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No. 19

THE RELATION BETWEEN TECTONICS AND SEDIMENTATION ALONG THE 'SILLARO LINE'

(Northern Apennines, Italy)

J. DE JAGER

STELLINGEN

1

De huidige positie van de 'Monghidoro-plak' in de Noord Apennijnen wordt niet verklaard door het 'rolling rug' mechanisme (MAXWELL – 1959) noch door de latere hypothese van HSÜ (1967).

HSÜ K. J. (1967) Origin of large overturned slabs of Apennines, Italy. Bull. Am. Ass. Petr. Geol. **51**: 67–72.

MAXWELL J. C. (1959) Orogeny, gravity tectonics, and turbidites in the Monghidoro area, Northern Apennines, Italy. Trans. N. Y. Acad. Sci. **21**: 269–280.

2

Laat-geosynclinale turbidiet sedimentatie wordt niet direct veroorzaakt door orogenetische bewegingen in internere delen van het orogeen, maar door een snellere verdieping van het geosynclinale bekken.

3

Er zijn geen duidelijke geologische argumenten dat de oer-atmosfeer van de aarde meer NH_3 bezat dan de huidige. Bovendien is zulks onwaarschijnlijk en onnodig.

BADA J. L. and MILLER S. L. (1968) Ammonium ion concentration in the primitive ocean. Science **159**: 423–425.

MOROZ V.I. and MUKHIN L. M. (1978) About the initial evolution of atmosphere and climate of the earth type planets. Acad. of Sc., USSR, Space Res. Inst.

SAGAN C. en MULLEN G. (1972) Earth and Mars: Evolution of atmospheres and surface temperatures. Science **177**: 52–56.

4

De Briançonnais zone van de West Alpen is geen goed voorbeeld van een 'Briançonnais-type miogeosynclinale rug' als bedoeld door AUBOUIN (1965).

AUBOUIN J. (1965) Geosynclines. Developm. in geotect. **1**.

5

De samenstelling van de Maan, zoals die nu bekend is, maakt theorieën volgens welke de Maan ontstaan is als een onafhankelijk lichaam door directe accretie in de zonne-nevel, zoals in de 'dubbelplaneet-hypothese' en de 'vang-hypothese' zeer onwaarschijnlijk.

RINGWOOD A. E. (1977) Basaltic magmatism and the bulk composition of the Moon (I and II). The Moon and the planets **16** (4): 389–464.

6

De naam 'Flysch Karpaten' voor de externe (noordelijke) Karpaten suggereert ten onrechte dat het merendeel van de turbidiet formaties aldaar als flysch gekenmerkt kan worden.

BIRKENMAJER K. (1974) Carpathian Mountains. in: Mesozoic-Cenozoic orogenic belts. ed: SPENCER A. M. the Geol. Soc. spec. publ. **4**: 127–157.

7

De anatexitische genese van de stollingsgesteenten van de Oslo-slenk uit de Pré-Cambrische korst wordt door de argumenten die RUTTEN (1969) gebruikt nauwelijks gestaafd.

RUTTEN M. G. (1969) The geology of western Europe. Elsevier Publ. Comp.

8

Het gebruik van letter-cijfer coderingen voor planktonzones is prematuur.

9

Het moet als zeer onwaarschijnlijk beschouwd worden dat de Silesische Cordillera in de Noord Karpaten de grootste leverancier is geweest van materiaal waaruit de Noord Karpatense turbidiet formaties bestaan.

KSIAZKIEWICS M. (1956) Geology of the Northern Carpathians. Geol. Rundsch. **45**: 369–411.

10

Veelal wordt over het hoofd gezien dat de zandsteen van een zandsteen-mergel afwisseling ten tijde van de plooiing nog niet gelithificeerd en dus niet competenter dan de mergel hoefde te zijn.

11

De veronderstelling van Mutch (1970) dat het ontbreken van gelaagdheid op de Maan pleit voor een 'Apennijns' type afzetting is onjuist.

MUTCH TH. A. (1970) Geology of the Moon, a stratigraphic view. Princeton Univ. Press, Princeton New Jersey.

12

Aangezien mensen die buitenaardse bemande voertuigen menen waar te nemen deze zelf al veelal u.f.o.'s (unidentified flying objects) noemen, lijkt enige scepsis ten aanzien van berichten over signaleringen van deze voertuigen gerechtvaardigd.

Stellingen behorende bij het proefschrift 'The relation between tectonics and sedimentation along the 'Sillaro line' (Northern Apennines, Italy) door J. de Jager.

Utrecht, 5 februari 1979.

THE RELATION BETWEEN TECTONICS AND SEDIMENTATION ALONG THE 'SILLARO LINE'

(Northern Apennines, Italy)

THE RELATION BETWEEN TECTONICS AND SEDIMENTATION ALONG THE 'SILLARO LINE'

(NORTHERN APENNINES, ITALY)

PROEFSCHRIFT
TER VERKRIJGING VAN DE GRAAD
VAN DOCTOR IN DE WISKUNDE EN NATUURWETENSCHAPPEN
AAN DE RIJKSUNIVERSITEIT TE UTRECHT,
OP GEZAG VAN DE RECTOR MAGNIFICUS
PROF. DR. A. VERHOEFF,
VOLGENS BESLUIT VAN HET COLLEGE VAN DECANEN
IN HET OPENBAAR TE VERDEDIGEN
OP MAANDAG 5 FEBRUARI 1979
DES NAMIDDAGS TE 4.15 UUR

DOOR

JAN DE JAGER

geboren op 11 maart 1950 te Oudenrijn

Promotor: Dr. E. ten Haaf

Mijn dank gaat uit naar ieder die geholpen heeft
bij de totstandkoming van dit proefschrift

Het foto- en tekenwerk is verzorgd door de tekenaars van het Geologisch Instituut onder leiding van dhr. F. Henzen.

Druk: Veenman Wageningen.

CONTENTS

SUMMARY

The transverse 'Sillaro line' is merely the outcrop line of the north-west dipping overthrust plane of the Liguride Complex. The anti-Apenninic strike of this overthrust plane is the result of the 'Sillaro flexure', a large synsedimentarily developed flexure of the basement of the Northern Apennines. The Liguride Complex overlies to the north-east progressively younger deposits dating from Early Miocene to eventually Pliocene: the Liguride overthrust is tectonically transgressive. The Liguride Complex is made up largely of clay, the Argille Scagliose; the overthrusting was gravitational. Just like glaciers and ice-sheets the Liguride Complex is a plastic mass capable of spreading under the influence of gravitation. Olistostromization played a role of varying importance in the advance of the Liguride Complex. During the Pliocene, when a shallow molasse basin was invaded, it was the dominating mechanism. Locally the Liguride Complex interfingers with autochthonous deposits, as a result of a temporary standstill of the thrust sheet. The deposits just below the overthrust plane often show the influence of the approaching Liguride Complex.

The Marnoso Arenacea is a turbidite formation of Miocene age. The configuration of its sedimentation basin was influenced strongly by an anticline that developed synsedimentarily, resulting in the Castel Vecchio 'high'. This 'high' divided the basin into an inner and outer basin in the south-west and north-east respectively, each with its own sedimentary history. The position of the inner basin axis was defined by the Rovigo syncline, which was also synsedimentary. A 'klippe sédimentaire', the Casaglia Member, invaded the basin from the south-west, and came to a standstill near the basin axis. It proved impossible to describe the turbidites in terms of inner and outer fan and basin plain deposits by analogy with recent turbidite sedimentation in oceanic environments. A purely descriptive classification in facies associations is introduced here. The thickest and coarsest turbidites were usually deposited in the deepest parts of the basin. Towards the 'high' they became thinner and finer grained, until only marls were deposited. Upon the Castel Vecchio 'high' moluscs are found in living position, indicating that occasionally this 'high' approached sea-level.

The Castel Guerrino Formation which constitutes a separate tectonic unit, was probably deposited in a more internal part of the Marnoso Arenacea basin. This formation thrust from the south-west upon the Marnoso Arenacea, after which it overthrust 5 km further to the north-east over a thin slice of Liguride rocks which had already overthrust the Marnoso Areacea.

SAMENVATTING

Langs de zuidwest-noordoost strekkende 'Sillaro lijn' dagzoomt het noordwest hellende overschuivingsvlak van het Liguride Complex. Deze anti-Apennijnse strekking is het gevolg van een grootschalige synsedimentaire flexuur, de 'Sillaro flexuur', waarin ook het basement betrokken is. Het Liguride Complex is alleen ten noordwesten van deze flexuur overschoven. In het zuidwesten overschuift het Liguride Complex vroeg-Miocene afzettingen, naar het noordoosten toe klimt het Complex in de stratigrafie, totdat uiteindelijk Pliocene afzettingen worden bedekt: de Liguride overschuiving is tektonisch transgressief. Het Liguride Complex dat in het bestudeerde gebied voor het grootste deel uit klei, Argille Scagliose, bestaat, overschoof gravitatief. Analoog aan gletsjers en ijskappen kan het Liguride Complex beschreven worden als een plastische massa die onder invloed van de zwaartekracht kan spreiden. Olistostromizatie speelde een rol van wisselende betekenis in de voortbeweging van het Liguride Complex. Gedurende het Plioceen, toen het dekblad een ondiep molasse bekken ingleed, was dit het belangrijkste mechanisme. Plaatselijk vervingert het Liguride Complex met autochtone afzettingen, als gevolg van tijdelijke stilstanden van het dekblad. Vlak onder het overschuivingsvlak zijn vaak invloeden te zien van het naderende Liguride Complex op de autochtone turbidietsedimentatie.

De Marnoso Arenacea is een Miocene turbidietformatie. De configuratie van zijn sedimentatie bekken werd sterk beïnvloed door een zich synsedimentair ontwikkelende antikline die het Castel Vecchio 'hoog' veroorzaakte. Dit 'hoog' verdeelde het bekken in een binnen en een buiten bekken in respectievelijk het zuidwesten en noordoosten, ieder met zijn eigen sedimentaire ontwikkeling. De eveneens synsedimentaire Rovigo syncline bepaalde de ligging van de binnen bekken-as. Het Casaglia Member vormt een 'klippe sédimentaire' die het bekken vanuit het zuidwesten ingleed, en nabij de bekken-as tot stilstand kwam. Het bleek onmogelijk de turbidiet-afzettingen te beschrijven naar analogie van recente turbidiet sedimentatie in de oceanen, in termen van inner fan, outer fan en basin plain afzettingen. Een zuiver beschrijvende classificatie in facies associaties is hier geïntroduceerd. Aan de hand van zowel tektonische en sedimentaire gegevens is tot een bekken analyse gekomen. De dikste en grofkorreligste turbidieten zijn in het algemeen in de diepste delen van het bekken afgezet. Naar het 'hoog' toe werden de turbidieten dunner en fijnkorreliger, totdat uiteindelijk alleen nog maar mergels werden afgezet. Het voorkomen van molusken in levenspositie op het Castel Vecchio 'hoog' toont

dat dit 'hoog' zo nu en dan het zeeniveau benaderde.

Waarschijnlijk is de Castel Guerrino Formatie, een onafhankelijke tektonische eenheid, afgezet in een meer intern deel van het Marnoso Arenacea bekken. Deze formatie is vanuit het zuidwesten opgeschoven over de Marnoso Arenacea, en is daarna nog 5 km doorgeschoven over een dunne zone van Liguride gesteenten, die zelf overschoven liggen op de Marnoso Arenacea.

1. INTRODUCTION

1.1. GENERAL

During the summers of 1973, 1974, 1976 and 1977 geological fieldwork was carried out in the Tuscan-Emilian Apennines. The area investigated covers approximately 1000 square kilometres, and is situated in the extreme north-east of Tuscany (province of Florence) and the southern part of Emilia-Romagna ('Romagna Fiorentina', province of Bologna) (fig. 1). The main towns in the area, which are to be found along the river Santerno, are Firenzuola in the south-west and Borgo Tossignano in the north-east.

Topographical maps were obtained from the 'Istituto Geografico Militare' in Florence. Use was made of the sheets 98 (Vergato) and 99 (Faenza), both on the scale 1 : 100.000, and of the 1 : 25.000 maps 98 I NE, SE, II NE, SE, NO, SO, 99 III NO, SO, IV NO, SO.

1.2. GENERAL GEOLOGICAL SETTING

The eugeosynclinal rocks of the Northern Apennines, commonly referred to as the 'Ligurids' or the 'Liguride Complex', comprise several sedimentary sequences representing the basin fills of, probably, several basins situated west of the present Northern Apennines. ABBATE and SAGRI (1970) distinguish the Helminthoid flysch sequences, the Vara Supergroup, the Calvana Supergroup and the Canetolo Complex. Whether the Canetolo Complex was really deposited in the eugeosynclinal domain is uncertain. It is younger than the other eugeosynclinal rocks, and in outcrop is always associated with the miogeosynclinal Tuscan Sequence. Therefore it is often called a sub-Liguride unit.

Sedimentation in the eugeosynclinal realm can be divided in a pre-flysch and a flysch period. The Jurassic and Lower Cretaceous pre-flysch sediments are radiolarian cherts, limestones and mudstones. The Vara Supergroup begins with ophiolites: serpentinites, gabbros and diabases. The flysch period starts during the Upper Cretaceous and continues through the Eocene.

The miogeosynclinal rocks are represented by the Tuscan Sequence in the south-west and the Umbrian Sequence in the east. Pre-flysch miogeosynclinal sedimentation starts during the Triassic with limestones and cherts, overlain by a pelagic formation, the Scaglia Toscana. The latter underlies the Tuscan

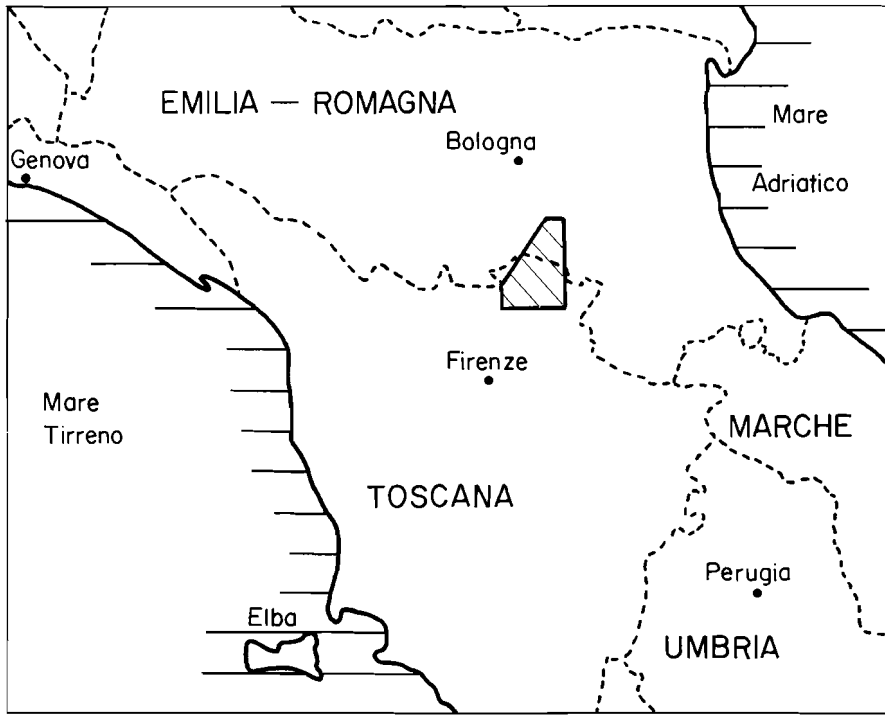

FIG. 1. Location of the area investigated in the Northern Apennines (Italy).

Macigno, a flysch series of Middle-Late Oligocene to Early Miocene age. The north-east part of this formation is often referred to as the Modino-Cervarola Sequence. In the more external Umbrian Sequence the flysch is called Marnoso Arenacea, and is of Miocene age. To the south-east the marly formations Bisciaro and Schlier are partly time-equivalent with the Marnoso Arenacea. The relationships between the various flysch formations are still a matter of debate.

The first major tectonic phase took place during the Eocene in the eugeosynclinal realm, and was probably of compressional character. Ligurian nappes began overthrusting the Tuscan Sequence during the Oligocene, and reached the Umbrian Sequence during the Miocene. Sedimentation in the miogeosyncline stopped because of the arrival of the Liguride Complex. Contemporaneous with or slightly before the overthrusting by the Ligurides, in our area the upper part of the Tuscan Sequence (the Scaglia Toscana, the Macigno and the Modino-Cervarola Sequence) started overthrusting to the north-east. During the Miocene and Pliocene the Liguride Complex overthrust the Umbrian Sequence. The Umbrian Marnoso Arenacea is usually regarded as autochthonous, although the south-west parts of it are very often thrust upon and sometimes even over more external parts. The driving force of the tectonics in the miogeosynclinal realm of the Northern Apennines was gravitation.

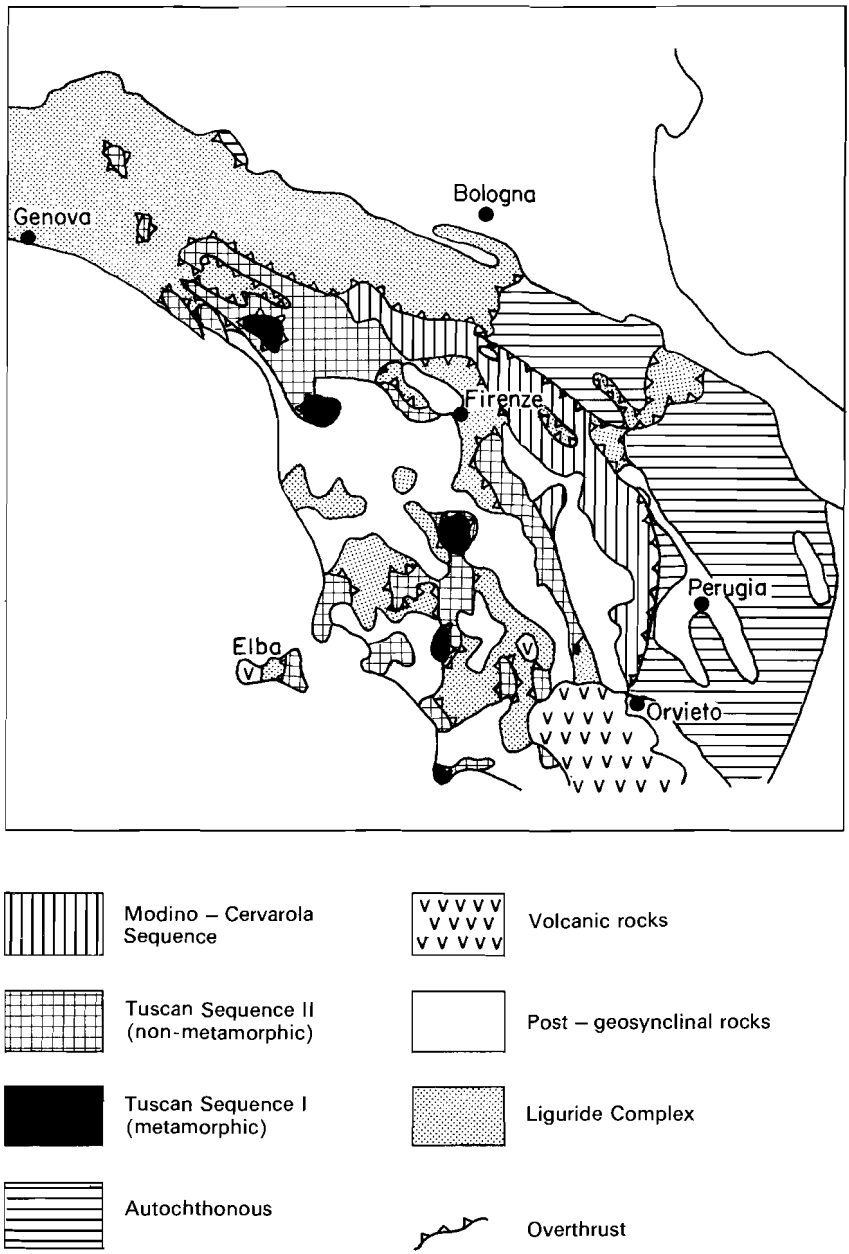

FIG. 2. Schematic tectonical map of the Northern Apennines.

All post-Tortonian deposits are considered as late or post-geosynclinal. Older late-geosynclinal rocks such as the Late Eocene to Middle Miocene Ranzano-Bismantova and Loiano Sequences had been deposited upon the Liguride Complex while it was overthrusting towards the north-east.

1.3. PREVIOUS AUTHORS

Transverse tectonic lines across the Apennines have been known to exist since 1933, when SIGNORINI and SACCO contemporaneously described several of these lines. They regarded the Sillaro line as the most obvious and important one. BOCCALETTI and GUAZZONE (1970) associated the Sillaro line with the counter-clockwise rotation of Italy. BORTOLOTTI (1966) saw the Sillaro line as a continuation of the Livorno-Pistoia line, which, according to him, existed already as early as the Jurassic. GROSCURTH (1971) and GROSCURTH and HEMMER (1973) saw the Sillaro line as a structure of synsedimentary origin. They found evidence that this line, which they regard as a transverse normal fault, already existed during the Oligocene. Most other investigators (GHELARDONI – 1965) see the Sillaro line as a transverse strike-slip fault which originated during the latest stages of orogenesis. Different opinions are expressed by BREUREN (1941) and BRUNI (1972). They state that after sedimentation had stopped the Liguride Complex overthrust the Marnoso Arenacea north-west of the Sillaro line. In this case the Sillaro line would be merely the outcropping overthrust plane.

SIGNORINI (1937, 1956) and BREUREN (1941) thought that they had evidence for overthrust phenomena within the Marnoso Arenacea. Later work carried out by GROSCURTH (1971) is in contradiction with their conclusions.

GROSCURTH (1971) is the only investigator who found evidence that the Sillaro line influenced the sedimentation of the Marnoso Arenacea. Others did not describe these features. RIZZINI and PASSEGA (1964) gave a fairly detailed description of the Marnoso Arenacea along the river Santerno, north of the Castel Vecchio-Palazzuolo structure. The most detailed study on the depositional cycle of the Marnoso Arenacea was carried out by RICCI LUCCHI and described in a number of papers (1965, 1967, 1968, 1969a, 1969b, 1973, 1975a). He described the deposits of the Marnoso Arenacea by comparing them to turbidites on the oceanic abyssal plains in terms of inner and outer fan, and basin plain deposits.

Important work on the overthrust mechanism of the Liguride Complex was done by MERLA (1952). He described the Liguride Complex as a series of orogenic landslides ('frane orogeniche') that slid to the north-east from successively rising 'tectonical ridges'. SIGNORINI (1956) described synthetic tiltings in the south-west and antithetic tiltings in the north-east part of the Liguride Complex. According to GÖRLER and REUTTER (1968) and GÖRLER (1975) the Liguride Chaotic Complex did not move as a whole, but was built up progressively by relatively small mudstreams. They computed that the front of the

Chaotic Complex in the Marecchia area advanced at an average speed of 3 metres per thousand years.

1.4. GEOMORPHOLOGY

Almost all of the area investigated lies north-east of the Apennine divide. Geomorphologically two different zones can be recognized: the Liguride Complex area north-west of the Sillaro line, and the area south-east of this line which is composed mainly of various turbidite formations. The turbidite formations, which are sandstone-marl alternations, resist erosion much better than the Liguride Complex, which is largely composed of clay, Argille Scagliose. Consequently, the highest mountains are found south-east of the Sillaro line. The mountains of the Liguride Complex area are not only less high, they are also less steep, except where major hard exotic components within the Liguride Complex protrude above their surroundings. Because of the plasticity of the Argille Scagliose and its ability to absorb vast amounts of water, many landslides (frane) occur within the Liguride Complex area.

Rivers that take their rise within the Liguride Complex area often follow a downstream course through the much harder turbidite formations area, making deep incisions. Because they do not visibly run along faults, it is thought that they chose their course before the Liguride Complex was eroded away from the turbidite formations there: so these rivers are antecedent. Good examples of antecedent rivers are the Santerno, the Diaterna and the Sillaro.

2. STRATIGRAPHY

2.1. INTRODUCTION

The rocks outcropping in the studied area can be divided into three sequences: the Umbrian, Tuscan and Liguride. The Umbrian Sequence is considered to be autochthonous, and is partly overthrust from the SW by the Tuscan Sequence. A separate position is occupied by the Castel Guerrino unit, a tectonic unit between the Umbrian and the Tuscan Sequence. All these rocks are considered to be miogeosynclinal (BORTOLOTTI et al. – 1970). They are overthrust by rocks belonging to the Liguride Sequence which had been deposited in the eugeosynclinal realm of the Northern Apennines geosyncline (ABBATE and SAGRI. – 1970).

The various formations will be described in order of tectonic superposition; i.e. first the Umbrian Sequence (Marnoso Arenacea, Gessoso Solfifera Formation, Casal Fiumanese Formation), then the Castel Guerrino unit (Castel

TABLE 1. Turbidite facies

facies	rock type	thickness	bed geometry	sand/marl ratio	sedimentary structures/ Bouma seq.
A	coarse sandstones conglomerates	0.5–15 m	lenticular	very high	none, or grading, lamination imbrication
B	medium to coarse sandstones	0.3–2 m	lenticular	> 1	grading, lamination (low angle, wavy, parallel), dish structures, cross lamination, cut and fills
C	medium to fine sandstones	0.5–3 m	even and parallel	> 1	classical proximal turbidites beginning with T_a
D_1	fine sandstones to siltstones	3–40 cm	even and parallel, local pinching out	low to very low, $\leqslant 1$	classical distal turbidites beginning with T_b or T_c
D_2	medium to very fine sandstones	up to 20 m	even and parallel	± 1	grading, plane parallel lamination
D_3	marlstones	0.5 m	even and parallel	zero	none or burrows
E	fine to coarse sandstones	3–20 cm	plane parallel to lenticular	medium to high, $\geqslant 1$	none or grading, cross lamination, T_{ae} or T_{ce}
F	chaotic deposits	up to 200 m	lenticular to sheet-like	–	slump folds, boudinage, contorted beds, etc.
H	breccias in very fine to coarse matrix	3 cm–1.5 m	plane parallel to lenticular	low, $\leqslant 1$	none or irregular grading, cross and parallel lamination

Guerrino Formation, Rifredo Marls), followed by the Tuscan Sequence (Scaglia Toscana, Cervarola Sandstone Formation), and the Liguride Formations (Chaotic Complex, Borgo Marls, Lame Limestones, Piotto Variegated Marls, Violla Marls, Monte Morello Formation, Sellustra Formation) and lastly the neo-autochthonous Mugello Formation.

In our description of the various turbidite formations a descriptive classification of turbidites and associated deposits compiled by MUTTI and RICCI LUCCHI (1972) is used. Their classification is based on features such as rock type, texture, bed thickness, bed geometry, sand/marl ratio and sedimentary features. The turbidites and associated deposits are classified as facies. The characteristics of each facies are summarized in table 1. The column 'rock type' refers only to the sandstone layer, but not in the case of facies D_3 and F. Facies H, which is not distinguished by Mutti and Ricci Lucchi, will be treated below. For more detailed information about the other facies the reader is referred to MUTTI and RICCI LUCCHI (1972) and RICCI LUCCHI (1975a).

Facies H (photo 2, 3) is a very distinct and easily recognizable one. BRUNI (1972) described layers belonging to this facies in the Marnoso Arenacea in the area N of Firenzuola (in the Diaterna valley). Angular components are embedded in a matrix that may be very fine (marl) to very coarse. The components have diameters of up to 1 m, but are usually in the range 0.5 to 3 cm. The thickness of the layers, which may be plane parallel to strongly lenticular, is 3 cm to 1.5 m. The bedding surfaces are irregular. The amount of angular

Photo 1. Gullies at the base of a coarse grained parallel laminated turbidite of facies B. The gullies have their axis in the direction of the supply as indicated by the flute casts superimposed upon the gullies.

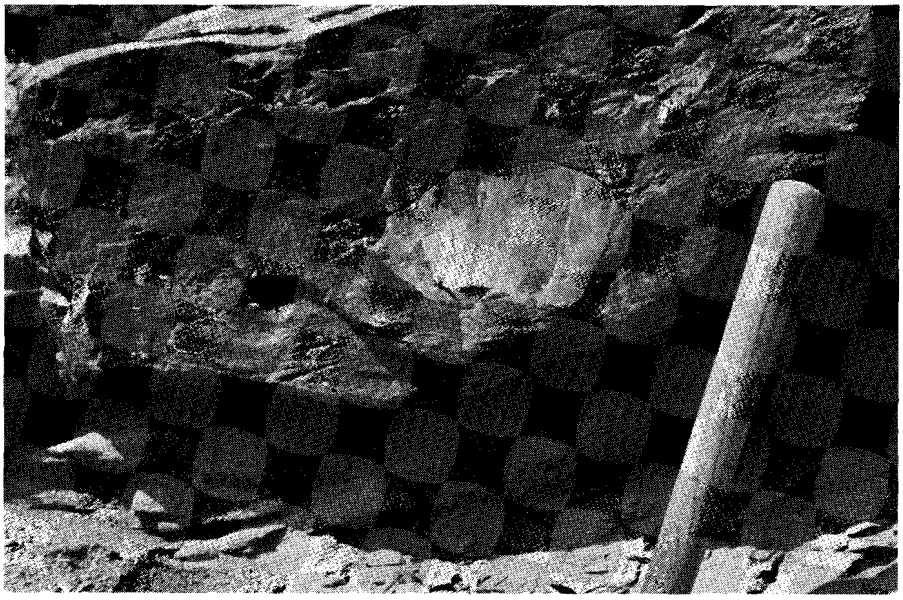

Photo 2. A Liguride limestone clast in a turbidite of the Marnoso Arenacea (Diaterna valley, 3 km N of Firenzuola): turbidite facies H.

Photo 3. Liguride clasts in the upper part of a turbidite of the Marnoso Arenacea (the same bed as on photo 2): turbidite facies H.

components may range from very small to very great. The polymict components can be recognized as Liguride, and are mostly concentrated in certain levels, very often also at the top of the layer. Grading is irregular, both normal and reverse grading occur. The matrix may be structureless or laminated (plane parallel, convolute, cross). Because of the irregularity of the sedimentary structures, the Bouma sequence is not applicable.

The turbidite facies as described above may combine to form larger facies associations. A purely descriptive classification is introduced and used here, for reasons that will be discussed later (section 3.3).

Facies association a: Composed mainly of the very coarse thick-bedded lenticular beds of facies A and B, but other facies may also occur. The coarser facies form lenticular bodies with a width of several hundreds of metres up to two kilometres, and a thickness of several metres up to several tens of metres. The base of these bodies is erosional. The sand/marl ratio is invariably high: from 1 to 10 (photo 4).

Facies association b: Megarhythms are distinct. Groups of thick turbidite beds, mainly facies B and C, alternate with groups of thinner turbidites, mainly facies D and C. Each group is at least ten metres thick. The sand/marl ratio of the thick bedded groups (2–5) is noticeably higher than in the thinner bedded groups ($\pm$ 1) (photo 5).

Facies association c: Megarhythms are absent. Thick bedded turbidites of facies C and thinner bedded turbidites of facies D are randomly distributed.

Photo 4. Thick massive sand beds as channel fill deposits in the Marnoso Arenacea (Fontanelice Member): facies association *a*.

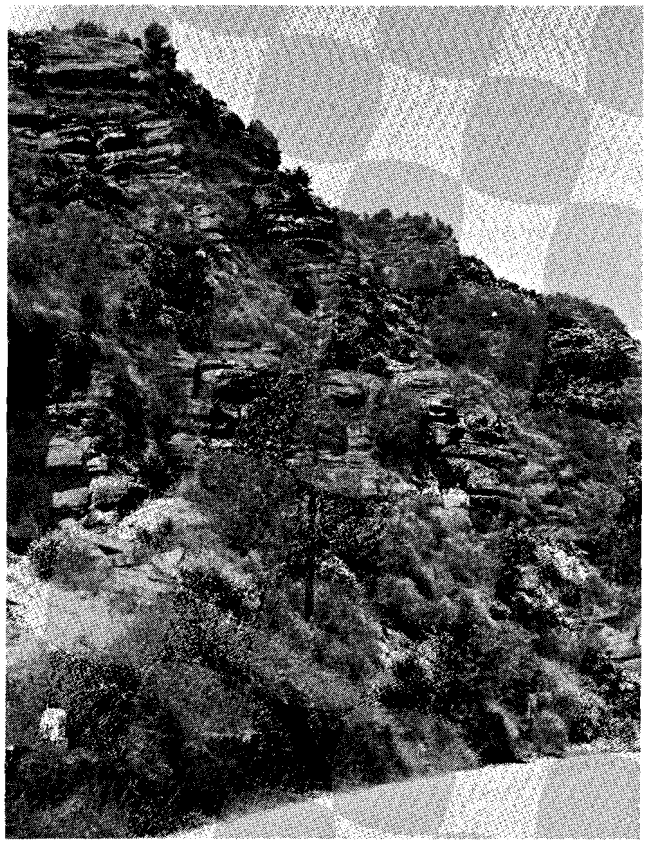

Photo 5. Megarhythms; groups of thick-bedded turbidites (facies B and C) alternate with groups of thin-bedded turbidites (facies C and D): facies association *b*.

The sand/marl ratio is approximately 1, but may vary between $^1/_2$ and $^3/_2$ (photo 6).

Facies association d: This facies association is also characterized by megarhythms. Groups of thick bedded turbidites, mainly facies C alternate with groups of thin bedded turbidites, mainly facies D. Each group is at least ten metres thick. The sand/marl ratio of the thick bedded groups is about 1–3, that of the thin bedded groups about $^1/_4$–1.

Facies association e: Thin bedded turbidites of facies D_1 and occasionally a turbidite of facies C. The marls of facies D_3 that occur in any facies association may become important here. The sand/marl ratio is low and varies between $^1/_5$ and 1. Often this facies association can be followed over great distances.

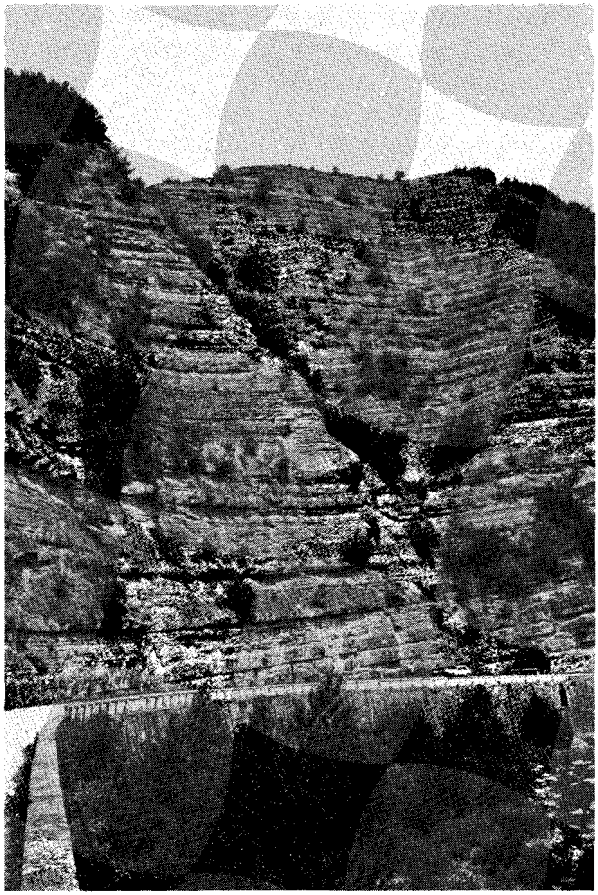

Photo 6. Thick- and thin-bedded turbidites (facies C and D) are randomly distributed: facies association *c*.

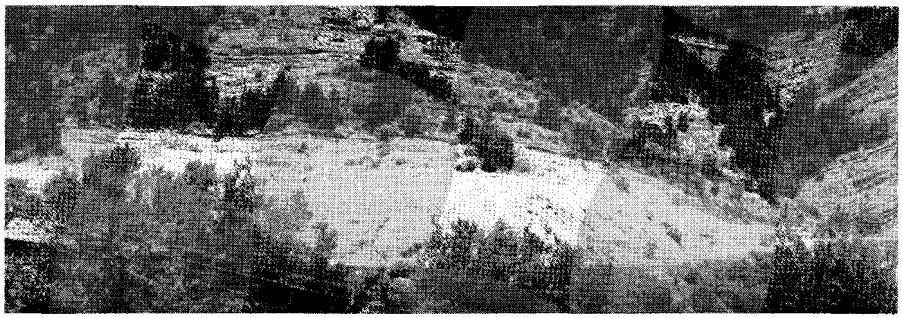

Photo 7. Marls of facies association *f* passing upwards rapidly into thick-bedded turbidites. (lower part of the Monte Coloreto Member near Casaglia).

Facies association f: A usually marly facies association. The bulk is composed of facies D_3 and D_1 in minor quantities. Intraformational slumps, facies F, may be important here. This facies association occurs only locally and has a lenticular form. Transitions to adjacent facies associations may be gradual as well as rather sudden. The sand/marl ratio is very low, less than 1. (photo 7)

Facies association g: Facies H is diagnostic. Usually this facies association is very marly (facies D_3), although locally more turbidites are present. All facies may occur. This facies association occurs only locally, and has a lenticular form. Transitions to adjacent facies associations may be gradual as well as rather sudden.

2.2. MARNOSO ARENACEA

2.2.1. General

The Marnoso Arenacea crops out on the Adriatic flank of the Northern Apennines in Umbria, Romagna, northern Marche and eastern Tuscany. The outcrops cover a band running NW-SE, 25 to 30 km wide and 180 km long. The base of the formation is not exposed in the area investigated. In the SE the formation overlies and grades to a calcareous-marly formation: the Bisciaro (Bortolotti et al. – 1970). In the NE the formation is stratigraphically overlain by the Gessoso Solfifera Formation. In the SW it is overthrust by the Castel Guerrino unit, and in the NW, along the 'Sillaro line' is the tectonically transgressive overthrust (section 4.6) by the Liguride Complex. The valley of the river Santerno is generally considered to be the type section of the Marnoso Arenacea. Other excellent outcrops can be found in the upper reaches of the river Senio.

The Marnoso Arenacea is composed mainly of alternations of calcareous sandstones and marls. Introducing the turbidite concept, Migliorini (1943, 1949, 1950) and Kuenen and Migliorini (1950) postulated that the Marnoso Arenacea is built up by resedimentated material transported by turbidity currents. At present we know that other sediment gravity flows such as fluidized sediment flows, grain flows and debris flows also played an important role (section 3.2). Recently the Marnoso Arenacea has been the subject of detailed sedimentological studies (see especially Ricci Lucchi – 1975a).

The Marnoso Arenacea sedimentation basin near the Sillaro line was divided into an inner and outer basin by a structural high. This structural high may be recognized at present in an anticlinal structure that runs from Castel Vecchio to Palazzuolo. SE of the Santerno this structure developed as a NE vergent reverse fault (and still further SE as an overthrust (Ten Haaf and Van Wamel – 1979)). The ancient high was high enough to prevent turbidity currents from flowing over it.

The Marnoso Arenacea ranks as a formation, but it is too voluminous and

variegated to be described as such. So we have split up the formation into 11 members. From top to bottom these are (see also fig. 3):
Borgo Tossignano Member
Fontanelice Member
Castel del Rio Member
Castel Vecchio Member
Monte Coloreto Member
Casaglia Member
Coniale Member
and furthermore we distinguish:
Piancaldoli Member
Ca Buraccia Member
Frena Member
Osteto Member
The last four members occupy a somewhat special position. They are lens-shaped bodies that occur at some places near the contact of the Marnoso Arenacea with the Liguride Complex.

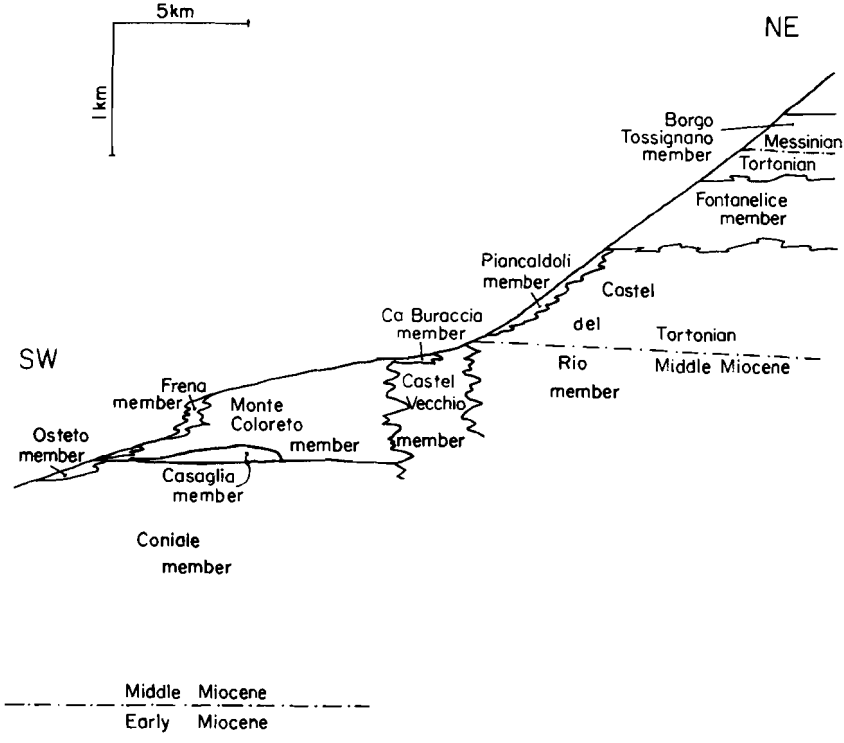

FIG. 3. Schematic palinspastic section showing the relations between the members distinguished within the Marnoso Arenacea.

The Marnoso Arenacea is usually considered to be an autochthonous sequence. However, there are some indications that more internal parts of the formation have undergone minor horizontal translations, and should therefore be called parautochthonous (TEN HAAF and VAN WAMEL – 1979).

Biostratigraphical work on the Marnoso Arenacea was done by CATI and

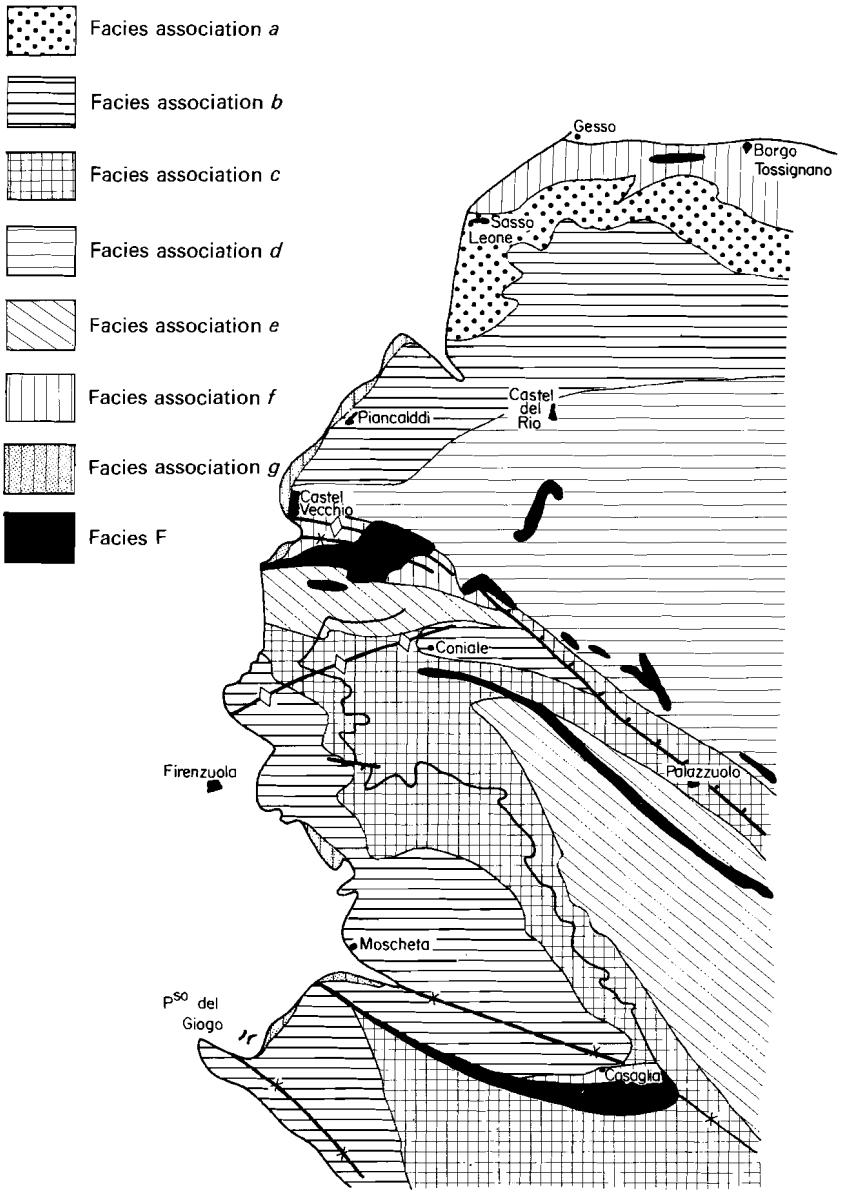

FIG. 4. Distribution of facies associations in the area investigated.

BORSETTI (1968) on several sections which are described sedimentologically by RICCI LUCCHI (1967). Two of these sections are situated in the area investigated here: the Sambucca section (SW of Palazzuolo) and the Castel del Rio section (along the river Santerno). Despite the scattered and ill-preserved state of the planktonic foraminifera Cati and Borsetti divided the formation into 6 cenozones. According to them the Orbulina date level (base of cenozone a Orbulina suturalis), which marks the beginning of the Langhian, is situated about 300 m below the intraformational slump that runs from Coniale to Acquadalto in the Coniale Member, and therefore in the lowermost deposits of the Marnoso Arenacea found in the area. This gives the base of the formation a Late Burdigalian age. The uppermost deposits of the Marnoso Arenacea, just below the Gessoso Solfifera Formation, have a Tortonian age according to CATI and BORSETTI (1968), but a Messinian age according to more recent work by RICCI LUCCHI (1977). So the Marnoso Arenacea can be dated as Late Burdigalian to Early Messinian. Assuming that the first appearance of Globigerina nepenthes (cenozone a Globigerinoides obliqua) marks the beginning of the Tortonian, then we can conclude that the base of the Tortonian is situated somewhere halfway along the Castel del Rio Member (CATI and BORSETTI – 1968) in the outer basin. No Tortonian deposits are known from the inner basin. The approximate positions of the boundaries are indicated in fig. 3. The Langhian-Serravallian boundary could not be located with any precision.

2.2.2. CONIALE MEMBER

The Coniale Member crops out as the oldest visible part of the Marnoso Arenacea in the area studied; accordingly the lower boundary is not known.

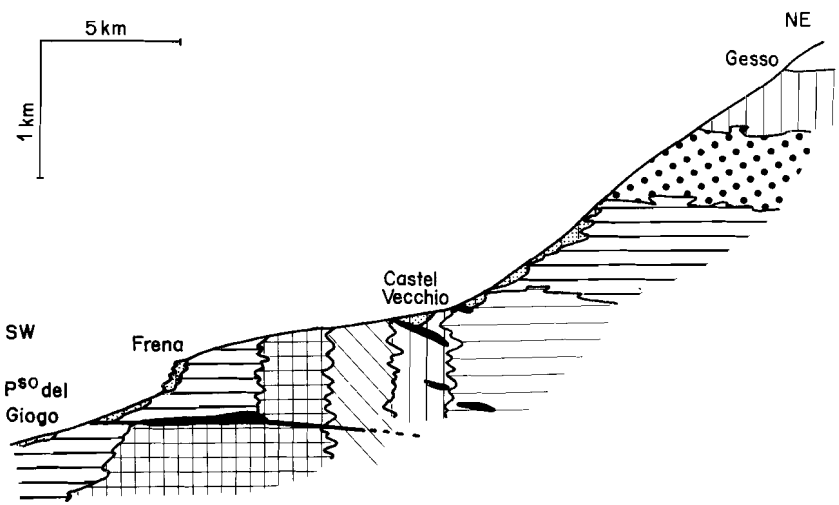

FIG. 5. Schematic palinspastic section through the Marnoso Arenacea showing the relations between the facies associations (see for legend fig. 4).

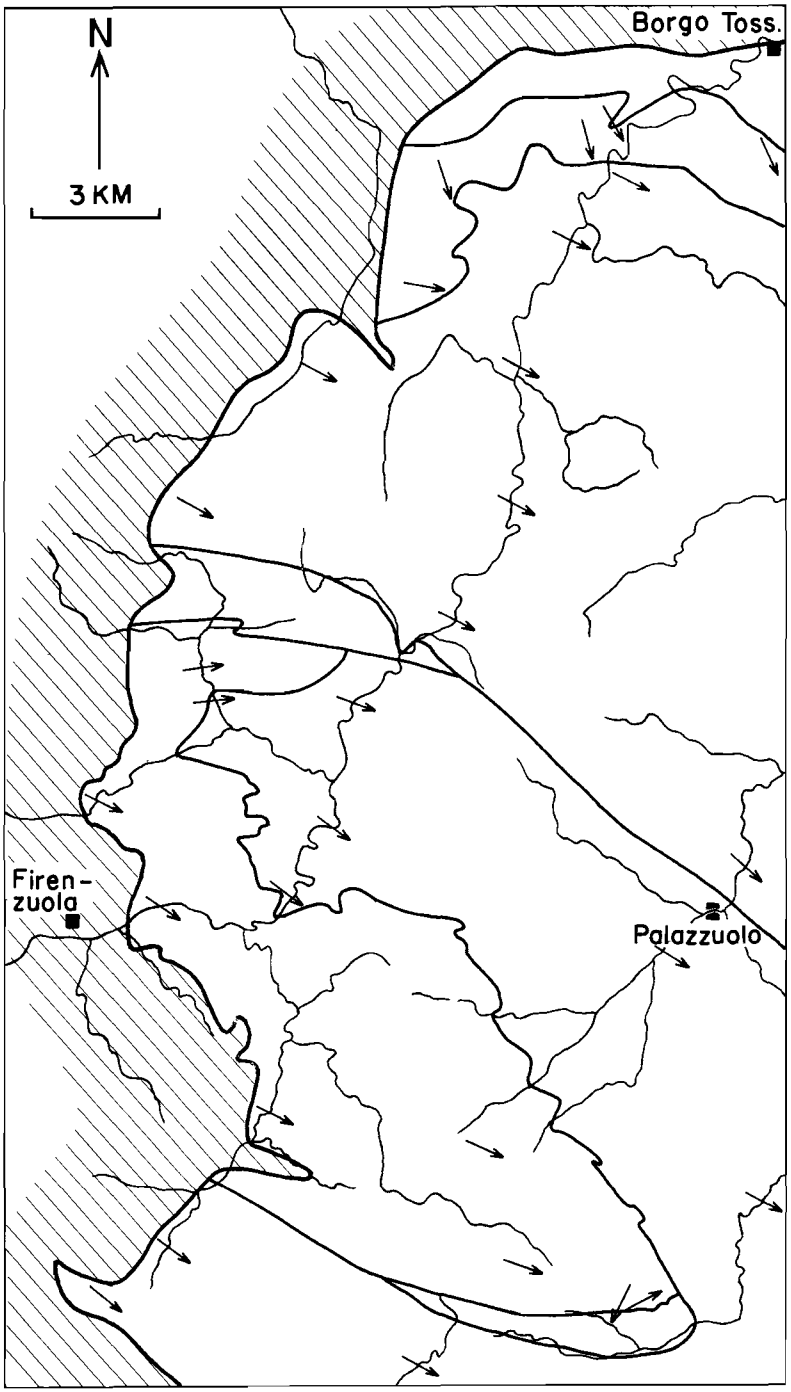

FIG. 6. Paleocurrents in the Marnoso Arenacea as measured on flute casts.

As upper boundary the top of an extensive intraformational slump running along San Pellegrino and Casaglia (the Casaglia slump, photo 8) has been chosen. In the SW, between Osteto and the P^so del Giogo, the Coniale Member partly interfingers with the Osteto Member. In the NE the Coniale Member is separated from the Castel del Rio Member by a reverse fault.

Important lateral variations occur in the Coniale Member (fig. 4, 5). Two trends interact:

1. A fining towards the NE, especially in the upper part of the member from facies association *b* to *e*.
2. A fining towards the SE, in the SW from facies association *b* to *c*, in the NE from facies association *c* to *e*.

Both the weathered and the fresh surface of the sandstones and marlstones have an ash-grey colour. The weathered colour of the sandstones may also sometimes be somewhat brownish. The marls are mostly rather calcareous.

The direction of supply of the turbidites of the Coniale Member generally do not deviate much from the average direction of the Marnoso Arenacea turbidites: 130° downcurrent. Near San Pellegrino in the Santerno valley a somewhat higher value is measured, i.e. 150°. In the uppermost parts of the member, W of Coniale, a value of 90° has been measured. If this value is corrected for an axial dip of about 15° towards the NW due to the Sillaro flexure (section 4.2), the direction of supply is 80° (fig. 6).

Ricci Lucchi (1975a) describes a turbidite with an opposite direction of

Photo 8. A large intraformational slump in the Marnoso Arenacea at the top of the Coniale Member: the Casaglia slump.

supply near Coniale. This would be his Contessa key bed: a calcarenite supplied from the SE with pale calcareous marls on top of it. The marls do indeed have a different colour from the other marls, but they are less, not more calcareous. The turbidite cannot be described as a calcarenite either, nor does it have an opposite direction of supply.

Two major intraformational slumps are present in the Coniale Member; one 40 m thick slump in the lower part, between Coniale and Acquadalto, with an outcrop length of 15 km, and another at the top of the member, with a maximum exposed outcrop length in the direction of the Apenninic strike of 20 km, the Casaglia slump (photo 8). The maximum thickness of the latter has been measured at 1 km E of Casaglia and found to be 60 m. Elsewhere the thickness is 30 to 50 m. N of San Pellegrino the thickness diminishes, until NW of Coniale the slump has disappeared. The real original extensions of both slumps are probably much greater. Fold axes are mostly orientated in a NW-SE direction, but deviations up to 90° occur.

The thickness of the Coniale Member is at least 3000 m. The thickness between the two above mentioned intraformational slumps between Acquadalto and Casaglia is about 2200 m, while N of Coniale the thickness between these slumps has diminished to about 1000 m.

2.2.3. Casaglia Member

The Casaglia Member is found at the top of the Coniale Member near Casaglia, and it underlies the Monte Coloreto Member. The member has a lenticular form.

The main part consists of light coloured calcareous marls, affected by strong fracture cleavage and containing many calcite veins (photo 9). Locally within these marls lenses with a different lithology are found: dark shales and marls alternating with dense limestones, siltstones and fine-grained sandstones. The thickness of these layers is usually some 10 to 20 cm. The weathered colour of the limestones is light yellowish or white, the fresh colour is dark grey. Both the weathered and the fresh colour of the siltstones and sandstones is dark grey or brown. The lenses are heavily tectonized. The maximum thickness of such lenses is 20 m. Breuren (1941) described some small appearances of these rocks NW of Casaglia lying directly upon the Casaglia slump and under the Monte Coloreto Member.

The maximum thickness of the Casaglia Member is 150 m, and its length several km.

The marls of the Casaglia Member contain Orbulina sp. This makes them at their oldest Middle Miocene. Groscurth (1971) found some Globorotalia in the limestones of the shaly lenses which point to a Paleocene-Eocene age.

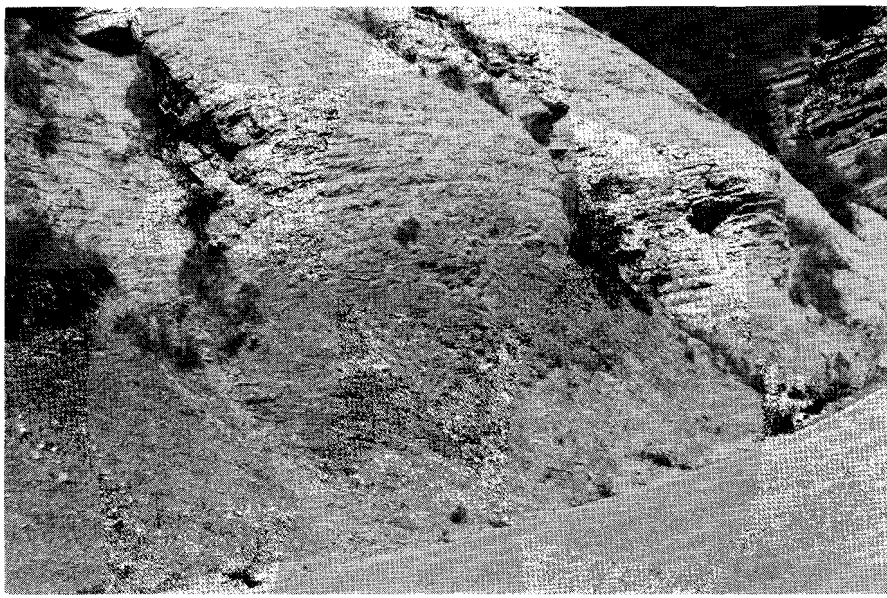

Photo 9. The contact of the strongly tectonized marls of the Casaglia Member (below) with the marls (facies association *f*) of the Monte Coloreto Member.

2.2.4. Monte Coloreto Member

The Monte Coloreto Member is deposited upon the Coniale Member, and in the S upon the Casaglia Member. In the N there is a lateral transition to the Castel Vecchio Member, near Osteto the Monte Coloreto Member interfingers with the Osteto Member, and near Frena with the Frena Member.

A general fining can be observed in the Monte Coloreto Member towards the NE, from facies association *b* via *c* to *e* (fig. 4, 5). Many intraformational slump phenomena are present in facies association *e* in the N of this member. At the base of the Monte Coloreto Member above the Casaglia Member a maximum of 30 m of facies association *f* is present, passing upwards quickly into facies association *b* (photo 7). To the E and N the thickness of facies association *f* diminishes. Petrographically the sandstones and marls of the Monte Coloreto Member are indistinguishable from those of the Coniale Member.

The direction of supply of the turbidites is in most places about 130°. Near Casaglia, however, in facies association *f* the few directions of supply that could be measured strongly deviated from that value. A different value was also measured in the northern-most part of this member. There the direction of supply is, corresponding to the directions of the Coniale Member just underneath it, about 90°, and corrected for the Sillaro dip: 80° (fig. 6).

The thickness of the Monte Coloreto Member varies greatly. Measured along the river Santerno, the thickness is 750 m, a thickness that also holds for the thickness in the N. Near Osteto, however, the thickness is only some tens of metres, diminishing to zero to the W, and increasing rapidly to 500 m to the E.

2.2.5. Castel Vecchio Member

The Castel Vecchio Member crops out only in a limited area, N of the Monte Coloreto Member. There are lateral transitions to both the Monte Coloreto Member in the S and to the Castel del Rio Member in the N. In the W the member interfingers with the Ca Buraccia Member. E of the river Santerno the Castel Vecchio Member disappears under the upthrust of the Coniale Member. N of the Castel Vecchio-Palazzuolo structure between the Santerno and the Senio in the Castel del Rio Member, some slumps are found that belong to the Castel Vecchio Member too.

The deposits of the Castel Vecchio Member have usually been considered to be the equivalent of the Borgo Tossignano Member, the top of the Marnoso Arenacea. Only Bruni (1972) described the Castel Vecchio deposits as occurring within the Marnoso Arenacea and having lateral transitions to the N (the Castel del Rio Member). Bruni called these deposits the Diaterna Member.

The Castel Vecchio Member, which exhibits many lateral facies variations, is made up mainly of facies association *f*. In the upper part, SW of Castel Vecchio, some highly bioturbate lenticular marly sandstones occur, containing moluscs. These moluscs are identified by Breuren (1941) and by Ricci Lucchi and Veggiani (1967) as Lucina. The lower part of the member, especially E of the river Santerno, is more calcareous than the upper part. Calcareous marls and limestones occur there. In this part, but also near Visignano, lenses are found where thin (up to 5 cm) siltstones alternate with grey marls. The fresh colour of the siltstones is dark-grey, the weathered colour is rusty. Internally the siltstone beds are almost completely cross laminated.

To the S the lateral transition to the Monte Coloreto Member is gradual, as more and thicker turbidites occur. The boundary is chosen where facies association *f* passes into facies association *e*, approximately where the sandstone/marl ratio becomes greater than $^1/_5$. To the N the lateral transition to the Castel del Rio Member is more abrupt. Over a short distance, maximum 300 m, facies association *f* of the Castel Vecchio Member passes into facies association *d* of the Castel del Rio Member. W of Castel Vecchio a 30 m thick intraformational slump can be observed near the transition to the Castel del Rio Member. Another 50 m thick intraformational slump can be seen in the Santerno valley near the transition to the Castel del Rio Member. Another 50 m thick intraformational slump is associated with an olistostrome consisting of Liguride material, which lies directly upon it. The olistostrome has a maximum thickness of 150 m. Its lateral extension is small.

The slumps intercalated in the Castel del Rio Member that belong to the Castel Vecchio Member, consist mainly of marls, and contain locally Lucinae.

The thickness of the Castel Vecchio Member is hard to estimate, as the lower boundary is not known. The maximum exposed thickness is 500 m. The lateral transitions to the adjacent members is coupled with an increase in thickness, obviously due to the appearance of more and thicker turbidites.

2.2.6. Castel del Rio Member

The Castel del Rio Member crops out N of the line Castel Vecchio to Palazzuolo. The lower part of this member is lateral-transitional to the Castel Vecchio Member. E of the river Santerno the Coniale Member is separated from the Castel del Rio Member by a reverse fault. In the N the Fontanelice Member overlies the Castel del Rio Member, and in the W, along the T. Sillaro, the latter member is lateral-transitional to the Piancaldoli Member.

The whole Castel del Rio Member is composed of megarhythms. Upwards there is a decrease in marl content of $^1/_3$ to 1 in the lower part up to 4 to 5 in the upper part. The lower part is classified as facies association *d*, the upper part as facies association *b*. The fresh colours of the sandstones and marls are ash-grey. The weathered colour of the marls is also ash-grey, that of the sandstones may be ash-grey, but is also often somewhat brownish. The marls are clearly less calcareous than those of the members mentioned before.

The lateral transition in the S to the Castel Vecchio Member is rather abrupt. Within a few hundred metres the marl content increases and the amount and thickness of the turbidites decrease until facies association *f* is formed. The boundary in the W with the Piancaldoli Member is chosen at the point where facies association *b* passes into facies association *g*.

The direction of supply of the turbidites as measured on the bottom features is in the whole member constantly SE (120°) (fig. 6).

The thickness of the Castel del Rio Member, measured along the river Santerno, is 2250 m. Because the base of the member is not exposed, the real thickness will be greater.

2.2.7. Fontanelice Member

The Fontanelice Member overlies the Castel del Rio Member, and is in turn overlain by the Borgo Tossignano Member.

The Fontanelice Member is made up entirely of facies association *a*. Three conglomerate lenses occur in this member. The largest has a length of 1 km, the two others are several hundred metres long. The thickness of the lenses is about 10 m. The conglomerates are polymict. Almost 50% of the pebbles

consists of igneous and metamorphic rocks: granite, granodiorite, gneiss, quartz porphyry, quartzite. The remainder consists of sedimentary rocks: limestone, dolomite, sandstone, chert, etc. Almost all these rock types are unknown as Apenninic formations (RICCI LUCCHI – 1969b). The pebbles are well rounded and usually about 2–5 cm across. Practically all pebbles are imbricated to the SE (150°–160°). In the marls just below the largest conglomerate lens several deep U-shaped burrows have been found. The thicker sandstones of the Fontanelice Member are often very poorly and irregularly cemented.

Lateral facies changes in the Fontanelice Member are numerous. E of the river Santerno the thicknesses of the sandstones and sandstone/marl ratio diminish.

Flute and groove casts are less often present than in most of the other members. If present they show directions of supply from a more northernly direction (towards 160°–170°) (fig. 6).

The thickness of the Fontanelice Member is about 500 m. E of the Santerno the thickness diminishes until eventually the member pinches out completely.

2.2.8. BORGO TOSSIGNANO MEMBER

The Borgo Tossignano Member overlies the Fontanelice Member, and is overlain by the Gessoso Solfifera Formation.

The Borgo Tossignano Member is composed almost entirely of light coloured clays with occasional thin turbidite intercalations. The deposits are classified as facies association *f*. In the lower part of the member two olistolithes occur. They are composed of coarse thick sandstones identical to those of the underlying Fontanelice Member. The maximum dimensions of the olistolithes are 75 by 30 m. The olistolithes, which occur on the same stratigraphic level, are embedded in a slump with a thickness of 20 m and a maximum observable length of 500 m.

The thickness of the Borgo Tossignano Member is about 500 m, and varies very little in the studied area. To the E, however, the thickness diminishes to zero.

2.2.9. OSTETO MEMBER

The Osteto Member is found lateral to the Coniale and the Monte Coloreto Members in the S of the investigated area between the P^{so} del Giogo and Osteto.

The Osteto Member is made up entirely of a very marly facies association *g*. Lateral to the Monte Coloreto Member the Osteto Member contains several

black chert lenses, with a maximum thickness of 5 cm and a maximum length of 3 m. Also some coal lenses several dm thick and up to 3 m long are present. Upwards the Osteto Member acquires a more slumped character. Some olistolithes of ash-grey sandstones occur, with a maximum volume of 10 m^3.

The transition of the Coniale and Monte Coloreto Members to the Osteto Member is gradual. The boundary is chosen where the sandstone/marl ratio becomes less than $^1/_5$.

The maximum thickness of the Osteto Member is 50 m.

Because of the partly slumped character of the Osteto Member, one might wonder whether or not these deposits are autochthonous. Because the lower contact of the member is sedimentary, and the member becomes gradually more disturbed upwards, one is probably justified in concluding that these deposits are at most parautochthonous. Besides, the age of the Osteto Member does not differ demonstrably from the ages of the adjacent members, as they all contain Orbulina suturalis (GROSCURTH – 1971), which makes them Middle Miocene or younger.

2.2.10. FRENA MEMBER

Deposits of the Frena Member crop out near Frena, SW of Firenzuola. The Frena Member is a lenticular body, interfingering in the N with the Monte Coloreto Member, and in the S with the Liguride Complex. Other deposits that belong to this member occur as lenses in the Liguride Chaotic Complex (fig. 7 and 8).

The Frena Member consists mainly of rather calcareous, blue-grey marls. Intercalated in the marls are sandlayers of facies H as well as marls of facies H (fig. 7 and 8). The deposits are classified as facies association *g*.

The maximum observed thickness is about 100 m.

2.2.11. CA BURACCIA MEMBER

The Ca Buraccia Member is a lenticular body, grading laterally to the Castel Vecchio Member. This member is named after a small village about 2 km SW of Castel Vecchio.

The Ca Buraccia Member does not differ much from the Castel Vecchio Member; both members are very marly. The Ca Buraccia Member however is made up of facies association *g*, and the Castel Vecchio Member of facies association *f*.

The thickness of this member does not exceed 10 m.

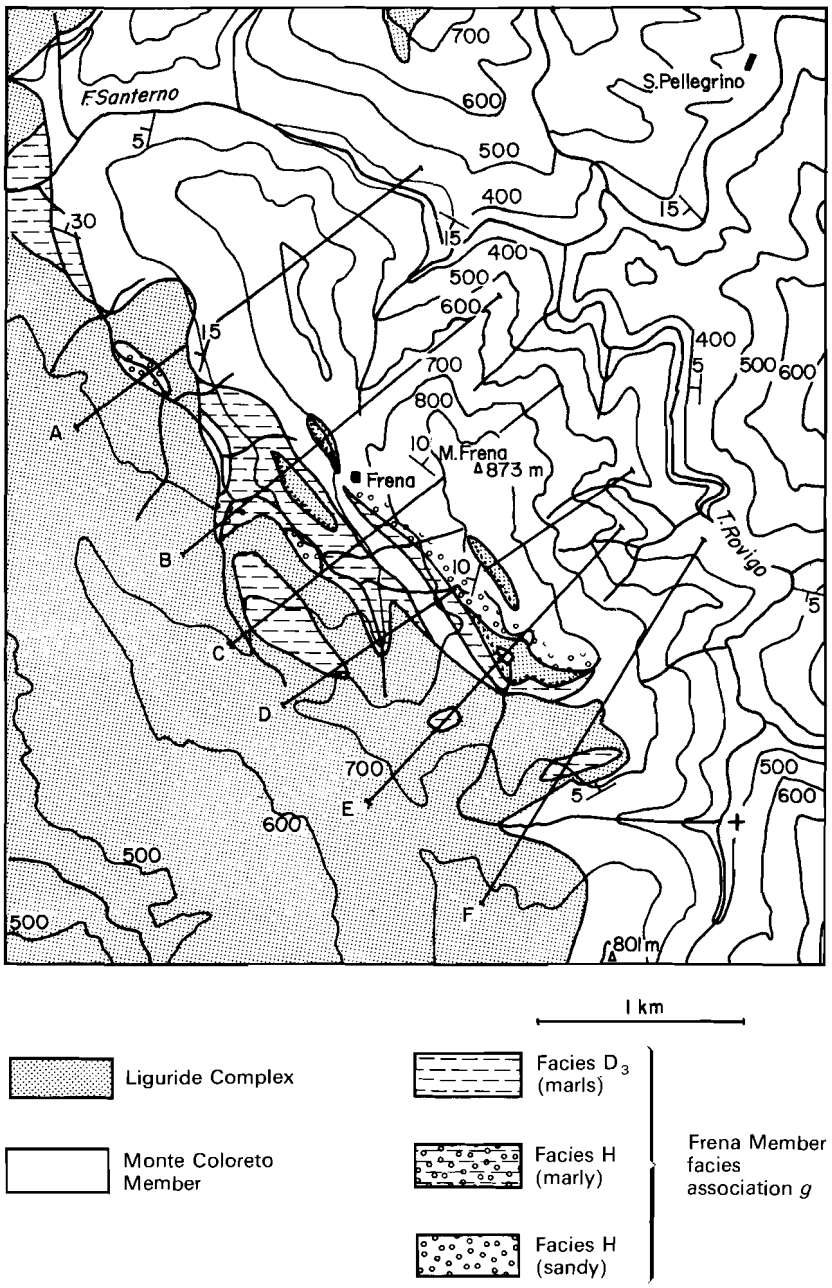

FIG. 7. Detailed map of the Frena area.

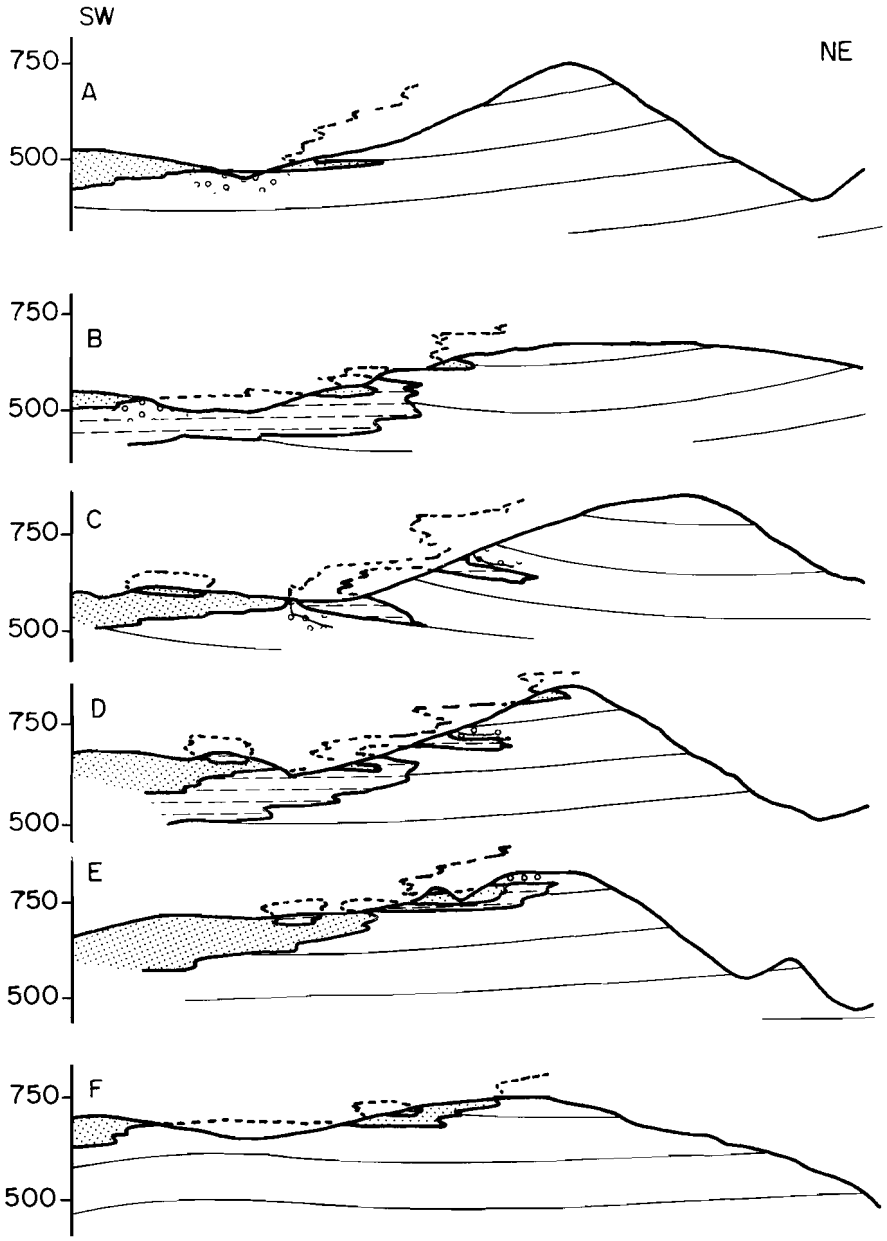

FIG. 8. Profiles belonging to fig. 7, showing the interfingering of the Liguride Complex with the Marnoso Arenacea.

2.2.12. Piancaldoli Member

The Piancaldoli Member is found almost everywhere between the Castel del Rio Member and the Liguride Complex. It grades laterally into the Castel del Rio Member.

The Piancaldoli Member consists mainly of a very marly facies association *g*. The colour of the marls is blue-grey. The boundary between the Castel del Rio Member and the Piancaldoli Member is chosen where facies association *b* and *d* of the Castel del Rio Member pass into facies association *g* of the Piancaldoli Member (at a sandstone/marl ratio of approximately $^1/_5$). Just N of the Castel Vecchio Member some slumped marls can be found.

The maximum thickness of the Piancaldoli Member is 40 m. The outcrop length is about 6 km.

2.2.13. Basin analyses

Crucial to the definition of the basin geometry of the Marnoso Arenacea basin is the interpretation of the depositional environment of the marly deposits of the Castel Vecchio Member (facies association *f*). These deposits are interpreted here as intrabasinal high deposits. The Castel Vecchio Member has lateral transitions to both the Castel Del Rio Member in the N, and to the Monte Coloreto Member in the S. The transition to these members is accompanied by a significant increase in the thickness of time-equivalent parts of these members. Also the number and the mean thickness of the turbidites increase considerably away from the Castel Vecchio Member. Obviously the sedimentation rate of the Castel Vecchio Member was lower than that of the adjacent members. The fact that only very few turbidites were deposited supports the view that the Castel Vecchio area was a high during the sedimentation in the Marnoso Arenacea basin. Turbidity currents could not surmount the high, with the result that only hemipelagic marls and the dilute upper parts of some turbidity currents could be deposited there. Because the sedimentation rate on an intrabasinal high is lower than elsewhere in the basin, after a certain time the high will cease to exist. The intrabasinal high deposits of the Castel Vecchio Member have a thickness of at least 500 m. This implies that some forces must have maintained the high in face of the influence of the faster sedimentation in the adjacent parts of the basin. The Castel Vecchio Member is situated on an anticline: the Castel Vecchio structure. This seems to justify the assumption that the Castel Vecchio high developed as a synsedimentary anticline. To the SE this anticline passes into a thrust (see section 4.3.2), under which the Castel Vecchio Member disappears. There is however an indication that the Castel Vecchio deposits (facies association *f*) extend further to the SE. NE of the upthrust, in the Castel del Rio Member, some slumps are found which consist of very marly rocks, locally containing molusces

of the genus Lucina. Because of the great resemblance of these rocks to the deposits of the Castel Vecchio Member, the slumps are probably descended from a continuation of the Castel Vecchio high there. The Castel Vecchio high thus divided the Marnoso Arenacea basin into an outer north-eastern and an inner south-western basin, which will be treated separately.

Sedimentation in the Marnoso Arenacea basin near the Sillaro line stopped because of the arrival of the Liguride Complex. The Liguride Complex, which overthrust the Marnoso Arenacea from SW to NE. Consequently, sedimentation stopped earlier in the SW than in the NE. The sediments of the inner basin (Coniale and Monte Coloreto Member) are as a whole older than those of the outer basin (Castel del Rio, Fontanelice and Borgo Tossignano Member).

Inner basin

The inner basin deposits comprise the Coniale, Casaglia, Monte Coloreto, Osteto and Frena Member. The presence of the Rovigo syncline and the Diaterna anticline means that the members crop out more than once in a direction perpendicular to the main Apenninic strike. This provides good opportunities to study lateral facies changes. The Casaglia slump is considered to have originated instantaneously, geologically speaking, so the upper boundary of the slump can be used as a synchronous marker.

The Casaglia Member is probably best described as a 'klippe sédimentaire'. The marls of this member are highly tectonized, and parts of it (the shaly intercalations) are older than the rocks in which the Casaglia Member is embedded. The origin of the rocks is uncertain. The appearance of the marls is very similar to the Vicchio Marls of the Tuscan Sequence, but the latter are usually considered older. The Paleocene-Eocene age of the shaly intercalations means that these may belong to the sub-Liguride Canetolo Complex, to which they bear some resemblance.

In the Monte Coloreto Member as well as in the upper part of the Coniale Member transversal facies variations can be seen to pass from facies association *b* or *c* in the SW to facies association *e* just S of the Castel Vecchio area (fig. 4 and 5). So towards the Castel Vecchio high the turbidites become thinner and finer grained, and the sandstone/marl ratio decreases. Facies association *e* is characterized by many small intraformational slump phenomena, indicating a somewhat inclined depositional surface. This is most probably the result of the presence of the Castel Vecchio high, which therefore must have existed already during the deposition of the upper part of the Coniale Member. On the other hand, the presence of facies association *b* near Coniale in the lowermost deposits of the Coniale Member (fig. 4) does not necessarily indicate that the Castel Vecchio high did not exist at the time of their sedimentation. This part of the formation has been thrust towards the NE (section 4.3.2), so the original position of the rocks was more to the SW.

Not far to the NE of the line Firenzuola-Casaglia a lateral transition from facies association *b* to *c* can be seen. Probably here the basin floor started to rise towards the Castel Vecchio high. An additional indication for this may be

that the Casaglia Member did not slide beyond that line. The position of these extraformational elements, and of facies association *b* roughly coincides with the axis of the Rovigo syncline, which also developed synsedimentarily. The coarse thick-bedded facies association *b* was in that case deposited in the deepest part of the basin, and the extraformational Casaglia Member came to a standstill near the basin axis. The intraformational Casaglia slump peters out towards the Castel Vecchio high. It is difficult to imagine that this slump could mount this high. A possible explanation is that this slump represents no more than the disturbed sediments of the basin floor after a severe earthquake. The preferential NW-SE axis of the slump-folds indicates than that any transport that did occur was perpendicular to the basin axis.

Along the strike, the upper part of the Coniale Member on the southern limb of the Rovigo syncline shows a lateral transition towards the SE from facies association *b* to *c*. The middle part of the Coniale Member on the northern limb of the Rovigo syncline shows a lateral transition in the same direction, from facies association *c* to *e*. As the direction of supply of the turbidites was from the NW, these represent a transition from more proximal to more distal sediments.

On top of the Casaglia sedimentary klippe 30 m of marls were deposited before the first turbidites occurred. Obviously at first the turbidity currents could not flow over the 150 m high rock sticking up out of the sea floor. As a result only marls were deposited on top of the klippe. These marls may in part be hemipelagic, but are probably also in part deposited from the uppermost dilute part of some turbidity currents. Only after the faster sedimentation around the sedimentary klippe had levelled out the differences in height could the sand carrying parts of the turbidity currents spread over the slump. These first turbidites however were still influenced by the basin floor irregularities, as can be deduced from the deviating directions measured on the bottom features.

Outer basin

The outer basin deposits comprise the Castel del Rio, Fontanelice, Borgo Tossignano and Piancaldoli Member. The beds are almost everywhere gently inclined N to NE. The absence of anticlines and synclines makes it impossible to study facies changes in a direction perpendicular to the main Apenninic strike.

The outer basin deposits, except for the Borgo Tossignano Member, exhibit a coarsening upwards from facies association *d* via *b* to *a*. Whether this coarsening upwards is also a lateral coarsening to the N (away from the Castel Vecchio high) cannot be decided because of the monoclinal attitude. Facies association *a* of the Fontanelice Member has been interpreted as consisting of channel fill sequences (Ricci Lucchi – 1968). The channels were broad (1–1.5 km) and shallow (some tens of metres), as can be deduced from the broad bodies consisting of turbidite facies A and B. On top of these sequences the marly deposits of the Borgo Tossignano Member (facies association *f*) were

deposited. The presence of two large olistolithes indicates that the depositional surface was inclined. The association with the channel fill sequences of the Fontanelice Member indicates that the Borgo Tossignano deposits must be considered as basin slope deposits.

Osteto, Frena, Ca Buraccia and Piancaldoli Member

As stated before, the Liguride Complex overrode the Marnoso Arenacea synsedimentary from SW to NE. Sedimentation was stopped by the arrival of the Liguride Complex. From SW to NE the Liguride Complex overlies progressively younger layers. Autochthonous sedimentation near the front of the advancing Liguride Complex was sometimes influenced by the Liguride Complex (fig. 12 and 15). This is reflected in the deposition of the Osteto, Frena, Ca Buraccia and Piancaldoli Member. These members are marly and in direct contact with the Liguride Complex. Except for the Osteto Member they also contain turbidite facies H, with a depositional mechanism as will be described in section 3.2. The marly development, or rather the reduced presence of turbidites, is probably the result of advanced lobes of the Liguride Complex that locally protected the depositional environment from turbidity currents (fig. 15). Usually the thickness of the members does not exceed 50 m. The Frena Member however probably has a thickness of more than 100 m. Besides this member interfingers with the Liguride Complex. This is best explained if we assume that the Liguride Complex temporarily did not advance at all or advanced very slowly. It is striking that this interfingering occurs in the Rovigo syncline, which has already been interpreted as a synsedimentary structure. The Liguride Complex, just like the extraformational Casaglia Member, obviously could not at first mount the northern limb of the Rovigo syncline. Only after some time, after a sufficient thickness had accumulated, did the Liguride Complex overthrust further to the NE (see section 4.6). The Marnoso Arenacea basin became very narrow, the more the Liguride Complex advanced. All turbidity currents were forced to follow a fairly restricted path. This may very well have caused the coarsening upwards that is observed in the deposits of the Castel del Rio Member.

Thickness and sedimentation rate of the Marnoso Arenacea

The total thickness of the Marnoso Arenacea is rather difficult to estimate. Individual members can be measured, but it would be a misrepresentation to state, as is often done, that the total thickness is the sum of the member thicknesses. Firstly, the members exhibit lateral transitions to each other. Secondly, the thicknesses of the members are liable to variations. The Marnoso Arenacea basin can be divided into an inner and an outer basin, each with its own sedimentary evolution. These two parts should be treated separately.

In the inner basin the Coniale and Monte Coloreto Member have been deposited. The sum of the maximum thicknesses of these two members is 3750 m. It should be borne in mind, however, that the lower boundary of the Marnoso Arenacea is not known, so the real thickness is greater.

In the outer basin the Castel del Rio, Fontanelice and Borgo Tossignano Members have been deposited. The sum of the maximum thicknesses of these members is 3250 m. This is the thickness as measured along the river Santerno from the Castel Vecchio-Palazzuolo structure to Borgo Tossignano. The thickness of the deposits will however diminish to the S (to the Castel Vecchio high) as well as to the N (to the edge of the basin). In the outer basin the lower boundary of the deposits is not known either. So no figure can be given, but the thickness must be variable on account of the synsedimentary tectonics, and is in the order of at least several kilometres in some places.

As the Orbulina date level (dated at 15–15.5 million years, IKEBE – 1978) is situated about 300 m below the intraformational slump running from Coniale to Acquadalto in the Coniale Member (CATI and BORSETTI – 1968, RICCI LUCCHI – 1967), the total thickness of the inner basin deposits above the Orbulina date level is about 3250 m. No late Miocene, which started 11.4–11.7 million years ago, deposits are known from the inner basin. So 3250 m of marls and sand were deposited during the Middle Miocene which lasted 3.2–4.1 million years. This implies that the mean sedimentation rate in the inner basin was probably about 0.70 m per thousand years. The maximum thickness of the outer basin deposits is 3250 m. Their sedimentation started during the Middle Miocene (more than 11.4–11.7 million years ago) and lasted until the Messinian (about 6 million years ago). So the mean sedimentation rate must have been lower than 0.60 m per thousand years, and was probably about 0.50 m per thousand years.

Depth of the Marnoso Arenacea basin

Usually a depth of 1000 to 2000 m is taken for granted for the Marnoso Arenacea basin. The most important argument for this seems to be that recent turbidite sedimentation on the oceanic abyssal plains also occurred and is still occurring at or beyond these depths. But the narrow, geosynclinal Marnoso Arenacea basin cannot be compared with oceanic environments. This makes it necessary to critically review all arguments concerning the depth of the basin.

Turbidity currents are born from accelerating sediment gravity flows. Whether a flow can become turbulent or not depends on the sediment involved, and on the length and inclination of the slope. It is difficult to evaluate these factors, but it is beyond doubt that the deeper the basin, the better the possibilities for the creation of extensive turbidite formations such as the Marnoso Arenacea. Recent turbidite sedimentation in lakes and fjords, however, shows that turbidity currents can originate at a depth of only some tens of metres. GOULD (1951) observed turbidity currents that travelled distances of up to 120 miles in lake Mead, at a depth of 460 feet. So a much shallower depth than the usually adopted 1000 to 2000 m seems possible.

A direct indication for the depth of the basin is supplied by the occurrence of moluscs (Lucina) in the Castel Vecchio Member. These moluscs live in very shallow water (RICCI LUCCHI and VEGGIANI – 1967). The Castel Vecchio deposits represent an intrabasinal high. This makes it impossible for the moluscs

to have been transported and deposited by means of sediment gravity flow. Also the moluscs are found in living position. This implies that occasionally the crest of the high reached littoral depth. The inclination of submarine slopes in environments with a high sedimentation rate is usually not more than 5°; steeper slopes are destroyed by slumps which can arise on submarine inclinations of as little as 1°–4° (LEWIS – 1971). As slumping did take place on the NE slope of the high, the inclination there was probably about 5°. This slope probably was steeper than the SW slope, since the Castel Vecchio-Palazzuolo structure is, and probably was, NE-vergent. If the SW slope was inclined about 3°, and if this inclination was constant from Castel Vecchio area to the hinge of the Rovigo syncline, which formed the basin axis, then the depth near the basin axis was about 500 m. Of course the depth of the Marnoso Arenacea basin did not remain constant. Sedimentation and subsidence balanced each other approximately, but it cannot be supposed that they always kept pace with each other. It was probably only an exception that the crest of the Castel Vecchio high reached the littoral zone.

Shallow water moluscs are also found in other places in the Marnoso Arenacea. RICCI LUCCHI and VEGGIANI (1967) describe a number of places where they found 'calcari a Lucina'. The question is how can these occurrences be explained. Ricci Lucchi and Veggiani state that the deposits 'a Lucina' originated as a result of repeated gravity-controlled resedimentation processes. Submarine mass transports started in the littoral area, successively reaching the deepest parts of the basin, transverse to its axis (i.e. from SW to NE). As we have seen however, not all 'a Lucina' deposits are slumps. But if we assume that the other occurrences are indeed slumps, then the locations can be arranged in the following 4 groups (fig. 9).

1. The marly deposits associated with the Castel Vecchio-Palazzuolo structure.
2. Scattered blocks below the overthrust plane of the Liguride Complex.
3. Occurrences along the Apenninic divide.
4. The Tortonian inner fan and slope deposits.

The Lucinae of group 1 lived on the Castel Vecchio high and are still in situ or slumped from the steep NE side of the high.

Because the Marnoso Arenacea was not always very deep, top parts of the overthrusting Liguride Complex may have reached a depth were moluscs could live. Slumps brought these deposits into the Marnoso Arenacea basin, where a little later they were overridden by the Liguride Complex. This explains the deposits of group 2.

Lucina limestones of group 3 are found near the Apenninic divide. Two other occurrences of this group are in two major slump horizons near Casaglia and Acquadalto. They can be related in an analogous manner to former shallow water deposits on top of an overthrusting unit – in this case probably the Tuscan and Castel Guerrino unit.

The deposits of group 4 are slope deposits and channel fill sequences. The slopes must have been directed SW (see also RICCI LUCCHI – 1975a). This is not

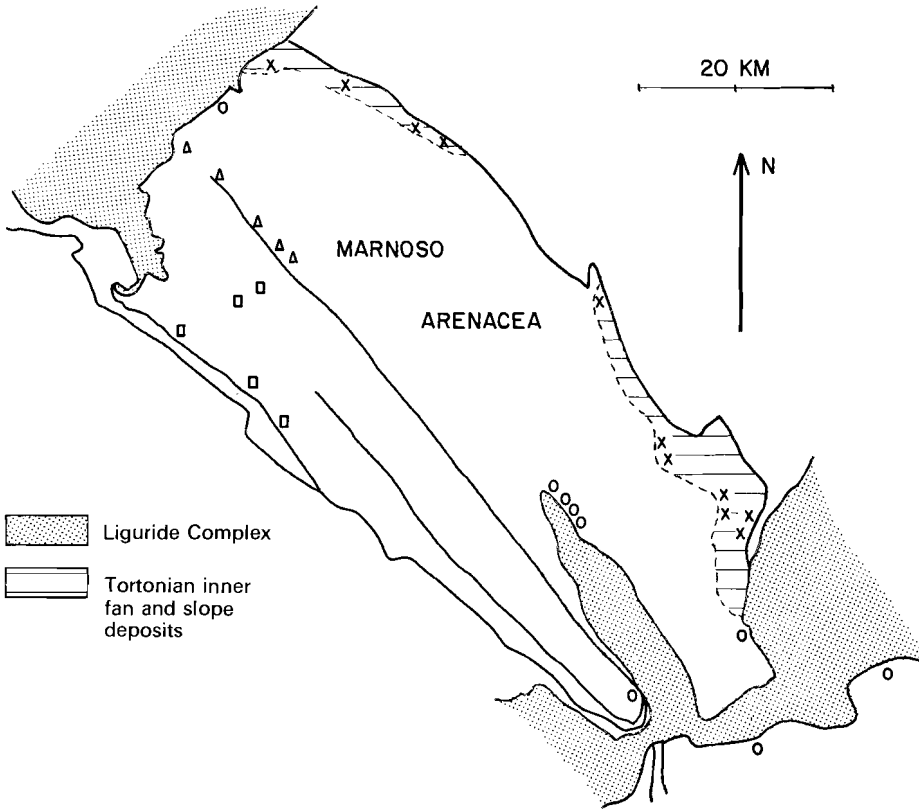

FIG. 9. Occurrences of deposits 'a Lucina' in the Marnoso Arenacea (after RICCI LUCCHI and VEGGIANI – 1967) (explanation in text). △ group 1, ○ group 2, □ group 3, × group 4.

in accordance with the view of Ricci Lucchi and Veggiani who supposed that the 'calcari a Lucina' came from the SW.

The Tortonian Fontanelice Member deposits present some problems. Most of the pebbles of the conglomerate occurring in this member are imbricated to the SE (150°–160°), suggesting deposition from shallow water running SE. In the marls just below this conglomerate U-shaped burrows are found; these are generally considered to be indicative for very shallow water deposits (HECKEL – 1972).

In conclusion it can be said that extensive thick turbidite deposits such as those of the Marnoso Arenacea are normally expected in rather deep water, but there are some indications that the basin was, at least not always, as deep as is commonly supposed. Occasionally the Castel Vecchio high reached the littoral zone, and then the deepest part of the basin may have been only about 500 m deep. There are also indications that the depositional environment of the Fontanelice Member was at least occasionally very shallow.

2.3. GESSOSO SOLFIFERA FORMATION

The local name for the Gessoso Solfifera Formation is 'vena del gesso', because of its conspicuous appearance as a E-W running seam through the landscape in the N of the investigated area (photo 10). The Gessoso Solfifera Formation is underlain by the Marnoso Arenacea, and overlain by the Casal Fiumanese Formation.

The Gessoso Solfifera Formation consists mainly of cristalline gypsum (selenite) alternating with thin carbonate and shale layers. The thickness of the gypsum layers ranges from a few metres to 30 metres. The selenite cristals are often very large, up to several decimetres. BATTISTA VAI and RICCI LUCCHI (1977) recognized 14 depositional cycles in this formation. The ideal cycle, from bottom to top, is as follows; shale, stromatolitic limestone, massive coarse selenite, banded selenite, nodular selenite and reworked selenite. In the area not all 14 cycles are present. Only some 8 to 10 cycles, probably corresponding to Ricci Lucchi's lower ones, could be found. These lower cycles are composed mainly of coarse selenite and nodular selenite, but reworked selenite may be important locally.

The thickness of the Gessoso Solfifera Formation between the Santerno and Sillaro is markedly variable. The thickness is about 100 m, but it may be much less, and locally the entire formation is not present. The mean thickness diminishes towards the W.

There is no disagreement about the Messinian age of the Gessoso Solfifera Formation.

Messinian evaporitic rocks occur all over the Mediterranean region. Several hypotheses have been proposed to explain the enormous amounts of evaporites that were deposited during the Messinian (see DROOGER – 1973). It would be beyond the scope of this thesis to discuss them. BATTISTA VAI and RICCI LUCCHI

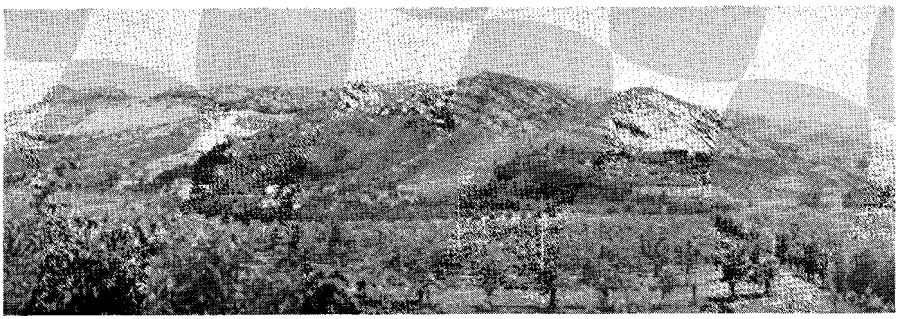

Photo 10. The Gessoso Solfifera Formation as a conspicuous seam in the landscape: the 'vena del gesso'.

(1977) interpreted the gypsum of the Vena del Gesso as a deposit in rather shallow lagoonal environments. According to him, each cycle represents a shallowing of the depositional environment from some tens of metres, becoming alluvial for the reworked selenite. Most of the gypsum, apart from the coarse massive selenite, probably accumulated as a result of resedimentation processes, indicating a slope and a certain instability. The direction of transport could not be established. SESTINI (1970a) states that in other places of the Messinian deposits of the Padan foretrough an external source could be established.

The variation in thickness of the Gessoso Solfifera Formation may in part be primary, but is probably mostly due to a mild erosion that occurred almost everywhere along the NE side of the Northern Apennines during the early Pliocene (LUCCHETTI et al. – 1962).

2.4. CASAL FIUMANESE FORMATION

The Casal Fiumanese Formation is called after the village Casal Fiumanese just N of the investigated area. In the literature this formation is usually referred to as 'Pliocene del Santerno'. The Casal Fiumanese Formation overlies the Gessoso Solfifera Formation, and is overlain outside the realm of the studied area by Quarternary deposits. In the W the formation interfingers with the Liguride Complex (photo 11). The whole formation dips about 30° to the N and NE.

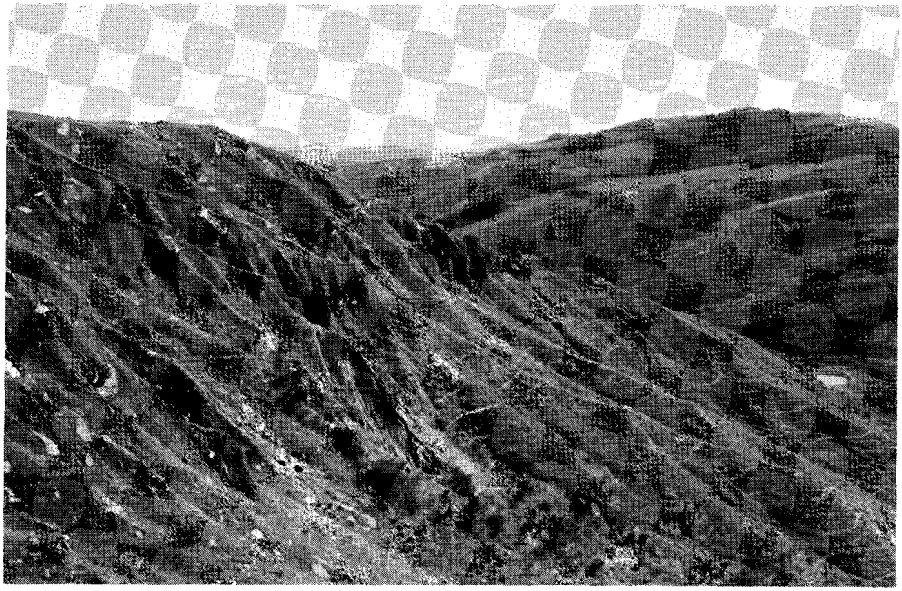

Photo 11. Overthrust contact of the Chaotic Complex on the Casal Fiumanese Formation.

The main constituent of the Casal Fiumanese Formation is clay of a light grey, sometimes bluish colour. Mostly it is not layered at all or is badly layered. Only in places where the clay is somewhat silty can some small scale sedimentary structures such as ripples and cross lamination be recognized. The lower contact with the Gessoso Solfifera Formation is mostly conformable, but exhibits in some places an unconformity. Some sand and conglomerate lenses are present in the Casal Fiumanese Formation. The amount of coarse clastics intercalated diminishes from W to E. Along the river Santerno sand-lenses occur only in the upper 200 m of the formation (fig. 10) (LUCCHETTI et al. – 1962). Near the contact with the Liguride Complex the formation contains not only more sand but also conglomerate lenses with thicknesses of up to several metres and lenghts up to several hundreds of metres, as well as large and small olistostromes derived from the Liguride Complex. The sand and conglomerate lenses often contain moluscs and gastropods, many of which are shattered. Large- and small-scale cross lamination is present in the sand and conglomerate lenses. The conglomerates are polymict. The pebbles could be recognized as being derived from the Liguride Complex and the Marnoso Arenacea. Boulders of up to 40 cm can be found, but usually the pebbles range from 15 to 20 cm.

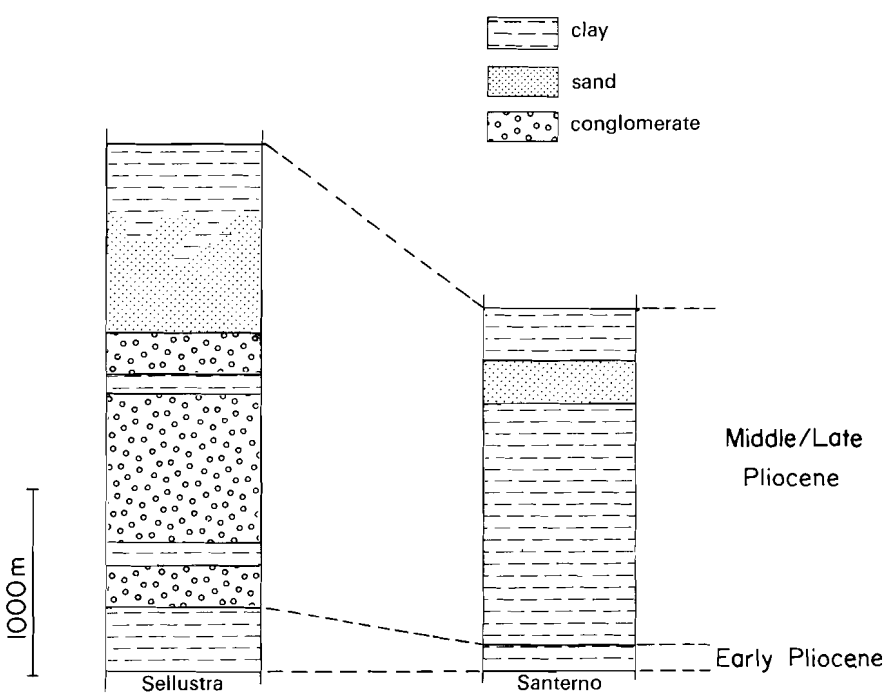

FIG. 10. Columnar sections of the Pliocene along the Sellustra and Santerno, only 5 km apart.

The thickness of the Casal Fiumanese Formation diminishes towards the E. Near its contact with the Liguride Complex in the W the thickness is about 2900 m, whereas along the river Santerno the thickness is only about 1000 m (fig. 10).

All authors agree on a Pliocene age of the Casal Fiumanese Formation. The lower 200 to 300 m are Early Pliocene, the rest Middle/Late Pliocene. There seems to be a slight hiatus between the underlying Messinian Gessoso Solfifera Formation and the base of the Casal Fiumanese Formation (LUCCHETTI et al. – 1962).

Sedimentary structures as well as the micro and macro fauna indicate a rather shallow neritic depositional environment (LUCCHETTI et al. – 1962, SESTINI – 1970a). Therefore the increase in thickness of the formation towards the W is ascribed to a faster subsidence of the western area. This faster subsidence apparently also determined the deposition of the coarser constituents of the Casal Fiumanese Formation in the western part. The sediment was obviously derived from the erosion of already emerged parts of the Northern Apennines, as can be deduced from the sedimentary structures indicating internal sources (S of W) and from the lithologies of the conglomerate pebbles.

The short hiatus in the Lower Pliocene, and the local unconformity at the base of the Casal Fiumanese Formation are interpreted as a short period of erosion that occurred at the beginning of the Pliocene. Unconformities between the Messinian and the Lower Pliocene are also known from beneath the Padan plain. The Pliocene sediments there are folded in broad, somewhat NE-vergent anticlines. On top of these anticlines unconformities are found between the Messinian and Lower Pliocene, as well as between the Lower and Middle Pliocene. In the synclines these unconformities do not seem to be present. Obviously the folds developed synsedimentarily. The Lower-Middle Pliocene unconformity could not be established in the area investigated.

2.5. CASTEL GUERRINO UNIT

2.5.1. GENERAL

The rocks of the Castel Guerrino unit crop out in a narrow zone along the Apenninic divide between the P^{so} della Futa and the Monte Falterona as a tectonic slice between the Marnoso Arenacea and the Cervarola Sandstone Formation (fig. 11). Whereas the Castel Guerrino unit is thrust upon the Marnoso Arenacea between the Monte Falterona and the P^{so} del Giogo, it overthrust the Liguride Complex between the P^{so} della Futa and the P^{so} del Giogo.

The reasons why this unit is treated separately are:

1. It constitutes a separate tectonic unit.

2. It is lithologically distinguishable from both the Marnoso Arenacea and the Cervarola Sandstone Formation. Groscurth treated the Castel Guerrino unit as a distinct unit as well, but it has generally been regarded as belonging to either the Marnoso Arenacea or the Cervarola Sandstone Formation.

The Castel Guerrino unit comprises two formations: the Castel Guerrino Formation, which makes up the bulk of the unit, and the Rifredo Marls.

2.5.2. Castel Guerrino Formation

The Castel Guerrino Formation is named after the highest mountain in the area studied: the Monte Castel Guerrino (1117 m). Groscurth, for rather vague reasons, regarded the most eastern part of the formation, a small thrust slice near Rifredo, as belonging to the Marnoso Arenacea.

The Castel Guerrino Formation is a turbidite formation which is as a whole more marly than the Cervarola Sandstone Formation and the adjacent parts of the Marnoso Arenacea. In the middle part of the formation several intervals of very calcareous marls with thicknesses of up to 20 m, showing great similarity with the Rifredo Marls, are found. The sandstones are lithologically more quartzose than those of the Cervarola Sandstone Formation, and less so than those of the Marnoso Arenacea. Some thin calcarenites occur. The fresh colour of the sandstones is ash-grey, the weathered colour is brown or grey-brown. The woolsack weathering, which can often be observed on the upper part of the layers, is due to the presence of clay mixed with the matrix. The main part of the deposits can be classified as facies association *c*. Lateral facies changes are not observed.

The thickness of the Castel Guerrino Formation can only be a minimum estimate, because all the contacts are tectonic, and many internal reverse faults occur. According to Groscurth (1971) the exposed thickness is 900 m, which seems reasonable.

The very poor fauna inhibits a precise dating of the Castel Guerrino Formation. Contrary to Groscurth, according to whom the age is Late Aquitanian to Early Burdigalian, the present author is of the opinion that the age is at least Langhian, because of the occurrence of Orbulina bilobata and Orbulina universa.

2.5.3. Rifredo Marls

The Rifredo Marls occur in several places near the P^{so} del Giogo, often apparently floating in the Chaotic Complex but generally in contact with the Castel Guerrino Formation.

The Rifredo Marls consist of brown, very calcareous marls that contain varying amounts of silt and micas. When the marls are weathered, the colour is light grey. The marls break up into large splinters.

The maximum observed thickness of the Rifredo Marls is 50 m.

The Rifredo Marls contain a very poor fauna. Because of the occurrence of Orbulina sp. the formation cannot be older than Langhian, and is probably of the same age as the Castel Guerrino Formation.

2.5.4. Stratigraphical status of the Castel Guerrino unit

The intermediate position of the Castel Guerrino unit between the Marnoso Arenacea and the Cervarola Sandstone Formation made Groscurth (1971) suppose that the unit was deposited either in the NE part of the Cervarola basin or in a narrow basin between the basins of the Cervarola Sandstone Formation and the Marnoso Arenacea. Units in a similar position to the Castel Guerrino unit are the Castiglione Sandstones (Hemmer – 1971) and the Mandrioli Sandstones (Frey – 1969) (fig. 11). The three units are similar as far as their tectonic position, their lithology and age are concerned. Frey (1969) supposed that the Mandrioli Sandstones were deposited in a transitional zone between the Marnoso Arenacea and the Cervarola Sandstone Formation. According to Hemmer (1971) the Castiglione Sandstones are an 'Umbro-Romagnolische Zwischeneinheit'. This would mean that this unit is also considered to have been deposited in a transitional zone between the Marnoso Arenacea and the Cervarola Sandstone Formation. The 'Berliner Schule', to which Frey, Groscurth and Hemmer belong, thus supposes that the three units mentioned are continuations of each other. The Castiglione Sandstones and the Castel Guerrino unit do indeed seem to form one tectonic unit, but this is not connected with the Mandrioli Sandstones (fig. 11). Near the Monte Falterona, the Castel Guerrino unit disappears under the overthrust of the Cervarola Sandstone Formation, while the thrust plane of the Mandrioli Sandstones degenerates towards the NW into a NE vergent fold. The Mandrioli Sandstones are consequently not a continuation of the other two units but are part of the Marnoso Arenacea – in fact, they are a continuation of the Coniale Member. There is no evidence for a separate Mandrioli basin. The lithological resemblance of the Castel Guerrino unit and the Castiglione Sandstones to the Mandrioli Sandstones makes it probable that these two formations were also deposited in a more internal part of the Marnoso Arenacea basin.

2.6. TUSCAN SEQUENCE

2.6.1. General

The Tuscan Sequence will be treated only briefly. It crops out in the S of the investigated area on the SW side of the Apenninic divide. The sequence is thrust upon the Castel Guerrino Formation, and is in turn overthrust by the Liguride Complex, and is also partly covered by the Mugello Formation.

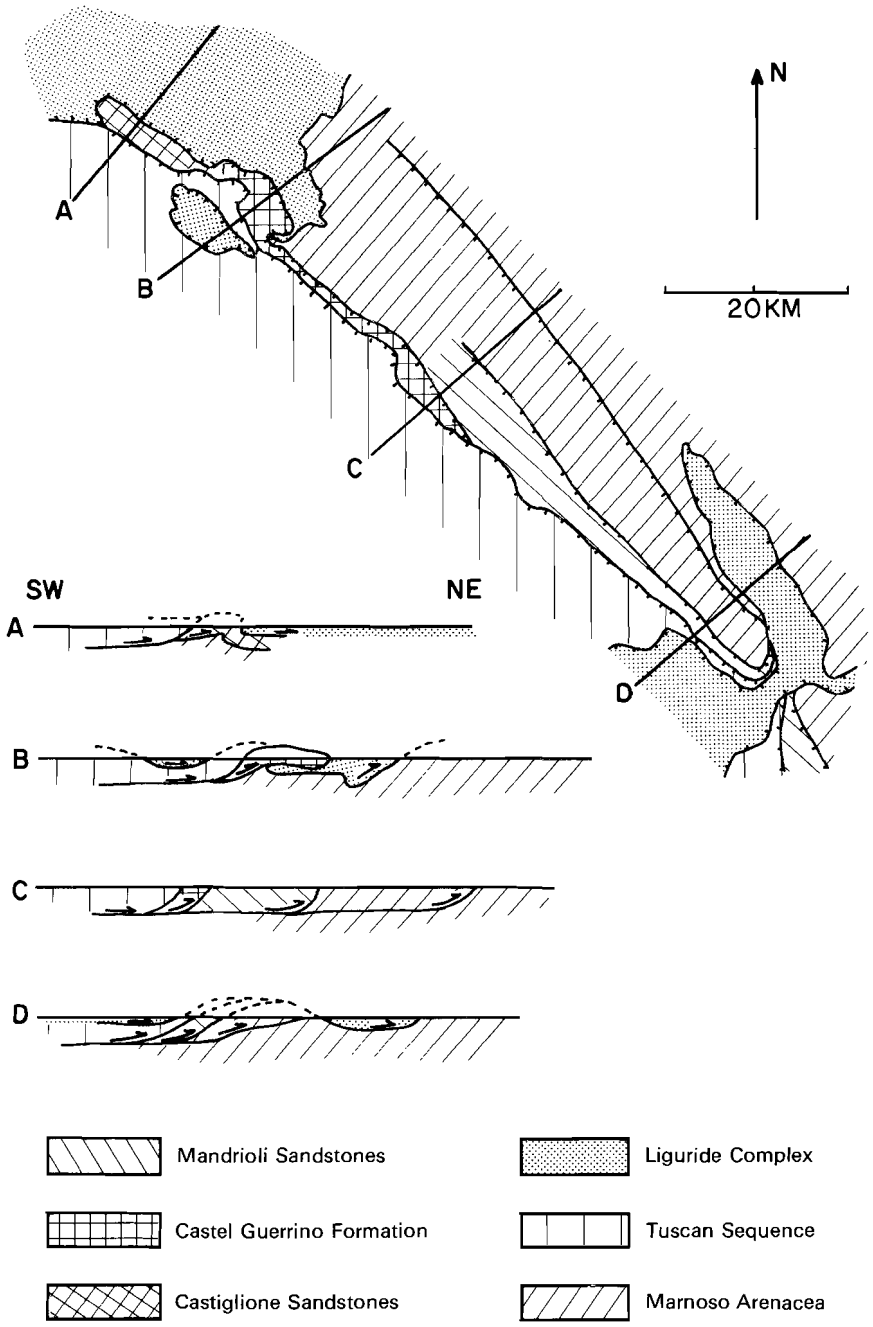

FIG. 11. Schematic tectonical map and sections showing the relations between the Mandrioli Sandstones, Castel Guerrino Formation, Castiglione Sandstones and the Marnoso Arenacea.

In the area investigated the Tuscan Sequence comprises two formations: the Scaglia Toscana and the Cervarola Sandstone Formation. The Vicchio Marls belong on top of the Cervarola Sandstone Formation (BORTOLOTTI et al. – 1970), but in the investigated area these marls do not crop out in normal stratigraphical contact with the Cervarola Sandstone Formation. Perhaps they are represented in the Casaglia 'klippe sédimentaire' (section 2.2.3).

2.6.2. SCAGLIA TOSCANA

Other names that have been given to the Scaglia Toscana in the past are: Scisti Varicolori, Scisti Policromi, Scisti Rossigni, Scisti Rossi, Scaglia Rossa and Scaglia. The Scaglia Toscana is a well known formation in the Northern Apennines and is usually found at the base of the Modino-Cervarola Sandstone Formation and the Macigno. In the area studied the Scaglia Toscana is always found below the Cervarola Sandstone Formation, just SW of the Apenninic divide, where it marks up- and overthrusts of the Tuscan Sequence.

Lithologically the Scaglia Toscana can be divided into two parts. The lower part consists of variegated shales in which red and green colours dominate, but yellowish and darker colours are also present. The upper part has a quite different appearance. Dark shales alternate with thin black siltstones and occasionally a sandstone layer. The thicknesses of the layers are generally between 5 and 15 cm. The upper part of the formation is generally more severely tectonized than the lower part.

Because the lower contact of the Scaglia Toscana is always tectonic, the thickness of the formation varies markedly, from zero to a maximum of 150 m.

As for the age of the Scaglia Toscana, most authors agree that it is Eocene-Aquitanian (GROSCURTH – 1971, BORTOLLOTI et al. – 1970).

2.6.3. CERVAROLA SANDSTONE FORMATION

Other names for the Cervarola Sandstone Formation are Macigno, Macigno B, Modino-Cervarola Sandstone Formation. This formation overlies the Scaglia Toscana, the upper boundary being the overthrust of the Liguride Complex. The transition of the Scaglia Toscana to the Cervarola Sandstone Formation is abrupt.

Lithologically the Cervarola Sandstone Formation is an alternation of coarse thick sandstones and thin marls. The sandstones are grauwackes, less quartzose than those of the Castel Guerrino Formation. Sedimentary structures point to deposition by turbidity currents. The direction of supply was to the SE (TEN HAAF – 1959). Both the fresh colour and the weathered colour are darker than the colour of the Marnoso Arenacea and the Castel Guerrino Formation. The weathered colour is often somewhat greenish.

The thickness of the Cervarola Sandstone Formation in the investigated area is at least 450 m.

The age of the Cervarola Sandstone Formation is Aquitanian to Langhian (BORTOLOTTI et al. – 1970).

2.7. LIGURIDE COMPLEX

2.7.1. GENERAL

The Liguride Complex consists of rocks that are supposed to have been deposited in the eugeosynclinal realm of the Northern Apennines. This eugeosynclinal realm was probably situated in the present Tyrrhenean and Ligurian Seas (ABBATE and SAGRI – 1970). At present the Liguride rocks are to be found in an overthrust position on the miogeosynclinal rocks. In the western part of the Northern Apennines a certain order of the various Liguride thrust sheets can still be recognized. To the E and NE the Liguride rocks become more and more chaotically intermixed. In the region studied, which is situated in the NE zone, the Liguride rocks can best be described as a melange on a large scale: a chaotically mixed assemblage of isolated rock fragments of different lithology and size, floating in clay or shale. The clay matrix of this melange is called Argille Scagliose. In the older Italian, and in the American literature the name Argille Scagliose is often used to describe the entire melange. In 1956, however, MERLA suggested that the term Argille Scagliose should refer only to the clays. The Argille Scagliose which contains rock fragments that are not sufficiently large to be mappable are given the 'formation' name Chaotic Complex. The larger rock fragments and dismembered thrust sheets floating in the Chaotic Complex are given the following formation names: Borgo Marls, Lame Limestones, Piotto Variegated Marls, Violla Marls, Monte Morello Formation and Sellustra Formation. Outside the realm of the area investigated many other formations occur, such as ophiolite and many different turbidite formations. Not all the rocks in the Liguride Complex were deposited in the eugeosynclinal realm of the Northern Apennines geosyncline. Those which were not should be called parautochthonous if they have been picked up from the footwall of the overthrusting Liguride Complex, and meso-autochthonous if deposited on the overthrusting Liguride Complex.

2.7.2. CHAOTIC COMPLEX

The Chaotic Complex makes up by far the largest part of the Liguride Complex. It overlies formations which become progressively younger from SW to NE: Cervarola Sandstone Formation, Marnoso Arenacea, Gessoso Solfifera Formation, Casal Fiumanese Formation (Early Miocene to Late Pliocene). It is said that the Chaotic Complex overlies tectonically transgressively its

footwall formations (section 4.6). It interfingers locally with the Marnoso Arenacea and the Casal Fiumanese Formation. In the SW the Chaotic Complex is overthrust locally by the Castel Guerrino unit.

The Chaotic Complex consists of slickensided, dark, sometimes red clays containing dispersed angular rock fragments of all sizes and lithologies. The mixing of the different elements of the Chaotic Complex has been so intense that the formation at first sight displays a great uniformity. However, a closer examination shows that there are differences within the formation, mainly because the various rock fragments are not distributed randomly. Some lithologies have areas of maximum occurrence, which are generally associated with the presence of larger slabs with the same lithologies. Most of the rock fragments belong to the following lithologies:
Limestones: mostly dense, with a conchoidal fracture, and containing many thin calcite veins, white or light coloured. Probably most of these clasts derived from the Monte Morello Formation. Some rather similar brown weathered limestones are probably derived from the Lame Limestones.
Marly limestones: mostly light to dark grey, and of Eocene age (GROSCURTH – 1971), probably partly derived from the sub-Liguride Canetolo Complex, and the Monte Morello Formation.
Sandstones: a great variety of sandstones is found, probably belonging to various North Apenninic turbidite formations, both eugeosynclinal and miogeosynclinal.
Marls: calcareous as well as clayey, mostly light grey. They probably have the same origin as the sandstones.
Siltstones: very hard grey siltstones with a rust coloured weathered surface of unknown age and origin.
Radiolarites: occasionally a red radiolarite clast is found. Their origin may by Liguride as well as Tuscan.
Ophiolites: occasionally ophiolite fragments occur. Just W of the investigated area they occur as mappable units. These are considered to be of diagnostic significance for the Liguride Complex.

Three main zones can be recognized within the Chaotic Complex.
1. the Firenzuola area
2. the Sillaro area
3. the Sellustra area

1. The Chaotic Complex of the Firenzuola area, in the S of the area investigated, is characterized by many marl clasts, which are probably mainly derived from the Borgo Marls and the Rifredo Marls. Also locally many Lame Limestone clasts occur. The Argille Scagliose are often rather marly, and have a somewhat lighter colour than usual. The occurrence of the mappable formations Borgo Marls, Violla Marls, Piotto Variegated Marls and Lame Limestones, floating within the Chaotic Complex, is restricted to this zone. Ophiolite clasts are present, but very scarce. The Chaotic Complex of this area inter-

fingers with the Monte Coloreto Member of the Marnoso Arenacea.

2. The Chaotic Complex of the Sillaro area, between Peglio and Gesso, is very uniform as the result of intense mixing. Monte Morello and sandstone clasts are abundant. Ophiolites and radiolarites have their maximum occurrence here. The lustrous slickensided character of the Argille Scagliose is best manifested in the Sillaro area. Near the base of the Chaotic Complex some marl lenses of the Piancaldoli Member of the Marnoso Arenacea are intercalated.

3. The Chaotic Complex of the Sellustra area, N of Gesso, is a superposition of distinct streaks or layers, several metres to several tens of metres thick. Each layer consists of clay (Argille Scagliose) containing rock fragments orientated parallel to the layering. The layers can be distinguished because of slight colour differences in the Argille Scagliose and differences in the amount and mean size of the rock fragments. The maximum size of most of the rock fragments ranges from a few cm to several dm. The olistolithes of this zone show less diversity than elsewhere in the Chaotic Complex. By far the greatest part consists of dense light grey, brown weathered limestones, with thin calcite veins. Clearly they are derived from layers that are 10 to 30 cm thick. Their origin however is unknown. They exhibit some similarity with the Lame Limestones, but their great distance from the Firenzuola area, to which the occurrence of the Lame Limestones seems to be restricted, makes it inadmissable to relate them to that formation. Near the contact with the Pliocene deposits locally many rust coloured hard siltstones are found. The thickness of these clasts is usually about 5 cm, their length 10 to 20 cm. Similar clasts have been found by the author in the Marecchia area, where the Chaotic Complex overlies Tortonian marls (near San Agata Feltria). The Sellustra Formation occurs only in the Chaotic Complex of the Sellustra area. Lenticular bodies of this formation, with a diameter of up to several hundreds of metres and a thickness of up to several tens of metres, are intercalated mainly in the lower part of the Chaotic Complex. Upwards the amount of these bodies diminishes rapidly. The Chaotic Complex of the Sellustra area interfingers strongly with the Pliocene Casal Fiumanese Formation.

The Chaotic Complex is always the highest tectonic unit, so only minimum thicknesses can be given. Due to the general dip to the NW, the depth of the base of the Chaotic Complex, and thus the thickness, increases in that direction. Near the NW edge of the map the thickness is already about 500 m (Masini – 1951). Beneath the Monghidoro slab, further to the NW, outside the area covered by the map, the thickness is at least 1200 m, as can be deduced from borehole data (Minucci – 1957, as quoted by Bruni – 1973). Elsewhere even greater thicknesses have been measured, up to 2000 m near Monteveglio, 17 km W of Bologna (Lucchetti et al. – 1962), and it is likely that thicknesses of up to 3000 m occur.

The age of most rocks appearing in the Chaotic Complex ranges from Creta-

ceous to Eocene (ABBATE and SAGRI – 1970), although older (Jurassic ophiolites and radiolarites) and younger (Paleocene to Miocene) ages do occur.

2.7.3. BORGO MARLS

Slabs of the Borgo Marls are found in the Liguride Complex as isolated lenses near its contact with the Marnoso Arenacea, N and W of Firenzuola. The occurrence of the Borgo Marls is restricted to the Firenzuola area.

The Borgo Marls consist of light grey marls which exhibit a brown, sometimes greenish weathered colour, alternating with more calcareous levels. Some thin (up to 5 cm) green sandstone layers occur. Both upper and lower surfaces of these layers are irregular. The Borgo Marls are always heavily tectonized.

The maximum observed thickness of the Borgo Marls is 40 m.

The very poor planktonic microfauna found in the Borgo Marls, prevented a precise dating of the formation.

2.7.4. LAME LIMESTONES

In the Firenzuola area two mappable units of the Lame Limestones are found as well as numerous unmappable rock fragments derived from this formation. One unit is located 1.5 km SW of Firenzuola, the other near Lame, 1 km S of Moscheta.

The Lame Limestones consist of grey marls and dark clays alternating with light grey, when weathered brown, dense limestones, some fine grained sandstones, and some marly limestones. The thickness of the layers ranges from 1 to 50 cm, but is usually 15 to 30 cm. In both locations the formation is heavily tectonized. Near Firenzuola the formation is completely broken up and the internal layering is completely destroyed.

The maximum observed thickness of the Lame Limestones is 25 m.

GROSCURTH (1971) found some radiolarians and diatoms in the Lame Limestones. The age of the formation could not be established. On lithological grounds Groscurth supposed the age to be Lower Cretaceous.

2.7.5. PIOTTO VARIEGATED MARLS

S of Firenzuola there are several occurrences of the Piotto Variegated Marls, floating as large olistolithes in the Chaotic Complex of the Firenzuola area.

The Piotto Variegated Marls consist mainly of clayey marls. The colour of these marls is light-grey, but green and red colours are also present in subordinate amounts. In the largest occurrence of this formation several thick coarse grained sandstone layers are found. The thickness of these sandstone layers is 0.5 to 2.5 m. The fresh colour is grey, the weathered colour brown. The sandstones show great similarity with the Cervarola Sandstones.

The maximum observed thickness of the Piotto Variegated Marls is 30 m.

The age of the Piotto Variegated Marls is not known. If the sandstones belong to the Cervarola Sandstone Formation, then the marls may belong to the Scaglia Toscana, and the age would then be Eocene to Early Miocene.

2.7.6. Violla Marls

The Violla Marls comprise several marly lenticular bodies that occur near the upper course of the T. Violla in the Firenzuola area. They comprise the Bunte Schiefertone von Collinaccia and the Mergel von Casetta di Rocco of Groscurth (1971).

The marls are generally somewhat clayey, but may be rather calcareous. The colour is light grey to brown. The marls may alternate with thin fine-grained brown sandstone layers and with grey mostly marly limestones. At one location several one-metre-thick olistostromes with white limestone olistolithes are intercalated. These olistostromes are probably descended from the Chaotic Complex. The Violla Marls are often tectonized, but never to such a degree that the bedding is completely destroyed.

The maximum observed thickness of the Violla Marls is 40 m.

Groscurth (1971) described a fauna of the Mergel von Casetta di Rocco which contains Orbulina bilobata, suturalis and universa, indicating a Langhian or younger age. Groscurth found the Bunte Schiefertone von Collinaccia to be completely sterile.

There are lithological differences between the different occurrences of the Violla Marls. The main resemblance is in the predominantly marly character. Although there is a possibility that some completely different, unrelated formations are taken together here, the author is of the opinion that most of the occurrences are at least genetically related.

A Langhian age for the Violla Marls means that they were deposited approximately in the course of the arrival of the Liguride Complex in the area where they are found at present, as the Liguride Complex of the Firenzuola area interfingers with Middle Miocene deposits of the Marnoso Arenacea. This implies that they cannot have been transported very far tectonically. This

opinion is supported by the fact that the occurrence of the Violla Marls is limited to a very small area. The short distance of transport did not allow an intensive mixing within the Liguride Complex. In fact the Violla Marls should not be called allochthonous but parautochthonous. Maybe their depositional history is similar to that of the Frena Member deposits. However, because similar deposits are not found in or in direct contact with the Marnoso Arenacea, the Violla Marls are treated here together with the Liguride rocks in which they occur at present.

2.7.7. Monte Morello Formation

Two large slabs of the Monte Morello Formation occur in the area studied, one near Monte Calvi in the SW (4 by 6 km large) and one near Monte Carpinaccio in the W (1 by 4 km large). This formation is also often referred to as Alberese.

The Monte Morello Formation consists mainly of dense, white limestones with a conchoidal fracture and marly yellow-grey limestones. Often the layers are parallel laminated. The layer thickness ranges from several cm to 3 m. The base of the thicker beds is sometimes sandy, and may exhibit flute casts, indicating deposition from turbidity currents. Subordinate are some blue-grey fine-grained mica-rich sandstones.

The minimum thickness of the slab near Monte Calva is 250 m (Groscurth – 1971), the one near Monte Carpinaccio having a minimum thickness of 200 m. Elsewhere the formation may be much thicker, e.g. up to 800 m in the Chianti Mountains (Abbate and Sagri – 1970).

The age of the Monte Morello Formation is Paleocene to Middle Eocene (Abbate and Sagri – 1970, Groscurth – 1971).

2.7.8. Sellustra Formation

The Sellustra Formation occurs as isolated lenses in the lower part of the Liguride Complex of the Sellustra area. The lenses have thicknesses between 5 and 50 metres, and lengths of 50 to 500 metres.

Different lithologies are met in this formation. Usually light grey clay, sometimes red, makes up the bulk of the formation. Locally the clay is marly. Thin sandlayers (cross laminated) as well as 10 to 30 cm thick white limestones, sometimes fossiliferous, occur in varying amounts. Generally the stratification is undisturbed, but locally slumplike folds are met.

Well preserved, often benthonic, microfaunas indicate ages of the Sellustra Formation varying between Late Eocene and Early Miocene.

Both microfauna and (scarce) sedimentary structures indicate that the Sellustra Formation is a shallow water deposit. Most probably the Sellustra Formation was deposited upon the overthrusting Liguride Complex, and is equivalent with the late geosynclinal Loiano and Bismantova series (SESTINI – 1970b), and may be called meso-autochthonous. Because the rocks occur at present intercalated in the Liguride Complex, they are treated here together with the Liguride rocks.

2.8. MUGELLO FORMATION

The Mugello Formation crops out in the extreme S of the area, where it discordantly overlies Liguride and Tuscan rocks.

The Mugello Formation consists of lacustrine deposits, mainly marls, sands and conglomerates (HEMMER – 1971).

The maximum thickness of the Mugello Formation is 700 m (GROSCURTH – 1971).

The age of the Mugello Formation is generally given as Villafranchien (Late Pliocene to Quarternary) (HEMMER – 1971, SESTINI – 1970a).

3. GRAVITATIONALLY EMPLACED SEDIMENT MASSES

3.1. INTRODUCTION

Gravitation has played a major role in the tectonics as well as in the sedimentation of the Northern Apennines miogeosyncline. The various types of gravitational transport phenomena form a series ranging from turbidity currents to the gravitational emplacement of large thrust sheets. One end of the series consists of phenomena commonly considered as (re)sedimentary, the other end consists of phenomena commonly considered as tectonic. The distinction between sedimentation and tectonics in this series is arbitrary and thus a matter of much debate. Non-genetic criteria, such as size and lithology, often play a part in these discussions. For this reason HOEDEMAEKER (1973) introduced a new comprehensive term for the gravitational emplacement of sediment masses: delapsion. Hitherto this term has not had many adherents. The term will not be used in the present paper either because it is often not easy to tell whether the emplacement of a thrust sheet has been gravitational or has been caused by compression, just as it is not always clear whether a certain layer is a sediment gravity flow or a 'normal' sediment. Hoedemaeker therefore has not solved the problem but has simply created another one. The question is really whether there is a genetic difference between small olistostromes and large gravitational emplaced melanges such as the Liguride Complex. According to the original definition of FLORES (1955) the former are sediments, whereas the latter are usually regarded as tectonic units.

The undisputedly sedimentary processes of gravitational mass transport, ranging from slumps to turbidity currents, can be classed together as sediment gravity flows.

3.2. SEDIMENT GRAVITY FLOWS

MIDDLETON and HAMPTON (1973) distinguished four classes of sediment gravity flows: debris flow, grain flow, fluidized sediment flow and turbidity current. As these four classes are members of one genetic series, all transitions between them are possible. In the case of the highly concentrated debris flows the concentration of the flow is about 1.5 to 2.4 g/cm^3, whereas with the turbidity currents the concentration decreases to about 1.03 to 1.2. Turbulence increases with decreasing concentration.

The genetic classification of sediment gravity flows of Middleton and Hampton can be compared with the descriptive classification of turbidites and associated deposits given in section 2.1., table 1. Both classifications correspond as follows:

Facies A – debris flow, grain flow, fluidized flow, high density turbidity current.
Facies B – fluidized flow, high density turbidity current.
Facies C – turbidity current.
Facies D_1 – turbidity current.
Facies D_2 – turbidity current.
Facies D_3 – very low density turbidity current, hemipelagic 'rainfall'.
Facies E – turbidity current.
Facies F – debris flow.
Facies H – debris or grain flow and turbidity current.

Although some of the deposits of facies F may be classified as debris flows, they generally fall outside the genetic classification. Olistostromes, slumps and klippes sédimentaires also belong to facies F.

Facies H is always found near the contact with the Liguride Complex, from which the angular components probably came. The transport and deposition can be visualized as follows (fig. 12). The turbidity currents flowing along the front of the overthrusting Liguride Complex sometimes 'touch' or overflow advanced lobes of it. These currents can pick up some of the smaller olistolithes and/or set in motion the upper part of the Liguride Complex which may then form one or more grain or debris flows. These flows mix with the deposits of the turbidity current. Fast grain and debris flows cause intercalations of angular components in lower levels of the resultant bed. Since turbidity currents however normally have higher speeds than the grain and debris flows, most

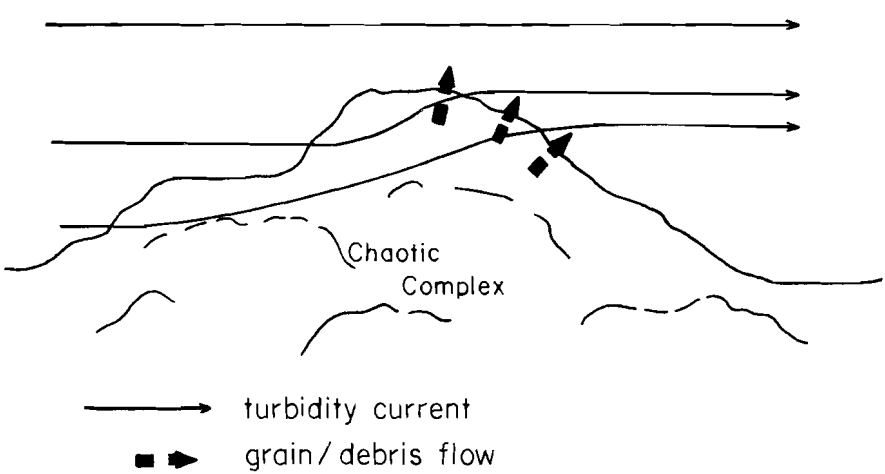

FIG. 12. Turbidity currents flowing over advanced lobes of the Chaotic Complex, causing grain and debris flows of Liguride material.

of the angular components are found on top of the resultant bed, where they may be covered by a thin layer of mud that has settled out from suspension later (photo 3). Facies H is often covered by polimict breccias of facies A. These layers represent grain flows that did not mix with a turbidite.

3.3. FACIES ASSOCIATIONS IN GEOSYNCLINAL BASINS

A number of detailed studies of the physiography and sedimentology of modern deep water fans have been carried out (NORMARK – 1970, 1974, NELSON and KULM – 1973, NELSON and NILSON – 1974). These fans consist mainly of turbidites and associated deposits. MUTTI and RICCI LUCCHI (1972), NELSON and KULM (1973), WALKER and MUTTI (1973), MUTTI (1974) and RICCI LUCCHI (1975a) developed a facies association model for ancient turbidite deposits. They distinguish, by analogy with modern deep water fans an inner fan, an outer fan and a basin plain facies association (fig. 13). However, geosynclinal basins differ essentially from the continental slope-abyssal plain environment. On the oceanic abyssal plains turbidites can spread out almost limitlessly causing radial symmetry. In most geosynclinal turbidite basins the direction of the turbidity currents was parallel to the basin axis, indicating that the narrow elongate form of these basins strongly controlled the course of the currents. It is therefore doubtful whether the morphological units of the modern deep water fans studied were also present in geosynclinal basins. Another complicating factor is that turbidites that are descended from more than one fan system may alternate. These considerations justify the use of a purely descriptive classification (section 2.1).

The particular facies association deposited at any one place within the basin depends on several factors such as the situation within the basin (near the basin axis, on a slope, on an intrabasinal high, proximal or distal) and the supply of sediment. Nevertheless, some speculations can be made a priori about the depositional environment of each facies association.

Facies association *a* is interpreted as being composed of channel fill sequences. As such it is almost identical with the inner fan facies association of the Mutti-Ricci Lucchi model but it differs by having no interchannel deposits and mouth bars. However, this does not seem to be a fundamental difference. Recent continental slopes and the slopes of geosynclinal turbidite basins are probably fairly comparable, although the latter may have been less deep.

Facies association *b* is lithologically comparable with the outer fan facies association of the Mutti-Ricci Lucchi model, but our interpretation of the depositional environment is different. This facies association may indicate proximity to the inner fan, but it usually has been deposited in the deepest parts of the basin, in which case it roughly indicates the position of the basin axis.

Facies association *c* is not comparable with any facies association of the Mutti-Ricci Lucchi model. It bears some resemblance to their fan fringe facies

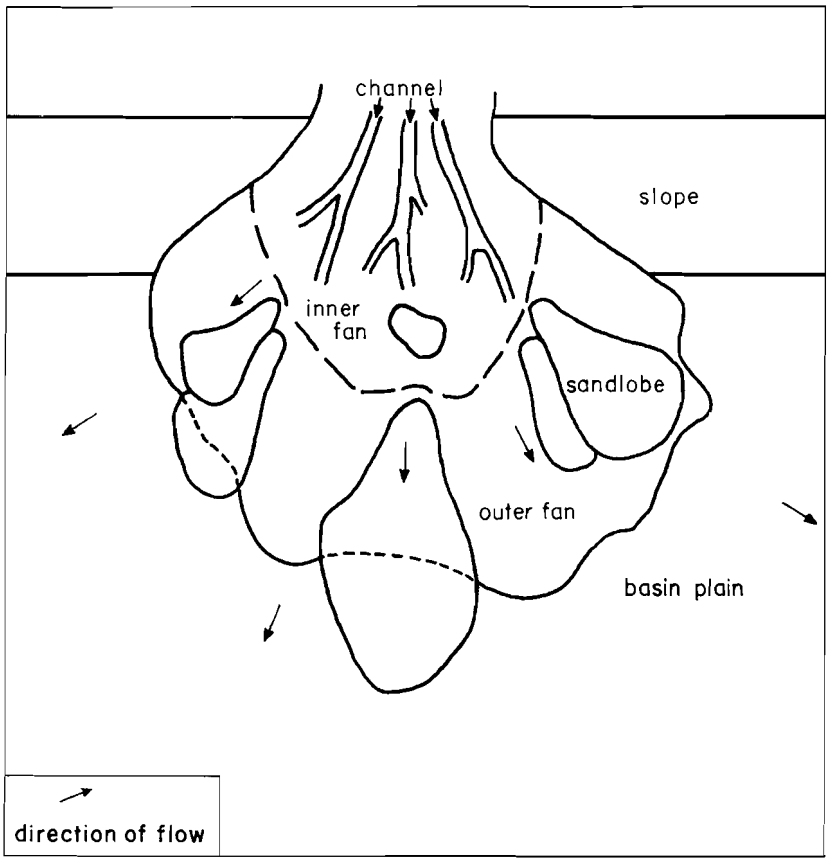

FIG. 13. Distribution model for turbidites in a wide basin, after MUTTI and RICCI LUCCHI (1972).

association which may be present in the transitional zone between the outer fan and the basin plain, but facies association *c* is generally coarser due to the presence of facies C. It may represent alternating deposits of two fan systems, a more proximal and a more distal one. It may also be found on the distal side or up-slope of facies association *b*.

Facies association *d*, which closely resembles facies association *b*, is a more distal or up-slope variant of facies association *b*.

Facies association *e* is lithologically identical with the basin plain facies association of the Mutti-Ricci Lucchi model. It can be found on the distal side of facies association *c*, as well as on the basin slopes in the transitional zone between facies associations *c* and *f*.

The marly facies association *f* is found on basin slopes and intrabasinal highs (fig. 14 a and b). So in places that are protected from turbidity currents.

Facies association *g* will generally not be present in turbidite formations. In the Marnoso Arenacea it occurs only just below the overthrust plane of the

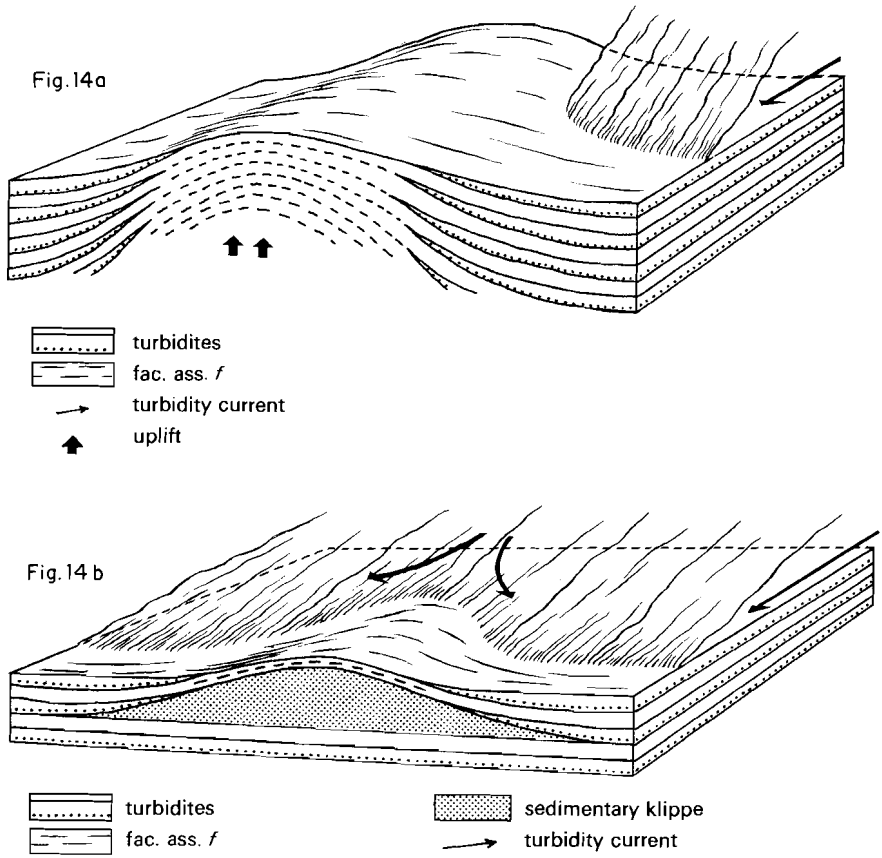

FIG. 14. Facies association *f* may represent intrabasinal high deposits.

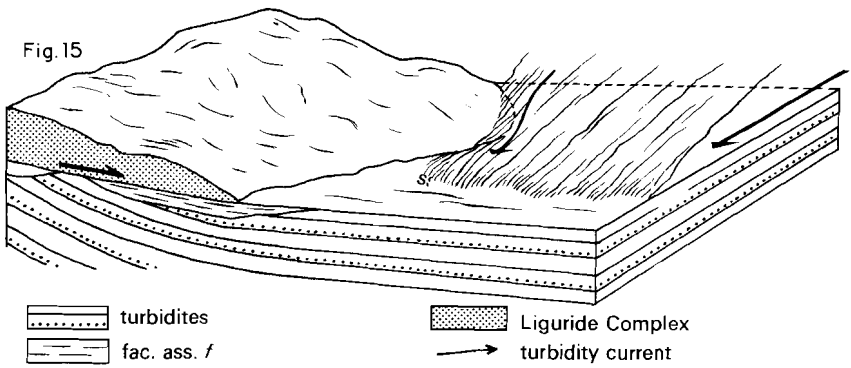

FIG. 15. Facies association *g* will be deposited behind advanced lobes of the Liguride Complex; in areas that are protected from turbidity currents, and where facies H may occur.

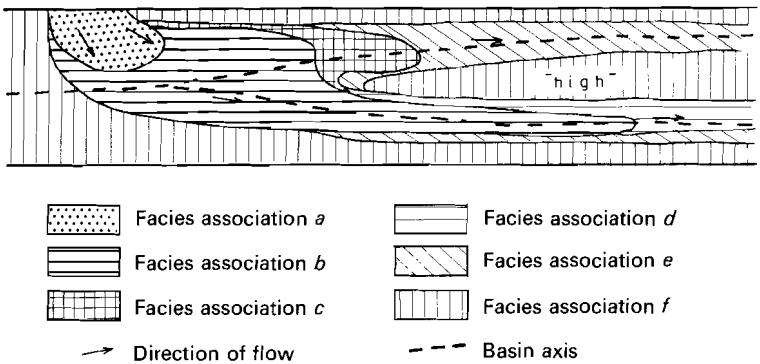

FIG. 16. Schematic model for the distribution of turbidite facies associations in an elongate basin.

Liguride Complex. Advanced lobes of the Liguride Complex locally protected the depositional environment from turbidity currents. The turbidites which were deposited in that environment after all were usually of facies H (section 3.2.) (fig. 15).

A schematic model for the distribution of turbidite facies associations in an elongate basin is given in fig. 16.

3.4. OLISTOSTROMES

The olistostrome concept was introduced by FLORES in 1955. The term olistostrome became widely accepted, although there has been much debate about the definition, which is partly genetic, but at the same time restricts the term olistostromes to deposits that have specific sizes and petrography. Flores states that 'in any olistostrome we distinguish a "binder" or "matrix" represented by prevalently pelitic, heterogenous material, containing dispersed bodies of harder rocks,' and that these rocks 'were accumulated as a semifluid body'. The most important controversies about olistostromes can be summarized in two questions:

1. Are there specific limits to the size of olistostromes? Is an olistostrome with a thickness of a few metres basically different from the Liguride Complex with a thickness of several kilometres?
2. Do olistostromes have to satisfy the petrography specified by Flores?

ad 1: At first glance the lithology of small olistostromes and the Liguride Complex in the Northern Apennines is very similar, although both are very different in size. This is not surprising as many of the small olistostromes originated from the advancing Liguride Complex. They therefore consist of the same material, and an important point is that this material was already chaotic before it became detached from the Liguride Complex. However, there are some differences. Although the small olistostromes may be structureless, some structures can often be observed; these include normal, reversed and irregular

grading of the olistolithes and alignment of the olistolithes parallel to the bedding planes (SAGRI – 1975, REUTTER – 1965, 1969, GÖRLER and REUTTER – 1968, SESTINI – 1976). These structures are not encountered in the Liguride Complex, although there is sometimes an alignment of the olistolithes near the base of the Complex and its large slabs may be imbricated (section 4.6). These differences may point to a genetic difference. The alignment of olistolithes parallel to the bedding planes is generally ascribed to a laminar type of flow. Grading will occur, according to HOEDEMAEKER (1973), in zones where the viscosity is lowered as a result of a high velocity gradient. It is likely that a concentration of olistolithes is formed at a level where the viscosity is such that the sinking comes to an end. The absence of grading in the Liguride Complex may indicate that such high velocity gradients were not present there; this may be the result of an overall lower velocity. At least the olistostromes which became detached from the advancing Liguride Complex and which are at present found intercalated in the sediments below the overthrust plane of the Liguride Complex, had a higher velocity than the Liguride Complex. The dynamics of the Liguride Complex will be treated in section 4.6.
ad 2: In gravitationally emplaced masses with a binder of cohesive pelitic material, plastic flow takes place exclusively through the binder. If there is not a minimum amount of binder present, another type of sediment gravity flow will occur (HOEDEMAEKER – 1973). On the other hand, the presence of olistolithes is not really necessary, as olistolithes do not actively participate in the transport mechanism. This means that mud flows may be included in the olistostrome concept.

In conclusion it can be said that there may be a genetic difference between small olistostromes and the Liguride Complex. Small olistostromes may have greater velocity gradients as a result of a higher velocity, but it is not known whether this is always the case. But if it is, in practice it will be hard to distinguish between the two types of flow. That is why we have adopted here the suggestion of ABBATE et al. (1970b), namely that there is an upper limit of 200 metres for olistostromes. 'This statement is not supported by any theoretical reason, but only by the habit of everyday practice' (ABBATE et al. – 1970).

Olistostromization can then be regarded as a sedimentary process, whereas the gravitational emplacement of sheets thicker than 200 metres can be considered to be a tectonic process.

4. TECTONICS

4.1. GENERAL

The rocks outcropping in the area studied have been assigned to three main stratigraphical sequences: the Liguride, Tuscan and Umbrian. The Tuscan Sequence overthrusts the Umbrian Sequence along a line running NW-SE that largely coincides with the Apenninic divide. This Tuscan unit is complicated by many internal upthrusts. The Umbrian Sequence in the area under investigation falls into two tectonic units: the Castel Guerrino unit and the (par)-autochthonous rocks comprising the Marnoso Arenacea, Gessoso Solfifera Formation and the Casal Fiumanese Formation. The Castel Guerrino unit is thrust upon the Marnoso Arenacea, which exhibits several NE vergent reverse faults and folds, all fading to the NW. The Tuscan and Umbrian rocks are overthrust by the Liguride Complex. A complication arises where the Castel Guerrino unit locally overthrusts rocks of the Liguride Complex. The overthrust of the Liguride Complex over the Umbrian Sequence dips to the NW, so that its outcrop line is perpendicular to the normal Apenninic strike. This is the result of a large scale transverse flexure: 'the Sillaro flexure'.

The development of the Sillaro flexure, the upthrusts and folds of the Marnoso Arenacea, the thrusts of the Castel Guerrino unit and of the Tuscan Sequence, and the overthrust of the Liguride Complex were all partly contemporaneous and synsedimentary.

SW of the Apenninic divide all strata are inclined 40° to 60° to the SW, towards the Mugello basin. This tilting probably dates from the late Pliocene when the Mugello basin was formed, as can be deduced from the Villafranchien lacustrine deposits of this basin.

4.2. THE SILLARO LINE

Along the Sillaro line the (par)autochthonous Miocene and Pliocene deposits are in contact with the allochthonous Liguride Complex. Field work made it clear that this is an overthrust contact which dips to the NW. In fact the area between the Sillaro line and the Marecchia depression represents an axial culmination. Not only the overthrust plane, but also the Miocene and Pliocene beds plunge to the NW, about 20° along the Sillaro line, and about 10° along the river Santerno (5–8 km more to the SE) (fig. 17). This general plunge

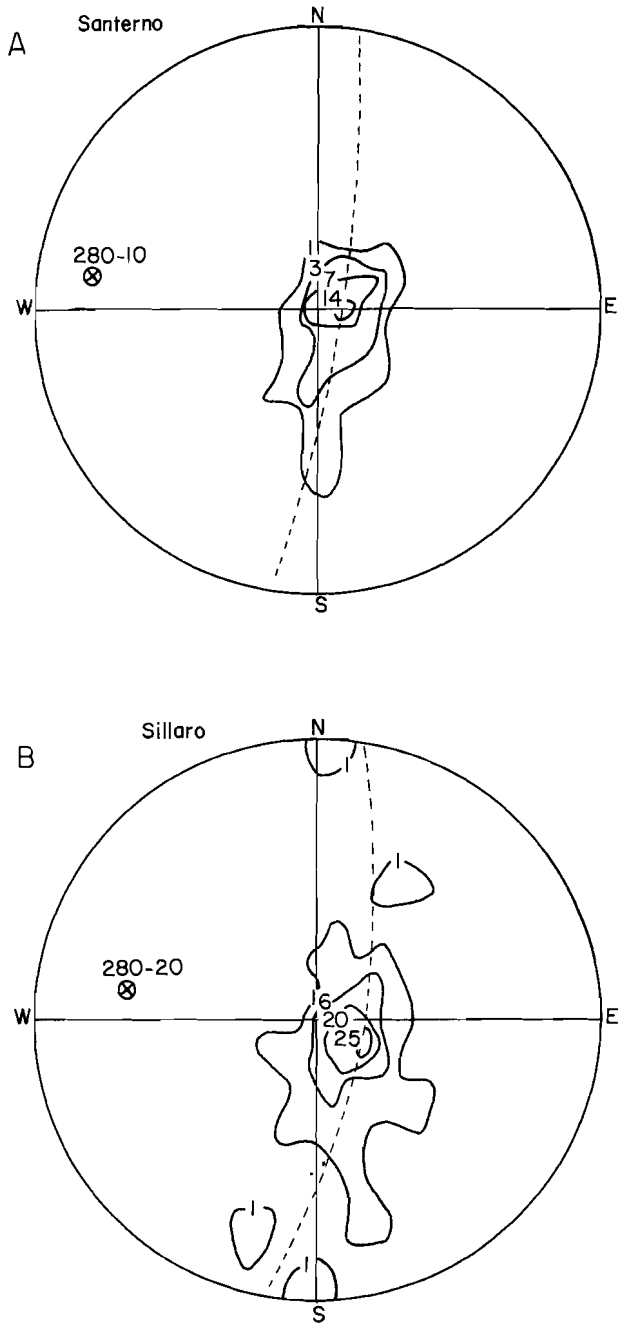

FIG. 17. Diagram of the normals on bedding planes, showing the general W-NW plunge along the Santerno (10°) and along the Sillaro (20°).

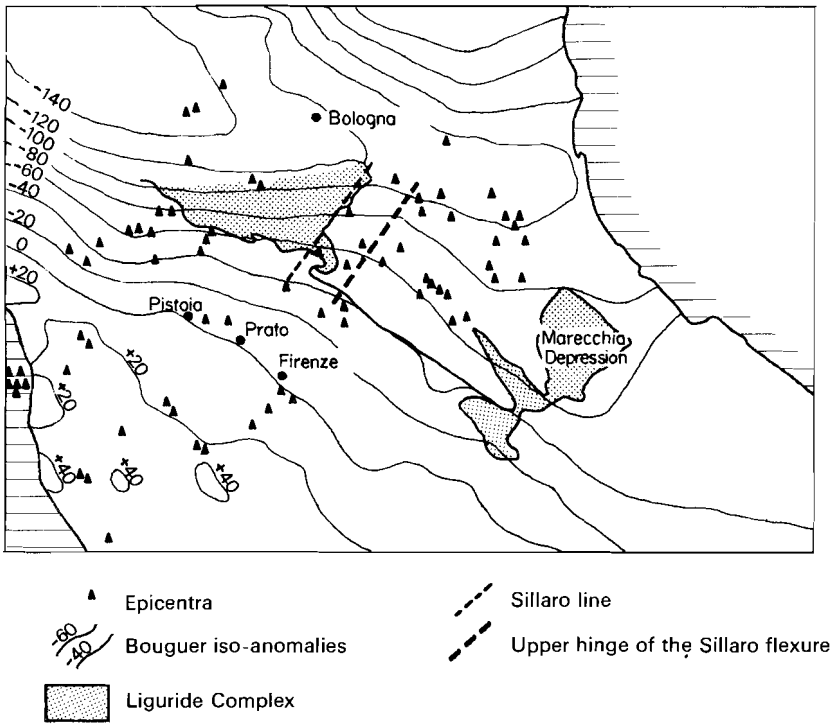

FIG. 18. Map of the Bouguer iso-anomalies and epicentra in the Northern Apennines. The upper hinge of the Sillaro flexure is marked by a distinct bending of the iso-anomalies.

accounts for the appearance of higher, Liguride units NW of the Sillaro line. It is not necessary to postulate a major fault or a number of minor faults along the Sillaro line.

Geophysical data also indicate the presence of a large scale flexure rather than a major fault. The Bouguer anomaly map shows a converging change in the direction of the iso-anomalies in the vicinity of the Sillaro line (fig. 18). This reflects a change in the direction of the slope of the basement: from NE to N. Moreover the distance between the iso-anomalies NW of the line is less than SE of it; this implies that NW of the line the dip of the basement is steeper. Apparently there has been a general subsidence of the basement NW of the line. In fact, the actual bend in the iso-anomalies is not at the Sillaro line, but some 15 to 20 km further to the SE, approximately along the river Senio. This is also where the superficial structures begin to plunge NW. So the actual flexural hinge is located there, while the Sillaro line, so conspicuous on a geological map, is merely the intersection of the Liguride overthrust plane with the present topography.

There are several indications that the Sillaro flexure developed synsedimentarily. First there is the fact that the Liguride Complex can be found in an over-

thrust position NW of the Sillaro line and not SE of it. (Except for the Marecchia depression, some 100 km to the SE). Theoretically it is possible that the Liguride Complex also overthrust part of the area between the Sillaro line and the Marecchia depression, but was eroded away as a consequence of greater uplift. Anyway, it can never have extended as far as the Po valley border, as it does not crop out there. And the absence of olistostromes and any other indications of the presence or nearness of Liguride material in that area makes it probable that the Liguride Complex never overthrust the (par)autochthonous very far. A reason for this may be that the slope towards that part of the basin was too slight to allow the Liguride Complex to invade the basin. The wedging out of Pliocene sediments towards the axial culmination (section 2.4) does indeed seem to indicate that at the time of the final advance of the Liguride Complex this area subsided at a lower rate than the surrounding parts of the basin. The difference in subsidence between the Santerno and the Sellustra, which are 5 km apart, was 900 m in 3 m.y. This caused a NW inclination of approximately 10° in the lowermost Pliocene deposits. The Messinian Gessoso Solfifera Formation and the two upper members of the Marnoso Arenacea (the Fontanelice and Borgo Tossignano Member) also peter out towards the axial culmination. However, it could not be established whether this is only a petering out of certain lithofacies or is a real thinning of the entire Upper Miocene. The transitions that can be seen in the Coniale Member of the Marnoso Arenacea from facies association *b* to *c* (in the SW) and from facies association *c* to *e* (in the NE) may also be caused by faster subsidence in the NW. But it is not certain whether these transitions are accompanied by a general decrease in thickness. GROSCURTH (1971) claims that a difference in thickness on both sides of the Sillaro line is observable in the Cervarola Sandstone Formation, 2000 m NW of the Sillaro line, and 1400 m SE of it. The deposits SE of the Sillaro line are also more marly than NE of it, which would be the result of the lower rate of subsidence. Although this argument can be adduced in favour of the proposed model, we are of the opinion that the information given by Groscurth must be regarded with some caution. The Cervarola Sandstone Formation is heavily tectonized, and truncated by a basal overthrust. Moreover, during its deposition it was still situated to the SW at an unknown but possibly considerable distance from the area we are considering.

Concluding one can say that in the rates and kinds of sedimentation during the Late Oligocene and the Miocene there is only slight evidence that the area NW of the Sillaro flexure subsided faster than the area SE of it, but there is clear evidence for such a difference in the rate of subsidence during the Pliocene. Besides, the overthrusting of the Liguride Complex which only occurred NW of the flexure makes it probable that differential subsidence had already started during the Miocene.

4.3. TECTONICS OF THE MARNOSO ARENACEA

4.3.1. General

The tectonics of the Marnoso Arenacea is characterized by NE vergent upthrusts, generally associated with a 'footwall syncline' (Ten Haaf and Van Wamel – 1979). In the direction of the Sillaro line, to the NW, these upthrusts pass into folds. The thrust which reaches furthest before this one too passes into a fold, the Castel Vecchio-Palazzuolo structure, will be treated more extensively in section 4.3.2. Other structures in the area investigated are the Diaterna anticline, which is probably genetic related to the Castel Vecchio-Palazzuolo structure (section 4.3.2) and the Rovigo syncline. Because of the presence of the Sillaro flexure all these folds plunge to the NW.

The sedimentology of the Marnoso Arenacea provided strong indications that the Castel Vecchio-Palazzuolo structure and the Rovigo syncline developed synsedimentarily (section 2.2.13). The Rovigo syncline is the continuation of the footwall syncline belonging to the upthrust of the Mandrioli Sandstones (fig. 11), so apparently the thrust of the Mandrioli Sandstones was also synsedimentary.

4.3.2. The Castel Vecchio-Palazzuolo structure

The Castel Vecchio-Palazzuolo structure can be followed over a distance of 75 km, from Castel Vecchio near the Sillaro line to the Marecchia area in the SE. The character of the structure changes along the strike. From a flat overthrust it passes towards the NW into an upthrust and finally an anticline. This anticline is first overturned to the N to NE, but still further to the NW it is not overturned any more (fig. 19). As the upthrust degenerates to a flat anticline, the strike changes from NW-SE to W-E. This is accompanied by a change in the direction of the sedimentary bottom features just S of the anticline (fig. 6) and by the formation of a secondary anticline (the Diaterna anticline) in the S, which has also a W-E strike. It has been concluded that the Castel Vecchio-Palazzuolo structure developed synsedimentarily (section 2.2.13). To explain the changing character of the structure along the strike, several lines of reasoning can be followed:

1. It is possible that near the Sillaro line where higher stratigraphic units crop out we see the cover of the synsedimentary structure. The younger deposits did not participate fully in the thrusting movement.
2. The fading of the thrusts towards the Sillaro line is general. It is possible that internal thrusting in the Marnoso Arenacea resulted in a tectonic thickening, and thus in a culmination, preventing the Liguride Complex from overthrusting. So it was the absence of thrusting that allowed the Liguride Complex to overthrust NW of the Sillaro line.
3. A third theory is that the presence of the Liguride Complex near the Sillaro line impeded the formation of a large upthrust there. The Liguride Complex

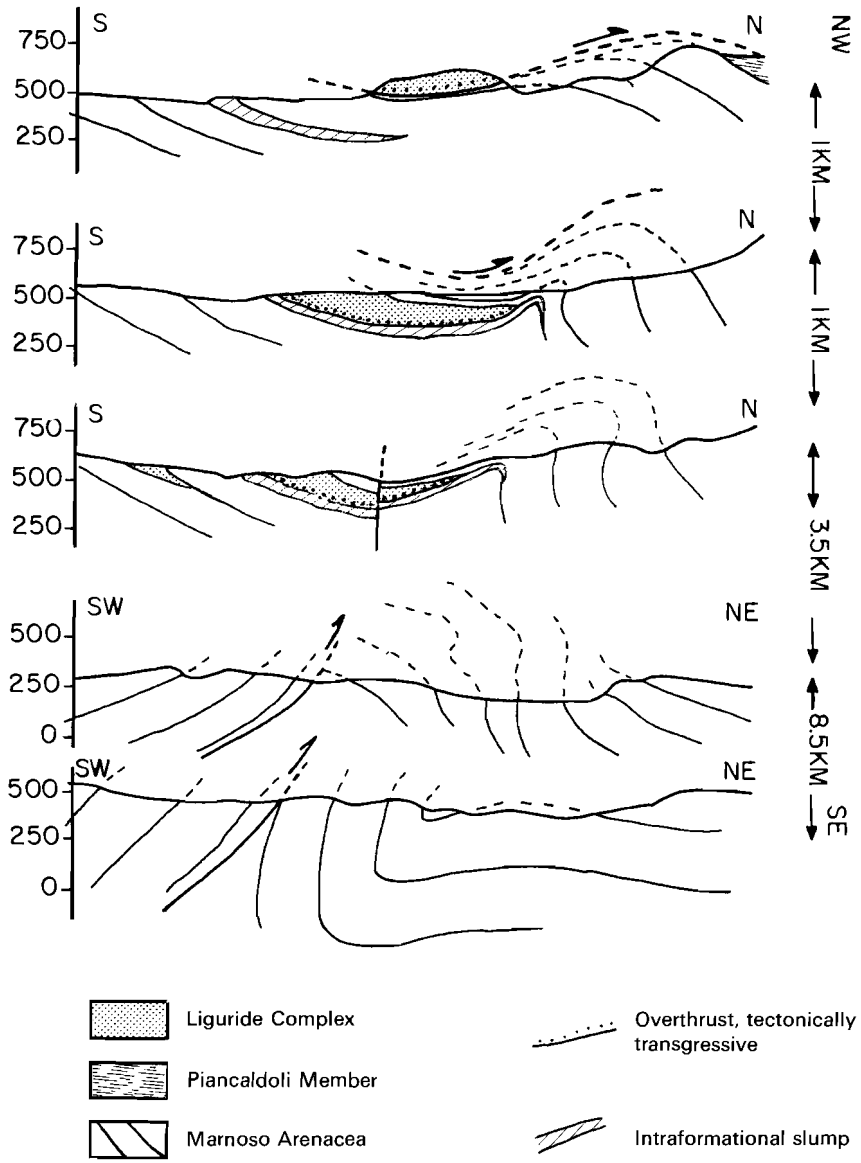

FIG. 19. The changing character of the Castel Vecchio-Palazzuolo structure along the strike, from a reverse fault in the SE (near Palazzuolo) to a fold in the NW (near the Sillaro line).

that overthrust the Marnoso Arenacea at an early stage of the formation of the Castel Vecchio-Palazzuolo structure may have acted as a heavy blanket that damped the structure.

The first theory is satisfactory as far as the synsedimentary character of the

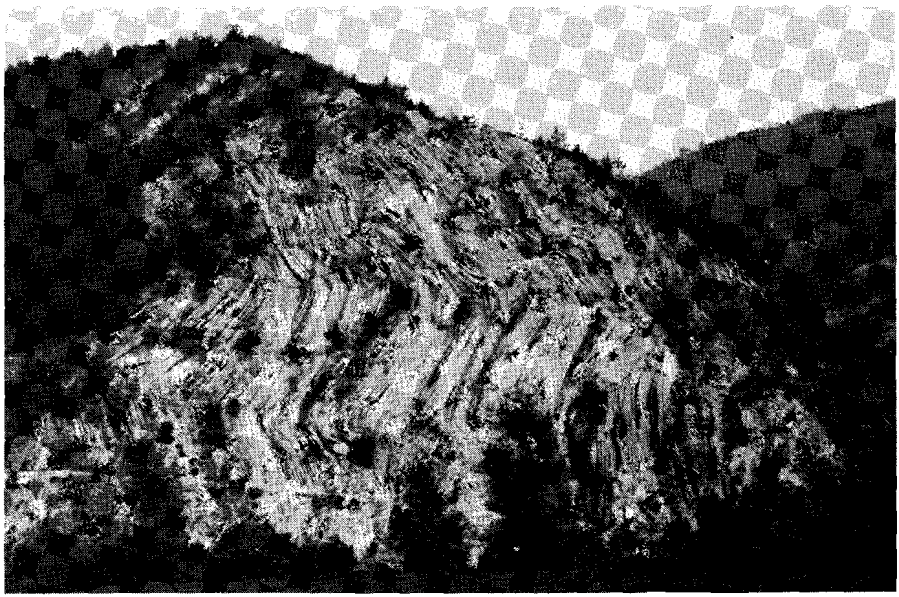

Photo 12. Nearly vertical beds in the Castel Vecchio-Palazzuolo structure in the Santerno valley.

structure is concerned. But is does not explain the deviating direction of the bottom features of the younger deposits just S of the Castel Vecchio area if they did not participate in the thrusting. The directions of these bottom features are best explained if one assumes that they rotated as a result of the fading of the thrust to the NW. The more south-eastern part of the hanging wall moved more to the NE, causing a rotation of the strike of the structure as well as of the direction of the bottom features. If this is true, it implies that the main thrusting took place after the deposition of these younger deposits, and that the thrust had an amplitude of several kilometres.

The second theory fails since the axial culmination between the Sillaro line and the Marecchia depression originates in the basement (section 4.2), and not in a superficial thrusting. Besides, the thrusts do not fade towards the Marecchia depression, and SE of its Liguride cover the same thrusts can be found again.

The third theory does not explain the general fading of the thrusts to the NW. However it is possible that the only structure that extends as far as the Sillaro line dies out very suddenly because of the presence of the Liguride Complex. Other thrusts pass into folds more gradually.

Although none of the theories is satisfactory, we are inclined to believe that the third one is most promising. This implies then that the Castel Vecchio-Palazzuolo structure was at the time of the overthrusting still an anticline, and that it developed into an upthrust after the overthrusting of the Liguride Complex.

4.4. TECTONICS OF THE CASTEL GUERRINO UNIT

Between the P^{so} del Giogo and the Monte Falterona, the Castel Guerrino unit thrusts upon the Marnoso Arenacea. This thrust is accompanied by a large NE vergent footwall syncline. Between the P^{so} del Giogo and the P^{so} della Futa the situation is more complicated. Here the Castel Guerrino unit overlies the Liguride Complex, which in turn overlies the Marnoso Arenacea (fig. 20). The underlying Marnoso Arenacea lies undisturbed, dipping about 10° N, while the Castel Guerrino unit is highly disturbed by reverse faults and overturns. The front of the Castel Guerrino unit NW of the P^{so} del Giogo is 5 km more to the NE. Whether this is the result of a faster erosion of the zone SE of the P^{so} del Giogo which emerged to greater heights, or of a larger amount of overthrusting NW of the P^{so} del Giogo could not be established. In the latter case a wrench fault must have been present to separate the two compartments. Such a fault could not be located, nor could its absence be proved. On a smaller scale, an analogous situation can be seen near Rifredo, 2 km N of the P^{so} del Giogo. There a wrench fault separates an area with overturned thrust slices from an area where overturns and faults are absent. The inferred origin of this situation is schematically indicated in fig. 21. Rocks belonging to the Liguride Complex and the Rifredo Marls are found along one of these internal upthrust planes. These rocks have probably been brought up from the footwall during the thrusting.

The overthrusting of the Castel Guerrino unit on the Liguride Complex must have occurred after the arrival of the Liguride Complex in this area.

4.5. TECTONICS OF THE TUSCAN SEQUENCE

Only the frontal zone of the Tuscan thrust sheet is exposed in the area studied. The Tuscan Sequence overthrusts the Castel Guerrino Formation of the Umbrian Sequence. The front of the Tuscan thrust sheet exhibits tectonic imbrication, which is characteristic for gravity tectonics. The thrust slices have thicknesses of several hundreds of metres. Lenses of tectonized Scaglia Toscana are found at the base of most thrust slices. This clayey formation served as a lubricant to facilitate the overthrusting. Besides, the clays of this formation may have acted as an impermeable seal to maintain a high pore pressure in the footwall formations. As HUBBERT and RUBBEY (1959) made clear, a high pore pressure considerably facilitates overthrusting.

The tectonic imbrication at the front of thrust sheets is explained as follows (fig. 22). Usually overthrust planes tend to curve smoothly upwards at the front of the thrust sheet to reach the topographic surface at a steep dip (± 60°). This frontal zone is called the toe (RAYLEIGH and GRIGGS – 1963). Pushing from behind forces the toe upwards. The volume (and the weight) of the toe will increase, causing an increase in the friction along the thrust plane beneath the toe, and a decrease in the potential energy of the thrust sheet as a whole.

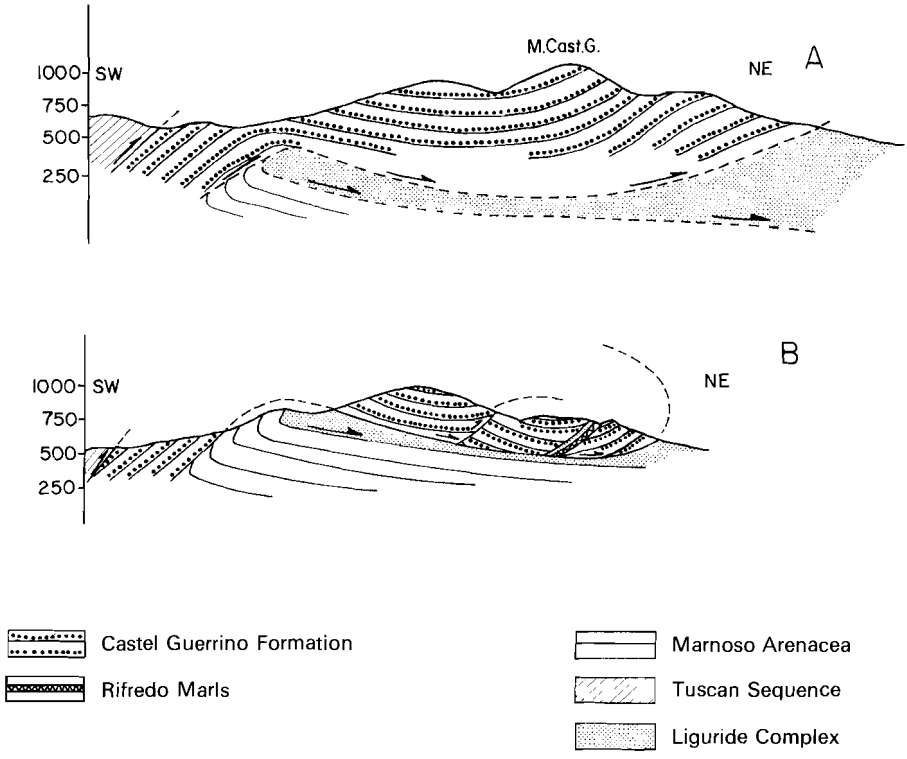

FIG. 20. Two sections through the Castel Guerrino unit. Section A over the Monte Castel Guerrino, and section B more to the SE showing overturned thrust slices.

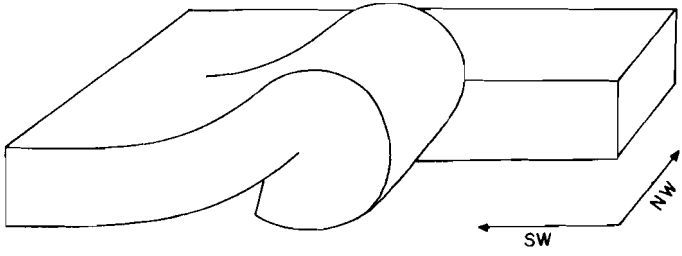

FIG. 21. Schematic representation of the inferred origin of the overturns as shown in fig. 20 B.

Eventually the thrust movement is either completely stopped or must seek a new break behind the old toe (RAYLEIGH and GRIGGS – 1963). Apparently this has happened repeatedly with the Tuscan overthrust in the area studied.

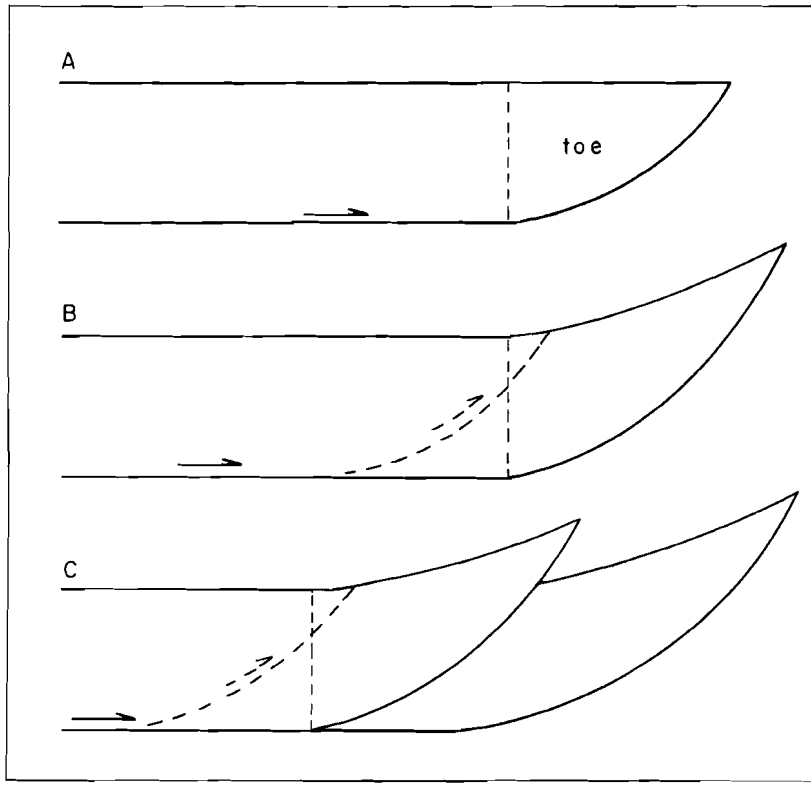

FIG. 22. Development of tectonic imbrication at the toe of a thrust sheet (after RAYLEIGH and GRIGGS – 1963, explanation in text).

4.6. SPECULATIONS ON THE DYNAMICS OF THE LIGURIDE COMPLEX

The Liguride Complex has overthrust the entire width of the Apennines. The incompetent nature of the Argille Scagliose, which makes up the bulk of the Liguride Complex, means that it is highly improbable that compression was the driving force. In fact, all current investigators agree that the driving force involved must have been gravitation. This is reflected in the different genetic names given to the Liguride Complex by the different authors: Fliessdecke (WIEDENMAYER – 1951), Frana orogenica (MERLA – 1952), Thrust flow and tilt flow (BUCHER – 1956), Olistostrome (GÖRLER and REUTTER – 1968), Olisthanite (DE RAAF – 1968), Olisthon (HOEDEMAEKER – 1973). As some of these names indicate, most authors think that the Liguride Complex behaved as a mass that could flow because of the large amount of clay it contains.

The behaviour of material is different under 'flowing' conditions and under rigid conditions in that in the former shear does not take place along specific

shear planes, but is general throughout the mass. This implies that the theories of HUBBERT and RUBBEY (1959) and HSÜ (1969) concerning the gravitational emplacement of rock masses are not fully applicable in this case. Their theories relate to rigid rock masses, but the Liguride Complex is not a rigid mass.

Experiments carried out by BUCHER (1956, 1962) and RAMBERG (1967) showed that under the influence of gravity rock masses can spread out from regions of higher topographic elevation to produce thrusts and recumbent folds. PRICE and MOUNTJOY (1970) used this theory of 'gravity spreading' to explain the structures of the foreland belt of the Canadian Rockies. ELLIOT (1976) developed this theory mathematically, making use of glaciological findings. NYE (1952) produced a theory concerning the flow of glaciers and ice-sheets, based on the assumption that glacier ice can be regarded as a perfectly plastic material. Shear stress beyond the yield limit causes plastic material to flow rather like a very viscous fluid. The results of Nye's computations agree very well with reality. ELLIOT (1976) states that this theory for plastic material is also applicable to rocks, provided the rock strength T_m is less than the weight of the overburden: $T_m < \rho gh$. As the rock strength of clay is only a few bar (HSÜ – 1969) this formula holds even at depths of some tens of metres. So there seems to be no reason why the theory of gravity spreading should not be applicable to the dynamics of the Liguride Complex.

Differences in the thickness (h) of a thrust sheet cause differences in lithostatic pressure. This gravity-induced pressure gradient is the force that causes the movement. When the surface slope of the mass (α) and the basal slope are slight, then the shear stress at the base is: $\tau = \rho gh\alpha$ (NYE – 1952). It is therefore the surface slope that determines both the sense and the magnitude of the shear stress at the base, irrespective of whether the dip of the base is in a direction that is the same or opposite to the surface (ELLIOT – 1976).

Changes in the thickness and surface slope of the spreading mass will cause differences in the basal shear stress. Decreasing basal shear stress in the direction of flow will cause compression: the flow is then referred to as being passive or compressing. Increasing basal shear stress will cause extension: then the flow is called active or extending. The higher parts of the flow will always have the greatest velocities, because the flow will encounter the greatest friction near its base. In a compressing flow the slip lines within the flow tend to be concave upwards, in an extending flow they tend to be concave downwards (NYE – 1952) (fig. 23). When the thickness of the flow is constant, an extending flow will be found on a convex footwall and a compressing flow on a concave footwall. Near the toe of the flow where the thickness decreases, the basal shear stress also decreases, so a compressing flow will be found there. Where both thickness and dip of the surface of the flow are constant, the basal shear stress will be constant, so neither extension nor compression will occur. In that case the slip lines will run parallel to the upper and lower surfaces.

Ideally, in plastic material the flow is general throughout the mass. One could imagine however that a zone of very concentrated shear strain or shear fracture or fault could be formed. This would be even more likely if high shear

strain had the effect of weakening the material for further shear. This can happen with ice, as well as with clay. The experiments of SKEMPTON (1964, as cited by HOEDEMAEKER – 1973) on the deformation of clay resulted in the shear strength-strain diagram of fig. 24. With an increasing shear strain the shear strength at first rises considerably, until a maximum value (M) is reached. Then the shear strength diminishes to a lower residual value (R). With further increasing strain the strength remains the same. This behaviour is explained as follows. In clay shear will cause an orientation of clay particles parallel to the flow lines. This will lower the friction, with the result that further strain will occur preferentially in zones already sheared. These zones of decreased strength will be determined by local irregularities. These irregularities may be external such as humps in the footwall, as well as internal. Internal irregularities may be:
a. large rigid slabs.
b. variations in the pore fluid pressure.
ad. a: The vertical velocity gradient within a large undeformed slab is zero. In the surrounding parts of the Liguride Complex the velocity will increase gradually with increasing height, with the result that a zone of increased shear will be created at the base of the olistolithe (fig. 25).
ad. b: The rock strength diminishes with increasing pressure of the pore fluids. So shear strain will first appear in zones of high fluid pressure, which causes a weakening of the rocks. Slip lines will follow zones of high fluid pressure, even if the latter do not follow precisely the maximum shearing stress. Zones of high fluid pressure will occur at the base of the Liguride Complex. The turbiditic sediments of the footwall were water-saturated, and were not compacted. The fluid pressure will rise under the weight of the Liguride Complex, because of the impermeability of the Argille Scagliose.

Competent slabs within the Liguride Complex will be orientated parallel to the slip lines. So in a 'fossilized' compressing flow the olistolithes will be imbricated, dipping upstream, and in a 'fossilized' extending flow they will be imbricated, dipping downstream. This is in agreement with the observation that in the Marecchia area almost all large slabs (mostly San Marino Formation) dip to the SW with inclinations of 10° to 60°. It is also consistent with the observations of SIGNORINI (1956), according to whom there is a strong tendency for the olistolithes of the Liguride Complex near the NE border of the Northern Appennines to be inclined to the SW. These observations point to a compressing flow, which is the type of flow that is to be expected near the front of the Liguride Complex. The larger olistolithes in the SW part of the Northern Appennines are mostly inclined to the NE (SIGNORINI – 1956), indicating an extending type of flow in that area. The position of the slip lines will probably be best indicated by the larger olistolithes. The smaller ones may very well be situated between two zones of increased shear-strain, while in between them the strain is negligible.

There can be no gravity sliding without a slope. The Liguride rocks have travelled 200 km or more. If we imagine a continuous slope of 5°, the altitude

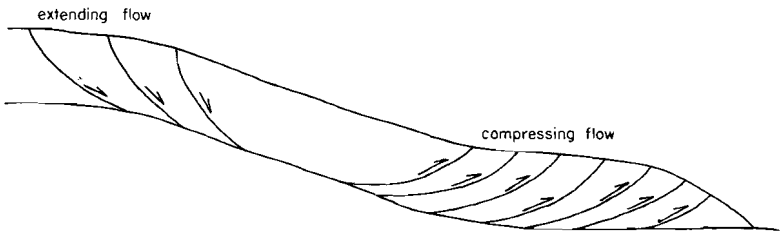

FIG. 23. The types of flow that may have occurred in the Liguride Complex: extending flow on a convex footwall, compressing flow on a concave footwall and at the toe.

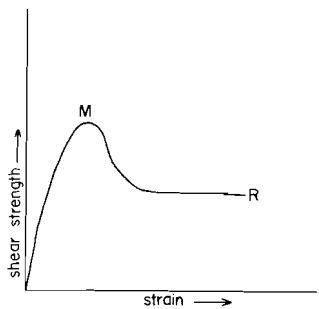

FIG. 24. Shear strength-strain curve for clay (after SKEMPTON as cited by HOEDEMAEKER – 1973). Explanation in text.

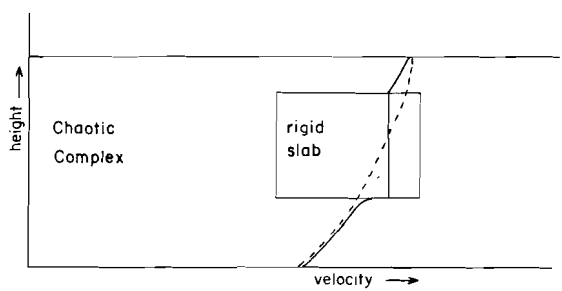

FIG. 25. Presumed velocity in the Liguride Complex (---), and when a large rigid slab is present (———).

of 'departure' would be more than 15 km. Even with a slope of only 2°, the altitude of departure would still be 7 km, which is considered to be too high. Besides, a continuous slope with a length of 200 km will never have existed.

According to the theory of gravity spreading the overthrust plane of the Liguride Complex need not to have been inclined. But because according to the theory of gravity spreading the upper surface must have been inclined, we are still left with the problem of length of slope versus difference in height. The only possible explanation seems to be that the Liguride rocks slid down the front of an advancing orogenic wave, as was proposed by MERLA (1952). This implies that the Liguride Complex probably moved generally down a slope. This does

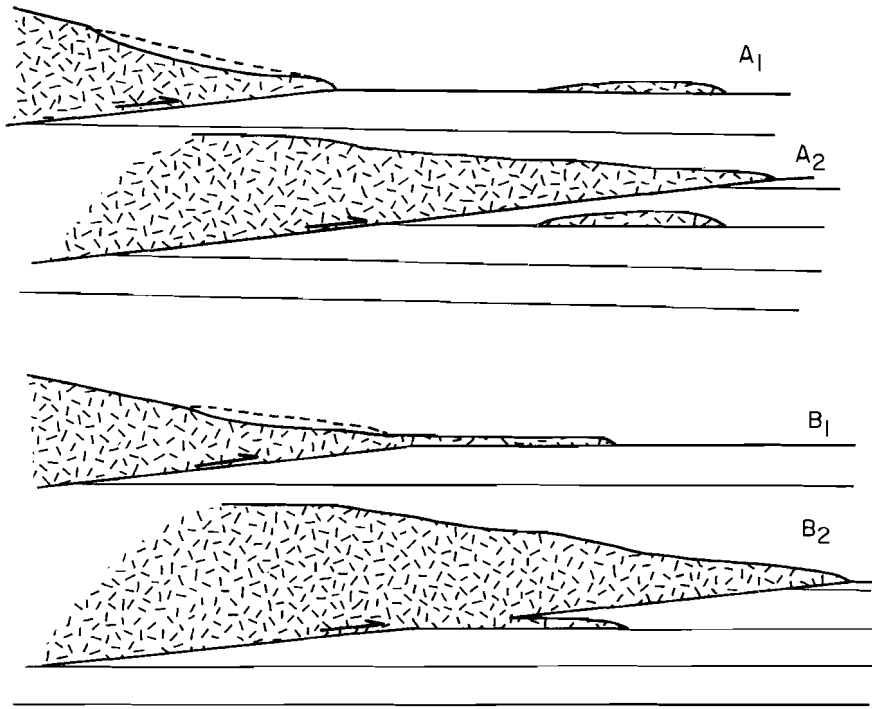

FIG. 26. Olistostromes may become completely detached from the Liguride Complex (A_1 and A_2), others may still be attached to it (B_1 and B_2).

not mean that the mechanism of gravity spreading was not active. Also when the Liguride Complex moves down a slope, the plasticity of the clay (Argille Scagliose) will allow it to spread by virtue of its own weight.

Gravity spreading and sliding are not the only mechanisms by means of which the front of the Liguride Complex advanced. Olistostromization also played an important role. Olistostromes consisting of material from the Liguride Complex are at present found intercalated in the underlying sediments (fig. 26 A). Not all olistostromes that occurred became completely detached from the Liguride Complex and involved in the autochthonous sedimentation. Probably many olistostromes can no longer be recognized as such because they were overridden again by the Liguride Complex, and possibly they even participated again in the spreading movement (fig. 26B). GÖRLER and REUTTER (1968) suppose that the Liguride Complex is composed entirely of small olistostromes and that Liguride Complex did not move as a whole, but that movement was restricted to the uppermost layer. The author does not share this opinion. Small olistostromes cannot possibly have transported the large exotic slabs that occur in the Liguride Complex. For example the sizes of the Monghidoro slab, just outside the investigated area are: $20 \times 15 \times 3$ km. It is more likely that olistostromization was superimposed on a general movement.

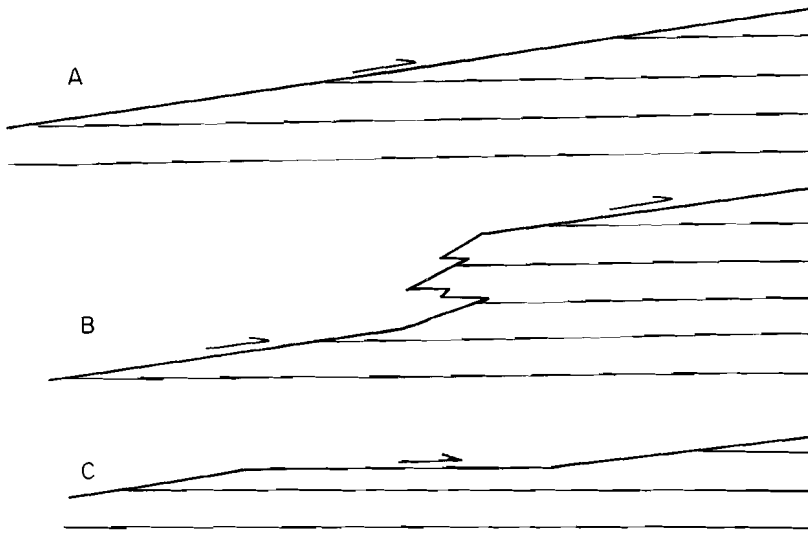

FIG. 27. When the velocity of the Liguride Complex and the sedimentation rate of the autochthonous deposits are constant, the overthrust 'climbs' gradually in the stratigraphy (A). A temporary standstill of the Liguride Complex causes interfingering (B). A fast progress causes the overthrust plane to follow the same stratigraphical level for a certain distance.

The foregoing is however not true for the Liguride Complex in the extreme N of the area studied, where it is found above the Pliocene deposits. While the Chaotic Complex above the Marnoso Arenacea is completely chaotic, above the Pliocene deposits it exhibits internal layering (section 2.7.2). Besides, the Liguride Complex there contains lenses of meso-autochthonous deposits, originally deposited upon the advancing mass. Their occurrence especially near the base of the Liguride Complex, and its internal layering suggest that the Liguride Complex there was progressively built up by olistostromes. Olistostromes could deposit the higher parts of the Liguride Complex (with meso-autochthonous deposits) in front of the main mass. Because olistostromization continued, the first olistostromes made up the base of the olistostrome pile that was formed.

The mechanism by which the Liguride Complex advanced is thus quite different from 'normal' overthrusting. Olistostromization, usually considered as a sedimentary process, may have sometimes been an important mechanism. The Liguride Complex overlies from SW to NE progressively younger deposits, with which it locally interfingers. MERLA (1952) therefore said that the Liguride Complex was tectonically transgressive. This view is upheld here, and for the notation on the geological map of the contact between the Liguride Complex and its footwall formations we have chosen the classical notation for an unconformity but in red, the colour of tectonics.

The velocity of the Liguride Complex

Near the P^{so} del Giogo the Liguride Complex overlies deposits of Middle Miocene age; near Gesso the deposits are of Messinian age. The distance between these two points is about 30 km, a distance covered in about 6 or 7 m.y. So the average velocity of the Liguride Complex was about 4 to 5 m/10^3 y. This implies that whenever both the velocity of the Liguride Complex and the sedimentation rate of the Marnoso Arenacea were constant, the Liguride Complex after advancing 100 m, overlies deposits at an 11 to 15 m higher stratigraphical level. A temporary stop in the progress of the Liguride Complex causes an abrupt rise in the stratigraphical level of the overthrust plane, whereas fast progress, geologically instantaneous, causes the overthrust plane to follow the same stratigraphical level for a certain distance (fig. 27). By giving attention to such features one must conclude that during the Middle Miocene the Liguride Complex experienced a major stagnation after its front had reached the Firenzuola-Frena line. The overthrust plane NE of this line is on a stratigraphical level that is about 300 m higher than it is SW of the line (fig. 8). The sedimentation rate there and then was probably somewhat higher than the mean sedimentation rate in the inner basin; even so, this implies that the stagnation lasted about 3.10^5 to 4.10^5 years. This is ascribed to the fact that NE of this line the Liguride Complex had to move uphill towards the Castel Vecchio high (section 2.2.13). Closer to the Castel Vecchio high it is difficult to prove there was a lowered velocity, because there the sedimentation rate was considerably lower than the mean value.

5. CONCLUSIONS

5.1. INTRODUCTION

Conclusions and assumptions are based on facts but this does not imply that the conclusions and assumptions are also facts. Only very rarely is a conclusion drawn in geology absolutely true. Generally the facts on which a conclusion is based leave room for alternatives, even if these are regarded as less probable. Therefore in this paper the following definitions will be given of the words 'conclusion' and 'assumption': A conclusion is based on many facts, and, in the eyes of the investigator, possible alternatives are very improbable. An assumption is not intrinsically more probable than alternative possibilities, but the investigator is inclined to choose the solution that best fits the general picture.

5.2. CONCLUSIONS

The Marnoso Arenacea basin, like all turbidite formations in the area studied, was filled with turbidites supplied from the NW (section 2.2). The course of the turbidity currents was determined by the elongate NW-SE form of the basin and by synsedimentarily developing anticlines and synclines such as the Castel Vecchio-Palazzuolo structure and the Rovigo syncline, which created 'highs' and 'lows' and by the synsedimentary overthrusting of the Liguride Complex (section 2.2.13). The coarsest and thickest turbidites were usually deposited in the deepest part of the basin (Rovigo syncline) or in the proximal inner fan environment (Fontanelice Member). Marls and very dilute turbidites were usually deposited on highs within the basin (Castel Vecchio Member, and upon the Casaglia Member) or on the basin slope (Borgo Tossignano Member), and in front of the Liguride overthrust (Osteto Member, Frena Member, Ca Buraccia Member, Piancaldoli Member) (section 2.2.13). The Castel Vecchio high divided the basin into an inner and outer basin (section 2.2.13). The Castel Vecchio high occasionally reached sea level (section 2.2.13). The inner basin was invaded during the Middle Miocene by a 'klippe sédimentaire' (Casaglia Member), which slid from the SW to the basin axis (section 2.2.13).

Synsedimentarily, a large scale transverse flexure developed near the present river Senio (the Sillaro flexure), involving the basement of the Northern

Apennines (section 4.2). Both the autochthonous sedimentation and the overthrusting of the Liguride Complex were influenced by this flexure. The thickness of the Casal Fiumanese Formation increases to the NW, and the Liguride Complex overthrust only NW of this flexure (section 4.2).

In the SE of the area studied the Castel Guerrino Formation first thrust upon the Marnoso Arenacea, and later overthrust Liguride rocks that had overthrust the Marnoso Arenacea in an earlier stage (section 4.4).

The Castel Guerrino Formation is overthrust by the Tuscan Sequence. The frontal upthrusts of this overthrust are to be found in the extreme S of the area studied (section 4.5).

The Castel Vecchio-Palazzuolo structure is an internal upthrust of the Marnoso Arenacea which fades into an anticline to the NW (section 4.3.2). This structure started developing in an anticline synsedimentarily (section 2.2.13 and 4.3.2). The development of the Diaterna anticline is connected with the Castel Vecchio-Palazzuolo structure (section 4.3.2).

After the deposition of the Messinian gypsum a short period of erosion occurred (section 2.3 and 2.4).

The overthrusting of the Liguride Complex over the Marnoso Arenacea was mainly a process of gravity sliding and spreading, but over the Casal Fiumanese Formation it was a process of olistostromization (section 4.6).

5.3. ASSUMPTIONS

The Castel Guerrino Formation was deposited in the same turbidite basin as the Marnoso Arenacea (section 2.5.4). The Rifredo Marls belong to the Castel Guerrino unit; they were deposited between the Castel Guerrino Formation and the Marnoso Arenacea on a 'high' created by the synsedimentarily developing upthrust between the two formations.

The Violla Marls were deposited in the front and on the frontal zone of the slowly advancing Liguride Complex, and later became intermixed with the Liguride Complex (section 2.7.6).

The 'olistostrome' intercalated in the Castel Vecchio Member should probably be regarded as an advanced lobe of the Liguride overthrust, which did not became detached from the main overthrusting mass.

The allochthonous marls of the Casaglia Member belong to the Tuscan Sequence (Vicchio Marls), and its shaly intercalations belong to the sub-Liguride Canetolo Complex (section 2.2.13).

The Castel Vecchio-Palazzuolo structure developed synsedimentarily in an anticline, after the overthrusting of the Liguride Complex it developed further into an upthrust (section 4.3.2).

The amount of overthrusting of the Castel Guerrino Formation to the NE is less in the SE.

5.4. GEOLOGICAL HISTORY OF THE SILLARO AREA

The geological history of the area investigated can be divided into nine stages, designated by a to i (fig. 28 a–i).

STAGE A – Late Burdigalian-Early Langhian

The oldest known autochthonous deposits are those of the Coniale Member of the Marnoso Arenacea, facies association *b* cropping out near Coniale and the P^{so} del Giogo, and passing to facies association *c* to the SE. Of the same age are the turbidites of the Castel Guerrino Formation (facies association *c*), which were deposited just SE of the P^{so} del Giogo in the SW part of the Marnoso Arenacea basin.

STAGE B – Early-Middle Miocene

tectonics

The Tuscan Sequence, carrying the Liguride overthrust on top of it, starts overthrusting the Castel Guerrino Formation. An anticlinal structure develops between the Castel Guerrino Formation and the Marnoso Arenacea, as well as further to the NE along the line Castel Vecchio-Palazzuolo.

sedimentation

The synsedimentary anticlines create 'highs' in the turbidite basin. Upon these highs, marls (facies association *f*) are deposited: the Rifredo Marls on the high in the SW, the Castel Vecchio Marls on the high in the NE. The basin in between these two highs, the inner basin, has its axis probably just NE of the P^{so} del Giogo. There facies association *b* is deposited, passing laterally (to the NE) and distally (to the SE) through facies association *c* into *e*.

STAGE C – Middle Miocene

tectonics

The Liguride overthrust advances to the NE, and overthrusts the Tuscan Sequence, the Castel Guerrino Formation and the south-western part of the Marnoso Arenacea. The basin NW of the river Senio subsides faster than SE of it as a result of a synsedimentary transverse flexure: the Sillaro flexure. The Liguride overthrust advances only NW of this flexure. After being overthrust by the Liguride Complex the anticline between the Castel Guerrino Formation and the Marnoso Arenacea develops into an overthrust of the Castel Guerrino Formation on the Marnoso Arenacea. The Castel Vecchio-Palazzuolo anticline continues to develop.

sedimentation

Just in front of the Liguride overthrust the Osteto Marls are deposited. Upon the Castel Vecchio-Palazzuolo anticline facies association *f* is still being

FIG. 28 a.

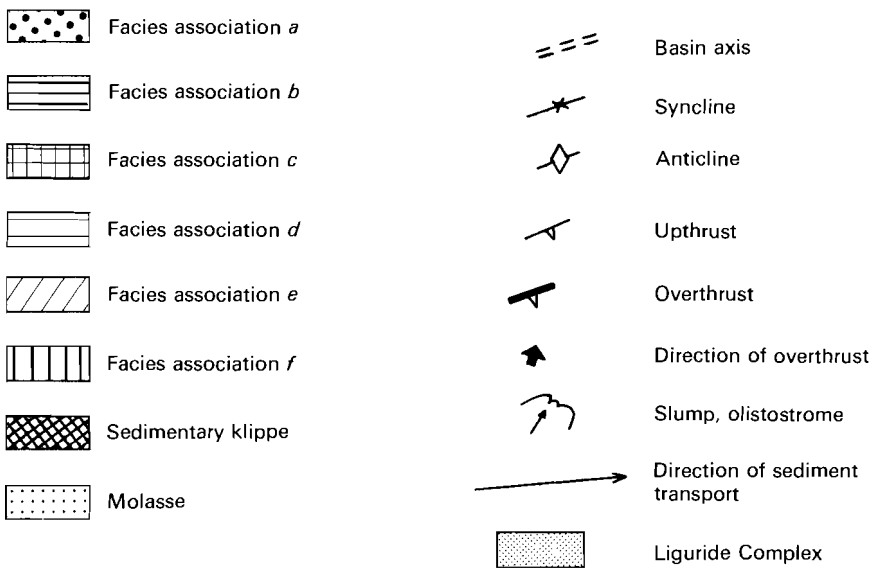

FIG. 28. Legend to fig. 28 a–i. Geological history of the Sillaro area.

FIG. 28 b.

deposited. The basin axis of the inner basin has shifted somewhat to the NE. In the deepest part we still find facies association *b*, passing to the SE and NW in facies association *c*. Just SW of the Castel Vecchio-Palazzuolo anticline facies association *e* is deposited. The oldest known deposits of the outer basin (NE of the Castel Vecchio-Palazzuolo anticline) date from this time; they have facies association *d*.

STAGE D – Middle Miocene

tectonics

The Sillaro flexure is still active and the Castel Vecchio-Palazzuolo anticline is still developing. The Castel Guerrino Formation now starts to overthrust the Liguride rocks and the Rifredo Marls in front of it. The amount of over-

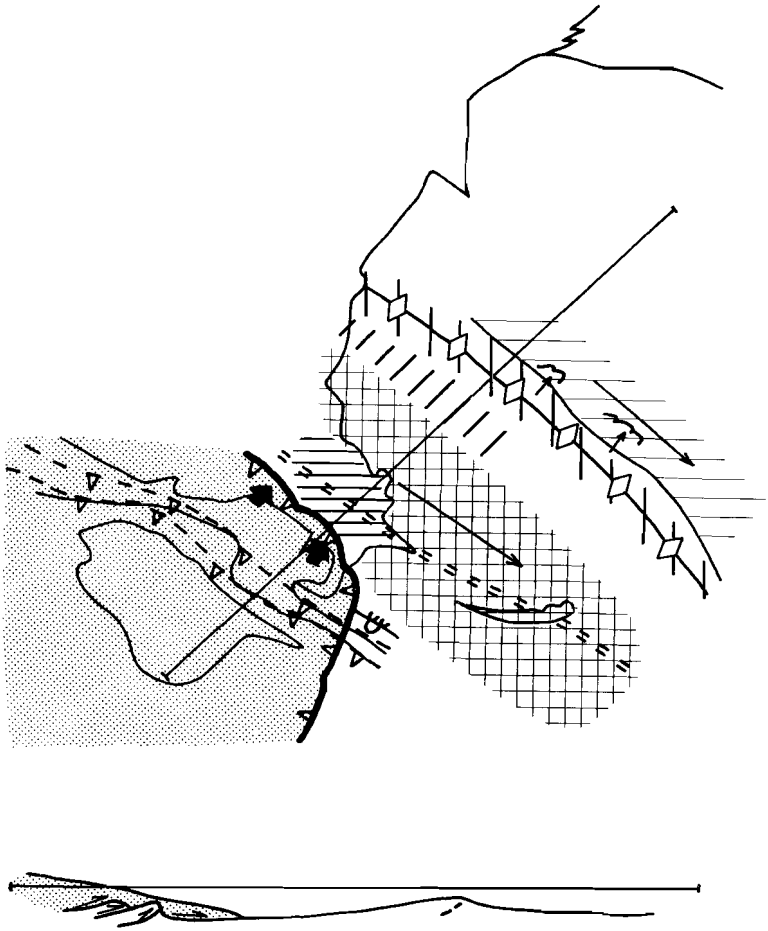

FIG. 28 c.

thrusting is less in the SE. From S of the P^{so} del Giogo a large slab of marls slides into the Marnoso Arenacea basin, and comes to rest near the basin axis. This slab is underlain by an extensive intraformational slump. It is tempting to suppose that this slide and slump were triggered by the overthrusting of the Castel Guerrino Formation, with the Tuscan Sequence on top of it.

sedimentation

Upon the marls of the allochthonous slab several tens of metres of marls (facies association *f*) are deposited, succeeded by facies association *b*, which passes again to the NE into facies association *e* and eventually *f* on the Castel Vecchio-Palazzuolo anticline. In the outer basin facies association *d* is still being deposited. Slumps descend from the Castel Vecchio-Palazzuolo anticline and alternate with these deposits.

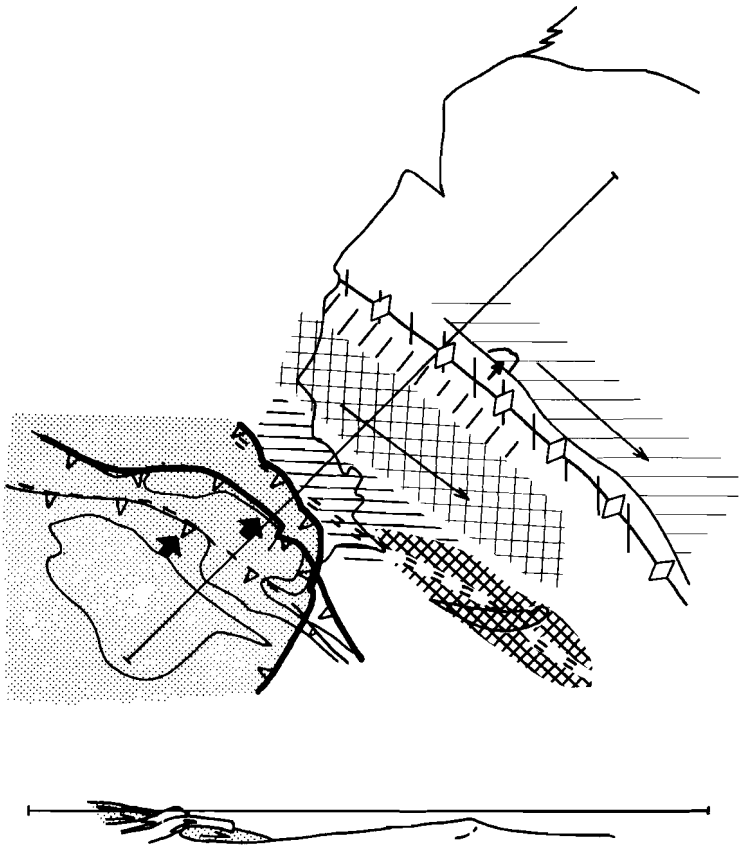

FIG. 28 d.

STAGE E – Middle Miocene

tectonics

The front of the Liguride overthrust advances slowly across the deeper parts of the inner basin. The Sillaro flexure and the Castel Vecchio-Palazzuolo anticline remain active.

sedimentation

In the deeper parts of the inner basin we find facies association *b* passing to the NE into facies association *c*, *e* and *f* as before. In the outer basin too the situation, as far as it can be reconstructed, remains the same. In front of the Liguride overthrust facies association *g* (the Frena Marls) is deposited.

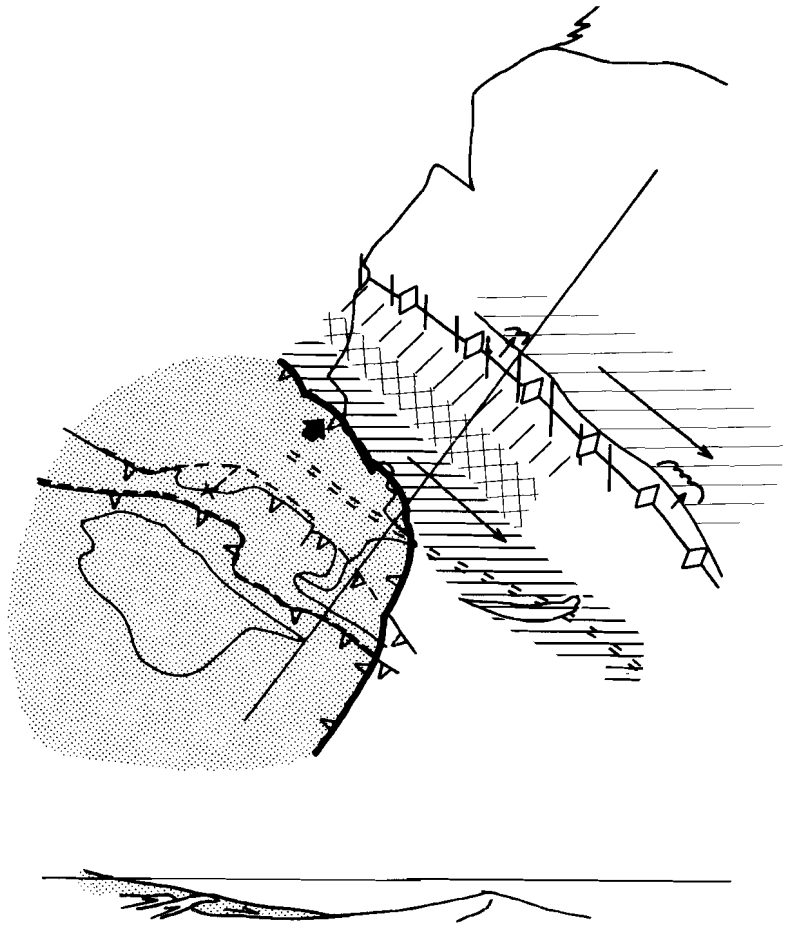

FIG. 28 e.

STAGE F – Late-Middle Miocene

tectonics

The Sillaro flexure is still active. The Liguride overthrust reaches the Castel Vecchio area, where an advanced lobe of the overthrust mass becomes intercalated in the autochthonous deposits. The anticline between the inner and outer basin continues to evolve.

sedimentation

On the anticline facies association *f* is still being deposited. With the arrival of the Liguride overthrust in the Castel Vecchio area, the inner basin is closed to turbidity currents from the NW. In the outer basin the deposits can still be classified as facies association *d*, although the turbidites slowly become thicker

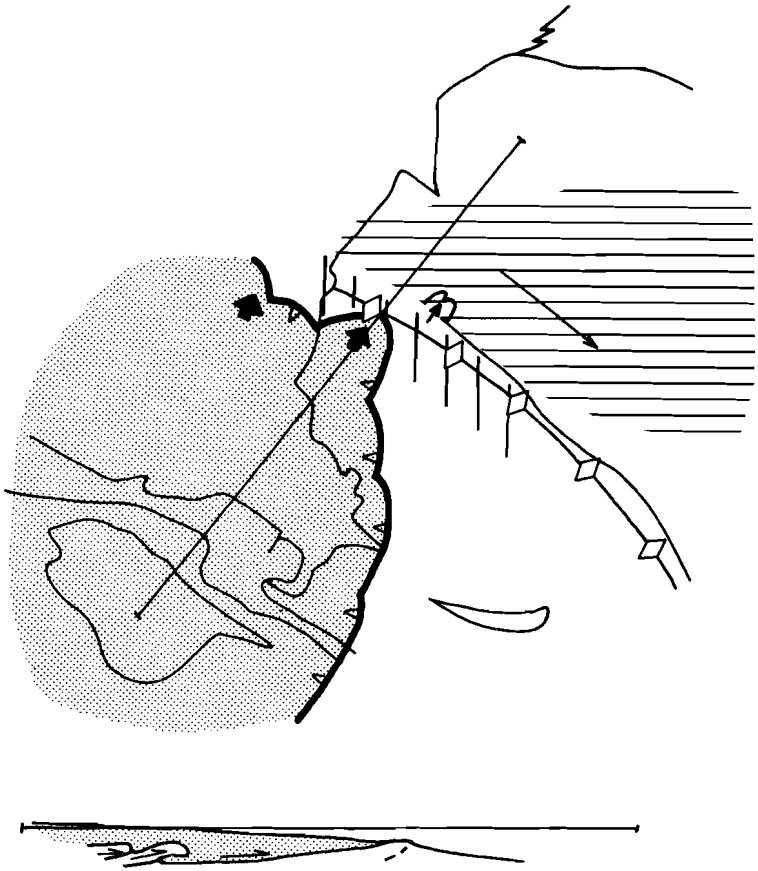

FIG. 28 f.

and coarser: facies association *b*. The outer basin is still invaded from time to time by slumps descending from the Castel Vecchio high. Just in front of the Liguride overthrust the Piancaldoli Marls are deposited (facies association *g*).

STAGE G – Tortonian-Messinian

tectonics

The Liguride Complex continues to advance.

sedimentation

On the NE slope of the Marnoso Arenacea basin an inner fan develops, resulting in facies association *a*, supplied from the NNE. Then quite suddenly the supply of coarse detritus stops, and only marls (facies association *f*) are

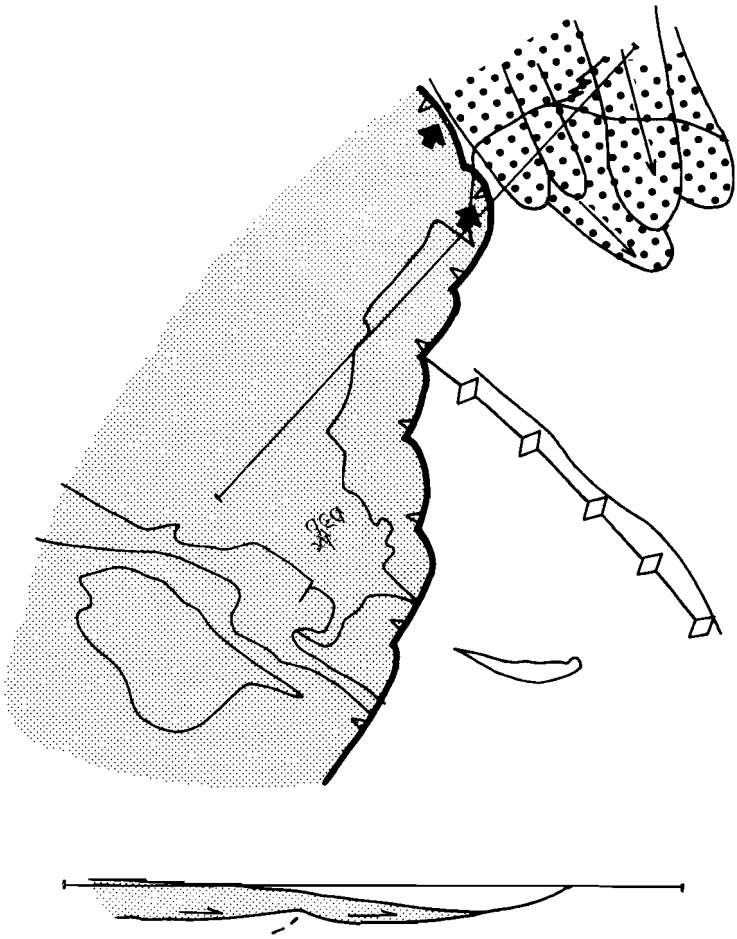

FIG. 28 g.

deposited. These marls are succeeded by several tens of metres of gypsum. Most of the gypsum is precipitated in situ, but reworked gypsum is also present.

STAGE H – Early Pliocene

tectonics

Vertical movements cause the entire area to emerge. Probably in this period the Castel Vecchio-Palazzuolo anticline evolves into an upthrust SE of the river Santerno. NW of the Santerno an anticline running in an E-W direction is created S of the Castel Vecchio area (the Diaterna anticline) with a syncline in between.

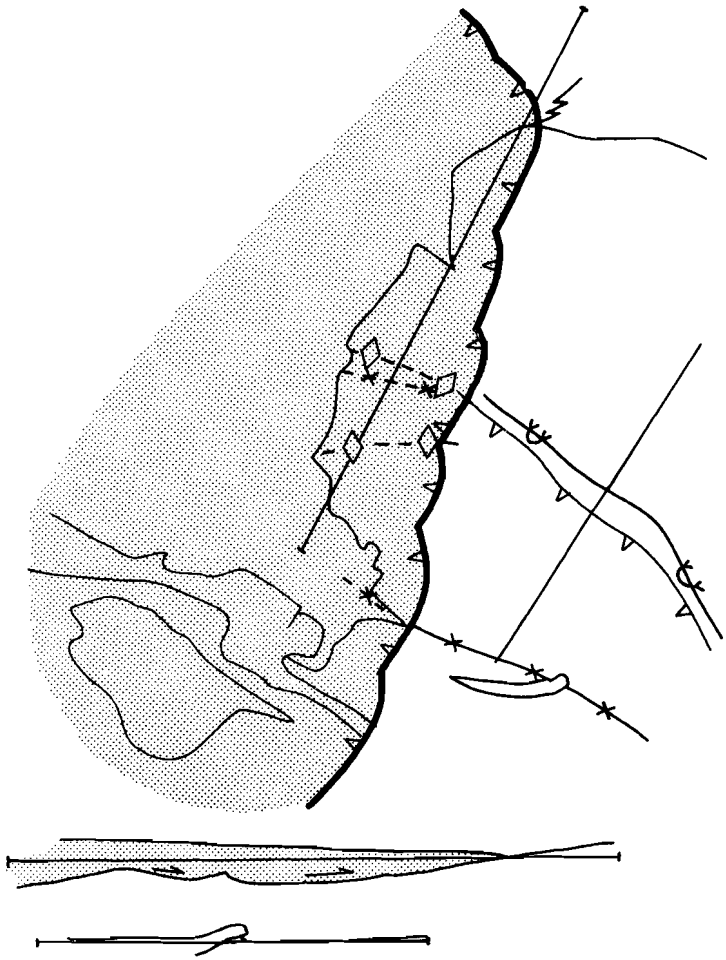

FIG. 28 h.

sedimentation

We found no evidence of sedimentation in the area studied. Mild erosion truncates the top of the gypsum.

STAGE I – Middle–Late Pliocene

tectonics

Differential uplift of the internal zones continues. Some parts lag behind (the Mugello area) and there lakes form. On the external side of the Northern Apennines a shallow sea is created. The front of the Liguride mass advances, no longer because of gravity sliding or spreading, but because of olistostromization. The upper parts of the former overthrust sheet (meso-autochthonous

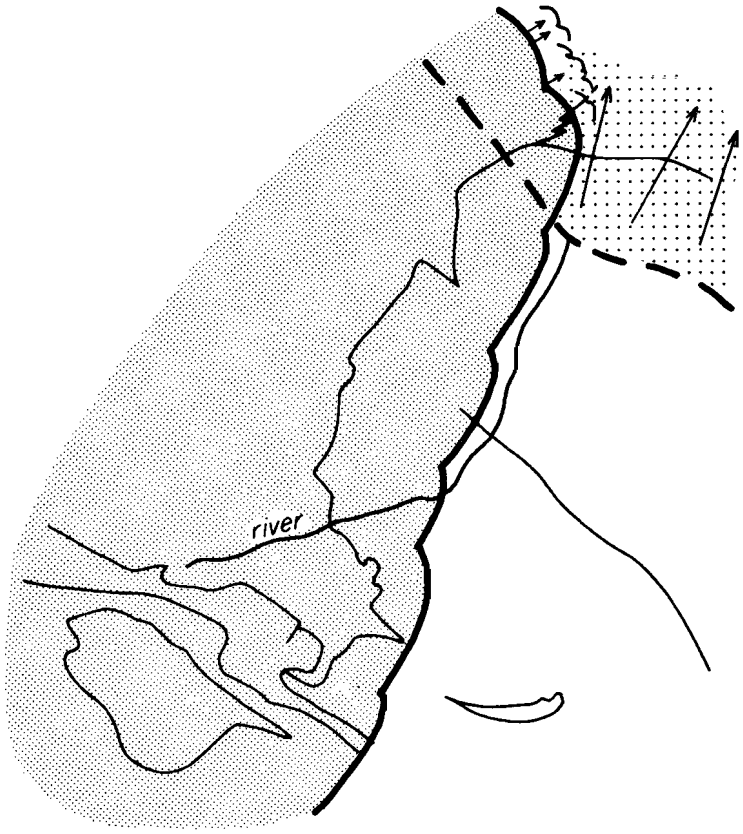

FIG. 28 i.

rocks) slide down the front, and become intercalated in the lower parts of the mass.

sedimentation

Developing rivers transport eroded material from the emerged land towards the NE, where it is deposited in the shallow sea. Olistostromes composed of Liguride material become intercalated between these deposits. Where the Liguride Complex stops moving, it is covered by neo-autochthonous marine sediments. The thickness of the shallow water deposits increases to the NW as a result of the persistently active Sillaro flexure.

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CURRICULUM VITAE

De schrijver van dit proefschrift behaalde in 1968 het einddiploma HBS-B aan de Rijks HBS te Utrecht. Na een half jaar wis- en natuurkunde werd in 1969 begonnen met de studie geologie aan de Rijksuniversiteit te Utrecht. Het kandidaats examen G3 werd in mei 1972 behaald, en in april 1976 volgde het doctoraal examen, hoofdvak strukturele en toegepaste geologie en bijvakken sedimentologie en petrologie. Van april 1976 tot januari 1979 was de schrijver werkzaam als wetenschappelijk medewerker bij de afdeling strukturele geologie van voornoemde universiteit. Sinds 26 januari 1979 is Shell Internationale Petroleum Maatschappij zijn werkgever.

GEOLOGICAL MAP OF THE SILLARO AREA

by J. de Jager
1979

Neo-autochthonous

Tuscan Sequence
- C.S.F. — Cervarola Sandstone Formation, *E.-M. Miocene*
- Sc.T. — Scaglia Toscana, *Eocene-E. Miocene*

Castel Guerrino unit
- R.M. — Rifredo Marls, *M. Miocene*
- C.G.F. — Castel Guerrino Formation, *M. Miocene*

Umbrian Sequence
- C.F.F. — Casal Fiumanese Formation, *Pliocene*
- G.S. — Gessoso Solfifero Formation, *Messinian*

Marnoso Arenacea
- P.m. — Piancaldoli Member, *M. Miocene-Tortonian*
- C.B.m. — Ca Buraccia Member, *M. Miocene*
- Fr.m. — Frena Member, *M. Miocene*
- O.m. — Osteto Member, *M. Miocene*
- B.T.m. — Borgo Tossignano Member, *L. Miocene*
- F.m. — Fontanelice Member, *Tortonian*
- C.d.R.m. — Castel del Rio Member, *M. Miocene-Tortonian*
- C.V.m. — Castel Vecchio Member, *M. Miocene*
- M.C.m. — Monte Coloreto Member, *M. Miocene*
- C. — Casaglia Member, *M. Miocene*, with shaly intercalations, *Paleocene (?)*
- C.m. — Coniale Member, *E.-M. Miocene*

M.F. — Mugello Formation, *Villafranchien*

Liguride Complex
- S.F. — Sellustra Formation, *L. Eocene-E. Miocene*
- M.M.F. — Monte Morello Formation, *Paleocene-M. Eocene*
- V.M. — Violla Marls, *M. Miocene (?)*
- P.V.M. — Piotto Variegated Marls, *Eocene-E. Miocene (?)*
- L.L. — Lame Limestones, *E. Cretaceous (?)*
- B.M. — Borgo Marls, *(?)*
- Ch.C. — Chaotic Complex, *Jurassic-Miocene*

Intraformational slump
Conglomerate
Overthrust
Overthrust, tectonically transgressive
Reverse fault
Unconformity

3 km
scale ± 1:110.000

N