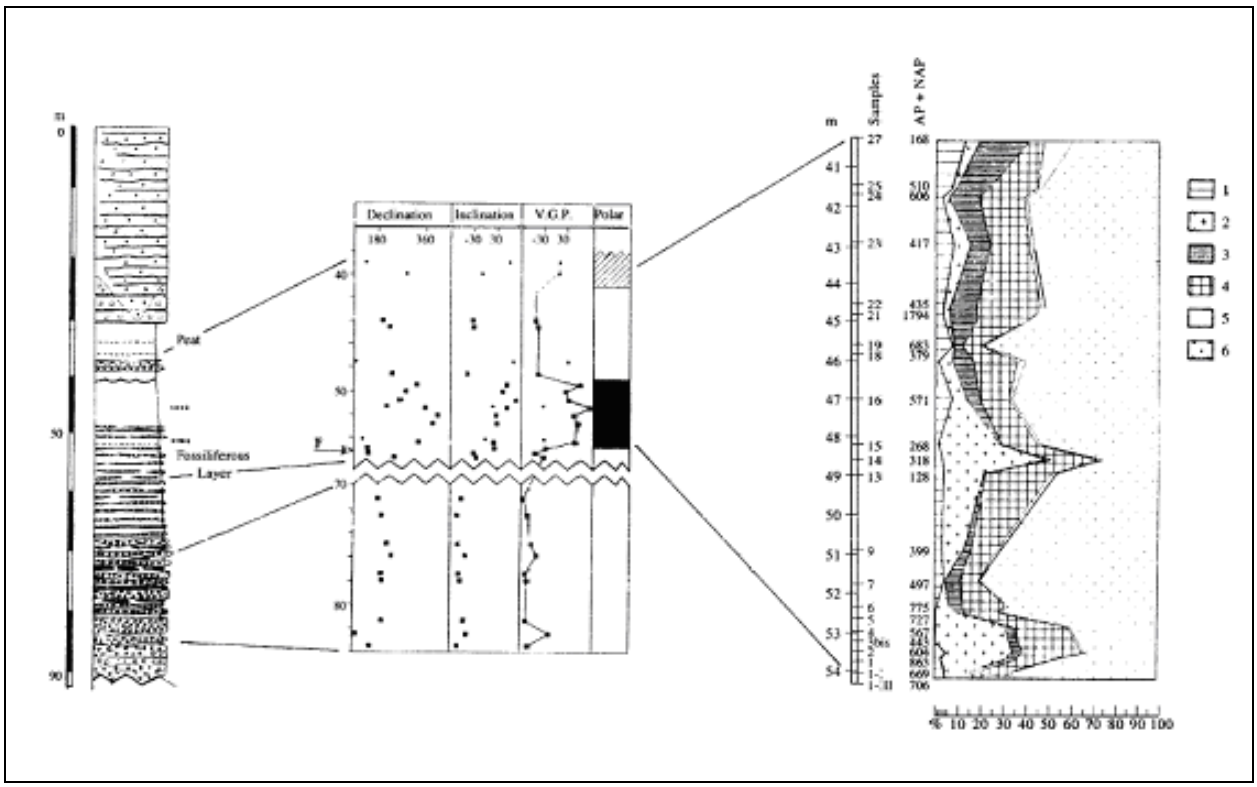

Figure 27 – The 90 m long sequence is plotted against the paleomagnetic characteristics and the main pollen association: 1- deciduous elements; 2- Tsuga and Cedrus; 3-Abies and Picea plus Betula and Fagus; 4-Pinus plus poorly preserved pineaceae pollen grain; 5-Other trees; 6-Herbs and other taxa.

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 an earliest Galerian Mammal Age (Upper part of the Early Biharian) (FICCARELLI et al., 1997). The assemblage is the most modern of those Local Faunas that, while maintaining some Villafranchian character, record in Italy the series of migratory and evolutionary events that mark the great faunal turnover which occurred during the Middle Pleistocene.

The intermediate clayey layer was investigated from a palinological point of view (FICCARELLI et al., 1997). In a synthetic palynological diagram (FIG 27) the pollen taxa are grouped according to their ecological and climatic requirements: deciduous elements (*Quercus, Carya, Carpinus, Ulmus-Zelkova*, etc.); *Tsuga* and *Cedrus; Picea* and *Abies* plus *Betula* and *Fagus*; *Pinus* plus Pinaceae poorly preserved pollen grains; other arboreal pollen plants (AP); non-arboreal pollen plants (NAP). The semidetailed pollen diagram shows the percentage of the most significant taxa characterizing the assemblage. Percentage of arboreal and non arboreal pollen are based on the total pollen sum. *Pteridofita, Zygnemataceae* (*Spirogyra, Zygnema*), *Concentricystes, Pediastrum, Botryococcus* are well represented particularly in the upper part of the section.

The successive abundance of the main pollen association define at least 4 pollen zones (FIG.28). The first one (I pollen zone; samples 1-III to 4) is characterised by a major episode where the dominating taxa are *Tsuga* and

Cedrus. Just at the base a high percentage of *Asteraceae Cichorioideae* and *Poaceae* is recorded.

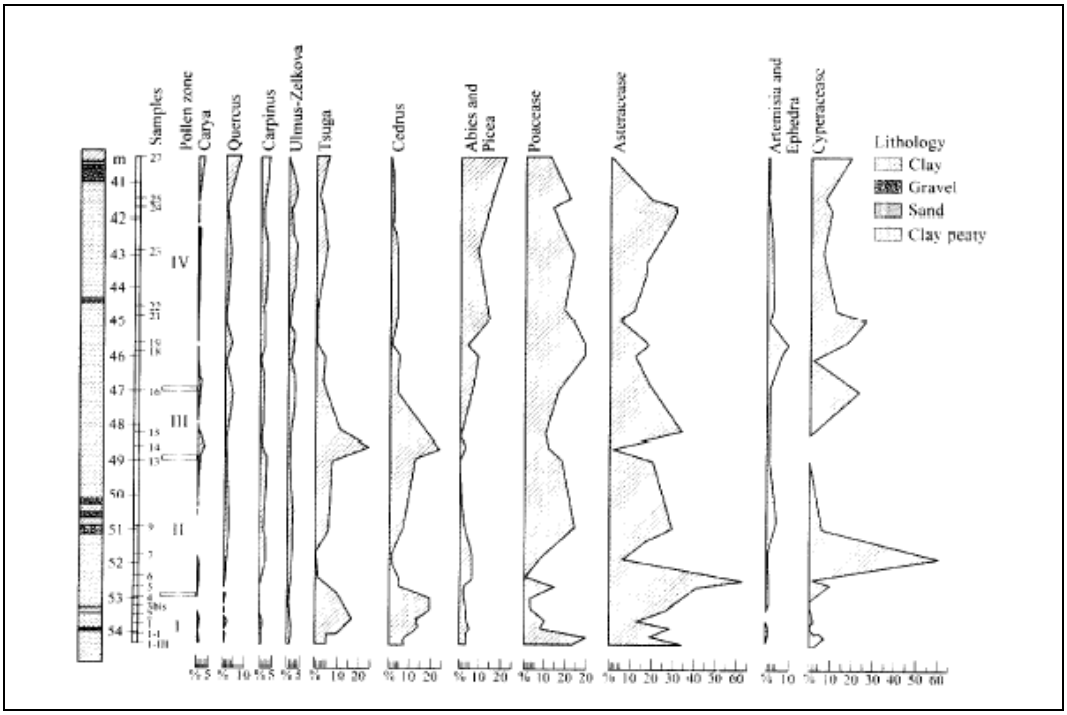

Figure 28 – Detail of the pollen diagram of the Colle Curti section

From sample 4 (II pollen zone) *Asteraceae Cichorioideae* followed by *Cyperaceae* show a strong increase. *Poaceae* are also well represented. *Artemisia* shows a peak in correspondence of sample 9. The development of an open vegetation with steppe-like character testifies to a climatic deterioration involving a decrease in humidity and probably in temperature. The high percentage of *Cyperaceae* probably reflects a contraction of the lake consequently to the establishment of drier conditions. This phase reaches sample 13. Between sample 13 and 16 a new expansion of forest occurs mainly with *Tsuga* and *Cedrus* (III pollen zone). A new phase (IV pollen zone) characterized by a new development of open vegetation and by a strong decrease of *Tsuga* and *Cedrus* starts from sample 16. In the lower part of this phase the mesophilous forest is replaced mainly by steppe vegetation. Montane elements such as *Picea* and *Abies* progressively increase. Near the top of the section *Picea* reaches 15% and *Abies* reaches 25%. At the same time, among the open vegetation elements, *Asteraceae Cichorioideae* record a strong reduction. In sample 19 pollen grains of *Koenigia islandica* and a new peak of *Artemisia* are recorded. *Pinus* is costantly and well represented along the section whereas the deciduous forest elements although constantly represented are attested on lower percentage values. Mediterranean xerophytes are sporadic. In the Colle Curti diagram, the striking feature is the alternation between a high percentage of open vegetation and high percentage of altitude coniferous forest pollen. In the Mediterranean region, the presence of an herbaceous formation, including *Artemisia*, has been correlated with late Neogene-Pleistocene glacial periods (BERTOLDI, 1977; SUC & ZAGWIJN, 1983; WATTS, 1985). The dominance of *Asteraceae Cichorioideae*, *Poaceae* and *Cyperaceae* in the Plio-Pleistocene section of Catalonia characterized a cold

steppe type with a no marked seasonality with respect to the *Artemisia* steppe (LEROY, 1990).

At Colle Curti, the open vegetation mainly represented by *Asteraceae Cichoroideae, Poaceae, Cyperaceae*, indicate dry and probably cool/cold climatic conditions. The occurrence, even if discontinuous, of *Artemisia* marked at least two periods of decreased aridity and temperature. The phase marked by this association seems to reflect the presence of an altitudinal steppe with cold winters and dry summers (i.e.glacial phase). Mesophilous taxa such as *Tsuga* and *Cedrus* and altitudinal microtermic elements such as *Abies* and *Picea* are present. The ecologic and climatic requirements of *Tsuga* are significant for the definition of phases I e III: this element grows under cool temperature and strong atmospheric humidity; values of 10-15% in the pollen diagrams of *Tsuga* do not seem consistent with the presence of a dry summer season. Phases I and III seem characterized by higher temperature with respect to phases II and IV and by an increase of humidity (i.e.Interglacial in altitudinal site or interstadial). The replacement of steppe by a forest dominated by mesophilous elements typical of the end of an Interglacial is problematic because of the lack of a progressive increase of deciduous trees. The palynological data, supported by sedimentologic and paleomagnetic evidence, indicate that the sequence can be related to large part of the Bavelian and part of the Menapian of the North European sequences. In the middle part, where clay layers outcrop, two cool and humid phases are followed by colder and drier periods.

This sequence was also investigated from a paleomagnetic point of view. The sequence covers a part of the Matuyama epoch which includes the Jaramillo event (FIG.27). The mammal fauna is contained inside the Jaramillo Chron.

Slightly to the north, inside the Colfiorito Basin, close to the village of Cesi, a palaeontological excavations exposed a sequence, about 20 m thick, that in the upper part was eroded to create an alluvial terrace (COLTORTI et al., 1998). Its lower outcropping part is represented by predominantly clayey rich sediments with pebbly lenses. At the top of this 8 m of massive clays, fossil bones are preserved in a fine gravel bed 20-50 cm thick, formed by subangular and angular siliceous gravels, resulting from decalcification of limestone clasts, which was locally affected by load casts. A thin pyroclastic layer (bed U.S. 4a) discontinuously occurs at the base of the gravels. Gravels weathered by a tropical reddish palaeosoil, of which only the flint elements remain, occur at the top of the sequence.

The sequence is similar to that previously described in Colle Curti, although only the middle and upper part crop out in Cesi. The sediments testify a lacustrine environment which, in its upper part, was reached by an alluvial fan prograding into the lake. However, the upper part of this sequence was severely eroded.

The taxa hitherto recognised from Cesi are *Elephas* sp., an advanced representative of *Stephanorhinus hundsheimensis*, a caballine equid, *Hippopotamus* sp., *Megaceroides solilhacus*, *Cervus elaphus*, *Dama clactoniana*, *Bison schoetensacki* and *Homotherium* sp. .

The palynological content of the 12 m sequence underlying the fossiliferous bed (U.S. 4), taken as reference (quote 0), has been studied (COLTORTI et al., 1998).

The Cesi pollen diagram is characterized by the dominance of herbaceous elements, especially represented by Poaceae and Asteraceae. Among the

Asteraceae, *Artemisia* is very abundant, sometimes in association with *Ephedra*, another steppe element. Ranunculaceae, Caryophyllaceae, Plantaginaceae and Polygonaceae pollen are also well represented. Arboreal plants are mainly represented by *Pinus* pollen that can reach frequencies of 73.65%. *Abies* and *Picea* are less abundant although always present, while *Tsuga* and *Cedrus* are sporadic. Deciduous broad-leaf trees, never more than the 6%, are mainly represented by *Quercus*, and sometimes also by *Acer* and *Carpinus.* Mediterranean xerophytes are rare.

Zygnemataceae and Chlorophyceae Algae and some local herbaceous plants provide indications of the history of the depositional environment. In particular, the constant presence of *Pedastrium*, in association with *Botryococcus, Spirogyra* and *Mougeotia* suggests a fresh-water environment, sometimes under mesotrophic conditions.

The Cesi pollen assemblages suggest the occurrence of a landscape dominated by open open vegetation characterized by taxa (Poaceae, Asteraceaeae, including *Artemisia, Ephedra, Chenopodiaceae*, etc.) typical of the glacial phases after 2,6 Ma in the Mediterranean region (SUC et al., 1995).

The markedly arid climate prevented the growth of plants demanding year-long humid conditions. Broad-leaf trees, mainly represented by *Quercus* pollen grain, occerred but always at low frequencies. Woodland was dominated by *Pinus*. The phases characterized by the expansion of *Pinus* are probably linked with slight increases of the temperature (wooded steppe).

In the upper part, from the sample immediately below the fossiliferous bed down to that at 6,5 m, the magnetizations have a simple normal component behavior. From 7 m downwards, a clearly reversed polarity occurs. The Cesi fauna is therefore much younger than the Colle Curti and can be attributed to the beginning of the Brunhes Epoch, ca 600 ka ago.

These sequence were deposited in tectonic depressions generated by the activation of a west dipping high angle normal fault that dammed a NE oriented drainage system (FIG. 29). Evidence of this first drainage network with paleovalleys oriented N-S and E-W with minor NW-SE and NE-SW direction is well preserved in the Colfiorito area, today located on the Tyrrhenian-Adriatic watershed. These valleys generate a "youth" landscape (sensu DAVIS, 1899), because the remnant of the planation surface are still very well recognisable in the area. After the activation of the fault, the watercourse run along the fault escarpment, and joined the paleo-Chienti in the Colfiorito basin that was still able to cut the ridge. Later, following more intense uplift movements and the renewed activity of the fault, probably couped with fault activity in the Foligno-Spoleto Graben, the drainage was reversed and captured to the Tyrrhenian side. During the beginning of the activity of the fault, the drainage system was slightly downcutted into the planation surface that was much closer to the sea level. In fact, further to the east, coastal deposits of the same age are preserved close to the present day coastline, at an elevation of 400-500 metres a.s.l.

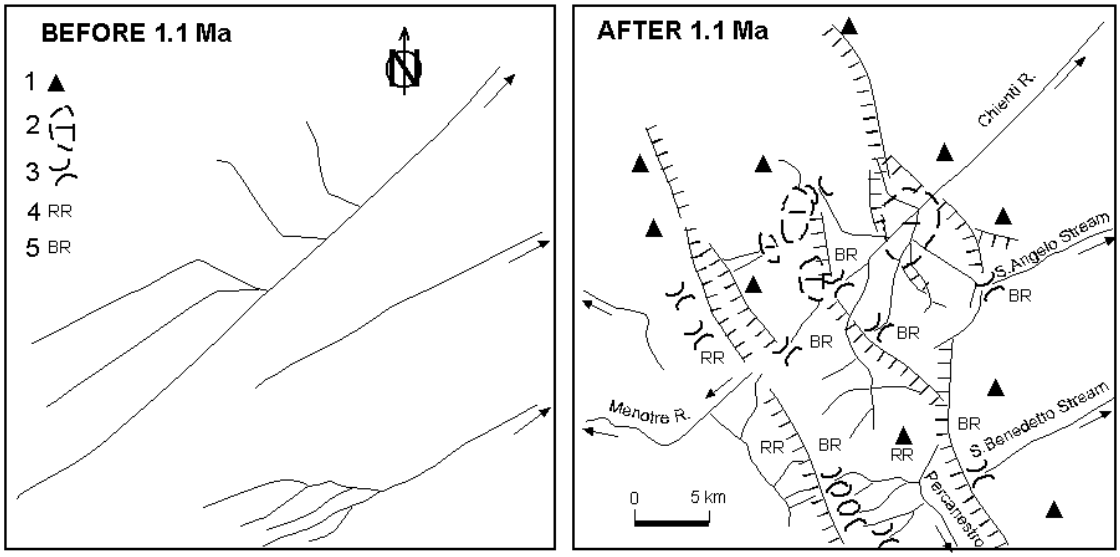

Figure 29 – The evolution of the drainage network in the Colfiorito area following the activation of extensional faults. 1- Main relief; 2-Closed depressions; 3-Wind gaps; 4-Reversed rivers; 5-Beheaded rivers.

6.1 Stop 7
A general overview

The Stop in the Colfiorito plain, near the old Abbey of Plestia, shows the characteristics of the landscape that constitute the Adriatic-Tyrrhenian watershed of the Apennine.

The higher parts of the area are characterised by well preserved remains of the Pliocene planation surface. This surface is faulted and displayed at different elevation around the Colfiorito plain by a series of fault escarpments. The steepness of the escarpments are more or less similar suggesting a comparable age for their activation. To the north, in correspondence with the ridge of M.Pennino, it is possible to observe a staircase of faults downthrowing the planation surface to the south. A similar staircase is recogniseable on the ridge to the west and also to the south and south east of Colfiorito.

The fault escarpments are very fresh and in some places they correspond to the fault plane, affected only by minor by erosional features (CALAMITA et al., 1998; 2000). This is one of the best evidence of an Holocene activity. The Colfiorito plain has been the epicentre of the 1997 earthquake but there is still a debate as to whether these faults have been reactivated during this event. In fact, a 10 cm scarplet has been observed only when the base of the scarp is made of loose debris suggesting the possibility that the earthquake had mostly a strong compactional effect. In correspondence of a trench that cut trasversally the continuation of these faults in the Colfiorito plain (fig.) no displacement of organic layer dated to the MIS 2 (MESSINA et al., 2002) has been observed. However, even in this case it is possible that the diplacement was absorbed by the loose material and did not reach the surface.

From this position it is also well recogniseable the old paleovalley of the Chienti that still constitute the source area of the river although it generate an important step in the longitudinal profile. To the west the valley is also affected by normal faults that generate a closed depression hosting a small lake (the Colfiorito swamp).

The dynamics that affected this area (FIG 29) are a model for the drainage captures and drainage inversion that occurred along the Apennine watershed since the late Early Pleistocene.

Figure 30 – The upper part of the Colle Curti sequence made of weathered flinty gravels.

6.2 Stop 8
The Colle Curti deposits

This section (FIG. 30), located close to the Colle Curti village, is dedicated to the sedimentological characteristics of the upper part of the sequence cropping out in the Colfiorito Basin. A series of road cutting allow us to observe the sedimentological characteristics of the coarse alluvial unit that reach a thickness of 30 metres in places. They are constituted by a series of angular and subangular gravels made of flint. As previously mentioned the exclusive flint composition is due to strong weathering processes. The original composition, included a large majority of limestone gravels that have been completely leached. These are not reworked materials because frequently the flint gravels are fragmented in situ along their weakness lines (i.e. the small fracture). The upper part of the sequence is usually completely weathered and only the thin gravel intercalations at the base of the sequence has again a predomint limestone composition. The coarse layers have an horizontal stratification (Gh), or they are deposited inside shallow channels (Gh). Some sandy and clayey intercalations have also been observed in places (Sh and Fm). They become more abundant in the lower part of the sequence where the clay layers increase in number and thickness before to reach the maximum thickness (up to 15-20 metres). Inside these thin intercalation a thin piroclastic layer has been dated at ca. 427 ka. This part of the sequence is within the Bruhnes Epoch. This indicate that the weathering process responsible for the complete leaching of the sequence occurred afterward and could be therefore associated with a long Interglacial (MIS 11 –13 –15 !!). It is also possible that the weathering is the result of more than a (Interglacial)

weathering cycle. The morphology of the upper part of the deposits is quite flat and only locally is buried under the younger alluvial fan coming from the local slope.

7 - Fluvial sedimentation in the Marchean river valleys (Coltorti M. & Farabollini P.*) *Dept. of Earth Sciences, University of Camerino

Almost all the Marchean rivers sources are located in Umbro-marchean Apennines and cross the whole region with a SW-NE trend, reaching the Adriatic sea. Their Quaternary evolution is due to the interaction between climatic changes and regional uplift, which has affected the Apennines since the Early Pliocene, and the Peri -Adriatic basin since late Early Pleistocene (CANTALAMESSA et al., 1986; COLTORTI et al., 1991; DRAMIS, 1992; BIGI et al., 1996). After this period, a flight of alluvial terraces were formed and are now preserved at progressive elevation on the thalweg (FIG.31).

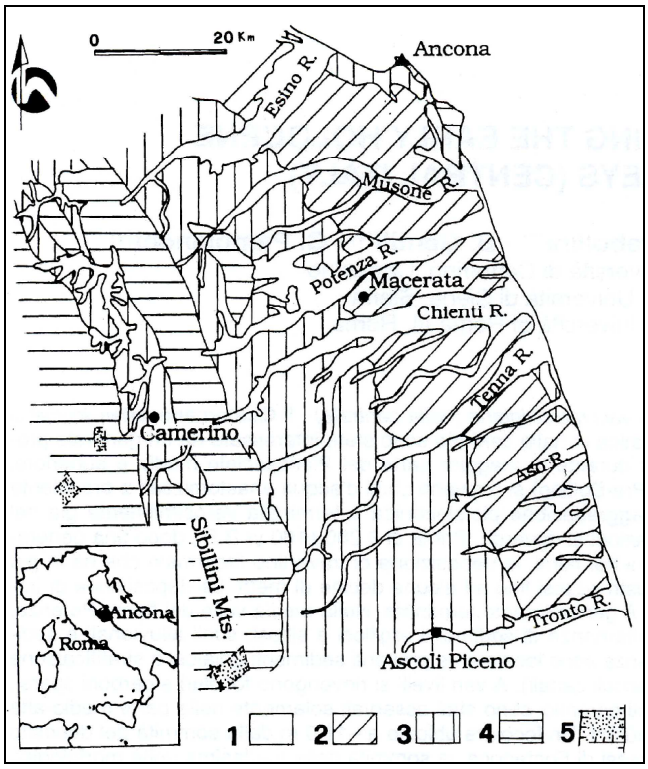

Figure 31 - Morphostructural scheme of the Marche region. Legend: 1- Pleistocene and Holocene continental and littoral deposits; 2- Pliocene-Early Pleistocene clays and sands (marine and littoral deposits); 3- Miocene turbidites; 4- Limestones and marly limestones; 5- Continental deposits of the intermontane tectonic depressions.

In the Marche Region, in correspondence to the limestone Ridges, narrow NE-SW oriented valleys were generated. The oldest fluvial deposits are missing or very scarce but the progressive lowering of the hydrographic network is locally marked by the presence of a series of karstic floors, very wide in places (CATTUTO, 1976; COLTORTI, 1981; BOCCHINI & COLTORTI,1990). In correspondence with the pelitic and arenaceous bedrock, that inside the chain are usually oriented NW-SE, the valleys can be many kilometres wide. In the Peri-Adriatic area, to the east of the limestone ridges, the valleys are wide and again oriented to the NE. The monoclinal setting of the bedrock led to the creation, in between the limestone ridges and the sea, of a cuesta landscape.

Close to the coastline, an Early-Middle Pleistocene series of marine and coastal deposits generated a sedimentary sequenze many hundres of meters in thickness. A series of stratigraphic and sedimentological investigations recognised a series of informal lithostratified Units later associated with sedimentological cycles (COLALONGO et al., 1979; NANNI et al., 1986;

CANTALAMESSA et al., 1986; MASSARI & PAREA, 1988) similar to what have been recognised on the Po plain margin (RICCI-LUCCHI et al., 1982). These gravelly and sandy deposits deposited in a beach, coastal lagoon, deltaic and alluvial plain environments are frequently indicated as "closing deposits" because they represent the end of the last marine cycle in the area (A4 and S4, COLALONGO et al., 1979; Qm1, CANTALAMESSA et.al., 1986). This is usually associated with the Sicilian marine stage, corresponding to the end of the Early Pleistocene. Locally, close to the coastline, the Pleistocene sequence is overlain by clayey lagoon and fluvial deposits (sequence A5, S5 ; Qc1 e Qc2). Close to the coastline, the top of the "Sicilian" sediments generate a series of wide marine terraces hanging over the present-day coastline at elevation between 130 (close to the coastline) and 450 m (in the inner part) a.s.l. This difference in elevation is partially associated to the lateral facies variation from coastal to alluvial fan deposits. However, part of this difference is associated with the presence of faults (NANNI et al., 1986; NANNI & VIVALDA, 1987; BIGI et al., 1996; CALAMITA et al.,1999). We remember that these deposits should be coheval of that recognised in the Colfiorito sequence.

After this moment the tectonic conditions that allow the creation of a wide marine Early Pleistocene basin along the Peri-Adriatic coastline ceased. The whole area emerged and the present day transversal drainage start to evolve with the deposition, of the alluvial terraces located at progressive elevation on the valley bottom. They testify the interference between the continuous uplift and the erosional and depositional phases in correspondence of the Quaternary climatic cycles (AMBROSETTI et al., 1982; COLTORTI et al., 1991).

7.1 The oldest fluvial deposits (Early-Middle Pleistocene)

Alluvial deposits inside a hydrographic network not so different from the present day one are widespread in the region (FIG.32). These deposits have been named F1 from LIPPARINI (1939) and VILLA (1942). They are usually represented by terraces of limited width, made of horizontal and cross-bedded coarse to medium size gravels (Gh, Gt). They are gravels bar and channel filling deposited in a wide braid-plain. The sandy interlayers are rare but become more frequent close to the longitudinal valley cut into pelitic and arenaceous terrain. In the southern part of the region they also represent wide alluvial fan that were directly fed from the limestone ridges. Wider deposits are present along the more important Rivers (Esino, Potenza, Chienti, and Tronto River). The thickness of the deposits varies according to the presence of paleovalley buried under the terrace. The top of th terrace is usally located at elebation ranging from 90 to 150 m a.l.m.. Usually the terraces, have a convergent longitudinal profile (COLTORTI et al., 1991; ELMI et al, 1987).

The chronological setting of these deposits are still unknown because chronological informations are missing. However, due to their climatic significance and to the attribution of the younger terraces to the MIS younger than 9 (MIS 8-6) most probably they could be attributed to MIS 10 and/or older.

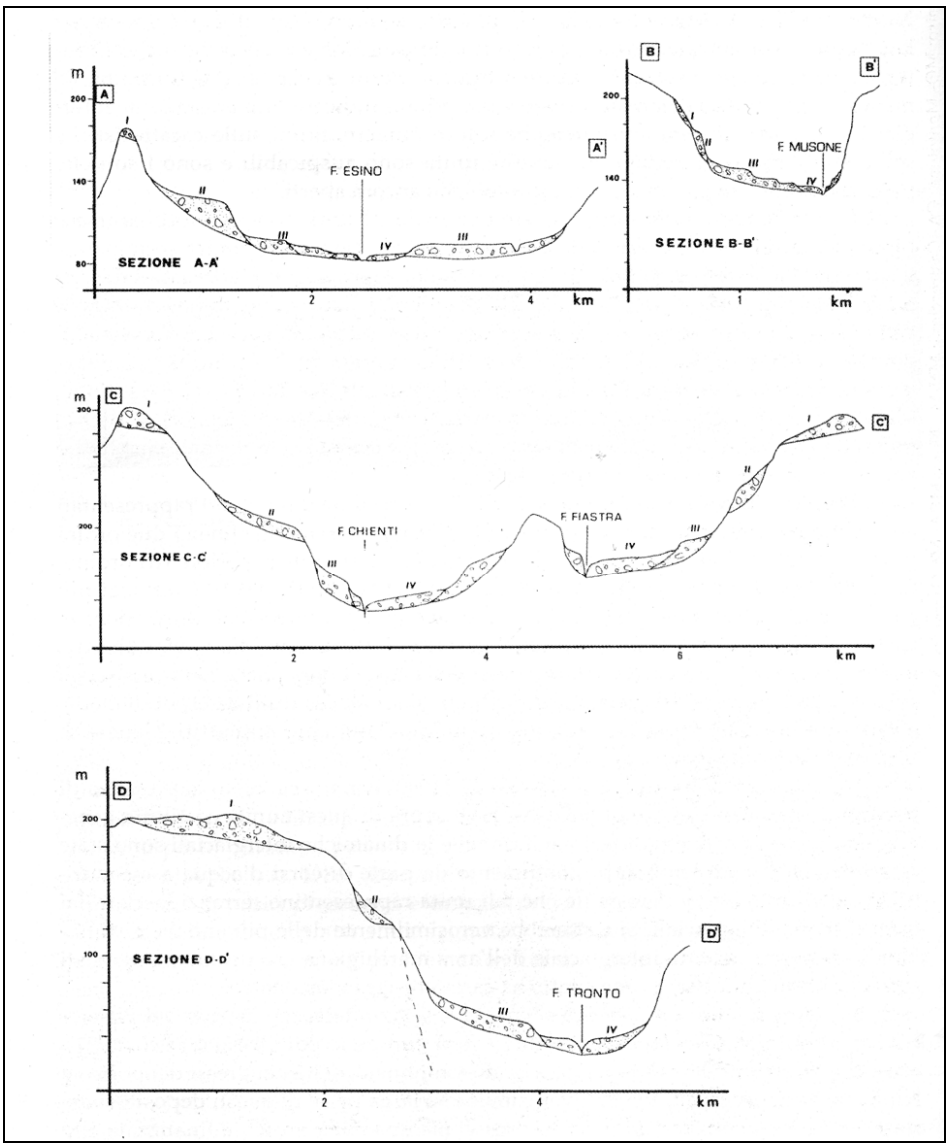

Figure 32 – Cross sections of the river terraces of the main valleys.

7.2 The late Middle Pleistocene alluvial deposits

The sediments belonging to this terraced unit are wider than the previous one and were deposited by hydrographic network similar to the present day. They were indicated as F2 (LIPPARINI, 1939; VILLA 1942). However, in many cases, this unit has been divided in two (COLTORTI & NANNI, 1987; NESCI & SAVELLI, 1990). From a sedimentological point of view, this unit is similar to the previous one and is made up of subangular and subrounded horizontally and cross bedded gravel or sandy gravels. However, a higher number of shallow channels filled with horizontal and cross bedded sands and silts increase. The Architecture was again a large braidplain. In the inner sector of the chain, these deposits are interlayered with clayey layers, many meters in thickness, that have been associated with a lacustrine environment due to the damming of the narrow valleys and gorges by alluvial fan and debris (COLTORTI, 1981) or by large size landslide (GENTILI & PAMBIANCHI, 1988). These deposits, contains Acheaulean reworked artifacts and Late Acheulean artifacts of levallois technique have been described at their top. At the top of this terrace an alfisoil with Bts horizon and an exclusive flint scheleton has been recognised. At its base sometimes there is a Cca horizon. All these

characteristics indicate that the terrace was deposited during MIS 6 (COLTORTI et al., 1981; COLTORTI, 1981). The discoidal and lamellar shape testify the reworking of debris produced by frost shattering on the limestone ridges. In fact, along the limestone slopes, estensive and thick stratified slope-waste deposits sometimes interlayered with the alluvial deposits have been observed (COLTORTI & DRAMIS, 1988; 1995).

7.3 The Late Pleistocene alluvial deposits

These are very well preserved and widespread alluvial terraces. Their attribution to the Late Pleistocene is well documented by many 14C datings and the presence, at their top, as well as in their inner part of Late Paleolithic tools (DAMIANI & MORETTI, 1969; ALESSIO et al.,1979; NESCI & SAVELLI, 1986; CALDERONI et al., 1991; 1996).

In the sequence of Mergaoni Stream (FIG. 33) they overlie a luvisol with Bt horizon that, slightly higher on the slope contain "in situ" Musterian artifacts.

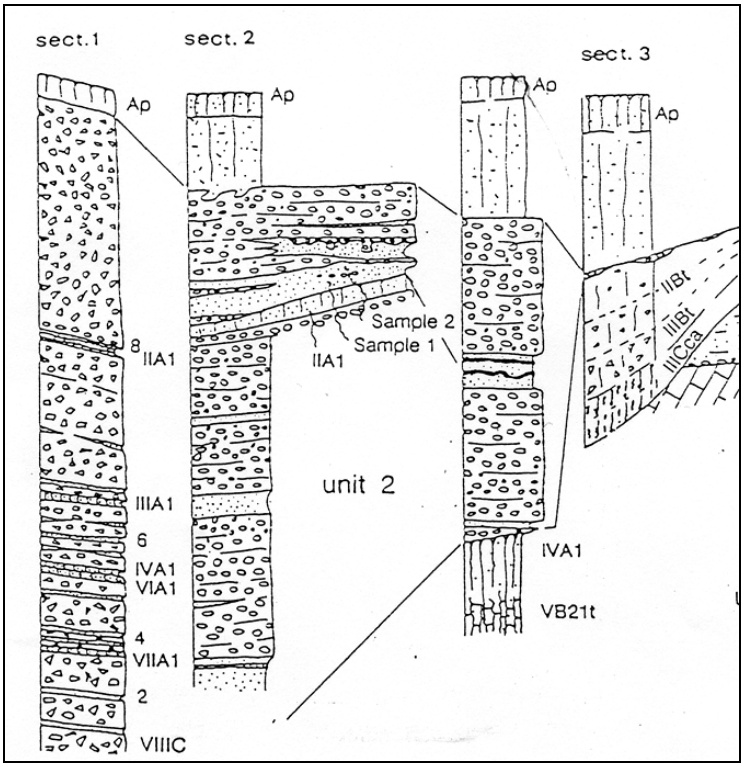

Figure 33 – The sequence along the Mergaoni stream

Above the basal part of the sequence, in Mergaoni Stream as well as in other sections of the Upper Esino valley, Late Paleolithic artifacts have been recognised. Therefore, large part of this terrace was deposited during MIS 3 and MIS 2 (COLTORTI & DRAMIS, 1988). At least 13 phases have been recognised in the Upper Esino River (CALDERONI et al., 1991) (FIG.34) but these events has been confirmed along the Tenna River (CILLA et al., 1996), along many other Marchean Rivers (CALDERONI et al., 1991) as well as along the stream alimented from the Ascensione Mt. (GENTILI et al. 1998). In the Upper esino River, where a detailed analysis have been carried out utilising the Architectural analysis proposed by MIALL (1985; 1996), the older layers (Phase 1) directly overlie the bedrock or on top of paleovalley (ALESSIO et al., 1979) In the region sediments belonging to the MIS 5 and the beginning of MIS 4 are therefore absent. The only exception is made by fluvial sediments outcropping

close to the present day coastline and attributed to the MIS 5e for the paleontological and palethnological content as well as for its morphostratigraphic position (SILVESTRINI et al., 2002). In fact, they are located at slight lower elevation of the MIS 6 terrace.

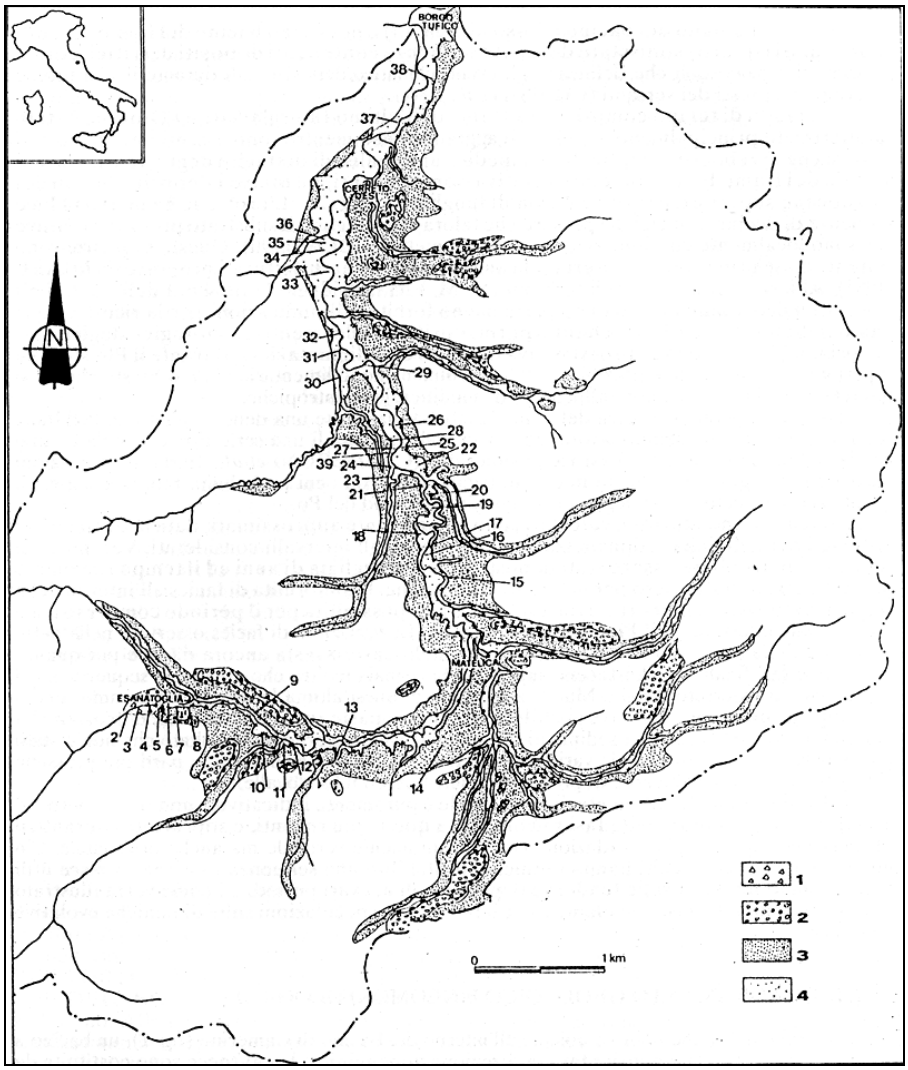

Figure 34 – Location of the studied sections in the Upper Esino valley: 1-travertines; 2-late Middle Pleistocene alluvial deposits; 3-Late Pleistocene alluvial deposits; 4-Holocene alluvial deposits.

The older deposits, that crop out downvalley of Matelica (FIG.35), are coarse gravels (phase 2, 4, 6 and 8) alternated with fine sediments few meters in thickness (phase 3, 5, 7, 9). The medium gravels show horizontal stratification with evident embrications (Gm) rarely with thin sandy (Sh) or silty (Fcf) layers. There are also shallow channels, few meters wide, filled with through cross bedded gravels (Gt). Slightly lager channels have been observed during phase 6.

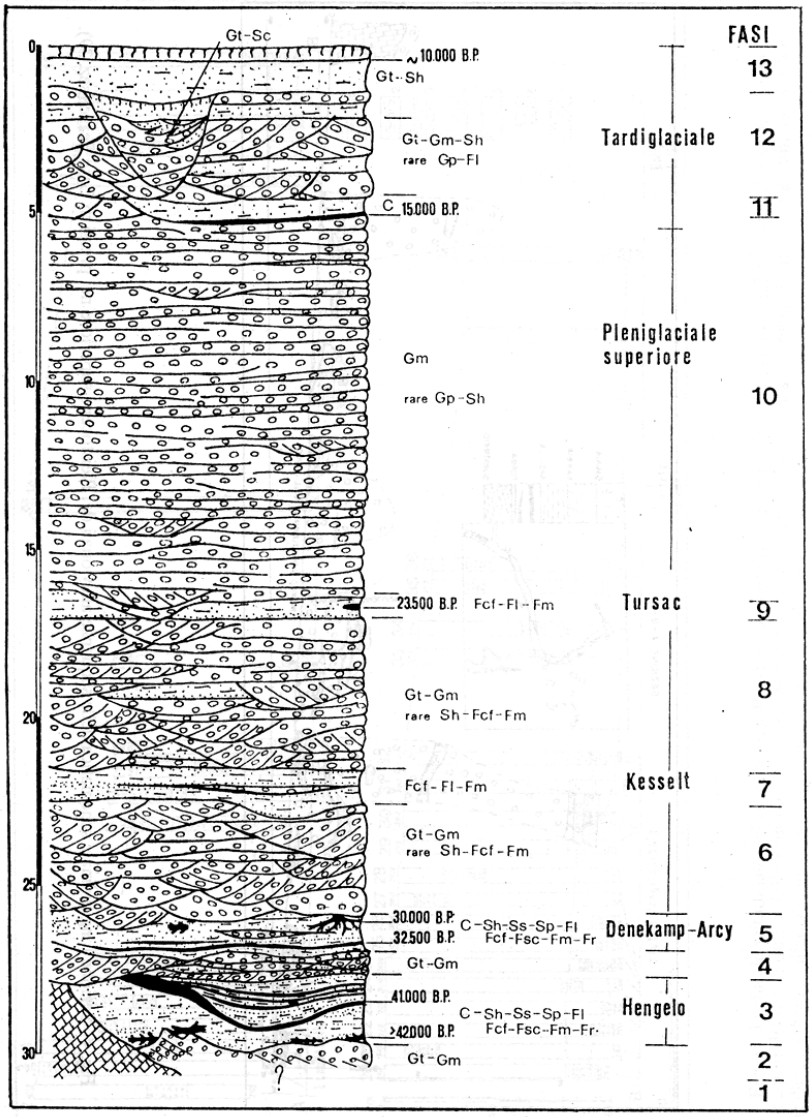

Figure 35 – Sinthetic section of the depositional phases, facies and chronology of the alluvial deposits in the Upper Esino valley.

The lower fine sediments (phase 3 and 5) are massive, due to bioturbation (Fm, Fsc, Fcf), or finely laminated (Fl e Fr). They are frequently interlayered with massive (Sh) or through cross bedded sands (Ss e Sp). Peat layers few centimeter thick, rarely reaching 20 cm, are also present. A large number of vegetion remains are sometimes preserved (leaves, seeds, needles, trunks sometimes in situ, etc.). These facies are frequently affected by load cast. The determination of the botanical content (EVANS, personal comm.) pointed out the presence of *Pinus* sp. (cfr.mugo) and *Betula* sp.(cfr dwarf), with rare erbaceous taxa (*Cyperaceae* e *Gramineae*) that represent a typical Taiga environment. These layers have been dated at ca. 40 ka and between 32 and 29 ka respectively.

The upper fine layers (phase 7 and 9), have a thickness ranging from few centimeter to almost one meters in thickness. They can be locally interrupted or eroded by the erosion of the overlying channels (FIG. 36). They are made by silty and clayey layers (Fl) with sandy intercalations (Sh, Sr). Layer 9 has been dated at ca. 23 ka.

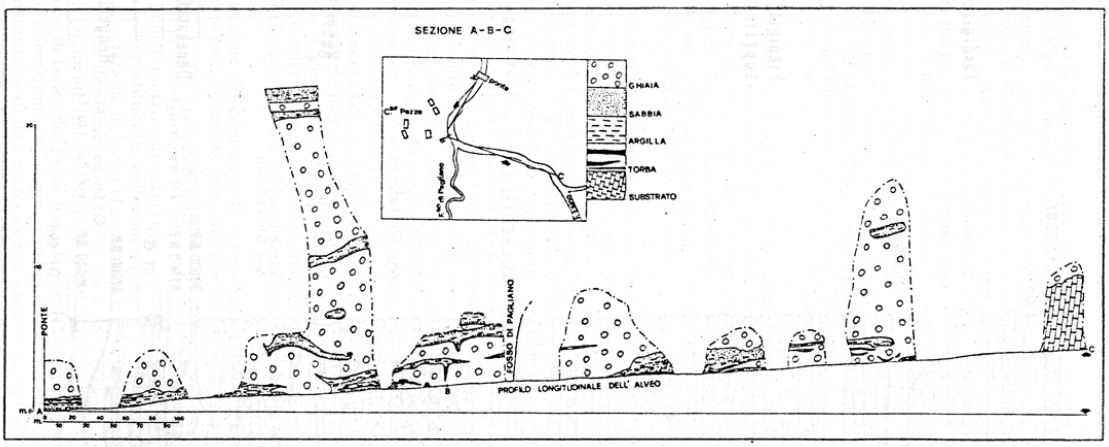

Figure 36 – Longitudinal section of the alluvial deposits recognised close to Case Pezze.

The facies association that characterise the four coarse layers (phase 2, 4, 6 and 8) is dominated by the deposition of gravel bars (GB) inside a large braid plain (Model 2 of MIALL, 1996). The association of the lower fine layers (phase 3 and 5) is dominated by overbank deposits inside a swampy environment (OF). These sediments testify the onset of an anastomosing plain with quite stable channels and depressed areas affected by swampy deposition (Model 8).

On the other hand, the layers of phase 7 and 9 indicate the onset of a sandy plain with small ponds characterised, for a short time, by a very arid environment with very scarce vegetation (Model 12).

Towards the upper part of the sequence (phase 10) the deposition is exclusively made with gravelly facies, up to 8 m in thickness (FIG. 35). These are massive gravels (Gm), and the through cross bedded gravels are very rare (Gt). Almost absent are the lenses of fine sediments (Sh e Fl). This is again due to the onset of a rapidly aggrading gravelly braidplain filled with thin gravel bars (GB) (Model 1).

At about 4-5 metres from the top of the terrace, through cross bedded gravels increase in number, with rare planar cross bedded gravels (Gp, phase 12). Horizontal sands and silty layers (Sh e Fl) become more frequent in placet. There are also some sandy layers with ripples (Sr). This suggest again the transition to a gravelly braidplain (Model 2) similar to the one that characterised the older phases.

In a lateral valley of the Giano River, a tributary of the Esino, at 4-5 metres from the top of the terrace, fine sandy and silty clayey layers have been observed (ALESSIO et al., 1979). They are horizontally layered or massive (Sh, Fl, Fm) with local intercalation of sands with ripples (Ss, Sr, Fr r) and peat. They have been dated at 15.250 ± 160 yr BP and 14.700 ± 150 BP corresponding to a Late Glacial Interstadial. Therefore, in some lateral valley there is again the evolution of an anostomosing architecture with stable channels and small swamps (Phase 11).

Close to the top of the of the terrace through cross bedded gravels (Gt) are interlayered with horizontal sands and silts (Sh, Fcf) that can be more than 1 m thick (Phase 13). Some wide and deep channels have been also observed (FIG. 37) filled at the base with through cross bedded gravels and by colluvial deposits at their top. It is possibile that the sandy silty facies previously described correspond to sporadic overbanks (LA) from these channels.. This is the transition to a more stable and deep channels. The general Pleistocene aggradation seems to have ended during the Late Glacial due to the progressive re-colonization of the slopes by a forest vegetation. This caused an important

downcutting of the river channels in the alluvial plain under a meander course, which slightly cut into the Late Pleistocene alluvial deposits (COLTORTI, 1991; 1997; COLTORTI & FARABOLLINI, 2004).

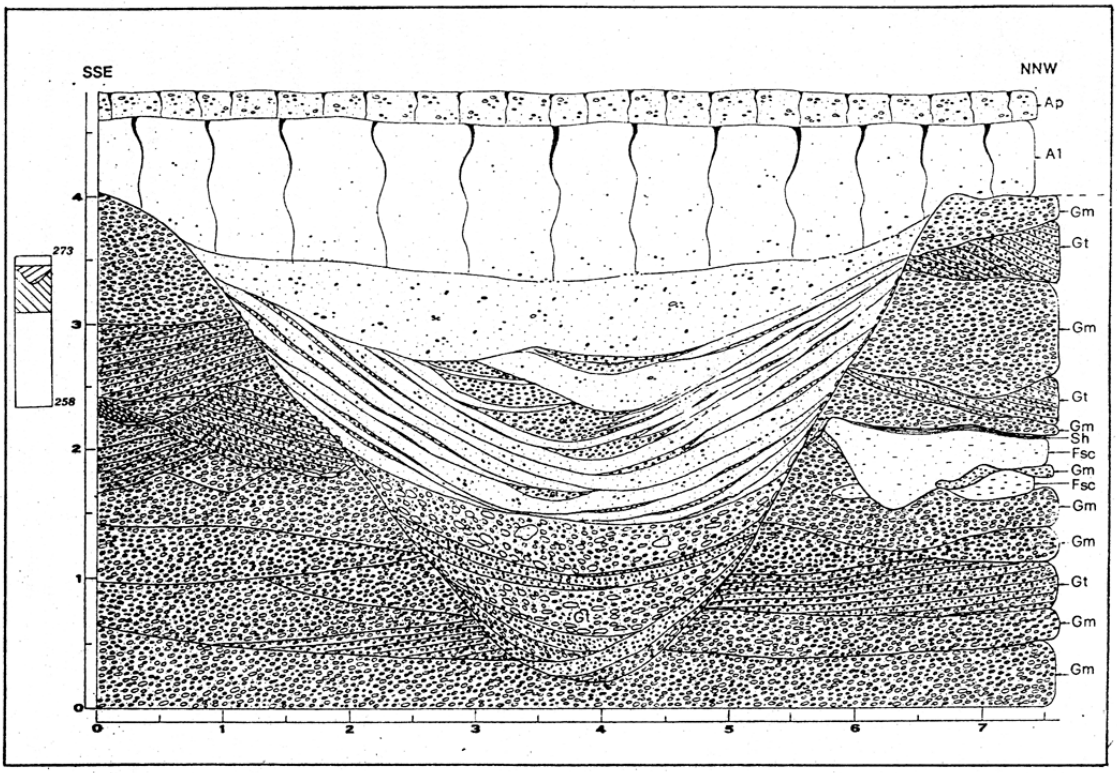

Figure 37 – A section close to the top of the terrace (Cerreto D'Esi).

Unfortunately there are not many datings on the onset of the incision and only along the Chienti, near Tolentino, there are some suggestions that the downcutting already started during the Allerod/Bolling Interstadials. In fact, a gravel carved with an anthropomorphic design, that in the rest of Europe is dated at ca. 11 ka, has been discovered at the top of the fluvial sequenze, inside cut off meander (Massi et al., 2000). Similar events have been recognized in many fluvial systems of central Europe (VANDENBERGHE, 1992; 2002 and ref. therein; MOL, 1997).

The sedimentation rates were evaluated without taking into account the increased width of the alluvial plain moving upward in the sequence. The rates are ca 1mm/yr during the Stadials, 0,5 mm/yr during the Interstadials and 1,5 mm/yr durino the Pleniglacial. A correlation with the events that affected the slopes of the area and recorded in the Ponte di Crispiero sequence has been proposed by COLTORTI & DRAMIS (1995).

7.4 The Holocene fluvial deposition

These sediments costitute the present day thalweg as well as terraces located at higher elevation on the valley floor, below the Late Pleistocene alluvial deposits (COLTORTI, 1991; 1997; CILLA et al., 1996). These sediments occupy large part of the valley close to the coastline where they can be many kilometres wide but landward they become narrow and rarely exceed some hundres metres in width. The Holocene evolution of the Marchean valley has been divided into 5 main phases (FIG. 38).

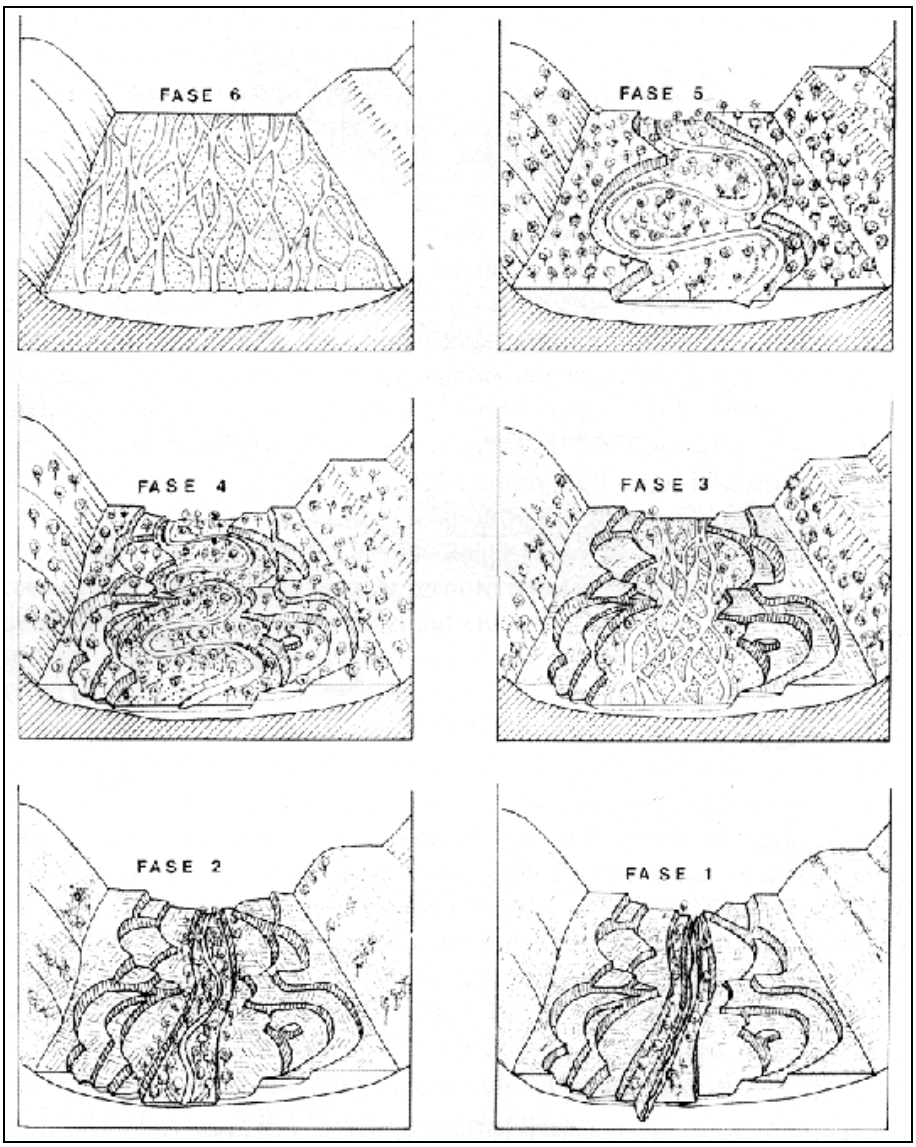

Figure 38 – The different phases of the evolution of the river architectures of the Marchean valleys during the Holocene

The Early Holocene was characterised by the downcutting of the valley under a meander course with large bends in the older phase (Phase 5) and smaller bends in the younger (Phase 4). This was followed by the evolution of an aggrading braid plain (Phase 3) that was later dissected (phase 2) and dissected again (phase 1).

In the inner part of the valley, the Holocene terrace have a limited thickness and sometimes correspond to straath on the bedrock. The evolution of these unpaired terraces was connected with a slow incision made by meander rivers. Along the Musone River, that has been investigated in detail, more than 7 terraces have been described in places. The incision under a meander architecture (phase 5 and 4) can be higher than 40 metres but, some cases are known where it can be ca. 90 metres. The sedimentation is made by through cross bedded gravels and sands that constitute the channel lag and through cross bedded sands and silts due to the evolution of point bars. Sequenze connected with cut off meanders have also been recognised. Charcoals associated with archaeological layers and buried soils in few section of the middle-lower part of the Tenna and Potenza Rivers (CILLA et al., 1996) allow us to establish that the downcutting of some of the older meanders is older than 8

ka BP. In many other alluvial terraces of the Marche and Abruzzi region, referred to the same stratigraphic and morphochronological unit, boreholes and excavations showed the presence of fine-grained sediments, sometimes more than 10m thick (CREMASCHI, 1979; MARABINI ET AL., 1987; FARABOLLINI, 1995; FARABOLLINI et al., 2004). During this period the mouth of the rivers is affected by a slow aggradation and progradation that started ca 11 ka (GORI, 1988; COLTORTI, 1997).In the middle part of the valley the slow incision by meanders, ceased ca 4 ka BP with the onset of a faster downcutting. During this period most of the travertine deposition in the mountain areas ended (CILLA et al.,1996). In correspondence with the valley mouth wide bays were progressively blocked by the evolution of a barrier beach that, between the 7th and the 3rd century BC, isolated large swamps and coastal lagoons. After the 4 ka BP the rivers, expecially in their lower reaches, developed a braided plain. Through cross bedded and horizontal gravels interlayered with channel filled with massive clay and silts. The dating of this braid plain architectural system was made thanks to the presence of many trunks and wood remains, expecially in the clay plug. They gave age between ca 2000 and 100 yr BP. However, some parts of the valley continued to evolve under a meander system up to the Medioeval times and modern times. Since the XI century AD, the lower reaches of the valleys are affected by a very fast aggradation

7.5 - STOP9
The Late Pleistocene fluvial sequence in the Upper Esino River Basin.

The Upper esino River is located inside a depression that extends for many dozens of kilometeres parallel to the Apennine (NO-SE). It is modelled over Miocene arenaceous and clayey layers bounded to the west and to the east by two important limestone ridges (Umbro-Marchean and Marchean Ridge). This part of the Apennine was not affected by an important glacial modelling but the slopes bear witness of intense crioclastic processes that led to the deposition of widespread and thick stratified slope waste deposits (COLTORTI & DRAMIS, 1995).

The Case Pezze Section (FIG.39) is located in the lower part of the Late Pleistocene terrace. The valley incision, that accelerated since the end of the Bronze Age (ca 3500 BP), allow us to study a series of sections cropping out along the river banks (CALDERONI et al., 1991).

This section costitute a summary of the events that have been recognised along the valley. The sequence is characterised by horizontal and through cross bedded gravels associated with the evolution of a rapidly aggradino braidplain with shallow channels. They are interlayered with horizontal sands and clays, sometimes with ripples (Sh, Fm, Fcf, Fr) an peat layers that contains abundant remains of the past vegetation. The lateral continuity of the peat layer indicate an anastomosing channels architecture. COLTORTI & DRAMIS, (1995) suggested that the onset of an anastomosing system was connected with the evolution of a steppe cover on the local slope during the Interstadials that was able to block the support of debris to the valley bottom. The gravel deposition occurred during Stadials. The lowerlying organic layer in this section contains early Late Paleolithic artifacts (Aurignacian) and was excavated from the Soprintendenza Archeologica delle Marche (BROGLIO et al., in press).

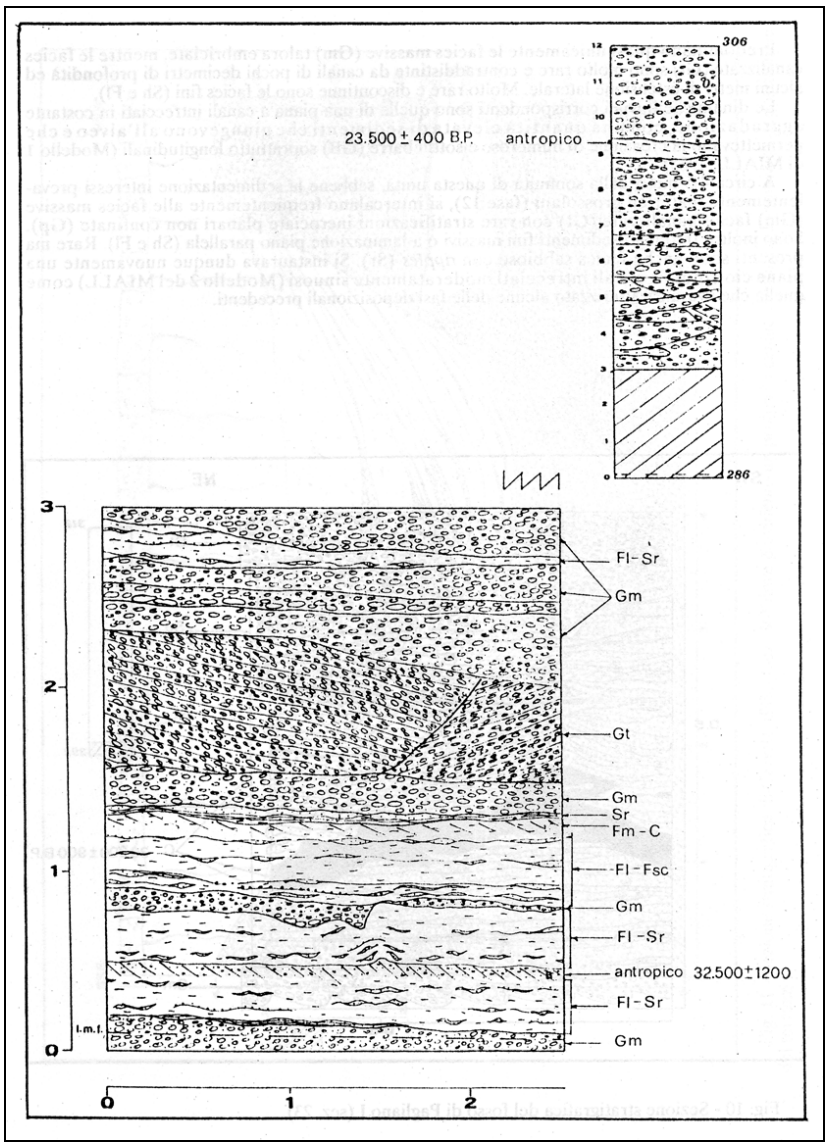

Figure 39 – The section of Case Pezze V.

The layer was dated at 32.500±1200 B.P. and represent (together with the nearby Case Felceto occupation layer dated at 31.800±1100 B.P.) the oldest evidence of *Homo sapiens* occupation in the area. The artifacts include well preserved blades, bladelets and instruments. The diagnostic elements are Dufours bladelets and endscrapers.

The vegetal remains discovered inside these layers belong to trunks of Pinus and Betulla. Some of the trunks reveal ring deformations usually associated with snow overload. The pollen analisys on the archaeological layer recognised the presence of Oak although this species has never been encountered among the wood remains. It is possibile that few individuals of Oak survived inside or close to the limestone Gorges that was affected by minor seasonal contrasts.

In the upper part of the sequenze the gravelly layers become exclusive whilst the fine layers with peat disappear. In this section the deposition of gravels is interrupted by two layers of sands and silts that bear witness of a limited stability of the plain. Charcoals associated with a fireplace have been discovered inside the in the higher layer and gave an age of ca.23 ka BP. Also this interval corresponds to a short Interval in the chronology of the Glacial (DANSGAARD et al., 1993).

7.6 STOP 10
The Late Pleistocene Ponte di Crispiero sequence

The sequence of Ponte di Crispiero is one of the most complete record of the depositional and erosional events that occurred during the Late Pleistocene along the Adriatic side of the Apennine. It has been studied for its stratigraphic significance but also for the presence of a rich Middle Paleolithic industry that was found in the basal layers (COLTORTI et al., 1980; CHIESA et al., 1990; COLTORTI & DRAMIS, 1995). The sequence is located at the top of a of a straath terrace that laterally cut gravels of the MIS 6 (CHIESA et al., 1990). After the Last Interglacial, this part of the valley was not affected by fluvial dynamics and record the various climatic changes that modelled the local slope during the Late Pleistocene.

At Ponte di Crispiero (FIG. 40), the base of the sequence is made by the remains of the strongly trunkated Last Interglacial soils, resting on the straath terrace. The Bts horizon, lying over a thin B/C horizon, is completely leached and enriched of residual elements like flint, clay and hydroxides of Fe.

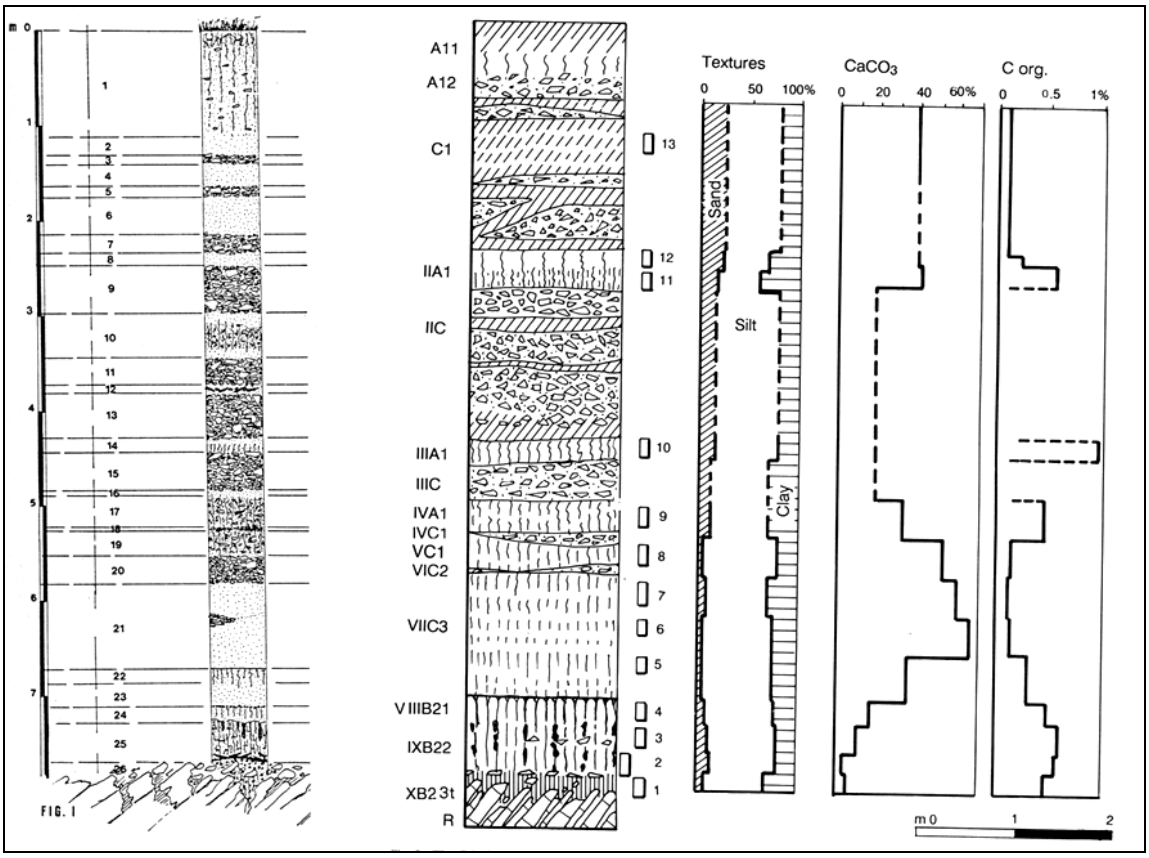

Figure 40 – The Ponte di Crispiero sequence.

A very rich Paleolithic archaeological layer has been discovered on the stone line that truncate the Bts horizon (COLTORTI et al., 1980) and that represent a period of slope erosion after the end of the Last Interglacial. The industry is characterised by the abundance of levallois flakes and toos, with an estreme abundance of side scrapers made on levallois flakes. The musterian point are also well represented. It is a Musterian industry of levallois technique.

A luvisol is present on top of the stone line and it is characterised by a brownish Bt horizon. It testify that, after the slope erosion erosion, a series of colluvial deposits were deposited at the feet of the slope and later weathered

under warm and humid conditions It is a soil similar to that recognised in the Fosso Mergaoni Sequence (FIG.33). Another horizon lie on top of the previous one. It is fairly leached but the presenco of CaCo3 concretions (pseudomiceli and nodules) indicate a stron seasonal contrast and the onset of more arid conditions. The lowermost soil has been attributed to the MIS 5e, and the overlying ones to the MIS 5c and 5e rispectively. Higher in the sequence there is an A1Ca horizon that represent another short living Interstadial at the beginning of the Last Glaciation (Odderadde, OIS 4; DANSGAARD et al., 1993).

A thick layer of loess is deposited on top of the previously decribed sequence. It is the thicker loess cover described in the Marchean region. It reveals that the MIS4 was characterised by a strong aridity, higher that the one experienced during the Pleniglacial (MIS2). The soil that develops on top of the aeolian deposits has a A1Ca profile, and it is characterised by this vein of soft carbonate concretions. It is a steppe soil (Cernozem) that indicate a climatic amelioration and the stability of the slope that was previously affected by deflaction and loess deposition. The dating of the organic matter gave an age of 36890 ± 170 Bp (Beta 194854) suggesting that the steppe evolved during the earlier part of MIS3. This layer was also dated with U/Th on the soft carbonate concretions and gave an age of ca. 48 ka (COLTORTI & CREMASCHI, 1990) although the dating was not realible as suggested by the U238/U234 ratio. Few rare blades coming from this layer indicate a human frequentation of the area.

This soil is buried by slope debris that indicate a climatic worsening and the onset of colder climatic conditions. However, the debris deposition is not able to bury all the feet of the slope and locally the evolution of the soil continue to evolve and later, during another climatic amelioration colonise also the debris layer. The dating on this second soil gave an age of 29780 ± 360 Bp (Beta 194854). After a new debris layer another steppe soil colonise again the slope. This soil was dated at 24870 ± 200 (Beta 194853).

The deposition of debris become almost exclusive upwards although local silty layers are present in the sequence, most probably coming from the colluviation of aeolian deposits. A Late Paleolithic industry made with very long blades, but no instruments, was found inside the debris layers.

In the upper èpart of the sequence a series of small and deep gully have been recognised. They testify a climatic episode but its significance is still unsolved. Few blades and bladelets belongig to the Late Paleolithic come also form the surface of the deposits that is weathered by a AbwC profile due to the soil developed since the beginning of the Holocene.

7.7 - STOP 11
Smorlesi Quarry: the Holocene deposits in the Potenza River.

In the middle part of the Potenza river, between Villa Potenza and Fontenoce (FIG.41) some fluvial terraces hanging at progressive elevation over the river valley floor are present. (CALDERONI et al., 1996; CILLA et al., 1996).

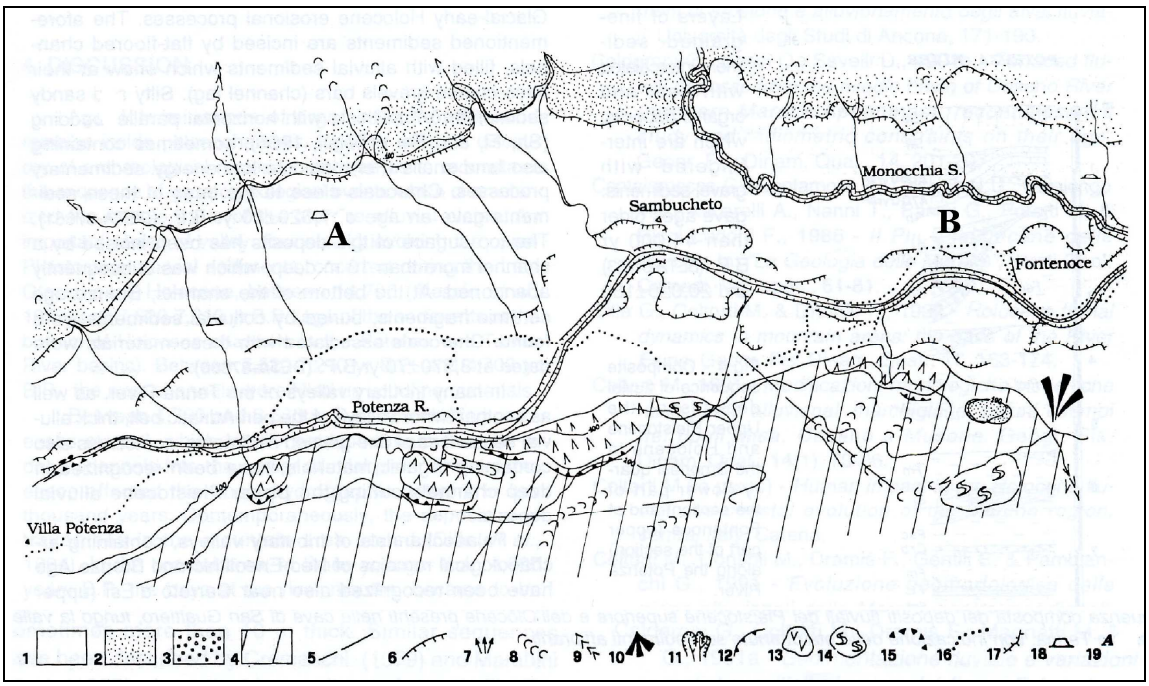

Fig.41 – Geomorphological scheme of the middle sector of the Potenza river valley. 1- Present-day and terraced alluvial deposits (Holocene); 2- Terraced alluvial deposits of Late Pleistocene; 3- Terraced alluvial deposits of Late Middle Pleistocene; 4- Siltites and sands with intercalated arenaceous and conglomeratic bodies (Pliocene – Early Pleistocene); 5- Faults; 6- Fluvial erosion escarpment; 7- Intense lateral erosion; 8- Through-floored small valley; 9- V-shaped small valley; 10- Alluvial fan; 11- Badlands; 12- Landslide scarp edge; 13- Landslide scree-tongue; 14- Plastic deformations; 15- Artificial levees; 16- Artificially cut channel; 17- Artificial embankments; 18- Quarry; 19- A- Smorlesi quarry, B- Fontenoce section.

A sequence more than 12m thick, mostly made of fine sediments outcrops at Smorlesi quarry (Fig.42). Massive and cross-bedded sands (*Sh* and *St facies* of MIALL, 1985) sometimes associated with shallow channels filled with gravels (*Gt facies*), are interlayered with massive clays and silts, sometimes with land snails (*Fm*, *Fl*, *Fsc*, *Fcf*). At 11,5m from the surface, fireplace charcoals with no archaeologifcal remains, gave an age of 7,210 ± 90 yr B.P. At the top of the sequence, within sandy layers (*Sh facies* of MIALL, 1985), Eneolithic ceramic fragment were found (FARABOLLINI, 1995; CALDERONI et al., 1996; CILLA et al., 1996).

Also a few km downstream, near Fontenoce (Fig.4) archaeological excavations showed Eneolithic layers (Fig.5) dated to 4,680 Yr B.P. and 4,700 yr B.P. (SILVESTRINI et al., 1992-3). These layers were separated by a few decimeter thick massive overflood sands and silts (*Sh* and *Fsc facies*).

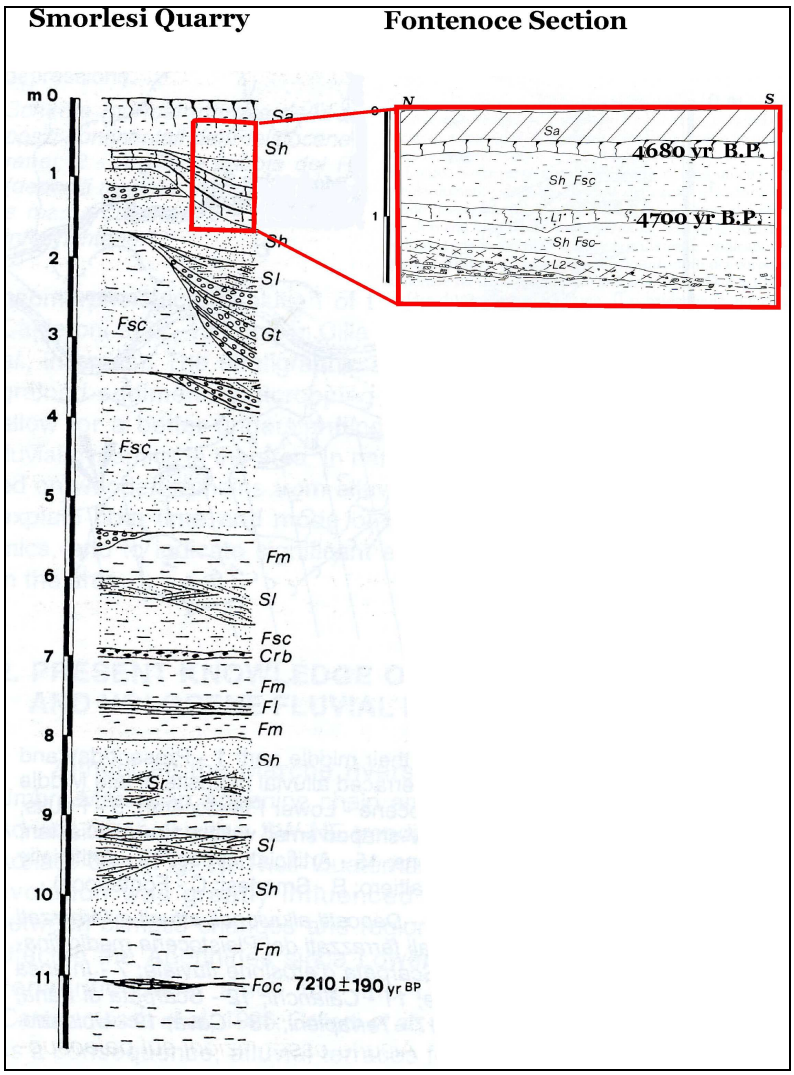

Fig.42 - Composite sequence of fluvial deposits of the Holocene in the Smorlesi quarry (left part) and at Fontenoce (right part) along the Potenza river.

7.8. The missing stop: the Tenna river valley

Along the Tenna river valley, between Servigliano and Molino di Monte San Martino (Fig.43) several quarries, open in the recent alluvial deposits, few meters above the present-day thalweg, allows the reconstruction of the events that characterize the fluvial dynamics during the Late Holocene. Layers of fine-grained sediments interbedded with peat and organic deposits, which are interfingered with gravelly sediments (Fig.44), gave ages older than 44,000 yr B.P. (BETA 87663) and 20,020±150 yr B.P. (BETA 87662) (CILLA et al., 1996).

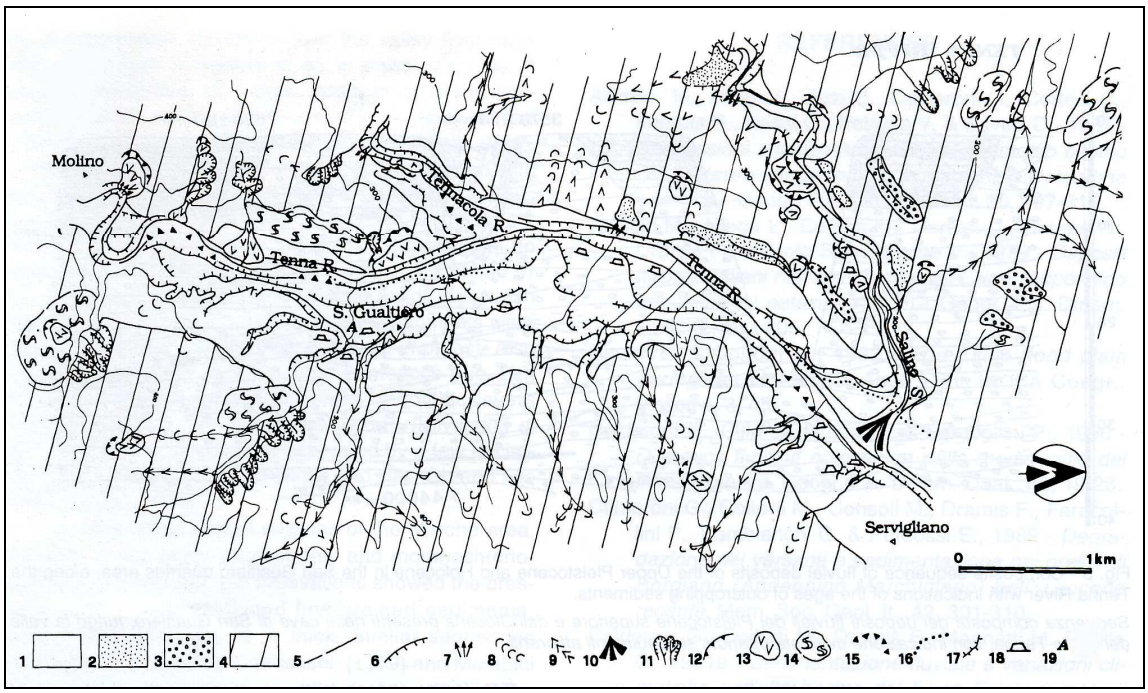

Fig.43 – Geomorphological scheme of the middle sector of the Tenna river valley. 1- Present-day and terraced alluvial deposits (Holocene); 2- Terraced alluvial deposits of late Pleistocene; 3- Terraced alluvial deposits of late Middle Pleistocene; 4- Siltites and sands with interbedded arenaceous and conglomeratic bodies (Pliocene – Early Pleistocene); 5- Faults; 6- Fluvial erosion escarpment; 7- Intense lateral erosion; 8- Through-floored small valley; 9- V-shaped small valley; 10- Alluvial fan; 11- Badlands; 12- Landslide scarp edge; 13- Landslide scree-tongue; 14- Plastic deformations; 15- Artificial levees; 16- Artificially cut channel; 17- Artificial embankments; 18- Quarry; 19- Study area (A = San Gualtiero).

In these sections, the thickness of Pleniglacial gravel deposits is reduced, due to Late Glacial – Early Holocene erosional processes. The afore-mentioned sediments are incised by flat-floored channels, filled with alluvial sediments which show at their base coarse gravels bars (channel lag). Silty and sandy sediments, 6-7 meters thick, with horizontal parallel bedding (*Sh, Fl* and *Fm facies* of MIALL, 1985), sometimes containing land snails (*Fcf facies*), indicate low energy sedimentary processes. Charcoals close to the base of these sediments gave an age of 7,620±80 yr B.P. (BETA 87661) (CILLA et al., 1996). The top surface of these deposits have been incised by a channel more than 10 m deep, which was subsequently abandoned. At the bottom of the channel, Bronze age pottery fragments, buried by colluvial sediments were found (Fig.2). Charcoals associated with these materials were dated at 3,570 yr B.P. (BETA 87660) (CILLA et al., 1996).

These ages testify a progressive deepening of the alluvial plain made by a single and sinuous channel up to 20 m higher than present between Late Glacial and 8,000 yr B.P.; processes of slow aggradation followed.

After the IV millennium B.P. the deposition of fine sediments ended and the rapid deepening of the thalweg was recorded. These modifications could be attributed to climatic changes as well as to the result of human impact following the widespread occupation of slopes for agriculture and farming purposes (CALDERONI et al., 1989; CILLA et al., 1996; COLTORTI, 1997; COLTORTI & FARABOLLINI, 2004).

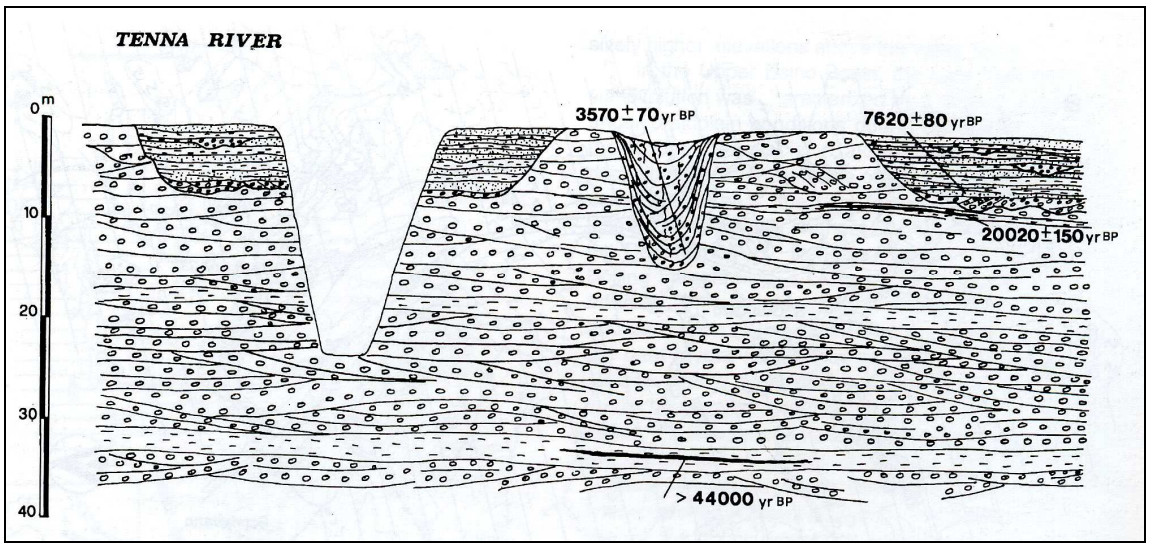

Fig.44 – Composite sequence of fluvial deposits along the late Pleistocene and Holocene in the several quarries along the middle sector of the Tenna river with indications of the ages of outcropping sediments.

REFERENCES

ABBATE, E., BALESTRIERI, M.L., BIGAZZI, G., NORELLI, P. & QUERCIOLI, C., 1994. Fission tracks datings and recent rapid denudation in Northern Apennines, Italy. Memorie Società Geologica Italiana, v. 48, p. 579-585.

ABBATE E., BALESTRIERI M.L., BIGAZZI G., VENTURA B., ZATTIN M. & ZUFFA G.G., 1999. An extensive Apatite fission-track study throughout the Northern Apennines nappe belt. Radiation measurements, 31, 673-676.

ABBAZZI L., ALBIANELLI A, AMBROSETTI P., ARGENTI P., BASILICI G., BERTINI A., GENTILI S., MASINI F., NAPOLEONE G. & PONTINI M.R., 1997. Paleontological and sedimentological recors in Pliocene distal alluvial fan deposits at Cava Toppetti (Todi, Central Italy). Bollettino della Società Paleontologica Italiana, 36, 5-22.

ALBANI A., 1962. L'antico Lago Tiberino. L'Universo. Rivista Istituto Geografico Militare 42, 4, 731-750.

ALESSIO M., ALLEGRI L., COLTORTI M., CORTESI C., DEIANA G., DRAMIS F., IMPROTA S. & PETRONE V., 1979. Depositi tardowurmiani nell'alto bacino dell'Esino (Appennino marchigiano) datazione al 14C. Geografia Fisica e Dinamica Quaternaria, 2, 203-205.

AMATO, A., CINQUE, A., SANTANGELO, N. & SANTO, A. 1992. Il bordo meridionale del Monte Marzano e la Valle del F. Bianco: geologia e geomorfologia. Studi Geologici Camerti, Volume Speciale 1, 191-200.

AMBROSETTI, P., CARBONI, M.G., CONTI, M.A., COSTANTINI, A., ESU, D., GANDIN, A., GIROTTI, O., LAZZAROTTO, A., MAZZANTI, R., NICOSIA, U., PARISI, G. & SANDRELLI F., 1978. Evoluzione paleogeografica e tettonica nei bacini tosco-umbro-laziali nel Pliocene e nel Pleistocene inferiore. Memorie Società Geologica Italiana, v. 19, p. 573-580.

AMBROSETTI P., CARRRARO F., DEIANA G. & DRAMIS F., 1982. Il sollevamento dell'Italia Centrale tra il Pleistocene Inferiore e il Pleistocene Medio. Contributi Conclusivi per la realizzazione della Carta Neotettonica d'Italia (II), CNR P.F. "Geodinamica", S:P: "Neotettonica", 356, 1341-1343.

AMBROSETTI P., BARBIERI M., BASILICI G., BOZZANO F., DE PARI P., DI FILIPPO M., DI MAIO R., DUDDRIDGE G., ETIOPE G., GAMBINO P., GRAINGER P., LOMBARDI S., MOTTANA A., PATELLA D., PENNACCHIONI E., RUSPANDINI T., SCARASCIA MUGNOZZA G., SORDONI G., TAZIOLI S., TORO B., VALENTINI G & ZUPPI G., 1993. Analysis of geoenvironmental conditions as morphological evolution factors of the sand-clay series of the tiberino valley and dunarobba forest preservation. Proceedings Progress Meeting in the Mirage Project, 3rd phase, Bruxelles, 7-8 October 1993, 17 pp.

AMBROSETTI P., BASILICI G., CIANGHEROTTI A.D., CODIPIETRO G., CORONA E., ESU D., GIROTTI O., LO MONACO A., MENEGHINI M., PAGANELLI A. & ROMAGNOLI M., 1995a. La foresta fossile di Dunarobba (Terni, Umbria, Italia Centrale): contesto litostratigrafico, sedimentologico, palinologico, dendrocronologico e paleomalacologico. Il Quaternario (8)2, 465-508.

AMBROSETTI, P., BASILICI, G., CAPASSO BARBATO, L., CARBONI, M.G., DI STEFANO, G., ESU, D., GLIOZZI, E., PETRONIO, C., SARDELLA, R., & SQUAZZINI, E., 1995b. Il Pleistocene inferiore nel ramo sud occidentale del Bacino Tiberino (Umbria): aspetti litostratigrafici e biostratigrafici. Il Quaternario, v. 8 (1), p. 19-36.

ARGENTI P., 2003. Plio-Quaternary mammal faunas in Umbria (Central Italy). Geologica Romana, 37: 1-18, Roma.

ARGNANI, A. & GAMBERI, F., 1996. Stili strutturali al fronte della catena appenninica nell'Adriatico centro-settentrionale. Studi Geologici Camerti, Volume Speciale 1995/1, p. 19-27.

BALESTRIERI M.L., BERNET M.T., BRANDON M.T., PICOTTI V., REINERS P. & ZATTIN M., 2003. Pliocene and Pleistocene exhumation and uplift of two key areas of the Northern Apennines. Quaternary International, 101-102, 67-73.

BALLY A.W., BURBI L., COOPER C. & GHELARDONI P., 1986. Balanced sections and seismic reflection profiles across the Central Apennines. Memorie della Società Geologica Italiana, 35, 257-310.

BARBERI F., BUONASORTE G., CIONI R., FIORDELISI A., FORESI L., IACCARINO S., LAURENZI M.A., SBRANA A., VERNIA L. & VILLA I.M., 1993. Plio-Pleistocene geological evolution of the geothermal area of Tuscany and Latium. Memorie Descrittive della Carta Geologica d'Italia, vol.XLIX, 77-134.

BARCHI, M.R., DE FEYTER, A., MAGNANI, M.B., MINELLI, G., PIALLI, G. & SOTERA, M., 1998. Extensional tectonics in the Northern Apennines (Italy): evidence from the CROP 03 deep seismic reflection line. Memorie Società Geologica Italiana, v. 52, p. 527-538.

BARTOLINI, C., 1980. Su alcune superfici sommitali dell'Appennino Settentrionale (Provincie di Lucca e Pistoia). Geografia Fisica Dinamica Quaternaria, 3, 42-60.

BASILICI G. , 1997. Sedimentary facies in an extensional and deep-lacustrine depositional system:the Plioceen Tiberino Basin, Central Italy. Sedimentary Geology, 109, 73-94.

BERGREEN W.A., HILGEN F.J., LANGEREIS C.G., KENT D.V., OBRADOVICH J.D., RAFFI I., RAYMO M.E. & SHAKLETON N.J., 1995a. Late Neogene (Pliocene-Pleistocene) Chronology: New perspectives in High Resolution Stratigraphy. GSA Bulletin, 107: 1272-1287.

BERGREEN W.A., KENT D.V., SWISHER III C.C. & AUBRY M.P., 1995b. A Revised Cenozoic Geochronology and Chronostratigraphy. In: Bergreen W.A., Kent D.V. and Hardenbol J. (Eds.), Geochronology, Time Scales and Global Stratigraphic Correlations: A Unified Temporal Framework for an Historical Geology. SEPM Special Volume 54.

BERNINI, M., CLERICI, A., PAPANI, G. & SGAVETTI, M., 1977. Analisi della distribuzione planoaltimetrica delle paleosuperfici dell'Appennino emiliano occidentale. L'Ateneo Parmense, Acta Naturalia, 13, 145-156.

BERNINI M. & PAPANI G., 2002. La distensione della fossa ettonica della Lunigiana nord-occidentale. Bollettino della Società Geologica Italiana, 121, 313-341.

BERTOLDI R., 1977. Studio palinologico della serie di Le Castella. Accademia Nazionale dei Lincei, Rendiconti, s.VIII, 62, 547-555.

BIGI, S., CALAMITA, F., CENTAMORE, E., DEIANA, G., RIDOLFI, M. & SALVUCCI, R., 1991. Assetto strutturale e cronologia della deformazione della "zona di incontro" tra le aree umbro-marchigiana e laziale-abruzzese (Marche meridionali e Lazio-Abruzzo settentrionale). Studi Geologici Camerti, Volume Speciale Crop 11, p. 21-26.

BIGI, S., CALAMITA, F., CELLO, G., CENTAMORE, E., DEIANA, G., PALTRINIERI, W. & RIDOLFI, M., 1996. The external Marchean-Abruzzi messinian foredeep in the frame of the evolution of the Central Apennines. Notes et Memoires du Service Geologique du Maroc, v. 387, p. 157-161.

BOCCALETTI M., CALAMITA F., CENTAMORE E., CHIOCCHINI U., DEIANA G., MORETTI G., MICARELLIA. & POTETTI M., 1986. Evoluzione dell'Appennino Tosco-Umbro-Marchigiano durante il Neogene. Giornale di Geologia, 48/1-2, 227-233. Bologna.

BOCCALETTI, M., BONINI, M., MORATTI, G. & SANI, F., 1995. Le fasi compressive neogenico-quaternarie nell'Appennino settentrionale: relazioni con l'evoluzione dei bacini interni e con la tettonica del basamento. Studi GeologiciCamerti, Volume Speciale 1, p. 51-72.

BOCCALETTI, M., BONINI, M., MORATTI, G. & SANI, F., 1999. Compressive Neogene-Quaternary tectonics in the Hinterland area of the Northern Apennines. Journal of Petroleum Geology, v. 22 (1), p. 37-60.

BOCCHINI A. & COLTORTI M., 1990. Il complesso carsico Grotta del Fiume - Grotta Grande del Vento e l'evoluzione geomorfologica della gola di Frasassi (Appennino umbro-marchigiano). Atti Conv.Naz. sul carsismo della Gola di Frasassi, 24-25/9/88, Memorie Istituto Italiano di Speleologia, 4, s.II, 155-180, Castellana Grotte.

BONCIO, P., BROZZETTI, F., PONZIANI, F., BARCHI, M., LAVECCHIA, G. & PIALLI, G., 1998. Seismicity and extensional tectonics in the Northern Umbria-Marche Apennines. Memorie Società Geologica Italiana, v. 52, p. 539-555.

BONINI M., 1998. Chronology of deformation and analogue modelling of the Plio-Pleistocene 'Tiber Basin': implcations for the evolution of the Northern Apennine (Italy). Tectonophysics, 285, 147-165.

BOSI C., COLTORTI M. & DRAMIS F., 1987. Carta del quaternario del Bacino di Gualdo Tadino. Tip. Errebi, Falconara.

BOSI, C., CAIAZZO, C., CINQUE, A., & MESSINA, P., 1996. Le superfici relitte della regione fucense (Appennino centrale) e loro possibile significato nella ricostruzione dell'evoluzione geologica. Il Quaternario, v. 9, p.381-386.

BROGLIO A., COLTORTI M., PERESANI M. & SILVESTRINI M., in press. Il Paleolitico delle Marche. Atti XXXVIII Riunione Istituto Italiano di Preistoria e Protostoria.

BROZZETTI F., 1995. Stile strutturale della tettonica distensiva nell'Umbria occidentale: l' esempio dei Massicci mesozoici perugini. Studi Geologici Camerti, volume speciale, 1, 105-120.

CALAMITA F. & DEIANA G., 1988. The arcuate shape of the Umbria-Marche-Sabina Apennines (Central Italy). Tectonophysics, 146, 139-147.

CALAMITA, F., CELLO, G., CENTAMORE, E., DEIANA, G., MICARELLI, A., PALTRINIERI, W. & RIDOLFI, M., 1991. Stile deformativo e cronologia della deformazione lungo tre sezioni bilanciate dall'Appennino Umbro-Marchigiano alla costa adriatica. In: Studi geologici preliminari all'acquisizione dati del profilo CROP 03 Punta Ala-Gabicce, Studi GeologiciCamerti, Volume Speciale, p.295-314.

CALAMITA F., COLTORTI M., FARABOLLINI P. & PIZZI A., 1994a. Le faglie normali quaternarie nella dorsale appenninica Umbro-Marchigiana: proposta di un modello tettonico di inversione. Studi Geologici Camerti Volume Speciale 1, 211-225.

CALAMITA F., CELLO G., DEIANA G. & PALTRINIERI W., 1994b. Structural styles, cronology-rates of deformation and time space relationships in the Umbria-Marche thrust system (Central Apennines, Italy). Tectonics, 13, 873-881.

CALAMITA F. & PIERANTONI P.P., 1995. Modalità della strutturazione neogenica nell'Appennino umbro-sabino (Italia centrale). Studi Geologici Camerti, v.s., 1995(1), 153-170.

CALAMITA F., COLTORTI M., PIERANTONI P.P., PIZZI A., TURCO E. & SCISCIANI V., 1998. Relazione tra le faglie quaternarie e la sismicità nella Dorsale Appenninica Umbro-Marchigiana: l'area di Colfiorito. Studi Geologici Camerti. XIV, 177-191, Camerino

CALAMITA, F., COLTORTI, M., PIERUCCINI, P. & PIZZI, A. (1999) - Evoluzione strutturale e morfogenesi plio-quaternaria dell'appennino umbro-marchigiano tra il pedappennino umbro e la costa adriatica. Bollettino Società Geologica Italiana, v.118, p.125-139.

CALAMITA F., COLTORTI M., PICCININI D., PIERANTONI P.P., PIZZI A., RIPEPE M., SCISCIANI V. & TURCO E.. 2000. Quaternary faults and seismicity in the Umbria-Marche Apennines (Central Italy): evidence from the 1997 Colfiorito earthquake. Journal of Geodynamics, 1-20, Elseveir, Amsterdam.

CALDERONI G., COLTORTI M., CONSOLI M., DRAMIS F., FARABOLLINI P., PAMBIANCHI G. & PERCOSSI S., 1989. Degradazione dei versanti e sedimentazione nei pressi di Borgiano (Appennino Marchigiano) nell'Olocene recente. Memorie della Società Geologica Italiana, 42, 301-310.

CALDERONI G., COLTORTI M., DRAMIS F., MAGNATTI M. & CILLA G., 1991. Sedimentazione fluviale e variazioni climatiche nell'alto bacino dell'Esino durante il Pleistocene superiore. In: "Fenomeni di erosione e alluvionamento degli alvei fluviali", Ancona, 1991, 171-190.

CALDERONI G., NESCI O. & SAVELLI D., 1991. Terrace fluvial deposits from the middle basin of Cesano River (northern Marche Apennines): reconisseance study and radiometric costraints on their age. Geografia Fisica e Dinamica Quaternaria, 14, 201-207.

CALDERONI G., CILLA G., DRAMIS F. & FARABOLLINI P., 1996. Dinamica fluviale olocenica nella media valle del fiume Potenza. Geografia Fisica e Dinamica Quaternaria, 19, 19-28.

CANTALAMESSA, G., CENTAMORE, E., CHIOCCHINI, U., COLALONGO, M.L., MICARELLI, A., NANNI, T., PASINI, G., POTETTI, M. & RICCI-LUCCHI, F., 1986. Il Plio-Pleistocene delle Marche. In:Centamore, E. & Deiana, G. eds.: La geologia delle Marche, Studi Geologici Camerti, Volume Speciale, p.61-81.

CARMIGNANI L., DECANDIA A:, FANTOZZI P., LAZZAROTTO A., LIOTTA D. & MECCHERI M., 1994. Tertiary extension in Tuscany (Northern Apennines, Italy). Tectonophysics, 238, 295-315.

CATTUTO C., 1976. Correlazioni tra piani carsici ipogei e terrazzi fluviali nella valle del fiume Esino (Marche). Bollettino della Società Geologica Italiana, 95, 313-326

CATTUTO C., CONTI M., NICOSIA U. & PARISI G., 1979. Relazione sulla neotettonica dei Fogli 130,131,137,138. Contr. prelim. alla real. della Carta Neotettonica d'Italia Progetto Finalizzato "Geodinamica"/Sottoprog. "Neotettonica" (C.N.R.), 13-28.

CATTUTO C. & GREGORI L., 1986. Elementi geomorfologici ed episodi di tettonica recente nei dintorni di Spoleto. Boll. Soc. Geol. It., 105, 173-184.

CATTUTO C., CENCETTI C. & GREGORI L., 1992. Il Plio-Pleistocene nell'area medio-alta del bacino del F.Tevere: possibile modello morfotettonico. Studi Geologici Camerti Volume Speciale/1, 103-109.

CENCETTI C., 1988. Evoluzione del reticolo idrografico in un tratto umbro-marchigiano dello spartiacque principale dell'Appennino. Geografia Fisica e Dinamica Quaternaria, 11, 11-24.

CENTAMORE E,, DEIANA G., DRAMIS F. & PIERUCCINI U., 1978. Guida alle escursioni nelle aree di Costacciaro-Gualdo Tadino e di Colfiorito (Appennino Umbro-Marchigiano). P.F. 'Geodinamica' C.N.R. , 181, 31-35.

CENTAMORE E, CHIOCCHINI M., CHOCCHINI U., DRAMIS F., GIARDINI G., IACOBACCI A., MARTELLI G., MICARELLI A. & POTETTI M., 1979. Carta Geologica d'Italia alla scala 1:50.000. F° 301 Fabriano. Servizio Geologico d'Italia, pp.51.

CENTAMORE E. & DEIANA G., 1986. La Geologia delle Marche. Studi Geologici Camerti, Volume Speciale, 145 pp.

CHIESA S., COLTORTI M., CREMASCHI M., FERRARIS M. & PROSPERI L., 1990 - Loess sedimentation and quaternary deposits in the Marche Province. In Cremaschi M. Ed."The Loess in Northern and Central Italy: a loess basin between the Alps and the Mediterranean regions". Quaderni di Geodinamica Alpina e Quaternaria, 103-130.

CIANGHEROTTI A. & ESU D., 2000. Paleoecologic and biochronologic meaning of the Early Pleistocene molluscan fauna from Anghiari basin (Tiber River Upper Valley, Central Italy). Bollettino della Società Paleontologica Italiana, 39: 217-224, Modena.

CILLA G., COLTORTI M., DRAMIS F., FARABOLLINI P., GENTILI B. & PAMBIANCHI G., 1996. Fluvial sedimentation during the Early Holocene in the Marchean valleys. Il Quaternario, 9(2), 459-464.

CINQUE, A., 1992. Distribuzione spazio-temporale dei movimenti tettonici verticali nell'Appennino Campano-Lucano: alcune riflessioni. Studi Geologici Camerti, Volume Speciale 1, 33-38

COLALONGO M.L., NANNI T. & RICCI LUCCHI F., 1979. Sedimentazione ciclica nel Pleistocene anconetano. Geologica Romana, 18.

COLTORTI M., 1981. Lo stato attuale delle conoscenze sul Pleistocene e il Paleolitico Inferiore e Medio della regione marchigiana. Atti del I Convegno dei Beni Culturali-Ambientali delle Marche, Numana, 8-10.5.81, Paleane Ed., 63-122.

COLTORTI M., 1996. Il Quaternario continentale e l'evoluzione del paesaggio della regione marchigiana. In: La preistoria e studi preliminari di preistoria nella provincia di Ascoli Piceno. Lab. Didattico di Ecologia del Quaternario, Cupra Marittima, 321-368.

COLTORTI M., 1997. Human impact in the Holocene fluvial and coastal evolution of the marche region, Central Italy. Catena, 30, 311-335.

COLTORTI M., CREMASCHI M. & PERETTO C. (1980) - Industria di facies levallois a Ponte di Crispiero (Marche). Atti XXIII Riunione Scientifica I.I.P.P., Firenze 7-9/5/80, 415-427, Firenze..

COLTORTI M. & NANNI T., 1983. Hydrology of the lower Esino basin, Central Italy. Papers Int. Conf. on Groundwater and Man, 3: 11-20.

COLTORTI M. & DRAMIS F., 1988. The significance of stratified slope waste deposits in the quaternary Umbria Marche Apennines (Central Italy). Zeit.fur Geomorphologie. N.F., suppl. Bol. 71, 59-70, Stuttgart.

COLTORTI M. & CREMASCHI M., 1990. Ponte di Crispiero: a type-sequence for the Upper Pleistocene of Central Italy, Adriatic side. In Riassunti 75° Congresso Società Geologica Italiana, 53.

COLTORTI M., CONSOLI M., DRAMIS F., GENTILI B. & PAMBIANCHI G., 1991. Evoluzione geomorfologica delle piane alluvionali delle Marche centro-meridionali. Geografia Fisica e Dinamica Quaternaria, 14(1), 87-100.

COLTORTI M & DRAMIS F., 1995. The chronology of Upper Pleistocene stratified slope-waste deposits in Central Italy. Permafrost and Periglacial Processes, vol 6, 235-242, Chichester.

COLTORTI M. & FARABOLLINI P., 1995. Quaternary evolution of the "Castelluccio di Norcia " basin (Umbro-Marchean Apennines, Central Italy). Il Quaternario, 8(1), 149-166, Verona.

COLTORTI M., PIERANTONI P.P. & PIERUCCINI P., 1995. I depositi fluvio-lacustri di Montebibico (M.ti Martani meridionali ed il loro significato nell'evoluzione tettonico-sedimentaria del Bacino Tiberino. Studi Geologici Camerti Volume Speciale, 1, 305-315.

COLTORTI M., FARABOLLINI P., GENTILI B. & PAMBIANCHI G., 1996. Geomorphological evidence for anti-Apennine faults in the Umbro-Marchean Apennines and in the peri-Adriatic basin, Italy. Geomorphology, 15, 33-45, Elsevier, Amsterdam.

COLTORTI, M & PIERUCCINI, P., 1997a. The southern East Tiber Basin (Spoleto, Central Italy): geology and stratigraphy of the Plio-Pleistocene sediments. Il Quaternario v.10 (2), p.159-180.

COLTORTI, M. & PIERUCCINI, P., 1997b. Middle-Upper Pliocene 'Compression' and Middle Pleistocene 'Extension' in the east Tiber Basin: from 'synform' to 'extensional' basins in the Tyrrhenian side of the Northern Apennines (Central Italy). Il Quaternario v.10 (2), p.521-528.

COLTORTI M., ALBIANELLI A., BERTINI A., FICCARELLI G., LAURENZI M., NAPOLEONE G. & TORRE D., 1998. The Colle Curti Mammal Site in the Colfiorito area Umbria-Marchean Apennine, Italy): geomorphology, stratigraphy, paleomagnetism and palynology. Quaternary International, v.47/48, 107-116, Elsevier, Amsterdam.

COLTORTI, M. & PIERUCCINI, P., 1999. A late Lower Pliocene planation surface across the Italian Peninsula: a key tool in neotectonic studies. Journal of Geodynamics, v.4 (2/2), p.1-6.

COLTORTI M. & PIERUCCINI P., 2002. The Late Lower Pliocene Planation Surface and Mountain Building of the Apennines (Italy). Studi Geologici Camerti Volume Speciale International Workshop "Large-Scale Vertical Movements and related gravitational processes" Camerino-Roma 21-26/06/1999.

COLTORTI M. & FARABOLLINI P., 2004. The fluvial-coastal system of the River Tronto: geomorphological changes and human impact. FLAG, Siena 7-11 september 2004.

COLTORTI M. & PIERUCCINI P., 2004. Pedostratigraphy as a key-tool in Quaternary Geology mapping in the Umbro-Marchean-Abruzzi Apenines. Abstract International Conference "Paleosols: memory of ancient landscapes and living bodies of present ecosystems" Florence, Italy 7-11 June 2004.

CONTI M. A. & GIROTTI O., 1977. Il Villafranchiano nel "Lago Tiberino", ramo sud-occidentale: schema stratigrafico e tettonica. Geologica Romana, 16, 67-80.

CONTI S., 1991. Caratteristiche geologico-strutturali delle placche epiliguri della coltre della Val Marecchia e loro riflessi sulla franosità. Giornale di Geologia, 53: 147-165

CREMASCHI, 1979. Alcune osservazioni sul paleosuolo delle conoidi wurmiane poste al piede dell'Appennino emiliano. Geografia Fisica e Dinamica Quaternaria, 2, 187-195.

CRONIN, T.M. & DOWSETT, H.J., eds, 1991. Pliocene climates. Quaternary Science Reviews,10, p.(2/3).

DAMIANI A.V. & MORETTI A., 1969. Segnalazione di un episodio wurmiano nell'alta valle del Chienti (Marche). Bollettino della Società Geologica Italiana, 87, 171-181.

DAMIANI A.V., MENCARELLI I. & PIAZZOLI S., 1993. Studi sulla catena "Preappeninica umbra". La Dorsale di M.Peglia.Nota 1a. Considerazioni e problematiche relative alla successione etrusca. Bollettino Servizio Geologico d'Italia, 109, 105-136.

DAMIANI A.V., TUSCANO F. & VECCHIA P., 1995. Assetto stratigrafico-strutturale del M.Subasio e delle circostanti successioni silico-clastiche mioceniche (Umbria). Studi Geologici Camerti Volume Speciale, 1, 349-359.

DANSGAARD W., JOHNSSEN S.J., CLAUSEN H.B., DAHL-JENSEN D., GUNDENSTRUP N.S., HAMMER C.U., HVIDBERG C.S., SREFFENSEN G.P., SVEINBJORNSDOTTIR A.E., JOUZEL J. & BOND G., 1993. Evidence for general instability of past climate from a 250-ky ice core record. Nature, 364, 218-220.

DAVIS W.M., 1899. The geographical cycle. Geography Journal, 14, 481-504.

DECANDIA F. A., 1982. Un fenomeno di tracimazione nell'area di Spoleto (prov. di Perugia) e le sue relazioni con la tettonica recente. Bollettino Società Geologica Italiana, 101, 85-88.

DECANDIA F.A. & GIANNINI E., 1977. Studi Geologici nell'Appennino Umbro Marchigiano.1- Evidenze di due fasi tettoniche terziarie a SE di Spoleto (Prov. di Perugia).Bollettino Società Geologica Italiana, 96, 713-722

DECANDIA F.A. & TAVARNELLI E., 1991a. Cronologia delle deformazioni nell'area di Spoleto. Studi Geologici Camerti Volume Speciale 1, 329-331

DECANDIA, F.A., LAZZAROTTO, A., LIOTTA, D., CERNOBORI, L. & NICOLICH, R, 1998. The CROP 03 traverse: insights on post-collisional evolution of northern Apennines. Memorie Società Geologica Italiana, v.52, p.427-439

DE FEYTER A.J., 1991. Gravity tectonics and sedimentation of the Montefeltro (Italy). Geologica Ultraiectina, 35: 1-168.

DEIANA G. & PIALLI G. (1994) - The structural provinces of the Umbro-Marchean Apennines. Memorie della Società Geologica Italiana 48, 473-484.

DEMANGEOT J., 1965. Geomorphologie des Abruzzes Adriatiques. C.N.R.S.,.Paris, 278pp.

DESPLANQUES, G., 1969. Champagne Ombriennes. C.N.R.S, Paris, p.544.

DRAMIS F., 1992. Il ruolo dei sollevamenti tettonici a largo raggio nella genesi del rilievo appenninico. Studi Geologici Camerti, volume speciale 1992/1, 9-15.

DRAMIS F., GENTILI B. & PAMBIANCHI G., 1988. Deformazioni gravitativo profonde e grandi frane in un tratto di valle trasversale dell'Appennino marchigiano (Fiume Chienti). Atti convegno 'I corsi d'acqua minori dell'Italia appenninica', Bollettino Museo Storia Naturale Lunigiana, 6-7, 29-33

DRAMIS, F., PAMBIANCHI, G., NESCI, O. & CONSOLI, M., 1991. Il ruolo di elementi strutturali trasversali nell'evoluzione tettonico sedimentaria e geomorfologica della regione marchigiana. Studi Geologici Camerti, Volume Speciale, CROP11, 2, p.283-286.

ELMI G., NESCI O., SAVELLI D. & MALTARELLO G., 1987. Depositi alluvionali terrazzati del margine adriatico appenninico centro settentrionale: processi geomorfologici e neotettonica. Bollettino della Società Geologica Italiana, 106 (4), 717-724.

ELTER P., GIGLIA G., TONGIORGI M. & TREVISAN L., 1975. Tensional and compressional areas in the recent (Tortonian to present) evolution of the Northern Apennines. Bollettino di Geofisica Teorica e Applicata, 65 (17), 13-19.

FARABOLLINI, 1995. Evoluzione geomorfologica quaternaria dell'area periadriatica tra Ancona e Vasto. Unpublished PhD thesis Università di Perugia.

FICCARELLI G., ABBAZZI L., ALBIANELLI A., BERTINI A., COLTORTI M., MAGNATTI M., MAZZA P., MEZZABOTTA C., NAPOLEONE G., ROOK L., RUSTIONI M., & TORRE D., 1997. Cesi, an early Middle Pleistocene site in the Colfiorito basin (Umbro-Marchean Apennine), Central Italy. Journal of Quaternary Science, 12 (6), 507-518, Chichester.

GE.MI.NA., 1963. Ligniti e torbe dell'Italia continentale. Ed. Ilte Roma. pubbl. Società Geomineraria Nazionale, Roma, 81-107.

GENTILI B. & PAMBIANCHI G., 1987. Morfogernesi fluviale e attività antropica nelle Marche centro meridionali. Geogr. Fis. Din. Quat., 10, 204-217.

GENTILI B. & PAMBIANCHI G., 1988. Segnalazione di alcuni episodi fluvio-lacustri nelle Marche centro-meridionali. Geografia Fisica e Dinamica Quaternaria, 11(1), 59-60.

GENTILI B., MATERAZZI M., PAMBIANCHI G., SCALPELLA G., ARINGOLI D., CILLA G. & FARABOLLINI P., 1998. I depositi di versante del M.Ascensione (Marche meridionali, Italia). Geografia Fisica e Dinamica Quaternaria, 21 (2), 205-214.

GIROTTI O. & PICCARDI E., 1994. Linee di riva del Pleistocene inferiore sul versante sinistro della media valle del fiume Tevere. Il Quaternario, 7, 525-536.

GLIOZZI E., ABBAZZI L., ARGENTI P., AZZAROLI A., CALOI L., CAPASSO BARBATO L., DI STEFANO G., ESU D., FICCARELLI G., GIROTTI O., KOTSAKIS T., MASINI F., MAZZA P., MEZZABOTTA C., PALOMBO M.R., PETRONIO C., ROOK L., SALA B., SARDELLA R., ZANALDA E. & TORRE D., 1997. Biochronology of selected mammals, molluscs and ostracods from Middle Pliocene to the Late Pleistocene in Italy. The state of art. Rivista Italiana di Paleontologia Stratigrafica., 103: 369-388, Milano.

GOLUBIC S., VIOLANTE C., FERRERI V. & D'ARGENIO B., 1993. Algal control and early digenesis in Quetarnary travertine formation (Rocchetta a Volturno, Central Apennines). Bollettino Società Paleontologica Italiana, Volume Speciale 1, 231-247.

GORI U., 1988, Contributo alla conoscenza della sedimentazione delle alluvioni quaternarie del Fiume Foglia (Marche). Geografia Fisica e Dinamica Quaternaria, 11 (2), 121-122.

GREGORI L., 1988. Il "Bacino di Bastardo": genesi ed evoluzione nel quadro della tettonica recente. Bollettino della Società Geologica Italiana, 107, 141-151.

GUEREMY, P., 1972. La Calabre centrale et septentrionale, guide d'excursion geomorphologique. Travaux de l'Istititute Geographique Reims, 128 p.

HANCOCK P.L,. CHALMERS R.M.L, ALTUNEL E. & CAKIR Z., 1999. Travitonics: using travertines in active fault studies. Journal of Structural Geology 21, 903-916.

LEROY S., 1990. Paleoclimate plio-pleistocene en Catalogne et Languedoc d'apres la palinologie de formation lacustre. These de l'Université de Louvain.

LIPPARINI T., 1939. I terrazzi fluviali delle Marche. Giornale di Geologia, 13, 5-22.

LOTTI B., 1917. L'antico Lago Tiberino e le ligniti dell'Umbria. La Miniera Italiana., 7 (1).

LOTTI B., 1926. Descrizione geologica dell'Umbria. Memorie descrittive della Carta Geologica d'Italia, 21, 320 pp.

MARABINI S., LENAZ R & VAI G.B., 1987. Pleistocene superiore e Olocene del margine appenninico romagnolo: rapporto preliminare. Rendiconti Società Geologica Italiana, 10, 33-37.

MARTÍNEZ-NAVARRO B. & ROOK L., 2003. Gradual evolution in the African hunting dog lineage. Systematic implications. C. R. Palevol., 2: 695-702, Paris.

MARTINI, I.P. & SAGRI, M., 1993. Tectono-sedimentary characteristics of Late Miocene-Quaternary extensional basins of the Northern Apennines, Italy. Earth Science Reviews, v.34, p.197-233.

MASSARI F & PAREA G.C., 1988. Progradational gravel beach sequences in a moderate to high energy microtidal marine environment. Sedimentology, 35, 881-913.

MAZZANTI, R. & TREVISAN, C., 1978. Evoluzione della rete idrografica nell'Appennino Centro Settentrionale. Geografia Fisica e Dinamica Quaternaria, v.1, p.55-62.

MESSINA P., GALADINI F., GALLI P. & SPOSATO A., 2002. Quaternary evolution and present tectonic regime in the area of the 1997-1998 Umbria-Marche seismic sequence (central Italy). Geomorphology, 42, 97-116.

MIALL A.D., 1985. Architectural element analysis: a new method of facies analysis applied to fluvial deposits. Earth Science Review, 22, 261-308.

MIALL A:D:, 1990. Principles of sedimentary basin analysis. 2nd edition Springer-Verlag.

MIALL A:D:, 1995. Description and interpretation of fluvial deposits: a critical perspective. Sedimentology, 42, 379-384.

MIALL A:D:, 1996. The geology of fluvial deposits. Springer-Verlag.

MINELLI G & MENICHETTI M., 1990. Tectonic evolution of the Perugia massif area (Central Italy). Boll.Soc.Geol.It., 109, 445-453.

MOL J., 1997. Fluvial response to Weichselian climate changes in the Niederlausitz (Germany). Journal of Quaternary Sciences, 12 (1), 43-60.

NANNI T., PENNACCHIONI E. & RAINONE M., 1986. Il Bacino quaternario marchigiano. Atti Riunione Gruppo Sedimentologia C.N.R., Ancona 5- 7.6.86.

NANNI T. & VIVALDA P., 1987. Influenza della tettonica trasversale sulla morfogenesi delle pianure alluvionali marchigiane. Geografia Fisica e Dinamica Quaternaria, 10, 180-192.

NESCI O. & SAVELLI D., 1986. Cicli continentali tardo quaternari lungo i tratti mediani delle Marche settentrionali. Geografia Fisica e Dinamica Quaternaria, 9, 192-211

NESCI O. & SAVELLI D., 1990. I terrazzi vallivi di I ordine nei bacini dei fiumi Metauro e Foglia (Appennino marchigiano). Geografia Fisica e Dinamica Quaternaria, 13 (1), 63-73.

OLLIER C.D., 1999. Geomorphology and Mountain Building. Proceedings of the Fourth International Conference on Geomorphology, Bologna (Italia) 28-8/3-9/1997, Supplementi di Geografia Fisica e Dinamica Quaternaria, 3 (3), 49-60.

OLLIER C.D. & PAIN C.F., 2000. The Origin of Mountains. Routledge, London.

ORI, G.G., ROVERI, M. & VANNONI, F., 1986. Plio-Pleistocene sedimentation in the Apenninic foredeep (Central Adriatic Sea, Italy). Special Publications International Association of Sedimentologists, v.8, p.183-198

PICKARD, J., ADAMSON, D.A., HARWOOD, D.M., MILLER, G.H., QUILTY P.G. & DELL, K.R., 1988. Early Pliocene marine sediments, coastline and climate of East Antarctica. Geology, 16, 158-162.

PIERUCCINI P. & PIZZI A., in prep. Neotectonics of the Martana fault (Southern Umbria Region, Central Italy): a contribution to its seismic hazard assessment.

RICCI LUCCHI F., COLALONGO M.F., CREMONINI G., GASPERI G., IACCARINO S., PAPANI G., RAFFI G. & RIO D., 1982. Evoluzione sedimentaria e paleogeografica del margine appenninico. In: Cremonini G. & Ricci Lucchi F. (eds) 'Guida alla Geologia del margine appenninico-padano', 17-46, Guide Geologiche Regionali, Società Geologica Italiana, Bologna.

RICCI LUCCHI, F., 1986. The Oligocene to Recent foreland basins of the northern Apennines. Special Publications International Association of Sedimentologists, v.8, p.105-139.

RICCI LUCCHI, F., 1987. Semi-allochtonous sedimentation in the Apenninic thrust belt. Sedimentary Geology, v.50, p.119-134.

ROSE J., TURNER C., COOPE G.R., BRYAN M.D., 1980. Channel changes in the lowland river catchment over the last 13,000 years. In "Callingford R.A., Davidson D.A. and Lewin J. (eds) , Timescale in Geomorphology", John Wiley and Sons Ltd, 159-175.

RUDDIMAN W.F. & RAYMO M.E., 1988. Northern emisphere climatic regimes during the past 3Ma: possible tectonic connections. Philosoph. Trans. R.Soc. London, 318, 411-430.

SERVIZIO GEOLOGICO D'ITALIA, in press. Carta Geologica d'Italia in scala 1:50.000, F° L'Aquila.

SESTINI, R., 1981. Un'antica superficie di erosione nei monti del Chianti. Rivista Geografica Italiana, 88, 214-220.

SIGNORINI, R., 1948. Trasgressioni post-mioceniche sul bordo esterno dell'Appennino. Bollettino Società Geologica Italiana, v.46, p.1-8.

SILVESTRINI M., CILLA G. & PINOCCHI D., 1992-93. La necropoli eneolitica di Fontenoce (Recanati). Picus, XII-XIII, 127-185.

SILVESTRINI M., BASSETTI M., BOSCATO P., COLTORTI M., ESU D., LEMORINI C., MAGNATTI M., PERESANI M., ROSSETTI P., 2001. An Acheulean site of the Last Interglacial at Boccabianca (Cupramarittima, Marche). Rivista di Scienze Preistoriche, 1,21-71

STARKEL L., 1983. The reflection of hydrologic changes in the fluvial environment of the temperate zone during the last 15.000 years. In: Starkel, Ed., Background to palaeohydrology, 213-237, Chichester.

SUC J.P. & ZAGWIJN W.H., 1983. Plio-Pleistocene correlation between the north-western mediterranean region and north-western Europe according to recent biostratigraphic and palaeoclimatic data. Boreas, 12, 153-166.

TORRE D., FICCARELLI G., MASINI F., ROOK L. & SALA B., 1992. Mammal dispersal events in the Early Pleistocene of Western Europe. Courier Forschungsinstitut Senckenberg, 153, 51-58.

VAIL P.R., MITCHUM R.M., TODD R.G., VIDMIER J.M., THOMPSON S., SANGREE J.B., BUBB J.N., & hatlelid w.g., 1977. seismic stratigraphy and

global changesof sea-level. In: Payton C.E. (ed.) Seismic stratigraphy – applications to hydrocarbon exploration. American Association of Petroleum Geologists, Memories 36, 49-212.

VANDENBERGHE J., 1992. Climatic change and landscape development: an example from the past. Catena Supplement 22, 73-83.

VANDENBERGHE J., 2002. The relation between climate and river processes, landforms and deposits during the Quaternary. Quaternary International, 91, 17-23.

VANDENBERGHE J., BOHNCKE S., LAMMERS W., ZILVERBERG L., 1987. Geomorphology and palaeoecology of the Mark Valley (southern Netherlands). I- Geomorphological development during the Weichselian and Holocene. Boreas 16,55–67.

VAN HUISSTEDEN J. & VANDENBERGHE J., 1988. Changing fluvial style of lowland river during the Weichselian pleniglacial in the eastern Netherland. Apennines. Journal of the Geological Society, London, 157, 629-633. Zeitschrift fur Geomorphologie, 71, 131-146.

VILLA G.M., 1942. Nuove ricerche sui terrazzi fluviali delle Marche. Giornale di Geologia, 16, 5-75.

WATTS W.A., 1985. A long pollen record from Laghi di Monticchio, southern Italy: a preliminary account. Journal of the Geological Society, 142, 491-499.

ZATTIN M., LANDUZZI A., RICOTTI V. & ZUFFA G.G., 2000. Discriminating between tectonic and sedimentary burial in a foredeep succession, Northern Apennines. Journal of the Geological Society of London, 157, 629-633.