GEOLOGICA ULTRAIECTINA

Mededelingen van de
Faculteit Aardwetenschappen der
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No. 35

GRAVITY TECTONICS AND SEDIMENTATION OF THE MONTEFELTRO, ITALY

ARNOUD JAN DE FEYTER

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GRAVITY TECTONICS AND SEDIMENTATION OF THE MONTEFELTRO, ITALY

GRAVITATIEVE TEKTONIEK EN SEDIMENTATIE VAN HET MONTEFELTRO, ITALIË

(met een samenvatting in het Nederlands)

PROEFSCHRIFT TER VERKRIJGING VAN DE GRAAD VAN DOCTOR
AAN DE RIJKSUNIVERSITEIT TE UTRECHT,
OP GEZAG VAN DE RECTOR MAGNIFICUS PROF. DR. J.A. VAN GINKEL,
INGEVOLGE HET BESLUIT VAN HET COLLEGE VAN DEKANEN
IN HET OPENBAAR TE VERDEDIGEN
OP MAANDAG 7 OKTOBER 1991 DES NAMIDDAGS TE 0.45 UUR

DOOR

ARNOUD JAN DE FEYTER

GEBOREN OP 28 MEI 1957 TE HEERLEN

PROMOTORES: PROF. DR. E. TEN HAAF
PROF. DR. G. PIALLI

ter nagedachtenis aan mijn vader
per Ita

...allevati per una bella carriera in geologia...

CONTENTS

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family of Monte Copiolo, and the Matteucci family of Cagli.

SUMMARY

The tectono-stratigraphic framework of the southern Montefeltro is illustrative of the interaction between thin-skinned shearing and sedimentation in the outer segment of the Apenninic orogenic system during the Neogene. Mesozoic through Paleogene evaporitic-carbonatic-marly terrains constitute the lower part of its autochthonous sedimentary cover. This Umbro-Marchean Sequence compares with deposits of the contiguous Adriatic Foreland. Structural style is characterized by longitudinal chains of brachyanticlines more or less intensely dissected by stepped thrusts rooted in a common sole fault.

The Neogene clastics of the overlying Umbro-Romagnan Sequence are arranged in broad synclines separated by complex highs developed over shallow detachment levels. The pronounced lateral variability of these foredeep deposits argues the synsedimentary origin of the structural pattern.

More internal detached Umbro-Romagnan terrains define a parautochthonous thrust sheet composed of synclinal elements bounded by listric reverse faults. The formation of these shallow structures was determined by migratory subsidence in consequence of crustal imbrication.

The synsedimentary orogenic paroxysm was complicated by the emplacement of a voluminous glide mass. This Montefeltro Colata principally consists of teleallochthonous terrains derived from the innermost segment of the Apenninic system. Semi- and peneallochthonous elements originated on the mobile allochthon as this advanced toward the Adriatic Foreland.

The phased progress of the Montefeltro Colata was guided by bulges bordering a transversal depression which resulted from the interaction between the mobile allochthon and the thin-skinned tectonics of its Umbro-Marchean-Romagnan substratum.

Neoautochthonous deposits suture the structural edifice on the outer side of the Apenninic orogen.

SAMENVATTING

De tektono-stratigrafische bouw van het zuidelijke Montefeltro weerspiegelt het samenspel van ondiepe deformatie en sedimentatie in het buitenste segment van het Apennijnse orogene systeem tijdens het Neogeen. Mesozoïsche en Paleogene evaporitisch-carbonatisch-mergelige gesteentes vormen het onderste deel van de autochtone sedimentaire bedekking. Deze Umbro-Marchigiaanse Sequentie is vergelijkbaar met de afzettingen van het aangrenzende Adriatische Voorland. Longitudinale ketens van brachyanticlines die in meer of mindere mate worden doorsneden door getrapte overschuivingen wortelend in een basaal décollement niveau kenmerken de structurele stijl.

De overliggende Umbro-Romagnoolse Sequentie bestaat uit Neogene klastische sedimenten. Deze vormen brede synclines gescheiden door complexe hoogs die zich ontwikkeld hebben boven ondiepe schuifzones. De uitgesproken laterale variabiliteit van deze voordiep afzettingen duidt op de synsedimentaire origine van het structurele patroon.

Meer interne ontwortelde Umbro-Romagnoolse gesteentes vormen een parautochtoon dekblad bestaande uit synclinale eenheden gescheiden door listrische opschuivingen. Het ontstaan van deze ondiepe structuren werd bepaald door migratoire daling ten gevolge van imbricatie van de korst.

De hoofdfase van synsedimentaire orogene activiteit viel samen met de aankomst van een volumineuze glijdmassa. Deze Montefeltro Colata bestaat grotendeels uit teleallochtone gesteentes afkomstig uit het meest interne segment van het Apennijnse orogene systeem. Semi- en peneallochtone elementen werden afgezet op het allochtoon terwijl dit zich voortbewoog in de richting van het Adriatisch Voorland.

De gefaseerde voortbeweging van de Montefeltro Colata werd gestuurd door drempels aan weerszijden van een transversale depressie. Deze structuren ontwikkelden zich ten gevolge van de wisselwerking tussen het mobiele allochtoon en de ondiepe tektoniek van de Umbro-Marchigiaanse-Romagnoolse ondergrond.

Neoautochtone sedimenten bedekken de structurele stapeling aan de buitenrand van het Apennijnse orogeen.

RIASSUNTO

La struttura tettono-stratigrafica del Montefeltro meridionale illustra l'interazione fra tettonica di scollamento e sedimentazione nella parte esterna del sistema orogenico appenninico durante il Neogene. Rocce evaporitico-carbonatico-marnose di età mesozoico-neogenica costituiscono la porzione inferiore della copertura sedimentaria autoctona. Tale Sequenza Umbro-Marchigiana si collega ai depositi del contiguo Avampaese Adriatico. Strutturalmente, essa è caratterizzata da catene longitudinali di brachianticlinali tagliate più o meno intensamente da sovrascorrimenti a gradini. Questi ultimi si collegano ad un livello di scollamento basale.

In sovrapposizione stratigrafica con la Sequenza Umbro-Marchigiana, si trova la Sequenza Umbro-Romagnola costituita da terreni clastici neogenici il cui stile tettonico è definito da ampie sinclinali separati da alti strutturali. Questi ultimi mostrano una elevata complessità dovuta al loro sviluppo sopra livelli di scollamento a bassa profondità. La marcata variabilità laterale di questi terreni di avanfossa riflette l'origine sinsedimentaria dell'assetto strutturale.

Nella parte interna della zona, si ha una falda parautoctona costituita da terreni umbro-romagnoli sradicati. Strutturalmente, essa è composta da unità sinclinaliche separate da faglie inverse listriche. La genesi di queste strutture superficiali è da ricercarsi in una subsidenza migratoria dovuta ad una embricazione crostale.

Il parossisma orogenico sinsedimentario veniva complicato dalla messa in posto di una potente massa di scivolamento formante la Colata del Montefeltro. Tale colata è costituita principalmente da terreni telealloctoni derivanti dalla zona più interna del sistema orogenico appenninico. Sono presenti inoltre elementi semi- e penealloctoni originatisi sull'alloctono mentre questo avanzava verso l'Avampaese Adriatico.

L'avanzamento discontinuo della Colata del Montefeltro era condizionato da dorsali poste ai bordi di una depressione trasversale formatasi come risultante dell'interferenza tra l'alloctono in movimento e la tettonica di scollamento del suo substrato umbro-marchigiano-romagnolo.

Nel margine esterno dell'orogeno appenninico, l'edificio strutturale viene suturato da sedimenti neoautoctoni.

CHAPTER I

INTRODUCTION

I.1. INTRODUCTION TO THE GEOLOGY OF THE NORTHERN APENNINES

The Northern Apennines belong to the Mediterranean system of Alpine chains, which resulted from the interaction between the African and European plates. The Apenninic cycle commenced with Triassic rifting of a Hercynian basement at the western margin of the Tethys (Grandjacquet & Glangeaud, 1962; Horvath & Channell, 1977; Laubscher & Bernouilly, 1977; Boccaletti *et al.*, 1980; Pieri & Mattavelli, 1986). This initially accelerated the continental coarse clastic deposition of the Permotriassic Verrucano Formation in the aftermath of the Hercynian cycle (Merla, 1964; Bortolotti *et al.*, 1970b; Dallan Nardi & Nardi, 1975; Bally *et al.*, 1986).

Wide-spread subsidence during the Late Triassic initiated marine Apenninic sedimentation, at first evaporitic-carbonatic. This was succeeded by the deposition of neritic and pelagic carbonates (Abbate *et al.*, 1970b). During the Late Jurassic, the continuing crustal divergence culminated with the formation of an elongate oceanic basin. This consisted of the Piemontese Zone, adjacent to the Sardo-Corsican continental margin of the European Plate, and the Ligurian Zone, bounded by the Adriatic continental margin of African afinity. As a first result of the onset of convergent plate motions, the Piemontese Zone was probably subducted under the Ligurian Zone during the Cretaceous (Boccaletti *et al.*, 1980). This is reflected by the hp-lT metamorphism of its sedimentary contents, which is mainly known from the Western Alps.

The Ligurian Zone initially accumulated radiolarites, Calpionella limestones, and shales with siliceous limestones (*e.g.*, Merla, 1951, 1959; Reutter & Sames, 1962; Abbate & Sagri, 1970). From the Late Cretaceous on, turbiditic sedimentation prevailed. In the internal part of the zone, this was determined by a Sardo-Corsican supply of siliciclastics (Bertini *et al.*, 1975; Sagri & Marri, 1980). More externall, that is closer to the Adriatic continental margin, mainly calcareous turbidites were deposited (Abbate & Sagri, 1981). Meanwhile, the Ligurian Zone had become involved in the subductive process due to the continuation of convergence after the Oiemontese Zone was completely consumed (Boccaletti *et al.*, 1971, 1980; Sagri, 1973). Its sedimentary cover essentially escaped metamorphism as it was obducted onto the continental margin facing the subduction zone, together with slivers of its ophiolitic substratum (Bortolotti, 1983).

The subduction of oceanic lithosphere was completed during the Late Eocene. The collision of the Sardo-Corsican and Adriatic continental margins was accompanied by intense shearing of the decoupled Ligurids. This Ligurian Phase represents the first paroxysm of Apenninic orogeny (*e.g.*, den Haan, 1979).

The orogenic cycle now entered its ensialic stage. Tectonic activity gradually encroached upon the Adriatic Foreland, which meanwhile became detached from the African Plate (*e.g.*, VandenBerg, 1979). The pelagic foreland sedimentation came to an end as elongate turbiditic foredeeps were generated successively from the Tuscan Zone to the more external Umbro-Marchean-Romagnan Zone. Clastics were mainly supplied longitudinally by Austro-Alpine sources. The resulting tripartite turbidite suites are distinguished by their arenaceous-marly component, reflecting the acme of migratory foredeep sedimentation (Sagri, 1972, 1973; Burger *et al.*, 1978; de Feyter *et al.*, 1990). Thus, the Inner Tuscan Foredeep accumulated the Oligocene Macigno Formation, and the Outer Tuscan Foredeep the Upper Oligocene to Lower Miocene M.Modino-M.Cervarola Formation. The Middle to Upper Miocene Marnoso-arenacea Formation originated in the Umbro-Romagnan Foredeep of the Umbro-Marchean-Romagnan Zone. In the southeastern part of this zone, the Laga Formation was deposited in the Laga Basin during the Late Miocene to Early Pliocene. The yet more external composite Padan-Marchean-Adriatic Foredeep is still partly active, receiving Apenninic detritus.

The eventual closure of the Apenninic foredeeps was generally provoked by thin-skinned shearing. Concomitantly, the Ligurids were translated towards the Adriatic Foreland as an incoherent thrust sheet, carrying active satellite basins.

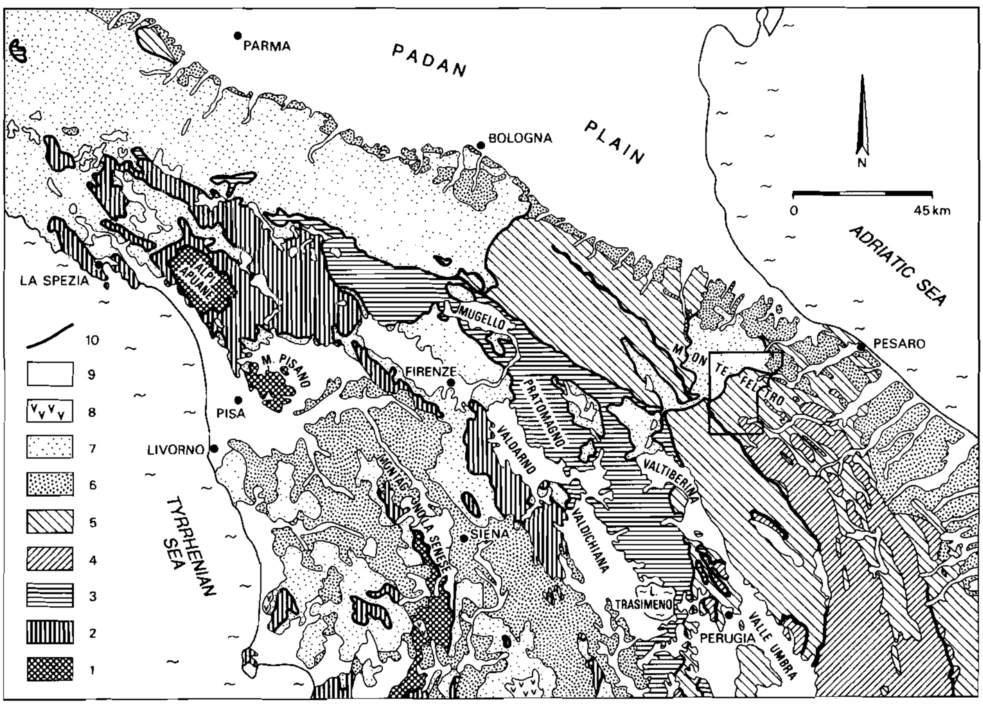


Fig. 1. Schematic geological map of the Northern Apennines, with indication of the area studied in detail (modified from Bortolotti et al.,

1970a; Ogniben, 1975).
Legend: Tuscan Complex: 1: metamorphic stationary elements; 2: Tuscan Nappe; 3: M.Modino-M.Cervarola Domain. Umbro-Marchean-Romagnan Complex: 4: foreland deposits; 5: Miocene foredeep deposits. 6: Upper Miocene to Pliocene deposits of Tuscan extensional basins and Pliocene foredeep deposits of Umbro-Marchean-Romagnan Zone; 7: Ligurids and associated terrains; 8: Pliocene and Quaternary volcanics; 9: intramontane "Villafranchian" and Quaternary deposits; 10: principal tectonic contact.

The Late Miocene Tuscan Phase of Apenninic orogeny caused major forward thrusting of the foredeep deposits and their sedimentary substrata (*e.g.*, Dallan Nardi & Nardi, 1975). The Tuscan Nappe, characterized by the fill of the Inner Tuscan Foredeep, experienced the largest displacement (Giannini *et al.*, 1962; Saggini, 1965). It overrode stationary elements which thereby became metamorphosed in greenschistfacies. The detached sedimentary contents of the Outer Tuscan Foredeep formed the M.Modino-M.Cervarola Domain (Guenther & Reutter, 1985). The internal part of the clastic wedge of the Umbro-Romagnan Foredep evolved into a similar superficial thrust sheet. The rootless character of the more external Umbro-Marchean-Romagnan terrains, which were affected by paroxysmal pulses until the Early Pliocene, gradually decreases towards the stable Adriatic Foreland hidden beneath the Padan Plain and Adriatic Sea.

Migratory uplift and extensional shearing related to the opening of the Tyrrhenian Sea, accompanied by Pliocene to Quaternary volcanism in the Tuscan Zone, completed the evolution of the Northern Apennines. Only in the external part of the orogenic belt, the structural edifice has remained largely unaffected by these processes (Fig. 1). Therefore, this is most suitable to study the interaction of detachment tectonics and foredeep sedimentation as well as the emplacement of far-travelled Ligurids.

I.2. STRUCTURAL CLASSIFICATION OF APENNINIC TERRAINS

Since the Tuscan and Umbro-Marchean-Romagnan foredeep wedges with their sedimentary substrata generally maintained their original relative position, they are referred to as the autochthonous assemblage of the Northern Apennines (cf. Beneo, 1949; Ricci Lucchi & Ori, 1985; Ricci Lucchi, 1986a). The ovelying Ligurids and associated terrains accordingly represent the allochthonous assemblage (Fig. 2). With reference to amount of displacement and structural affinity, a more detailed classification is possible. Hereby, one has to bear in mind that the peculiarities of Apenninic geology warrant a terminology which does not necessarily conform to the definitions given by Bates & Jackson (1980; cf. den Haan, 1979).

The rarity of truly autochthonous terrains is characteristic of essentially thin-skinned fold-and-thrust belts such as the Northern Apennines. Harris & Milici (1977) and ten Haaf (1985) therefore extended the concept of autochthony to the lowest and least displaced tectonic elements of such belts. This implies that the autochthon comprises both the stable foreland and terrains which are merely separated from it by minor discontinuities of limited dimensions. The sedimentary cover of the Umbro-Marchean-Romagnan Zone largely fulfils this condition (Decandia & Giannini, 1977).

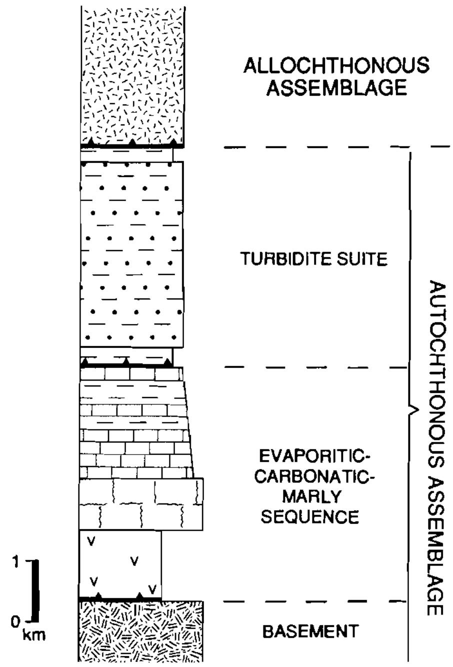


Fig. 2. Schematic columnar section of the structural edifice of the Northern Apennines.

A subdivision of autochthonous terrains is possible according to their position relative to the allochthonous assemblage (Fig. 3). Those which originated prior to the emplacement of allochthonous elements can be designated as palaeoautochthonous (Amadesi, 1963; Castellarin & Pini, 1987; Conti, 1989). Neoautochthonous terrains on the other hand were deposited after the allochthon ceased movement (*e.g.*, de Wijkerslooth, 1934; Merla, 1951; Ruggieri, 1958).

Parautochthonous elements experienced marked relative displacements within the autochthonous assemblage. This applies to the Tuscan Nappe and the M.Modino-M.Cervarola Domain as well as to the internal portion of the Umbro-Romagnan clastic wedg (Frey, 1969; ten Haaf & van Wamel, 1979).

The Ligurids, which make up the bulk of the allochthonous assemblage, have been addressed as teleallochthon (R.Selli *in* Selli, 1967; Castellarin & Pini, 1987). This attests to the impressive translations they underwent in the course of Apenninic orogeny. The deposits of the satellite basins carried by them display a variable degree of allochthony. Semiallochthonous ones as a rule originated at a considerable distance from the site where the allochthon eventually came to rest (Merla, 1951; Ruggieri, 1970; Ricci Lucchi, 1987). Peneallochthonous terrains on the other hand were only slightly displaced. Unlike the semiallochthon, they typically resemble the coeval autochthon. Peneallochthony thus appears a stage intermediate between neoautochthony and semiallochthony.

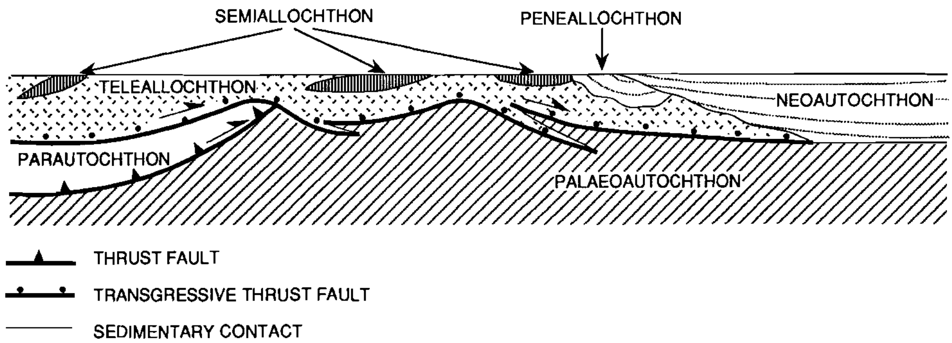


Fig. 3. Structural classification of Apenninic terrains. The tectonically transgressive and interdigitating aspect of the allochthonous assemblage is indicative of an intermittent synsedimentary advance.

I.3. OUTLINE OF THE SOUTHERN MONTEFELTRO

The Mesozoic through Lower Miocene Umbro-Marchean Sequence forms the lower part of the sedimentary cover of the Umbro-Marchean-Romagnan Zone. It comprises evaporitic-carbonatic-marly foreland deposits. The overlying Umbro-Romagnan Sequence is characterized by the Middle to Upper Miocene turbidite suite of the Umbro-Romagnan Foredeep. This is overlain to the NE by Upper Miocene and younger evaporites and molasse deposits.

Longitudinal basins and chains are the principal structures of the essentially thin-skinned autochthon of the Umbro-Marchean-Romagnan Zone (Fig. 4). In the internal part of the zone, the autochthon solely comprises the Umbro-Marchean Sequence, while the Umbro-Romagnan terrains constitute a parautochthonous thrust sheet. The structural pattern of the autochthonous assemblage is masked by "*colate*", *i.e.* flows of allochthonous material, NW of the Sillaro Line and in the Montefeltro (*e.g.*, Dal Piaz, 1943; Merla, 1951). The Montefeltro Colata forms the boundary between the Romagnan Apennines to the NW and the Umbro-Marchean Apennines to the SE.

The southern Montefeltro, which has been studied in detail, displays palaeo- and neoautochthonous, parautochthonous, and allochthonous terrains (Fig. 5). The Umbro-Romagnan parautochthon is represented by the Borgo Pace Unit and the frontal M.Vicino Unit. The contiguous autochthon is characterized by broad synclines and complex highs. These are the counterparts at the level of the Umbro-Romagnan Sequence of the principal basins and anti-

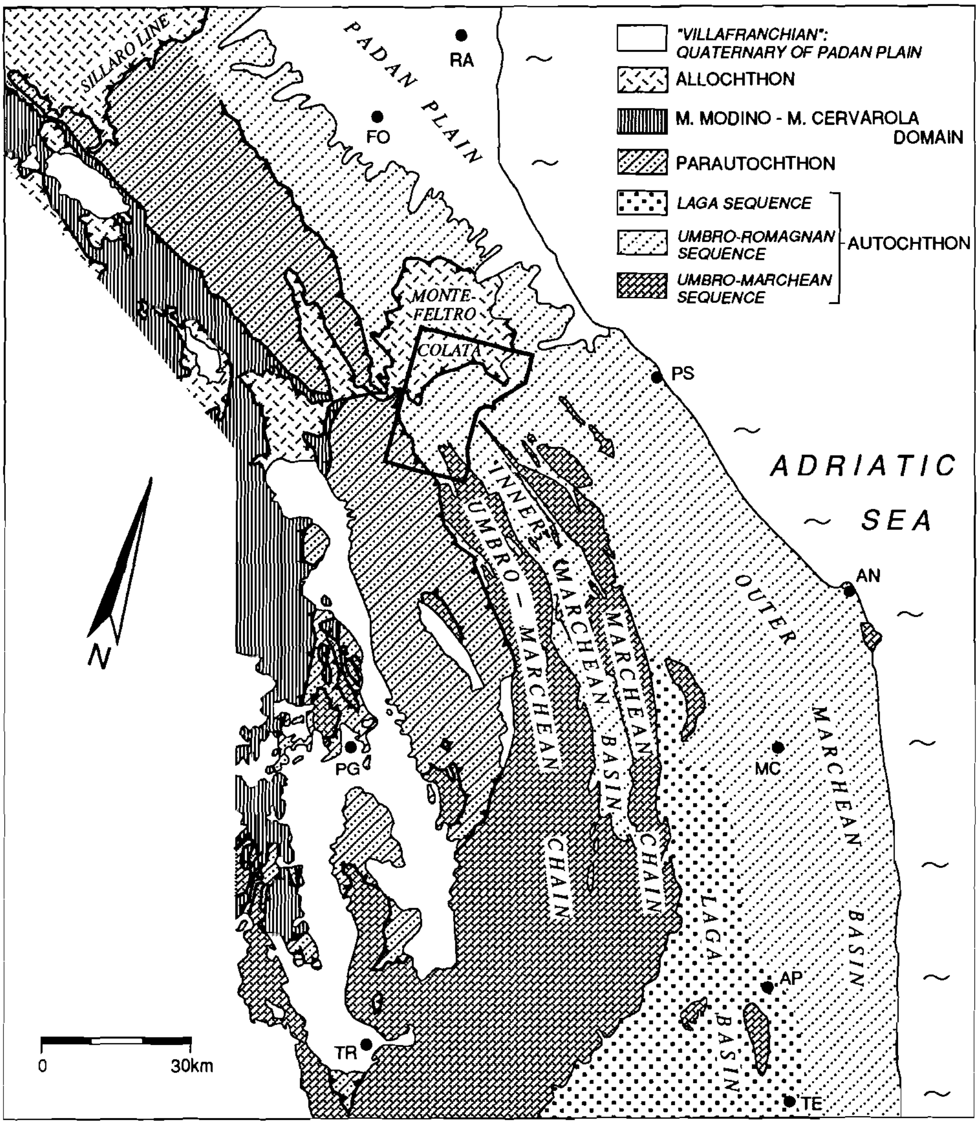


Fig. 4. Geological scheme of the Umbro-Marchean-Romagnan Zone SE of the Sillaro Line, with indication of the area studied in detail. Barbed thrust faults are only shown as far as they constitute major structural boundaries.
AN: Ancona; AP: Ascoli Piceno; FO: Forlì; MC: Macerata; PG: Perugia; PS: Pesaro; RA: Ravenna; TE: Teramo; TR: Terni.

clinal chains forming the backbone of the Umbro-Marchean Apennines further to the SE. Thus, the Piandimeleto High, situated between the Casale-Campo and Pietrarubbia synclines, lines up with the Montiego Anticline of the Umbro-Marchean Chain. The more external Macerata Fèltria High, which is bordered by the secondary M.S.Leo High and Castellina Anticline, relates to the Marchean Chain.

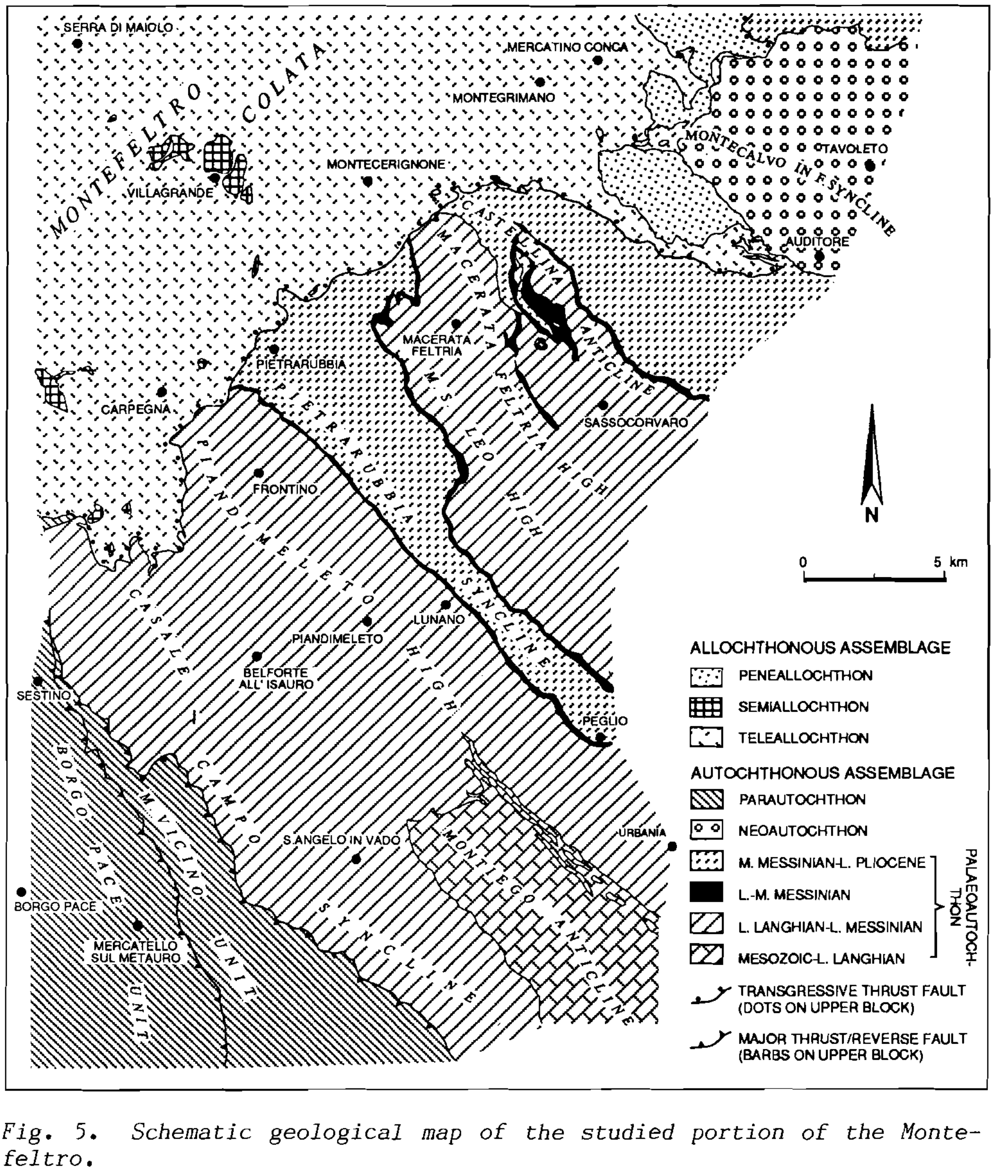


Fig. 5. Schematic geological map of the studied portion of the Montefeltro.

The Montefeltro Colata transversely covers the longitudinal structures of the autochthonous assemblage (Fig. 5). It is largely of teleallochthonous affinity. The allochthon incorporates few semiallochthonous elements. More voluminous peneallochthonous basins are displayed by its frontal portion. This is overlain by neoautochthonous deposits in the Montecalvo in Foglia Syncline.

I.4. SCOPE AND METHODS OF RESEARCH

The southern Montefeltro is exemplary of the complexity of Apenninic

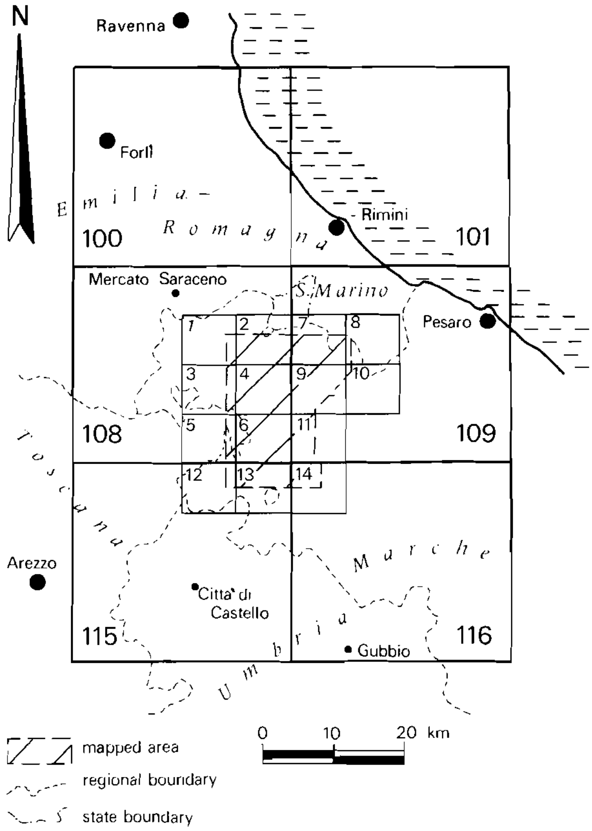


Fig. 6. Situation of 1:25.000 topographic maps and 1:100.000 geological maps (listed with date of most recent edition) with respect to the investigated area.
1: 108 I SO Novafèltria

2:	*108*	*I SE*	*S.Leo*
3:	*108*	*II NO*	*Pennabilli*
4:	*108*	*II NE*	*Macerata Fèltria*
5:	*108*	*II SO*	*Sestino*
6:	*108*	*II SE*	*Piandimeleto*
7:	*109*	*IV SO*	*Monte Grimano*
8:	*109*	*IV SE*	*Saludecio*
9:	*109*	*III NO*	*Sassocorvaro*
10:	*109*	*III NE*	*Montecalvo in Foglia*
11:	*109*	*III SO*	*Urbania*
12:	*115*	*I NO*	*Borgo Pace*
13:	*115*	*I NE*	*S.Angelo in Vado*
14:	*116*	*IV NO*	*Piòbbico*

100	*Forlì (1968)*
101	*Rimini (1967)*
108	*Mercato Saraceno (1969)*
109	*Pesaro (1969)*
115	*Città di Castello (1969)*
116	*Gubbio (1952)*

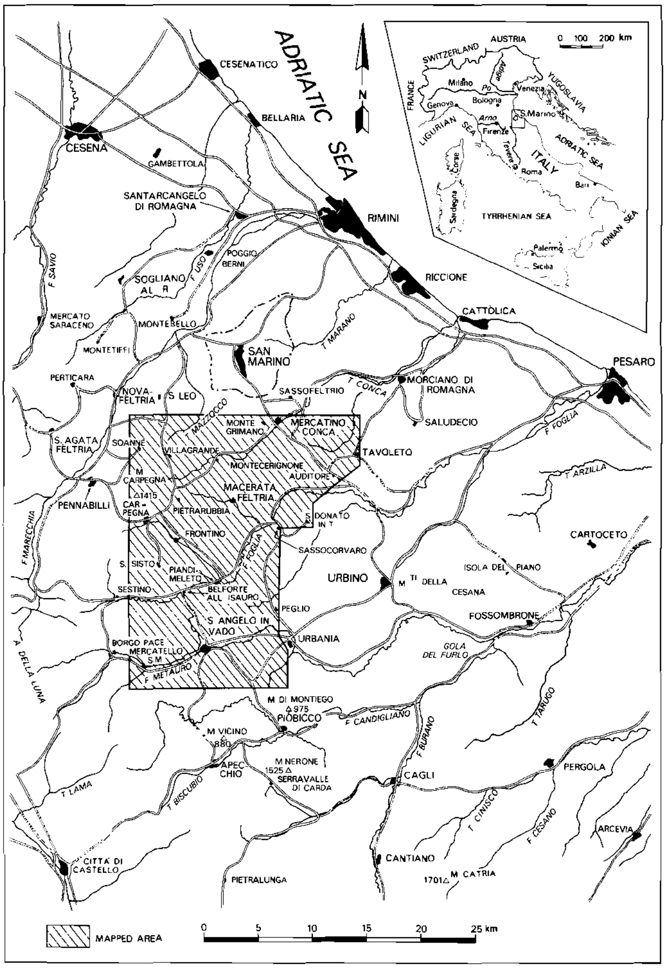


Fig. 7. Location and road map of the Montefeltro and its surroundings.

foredeeps due to the interaction of sedimentation and tectonics. Therefore, research was primarily aimed at the integration of stratigraphic and structural data. The main features of the autochthonous assemblage in this context are its structural style, indicative of a thin-skinned deformation mechanism, and the variability of the Umbro-Romagnan Sequence as a function of structural position, demonstrating that sedimentation patterns were tectonically conditioned. Anomalies resulted from the emplacement of the Montefeltro Colata. Mode and timing of the phased progress of the allochthon are attested to by internal structure and position relative to autochthonous stratigraphy.

Although major errors appear absent on the official 1:100.000 geological maps of the Servizio Geologico d'Italia in Roma, the goal set for this study necessitated detailed mapping of the southern Montefeltro (cf. enclosures). This was essentially carried out on 1:25.000 topographic maps of the Istituto Geografico Militare in Firenze (Fig. 6). In particularly complicated sectors, additional use was made of 1:10.000 orthophotographic maps of the Regione Marche in Ancona.

With regard to the efficiency of the field investigations, the rather low degree of exposure of the southern Montefeltro was somewhat compensated for by its excellent accessibility (Fig. 7).

I.5. PREVIOUS AUTHORS

The earliest contributions on the geology of the Montefeltro and its surroundings date from the nineteenth century (*e.g.*, Scarabelli Gommi Flamini, 1851, 1880; Pareto, 1865). They are merely of historical value, depicting a general autochthony. The mineralogical analyses Salmojraghi (1903, 1909) conducted on semiallochthonous deposits yielded the first indications of a pronounced allochthony.

The allochthonous character of the Ligurids in the Northern Apennines was first recognized by Steinmann (1907). Bonarelli (1929) accordingly interpreted the Montefeltro Colata as a gravitationally emplaced allochthonous element of southwestern provenance. This was placed in a wider perspective by de Wijkerslooth (1934) and Merla (1951), who introduced the concept of semiallochthony. Principi (1925b, 1934, 1938, 1939) and Sacco (1937, 1940) on the other hand continued to advocate a diapiric autochthonous origin for the Montefeltro Colata. Diapirism, albeit redistributing allochthonous material, was still invoked by Lipparini (1947-1948-1949), Behrmann (1958) and Richter (1962). Ruggieri (1953a, 1953b, 1953c, 1956a, 1956b, 1958) however demonstrated the phased superficial emplacement of the Montefeltro Colata during the Late Miocene to Early Pliocene.

Principi (1925a, 1931, 1939) and Renz (1933) conducted the first systematic studies on the autochthonous assemblage of the southern Montefeltro. They focused on the lithostratigraphy of the Umbro-Romagnan and the Umbro-Marchean Sequence, respectively. The structural style of the Umbro-Romagnan parautochthon was documented by Signorini (1941). Selli (1949, 1952) exhaustively described the autochthon, arguing its thin-skinned character.

The regional syntheses by Selli (1952) and Ruggieri (1958) provided the basis for modern research. Thus, Venzo (1954), Amadesi (1962) and Stern (1969) studied in detail segments of the southern Montefeltro. Its general constitution was outlined on the Mercato Saraceno and Pesaro sheets of the 1:100.000 geological map of Italy (Ruggieri, 1970; Carloni *et al.*, 1971).

Thematic studies mainly addressed biostratigraphical and sedimentological issues. Hereby, the allochthonous terrains of the Montefeltro Colata received little attention (Ricci Lucchi, 1964; Capuano *et al.*, 1982). The upper par of the Umbro-Marchean Sequence was examined by Baumann & Roth (1969), Guerrera (1977) and Guerrera *et al.* (1988), and various aspects of the overlying Umbro-Romagnan turbidite suite by Ardanese *et al.* (1982-1983, 1987) and Capuano *et al.* (1987a). Particularly intense research was devoted to the Messinian portion of the Umbro-Romagnan Sequence (Farabegoli & Ricci Lucchi, 1973; Bellagamba, 1978; Savelli & Wezel, 1978; Ardanese & Martelli, 1983-1984; de Feyter & Molenaar, 1984; Molenaar & de Feyter, 1985). This interest was excited by the recognition of the Messinian Salinity Crisis as a Mediterranean event (*e.g.*, Hsü *et al.*, 1973). Capuano *et al.* (1986b, 1987d) evaluated the sedimentary evolution of the neoautochthon.

The structural framework of the Montefeltro has only recently become the target of systematic research. With reference to seismic sections, Bally *et al.* (1986) envisaged a typical thin-skinned geometry for the autochthonous assemblage. Rootless autochthonous features were studied in detail by de Feyter & Menichetti (1986). An interdisciplinary approach enabled de Feyter *et al.* (1986, 1990) to demonstrate the superficial deformation style of the Umbro-Romagnan parautochthon, in conformity with the concept outlined by ten Haaf & van Wamel (1979). Thorough structural analyses of the Montefeltro Colata have only been carried out by Mannori & Sani (1987). Their findings essentially comply with a gravitational emplacement mechanism.

Capuano *et al.* (1986a, 1987c) presented schematic palinspastic reconstructions of the frontal portion of the Montefeltro Colata. They elucidated how it affected the sedimentary evolution of the autochthon. The general constitution of the allochthon was reviewed by Veneri (1986). Finally, Conti *et al.* (1987) and Conti (1989) attributed an imbricated style to the Montefeltro Colata, which they considered contradictory to its presumed gravitational emplacement.

CHAPTER II

LITHOSTRATIGRAPHY AND SEDIMENTARY EVOLUTION OF THE AUTOCHTHONOUS ASSEMBLAGE

II.1. INTRODUCTION

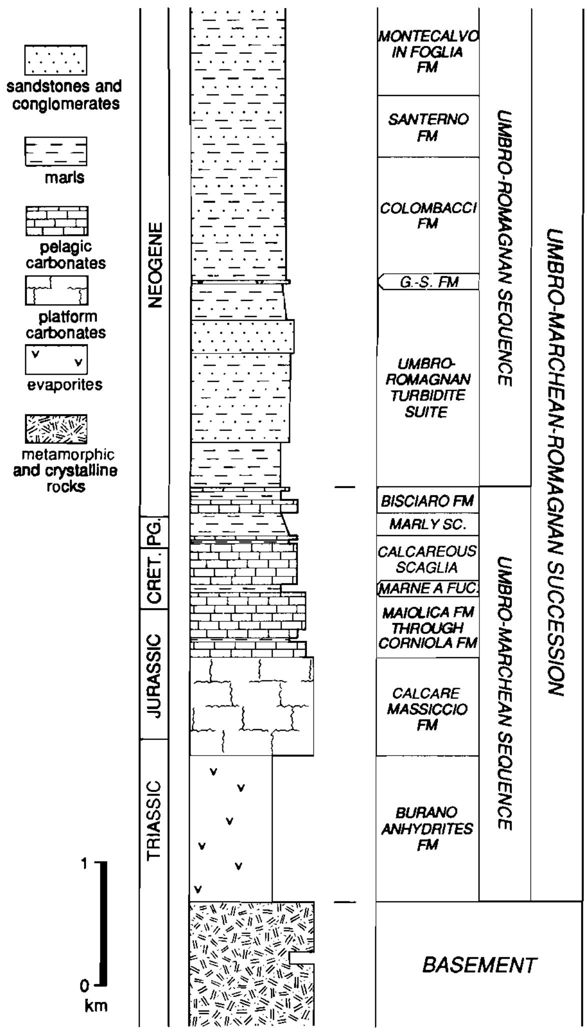


Fig. 8. Schematized stratigraphy of the autochthonous assemblage of the southern Montefeltro.

The bipartite Umbro-Marchean-Romagnan Succession constitutes the autochthonous assemblage of the southern Montefeltro (Fig. 8). Its lower part consists of the evaporitic-carbonatic-marly Umbro-Marchean Sequence, which is confined to the palaeoautochthon. Only the upper portion of these foreland deposits is exposed in the area studied in detail. The character of the basal part of the sedimentary cover and the basement must however also be considered in the context of an accurate thin-skinned evolutionary model.

The Umbro-Romagnan Sequence comprises the fill of the Umbro-Romagnan Foredeep. Both the parautochthon and the palaeoautochthon display its voluminous tripartite turbidite suite. The overlying evaporitic and molasse deposits range from the palaeoautochthon to the neoautochthon.

II.2. THE BASEMENT

The rigid basement of the Umbro-Marchean-Romagnan Zone is here considered to encompass the actual Hercynian basement as well as its Permotriassic tegument. It has only been evidenced by exploratory drilling. Thus, the Perugia 2 well, 50 kilometers S of the Montefeltro, penetrated sheared phyllites (Ghelardoni, 1962; Martinis & Pieri, 1964). These probably constitute an equivalent of the detrital Verrucano Formation exposed in the Tuscan Zone (Bortolotti *et al.*, 1970b; G. Giglia & L. Trevisan *in* Desio, 1973; Lavecchia & Pialli, 1981a). The Alessandra 1 well, situated in the Adriatic Sea some 100 kilometers E of the Montefeltro, encountered a thick series of red sandstones of similar affinity (Bally *et al.*, 1986).

II.3. THE UMBRO-MARCHEAN-ROMAGNAN SUCCESSION

II.3.1. THE UMBRO-MARCHEAN SEQUENCE

II.3.1.1. BURANO ANHYDRITES FORMATION (NORIAN)

The Upper Triassic Burano Anhydrites Formation is the basal component of the sedimentary cover of the Umbro-Marchean-Romagnan Zone. In the Massicci Perugini, 50 kilometers S of the Montefeltro, as well as in more internal sectors of the Northern Apennines, it is represented in outcrop by a kind of dissolution breccia known as *"calcare cavernoso"* (*e.g.*, Merla, 1951; Dessau, 1962). its presence in the subsurface of the Umbro-Marchean Apennines has been demonstrated by exploratory drilling. In the vicinity of the Montefeltro, the Burano Anhydrites Formation was penetrated by wells near Fossombrone and in the F.Burano gorge between Cagli and Cantiano, from which its name has been derived (Martinis & Pieri, 1964).

The Burano Anhydrites Formation is constituted by irregularly alternating carbonates and sulphates, with subordinate marls. It is generally attributed to evaporitic deposition in a coastal setting (*e.g.*, Bortolotti *et al.*, 1970b; Colacicchi & Passeri, 1981; Centamore *et al.*, 1986). Commonly, the incompetent terrains are severely contorted (Ciarapica & Passeri, 1976). This attests to the mobility of the Burano Anhydrites Formation in the course of Apenninic orogeny (Bally *et al.*, 1986). Its original thickness therefore can only be estimated. Clearly, tectonic accumulations must be invoked for the thicknesses of several kilometers encountered in drilled anticlinal cores (Martinis & Pieri, 1964). An original thickness of about 1200 meters seems reasonable for the southern Montefeltro. This is in line with the average value Lavecchia & Pialli (1980) determined for the northern part of the Umbro-Marchean Apennines.

II.3.1.2. CALCARE MASSICCIO FORMATION (UPPER NORIAN-SINEMURIAN)

In the internal part of the Northern Apennines, the Burano Anhydrites Formation is overlain by dark limestones and marls. These *Rhaetavicula contorta* Beds supposedly originated in a shallow-marine environment (Bortolotti *et al.*, 1970b). In the subsurface of the Umbro-Marchean Apennines, they are reduced to a thin and discontinuous interval at the base of the Calcare Massiccio Formation (Crescenti *et al.*, 1969; Jacobacci *et al.*, 1974; Centamore *et al.*, 1986).

The Calcare Massiccio Formation is characterized by massive platform carbonates. It displays two distinctive varieties (Centamore *et al.*, 1971, 1986; Jacobacci *et al.*, 1974). The essentially micritic Burano Calcare Massiccio resulted from subtidal sedimentation, which lasted until the Early Sinemurian. The M.Nerone Calcare Massiccio on the other hand contains oolitic and stromatolitic intercalations, attesting to a tidal flat environment. Its deposition persisted until the end of the Sinemurian.

The lateral variability of the Calcare Massiccio Formation was determined by differential subsidence in consequence of incipient listric extensional faulting (Parotto & Praturlon, 1975; Alvarez, 1989). Hereby, the Burano Anhydrites Formation probably acted as the basal detachment zone (Bally *et al.*, 1986).

The Calcare Massiccio Formation is presumed to maintain a rather invariable thickness of over 750 meters in the subsurface of the southern Montefeltro, irrespective of its affinity. Such a value is in line with the ones commonly envisaged for the northern part of the Umbro-Marchean Apennines (cf. Crescenti *et al.*, 1971)-

II.3.1.3. JURASSIC PELAGIC SUBSEQUENCE

II.3.1.3.1. INTRODUCTION

The progressive drowning of the Liassic carbonate platform determined the onset of carbonatic-marly pelagic sedimentation. This was accompanied by the accentuation of the irregularities resulting from extensional tectonics (Alvarez, 1989). Elevated segments of the platform thus evolved into seamounts. This initially permitted the prolonged deposition of the M.Nerone Calcare Massiccio. At a later stage, they received the condensed pelagic series which Centamore *et al.* (1971, 1986) and Jacobacci *et al.* (1974) named Bugarone Formation. Only during the Early Cretaceous, the submarine topography was leveled by the differential sedimentation.

A pronounced lateral variability thus characterizes the Jurassic Pelagic Subsequence, which ranges into the Lower Cretaceous. The limited areal extent of the condensed series displayed by the nearby Furlo and M.Nerone-M. Catria anticlines warrants the assumption of a complete series in the subsurface of the southern Montefeltro.

II.3.1.3.2. CORNIOLA FORMATION (UPPER SINEMURIAN-PLIENSBACHIAN)

The Middle Liassic Corniola Formation forms the basal part of the complete variety of the Jurassic Pelagic Subsequence. It is constituted by thin-bedded beige cherty micritic limestones with marls (*e.g.*, Carloni *et al.*, 1971; Centamore *et al.*, 1971, 1986; Jacobacci *et al.*, 1974). In the vicinity of seamounts, the lower part of the formation contains lenticular detrital intercalations. This *"marmarone"* is also exposed in the Montiego Anticline just a few kilometers S of the investigated area (Crescenti *et al.*, 1969). Differential subsidence is furthermore attested to by numerous slumps, olisthostromes and olistholiths (Alvarez, 1989).

The Corniola Formation attains thicknesses of more than 200 meters in the northern part of the Umbro-Marchean Apennines (Centamore *et al.*, 1986; Alvarez, 1989). In the subsurface of the southern Montefeltro, it is probably about 100 meters thick, considering the thicknesses of 50 to 150 meters reported from the Montiego Anticline (Crescenti *et al.*, 1969; Jacobacci *et al.*, 1974) and the 50 meters of Corniola Formation perforated by the Gambettola 1 well, some 25 kilometers to the N (De Francesco & Veggiani, 1967).

II.3.1.3.3. BOSSO FORMATION (TOARCIAN-BAJOCIAN)

The top part of the Corniola Formation and the lower part of the overlying Bosso Formation are locally substituted by the M.Serrone Marls (Pialli, 1969). This chert-bearing marly-calcareous interval is largely composed of detritus supplied by nearby seamounts. Considering its patchy distribution in the Umbro-Marchean Apennines, its significance for the subsurface of the southern Montefeltro is negligible.

The bipartite Bosso Formation consists of more or less nodular limestones and marls (Jacobacci *et al.*, 1974; Centamore *et al.*, 1986). Its Upper Liassic portion, known as Rosso Ammonitico, is characterized by a conspicuous red colour and abundant ammonites. The *Posidonia* Marls of Early Dogger age are predominantly greyish. They contain some calcarenitic intrcalations and, near the contact with the overlying Calcari Diasprini Formation, also nodular chert.

The complessive thickness of the Bosso Formation in the northern part of the Umbro-Marchean Apennines is about 50 meters (Jacobacci *et al.*, 1974; Centamore *et al.*, 1986). In the Gambettola 1 well N of the Montefeltro on the other hand, the Rosso Ammonitico alone measures already 50 meters (De Francesco & Veggiani, 1967). An average thickness of 65 meters is therefore feasible for the subsurface of the investigated area.

II.3.1.3.4. CALCARI DIASPRINI FORMATION (BATHONIAN-LOWER TITHONIAN)

Bortolotti *et al.* (1970b), Jacobacci *et al.* (1974) and Centamore *et al.* (1986) described the Middle to Upper Jurassic Calcari Diasprini Formation as an approximate equivalent of the Aptychi Limestones referred to by previous authors (*e.g.*, Selli, 1952; Ceretti, 1964). It is constituted by thin-bedded green to reddish granular limestones with abundant chert and marl partings. Radiolaritic intercalations characterize its middle part (Alvarez, 1989). Thicknesses envisaged for this formation in the surroundings of the Montefeltro typically vary from 50 to 100 meters (*e.g.*, Selli, 1952; De Francesco & Veggiani, 1967; Jacobacci *et al.*, 1974). An average of 75 meters can therefore be assumed for the subsurface of the area studied in detail.

II.3.1.3.5. MAIOLICA FORMATION (UPPER TITHONIAN-LOWER APTIAN)

The Maiolica Formation concludes the Jurassic Pelagic Subsequence, marking the establishment of sedimentary uniformity. It is composed of thin-bedded white micritic limestones with nodular chert. Detrital intercalations and slumps attest to the final stages of differential subsidence. Dark bituminous argillite partings, reflecting anoxic conditions, are not uncommon in the upper part of the formation. Interlaminated with variegated argillites and radiolarites, they even constitute an up to several meters thick interval near the contact with the overlying Marne a Fucoidi Formation. This Selli Level has been recognized throughout the Umbro-Marchean Apennines (Wezel, 1985; Coccioni *et al.*, 1987).

The average thickness of the Maiolica Formation in the northern part of the Umbro-Marchean Apennines is about 300 meters. Such a value seems appropriate for the subsurface of the southern Montefeltro as well.

II.3.1.4. CRETACEOUS-PALEOGENE PELAGIC SUBSEQUENCE

II.3.1.4.1. INTRODUCTION

General rather than differential subsidence determined the sedimentary evolution from the Early Cretaceous on. The ensueing pelagic deposits accordingly do not display marked lateral facies and thickness changes. Still, minor articulation did occur at times, as evidenced primarily by slumps (cf. Alvarez *et al.*, 1985; Guerrera *et al.*, 1988).

The Cretaceous to Paleogene Pelagic Subsequence, for convenience inclusive of the Lower Miocene top part of the Umbro-Marchean Sequence, is largely constituted by the Scaglia. This alternation of scaly limestones and marls displays a predominantly calcareous lower part and a more marly upper part. Besides marl contents, colour and the presence of chert are the main criteria for the distinction of its component formations (Renz, 1936; Selli, 1952).

II.3.1.4.2. MARNE A FUCOIDI FORMATION (MIDDLE APTIAN-UPPER ALBIAN)

At the base of the Cretaceous to Paleogene Pelagic Subsequence is the Marne a Fucoidi Formation. It consists of varicoloured marls with thin-bedded micritic limestones. Nodular chert occurs in the upper part of the formation. The characteristic presence of dark bituminous argillites relates to the significance of the Middle Cretaceous in the context of global anoxic events (Wezel, 1985).

The Marne a Fucoidi Formation probably measures about 70 meters in the subsurface of the southern Montefeltro, in conformity with the thickness Alvarez (1989) determined just a few kilometers to the S near Piõbbico.

II.3.1.4.3. SCAGLIA BIANCA FORMATION (UPPER ALBIAN-LOWER TURONIAN)

The Scaglia Bianca Formation is the lowest component of the Umbro-Marchean-Romagnan Succession exposed in the investigated area. A small outcrop in the F.Bottrina bed, 5.5 kilometers S of Peglio, displays its top at the hinge of the Montiego Anticline. More representative outcrops of this formation, which forms part of the Calcareous Scaglia, occur not much further S, in the surroundings of Piõbbico. It is largely composed of thin-bedded white to occasionally reddish micritic limestones, resulting from the diagenesis of a bathyal chalk ooze (de Boer, 1983). Greenish marl partings are typical of the lower part of the formation. Black and red nodular chert on the other hand is most common in its upper part (Selli, 1952; Ceretti, 1964).

A few meters below the contact with the Scaglia Rosata Formation, the Scaglia Bianca Formation contains an approximately one meter thick argillitic-radiolaritic laminar interval, which closely resembles the Selli Level developed at the top of the Maiolica Formation (cf. II.3.1.3.5.). This Bonarelli Level, which does not surface in the investigated area, has been distinguished throughout the Umbro-Marchean Apennines (*e.g.*, Renz, 1936; Selli, 1952; Ceretti, 1964; Bortolotti *et al.*, 1970b; Koopman, 1983; Centamore *et al.*, 1986). It supposedly marks the Cenomanian/Turonian boundary (Premoli Silva *et al.*, 1980; de Boer, 1983; Alvarez *et al.*, 1985).

The thickness of the Scaglia Bianca Formation averages 80 meters in the surroundings of Piõbbico. This value, which complies with the one de Boer (1980) determined further S, may also apply to the southern Montefeltro.

II.3.1.4.4. SCAGLIA ROSATA FORMATION (LOWER TURONIAN-LOWER THANETIAN)

The upward disappearance of black chert nodules marks the transition

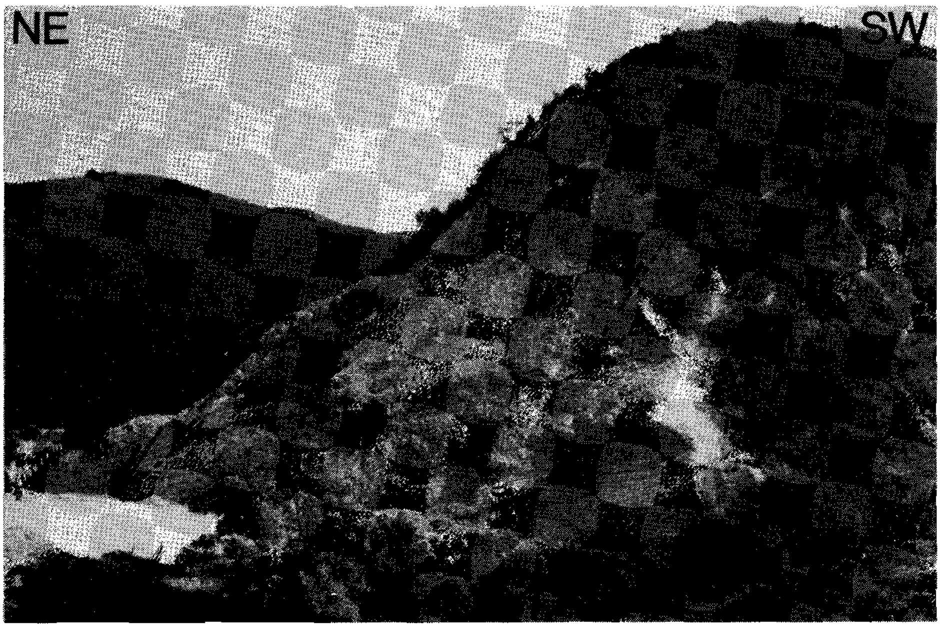


Fig. 9. The middle part of the Scaglia Rosata Formation in the quarry

of Ca Madonna, F. Bottrina valley, 5 km S of Peglio. Minor longitudinal normal faults are indicated.

from the Scaglia Bianca Formation to the Scaglia Rosata Formation. This predominantly reddish formation was distinguished as such by Dallan Nardi & Nardi (1975). It is well exposed in the Montiego Anticline, by natural outcrops as well as quarries for the production of road gravel (Fig. 9).

The Scaglia Rosata Formation consists largely of thin-bedded micritic limestones. These are similar to the ones of the Scaglia Bianca Formation, except for colour and subtle compositional differences (Vannucci *et al.*, 1979). Likewise, they resulted from the recrystallization of a bathyal chalk ooze (Premoli Silva *et al.*, 1974). Colour mottling generally attests to bioturbation (cf. Arthur, 1976). The limestone beds contain red nodular chert in

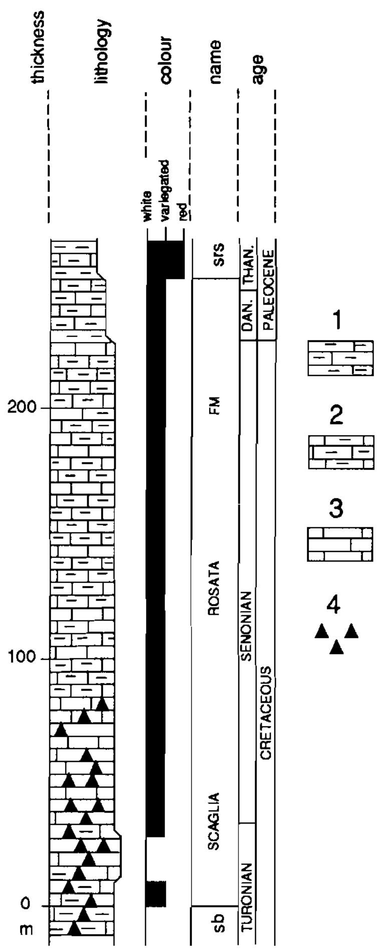


Fig. 10. Schematic columnar section of the Scaglia Rosata Formation in the Montiego Anticline. Legend: 1: marls and marly limestones; 2: micritic and marly limestones with marl partings; 3: micritic limestones; 4: nodular chert.

the lower part of the formation (Fig. 10). Marl and marly limestone beds are most common in its uppermost portion. They reflect a substantial decrease in carbonate productivity at the Cretaceous/Tertiary boundary (*e.g.*, Lowrie *et al.*, 1982; Napoleone *et al.*, 1983).

More or less thin-bedded calcarenites, characterized by a conspicuous white colour, occur throughout the Scaglia Rosata Formation. They are particularly evident in an about 20 meters thick interval in the basal part of the formation. The calcareous detritus presumably was derived from the Laziale-Abruzzese Carbonate Platform at the southeastern margin of the Umbro-Marchean-Romagnan Zone (Colacicchi & Baldanza, 1986). Volcaniclastic intercalations also stand out thanks to a white colouring. They are most evident near the Cretaceous/Tertiary boundary (cf. Vannucci *et al.*, 1979).

The Scaglia Rosata Formation measures about 250 meters in the F. Bottrina valley, 5.5 kilometers S of Peglio. This is in line with the thickness it attains in secions measured elsewhere in the northern part of the Umbro-Marchean Apennines (Renz, 1936; Selli, 1952; Arthur, 1976).

II.3.1.4.5. SCAGLIA ROSSA FORMATION (LOWER THANETIAN-LOWER LUTETIAN)

The top part of the Calcareous Scaglia is constituted by the Scaglia Rossa Formation (cf. Dallan Nardi & Nardi, 1975). It is markedly bipartite (Fig. 11). Brick-red platy marls and marly limestones dominate its Upper Paleocene portion. This is overlain by thin-bedded red micritic and marly limestones with abundant red nodular chert. Colour mottling is not uncommon.

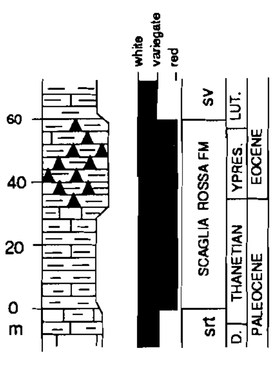


Fig. 11. Schematic columnar section of the Scaglia Rossa Formation in the Montiego Anticline. Legend as in Fig. 10.

The thickness of the Scaglia Rossa Formation in the Montiego Anticline is about 40 to 70 meters. Similar values have been inferred in neighbouring segments of the Umbro-Marchean Apennines (*e.g.*, Renz, 1936; Selli, 1952; Lowrie *et al.*, 1982; de Feyter *et al.*, 1986).

II.3.1.4.6. SCAGLIA VARIEGATA FORMATION (LOWER LUTETIAN-LOWER PRIABONIAN)

A marked increase in terrigenous supply during the Middle Eocene determined the onset of the deposition of the Marly Scaglia. The relative transition from the Scaglia Rossa Formation to the Scaglia Variegata Formation coincides with the upward disappearance of nodular chert (cf. Lowrie *et al.*, 1982; Guerrera *et al.*, 1988). The Scaglia Variegata Formation is composed of more or less calcareous marls with thin-bedded marly limestones (Fig. 12). Its marl contents decreases upwards.

Fig. 12. Variegated marls with thin white marly limestone beds in the lower part of the Scaglia Variegata Formation near M.Ciolino, 4.5 km ESE of S.Angelo in Vado. Hammer in centre of photograph for scale.

The name of the Scaglia Variegata Formation refers to its red to white variegated aspect. This characterizes the lower and the uppermost part of the formation. In between is an entirely greyish interval (Fig. 13; Guerrera *et al.*, 1988).

In the Montiego Anticline, the Scaglia Variegata Formation averages 75 meters in thickness (cf. Guerrera *et al.*, 1988). Similar values hava been determined elsewhere in the northern part of the Umbro-Marchean Apennines by Renz (1936) and Selli (1952).

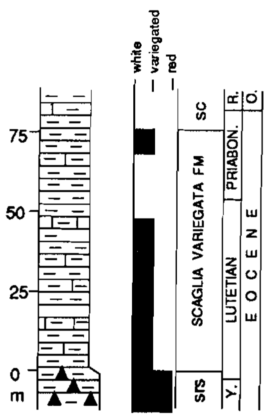


Fig. 13. Schematic columnar section of the Scaglia Variegata Formation in the Montiego Anticline. Legend as in Fig. 10.

II.3.1.4.7. SCAGLIA CINEREA FORMATION (LOWER PRIABONIAN-LOWER BURDIGALIAN)

The Scaglia Cinerea Formation is the major constituent of the Marly Scaglia. As indicated by its name, it is characterized by an ash-grey colouring. Only some 25 meters above the contact with the Scaglia Variegata Formation, it displays a thin reddish interval (Fig. 14; Guerrera *et al.*, 1988). This has also been encountered in adjacent areas (de Feyter *et al.*, 1986).

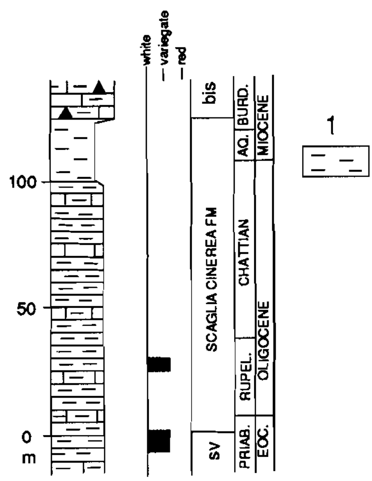


Fig. 14. Schematic columnar section of the Scaglia Cinerea Formation in the Montiego Anticline. Legend: 1: clayey marls; otherwise as in Fig. 10.

The Scaglia Cinerea Formation is composed of marls with thin marly limestone beds. Both the frequency of the latter and the carbonate contents of the marls typically decrease upwards (*e.g.*, Renz, 1932, 1933; Selli, 1952; Carloni, 1962; Ceretti, 1964; Baumann & Roth, 1969; Lowrie *et al.*, 1982; Centamore *et al.*, 1986). Thin coarse calcarenite beds are not uncommon in the lower part of the formation. They display a characteristic friable weathering. Similar bioclastic intercalations, albeit of greater thickness, occur in the southern segment of the Umbro-Marchean Apennines (*e.g.*, Baumann, 1970). Therefore, the Laziale-Abruzzese Carbonate Platform probably supplied the detritus (Centamore *et al.*, 1986).

The upper 25 meters of the Scaglia Cinerea Formation consist of blue-greyish clayey marls which are virtually devoid of calcareous intercalations (cf. Micarelli, 1969; Jacobacci *et al.*, 1970, 1974). They bear a striking resemblance to the clayey marls of the Schlier Formation at the base of the Umbro-Romagnan turbidite suite (cf. II.3.2.1.2.). Thus, they mark the onset of foredeep activity replacing pelagic foreland sedimentation.

The thickness of the Scaglia Cinerea Formation in the Montiego Anticline ranges from 80 to 170 meters, averaging 125 meters. Such a variability occurs throughout the Umbro-Marchean Apennines (*e.g.*, Selli, 1952; Ceretti, 1964; Centamore *et al.*, 1986). On a regional scale, this may be a primary feature (Carloni, 1962; Baumann, 1970). With reference to the investigated area however, it essentially relates to orogenic shearing.

II.3.1.4.8. BISCIARO FORMATION (LOWER BURDIGALIAN-LOWER LANGHIAN)

The Lower Miocene Bisciaro Formation forms the top part of the Umbro-Marchean Sequence. It consists of yellow more or less marly limestones and blue-greyish clayey marls. In local dialect, these lithotypes are known as respectively *"bisciaro"* and *"genga"* (Principi, 1931, 1939; Selli, 1952). The variability of their ratio determines the tripartite character of the formation

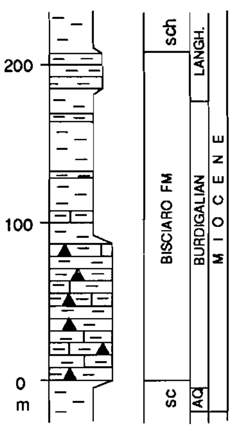


Fig. 15. Schematic columnar section of the Bisciaro Formation in the Montiego Anticline. Legend as in Figs. 10 and 14.

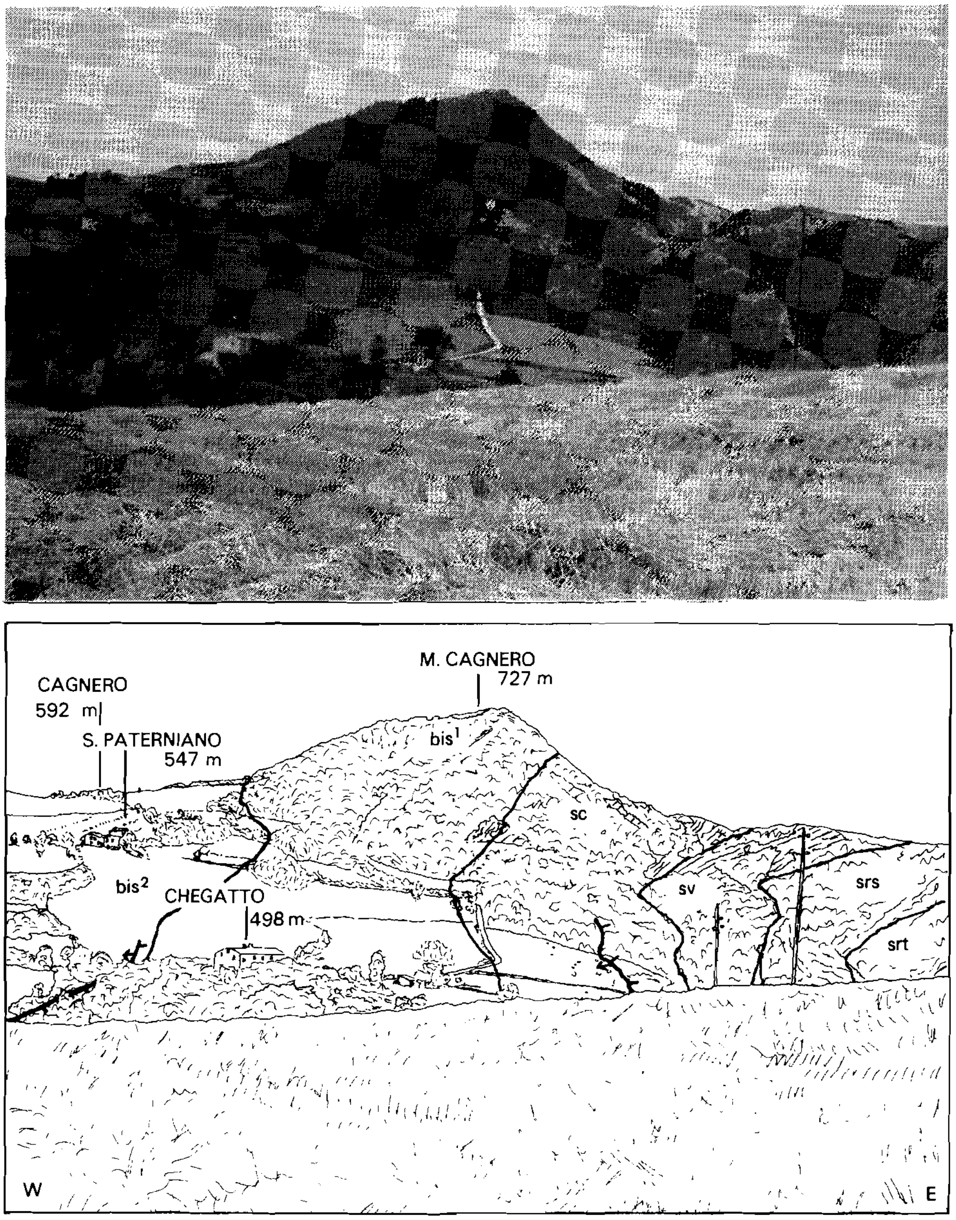


Fig. 16. Photograph and sketch of M.Cagnero, 4.5 km SE of S.Angelo in Vado. Key: srt: Scaglia Rosata Formation; srs: Scaglia Rossa Formation; sv: Scaglia Variegata Formation; sc: Scaglia Cinerea Formation; Bisciaro Formation: bis^1: lower member; bis^2: middle member.

(Fig. 15). This has also been recognized in neighbouring areas (Jacobacci *et al.*, 1974; Guerrera, 1977; de Feyter *et al.*, 1986).

The about 90 meters thick lower member of the Bisciaro Formation as a rule stands out topographically (Fig. 16). It is largely composed of thin- to medium-bedded pelagic limestones with dark chert lenses and nodules (Fig. 17). Calcarenitic intercalations of turbiditic origin, identified in the vicinity of the southern Montefeltro by Guerrera (1977, 1978) and Capuano *et al.* (1987b) have not been encountered (cf. Ricci Lucchi, 1981a).

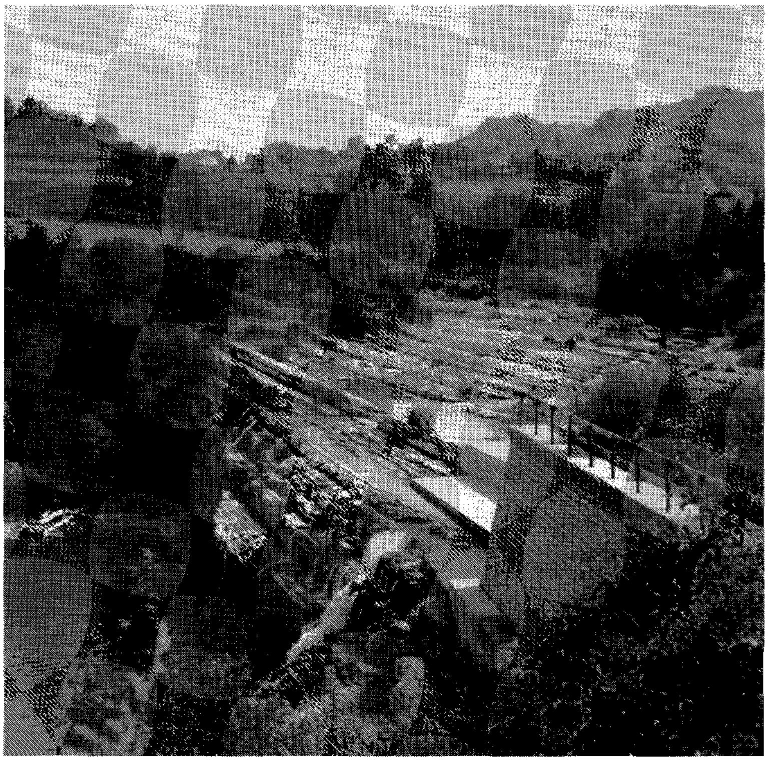


Fig. 17. The regularly bedded calcareous lower member of the Bisciaro Formation in the F.Metauro bed near il Sasso, 2 km E of S.Angelo in Vado. The ledge of the fall is underlain by the soft top part of the Scaglia Cinerea Formation.

Clayey marls prevail in the middle member of the Bisciaro Formation. They resemble the ones constituting the top part of the Scaglia Cinerea Formation and the Schlier Formation. Likewise, they are supposedly of turbiditic origin (cf. II.3.2.1.2.). The thickness of this member varies according to structural position. It measures well over 100 meters in the limbs of the Montiego Anticline, but is reduced to 50 meters or less in the anticlinal crest. The Bisciaro Formation displays similar systematic thickness variations throughout the Umbro-Marchean Apennines (*e.g.*, Guerrera, 1977; Cantalamessa *et al.*, 1986). They generally argue the synsedimentary nucleation of the principal longitudinal structures (Toni & Ardanese, 1981).

In the poorly exposed uppermost part of the Bisciaro Formation, the clayey marls alternate with thin-bedded pelagic limestones. This member attains a thickness of approximately 25 meters. Mainly due to the thinning of its marly constituents, this is reduced to a few meters in the crestal zone of the Montiego Anticline.

Dark and ochreous volcaniclastic intercalations are a distinguishing feature of the Bisciaro Formation. They determine its generally brownish weathering. Volcaniclastics furthermore occur dispersed in the regular lithotypes. Most common are more or less altered glass shards and plagioclase crystal fragments. They are of rhyodacitic character, resulting from essentially submarine explosive calcalkaline volcanism (Mezzetti & Olivieri, 1964; Mezzetti, 1969; Ardanese & Grandi, 1977; Guerrera, 1977, 1978, 1982a, 1982b; Borsetti *et al.*, 1979, 1982-1983). Similar volcaniclastic intercalations have been reported from the temporal equivalents of the Bisciaro Formation throughout the Northern Apennines (cf. Guerrera & Veneri, 1989). The pertinent major eruptive centres supposedly were situated at the internal margin of the Tuscan Zone (Mezzetti, 1969; Borsetti *et al.*, 1979, 1982-1983; Guerrera, 1982a, 1982b).

The massive supply of volcaniclastics probably was the prime cause of blooms of calcareous and siliceous microplankton. Accordingly, the Early Miocene volcanism indirectly determined the constitution of the Bisciaro Formation through periodically increased pelagic settling superimposed on continuous clayey marly turbiditic sedimentation.

II.3.2. THE UMBRO-ROMAGNAN SEQUENCE

II.3.2.1. MIDDLE-UPPER MIOCENE TURBIDITIC SUBSEQUENCE

II.3.2.1.1. INTRODUCTION

The Middle to Upper Miocene Turbiditic Subsequence comprises the tripartite turbidite suite of the Umbro-Romagnan Foredeep. It is characterized by a thick arenaceous-marly orthoturbiditic series which is enclosed by essentially marly proto- and cataturbiditic deposits (cf. Sagri, 1973).

The Umbro-Romagnan turbidite suite displays a systematic lateral variability clearly related to synsedimentary tectonics. Thus, its component intervals are markedly diachronous at principal longitudinal structures such as the thrust front of the Umbro-Romagnan parautochthon (de Feyter *et al.*, 1986, 1990). Its essence however does not change. Accordingly, the Middle to Upper Miocene Turbiditic Subsequence encompasses both parautochthonous and palaeoautochthonous terrains.

II.3.2.1.2. SCHLIER FORMATION (LOWER LANGHIAN-MIDDLE SERRAVALLIAN)

The prototurbiditic Schlier Formation constitutes the basal part of the Middle to Upper Miocene Turbiditic Subsequence. It is composed of faintly laminated blue-greyish clayey marls with dark partings of organic material and more calcareous intercalations. The name of the formation refers to its resemblance to coeval deposits of the Vienna Basin (Selli, 1952; R.Selli *in* Selli, 1967).

The top of the Schlier Formation youngs systematically towards the NE at the expense of the orthoturbiditic Marnoso-arenacea Formation, reflecting the migratory character of foredeep activity (de Feyter *et al.*, 1990). It is of Middle Langhian age in the internal segment of the Umbro-Romagnan parautochthon. More external parautochthonous units contain an about 200 meters thick Schlier Formation which also encompasses part of the Upper Langhian. Roughly in correspondence with the thrust front of the parautochthon, its top attains a Middle Serravallian age (cf. Centamore & Jacobacci, 1968; Micarelli, 1969; Jacobacci *et al.*, 1974; Centamore *et al.*, 1977, 1978; de Feyter *et al.*, 1986).

In the palaeoautochthon of the investigated area, the well-exposed Schlier Formation averages 375 meters in thickness (Fig. 18). Thin-bedded whitish calcareous marls, concentrated in intervals, are not uncommon in its middle to upper part (Fig. 19). At the adjoining Montescudo-Serrungarina High, the top of the Schlier Formation again displays a pronounced diachroneity, resulting in Tortonian to Messinian ages more externally (Angeli & Veggiani, 1967; Carloni *et al.*, 1967; Cantalamessa *et al.*, 1986).

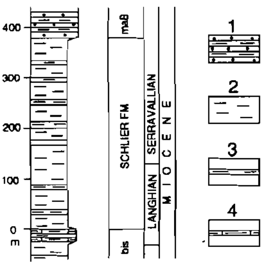


Fig. 18. Schematic columnar section of the Schlier Formation in the palaeoautochthon of the investigated area. Legend: 1: alternating sandstones and clayey marls; 2: clayey marls; 3: calcareous marls; 4: marly limestones.

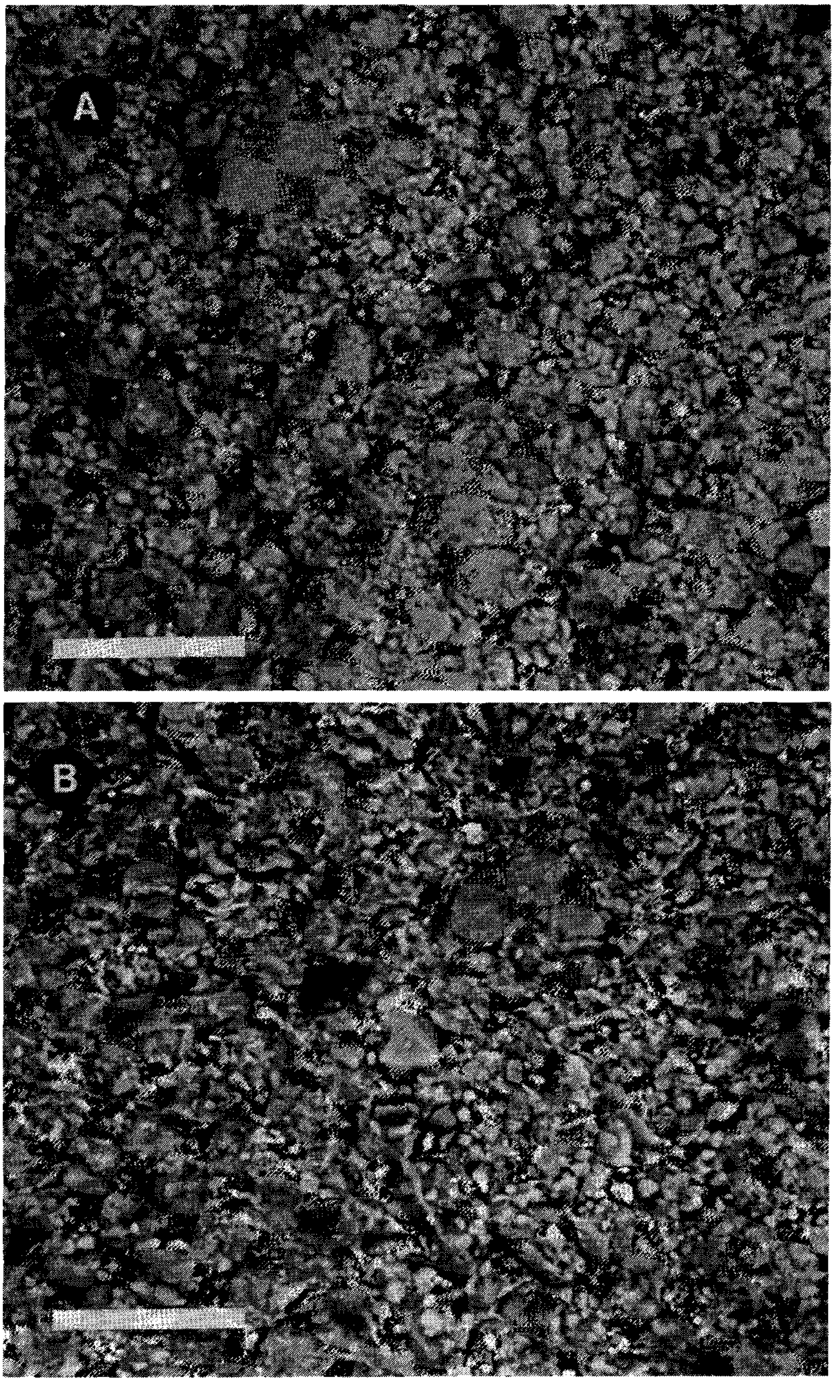
The Schlier Formation generally contains rich planktonic microfaunas attesting to a bathyal depositional environment (Carloni *et al.*, 1967; Cantalamessa *et al.*, 1986). This is compatible with the macrofaunal elements, in particular molluscs such as pteropods, reported by Principi (1925a, 1931, 1939), Ruggieri (1970) and Carloni *et al.* (1971).

The homogeneous clayey marls of the Schlier Formation are comparable with the unifites described by Feldhausen *et al.* (1981) and Stanley (1983). Accordingly, they must have been emplaced by low-density turbidity currents which had assimilated settling hemipelagic material rather than in the purely hemipelagic fashion inferred by Selli (1949, 1952), Ricci Lucchi & Pialli

(1973), Ricci Lucchi (1975a), Centamore *et al.* (1977) and Conti (1989). Sedimentation from the dilute marginal parts of regular turbidity currents primarily applies to marked heteropies, representing a mere slope effect. De Feyter *et al.* (1986) documented this for the frontal sector of the Umbro-Romagnan parautochthon immediately S of the investigated area, where the northeastward petering out of the arenaceous portion of orthoturbidites accounts for the diachronous aspect of the top of the Schlier Formation. As indicated by the relative uniformity of the Schlier Formation in the Umbro-Marchean-Romagnan Zone, the deposition of the bulk of the clayey marls however must be attributed to separate low-density turbiditity currents, according to the concept of turbid layers outlined by Moore (1969). The general southeastward thinning of the formation argues an essentially longitudinal dispersal (Crescenti *et al.*, 1969). In conformity with the Marnoso-arenacea

Fig. 19. Clayey marls alternating with more resistant thin-bedded calcareous marls in the middle part of the Schlier Formation between Ca Braccio and M.Ca Bertino, 2 km S of Sassocorvaro.

Fig. 20. Scanning electron microscope back-scattered electron images of Lower to Middle Serravallian calcareous marls of the Umbro-Romagnan turbidite suite. Their turbiditic origin is indicated by the random orientation of clay mineral flakes (cf. O'Brien et al., 1980). Scale bars are 30 microns. A: prototurbidite of Schlier Formation near Ca Vieto, 4 km SSE of S.Angelo in Vado; B: top part of orthoturbiditic colombina of Marnoso-arenacea Formation A near Metola, 3.5 km SW of S.Angelo in Vado.


A
B

Formation, the terrigenous fines probably were supplied from the NW by Austro-Alpine sources (cf II.3.2.1.3.).

The turbiditic origin of the thin-bedded calcareous marls characterizing the Lower to Middle Serravallian portion of the Schlier Formation in the palaeoautochthon is indicated by their resemblance to the marly intervals of coeval calcareous orthoturbidites displayed by the Marnoso-arenacea Formation A more internally (Fig. 20). By analogy, the Laziale-Abruzzese Carbonate Platform SE of the Umbro-Romagnan Foredeep must have supplied the calcareous clastics (cf. II.3.2.1.3.). This is corroborated by the presence of proximal calcarenitic intercalations, or "*cerrogne*", in the southeastern part of the Umbro-Marchean Apennines (*e.g.*, Cantalamessa *et al.*, 1986). In this context, the longitudinal dispersal pattern can be classified as asymmetric bimodal (Fig. 21A).

II.3.2.1.3. MARNOSO-ARENACEA FORMATION (UPPER LANGHIAN-LOWER TORTONIAN)

The widely exposed orthoturbiditic Marnoso-arenacea Formation is the principal constituent of the Umbro-Romagnan turbidite suite. It displays a systematic lateral variability, primarily determined by the intermittent shifting of foredeep activity towards the Adriatic Foreland (Fig. 21). Thus, an Upper Langhian to Lower Tortonian variety characterizing the external por-

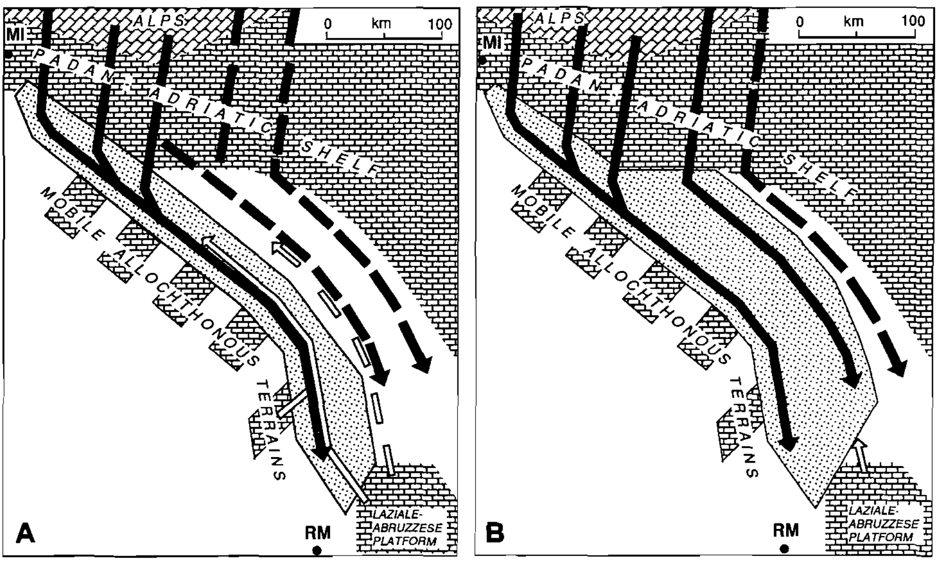


Fig. 21. Turbiditic dispersal patterns of the Umbro-Romagnan Foredeep during the Late Langhian to Middle Serravallian (A) and during the Middle Serravallian to Early Tortonian (B), inspired by Ricci Lucchi (1975a,

1978; 1981b). Interrupted and continuous trajectories relate to proto-turbiditic and orthoturbiditic sedimentation, respectively. The depicted geometry of the foredeep reflects the actual position of its clastic wedge. Approximate locations of Milano (MI) and Roma (RM) are indicated for reference.

tion of the Umbro-Romagnan parautochthon can be distinguished from a Middle Serravallian to Lower Tortonian one in the contiguous palaeoautochthon. With reference to Signorini (1943, 1945a), who labeled the major orthoturbiditic series of the Northern Apennines alphabetically, these varieties are designated as Marnoso-arenacea Formation A and Marnoso-arenacea Formation B, respectively (de Feyter *et al.*, 1986).

The Marnoso-arenacea Formation A attains a thickness of several kilometers in the investigated area. Its bathyal character is indicated by trace fossils such as *Paleodictyon* (Principi, 1931, 1934). In accordance with the general picture outlined by ten Haaf (1959), most turbidites were supplied longitudinally from the NW (Fig. 22). They consist of brown-greyish quartzose-feldspathic litharenites grading upward into blue-greyish clayey marls. Vegetal matter is not uncommon in the uppermost part of the sandstone beds (Fig. 23). The thickness of the turbidites ranges from less than 40 centimeters for thin-bedded ones to more than a meter in the case of megaturbidites (cf. Ricci Lucchi, 1981b). They generally display base-missing and incomplete Bouma sequences (Fig. 24).

The composition of the northwesterly derived turbidites, marked by the presence of detrital dolomite, attests to an Austro-Alpine provenance (Gazzi, 1965; Gandolfi *et al.*, 1983; Valloni & Zuffa, 1984). Submarine canyons supposedly permitted the transport of clastics across the Padan-Adriatic Shelf (Fig. 21A). At their entry points in the northwestern segment of the Umbro-Romagnan Foredeep, these developed submarine fans (*e.g.*, Ricci Lucchi, 1975a, 1975b, 1978, 1981b).

The lower part of the Marnoso-arenacea Formation A comprises minor amounts of turbidites emplaced by currents flowing longitudinally towards the NW (Fig. 22). Two particularly conspicuous megaturbidites of Early Serravallian age, situated about 575 meters vertically apart, display 5 to 6 meters of brown-greyish calcareous-quartzose litharenite grading upward into 8 meters of blue-greyish clayey marl (Fig. 25). Both are characterized by a base-missing Bouma sequence (T_{b-e}). The upper layer represents the Contessa megaturbidite defined by Ricci Lucchi & Pialli (1973) in the surroundings of Gubbio, 25 kilometers S of the investigated area (cf. Ardanese *et al.*, 1982-1983). It constitutes a marker of regional significance (*e.g.*, Bortolotti *et al.*, 1970b; Ricci Lucchi, 1975a, 1978, 1981b; Ricci Lucchi & Valmori, 1980; van Wamel & Zwart, 1990). The lower layer, which occurs some 250 meters above the base of the Marnoso-arenacea Formation A, has a similar extent. This warrants the designation of these megaturbidites as Contessa II and Contessa I, respectively.

The Contessa megaturbidites are typified compositionally by shallow-marine fossil fragments and glauconite, indicative of an Apenninic provenance (Centamore *et al.*, 1977; Ardanese *et al.*, 1982-1983; Gandolfi *et al.*, 1983). Clastics were presumably supplied by mobile allochthonous terrains at the inner margin of the Umbro-Romagnan Foredeep (Fig. 21A; Ricci Lucchi &

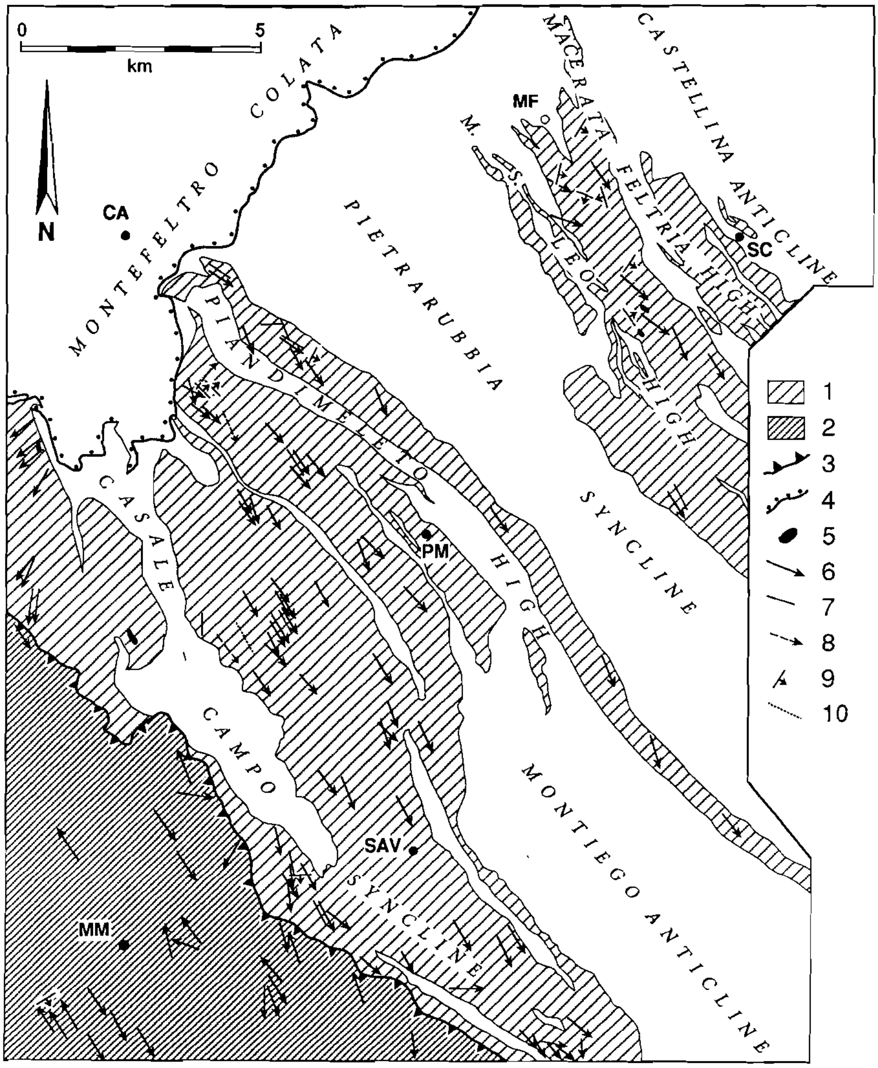


Fig. 22. Palaeocurrent pattern of the Marnoso-arenacea Formation in the investigated area. Legend: 1: Marnoso-arenacea Formation A; 2: Marnoso-arenacea Formation B; 3: major thrust fault, with barbs on upper block; 4: transgressive thrust fault, with dots on upper block; 5: slump; 6: palaeocurrent direction indicated by flute casts; 7: palaeocurrent direction indicated by groove casts; 8: palaeocurrent direction indicated by current ripples; 9: attitude of systematically heeling crests of convolute lamination; 10: direction of aligned entrapment burrows. Location labels: CA: Carpegna; MF: Macerata Fèltria; MM: Mercatello sul Metauro; PM: Piandimeleto; SAV: S.Angelo in Vado; SC: Sassocorvaro.

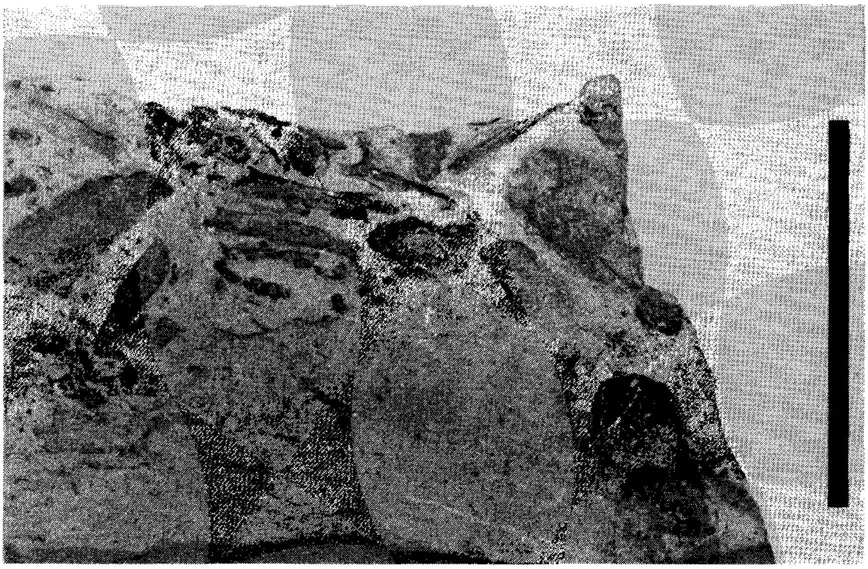


Fig. 23. Top part of turbiditic sandstone bed of northwestern origin, containing abundant vegetal matter. Sample from Marnoso-arenacea Formation A near Ca Franceschino, 2 km NE of Mercatello sul Metauro. Scale bar is 4 cm.

Pialli, 1973; Ricci Lucchi, 1975a, 1975c, 1978, 1981b, 1987). According to the concept of seismoturbidites, tectonically induced gravitational failures must have originated the turbidity currents which deposited the Contessa layers (Mutti *et al.*, 1984).

Light-coloured calcareous turbidites form another lithotype of southern derivance. These "*colombine*" are most evident in the Lower to Middle Serravallian portion of the Marnoso-arenacea Formation A, overlying the Contessa II megaturbidite. The generally megaturbiditic layers are constituted by calcareous litharenites grading upward into calcareous marls. Base-missing and incomplete Bouma sequences prevail (cf. Statera & Ricci Lucchi, 1981; Ardanese *et al.*, 1982-1983). The *colombine* are largely composed of shallow-marine fossil fragments and micrite clasts (Parea, 1967; Ardanese *et al.*, 1982-1983; Gandolfi *et al.*, 1983). They were supplied by the Laziale-Abruzzese Carbonate Platform SE of the Umbro-Romagnan Foredeep (Fig. 21A; Ricci Lucchi & Pialli, 1973; Ricci Lucchi, 1975a, 1978, 1981b). Like the Contessa megaturbidites, the *colombine* represent seismoturbidites unrelated to submarine fans. The concentration in intervals they have in common with the coeval prototurbidites of Laziale-Abruzzese provenance contained by the Schlier Formation more externally (cf. II.3.2.1.2.) accordingly reflects periodical increases of tectonic activity.

The bimodal dispersal pattern and the impressive lateral continuity displayed by individual turbidites in the lower part of the Marnoso-arenacea Formation A are diagnostic of sedimentation on a subhorizontal basin plain

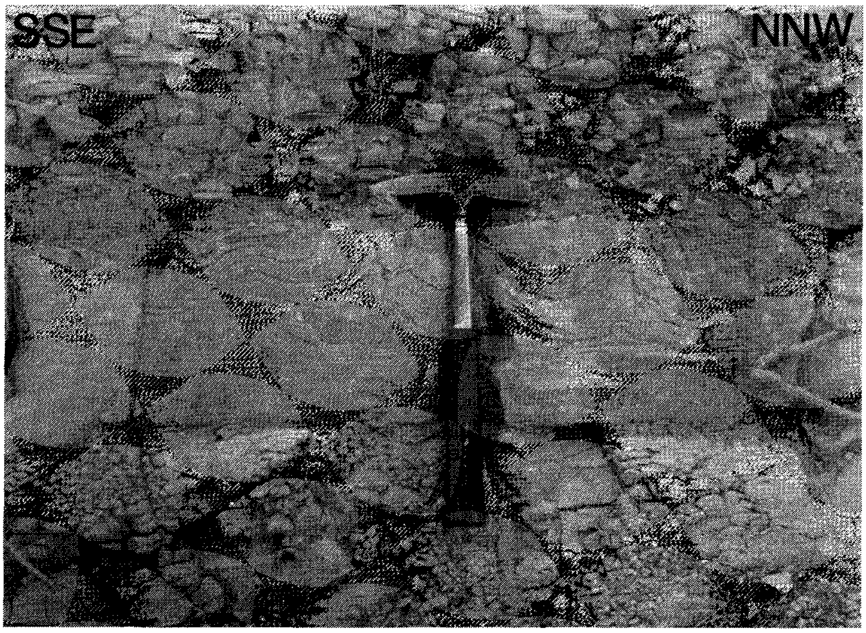


Fig. 24. Typical thin-bedded turbidite of northwestern origin, displaying a base-missing incomplete Bouma sequence (T_{ce}). In its arenaceous portion, current rippling grades upward into convolute lamination. The crests of the latter heel with the inferred palaeocurrent direction. Marnoso-arenacea Formation A in bed of T.S.Antonio near Cse.Sassorotto, 2 km SW of Mercatello sul Metauro.

(Parea, 1967; Ricci Lucchi, 1975c; Ricci Lucchi & Valmori, 1980). This agrees with the general downcurrent heeling of the crests of convolute laminations (Fig. 24; cf. Statera & Ricci Lucchi, 1981).

Minor slumps restricted to the frontal segment of the Umbro-Romagnan parautochthon mark the outer margin of the basin plain (Fig. 20). Up to 20 centimeters thick ochreous siltite lenses, resulting from grain flows, also document a pronounced palaeoslope (cf. Lowe, 1976, 1979, 1982). The concomitant northeastward thinning of orthoturbidites relates primarily to their arenaceous portions (Fig. 27). This determines the heteropical contact with the Schlier Formation (cf. II.3.2.1.2.).

In terms of the descriptive classification de Jager (1979) elaborated from the facies model outlined by Mutti & Ricci Lucchi (1972), the Marnoso-arenacea Formation A pertains entirely to facies association *e* in the investigated area. This is characterized by the predominance of laterally rather consistent turbidites displaying base-missing and incomplete Bouma sequences and a sandstone/marl ratio of less than 1. The application of a more refined classification permits to accurately identify the outer margin of the basin plain (Fig. 27, 28). Facies association e_1, representing the basin plain depositional environment, is largely constituted by turbidites whose basal part con-

Fig. 25. Arenaceous portion of Contessa I megaturbidite, showing large flute casts, in the lower part of the Marnoso-arenacea Formation A along the F.Metauro, 1 km E of Mercatello sul Metauro. Hammer for scale.

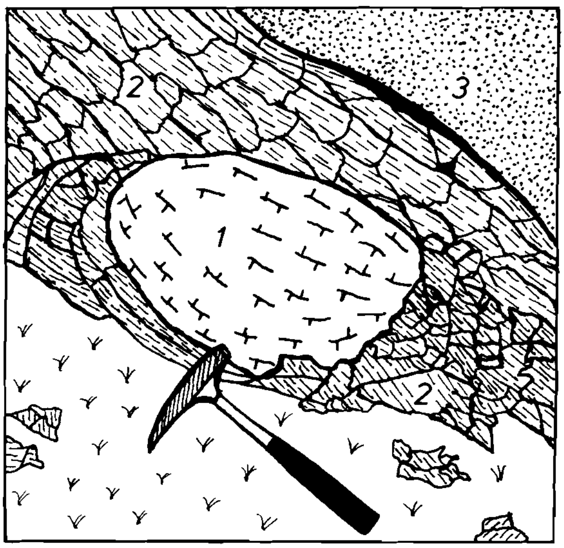


Fig. 26. Drawing from photograph of slump ball of calcareous marl (1)

embedded in clayey marls (2), which are overlain by a colombina (3). Marnoso-arenacea Formation A along T.Morsina near Metola, 3.5 km SW of S.Angelo in Vado.

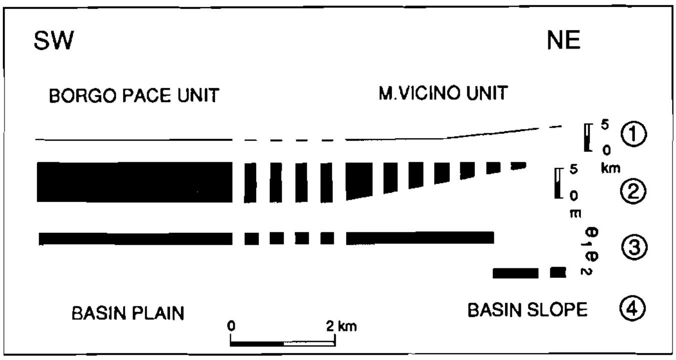


Fig. 27. Palinspastically restored Late Langhian to Middle Serravallian foredeep morphology and related sedimentary features of the Marnoso-arenacea Formation A in the investigated area, interrupted where inferred. Key: 1: submarine topography; 2: thickness of arenaceous portion of Contessa II megaturbidite; 3: facies associations; 4: depositional environments.

sists of the *c*-interval of the Bouma sequence, while the sandstone/marl ratio varies from 0.25 to 1. The basin slope is indicated by facies association e_2, with more turbidites commencing with the *d*-interval of the Bouma sequence and a sandstone/marl ratio of 0.1 to 0.25.

The upper part of the Marnoso-arenacea Formation A contains solely turbidites of northwestern origin. These exhibit a lateral variability indicative of the gradual articulation of the basin plain from the Middle Serravallian on. Thus, relatively arenaceous lithofacies occupy the principal synclinal hinge of the Borgo Pace Unit near Guinza, just SW of the investigated area.

A more pronounced lateral variability as a function of structural position, superposed on a general northeastward fining and thinning, is displayed by the coeval Marnoso-arenacea Formation B. Maximum thicknesses of more than 2 kilometers in the Casale-Campo Syncline and more than 1 kilometer in the Pietrarubbia Syncline contrast with minima of less than 250 meters at major highs.

With respect to sedimentary features, the Marnoso-arenacea Formation B does not differ substantially from the upper part of the Marnoso-arenacea Formation A. It is also entirely composed of turbidites of Austro-Alpine provenance, supplied longitudinally from the NW (Fig. 21B, 22). Poorly cemented sandstone beds with spheroidal concretions are not uncommon in the upper part of the formation. Megaturbidites typically show diffuse laminations (Fig. 29; cf. Ricci Lucchi, 1981b). The irregular motion of laterally confined voluminous turbidity currents probably determined the longitudinally discontinuous character of these layers (cf. enclosed map).

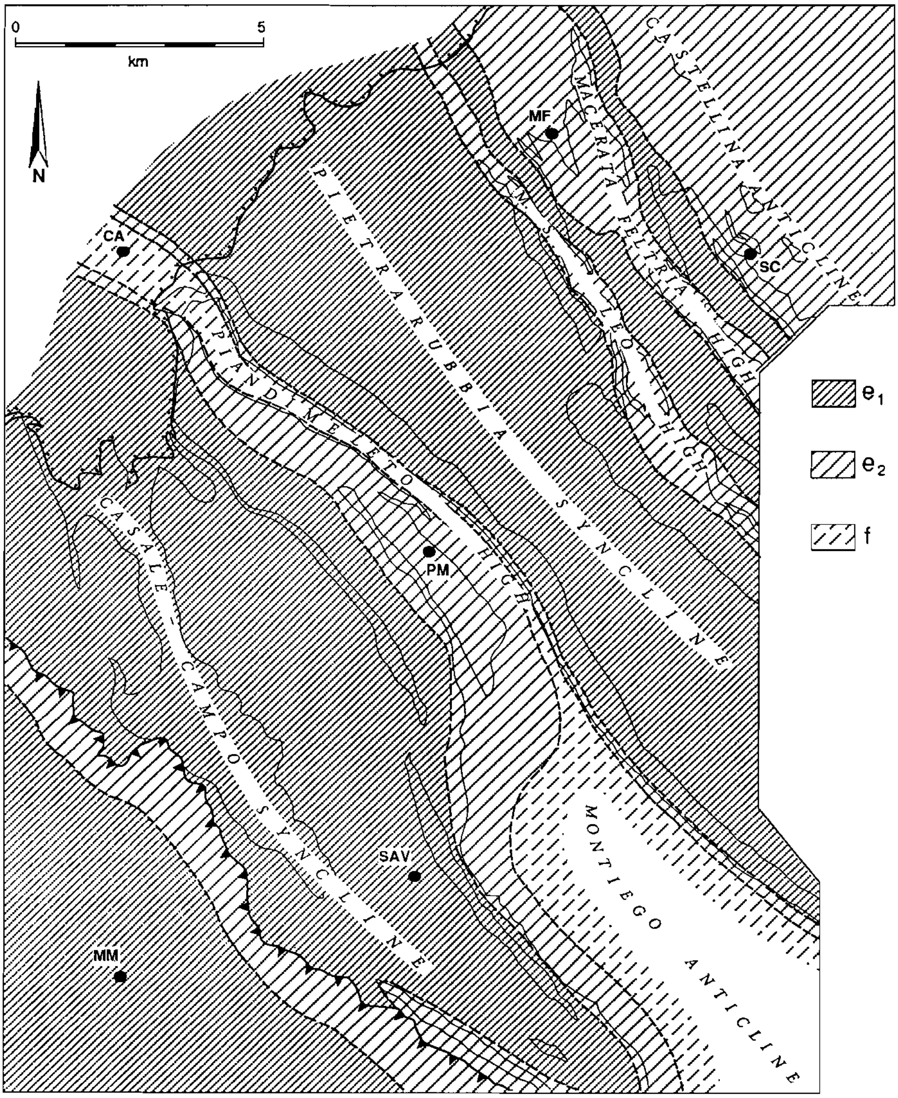


Fig. 28. Distribution of facies associations of the Marnoso-arenacea Formation in the investigated area. Structural symbols and location labels as in Fig. 22.

The heeling direction of the crests of convolute laminations and the palaeocurrent direction indicated by current ripples essentially reflect the inferred palaeotopography (Fig. 22). Longitudinal reliefs apparently induced transversal flow in the tail of turbidity currents (cf. Kuenen, 1967; Statera &

Ricci Lucchi, 1981). Slumps are relatively rare.

Fig. 30. Megaturbidite of northwestern origin in the Marnoso-arenacea Formation B along il Fossato, 1.5 km S of Belforte all'Isauro. The wavy aspect of the prominent diffuse lamination displayed by its 4.7 m thick arenaceous portion relates to fluid-escape structures.

The articulated foredeep morphology is reflected by the distribution of facies associations (Fig. 28). Facies association e_1 attests to the depocentral setting of the major synclines. The bordering slopes are marked by facies association e_2. The condensed series with sandstone/marl ratio of less than 0.1 characterizing the Piandimeleto High represent facies association *f* (cf. de Jager, 1979).

II.3.2.1.4. S.PAOLO AND CAMPO MARL FORMATIONS (LOWER TORTONIAN)

Lower Tortonian cataturbiditic terrains largely composed of greyish clayey marls conclude the Umbro-Romagnan turbidite suite SW of the Pietrarubbia Syncline. They reflect the outward shifting of orthoturbiditic sedimentation in consequence of accelerated foredeep differentiation. The relative S.Paolo and Campo Marl formations accordingly display a marked lateral

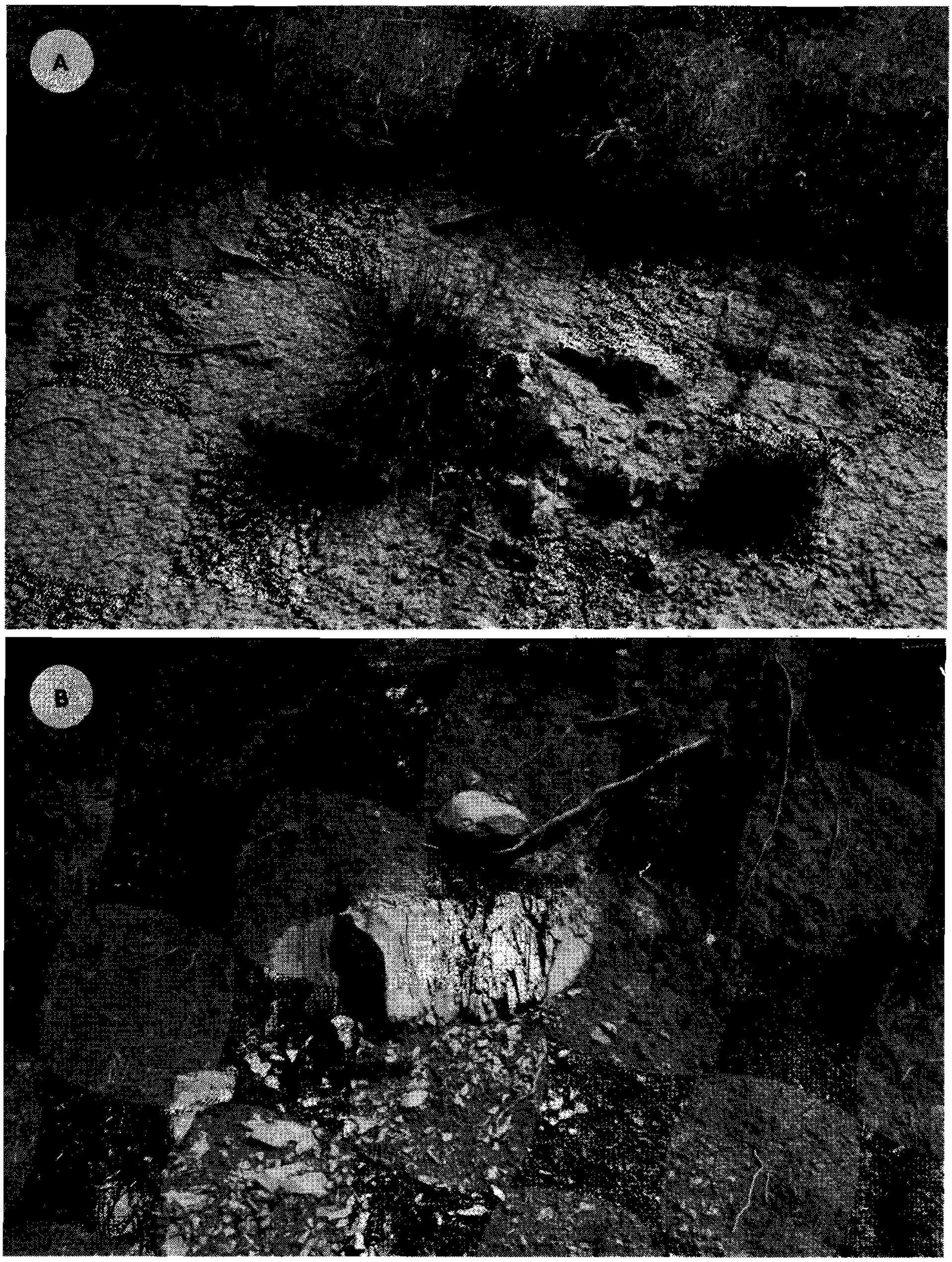


Fig. 30. Slumped siltites embedded in clayey marls of the Campo Marl Formation near Campo, 3 km SW of Belforte all'Isauro (A), and at M.del Prete, 4 km NW of S.Angelo in Vado (B). Hammer for scale.

variability. Thick bathyal deposits of major synclinal depocentres grade into shallower condensed series at intervening highs. Ochreous siltites emplaced by minor grain flows also attest to an articulated morphology (Fig. 30). Moreover, slumps are particularly common.

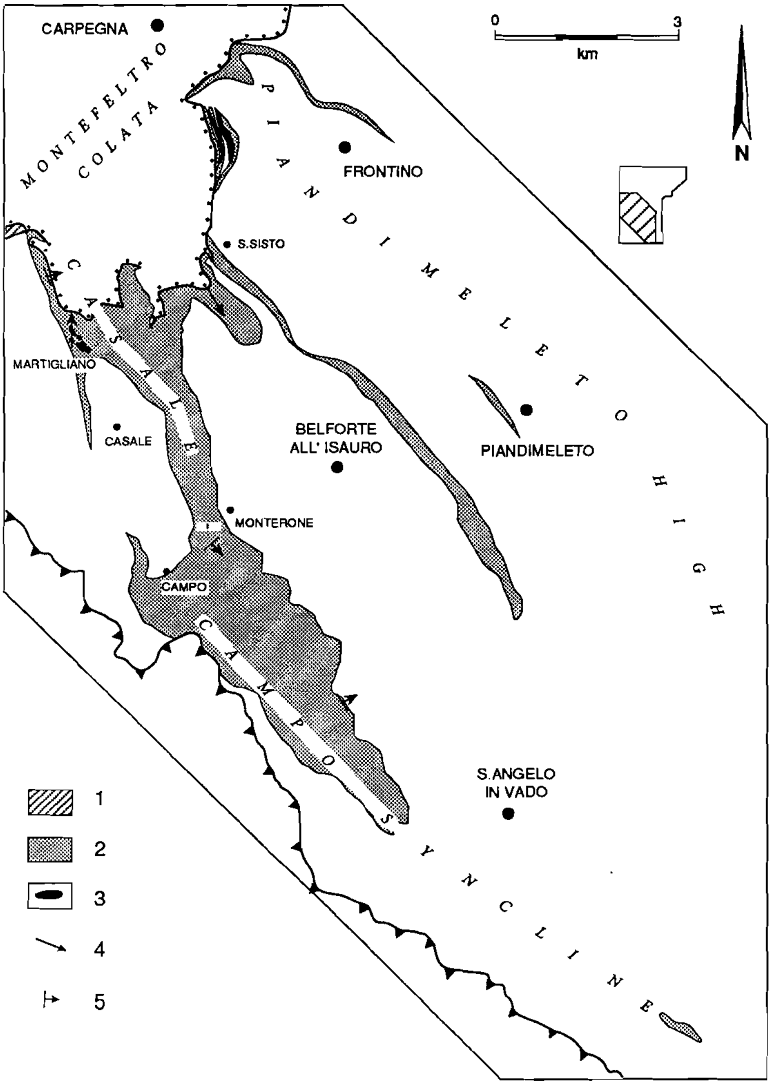


Fig. 31. Aspects of Early Tortonian cataturbiditic sedimentation in the investigated area. Legend: 1: S.Paolo Marl Formation; 2: Campo Marl Formation; 3: allolisthostrome; 4: palaeocurrent direction indicated by flute casts; 5: attitude of slump structure. Structural symbols as in Fig. 22.

The S.Paolo Marl Formation overlies the Marnoso-arenacea Formation A. It has been defined in the Romagnan Apennines some 25 kilometers W of the investigated area, where its thickness amounts to several hundreds of meters (Di Napoli Alliata, 1943; Signorini, 1943; Ruggieri, 1958). A relatively light colour and the virtual absence of arenaceous intercalations characterize the poorly exposed S.Paolo Marl Formation in the frontal portion of the Umbro-Romagnan parautochthon near C.Nuova dei Prati, 4 kilometers SW of Carpegna.

The Campo Marl Formation, which overlies the Marnoso-arenacea Formation B in the contiguous palaeoautochthon, intertongues with the ortho-turbiditic Urbania Sandstone Formation at the Piandimeleto High. It primarily occupies the Casale-Campo Syncline, attaining a thickness of 750 meters near Campo, 3 kilometers SW of Belforte all'Isauro. Dispersed organic material accounts for the rather dark appearance of the formation. Poorly cemented greyish sandstone beds may be several meters thick. The average sandstone/marl ratio however does not exceed 0.05.

The palaeocurrent pattern of the Campo Marl Formation in the Casale-Campo Syncline is bimodal longitudinal (Fig. 31). This relates to the activation of Apenninic sources supplementing the decreased supply of Austro-Alpine material. Presumably, clastics derived from the Tuscan Zone were distributed longitudinally in the synclinal depocentre. A similar mechanism has been outlined for the coeval cataturbiditic M.Vicino Formation, which overlies the Marnoso-arenacea Formation A in the M.Vicino Unit S of the investigated area (Centamore *et al.*, 1977, 1978; Chiocchini & Cipriani, 1984).

Allochthonous elements are most evident in the upper part of the Campo Marl Formation in the vicinity of the Montefeltro Colata. They range from voluminous olisthostromes to clastic components of regular cataturbidites, which were apparently picked up from allochthonous terrains by overflowing currents (cf. de Jager, 1979).

II.3.2.1.5. CALCARE A LUCINA

Marly limestones and calcarenites, generally containing shallow-marine molluscs, constitute the Calcare a Lucina. This lithofacies predominantly occurs associated with the cataturbiditic terrains of the Umbro-Romagnan turbidite suite as displaced bodies derived from major highs (Ricci Lucchi & Veggiani, 1966). Repeated resedimentation is attested to by admixed littoral to neritic and bathyal faunal elements.

In the investigated area, the Calcare a Lucina is represented by coarse calcarenitic bodies with more or less abundant bivalves and gastropods lining the major reverse fault which borders the hinge zone of the Casale-Campo Syncline (Fig. 32, 33, 34). This is indicative of their emplacement at irregularities resulting from incipient faulting.

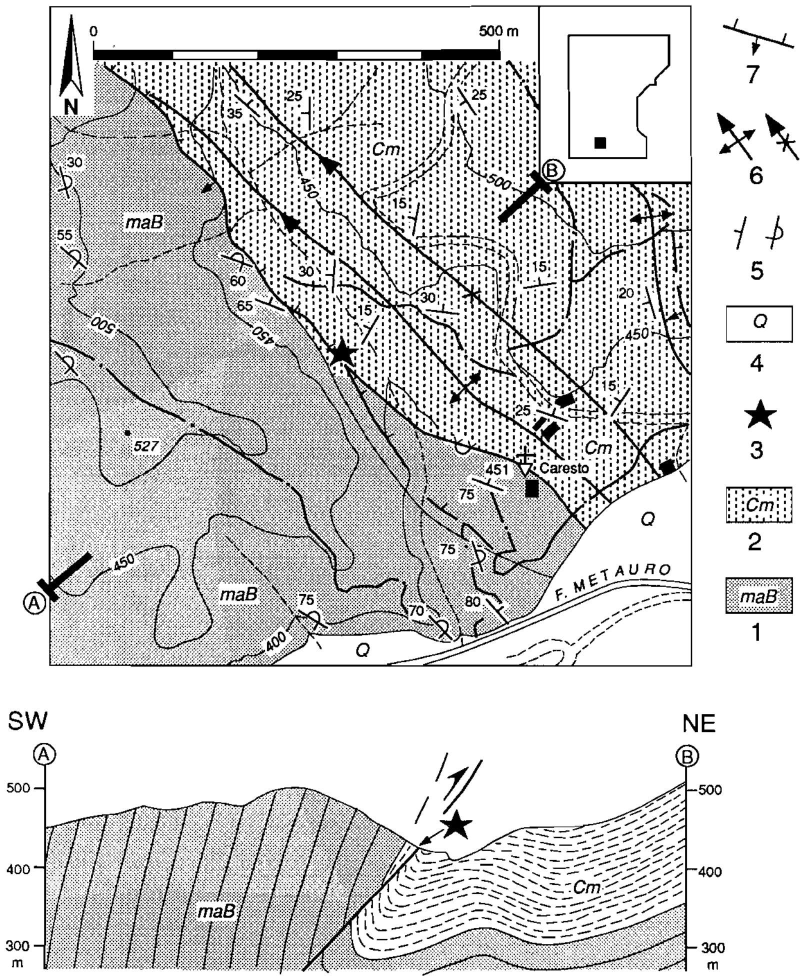


Fig. 32. Geological map and cross-section of the hinge zone of the Casale-Campo Syncline in the surroundings of Caresto, 2 km W of S.Angelo in Vado. Legend: 1: Marnoso-arenacea Formation B; 2: Campo Marl Formation; 3: Calcare a Lucina; 4: Quaternary; 5: bedding symbols, with indicated angle of dip; 6: axial traces of plunging folds; 7: reverse fault, with hachures on downthrown block and arrow indicating sense of dip.

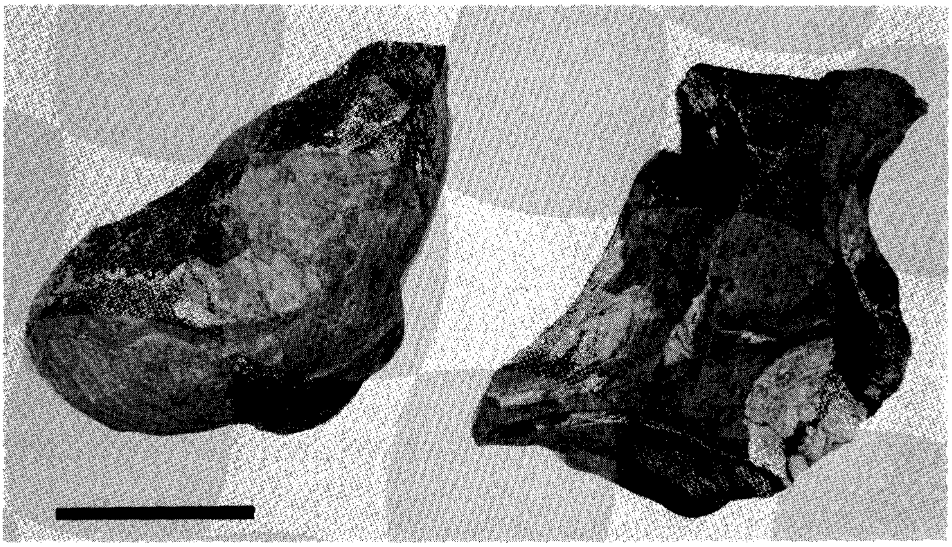


Fig. 33. Specimen of Modiolus and typical aggregate of more or less fragmented shells with a coarse calcarenitic matrix. Samples from Calcare a Lucina near Caresto, 2 km W of S.Angelo in Vado. Scale bar is 5 cm.

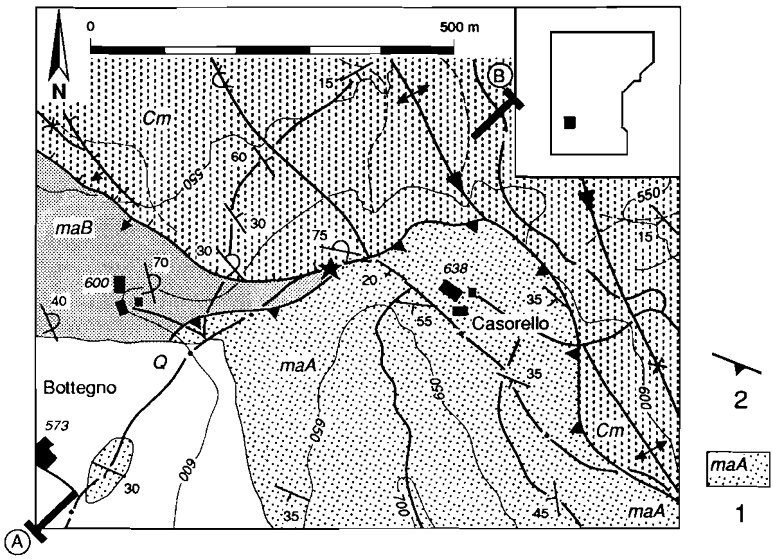
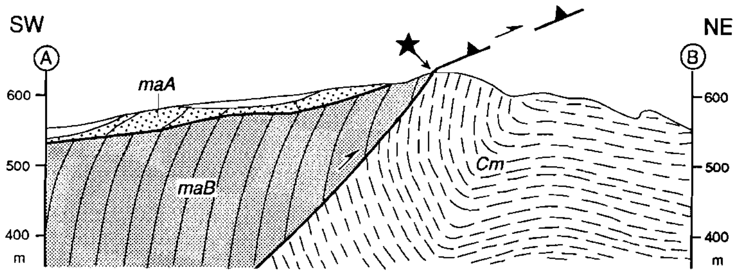


Fig. 34. Geological map and cross-section of the hinge zone of the Casale-Campo Syncline in the surroundings of Casorello, 4 km SW of Belforte all'Isauro. Legend: 1: Marnoso-arenacea Formation A; 2: major thrust



fault, with barbs on upper block. Otherwise as in Fig. 32.

II.3.2.1.6. URBANIA SANDSTONE FORMATION (LOWER-MIDDLE TORTONIAN)

More or less amalgamating sandstone bodies separated by regular orthoturbiditic deposits characterize the Urbania Sandstone Formation, which overlies the Marnoso-arenacea Formation B between the Piandimeleto High and the Montescudo-Serrungarine High (Centamore *et al.*, 1978). Like the Marno-

Fig. 35. Sandstone body of Urbania Sandstone Formation near Frontino, showing prominent cogoli in subvertical intervals parallel to bedding. Stratigraphic top to left.

so-arenacea Formation B, it displays a lateral variability according to structural position superposed on a general northeastward fading. Thus, its average thickness ranges from 450 meters in the internal part of the Pietrarubbia Syncline to 50 to 200 meters more externally.

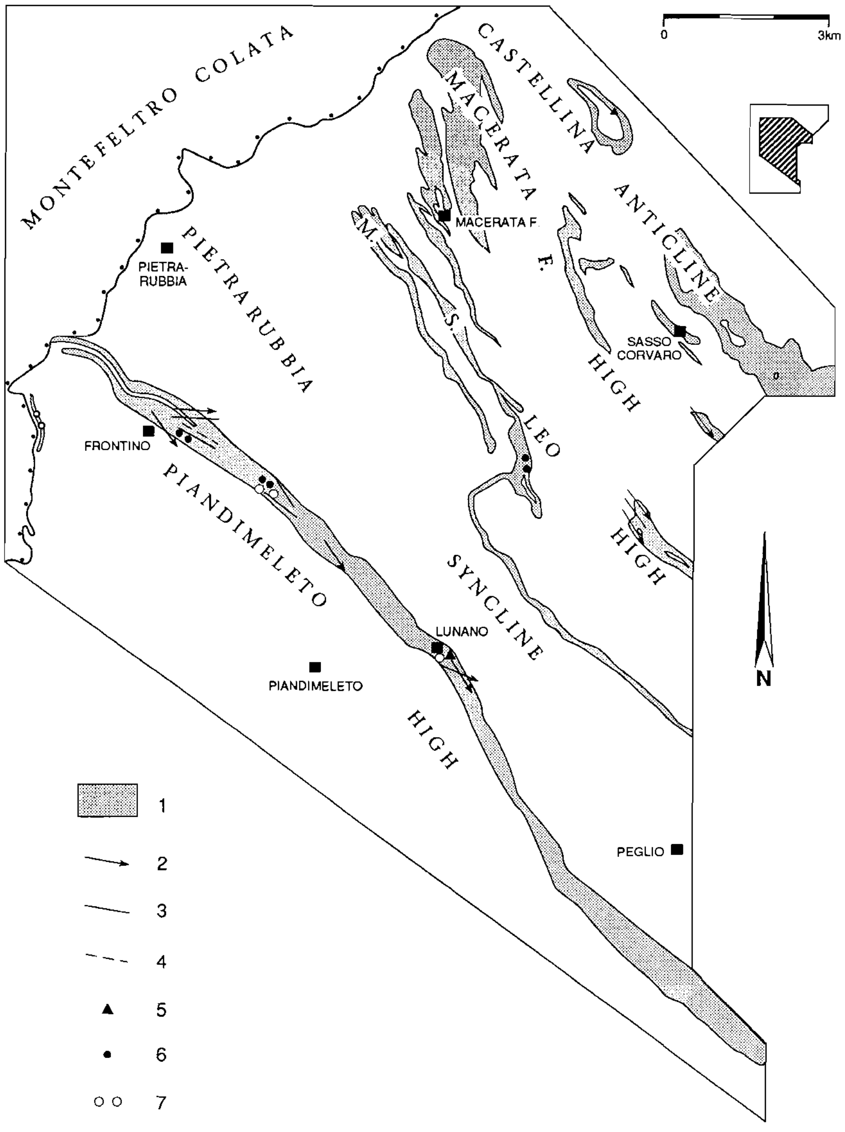


Fig. 36. Aspects of the Urbania Sandstone Formation in the investigated area. Legend: 1: Urbania Sandstone Formation; 2: palaeocurrent direction indicated by flute casts; 3: palaeocurrent direction indicated by groove

casts; 4: channel direction; 5: marl intraclasts; 6: granitello; 7: gravel of Apenninic provenance. Structural symbols as in Fig. 22.

The brown-greyish bodies of stacked sandstone beds commonly are several tens of meters thick. They are rather poorly cemented, exhibiting conspicuous spheroidal concretions. These "*cogoli*" typically occur concentrated in intervals parallel to bedding (Fig. 35).

The compositional similarity of the Urbania Sandstone Formation to the Marnoso-arenacea Formation B argues the Austro-Alpine provenance of both its arenaceous and arenaceous-marly lithotypes (Centamore *et al.*, 1987; Gandolfi *et al.*, 1983; Ardanese *et al.*, 1987). Likewise, they were supplied longitudinally from the NW (Fig. 36). The composite sandstone beds resulted from voluminous turbidity currents which virtually flooded the narrow active segment of the Umbro-Romagnan Foredeep (Ricci Lucchi, 1975a, 1981b). Their violent character is indicated by marl intraclasts (Fig. 37). Moreover, intercalated gravel is not uncommon in the depocentral internal portion of the Pietrarubbia Syncline (Fig. 36). More or less rounded pebbles to cobbles of calcareous and marly lithotypes occur admixed with molluscan fragments in pockets and stringers (cf. Ardanese *et al.*, 1987; Capuano *et al.*, 1987a). They were probably derived from autochthonous and allochthonous terrains through minor debris flows generated at the Piandimeleto High by the oversized turbidity currents (cf. de Jager, 1979).

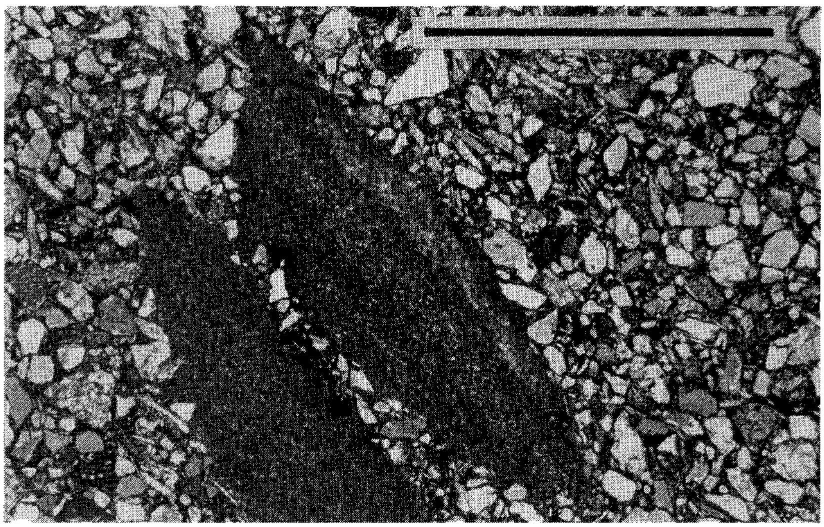


Fig. 37. Photomicrograph of sandstone containing marl intraclasts. Sample from Urbania Sandstone Formation near Castellina, 3 km ENE of Macerata Fèltria. Crossed nicols. Scale bar is 2 mm.

Finer-grained conglomerate lenses at the base of sandstone beds constitute the "*granitello*" (Venzo, 1954). This is characterized by clasts of sedimentary, metamorphic and igneous lithotypes, derived from the same Austro-Alpine sources as the sand particles (Ricci Lucchi, 1975a; Ardanese *et al.*, 1987; Capuano *et al.*, 1987a). NW of the Montefeltro Colata, the Urbania Sandstone Formation contains coarser intercalations of *granitello* in the surroundings of Mercato Saraceno (Veggiani, 1953; Veggiani & De Francesco, 1968). They reflect the vicinity of the entry point of the voluminous turbidity currents (Ricci Lucchi, 1975a, 1981b). In the northwestern part of the Romagnan Apennines, similar lenticular conglomerates are displayed by the Fontanelice Sandstone Formation (Ricci Lucchi, 1968; de Jager, 1979). This counterpart of the Urbania Sandstone Formation relates to a separate entry point, underlining the regional significance of the massive supply of Austro-Alpine clastics to the Umbro-Romagnan Foredeep during the Early to Middle Tortonian (cf. Ricci Lucchi & Ori, 1985; Ricci Lucchi, 1986a).

II.3.2.1.7. MARNE DI LETTO FORMATION (MIDDLE TORTONIAN-LOWER MESSINIAN)

The Marne di Letto Formation constitutes the upper part of the Middle to Upper Miocene Turbiditic Subsequence NE of the Piandimeleto High (cf. Farabegoli & Ricci Lucchi, 1973). It displays a prominent fining-upward trend, reflecting the gradual cessation of orthoturbiditic sedimentation (Fig. 38). The succession of depositional stages is documented by the arrangement of facies associations defined with reference to sandstone/marl ratio (cf. II.3.2.1.3.). Their transversal diachroneity attests to the migratory aspect of foredeep activity.

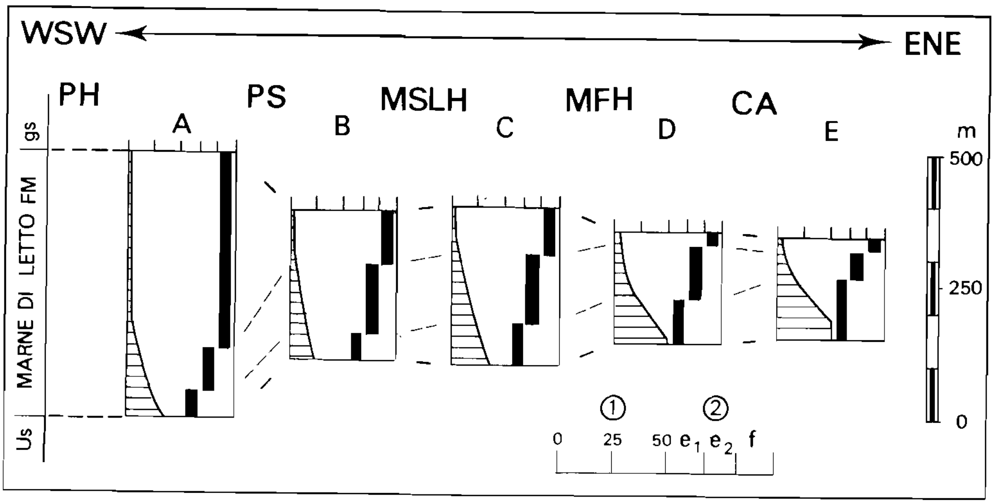


Fig. 38. Transversal variability of the Marne di Letto Formation with

respect to sandstone percentage (1) and facies associations (2). The columnar sections relate to the surroundings of Monastero, 1 km E of Frontino (A), C.Fagnano, 2 km SW of Macerata Fèltria (B), Cse.Giacomo, 2.5 km N of Macerata Fèltria (C), Sassocorvaro (D), and S.Donato in Taviglione, 2.5 km E of Sassocorvaro (E). Labels of major longitudinal structures: PH: Piandimeleto High; PS: Pietrarubbia Syncline; MSLH: M.S. Leo High; MFH: Macerata Fèltria High; CA: Castellina Anticline.

The lower part of the Marne di Letto Formation is characterized by regular arenaceous-marly orthoturbidites of Austro-Alpine provenance, like their counterparts of the Marnoso-arenacea and Urbania Sandstone formations supplied longitudinally from the NW. Poorly cemented sandstone beds are most common NE of the Macerata Fèltria High, attaining thicknesses of several meters. Erosional contacts, marl intraclasts, and lenses of *granitello* add to their similarity to the arenaceous lithofacies of the Urbania Sandstone Formation (cf. II.3.2.1.6.).

Blue-greyish clayey marls prevail in the cataturbiditic upper part of the Marne di Letto Formation. This differs from the comparable Campo Marl Formation in the absence of clastic elements of Apenninic derivance.

The general thinning of the Marne di Letto Formation from an average of more than 500 meters in the depocentral segment of the Pietrarubbia Syncline to 100 to 300 meters more externally relates primarily to its cata-

Fig. 39. Contorted bedding in toe of slump of southwestern origin in the lower part of the Marne di Letto Formation near Paganica, 1.5 km SSE of Lunano.

turbiditic portion (Fig. 38). Markedly reduced thicknesses at major longitudinal reliefs denote the progressive structural articulation of the Umbro-Romagnan Foredeep during the Late Miocene (cf. Conti *et al.*, 1987; Conti, 1989). This is corroborated by prominent slumps (Fig. 39, 40).

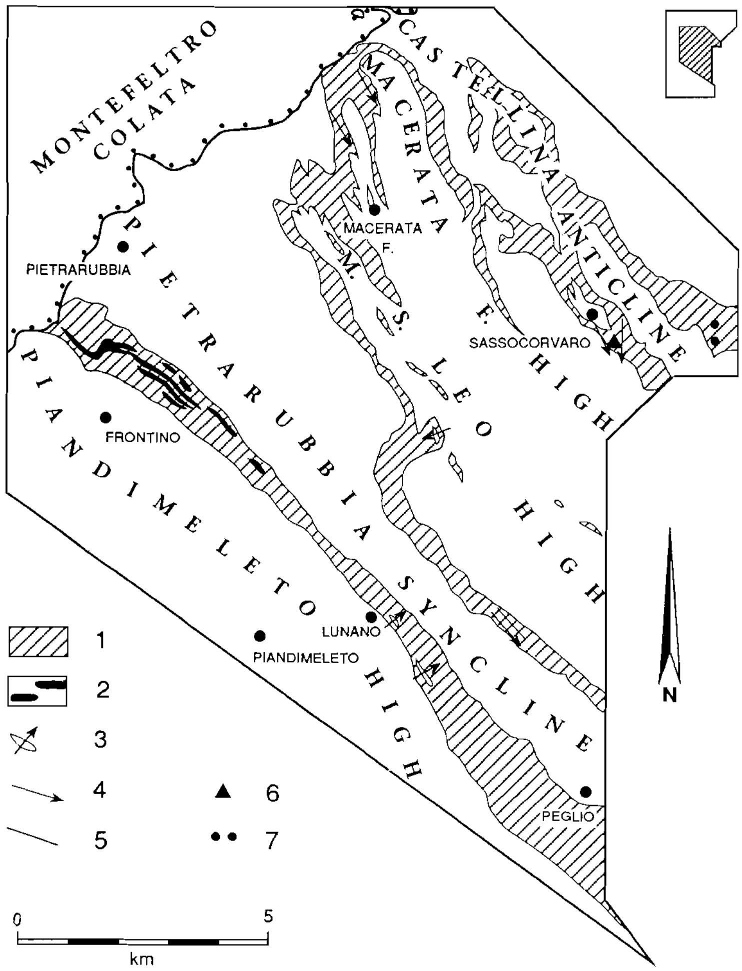


Fig. 40. Aspects of the Marne di Letto Formation in the investigated area. Legend: 1: Marne di Letto Formation; 2: allolisthostromes; 3: slump with inferred sense of emplacement; 4: palaeocurrent direction indicated by flute casts; 5: palaeocurrent direction indicated by groove casts; 6: marl intraclasts; 7: granitello. Structural symbols as in Fig. 22.

Near Frontino, the Marne di Letto Formation comprises voluminous olisthostromes derived from more internal allochthonous terrains (Fig. 40). Their longitudinal emplacement in the depocentral segment of the Pietrarubbia Syncline is suggested by trailing olistholiths.

II.3.2.2. GESSOSO-SOLFIFERA FORMATION (LOWER-MIDDLE MESSINIAN)

Heralded by a marked shallowing and microfaunal impoverishment, evaporitic sedimentation set in during the Early Messinian as a function of the Messinian Salinity Crisis of the Mediterranean (*e.g.*, Borsetti & Cati, 1974; Borsetti *et al.*, 1974b). The resulting Gessoso-solfifera Formation comprises variety of lithotypes, distributed more or less systematically according to structural position (Fig. 41, 42).

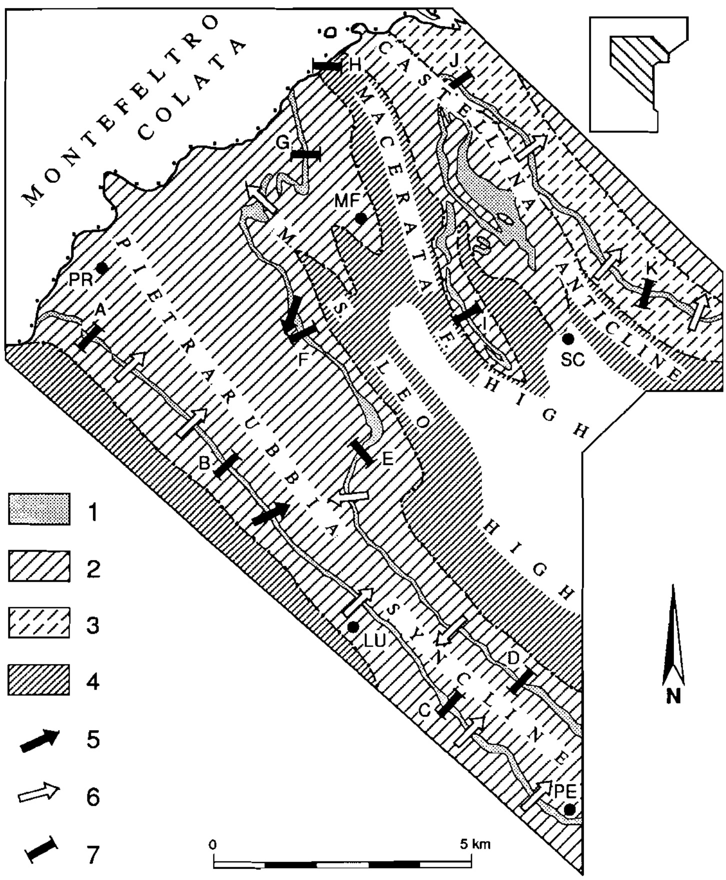


Fig. 41. Lateral variability of the Gessoso-solfifera Formation in the

investigated area. Legend: 1: Gessoso-solfifera Formation; 2: complete basinal series; 3: disrupted marginal series; 4: condensed interbasinal series; 5: palaeocurrent direction indicated by current ripples; 6: inferred sense of gravity displacement; 7: location of measured section. Structural symbols as in Fig. 22. Location labels: LU: Lunano; MF: Macerata Fèltria; PE: Peglio; PR: Pietrarubbia; SC: Sassocorvaro.

An average thickness of 25 to 50 meters characterizes the basinal series in the major synclines, consisting predominantly of more or less bituminous dark fissile marls with laminar and dispersed crystalline whitish gypsum. These euxinic deposits are commonly referred to as "*ghioli*" (Sacco, 1937; Lipparini, 1969). They reflect anoxic conditions due to stagnant brines (cf. Schmalz, 1969). This is corroborated by rare "*tripoli*" of whitish laminated diatomite in the lower part of the formation (Fig. 43). Progressive restriction was enhanced by the accentuation of longitudinal reliefs (Savelli & Wezel, 1978).

Arenaceous intercalations are essentially confined to synclinal depocentres. Sparse thin-bedded turbidites of Austro-Alpine provenance, supplied longitudinally from the NW similar to their counterparts of the Middle to Upper Miocene Turbiditic Subsequence, attest to protracted cataturbiditic activity. Intrabasinal reworking originated thin-bedded brownish hybrid sandstones with vegetal matter (Fig. 44).

The lower part of the Gessoso-solfifera Formation usually contains the "*calcare di base*" of more or less thick-bedded foetid dolomitic limestones of reefalgal origin. Light-coloured rather massive "*cagnini*" are most common in the marginal segments of the synclinal basins. They are particularly rich in microcrystalline sulphur (Sacco, 1937; Selli, 1952, 1973; Lipparini, 1969).

A somewhat laminated dark variety of the *calcare di base* typifies the condensed series of interbasinal highs (Conti *et al.*, 1987). Due to the near-surface dissolution of sulphates, this generally appears as cavernous "*calcare perciuliato*" (cf. Ogniben, 1957, 1963).

The acme of cyclic evaporation is reflected by interlaminated greyish to reddish dolomicrite and whitish microcrystalline gypsum in layers up to a few meters thick (Fig. 45). Basinal series maximally display 13 of these "*balatini*". This is compatible with the average cyclicity identified in neighbouring areas (Borsetti *et al.*, 1974a; Carloni *et al.*, 1974; Vai & Ricci Lucchi, 1977).

The laminar gypsum essentially represents a primary precipitate. Effects of recrystallization range from contorted laminae of "*serpentino*" to massive alabaster (cf. Ogniben, 1955, 1957; Parea & Ricci Lucchi, 1972; Borsetti *et al.*, 1974a, 1974b; Carloni *et al.*, 1974).

The shallow lagoonal origin of the *balatini* is indicated by greyish stromatolitic intercalations (Ardanese & Martelli, 1983-1984). Palaeopedogenic features denote more or less prolonged exposure (Fig. 46).

Numerous gravity displacement structures add to the lateral variability of the Gessoso-solfifera Formation. Slumping and sliding affected in particular the *calcare di base* (*e.g.*, Fig. 47). *Cagnini* and *calcari perciuliati* accordingly are the most prominent lithotypes of the disrupted series displayed by the northeastern limb of the Castellina Anticline (Savelli & Wezel,

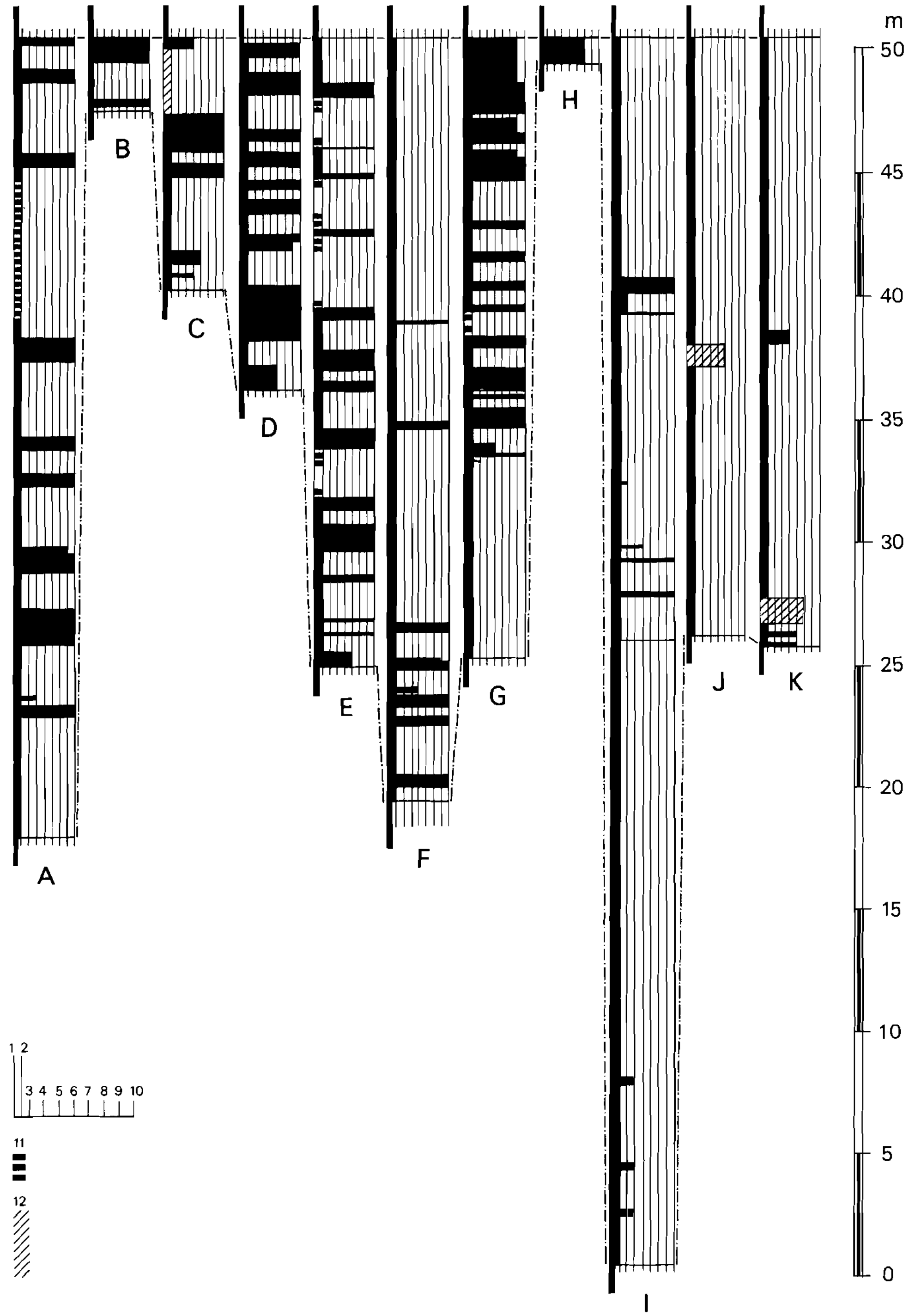


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Fig. 42. Schematic columnar sections of the Gessoso-solfifera Formation in the investigated area. Locations are indicated in Fig. 41. Key: 1: not exposed; 2: marly lithofacies of Marne di Letto and Colombacci formations; 3: ghioli; 4: hybrid sandstone; 5: siliciclastic sandstone; 6: stromatolite; 7: cagnino; 8: calcare perciuliato; 9: alabaster; 10: balatino; 11: inferred; 12: displaced.

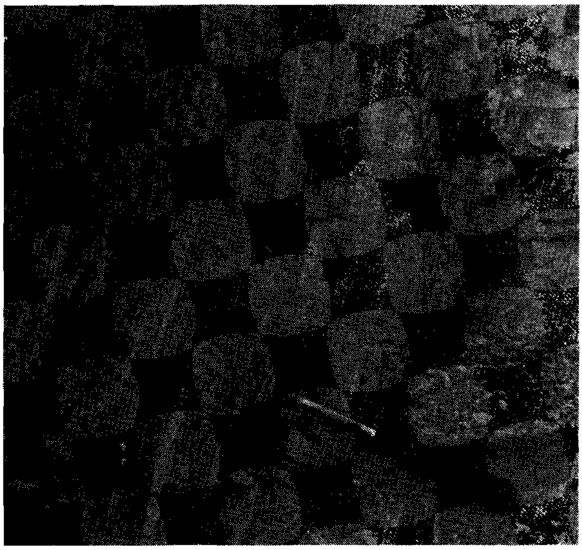


Fig. 43. Dark ghioli with thin lighter-coloured tripoli, overlying a calcareous marl bed in the lower part of the Gessoso-solfifera Formation along the T. Apsa near Mondagano, 2.5 km SE of Macerata Fèltria. Stratigraphic top to left.

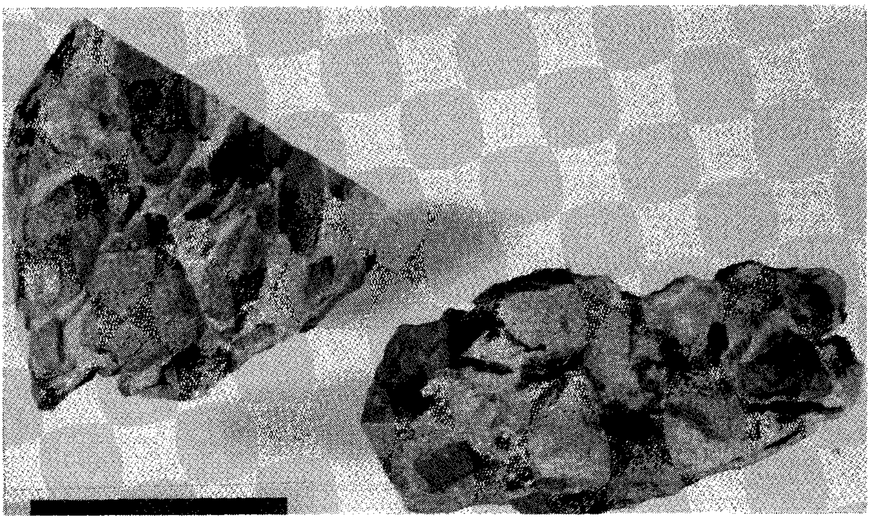


Fig. 44. Hybrid sandstone with abundant vegetal matter.

Samples from upper part of Gessoso-solfifera Formation near Castelluccio, 3 km N of Sassocorvaro. Scale bar is 4 cm.

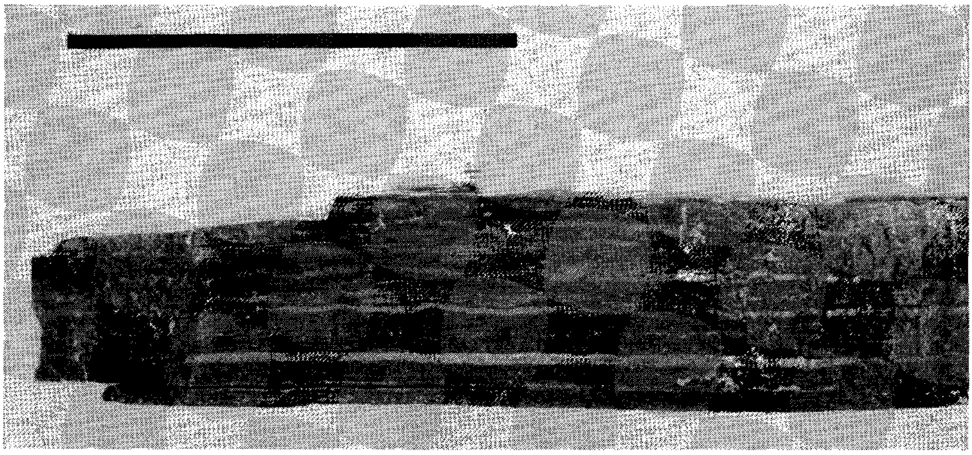


Fig. 45. Typical balatino of interlaminated dark dolomicrite and whitish gypsum. Sample from lower part of Gessoso-solfifera Formation near C.Fagnano, 2 km SW of Macerata Fèltria. Scale bar is 5 cm.

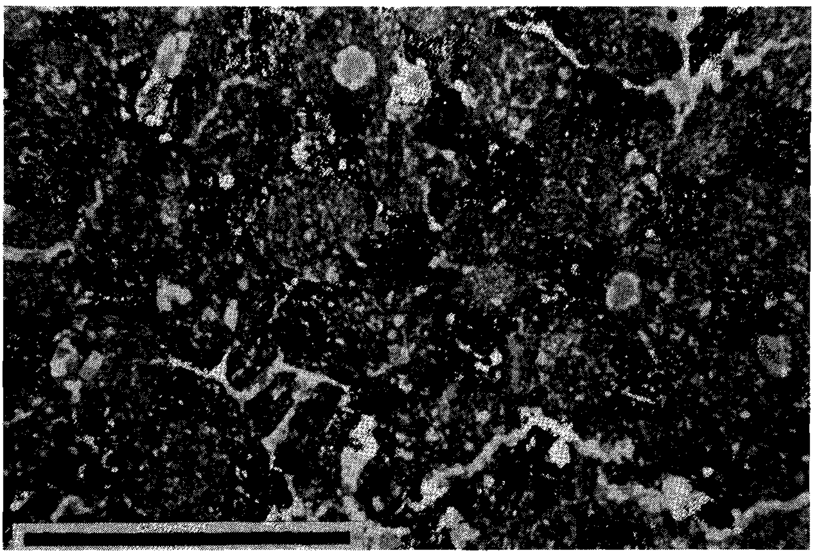


Fig. 46. Photomicrograph of palaeosol showing cracks and root channels rimmed with calcitans. Sample from upper part of Gessoso-solfifera Formation near Certalto, 2 km E of Macerata Fèltria. Crossed nicols.

Scale bar is 2 mm.

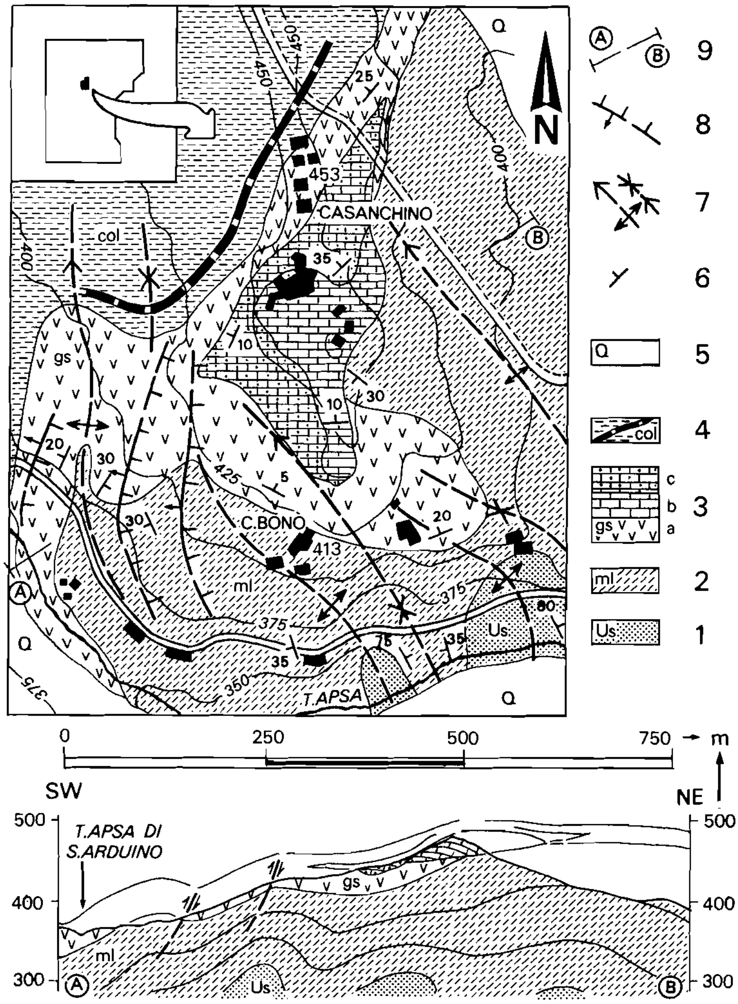


Fig. 47. Geological map and cross-section of the surroundings of Casanchino, 2 km W of Macerata Fèltria. Legend: 1: Urbania Sandstone Formation; 2: Marne di Letto Formation; 3: Gessoso-solfifera Formation, with regular basinal deposits (a) enclosing a displaced body of calcare perciuliato (b) surrounded by brecciated carbonates (c); 4: Colombacci Formation, with a colombaccio; 5: Quaternary; 6: bedding symbol, with indicated angle of dip; 7: axial traces of plunging folds; 8: reverse fault, with hachures on downthrown block and arrow indicating sense of dip; 9: trace of cross-section.

1978).

II.3.2.3. UPPER MIOCENE-PLIOCENE CONTINENTAL TO SHALLOW-MARINE SUBSEQUENCE

II.3.2.3.1. INTRODUCTION

The Upper Miocene to Pliocene top part of the Umbro-Romagnan Sequence consists of continental to shallow-marine terrigenous deposits. These were essentially supplied by local sources as an Apenninic molasse (Ricci Lucchi, 1986b). Their common cyclicity accordingly reflects the intermittent paroxysmal accentuation of orogenic structures (Savelli & Wezel, 1978).

II.3.2.3.2. COLOMBACCI FORMATION (MIDDLE-UPPER MESSINIAN)

The lower part of the Umbro-Romagnan molasse is formed by the Colombacci Formation. It attains a thickness of about a kilometer in the Pietrarubbia and Montecalvo in Foglia synclines. The indicated massive terrigenous supply was determined by a markedly increased continental run-off. This brought about the brackish *"lago-mare"* conditions which ended evaporitic sedimentation in the Adriatic sector of the Mediterranean during the Middle Messinian (cf. Ruggieri, 1967).

The Colombacci Formation is largely composed of blue-greyish clayey marls with brownish coarse clastic intercalations pertaining to 8 discrete cycles. These are most evident in the Pietrarubbia Syncline as fanglomeratic bodies up to 100 meters thick (Farabegoli & Ricci Lucchi, 1973; de Feyter & Molenaar, 1984). They originated as coalescent fans spreading from the Piandimeleto High. Clastics were supplied by torrential streams draining autochthonous and allochthonous terrains. An aberrant gypsiferous body fringing the M.S.Leo High denotes the significance of additional sources (cf. Venzo, 1954; Ruggieri, 1970; Ardanese & Martelli, 1983-1984; Conti, 1989).

The physiography of the fans is reflected by the distribution of sedimentary facies (Fig. 48, 49). Unsorted massive conglomerates resulting from viscous debris flows are typical proximal deposits. The lower fan reaches and the contiguous floodbasin predominantly accumulated water-laid pebbly sandstones with conglomeratic intercalations. Turbiditic pebbly sandstones are distal features of an initial fan-deltaic stage.

The general sedimentary pattern is characteristic of alluvial fans in a semi-arid environment (Blissenbach, 1954; Bull, 1972; Rust, 1979). Palaeopedogenic features such as colour mottling and caliche nodules are compatible with this.

The rapid terrestrial sedimentation in the Pietrarubbia Syncline tended to suppress the effect of confining longitudinal reliefs. This resulted in a

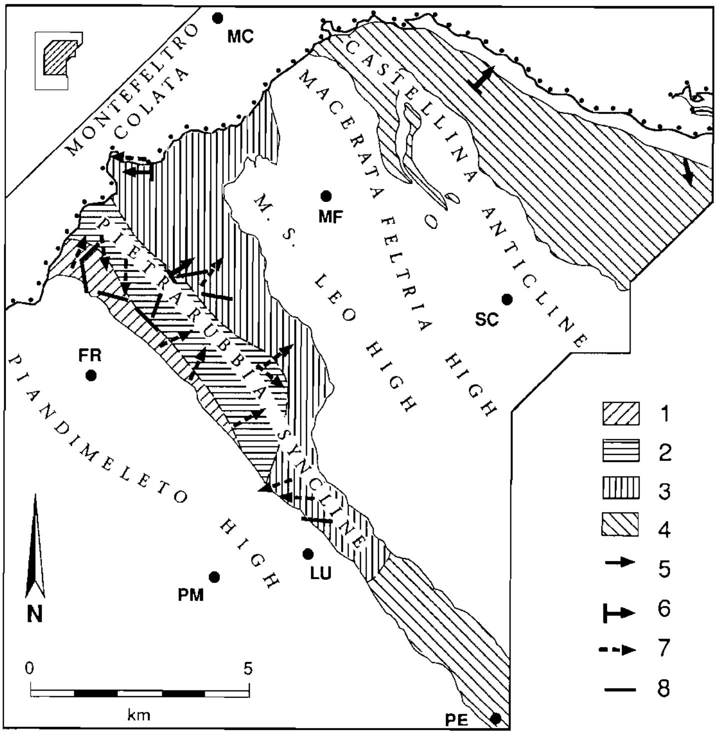


Fig. 48. Sedimentary environments of the Colombacci Formation in the investigated area. Legend: 1: upper fan; 2: midfan; 3: lower fan to floodbasin; 4: deltafront to prodelta; 5: palaeocurrent direction indicated by flute casts; 6: palaeocurrent direction indicated by ripples; 7: palaeocurrent direction indicated by imbrication; 8: channel direction. Location labels: FR: Frontino; LU: Lunano; MC: Montecerignone; MF: Macerata Fèltria; PE: Peglio; PM: Piandimeleto; SC: Sassocorvaro. Structural symbols as in Fig. 22.

mounting transversal component of clastic dispersal. Frontal fan-deltaic deposits NE of the Macerata Fèltria High thus primarily relate to the upper fan cycles. They appear as partly turbiditic sandstone bodies a few tens of meters thick with sparse pebbly intercalations (cf. Conti *et al.*, 1987; Conti, 1989).

More or less dolomitic whitish micritic limestones in discontinuous lithozones averaging 50 centimeters in thickness constitute the *"colombacci"* (Selli, 1952). They occur at 5 stratigraphic levels. These reflect phases of markedly reduced terrigenous input (Molenaar & de Feyter, 1985).

Prominent gravity displacement structures underline the synorogenic character of the Colombacci Formation. Elements derived from the Gessoso-solfifera Formation prevail in its lower part. They are particularly evident in the Pietrarubbia Syncline near Peglio (Fig. 50). Dispersed olisthostromes of allochthonous affinity on the other hand are most common in the upper part

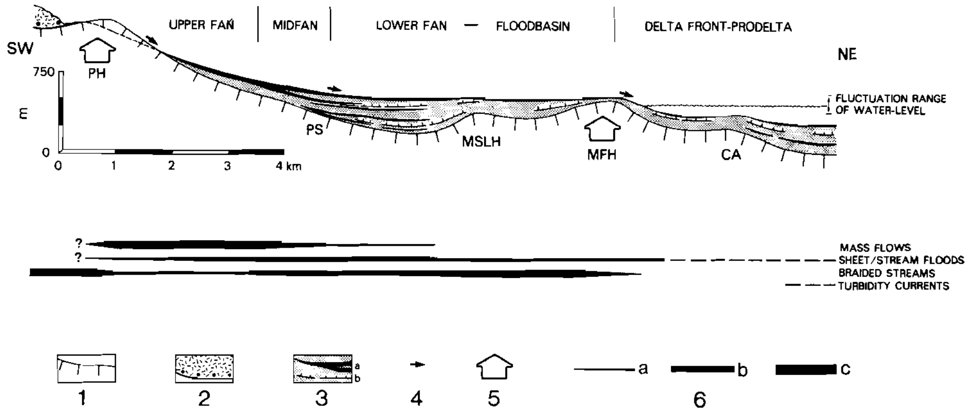


Fig. 49. Reconstructed setting of the fifth fan cycle of the Colombacci Formation in the investigated area. Legend: 1: pre-molasse terrains; 2: allochthonous terrains; 3: Colombacci Formation, with coarse clastics (a) and colombacci (b); 4: sediment transport; 5: relative uplift; 6: importance of depositional mechanisms, dashed where inferred: a: minor; b: intermediate; c: major. Labels of major longitudinal structures: PH: Piandimeleto High; PS: Pietrarubbia Syncline; MSLH: M.S.Leo High; MFH: Macerata Fèltria High; CA: Castellina Anticline.

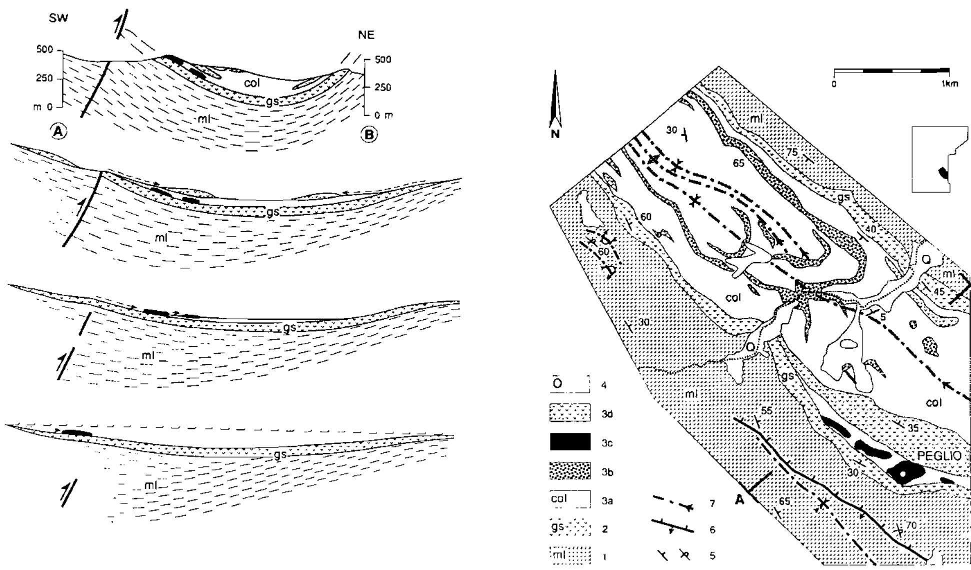


Fig. 50. Geological map and cross-section of the Pietrarubbia Syncline in the surroundings of Peglio. Palinspastically restored sections elucidate the mechanism of repeated resedimentation during the Middle

Messinian. Legend: 1: Marne di Letto Formation; 2: Gessoso-solfifera Formation; Colombacci Formation: 3a: marly lithofacies; 3b: arenaceous lithofacies; 3c: olistholithic massive gypsum; 3d: other displaced material derived from Gessoso-solfifera Formation; 4: Quaternary; 5: bedding symbols, with indicated angle of dip; 6: reverse fault, with hachures on downthrown block and arrow indicating sense of dip; 7: axial trace of plunging fold.

of the formation in the vicinity of the Montefeltro Colata.

II.3.2.3.3. SANTERNO FORMATION (LOWER PLIOCENE)

NE of the Macerata Fèltria High, the Colombacci Formation is overlain by the Lower Pliocene Santerno Formation, which attains a thickness of about 500 meters in the Montecalvo in Foglia Syncline. Its conformable base is marked by the appearance of predominantly planktonic microfaunas denoting normal marine conditions (Borsetti & Cati, 1974; Borsetti *et al.*,1974a; Bellagamba, 1978; Cremonini *et al.*, 1978; Conti *et al.*, 1987). This reflects the Atlantic invasion of the Mediterranean at the beginning of the Pliocene (Hsü *et al.*, 1973). The transgressive effect of this event was rather limited, as indicated by the similarity with regard to terrigenous lithofacies of the Santerno Formation to the underlying Colombacci Formation (cf. Selli, 1952, 1973; Borsetti *et al.*, 1974a; Carloni *et al.*, 1974; Cremonini *et al.*, 1978).

The Santerno Formation is primarily constituted by blue-greyish clayey marls, analogous to its type section in the northern part of the Romagnan Apennines (G.Cremonini, C.Elmi & A.Monesi *in* Selli, 1967). Brownish sandstone bodies of fan-deltaic origin, supplied transversally by Apenninic sources, are restricted to the upper part of the formation. In the southwestern limb of the Montecalvo in Foglia Syncline, they appear as frontal lobes tens of meters thick (cf. Conti *et al.*, 1987; Conti, 1989). Dispersed fresh-water gastropods attest to a regressive coastal setting. More externally, prodeltaic turbidites prevail.

Gravity displacement structures are mainly represented in the lower part of the Santerno Formation, as minor olisthostromes of allochthonous material in the vicinity of the Montefeltro Colata.

II.3.2.3.4. MONTECALVO IN FOGLIA FORMATION (MIDDLE PLIOCENE)

The Umbro-Romagnan molasse in the Montecalvo in Foglia Syncline is concluded by the Montecalvo in Foglia Formation of Middle Pliocene age (Amadesi, 1962). This constitutes the neoautochthon sealing the Montefeltro Colata. Basal lags of shallow-marine molluscan remains denote its slightly transgressive position (cf. Selli, 1949, 1952; Carloni *et al.*, 1971).

The maximum thickness of the Montecalvo in Foglia Formation amounts to 700 meters. It comprises greyish clayey to sandy marls with dispersed gravel and shell debris, and more or less tabular coarse clastic bodies corresponding with 9 discrete cycles. These were emplaced as fan-deltas spreading radially from the southwestern margin of the Montecalvo in Foglia Syncline (Capuano *et al.*, 1986a, 1986b, 1987c, 1987d). Their proximal portions of tens of meters thickness typically coarsen upward, grading from interbedded greyish sandstones and marls to brownish channelized sandstones with gravel and shell lags (Fig. 51). Massive conglomerates up to a meter thick resulting from debris flows are restricted to the upper cycles.

Fig. 51. Proximal portion of coarse clastic body representing the uppermost fan-deltaic cycle of the Montecalvo in Foglia Formation near S.Giovanni, 1.5 km NW of Auditore. Its arenaceous-marly lower part displays a prominent wave-cut surface. The channelized arenaceous upper part is about 8 meters thick. This is overlain by predominantly marly deposits.

U-shaped burrows and wave-generated structures argue the coastal origin of the proximal deposits. Palaeopedogenic features such as colour mottling and caliche nodules characterize the upper fan-deltaic cycles, pointing to a general regressive trend. This is corroborated by fresh-water gastropods contained by the intervening marls.

The thinner and finer prodeltaic portions of the coarse clastic bodies are less distinctive (Fig. 52). They consist of intensely bioturbated greyish sandstones with gravel and shell lags interbedded with marls. Sparse granules of Austro-Alpine affinity attest to the activity of more external clastic sources. They probably resulted from repeated resedimentation from the Adriatic Foreland, analogous to exotic elements reported from adjacent areas (cf. Selli, 1949, 1952; Veggiani & De Francesco, 1968; Castellarin & Stewart, 1989).

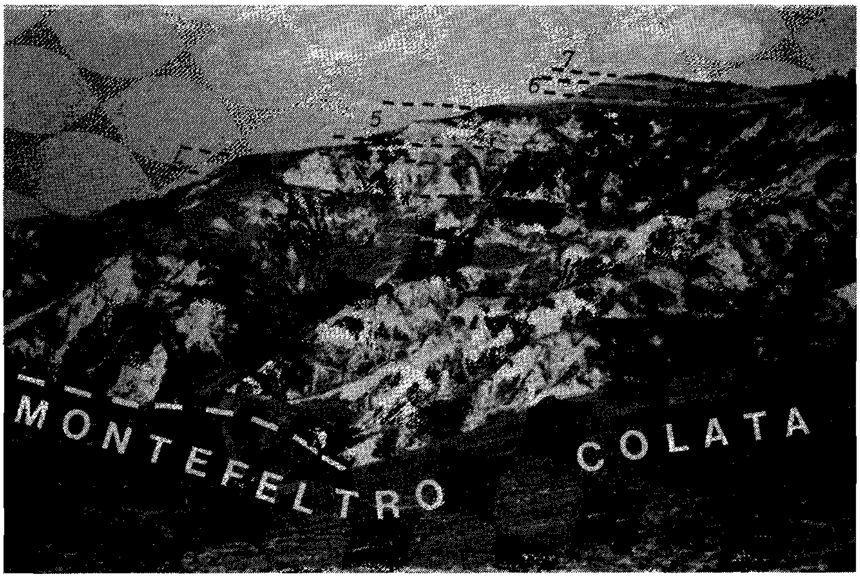


Fig. 52. Distal aspect of the Montecalvo in Foglia Formation, overlying tele- and peneallochthonous terrains of the Montefeltro Colata at M. Croce, 4 km E of Mercatino Conca. Numbered fan-deltaic cycles appear as somewhat arenaceous intercalations forming minor escarpments.

CHAPTER III

LITHOSTRATIGRAPHY AND SEDIMENTARY EVOLUTION OF THE ALLOCHTHONOUS ASSEMBLAGE

III.1. THE PIETRAFORTE-ALBERESE SEQUENCE

III.1.1. INTRODUCTION

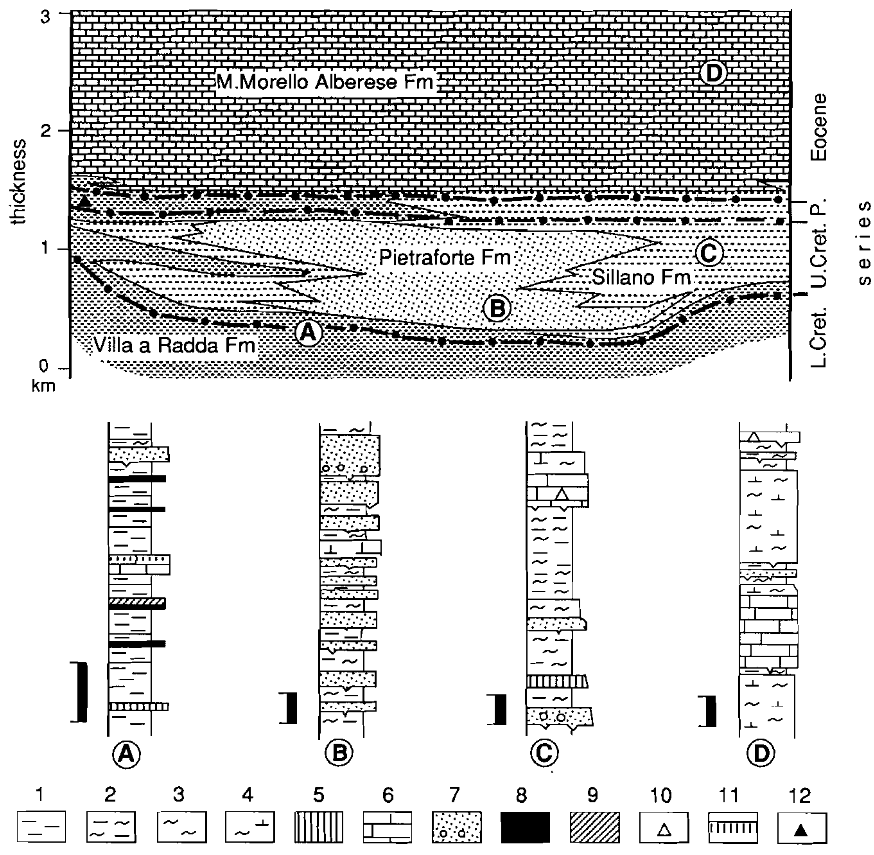


Fig. 53. Restored transverse scheme of the Pietraforte-Alberese Sequence covering about 25 km. Synthetic columnar sections illustrate the lithostratigraphy of its constituent formations. Scale bars are 1 m.

Legend: 1: cretoni; 2: gaioli; 3: marls; 4: calcareous marls; 5: siltites; 6: limestones; 7: sandstones, with lenses of cicerchina; 8: umbers; 9: beef calcite; 10: chert; 11: hardground; 12: ophiolitic fragments.

The teleallochthonous portion of the Montefeltro Colata is composed of pelagic and turbiditic oceanic deposits of Cretaceous to Paleogene age. These belong to the Pietraforte-Alberese Sequence, which originated in the outer part of the Ligurian Zone adjacent to the Adriatic continental margin (*e.g.*, Abbate & Sagri, 1970, 1982). Its stratigraphic framework to a major extent has been obliterated by translative tectonics. The pattern depicted in Fig. 53 therefore largely relates to less dismembered series in the Tuscan Zone (Merla, 1956; Bortolotti, 1962; Merla & Bortolotti, 1967; Bettelli *et al.*, 1980a; Manganelli, 1982).

III.1.2. VILLA A RADDA FORMATION (APTIAN-YPRESIAN)

Pelagic deposits of abyssal origin are most evident in the lower part of the Pietraforte-Alberese Sequence. They characterize the Villa a Radda Formation, which has been defined near Radda in Chianti, 30 kilometers S of Firenze (Merla & Bortolotti, 1967). Its thickness in the investigated area reaches many hundreds of meters.

The bulk of the Villa a Radda Formation is constituted by conspicuous reddish and green-greyish clays with rare dark bituminous patches (Fig. 54). They are known as "*cretoni*" (Principi, 1925b, 1934). Intercalated dark fine-grained beds up to 10 centimeters thick composed mostly of ferromanganese hydroxyoxides stand out in the lower part of the formation. These umbers are typical precipitates of hydrothermal affinity (cf. Dymond *et al.*, 1973; Robertson & Hudson, 1973). The pertinent metalliferous suspensions accordingly originated at Tethyan oceanic spreading centres as Mesozoic crustal divergence came to an end.

Serpentinite debris is restricted to the upper part of the Villa a Radda Formation near Serra di Maiolo. It correlates with prominent olisthostromes in other sectors of the Northern Apennines. These were derived from ophiolitic reliefs which evolved in the Ligurian Zone as a function of progressive crustal convergence from the Late Cretaceous on (*e.g.*, Bortolotti, 1963, 1983; Bertini *et al.*, 1975; Abbate & Sagri, 1982).

Subordinate turbiditic intercalations in the lower part of the Villa a Radda Formation are mainly represented by yellowish siltites and green micritic limestones known as "*pietra paesina*" (Bortolotti, 1962). The brown-greyish calcareous-quartzose litharenites which prevail in its middle part compare with the principal lithotype of the Pietraforte Formation (cf. III.1.4.). More common in the upper part of the formation are whitish to slightly variegated cherty limestones, to a variable extent burrowed and capped with hardgrounds (Fig. 55). These resemble the prominent calcareous turbidites of the overlying M.Morello Alberese Formation (cf. III.1.5.).

Diagenetic features add to the distinctive character of the Villa a

Fig. 54. Colour banding in cretoni exposed in badlands of the Villa a Radda Formation near Mo.Montalino, 2 km SW of Montecerignone.

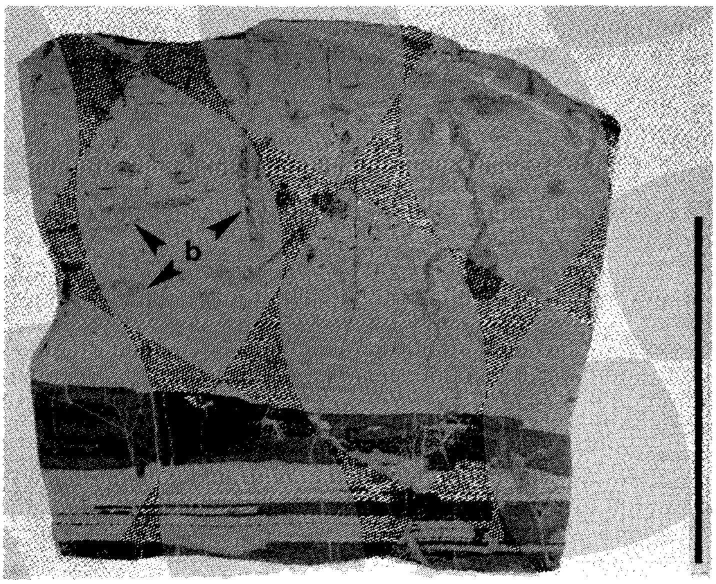


Fig. 55. Clay-filled burrows (b) and replacive chert (c) in a turbiditic limestone bed. Sample from upper part of Villa a Radda Formation be-

tween Pgio.Forco and M.della Valle, 2 km N of Montecerignone. Scale bar is 10 cm.

Radda Formation. Veins up to a few centimeters thick of fibrous beef calcite resulted from hydraulic fracturing parallel to bedding (cf. Shearman *et al.*, 1972). Millimetric spheroidal aggregates of crystalline pyrite stand out in turbiditic lithotypes (Fig. 56).

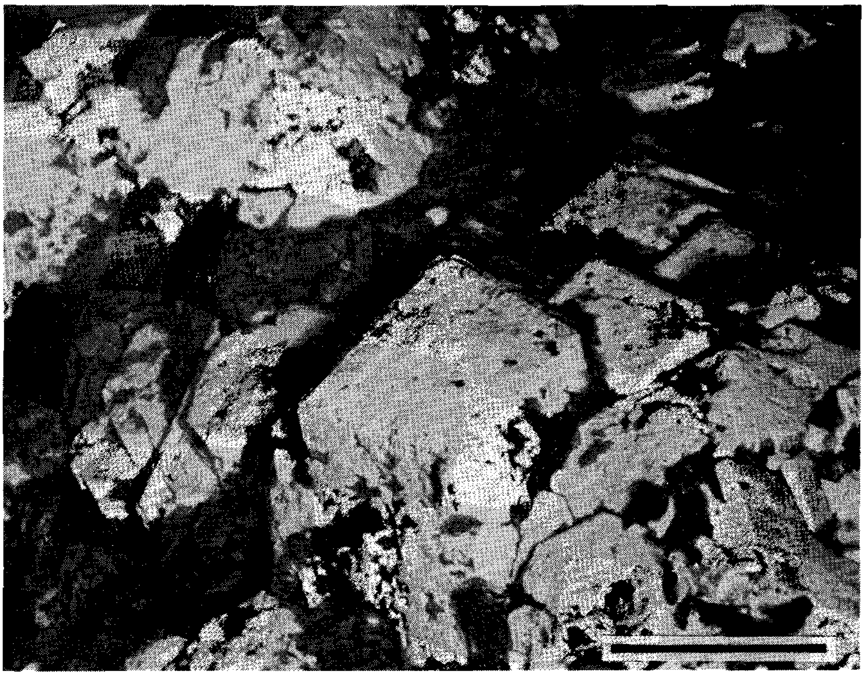


Fig. 56. Scanning electron microscope back-scattered electron image of part of spheroidal aggregate of crystalline pyrite in turbiditic sandstone. Sample from lower part of Villa a Radda Formation near Petorno, 2 km SE of Villagrande. Scale bar is 100 microns.

III.1.3. SILLANO FORMATION (CENOMANIAN-YPRESIAN)

Bortolotti (1962) distinguished the Sillano Formation near Greve, some 20 kilometers S of Firenze, as an intermediate element relative to the principal pelagic and turbiditic portions of the Pietraforte-Alberese Sequence. In the investigated area, this formation attains a thickness of more than 750 meters.

The Sillano Formation is largely composed of lead-grey to brownish clayey marls known as "*gaioli*" (Principi, 1925b, 1934). These contain sparse dark siltitic intercalations. Brown-greyish calcareous-quartzose litharenites stand out as coarse turbiditic lithotype in the lower part of the formation, fringing the Pietraforte Formation (Fig. 57). Whitish to slightly variegated cherty limestones grading into greyish to reddish calcareous marls are more common in its upper part, which underlies the M.Morello Alberese Formation.

Fig. 57. Overturned thin marly limestone bed pierced by bulbous flute casts of arenaceous megaturbidite in Sillano Formation near Pugliano, 3 km NNE of Villagrande.

III.1.4. PIETRAFORTE FORMATION (UPPER TURONIAN-MAASTRICHTIAN)

The Upper Cretaceous Pietraforte Formation essentially represents a lenticular arenaceous intercalation in the pelitic series of the Pietraforte-Alberese Sequence (*e.g.*, Losacco, 1963; Bortolotti, 1967; Abbate & Sagri, 1970, 1982; Bettelli *et al.*, 1980a; Sagri & Marri, 1980; Civitelli & Corda, 1982). Its thickness amounts to some 800 meters. The bulk of the formation consists of turbidites supplied longitudinally from Adriatic sources of Austro-Alpine affinity (Bortolotti, 1967; Abbate & Sagri, 1970, 1982; Cipriani *et al.*, 1976; Sagri & Marri, 1980; Wildi, 1985). Brown-greyish calcareous-quartzose litharenites prevail as "*pietraforte*" (Fig. 58). Megaturbiditic beds in more internal sectors of the Northern Apennines contain conglomeratic lenses known as "*cicerchina*" (*e.g.*, Azzaroli, 1946; Losacco, 1958, 1963; Bortolotti, 1962). In the investigated area, this appears solely reworked in the Umbro-Romagnan molasse.

Most evident in the middle part of the formation are marly intervals similar to the *gaioli* of the Sillano Formation (cf. III.1.3.). Dispersed calcareous turbidites compare with the principal lithotypes of the M.Morello Alberese Formation (cf. III.1.5.).

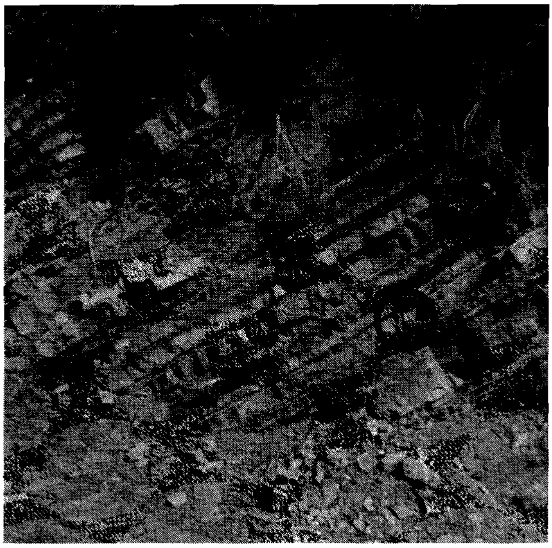


Fig. 57. Overturned predominantly arenaceous turbidites in upper part of Pietraforte Formation near Monte Altavelio, 2 km S of Mercatino Conca. Hammer for scale.

III.1.5. M.MORELLO ALBERESE FORMATION (YPRESIAN-LOWER PRIABONIAN)

The upper part of the Pietraforte-Alberese Sequence consists of the Eocene M.Morello Alberese Formation, the stratotype of which is situated near Firenze (Bortolotti, 1964; Curcio & Sestini, 1965). This formation essentially comprises calcareous turbidites supplied longitudinally by intrabasinal sources (cf. Parea, 1965; Valloni & Zuffa, 1984; Wildi, 1985). Concluding marly deposits of cataturbiditic affinity appear only at its type locality (Abbate & Sagri, 1970). In the investigated area, the thickness of the formation amounts to 1500 meters.

Generally megaturbiditic whitish cherty limestones, known as *"calcare alberese"*, and greyish calcareous marls compose the bulk of the M.Morello Alberese Formation (Fig. 59). Macroforaminiferal breccias form a distinctive minor lithotype (cf. Amadesi, 1962; Conti, 1989). Transitional features in the lower part of the formation are represented by somewhat reddish levels and by relatively common dark clayey marls and brown-greyish calcareous-quartzose litharenites (Brandi *et al.*, 1966; Capuano *et al.*, 1982). These compare with the *gaioli* of the Sillano Formation and the *pietraforte* of the Pietraforte Formation, respectively (cf. III.1.3., III.1.4.).

Fig. 59. More or less thick-bedded calcareous turbidites in upper part of M.Morello Alberese Formation at M.Carpegna Est, 2 km N of Carpegna.

III.2. THE S.MARINO-M.FUMAIOLO SEQUENCE

III.2.1. INTRODUCTION

The semiallochthon of the investigated area is constituted by the S. Marino-M.Fumaiolo Sequence of Middle to Late Miocene age (Fig. 60). Compositional parameters argue its internal Apenninic origin (Salmojraghi, 1903, 1909; Signorini, 1945b; Malesani & Manetti, 1967). In fact, it correlates with neoautochthonous terrains of the Tuscan Zone (*e.g.*, Merla & Bortolotti, 1967; Sestini, 1970). Sedimentation in irregular satellite basins is reflected by a marked lateral variability (Frey, 1969; Stern, 1969; Conti, 1989).

Semiallochthonous deposits pertaining to a terminal Miocene cycle are represented in the northern part of the Montefeltro Colata. Massive fanglomerates characterize the Acquaviva Formation (Ruggieri, 1970). They correlate with the terrestrial coarse clastics displayed by the Umbro-Romagna Colombacci Formation in the Pietrarubbia Syncline (cf. II.3.2.3.2.).

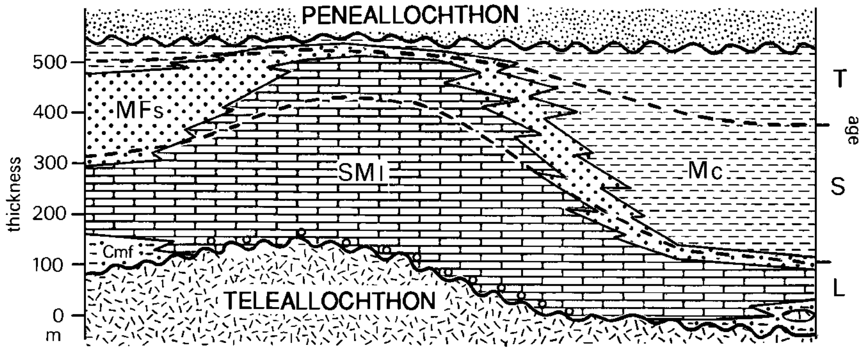


Fig. 60. Stratigraphic framework of the S.Marino-M.Fumaiolo Sequence, comprising the Cantoniera Marl Formation (Cmf), the S.Marino Limestone Formation (SMl), the M.Fumaiolo Sandstone Formation (MFs), and the Montebello Clay Formation (Mc), which is not exposed in the investigated area. Key: L: Langhian; S: Serravallian; T: Tortonian.

III. 2.2. CANTONIERA MARL FORMATION (LANGHIAN)

The Cantoniera Marl Formation forms a neritic intercalation at the base of the S.Marino-M.Fumaiolo Sequence (cf. Ruggieri, 1970). At its type locality, 3 kilometers W of Carpegna, this formation attains a thickness of about 75 meters. It is composed of blue-greyish sandy marls with thin-bedded brownish sandstones. Their depocentral affinity is corroborated by reworked marginal elements and olisthostromes of teleallochthonous material.

III.2.3. S.MARINO LIMESTONE FORMATION (LANGHIAN-SERRAVALLIAN)

Massive platform carbonates dominate the S.Marino Limestone Formation of Middle Miocene age (Fig. 61). This formation, defined 8 kilometers NW of Mercatino Conca at M.Titano in the Republic of S.Marino, attains a maximum thickness of a few hundreds of meters (Ricci Lucchi, 1964). The yellow-greyish somewhat dolomitic fossiliferous limestones form the *"pietra francescana"* described by Sacco (1940) and Signorini (1940). The transgressive base of the formation in marginal segments of satellite basins is marked by conglomeratic lenses attesting to erosion of its teleallochthonous substratum (cf. Principi, 1925b, 1934, 1938, 1939; Ruggieri, 1953a, 1953c, 1958, 1970; Ricci Lucchi, 1964; Stern, 1969; Conti, 1989). The transitional aspect of its upper part is indicated by a thinning-upward trend reflecting gradual deepening and concomitant reworking (Ricci Lucchi, 1964; Ruggieri, 1970; Sestini, 1970).

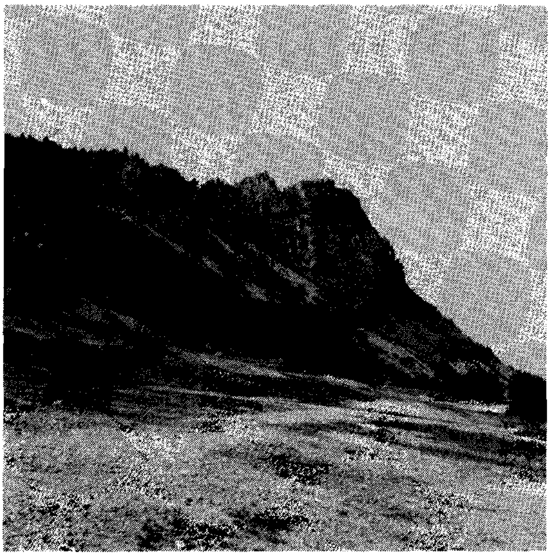


Fig. 61. Massive platform carbonates of the S.Marino Limestone Formation in the ridge of Castello di Monte Copiolo near Villagrande.

III.2.4. M.FUMAIOLO SANDSTONE FORMATION (UPPER LANGHIAN-LOWER TORTONIAN)

The M.Fumaiolo Sandstone Formation essentially is a heteropy of the upper part of the S.Marino Limestone Formation (Ricci Lucchi, 1964; Conti, 1989). It attains a maximum thickness of several hundreds of meters at its type locality, some 15 kilometers W of the investigated area (Malesani & Manetti, 1967; Frey, 1969; Sestini, 1970). The most prominent lithotype of the formation is represented by green-greyish biocalcarenites of coastal origin, composed mostly of clastics derived from the S.Marino Limestone Formation (Ricci Lucchi, 1964, 1981b). Marly intercalations are most common in its upper part, grading into the Montebello Clay Formation (Frey, 1969; Stern, 1969).

III.3. THE VALLE DI TEVA-ONFERNO SEQUENCE

III.3.1. INTRODUCTION

Peneallochthonous terrains constituted by the Lower Pliocene Valle di Teva-Onferno Sequence characterize the frontal portion of the Montefeltro Colata in the investigated area (Fig. 62, 63). They comprise clastic wedges whose internal variability reflects sedimentation on a mobile substratum (Fig.

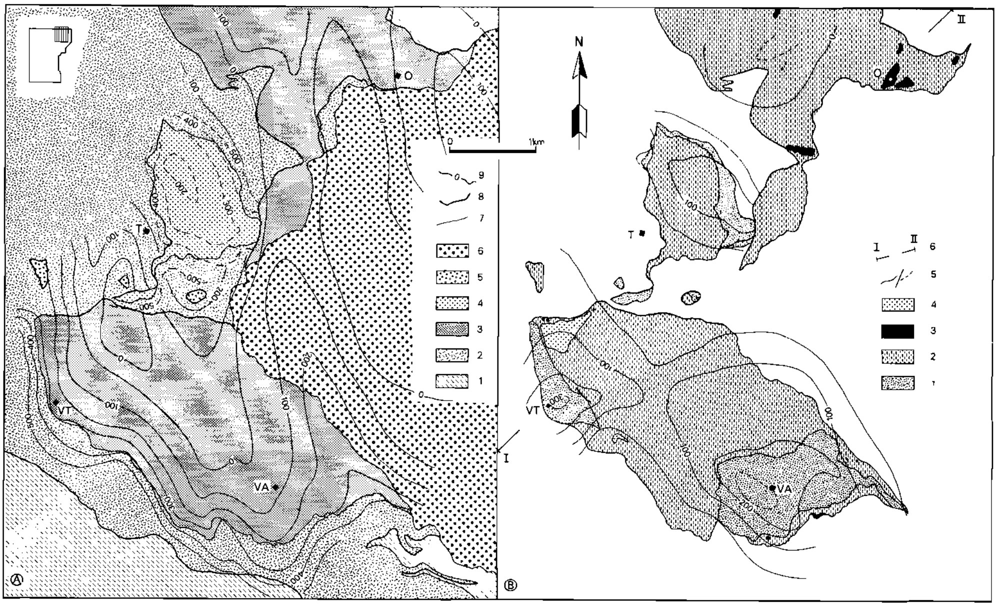


Fig. 62. Peneallochthonous terrains of the investigated area. Location labels: O: Onferno; T: Trebbio; VA: Valle Avellana; VT: Valle di Teva. A: Structural framework. Legend: 1: palaeoautochthon; 2: teleallochthon; 3: peneallochthonous body of Valle Avellana-Onferno; 4: peneallochthonous body of Trebbio; 5: minor peneallochthonous slices; 6: neoautochthon; 7: sedimentary contact; 8: tectonic contact; 9: isobaths of base of peneallochthon (contour interval 100 m). B: Distribution of the Valle di Teva-Onferno Sequence. Legend: 1: Valle di Teva Sandstone Formation; 2: Ca Bertoldo Marl Formation; 3: Onferno Evaporites; 4: C.M.Sabatino Conglomerates; 5: isopachs of Valle di Teva Sandstone Formation (contour interval 100 m/50 m); 6: trace of cross-section of Fig. 63.

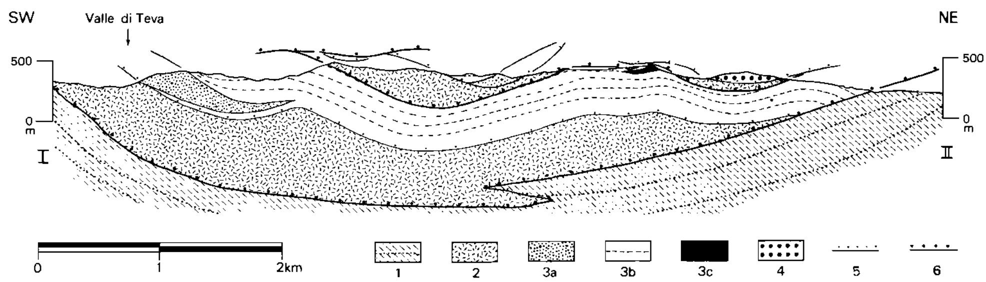


Fig. 63. Section across the frontal portion of the Montefeltro Colata, illustrating the configuration of the peneallochthon. Its trace is indicated in Fig. 62B. Legend: 1: palaeoautochthon; 2: teleallochthon; peneallochthon: 3a: Valle di Teva Sandstone Formation; 3b: Ca Bertoldo

Marl Formation; 3c: Onferno Evaporites; 4: neoautochthon; 5: unconformity; 6: tectonic contact.

64). A local origin is indicated by their similarity to the coeval Umbro-Romagnan molasse of the palaeoautochthon (cf. II.3.2.3.3.). Accordingly, the principal peneallochthonous bodies preserve the geometry of individual satellite basins (Fig. 62).

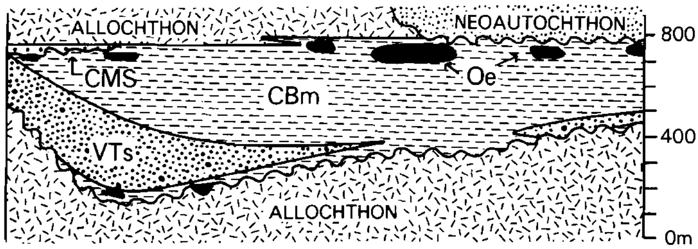


Fig. 64. Lithostratigraphy of the Valle di Teva-Onferno Sequence, comprising the Valle di Teva Sandstone Formation (VTs), the Ca Bertoldo Marl Formation (CBm), the Onferno Evaporites (Oe), and the C.M.Sabatino Conglomerates (CMS).

III.3.2. VALLE DI TEVA SANDSTONE FORMATION (LOWER PLIOCENE)

Fig. 65. Conspicuous fan-deltaic body of Valle di Teva Sandstone Forma-

tion unconformably overlying teleallochthonous terrains near Ca Antoniuccio, 2 km E of Mercatino Conca.

The Valle di Teva Sandstone Formation constitutes prominent lenticular bodies of fan-deltaic origin in the lower part of the Valle di Teva-Onferno Sequence (Fig. 65). These attain thicknesses of several hundreds of meters at Valle Avellana and at Valle di Teva (Fig. 62B). Clastics were supplied by torrential streams draining the accidented frontal portion of the Montefeltro Colata. The brown-greyish sandstones contain gravel and shell lags of coastal affinity. Less common continental aspects include dispersed fresh-water gastropods and palaeopedogenic features such as colour mottling and caliche nodules.

III.3.3. CA BERTOLDO MARL FORMATION (LOWER PLIOCENE)

The principal constituent of the Valle di Teva-Onferno Sequence is the Ca Bertoldo Marl Formation. This attains a thickness of about 400 meters at its type locality between Valle Avellana and Valle di Teva. The bulk of the formation consists of greyish clayey marls, which are relatively sandy at the fringes of the fan-deltaic bodies of the Valle di Teva Sandstone Formation. Their microfaunal contents is indicative of a coastal origin. This is corroborated by intercalated gravel and shell lags and thin-bedded sandstones possibly representing storm layers (cf. Ricci Lucchi *et al.*, 1981). Rare sandstone balls resulted from slumping.

III.3.4. ONFERNO EVAPORITES

Conspicuous displaced bodies of evaporitic material, the thickness of which reaches 50 meters at Onferno, occur at the base and at the top of the Valle di Teva-Onferno Sequence. These Onferno Evaporites correlate with the Messinian Gessoso-solfifera Formation of the palaeoautochthon (cf. II.3.2.2.). They comprise carbonatic and gypsiferous elements relating to a regular series, as evidenced by the voluminous olisthoplakha of Sassofeltrio, 3 kilometers NE of Mercatino Conca (Ruggieri, 1958; Amadesi, 1962; Carloni *et al.*, 1971).

Olistholiths of foetid dolomitic limestones in the surroundings of Valle Avellana represent the *calcare di base*, with yellowish rather massive *cagnini* and brownish cavernous *calcare perciuliato*. The most prominent lithotype of the Onferno Evaporites is formed by the "*spicchiolino*" of megacrystalline selenite in massive layers of swallow-tail twins (cf. Schreiber, 1973). The latter are typically perpendicular to bedding, with apices pointing downward (Amadesi, 1962). According to Mottura's Rule, this implies an upright position (*e.g.*, Ogniben, 1955, 1957; Selli, 1967). The secundary selenite commonly displays an internal zonation parallel to bedding (Fig. 66). This reflects an original laminar fabric (cf. G.C.Carloni & E.Ceretti *in* Selli, 1967; Schreiber, 1973; Longinelli, 1979). Primary laminites are partly preserved at the base of some gypsiferous layers (Fig. 66).

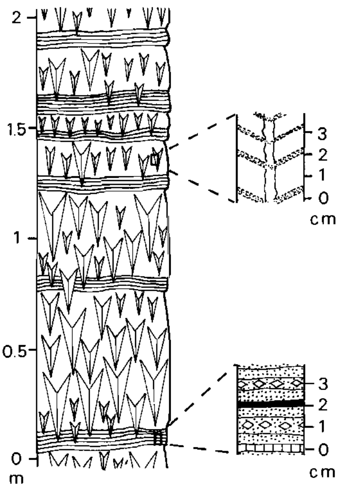


Fig. 66. Schematic columnar section of gypsiferous lithofacies of the Onferno Evaporites at C.Nuova, 3 km ENE of Mercatino Conca. Layers are composed of primary gypsiferous-carbonatic laminites grading upward into secundary spicchiolino of banded selenite.

III.3.5. C.M.SABATINO CONGLOMERATES (LOWER PLIOCENE)

The C.M.Sabatino Conglomerates unconformably cap the Valle di Teva-Onferno Sequence in the surroundings of Valle di Teva, attaining a thickness of a few tens of meters at their type locality. They accumulated as a minor fan-delta fed from the NW by a torrential stream draining tele- and semiallochthonous terrains. Unsorted massive conglomerates resulting from viscous debris flows alternate with water-laid arenaceous-conglomeratic deposits. Their coastal origin is indicated by the co-occurrence of shell lags and palaeopedogenic features such as colour mottling.

CHAPTER IV

QUATERNARY DEPOSITS

Extensive Quaternary alluvium characterizes the major stream valleys of the southern Montefeltro. Terraced deposits labeled as T1 to T4 correlate with successive Pleistocene interglacials (Lipparini, 1939; Villa, 1942). Their distribution vaguely reflects Quaternary tectonic activities (*e.g.*, Carloni *et al.*, 1971; Guerrera *et al.*, 1978; Nesci *et al.*, 1978). Mottled gravels and sands of T1 are present at relative altitudes of about 100 meters in the F. Foglia valley downstream of Lunano and of more than 150 meters in the F. Metauro valley near S.Angelo in Vado. The less distinctive younger terraces occur at relative altitudes of 3 to more than 75 meters.

Rock waste is particularly common in the investigated area. Some calcareous reliefs are flanked by talus slopes. Voluminous ancient landslides descend from protruding bodies constituted by the M.Morello Alberese and S.Marino Limestone formations. They illustrate the general instability of rigid terrains on a plastic substratum (cf. Braga & Marchetti, 1978). Conspicuous landslides lining the thrust front of the Umbro-Romagnan parautochthon have a similar significance. Recent landsliding especially affects the weak components of the teleallochthon. The Sillano Formation gives rise to numerous superficial sheet slides, while the Villa a Radda Formation is mainly degraded by earth flows (cf. Zâruba & Mencl, 1969).

CHAPTER V

TECTONICS OF THE AUTOCHTHONOUS ASSEMBLAGE

V.1. INTRODUCTION

The structural style of the external portion of orogens especially relates to the distribution of detachment zones and the material properties of the deformed terrains (Dahlstrom, 1970; Laubscher, 1978; Bally *et al.*, 1986; Vann *et al.*, 1986). Lithostratigraphy exerts a major control on the style of deformation. Conversely, the lateral variability of Apenninic terrains is largely a result of synsedimentary tectonics. This warrants the distinction of tectono-stratigraphic domains (cf. Lavecchia *et al.*, 1987). Individual elements composing a domain represent tectono-stratigraphic units (de Feyter *et al.*, 1990).

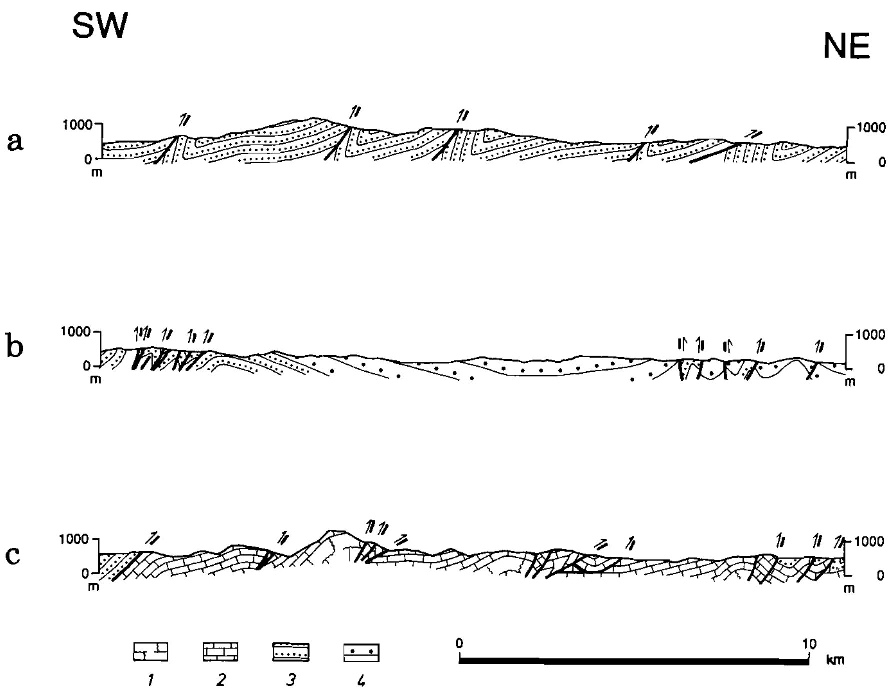


Fig. 67. Schematic cross-sections to illustrate the various structural styles displayed by the autochthonous assemblage of the Umbro-Marchean-Romagnan Zone: a: Umbro-Romagnan parautochthon between the Val Tiberina N of Città di Castello and S.Angelo in Vado (partly after Signorini,

1941); b: Umbro-Romagnan terrains of the autochthon between Sassocorvaro and Saludecio; c: Umbro-Marchean terrains of the autochthon SW of Pergola (partly after Barchi et al., 1989). Legend: Umbro-Marchean Sequence: 1: Calcare Massiccio Formation; 2: Corniola Formation through Bisciaro Formation; Umbro-Romagnan Sequence: 3: turbiditic subsequence; 4: continental to shallow-marine subsequence.

The autochthonous assemblage of the Umbro-Marchean-Romagnan Zone displays marked variations in structural style (Fig. 67). The autochthon, comprising the palaeoautochthon and the neoautochthon of the investigated area, forms a major tectono-stratigraphic domain contiguous with the stable Adriatic Foreland to the NE. Its Umbro-Marchean and Umbro-Romagnan terrains are characterized by different deformational geometries. In the internal part of the zone, the Umbro-Romagnan parautochthon constitutes a separate superficial domain.

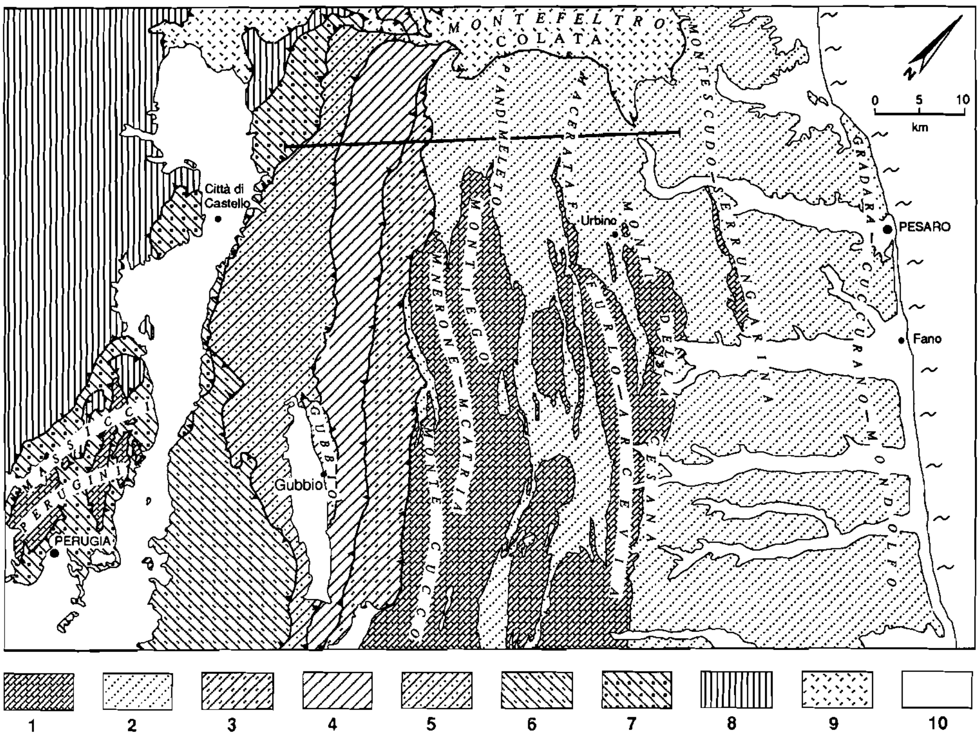


Fig. 68. Tectonic map of the northwestern part of the Umbro-Marchean Apennines, with indication of the trace of the cross-section of Fig. 69. Major anticlines and highs of the autochthon are named. Thrust faults have barbs on their upper block. Legend: autochthon: 1: Umbro-Marchean Sequence; 2: Umbro-Romagnan Sequence; Umbro-Romagnan parautochthon: 3: M.Vicino Unit; 4: Borgo Pace Unit; 5: Pietralunga Unit; 6: M.delle Por-

tole/M.Salaiole-M.Urbino Unit; 7: M.Nero Unit; 8: M.Modino-M.Cervarola Domain; 9: allochthonous assemblage; 10: "Villafranchian" and Quaternary.

Umbro-Romagnan terrains belonging to the parautochthon and to the autochthon are exposed together with the underlying Umbro-Marchean terrains in the northwestern part of the Umbro-Marchean Apennines (Fig. 68). The relationship between their distinctive styles of deformation is elucidated by the plunging structures of this sector (Fig. 69).

V.2. TECTONICS OF THE AUTOCHTHON

V.2.1. STRUCTURAL FEATURES

V.2.1.1. GENERAL ASPECTS

The autochthon of the Umbro-Marchean Romagnan Zone is characterized by more or less continuous NW-SE trending longitudinal chains. At the level of the Umbro-Marchean Sequence, which is extensively exposed in the Umbro-Marchean Apennines SE of the investigated area, it displays anticlinal chains and intervening synclinal zones with isolated positive structures (Fig. 67c, 68). The essentially rootless deformation style of this arcuate belt with outward convexity has been referred to by numerous authors, who emphasized the role of the incompetent Burano Anhydrites Formation as a detachment or disharmony level between the basement and the deformed sedimentary cover (*e.g.*, Selli, 1952; Scarsella, 1964; Centamore & Jacobacci, 1968; Carloni *et al.*, 1971; B.Accordi *in* Desio, 1973; Decandia & Giannini, 1977).

The backbone of the Umbro-Marchean Apennines is formed by the Umbro-Marchean Chain and the more external Marchean Chain, which are separated by the Inner Marchean Basin (Fig. 4). The Umbro-Marchean Chain consists of brachyanticlines arranged systematically in a right-hand *en-échelon* pattern with zig-zag linkage (cf. Campbell, 1958). This fabric is known as "graduated virgation" (Scarsella, 1951; Parotto & Praturlon, 1975; Lavecchia & Pialli, 1980; Lavecchia, 1981). In the northwestern part of the Umbro-Marchean Apennines, it comprises the M.Cucco Anticline, the M.Nerone-M.Catria Anticline, and the Montiego Anticline (Fig. 68).

Unlike its Umbro-Marchean counterpart, the Marchean Chain displays a longitudinally rather continuous anticlinal structure, labeled as Furlo-Arcevia in the northwestern part of the Umbro-Marchean Apennines (Fig. 68).

The major anticlines are typically more or less box-shaped asymmetric NE-facing features. A concentric geometry is most evident at the level of the pelagic series of the Corniola Formation through the Calcareous Scaglia, whose pronounced mechanical anisotropy permitted a flexural-slip mode of deformation (cf. Donath & Parker, 1964; Chapple & Spang, 1974). Instead, the underlying rigid Calcare Massiccio Formation is more severely fractured (*e.g.*,

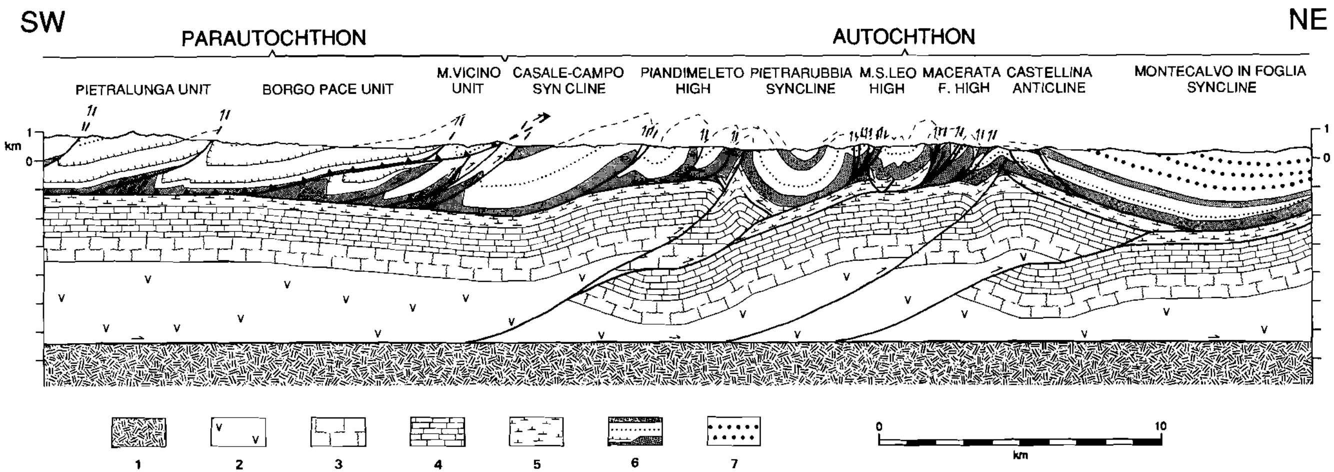


Fig. 69. Balanced section across the investigated area and its surroundings. Its trace is indicated in Fig. 68. The contact between the autochthon and the overriding Umbro-Romagnan parautochthon is barbed. Legend: 1: putative basement; Umbro-Marchean Sequence: 2: Burano Anhydrites Formation; 3: Calcare Massiccio Formation; 4: Corniola Formation through Calcareous Scaglia; 5: Marly Scaglia and Bisciaro Formation; Umbro-Romagnan Sequence: 6: turbidite suite, with shaded proto- and cataturbiditic intervals and indicated Contessa megaturbidites; 7: molasse.

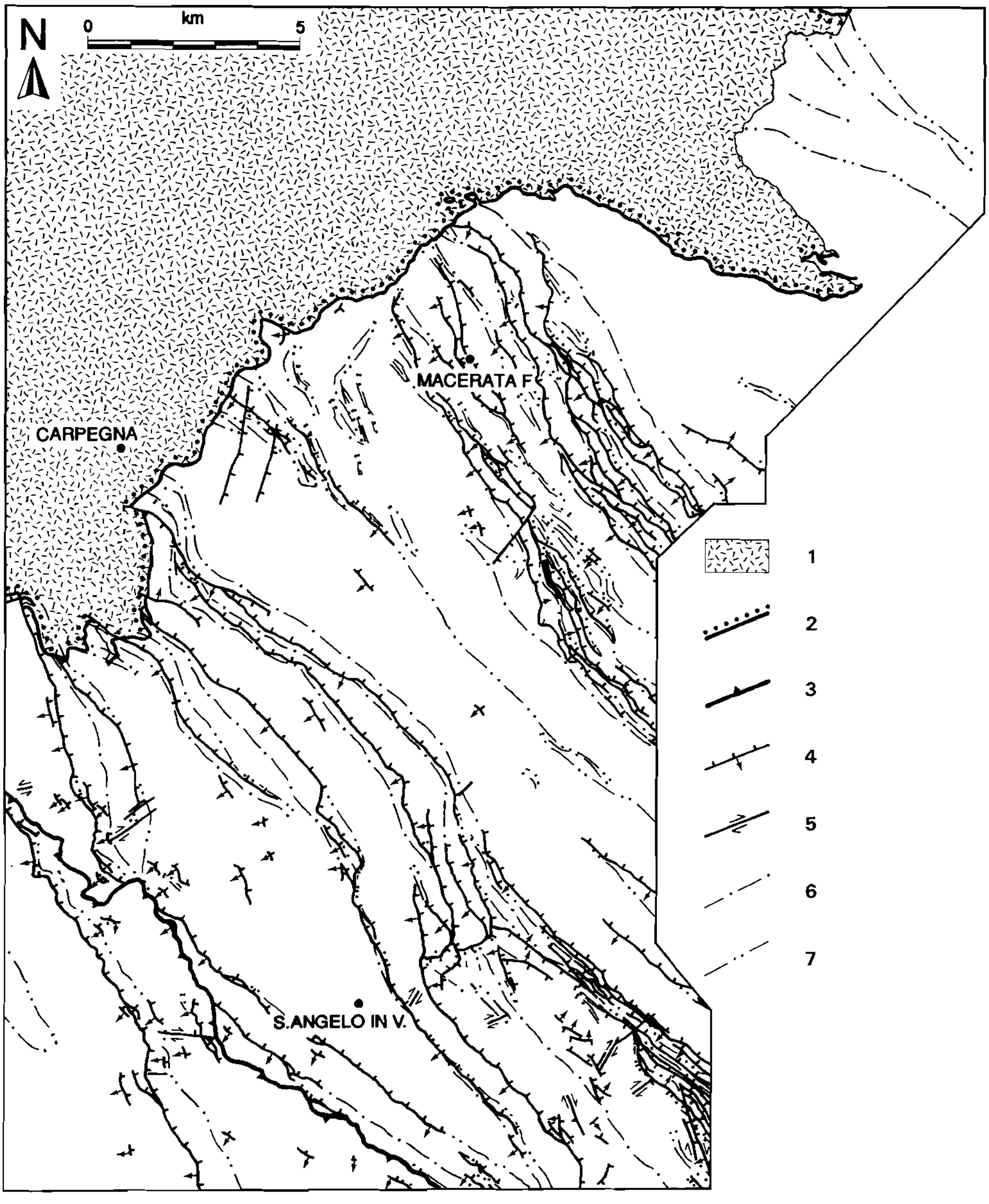


Fig. 70. Structural framework of the autochthonous assemblage of the investigated area. Legend: 1: allochthonous assemblage; 2: transgressive thrust fault, with dots on upper block; 3: major thrust fault, with barbs on upper block; 4: reverse fault, with hachures on downthrown block and arrow indicating sense of dip; 5: strike-slip fault, with indicated sense of slip; 6: anticlinal axial trace; 7: synclinal axial

trace.

Selli, 1949, 1952; Ceretti, 1964; Lavecchia, 1981; Koopman, 1983).

The anticlines to some extent override the adjoining synclines to the NE on stepped thrust faults. Their staircase trajectory relates to lithostratigraphy rather than to geometry, with flats and subsidiary detachments parallel to bedding preferentially developed in incompetent intercalations such as the Bosso Formation, the Marne a Fucoidi Formation, the lower part of the Scaglia Rossa Formation, and the Marly Scaglia (Baldacci *et al.*, 1967; Lavecchia *et al.*, 1984b; Bally *et al.*, 1986). The base of the Schlier Formation constitutes a particularly important disharmony level (Pieri & Mattavelli, 1986). The overlying Umbro-Romagnan terrains display complex highs (Fig. 67b). These are the expression of the subsurface prolongation to the NW of

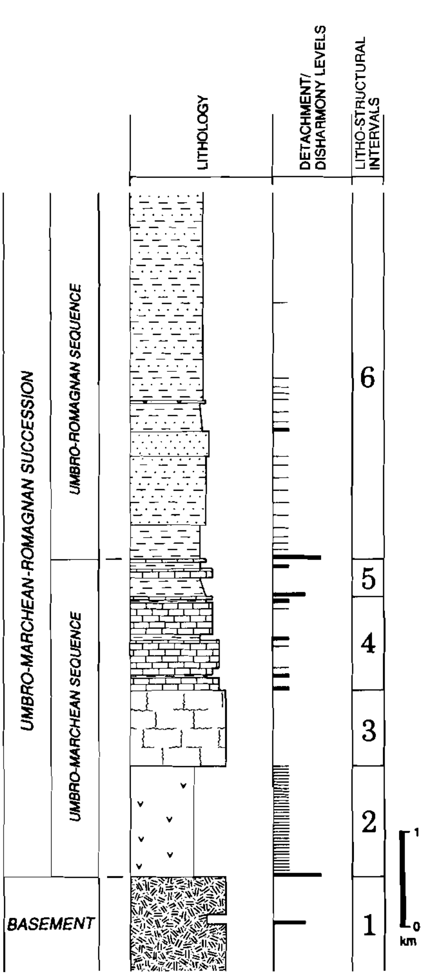


Fig. 71. Schematized lithostratigraphy of the autochthon of the investigated area, with indication of litho-structural intervals. The importance of detachments and disharmony levels increases with length of bar. Note how their distribution relates to competence contrasts. Legend as in Fig. 8.

the anticlinal chains of the Umbro-Marchean Apennines (Fig. 68, 69). The intervening broad synclines are markedly less articulated, as illustrated by the structural pattern of the investigated area (Fig. 70).

Taking into account lithological and consequently mechanical parameters, litho-structural intervals can be defined in the autochthon of the Umbro-Marchean-Romagnan Zone (Fig. 71; cf. Koopman, 1983; de Feyter *et al.*, 1986). The lowest one is represented by the basment. The incompetent Burano Anhydrites Formation and the competent Calcare Massiccio Formation constitute intervals 2 and 3. Interval 4 comprises the pelagic series of the Corniola Formation through the Calcareous Scaglia. Interval 5 is composed of the Marly Scaglia and the Bisciaro Formation. The Umbro-Romagnan Sequence forms the uppermost litho-structural interval.

V.2.1.2. FORETHRUSTS AND BACKTHRUSTS

The thrust faults associated with the major anticlines of the Umbro-Marchean Apennines can be classified according to structural position and sense of slip (Fig. 72). Backlimb and forelimb thrusts affect the inner and the steeper outer fold limbs, respectively (Dahlstrom, 1970; Butler, 1982). The sense of slip of forethrusts matches the direction of principal tectonic transport, while the opposite holds for backthrusts (Bick, 1973; Butler, 1986).

Forelimb forethrusts are upward splays from the main fault steps on which anticlines were thrust toward the NE. In the investigated area, dislocations of this kind cut the Urbania Sandstone and Marne di Letto formations SW of Peglio (Fig. 50; enclosed map and cross-sections I-I' and J-J'). They presumably pertain to a rootless element underlying the Montiego Anticline. This surfaces SE of the investigated area as the Forntino di Naro Anticline (de Feyter & Menichetti, 1986).

Forethrusts in the backlimbs of anticlinal structures generally represent flats associated with stepped faults rooted more internally. The Marly Scaglia of the Montiego Anticline hosts such a detachment level, as evidenced by the pervasively sheared Scaglia Variegata Formation exposed along the road Urbania-Piòbbico in the F.Bottrina valley (cf. Guerrera *et al.*, 1988). It is most likely connected with a stepped thrust fault emanating from the more internal M.Nerone-M.Catria Anticline (de Feyter *et al.*, 1986). In the crestal zone of the Montiego Anticline, leading-edge shortening was accommodated disharmonically by the overlying Bisciaro Formation in a pattern similar to the one reported from the Canadian Rocky Mountains by Fitzgerald & Braun (1965).

Backthrusts dissecting rootless anticlines are characteristic elements of the outer part of orogens. Backlimb backthrusts are typically minor curved reverse faults convex to the principal thrusting direction, which are generated in thrust sheets forced through the synformal zone at the base of a ramp (Morse, 1977; Serra, 1977; Wiltschko, 1979). In the Umbro-Marchean Apennines, they are of negligible importance. Forelimb backthrusts on the other hand are relatively common (de Feyter & Menichetti, 1986). These listric featurs are concave to the principal thrusting direction and show more continuity and offset than average backlimb backthrusts. They supposedly grow upward from the tip of forethrusts propagating along a flat, thus accommodating the shortening at advancing thrust fronts (Fig. 73). This results in a decoupling along the pertinent litho-structural horizon (Morley, 1986). Eventually, a forelimb back thrust may cease activity and be translated passively with the progressing rootless anticline. Shallow back- and forethrusting may then take place more externally.

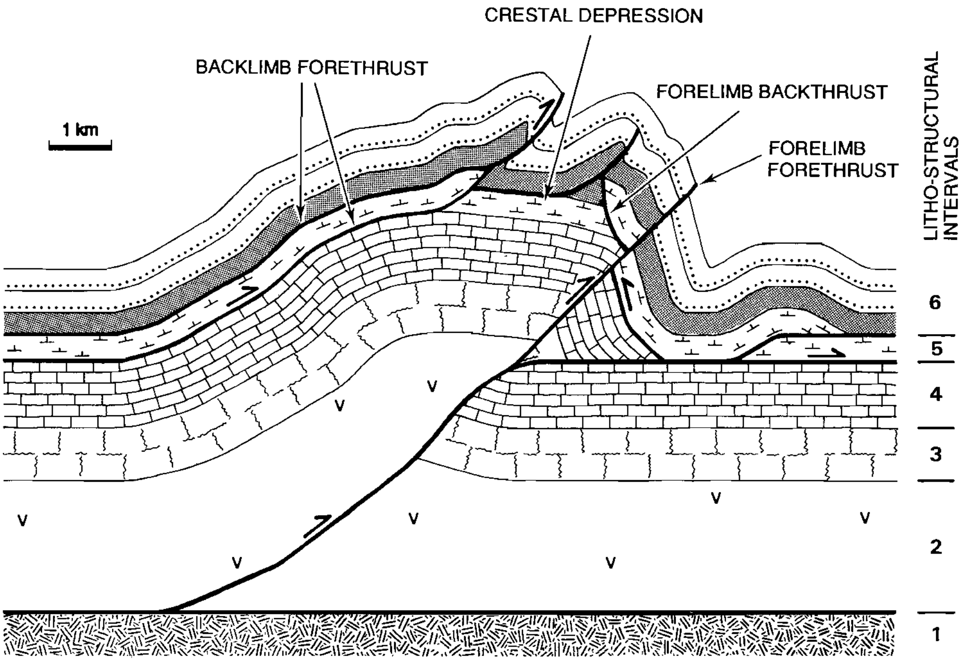


Fig. 72. Terminology of thrust faults associated with rootless anticlines of the Umbro-Marchean Apennines. Structural elements as displayed by the Montiego Anticline in the investigated area. Litho-structural intervals are defined in Fig. 71.

Forelimb backthrusts evidently occupy the same litho-structural position as the associated forethrusts at depth. The conspicuous forelimb backthrusts emanating from the Marly Scaglia in front of the Montiego Anticline thus indicate that the shortened terrains were sheared off the backlimb of the buried Frontino di Naro Anticline at the same level (enclosed map and cross-sections I-I' through K-K').

The crestal zone of a rootless anticline may become enclosed between backlimb forethrusts and forelimb backthrusts (Gwinn, 1964; Dahlstrom, 1970). A crestal depression generated accordingly is displayed by the Montiego Anticline at the level of the lower member of the Bisciaro Formation (Fig. 72; enclosed map and cross-section I-I'). Cut-off relationships attest to the general penecontemporaneity of backlimb forethrusts and forelimb back-

I

II

III

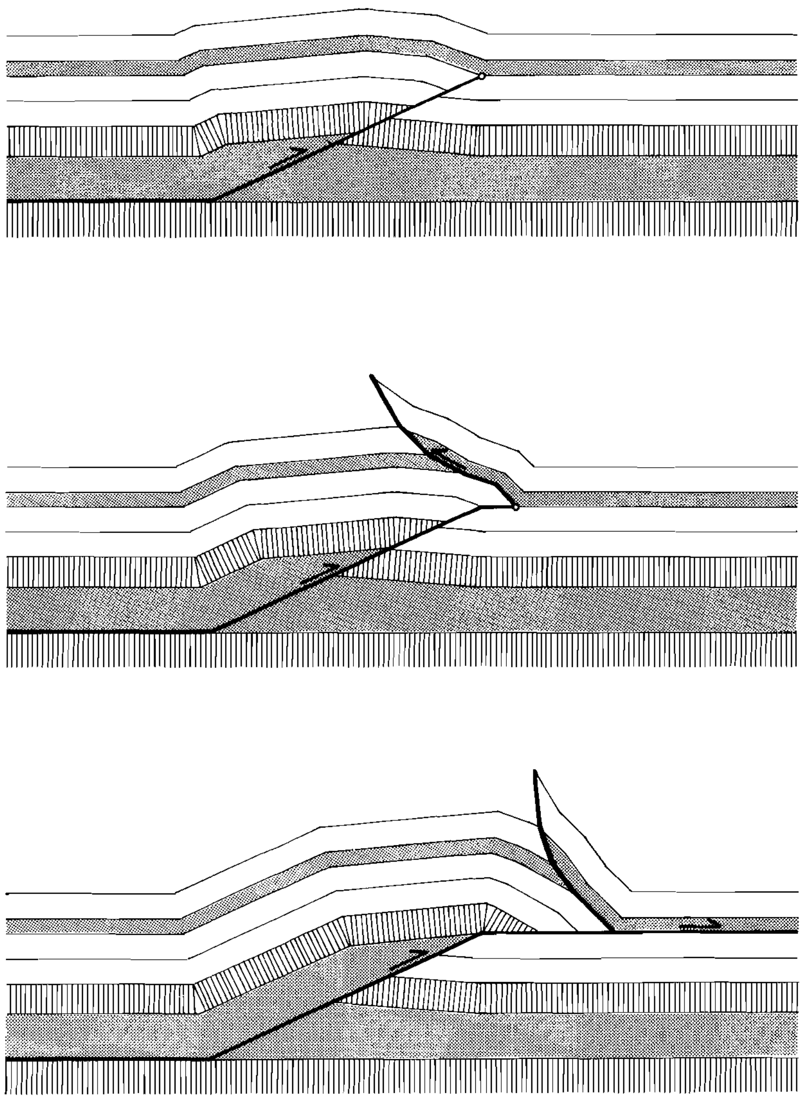


Fig. 73. Evolution of a forelimb backthrust from the tip of a stepped thrust fault propagating along a flat. With reference to the Umbro-Marchean Apennines, shaded incompetent intervals represent the Burano Anhydrites Formation and the Marly Scaglia, and hatched competent ones the basement and the Calcare Massiccio Formation. Effects of erosion are omitted.

thrusts. This implies that the major anticlines of the Umbro-Marchean Chain experienced the relative deformational stage simultaneously, if not in an overstep sequence.

Shallow backthrusts dissecting the Urbania Sandstone and Marne di Letto formations in the forelimb of the Castellina Anticline E of Sassocorvaro correlate with a system individuated in the forelimb of the Furlo Anticline near Urbino by Guerrera (1977). With reference to the concept outlined by Calamita (1986) and Calamita & Deiana (1986a), Capuano & Giampieri (1989) described this as a major feature suturing the Marchean Chain. Dislocations of this kind are not uncommon at the front of orogenic belts, where they characterize triangle zones (Dahlstrom, 1970; Jones, 1982, 1987). Their presence in the Umbro-Marchean Apennines however is unlikely (Bally *et al.*, 1986).

V.2.1.3. SUPERFICIAL HIGHS

The Umbro-Romagnan terrains of the autochthon constitute broad synclinal zones separated by complex highs of tight anticlines and anastomosing reverse faults (Fig. 67b, 70). With respect to the geometric configuration of the underlying Umbro-Marchean terrains, the main synclines typically are rather passive features, while the disharmonic highs mark the crestal zone of anticlinal structures. Their origin must generally be attributed to splaying at the leading edge of stepped thrust faults emanating from more internal anticlines (de Feyter & Menichetti, 1986; de Feyter *et al.*, 1986). The Macerata Fèltria High, which overlies the crestal zone of a rootless anticline pertaining to the Marchean Chain, accordingly represents the leading edge of a stepped thrust system rooted beneath the more internal structures of the Umbro-Marchean Chain. As this high involves the Scaglia Cinerea and Bisciaro formations just SE of the investigated area, the main flat probably is developed at the base of the Marly Scaglia. The M.S.Leo High may then relate to a subsidiary flat of this system at the base of the Schlier Formation (Fig. 69).

The Piandimeleto High is of a more hybrid character (Fig. 69; enclosed map and cross-sections E-E' through J-J'). Its principal component is a conspicuous anticline at the tip of the stepped thrust fault on which the Montiego Anticline overrides the backlimb of the Frontino di Naro Anticline. More internal structures are situated on the crestal zone of the Montiego Anticline above a flat at the base of the Schlier Formation. S of the F. Metauro valley, this largely belongs to the stepped thrust system associated with the M.Nerone-M.Catria Anticline (de Feyter *et al.*, 1986). Towards the NW, it becomes increasingly linked up with the thrust front of the Umbro-Romagnan parautochthon. The frontal footwall structures of the latter to some extent resemble the more external highs. This is most evident S of the F.Metauro valley, where they are developed on the crestal zone of the fading M.Nerone-M.Catria Anticline.

The anticlinal structures of the superficial highs have been described as "pseudodiapirs" (Selli, 1949, 1952; Ceretti, 1964; R.Selli *in* Selli, 1967; Carloni *et al.*, 1971; Carloni & Zecchi, 1979). This refers mainly to their conspicuous tight shape with thickened hinges at the level of the Schlier Formation. Except for the anticlinal cores, the hinges are not significantly disrupted (Fig. 74). The forelimbs instead are commonly cut by high-angle reverse faults. This structural style is typical of hybrid features resulting from a combination of detachment and fault-propagation folding (Suppe & Medwedeff, 1984; Mitra, 1986; Jamison, 1987). With respect to the first mechanism, their tightness is indicative of rather thin "efficient" detachment zones (Dahlstrom, 1990). This concentrated shearing required high pore-fluid pressures counteracting frictional resistance at the mechanical inhomogeneities represented by the base of the Marly Scaglia and the base of the Schlier Formation (cf. Hubbert & Rubey, 1959; Gretener, 1972, 1981). Splaying took place where the propagation of these detachments was obstructed by lower pore-fluid pressures. The position of the major highs on the crestal zone of rootless anticlines, implying the pre-existence of the latter (cf. Dahlstrom, 1970), accordingly reflects the light overburden of synsedimentary reliefs. The

Fig. 74. Anticlinal hinge in the Marnoso-arenacea Formation B near M.S. Leo, 3 km ENE of Lunano.

rapid upward steepening of the splays is compatible with a shallow origin (Price, 1977).

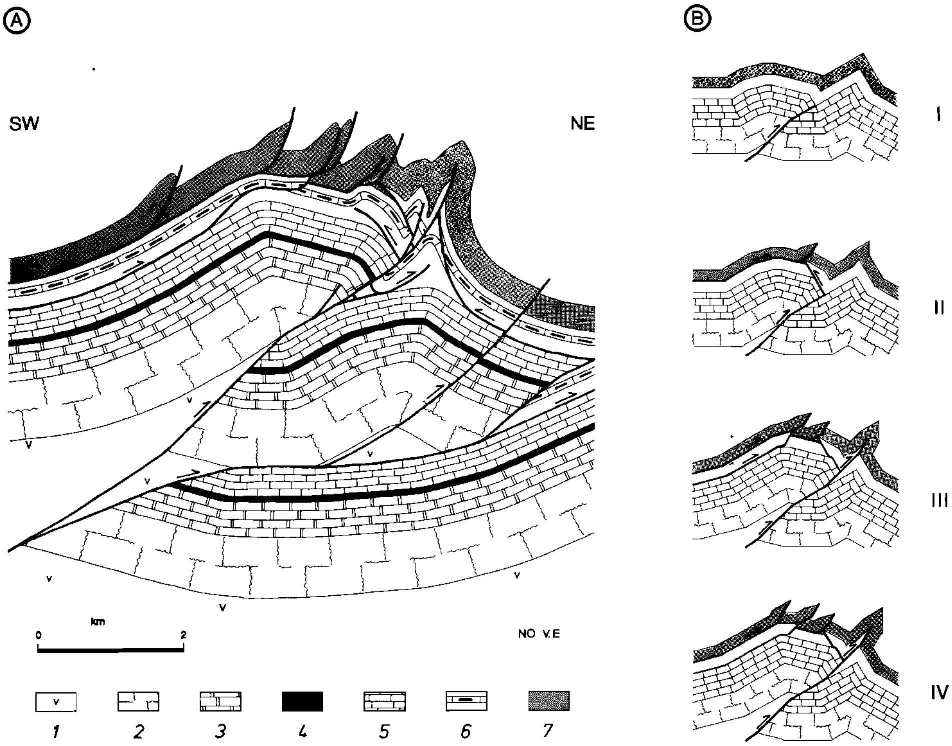


Fig. 75. A) Synthetic section across the Montiego Anticline and the superficial Piandimeleto High. The underlying element represents the Frontino di Naro Anticline. Legend: 1: Burano Anhydrites Formation; 2: Calcare Massiccio Formation; 3: Corniola Formation through Maiolica Formation; 4: Marne a Fucoidi Formation; 5: Calcareous Scaglia; 6: Marly Scaglia and Bisciaro Formation; 7: Schlier Formation. B) Evolution of shallow structures. Backlimb forethrusting and forelimb backthrusting interact to produce a crestal depression, while at a higher level the Piandimeleto High grows through overstep imbrication. Symbols of Calcare Massiccio Formation and Schlier Formation as in A); Corniola Formation through Calcareous Scaglia shown as single interval, and Marly Scaglia and Bisciaro Formation in white.

The composite superficial highs can be regarded as imbricate stacks. Both forward, or piggyback, and hindward, or overstep, sequences of fault emplacement have been envisaged for such structures (Dahlstrom, 1970; Boyer & Elliott, 1982; Jones, 1984, 1987). The hindward oversteepening of imbricates, which is most evident in the M.S.Leo High, appears indicative of a piggyback mechanism (cf. Mulugeta & Koyl, 1987). However, shortening at depth combined with gravitative backtilting may produce a similar effect (Balkwill, 1972). The irregularity of the superficial highs, with locally anomalous structural relationships, in fact demonstrates that overstep imbrication prevailed (cf. Butler, 1982; Morley, 1988). This is confirmed by the concentrated disharmonic shortening on the crestal zone of anticlinal structures (Fig. 75). As imbricate stacks evolved through the accretion of backlimb material, lithofacies indicative of the crest of the embryonic synsedimentary reliefs typically occur at their leading edge (*e.g.*, Fig. 28, 48).

V.2.1.4. LONGITUDINAL, TRANSVERSAL, AND OBLIQUE FAULTS

Longitudinal, transversal, and oblique high-angle faults form an integral part of the structural style of the autochthon of the Umbro-Marchean-Romagnan Zone (Selli, 1952). In the investigated area, they are most evident in the Umbro-Marchean terrains of the Montiego Anticline. A major WSW-dipping dextral strike-slip fault cuts its forelimb (Fig. 76). This oblique feature has an impressive continuity, crossing the entire anticline to reach the more internal M.Nerone-M.Catria Anticline near Piõbbico (de Feyter *et al.*,

Fig. 76. Major NNW-SSE trending oblique dextral strike-slip fault dissecting the lower member of the Bisciaro Formation in the forelimb of the Montiego Anticline near Porreo, 6 km S of Peglio.

1986). Its rectilinearity points to a rather late emplacement. This is corroborated by a more or less pronounced reverse component of slip and disparate forelimb deformation patterns on either side of it. Relative to these superficial structures alone, it therefore constitutes a primary tear fault (Dahlstrom, 1969, 1970). Similar NNW-SSE trending faults, albeit of more modest dimensions, occur throughout the Montiego Anticline. This oblique fault set is the most important one of the Umbro-Marchean Apennines (Lavecchia & Pialli, 1980; Lavecchia *et al.*, 1980, 1984a; Lavecchia, 1982; Marshak *et al.*, 1982; Calamita & Deiana, 1986a, 1986b). Their conjugate set of E-W trending sinistral strike-slip faults is not represented in the investigated area.

The NE-SW to ENE-WSW trending faults invariably are characterized by a dextral component of strike-slip. The most conspicuous one dissects the forelimb of the Montiego Anticline in the F.Bottrina valley as a primary tear fault (de Feyter & Menichetti, 1986). A transversal fault terminating against a backthrust surface 1 kilometer further SE has a similar significance (cf. Dahlstrom, 1970). The transversal faults associated with the Montiego Anticline are clearly shallow secondary features which do not substantiate the prominent role transversal lineaments allegedly played in the evolution of the Umbro-Marchean Apennines (Bocaletti *et al.*, 1983).

The few individuated longitudinal normal faults are minor features concentrated in the crestal zone of the Montiego Anticline (Fig. 9). They represent local relaxation effects not necessary related to the important late extensional tectonics recognized elsewhere in the Umbro-Marchean Apennines (*e.g.*, Lavecchia *et al.*, 1987).

Interrupted lithostratigraphic and structural trends commonly are the sole indications of major faults dissecting Umbro-Romagnan terrains. In this respect, there is no trace of the large oblique faults Ruggieri (1958, 1970) and Ardanese & Martelli (1983-1984) considered to cross the Pietrarubbia Syncline, nor of the transversal faults Conti (1989) ascribed to the Castellina Anticline. Prominent transversal faults displaying a dextral component of strike-slip do however affect the Casale-Campo Syncline near Campo, 2 kilometers SW of Belforte all'Isauro, and the M.S.Leo High near Caprazzino-Strada, 4 kilometers NNE of Lunano. They represent primary tears separating different structural patterns and must be rooted in a shallow detachment level, probably at the base of the Schlier Formation. Minor faults evidenced by fresh outcrops such as in the F.Foglia and F.Metauro beds predominantly pertain to the same system.

Conspicuous longitudinal and transversal normal faults whose formation was determined by the emplacement of the Montefeltro Colata occur in the surroundings of Frontino (cf. VI.4.). These of course do not relate to the regular structural style of the autochthon.

V.2.1.5. COLLAPSE STRUCTURES

Minor folds and reverse faults complicate the Pietrarubbia Syncline at the level of the Colombacci Formation between Frontino and Macerata Fèltria (Fig. 70). Venzo (1954), Ruggieri (1970) and Conti (1989) depicted these as normal components of the structural style of the autochthon. However, the regularity of the underlying Gessoso-solfifera Formation indicates that they in reality are disharmonic superficial features (enclosed cross-sections D-D' through F-F'). Considering the open shape of the Pietrarubbia Syncline and the NE-vergence of the minor structures irrespective of their position relative to the putative synclinal hinge, this geometry clearly does not reflect the out-of-the-syncline expulsion of material from a tightening fold (cf. Dahlstrom, 1970). Instead, it must have resulted from gravitational sagging within the evolving Pietrarubbia Syncline in a pattern compatible with the asymmetry of the latter.

Less conspicuous collapse structures are displayed by the Marne di Letto Formation in the steep inner limb of the Pietrarubbia Syncline (enclosed map and cross-section I-I'). They range from small listric reverse faults to recumbent folds. The origin of such cascades is generally attributed to gravitational gliding on incompetent horizons down steepening fold limbs (Harrison

Fig. 77. Shear zone subparallel to bedding in the Colombacci Formation in the forelimb of the Castellina Anticline near C.Belvedere, 2 km SW of Auditore. The topmost colombaccio is visible in the upper left of the photograph. It is cut by a small secondary normal fault. The main shear zone is characterized by bands of tension gashes, of which a particularly conspicuous one is outlined above the hammer. It is dissected by subhorizontal shear fractures. Both the curvature of the tension gashes (Cloos, 1955; Shainin, 1960; Roering, 1968; Lajtai, 1969; Durney & Ramsay, 1973) and the attitude of calcitic slickenfibers on the shear surfaces bounding this band attest to the indicated NE-ward downlimb tectonic transport.

& Falcon, 1934, 1936).

The only shear zone encountered in the investigated area on which downlimb gravitational gliding took place is developed in the uppermost part of the Colombacci Formation in the forelimb of the Castellina Anticline (Fig. 77).

V.2.2. STRUCTURAL EVOLUTION

V.2.2.1. TIMING OF SUPERFICIAL DEFORMATION

The lateral variability of synorogenic deposits generally permits to specify the age and intensity of tectonic events. Thus, the shortening of the autochthon of the Umbro-Marchean-Romagnan Zone is commonly considered in terms of more or less discrete phases, notably a Late Miocene one and an Early Pliocene one (*e.g.*, Selli, 1949, 1952; Giannini & Tongiorgi, 1962; Centamore *et al.*, 1980b). The data gathered in the investigated area however tell a different tale. The somewhat reduced thickness of the Bisciaro Formation in the crestal zone of the Montiego Anticline suggests that the major anticlinal chains nucleated during the Early Miocene (cf. II.3.1.4.8.). There is no proof of their pre-existence as advocated by Centamore & Jacobacci (1968), Centamore *et al.* (1972, 1980a) and Jacobacci *et al.* (1974).

After an Early Langhian to Middle Serravallian period of relative tectonic quiescence, reflected by the uniformity of the Schlier Formation, the anticlinal chains experienced a renewed rather continuous growth which consequented the systematic lateral variability of the ortho- and cataturbiditic components of the Umbro-Romagnan turbidite suite (cf. II.3.2.1.). The structural accentuation of the Umbro-Marchean Chain culminated during the Late Tortonian to Messinian with the formation of the Piandimeleto High. This is indicated by the conspicuous gravity displacement structures and coarse clastic intercalations of local provenance in the Marne di Letto, Gessoso-solfifera, and Colombacci formations in the Pietrarubbia Syncline (cf. II.3.2.1.7. through II.3.2.3.2.). Similar features displayed by the Gessoso-solfifera, Colombacci, and Santerno formations more externally attest to accelerated shortening at the Marchean Chain with the formation of the Macerata Fêltria High during the Messinian to Early Pliocene (cf. II.3.2.2. through II.3.2.3.3.). The neighbouring M.S.Leo High and Castellina Anticline must have attained their actual shape roughly at the same time. The coarse clastic bodies of the but gently folded Montecalvo in Foglia Formation indicate that weak tectonic pulses still affected the northeastern part of the investigated area during the Middle Pliocene (cf. II.3.2.3.4.; Amadesi, 1962; Capuano *et al.*, 1986b; Conti *et al.*, 1987; Conti, 1989). Subsequent regional uplift completed the structural evolution (*e.g.*, Ruggieri, 1958; Centamore *et al.*, 1980b; Ambrosetti *et al.*, 1981).

The picture emerging is that of more or less continuous Neogene shortening with momentary intensifications. The gradual outward shifting of these peaks agrees with the migratory pattern generally ascribed to Apenninic tectonics (*e.g.*, Merla, 1951; Lavecchia *et al.*, 1987). The validity of this concept is demonstrated by the Middle Pliocene age of the orogenic paroxysm in more external segments of the Umbro-Marchean-Romagnan Zone (Lucchetti, 1959; Bongiorni, 1964; Ori *et al.*, 1986). In fact, shallow tectonic activity still takes place beneath the Adriatic Sea NE of the investigated area (Malaroda & Raimondi, 1957; Lipparini, 1969; Carloni & Zecchi, 1979).

V.2.2.2. ROLE OF THE BASEMENT

Scarsity of data seriously hampers an assessment of the effect basement tectonics had on the structural evolution of the autochthon of the Umbro-Marchean-Romagnan Zone. Gravimetric surveys typically yield a regional Bouguer iso-anomaly pattern which is discrepant to superficial trends (Fig. 78; Scarsella, 1955; Pinna & Giannessi, 1981). A general disharmony between the basement and its sedimentary cover has also been pointed out by refraction seismic and aeromagnetic studies (Alfano *et al.*, 1982; Lavecchia *et al.*, 1984b; Cassano *et al.*, 1986).

Lavecchia *et al.* (1980, 1984a) and Lavecchia (1982) regarded the tectonics of the Umbro-Marchean Apennines as a superficial effect of basement wrenching. Indeed, the individuated structural pattern of the basement is compatible with the maximum principal stress direction inferred for its deformed cover (cf. Wilcox *et al.*, 1973). It is however doubtful whether large stresses could have been transmitted systematically by the thick incompetent Burano Anhydrites Formation. A comparison with the Zagros Orogenic Belt of southwestern Iran suggests that the observed compatibility is better explained by independent shearing of basement and cover in response to a regional stress field (Colman-Sadd, 1978).

The depth of the basement of the Umbro-Marchean Apennines is still a matter of debate. Thin-skinned fold-and-thrust belts are generally considered to have formed on a hindward-dipping monoclinal basement (Chapple, 1978). This is one of the principles originally formulated for the Canadian Rocky Mountains (Bally *et al.*, 1966; Dahlstrom, 1969, 1970; Jones, 1971; Price, 1973). Its validity has been demonstrated for the offshore portion of the Umbro-Marchean-Romagnan Zone by reflection seismic investigations (Bally *et al.*, 1986). Extrapolation towards the Umbro-Marchean Apennines would result in basement depths of more than 10 kilometers. This implies extensive structural duplications, with a superficial shortening of more than 60 % (Roeder, 1984; Bally *et al.*, 1986; Calamita, 1986; Calamita & Deiana, 1986a, 1986b; Hill & Hayward, 1988). However, the demonstrable shortening of the autochthon in the investigated area does not exceed 30 %, implying a basement depth of about 6.5 kilometers (Fig. 69). A basement step must thus be assumed between the investigated area and the Adriatic Sea. This possibly resulted from crustal imbrication beneath the Montescudo-Serrungarina High (cf. Passerini, 1961; Philip, 1987; Royden *et al.*, 1987). Like the basement thrusts of the more internal Massicci Perugini depicted by Alfano *et al.* (1982), Damiani *et al.* (1982-1983) and Lavecchia *et al.* (1984b, 1987), this deep-seated shearing probably post-dated the paroxysmal emplacement of superficial structures.

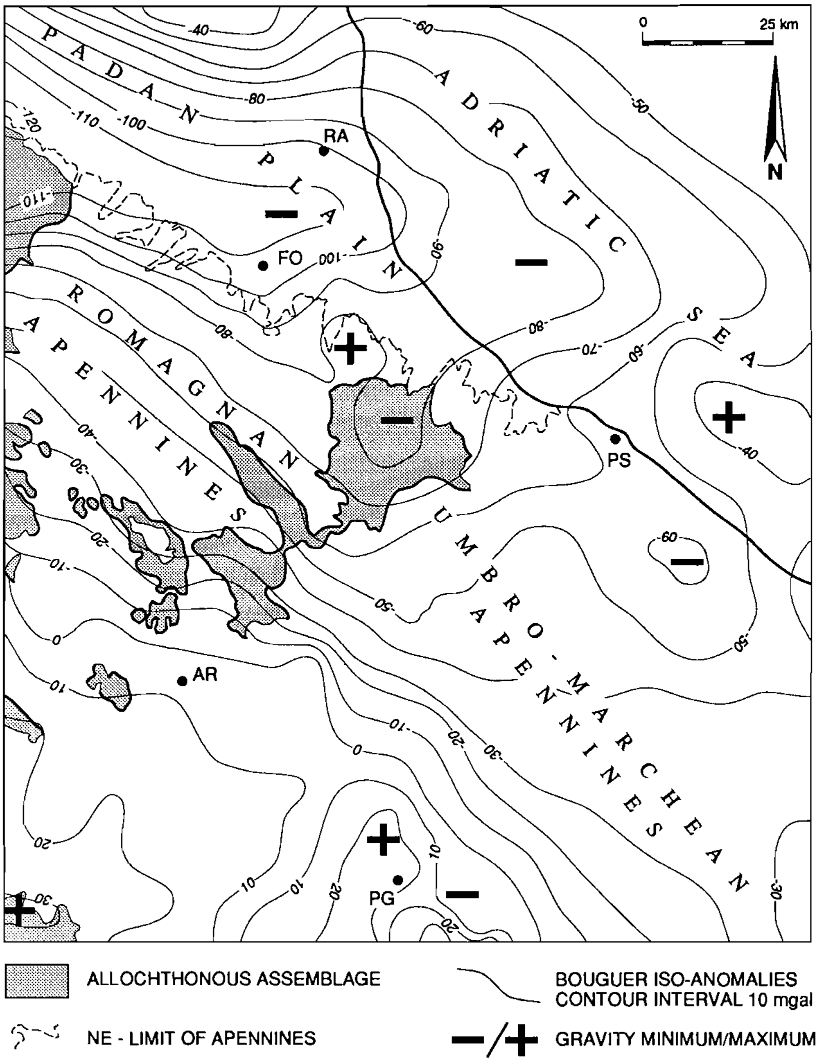


Fig. 78. Gravity map of the central part of the Umbro-Marchean-Romagnan Zone, compiled from Morelli (1955a, 1955b), Ogniben (1975) and Lavecchia & Pialli (1981a). Pronounced minima mark the Padan-Marchean-Adriatic Foredeep. The Montefeltro Colata coincides with a minor depression. Bordering highs appear as relative maxima. More internally, the pattern is complicated by late basement tectonics.

V.2.2.3. ROOTLESS FOLDING AND STEPPED THRUSTING

Rich (1934) first described rootless folding as a result of stepped thrusting of unfolded "passive" strata, a concept which has been markedly refined in recent years (*e.g.*, Berger & Johnson, 1980; Suppe, 1983; Williams, 1987). Koopman (1983) envisaged such fault-bend folding for the Umbro-Marchean Apennines. This is however contradicted by a generally recognized posteriority of stepped thrusting to folding (*e.g.*, Selli, 1949, 1952; Calamita & Deiana, 1982; Lavecchia *et al.*, 1987, 1988; Barchi *et al.*, 1989). As suggested by Lavecchia (1981), initial shortening probably occurred through buckling. In particular, the simultaneous nucleation of anticlinal chains during the Early Miocene can be attributed to this mechanism. The inferred gentle folding required a ductile detachment zone with mobile material to fill anticlinal cores (Dahlstrom, 1990). This is corroborated by the thoroughly deformed state of the Burano Anhydrites Formation at the base of the sedimentary cover (cf. II.3.1.1.). Its structural style is compatible with the viscous shearing model outlined by Kehle (1970).

The concept of buckling of a detached multilayer predicts a linear relationship between the thickness a_D of its "dominant member", which conditions the structural geometry, and the wavelength $_D$ of incipient folds (Biot, 1961, 1964; Ramberg, 1963). Currie *et al.* (1962) determined empiric-

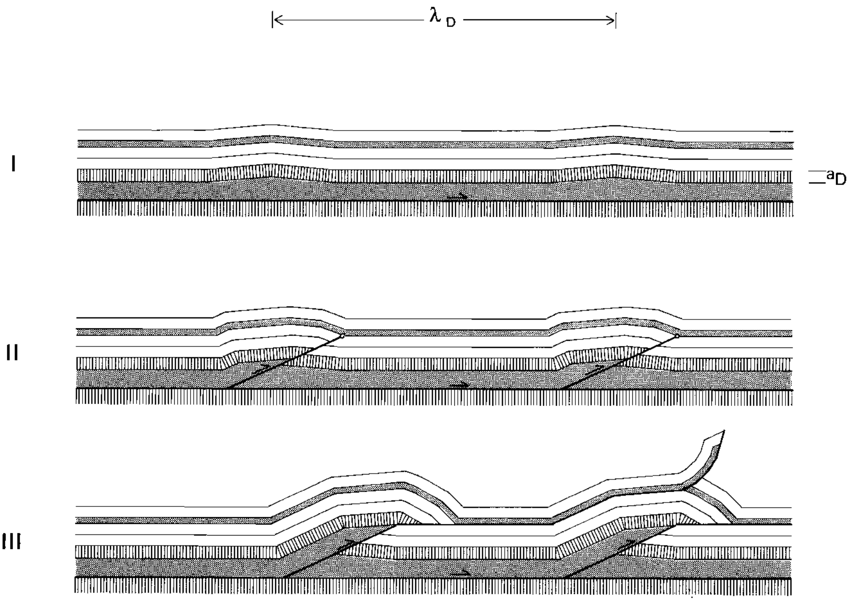


Fig. 79. Schematized Neogene structural evolution of the autochthon of the Umbro-Marchean-Romagnan Zone through buckling followed by stepped thrusting. Shaded incompetent intervals correspond with the Burano Anhydrites Formation and the Marly Scaglia through Schlier Formation, and hatched competent ones with the basement and the Calcare Massiccio Formation. The thickness a_D of the latter determined the wavelength $_D$ of incipient folds.

ally that the multiplicand approximates 27. As Price (1967) pointed out, this applies to the spacing of anticlinal chains rather than individual anticlines.

The rigid Calcare Massiccio Formation undoubtedly constitutes the dominant member of the sedimentary cover of the Umbro-Marchean-Romagnan Zone (Lavecchia, 1981). Its average thickness of 750 meters would imply a fold wavelength of about 20 kilometers. However, the more or less complete longitudinal chains are only 10 to 15 kilometers apart in the northwestern segment of the Umbro-Marchean Apennines (Fig. 68). As schematized in Fig. 79, this reflects the continuation of shortening from the Middle Miocene on. Progressive failure of the Calcare Massiccio Formation in tightening anticlinal cores determined the onset of stepped thrusting (cf. Gretener, 1972; Eisenstadt & De Paor, 1987). The further growth of the rootless anticlines mainly resulted from fault-propagation folding, whereby diapiric behaviour of the Burano Anhydrites Formation locally may have played a minor role (Coli, 1980; Bally *et al.*, 1986). The migratory aspect of Apenninic tectonics primarily relates to the final splaying which originated superficial highs.

The validity of the outlined evolutionary model is demonstrated by the similarity of the deformation pattern of the Umbro-Marchean Apennines to structures generated experimentally by Blay *et al.* (1977). These authors showed that initial superficial buckling yields widely spaced *en-échelon* arrays of anticlines (Fig. 80). With continued shortening, these may coalesce into

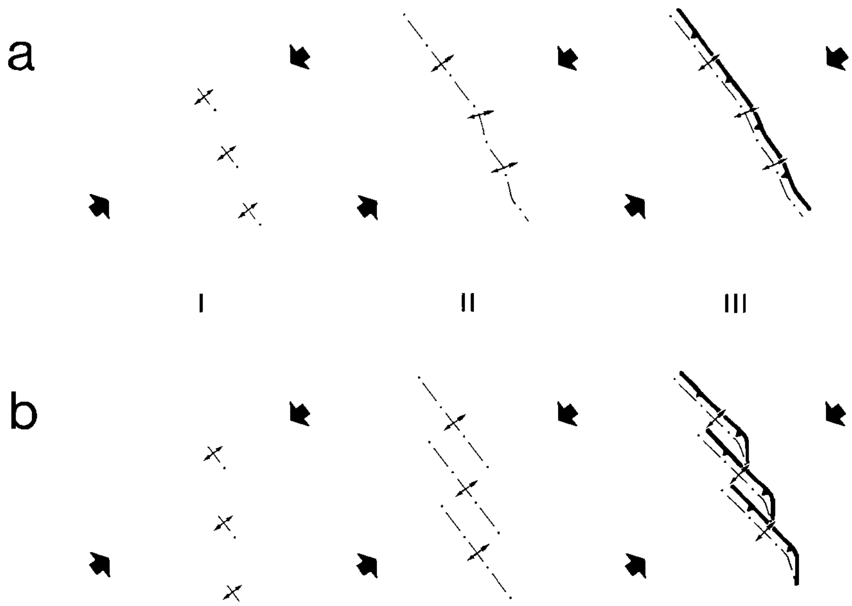


Fig. 80. Patterns of superficial tangential shortening, based on experiments carried out by Blay et al. (1977). Initial en-échelon anticlines may coalesce into a simple longitudinal chain ultimately accentuated by thrust faulting (a), or they may be amplified and eventually thrust faulted as individual brachyanticlines within a composite chain (b).

simple longitudinal features such as the Marchean Chain. On the other hand, the fabric resulting from the amplification of individual anticlines resembles the graduated virgation displayed by the Umbro-Marchean Chain. Precocious failure of the Calcare Massiccio Formation may have prevented the unification of brachyanticlines (de Feyter *et al.*, 1986). Brittle behaviour was possibly induced by a relatively high strain rate (cf. Biot, 1961). The same may apply to the more internal Gubbio-M.Subasio-M.Martani Chain, which also displays graduated virgation (Parotto & Praturlon, 1975; Lavecchia *et al.*, 1984b). Like the neighbouring Massicci Perugini, this discontinuous feature was not accentuated during the Middle to Late Miocene, probably because of the thick damping cover of Umbro-Romagnan foredeep deposits (cf. Morley, 1987). Only during the Pliocene, actual geometries were attained as a superficial effect of late basement shearing (de Feyter *et al.*, 1990).

The superficial deformation of the autochthon of the Umbro-Marchean-Romagnan Zone during the Neogene has frequently been attributed to gravity tectonics (*e.g.*, Giannini & Tongiorgi, 1962; Baldacci *et al.*, 1967; Elter & Trevisan, 1973; Boccaletti *et al.*, 1980). Indeed, its kinematics are compatible with a mechanism of gravitational spreading (Dal Piaz, 1942; Bucher, 1956; Elliott, 1976; Cooper, 1981). Van den Berg (1990), elaborating the concept of critical taper introduced by Chapple (1978) and Davis *et al.* (1983), showed how the shortening of the outer part of the Northern Apennines could take place at the front of a spreading orogenic wedge. Ultimately, this process must have been controlled by tangential loading in the internal part of the orogen (cf. Price, 1973).

V.3. TECTONICS OF THE UMBRO-ROMAGNAN PARAUTOCHTHON

V.3.1. STRUCTURAL STYLE

The Umbro-Romagnan parautochthon constitutes a superficial tectono-stratigraphic domain characterized by broad NE-facing asymmetric synclines separated by listric reverse faults of impressive longitudinal continuity (Fig. 67a). Signorini (1940, 1941, 1956) designated this style as "Romagnan structure".

The shortening between the synclines typically is in the order of a few kilometers. The displacement on the listric reverse faults diminishes upwards, which is indicative of a mechanism of fault-propagation folding (Chapman & Williams, 1984; Suppe & Medwedeff, 1984; Jamison, 1987). Indeed, the dislocations fade along strike into NE-facing asymmetric anticlines. This is most evident in the northwestern part of the Romagnan Apennines, where the lateral variability displayed by the Umbro-Romagnan turbidite suite attests to the synsedimentary origin of these structures (de Jager, 1979; ten Haaf, 1985). The synclinal blocks can therefore be classified as tectono-stratigraphic units (ten Haaf & van Wamel, 1979).

Signorini (1940), Selli (1949, 1952), Centamore *et al.* (1972) and Fazzini (1973) attributed the tectonics of the parautochthon to imbrications of the rigid Calcare Massiccio Formation at depth. However, reflection seismic surveys have demonstrated that it forms a shallow thrust sheet shortened independently of its autochthonous substratum (Bally *et al.*, 1986; Hill & Hayward, 1988). The pertinent detachment zone, situated at the base of the Schlier Formation, is exposed at the thrust front of the parautochthon SE of the investigated area as well at more internal tectonic windows formed by autochthonous structures like the Gubbio Anticline and the Massicci Perugini (de Feyter *et al.*, 1986, 1990). It consists of a relatively thin array of discrete slip surfaces. High pore-fluid pressures must have determined this concentrated shearing, as already suggested for the comparable efficient detachment zones developed at the base of the Marly Scaglia and at the base of the Schlier Formation in the autochthon (cf. V.2.1.3.).

The external portion of the Borgo Pace Unit and the M.Vicino Unit represent the Umbro-Romagnan parautochthon in the investigated area. Their general geometry respects the Romagnan structure (Fig. 69; enclosed cross-sections F-F' through H-H'). Among the structural elements of smaller magnitude are NE-vergent reverse faults attended by asymmetric folds. These mesoscopic features typically affect the long outer limb of the principal synclines. They mark the leading edge of minor bedding-plane detachments, which are generally common in thrust faulted turbiditic terrains (Passerini, 1973). Their formation must be attributed to the expulsion of material from the tightening synclinal cores, a process known as out-of-the-syncline thrusting (Dahlstrom, 1970). Furthermore, transversal and oblique faults occur in the parautochthon. The most prominent one is an E-W trending oblique dextral strike-slip fault dissecting the M.Vicino Unit 2 kilometers NE of Mercatello sul Metauro. This clearly is a primary tear fault which does not extend down into the autochthon.

V.3.2. OUT-OF-SEQUENCE FRONTAL STRUCTURES

The low-angle thrust surface on which the Umbro-Romagnan terrains of the parautochthon overrode their autochthonous counterparts to the NE was first identified in the Romagnan Apennines (Ruggieri, 1970; ten Haaf & van Wamel, 1979). Like the listric reverse faults of the Romagnan structure, it fades towards the NW into a NE-facing asymmetric anticline of synsedimentary origin.

SE of the investigated area, the thrust front of the parautochthon is situated at the inner margin of the Umbro-Marchean Chain (Fig. 4, 68). The brachyanticlinal backlimbs forming its autochthonous substratum are intensely deformed. Of particular significance is the local superposition of younger parautochthonous terrains on older autochthonous ones. This apparent anomaly must have resulted from the thrust front propagating along down-section trajectories with respect to footwall stratigraphy. Down to the level of the Calcareous Scaglia, slivers of autochthonous material thus were sheared off the backlimbs opposing the advancing parautochthon and piled up ahead of it as imbricate stacks (de Feyter *et al.*, 1986, 1990). This mechanism is indicative of out-of-sequence thusting (Morley, 1988). With reference to the structural pattern of the M.Nerone-M.Catria Anticline a few kilometers SE of the investigated area, de Feyter *et al.* (1986) demonstrated that the emplacement of the parautochthon clearly did not precede the formation of the more external Umbro-Marchean Sequence, as prescribed by a piggyback deformation sequence.

In the investigated area, the progress of the Umbro-Romagnan parautochthon was not obstructed by autochthonous brachyanticlines. Frontal structures accordingly do not involve Umbro-Marchean terrains. They appear as an array of listric reverse faults and tight folds, accounting for more than 6 kilometers of shortening, which is dissected by a low-angle thrust surface on which the parautochthon was translated another 3 kilometers towards the NE (Fig. 69; enclosed cross-sections F-F' through J-J'). A similar configuration characterizes this thrust front in the Romagnan Apennines (van Wamel & Zwart, 1990).

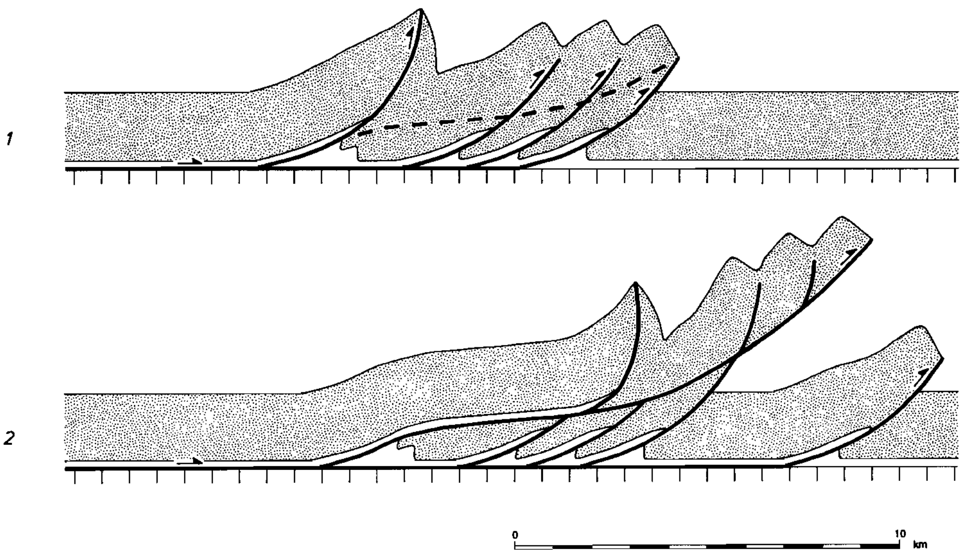


Fig. 81. Schematized evolution of the thrust front of the Umbro-Romagnan parautochthon in the investigated area. The initial weakly emergent thrust front (1) is truncated by an out-of-sequence strongly emergent one attended by distal footwall shortening (2). Footwall subsidence induced by superficial loading is not shown.

The evolution of the frontal deformation pattern of the Umbro-Romagnan parautochthon can be described as a late strongly emergent thrust front overriding an early weakly emergent one (Morley, 1986). Forward displacement at the leading edge of the embryonic parautochthon was absorbed by gentle folding from the Middle Miocene on, as indicated by the lateral variability of the Umbro-Romagnan turbidite suite (cf. II.3.2.1.). During the Late Miocene, superficial shortening accelerated. Fault-propagation folding shaped steep frontal structures, including the M.Vicino Unit. Waning pore-fluid pressures as a result of imbricates reaching the earth surface ultimately caused this weakly emergent thrust front to lock. The final forward translation of the parautochthon therefore necessitated the out-of-sequence formation of the cross-cutting principal thrust surface (cf. Price, 1977; Morley, 1986, 1988). Increased pore-fluid pressures due to loading by the overriding toe of the parautochthon determined the propagation of the detachment at the base of the Schlier Formation beyond the locked frontal structures (Fig. 81). Thence, faulted pseudodiapiric anticlines were accreted onto the Piandimeleto High as a distal leading-edge phenomenon (cf. V.2.1.3.). Furthermore, the superficial loading could have provoked footwall subsidence as the Burano Anhydrites Formation was squeezed from under the Casale-Campo Syncline.

V.3.3. CRUSTAL IMBRICATION AND GRAVITATIONAL SPREADING

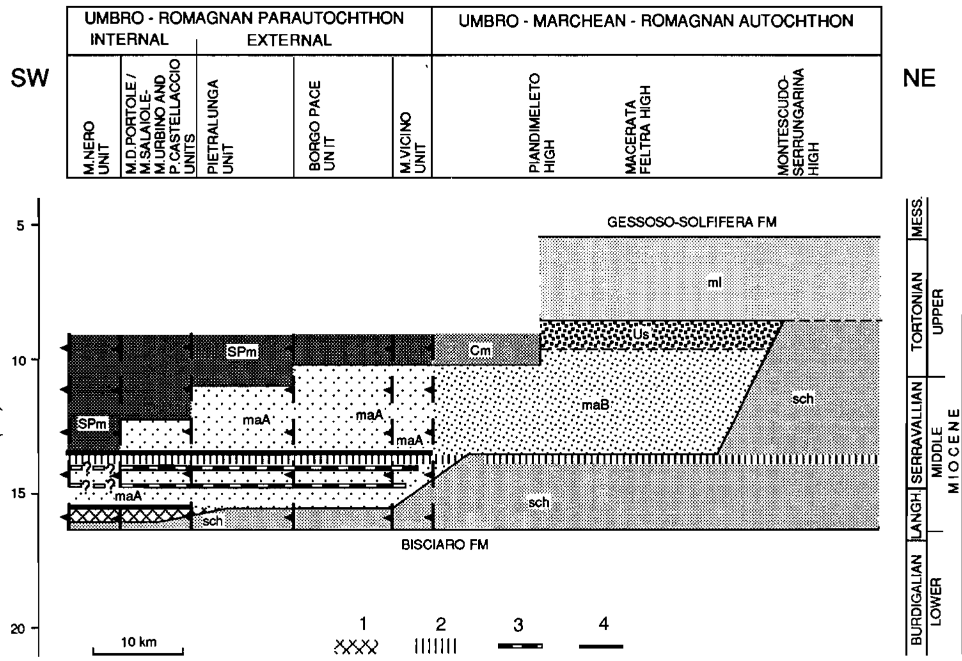


Fig. 82. Tectono-stratigraphic framework of the Umbro-Romagnan turbidite suite. Formation labels; sch: Schlier Formation; maA: Marnoso-arenacea Formation A; maB: Marnoso-arenacea Formation B; Us: Urbania Sandstone Formation; SPm: S.Paolo Marl Formation; Cm: Campo Marl Formation; ml: Marne di Letto Formation. Legend: 1: interval with arkosic and bioclastic turbidites; 2: interval with calcareous turbidites; 3: Contessa megaturbidite; 4: interval with gravity displacement structures (distinguished only for orthoturbiditic stage).

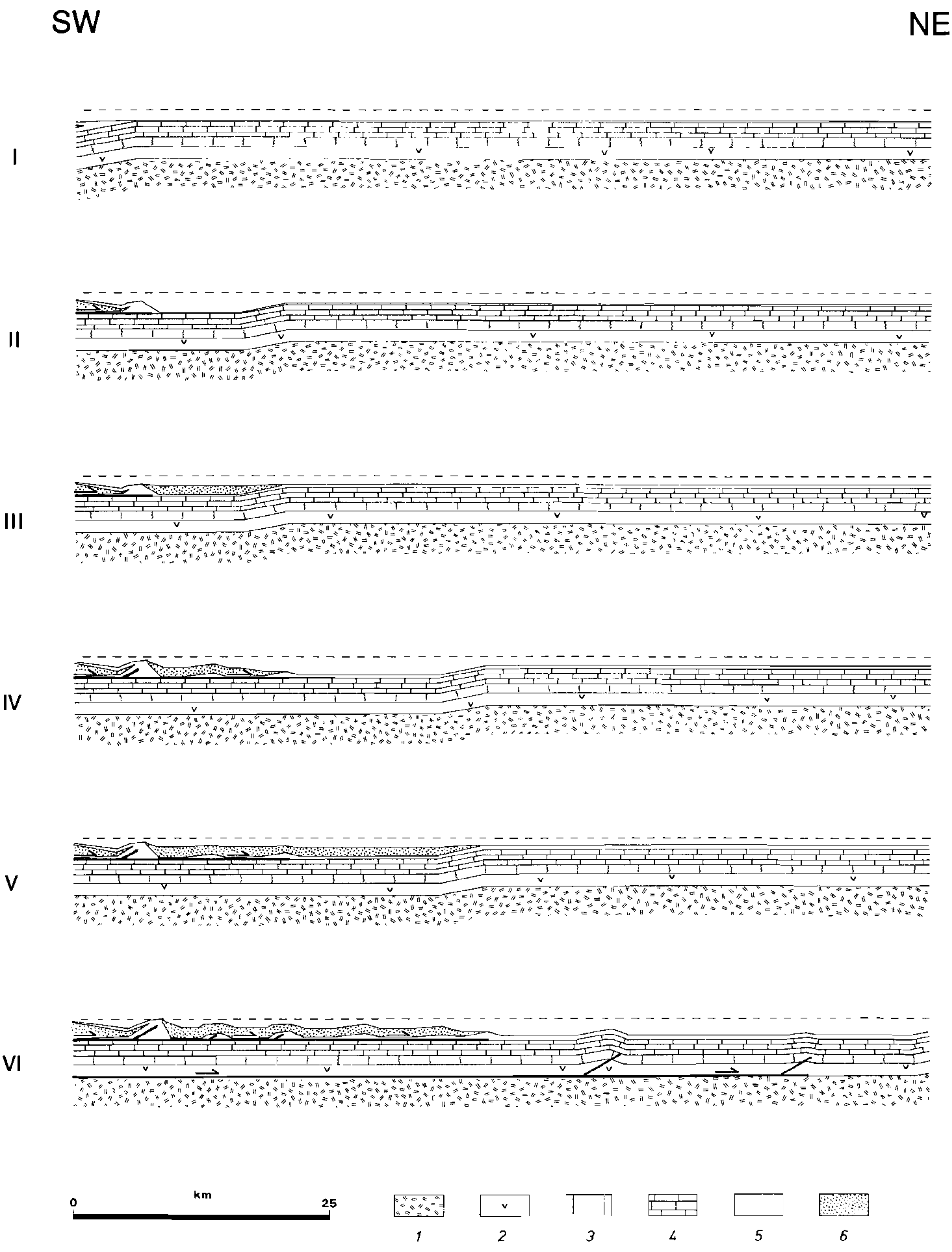


Fig. 83. Schematized evolution of the Umbro-Romagnan Foredeep from the Middle Langhian to the Middle Serravallian. Stage I (Middle Langhian): Unlike the Outer Tuscan Foredeep, the Umbro-Romagnan Foredeep is not yet

active; Stage II (Middle Langhian): Relative subsidence of inner foredeep segment and consequent gravitational spreading more internally; Stage III (Middle to Late Langhian): Orthoturbiditic sedimentation in inner foredeep segment; Stage IV (Late Langhian): Relative subsidence of central foredeep segment and consequent gravitational spreading more internally; Stage V (Late Langhian to Middle Serravallian): Orthoturbiditic sedimentation in inner and central foredeep segments; Stage VI (Middle Serravallian): Relative subsidence of outer foredeep segment and consequent gravitational spreading more internally. Less shallow detachment tectonics emerge as well. Legend: 1: basement; 2: Burano Anhydrites Formation; 3: Calcare Massiccio Formation; 4: Corniola Formation through Bisciaro Formation; 5: proto- and cataturbiditic foredeep deposits; 6: orthoturbiditic foredeep deposits.

The tectono-stratigraphic evolution of the Umbro-Romagnan parautochthon should be considered in a regional context. The spatial and temporal distribution of sedimentary features in the Umbro-Romagnan turbidite suite is of particular palaeotectonic significance (de Feyter *et al.*, 1990). Thus, the step-like younging towards the NE of the base of the Marnoso-arenacea Formation attests to a discontinuously shifting orthoturbiditic sedimentation pattern (Fig. 82; cf. II.3.2.1.). This implies subsidence of successive longitudinal foredeep segments. Therefore, tectono-stratigraphic subdomains constituted by units derived from a specific foredeep segment can be distinguished.

The base of the Marnoso-arenacea Formation A, reflecting the onset of relative subsidence, is of Middle Langhian age in the internal subdomain of the Umbro-Romagnan parautochthon. In the Umbro-Marchean Apennines, this is composed of the M.Nero Unit and the M.delle Portole/M.Salaiole-M.Urbino Unit. In the external subdomain of the parautochthon, comprising the Pietralunga, Borgo Pace, and M.Vicino units, the base of the Marnoso-arenacea Formation A is of Late Langhian age. The Middle Serravallian age of the base of the Marnoso-arenacea Formation B in the contiguous autochthon indicates that this accumulated in a third segment of the Umbro-Romagnan Foredeep.

The sequential activation of longitudinal foredeep segments must be attributed to migratory crustal downwarping. Lavecchia & Pialli (1981b) and Lavecchia *et al.* (1987) regarded this as the isostatic response to thrust loadinng. Superficial loads alone however cannot account for the inferred differential movements, notwithstanding the additional effect of rapid orthoturbiditic sedimentation in subsiding foredeep segments (Dal Piaz, 1942; Gretener, 1972, 1981). In fact, Apenninic foredeeps generally were activated well before important thrusting took place at their inner margin. Moreover, extensive tectonic duplications are uncommon in the autochthonous assemblage of the Northern Apennines. Tangential forces must therefore have determined the evolution of the Umbro-Romagnan Foredeep. Sequential subsidence probably resulted from crustal imbrication encroaching upon the Adriatic Foreland (cf. Sagri, 1973; Kligfield, 1979; Boccaletti *et al.*, 1980; Reutter, 1981; Gasperi *et al.*, 1986). For the Outer Marchean Basin of the more recent Padan-Marchean-Adriatic Foredeep, Royden *et al.* (1987) demonstrated that foredeep geometry was conditioned by such shearing.

The Marnoso-arenacea Formation A displays several intervals with gravity displacement structures such as slumps and olisthostromes (*e.g.*, Pialli, 1966; Jacobacci *et al.*, 1970; Ricci Lucchi, 1981b; Damiani *et al.*, 1982-1983). These generally are of southwestern derivation and concentrated at the major synclinal hinges. The intervals can therefore be ascribed to the nucleation and accentuation of the rootless synclines which eventually yielded the tectono-stratigraphic units of the Umbro-Romagnan parautochthon (de Feyter *et al.*, 1990). Accordingly, it is possible to distinguish a Late Langhian deformation pulse exclusive to the inner foredeep segment, and a Middle Serravallian one which also affected the central foredeep segment (Fig. 82). They reflect gravitational spreading induced by the migratory crustal downwarping (Fig. 83). Superficial shearing remained confined to foredeep segments whose sedimentary fill was thick enough to maintain high pore-fluid pressures in the evolving detachment zone at the base of the Schlier Formation (cf. Gretener, 1981). The leveling of the basement during the Middle Serravallian could also have provoked the expansion of the less shallow detachment tectonics which shaped the anticlinal chains of the autochthon (cf. V.2.2.1.).

Wide-spread gravity displacements in the cataturbiditic terminal stage of foredeep activity heralded the Late Miocene main folding and thrusting of the Umbro-Romagnan parautochthon (cf. II.3.2.1.4.). Taking into account structural style and out-of-sequence emplacement, these paroxysmal tectonics must be attributed to gravitational spreading as well (Elliott, 1976; ten Haaf & van Wamel, 1979; Cooper, 1981; Jones, 1987). Unlike the previous mild deformation pulses, this was probably incuced by substantial loading in the internal part of the Apenninic orogen (Giannini & Tongiorgi, 1962; Giannini *et al.*, 1962; Baldacci *et al.*, 1967; Reutter & Groscurth, 1978; Boccaletti *et al.*, 1980, 1983).

CHAPTER VI

COLATA TECTONICS

VI.1. FLOATING SLABS AND IMBRICATE SLICES

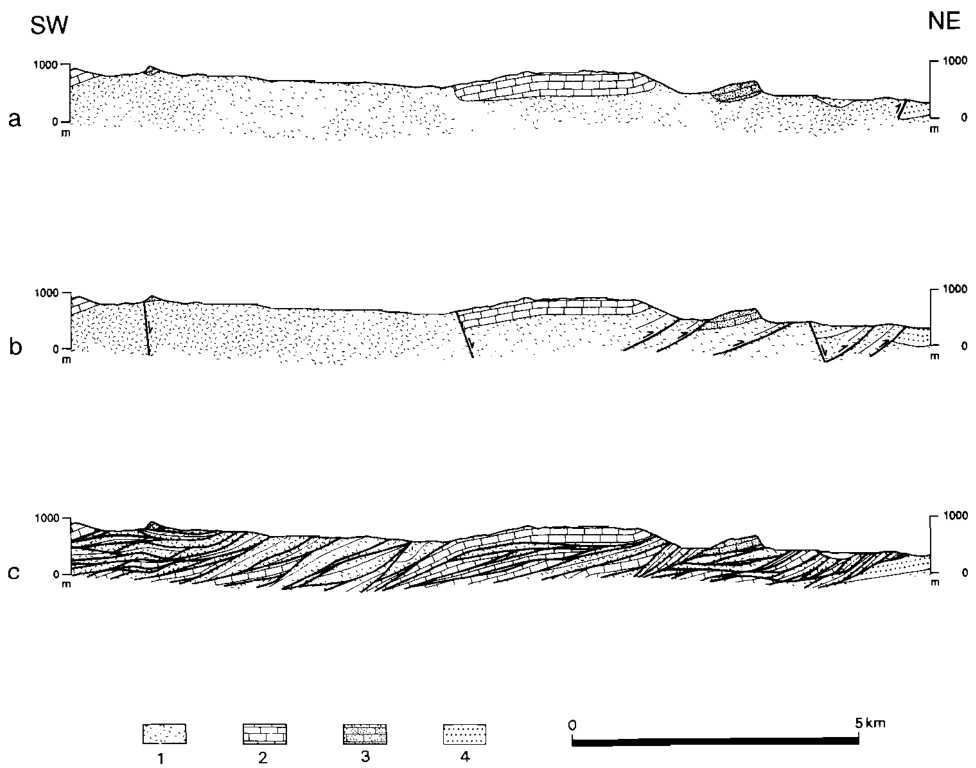


Fig. 84. Synthetic cross-sections illustrating the different structural styles envisaged for the Montefeltro Colata; a: clayey complex with floating slabs and satellite basins (inspired by Ruggieri, 1970); b: imbricate thrusts and associated normal faults dissecting clayey complex with floating slabs and satellite basins (modified from Conti, 1989); c: imbricate slices with floating slabs and satellite basins (based on enclosed cross-sections C-C' through E-E'). Legend: 1: incompetent teleallochthon; 2: competent teleallochthon; 3: semiallochthon; 4: peneallochthon.

The classic picture of the allochthonous assemblage in the external part of the Northern Apennines is that of a melange with more or less voluminous exotics (Merla, 1951, 1959, 1964; Maxwell, 1959b; R.Selli *in* Selli, 1967; Hsü, 1968; Price, 1975; Page, 1978; ten Haaf, 1985). The pervasively sheared character attributed to its weak matrix of slickensided clays, known as *"argille scagliose"*, has been considered indicative of a plastic flow, or *"colata"*, mode of emplacement (Dal Piaz, 1942; Bucher, 1956; Signorini, 1956; Kehle, 1970; de Jager, 1979). Accordingly, Bonarelli (1929), Ruggieri (1958, 1970), Amadesi (1962), Stern (1969) and Veneri (1986) depicted the Montefeltro Colata as a chaotic clayey complex carrying slabs of relatively undisturbed terrains (Fig. 84a).

A more systematic structure was already implied for the allochthon NW of the Sillaro Line by Pareto (1861). In fact, Castellarin *et al.* (1986), Castellarin & Pini (1987) and Pini (1987) revealed the presence of regular shear zones bounding lenticular bodies. Conti *et al.* (1987) and Conti (1989) applied this concept to the Montefeltro Colata, individuating a number of imbricate thrusts partly accompanied by antithetic normal faults. However, they did not abandon the principle of rigid slabs riding on contorted clayey terrains (Fig. 84b).

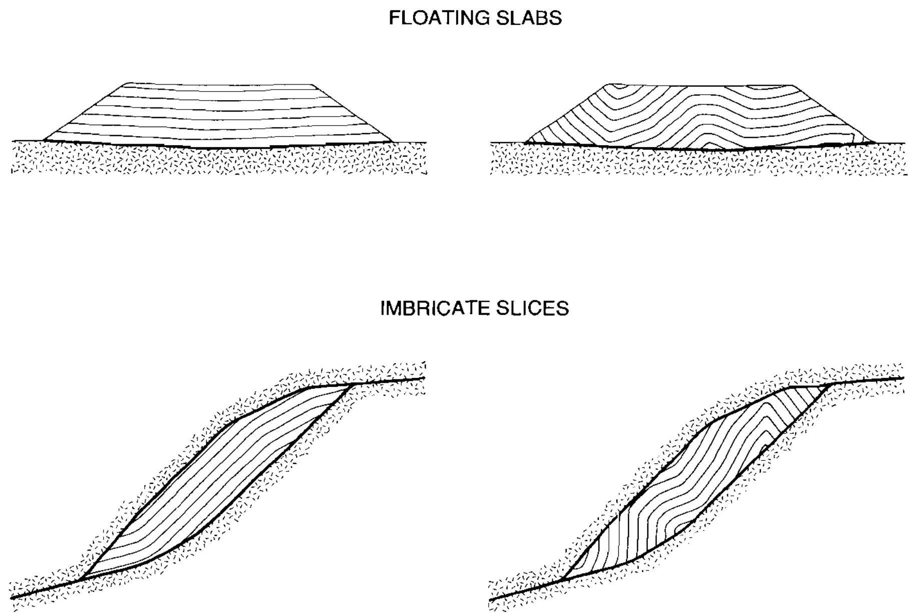


Fig. 85. Allochthonous structural elements classified according to position and internal geometry.

The visible chaoticity of pelitic lithotypes in the studied portion of the Montefeltro Colata actually appears to be a largely superficial feature resulting from recent landsliding rather than from distributed shearing (cf.

Manfredini, 1963). Thus, even the incompetent Villa a Radda and Sillano formations constitute individual teleallochthonous bodies. The Montefeltro Colata accordingly forms an array of juxtaposed and superposed lenticular bodies (Fig. 84c). Apart from the satellite basins of the peneallochthon, its basic elements can be classified with respect to structural position as floating slabs and imbricate slices. Their internal geometry represents an additional discriminatory feature (Fig. 85). Bedding usually conforms to the shape of the elongate bodies, reflecting their origin as detached monoclinal panels. Their principally monoformational character supports this concept (Stern, 1969).

The bulk of the Montefeltro Colata is made up of monoclinal imbricate slices (enclosed cross-sections B-B' through E-E'). Conformable floating slabs are less common elements (Fig. 86). Especially the larger ones display a more or less pronounced saucer shape (Fig. 87). This configuration in general must be ascribed to compaction and squeezing sideways of underlying weak material (Passerini, 1958).

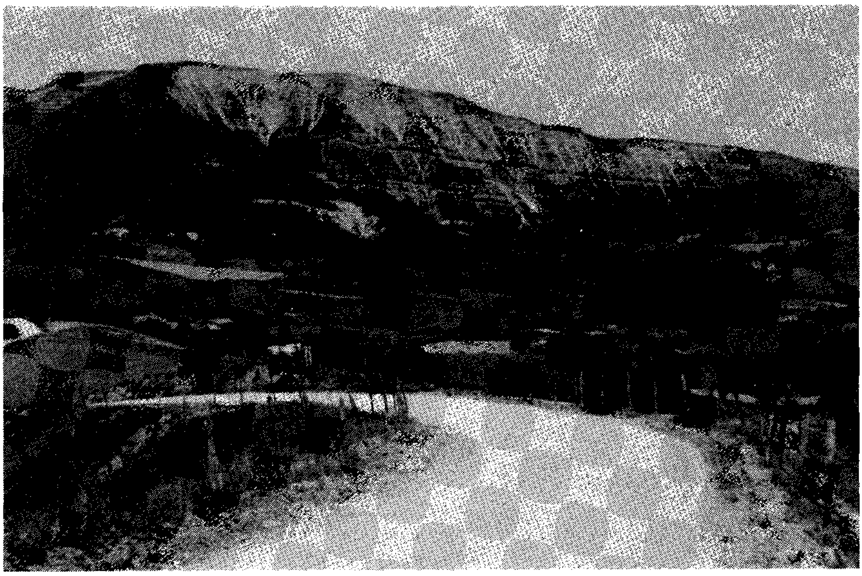


Fig. 86. Regularly bedded M.Morello Alberese Formation of the conformable M.Carpegna Slab in the Costa dei Salti, 3 km NE of Carpegna.

Some tabular bodies consist of overturned teleallochthonous terrains (Fig. 88). Lazzarotto & Mazzanti (1964) interpreted their equivalents in more internal parts of the Northern Apennines as remnants of disrupted recumbent folds. Primarily with reference to the enigmatic Monghidoro SLab of the allochthon NW of the Sillaro Line, such structures have been attributed to gravitational collapse or dragging by mobile plastic terrains (Migliorini, 1948; Valduga, 1950; Signorini, 1956; Maxwell, 1959a; Hsü, 1967; ten Haaf, 1985).

This probably resulted from the instability of the embryonic allochthon in the wake of the Eocene Ligurian Phase of Apenninic orogeny.

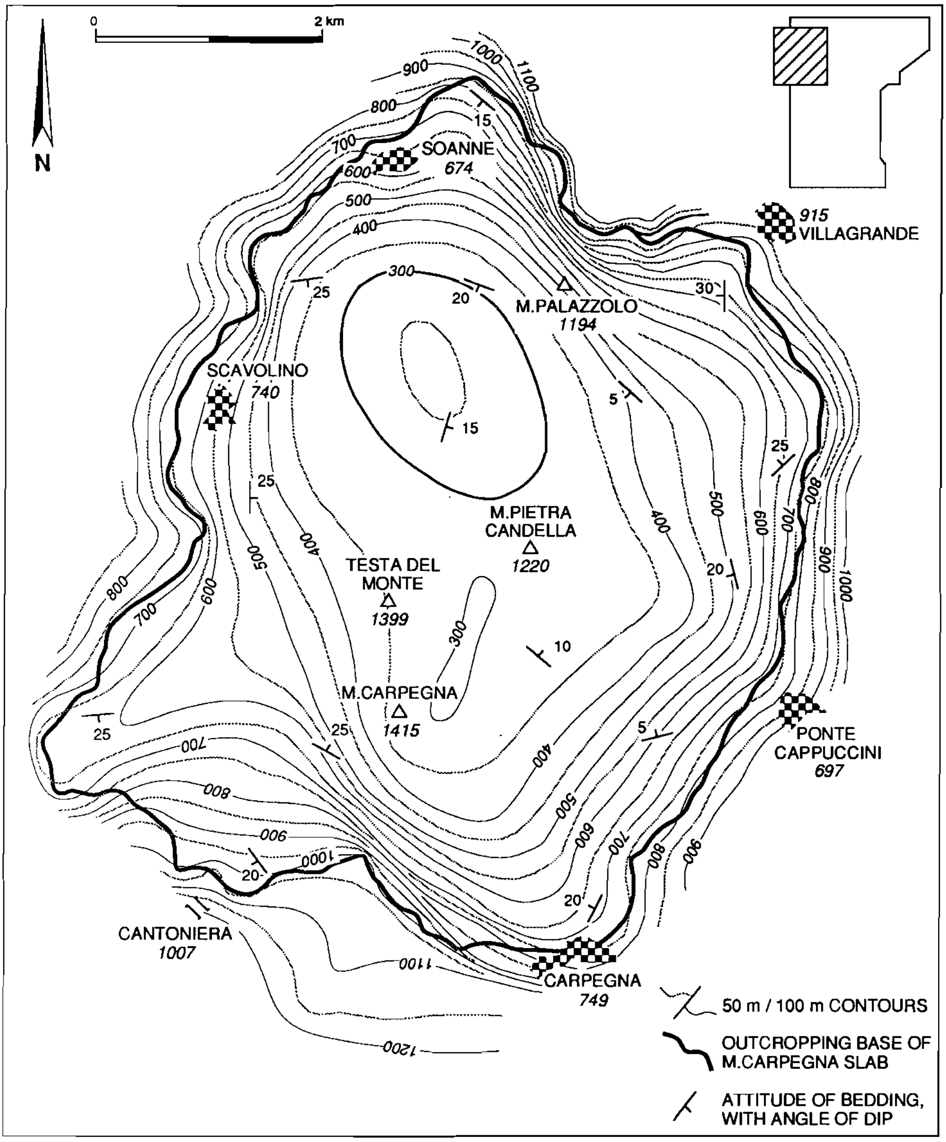


Fig. 87. Structure-contour map of the base of the M.Carpegna SLab.

Markedly deformed teleallochthonous terrains compose several slabs and slices (Fig. 89, 90). Structural trends are generally parallel to the long axes of these bodies. They primarily date from the Ligurian Phase of Apenninic orogeny (cf. Coli & Fazzuoli, 1983). To some extent, internal shearing however may have affected coherent bodies within the mobile allochthon. This applies especially to minor asymmetric folds and reverse faults preferentially developed in their marginal portions as drag structures. The gentle folds displayed by the peneallochthon between Mercatino Conca and Auditore demonstrate that additional shortening accompanied the emplacement of the Montefeltro Colata (Fig. 63).

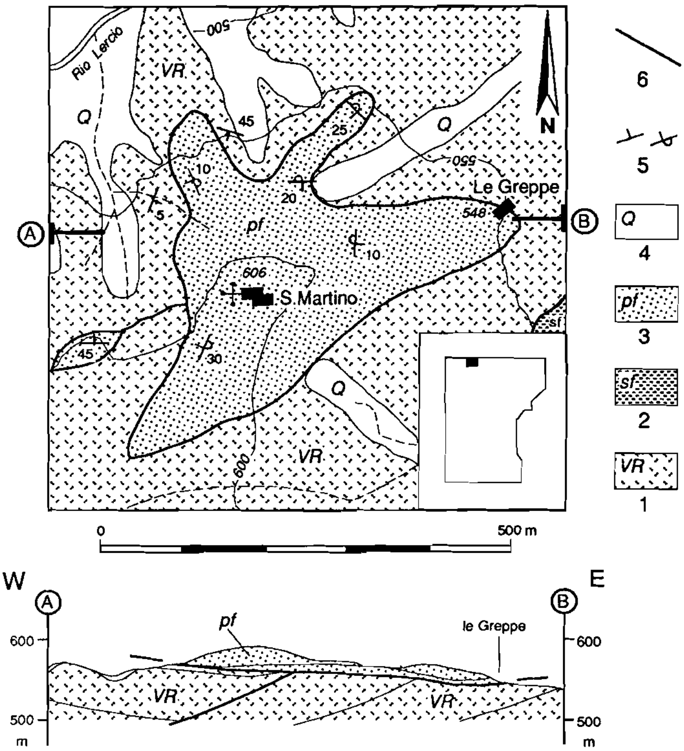


Fig. 88. Detailed geological map and cross-section of the S.Martino Slab, 4 km NNE of Villagrande. Legend: 1: Villa a Radda Formation; 2: Sillano Formation; 3: Pietraforte Formation; 4: Quaternary; 5: bedding symbols, with indicated angle of dip; 6: tectonic contact.

Some piles of imbricates have been folded together with their autochthonous substratum. Thus, the structural pattern of the Montefeltro Colata at Val di Meola, 4 kilometers SSW of Carpegna, clearly relates to the Casale-Campo Syncline (enclosed map and cross-sections E-E' and F-F'). Folding of imbricates due to piggyback shearing within the allochthon on the other hand is most evident in the surroundings of Petorno, 2 kilometers SE of Villagrande (enclosed map and cross-section C-C').

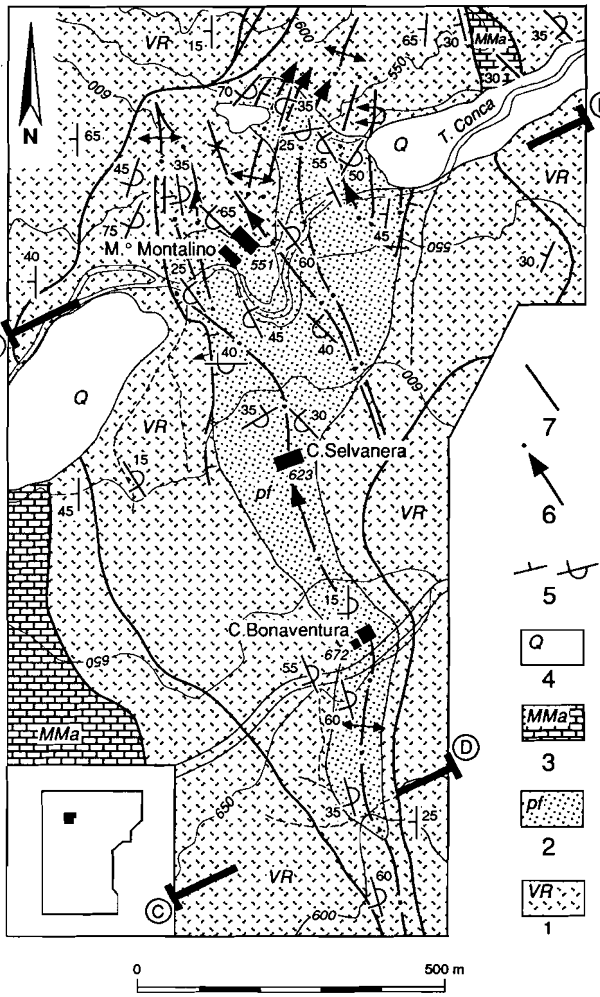


Fig. 89. Detailed geological map of the Mo.Montalino Slice, 2 km SW of Montecerignone. Traces of cross-sections of Fig. 90 are indicated. Legend: 1: Villa a Radda Formation; 2: Pietraforte Formation; 3: M. Morello Alberese Formation; 4: Quaternary; 5: bedding symbols, with indicated angle of dip; 6: axial trace of plunging fold; 7: tectonic contact.

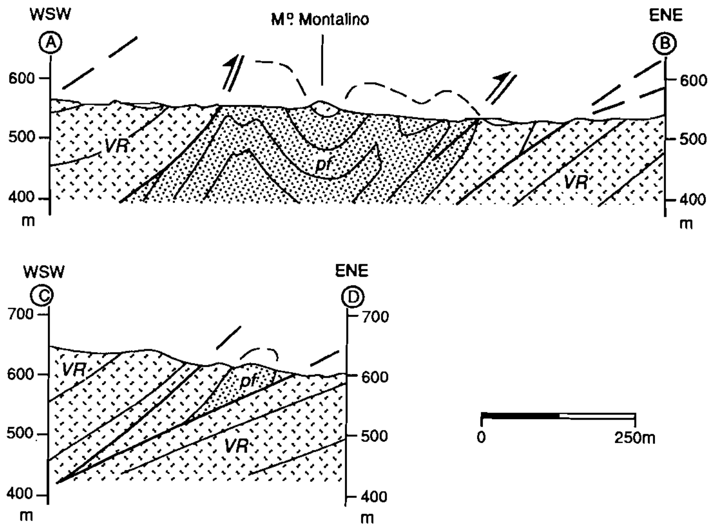


Fig. 90. Cross-sections to Fig. 90.

VI.2. EMPLACEMENT MECHANISM

Its classic description as orogenic landslides demonstrates that the translation of allochthonous material towards the Adriatic Foreland since long has been regarded as a gravitational process (Migliorini, 1933, 1945, 1948; Signorini, 1940; Merla, 1951, 1959, 1964; Beneo, 1949, 1956). Most of the proposed emplacement models thus assume an essentially submarine passive allochthon, structurally independent of its autochthonous substratum.

R.Facca (*in* Beneo, 1956) and Rigo de Righi (1956) designated voluminous allochthonous masses in general as olisthostromes. However, olisthostromes originally were defined as sedimentary intercalations characterized by more or less intimately admixed heterogenous material which accumulated in a semifluidal state (G.Flores *in* Beneo, 1955). This obviously is not the same as allochthonous masses whose dimensions are compatible with a tectonic rather than a sedimentary mechanism (Jacobacci, 1965; Hsü, 1968; Abbate *et al.*, 1970a; Richter, 1975; de Jager, 1979; Abbate & Sagri, 1981). In the case of the Montefeltro Colata, structural style provides another argument against the olisthostromic origin implied by Ruggieri (1958), Frey (1969) and Stern (1969). Emplacement through numerous small olisthostromes, a model formulated by Reutter (1965), Goerler & Reutter (1968) and Goerler (1975), must be rejected for the same reason (Fig. 91A).

Downslope gravitational gliding, schematized in Fig. 91B, has been advocated by numerous authors (*e.g.*, de Wijkerslooth, 1934; Dal Piaz, 1942; Veneri, 1986). On the other hand, de Jager (1979), ten Haaf & van Wamel (1979) and van den Berg (1987, 1990) adhered to the concept of gravitational

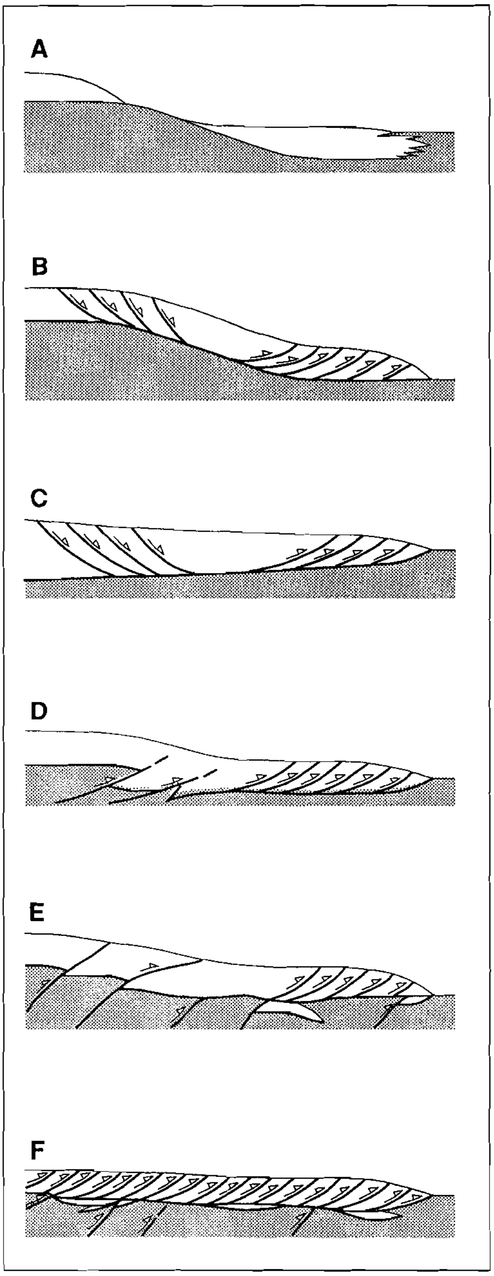


Fig. 91. Proposed models for the emplacement of far-travelled allochthonous masses. Sedimentary: A: stacking of olisthostromes. Tectonic: B: gravitational gliding (after de Jager, 1979); C: gravitational spreading (after Elliott, 1976); D: active shearing plus gravitational gliding (after Ricci Lucchi & Ori, 1985); E: active shearing (after Castellarin & Pini, 1987); F: essentially gravitational gliding, as inferred for the Montefeltro Colata.

spreading elaborated by Elliott (1976). It presumes that the surface slope determined the motion of the allochthon irrespective of the attitude of its base (Fig. 91C). In either case, migratory vertical differentiations must be invoked in view of the magnitude of the superficial translations, amounting to several hundreds of kilometers from the inner margin of the Tuscan Zone to the most distal allochthonous terrains of the Montefeltro Colata and NW of the Sillaro Line (*e.g.*, de Wijkerslooth, 1934; Migliorini, 1948; Merla, 1951, 1959, 1964).

Tectonic activities in substrata adding to the essentially passive progress of the allochthonous assemblage were already envisaged by Merla (1951) and Bortolotti (1964). Likewise, Ricci Lucchi & Ori (1985) and Ricci Lucchi (1986a) depicted gravitational gliding as a function of active shearing (Fig. 91D). Embroidering on the findings of Castellarin *et al.* (1985, 1986), Castellarin & Pini (1987) and Pini (1987) in the allochthon NW of the Sillaro Line, Conti *et al.* (1987) and Conti (1989) even proposed an active emplacement mechanism with merely secondary gravitational effects for the Montefeltro Colata (Fig. 91E). They considered thrust faults dissecting the allochthon to emanate from its autochthonous substratum. However, no major shear zones extend from the autochthon into the allochthon in the investigated area. Some are cut off by the Montefeltro Colata, as in the case of the leading edge of the Macerata Fèltria High near S.Maria Valcava, 2 kilometers SE of Montecerignone (enclosed map and cross-section D-D'). Other structures fade in its periphery. The same phenomenon characterizes the Sillaro Line at the north-

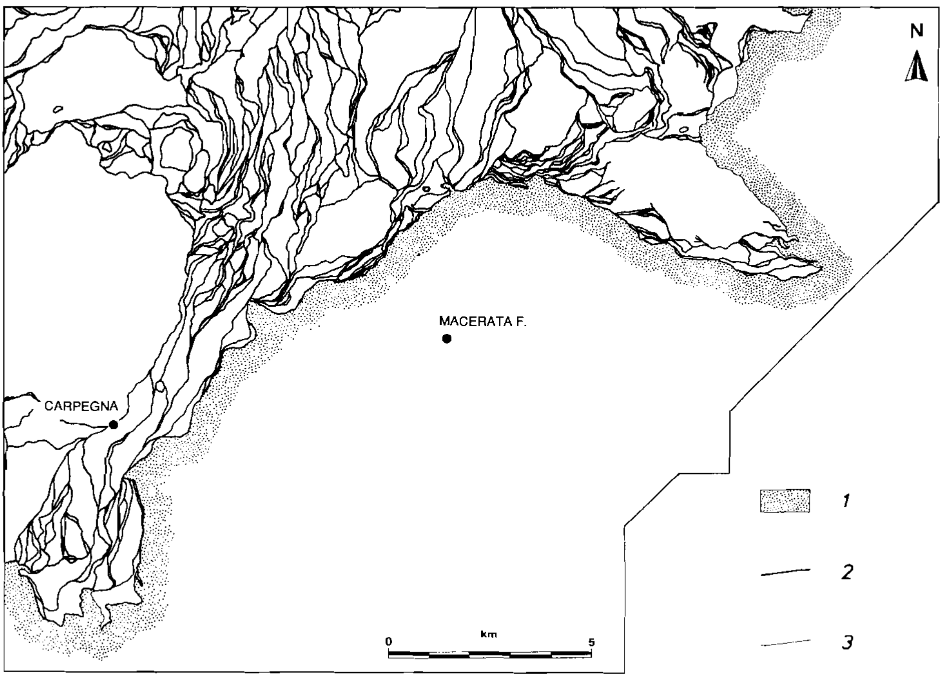


Fig. 92. Structural pattern of the Montefeltro Colata in the investigated area. Legend: 1: autochthonous assemblage; 2: tectonic contact; 3: sedimentary contact.

western margin of the Romagnan Apennines (de Jager, 1979). Anyway, the presence of imbricate thrusts alone does not argue an active emplacement mechanism. Internal shortening, primarily through imbrication, actually is a common toe effect of gravitational translations (Dal Piaz, 1942; Signorini, 1956; Price, 1977; de Jager, 1979). The arcuate structural pattern of the Montefeltro Colata is compatible with a passive origin (Fig. 92). In fact, it resembles the outward convexity of pressure ridges developed by the frontal portion of sedimentary mass flows (J.D.Collinson *in* Reading, 1978). The systematic arrangement of imbricates accordingly attests to the direction of superficial transport, similar to the attitude of olistholiths in olisthostromes (Goerler & Reutter, 1968; Bertini *et al.*, 1975; Goerler, 1975; Castellucci & Cornaggia, 1980). On a scale sufficiently large, colata tectonics thus may be compared with plastic flow.

The imbricated structure of the Montefeltro Colata must have developed from its front hindwards (Fig. 91F). Such an evolution is typical of gravitational gliding (cf. Price, 1977). Earlier arrays of imbricates apparently carried the more or less intact satellite basins of the peneallochthon. The same applies to not significantly disrupted semiallochthonous basins in the northern part of the Montefeltro Colata (cf. Abbate & Sagri, 1981; Conti, 1989). Other large floating bodies, such as the M.Carpegna Slab, instead seem to have glided individually (Stern, 1969).

Initial extensional shearing is inherent to the concept of gravitational gliding. Mannori & Sani (1987) demonstrated this for the northern part of the Montefeltro Colata. Some of the normal faults displayed by teleallochthonous bodies in the investigated area may have originated accordingly.

VI.3. PHASED PROGRESS

The allochthon of the Northern Apennines overlies progressively younger terrains towards the Adriatic Foreland, a situation known as "tectonic transgresion" (*e.g.*, Merla, 1951, 1959, 1964; de Jager, 1979; ten Haaf, 1985). Related sedimentary features of course display a similar trend. The most obvious ones are precursory olisthostromes supplied to the Apenninic foredeeps by the advancing glide masses (Elter & Trevisan, 1973; Bertini *et al.*, 1975; Richter, 1975; de Jager, 1979; Castellucci & Cornaggia, 1980; Abbate & Sagri, 1981). Satellite basins on the other hand mainly were active on temporarily stationary allochthonous terrains (Stern, 1969; Ricci Lucchi, 1981b, 1986a; Ricci Lucchi & Ori, 1985; Mannori & Sani, 1987). Thus, the S.Marino-M.Fumaiolo Sequence of the semiallochthon reflects a prolonged Middle Miocene standstill in the Tuscan Zone (cf. III.2.). According to Merla & Abbate (1967), Stern (1969) and Dallan Nardi & Nardi (1975), the front of the allochthon then was situated near the inner margin of the Umbro-Romagnan Foredeep. Thus, it could provide intercalations of allochthonous material that characterize the Middle Miocene portion of the Marnoso-arenacea Formation A. These appear in the form of detrital components, as in the case of Contessa megaturbidites, as well as olisthostromes (cf. II.3.2.1.3., V.3.3.).

Subsequently, the allochthon invaded the Umbro-Romagnan Foredeep. The spatial and temporal distribution of allochthonous material in the investigated area exemplifies the discontinuous aspect of this essentially submarine process (Fig. 93). Clearly, it was conditioned by the growing longitudinal structures of the autochthonous assemblage. As a rule, the Montefeltro Colata occupies a specific stratigraphic position in synclinal sectors, slightly interfingering with its substratum and accompanied by precursory allolisthostromes. This is indicative of rapid synsedimentary progress. At major highs instead, the base of the allochthon typically climbs in the footwall stratigraphy, implying temporary standstills. Frontal instability could however generate allolisthostromes.

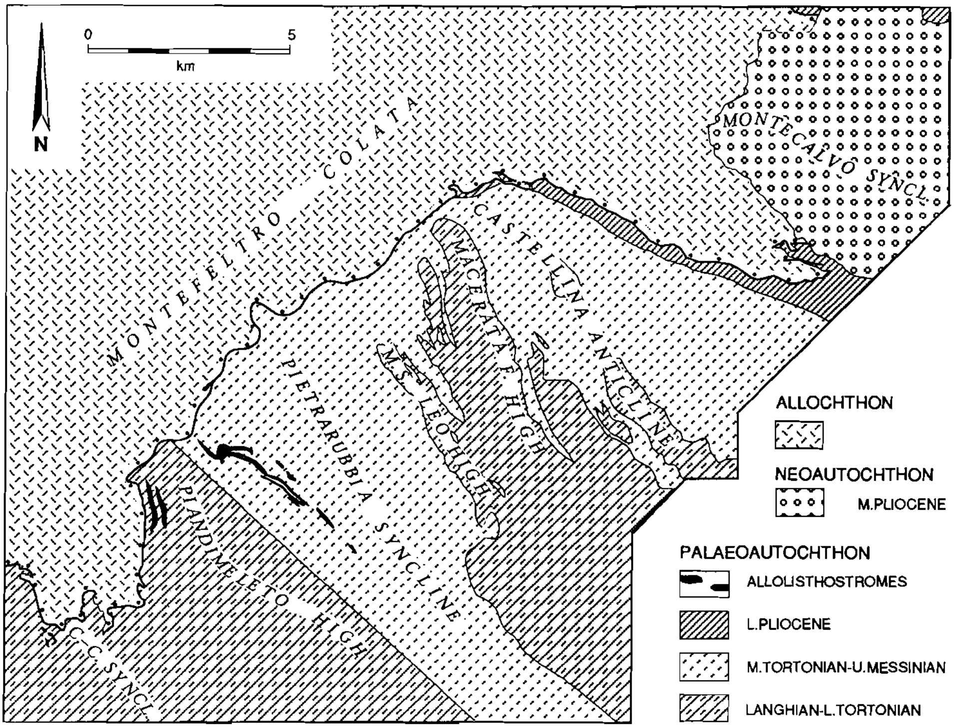


Fig. 93. Schematic map to elucidate the phased emplacement of the Montefeltro Colata and associated allolisthostromes in the investigated area as a function of footwall morphology. Subdivisions within the autochthon are based on age relative to the principal translations of the allochthon.

These criteria make it possible to accurately reconstruct the progress of the Montefeltro Colata. Its arrival in the investigated area at the end of the Early Tortonian was heralded by the deposition of the allolisthostromes intercalated in the Campo Marl Formation in the Casale-Campo Syncline (cf. II.3.2.1.4.). The allochthon eventually reached the Piandimeleto High during this phase. Detached frontal slices emplaced individually produced apparent hindward interfingering with the autochthon in the Casale-Campo Syncline (enclosed map and cross-sections D-D' and E-E').

The Montefeltro Colata certainly did not surmount the Piandimeleto High prior to the end of the Messinian. It did however interfere with the sedimentary evolution of the more external autochthon. The voluminous allolisthostromes displayed by the Marne di Letto Formation in the Pietrarubbia Syncline attest to a marked frontal instability of the allochthon during the Late Tortonian (cf. II.3.2.1.7.). Similar intercalations occur in an identical position NW of the Montefeltro Colata (Ruggieri, 1958, 1970; Ricci Lucchi & D'Onofrio, 1966). The frontal shedding of particularly abundant debris and minor olisthostromes by the allochthon during the Middle to Late Messinian is reflected in the constitution of the Colombacci Formation in the Pietrarubbia Syncline (cf. II.3.2.3.2.).

The remobilization of the Montefeltro Colata is commonly dated to the later part of the Early Pliocene (*e.g.*, Ruggieri, 1953a, 1953b, 1956a, 1958, 1970; De Francesco & Veggiani, 1967; Veneri, 1986; Mannori & Sani, 1987). Yet, it locally interfingers with the top of the Messinian Colombacci Formation in the Pietrarubbia Syncline. At the same stratigraphic level, allolisthostromes occur in the Montecalvo in Foglia Syncline (Fig. 93). Therefore, an advance of the allochthon from the Piandimeleto High to the Macerata Fêltria High at the end of the Messinian must be assumed. This implies subaerial gliding, in view of the continental character of the Colombacci Formation in the Pietrarubbia Syncline (cf. II.3.2.3.2.). Clearly, the paradigm of submarine emplacement specified for the Montefeltro Colata by Ricci Lucchi *et al.* (1982) is untenable.

After a brief standstill at the beginning of the Early Pliocene, the front of the Montefeltro Colata shifted to the Castellina Anticline. It then entered the Montecalvo in Foglia Syncline, preceded by allolisthostromes. The satellite basins of the peneallochthon must have originated at the same time. Their allochthony consequently is in the order of a few kilometers (cf. Bartolini *et al.*, 1982; Capuano *et al.*, 1986a; Conti, 1989).

The most external coherent allochthon emplaced during the late Early Pliocene has been encountered in borings W of Rimini (AGIP Mineraria, 1959; Ruggieri, 1958, 1970; L.Lucchetti *in* Lucchetti *et al.*, 1962; De Francesco & Veggiani, 1967; Lipparini, 1969). More recent activities of the Montefeltro Colata were envisaged by Ricci Lucchi (1975a) and Elmi *et al.* (1981). Indeed, the neoautochthon in front of it is somewhat disturbed. This should however be attributed to secondary frontal instability rather than to renewed translations of the entire allochthon (Colalongo *et al.*, 1982).

In view of its phased emplacement, the average advance rates of several meters to several tens of meters per 1000 years estimated for the Montefeltro Colata by Goerler & Reutter (1968), Stern (1969), Conti *et al.* (1987) and Conti (1989) have little significance. The three main pulses at the end of the Early Tortonian, at the end of the Messinian, and during the late Early Pliocene can be considered geologically instantaneous. This implies true gliding velocities exceeding the average ones by an order of a magnitude or more.

VI.4. STRUCTURAL EVOLUTION

The structural framework of the Montefeltro Colata reflects the kinematics of its phased emplacement. Irregularities in the arcuate pattern of imbricates attest to differential movements within the allochthon. It is possible to individuate marginal elements which apparently did not participate fully in the forward translations (Fig. 94). Differential gliding to some extent however also took place within the main body of the mobile allochthon. For instance, the voluminous M.Carpegna Slab seems to have slightly pushed up the imbricates in front of it (enclosed map and cross-section B-B').

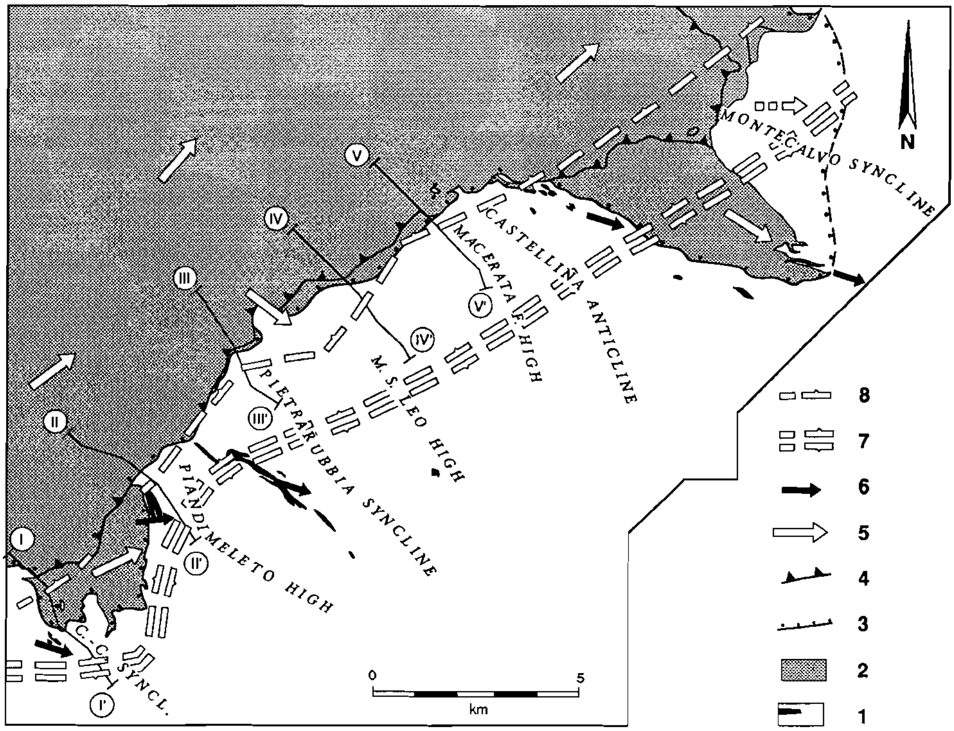


Fig. 94. Synopsis of structural elements relating to the emplacement of the Montefeltro Colata in the investigated area. Cross-sections I-I' to V-V' are shown in Fig. 97. Legend: 1: autochthonous assemblage; 2: allochthonous assemblage and allolisthostromic intercalations in autochthon; 3: transgressive thrust fault, with dots on upper block; 4: principal tectonic contact within allochthon; 5: inferred movement direction of allochthon; 6: inferred movement direction of allolisthostromes; 7: crest of marginal bulge; 8: upper hinge of steep segment of marginal bulge.

The Montefeltro Colata displays defunct marginal elements in correspondence with major footwall synclines (Fig. 94). Basically, two genetic types can be distinguished (Fig. 95). Components of the coherent allochthon could become trapped in synclines as the basal gliding surface shifted to a higher position. Structures of this kind are represented between the Piandimeleto High and the Macerata Fèltria High. Alternatively, the allochthon could advance over detached frontal slices emplaced ahead of it, as inferred for the Casale-Campo Syncline (cf. VI.3.). This mechanism is comparable with the overriding of precursory allolisthostromes by allochthonous masses depicted by Bortolotti (1964).

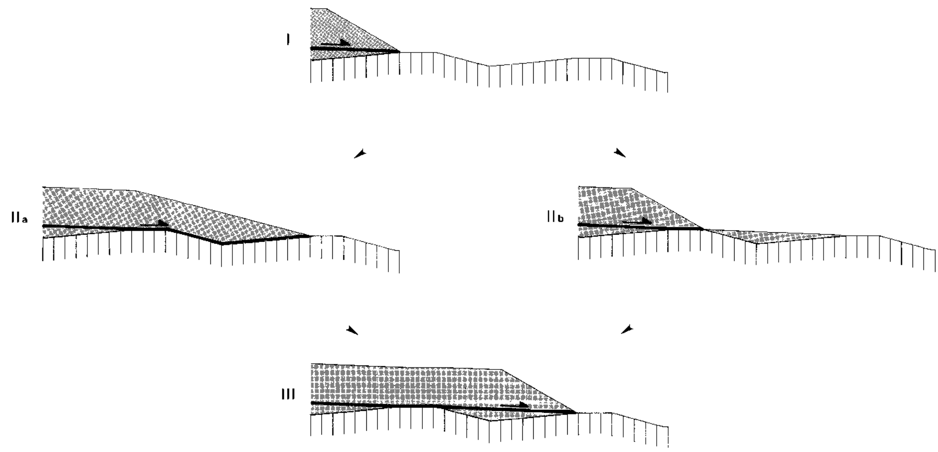


Fig. 95. Basic mechanisms for the formation of defunct elements by a mobile allochthon to overcome footwall irregularities. A coherent allochthon may be differentiated as its active base shifts upwards (I-IIa-III), or it may advance over precursory material (I-IIb-III).

The development of defunct elements by the advancing allochthon clearly prevented major footwall disruptions. Yet, not everywhere the base of the Montefeltro Colata conforms exactly to footwall stratigraphy (Fig. 96; Signorini, 1940, 1941; Ruggieri, 1953a). The only indication of footwall plucking encountered in the investigated area is represented by a sliver of gypsiferous material at C.Raggio, 2.5 kilometers NE of Pietrarubbia. This was apparently derived from the Gessoso-solfifera Formation and dragged along at the base of the allochthon for about 4 kilometers (cf. Ruggieri, 1958). Elsewhere in the Umbro-Marchean-Romagnan Zone, the allochthonous assemblage did not extensively deform its substratum either (Masini, 1951; Centamore *et al.*, 1972). This suggests that it generally comprises defunct elements. These must however be absent at the northeastern segment of the Sillaro Line, in view of the intense footwall shearing described by Castellarin & Pini (1987).

Ruggieri (1953a, 1958) already diagnosed secondary sideward gliding at the northwestern margin of the Montefeltro Colata. Similar aberrant marginal movements took place in the investigated area (Fig. 94). They resulted in part from the expulsion of defunct elements by the advancing allochthon. More consistent sideward transport must be invoked for the major outlier of the Montefeltro Colata in the Montecalvo in Foglia Syncline. Amadesi (1962) designated this as a lateral flow. Indeed, its structural framework is indicative of frontal elements forced aside to glide freely on a southeasterly inclined substratum. This is corroborated by the dispersal pattern of allolisthostromes. Allochthonous material subjected to such translations had surmounted a more or less pronounced marginal bulge which guided the principal forward progress of the Montefeltro Colata (Fig. 94, 97). The main body of the allochthon actually appears confined by a marked break of the northwestern slope of the asymmetric bulge. This suggests that its original outlines have not been substantially obliterated by erosion. Clearly, the Montefeltro Colata is not a remnant of a much more extensive cover, as inferred by Dal Piaz (1943).

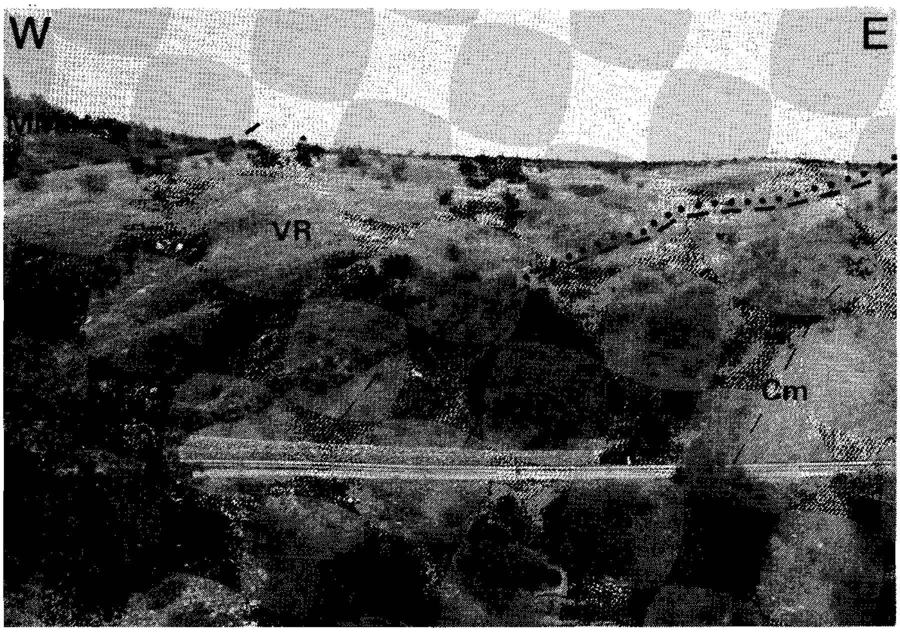


Fig. 96. The tectonically transgressive base of the Montefeltro Colata (dotted) near S.Sisto, 3.5 km S of Carpegna. It does not exactly conform to the indicated footwall bedding of the Campo Marl Formation (Cm). The structural geometry is essentially intact, despite recent landsliding of the Villa a Radda Formation (VR). This is overlain by a slice of the more resistant M.Morello Alberese Formation (MMa).

The marked depression containing the Montefeltro Colata involves the basement, as indicated by its coincidence with a gravimetric minimum (Fig.

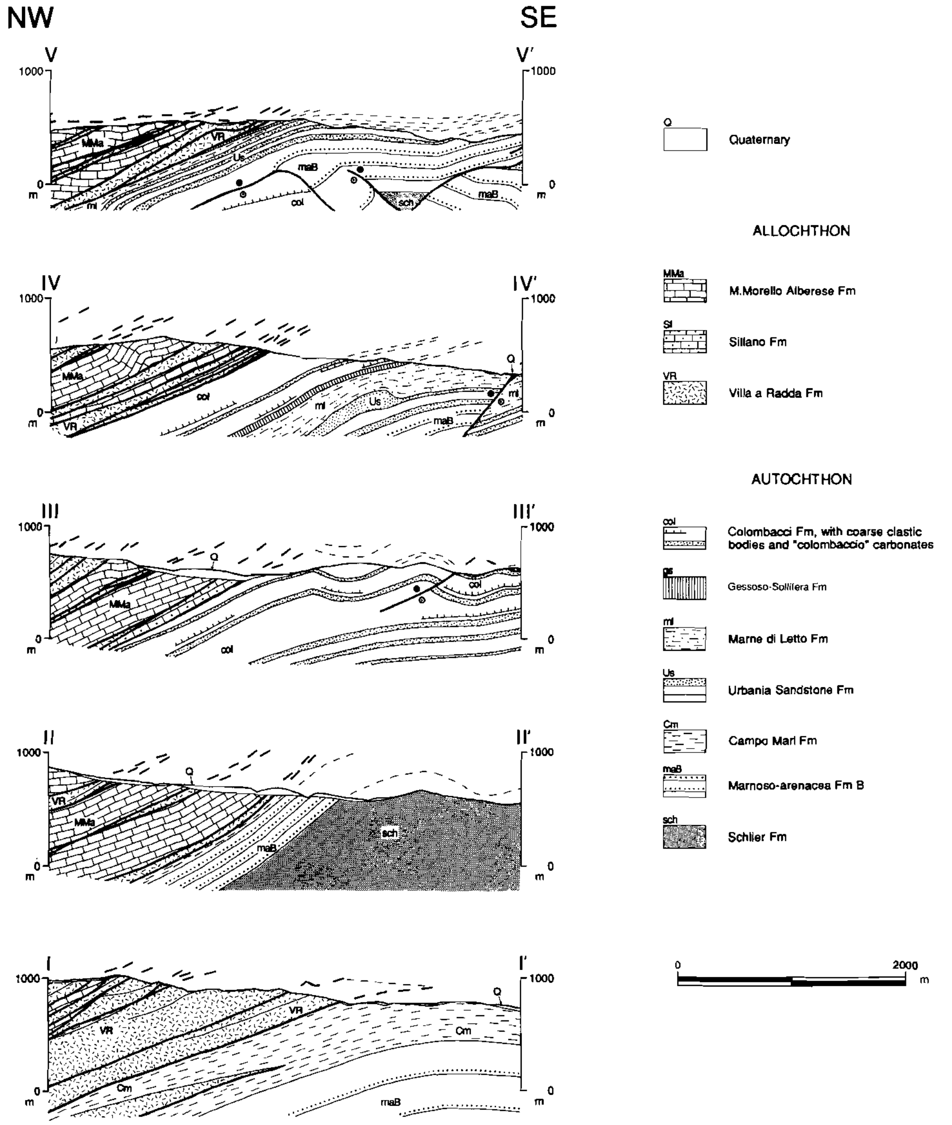


Fig. 97. Cross-sections to Fig. 94, showing the marginal bulge which borders the Montefeltro Colata.

78). It has been regarded as part of a major lineament traversing peninsular Italy (Bettelli *et al.*, 1980b; Conti, 1989). Activities ascribed to this feature range from sinistral wrenching (Boccaletti *et al.*, 1977; Lavecchia & Pialli, 1981a) and dextral wrenching (Fazzini & Gelmini, 1982) to normal faulting (Lavecchia & Pialli, 1980). Yet, the main longitudinal structures of the

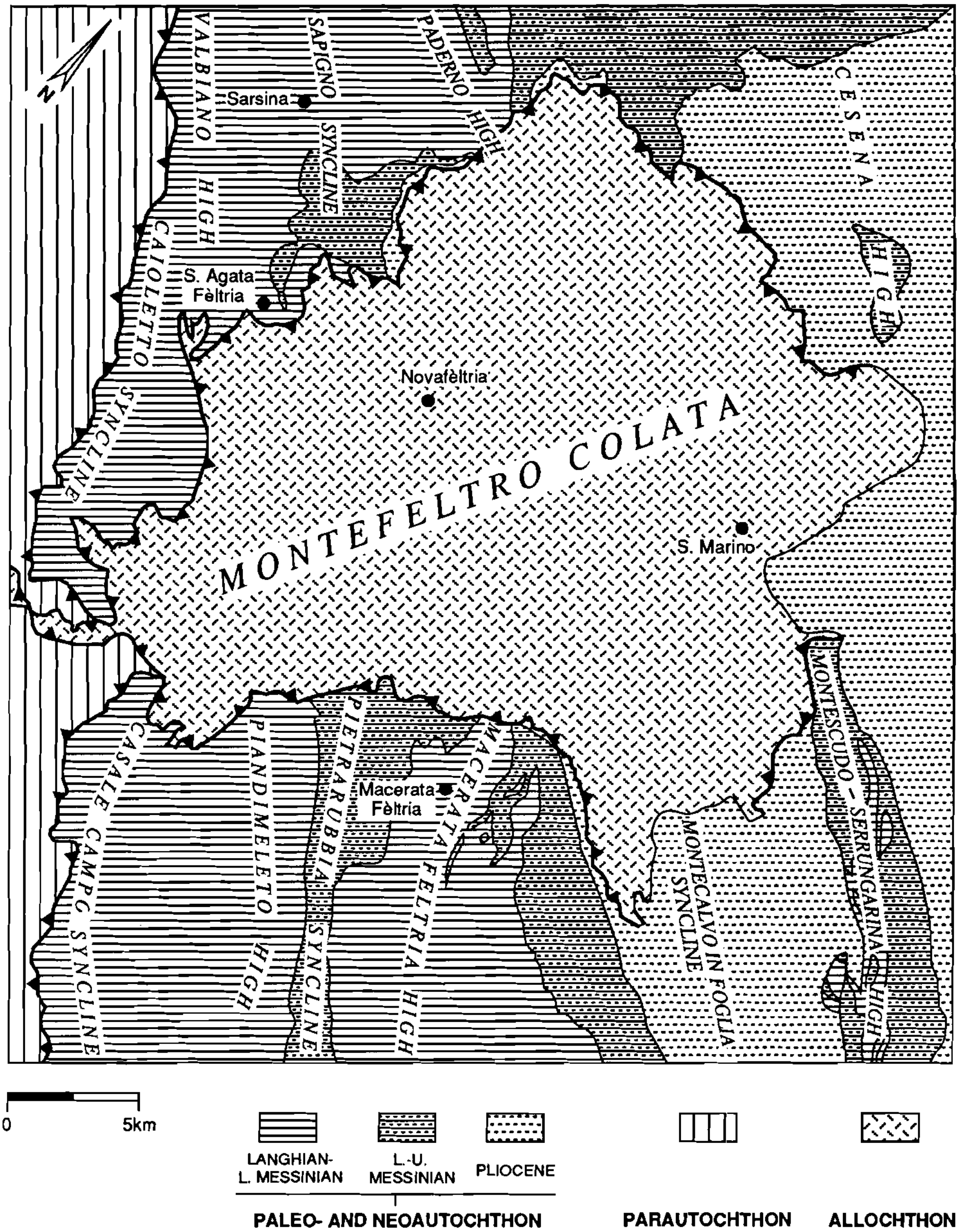


Fig. 98. Outline of the Montefeltro Colata and its surroundings.

autochthonous assemblage on either side of the Montefeltro Colata seem to be roughly in line (Fig. 98). This suggests that the allochthon does not conceal a significant transversal discontinuity. Apparently, a mechanism other than deep-seated shearing caused the primary depression. Gretener (1972, 1981) illustrated that superficial loads of limited dimensions, such as the Montefeltro Colata, can provoke crustal downwarping. In accordance with the scenario outlined for the Sillaro Line by Ceretti & Colalongo (1984), isostatic subsidence must have taken place in front of the mobile allochthon. This accelerated as allochthonous elements invaded the embryonic depression (cf. Ricci Lucchi & ori, 1985; Ricci Lucchi, 1986a; Veneri, 1986a). Footwall material forced aside hereby shaped marginal bulges. Probably, this process also originated the oblique faults dissecting the Pietrarubbia Syncline N of Frontino (enclosed map and cross-section M-M'). The transversal depression was further accentuated because the overburden of the Montefeltro Colata hindered detachment tectonics to thicken its substratum (cf. Migliorini, 1948; de Jager, 1979; van den Berg, 1990). The pattern of recent seismic activity in the external part of the Northern Apennines indicates that allochthonous masses indeed have a suppressive effect on footwall shearing (Elmi *et al.*, 1981). Marginal bulges thus mark the boundaries between differentially shortened transveral sectors of the autochthonous assemblage, not unlike tear faults.

The emplacement of the Montefeltro Colata interfered with the superficial deformation of the autochthon to produce the longitudinal normal faults dissecting the inner flank of the Piandimeleto High W of Frontino (enclosed map and cross-sections D-D' and E-E'). It furthermore determined fading and dextral bending of regular longitudinal structures. Actual transversal tear faults have not been encountered. They may however very well be concealed by marginal portions of the allochthon.

VI.5. SYNOPSIS

The essence of colata tectonics consists in the intermittent gravitational progress of an aggregate of lenticular bodies like a giant mass flow. In this respect, the well-preserved Montefeltro Colata may be taken as an archetype. The dynamics of its emplacement are indicated by internal geometry and position relative to footwall stratigraphy.

The arcuate pattern of imbricates generally attests to the direction of tectonic transport (Fig. 99). This permits the identification of secondary marginal features, such as defunct elements and lateral flows.

Through loading, the Montefeltro Colata created the transversal depression bordered by marginal bulges which guided forward transport. Its stepped positional younging towards the Adriatic Foreland partly reflects the temporary obstructive effect of the evolving longitudinal highs of the autochthon (Fig. 100, 101). Van den Berg (1990) related phases of allochthonous progress in the Northern Apennines to sea-level drops. He documented the mechanical feasibility of emergence inducing gravitational gliding or spreading. The concomitant weakening of clay minerals in consequence of meteoric water reducing pore-fluid salinities could have stimulated the mobility of superficial terrains (cf. Goerler & Reutter, 1968). However, the principal advances of the Montefeltro Colata at the end of the Early Tortonian, at the end of the

Messinian, and during the late Early Pliocene manifestly do not correlate with regressive cycles. They must therefore have been determined by tectonic rather than eustatic pulses. Their coevality with intensified autochthonous shearing is particularly significant in this context (Fig. 101). Presumably, migratory orogenic loading and relative tilting impelled both the paroxysmal detachment tectonics of the autochthonous assemblage and the phased progress of the Montefeltro Colata.

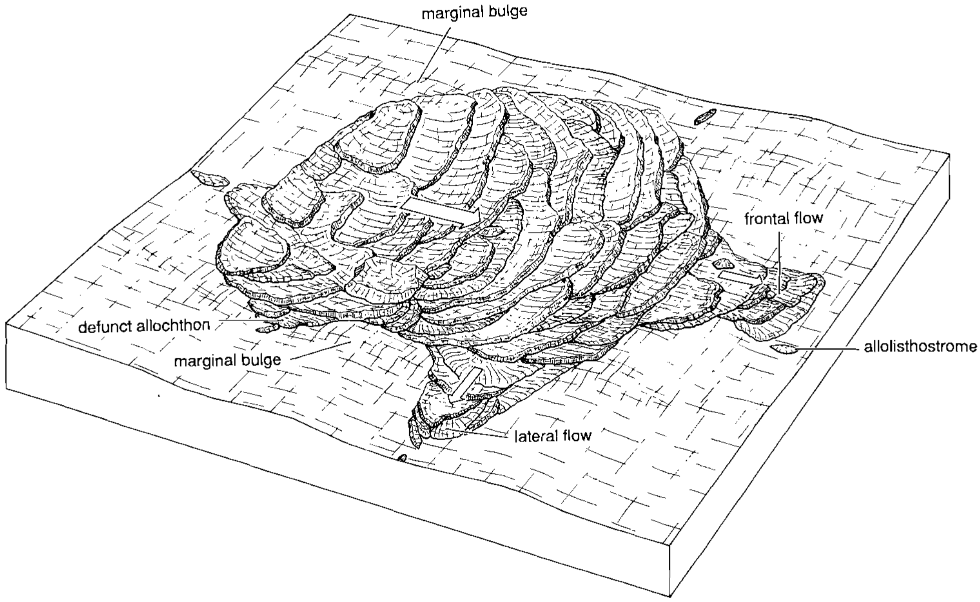


Fig. 99. Anatomy of a colata (based on the structural framework of the Montefeltro Colata).

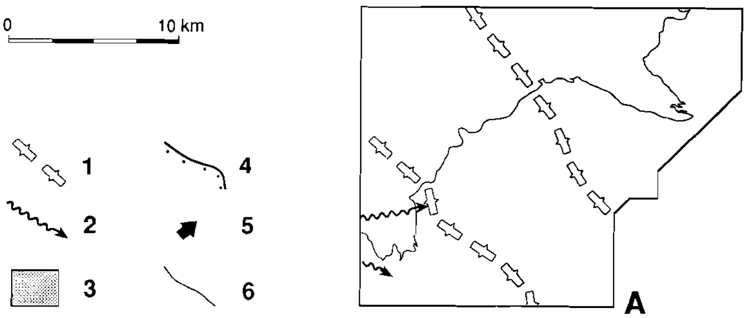
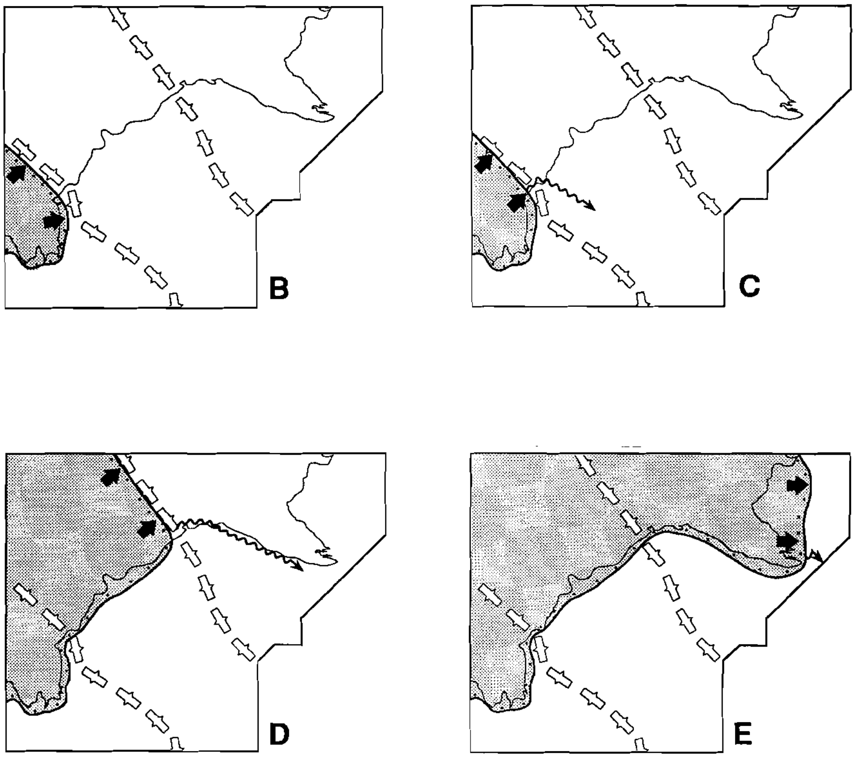


Fig. 100. Reconstruction of the emplacement of the Montefeltro Colata in the investigated area. Legend: 1: crest of major autochthonous high (in present-day position); 2: dispersal of material detached from front



of allochthon; 3: coherent allochthon; 4: transgressive thrust fault, with dots on upper block; 5: frontal progress of coherent allochthon; 6: present-day margin of Montefeltro Colata. Evolutionary stages: A (Early Tortonian): allolisthostromes and detached frontal slices accumulate in Casale-Campo Syncline as precursors of approaching Montefeltro Colata; B (end of Early Tortonian): Montefeltro Colata advances to Piandimeleto High; C (Late Tortonian): voluminous allolisthostromes accumulate in Pietrarubbia Syncline due to frontal instability of Montefeltro Colata; D (end of Messinian): Montefeltro Colata advances to Macerata Fèltria High, shedding allolisthostromes into Montecalvo in Foglia Syncline; E (late Early Pliocene): Montefeltro Colata invades Montecalvo in Foglia Syncline, preceded by allolisthostromes.

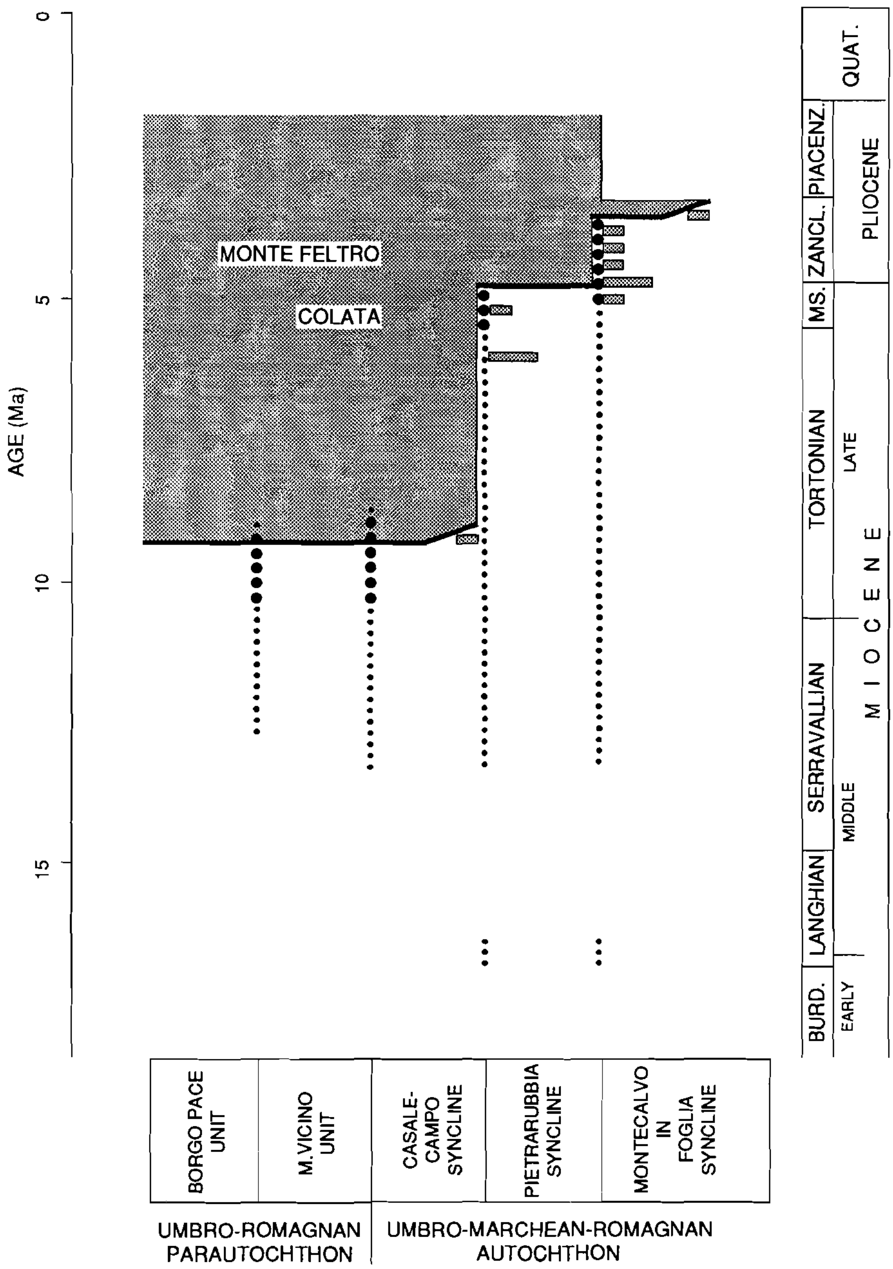


Fig. 101. Diagram illustrating the contemporaneity in the investigated area of the phased progress of the Montefeltro Colata and associated precursory elements with the detachment tectonics of the autochthonous assemblage (indicated by dots sized according to intensity).

CHAPTER VII

CONCLUSIONS AND DISCUSSION

VII.1. GRAVITY TECTONICS AND SEDIMENTATION

The structural and stratigraphic features of the southern Montefeltro reflect the complex interaction between tectonics and sedimentation in the Umbro-Marchean-Romagnan Zone during the Neogene. Typically, superficial shearing affected active basins according to the piggyback concept outlined by Ori & Friend (1984).

The elongate Umbro-Romagnan Foredeep, which came into existence at

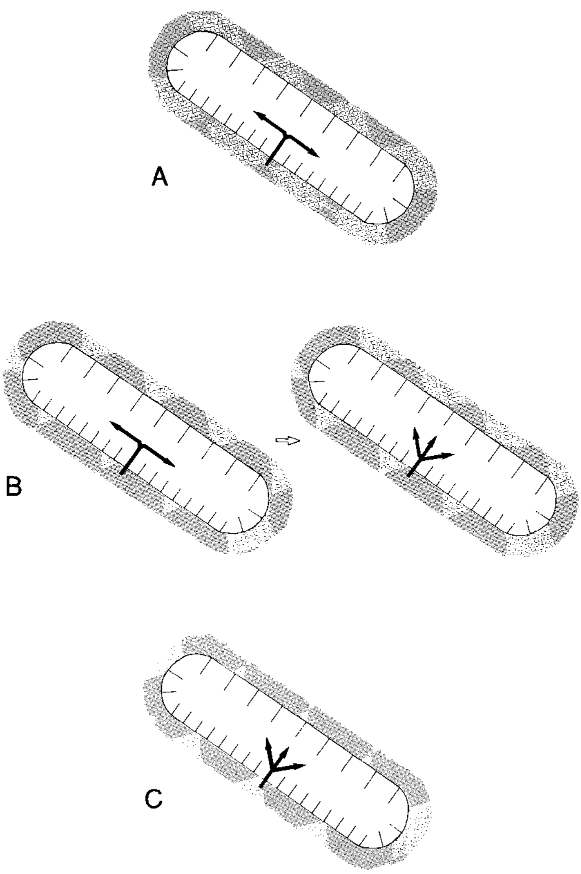


Fig. 102. Dispersal patterns in elongate Apenninic foredeeps as a function of the rate of transversal clastic supply relative to basin topography. A: minor relative supply (e.g., Campo Marl Formation partly in Casale-Campo Syncline); B: intermediate relative supply (e.g., Colombacci Formation in Pietrarubbia Syncline); C: major relative supply (e.g., Montecalvo in Foglia Formation in Montecalvo in Foglia Syncline).

the end of the Early Miocene, underwent two distinctive depositional stages. The conspicuous turbidite suite it accumulated at first was largely supplied longitudinally by Austro-Alpine sources. The relative subsidence of successive foredeep segments during the Middle Miocene determined the systematic shifting of the turbiditic sedimentation pattern toward the Adriatic Foreland. It furthermore provoked intermittent shallow gravitational spreading which generated synformal subbasins. These evolved into tectono-stratigraphic units during the Late Miocene, when the sedimentary fill of the inner and central foredeep segment was translated toward the Adriatic Foreland as a parautochthonous thrust sheet.

Gravitational spreading of the sedimentary cover as a whole caused the growth of longitudinal structures within the outer foredeep segment. This peaked during the Late Miocene to Early Pliocene. The migratory orogenic paroxysm brought on the second foredeep stage, characterized by the continental to shallow-marine deposition of clastics supplied transversely by Apenninic sources. Its onset roughly coincided with the evaporitic event of the Messinian Salinity Crisis of the Mediterranean. Dispersal patterns chiefly depended on the rate of detrital supply relative to shape of evolving synformal basins (Fig. 102).

At the same time as the paroxysmal shearing of the autochthonous assemblage of the Umbro-Marchean-Romagnan Zone, prominent allochthonous masses such as the Montefeltro Colata glided gravitationally toward the Adriatic Foreland. This discontinuous process markedly complicated foredeep evolution.

VII.2. GEODYNAMIC SETTING

The pattern of migratory synsedimentary thin-skinned tectonics outlined for the southern Montefeltro is exemplary of Apenninic orogeny following the Late Eocene collision of the Sardo-Corsican and Adriatic continental blocks. Irrespective of gravitational aspects, this general superficial shortening demands the disappearance of substantial amounts of continental lithosphere (Bally *et al.*, 1986). Many authors envisaged the subduction of Adriatic lithosphere under the Sardo-Corsican block (*e.g.*, Boccaletti *et al.*, 1971; Biju-Duval & Montadert, 1977; Reutter & Groscurth, 1978). Yet, geophysical surveys have revealed that to some extent the Adriatic continental crust actually overlies the Sardo-Corsican one (Morelli *et al.*, 1977; Giese *et al.*, 1978; Cassinis, 1981). This has been attributed to secondary obduction by Reutter *et al.* (1978), Boccaletti *et al.* (1980), Nicolich (1981) and Reutter (1981). Apenninic foredeeps would then represent a peripheral foreland basin system related to the progressive decoupling and imbrication of the Adriatic crust in consequence of a retreating subduction zone or as a synthetic effect of a stationary subduction zone (Fig. 103a). The scarcity of relevant data however does not permit to reject *a priori* the alternative option of the subduction of Sardo-Corsican lithosphere under the Adriatic block. In this case, Apenninic foredeeps would constitute a retroarc foreland basin system determined by antithetic shearing of the Adriatic crust (Fig. 103b). Such a geodynamic framework for the ensialic stage of Apenninic orogeny in fact appears compatible with the general tectono-stratigraphic criteria listed by Ingersoll (1988).

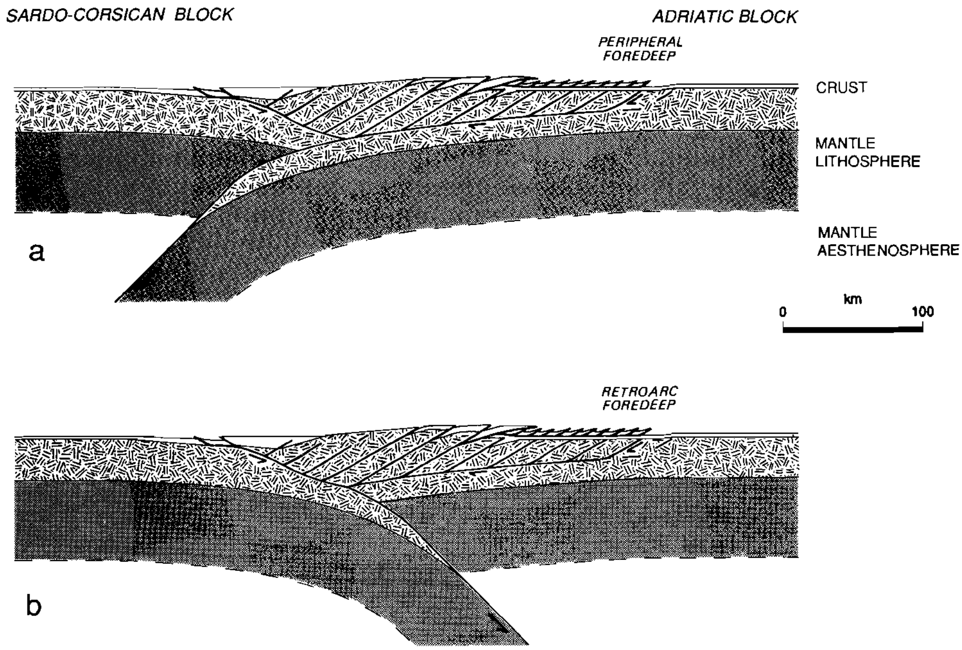


Fig. 103. PLate-tectonic models for the ensialic stage of Apenninic orogeny.

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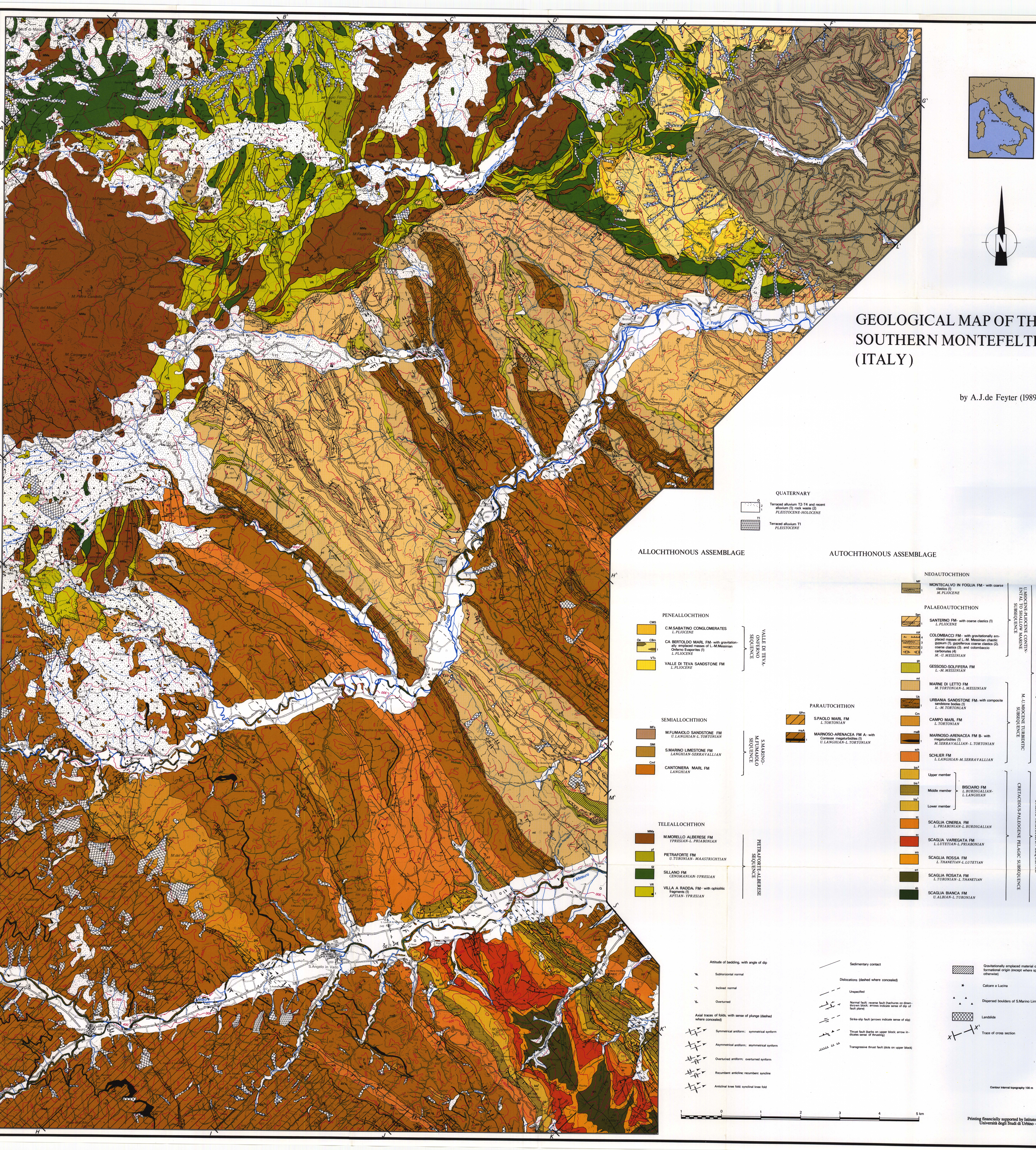
CURRICULUM VITAE

De schrijver van dit proefschrift werd geboren op 28 mei 1957 te Heerlen. Op 20 juni 1975 behaalde hij cum laude het diploma Atheneum B aan het Eijkhagencollege te Schaesberg. In hetzelfde jaar werd begonnen met de studie geologie aan de Rijksuniversiteit te Utrecht. Op 13 maart 1978 werd het kandidaatsexamen G3 afgelegd, gevolgd op 16 maart 1981 door het doctoraalexamen met hoofdvak structurele geologie en bijvakken sedimentologie, economische geologie en exploratie geofysica.

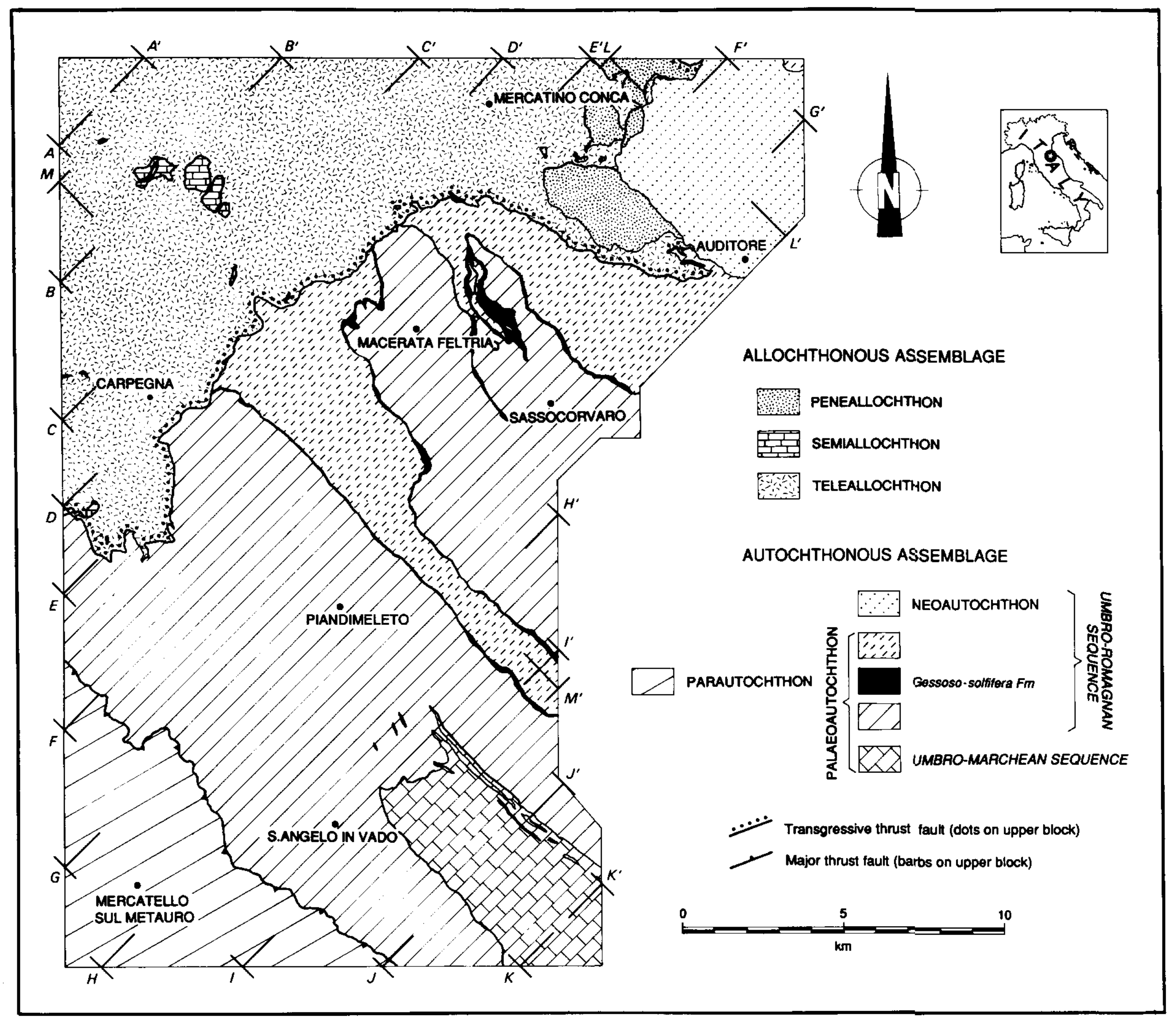
Na zijn afstuderen heeft de schrijver periodiek wetenschappelijk onderzoek verricht. Inmiddels is hij werkzaam bij AGIP S.p.A. te S.Donato Milanese (Italië).

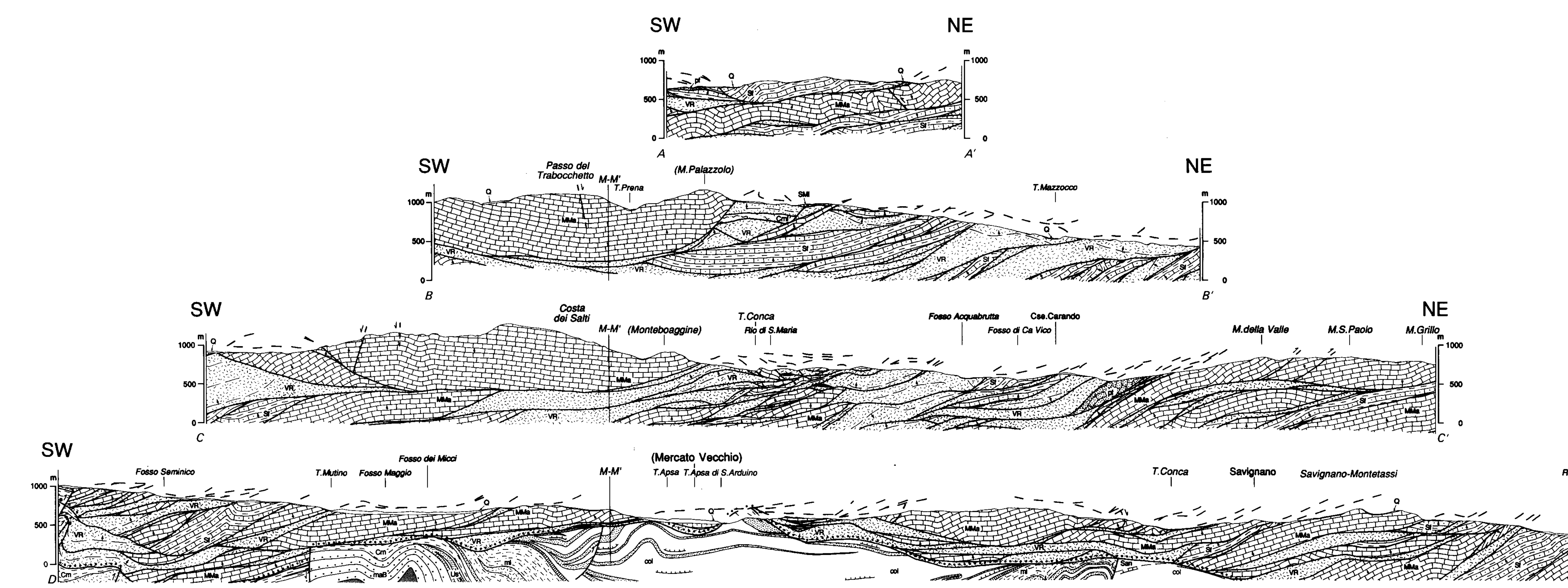


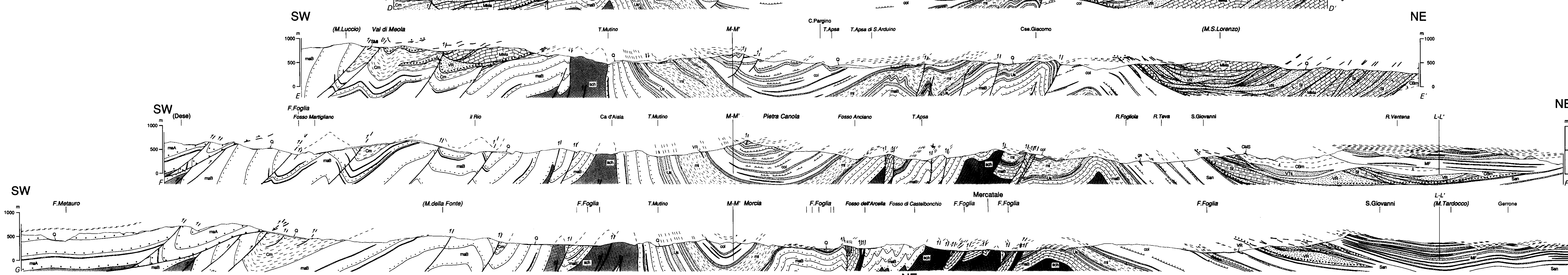
GEOLOGICAL MAP OF TH
SOUTHERN MONTEFELT
(ITALY)
by A.J.de Feyter (1989
QUATERNARY
Terraced alluvium T2-T4 and recent alluvium (1); rock waste (2)
PLEISTOCENE-HOLOCENE
Terraced alluvium T1
PLEISTOCENE
ALLOCHTHONOUS ASSEMBLAGE
AUTOCHTHONOUS ASSEMBLAGE
PENEALLOCHTHON
C.M.SABATINO CONGLOMERATES
L.PLIOCENE
CA BERTOLDO MARL FM- with gravitationally emplaced masses of L.-M.Messinian Onferno Evaporites (1)
L.PLIOCENE
VALLE DI TEVA SANDSTONE FM
L.PLIOCENE
VALLE DI TEVA-ONFERNO SEQUENCE
SEMIALLOCHTHON
M.FUMAIOLO SANDSTONE FM
U.LANGHIAN-L.TORTONIAN
S.MARINO LIMESTONE FM
LANGHIAN-SERRAVALLIAN
CANTONIERA MARL FM
LANGHIAN
S.MARINO-M.FUMAIOLO SEQUENCE
TELEALLOCHTHON
M.MORELLO ALBERESE FM
YPRESIAN-L.PRIABONIAN
PIETRAFORTE FM
U.TURONIAN-MAASTRICHTIAN
SILLANO FM
CENOMANIAN-YPRESIAN
VILLA A RADDA FM- with ophiolitic fragments (1)
APTIAN-YPRESIAN
PIETRAFORTE-ALBERESE SEQUENCE
PARAUTOCHTHON
S.PAOLO MARL FM
L.TORTONIAN
MARNOSO-ARENACEA FM A- with Contessa megaturbidites (1)
U.LANGHIAN-L.TORTONIAN
NEOAUTOCHTHON
MONTECALVO IN FOGLIA FM- with coarse clastics (1)
M.PLIOCENE
PALAEOAUTOCHTHON
SANTERNO FM- with coarse clastics (1)
L.PLIOCENE
COLOMBACCI FM- with gravitationally emplaced masses of L.-M. Messinian chaotic gypsum (1), gypsiferous coarse clastics (2), coarse clastics (3), and colombaccio carbonates (4)
M.-U.MESSINIAN
U.MIOCENE-PLIOCENE CONTINENTAL TO SHALLOW MARINE SUBSEQUENCE
GESSOSO-SOLFIFERA FM
L.-M.MESSINIAN
MARNE DI LETTO FM
M.TORTONIAN-L.MESSINIAN
URBANIA SANDSTONE FM- with composite sandstone bodies (1)
L.-M.TORTONIAN
CAMPO MARL FM
L.TORTONIAN
MARNOSO-ARENACEA FM B- with megaturbidites (1)
M.SERRAVALLIAN-L.TORTONIAN
SCHLIER FM
L.LANGHIAN-M.SERRAVALLIAN
M.-U.MIOCENE TURBIDITIC SUBSEQUENCE
Upper member
Middle member
Lower member
BISCIARO FM
L.BURDIGALIAN-L.LANGHIAN
SCAGLIA CINEREA FM
L.PRIABONIAN-L.BURDIGALIAN
SCAGLIA VARIEGATA FM
L.LUTETIAN-L.PRIABONIAN
SCAGLIA ROSSA FM
L.THANETIAN-L.LUTETIAN
SCAGLIA ROSATA FM
L.TURONIAN-L.THANETIAN
SCAGLIA BIANCA FM
U.ALBIAN-L.TURONIAN
CRETACEOUS-PALEOGENE PELAGIC SUBSEQUENCE
Attitude of bedding, with angle of dip
Subhorizontal normal
Inclined normal
Overturned
Axial traces of folds, with sense of plunge (dashed where concealed)
Symmetrical antiform; symmetrical synform
Asymmetrical antiform; asymmetrical synform
Overturned antiform; overturned synform
Recumbent anticline; recumbent syncline
Anticlinal knee fold; synclinal knee fold
Sedimentary contact
Dislocations (dashed where concealed)
Unspecified
Normal fault; reverse fault (hachures on downthrown block; arrows indicate sense of dip of fault plane)
Strike-slip fault (arrows indicate sense of slip)
Thrust fault (barbs on upper block; arrow indicates sense of thrusting)
Transgressive thrust fault (dots on upper block)
Calcare a Lucina
Landslide
Trace of cross section
Contour interval topography 100 m
1 0 1 2 3 4 5 km
Printing financially supported by Istituto
Università degli Studi di Urbino -

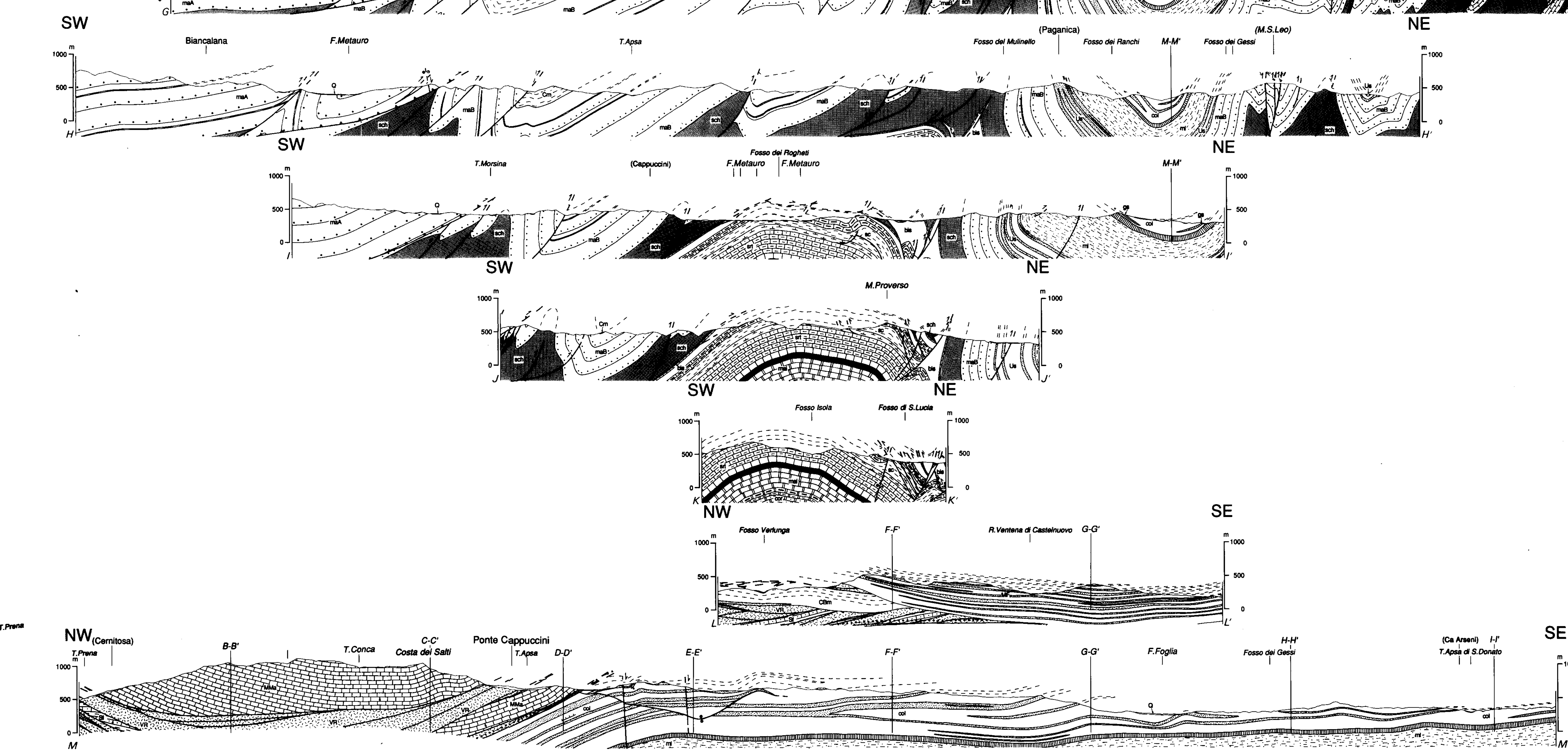
STRUCTURAL SECTIONS ACROSS THE SOUTHERN MONTEFELTRO (NORTHERN APENNINES, ITALY)

A.J. de Feyter (1990)









Basal thrusts of allochthon and parautochthon are marked by dots and barbs, respectively.