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THE STRUCTURAL AND SEDIMENTOLOGICAL EVOLUTION OF THE LAGONEGRO ZONE, SOUTHERN ITALY

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Wolfson College

Thesis presented to the University of Oxford for the Degree of Doctor of Philosophy.

Michaelmas 1979

Volume I : TEXT

FRONTISPIECE

Jurassic calciturbidites and cherts of the Lagonegro Formation overlying dolomitised Upper Triassic cherty limestones of the Sirino Formation, Lagonegro Unit II, S. Fele.



"Se non é vero, é molto ben trovato"

Anon.

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SHORT ABSTRACT

The Lagonegro Zone comprises three structural units which outcrop within the pile of décollement nappes forming the southern Italian Appenines. Stratigraphic, structural and sedimentological evidence suggests that the zone represents the relics of a Mesozoic basin that developed amidst the carbonate platforms bordering the southern margin of Tethys. Mapping has confirmed that the Lagonegro I, Lagonegro II and Monte Foraporta Units are separated by thrusts and are stacked in ascending order. Middle Triassic fine-grained terrigenous clastic sediments and neritic limestone olistoliths of Lagonegro Unit II record the disintegration of a young carbonate platform under the influence of extensional tectonics. Deposition and redeposition of hemi-pelagic lime-mud washed from the adjacent shallow-water platforms during the Late Triassic is manifested in both Lagonegro units by the cherty limestones of the Sirino Formation; a gradual transition from calcareous to siliceous deposition at the top of the formation reflects the subsidence of the basin-floor beneath the Calcite Compensation Depth. Contrasting patterns of siliceous and calciclastic sedimentation developed during the Jurassic, probably imposed by renewed extensional tectonics; coeval calciclastic deposition in a small, and at times anoxic marginal basin is recorded in the Monte Foraporta Unit. Dolomitisation of basinal carbonate rocks took place during the Jurassic due to mixing of saline pore-fluids with meteoric water recharged from the adjacent carbonate platforms. Deposition of terrigenous shales below the Calcite Compensation Depth occurred during the Lower Cretaceous, but calcareous sedimentation was resumed in the Late Cretaceous and Palaeogene. The progressive deepening of the basin documented by these facies transitions is attributed to regional subsidence, caused by crustal extension and attenuation, and relative accretion of the surrounding platforms; comparable basal stratigraphies in several ophiolitic zones of the Alpine-Mediterranean region suggest that many Tethyan 'oceanic' basins may have initially developed in a similar manner.

LONG ABSTRACT

The Lagonegro Zone, composed predominantly of Lower Mesozoic basinal pelagic sedimentary rocks, outcrops within the pile of décollement nappes forming the southern Italian Apennines. These nappes, which were emplaced during the Miocene, are believed to represent the deformed remnants of the southern continental margin of Tethys; the Mesozoic palaeogeography of the area has been likened to the present day configuration of the Bahamas. The Lagonegro Zone itself comprises the relics of one of several basins, the nature of whose original basement is unknown, that developed amongst the carbonate platforms of the Tethyan region during the Triassic. Stratigraphic, structural and sedimentological evidence is considered with a view to establishing the nature of origin of the Lagonegro Basin, and assessing its significance in relation to the Mesozoic evolution of the Alpine-Mediterranean region as a whole.

The zone comprises three structural units, the Lagonegro I, Lagonegro II and Monte Foraporta Units. The oldest rocks, those of the Middle Triassic Monte Facito Formation, outcrop only in Lagonegro Unit II. The formation is constituted by about 500m of fine-grained terrigenous clastic sediments and neritic limestone olistoliths, overlying a succession of alternating neritic limestones and shales. The succeeding Sirino Formation, of Late Triassic age, outcrops in both Lagonegro Units and is composed of up to 400 m of grey, cherty limestones. The Jurassic Lagonegro Formation, also common to both units, comprises a succession of siliceous mudstones, radiolarian cherts and calciturbidites that attains thicknesses of 60 and 300 m in Lagonegro Units I and II respectively; grey limestones and dolomites of similar age characterise the formations of the Monte Foraporta Unit. About 500 m of Lower Cretaceous grey and black shales and silicified allodapic limestones of the Brusco Formation in both Lagonegro Units I and II are succeeded by Upper Cretaceous and Palaeogene pelagic limestones, marls and cherts.

Detailed mapping of a 40 km^2 segment of the zone has confirmed that the Lagonegro I, Lagonegro II and Monte Foraporta Units are stacked in ascending order. The orientation and vergence of asymmetric minor folds associated with the intervening thrusts imply that the units were emplaced from west to east. It is considered, therefore, that the Monte Foraporta Unit originated on the western flank of the Lagonegro Basin and that Lagonegro Units I and II occupied the eastern and western parts respectively of the basin itself. Folding of all units of the Zone, together with the intervening thrusts, is thought to have occurred as they were overidden by the Alburno-Cervati Carbonate Platform Unit during its emplacement onto the Matese -Monte Maggiore Carbonate Platform during the Tortonian. Late-stage normal faulting occurred during Plio-Pleiocene uplift of the Italian peninsula. The disposition of outcrops of the zone suggests that the basin terminated to the north during the Triassic, but may have extended southwards to join up with the coeval Imerese and Sicani Basins of Sicily

The basal neritic limestones and shales of the Monte Facito Formation are interpreted as the products of episodic blanketing of an embryonic carbonate platform by terrigenous sediments reworked from the underlying basement during an early block faulting phase. Palaeokarst

features in the limestones suggest short periods of emergence. Channellised, planar and trough cross-bedded calcarenites and calcirudites are considered to be the product of deposition under restricted tidal-current regimes; finer-grained calciclastics, interbedded with the apparently structureless shales comprising the remainder of the formation, may reflect redeposition and reworking of shallow-water carbonates by storms. The isolated blocks of neritic limestone which characterise the formation are interpreted as olistoliths derived from the progressive collapse of the carbonate platform. Some blocks retain vestigial pelagic veneers that accumulated during a transitional stage of emplacement; neptunian dykes and fissures attest to extensional stresses associated with disintegration of the platform. Nodular pelagic limestones developed throughout the basin in the Late Ladinian and Early Carnian during a period of sediment starvation prior to the deposition of grey cherty limestones in the Late Triassic. Thus extensional tectonics and minor block faulting resulted in early differentiation of platform and basinal facies.

The Upper Triassic Sirino Formation comprises carbonate sediments deposited during a period of relative tectonic quiescence. The limestone/marly limestone/shale cycles of the Gianni Griecu Member point to the influence of storm-reworking during the earliest phases of sedimentation. Deposition of lime mudstones at greater depths higher in the succession bears witness to the progressive deepening of the basin during the Late Triassic, and a gradual transition from calcareous to siliceous deposition at the top of the formation records the subsidence of the basin-floor beneath the Calcite Compensation Depth at the end of the Period. Intrabasinal redeposition is evinced throughout the formation by slumps and calcarenites, and the resumption of block-faulting in the Late Norian is marked by the extensive development of both intra- and extra-formational calcirudites. Evidence for early lithification, the rarity of preserved calcareous nanno-organisms and the abundance of intrabasinal redeposition suggest that most of the parent lime-mud for these cherty limestones was derived from the shallow-waters of the adjacent carbonate platforms, mainly in the form of aragonite and high-magnesian calcite. With the exception of a paradoxical absence of bank-derived calciturbidites, the patterns of sedimentation proposed for the Lagonegro Basin during the Late Triassic show many features in common with those of the modern deep channels of the Bahama Platform.

The Triassic-Jurassic boundary is marked by a transition from the predominantly calcareous deposits of the Sirino Formation to the siliceous mudstones and cherts of the Lagonegro Formation. Contrasting patterns of sedimentation developed in the Lagonegro units, probably imposed by extensional tectonics; whilst Unit I comprises a relatively monotonous succession of siliceous pelagic sediments, Unit II bears the record of abundant redeposition, of both intra- and extra-formational origin, in the form of radiolarites and calciturbidites respectively. Structureless siliceous mudstones at the base of the Formation in Unit I reflect slow deposition below the Calcite Compensation Depth, whereas the overlying cherts denote higher rates of sedimentation following an increase in surface productivity of Radiolaria in the Late Jurassic; vitreous cherts are interpreted as silicified calciturbidites

derived from the basin flanks following their emergence from beneath the Calcite Compensation Depth as it deepened towards the end of a Period. An overall scarcity of redeposited sediments, however, suggests that the unit was tectonically stable. In Unit II, on the other hand, major block faulting caused the introduction of calciturbidites from the adjacent platforms, flows being ponded in the deeper, northern parts of the basin. Structural instability is also reflected in the thick sequence of rhythmically bedded cherts and siliceous mudstones. In contrast to other basinal sequences of the Tethyan region, pelagic carbonate sedimentation was not re-established until the Late Cretaceous, and the Lower Cretaceous Brusco Formation is characterised by dark-grey and black shales and allodapic limestones that accumulated below the Calcite Compensation Depth. The pelagic limestones, marls and calciturbidites of the 'flysch rosso' denote a resumption of calcareous sedimentation which continued through to the Early Miocene.

Coeval calciclastic deposition in a small, and at times anoxic, marginal basin is recorded in the Jurassic formations of the Monte Foraporta Unit. This basin is thought to have formed by foundering of the margin of the Campania-Lucania carbonate platform along listric faults, the sediments of the Monte Foraporta Unit onlapping the tilted surface of the downthrown block.

A combination of petrographic and geochemical techniques have been employed in order to elucidate the late diagenetic history of the carbonate lithologies of the Zone. In the hemi-pelagic lime-mudstones of the Sirino Formation, aggrading neomorphism has obliterated primary textures, whilst solution and re-precipitation of biogenous silica have fostered the development of nodular cherts. Such is the degree of diagenetic modification that little evidence remains of the primary constitution of the lime-mud. Original fabrics are further obscured by dolomitisation, which has affected limestones of both the Sirino and Lagonegro Formations in the northern outcrops of Unit II. Stratigraphic, petrographic and geochemical data suggest that dolomitisation was effected by mixing of sea-water with meteoric water recharged from emergent parts of the adjacent carbonate platforms. Regional δO^{18} variations are thought to reflect contrasting depths in different parts of the basin, the shallower tracts impinging upon less-saline parts of the mixing zone.

The basinal sediments of the Lagonegro Zone thus demonstrate a steady increase in water-depth from near sea-level in the Middle Triassic to depths in excess of 4 km in the Early Cretaceous. Meanwhile, a continuation of shallow-marine platform carbonate sedimentation is manifested in adjacent units. Comparison of rates of subsidence evinced by the vertical facies changes of the Lagonegro units with those implied by the thicknesses of coeval sediment in the structurally adjacent carbonate platform units indicates that the platforms and basins of the southern Apennines subsided almost in sympathy during the Mesozoic. Relative accretion of the platforms, following early basinal and shallow-water facies differentiation in the Middle Triassic, is therefore held responsible for accentuating relief such that the basin attained bathyal depths in the Jurassic and Cretaceous.

Regional subsidence is thought to have been initiated by crustal

extension and attenuation caused by relative movement of the African and European continents prior to the opening of Tethys in the Liassic. It is considered most likely that the network of basins and platforms were founded upon continental rather than oceanic basement, as has been suggested most recently for the analogous modern Bahama Platform. The origin of the basins is thus only indirectly linked to the formation of the Tethys Ocean, and owes little or nothing to Liassic rifting. Moreover, comparable stratigraphies may be found in other parts of the Alpine-Mediterranean region, suggesting that basins of this type were an important feature of the area during the Triassic. It is possible that several of the Tethyan 'oceanic' basins, represented by the ophiolite sones of the eastern Mediterranean (e.g. Pindos, Othris, Antalya, Mamonia, Baër-Bassit), may have evolved initially along similar lines.

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CHAPTER ONE

INTRODUCTION

1.1 Preface

The rocks of the Lagonegro Zone are exposed over an area of about 600 km^2 in the southern Italian Apennines, southeast of Naples (Fig.1.1). Composed predominantly of Lower Mesozoic basinal pelagic sedimentary rocks, the zone outcrops within a series of décollement nappes which represent the deformed remnants of the southern continental margin of Tethys* (Bernoulli & Jenkyns, 1974; Catalano et al., 1976; Channell & Horváth, 1976; Laubscher & Bernoulli, 1977). The zone is exposed through a network of tectonic windows and half-windows in the overlying units, the principle tracts, the Vulturino and Sirino Windows, being situated between Potenza and Lauria (Fig.1.1); subsidiary outcrops can be found to the northwest and west around San Fele and north of Campagna.

The zone comprises the relics of one of several Mesozoic basins which developed in the Tethyan region during the Triassic, and that are now represented also by the Imerese and Sicani (or Sclafani and Campofiorito - Monte Cammarata) Zones of Sicily, the Pindos, Budva and Cukali Zones of Greece and Yugoslavia, the Antalya Complex of

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*In this study, the term 'Tethys' or 'Tethyan Ocean' is applied to the ocean that developed between Africa and Europe during the breakup of Pangea in the Jurassic and Cretaceous, as distinct from 'Palaeotethys', which refers to the triangular embayment that palaeocontinental reconstructions indicate existed between Arabia and Russia in the Triassic (Smith & Briden, 1977). This follows the terminology of Laubscher & Bernoulli (1977), and is in contrast to the usage of 'Tethys' and 'Mesogea' in the French geological literature to describe 'Palaeotethys' and 'Tethys' respectively (Biju-Duval, 1977). For a review of the use of the word 'Tethys', the reader is referred to Jenkyns (1980, in press).

southern Turkey, the Mamonia Complex of Cyprus, the Baër-Bassit Complex of Syria, the Neyriz Complex of Iran and the Hawasina Complex of Oman (Broquet et al., 1966; Aubouin, Bonneau et al, 1970; Aubouin Blanchet et al, 1970; Lapierre, 1975; Marcoux, 1976; Lapierre & Parrot, 1972; Glennie et al., 1973; Bernoulli & Jenkyns, 1974; Smith & Moores, 1974; Channell & Horváth, 1976; Hallam, 1976; Fleury, 1977; Laubscher & Bernoulli, 1977; Catalano & d'Argenio, 1978; Robertson & Woodcock, 1979). A common problem concerning the origin of these basins, the nature of whose original basement is unknown, centres upon whether or not their inception was directly related to the opening of the Tethyan Ocean, the effects of which have been so extensively documented throughout the Mediterranean region (for reviews, see Smith, 1971; Dewey et al., 1973; Bernoulli & Jenkyns, 1974; Channel & Horváth, 1976; Biju-Duval et al., 1977; Laubscher & Bernoulli, 1977). It has been suggested that these basins owe their origin to an early, Middle Triassic phase of rifting, and that they were founded upon thinned continental crust (Scandone, 1975). However, some zones contain, or are associated with other zones that contain ophiolites, and furthermore their imbricate structure suggests that the basinal deposits may have been stripped from subducted oceanic crust in the form of an accretionary prism (Smith, 1976); the inferred oceanic basement for these successions must have belonged either to a pre-existing ocean or to the oldest parts of the Tethys. It was in an attempt to reconcile these conflicting views that this study of the Lagonegro Zone was initiated.

1.2 Regional Geology

The décollement nappes of the southern Apennines have been grouped into a series of almost coeval structural-stratigraphic units that derive from the destruction of a network of basins and carbonate

platforms bordering the southern margin of Tethys (Fig.1.2) (d'Argenio et al., 1973, 1975; Ippolito et al., 1975); the palaeogeography of the region has been likened to the present-day configuration of the Bahamas (Bernoulli, 1972; d'Argenio, 1970a; d'Argenio,de Castro et al 1975; These allochthonous units are bounded to the north by the Anzio-Ancona Line, although correlative platforms and basins can be found in the autochthonous and allochthonous Tuscan and Umbrian units of the northern Apennines (Fig.1.3) (d'Argenio & Pialli, 1975). At the base of the nappe pile lies the autochthonous Apulo-Gargano Unit, which represents the stable foreland upon which the units are stacked (Fig.1.4); this unit, together with the Paxos Zone of western Greece and the Iblean Zone of southeastern Sicily, probably forms part of the African continent (Catalano et al., 1976; Channel & Horváth, 1976). The southern Apennines units are themselves overlain structurally by the Calabrian Complex, which has been interpreted as a continuation of the Alpine chain (Alvarez, 1976;Amodio-Morelli et al.,1979); this comprises an assemblage of crystalline and ophiolitic nappes, the latter of which most likely constitute remnants of the Tethyan Ocean (Dietrich & Scandone, 1972).

From the base, the structural-stratigraphic units of the southern Apennines include the following (Fig.1.5):

i) The Apulo-Gargano Unit

This autochthonous unit comprises 4-6km of only mildly deformed Upper Triassic to Lower Tertiary carbonate platform deposits, with evaporites at the bases, overlain to the southwest by up to 3km of Pliocene and Pliestocene clastic sediments of the Bradanic Foretrough (Fig.1.5) (d'Argenio et al., 1973, 1975; Ippolito et al., 1975). These platform carbonates originally belonged to the Apulian carbonate platform (Fig.1.2).

ii) The Frosolone Unit

This basinal unit consists of Upper Triassic and Lower Jurassic cherty dolomites, overlain by Middle and Upper Jurassic shales and radiolarites, Cretaceous and Lower Tertiary calciturbidites, pelagic limestones and marls, and Miocene terrigenous clastic deposits (Fig.1.5) (Pescatore, 1965; d'Argenio et al., 1973, 1975; Ippolito et al, 1975). The unit originated in the Molise Basin palaeogeographic domain (Fig.1.2).

iii) The Matese - Monte Maggiore Unit

Deriving from the Abruzzi-Campania Platform, this unit is made up of over 3km of shallow marine carbonate sediments of Late Triassic to Cretaceous age that are unconformably overlain by Miocene reef limestones, marls and terrigenous clastic deposits (Fig.1.5)(d'Argenio et al., 1973, 1975; Ippolito et al., 1975; d'Argenio, 1976). The Middle Cretaceous in this unit is characterised by a bauxite horizon, and the platform is thought to have been emergent during this period (d'Argenio, 1970b). Outcropping principally in Campania, north of Naples, an isolated fragment of the unit is also to be found at Monte Alpi, east of Lagonegro (Fig.1.3) (Ortolani & Torre, 1971).

iv) The Monte Croce Unit

This unit, which is exposed in the Picentini Mountains near Campagna, is chiefly constituted by Upper Triassic dolomites, unconformably overlain by limestones and breccias of Late Jurassic age that are themselves succeeded disconformably by Eocene to Aquitanian calcarenites, and Tortonian marls and sandstones (Fig.1.5) (d'Argenio et al., 1973, 1975; Ippolito et al., 1975). Probably once forming part of the inner margin of the Abruzzi-Campania Platform, this unit is structurally overlain by the Lagonegro Unit I in the

Campagna tectonic window (Figs. 1.2 & 1.6) (Scandone et al., 1967; Turco, 1976).

v) The Lagonegro Units

Comprising two individual thrust sheets, the Lagonegro Units contain Middle Triassic shales, sandstones and limestone olistoliths, Upper Triassic cherty limestones and dolomites, Jurassic radiolarian cherts, siliceous mudstones and calciturbidites, Lower Cretaceous grey and black shales, also containing calciturbidites, and Upper Cretaceous to Palaeogene pelagic limestones, marls and calcarenites (Fig. 1.5) (Scandone, 1967b, 1972; d'Argenio et al., 1973, 1975; Ippolito et al., 1975). These units, which are over up to 1900m. thick, and are structurally overlain by the carbonates of the Alburno-Cervati, Monti della Maddalena and Monte Foraporta Units to the west, disappear eastwards beneath the Miocene flysch of the synorogenic Irpinian basins (Fig.1.3); they derive from the floor of the Lagonegro basin (Fig.1.2).

vi) The Monti della Maddalena Unit

Originating in the external margin of the Campania-Lucania Platform, the Monti della Maddalena Unit is constituted by up to 100m of Triassic white dolomites, transgressively overlain in some areas by Jurassic and younger limestones that are in turn succeeded unconformably by Miocene calcarenites and flysch (Fig.1.5) (Scandone & Bonardi, 1968; d'Argenio et al., 1973, 1975; Ippolito et al., 1975). In the Lagonegro area, the Triassic dolomites are tectonically overlain by the limestones and dolomites of the Monte Foraporta Unit, but generally the unit directly underlies the Alburno-Cervati Unit (Fig.1.6) (Boni et al., 1974).

vii) The Monte Foraporta Unit

This unit, which occupies a structural position between the Lagonegro II or Monti della Maddalena and Alburno-Cervati Units, is made up of 70-80m of grey and black Lower Liassic dolomites and about 300m of grey or black calcilutites, calcarenites and calcirudites of Middle and Late Liassic, Dogger and Malm age (Fig.1.6) (Boni et al., 1974; d'Argenio et al., 1975; Ippolito et al., 1975). Their original palaeogeographic position, like that of the Monti della Maddalena Unit, lay on the external margin of the Campania-Lucania Platform (Fig.1.2).

viii) The Alburno-Cervati Unit

Representing the bulk of the Campania-Lucania Platform, this unit outcrops along almost the entire length of the southern Apennine chain, from the Anzio-Ancona Line south to Calabria (Fig.1.3) (d'Argenio et al., 1973, 1975; Ippolito et al., 1975; d'Argenio, 1976): correlative units are exposed in tectonic windows through the Calabrian Complex in the Lungro-San Donato region (Bousquet & Dubois, 1967; Amodio-Morelli et al., 1979; Dietrich, 1979). The succession is more than 4km thick and consists of massive Upper Triassic dolomites, and shallow-water carbonate sediments of Jurassic to Eocene age, transgressively overlain by Lower Miocene calcarenites, sandstones and shales (Fig.1.5). The carbonate facies are chiefly characteristic of reef and back-reef environments (d'Argenio, 1976).

ix) The Bulgheria-Verbicaro Unit

This unit is characterised by Upper Triassic to Lower Liassic dolomites, and Upper Liassic to Lower Miocene limestones of fore-reef to basinal environments, that are succeeded unconformably by Miocene flysch (Fig.1.5) (d'Argenio et al., 1973, 1975; Ippolito et al., 1975).

The total thickness of the unit, which originally formed the internal margin of the Campania-Lucania Platform, ranges from a few hundred metres to 2km.

x) The internal units

Overlying the external units of the southern Apennines are a number of internal nappes, including the Sicilide, Cilento and Frido Units, that are chiefly constituted by Middle and Upper Cretaceous and Palaeogene flysch, black shales, tuffs and ophiolitic olistostromes (Vezzani, 1969; d'Argenio et al., 1973, 1975; Ippolito et al., 1975). These sediments are generally chaotic, and may in some cases be interposed between the external thrust sheets. They belong to the Liguride and Sicilide Complexes of Ogniben (1969), and can be equated with the Liguride nappes, of similar stratigraphy and age, in the Ligurian northern Apennines (Decandia & Elter, 1969; Abbate & Saggri, 1970); as such, they are thought to derive from the Tethyan Ocean basin (cf. Bernoulli & Jenkyns, 1974; Laubscher & Bernoulli, 1977).

During the Miocene, the structural-stratigraphic units of the southern Apennines became uncoupled from their basement and successively overthrust each other from the west, beginning with the most internal zones (Scandone, 1972; d'Argenio et al., 1973, 1975; Ippolito et al., 1975; Catalano et al., 1976). The Calabrian complex was first of all emplaced over the external palaeogeographic domains during the Aquitanian, and was transported eastwards towards the stable foreland as the underlying units overode each other (Pescatore, 1970; Grandjacquet et al., 1972; Scandone, 1972; d'Argenio et al., 1973, 1975; Pescatore & Ortolani, 1973; Catalano et al., 1976; Amodio-Morelli et al., 1979). The progressive superposition of these units is clearly documented by the successive transgression and subsidence of the carbonate platforms from west to east, and by the eastward migration

of the Irpinian flysch basins, which evolved ahead of the advancing nappe fronts between the Langhian and Late Tortonian (Fig.1.7) (Cocco et al., 1972, 1974; d'Argenio et al., 1973; Pescatore & Ortolani, 1973; Ippolito et al., 1975); further compressional tectonic events took place in the Messinian and Pliocene (Pescatore et al., 1970; Pescatore & Ortolani, 1973; di Nocera et al., 1976). The southern Apennines were elevated to their current level during a period of extensive normal faulting in the Late Pliocene and Pleistocene (d'Argenio et al., 1975).

1.3 Previous work on the Lagonegro Zone

For the purposes of this study, the Lagonegro Zone is taken to comprise the Lagonegro I, Lagonegro II and Monte Foraporta Units.

Prior to the nineteen-sixties, the only significant synthesis of the geology of the southern Apennines was that of Giuseppe de Lorenzo, who devoted much of his career to describing the geology of his native Lucania (de Lorenzo, 1896a). He considered that the rocks of the Lagonegro area formed part of a single, autocthonous succession that was entirely of Triassic age (for review, see Scandone, 1967b). In his earlier works, the succession he described was as follows:

'Dolomie bianche'

'Scisti silicei'

'Calcari con liste e noduli di selce'

'Calcari dolomitici di scogliera'

Later, however, he modified his original ideas, and interpreted the 'calcari di scogliera' as lenticular masses within, and coeval to, the 'scisti silicei' (de Lorenzo, 1894). It was not until 1939 that the presence of a thrust between the 'dolomie bianche' and

'scisti silicei' was identified, although several authors had doubted the total autocthony of the southern Apennines for some time (Grzybowski, 1921; Anelli, 1939; Signorini, 1939). Subsequently, Lucini (1956) indicated that the Middle Triassic 'calcari di scogliera', later to become known as the Monte Facito Formation, was also in tectonic contact with the 'scisti silicei', a possibility that had been put forward by Mojsisovics (1896) sixty years previously. Lucini was in addition responsible for the recognition of a continuity of sedimentation in the zone up to the Cretaceous, a fact subsequently confirmed by Scarsella (1957), and the ensuing decade saw vindication of his ideas as dating of the 'scisti silicei' confirmed their Jurassic age (Tacoli & Zoja, 1957; Richetti, 1961; de Castro, 1963; Scandone, 1967b; Luperto, 1964, 1966; Crescenti, 1966).

In 1967, following a series of papers refining the stratigraphy of the 'serie calcareo-silico marnosa', and in particular of the Monte Facito Formation, Scandone published what remains as the most comprehensive description of the geology of the zone available (Scandone, 1961, 1963, 1964a,b, 1965, 1967a,b; Scandone & de Capoa, 1966). Based upon the recognition of two, distinct, allocthonous structural units, Scandone was the first to attempt a detailed analysis of the sedimentary facies of the zone, and he interpreted the Lagonegro Units I and II as the relics of the axial and marginal parts respectively of a Mesozoic basin that was destroyed during a phase of compressional tectonic in the Miocene. In the same year, Scandone et al. (1967) described further outcrops of the zone in the Campagna tectonic window, where the basinal units occupy a structural position between the Alburno-Cervati and Monte Croce Units, thus confirming that they are allocthonous. The presence of eruptive basic igneous rocks of Cretaceous age in the zone was reported by Ietto & Cocco (1965),

and a refinement of the Triassic stratigraphy of the Lagonegro Units has been provided by de Capoa Bonardi (1970).

Many of the ideas expressed by Scandone concerning the relevance of the zone to the overall development of the southern Apennines were crystallised in a subsequent work that accompanied the publication of a map on a scale of 1:100,000 (Scandone, 1972); the broader significance of the zone in relation to the rest of the Mediterranean has also been recently evaluated (Scandone, 1975). However, with the exception of a more comprehensive account of the stratigraphy and structure of the Monte Foraporta Unit (Boni et al., 1974), and the mapping of the Campagna tectonic window (Turco, 1976), the only new information of recent years has been provided by Cocco et al. (1974), who describe in more detail than hitherto the Upper Cretaceous and Palaeogene deposits of the basin in a newly mapped area north of Monte Marzano, between San Fele and Campagna.

1.4 Aims and scope of the study

The objectives of this thesis are essentially threefold:

i) To reconstruct the pre-orogenic morphology and palaeogeography of the Lagonegro Basin.

ii) To document the Early Mesozoic history of the basin in an attempt to put constraints upon its probable mode of origin.

iii) To assess the significance of the basin in terms of the Mesozoic development of the Tethyan region as a whole.

In order to achieve these ends, field studies have been carried out at three levels:

i) On the largest scale, an area of 40km^2 in the Sirino Window north-east of Lagonegro has been mapped in detail to determine the structural and stratigraphic relationships between the diverse elements of the zone; the results are chiefly incorporated into Chapters 2 and 3.

This area was chosen because it provides a representative cross-section through the three units, and because the relief and exposure are sufficient to allow a three dimensional appreciation of the structure; it also contains a large number of well-exposed stratigraphic sections.

ii) Viewing the basin in its entirety, a large number of key localities throughout the zone were visited, and the various formations examined from a sedimentological standpoint in order to elucidate the complexities of the basin's history and palaeogeography from its formation in the Middle Triassic, up until the Cretaceous (Fig.1.8). Representing the bulk of the thesis, these aspects are discussed in Chapters 4, 5, 6, 7 and 8. Appendix 1 contains detailed stratigraphic descriptions of all the sections measured, and a summary of these sections, together with location maps, is given in Chart I.

iii) Finally, in order to gain a perspective for interpretation of the basin's significance in relation to the Tethyan region as a whole, brief reconnaissance visits were made to four other, coeval basinal zones in Sicily and Greece; these trips have chiefly inspired the writing of the final chapter.

A total of seven months fieldwork has been backed up by laboratory analysis of a selection of rock samples in Durham, Oxford and Zürich. Over 300 thin sections were prepared, and, in addition, acetate peels were taken from polished and etched surfaces of limestones and cherts using the methods of McCrone (1963), Katz & Friedman (1965) and Price (1975). Both thin sections and etched surfaces of limestones were stained in solutions of Alizarin Red-S and Potassium Ferricyanide according to the techniques outlined by Friedman (1959) and modified by Dickson (1965). It has been found most convenient in describing the carbonate lithologies to use the terminologies of

Grabau (1904) and Dunham (1962) in the field and in hand specimen, and Folk's (1959, 1962) classification for thin sections; for cherts, the usage of Folk & Weaver (1952) has generally been followed. Further samples were studied under the scanning electron microscope and analysed by a variety of geochemical techniques, details of which are given in Chapter 8.

CHAPTER TWO

STRATIGRAPHY

2.1 Introduction

In the course of mapping a segment of the Lagonegro Zone, a lithostratigraphic framework has been erected (Fig.2.1), based upon the work of Selli (1962) and Scandone (1967b,1972). The procedure outlined by Harland et al. (1972) has been followed where possible; thus that status of formation has been limited to distinctive lithological units mappable throughout the area.

The Monte Facito Formation is as originally described by Scandone (1965); the names of the Sirino, Lagonegro and Brusco Formations, however, have been adopted from Selli (1962) and are used in preference to the alternative 'calcari con selce', 'schisti silicei', and 'flysch galestrino' of other authors (de Lorenzo, 1892 a,b; Scarsella, 1957; Scandone, 1967b,1972). In order to avoid proliferation of stratigraphic nomenclature, correlative successions in the two Lagonegro units have been given the same name. No rocks younger than those of the Brusco Formation were encountered in the area mapped, and all other lithostratigraphic names are taken from Scandone (1967b,1971, 1972). The formations of the Lagonegro units constitute the Lucanian Group, but are more commonly referred to as the 'serie calcareo - silico - marnosa' in the Italian literature.

Type sections of the formations are generally those of Scandone (1967b)and are not necessarily located within the area mapped in this study. Due to intense late-orogenic normal faulting, it is rarely possible to give a single type succession for any particular formation; in such cases, two or more sections are described.

The stratigraphy, dominant lithology and palaeontological data for the formations of each unit are described in sections 2.2 to 2.4 and a summary of the stratigraphy of the Lagonegro Zone, together with details of section localities, is given in Chart 1. More detailed lithological logs of each section measured, together with a brief description of their salient features, are given in Appendix 1. For a comprehensive catalogue of outcrops of each formation, the reader is referred to Scandone (1967b).

2.2 Lagonegro Unit I

The Lagonegro Unit I is the lowest of the three structural units of the Zone (see Chapter 3), and the oldest exposed rocks are those of the Sirino Formation. It is possible, however, that the Monte Facito Formation of Unit II originally formed the base of the succession in this unit as well, but that décollement occurred at a higher stratigraphic level.

2.2.1. The Sirino Formation

2.2.1.1. Name and distribution

The Sirino Formation, whose name derives from that of the mountain overlooking the town of Lagonegro, is extensively exposed in the Sirino Window; notable outcrops are to be found on Monte Sirino itself, in the two anticlinal ridges of Costa del Alto - Monte Nicola - Mizzo Milego and Gianni Griecu - Monte Gurmara - Monte Castagnereto, north of Lagonegro, and at Monte Farno. Further outcrops can be found in the Vulturino Window, principally on Monte Vulturino and north of Sasso di Castalda. Comparable lithologies ascribed to this unit have also been described from Campagna (Scandone et al., 1967; Turco, 1976).

2.2.1.2. Type sections and thickness

Scandone (1967b), describes two type sections (in the Vulturino

Window), at Sorgente Acero and Monte Lama, that represent the base and top of the formation respectively. Detailed examination of the former, however, has revealed that it crosses the core of a faulted anticline, and that the cherty limestones interpreted as representing the base of the formation are a tectonic repeat of a higher member. The Monte Lama Section, although apparently continuous, is poorly exposed and difficult to follow in detail; in the lower part it is also interrupted and repeated in places by folding. The top and bottom of the formation are best exposed in the Lagonegro and Gianni Griecu Sections, and the latter is proposed as an alternative type succession for the base.

Without a complete section through the formation, it is not possible to give a precise value for its thickness; the 500m quoted by Scandone (1967b) is probably an overestimate due to the tectonic repetition mentioned above. The apparently complete but inaccessible section on the south face of Gianni Griecu is about 400m thick.

2.2.1.3. Members

The Sirino Formation in Unit I has been divided into three members; the Gianni Griecu, Monte Sirino and Carboncello Members. Where exposure has allowed, these have been mapped as individual stratigraphic units, but this has not been possible throughout the area.

a) The Gianni Griecu (Limestone/Marl) Member (New name)

This corresponds to the Halobia superba argillaceous level of Scandone (1967b) from Sorgente Acero, and is best exposed in the Gianni Griecu Section. Lithologically, the member comprises an alternation of grey and black pencil shales and thinly bedded limestone/marly limestone couplets, with rare planar- and cross-laminated

calcilutites and calcisiltites. The limestones - marly limestone couplets are typically 4-15cm thick and comprise a strongly bioturbated yellowish marly limestone overlain by grey calcilutite containing rare ammonites and abundant pelagic bivalves, the latter being commonly oriented convex upwards. Trace fossils are commonly preserved on the undersides of the beds. The base of the member is not exposed but the top is marked by the final disappearance of the limestone - marly limestone couplets.

Several other outcrops of the member may be found in both the Sirino and Vulturino Windows, and in each case, as at Gianni Griecu, they outcrop in the cores of major anticlines. Notable exposures occur beside the road from Lagonegro to Lago Remmo east of M.Niella, and at Sorgente Acero (Scandone, 1967b). Thicknesses are difficult to compare but they amount to 12.5m in the Gianni Griecu Section. The figure of 80m reported by Scandone (1967b) is certainly an exaggeration due to the abundant tectonic repetitions mentioned above.

b) The Monte Sirino (Cherty Limestone) Member (New name)

Comprising the majority of the formation, this member outcrops extensively throughout both of the principle tectonic windows, forming the bulk of the mountains of M.Sirino and M.Vulturino. It consists of a monotonous sequence of thinly bedded grey calcilutites containing bed-parallel bands and nodules of grey or black chert, lithologically indistinguishable from the cherty limestones of the Sirino Formation of Unit II. This member incorporates the Halobia charlyna level of Scandone (1967b); the Halobia styriaca level, interpreted as occurring below that of H.superba by that author, can also be included.

Although poorly exposed, the section at Monte Lama is the thickest measured in this member. It is not, however, complete since the contact

with the Gianni Griecu Member is not exposed; the contact with the overlying member, on the other hand, is clearly visible and is marked by the appearance of thin shaley partings between the calcilutite beds. The base of the member is most readily observed in the Gianni Griecu Section. Other good but incomplete sections have been measured at Costa del Alto, M. Nicola and M. Sirino (See Appendix 1).

The thickness of the member at M. Lama is at least 150m, but is probably as much as 300m in total (Scandone, 1967b).

c) The Carboncello (Limestone Shale) Member (New name)

The highest member, and its contact with the overlying Lagonegro Formation, is best exposed at Lagonegro, although an equally complete section can be observed in less detail at M. Lama. This member includes the H. insignis and H. sicula levels of Scandone (1967b) and de Lorenzo (1893). Thinly bedded grey calcilutites with shale interbeds, which gradually increase in thickness and abundance through the member, characterise the lithology. The calcilutites resemble those of the Monte Sirino Member except that the chert bands and nodules are predominantly black. The shales, generally green in colour, are locally red and become progressively more siliceous towards the top. The disappearance of the calcilutites marks the top of the formation.

In the type section at M. Lama the member is 65m thick; at Sasso di Castalda and Lagonegro, 75 and 55m are exposed respectively. Since the presence of shale intercalations is considered to be as much a function of diagenesis as of primary deposition, and the transition to the Lagonegro Formation so gradual, these thickness variations probably have no real stratigraphic significance.

2.2.1.4. Palaeontology, age and correlation

The Late Triassic age of the formation was first recognised by de

Lorenzo (1892a,b, 1893) who specifically ascribed it to the Trachyceras aon (Julic) and T. aonoides (Cordevolic) zones of the Carnian stage. Scandone and de Capoa (1966) also gave a Carnian age, but de Capoa Bonardi (1970) more specifically attributes the formation to the Lower Carnian (Cordevolic) - Middle Norian (Alaunic) interval. The biostratigraphy of the latter author is complicated by the fact that it is based upon the sections of Scandone (1967b) which, as discussed above, are in some cases tectonically repeated; thus the Halobia styriaca of the section at Sorgente Acero is higher than the H. superba level and not lower as previously described (Scandone, 1967b; de Capoa Bonardi, 1970).

As noted by de Capoa Bonardi (1970), the stratigraphic range of Halobia superba probably extends as far down as the Lower Carnian (Cordevolic), as indicated by its co-existence with H. cassiana and H. styriaca; it is not possible, therefore, to give a more specific age than Middle or Lower Carnian (Julic or Cordevolic) for the Gianni Griecu Member. The Monte Sirino Member extends into the Lower or, possibly Middle Norian, while the Carboncello Member, which incorporates the H. halorica and H. norica levels of de Capoa Bonardi (1970), is of Middle Norian (Alaunic) age or younger (de Capoa Bonardi, 1970). The top of the formation is almost 50m above the level of H. norica, and the calcilutites in the top 25m are apparently devoid of Halobia sp. shells. It is possible, therefore, that the youngest rocks of the Carboncello Member are of Early Liassic age.

Correlation with the Sirino Formation of Unit II on the basis of the Halobia fauna is not easy since the H. superba level is apparently younger in this unit (de Capoa Bonardi, 1970); an Early Carnian (Cordevolic) age for the Gianni Griecu member would possibly imply a partial equivalence with the Valle del Pesce Member of the Monte

Facito Formation (see section 2.3.1.4). Correlation of the remaining members is precluded by the paucity of faunal data from the higher parts of the formation in Unit II.

2.2.2. The Lagonegro Formation

2.2.2.1. Name and distribution

The town of Lagonegro, from which the formation takes its name, was the birthplace of the celebrated Italian geologist Giuseppe de Lorenzo, who first described the 'schisti silicei' here in the late nineteenth century (de Lorenzo, 1892 a,b). As with the Sirino Formation, however, the terminology is that of Selli (1962). Exposure, particularly of the Pietra Member, is good, and outcrops forming features of positive relief can be found in both the Sirino and Vulturino Windows as well as in the village of Campagna (Scandone et al., 1967; Turco, 1976).

2.2.2.2. Type sections and thickness

Two type sections have hitherto been described, at Lagonegro and Sasso di Castalda (Scandone, 1967b, 1972). The latter is now poorly exposed due to recent afforestation. The Lagonegro Section, on the other hand, is clear and easily accessible, but the topmost part of the formation and the contact with the overlying Brusco Formation is most readily observed in the Ponte della Pietra Section. The thickness of the formation is about 60m at Lagonegro.

2.2.2.3. Members

The formation can be subdivided into two members that are usually readily distinguishable in the field; the Cararuncedde and Pietra Members.

a) The Cararuncedde (Siliceous Mudstone) Member (New name)

Well exposed in the Lagonegro section, the member comprises red, green, grey and black siliceous mudstones and cherts, thinly bedded

silicified fine-grained limestones and rare beds of partially to totally silicified carbonate microbreccias. The siliceous mudstones dominate the succession and cherts are only rarely developed; the silicified limestones occur only at the base. The microbreccias are rarely graded and the clasts possess shallow water carbonate fabrics and faunas.

The base of the member corresponds to the top of the Sirino Formation and the top is marked by the transition from red siliceous mudstones to the green and grey cherts of the overlying Pietra member. A thickness of 25.5m has been measured in the Lagonegro section.

Although not as well exposed as the Pietra Member, other good sections can be found at Monte Lama and Mangoso. Several additional outcrops in both the Sirino and Vulturino Windows have been reported (Scandone, 1967b); of particular note is that at Coste Roberto on M. Vulturino, where the intercalation of calcareous microbreccias with the siliceous mudstones is readily observed. An overall increase in the thickness of the member and in the proportion of carbonate microbreccias is apparent northwards.

b) The Pietra (Chert) Member (New name)

This member is exposed in its entirety in the Ponte della Pietra Section where it measures 33m in thickness. It comprises a monotonous succession of green and grey, highly silicified radiolarian cherts and grey or white vitreous cherts with rare, more thickly bedded and cross-laminated brown cherts. The transition to the silicified limestones and black shales of the Brusco Formation is abrupt and corresponds with a marked reduction in the degree of silicification.

Other, incomplete sections have been measured at Mangoso, Lagonegro, Monte Lama and Sasso di Castalda, and an abundance of other exposures are documented throughout the outcrop of the unit (Scandone,

1967b); cherts of identical aspect have been described from the village of Campagna (Turco, 1976).

2.2.2.4 Palaeontology, age and correlation

The paucity of palaeontological data, makes precise dating impossible. The fauna described in the literature are entirely derived from calcareous microbreccias which occasionally punctuate the succession. Since these are interpreted as having been redeposited from shallow water environments of other units they can only give a maximum age.

Scandone (1967b,1972) describes a microfossil assemblage from a carbonate microbreccia 10m above the base of the formation at Sasso di Castalda which, although containing no index species, he tentatively ascribes to the Lias. The presence of a 75cm band of black laminated chert, which contains abundant pyrite and organic matter, may reflect anoxic bottom conditions during deposition. The widespread occurrence of bituminous shales of Toarcian age in deep water successions has led some authors to speak of a world-wide anoxic event at that time (Hallam and Bradshaw, 1979). The black cherts of the Lagonegro Section may be a testimony to this Toarcian event, implying a Late Liassic-Dogger age for the Cararuncedde Member.

The widespread occurrence of carbonate breccias of Liassic age elsewhere in the Alpine Mediterranean region, that relate to the widespread initiation of block faulting at that time, lends some credence to this hypothesis (Bernoulli and Jenkyns, 1974). Equally, the abundance of radiolarian cherts in the Tethyan Late Jurassic may point to a similar age for the Pietra Member (Garrison and Fischer, 1969; Bernoulli and Jenkyns, 1974; Bosellini and Winterer, 1975). Upper and Lower limits can be set by the Late Triassic/Early Liassic and Early Cretaceous ages of the under- and overlying formations

respectively. A period spanning most or all of the Jurassic is therefore favoured (Crescenti, 1966; Scandone, 1967b,1972).

Dating of the Lagonegro Formation of Unit II is similarly imprecise but the succession is probably of the same age as in Unit I. The Formation is co-eval with both formations of the Monte Foraporta Unit, and the structurally adjacent Alburno-Cervati and Matese-Monte Maggiore carbonate platform units expose a continuity of shallow-water carbonate sediments, principally characteristic of black reef environments, during this time period (Boni et al. 1974; Ippolito et al 1975).

2.2.3. The Brusco Formation

2.2.3.1. Name and distribution

The Brusco Formation, or 'flysch galestrino' of other authors, has never been formally defined. By virtue of the incompetence of the dominant lithology, it is invariably both highly deformed and poorly exposed, and it is thus impossible to give a type section except for the base. It represents the youngest formation of Unit I in the area mapped, where it is structurally situated below the basal thrust of Unit II. Although subject to intense erosion, the formation outcrops extensively around the bases of M. Sirino and M. Vulturino as well as in the vicinity of Sasso di Castalda. Sections have been measured at M. Farno, Bersaglio and Fiume Torbido.(See Appendix 1).

2.2.3.2. Type section and thickness

The only type section that can be given covers the basal few metres and the contact with the Lagonegro Formation, and is to be found in the Ponte della Pietra section. It is not possible, therefore, to give a value for the thickness of the formation, although Scandone (1971) quotes an estimate of more than 500m in the area south of M. Sirino.

2.2.3.3. Lithology

The formation is composed of a succession of grey to black fissile shales punctuated by beds of grey, white and, rarely, pink calcilutite, up to 1.5m thick and often markedly silicified. Also present are carbonate microbreccias, that are commonly silicified and which may be graded and exhibit sole marks; some quartzose sandstones and siltstones have also been reported (Scandone, 1967b,1972). At outcrop, the shales and calcilutites often bear a pronounced black Fe/Mn oxide-hydroxide patina. The fissility in the shales is usually parallel to the bedding of the limestones, but angular bedding/cleavage relations may be observed near the axes of minor folds. Scandone (1967b) has distinguished two regional facies types within the formation based on the proportion of siliciclastic beds; since, however, it cannot be shown that the two facies are coeval, their validity as regional variants is doubtful.

2.2.3.4. Palaeontology, age and correlation

In common with the Lagonegro Formation, only the carbonate microbreccias have furnished any palaeontological data, the shales being devoid of fauna with the exception of rare radiolarians. Furthermore there are no data extant for the age of the formation in Unit I, but by comparison with Unit II, a Late Malm/Early Cretaceous age may be inferred for the base.

No deposits of Early Cretaceous age are present in Monte Foraporta Unit, but shallow-water carbonate sediments of this age follow conformably upon the Jurassic in the Alburno-Cervati Unit. In contrast with the latter, the Matese-Monte Maggiore Unit is marked by an unconformity below bauxites of Middle Cretaceous age. In the more internal Liguride Units, the Early Cretaceous saw the inception of true flysch deposition (c.f. The Frido Formation (Vezzani, 1969); The Crete Nere Formation

(Selli, 1962); the S. Venere Formation (Ietto et al. 1965)).

2.2.3.5. The 'Scisti rossi di Pecorone'

Scandone (1967b,1972) has indicated the existence of a possible continuation of sedimentation in the unit at least into the Late Cretaceous. A small outcrop comprising an alternation of red cherts thin grey silicified carbonate breccias, and red or green shales and marls near Pecorone, between Lagonegro and Lauria, has been dated as Upper Senonian (Scandone, 1972). The facies bears a close resemblance to the Upper Cretaceous part of the Crisanti Formation in the Imerese Zone of Sicily, where basinal deposits analogous to those of the Lagonegro Zone can be followed without a break up in the Eocene. Although it is impossible at Pecorone to identify any stratigraphic relations with the Brusco Formation, other outcrops are documented where contacts can be observed (Scandone, 1972). This formation can be loosely correlated with the 'Flysch Rosso' of Lagonegro Unit II (Scandone, 1972).

2.3 Lagonegro Unit II

The Lagonegro Unit II preserves a more extensive stratigraphy than Unit I, extending down to the Middle Triassic, although the post-Jurassic parts of the succession are poorly known.

2.3.1. The Monte Facito Formation

2.3.1.1. Name and distribution

For many years the rocks of this formation were equated with those of the Lagonegro Formation (see Scandone, 1967b), a confusion which has persisted until quite recently (Ogniben, 1969). Recognition of a distinct formation underlying the cherty limestones of the Sirino Formation and its formal definition was made by Scandone (1963, 1965, 1967b). The type section on M. Facito, from which the formation takes

its name, however, is no longer visible due to afforestation.

The formation occurs throughout the two principle unit outcrops, as well as in the Campagna tectonic window (Scandone et al. 1967b, Scandone, 1972; Turco, 1976). Its proximity to the basal thrust of the unit has led to extensive tectonic modification of the primary depositional relations and it is seldom possible to follow vertical or lateral facies changes.

2.3.1.2. Type section and thickness

A type section for the formation on M. Facito itself has been described by Scandone (1967b)that, for the reasons stated above, is no longer extant; furthermore, poor exposure and complicated tectonics have prevented recognition of a suitable alternative. Facies variations, however, are so marked that no individual section could be considered to be representative of the formation as a whole and thus the concept of a 'type section' is inappropriate; none, therefore, have been defined except for the very uppermost part extending to the contact with the Sirino Formation.

The overall thickness of the formation is difficult to establish and likely to be extremely variable. A cross-section through a relatively undeformed tract south of La Cerchiara (see Chapter 4) that encompasses the two lower members of the formation indicates a value of more than 500m. This is greatly in excess of the 200m measured in the type section; this, however, was incomplete, lacking both the lower and upper members (Scandone, 1967b, 1972). A more precise estimate cannot be given.

2.3.1.3. Members

The 'Organogenic' and 'Terrigenous' Members distinguished by Scandone (1967b, 1972) are not considered to have any stratigraphic

validity since they were deposited coevally. They are, therefore, rejected in favour of three new members which effectively comprise three distinctive and vertically sequential facies associations, the La Cerchiara Member, the Pietra Maura Member and the Valle del Pesce Member. Only the last two outcrop in the mapped area, the lowest member being exposed only at the locality south of La Cerchiara described in Chapter 4.

a) The La Cerchiara (Limestone/Shale) Member (New name)

The member comprises a series of three tabular limestone packets, varying in thickness from 4 to 50m and separated by structureless red or greeny/yellow shales and marls. The limestones, which are light grey in colour, are massive and commonly cleaved; the uppermost packet is laterally discontinuous. The base of the sequence is faulted, but the contact with the next member, coinciding with the top of the highest limestone packet can be observed. The total thickness is about 100m. Although not individualised as a separate member, the tabular limestones of this outcrop are cited by Scandone as one of three characteristic forms of the 'Organogenic Member'.

b) The Pietra Maura (Olistostrome) Member (New name)

Areally and volumetrically the largest member, it comprises 400m or more of red, green, grey, yellow and khaki sandstones, shales and marls with interbedded grey to yellow calcarenites and calcisilites. These predominantly fine-grained clastics enclose blocks of grey limestone of various shapes that range from a few metres up to 2km across. These blocks are distributed at random, both areally and stratigraphically, imparting a characteristically 'hummocky' aspect to the terrain; only in several outcrops south-east of Tramutola, where

the calcarenites and calcisiltites observed elsewhere are also lacking, and the shales, which are notably more siliceous, contain rare cherts, are they entirely absent.

c) The Valle del Pesce (Dolomitic Marl) Member

Amongst the shales and marls at the top of the formation, bed-parallel nodules and bands of pink or grey fine-grained limestone are present; beds of structureless pink, green or yellow micaceous sandstone also occur. The limestone nodules, which are commonly only separated by thin clay seams, may be replaced by pink chert. This facies can be recognised throughout the formation outcrop at the contact with the overlying Sirino Formation. Its thickness in the type section at Valle del Pesce, from the first nodular limestone at the base to the appearance of the cherty limestones of the Sirino Formation at the top, is 15m. An almost identical sequence is exposed in the Armizzone section, and several others are documented by Scandone (1967b).

2.3.1.4. Palaeontology, age and correlation.

No identifiable fossils have been collected from the basal member and its precise age is unknown; Scandone (1967b), however, records several poorly preserved examples of Daonella sp. An abundant brachiopod fauna collected from the claystones surrounding the limestone block of Pietra Maura include Spiriferina fragilis, and is thought to indicate a Late Anisian (M.Trias) age (Scandone, 1965; Taddei-Ruggiero, 1968). These claystones are equivalent to the oldest beds of the type succession described by Scandone (1967b,1972). It may be supposed, therefore, that the La Cerchiara Member is at least as old as Early Anisian.

A pelagic-bivalve fauna collected 70m higher in the succession

furnished Daonella taramellii, which is indicative of the Early Ladinian (Fassanian) Stage (de Capoa Bonardi, 1970). The base of the Valle del Pesce Member has been given a Late Ladinian age on the basis of the presence of Daonella lommelli (de Capoa Bonardi, 1970). The sedimentology of this member, however, suggests that sedimentation rates were slow and it may, therefore, be stratigraphically condensed, in which case it could extend well into the Early Carnian; the proximity of the Halobia styriaca level, generally considered to be of Middle Carnian (Julian) age, in the overlying Sirino Formation supports this contention. It is thus possible that the V. delle Pesce Member is at least partially equivalent to the base of the Sirino Formation in Unit I.

The age of the limestone blocks is more obscure: de Lorenzo (1896b) attributes a rich fauna collected from one such block near Lagonegro to the Ladinian, an age confirmed by several other authors (Scandone, 1964b; Scandone and de Capoa, 1966; Donzelli and Crescenti, 1970). Many of the species described by de Lorenzo, however, are of pelagic affinity and were probably collected from neptunian dykes; they do not, therefore, represent the true age of the block itself which may be slightly older. The suggestion of Azzaroli (1962) that other blocks may be as old as Permian, although doubted by Scandone (1964b) is lent credence by the recognition of a reworked Permian algal and foraminiferal fauna in a calcarenite collected close to M.Facito (Donzelli and Crescenti, 1970). These calciclastic facies are generally composed of lithic fragments of limestones formed in shallow water environments and their fauna is thus not necessarily indicative of the age of the calcarenite but only of the parent limestone. Discontinuous blocks of limestone of definitive Permian age are present in analogous sequences in the Lercara Formation of the Sicani Zone in Sicily

(Mascle, 1967).

Rocks of Middle Triassic age are not found in the structurally adjacent platform carbonate units. However, alternating siliciclastic and shallow marine carbonate sedimentation units similar to those of the La Cerchiara Member, are documented from the Anisian at the base of the S. Donato carbonate platform unit visible beneath the Calabrian Complex in the S. Donato-Lungro region (Bousquet and Dubois, 1967; Amodio-Morelli et al. 1979; Dietrich, 1979). This unit has been equated with the units derived from the Campania-Lucania carbonate platform of the southern Italian Apennines (Amodio-Morelli et al. 1979).

2.3.2 The Sirino Formation

2.3.2.1. Name and distribution

At risk of causing confusion, this formation has been given the same name as that of the broadly coeval and lithologically similar formation of the lower unit. Scandone (1967b,1972) distinguishes four 'facies' in the formation, the S.Fele, Pignola-Abriola, Armizzone and Lagonegro-Sasso di Castalda Facies, the latter corresponding to the formation in Unit I, and the others, in Unit II. Since the lithological differences between those 'facies' of Unit II are largely of a secondary nature (dolomitization), this definition has been discarded.

The most extensive outcrops are to be found in the northern parts of the Sirino and Vulturino Windows, with other notable exposures south and west of Lagonegro, in the vicinity of Castelsaraceno, north of Viggiano, south of Vietri di Potenza and in the north at S. Fele, M. Pierno and Campagna (Scandone, 1967b, Turco, 1976). Sections incorporating parts of the formation are described from M. Armizzone, Torrente Bitonto, Pignola Abriola, Rupe del Corvo and La Ralla (see Appendix 1).

2.3.2.2. Type sections and thickness

Three type sections have previously been described for the three 'facies' of the formation respectively (Scandone, 1967b,1972). None of these, however, are complete, and quoted thicknesses are necessarily inexact; they range from 165m at M. Armizzone to 200m at M. Pierno and 230m at Pignola-Abriola. Re-examination of these and other sections has failed to give anything other than minimum values; 190m at M. Armizzone, 100m at Torrente Bitonto, 145m at Pignola-Abriola and 153m at Rupe del Corvo. It does appear, however, that the formation may not attain the thickness it does in Unit 1.

2.3.2.3. Lithology

No individual members have been distinguished in the Sirino Formation of Unit II, which lithologically resembles the M. Sirino Member of Unit I; the dolomitisation that affects the formation north of Marsico Nuovo is secondary, the contact with the limestones being strongly discordant to bedding, and it is thus innappropriate to consider the dolomites as a separate member (see Chapter 7).

Thinly bedded grey calcilutites with chert nodules are the dominant rock type, being more thickly bedded in the central portion of the succession. Thin intercalations of green shale characterise both the basal and uppermost parts of the formation. Calcarenite beds, which may be graded and contain large number of Halobia sp. shells, occur sporadically throughout. For a short distance above the base, the limestones have a slightly nodular aspect imparted by anastomosing clay seams and irregularity of bedding.

Extraformational carbonate breccias occurring at the top of the succession are present throughout the northern part of the zone, achieving maximum development in the north where they are dolomitised;

further south, they are replaced by intraformational calcirudites at a similar stratigraphic level. The proportion of dolomitisation follows a similar trend, gradually decreasing from totality at S. Fele to its final disappearance at Marsico Nuovo. Breccias in the dolomites may be recognised in the field by the presence of angular chert fragments.

The base of the formation is defined as the base of the lowest cherty limestone bed. The uppermost such bed is not so easily recognisable due to the presence of thinly bedded limestones in the base of the overlying formation, although these differ in colour and possess sedimentary structures generally absent in the limestones of the Sirino Formation. The top of the uppermost intraformational calcarenite or calcilutite has therefore been chosen.

2.3.2.4. Palaeontology, age and correlation

At M. Armizzone, levels of Halobia styriaca, H. austriaca and H. superba have all been found in the lower 20m of the formation (de Capoa Bonardi, 1970). The latter two species are considered to indicate a Late Carnian (Tuvalic) age, and the first an Early Carnian (Cordevolic) age. Their stratigraphic proximity implies that either sedimentation rates in the Carnian were very low or that parts of the succession are missing. However, the Lower Carnian age given for the H. styriaca level is based solely on its occurrence 10m above the Upper Anisian Daonella lomelli level of the Monte Facito Formation and, by comparison with other areas, its presence should in fact indicate a Middle Carnian (Julic) age (de Capoa Bonardi, 1970). In view of the intervening nodular limestone facies, a feature common to many stratigraphically condensed sequences in Tethyan pelagic facies, a Middle Carnian age is, perhaps, more likely, implying that the base of the formation may be younger in Unit II than in Unit I. The

appearance of H. Cassiana at the base of the formation in the Pignola-Abriola section also argues in favour of a Middle Carnian age (de Capoa Bonardi, 1970); here it occurs 65m below the closely separated levels of H. austriaca and H. superba.

No palaeontological data are available for the upper part of the formation that could provide points of correlation between the formations in the two units. Shallow water carbonate rocks of Late Triassic age, although invariably dolomitised, are present in both the Alburno-Cervati and Matese-Monte Maggiore carbonate platform units (Ippolito et al. 1975).

2.3.3. The Lagonegro Formation

2.3.3.1. Name and distribution

As for the Sirino Formation, the three 'facies' differentiated within this formation by Scandone (1967b, 1972) on the basis of the proportion of carbonate breccias and cherts in the succession have not been retained. Outcrops can generally be found in association with those of the Sirino Formation; of particular note are those at La Ralla, on the Pignola-Potenza and Pignola-Abriola roads, beside the road from Padula to Paterno, at Monte Armizzone and Torrente Bitonto, and sections have been measured at several of these localities. A further section, similar to that at La Ralla, is described from the area north of Monte Marzano (Cocco et al. 1974). Exposure of the rocks of the formation is generally not as good as in Unit I due to the predominantly less resistant lithologies and a greater degree of structural deformation.

2.3.3.2. Type sections and thickness

Of the type sections given by Scandone (1967b), that at Pignola-Abriola is far from complete, but the others, at La Ralla and Torrente

Bitonto, are exposed almost in their entirety and without any obvious stratigraphic gaps or repeats. Situated at opposing ends of the zone, they are very different, and represent end members of a spectrum of intermediate examples. It is considered appropriate, therefore, to retain both of these in conjunction as type sections, since neither is truly representative of the formation as a whole.

It is clear from the type sections that the formation is considerably thicker in this unit than in Unit I, the successions at La Ralla and Torrente Bitonto measuring 273m and 203m respectively. This is due only in part to the presence of large numbers of calciturbidites that are almost absent in the other unit, there being in addition a greater thickness of cherts and siliceous shales; the difference in thickness between the two type sections may, however, be accounted for by the greater proportion and thickness of the calcareous breccia beds in the north. Thicknesses at intervening localities are difficult to assess due to the lack of complete sections, but a minimum of 215m can be observed in the Pignola-Potenza and Pignola-Abriola road sections.

2.3.3.3. Lithology

The formation does not outcrop extensively in the mapped area, and is strucutrally highly deformed; no individual members have, therefore, been distinguished. Furthermore, the lithological features upon which such a distinction could be made vary laterally to such an extent that any lithostratigraphic sub-divisions would have limited applicability outside the immediate vicinity of that area. Taken as a whole, the formation comprises red, grey, green and yellow shales, siliceous mudstones and thinly bedded cherts, interbedded in varying proportions but showing a progressive increase in silicification up section. These siliceous deposits are intercalated with calciturbidite beds, up to

20cm thick in the south and as much as 6m in the north, which may be partially or totally silicified or dolomitised, exhibit normal grading and rarely possess sole marks. Dolomitisation, which affects only the lower parts of the formation, has not been observed south of Marsico Nuovo (Scandone, 1967b).

The formation extends from the top of the highest intraformational carbonate breccia of the Sirino Formation to the level at which the green and red cherts and siliceous mudstones are replaced by the grey/black shales of the Brusco Formation.

2.3.3.4. Palaeontology, age and correlation.

Although there are more palaeontological data available for the formation in this unit than in Unit 1, they are all derived from extra-formational carbonate breccias and thus can only give maximum ages. Furthermore, dating was complicated until recently by the confusion of the 'scisti silicei' with the shales of the Monte Facito Formation, both of which were originally thought to be Triassic (de Lorenzo, 1893, 1894). Tacoli and Zoja (1957) collected Conisconus alpinus and Trocholina elongata 90m below the top of the section on La Ralla, upon the basis of which they gave an Early Cretaceous age for the entire succession; Crescenti (1966) has pointed out that the stratigraphic range of these species extends down into the Malm. Almost a decade later, Luperto (1966) reported a Malm fauna at almost the same stratigraphic level as Tacoli and Zoja, and Scandone (1963) found Dictyoconus alpinus, of Aalenian-Bajocian age, within 40m of the base of the formation. Elsewhere in the zone, Richetti (1961) and Luperto (1966) have documented faunas in the top of the Pignola-Abriola section that they both ascribe to the Malm, and de Castro (1963) records the Middle Lias at Giffoni Vallepiana.

In an attempt to date the formation more precisely, the indigenous radiolarian faunas of some cherts and siliceous mudstones (etched in hydroflouric acid) were examined under the scanning electron microscope. In general the quality of preservation of these siliceous micro-organisms does not even allow indentification at the genus level, but one possible example of Hagiastrum plenum RÜST was found in a radiolarian chert from the Torrente Bitonto section 280m above the base (Pl.2.1). Examples of this species are documented from the Upper Kimmeridgian-Lower Tithonian of the California Coast Ranges (Pessagno, 1977).

Richetti (1961) tentatively proposed a Dogger-Lower Cretaceous age for the formation but Crescenti (1966), in a review of all the available micropalaeontological data available, considered that it encompassed the entire Jurassic and possibly a part of the Late Triassic; Scandone (1972) restricts it solely to the Jurassic. The age of the base of the formation has not been determined but the top can be bracketed between the Lower Cretaceous Brusco Formation and the carbonate breccias of Malm age described above. Since both of these are maximum ages, and in view of the considerable thickness of cherts (90m) overlying the Malm breccias at S.Fele, the possibility that the formation extends into the Lower Cretaceous at some localities cannot be excluded.

2.3.4. The Brusco Formation

2.3.4.1. Name and distribution

Exposure is essentially limited to the northern part of the zone, although minor isolated outcrops occur within the mapped area. These are commonly found in association with Upper Cretaceous and younger formations, rather than with those of the Lower Mesozoic, possibly implying that the upper parts of the succession have become structurally

uncoupled at the level of the rheologically incompetent Brusco Formation.

2.3.4.2. Type section and thickness

No sections have been measured and, for the same reasons as given for Unit I, there is no type section for the formation; it is thus not possible to give an estimate of its thickness.

2.3.4.3. Lithology

The formation comprises an alternation of silicified calcilutites, calcarenites, carbonate breccias, shales and rare marls. The shales are dark grey to black in colour and show a characteristic prismatic fracture; they are devoid of fauna. The lighter-coloured grey, green or white calcilutites are commonly completely silicified, and beds may grade at the base into calcarenites or breccias; sedimentary structures, with the exception of grading, parallel lamination and rare sole structures on the bases of some calcarenite and breccia beds, are absent.

2.3.4.4. Palaeontology, age and correlation

Several biostratigraphic determinations of the faunas contained in the calcarenites and breccias have been made (Ricchetti, 1961; Brönnimann et al. 1971; de Stasio, 1971). There is general agreement on the Lower Cretaceous age of the formation, the base being dated as Portlandian-Valanginian (de Stasio, 1971) or Upper Berriasian-Valanginian (Brönnimann et al. 1971). The latter age is based upon the recognition of Neocomian calpionellids whose pelagic affinity may imply a true, rather than simply a maximum, age. The top of the formation is not so well dated, although the overlying 'flysch rosso' contains Upper Cretaceous microfaunas near the base (Cocco et al. 1974).

2.3.5. The 'flysch rosso' and 'flysch numidico'

The most complete stratigraphic record of the Late Mesozoic and Tertiary history of the Lagonegro Basin is preserved in Unit II, particularly in the most northerly outcrops. A poorly-exposed and highly structurally deformed succession of pink and white fine-grained pelagic limestones, marly limestones, marls and mudstones, interbedded with pink or grey calcarenites that are commonly graded, can be found beneath the basal thrust of the Alburno-Cervati Unit along the Torrente Pergola north of Brienza. The stratigraphic relationship of this sequence, called the 'flysch rosso' (Scandone, 1967), or the Toppo Camposanto Unit (Scandone, 1971), to the Brusco Formation cannot be established in this area but conformable contacts are documented from the region surrounding Pescopagno, west of S. Fele (Cocco et al. 1974). Here a 350m succession is described of similar aspect to that at Brienza, but also containing red or green cherts and siliceous mudstones near the base and, higher up, calcirudites; slumped horizons and slump breccias are also present (Cocco et al. 1974). This sequence, of Late Cretaceous-Aquitanian (Early Miocene) age, is succeeded by the 'flysch numidico', comprising up to 60m of thickly bedded, structureless quartzose sandstones that are dated as Langhian (Middle Miocene)(Cocco et al. 1974)

2.4. The Monte Foraporta Unit

The Monte Foraporta Unit outcrops along the western flank of the Sirino Window, where it occupies a structural position between the overlying Alburno-Cervati Unit and the Lagonegro Unit II. The principle exposures are to be found in the valley of the F. Noce between Tempa Pertusata and Serra Luceta. The recognition of a distinct structural-stratigraphic unit comprising these deposits was

initially made by Scandone (1972) and subsequently refined by Boni et al (1974), although the existence of Liassic deposits distinctive from the coeval limestones of the Alburno Cervati Unit, had been acknowledged since the turn of the century (de Lorenzo, 1894; Greco, 1900). Two formations have been described, the La Calda (Dolomite) Formation and the Serra del Palo (Limestone) Formation (Boni et al. 1974); outcrops of both formations have been found in the area mapped.

2.4.1. The La Calda Formation

The lower of the two formations is well exposed in road cuttings beside the Superstrada del Noce SS18 between its intersection with the Autostrada A3 and the junction for Lagonegro. The original base of the formation is unknown, since it now everywhere overlies a thrust, but it appears to be conformable with the overlying limestones of the Serra del Palo Formation, although the contact is not exposed; no type section has been described. Incomplete sections have been measured at intervals along the Superstrada (Marea and Malamugliera Sections). (See Appendix 1).

Two members have been identified, a lower Dolomitic Member overlain by a Limestone/Dolomite Member (Boni et al. 1974). The lower of the two comprises thinly bedded (5-50cm) grey, black or brown dolomites that commonly show planar laminations and, more rarely, grading. Sapropelic horizons are intercalated with the dolomites, and some chert bands are also present. The overlying member can be distinguished by a lower proportion of dolomite beds and the presence of graded breccias up to 50cm thick interbedded with the more fine-grained limestones and dolomites.

The thickness of the formation has not been measured, but estimates of 100-120m and 30-50m have been given for the Lower and Upper Members

respectively. It is thought to be of Late Triassic-Early Liassic age, although no biostratigraohic data have been published (Boni et al. 1974).

2.4.2. The Serra del Palo Formation

The Serra del Palo Formation outcrops continuously between Serra del Palo itself and Tempa Pertusata in the north, as well as on M. Foraporta, Serra Luceta, M. Iatile and in several scattered klippen within the Sirino Window (e.g. Buonaficenza, Giardini dei Tuori). It is structurally bounded above by the Alburno-Cervati Unit.

The two type sections described by Boni et al. (1974) are poorly exposed and cannot be measured with any accuracy; a continuous section, however, is described throughout the upper part of the formation of Serra Luceta. The formation is divided into three members, the Lower Limestone Member, the Limestone/Marl Member and the Upper Limestone Member (Boni et al. 1974). The former comprises an estimated 100m of thinly bedded dark grey calcarenites and calcilutites, that commonly show parallel laminations and may be graded; they are intercalated with rare marls and claystones. These are succeeded by approximately 50m of alternating yellow limestones, marly limestones and claystones with rare intraformational conglomerates that comprise the Limestone/Marl Member; these are well exposed in the Canale del Torno. The highest member, something over 100m in thickness, includes dark grey mud- to sand-grade limestone beds, the coarser examples commonly being laminated and rarely graded.

Rich faunas in the calcarenite beds have yielded Middle Lias, Upper Lias-Lower Dogger and Dogger ages for the three members respectively; the latter may, however, extend into the Malm (Boni et al. 1974).

CHAPTER THREE

STRUCTURE

3.1 Introduction

In order to determine the stacking sequence of the three units comprising the Lagonegro Zone, a northwest-southeast segment of the Sirino Window, between Monte Farno and Monte Sirino, has been mapped on a scale of 1:10,000 (See enclosures). This 40km^2 area encompasses major contacts between all three units, and the considerable topographic relief allows an appreciation of the three-dimensional geometry of the area (Pl. 3.1). Essentially, the segment comprises a series of generally north-south-trending anticlines and synclines which are commonly asymmetric with an easterly vergence (Fig.3.1). These major folds are dissected by a complex array of normal faults which hamper recognition of the primary structural relationships between the three units. Primary thrusts, which are both folded and faulted during later deformation, can, however, be identified and it is possible to establish that the Lagonegro I, Lagonegro II and Monte Foraporta Units were originally stacked in ascending order (Fig.3.1).

In addition to simply determining the geometrical relationships between the units, major and minor structures have been studied in an attempt to deduce emplacement vectors for each thrust sheet, an important prerequisite for making a palinspastic reconstruction of the pre-orogenic morphology of the basin. It emerges that all of the units were emplaced from west to east, implying that the uppermost sheet, the Monte Foraporta Unit, lay to the west of the Lagonegro basinal units, and that the Lagonegro Unit II occupied the western part of the basin.

The megascopic and mesoscopic structures of the three units are

described and interpreted in Sections 3.3, 3.4 and 3.5, and the structural relationships of the zone to other units, in Section 3.6. A model for the deformational history of the zone is presented and discussed in the final section.

3.2 Analytical techniques

3.2.1. The eigenvalue method

Fabric orientations have been analysed using an eigenvalue technique, and have been plotted on conventional equal-area projections (Scheidegger, 1965; Watson, 1966; Woodcock, 1977). The eigenvalues and eigenvectors computed by this technique have been used to specify fabric distributions as outlined by Woodcock (1977). Briefly, the eigenvector v_1 gives an estimate of the distribution mean, while v_3 is an estimate of the pole to the best-fit girdle of the distribution; eigenvector v_2 is mutually perpendicular to v_1 and v_3. These vectors have been plotted on the equal area projections as open circles and marked 1, 2 and 3 appropriately.

The eigenvalues, λ_1, λ_2, and λ_3, or in their normalised form, S_1, S_2 and S_3, are a measure of the degree of clustering of the distribution about the corresponding eigenvectors. Woodcock (1977) has shown that a plot of $\ln S_2/S_3$ against $\ln S_1/S_2$ can be used to distinguish populations that are clustered from those which form great-circle girdles (Fig.3.2). This tendency can be quantified by parameter K, where

$$K = \frac{\ln(S_1/S_2)}{\ln (S_2/S_3)}$$

Clustered distributions have a K value greater than 1, whilst for girdles, K is between 0 and 1. This parameter is quoted together with the number of data elements in the distribution, N, for all the

equal area projections.

3.2.2 The separation-arc method

The separation-arc method of Hansen (1965, 1971) has been used to determine emplacement vectors for each of the thrust sheets from the orientation and sense of rotation of asymmetrical folds associated with the thrusts. Such folds commonly plot as girdles whose orientation corresponds with that of the thrust plane, and their senses of asymmetry, or overturn, are recorded on equal-area projections (Fig.3,3); conventionally, the sense recorded is that observed by looking along the fold axis down-plunge. The bisector of the separation angle between the tips of the girdle can be used as an estimate of the emplacement vector, the sense of shear impled by the fold asymmetry being used to determine the direction of transport. This method can only be used in cases where the folds are directly related to a thrust and to no other major structure, since drag folds on the limbs of a megascopic fold could be indistinguishable from those developed in a thrust zone (Naylor & Harle, 1977).

3.2.3 Fold shapes

Fold shapes have been described in the field using the classification adopted by the Data for Orogenic Studies Project (Spencer, 1974) (Fig. 3.4). In some cases, dip isogons have been constructed from fold profiles and the classification of Ramsay (1967) has been used.

3.3 Lagonegro Unit I

3.3.1 Description

The Lagonegro Unit I, the lowest of the zone, outcrops principally in the southern part of the mapped area, although an isolated tract also occurs in the north around Monte Farno (Fig.3.1.). The base of

unit is not exposed, and the basal limestones and shales of the Gianni Griecu Member invariably occupy the cores of major anticlines. The unit is bounded at the top, however, by the Tempa la Secchia Thrust which separates it from Lagonegro Unit II in this area.

3.3.1.1 Megascopic structures

Poles to bedding planes plot as girdles which define the orientation of megascopic fold structures with horizontal axes lying approximately north-south (Fig.3.5). The axial traces of these folds are arcuate, trending slightly west of north in the Monte Farno fault block, and slightly east of north in the south (Fig.3.1). Several major anticlines can be distinguished which are responsible for exposing the unit in most of the area. In the Monte Farno fault block, these take the form of symmetrical open folds with a wavelength of about 500 m (Fig.3.5a), but the Gianni Griecu and Monte Gurmara blocks are dominated by an asymmetrical anticline that bifurcates before disappearing beneath the Tempa la Secchia Thrust north of Gianni Griecu itself (Fig.3.1): this structure is slightly overturned to the east, and fold axes plunge gently to the north (Fig.3.5c & d). Along the western edge of the area, the Brusco Formation is exposed in a syncline which separates the Gianni Griecu-Monte Gurmara anticline from a similar structure forming the Costa del Alto-Monte Nicola- Mizzo Milego ridge (Fig.3.5b).

A more symmetrical anticline outcrops in the Monte Niella and Bramafarina blocks (Fig. 3.5e & f), but on the summit of Monte Sirino, a series of tight, almost isoclinal folds can be distinguished, whose axes plunge gently northwest, and whose axial planes dip both southwest and northeast (Fig.3.5g). Lack of exposure prevents these structures from being followed northwestwards. The easternmost ridge

of Monte Sirino constitutes a monocline whose northeast-southwest axis is almost at right-angles to the dominant fold trend (Fig. 3.5h).

The overlying Tempa la Secchia thrust is also folded by these anticlines and synclines, but the stratigraphic position of the thrust changes across strike; in the Gianni Griecu- Monte Gurmara anticline it follows the top of the Lagonegro Formation, whereas in the syncline between Monte Gurmara and Monte Niella it occurs at a higher level within the Brusco Formation. On Monte Niella, a minor thrust fault is present that dips steeply to the west at a high angle to the Tempa la Secchia thrust, cutting through the western limb of the Monte Niella anticline.

3.3.1.2 Mesoscopic structures

Parasitic drag folds, whose axial orientation and asymmetry sense mimic those of the major anticlines, are abundantly developed in the limestones of the Sirino Formation (Fig.3.6d, e & f). Shapes vary from D1 and D2 to E1 and E2, fold hinges are generally well rounded and folding is most commonly harmonic. On the south side of Gianni Griecu, however, in the upright limb of the major anticline, strongly disharmonic monoclines are present which overlie undeformed strata along horizontal minor thrust planes.

A distinctive style of folding characterises outcrops of the Gianni Griecu Member, which are invariably strongly deformed. In the tract exposed at the core of the Gianni Griecu-Monte Gurmara anticline, where it is dissected by the Fiume Pietra, moderately large, straight-limbed folds with angular hinge zones, whose shapes include D3, D4, E2, E3 and E4, are developed (Pl.3.2 & 3.3). The axial planes to these folds, which trend almost north-south and are overturned towards the east, consistently dip westwards. Deformation is much more intense

than in the overlying limestones of the Sirino Member.

A further outcrop of the member 1 km southeast of Monte Niella shows similarly strong deformation, and here too the folds, though smaller, show a consistent sense of asymmetry (Pl. 3.4 & 3.5). Due to the alteration of rheologically competent limestone beds with less resilient shale horizons, the folding is strongly disharmonic between layers; multilayer fold profiles indicate that the folds belong to class 2 of Ramsay (1967), although the limestones and shales individually form class 1B or 1C and class 3 folds respectively (Fig. 3.7; Pl.3.6). An axial-planar cleavage is commonly developed in the hinge zone that is divergent in the shale horizons and convergent in the limestones (Pl.3.4). The axes of these folds form a well-defined cluster (K=8.32) plunging gently north-north-east (Fig.3.6a & c); axial planes dip west-north-west (Fig. 3.6b).

At Sorgente Acero, in the Vulturino Window, the Gianni Griecu Member also outcrops in the core of an asymmetrical anticline that is almost isoclinal and is overfolded to the east; minor asymmetrical folds show a similar vergence to the major structure (Fig.3.9 g & h).

Minor folds are also abundant in the cherts near the top of the unit close to the Tempa la Secchia Thrust, especially in the Pietra Member. They are typically straight-limbed with highly angular hinge zones, fold shapes D2, D3, E2, E3, E4, F3 and F4 being most common (Pl.3.8). Folding is strongly disharmonic, and highly contorted strata may be separated from flat-lying beds beneath by a single minor shear plane (Pl.3.9); alternatively, minor thrusts may develop at an angle to bedding along the axial planes of folds (Pl.3.10). Fold axes are generally horizontal, and when plotted for individual outcrops tend to form clusters ($0.83 < K < 4.71$)(Figs 3.8a,

c, e, g & h, 3.9a, c and e). Plotted together, however, the distribution forms a horizontal girdle (K=0.61)(Fig.3.6g). The sense of asymmetry of the folds is consistent throughout the area, irrespective of whether they are situated on the upright or inverted limbs of the major anticlines, thus ruling out the possibility that they are parasitic drag folds.

3.3.2 Interpretation

Since the base is not exposed, it is not possible to prove that the unit is allochthonous, as has been demonstrated in the Campagna tectonic window (Scandone, et al., 1967; Turco, 1976). The presence of deformed rocks of the Gianni Griecu Member in the cores of major anticlines, however, and the structural style of the unit, suggest that décollement has occurred, and by extrapolation from the evidence obtained at Campagna, it is likely that the unit is also allochthonous in the Lagonegro area.

The easterly vergence of the magascopic structures towards the stable foreland on the eastern side of the Italian peninsula, and the common development of minor thrust planes in association with parasitic mesoscopic folds, suggests that folding deformation was related to the emplacement of the Lagonegro I and overlying sheets onto the more external terrains; since the Tempa la Secchia Thrust is itself folded, it is clear that this deformation occurred after superimposition of the two Lagonegro units. The divergence of fold-axis orientations in adjacent fault blocks on Monte Sirino may be the product of rotation during emplacement rather than polyphase folding, since neither block shows evidence of re-folding of fold axes (c.f. Scandone, 1967b, 1972). The overall easterly vergence of both major and minor folds implies that tectonic translation was broadly from west to east.

The features of the asymmetric folds developed in the limestones and shales of the Gianni Griecu Member are typical of buckling deformation in layered sequences (Ramsay, 1967). The competent limestone beds appear to have undergone longitudinal tangential strain whilst the shales have deformed by flexural slip. The orientations and senses of asymmetry of these folds are coincident with those of the megascopic structures in the area, implying that they too relate to the emplacement of the unit. The translation vector determined by the separation-arc method of Hansen (1965, 1971) is towards the southeast (Fig. 3.6c, 3.9h).

The identical vergence of the asymmetrical folds in the cherts of the Pietra Member close to the Tempa la Secchia Thrust in both the upright and inverted limbs of the asymmetrical anticlines implies that their formation predated the major folding phase, and their spatial association with the thrust favours the contention that they formed in response to shear stresses which acted during the overlapping of the two Lagonegro units; the development of minor thrusts along the axial planes of some of these folds also militates in favour of their having originated in this way. As is common in the case of deformation associated with the translation of single sheets, fold axes tend to be distributed around a horizontal plane rather than in a cluster, although this pattern may have been obscured to some extent by the later folding (c.f. Hansen, 1965, 1971). Considering single outcrops in isolation, the determined tectonic vectors vary between northeast and southeast, whereas plotted together, the fold distribution suggests a northeasterly vector (Figs. 3.6h, 3.8b, d & f, 3.9b, d & f). There is no significant difference, therefore, in the direction of transport during the initial superimposition of the two Lagonegro units and the later folding events.

3.4 Lagonegro Unit II

3.4.1 Description

In the area mapped, the Lagonegro Unit II outcrops principally in an east-west belt between Tempa la Secchia and Tempa di Roccarossa and in the syncline between the Gianni-Griecu-Monte Gurmara and Monte Niella-Bramfarina anticlines (Fig. 3.1). It is separated from Unit I around Monte Farno by a major east-west fault, but is bounded in the south by the Tempa la Secchia Thrust. On Giardini dei Tuori, it is possible to observe the top of the unit, which is marked by the Giardini dei Tuori Thrust (Pl. 3.11).

3.4.1.1 Megascopic structures

The overall structure of Unit II mirrors that of the lower unit, and two north-south anticlines can be distinguished which correspond directly to the Gianni Griecu-Monte Gurmara and Monte Niella-Bramafarina anticlines on Unit I; an additional anticline developed 500m west of Rocca Rossa represents a branch of the former. A plot of poles to bedding indicate that the structures plunge gently north-north-east (Fig. 3.10a). Although possessing an eastward vergence, the megascopic folds of Unit II are more open than in Unit I to the south, their style more closely resembling the folds of the Monte Farno block.

3.4.1.2 Mesoscopic structures

Coherent minor folds are generally less common than in Unit I, due largely to the less competent nature of the dominant lithologies (Scandone, 1967b, 1972). None, therefore, have been found in the area mapped which can be related unequivocally to the thrust planes at the top and base of the unit. In an exposure of the Lagonegro Formation west of Rocca Rossa, however, chevron folds are extensively

developed with axes oriented approximately north-south and axial planes that dip to the west (Fig.3.10b, c & d)(Pl. 3.12). Although fold axes show a slight tendency to cluster (K=1.09), they are relatively dispersed and may possibly have been refolded (Fig. 3.10d). The outcrop occurs between the two branches of the Gianni Griecu-Monte Gurmara anticline, but the orientation of the fold axes does not correspond to the axis of the major structures in the unit (Compare v_3 of Fig. 3.10a with v_3 of Fig. 3.10b and v_1 of Fig. 3.10c).

A tectonic window through the Monti della Maddalena Unit east of Padula exposes the Lagonegro and Brusco Formations of Unit II directly beneath the thrust separating the two units (Fig.3.11). The rocks outcrop in a north-north-west-trending anticline and are deformed by a set of abundant asymmetrical folds which show a consistent easterly vergence on both limbs of the major structure (Pl.3.13). Folds are open to closed, but are commonly straight-limbed with angular hinges, and fold shapes include C2, D2, D3, E2, E3 and E4; fold axes are strongly clustered about an almost horizontal, north-south principle eigenvector (K=8.27)(Fig.3.10e). Shearing along axial planes, which dip to the west, is common, and in one case, conjugate shear planes are developed in the horizontal limb of a fold (Pl. 3.14 & 3.15).

3.4.2 Interpretation

The continuity of the major structures of Unit I through the outcrop area of Unit II gives additional evidence in support of the contention that folding post-dated super-position of the units. The more open style of the anticlines in this tract, as well as in the Monte Farno block, is considered to be a response to the bifurcation of the Gianni Griecu-Monte Gurmara anticline, the stratal

shortening being accomodated by two smaller folds rather than a single large one.

The obliquity of the axes of the chevron folds to the orientation of the major structures suggests that they did not develop during the principle folding phase. On the other hand, the v_1 eigenvector of the distribution is almost coincident with that of the folds from the Padula tectonic window, which also does not correspond to the major structure. In view of the consistent sense of asymmetry shown by the latter folds, their close spatial association with the thrust, the high incidence of shearing along axial planes and the lack of apparent relationship to the major anticline, it is considered most likely that both fold sets developed during overthrusting of the Campania-Lucania platform units onto Lagonegro Unit II. The easterly tectonic vector that can be deduced from these data imply that the emplacement direction was the same as during the ensuing phases (Fig.3.10f).

3.5 Monte Foraporta Unit

Exposure of the Monte Foraporta Unit in the area mapped is limited to two small tracts on Buonaficenza and Giardini dei Tuori (Fig. 3.1). In the former, the unit overlies flysch deposits of uncertain affinity, but which may belong to the Sicilide Unit, along a thrust. In the latter, however, the unit is situated structurally above the cherts of the Lagonegro Formation of Unit II along a thrust (Pl. 3.11). The outcrops are too small to allow identification of major structures, but the Giardini dei Tuori Thrust, together with the bedding in the La Calda Formation dolomites above, is folded into a gentle anticline along with the underlying cherts. Both outcrops are devoid of minor structures, although there is abundant evidence of shearing in the dolomites above the thrust.

On the south side of Monte Foraporta itself, 2.5 km northwest of Lagonegro, the basal thrust of the unit is also well exposed (Pl. 3.16). Here, however, grey dolomites of the La Calda Formation overlie Triassic white dolomites of the Monti della Maddalena Unit, and there is an angular relationship between bedding in the upper unit and the contact. This contact is extremely abrupt, occurring across as little as 1 cm, and although shearing is abundant both above and below, no asymmetrical folds have been observed (Pl. 3.17).

This surface has hitherto been interpreted as a major thrust, the rocks of the two units being thought to have originated in different palaeogeographical units (Scandone, 1972; Boni et al., 1974); the deposits of the Monte Foraporta Unit are considered to have accumulated in a marginal basin on the edge of the Campania- Lucania Platform that was separated from the Lagonegro Basin by a structural high, now represented by the Monti della Maddalena Unit. Although the contact is manifestly tectonic, its extreme sharpness, and the absence of a demonstrable stratigraphic jump across it, militates against its being a major thrust, and the inclination of the bedding in the upper unit suggests that it may represent a tectonised onlap. An alternative model for the origin of the Monte Foraporta basin is presented in Chapter 7.

3.6 Relationship of the Lagonegro Zone to other units

3.6.1 Overlying units

On the west side of the Torrente Pergola, directly opposite the Castello in the ancient village of Brienza, Triassic dolomites at the base of the Monti della Maddalena Unit can be seen thrust over Late Cretaceous or Palaeogene deposits of the Lagonegro Basin (Pl. 3.18). The red shales, marls and pelagic limestones of the

'flysch rosso' that lie beneath the westerly dipping thrust plane are intensively brecciated and sheared, and the clasts, elongated in a northeasterly direction, impart a mullion-like texture (Pl. 3.19). Bedding in the dolomites cannot be discerned, but the underlying strata are parallel to the thrust plane. Below the zone of shearing, asymmetrical folds with an easterly vergence, associated with minor thrusts, are commonly developed (Pl. 3.20). Although there are insufficient data from which to determine a tectonic vector, the orientation of the thrust plane and the folds beneath it, as well as the elongation of the clasts, provide support for the postulated westerly provenance of the Monti della Maddalena Unit implied by the data for the Padula tectonic window.

At several localities in the Campagna tectonic window, and on the eastern flank of Monte Raparo, 17 km east of Lagonegro, the thrust contact between the Lagonegro Unit II and the Alburno-Cervati Unit can be observed (Scandone, 1967b, 1972; Scandone et al., 1967; Ortolani & Torre, 1971; Turco, 1976).

3.6.2 Underlying units

With the exception of those exposures documented from the Campagna tectonic window, where the Lagonegro Units can be observed structurally overlying the Monte Croce Unit, there are no outcrops of the basal thrust of the zone (Turco, 1976). There is no direct evidence in the remainder of the region, therefore, for believing that the basinal units were ever thrust onto the external Abruzzi-Campania Platform at all. Nevertheless, the hypothesised continuity of the Matese-Monte Maggiore Unit, which represents the remains of this platform, as far south as Monte Alpi, suggests that the Lagonegro Units must have overlapped the external carbonate unit

by at least 50 km, if the platform margin is assumed to have been straight (Figs. 1.2 & 1.2a) (Ortolani & Torre, 1971; Pescatore & Ortolani, 1973). It is possible, however, that a major embayment existed in this margin between Campagna and Monte Alpi, in which case the Lagonegro Units may simply have been rucked-up against the edge of the platform as it was overidded by the Alburno-Cervati and Monti della Maddalena Units (Fig.3.12b). It is suggested that the observed change in the orientation of major fold axes in the outcrop of the zone from north-south in the south to almost east-west in the north, contrary to the regional trend, is due to structural accommodation of the deformed basinal units against the indented platform margin (Fig. 3.12a) (c.f. Scandone, 1972; Pescatore & Ortolani, 1973).

3.6.3 Late orogenic normal faulting

The southern Apennine tectonic units are not invariably separated by thrusts, and are more commonly in faulted contact. In the area mapped, normal faulting is abundant, the faults cutting both thrusts and major folds. They represent, therefore, the ultimate phase in the structural history of the zone, probably relating to the Plio-Pleistocene uplift of the Apennine chain (Scandone, 1967b, 1972; d'Argenio et al., 1973, 1975; Pescatore & Ortolani, 1973; Ippolito et al., 1975). Two principle trends can be distinguished, oriented north-north-east - south-south-west and west-north-west - east-south-east. Examples can be found of faults of either set cutting those of the other, demonstrating their contemporaneity. They show a pronounced relationship to the orientation of the earlier structures, however; for example, a major fault follows the steeply dipping eastern limb of Gianni Griecu Anticline.

3.7 Deformational history of the Lagonegro Zone

Detailed mapping has therefore confirmed the stacking sequence and emplacement directions of the thrust sheets of the zone previously advanced by Scandone (1967b, 1972) and other workers in the southern Apennines (Pescatore & Ortolani, 1973; d'Argenio et al., 1973, 1975; Ippolito et al., 1975; Catalano et al., 1976). It is clear, however, that the structural history of the basin has hitherto been greatly oversimplified, and the stacking sequence of the structural-stratigraphic units is not necessarily consistent throughout the chain.

3.7.1 The model

By restoring the thrust sheets to their original positions, it is possible to construct a pre-Miocene palinspastic model of the basin (Fig.3.13). The shape of the basin can be approximately determined from the outcrop pattern of the basinal units themselves and of the adjacent platform units. Since no outcrops of the zone occur northwest of Campagna, and because the Alburno-Cervati and Matese-Monte Maggiore units are in direct structural contact north of Naples, it is assumed that the basin terminated in this region; it is conceivably possible, however, that the Lagonegro Units continue beneath the Alburno-Cervati Unit out into the Tyrrhenian Sea. The Morphology of the Campania-Lucania Platform, which formed the internal margin of the basin, can be reconstructed with some accuracy, since marginal facies outcrop in linear belts down each side of the Alburno-Cervati Unit along its entire length in the Bulgheria-Verbicaro and Monti della Maddalena Units (Fig.1.2) (Ippolito et al., 1975). From the disposition of these facies, it appears that the external margin of the platform was almost straight.

The internal margin of the Abruzzi-Campania Platform, however, which enclosed the basin to the east, is almost entirely hidden beneath overlying thrust sheets, and its shape is therefore unknown. As discussed in Section 3.6.2, the structure of the Lagonegro Units, and the morphology of the syn-orogenic Irpinian Basin, give the impression that this margin was indented between Campagna and Monte Alpi. The depicted morphology of this margin is, therefore, largely conjectural. To the south, the zone disappears once more beneath the Alburno-Cervati Unit, but the presence of similar basinal units in the Sicani and Imerese Zones of Sicily suggests a continuity of the zone beneath the Calabrian Complex (Scandone, 1975; Scandone et al., 1974; Catalano et al., 1976).

The super-position of the units derived from the Campania-Lucania Platform and the Lagonegro Basin is known to have taken place during the Langhian (Early M. Miocene), the Irpinian flysch basin having developed at that time between the advancing nappe fronts and the western margin of the Abruzzi-Campania Platform (Fig.3.14) (Scandone, 1967b, 1972; Cocco et al., 1972; d'Argenio et al., 1973, 1975; Pescatore & Ortolani, 1973; Ippolito et al., 1975). The timing is witnessed by the conformable deposition of flysch upon the youngest deposits of the Lagonegro Basin in the east, while in the west, the units derived from the Campania-Lucania Platform and the Lagonegro Basin are stacked without flysch interposed between them (Pescatore & Ortolani, 1973); furthermore, olistostromes containing material derived from the front of the Campania-Lucania Platform are to be found amongst the oldest sediments of the Irpinian Basin (Ippolito et al., 1975). Much of the Langhian flysch deposition occurred in an area corresponding to the postulated embayment in the

Abruzzi-Campania Platform margin, and it is herewith suggested that the flysch basin represented a gap left after collision of the linear Campania-Lucania Platform margin with the indented margin of the still-rooted Abruzzi-Campania Platform (Fig. 3.14).

During the Serravalian (Late M. Miocene), the external platform subsided rapidly, allowing flysch-deposition to encroach eastwards; the Tortonian (Late Miocene), however, saw the final elimination of the flysch basin as the units of the Campania-Lucania Platform, as well as those of the Irpinian Basin itself, overode the now-foundered external platform (Fig. 3.15) (Scandone, 1972; d'Argenio et al., 1973, 1975; Pescatore & Ortolani, 1973; Ippolito et al., 1975). It has hitherto been assumed that the folding of the Lagonegro Units occurred during the ultimate, Pliocene thrusting event, during which all the allochthonous units were emplaced onto the margin of the autochthonous Apulian Platform (Scandone, 1967b, 1972; d'Argenio et al., 1973, 1975; Pescatore & Ortolani, 1973; Ippolito et al., 1975). The evidence for this dating is meagre, however, the orientation of the folds simply being co-incident with other folds developed in Early Pliocene deposits elsewhere in the basin (Pescatore & Ortolani, 1973). It is considered more likely that this folding, which is notably more intense in the Lagonegro Units than in either of the adjacent platform-carbonate units, was a consequence of Tortonian movement along the basal thrust of the Campania-Lucania Platform units as they were emplaced onto the external platform, leaving the basinal units behind (Fig.3.15).

3.7.2 Discussion

The 'piggyback' mode of emplacement of the southern Apennine structural-stratigraphic units, whereby the uppermost sheets were

emplaced first, is a typical feature of thrust belts, and the currently favoured 'gravitational spreading' hypothesis is commonly advanced as a mechanism for their development (Elliott, 1976). A consistent stacking sequence and the deposition of a clastic wedge ahead of the nappe fronts are considered to be diagnostic of such edifices, which develop due to sudden increases in pore pressure in the autochthon following emplacement of an overlying thrust sheet; this leads to de-coupling of the autochthon from its basement, generating a new thrust. Provided suitable décollement horizons are present within the foreland sequence, therefore, it can be seen that thrusting is self-perpetuating, the thrust zone migrating laterally for considerable distances; all that is required to set the process in motion is an initial shunt, which is probably provided in most cases by gravity (Elliott, 1976).

In the case of the southern Apennines, emplacement of the Calabrian Complex in the Early Miocene probably initiated the 'piggy back' process; comprising both ophiolitic and crystalline nappes, this complex is interpreted as a part of the Alpine chain, and, as such records the continental collision of Africa and Europe in the Palaeogene (Alvarez, 1976; Amodio-Morelli et al., 1979). The back-thrusting of the Calabrian nappes, which have a European vergence, onto the southern margin of Tethys in the Miocene possibly reflects the overlapping of irregularities in the margins of the converging continents.

Although gravitational spreading can explain the observed structural history of the external zones, the internal units, which include the Liguride and Sicilide Complexes, are commonly interposed between the external units; thus the criterior of consistent

stacking order is violated (Elliott, 1976). These units are generally composed of rheologically incompetent flysch deposits, however, and their structural aspect is invariably chaotic, resembling a 'mélange'. It is more likely, therefore, that these units were emplaced by gravitational sliding off the front of the Campania-Lucania Platform units during its emplacement.

CHAPTER FOUR

EXTENSIONAL TECTONICS AND THE BIRTH OF THE LAGONEGRO BASIN

4.1 Introduction

The oldest deposits of the Lagonegro basin, those of the Middle Triassic Monte Facito Formation, are characterised by discontinuous blocks of neritic limestone surrounded by shales (Fig.4.1). Although the formation is only exposed in Lagonegro Unit II, décollement in Unit I occurred at a higher stratigraphic level and it is probable that similar successions originally underlay the entire basin. The proximity of the formation outcrops to the basal thrust of the upper unit has led to extensive tectonic modification of the primary depositional relationships.

Scandone (1967b, 1972) contended that the limestone blocks were patch reefs situated within a predominantly terrigenous shallow-marine environment and he interpreted the associated breccias as reef talus deposits: he did, however, acknowledge the possibility that some of the blocks may have had an allochthonous origin. The absence of reef lithofacies, and the clearly discordant relationships shown by many blocks to the surrounding clastics, argue strongly in favour of the latter hypothesis. Moreover, re-examination of the supposed reef talus breccias has revealed evidence to suggest that palaeokarstification was responsible for their genesis. The formation, therefore, is re-interpreted here as the product of disintegration of a young carbonate platform under the influence of an extensional tectonic regime associated with crustal thinning and regional subsidence of the Hercynian basement. The occurrence of similar deposits of Triassic age throughout the Tethyan region is considered to be a reflection of the fragility of the newly-formed carbonate platforms and the widespread inception of extensional

tectonic regimes at that time (Cros, 1967; Mascle, 1967; Orombelli & Pozzi, 1967; Glennie et al., 1973; Bechstädt et al., 1978).

The sedimentary facies of the Monte Facito Formation are described in Section 4.2, followed by a brief review of the igneous rocks of the formation in Section 4.3. In sections 4.4 and 4.5, an evolutionary model for the Lagonegro Basin during the Middle Triassic is presented and discussed in relation to development of the Tethyan region at that time.

4.2 The Sedimentary Facies

4.2.1. The Neritic Limestone - Shale Facies

i) Description

The lowest member of the Monte Facito Formation, the La Cerchiara Member, is exposed in a relatively undeformed tract between 1 and 1.5 kms south of La Cerchiara and 8 kms southof the village of Tito (Fig.4.1). The area has been mapped at a scale of 1:10,000 (Fig.4.2) and is bounded to the north, west and south by normal faults. A 400 m succession of monotonous yellow/brown shales containing large blocks of neritic limestone is underlain by a series of three tabular limestone units separated by structureless red or greeny/yellow shales; these comprise the Neritic Limestone - Shale Facies (Fig.4.3; Pl.4.1). The lowest unit is at least 50 m thick and the second and third are 4 and 10 m thick respectively.

These rocks are extensively cleaved and contain abundant calcite veins which may obliterate the primary fabrics. The limestones comprise intrabiosparmicrites with a variety of abraded and partially micritised skeletal fragments and micrite peloids, many of which support oncolitic overgrowths (Pl.4.3). Some allochems have been leached and replaced by sparry calcite, but micrite envelopes developed round their margins

are preserved intact. These originally unstable components, probably once composed of aragonite, may represent molluscan fragments: dascycladacean and solenoporacean algae, bryozoans, echinodermal plates, benthonic foraminifera, ostracods and rare corals can also be found. These particles are organised into poorly graded laminae about 1 cm thick but show no preferential orientation. Grains at the tops of laminae are set in a micritic matrix which gives way beneath to a coarse sparry calcite cement: successive beds incorporate intraclasts of biomicrite identical to that forming the tops of underlying laminae.

The upper tabular limestone unit is pervaded by an abundance of irregularly shaped marl- and calcite-filled cavities and is capped by a breccia of sub-rounded clasts set in a marly matrix (Pl.4.2). The unit itself is discontinuous and cavities are especially densely distributed close to the margins of the isolated fragments. In thin section, the primary depositional fabrics are almost entirely obliterated by the development of a honeycomb of irregularly shaped pores; only those particles which have been leached and replaced by sparry calcite can still be recognised (Pl.4.4). This pore network has been incompletely filled by a radiaxial fibrous calcite cement, marl or, more rarely, by sparry calcite that occupies the final voids. The radiaxial fibrous cement, however, does not exhibit good crystal terminations and cement-substrate contacts are commonly truncated.

ii) Interpretation

These bioclasts and pelletal intrasparites and intramicrites are interpreted as a skeletal sand facies deposited in a back-reef lagoon and stabilised by blue-green algae. The graded laminae probably represent discrete depositional events by gravitational settling, and the apparently poor grading, most notable in the case of echinodermal fragments, may be a

function of grain density differences. The biomicrites forming the upper parts of laminae are the products of sediment binding by blue-green algae to form algal mats, some of which have been reworked and incorporated as intraclasts at the base of successive beds. Reef derived faunal elements testify to the proximity of either a marginal reef or, more likely, localised patch reefs; the dascycladacean algae, particularly Diplopora, may have preferentially colonised back-reef lagoons and probably represent the indigenous flora (Elliot, 1978). The absence of typical inter- and supra-tidal lithologies, such as vadose pisolites, stromatolites, birds-eye structures and sheet cracks (Bosellini & Rossi, 1974) or Lofer cyclothems (Fischer, 1964) suggests a subtidal environment. Skeletal sand facies of this type, stabilised by blue-green algae, are typical of modern back-reef lagoons (Bathurst, 1975); it is also a facies common to many Triassic carbonate build-ups of the Tethyan region where they are also identified with near back-reef settings (Bosellini & Rossi, 1974; Abate et al., 1977, Fürsich & Wendt, 1977; Toschek, 1968).

The marl- and cement-filled pore network of the upper unit is a typical dissolution fabric, and is interpreted as the product of palaeokarstification. The preservation of spar-replaced skeletal fragments indicates that the rock had already attained an advanced stage of fresh-water diagenesis prior to the onset of dissolution (Stage IV of Land et al., 1967). The observation that this solution-reprecipitation process, whereby unstable aragonitic skeletal fragments are replaced by low-magnesian calcite, is locally more than 90% efficient suggests that this early vadose cementation could have occurred without an external source of $CaCO_3$ (Harris & Matthews, 1968). Karstic solution would have yielded pore waters supersaturated with respect to

low-magnesian calcite which may have contributed towards the cementation of the underlying tabular limestone units: the radiaxial fibrous calcite cement which filled the solution cavities, however, is probably replacing an acicular marine cement (Kendall & Tucker, 1973). Palaeokarstification must therefore have ceased abruptly due to a rapid rise in relative sea-level. Truncation of the first marine cement and the filling of the late-stage cavities by marl or sparry calcite attest to a second phase of emergence, dissolution and fresh-water diagenesis. Palaeokarst features of Upper Anisian - Lower Ladinian age are common in the Alps (Cros & Lagny, 1969; Bosellini & Rossi, 1974; Epting et al., 1976; Bosellini et al., 1977); here too the cavities are commonly lined by a radiaxial fibrous calcite cement.

The Neritic Limestone - Shale Facies is thus thought to be the product of repeated suffocation of extensive shallow sub-tidal carbonate environments by fine grained terrigenous clastic sediments reworked from the underlying basement. The rapid facies transitions and the evidence for short periods of emergence in the limestones reflect the onset of synsedimentary tectonic activity. Alternating siliciclastic and shallow-marine carbonate sedimentation phases are documented from the Anisian at the base of the S. Donato Unit in Calabria where they represent the earliest deposits of a carbonate platform (Bousquet & Dubois, 1967; Ippolito et al., 1975; Amodio-Morelli et al., 1979; Dietrich, 1979); this unit has been equated with the units derived from the Campania-Lucania carbonate platform of the southern Apennines (Amodio-Morelli et al., 1979). The Neritic Limestone - Shale Facies of the Monte Facito Formation preserves the relics of an embryonic carbonate platform which failed to keep pace with the rapid regional subsidence and evolved, by default, into a basin.

4.2.2. Terrigenous Clastic Facies

i) Description

The incompetence of the shales in the overlying Pietra Maura Member and the intensity of the deformation which affects the formation as a whole, conspire to make description of a type succession impossible: that previously documented by Scandone (1967b, 1972) has since been afforested and is no longer visible (see Chapter 2). A number of different lithologies, however, can be distinguished.

The shales, like those of the Neritic Limestone - Shale Facies, have retained no primary depositional structures. They comprise quartzose and subordinate calcareous mudstones and siltstones with a clay mineral assemblage consisting of illite and some chlorite; smectites are not represented. Coarser grained terrigenous clastics are rarely found. Around the base of Tempa di Roccarossa, one of the largest neritic limestone blocks that lies 8 km north-east of Lagonegro, are scattered outcrops of poorly-exposed fine- to medium-grained micaceous sandstones. The beds, devoid of sedimentary structures, are about 10 cm thick and the rock comprises an annealed assemblage of quartz, minor plagioclase feldspar and muscovite with rare epidote, sphene and zircon. At one locality they occur at the top of a 10 m thick coarsening-upwards sequence of red and green mudstones, siltstones and sandstones. Similar rocks are exposed near the top of the formation in some parts of the basin and quartzarenites, with both symmetrically and asymmetrically rippled tops, have been recorded from the type succession by Scandone (1967b). He also describes calcite-cemented polygenic breccias and conglomerates, containing normally-graded pebble- to cobble-grade clasts of most of the other basinal lithologies, that fill erosional channels 70-80 cm deep and 2-3 m wide in the underlying shales.

At several localities close to the basal thrust and in the vicinity of large neritic limestone blocks, it is possible to find green chloritic claystones containing broken and often rotated fragments of parallel- and cross-laminated calcarenite, structureless, micaceous sandstone cobbles and rounded neritic limestone fragments. The long axes of the more elongate clasts are generally parallel or sub-parallel to the shaly parting although some are strongly oblique. The well-developed fissility of the shales may either abut against clasts or wrap around them.

ii) Interpretation

The paucity of depositional structures in the terrigenous clastics makes facies interpretation difficult. The polygenic breccias and conglomerates, however, highlight the importance of catastrophic intra-basinal reworking that may reflect synsedimentary tectonic activity and differential subsidence. The associated calciclastic facies described below, however, suggest that depths were relatively shallow. The mineralogy of these sediments implies derivation from either a high-grade metamorphic or granitic terrain, implying that the basin was founded on continental rather than oceanic basement: the possibility, however of an interposed clastic wedge masking an oceanic basement cannot be excluded (see Chapter 8).

The spatial association of the green scaly claystones with large, flat-lying neritic limestone blocks close to the basal thrust is considered generic. The epimetamorphic appearance of the shales, their intense fissility, the rotation of elongate calcarenite clasts and the rounding of the neritic limestone fragments suggest that these deposits have been strongly sheared: they are interpreted as the products of local deformation of terrigenous and calciclastic sediments at the base of the large limestone blocks during tectogenesis.

4.2.3. The Calcarenite-Calcisiltite Facies

i) Description

The shales of the Pietra Maura Member are commonly interbedded with calcarenites and calcisiltites, particularly in places close to the neritic limestone blocks. Occurring in beds varying in thickness from 1 to 10 cm, which often pinch out over short lateral distances, they comprise silt and fine sand grain sizes. Although poor exposure precludes the description of vertical or lateral sequences, it is clear that the fine-grained terrigenous and calciclastic facies are strongly heterolithic. Proportions of terrigenous material in the calcisiltites and calcarenites are generally low but are variable both between and within individual beds whose upper parts tend to be better sorted (Pl.4.5). These deposits are parallel- or, more commonly, cross-laminated and exhibit both symmetrical and asymmetrical ripple form-surfaces. Internal sedimentary structures of the well-sorted lithologies include form discordant surfaces, irregular and undulating lower set boundaries, bi-directional cross-lamination, chevron and bundled upbuilding, swollen lens-like sets and offshooting and draping foresets. Many of the poorly-sorted deposits, however, show graded bedding, planar-, cross- and convolute lamination (Pl.4.6), and rare flute casts and tool marks. Some beds, with well-sorted tops, may incorporate many of the latter features nearer the base.

ii) Interpretation

The sedimentary structures exhibited by the well-sorted lithologies are considered to be diagnostic of wave-generated structures (de Raaf et al., 1977); many beds, however, show several features typical of calci-turbidites (Middleton & Hampton, 1976) and these structures commonly characterise the bases of layers whose tops show evidence of wave action. One might conclude, therefore, that carbonate material was being transported by turbidity currents and subsequently reworked by wave

oscillations. However, turbidite-like sublittoral sand sheets interbedded with mudstones and showing post-depositional wave-influenced re-sedimentation have also been interpreted as the products of deposition from waning storm-generated currents (Johnson, 1978). Nevertheless, the lack of mixing of calcareous and terrigenous components in the calciclastic and shale facies militates against such an origin for these rocks. It is impossible, due to the fine grain sizes, to identify the provenance of the carbonate material but the surrounding carbonate banks are the most likely source.

The bathymetric significance of this facies is, on first inspection, ambiguous. The contention that turbidity currents were responsible for introducing carbonate material into the basin from the neighbouring carbonate build-ups should imply the existence of considerable relief which is at variance with the shallow depths implied by the wave-generated structures. If the turbidite flows were initiated by storm-provoked turbulence, however, they may have possessed sufficient initial maturity to begin graded deposition after quite short translations, both laterally and vertically. If the wave action is attributed to the later stages of storm activity, the apparent dichotomy can be resolved. Thus it can be concluded that these sediments were deposited at some depth above the ambient storm-wave base.

The observed association of this facies with the olistoliths only implies a common provenance and is probably not generic, particularly since no transitional facies between the two end members have been recognised.

4.2.4. The Calcarenite-Calcirudite Facies

i) Description

Within the studied area at La Cerchiara, a number of lenticular

outcrops of more coarsely grained calciclastic rock occur within the shales of the Pietra Maura Member. The most spectacular of these can be seen occupying a gap in the palaeokarstified upper tabular limestone unit of the La Cerchiara Member (Pl.4.7). The field relations are complicated by a series of low-angle reverse faults which have imbricated the units to some extent: it is unclear whether this disturbance is of tectonic origin or the result of post-depositional slumping.

The rocks of this outcrop comprise tabular cross-bedded calcarenites and calcirudites in sets 20-30 cm thick, the entire unit being no more than 10 m thick and 30 m in lateral extent (Pl.4.8 & 4.9). The margins are poorly exposed but the morphology of the packet is evocative of a channel cross-section. The angle of cross-stratification varies between 10 and 30 degrees and is picked out by the pronounced imbrication of the lithic clasts: the cross-bedding dip directions are unimodal. The sediments themselves are composed of medium sand to pebble-grade intraclasts derived largely from the neighbouring shallow-water carbonate banks, although some are of pelagic affinity (Pl.4.10): larger lithic fragments of calcarenite and calcisiltite reworked from within the basin itself are also common (Pl.4.11). The average grain size and proportion of lithic fragments in successive beds appears random and there is no evidence of overall coarsening or fining upwards, either within individual sets or throughout the sequence as a whole; cobble-size lithic clasts do, however, form a lag at the base of some sets. Successive beds are commonly separated by thin silt and fine sand drapes.

Higher in the formation, a number of other channellised sand-bodies are exposed that show many of the features of the outcrop described below. These, however, are trough cross-bedded, are generally finer-

grained and lack the coarse lithic clasts. The dimensions of the packets cannot be determined due to lack of exposure but are of the order of 5 m deep. Set thicknesses rarely exceed 15 cm, and cross-bedding dip directions are distinctly bimodal. These lenticular sand-bodies are surrounded by red shales and are not apparently related to any of the olistoliths or tabular limestone units.

ii) Interpretation

The large-scale sedimentary structures and the coarse grain size of the lower packet reflect the influence of strong traction currents which fashioned large bed-forms. These beds were most probably deposited by migrating sand waves, successive sets representing discrete depositional events. Occupying as they do a position between the isolated fragments of the upper tabular limestone unit, it is feasible to invoke constricted tidal currents as the influence under which the sand waves developed. The same currents were probably responsible for cutting the channel into the siltstones and mudstones which initially filled the gap between the limestone fragments. Although it is more difficult in the case of the other packets to identify a unique cause of tidal current constriction, it is likely that these too are the products of such currents, particularly in view of the bimodal palaeocurrent distributions.

The deposition of steeply-dipping planar cross-beds from tidal current-generated sand waves has been documented from a tidal inlet by Kumar & Sanders (1974) and bi-directional trough cross-bedded sands overlying similar large-scale planar sets are incorporated in the idealised tidal inlet sequence described by Hayes & Kana (1976). In such cases, channel depths rarely exceed 10 m: the importance of reversing tidal currents in effecting bed-load transport in modern submarine canyons, on the other hand, suggests that the necessary conditions for moving coarse-grained

sediments as sand waves can exist at much greater depths, provided that currents are constricted in some way (Knebel & Folger, 1976; Keller & Shepard, 1978).

4.2.5. The Olistoliths

4.2.5.1. General description

The ubiquitous, isolated blocks of neritic limestone that characterise the formation are distributed randomly throughout the formation outcrop amongst the shales and calciclastics overlying the neritic limestone-shale facies (Fig.4.1). Their dimensions vary from a few metres to as much as 2 km and, with the exception of the extensive, flat-lying oblate units close to the basal thrust, their shapes show no obvious pattern. The limestones are commonly recrystallised and veined such that bedding can only rarely be discerned, making recognition of their orientation with respect to the surrounding facies difficult; furthermore, the majority of primary stratigraphic contacts have been tectonised. Nonetheless, it is still possible in several cases to recognise strongly discordant primary relationships. Individual units are commonly surrounded by a wealth of smaller fragments that are clearly derived from the parent block; coarse, angular meggabreccias are conspicuously absent.

The limestone facies strongly resemble those of the neritic tabular limestones beneath. They generally comprise poorly- to well-sorted assemblages of partially to totally micritised allochems, peloids, grapestone particles and rare ooids. Oncolitic overgrowths are common and the grains, some of which have been leached and replaced by sparry calcite, are surrounded by a finely laminated micrite or cemented by sparry calcite. Marl-, sparry calcite- and radiaxial fibrous calcite-filled irregular cavities identical to those described above can also be found.

4.2.5.2. Representative localities

The principle features of the limestone blocks can be observed at four localities that are most conveniently documented and interpreted individually. These characteristics are not unique to single outcrops, however, and similar examples can be found throughout the basin.

a) Pietra Maura

i) Description

The imposing, composite limestone mass of Pietra Maura, which emerges from the shales of the formation 4.5 km north-north-west of Marsico Nuovo, is 250 m across and 75 m high (Pl.4.12). A breccia of sub-rounded, neritic limestone fragments in a marl matrix is preserved on the western flank of the block (Pl.4.13); way-up structures in the adjacent deposits reveal that this was the bottom surface prior to tectonic rotation. The breccia, which grades downwards into the underlying claystones, merges gradually into the block margin which is riddled with marl-filled cavities (Pl.4.14 & 4.15). These cavities are distributed along a closely spaced network of fissures that impart a pseudonodular aspect of the rock, and become increasingly rare away from the margin of the block.

ii) Interpretation

As in the case of the tabular limestone units of the Neritic Limestone-Shale Facies, the marl-filled cavities are attributed to palaeokarstification. In this case, dissolution has picked out and emphasised pre-existing cracks which are probably of extensional origin. The gradual transition from the pseudo-brecciated block surface to the breccia itself suggests that palaeokarstification was also responsible for the generation of the rounded limestone clasts. This surface may, therefore, represent the original upper surface of the block which is now inverted upon its own palaeokarst breccia.

b) Fontana d'Eboli

i) Description

A block of neritic limestone capped by a 2 m veneer of nodular pelagic limestone is exposed about 1 km along the mule track which leads from Fontana d'Eboli towards Rocca Rossa (Pl.4.16). The contact between the two lithologies is not easily distinguished, both being light grey in colour, and the nodular fabric is poorly developed near the boundary. These pelagic deposits are typical radiolarian biomicrites with abundant thin-shelled pelagic bivalve test fragments, probably of Halobia sp. or Daonella sp. (Pl.4.23); calcitised radiolaria too are abundant. The chert matrix is constituted by a green siliceous claystone containing both silicified and calcitised radiolarian ghosts and is pervaded by disseminated silica (Pl.4.24). Gradational nodule-matrix contacts are rare and the nodules are closely packed, with chert filling the interstices of the nodule-supported framework (Pl.4.17). Individual nodules are often separated only by thin stylolitic seams.

ii) Interpretation

Red nodular limestones of the Ammonitico Rosso and Hallstatt Facies are widely documented throughout the Mediterranean region where they commonly occur within condensed pelagic sequences overlying shallow-water carbonate units (Bernoulli & Jenkyns, 1974). The nodular fabric is considered to have developed by diagenetic redistribution of $CaCO_3$ close to the sediment-water interface and to be a reflection of slow rates of deposition (Lucas, 1955; Jenkyns, 1974). The absence, in this case, of faunal evidence of true stratigraphic condensation, the unique siliceous nature of the matrix and the lack of gradational nodule margins prevent direct comparison. This veneer does, however, preserve the record of transient residence of the block in a pelagic environment and the nodular fabric may imply slow rates of deposition.

c) Murge del Principe

i) Description

In a small valley immediately to the south of Murge del Principe, and 2 km south-west of outcrop (b), a block several tens of metres across is exposed which retains the vestiges of a now destroyed pelagic veneer; it is probably a fragment derived from the adjoining mass of Tempa di Roccarossa, which may have originally supported an extensive pelagic cover. The pelagic sediments are located in a network of fissures that permeate the block (Pl.4.18). Some of the fissures have been enhanced and have coalesced to form irregular cavities that are lined by a radiaxial fibrous calcite selvedge (Pl.4.19, 4.20, 4.21). These early cements and their substrates are commonly fractured, and a later isopachous sparry calcite cement has developed on the truncated surfaces prior to filling of the fissures by pink micrite (Pl.4.22). These micrites are barren but similar examples from other blocks elsewhere in the region abound with pelagic bivalves and radiolaria (Pl.4.25).

Some of the larger fissures contain polygenic breccias that comprise fragments of neritic limestone and a variety of pelagic limestones and mudstones in a red siliceous mudstone matrix (Pl.4.26). Individual clasts are themselves composed of an amalgam of smaller clasts and up to four successive breccia generations can be recognised (Pl.4.27); it is also possible to observe cross-cutting relationships between fissures formed at different times. Some of the red pelagic limestone clasts display a nodular fabric more typical of the 'Ammonitico Rosso' facies while others may preserve relics of the margin of the original fissure.

ii) Interpretation

These fissures developed in response to impinging extensional stresses during subsidence. The solution fabrics at the margins of the early cracks testify to a period of karstification prior to rapid

submergence, as evinced by the filling of the cavities by pelagic rather than neritic carbonate sediment. The pelagic components of the breccias were incorporated during successive fissuring phases and represent the relics of veneers that accreted in the interim: those deposits that had undergone early submarine lithification formed angular clasts whereas unlithified sediments provided the matrix. Feint laminations in the barren micrites may be flow textures, suggesting that the particles were injected through very narrow apertures excluding all but the finest material.

Neptunian dykes and sills are common in shallow-water limestone units underlying condensed pelagic sequences of the Tethyan Jurassic (Castellarin, 1965; Wendt, 1971; Bernoulli & Jenkyns, 1974), and Triassic (Schöll & Wendt, 1971). These features characterise units that are interpreted as fossilised submarine-highs and seamounts produced by block faulting and regional subsidence of a carbonate platform (Bernoulli & Jenkyns, 1974).

d) Tempa la Secchia

i) Description

The limestone block at Tempa la Secchia, 500 m north-east of the road to Moliterno 6.5 km north of Lagonegro, is capped by 50 cm of polygenic breccia overlain by 2 m of massive red siliceous mudstone (Pl.4.28). The breccias, identical to those found at Fontana d'Eboli, have a matrix of similar composition to the overlying mudstone (Pl.4.29). These siliceous deposits comprise radiolarian ghosts set in a dark red clay matrix along with scattered euhedral rhombs of dolomite; they are otherwise lime free. Always representing the final phase of pelagic deposition on the blocks, these mudstones are common throughout the basin.

ii) Interpretation

The breccias at this outcrop, generated during fissuring, have been preserved on the surface of the block, rather than in neptunian dykes, possibly because the unit remained horizontal during faulting. The mudstones are not visibly brecciated although diffuse micro-faults discernible in thin-section suggest some deformation, and mudstone clasts are present in the breccia beneath. At the time of their deposition, any calcareous material was either being leached out or was not reaching the sediment-water interface in the first place; small amounts may however have been dissolved and re-precipitated as dolomite within the sediment.

4.2.5.3. Discussion

The neritic limestone facies of the blocks indicate that they once formed part of an extensive carbonate edifice, probably similar to the build-ups of the Dolomites (Bosellini & Rossi, 1974), rather than localised patch reefs. Their allochthonous origin is demonstrated by their discordant relationship to the surrounding sediments and they are interpreted as olistoliths that owe their origin to the destruction of a carbonate platform. There is evidence that these fragments of the platform were subjected to periods both of emergence and residence in a pelagic environment. The fact that the nodular veneers parallel the bedding of the host limestone, rather than that of the surrounding shales, confirms the fact that pelagic sedimentation occurred prior to the final emplacement of the faulted platform fragments as olistoliths onto the basin floor. Minor block faulting and differential subsidence are held responsible for creating submarine highs that accumulated the pelagic sediments. These deposits were rarely preserved in situ, however, and extensional stresses led to their incorporation as breccias into neptunian dykes. The pink micrites found filling other fissures may have been injected

by some process of suction during faulting.

The source of fine-grained pelagic carbonate is not clear; although some Triassic calcareous nannoplankton have been described (Fischer, Honjo & Garrison, 1967; Zankl, 1971; di Nocera & Scandone, 1977) they are not generally considered to have existed in rock-forming quantities and none have been observed in these rocks (Black et al., 1967). Lime mud washed from the neighbouring carbonate banks may, therefore, have been the major contributor (cf. Busby, 1962; Pilkey & Rucker, 1966) (see Chapter 5). Furthermore, the notable lack of traces of ammonites, whose unstable aragonitic tests probably provided the majority of the $CaCO_3$ necessary for the development of limestone nodules in the 'Ammonitico Rosso' facies (Jenkyns, 1974), presents a further problem regarding the genesis of the nodular fabrics. The abundant pelagic bivalve tests probably possessed an aragonitic component (cf. Jefferies & Minton, 1965), however, and a high proportion of any material derived from the carbonate banks would have been composed either of aragonite or high magnesian calcite.

The commonly observed transition from calcareous to siliceous pelagic sedimentation recorded in many olistoliths is also problematic. Lime-free radiolarian mudstones and cherts in mountain belts are now widely recognised as lithified equivalents of the siliceous oozes accumulating below the CCD in modern deep ocean basins. Garrison & Fischer (1969) postulated that the CCD prior to the late Jurassic was at a much shallower depth than today; at that time, the dramatic development of calcareous nannoplankton and foraminifera led to a net transfer of $CaCO_3$ from shallow seas into the deep oceans and a concomitant depression of the CCD. Although there are no extant data relating to dissolution depths in the Middle Trias, the onset of

siliceous sediment accumulation may reflect the subsidence of the faulted carbonate platform below an anomalously shallow CCD, as depicted in Fig. 4.4(a). This interpretation, however, is at variance with the occurrence of marls and apparently uncorroded calciclastic beds amongst the basin floor deposits. It is conceivable, however, that these marls are only found at the base of the formation, and that they give way to lime-free terrigenous clastics deposited at greater depth higher in the succession; one calcisiltite bed collected from the top of the type succession of Scandone (1967b), close to the summit of Monte Facito, has an irregular structure-discordant upper surface that may be due to corrosion.

A second model, illustrated in Fig.4.4(b), is based upon the assumption that the 'pelagic' carbonate is of shallow-water derivation. During the early stages of platform disintegration downfaulted blocks accumulated lime mud from the adjoining undisturbed carbonate banks. As continued block faulting led to an increase in the areal extent of the basin and the outward migration of the platform margins, the oldest blocks, now at some distance from a potential source of carbonate mud would have begun to accumulate truly pelagic sediments by default; in the absence of calcareous nannoplankton this would have been almost entirely siliceous. This model obviates the need to invoke an abnormally shallow CCD for which there is scant evidence. The lack of any independent features which might suggest submarine volcanism fails to lend credence to the alternative possibility that hydrothermal activity had lowered the seawater pH sufficiently to encourage whole-scale leaching of calcareous material.

Irrespective of their mode of genesis, it was manifestly possible for siliceous pelagic deposits to accumulate on submarine highs at this time. Such environments provide a viable alternative to mid-ocean ridges as the sources of redeposited Triassic cherts described from

other Tethyan continental margins such as Othrys, Greece (Nisbet & Price, 1974; Smith et al., 1975).

4.2.6. The Basinal Pelagic Facies

i) Description

The Monte Facito Formation is succeeded by the Sirino Formation, a 250 m succession of grey, thinly laminated micritic cherty limestones (see Chapter 5). The transition between the two formations is well exposed in a small gulley immediately to the west of Tempa di Roccarossa, and is marked by a characteristic nodular limestone and mudstone facies (Pl.4.30, 4.32). This, the Daonella lomelli level of de Capoa Bonardi (1970), is found throughout the basin and consists of about 4 m of red and green marls and mudstones with several horizons of grey or pink nodular micrite (Pl.4.33). The limestone is a typical radiolarian biomicrite, similar to that forming the nodules found capping some of the olistoliths, but is commonly replaced by chert which, unlike the micrite, has retained a red colour; in thin section, chalcedonic ghosts of pelagic bivalves, and their umbrella cements, and radiolaria confirm the replacement nature of the chert. The red marly matrix is commonly pervaded by euhedral rhombs of dolomite while adjacent nodules are unaffected (Pl.4.31). The undolomitised matrix types include red and green claystones devoid of lime as well as the predominant marly mudstones. In plan view the nodules are often elongate and they impart a characteristic mullion-like appearance to the rock.

ii) Interpretation

Both stratigraphically condensed and expanded pelagic facies can be found in the Alpine-Mediterranean Triassic: slow rates of deposition in starved basins surrounded by carbonate platforms are invoked to explain the development of the red and often nodular limestone sequences, whereas the more expanded grey cherty biomicrites are thought to reflect higher

sedimentary rates (Bernoulli & Jenkyns, 1974). It has been suggested that these pelagic limestones, like their younger counterparts, were primarily composed of calcareous nannoplankton remains that have now disappeared as a result of intensive recrystallisation (di Nocera & Scandone, 1977); on the other hand, it has been shown that the analogous modern deep basins of the Bahamas Platform derive much of their lime mud from the adjacent shallow-water carbonate banks (Busby, 1962; Pilkey & Rucker, 1966; Rucker, 1968)(see Chapter 5).

The appearance of a nodular pelagic facies at the top of the formation is, therefore, considered to mark a temporary period of sediment starvation. The abundance of Radiolaria in the sediments suggests that the basin allowed relatively unimpeded access to ocean waters; had coccoliths been present in abundance, such conditions would hardly have been conducive to sediment starvation. Once the tectonic instability and concomitant intra-basinal reworking of the Middle Trias had come to an end, however, rates of sediment deposition in the absence of a nannoplanktonic carbonate contribution may have dropped to a level which would allow nodule development; this situation would have continued until such time as the platforms began contributing lime mud to the basin. If, as has been suggested above, the CCD was abnormally shallow at this time, any large-scale transfer of $CaCO_3$ from the platforms to the basins during the more stable Late Trias would have led to depression of the CCD and allowed the deposition first of condensed limestone facies followed later by the much thicker grey cherty biomicrites of the Sirino Formation.

4.3. Igneous Rocks

The presence of igneous rocks of Middle Triassic age in the Lagonegro Basin has been noted by a number of authors (Ietto & Cocco, 1965; Dietrich & Scandone, 1972). The 'diabases' described by Ietto & Cocco near Tito and

subsequently re-examined by Dietrich & Scandone are contained in isolated exposures of highly altered agglomerate of indeterminable age. The author has found no other igneous rock outcrops within the confines of the zone, but pillow lavas have been reported from the Monte Facito Formation north-east of La Cherchiara (Scandone, pers. comm).

The paucity of Middle Triassic volcanism in the Lagonegro Basin is in marked contrast to many other co-eval basins of the Alpine-Mediterranean region where it is widely held to be a reflection of early rifting (Obradovic & Stojanovic, 1972; Bernoulli & Jenkyns, 1974; Scandone, 1975; Bosellini et al., 1977; Laubscher & Bernoulli, 1977; Bechstädt et al., 1978; Castellerin, Rossi et al., 1978).

4.4. Model for the Monte Facito Formation

By collating the data obtained from a large number of outcrops, it is possible to build up a picture of the primary depositional relationships between the diverse elements of the formation prior to tectonic modification (Fig. 4.5). Although many of the neritic limestone blocks are clearly allocthonous, the largest units, situated close above the basal thrust, may have been stripped from their basement during Miocene thrusting; such blocks have been excluded from this diagram. For simplification, a two-dimensional ideal section through one fault block is illustrated. The position which the igneous rocks occupy within this time-space framework is not known and they have been omitted.

Any attempt to construct a model for the development of the formation must take into account the following:

i) The autochthonous limestone units of the Neritic Limestone-Shale facies

ii) The reworking of the terrigenous clastics

iii) The three-dimensional arrangement of the olistoliths

iv) The random outcrop distribution of the olistoliths

v) The local presence of pelagic veneers

vi) The neptunian dykes, fissures and associated polygenic breccias

vii) The palaeokarstification of the upper tabular limestone units of the Neritic Limestone-Shale Facies and many of the olistoliths

viii) The apparent co-existence of tidal and turbidity current influences

ix) The condensed pelagic interval at the top of the formation

x) The alkali basalt pillow lavas

Since the olistoliths are not concentrated near the margins of the basin, it is clear that they have not simply fallen from the bordering carbonate platforms, as is apparently the case in the St. Cassian Formation in the Dolomites (Fürsich & Wendt, 1977). Furthermore, the fact that the blocks do not occur locally at any one particular stratigraphic level implies that olistolith formation was spasmodic during the genesis of the basin. The allochthonous nature of the blocks is not open to question, however, except for those larger units which may be relict seamounts. One must therefore envisage some mechanism for the development of local sources throughout the area for which block faulting must be the most potent candidate. The various indications of extensional tectonics outlined above lend credence to this hypothesis.

The Monte Facito Formation is therefore interpreted as the result of the destruction of an embryonic carbonate platform under the influence of an extensional tectonic regime during the Middle Trias. The progressive development of the basin is illustrated in a series of block diagrams which depict the various identifiable stages (Fig.4.6). The width of the basin as shown is strictly schematic and by the end of the

Ladinian was probably of the order of 50km. The sequence of events affecting the basin margins may equally well have pertained around the numerous seamounts which doubtless survived until the latter phases of subsidence; hence the random distribution of the olistoliths over the basin floor.

During the Early Anisian, the peneplained Hercynian basement began to subside under the influence of crustal thinning and, once submerged, accumulated a cover of shallow-marine carbonate sediments. It is not known whether these first carbonate sediments were deposited directly on the metamorphic basement or whether, as seems more likely, a clastic wedge was interposed between the two. As subsidence progressed, early extensional stresses led to minor block faulting; concomitant reworking of the basement or its clastic cover provided the terrigenous clastic sediments that blanketed the downthrown surfaces of the platform. Once relief was thus eliminated, shallow sub-tidal carbonate sedimentation was renewed. This process was repeated several times and was responsible for the production of the tabular limestone units of the Neritic Limestone-Shale Facies.

Later in the Anisian, parts of the young platform became emergent for a short period leading to karstification and dissection of the upper tabular limestone unit. After re-submergence, the gaps between these fragments became the sites of deposition of calcarenites and calcirudites from migrating sand waves generated by constricted tidal currents. Further faulting led to the establishment of considerable vertical relief, and many of the karstified remnants of the tabular limestone units slipped and were redeposited as the first olistoliths. Turbidity currents were responsible for transporting shallow-water carbonate material from the banks into the basin, where they were re-worked by the ambient tidal currents and wave action. Re-mobilisation of

terrigenous clastics continued, leading to the final suffocation of extensive areas of the platform: these areas acted as a template upon which the ultimate basin morphology was modelled. Extrusive volcanism may have accompanied these early phases of faulting.

Even into the Early Ladinian, many of the remnants of the young platform managed to continue acting as hosts to shallow-water carbonates, and sedimentation in these areas was able to keep pace with the rapid rate of regional subsidence. The basin, however, became progressively deeper by default and further minor block faulting led to continued foundering of the platform margins. The marginal blocks and seamounts thus formed, once divorced from the neritic realm, became the site of pelagic sedimentation. In the majority of cases, pelagic carbonate deposition was succeeded by the accumulation of red siliceous mudstones, either as the blocks subsided below the CCD or else as a result of foundering of the shallow-marine carbonate source areas. Due to the incompetent nature of the terrigenous clastic horizons at their bases, however, these blocks developed sediment-filled fissures, and fragments began to break off and fall onto the basin floor: successive faulting episodes thus led to the destruction of many of the pelagic veneers which are commonly only preserved in breccias that fill fissures in the olistoliths. The limestone blocks were finally emplaced along with their associated karst breccias amongst the terrigenous clastics covering the basin floor.

These catastrophic events were succeeded in the Late Ladinian by a period of tectonic quiescence, and re-working of the terrigenous clastic sediments in the now well-established basin was reduced dramatically. The surrounding carbonate platforms, once stabilised, began contributing lime mud to the basin, and the CCD may have migrated down to the level of the basin floor. Slow overall rates of sedimentation, coupled with a

relative increase in the amount of hemipelagic carbonate deposition, fostered the development of a nodular limestone facies within the marls which mark the final phase of the formation. Accumulation of lime mud at the foot of the reef scarps, and its redeposition out onto the basin floor, characterised the sedimentation of the overlying Sirino Formation of the Late Trias (see Chapter 5).

Outcrop of the formation does not allow construction of a detailed palaeogeographical map of the basin at the end of the Middle Triassic. Since only relative minor block faulting may have been required to individualise the basinal and shallow-water areas, however, it is unlikely that the basin was of regular shape; there is no reason for believing that it was linear, for example, and an almost amorphous network of troughs dissecting the platforms can be envisaged, such as characterises the Bahamas region today (cf. Scandone, 1975). Nevertheless, the abundance of Radiolaria in the pelagic deposits of the formation, and the absence of any indications of anoxia during sedimentation, suggest that the basin must have had an outlet into the Palaeotethys Ocean to the east (Laubscher & Bernoulli, 1977).

4.5. Discussion

Evidence of Triassic, syn-sedimentary, extensional tectonics is now widely recognised (Dewey et al., 1973; Bosellini & Hsu, 1973; Bernoulli & Jenkyns, 1974; Scandone, 1975; Bosellini et al., 1977; Laubscher & Bernoulli, 1977): in particular, Bëchstadt et al (1978) have stressed the importance in the Alps of a distinct phase in the Middle Triassic, separated from the well-documented Liassic rifting events by a period of tectonic quiescence. The 'Cipit' blocks of the Middle and Upper Triassic St. Cassian Formation of the Southern Alps, up to several hundred cubic feet in volume, have been variously

interpreted (for review, see Fürsich & Wendt, 1977, pp.302-305). The facies types represented in the blocks are typical of platform carbonate environments, and they too probably represent re-deposited relics of such a platform; their close spatial relationship to the platform margin is adequate testimony of their provenance (Bosellini et al., 1977; Fursich & Wendt, 1977). In common with the olistoliths of the Monte Facito Formation, several of these blocks bear evidence of palaeokarstification in the form of spar-filled solution cavities (Bosellini et al., 1977). Other occurrences of displaced shallow-water limestone blocks within the Tethyan Triassic are documented from Rhodes (Orombelli & Pozzi, 1967), Sicily (Mascle, 1967) and the Oman (Glennie et al., 1973), and Bassoulet et al. (1978) describe a Late Permian block which incorporates pelagic components of Scythian age from the Himalayas. The widespread development of olistoliths during this period was largely a consequence of the limited thickness of the young platform and the incompetence of the underlying clastic deposits: by the time block-faulting was resumed in the Lias, the thickness and rigidity of the platforms made the process of olistolith generation much less viable, although olistoliths of carbonate platform material, associated with rudites and meggabreccias, have been described from the Kokkinovrakhos Formation of central southern Greece (Johns, 1978).

It is difficult to find modern analogues for the Monte Facito Formation, but the envisaged process by which the olistoliths were emplaced is similar to that which causes periodic subsidence of the Chalk cliffs of southern England. An illustration of the Axmouth landslip of 1831 by J. Ruskin depicts fissured blocks of chalk that have broken off from the cliff and subsided as a result of the incompetence of the underlying Greensand and Gault Clay (Pl.4.34). Although in a

submarine setting, the margins of the embryonic carbonate platforms of southern Italy must have suffered a similar fate, the foundered blocks accumulating pelagic sediment prior to finally collapsing onto the basin floor.

CHAPTER FIVE

TECTONIC STABILITY AND HEMI-PELAGIC CARBONATE DEPOSITION OF THE LATE TRIASSIC: THE SIRINO FORMATION

5.1 Introduction

The lithofacies of the Sirino Formation chiefly comprise carbonate sediments whose sedimentology is treated in two parts: those pertaining to the Gianni Griecu Member of Unit I are discussed in Section 5.2, and the remainder in Section 5.3. The limestone/marly limestone/shale cycles of the Gianni Griecu Member point to the influence of storm reworking during the earliest phases of carbonate sedimentation. Deposition of lime mudstones at greater depths higher in the succession bears witness to the progressive deepening of the basin during the Late Triassic, and a gradual transition from calcareous to siliceous sedimentation at the top of the formation records the subsidence of the basin floor beneath the CCD. Intrabasinal redeposition is evinced throughout the period by slumps and calciclastic facies, and the resumption of block faulting in the Late Norian is marked by the extensive development of both intra- and extra-formational calcirudites. An overview of Late Triassic carbonate sedimentation in Section 5.4 is followed by a depositional model for the Sirino Formation in Section 5.5. An appraisal of bathymetry and possible modern analogues is presented at the end of the chapter.

5.2 Sedimentology and diagenesis of the Gianni Griecu Member

The Gianni Griecu Member comprises an association of three lithofacies that are rhythmically interbedded throughout: the Limestone/Marly Limestone, Calcarenite/Calcisiltite and Pencil

Shale Facies. The characteristic features of these facies are summarised in Fig.5.1. All exposures of the member are complicated by the intense structural deformation inherent in a major décollement horizon, and shearing along bedding planes, coupled with diagenetic modification, has obscured the primary depositional relationships of the two facies.

5.2.1. Facies descriptions

5.2.1.1 The Limestone/Marly Limestone Facies

Beds composed of dark-grey, fine-grained limestone underlain by a lighter-coloured, yellowish marly-limestone characterise the facies (Pl.5.1). These limestone/marly limestone couplets are variable in thickness from 2 to 12 cm but may show marked lateral variations. The ratio of marly limestone to limestone is also highly variable, but where both are present, as is generally the case, the marly limestone is invariably overlain by the limestone.

The limestone intervals are generally structureless, and comprise lime mudstones with a variable clay content, those occurring in couplets tending to contain a lower argillaceous component than beds lacking a marl division. Polished faces reveal abundant evidence of bioturbation, but planar lamination is commonly preserved at the tops of beds (Fig.5.2). The laminated intervals are constituted by thin-shelled bivalve packstones and grainstones, and beds may thus show a crude overall inverse grading. However, the laminated and bioturbated components are separated by a sharp undulose contact. Small ammonites, partially filled with lime mud, may be found whose geopetal structures are either right-way-up, at an angle to bedding or completely inverted. The abundant thin-shelled bivalves that are also present, probably Halobia sp. and Daonella sp., are almost entirely oriented with their shells convex upwards. The tops of the limestones are usually in sharp contact with pencil shales or other limestone-marl couplets,

although some gradational contacts through a finely-laminated calcareous shale are also common.

In contrast, the marly-limestone intervals are completely devoid of body fossils, and show no sedimentary structures with the exception of rare grading at the bases of some couplets. The contact with the limestone is generally sharp and parallel or sub-parallel to bedding; in some cases it may be undulose over lateral distances of several metres. The marls are pervaded by a complex network of trace fossils that introduce grey calcilutite from above. These burrows appear to be randomly oriented except at the bases of couplets where they lie horizontally. They are circular in section and their widths vary between 0.25mm and 1cm, although there are two modal ranges of 0.25-1mm and 2-3mm corresponding to Chondrites and Planolites respectively. The undersides of some couplets show the imprints of these trace fossils clearly and the characteristic, dendritic pattern of the ramifying Chondrites can be seen in the plane of the bed (Pl.5.3); the Planolites burrows either terminate at the base or may extend laterally for several centimetres. The depth of penetration of the burrows into the marly limestone is variable and in apparent inverse proportion to the thickness of the overlying limestone. Rare trace fossils up to 1cm wide, probably Thalassinoides, may also be present. The Chondrites-Planolites-Thalassinoides trace fossil assemblage is common to many limestone-marl rhythmic sequences of the European Jurassic and Lower Cretaceous (Kennedy, 1975).

In thin-section, the limestones comprise radiolarian/pelagic-bivalve biomicrites and pelagic-bivalve biosparites (Pl.5.5). The micrite component is made up of an intergrown mosaic of calcite grains about 5μ in diameter which constitute about 98% of the rock; the

insoluble residue consists predominantly of quartz with minor clay minerals. Radiolaria, which are calcitised and filled either by micrite or a void-filling sparry calcite, are distributed at random and exhibit geopetal structures in a variety of orientations. The pelagic-bivalve tests are either preserved intact, in which case the cavity beneath is filled by sparry calcite umbrella cement, or else, as in the case of the more-obviously laminated limestones, they are completely flattened parallel to bedding (Pl.5.6). In the latter case, fragmented shells may be coalesced into spherular aggregates a few millimetres in diameter. The proportion of shells in these laminated biomicrites is greater than in the more massive radiolarian biomicrites, and they may be concentrated in certain laminae to the extent of forming lumachelles.

The marly limestones, in contrast, are devoid both of Radiolaria and thin-shelled bivalves, and comprise an admixture of microspar and clay minerals; the $CaCO_3$ content averages about 93%. The calcite component is notably more coarse-grained than in the biomicrites, averaging 10-20μ. The burrows contain biomicrite similar to that of the limestones, but lack the coarser bioclastic components. The contact with the limestone is either transitional over a scale of millimetres or is marked by a single or anastomosing group of stylolites; clay minerals are concentrated along these seams (Pl.5.6).

A notable diagenetic feature of the couplets is that the limestone portion is commonly pervaded by a dense, parallel network of thin, calcite veins lying normal to bedding (Pl.5.6). These veins, between a few microns and 1mm across, die out near the contact with the marl and only rarely continue below it. Several different generations can be distinguished that either abut against or offset the horizontal

stylolites.

X-ray diffration analysis of the clay fraction from one of the marly limestones reveales the presence of a strong, well-defined peak at 7Å, and two broader, less pronounced peaks at about 10Å and 14Å (Fig.5.2). Treatment with ethylene glycol and heating to 500°C produced negligible change except that the latter led to the complete supression of the 7Å peak. These peaks indicate the presence of kaolinite, illite and chlorite respectively. The illite peak shows an asymmetry towards broader d-spacings, implying some mixed layering.

5.2.1.2. The Calcisiltite/Calcarenite Facies

Towards the top of the member, beds of very fine sand- and silt-grade limestone, up to 20cm thick and showing low-angle cross-lamination throughout, become increasingly common. They lack a basal marly limestone interval and usually directly overlie a limestone/marly limestone couplet without any intervening shale horizon (Pl.5.4). The origin of the grains, which now form a calcite mosaic, cannot be determined, but these beds are well sorted and contain no bioclasts; the lamination is produced by concentrations of terrigenous grains.

5.2.1.3. The Pencil Shale Facies

The pencil shales interbedded with the limestone/marly limestone couplets are predominantly grey in colour, are up to 15cm thick and comprise highly fissile mudstones and claystones. Cleavage is generally parallel to bedding except near the axes of folds where intersecting bedding partings and cleavage produce the characteristic 'pencil' fracture. At the base, the shales may pass into finely-laminated,

dark-grey, calcareous shales up to 2cm thick, overlying a laminated limestone forming the top of a limestone/marly limestone couplet. These calcareous shales contain large numbers of pelagic bivalve shells whose imprints are commonly visible on upper bedding surfaces.

Thin-sections of the calcareous shales reveal a close-packed mass of filamentous shells set in a dark micritic matrix that comprises micas, clay minerals and flecks of pyrite; scattered Radiolaria may also be present.

X-ray diffraction of shale samples shows dominant quartz with minor clay minerals, and in one case, calcite. Clay-mineral analysis revealed peaks at about 7Å, 10Å and 14Å typical of kaolinite, illite and chlorite respectively (Fig.5.3). One sample, however, showed a broad peak at 11Å rather than 10Å which shifted to 9.5Å with ethylene glycol and back to 10.6A after heating to 500°C. This may represent a mixed layer species composed predominantly of mica, but has not been more precisely identified (Warshaw & Roy, 1961). The clay mineral assemblage is similar to that found in the marly-limestones; in both cases, smectites are entirely absent.

5.2.2. Interpretation

The limestone-marly limestone couplets are interpreted as the products of late diagenetic modification of a primary, or very early diagenetic, bipartite fabric. The piping of limestone into marly limestone by trace fossils provides compelling evidence for an early differentiation of the two components. The limited penetration of these burrows suggests that only limited time was available for biological activity before deposition of the overlying couplet. Where parallel-lamination is preserved at the tops of couplets, some sedimentation must have occurred after bioturbation. It is clear, therefore,

that each couplet does not represent a single depositional event, and that the limestone and marly limestone components must have been differentiated during or before the final sedimentation of the limestone intervals.

Before dicussing possible modes of deposition of the couplets, it is necessary to examine the extent to which diagenesis may have modified their primary features. The possibility of dissolution, migration and reprecipitation of $CaCO_3$ close to the sediment-water interface, leading to enhanced segregation of lime-rich and lime-poor intervals, has been suggested for similar rhythms in the Blue Lias of Dorset by Hallam (1964). A similar process has also been envisaged by Jenkyns (1974) to account for the origin of red nodular limestones of the Tethyan Jurassic. Such a mechanism could have been responsible for the removal of body fossils from the marly limestone intervals, post-diagenetic compaction having subsequently eliminated the resultant voids. Furthermore, the preservation of unbroken, arcuate pelagic-bivalve shells in the overlying limestone intervals attests to some early lithification. The conditions under which such a redistribution took place may have been similar to those envisaged by Jenkyns (1974) for the development of the nodular 'Ammonitico Rosso'. The predominantly aragonitic Daonella sp. and Halobia sp. shells which survived into the burial stage were possibly dissolved below the sediment-water interface to yield pore-waters supersaturated with respect to calcite. These solutions may have supplied the $CaCO_3$ necessary for lithification of the limestone. A necessary condition for such a process of early lithification would have been slow overall rates of sedimentation (Jenkyns, 1974). The calcium and bicarbonate ions derived from the marly limestone are unlikely to have migrated downwards, particularly since the underlying shales would have prevented

the passage of pore fluids. Furthermore, in the majority of cases, it is the base of the limestones which show most evidence of early lithification in the form of unbroken filamentous shells with umbrella cements. Transport paths were most probably upwards, therefore, pore-fluids percolating only short distances into the limestone before calcite reprecipitated upon stable nuclei, such as the fibrous calcite component of pelagic bivalve shells. This would have effectively reduced the concentration of ions in solution and prevented further cementation higher in the bed. However, since compaction can be demonstrated for a large proportion of the limestones, and in view of the small differences in argillaceous content between the limestones and marly limestones, redistribution of $CaCO_3$ is not considered to have been of major significance in differentiating the two components.

The fate of the SiO_2 is more problematic. The complete absence of replacement cherts in the member implies that silica derived from the dissolution of opaline radiolarian tets, both in the limestones, where calcitised moulds are preserved, and also possibly in the marls, has been lost into the overlying water column. This, too, suggests that rates of sedimentation were low. The presence of sharp concentration gradients of SiO_2 across the sediment/water interface in recent oceans implies a considerable flux of silica, and more than 90% of biogenous opal deposited may be lost to the water column as a result (Hurd, 1973; Calvert, 1974; Wollast, 1974); furthermore, Kastner et al. (1977) have shown that the transformation from opal-A to opal-CT is strongly retarded in the presence of clay minerals, which act as a sink for the magnesium and hydroxyl ions that could provide potential nuclei for precipitation. Thus the possibility of precipitation of a chert precursor would have been more remote in these argillaceous

lithologies than in the purer calcilutites higher in the formation.

A more plausible and primary mechanism whereby the two components may have been segregated involves the winnowing out of the argillaceous component from the limestone. The most compelling evidence for reworking is provided by the presence of low-angle trough cross-laminations in the calcarenite/calcisiltite beds and scour surfaces within some of the limestone intervals. The absence of sedimentary structures from the marly limestones suggests that resedimentation may have been of prime importance in differentiating the two components of the couplets. Furthermore, multiple phases of reworking are indicated by the scour surfaces within the limestone intervals themselves, and a protracted process of winnowing and resedimentation must be invoked.

The scale of cross-lamination in the calcisiltite/calcarenite beds suggests that relatively strong currents were sporadically active. Although the grainsize of these calciclastics does not allow recognition of their source, it is most likely that they originated in the adjacent shallow-water carbonate banks and may have been introduced by turbidity currents. Their infrequent distribution through the member suggests that they may represent the products of exceptionally severe storms, and storm-induced currents may have been responsible for reworking the beds after deposition to produce the cross-lamination. Turbidity currents with velocities of more than 1 m/sec that are thought to have been generated by storms are documented from modern submarine canyons, and wave-generated oscillation ripples in coarse sands have been observed at depths of up to 200 m on the continental shelf of western North America, where current velocities of 30 cm/sec are recorded (Reimnitz, 1971; Komar et al., 1972; Shepard & Marshall, 1973;

Yorath et al., 1979). In an interplatform basin such as this, however, it is unlikely that storm swells or surges could have effected reworking at anything but relatively modest depths.

5.2.3. Depositional model for the facies of the Gianni Griecu Member

The sequence of events envisaged for the development of alternations of limestone/marly limestone couplets and shales is summarised in Fig.5.4. It commences with deposition of an undifferentiated marly limestone that contains scattered pelagic bivalves and Radiolaria. These may represent the normal background sedimentation of the basin, involving deposition of pelagic or hemipelagic lime muds and terrigenous clays from the water column. The clay-mineral assemblage of the marly limestones gives no indication of a possible source for the clays, but it is compatible with derivation by weathering of continental basement; the argillaceous material may have been introduced into the basin by wind transport. The fact, however, that some of the marls show grading in the lower few centimetres, and are generally more coarse-grained than the limestones, suggests that they may have been deposited in a more dramatic manner. The absence of any other sedimentary structures precludes any concrete interpretation, but the possibility that the marly limestones represent the deposits of very low density flows of some kind, cannot be ruled out. The cross-laminated calcisiltite/calcarenite beds may also have been introduced into the basin by mass-flow processes since there is no obvious indigenous source of fine carbonate sand. The fact that these beds commonly overlie limestone/marly limestone couplets suggests that any intervening terrigenous muds were eroded during their deposition.

Deposition of the marly limestone was followed by a period of erosion and mobilisation of sediment. This may have been effected by storm-generated currents in an analagous manner to the process suggested for the development of sub-littoral sheet sandstones (Johnson, 1978). Such currents may occasionally have been of sufficient power to form ripples that gave rise to the low-angle trough cross-laminations of the cacisiltite/calcarenite beds. The planar-laminated horizons of the limestone intervals, on the other hand, reflect lower flow regimes, and the pencil shales, comprising the argillaceous component of the original bed or beds reworked during the storm, were probably deposited from suspension. Bioturbation destroyed the laminations of some limestone beds and in some cases extended down to the base of the couplet. Meanwhile, dissolution of unstable components in the marly limestone induced early lithification at the base of the limestones, while dissolved silica was lost to the water column. The presence of scours within some limestones, above which primary laminations are preserved, suggests that significant post-storm reworking took place in the case of some cycles, possibly by later storms of lower intensity; repeated winnowing could thus have produced the observed crude coarsening-upwards sequence which is in contrast to the normal grading that one might expect from a single resedimentation event. The preservation of lamination at the tops of these beds and the high organic contents of the calcareous shales attest to temporarily anoxic bottom conditions after passage of the storm. Renewed marly limestone deposition then began another cycle. The sequence of events described pertains to a typical marly limestone/limestone/shale cycle, but any of the stages may be missed out and the bed 'frozen' before it had completely developed, probably as a result of a high frequency of storms.

5.2.4. Discussion

Alternations of limestone/marl rhythms and shales constitute a facies association common to many basinal sequences of the European Mesozoic. Typical examples include the Upper Muschelkalk of south-west Germany, the Kössen beds of the Alpine Rhaetic, the Blue Lias of Dorset, the Plenus Marls of the North European Chalk and the Upper Liassic of the High Atlas Mountains, Morocco (Hallam, 1960, 1964; Jeffries, 1963, Fabricius, 1966; Kennedy, 1967; Sellwood, 1970; Ager, 1974; Evans & Kendall, 1977; Aigner et al., 1979). Several types of small-scale cyclic sedimentation have also been documented at a variety of stratigraphic levels in Deep Sea Drilling Project cores from the Atlantic Ocean (Dean et al., 1978). Although some authors consider that differentiation of the cycles is a partially diagenetic feature, many favour a primary origin and have stressed the importance of sediment winnowing in concentrating bioclasts in the limestone intervals. The origin of these deposits, often referred to as 'tempestites', is commonly attributed to storm action. In contrast, pulsing input of skeletal or terrigenous material, and cyclic variations in $CaCO_3$ dissolution due to fluctuations of the CCD have been suggested by Dean et al. (1978) to explain the deep-sea occurrencies. It is evident that superficially similar limestone-marl couplets can be developed under a wide range of depositional and diagenetic conditions.

5.3 Sedimentology of the remaining members of the Sirino Formation

The bulk of the Sirino Formation is constituted by three principle limestone facies, termed the Calcilutite, Calcarenite and Calcirudite Facies; a subsidiary Fine-grained Terrigenous Clastic Facies is also developed chiefly towards the top of the succession in both units. The sedimentology and early diagenesis of these rocks is described

and interpreted in the foregoing sections, but their ultrastructure and late diagenesis is discussed along with geochemical data Chapter 8.

5.3.1. The sedimentary facies

5.3.1.1 Calcilutite Facies

i) Description

Beds of massive, lead-grey or greeny grey calcilutite, up to a maximum of 2.5m in thickness but generally in the 5-20cm range, characterise the facies (Pl.5.7, 5.8 and 5.9). Bedding is regular and laterally persistent over hundreds of metres, although pinching and swelling can be observed in some beds, particularly in the vicinity of chert nodules. Beds are separated along undulose partings that may or may not contain paper-thin shales,; the shales are most common near the top of the formation (Section 5.3.1.2). In general, vertical bed-thickness variations show no obvious pattern, although in the Rupe del Corvo, Armizzone, Torrente Bitonto and Monte Lama sections, average thicknesses are greatest in the middle of the formation. In the Lagonegro Section, thick, massive beds alternate with packets of thinner beds, separated by shaley partings, in the Carboncello Member. Correlation of thick beds between sections is not possible however, since single units may develop laterally into packets of thinner beds. Bed-parallel stylotites are abundant.

In the section on Monte Nicola, a 50cm bed of calcilutite showing soft sediment deformation structures is present 5m above the base (Pl.5.10). Composed predominantly of laminated calcilutite, but also containing an abundance of structureless calcilutite intraclasts, this bed can be followed laterally for at least 50m. The laminations are picked out by terrigenous silt-rich layers that are slightly more resistent to erosion. These are cut by a later set of more widely-

spaced vertical, calcite stringers. Both overlying and underlying beds are undisturbed, and there is no evidence of lateral incorporation of adjacent strata. The orientations of slump fold axes can not be determined, but fold vergences show a consistent sense of overturning towards the east.

Rarely, the limestones contain up to 15% of terrigenous silt and are finely laminated; such intervals, darker in colour than the normal calcilutites, may constitute the upper few centimetres of an otherwise massive bed and may incorporate large intraclasts of the underlying limestone (Pl.5.11). Also, massive calcilutites can be observed that grade downwards into packstones at the base of the bed (Pl.5.12).

In the Monte Nicola Section, both parallel-sided and tapering fissures have been observed in the calcilutite beneath a calcarenite bed that are filled by the overlying sediment (Pl.5.13 & 5.14). In one case the walls of the fissure are cleanly fractured, and the calcarenite filling it is structureless. The margins of the tapering 'neptunian dyke' however, are rounded, and the overlying deposits appear to have sagged into it. In both cases the bed above thins considerably over the fissure.

These limestones are similar in appearance to the stratigraphically-expanded sequences of a number of Tethyan continental margin basins of Triassic and Jurassic age (Bernoulli and Jenkyns, 1974). Of especial note are those of Late Triassic age in the Budva, Pindos and Othrys Zones of Greece and Yugoslavia, the Antalya Complex of Turkey, the Mamonia Complex of Cyprus, the Sicani and Imerese Zones of Sicily and the Hallstatt limestones of the Eastern Alps and Carpathians (Schmidt di Friedberg et al., 1960; Broquet et al., 1966; Schlager, 1969; Aubouin, Blanchet et al., 1970; Aubouin, Bonneau et al., 1970; Zankl , 1971;

Lapierre, 1975; Smith et al., 1975; Marcoux, 1976; Mišík & Borza, 1976; Fleury, 1977). They are commonly referred to in the literature as 'Halobia Limestones.'

Thin sections of the calcilutites generally disclose pelagic bivalve/radiolarian biomicrites, whose bioclast content ranges from almost nil to abundant (Pl.15). Preservation of radiolarians is poor since they are invariably calcitised, and their moulds are filled with sparry calcite or, very rarely, pyrite or chalcedonic quartz; they may also be geopetally filled, totally or partially, by micrite. The pelagic bivalves appear as thin calcitic filaments, and are probably representative of Halobia sp. (Scandone & de Capoa, 1966; de Capoa Bonardi, 1970). They are rarely intact, seldom exceed 5mm in length and are oriented at random. Where best preserved, they are often supported by a sparry-calcite cement, particularly if their arcuate shells are oriented convex-upwards. These cements have developed on the undersides of the bivalve tests and fill the cavities from above; they are not, however, epitaxial overgrowths on the original shell. In some cases it can be seen that these umbrella cements developed prior to fracture, adjacent but disoriented fragments possessing identical isopachous rims of sparry calcite.

The micrite matrix to these bioclasts is variable in appearance, being either homogeneous or clotted, and may contain scattered euhedral crystals of quartz and plagioclase feldspar. Some samples show evidence of burrow mottling imparted by a differentiation of bioclast-rich and pure micrite patches. Trace fossils, however, have only been identified in the Monte Nicola Section, where burrows ressembling Planolites can be observed piping calcarenite a limited distance into the calcilutite beneath (Pl.5.16).

X-ray diffraction analysis of the lime mudstones reveals calcite as the dominant mineral with very minor quartz, plagioclase feldspar and clay minerals. Insoluble residues comprise less than 2% of the rock, but precise determination of the clay mineral assemblage was not possible; illite is the only species that has been recognised

ii) Interpretation

The fine grain-size and lack of sedimentary structures in the calcilutites hinder interpretation of the processes by which they were deposited. The observation that some beds grade down into redeposited calcarenites suggests that they may represent the final deposits of waning turbidity currents, but for the majority, which are massive, no such inference can be made. The random distribution of bioclasts and the consistent orientation of geopetal fills in radiolarian moulds that characterise most beds suggests steady, particle by particle deposition from the water column. However, the distinction of fine-grained carbonate turbidites from pelagic layers in deep-sea sequences deposited above the CCD is difficult, and relies upon textural and faunal criteria that cannot be applied to these rocks (Hesse & Butt, 1976). The observation that burrows only penetrate a limited distance from the tops of beds cannot be taken as evidence for deposition from turbidity currents *per se* but simply as an indication of episodic and relatively rapid sedimentation. Furthermore, the absence of bioclasts of shallow water derivation, either in the calcilutites or the calcarenites, indicates that any redeposition that has occured was uniquely intra-basinal. This is compatible with the model for derivation of the lime mud from the neighbouring carbonate banks discussed below.

Cyclic sedimentation may also be manifested by the thin rhythmic nature of the bedding, particularly where partings contain paper-thin shales. However, the abundance of bed-parallel stylotites in some

sections suggests that pressure-solution may have exerted a significant influence in the development of bedding. The penetration of chert nodules from one bed into the base of the overlying layer points to a secondary origin, whereas the truncation of strata at the base of some calcarenite and calcirudite beds argues in favour of relatively early development. Dramatic truncation surfaces in similar, thinly bedded lime mudstone sequences from the Sverdrup Basin suggest that bedding in their case is primary (Davies, 1977). Wachs & Hein (1974), however, favour the diagenetic option; they note that bedding plates in the limestones of the Franciscan Formation of California do not necessarily coincide with boundaries between distinctive facies types and may even truncate primary sedimentary features. It is probable that the rhythmic bedding of the Sirino Formation reflects primary sedimentary cycles that have been enhanced and subdivided diagenetically.

The soft-sediment deformation evinced by the slump on Monte Nicola is reminiscent of the phacoidal structures described from slump sheets in the Turonian Limestones of the Teutoburger Wald which are interpreted as submarine slide deposits on the margins of the Munster Basin (Voigt, 1962, 1977). The lamination of the slumped bed gives the impression of highly visco-plastic deformation in sediments that were only weakly consolidated, and was probably achieved by laminar flow (c.f. Voigt , 1962). The nature of initial failure is unknown, but may have resembled the sigmoidal fracture pattern figured by Voigt ; early diagenetic tensional stresses are clearly evinced by the calcarenite-filled fractures observed in some calcilutites of the same section.

Slumping is generally accepted as an indicator of slopes, although the angles of inclination required may be as little as one degree (Morganstern, 1967). The most common cause of slumping is slope

over-steepening, either as a result of rapid sedimentation or tectonic tilting, but earthquakes may exercise an important influence by inducing liquification. For reasons discussed below, tectonic effects are not considered to have exercised a major control on deposition in the Late Triassic. Passive slope oversteepening by differential accumulation of sediment is favoured, therefore, particularly in view of the other evidence of extensional stresses in the sediment pile mentioned above. The cause of slope oversteepening cannot be adequately explained by the accumulation of turbidites since their frequency in the section is low. Uneven primary deposition of lime mud over the basin as a whole is therefore envisaged. The paucity of measurable slump folds precludes determination of palaeoslope orientation but the consistent eastward fold vergence may imply an easterly dip (c.f. Woodcock, 1979).

The origin of the rare, laminated, silty intervals that form the tops of some beds is not clear. Petrographic evidence suggests that they have undergone considerable compaction and pressure solution. That they are preserved at all may only be due to the incorporated intraclasts which have effectively supported the overlying bed and prevented complete elimination of the clay-rich layer. The presence of ripped-up clasts from the underlying bed attests to an erosive process, and passing turbidity currents,from whose tails the silty layers were deposited, are suggested as the most likely agents.

Evidence for early lithification of the calcilutites is abundant, particularly in the Monte Nicola and Monte Sirino Sections. The incorporation of intraclasts of calcilutite into the laminated silt layers indicates that underlying beds were already partially cemented before deposition of the marly layer. Similarly, the calcarenite-filled

fissures suggest that some calcilutite beds were cemented sufficiently to fracture brittly shortly after deposition of the succeeding layer. Finally, evidence for lack of compaction in the form of unbroken fragile bivalve shells could be taken as a likely reflection of early cementation. However, doubt has been expressed, not only about the ability of calcilutites to compact, but also upon the likelihood that such a process would lead to fracture of fragile bioclasts (Pray, 1960; Shinn et al., 1977). The significance of deep sea lithification, particularly in post-platform oozes is discussed at more length in Section 5.4.

5.3.1.2. The Terrigenous Clastic Facies

i) Description

At a variety of levels in both units, partings of green, grey , beige or red shales up to a few centimetres thick are interposed between the limestone beds. In some localities these fissile mudstones and claystones may attain greater thicknesses and dominate the succession for several metres. With the exception of Radiolaria, they are barren and those from the top of the formation are lime-free.

In the upper parts of the formation, green, paper-thin shales appear in the partings between the limestone beds, and they become progressively thicker through the succession. The top few metres are dominated by red and green shales, siliceous mudstones and rare cherts that are interbedded with thin calcilutites (Pl.5.17). Near the contact with the overlying formation, the limestones become thinner and less common, and finally disappear. Radiolaria are common in the siliceous mudstones and cherts, and the latter may exhibit normal grading that reflects the density of Radiolaria rather than grainsize (Pl.5.18). Gradational contacts can be observed from pure chert, up

through siliceous mudstone to radiolarian biomicrite; contacts between the calcilutites and overlying shales are generally sharp, however. The siliceous mudstones are composed of microcrystalline quartz, hematite and clay minerals and the Radiolaria are preserved as moulds filled with ferroan calcite and chalcedony (Fig.5.19); the cherts comprise microcrystalline quartz and contain 'ghosts' of radiolarians.

A 1.5m interval 68 metres above the base of the Pignola-Abriola Section II comprises green, yellow, brown or black mudstones, siltstones, rare sandstones and marls. X-ray diffraction analysis reveals quartz, calcite and glauconite and a clay mineral assemblage consisting solely of illite; a sandstone band contains an abundance of degraded alkali and plagioclase feldspar and chlorite, and Scandone (1967b) reports zircon, zoisite, apatite, wollastonite and clinozoisite from one sample. These deposits are finely laminated, often graded, and have an ash-like aspect.

ii) Interpretation

The increasing abundance and thickness of shale and siliceous mudstone intercalations towards the top of the formation may reflect the onset of dissolution of lime mud as the depth of the basin approached the CCD. The absence of pelagic bivalve tests, which may have been largely aragonitic, in the calcilutites at this level suggests that deposition was already occurring below the Aragonite Compensation Depth (ACD), and alternating lime-rich and lime-poor intervals may have accumulated in response to minor depth fluctuations of the dissolution facies (Fig.5.5). Sediments deposited below the calcite lysocline would have contained a low carbonate component, but episodic depression of this level of increased dissolution would have restored normal deposition of lime mud. Diagenetic compaction may have been

responsible for accentuating these alternations by removing $CaCO_3$ from the lime-poor intervals to produce the shale partings. The steady transfer of calcium carbonate from the platforms to the basins during the late Triassic may have pushed the dissolution facies to greater depths by increasing the concentrations of calcium and bicarbonate ions in solution (Section 5.5). If the rate of subsidence of the sediment-water interface was greater than the rate of depression of the lysocline, however, a gradual reduction in the proportion of calcite deposition with time would be expected; this is precisely what is observed in the Late Norian or the Sirino Formation. Episodic delivery of lime mud could have caused a stepwise descent of the lysocline that, superimposed upon the steady subsidence of the basin floor, may have produced cycles of lime-rich and lime-poor sedimentation over a protracted period of time (Fig.5.6).

The limestone-shale rhythms of the late Jurrasic Oberalm Limestones of the Northern Calcareous Alps have also been ascribed depositional depths between the ACD and CCD (Garrison & Fischer, 1969). However, periodic fluctuations in calcareous nannoplankton productivity were held responsible for producing these cycles (Garrison, 1967). On the other hand, fluctuations in the depth of the lysocline has been argued as a likely mechanism for producing calcareous ooze - clay cycles in the early Neogene sediments of the Sierra Leone Rise of the Atlantic Ocean drilled during Leg XLI of the Deep Sea Drilling Project (Dean et al., 1978). The periodicities of these cycles leads the authors to explain the lysocline fluctuations in terms of climatic changes, related to the earth's orbital cycles. These are thought to have affected the flow of Antarctic Bottom Water, to whose top the lysocline relates in the modern Atlantic Ocean.

The absence of smectites from the clay mineral assemblages of the 1.5m laminated mudstone, siltstone and sandstone succession in the Pignola-Abriola Section II does not support the contention that they represent volcanoclastic deposits (c.f. Dietrich & Scandone, 1972). However, it is difficult to explain their graded, varve-like and ashy appearance or their mineralogy in any other way than as submarine ash bands derived from moderately distant eruptions.

5.3.1.3. The Calcirudite Facies

i) Description

In several sections from Unit II, calcirudite beds up to 1m thick are commonly found near the top of the formation. The thickest and most coarse-grained examples are located at Torrente Bitonto, where the largest clasts attain cobble grade (Pl.5.20). Rounded but elongate fragments of intraformational calcilutite are dispersed through a lighter-coloured, fine grained, marly limestone matrix. The beds are poorly-sorted, show no grading or imbrication of clasts and possess no sedimentary structures (Pl.5.21). Chert nodules are present that replace both matrix and clasts, and beds are interrupted by abundant, horizontal stylotites. The base of one bed was observed to have cut down several tens of centimetres into the underlying strata (Pl.5.22) Thinner beds of similar aspect, but comprising pebble and granule-grade clasts, are present in the Rupe del Corvo and Monte Armizzone Sections. In some cases it is possible to observe chert nodules that are truncated at the margins of their host pebble and that are not elongate parallel to bedding. Intensive stylotitisation in the matrix has brought many clasts into close juxtaposition (Pl.5.23).

Pervasive dolomitisation of the upper part of the formation in outcrops north of Marisco Nuovo hampers recognition of this facies in the La Ralla and Pignola-Abriola Sections, but brecciation of some form is

evinced by the presence of fractured chert nodules (Pl.5.24). In the Pignola-Abriola Section II, these occur in massive beds of dolomite that are themselves composed of matrix-supported, angular clasts. These beds are several metres thick and are separated by thin, well-bedded dolomites with undisturbed chert nodules. The fractured cherts may remain virtually intact or fragments may be dispersed locally within the matrix; isolated clasts are scattered throughout. In polished faces, concentric and dome shaped laminae are discernible in the dolomitic clasts but the cherts, which occur solely in the matrix, are structureless (Pl.5.25). In thin section, both matrix and clasts are composed of saccharoidal dolomite rhombs and primary fabrics are entirely destroyed. Neither are relict structures preserved in the cherts which simply comprise homogeneous, microcrystalline quartz; those from the La Ralla section, on the other hand, retain chalcedonic ghosts of thin-shelled bivalves and Radiolaria.

ii) Interpretation

In contrast to the lime mudstones, this coarser grained facies shows abundant evidence of deposition from gravity flows. The features that characterise the calcirudites, particularly the random dispersion of pebbles in a fine-grained matrix, lack of grading or preferred orientation of clasts, and thick, massive beds, are typical of debris-flow deposits (Middleton & Hampton, 1976). The observation in the Torrente Bitonto Section that the base of one flow sharply truncates the strata beneath attests to filling a pre-existing channel. The importance of debris flows in producing turbidity currents by mixing of water with material eroded from the front of the flow has been demonstrated by Hampton (1972). Turbidity currents generated in such a way, passing in advance of the main flow, may have eroded the channel prior to

deposition of the calcirudite.

It seems probable that these debris flows developed from slumps by incorporation of water into the slide (Fig.5.7a)(Hampton, 1972; c.f. Cook & Taylor, 1977). Slumps on the scale of that seen on Monte Nicola, however, are insufficient to produce debris flows of this size and the variability and texture of the clasts indicate that a large number of calcilutite beds have been incorporated. Failure of a considerable thickness of the lime mudstone succession must therefore be invoked, possibly on as dramatic a scale as is evinced by truncation surfaces in sequences of the Sverdrup Basin, the Rancheria Formation of New Mexico and the Turonian chalks of Normandy (Juignet & Kennedy, 1974; Davies, 1977, Yurewicz, 1977).

It is unlikely that similar processes were responsible for generation of the dolomitic breccias of the Pignola-Abriola II and La Ralla sections. The irregular laminations observed in the clasts are interpreted as algal structures typical of a shallow-water limestone. Although dolomitisation precludes identification of shallow-water organisms or fabrics in thin section, these clasts are considered to be derived from an adjacent carbonate platform. The identification of ghosts of pelagic bivalves and Radiolaria in the cherts indicates a basinal origin for the matrix, but it is evident that the clasts are extraformational. Since the breccias are matrix-rich and show random dispersion of ungraded clasts, deposition from debris flows is favoured, but fracturing and scattering of chert nodules indicates that mass transport occurred subsequent to their genesis. One must infer, therefore, that a time-lapse sufficient for the development of the nodules or their precursors preceded initiation of the debris flow. The peri-platform mudstones which hosted these cherts almost certainly

contained a high proportion of high-magnesian calcite (see Section 5.4), and it has been shown that the solution-repreciptation mechanism involved in the transformation of biogenous opal-A to opal-CT, a chert precursor, is greatly accelerated in the presence of magnesium ions (Kastner et al., 1977). Early nodule development may, therefore, have been rapid and as gravity-induced transport commenced, the precursor-nodules were fractured and became dispersed through the lime mud matrix.

Detachment of carbonate platform material by earthquakes, and its emplacement by rockfall into the lime muds of the basin is considered to have been responsible for generating these breccias. As they accumulated around the basin margins, movement of debris flows was initiated, inducing mixing of angular platform carbonate clasts with the indigenous muds. This was caused either by slope overstepping or by the development of high pore pressures in the underlying sediments (Fig.5.7b). These gravity flows may have travelled laterally for several kilometres before finally coming to rest (c.f. Mountjoy et al., 1972; Conaghan et al., 1976). The absence of large isolated blocks or olistoliths of platform carbonate rock, however, suggests that the sequences described lay some distance from the basin margin itself, (c.f. Mountjoy et al., 1972; Cook et al., 1972; Johns, 1978). Generation of intraformational debris flows in more distal environments by large-scale slumping was also probably triggered tectonically.

Allochthonous carbonate breccias and meggabreccias containing blocks and clasts of bank-derived carbonate in a lime-mud matrix commonly occur in deeper water basinal successions adjacent to carbonate bank complexes (Mountjoy et al., 1972; Cook et al., 1972;

Enos, 1974, 1977; Carrasco-V,1977; Davies, 1977; Hubert et al., 1977; McIlreath, 1977; Yurewicz, 1977). Although the dolomitised breccias of the Sirino Formation lack the coarse clasts (greater than 1 m) of a true meggabreccia, they nonetheless possess many features typical of these sediments, which are generally interpreted as the products of debris flow. Genesis of these mass flows is attributed to a number of factors, including progradation of carbonate buildups over unconsolidated basin sediments, making the margins unstable (Cook et al., 1972). However, the widespread renewal of block faulting and extensional tectonic regimes in the Late Triassic and Early Liassic throughout the Mediterranean region suggests that detachment of portions of the platform margins by earthquakes is more likely to have pertained in this case (Bernoulli & Jenkyns, 1974; c.f. Conaghan et al., 1976). Jurassic scarp breccias are common in the Alps, and carbonate rudites and meggabreccias of probable Liassic age are also documented from Central Greece (Finger, 1975; Trümpy, 1975; Johns, 1978). Dolomitised breccias additionally occur at a similar stratigraphic level in the Fanusi Formation of the Madonie Mountains, Sicily (Scandone et al., 1972). In all these cases, carbonate platform sequences can be found close by or in adjacent structural units.

Deep-sea drilling off the eastern side of the Bahama Banks has revealed pebbly mudstones comprising shallow-water limestone clasts in a nannoplankton chalk matrix (Beall & Fischer, 1969). These, too have been interpreted as comprising rubble fallen from the bank margins that became incorporated with the indigenous oozes and was transported seawards as mudflows.

5.3.1.4. The Calcarenite Facies

i) Description

Occurring sporadically through the majority of sections in the formation of both units are packstone, wackestones and grainstones, that comprise sand-sized carbonate particles and bioclasts. They can usually be recognised in the field by their grainy texture or by the presence of lamination. Predominantly bioclastic and intraclastic subfacies can be distinguished.

a) The Bioclastic Calcarenite Sub-facies

The most common sub-facies comprises packstones and grainstones that are usually laminated and are constituted by grain-supported thin-shelled bivalve tests, probably of Halobia sp. and Daonella sp. These are set in a lime mudstone matrix, a sparry calcite cement, or most commonly, a combination of the two. Often forming the tops of beds, they may be underlain by intraclastic calcarenites; alternatively, they may occur at the base of a thicker bed of calcilutite. Beds or packets of beds are commonly interbedded with green shales at certain levels in several sections (Pl.5.26). Replacement chert nodules and stringers, which occur predominantly at the tops of beds, may preserve parallel- and low-angle cross-lamination of the limestones (Pl.5.27). Individual sets are lens-shaped, have irregular lower bounding surfaces, and the cross-laminae show bimodal dip directions. Imprints of Halobia sp. shells are common on bedding surfaces, though not within the shales (Pl.5.28).

Microscopic examination reveals two distinct micro-facies: an intrasparite, composed of arcuate bivalve shells containing geopetal micrite fills, and biosparites, that comprise closely-packed, spar-cemented shells flattened parallel to bedding (Pl.5.29 & 5.30).

It is only the latter which are strongly laminated. The geopetal fills of the intrasparites display a variety of orientations, and incorporate smaller fragments of pelagic bivalve shells, calcitised Radiolaria, flecks of pyrite and euhedral, cubic crystals of quartz. Rare micritic intraclasts are also present. The microspar matrix is more coarse-grained than the micrite of the intraclasts, and cavities sheltered beneath bioclasts are filled by sparry calcite.

The biosparites differ from the intrasparites by virtue of their lamination, imparted by parallel orientation of filamentous bivalve shells. This lamination is not always planar and may appear crenulated in thin section. These accumulations of shells commonly constitute lumachelles, tests lying in contact with each other and being cemented by sparry calcite. A partially silicified sample from the Pignola-Abriola Section II shows an early, isopachous, calcite cement lining the cavities between bioclasts, and a second generation of void-filling chalcedony (Pl. 5.31, 5.32 & 5.33); in this example, the shells are not flattened.

b) The Intraclastic Calcarenite Sub-facies

Other types of calcarenite which contain a much lower proportion of bioclasts may also be encountered. These include grainstones, packstones and, rarely wackestones of well-rounded, sand- to granule-sized grains of lime mudstone. They occur in beds up to 30 cm thick, though seldom exceeding 10 cm, are laterally continuous and generally of constant thickness (Pl.5.34). They are commonly graded, may pass upwards into bioclastic calcarenites and may themselves exhibit small-scale cross-laminations in the upper few centimetres (Pl.5.35). These structures, which are only visible in polished faces occur in tabular sets about 3cm thick and show consistent dip directions.

In the Monte Nicola Section, three such calcarenite beds have been identified. One of these fills a shallow depression 2.5m wide and 30cm deep at the centre, but overlaps the margins and can be followed laterally over several tens of metres (Pl.5.36). The depression cuts down 20cm into the underlying strata and is lined by a finely laminated mudstone about 3cm thick; the lamination is imparted by closely-spaced stylotites (Pl.5.37). The calcarenite itself passes from medium sand-grade at the base to fine sand at the top and lacks internal sedimentary structures with the exception of parallel-lamination in the upper part. The top surface is planar. A sample collected from a 10cm bed at a similar stratigraphic level below the base of the Lagonegro Section in the Fiume Carboncello Gorge, shows similar features. However, it grades from granules to fine sand, is matrix supported at the base and the top 2cm are constituted by laminated biosparites (Pl.5.38).

Petrographically, the intraclastic calcarenites can be classified as intrabiosparites and intrabiomicrites, the rounded grains chiefly consisting of radiolarian biomicrite identical to that of the calcilutite facies (Pl.5.39). In addition there are benthonic foraminifera, crinoid ossicles and plates with epitaxial calcite overgrowths, ostracodes and thin-shelled bivalves (Pl.5.40 & 5.41). These grains are either spar-cemented or supported in a micrite matrix.

ii) Interpretation

Grading of calcarenite beds that also show structurless, planar- and cross-laminated intervals suggests that they are turbidites (Bouma, 1962). Complete Bouma sequences are rarely developed, however, beds containing BCDE and CDE divisions being most common; the A, B and

C intervals are chiefly developed in intraclastic calcarenites, whilst bioclastic facies characterises the C and D divisions. Beds that show no current structures or grading may have been deposited by other types of gravity flow; structureless and ungraded intraclastic calcarenites that lack matrix, for example, were probably deposited from grain flows (Middleton & Hampton, 1976).

Because of the absence of faunas and textures typical of shallow-water carbonate environments among the clasts, the calcarenites are interpreted as redeposited intrabasinal sediments. The thin-shelled bivalves and Radiolaria contained in the intraclasts are both pelagic organisms, and other bioclasts present as grains have no bathymetric significance; stalked crinoids, for example, have been observed at depths of 600-700m in the Straits of Florida (Neumann et al., 1977). Furthermore, micritisation of bioclasts, a process normally attributed to boring by endolithic algae in shallow marine environments, has also been shown to be a significant process at depths of as much as 5000m (c.f.Zeff and Perkins, 1979; Bathurst, 1966, 1967; Alexanderson, 1972). It is suggested, therefore that these calciturbidites were generated within the basin, probably as slumps, and that they picked up a variety of bioclasts during transport which were incorporated into the flow. The pelagic bivalves that commonly dominate the C and D intervals may either have been hydrodynamically sorted from the original slumped sediment or, like the other bioclasts, entrained during passage of the turbidite. The small slump on Monte Nicola is probably typical of the type which developed into turbidites, and an envisaged genetic sequence relating such slumps to the observed types of calcarenite is given in Fig.5.8.

The planar- and cross-laminations of bioclastic calcarenites that lack other features of a turbidite may attest to bottom currents which winnowed out the finer sediments and produced small-scale ripples. Such beds can usually be distinguished from turbidites by their good sorting and sharp top and bottom contacts (Middleton & Hampton, 1976). These currents may have been responsible for sorting and breaking bioclasts and finally concentrating them into discrete beds. There is evidence that such currents are effective in achieving bed-load transport in the present-day deep channels of the Bahama Paltform. Currents of up to 60 cm/sec are evinced by ripples observed in the abyssal sediments of the Straits of Florida and currents in excess of 50 cm/sec have actually been recorded at a depth of 320m (Hurley & Fink, 1963; Neumann & Ball, 1970). In addition, symmetrical ripple marks are recorded at depths of 2000m in Tongue of the Ocean and current velocities of 28-90 cm/sec relating to tidal oscillations have been invoked to explain their formation (Busby, 1962).

An altogether different mechanism has been suggested for the development of identical lumachelles in coeval basinal limestone successions of the Sicani Mts. of Sicily (Montanari & Renda, 1976). Changing physico-chemical conditions are thought to have produced accumulations of small, juvenile shells, whilst stratigraphic condensation has been invoked for concentrating larger, adult forms. However, in the absence of independent evidence for periods of slow sedimentation, such as can be found in condensed sequences of the Tethyan Jurassic for example, the latter process is considered implausible; indeed, both the sedimentology and stratigraphy of the formation suggest that deposition was rapid (Broquet, Caire & Mascle, 1966 and author's personal observation). Size sorting by tractive currents is considered a much more likely mechanism, particularly in view of the abundance of other evidence for intrabasinal redeposition in the

form of intraclastic calciturbidites, and the parallel- and cross-laminations of the lumachelles themselves.

The unusual nature of the calcarenite-filled depression in the Monte Nicola Section calls for comment. The presence of a laminated mud drape beneath the calcarenite suggests that more than one depositional event was involved; it is difficult to envisage how a single turbidity current could have cut the channel, and then deposited the mud drape and the calcarenite. Furthermore, the overlapping of the bed over the sides of the depression suggests that the channel was already in existence at the time of deposition. Erosion of the gulley was probably effected by a non-depositing turbidity current; alternatively it may represent a small slump scar (Fig.5.9). The sides were then covered by a mud drape, deposited by grain-by-grain settling from the suspension cloud left by the turbidity current or slump. The gully was finally filled by a second turbidity current that entrained a certain amount of the unconsolidated mud drape into its base, producing rapid deposition. This is preferred to an alternative model, whereby the calcarenite subsided under its own weight into the mud filled channel, because the top of the calciturbidite bed is planar.

Generation of micrite intraclasts, and the abundance of inverted geopetal structures in redeposited bivalve shells, provides further evidence for early lithification. Furthermore, the lack of compaction evinced by coquinas whose constituent fragile, thin pelagic bivalve shells have remained intact implies that early cementation provided support as overburden increased. Similar features are recorded in the Upper Triassic thinly bedded Hallstatt Limestones of the Northern Calcareous Alps (Zankl, 1969).

Allodapic limestones are a common feature of basinal carbonate successions but in general they contain bioclasts that attest to shallow-water provenance (Meischner, 1964; Bernoulli & Jenkyns, 1974; Cook & Enos, 1977 & reference therein). Crinoidal turbidites, which lack shallow-water allochems and are considered to have originated in a slope environment seaward of a carbonate platform margin, are, however, described from the Upper Palaeozoic limestones of the Sverdrup Basin (Davies, 1977). Further examples may be found in Devonian pelagic sequences of Europe and in the Tethyan Jurassic (Bernoulli & Jenkyns, 1970; Tucker, 1973).

5.4 Cherty lime mudstones of the Tethyan Late Triassic - An overview

Rhythmically bedded grey or white cherty calcilutites are a lithology common to many ancient deep-water carbonate sequences, and they may occur in units of both oceanic and continental-margin affinity (Wilson, 1969; Cook & Enos, 1977 and references therein; Jenkyns, 1978). Of especial note are those of the Tethyan region (Garrison & Fischer, 1969; Bernoulli, 1972; Bernoulli & Jenkyns, 1974), the Franciscan Formation of California (Bailey, Irwin & Jones, 1964; Wachs & Hein, 1974, 1975), the Sambosan Belt of Japan (Kanmera, 1969) and the Tamabra limestones of Mexico (Enos, 1974, 1977; Carrasco-V,1977). Numerous examples are also documented from the Palaeozoic of North America (Cook & Enos, 1977, and references therein). The Lagonegro Zone is only one of a number of Palaeogeographic units of basinal affinity in the Tethyan region that are characterised in particular by cherty limestones of Late Triassic age (Fig.5.3). It is the purpose of this overview to discuss the possible origins of the lime muds from which these 'Halobia Limestones' developed.

Deep-water lime mudstones have generally been interpreted as pelagic due to the absence of benthonic organisms and the presence of ammonites, thin-shelled bivalves, planktonic Foraminifera and siliceous or calcareous nannoplankton. Those of Late Jurassic age or younger, in common with the calcareous oozes of modern oceans, derived the majority of their lime mud from calcite-secreting micro- and nanno-organisms such as coccoliths and planktonic Foraminifera (Garrison & Fischer, 1969; Bosellini & Winterer, 1975); basinal limestones of Liassic age from the Glasenbach Gorge, Austria, have also been shown to contain calcareous nannoflora belonging to the species Schizosphaerella in rock-forming quantities (Bernoulli & Jenkyns, 1970). However, it is not generally thought that these organisms contributed significantly to pelagic carbonate sedimentation prior to their proliferation in the Late Jurassic and Early Cretaceous, even though sporadic examples of their remains are documented from the Liassic, Triassic and Late Palaeozoic (Bramlette, 1958; Nöel, 1965; Fischer, Honjo & Garrison, 1967; Pirini Radrizzani, 1971; Gartner & Gentile, 1972; Mišík & Borza, 1976; di Nocera & Scandone, 1977; Minoura & Chitoku, 1979). An abundance of Radiolaria, as well as thin-shelled bivalves of the Posidoniidae family, that are thought to have been either nekto- or pseudo-planktonic, and the absence of benthonic organisms, all lend support to a postulated nannoplanktonic origin for the cherty limestones of the Sirino Formation (c.f. Jeffries & Minton, 1965; Stanley, 1972, 1974). However, although Nannoconus sp. and other coccoliths of unknown taxonomic affinity are documented from these and similar rocks from Sicily, examination of a large number of samples under the SEM failed to confirm the contention that nannofossils were abundant and comprised the majority of micrite (c.f. di Nocera & Scandone, 1977). It is possible, on the other hand, that diagenesis and neomorphism

may have led to their recrystallisation and absorption into the micritic mosaic that characterises the ultrastructure of these limestones; (c.f. Wachs & Hein, 1974)(see Chapter 8). It is also conceivable that the degree of calcification, and hence preservation potential, of primitive nanno-fossils was low but that they contributed significantly to sedimentation in the form of disaggregated crystals (Jenkyns, 1971). Certainly the rates of deposition evinced by the Sirino Formation limestones (about 10-25 m/million years) are of the same order as those of typical post-Tithonian nannofossil limestones of the Alps and Apennines (17-51 and 8-10 m/million years respectively) (Schlager, 1974; di Nocera & Scandone, 1977). Although coccoliths cannot be ruled out as a contributor to sedimentation, therefore,supportive evidence is highly circumstantial and more plausible alternatives must be sought.

The carbonate platforms which surrounded the Lagonegro Basin during the Late Triassic represent a more probable source of the lime mud. Bahamian shallow-water environments are known to produce large quantities of such material, chiefly in the form of aragonite or high-magnesian calcite, and there is evidence that storm surges may be responsible for carrying considerable amounts off into adjacent basins (Bathurst, 1975). High tide during the passage of Hurricane Donna, for example, was more than a metre above its normal level, and muddy sediment carried from the platform with the ebb was deposited locally in the Straits of Florida to form a 15cm layer (Ball et al., 1967). Furthermore, the contribution of material from the Bahama Banks to sedimentation in the intervening deep channels, such as Tongue of the Ocean and Exuma Sound, has also been conclusively demonstrated (Busby, 1962; Pilkey & Rucker, 1966; Rucker, 1968; Schlager & James, 1978; Schlager & Chermak 1979). This is

achieved not only in the form of calciturbidites but also by grain-by-grain deposition from suspension. The high initial contents of aragonite and high-magnesian calcite in these peri-platform oozes militates against their being entirely nannoplanktonic. Moreover, these metastable components survive only a short time before either being dissolved or stabilised as low-magnesian calcite; early lithification is therefore a major feature of the diagenesis of these deposits (Schlager & James, 1978). True nannoplanktonic oozes, on the other hand, may remain unlithified for as much as 100 MY due to their relatively low diagenetic potential (Schlanger & Douglas, 1974; Matter et al., 1975). It is suggested that evidence for early lithification in the cherty limestones of the Sirino Formation bears witness to high initial contents of metastable carbonate and argues strongly in favour of their having originated as peri-platform oozes comparable to those of the Bahamas.

The association of the majority of sequences containing Upper Triassic cherty limestones in the Alpine-Mediterranean region with platform carbonate units certainly attests to a genetic relationship between the two (Bernoulli, 1972; Bernoulli & Jenkyns, 1974). Furthermore, this association has also been noted elsewhere in the geological record (Wilson, 1969, 1974; Cook & Enos, 1977, and references therein). These basinal deposits commonly occupy an intermediate position between lime-poor basinal shales and shallow-marine carbonates, and in many cases thin dramatically away from the edge of the bank. Although the latter observation may be considered adequate testimony to their provenance, typical sequences contain a high proportion of calciturbidites and breccias which would have been thicker and more numerous around the edge of the

basin. Moreover, carbonate dissolution levels in Palaeozoic and Early Mesozoic basins may have been at much shallower depths than at present, and the transition from lime mudstones to shales may thus relate to depth rather than to proximity of the platform margin. Caution should therefore be exercised in interpreting the association of cherty limestone successions with carbonate bank sequences in these older basins.

A third alternative for the origin of the micrite is that much of it was inorganically precipitated as high-magnesian calcite lutite, such as is known to have formed recently in the Mediterranean and Red Sea basins under conditions of elevated salinity and temperature (Gevirtz & Friedman, 1966; Milliman et al., 1969; Müller & Fabricius, 1974; Milliman & Müller, 1973, 1977; Sartori, 1974). Sediment cores from the eastern Mediterranean contain 20-50% high-magnesian calcite, mainly in the silt and mud fraction, which is thought to have precipitated inorganically at the sediment-water interface under ambient salinities of 38-39‰ and temperatures of 13-14°C (Milliman & Müller, 1973). Where rates of sedimentation are sufficiently low, cementation of this lutite into nodules may occur (Müller & Fabricius, 1974). In the Red Sea, under salinities of 40‰ and temperatures greater than 21°C, similar muds are being formed at rates of 10 to 50 m/MY (Milliman et al., 1969). Even allowing for considerable compaction, these values are of the same order as rates of sedimentation for the Sirino Formation limestones (about 10-25 m/MY). Probably constituting a small, pelagic basin of relatively restricted circulation and lying in topical latitudes, conditions in the Late Triassic Lagonegro Basin may have been similar to those of the Mediterranean and Red Sea basins today.

Although the geochemistry of the limestones cannot provide any positive support for the contention that they originated as high-magnesian calcite lutites in this way, the presence of dolomitic chert nodules in some sections may attest to high contents of magnesium in the original sediment (See Chapter 8). Evidence for early lithfication also may point to inorganic precipitation as a significant process, the importance of high-magnesian calcite and, less commonly, aragonite as submarine cements now being widely recognised (Friedman, 1964; Milliman, 1966; Fischer & Garrison, 1967; Thompson et al., 1968; Neumann et al., 1977).

Study of the ultrastructure and geochemistry of the Sirino Formation cherty limestones has failed to find a solution to the problem of their origin, and data are insufficient to allow discrimination between the three possibilities proposed above (See Chapter 8). It is possible that all have made a significant contribution, but it seems most likely that the bulk was derived from the neighbouring carbonate banks.

A further paradoxical feature of the majority of Late Triassic cherty limestone successions of the Alpine-Mediterranean region is the paucity or complete absence of calciturbidites containing displaced shallow-water faunas and clast fabrics. This is commonly the case even in sequences that contain an abundance of redeposited platform carbonate material higher in the succession, and which were therefore in the manifest vicinity of a platform margin. The Bahamian lesson demonstrates that tectonic stimuli are not necessarily required to generate mass flows, so simply to explain their absence in terms of tectonic quiescence is clearly inadequate. Although the Triassic buildups of the Alpine-Mediterranean region, lacking true fringing reefs, did not constitute 'reefs' in the ecological sense,

it is evident that they acted as a copious source of skeletal sands for the adjacent basinal carbonates of the San Cassian Formation (Bosellini & Rossi, 1974). The interior margins of the Bahama Banks likewise lack fringing reefs, but intraplatform deep basins, such as Tongue of the Ocean, nonetheless contain large numbers of bank-derived calciturbidites (Busby, 1962; Rusnak & Nesteroff, 1964; Gibson & Schlee, 1967; Schlager & Chermak, 1979). Only a narrow zone around the margin of the basin which is by-passed by the majority of turbidity currents, lacks abundant redeposited calciclastics. It seems highly probable, therefore, that potential sources of skeletal sand must also have existed around the Late Triassic deep basins of the Alpine-Mediterranean region, but that this material was unable to reach the more distal areas. This was evidently the case in the Othrys Basin of Greece where Triassic and Jurassic turbidite-free lime mudstones of the Neokhorion Formation can be observed interfingering with the distal deposits of a carbonate submarine fan (Price, 1976, 1977a). The morphology and slope of the Late Triassic platform margins may have been incapable of producing mass flows of sufficient magnitude to cover wide areas of the basin floor. Liassic rifting, however, possibly rendered the platform margins less stable such that re-deposition in the Jurassic occurred on a much greater scale, affecting even the most distal parts of the basin. This perplexing problem is one for which it is difficult to find a more convincing answer.

In view of the apparent lack of tectonic disturbance during most of the Late Triassic, the ubiquity of intrabasinally redeposited beds throughout the basin and at all stratigraphic levels requires

explanation. The most likely generative process was slope over-steepening, only small angles of inclination being necessary to induce slumping. Any differential relief produced by block faulting during the Middle Triassic, however, would have been gradually eliminated as the carbonate oozes of the Sirino Formation blanketed the basin floor. Despite this, redeposition persisted, even in the deepest parts of the basin. The continuity of re-sedimentation can most easily be accounted for by invoking preferential accumulation of lime- mud around the basin margins. This admittedly circumstantial evidence perhaps provides the most convincing testimony to the provenance of the Late Triassic calcareous oozes of the Lagonegro Basin.

5.5 Model for the Sirino Formation

In constructing a depositional model for the Sirino Formation, the following factors must be accounted for:

i) The probable time equivalence of the Gianni Griecu Member and the Valle del Pesce Member of the Monte Facito Formation.

ii) The greater thickness of the formation in Unit I than in Unit II.

iii) The presence of extraformational carbonate breccias in the more northerly sections of Unit II (Pignola-Abriola II & La Ralla Sections).

iv) The presence of intraformational carbonate breccias in the more southerly sections of Unit II (Rupe del Corvo, M.Armizzone, T. Bitonto).

v) The presence of intraformational calcarenites at all stratigraphic levels in both units.

vi) The evidence of slumping at the top of the Monte Sirino Member of Unit I (Monte Nicola Section).

vii) The evidence of storm activity in the Gianni Griecu Member.
viii) The evidence of carbonate dissolution near the top of the formation.

Fig.5.10 is a postulated cross-sectional reconstruction of the basin in the Late Norian. Typical stratigraphic sequences are superimposed to illustrate how the different successions relate to the overall development of formation. Due to the lack of palaeocurrent and palaeoslope data, the morphology of the basin cannot be determined. The sedimentology of Unit II in the north, however, suggests proximity to the platform margins, whilst the more distal environments of Unit I are representative of the central parts of the basin. The absence of proximal facies in Unit I may imply that only distal environments are represented. Alternatively Unit I may have been juxtaposed with a stable platform margin from which no extraformational breccias were derived.

Consecutive stages in the development of the basin during the Upper Triassic are depicted in Fig.5.11. At the end of the Ladinian, the final stages of deposition of the Monte Facito Formation in Unit II saw the development of nodular limestones and dolomitic marls; there is no record of coeval sedimentation in Unit I. In the Early Carnian, the basinal calcilutites, marly limestones and shales of the Gianni Griecu Member were deposited, whilst condensed sedimentation continued in Unit II (Fig.5.11a). The two facies probably reflect different depths of deposition inherited from differential subsidence during the Middle Triassic. The Valle del Pesce Member developed in shallower, well-oxygenated waters where storm-generated currents caused sediment winnowing and concomitant slow rates of deposition. Material removed from the higher levels of the basin were redeposited, either as turbidity currents or from

suspension, in the deeper parts of the basin, under conditions of more rapid sedimentation; these were finally reworked to produce the limestone/marly limestone/shale cycles of the Gianni Griecu Member. More reducing conditions during diagenesis reflect the higher rates of deposition, and the sediments are therefore grey rather than red in colour. The thickness differences between the two units may be a function of differential relief of the basin floor.

By the Middle Carnian, topographic irregularities had been eliminated and sedimentation became homogeneous throughout the basin (Fig. 5.11b). Lime-mud, derived largely from the neighbouring platforms, accumulated around the basin margins and slope over-steepening led to slumping and redeposition of the calcilutites as thin, intraformational calcarenites. Small cracks developed in some calcilutite beds, attesting both to early lithfication and the development of extensional stresses in the sediment pile shortly after deposition. Regional subsidence and relative accretion of the surrounding carbonate platforms had increased the depth of the basin such that storm influences were no longer directly felt by the basinal sediments. Rythmic bedding on the other hand, could be a reflection of episodic deposition induced by intermittent delivery of lime-mud to the water-column by storm-surges on the platforms.

As subsidence continued into the Middle and Late Norian, the basin floor passed gradually beneath the Aragonite Compensation Depth and Calcite Lysocline. Introduction into the basin of large quantities of lime-mud, composed predominantly of metastable carbonate minerals, caused stepwise suppression of these dissolution levels; coupled with the subsidence of the basin floor, this led to

alternating periods of calcareous and siliceous deposition, the latter becoming successively more prolonged until the basin finally sank below the CCD.

The renewal of tectonic instability in the Late Triassic is evinced by the development of both intra- and extra-formational breccias in Unit II (Fig. 5.11c). Slumping of considerable thicknesses of basinal mudstone was responsible for generation of the former, whilst rockfall and subsequent debris-flows are invoked to explain the origin of the coarse, angular breccias containing clasts apparently derived from collapse of the platform margins.

5.6. Discussion

'Internal evidence of depositional depth in a pure calcite mudstone (Micrite) must ..... be scarce to non-existent'. This statement by Bathurst (1967) is as true now as it was a decade ago and moreover many of the features considered at that time to have been diagnostic of shallow-water have now been shown to be ambiguous. Stromatolites, ooids, algal borings and reef complexes have all been shown to develop in pelagic environments at depths of 200 m or more (Hallam, 1970; Jenkyns, 1971, 1972; Neumann et al., 1977; Zeff & Perkins, 1979). Some constraints may, however, be put upon the depositional depths of the lime mudstones of the Sirino Formation.

Recognition of algal stromatolites in the condensed Jurassic sequences of western Sicily has been used as evidence that deposition occured in the photic zone at depths of no more than 200 m (Jenkyns, 1971). No such features have been observed in the nodular limestones of the Valle del Pesce Member and greater depths are possible, although similar values have been suggested for

comparable facies in the Norian Hallstatt limestones of the Alps (Zankl, 1971). The depth of deposition of the Gianni Griecu Member may have been as much as 50-100 m greater, however, judging by the difference in thickness of the Early Carnian succession in each unit. In view of the evidence of storm reworking in the sediments of the Gianni Griecu Member, it is considered unlikely that the deepest parts of the basin were more than about 250 m below sea level at that time.

The transition from carbonate to siliceous deposition at the end of the Jurassic reveals that the basin floor had reached the CCD. It has been suggested that, prior to its depression at the end of the Jurassic, the CCD in the Tethyan region was at 2500 m (Boselli & Winterer, 1975). Assuming 50% compaction of the lime mudstones of the Sirino Formation and a depth of 250 m at the beginning of the Late Triassic, a basement subsidence of 3000 m is implied during the 15 MY before the beginning of the Jurassic. This subsidence rate of 200 m/MY is considerably greater than the 100-150 m/MY maximum implied by the thickness of shallow marine sediments of the neighbouring carbonate platforms (d'Argenio, 1976). In the absence of evidence for block faulting during this period, which could have produced differential subsidence rates for the platforms and basins, it is suggested that CCD in the Lagonegro Basin at the beginning of the Jurassic was shallower than 2000 m, and that it was pushed to progressively greater depths throughout the Early Mesozoic due to the introduction of lime mud from the carbonate banks.

The similarity between the postulated palaeogeography of the southern Apennine platforms and basins and the modern configuration of the Bahamas has already been mentioned (d'Argenio, 1970a;

Bernoulli,1972;d'Argenio, de Castro et al.,1975;Catalano et al.,1976). Deep basins, such as Tongue of the Ocean and Exuma Sound, might therefore be suitable places to find modern analogues for the lime mudstones of the Sirino Formation. Site 98 of the Deep Sea Drilling Project, situated in Northeast Providence Channel in 2750 m of water, encountered 357 m of white-to cream-coloured foraminiferal-nannoplanktonic carbonate oozes and chalks of Tertiary and Late Cretaceous age (Fig.5.12) (Hollister, Ewing et al., 1972). A high proportion of these sediments may have been derived from the adjacent platforms; aragonite needles are abundant in Holocene and Pleistocene samples, and high-magnesian calcite can be found in deposits of Pliocene and younger age. Bioclastic limestones bearing displaced reef debris are also common. The bulk of the sediment, however, is composed of calcareous nanno-plankton, and the rates of sedimentation, between 5 and 40 m/MY, are comparable to those of calcareous oozes in modern ocean basins (Berger, 1974). These alternations of predominantly nannoplanktonic chalks and bioclastic turbidites are more evocative of the Upper Cretaceous 'Scaglia Limestones' of the Alpine-Mediterranean region than the Upper Triassic limestones of the Sirino Formation (Bernoulli, 1972).

Carbonate oozes from the Tongue of the Ocean, on the other hand, contain much higher proportions of bank-derived material, comprising aragonite, calcite and high-magnesian calcite in a ratio of 3:2:1 (Gibson & Schlee, 1967; Schlager & James, 1978; Schlager & Chermak , 1979). Rates of sedimentation in the axial parts of the basin vary between 4 and 17 m/MY but may be at least an order of magnitude higher near the flanks (Fig.5.12)(Busby, 1962). Turbidites derived mainly from the carbonate banks are abundant and

may make up 70% to 90% of the sediment (Rusnak & Nesteroff, 1964). However, the frequency of turbidites is insufficient to explain the greater sedimentary rates near the flanks, and it seems that lime-mud is preferentially accumulating around the edges of the basin (c.f. Schlager & Chermak, 1979). Indeed, turbidites are generally more common in the axial parts of the basin than on the flanks, which are by-passed by the majority of turbidity currents. The blocky and precipitous topography of the basin margins, where cliffs tens of metres high may be exposed, has been interpreted as the product of slumping and erosion by turbidity currents, and, together with the presence of intrabasinal turbidites in the axial regions, suggests that considerable redistribution of lime mud is taking place (Busby, 1962; Gibson & Schlee, 1967). Episodic erosion of the channel floor has been invoked to explain the preservation of the steep, V-shaped profile near the mouth of the basin (Andrews et al., 1970; Schlager et al., 1976; Schlager & Chermak, 1979)(Fig.5.12).

Patterns of sedimentation in the Bahama channels thus show a number of features in common with those proposed for the Lagonegro Basin during the Late Triassic. As well as demonstrating that the carbonate platforms may indeed have been the major source of lime mud, they also underline the importance of intrabasinal redeposition in transferring this material from the flanks to the axial regions. It is interesting to note that average rates of accumulation, excluding the contribution made by calciturbidites, in the Tongue of the Ocean (15-30 m/MY), are of much the same order as those of the Sirino Formation (Busby, 1962). The absence of bank-derived calciturbidutes, however, constitutes a major limitation in putting an actualistic interpretation upon these Upper Triassic limestones.

CHAPTER SIX

RENEWED EXTENSIONAL TECTONICS AND BATHYAL DEPOSITION OF THE JURASSIC AND CRETACEOUS - THE LAGONEGRO AND BRUSCO FORMATIONS

6.1 Introduction

The Triassic-Jurassic boundary saw a transition from predominantly calcareous to siliceous deposition in the Lagonegro Basin that continued until the Early Cretaceous. Contrasting patterns of sedimentation developed in the Lagonegro units probably imposed by extensional tectonics. Unit I comprising a relatively monotonous succession of siliceous pelagic sediments whilst Unit II bears the record of abundant redeposition, of both intra- and extra-formational origin, in the form of radiolarites and calciturbidites respectively. Coeval calciclastic deposition in a small, and at the times anoxic, marginal basin is recorded in the formations of the Monte Foraporta Unit (See Chapter 7). In contrast to other basinal sequences of the Tethyan region, pelagic carbonate sedimentation was not re-established at the end of the Jurassic, and the Brusco Formation is characterised by dark grey and black terrigenous shales and allodapic limestones that accumulated below the Calcite Compensation Depth.

The sedimentology of the Lagonegro and Brusco Formations are described and interpreted in Sections 6.2 and 6.3 respectively. This is followed by a discussion of the Jurassic and Cretaceous evolution of the basin in terms of its palaeooceanography and palaeobathymetry.

6.2 Sedimentology of the Lagonegro Formation

The Lagonegro Formation is chiefly constituted by siliceous mudstones and cherts, with minor marls, marly limestone, calcilutites and

terrigenous mudstones; calcarenites and calcirudites of turbidite origin are also present which achieve maximum development in the northern part of Unit II.

The facies of Unit I show a remarkable uniformity throughout the area and the Cararuncedde and Pietra Members, composed dominantly of siliceous mudstones and radiolarian cherts respectively, each preserve a record of steady pelagic deposition during the Jurassic. Facies transitions in Unit II, on the other hand, are common both vertically and laterally. Beds of siliceous mudstone and radiolarian chert alternate in varying proportions at all levels, but an overall increase in the abundance of radiolarian cherts can be perceived through the succession. Superimposed upon this background of siliceous pelagic sedimentation were periodic incursions of shallow-water calciclastic material that seem to have occurred with most notable frequency and effect in the La Ralla Sections; they, nonetheless, affected the entire basin to some extent. Emphasising the contrast in sedimentary patterns in the two units is a considerable difference in thickness (Chapter 2). It should therefore be borne in mind that although identical facies occur in both units, their relationship to each other, and the extent to which they are developed, are both highly variable.

6.2.1 The Facies

6.2.1.1. The Siliceous Mudstone Facies

i) Description

Red and green mudstones, that are more or less siliceous, particularly characterise the lower parts of the formation. They may be interbedded with other facies or may constitute monotonous packets, a feature largely unique to Unit I.

In the Lagonegro Section, thick packets of red or green siliceous mudstone dominate the Cararuncedde Member (Pl.6.1). At the base they alternate with sequences of highly silicified calcilutite, but at the top they are in sharp contact with the green and grey cherts of the Pietra Member (Pl.6.2). These structureless deposits are either massive or cleaved parallel or sub-parallel to bedding. Red varieties show green reduction spots and surfaces along joints, but colour changes are not abundant through the section as a whole, and thick packets of red, and less commonly green, sediments predominate.

In the Torrente Bitonto Section, siliceous mudstones are present almost throughout the formation, but they rarely form distinct packets and are interbedded with radiolarian cherts, vitreous cherts and silicified calciturbidites (Pl.6.3). The mudstones may be either fissile or massive, and in the latter case individual beds may be recognised. They are predominantly red in colour but isolated green beds may also be found. Near the top of the formation, siliceous mudstones may alternate rhythmically with radiolarian cherts and are commonly reduced to shaley partings between radiolarite beds (Pl.6.4 & 6.6). This is a common aspect of chert sequences in the Tethyan Jurassic and Triassic (Garrison & Fischer, 1969; Bernoulli & Jenkins, 1974; Nisbet & Price, 1974).

In thin section, the red siliceous mudstones are almost opaque and comprise a fine-grained admixture of silica, clay minerals and hematite. An extremely fine network of anastomosing hematite- and clay-rich stringers separating clay and silica beds can be distinguished. Green varieties are less opaque and comprise a more obvious fine-grained quartz component that is mixed with green clay

minerals and opaque fleck, probably of pyrite. Chalcedonic radiolarian moulds are rarely found and are usually flattened parallel to bedding, preserving no primary structure. Mineralogically, the red siliceous mudstones comprise dominant quartz with minor hematite and a clay-mineral assemblage consisting mainly of illite with some kaolinite and chlorite (Fig.6.1); some calcite is locally present. The green varieties are similar but contain a higher proportion of kaolinite and chlorite, and lack hematite.

ii) Interpretation

Red and green siliceous mudstones are interpreted as the products of normal pelagic sedimentation below the calcite compensation depth and probably represent the lithified equivalents of the red clays currently accumulating at such depths in modern oceans (Berger, 1974; Jenkyns, 1978). The monotonous packets of Unit I, which lack sedimentary structures, are though to have accumulated grain by grain from pelagic suspension. Rhythmicity of bedding and gradational transitions into underlying radiolarite beds, on the other hand, suggest that a large proportion of the siliceous mudstones in Unit II represent the fine-grained components of siliceous turbidites. A similar origin has been suggested for inter-radiolarite mudstones from a number of other chert sequences (Nisbet & Price, 1974; Price, 1977D; Leggett, 1978; Barrett, 1979).

The colour differences are considered to reflect post-depositional but early diagenetic Eh conditions. Hematite, which imparts a red colour, developed in an oxidising diagenetic environment, possible engendered by slow rates of deposition, whereas reducing conditions fostered the development of pyrite and chlorite minerals, producing a green colouration. Alternating packets of red and green mudstones may

reflect fluctuations in input of organic material or variations in sedimentary rate. The relationship of reduction spots and surfaces to bed partings and veins, however, suggests that in some cases reducing conditions were realised at a late stage of diagenesis. Compaction and dissolution are evinced by the flattening of radiolarian moulds and the development of anastomosing, clay- and hematite-rich stringers.

The clay minerals are probably terrigenous and are thought to be wind-derived chemical weathering products of continental terrains, possibly located in the African continent (Griffen & Goldberg, 1963). Although chlorite is often held to be an indicator of submarine alteration of basalts or basaltic volcanism, the absence of smectites or independent evidence for Jurassic volcanism in the Lagonegro Zone do not support a volcanic origin (Nisbet & Price, 1974; Barrett, 1979; c.f. Scheidegger & Stakes, 1977). Hematite, which is produced by dehydration of goethite, probably originated as iron-manganese oxide/hydroxides such as constitute the majority of the x-ray amorphous component of modern oceanic red clays (Berger, 1974).

The paucity of Radiolaria in these deposits is in marked contrast to their abundance in the Radiolarian Chert Facies, and the dominance of radiolarian-free sediments throughout the Cararuncedde Member of Unit I is especially puzzling. It has been shown, however, that in areas below the CCD, considerable dissolution of biogenous opal occurs close to the sediment-water interface, and only about 10% of the total deposited is actually preserved in the sediment (Hurd, 1973; Calvert, 1974; Wollast, 1974; Johnson, 1976). Depositional rates in the Lagonegro Basin during the Early Jurassic were probably sufficiently low that, under normal conditions, the majority of radiolarians would have been dissolved before they could be buried;

only when concentrated by redepositional processes did they survive, in a manner analogous to the preservation of calciturbidites in sequences deposited below the CCD. The radiolarian-free siliceous mudstones of Unit I, therefore, evince minimal resedimentation in those parts of the basin now represented by that Unit.

6.2.1.2. The Marl Facies

i) Description

Thin beds of pink and red marl are interspersed with the siliceous mudstones and radiolarites of the Torrente Bitonto Section at several levels. They are either parallel-laminated or structureless and stongly bioturbated, and are consituted by a fine-grained mixture of quartz, calcite and clay minerals. Radiolaria are rare and calcitised, but abundant angular flecks of silica may represent their macerated remains; radiolarian spines are plentiful. Thin, filamentous shell fragments, probably of pelagic bivalves (e.g. Bositra) are also rarely present; they are invariably oriented parallel to bedding and are responsible for imparting the parallel lamination.

In the Pignola-Abriola Section II, grey marls and marly limestones are present in all but the highest part of the succession. They are medium-bedded, structureless and intensely bioturbated.

ii) Interpretation

The thin, laminated marls of the Torrente Bitonto Section were probably redeposited from shallower parts of the basin that lay close to or above the CCD. Their appearance near the top of the formation provides the first indication that the basin floor was becoming shallower in relation to calcite dissolution levels towards the end of the Jurassic, probably as a result of the depression of these levels at about that time (Bosellini & Winterer, 1975). Deposition by

currents is witnessed by the parallel-lamination and fracturing of bioclasts; recognition of reworking in these beds, which also do not contain abundant Radiolaria, suggests that some of the co-existing siliceous mudstones may also be allochthonous.

The marls and marly limestones of the Pignola-Abriola Section II, on the other hand, may reflect *in situ* deposition of calcareous material, suggesting that the basin floor in that area had not subsided beneath the CCD. Although stratigraphic control is lacking, their presence over 100 m above the base of the formation implies that sedimentary rates were sufficiently high to more than keep pace with subsidence. The dominance of distal calciturbidites, which may have been responsible for maintaining high rates of sedimentation, in the underlying part of the sequence lends some weight to this idea.

6.2.1.3 The Radiolarian Chert Facies

i) Description

Banded cherts dominate the upper part of the formation in both units but may also be common near the base where they are interbedded with siliceous mudstones. It is possible to distinguish two groups; those that are associated and commonly rhythmically interbedded with siliceous mudstones, and those which constitute homolithic packets (c.f. Price, 1977b). The first group is dominant in Unit II, whereas more monotonous cherts particularly characterise the Pietra Member of Unit I.

a) Radiolarian cherts associated with siliceous mudstones

In Unit I, the succession of red and green siliceous mudstones of the Cararuncedde member is punctuated at irregular intervals by rare chert beds up to 5 cm thick. These radiolarites can be distinguished

by their massive appearance, conchoidal fracture and paler colour on weathered surfaces. They contain abundant Radiolaria, but, with the exception of indistinct parallel lamination, are structureless. They are invariably red but may contain green reduction spots and surfaces similar to those of the host mudstones. They show uniform thickness, and both upper and lower contacts with the mudstones are sharp.

Similar beds, also interbedded with siliceous mudstones, are common in the Torrente Bitonto Section of Unit II and become increasingly abundant up the section. In the upper part, they dominate the succession and may be separated only by thin shale partings (Pls. 6.4 & 6.6). Most commonly, however, they alternate rhythmically with more thinly bedded siliceous mudstones that can be distinguished by their darker colour and smooth texture (Pl.6.3). The beds show dramatic lateral thickness changes and may swell by as much as five times over as little as 5 cm; beneath these swollen cherts, the siliceous mudstones appear to be compressed rather than truncated. Rarely, the cherts may contain fewer radiolarians at the top and, in some cases, show pronounced parallel lamination (Pl.6.4). They may also show gradational transitions at the top to siliceous mudstone, but contacts are generally sharp. Fig.6.2 summarises the features of these radiolarite beds.

Similar successions may be found throughout the formation in Unit II and are invariably most common towards the top. In the Pignola-Abriola I and Pignola-Potenza Sections, packets of beds several metres thick are slumped (Pl.6.5). Slump folds are insufficient in number to determine palaeoslope directions, but axes are oriented northeast-southwest and overfolding is to the northwest.

The radiolarites are composed almost entirely of radiolarian ghosts, but hematite- and clay-rich stringers may emphasise the primary parallel-lamination (Pl.6.11). Grading has not been observed in thin section, either in beds as a whole or within individual laminae, the Radiolaria being of almost uniform size, and distributed with equal density throughout. Contacts with the siliceous mudstones may be highlighted by corrosion of siliceous radiolarian spheres. Preservation of radiolarian tests is generally poor, however, and they may be flattened parallel to bedding, or in rare instances, to cleavage. Only rarely is their structure preserved, and their moulds are filled either by chalcedony or, more rarely, by ferroan sparry calcite (Pl.6.12); in the more clay-rich beds, some moulds contain geopetal fillings of matrix material. These siliceous spheres are set in a matrix of microcrystalline quartz or dark, hematite-rich siliceous mudstone, and the latter is pervaded by anastomosing opaque stringers similar to those observed in the siliceous mudstones. Corrosion of radiolarian ghosts is commonly evinced along these seams, and contacts between them may be pressure-solved. The more siliceous radiolarite beds are composed almost entirely of microcrystalline quartz, and the outlines of Radiolaria are preserved only by diffuse spherical dustings of clay minerals.

Thin, dark bands about 2 mm thick, that are composed largely of hematite and quartz, have also been observed in the middle of some beds, while others are commonly pervaded by rhombic crystals of calcite and dolomite about 200μ in diameter. The latter are preferentially situated along vertical chalcedonic veins and are often corroded and partially replaced by chalcedony. They appear to have grown epitaxially upon calcite which originally filled the veins but was later converted to chalcedony.

X-ray diffraction analysis of the red cherts reveals dominant quartz, minor calcite and hematite, plus kaolinite (Fig.6.1.). Attempts to identify the clay mineral assemblage more precisely by separation techniques proved unsuccessful.

b) Radiolarian cherts not associated with siliceous mudstones.

The Pietra Member of Unit I is chiefly constituted by finely laminated and extremely siliceous cherts in beds up to 20 cm thick (Pl.6.7). The apparent rhythmicity of the beds at outcrop is a superficial feature, and adjacent 'beds' are of identical composition and of random thickness; contacts between them are true partings, lacking thin shale intercalations. Weathered faces are pale green, brown or cream coloured, but fresh surfaces are green, or grey near the margins of the vitreous chert beds that punctuate the succession (Pls. 6.8, 6.19 & 6.20). They are extremely hard, and fracture conchoidally or, more rarely, along the horizontal laminae. With the exception of the lamination, however, these cherts are structureless and monotonous; moreover, with the exception of those in the Sasso di Castalda Section II, which may also be red, they show a remarkable conformity, both in appearance and thickness, throughout the unit.

These deposits also contain abundant Radiolaria that are even visible in hand specimen. These organisms are preserved as moulds which are more or less closely packed, and are filled by chalcedony, either in a granular form or as sheaf-like bundles; in some cases they preserve traces of primary structure (Pl.6.14 & 6.15). The matrix is constituted by microcrystalline quartz and clay minerals, and anastomosing, clay-rich stringers are responsible for imparting the horizontal lamination; radiolarian moulds may be strongly

flattened in the vicinity of these seams. In some beds, radiolarians are preserved only as indistinct, spherical or ellipsoidal phantoms composed of clay-free microcrystalline quartz. The green colour derives from the disseminated clay flakes, probably of illite or chlorite (Fig.6.1).

Packets consisting almost uniquely of cherts are also present in the upper parts of sections in Unit II, but they are of limited thickness and do not constitute the dominant lithology. Beds in one such packet at the top of the Pignola-Abriola Section I differ from those in Unit I in that they are more massive and lack fine lamination. Beds may be up to 20 cm thick, but seldom exceed 5 cm, and commonly show colour banding (Pl.6.9). They are interbedded with packets of chert and siliceous mudstone, vitreous cherts and silicified calciturbidites. Bioturbation is evinced in some of these Unit II cherts by green and yellow burrows, both within beds and at their base (Pl.6.10). In thin section, they differ from their counterparts in Unit I only in their more homogeneous texture and, in the red varieties, a higher proportion of hematite in the matrix. Preservation of Radiolaria is variable, but is commonly at its best in the more clay-rich red cherts (Pl.6.13).

Studies of chert ultrastructure under the scanning-electron microscope reveals that the microcrystalline variety is composed of more or less euhedral crystals about 1μ across that are mixed with hematite and clay minerals in the groundmass to form an amorphous and featureless mass (Pl.6.16). Well-preserved radiolarian tests are composed of similar crystals that are fused together into a faithful replica of the original trellis structure of the shell. The interiors of the moulds, however, contain chalcedony that commonly

displays a radial fabric in the centre and a more amorphous, fine-grained texture near the margins (Pl.6.12).

ii) Interpretation

The rarity of grading, either in the size, or density of packing of bioclasts, mitigates against deposition from turbidity currents, an origin that has nonetheless often been advanced for banded cherts of the Tethyan region (Garrison, 1974; Nisbet & Price, 1974; Price, 1977b; Barrett, 1979). Lack of grading, however, is characteristic of these supposedly redeposited radiolarites, and several features of the rhythmically bedded cherts of the Lagonegro Zone are consistent with their being turbidites; these include parallel-lamination, gradational contacts with overlying mudstones, and the rarely observed increase in the proportion of matrix near the tops of some beds. Moreoever, settling velocities of radiolarian tests ranging in size by a factor of five probably vary by as little as 1 cm/sec, implying that grading would not be readily achieved by deposition from a radiolarian-rich turbidity current (Lorsong, in Barrett, 1979). Indeed, the most dramatic grading in redepositioned radiolarites of the Ligurian Apennines is shown by the concentration of incorporated clastic fragments near the bases of beds (Garrison, 1974; Barrett, 1979). Although the textural evidence is equivocal, therefore, the rhythmically bedded cherts which particularly characterise the Lagonegro Formation in Unit II are believed to have been deposited from low-density turbid flows; their scarcity in Unit I provides further evidence for passive conditions of sedimentation in those parts of the basin represented by that unit.

An additional testimony to an allochthonous origin is given by the presence of calcitised Radiolaria in some of these beds. The associated mudstones are lime-free and it is believed that the basin floor

during most of the Jurassic was below the CCD. The survival of $CaCO_3$ into the burial stage implies very rapid deposition in order to avoid dissolution, either during sedimentation or at the sediment-water interface. Such a process may have been achieved by deposition of turbidites whose source lay close to the CCD. The presence of scattered calcite and dolomite rhombs in some of the purer cherts suggests that they too may have contained a primary calcareous component (c.f. Dietrich et al, 1963; Jacka, 1974; Price, 1977b).

Despite the similarity of the Lagonegro cherts to those encountered elsewhere in the Tethyan Jurassic, an altogether different depositional environment must be envisaged to that proposed for the supposedly 'oceanic' occurrencies of the Ligurian Apennines and Othrys, Greece. Association of chert sequences with ophiolites and basic volcanics has led several authors to attribute redeposition in these areas to sediment ponding on the flanks of active, mid-oceanic ridges (Fig.6.3)(Garrison, 1974; Nisbet & Price, 1974; Price, 1977b; Barrett, 1979). No such relationship pertains in the Lagonegro Zone, so a similar model is clearly inappropriate. Evidence of extensional tectonic activity at the end of the Triassic suggests that differential relief may once again have characterised those parts of the basin floor now represented by Unit II. Areas of positive relief would have accumulated siliceous pelagic sediments that, particularly on the slopes, would have been unstable. Tectonic disturbances, or simply slope oversteepening, may have triggered turbidity currents which transported the sediment to the deeper parts of the basin. As one would thus expect, therefore, sections in Unit II, in which radiolarian chert and siliceous mudstone alternations are most abundant, are much thicker than those of Unit I (See Chapter 2). The relatively monotonous aspect of the siliceous mudstones and cherts

of the formation of the latter unit is more evocative of depostiion in a stable environment of low relief in which redeposition was only of minor importance. The presence of sequences comprising alternations of siliceous mudstones and cherts in a unit of non-oceanic affinity suggests that caution should be exercised in advancing a mid-ocean ridge source for redeposited siliceous deposits simply on the basis of their association with ophiolites of dubious age in adjacent units (c.f. Nisbet & Price, 1974). In the Early Mesozoic in particular, when the CCD was shallower than it is today, deep-water continental margins may have been accumulating siliceous oozes in abundance, and source areas for redeposition may have been commensurately numerous.

The contrasting, monotonous aspect of structureless chert successions that are <u>not</u> associated with siliceous mudstones suggests that they have not been redeposited but represent the lithified equivalents of radiolarian oozes that accumulated below the CCD as a steady pelagic rain. The high radiolarian content possibly attests to gentle reworking of radiolarian ooze and winnowing of the clay fraction by bottom currents, but the absence of rhythmic bedding argues against major resedimentation. Furthermore, the anastomosing clay-rich stringers, which are responsible for producing the lamination, are the products of dissolution, as is evinced by the corrosion of radiolarian silica spheres along them. The lamination is thus a diagenetic feature, and was probably caused by compaction.

The presence of green burrows in bioturbated red cherts suggests that organic content was a major influence on colouration, and the predominantly green colour of the cherts of the Pietra Member in Unit I probably reflects the large contribution of organic material made by the abundant Radiolaria present; reducing conditions below the sediment-water interface may also have been maintained by high

rates of sedimentation. The fact that the redeposited radiolarites of Unit II are invariably red probably indicates that they were derived from sources where their organic component had already been oxidised.

An overall pattern of abundant redeposition in Unit II and steady pelagic sedimentation in Unit I therefore emerges from the cherts and siliceous mudstones of the formation. Before discussing the remaining, dominantly calciclastic facies, whose distribution amplifies this pattern, it is pertinent to consider briefly the diagenetic history of the siliceous deposits.

Understanding of the diagenesis of oceanic sediments has been greatly enhanced during the last decade by the plethora of new data made available by the Deep Sea Drilling Project. It is now generally considered that cherts develop by diagenetic conversion of biogenous opal, usually in the form of radiolarian and diatom tests, via a disordered-cristobalite intermediary structure, to chalcedonic quartz (Heath & Moberly, 1971; Berger & von Rad, 1972; Weaver & Wise, 1972; Robertson, 1977; for review, see Wise & Weaver, 1974). Paradoxically, bedded radiolarian cherts like those of the Tethyan Late Jurassic, as opposed to more nodular varieties, have yet to be encountered in the modern oceans. The mechanism for producing thick successions of bedded chert remains elusive, therefore, but it seems probable that these deposits underwent a similar 'maturation' sequence, even though precursor silica minerals are seldom preserved (Robertson, 1977; Barrett, 1979).

Contrasting preservation of radiolarian structure in matrix-rich and matrix-poor cherts of the Lagonegro Formation suggests that their diagenetic histories may not be identical. In the case of the purer cherts, dissolution of opaline radiolarian tests and precipitation of opal-CT lepishpheres in resultant voids probably produced a

homogeneous procellanite which was subsequently converted to the presently featureless microcrystalline quartz (c.f. Heath & Moberly, 1971; Wise & Weaver, 1974; Kastner et al., 1977; Robertson, 1977; Hein et al., 1978). Radiolarian moulds are therefore difficult to distinguish, and preserve none of the primary structure of the tests. Replication of structure by chalcedonic microgranules in some of the red, more matrix-rich cherts, however, argues against complete dissolution and reprecipitation (c.f. Thurston, 1972). Moreover, radiolarian tests composed of fused granules of disordered cristobalite about 5μ across, that are thought to represent diagenetically aged biogenous opal, have also been observed in younger cherts from Cyprus (Robertson, 1977). A process of internal reorganisation rather than solution-reprecipitation is documented in these cherts, and it is suggested that the well-preserved chalcedonic radiolarian moulds of the Lagonegro samples may have passed through a similar state. Clay minerals, which tend to act as a sink for the cationic nuclei upon which opal-CT probably precipitates, is known to have an inhibitory effect upon lepisphere formation (Kastner, et al., 1977). Since preservation of radiolarian test structure is limited to clay-rich red cherts, clay minerals are held responsible for slowing the diagenetic process sufficiently to allow reorganisation of biogenous opal-A into disordered cristobalite microgranules and finally microcrystalline quartz.

The characteristic radial fabrics of many of these chalcedonic radiolarian moulds may also bear witness to an opal-CT precursor. Heath & Moberly (1971) suggested that this texture is produced by solid-state inversion, but chalcedony produced experimentally from opal-CT by dissolution and reprecipitation also takes the form of fibrous bundles (Stein & Kirkpatrick, 1976). The sheaf-like fabrics of

the radiolarian moulds may thus have formed during late-stage precipitation of quartz. The abundance of clay- and hematite-rich stringers in these cherts suggests that dissolution of opal-CT during compaction may have generated a sufficiency of free silica to cement up any remaining pores. Moreover, the importance of redistribution of silica is clearly evinced by the pinching and swelling of radiolarite beds, but deformation of siliceous mudstones around the swelling suggests that these may be early, pre-compactional features. Much of the redistribution may therefore have occurred during the earlier conversion of opal-A to opal-CT.

Diagenesis of the cherts of the Lagonegro Formation can thus be broadly explained in terms of the quartz maturation theory, even though no precursor silica minerals are preserved. Following the philosophy of Wise & Weaver (1974), the abundance of radiolarian moulds in these rocks suggest that biogenous opal provided the source of silica for chertification and obviates the need to invoke more exotic alternatives.

6.2.1.4. The Vitreous Chert Facies

i) Description

The green radiolarian cherts of the Pietra Member are interbedded with bands of vitreous grey or milk-white chert up to 35 cm thick that preserve parallel- and low-angle cross laminated intervals, largely near the top, and may be erosive at the base (Pls. 6.17, 6.19 & 6.20). These beds possess opaque, pale green or white selvedges which contain Radiolaria visible in hand specimen, but the vitreous portions are almost featureless. Weathered surfaces are commonly a dirty brown or black and may bear an Fe/Mn patina (Pl.6.18). Bed thicknesses are generally regular but there is an overall increase in the proportion of vitreous cherts towards the top of the formation.

Similar beds may be found interbedded with the cherts and siliceous mudstones of the Pignola-Abriola Section I. Although commonly red and green in colour, they possess identical sedimentary structures to those of the Pietra Member and, in addition, the top few centimetres are in some cases convolutely laminated (Pl.6.21 & 6.24). Many beds are brown or black on weathered surfaces, are notably fragile and are partially calcareous. Rarely they appear to be channellised and may swell at the base by as much as three times over a distance of only 20 cm (Pl.6.22); underlying beds, however, are compressed rather than truncated. As in Unit I, these cherts are only found near the top of the formation. In the Torrente Bitonto Section, similar beds of less vitreous aspect are abundant at the same level (Pl.6.23). The latter, which are interbedded with siliceous mudstones and radiolarian cherts, are also commonly calcareous.

The vitreous cherts of the Pietra Member are composed almost entirely of microcrystalline quartz, although rare clay minerals and minute opaque grains may impart a feint lamination (Pls. 6.14 & 6.15). Undulose clay-rich stringers resembling stylotites are also found. In the less-vitreous selvedges, however, closely-packed chalcedonic ghosts of Radiolaria can be recognised that may themselves be organised into laminae along with the clay minerals. The transition to the vitreous interval is gradational and is marked by a progressive decrease in the proportion of chalcedonised Radiolaria in favour of spheres of microcrystalline quartz. The cherts which show brown or black weathering are distinguished in thin section by the presence of euhedral rhombs of calcite and dolomite about 100μ in diameter that are zoned and commonly opaque in the core (Pl.6.25). Staining reveals that the clear rhombs are dolomite whereas those with opaque cores are calcite; the latter are particularly abundant in the vicinity

of vertical quartz veinlets. Although these rhombs are organised into laminae, they show no evidence of abrasion and are commonly intergrown.

The vitreous cherts of the Pignola-Abriola I and Pignola-Potenza Sections resemble their counterparts in the lower unit, but in addition to rhombs of calcite and dolomite, laminae may also contain calcitised Radiolaria whose structure is well preserved and which are completely surrounded by microcrystalline quartz; they are commonly filled by single calcite crystals. Chalcedonic radiolarian ghosts are abundant in the groundmass, the quartz taking the form of sheaf-like bundles of acicular crystals, or a radial, void-filling fabric. Parts of some beds may even comprise an intergrown mosaic of chalcedony and neomorphic sparry calcite, the latter commonly forming laminae and surrounding scattered dolomite rhombs (Pl.6.26). These laminated calcareous zones may pass laterally into purer cherts in whichthe laminae are preserved only by dolomite rhombs (Pl.6.27). Comparable beds in the Torrente Bitonto Section show identical features.

ii) Interpretation

The parallel-, cross- and convolute-lamination and low clay content of the vitreous cherts attests to reworking of sediment, probably by turbidity currents. Moreover, the rhombs of calcite and dolomite, which pass laterally into laminae of neomorphic spar, are testimony to a primary carbonate component that was dissolved and reprecipitated during silicification, rhombic inclusions of carbonate being a common feature of replacement cherts in limestone successions (Dietrich et al., 1963; Mišík, 1973; Jacka, 1974; Price, 1977b). The lack of clay minerals and presence of seams

reminiscent of stylotites suggests that these beds were initially composed uniquely of carbonate material and Radiolaria. Although the dolomite rhombs are organised into laminae, they are not considered to be detrital because of their sharp outlines and intergrown texture. It is more likely that they are the relics of primary carbonate-rich laminae that are now otherwise preserved as neomorphic sparry calcite in some beds of the Pignola-Abriola I and Torrente Bitonto Sections; the more quartzose laminae may have been composed entirely of radiolarian tests. The calcite rhombs are probably dedolomite, the opaque iron oxides representing excess iron expelled during conversion of the ferroan dolomite to calcite (c.f. Evamy , 1967; Shearman et al., 1961). Although outlines of carbonate grains can seldom be recognised, the calcitised radiolarian moulds may have been filled with calcite prior to redeposition and have subsequently resisted silicification. The majority of the carbonate in these beds, however, was probably in the form of lime mud.

It is suggested that the vitreous cherts are the silicified equivalents of thin intrabasinal calciturbidites that originated above the CCD. Because of their high biogenous-opal and calcite component, they were silicified more rapidly than the host radiolarian oozes and the swelling of the bases of some beds testifies to migration of silica from the underlying clay-rich deposits. Kastner et al. (1977) have shown that the transformation of biogenous opal-A to opal-CT is most readily achieved in the presence of carbonate which can supply the OH^- nuclei necessary for precipitation of the disordered cristobalite intermediary. Concomitant reduction in alkalinity enhances dissolution of the host limestone, and a build-up of Ca^{++}, Mg^{++}

and HCO_3^- concentration in solution might be expected to encourage precipitation of calcite and dolomite; rhombic carbonate crystals have actually been documented from the margins of replacement cherts drilled in modern oceans (Wise & Weaver, 1974). Diagenetic replacement of radiolarian tests is thus likely to have been achieved most rapidly in the heart of the beds, well away from the inhibiting influence of clay minerals, and probably explains the complete destruction of primary fabrics. In the selvedges, however, production of opal-CT occurred much more slowly and radiolarian moulds were incompletely filled with lepispheres; late stage filling by chalcedony ensured their preservation.

Preferential silicification of calciturbidites interbedded with pelagic limestones and chalks has also been observed in the Late Jurrasic and Early Cretaceous Oberalm Beds of Austria, and in the Upper Palaeocene and Lower Eocene sediments overlying the Troodos Massif of Cyprus (Garrison, 1967; Robertson, 1977). Concentration of Radiolaria and high initial porosities at the bases of these beds are considered to have been responsible for enhancing the development of chert in the calciturbidites relative to the completely unsilicified, interbedded chalks and limestones.

The abundance of siliceous micro-organism phantoms suggests that these calciturbidites were generated on the flanks of the basin, above the CCD, rather than in the carbonate platforms. Although no Late Jurassic pelagic carbonates have been recognised in the Lagonegro Zone, it is plausible that depression of calcite dissolution levels at that time had restored pelagic carbonate deposition to the lower slopes of platform margins which, until then had been accumulating

siliceous sediments. This may explain why the vitreous cherts appear only in the upper parts of the formation in both units.

6.2.1.5. The Calcilutite Facies

i) Description

At the base of the formation in both units, grey, white or cream-coloured calcilutites are commonly interbedded with shales and siliceous mudstones. In Unit I, they occur as single beds or may constitute discrete packets, and are invariably parallel-laminated. Beds seldom exceed 10 cm in thickness, and in the packeted successions of the Lagonegro Section, they average about 3 cm. Chert nodules are common, although the packeted beds are incipiently silicified throughout and only the outlines of chert nodules can be distinguished. In Unit II, the calcilutites show similar features, but packets of thinly bedded lime-mudstones are absent. In the Pignola-Abriola Section II, both laminated and massive calcilutites may grade down into calciturbidites, but the calcilutites of the Torrente Bitonto Section, on the other hand, occur only as isolated, parallel-laminated beds that show no grading.

In thin section, the calcilutites are seen to comprise radiolarian biomicrites containing abundant calcitised Radiolaria and filamentous shell fragments set in a micritic matrix. Clay-minerals are more or less abundant, and the bioclasts, together with micritic intraclasts, may be organised into laminae. The incipiently chertified beds of Unit I contain fewer Radiolaria and are pervaded by microcrystalline quartz that forms a mosaic with the micrite.

Some of the beds in the Pignola-Abriola Section show a biparite

fabric, the lower part consisting a barren, light-coloured micrite, and the upper part, a darker radiolarian biomicrite containing ammonite moulds with geopetal micrite fills. Burrows penetrate down from the tops of beds through the biomicrite and a short distance into the basal layer, piping down biomicrite from above.

ii) Interpretation

The association of calcilutites with calciturbidite beds in the Pignola-Abriola Section II suggests that they may have a common origin, but for the majority in other sections, no such relationship pertains. Distinction of fine-grained calcareous turbidites from pelagic lime-mudstones in successions deposited above the CCD is difficult and may rely solely upon faunal criteria that cannot be applied to these calcilutites (Hesse & Butt, 1976). It is thus difficult to establish whether the limestones are redeposited or represent in-situ accumulations of pelagic ooze, particularly in the case of the more massive beds. The preservation of ammonites, whose aragonitic phragmacones must have been buried sufficiently quickly to avoid dissolution, provides some evidence of rapid sedimentation that is consistent with deposition from turbidity currents. Moreover, the limited penetration of burrows suggests that only limited time was available for bioturbation prior to deposition of the succeeding bed, also a common feature of turbidite sequences (Seilacher, 1962; Scholle, 1971a; Hesse,1975; Hesse & Butt, 1976). Since lamination is commonly developed at the bases of beds, laminated and structureless intervals may be interpreted as the D and E intervals respectively of a typical Bouma sequence (Bouma, 1962); examples of lime mudstone turbidites containing these intervals are also documented from several Cretaceous successions in the Alps (Scholle, 1971b;Hesse, 1975).

Furthermore, interbedding of the calcilutites with shales and siliceous mudstones, which are thought to have been deposited below the CCD, also testifies to an allochthonous origin. Although the uniquely pelagic nature of the faunas implies derivation from within the basin, the presence of ammonites suggests that the carbonate mud had a relatively shallow-water provenance above the ACD. Their presence solely at the base of the formation therefore suggests that their source areas lay in parts of the basin that were still above the CCD during the Early Jurassic, probably high on the platform slopes. On the other hand, the possibility that they represent distal flows that were initiated in the platforms themselves, and that the pelagic organisms were entrained during transport, cannot be ruled out.

6.2.1.6 The Calciturbidite Facies

i) Description

Graded carbonate breccias which embody features typical of turbidites are present in both units, but they achieve maximum development in the La Ralla Section of Unit II. In this section, they occur in beds up to 6 m thick that show more or less complete Bouma sequences and rarely possess load-modified flute marks on their bases (Pls. 6.28, 6.29 and frontispiece). The thickest beds comprise Bouma A, B and C or A and B intervals and grade from granules to fine sand within a single unit; some units however, may contain several amalgamated graded intervals. Although bed thicknesses may vary laterally erosional bases have not been observed. The calciturbidites are interbedded with shales, siliceous mudstones and cherts, but thin, laminated calcilutite and calcisiltite beds are also rarely present. In the lower part

of the succession, the calciturbidites occur in packets and may be amalgamated, but bed thickness and grainsize variations follow no regular pattern from bed to bed (Pl.6.30). In the upper parts of the section, beds characterisitically occur in isolation within the cherts that dominate the succession (Pls. 6.31 & 6.32).

Many calciturbidite beds are silicified, either partially in the form of nodules and bands along the top and bottom, or completely; in the lower part of the formation, they may also be dolomitised (Chapter 8). These cherts commonly preserve primary lamination, although swelling may be evinced by distension of laminae, and by the bulbous protrusion of nodules on the bases of some beds (Pl. 6.33 & 6.34).

The constituent grains fall into two categories; the coarser varieties comprise lithic fragments of typical shallow-water carbonate lithologies, while the sand-sized grains are composed of a wide variety of bioclasts, peloids and ooids (Pl.6.35 & 6.36); fragments of siliceous mudstone may also be incorporated into the bases of some beds. The lithic clasts are composed of micrites, biopelsparites and oosparites, and the constituent grains are all highly micritised; fragments of molluscs, brachiopods, echinoderms, Foraminifera, sceleractinian corals, Bryozoa and green and red algae can all be recognised amongst the allochems. In addition to the neritic lithologies, some clasts are constituted by radiolarian/pelagic bivalve biomicrites. Within the clasts, grains are generally surrounded by early, isopachous rim cement, and remaining voids are filled by sparry calcite; alternatively, they may have been bound by encrusting organisms into grapestones. The sand-sized grains that characterise the finer-grained beds are also more or less

micritised and chiefly comprise ooids; the bioclast assemblage mirrors that of the lithic fragments.

The calciturbidites are generally cemented by either isopachous or radiaxial-fibrous calcite, and late-stage porosity is eliminated by sparry calcite or chalcedonic quartz; the tops and bottoms of some beds, however, are entirely cemented by chalcedony. Silicification may also affect the clasts themselves, which are commonly leached and replaced by bundles of fibrous chalcedonic quartz crystals. Even where silicification is complete, however, ghosts of the primary fabrics are generally preserved, the clasts as microcrystalline quartz and the cements as chalcedony (Pl.6.37 & 6.38). Furthermore, in beds that are otherwise dolomitised, only rarely are ghosts or relict dolomite rhombs preserved in the cherts, and primary grain fabrics can still be distinguished; smaller, and usually opaque, rhombs may be present, but these are also found in nodules from otherwise completely undolomitised calciturbidites.

Calciturbidites are also developed at the base of the Pignola-Abriola Section I, but here they are much thinner and fewer in number in comparison to those at La Ralla. They are up to 30 cm thick, are interbedded with laminated calcilutites, grade from sand at the base to silt and mud at the top, and show more or less complete Bouma sequences. The coarser fraction contains lithic clasts and allochems similar to those from the La Ralla calciturbidites, but calcitised Radiolaria are commonly found in the mudstone intervals. Higher in the section, as well as in the Torrente Bitonto Section, thinner calciturbidites, that are partially or completely silicified, are rarely intercalcated with siliceous mudstones and cherts.

Up to 20 cm thick, these beds preserve B, C and D Bouma divisions and commonly appear to be channellised, although underlying siliceous deposits are compressed rather than truncated. The coarsest grains are of neritic origin, but finer-grained beds contain an abundance of Radiolaria.

Carbonate microbreccias are also present in the Cararuncedde Member of Unit I, although they are rare in sections from the Sirino tectonic window. Red, green or grey in colour, they are generally structureless and ungraded. They comprise clasts of sparry calcite, radiolarian/pelagic bivalve biomicrite and a variety of lithic fragments containing spar-cemented ooids, unidentified peloids, forams and algae; echinoid spines have also been documented (Scandone, 1967b). These grains, many of which are recrystallised, are tightly packed, and contacts are pressure solved along thin solution seams. The matrix is micritic but contains a high content of clay and, in the case of the red microbreccia, hematite.

ii) Interpretation

The most distinctive features of the turbidites at La Ralla are their immense thickness and their occurrence within a succession principally constituted by siliceous pelagic sediments. In terms of their dimensions, grainsizes and sedimentary structures, they satisfy all of the criteria of proximality defined by Walker (1967). The validity of these features as indicators of distance from source is doubtful, however, and it is clear that under certain conditions 'proximal' calciturbidites may occur within sequences that are at a considerable distance from their point of origin; this is elegantly demonstrated by the punctuation of the deep-sea plain turbidite successions of the Marnoso-Arenacea Basin of the northern Apennines

by single turbidite beds that are as much as an order of magnitude thicker and more coarse grained than the ambient deposits (Parea & Ricci-Lucchi, 1975). Nonetheless, the relative scarcity of grading and poor organisation of the La Ralla turbidites implies that the depositing flows were immature and had not travelled a significant distance prior to deposition.

It is clear from the sporadic distribution of these turbidites through the section, and the absence of characteristic vertical sequences of beds thinning and fining, or thickening and coarsening upwards, that they cannot be interpreted in terms of a submarine fan model or well-established turbidite depositional system (c.f. Nelson & Nilson , 1974; Mutti, 1974; Ricci-Lucchi, 1975; Rupke, 1977). Furthermore, the paucity of thin-bedded carbonates in the section does not even suggest the proximity of a fan system, and the calciturbidites must, therefore, be interpreted as the products of discrete depositional events. The demonstrably deep-water origin of the host siliceous sediments, and the paucity of slumps, is evocative of a level, basin-plain environment in which slopes were negligible. Periodic incursions of high density turbidity currents upset this tranquil scene and deposited thick calciclastic beds sufficiently rapidly to evade dissolution. Paucity of palaeocurrent data precludes identification of their source direction, although rare flute marks denote a north-westerly provenance. In the absence of a fan system, however, it is unlikely that these flows had a point source, and the palaeocurrents most probably indicate the orientation of the axis of the basin (c.f. Ricci-Lucchi, 1975).

The 'proximal' aspect of these turbidites implies rapid deposition

of load once the flows reached the basin floor. Under normal conditions, the momentum of such large flows would have ensured that their load was spread out and deposited over a wide area, and it should be possible, therefore, to trace a lateral transition of bed thickness, grainsize and sedimentary structure downcurrent, such as can be observed in fan sequences with good, lateral control. The Pignola-Abriola Section I, less than 40 km away, contains relatively abundant calciturbidites, especially near the base, but they are much more fine-grained and thinly bedded than one would expect the lateral equivalents of the La Ralla turbidites to be. The depositing flows must, therefore, have decayed rapidly, suggesting that slopes were negligible or even reversed; this introduces the possibility that the basin was deeper at La Ralla than at Pignola and that the flows were being ponded in the north.

Ponding of turbidity currents has been mooted by several authors to account for the development of abnormally large turbidite beds (Parea & Ricci-Lucchi, 1975; Rupke, 1976; Ricci-Lucchi, 1978). The Contessa-like beds, which punctuate the basin plain successions of the Marnoso-Arenacea Basin in the northern Apennines, may show lateral continuity over distances as great as 175 km with only relatively minor changes in thickness (Parea & Ricci-Lucchi, 1975; Ricci-Lucchi, 1978); their persistence throughout the basin makes them excellent stratigraphic markers. They are up to 7 m thick and are commonly overlain by an equally thick pelite interval, their calcareous composition contrasting with the predominantly siliciclastic nature of the thinner bedded turbidites that chiefly characterise the basin. Their abnormal thicknesses and the presence of thick pelite beds above the sandstone intervals are thought to reflect the constriction

of flows within the limited area of the basin. As a result, they were unable to thin laterally and deposit their finer-grained components in more distal environments. The huge volumes of material involved and their calcareous composition suggests that they represent exceptional depositional events triggered by tectonic disturbances, possibly in the Latium-Abruzzi Platform to the south (Ricci-Lucchi, 1978).

Mega-beds up to 41 m thick that comprise discrete calcarenite and marlstone intervals, and can be followed only 16 km along strike, are also documented from the southwestern Pyrenees (Rupke, 1976). These beds, which are invariably underlain by slumps, are also thought to have been induced by earthquakes and to have been ponded behind basement highs.

The most particular feature of these mega-beds is the presence of a very thick, fine-grained layer overlying the sandstone interval. There are no such layers associated with the thick turbidites of the La Ralla Section, which grade only to fine calcareous sand at the top. However, gravitational settling from suspension of the mud component of these flows may not have occurred sufficiently rapidly to avoid dissolution, since the ambient siliceous sediments clearly indicate that deposition was occurring well below the CCD. The absence of thick marlstone layers may therefore be a function of depth and does not exclude the possibility that the thicknesses of these calciturbidites are due to sediment ponding. The abrupt nature of the transition from calcareous to siliceous deposition in the La Ralla Section suggests major downfaulting, and the basin may therefore have been deeper in this area than it was further south. As a result, turbidity currents following the axis of the basin from the northwest would have encountered an adverse gradient, causing rapid decelleration

and ponding of the flows. Differential downfaulting of the basin floor during the Early Liassic may thus have prevented lateral thinning of turbidites and restricted their deposition to relatively small areas, causing the development of abnormally thick beds.

The fabrics of the clasts indicate that the redeposited material must have originated in the surrounding carbonate platforms, but normal sediment supply cannot account for the immense volumes involved. Although prolonged hiatuses in the Jurassic sedimentary record of the Monti della Maddalena Unit suggests that the adjacent margin of the Campania-Lucania Platform was emergent for much of this period, tropical processes of subaerial erosion are considered unlikely to have produced coarse sediment in such large quantities (Scandone & Bonardi, 1968). Moreover, fracture of lithified sedimentary units must be invoked to explain the generation of lithic clasts.

Evidence for synsedimentary tectonic activity in the margins of the adjacent carbonate platforms during the Liassic and Dogger suggest that periodic faulting may have been responsible for the intermittent generation of these flows, which do not therefore relate to any established depositional system. (Scandone & Bonardi, 1968; d'Argenio & Scandone, 1970; d'Argenio & Sgrosso, 1974; d'Argenio, 1976). However, caution should be exercised in advancing a direct shallow-water origin for these calciturbidites simply upon the nature of their allochems and clast fabrics. Spar-cemented grainstones, such as characterise many of the lithic fragments in these beds, are not unique to platform carbonate (s.s) environments, but may also be found in redeposited platform margin facies (c.f. Bathurst, 1975); oolitic calcarenites, for example, are common in the Serra del Palo

Formation of the Monte Foraporta Unit (See Chapter 7). Moreover, rare clasts of radiolarian biomicrite also attest to some derivation of material from within the basin. Many of the calciturbidites may therefore have been generated by faulting of the basin margins and attest to remobilisation of sediment already redeposited from the shallow waters of the platform itself. Hence a large proportion of the calciclastic material of the La Ralla Section may have resided transiently on the basin slopes before finally reaching the deeper tracts.

Although the carbonate microbreccias of Unit I lack typical turbidite structures, their content of allochems and clasts of shallow-water origin confirms that they are allochthonous. The abundance of radiolarian biomicrite clasts, however, suggest that the flows originated on the platform margins, or else that the pelagic limestone clats were incorporated during trasport. The high clay-and hematite-contents of the matrix of some beds attest to entrainment of basinal siliceous ooze prior to deposition.

Silicification of the calciturbidites apparently occurred at a relatively early diagenetic stage and, in some cases, even before the development of a calcareous cement. Since deposition was occurring well below the CCD, this is perhaps not surprising; indeed, it is difficult to envisage how calcite cementation was achieved at all, particularly since the isotope data suggest that the precipitating fluids had a typically marine composition (Chapter 8). The necessary $CaCO_3$ must have been derived from the dissolution of the less-stable components of the sediment, particularly aragonite, and pore waters in the centres of the beds must have become sufficiently oversaturated with respect to calcite to cause precipitation. At the margins, on the other hand, pore-waters were initially undersaturated

with respect to both calcite and opal-CT with the result that calcedonic cements developed. Subsequent replacement of grains probably did not occur until dissolution of biogenous opal in the surrounding cherts had augmented concentrations of dissolved silica sufficiently to allow precipitation of opal-CT. The distension of primary laminae within chert nodules, and the development of bulbous cherts on the bases of calciturbidite beds, attests to a considerable influx of silica from the host sediments. It is suggested that the presence of carbonate may have encouraged the nucleation of opal-CT such that, with continual precipitation and nucleation, these calcareous beds acted like a sponge for dissolved silica (c.f. Kastner et al., 1977). As a result, the process of silicification was completed prior to the dolomitisation which subsequently affected many of the beds (See Chapter 8). The only dolomite rhombs present in the nodules were probably formed during chertification (c.f. Dietrich et al., 1963; Mišík, 1973; Jacka, 1974; Price, 1977b). Dissolution of calcite may have led to pore-waters becoming supersaturated with respect to dolomite, which thereafter began to precipitate along with opal-CT; alternatively, rhombs may have formed during the conversion of opal-CT to quartz due to the release of Mg^{++} ions (c.f. Kastner et al., 1977).

6.2.2 Model for the Lagonegro Formation

Deposition in the Lagonegro Basin during the Jurassic occurred largely below the CCD, but contrasting patterns of sedimentation developed in different parts of the basin in response to synsedimentary tectonic control. The principle features which must be taken into account in constructing an integrated depositional model are:

i) The rapid transition from calcareous to siliceous deposition in Unit II in the Early Jurassic, above the level of the intra-

and extra-formational carbonate breccias of the Sirino Formation;

ii) The difference in thickness of the formation in Units I and II;

iii) The uniformity of thickness and facies in Unit I;

iv) The rapid lateral and vertical thickness and facies changes in Unit II;

v) The paucity of redeposited siliceous sediments in Unit I and their abundance in Unit II;

vi) The presence of marls in the Pignola-Abriola Section;

vii) The presence of thick calciturbidites in the La Ralla Section, and their possible distal equivalents in the Pignola-Abriola Section;

viii) The abundance of vitreous cherts and thin pelagic limestone turbidites at the top of the formation in both units;

ix) The abrupt transition from siliceous mudstone to radiolarian chert deposition between the Cararuncedde and Pietra Members of Unit I, and the greater abundance of radiolarian cherts at the top of the formation in Unit II.

It is clear from the contrasting sedimentary features of the two units that the effects of synsedimentary tectonics were felt principally in Unit II, and that the depositional history of the lower unit reflects a more stable regime. Assuming that Unit I originally lay to the east of Unit II, as is indicated by their structural relationships, it can be seen that it was principally the western margin that was involved in the faulting. The palaeogeography of the basin during the Jurassic is discussed more fully in Chapter 9.

It is not possible to incorporate all of the Jurassic depositional environments of the basin onto a single diagram, and Figs. 6.4 and 6.5

depict the contrasting settings of the eastern and western margins during the Early and Late Jurassic respectively. Subsidence of the basin floor in the more-stable, eastern part continued through the Triassic-Jurassic boundary such that pelagic lime ooze deposition was finally eliminated. Thinly bedded, laminated, silicified limestones at the base of the Cararuncedde Member constitute redeposited lime muds from the shallower margins which were still above the CCD. Their gradual replacement by siliceous mudstones reflects the eradication of calcareous sedimentation in these marginal tracts as they too subsided. Periodic injection of carbonate microbreccias records minor disturbances in the platform, although a high proportion of pelagic limestone clasts suggest that much of the material was derived from the basin margins. Siliceous deposition was largely achieved by gravitational settling, although rare radiolarite beds imply some redeposition, again possibly from the basin flanks.

An altogether different story is told by the Torrente Bitonto Section, where faulting and subsidence at the end of the Triassic rapidly lowered the basin floor to depths below the CCD. Laminated and silicified limestones also occur in this part of the basin, but sedimentation and resedimentation of siliceous deposits in the form of siliceous mudstones and radiolarian cherts dominated. Rates of deposition were therefore correspondingly greater than in Unit I.

Meanwhile, on the other side of the basin and further to the north, events took on a much more dramatic appearance (Fig.6.4b). Subsidence in some parts was considerable, and the extraformational carbonate breccias at the top of the Sirino Formation were directly overlain by siliceous mudstones. Extensive faulting of the platform margins caused the introduction of large numbers of calciturbidites at

progressively less frequent intervals. Turbidite dispersal patterns were probably axial, and ponding of the flows may have been responsible for the rapid distal thinning of the beds to the south. Redeposition of indigenous siliceous deposits is once more manifested in the deeper parts of the basin, whereas marls and marly limestones occasionally accumulated in the shallower tracts, where less dramatic subsidence and high rates of deposition of calciturbidites conspired to maintain the basin floor at, or just above, the CCD.

The difference in sedimentary patterns in the two units are much less marked in the Late Jurassic when siliceous deposition predominated (Fig.6.5). In Unit I, siliceous mudstones were suddenly overlain by radiolarian oozes, and highly silicified pelagic carbonate turbidites once more punctuate the successions; the latter reflect the re-emergence of the basin flanks from below the CCD, which is believed to have undergone rapid depression at the end of the Jurassic (Garrison & Fischer, 1969; Bosellini & Winterer, 1975). In the deeper, rifted areas, redeposition of siliceous oozes continued, but a progressive degrease in the number of calciturbidites reflects a gradual cessation of tectonic activity. Here, too, the re-appearance of thin, intrabasinal lime-turbidites documents deepening of the CCD, although ambient sedimentation was primarily radiolarian in character. This essentially moderate pelagic depositional regime reigned supreme in the basin until the Early Cretaceous, when its position was usurped by sediments of an altogether different nature.

6.3 Sedimentology of the Brusco Formation

The cherts of the Lagonegro Formation are succeeded in both units by alternations of grey or black shales, silicified calcilutites and calciturbidites of the Brusco Formation. This formation is of

unknown thickness, but is believed to be of Lower Cretaceous age (See Chapter 2). Owing to the dominance of rheologically incompetent shales, the formation is invariably both highly deformed and poorly exposed. Sedimentological analysis is difficult, therefore, and only a brief description of the lithologies and their significance is possible.

i) Description

The fissile shales that dominate the formation are most commonly grey in colour, but may also be black, and are of clay or fine-silt grade. They are quartzose and contain a clay mineral assemblage that includes kaolinite, illite and chlorite (Fig.6.7); calcite is absent, and the shales are generally siliceous. Although commonly cleaved, particularly in highly deformed outcrops, sedimentary structures are not apparent. The darker shales appear to be relatively carbonaceous. Trace fossils have not been observed and the shales are seemingly barren.

Thick beds of calcilutite and carbonate breccia punctuate the majority of successions and impart a rhythmic appearance to which the formation owes the name 'flysch galestrino'. The calcilutite beds, which may be up to 1.5 m. thick, are massive and in sharp contact with the shales above and beneath. The upper few centimetres, however, may be strongly bioturbated. They are commonly silicified throughout, and contain appreciable amounts of fine-grained terrigenous material as well as both silicified and calcitised radiolarians; at the bases of beds, they may be distinctly pelletal.

The carbonate breccia beds, which may pass up into calcilutite at the top, are commonly graded and may bear flute marks on their bases. They comprise rounded grains set in a sparry calcite matrix, but

silicification has destroyed the primary textures of the majority of clasts; however, outlines of molluscs, red algae, benthonic Foraminifera and ooids may rarely be distinguished. Both clasts and cements are replaced by chalcedony or microcrystalline quartz.

In the southern part of the zone, Scandone (1967b, 1972) reports rare siliciclastic beds, comprising terrigenous silt and fine-sand, that are not found elsewhere in the formation. At Pecorone, abo in the south, the Brusco Formation passes up into Upper Cretaceous pelagic limestones, marls and cherts of the 'schisti rossi'.

ii) Interpretation

The non-calcareous composition of the grey and black shales of the Brusco Formation attests to lime-free deposition below the CCD. Estimates of the thickness of the formation, though inexact, imply that rates of sedimentation were much greater during the Lower Cretaceous than hitherto. This is thought to reflect an input of fine-grained terrigenous material which overwhelmed the background pelagic sedimentation of the basin. The absence of smectites from the clay mineral assemblages argues agains a significant volcanic contribution.

The evidence for bioturbation at the tops of calcilutite beds indicates that they were deposited rapidly. Moreover, their association with graded carbonate breccias suggests that all the calc-areous deposits of the formation were introduced by turbidity currents. The more coarse-grained calciturbidites contain clasts that indicate derivation from the adjacent carbonate platform, whereas Radiolaria in the calcilutites testify to intrabasinal redeposition. These beds are therefore thought to represent periodic interruptions of the

ambient fine-grained terrigenous sedimentation by incursion of extraneous material from the margins of the basin. The sharp contacts between calcilutites and shales argues against fluctuating deposition of either lime mud or terrigenous material as an alternative mechanism for their formation.

The origin of the terrigenous material is problematic since there is no evidence that any of the palaeogeographic units of the southern Apennines were subaerially exposed during the Early Cretaceous. Coeval pelitic deposits are present in the internal, Liguride Units of Lucania (Frido and Crete Nere Formations), as well as in the Palombini and Lavagna Shales of the northern Apennines, and at the base of the Monte Soro Flysch in Sicily (Vezzani, 1968, 1969; Bernoulli & Jenkyns, 1974); deposits comparable to the latter may also be found in Algeria, Morocco and the Canary Islands (Grandjacquet & Mascle, 1977; Robertson & Stillman, 1979). In the majority of cases, Lower Cretaceous shales pass upwards into flysch, but sequences in North Africa and the Canaries themselves contain high proportions of coarse siliciclastic material. These flysch deposits are thought to have originated through uplift and erosion of the Anti-Atlas Mountains (Robertson & Stillman, 1979). The grey and black shales that characterise units pertaining to the Tethyan oceanic basin as far north as Liguria may therefore represent the distal equivalents of Lower Cretaceous flysch deposited in submarine fans along the passive northern continental margin of Africa.

Widespread occurrence of black shales of Aptian-Albian age in not only the Alpine-Mediterranean region but also the world oceans has been interpreted as evidence for a world-wide oceanic anoxic event (Schlanger & Jenkyns, 1976; Fischer & Arthur, 1977; Jenkyns, in press). The black shales of the Brusco Formation may likewise reflect anoxic

conditions engendered in the basin during this episode (Jenkyns, in press).

Calciclastic deposits of the Brusco Formation are equally common in both of the Lagonegro Units and their presence in sequences of Unit I that lack such beds in the Lagonegro Formation may attest to proximity of the Abruzzi-Campania Carbonate Platform. The paucity of calciturbidites lower in the succession in Unit I cannot therefore be taken as evidence for this unit having occupied a distal part of the basin (c.f. Scandone, 1967b, 1972).

The lime-free nature of background sedimentation in the Lagonegro Basin during the Lower Cretaceous is in marked contrast to other Tethyan basins, the majority of which began to accumulate calcareous oozes at that time. Evidently the basin was sufficiently deep to remain beneath the CCD until the Late Cretaceous. A possible lateral equivalence of the Palombini Shales and the Calpionellid Limestones of the Ligurian northern Apennines may suggest that the deepest parts of the Tethyan Ocean also remained below the CCD at the end of the Jurassic (Bernoulli & Jenkyns, 1974; Barrett, 1979).

6.4 Discussion

One of the more puzzling aspects of Jurassic sedimentation in the Lagonegro Basin is the sudden transition from deposition of siliceous clays to radiolarian oozes in the Middle or Late Jurassic. This relationship, which is particularly clearly expressed in Unit I, contrasts markedly with the observation often made in chert sequences overlying ophiolites that the proportion of siliceous mudstones increases upwards (e.g. Leggett, 1978); this has been interpreted as a reflection of smoothing of topography, with a concomitant reduction in redeposition. Besides, the radiolarian cherts of the Pietra Member

show no evidence of redeposition, and a primary mechanism for producing concentrations of Radiolaria must be sought.

Cherts of Middle and Late Jurassic age are common in both oceanic and continental margin sequences of the Alpine-Mediterranean region (Bernoulli & Jenkyns, 1974). The association of radiolarites and ophiolites, that is so typical of the Tethyan oceanic domains and yet is so anomalous when compared with modern oceans, has been interpreted by several authors as evidence for a generic connection between submarine volcanism and chert formation (Steinmann, 1905; Grunau, 1965; Thurston, 1972). However, the presence of cherts in continental margin settings argues against such a relationship; besides, it has been shown that the hydrothermal contribution of SiO_2 to oceanic sedimentation, either by direct precipitation or by stimulating surface productivity of Radiolaria, is negligible (Garrison & Fischer, 1969; Garrison, 1974; Barrett, 1979). It is unlikely, therefore, that the radiolarian cherts of the Pietra Member owe their existence to a radiolarian bloom connected with submarine volcanism in the newly established Tethyan Ocean.

The distribution of radiolarian oozes in modern oceans is principally related to areas of upwelling and high plankton productivity located in sub-polar and equitorial regions (Ramsay 1973; Calvert, 1974). The sudden and widespread development of radiolarites in the Late Jurassic of the Tethyan region may therefore reflect changing patterns of oceanic circulation (Hsu, 1976). The opening of Tethys may have allowed nutrient-rich equatorial currents from the Pacific to invade the continental margin basins and stimulate surface productivity of Radiolaria. Basins that were deeper than the

CCD at that time would therefore have experienced a sudden transition from red clay to radiolarian ooze accumulation, such as is recorded by the Cararuncedde and Pietra Members of the Lagonegro Formation (Fig. 6.6). Rates of deposition following this event must have been sufficiently high to prevent dissolution of Radiolaria at the sediment-water interface. Furthermore, high rates of accumulation may also have conspired with higher inputs of organic carbon to foster reducing conditions in the underlying sediments and thus cause a change in their dominant colour from red to green.

Many Tethyan basins also show a transition from siliceous to calcareous sedimentation during the Tithonian that has been explained in terms of depression of the CCD at that time (Garrison & Fischer, 1969; Bosellini & Winterer, 1975). This, too, was probably related to the influx of nutrient-rich waters which in addition to Radiolaria, also engendered a proliferation of calcareous Foraminifera and nannoplankton, causing a dramatic increase in $CaCO_3$ supply to the ocean. The onset of calcareous sedimentation, however, would have lagged behind the transition from red clay to radiolarian ooze accumulation by several million years whilst the CCD was in the process of subsiding to the depth of the basin floor. Increased surface productivity is therefore recorded by a widespread occurrence of cherts overlain by pelagic limestones. The thickness of the chert succession should reflect water depth, since the shallower basins would have emerged from beneath the CCD relatively rapidly, allowing only a limited time for radiolarian accumulation. The Lagonegro Basin, on the other hand, was particularly deep and radiolarian cherts achieve notable development in the Pietra Member, but the transition to calcareous sedimentation is masked by swamping of the basin by fine-grained terrigenous clastics in the Lower Cretaceous. The sequential deposition of red clays,

radiolarian oozes and calcareous oozes observed in other Tethyan basins, however, is interpreted as a reflection of increased oceanic circulation during the Late Jurassic brought about by the opening of Tethys.

Depths of deposition can only be monitored in relation to the CCD which, as stated above, was not static during the Jurassic. Applying the model of Bosellini & Winterer (1975), lime-free radiolarites prior to the Malm were deposited below 2,500 m, but it seems unlikely that the CCD remained constant even before then; as discussed in Chapter 5, the continual introduction of lime mud from the carbonate platforms during the early Mesozoic had probably caused a progressive depression of dissolution levels throughout the period and, at the beginning of the Lias, the CCD may have been no deeper than 2,000 m. Rates of sedimentation in the Pignola-Abriola Section I were sufficiently high during the early Jurassic to maintain depths at or around the CCD, and marls still crop out over 100m from the base. Average rates of sedimentation of 1 and 6 m/MY calculated for Units I and II respectively, however, are at least an order of magnitude lower than the rates of regional subsidence implied by the Jurassic thicknesses of the carbonate platforms, and assuming that the basin was subsiding at a similar rate, it must have continued to deepen (d'Argenio & Scandone, 1970; Scandone, 1972; d'Argenio, 1976). Moreover, evidence for Liassic block faulting in Unit II suggests that some part of the basin may have subsided even more rapidly than the platforms. It is thus perhaps not surprising that, in contrast to the majority of Tethyan sequences, the Lagonegro Basin did not experience a resumption of calcareous pelagic depositions at the end of the Jurassic and must, therefore, have remained deeper than CCD. If Early Cretaceous dissolution levels in the Tethys Ocean were comparable to those of the other ocean basins, depths in excess of 4km are implied (van Andel, 1975)(See Chapter 8).

CHAPTER SEVEN

JURASSIC SEDIMENTATION IN A MARGINAL BASIN - THE MONTE FORAPORTA UNIT

7.1 Introduction

A further perspective of sedimentary patterns in the Lagonegro Zone during the Jurassic is provided by the marginal facies of the Monte Foraporta Unit. This Unit is elongate in outcrop and structurally overlies either the Lagonegro II or Monti della Maddalena Units (Figs. 1.1 & 1.6)(See Chapter 1). It has been interpreted as a relict marginal basin to the Campania-Lucania Platform, separated from the Lagonegro Basin by a horst, that is now represented by the Monti della Maddalena Unit (Boni et al., 1974). On the basis of sedimentological and structural considerations, this simple model has been modified to account for a possible stratigraphic contact between the two units.

7.2 The Facies of the La Calda Formation

7.2.1. The Calcilutite/Calcisiltite Facies

i) Description

Thinly bedded, dark-grey limestones and dolomites dominate both members of the formation, although deposits of the lower member are entirely dolomitic. These beds are up to 10 cm. thick, are almost invariably parallel-laminated and some may be markedly fissile (Pl.7.1). The lamination is primarily one of grain-size, which varies from silt to mud, and some beds may be graded within these limits. Individual beds commonly comprise several thin, graded laminae whose thickness and average grainsize vary at random. Those that are not dolomitic comprise micrite or micritic intraclasts and calcitised or silicified Radiolaria. The dolomites are composed of fine-grained dolomite in which no trace of primary

structure is preserved, although alternating dolomitic and calcareous layers may give the appearance of lamination. Replacement chert nodules are locally abundant.

These rocks are reduced and give off a pungent odour of H_2S when struck with a hammer. Black sapropelic horizons up to 1cm thick occur sporadically through the sequence and are associated with brown, argillaceous bands which may originally have contained a now-oxidised organic carbon component.

ii) Interpretation

The presence of Radiolaria in these deposits may attest to a pelagic origin, but the grading and varve-like grainsize laminations suggest that they are redeposited. The absence of cross-lamination argues against strong current deposition, and gravitational settling from very low density flows derived from the platform margin is favoured. A highly reducing depositional milieu is indicated by the foetid aroma of the carbonates and by the presence of sapropelic horizons containing unoxidised organic matter. Anoxic bottom conditions in modern oceans pertain where the oxygen minimum layer impinges on the continental slope (Fischer & Arthur, 1977). These limestones may have been deposited within this zone, but the absence of reduced sediments in other marginal sequences of the southern Apennine carbonate platforms of this age mitigates against this hypothesis. Similarly, the relatively oxidised aspect of coeval sediments in the Lagonegro units argues against invocation of an expanded oxygen minimum layer, such as is thought to characterise the world oceans during times of elevated sea level (Schlanger & Jenkyns, 1976; Fischer & Arthur, 1977). The preferred model involves deposition in a restricted basin in which anoxic bottom conditions arose through stagnation

fostered by a lack of circulation; in the Black Sea, for example, anoxic conditions are currently endemic below about 200 m as a result of density stratification, induced during the Holocene by the introduction of more saline water from the Mediterranean Deuser, 1974). The transition from the reduced rocks of the La Calda Formation to the more-normal Serra del Palo limestones may either reflect filling of the basin or the removal of whatever feature was causing the restriction; the latter may have been achieved by downfaulting of the basin threshold.

7.2.2. The Calcarenite/Calcirudite Facies

i) Description

Turbidite beds 4-60 cm thick, showing grading and comprising structureless and parallel laminated intervals, are commonly encountered (Pl.7.2 & Pl.7.3). Grain sizes range from pebble and coarse sand grade at the base of the coarsest beds to fine sand or silt. Both normal and coarse-tail grading are developed, and elongate clasts are oriented parallel to bedding, imparting a crude parallel lamination. The C, D and E divisions of a typical Bouma sequence are only seldom developed, and in some cases even the B division is absent. The bases of even the finest grained beds may be strongly erosive, but channelling on the scale of outcrop has not been observed. Graded units may be amalgamated or separated by laminated calcilutite/calcisiltite beds (Pl.7.4). The coarser and thicker beds occur sporadically throughout and may be overlain by packets of thin, graded beds in a fining upwards sequence.

The coarsest beds are almost entirely dolomitised, although less so in the upper member, and recognition of clast fabrics is

difficult. Large bioclasts are rare, and beds are chiefly composed of intraformationally derived mudstone and oxidised sapropel clasts. The matrix comprises a mosaic of fine-grained dolomite, probably replacing micrite. The majority of allochems have been leached and replaced by dolomite.

Slumping is abundant and may take a variety of forms. Individual beds, or groups of beds may appear chaotic and contain slump balls (Pl.7.5). Alternatively, coherent isoclinal folds may be developed, but they are insufficient in number to be used in the determination of regional palaeoslope orientations (Pl.7.6). A third form of soft sediment deformation is manifested by chaotic units which thicken laterally as successively older beds are incorporated (Pl.7.7).

ii) Interpretation

Bioclasts of shallow-water origin suggest that these sediments were redeposited from the adjacent carbonate platform. However, the incorporation of intraclasts in the coarser beds attests to intrabasinal derivation, and the abundance of slump sheets suggests that slumping may have been responsible for their genesis; sequences of thining and fining upwards beds above some of the coarse turbidites may possibly be the products of retrogressive flow sliding (Pickering, 1979). There is no evidence, therefore, to suggest that any of these calciturbidites derived directly from the platform margins. The absence of Bouma C, D and E divisions and coarse tail grading that typify these beds are characteristic of liquified flow deposits (Lowe, 1976). The crude lamination developed at the tops of some beds, imparted by preferential orientation of elongate clasts, may attest to deposition from flows transitional between liquefied flows and turbidity currents, but

unequivocal current structures, such as parallel- or cross-lamination are not developed.

The abundance of slumps and the nature of resedimentation are evocative of a slope environment but there is no evidence for introduction of coarse material directly from the platforms. On the deep fore-reef and island slope of Bermuda, infrequent and catastrophic movements of sediment are separated by long periods of suspended sediment accumulation, and only a small proportion of coarse material actually finds its way down to the lower island slope due to trapping on the fore reef (Moore et al, 1976). It is envisaged that most of the calcilutites and calcisiltites of the La Calda Formation were likewise passively introduced into the basin, and that resultant slope oversteepening led to their remobilisation in the form of intraformational slumps, liquefied flows and turbidity currents.

7.3. The Facies of the Serra del Palo Formation

7.3.1. The Calcilutite/Calcisiltite Facies

i) Description

Laminated and massive fine-grained limestones make up the bulk of the formation. They are dark grey in colour and may be either thinly bedded, in which case laminations are well preserved, or else massive (Pl.7.8 & 7.9). The structureless beds show evidence of bioturbation and commonly contain horizontal stylotites. Slumps are more or less abundant, particularly in the lower member (Pl.7.10) (Boni et al., 1974). The laminated beds comprise thin graded units about 5 mm thick that are highlighted by their variable organic content (Pl.7.11). The bases of these laminae are distinctly pelletal and may attain fine-sand grade, ostracods

and Foraminifera often forming a spar-cemented lag; the mudstone tops are free of bioclasts other than Radiolaria. Fragments of pelagic bivalves and ammonites are also present. The structureless beds generally show mixing of these components due to bioturbation. Calcareous nannoplankton are documented from the micrite (Boni et al., 1974).

ii) Interpretation

Deposition by gravitational settling, either from the water column or extremely low-density flows, is envisaged for this facies. Alternating periods of oxidising and reducing conditions are reflected by packets of thick, massive beds and more thinly bedded laminated limestones respectively. Although the majority of beds are resedimented, the structureless micrites with rare radiolarians and ostracods may have accumulated in situ. The faunas indicate a dominantly pelagic provenance, and redeposition was probably largely intrabasinal; frequent slumps are indicative of slopes. The dark-grey colour probably attests to reducing conditions during diagenesis, but oxygen levels near the sediment-water interface were at most times sufficiently high to support an infauna, as evidenced by the bioturbation. The dominance of this facies in the Lower Limestone Member reflects a continuation of sedimentary patterns established in the Early Liassic, but under less reducing conditions.

7.3.2. The Calcarenite Facies

i) Description

Poorly-sorted wackestones, packstones and grainstones occur in beds 10-30 cm thick that commonly grade up into parallel and convolutely laminated calcisiltites and calcilutites (Pl.7.12); in many cases, they may be bioturbated. Structures are inconspicuous,

however, and the beds are indistinguishable from massive calcilutites on weathered surfaces. They contain a mixed neritic and pelagic fauna including, either as complete shells or as fragments, echinoids, bivalves, gastropods, brachiopods, corals, red and green algae, ammonites, Foraminifera, ostracods, Radiolaria and sponge spicules (Pl.7.13 & 7.14). All neritic bioclasts are micritised to some extent, the largest possessing micrite envelopes; molluscan fragments are leached and replaced by sparry calcite. An abundance of unidentified micritised peloids and grapestone clasts, however, make up the bulk of the sediment. Grains are either spar-cemented or set in a fine-grained lime matrix, which may contain calcareous nannoplankton (Boni et al., 1974).

ii) Interpretation

The grading and parallel-lamination of these beds suggests that they were deposited by turbidity currents which, judging by their high content of neritic bioclasts, were derived from the neighbouring platform. The incorporation of pelagic organisms and micrite into the flows, typical of the wackestones and packstones, was probably achieved during transport by erosion of underlying beds; the spar-cemented grainstones, on the other hand, represent the deposits of 'cleaner' flows that did not entrain significant amounts of lime mud.

7.3.3. The Calcirudite Facies

i) Description

At intervals in the Limestone-Marl and Upper Limestone Members, beds of intraformational conglomerate are developed that comprise matrix-supported clasts of calcarenite, calcisiltite and calcilutite. The matrix, which is generally marly, is commonly laminated, the laminae showing evidence of soft-sediment deformation (Pl.7.15 & 7.16).

Clasts, which vary in size from granules a few centimetres across to cobbles and boulders, are commonly elongate parallel to bedding and are dispersed at random through the beds, which may be several metres thick. In the finer-grained and less matrix-rich varieties, clasts are commonly in contact and may be separated along anastomosing stylotites. There is no evidence of grading or imbrication of pebbles, which are structureless and well-rounded.

ii) Interpretation

The random dispersion of intraformationally derived clasts in a marly matrix is typical of debris flow deposits (Middleton & Hampton, 1976). These mass flows were probably generated by slumping of packets of beds induced by failure and subsequent liquefaction of the matrix. Lamination developed in response to laminar flow, and rounding of the clasts may have occurred during transport. The triggering mechanism for these submarine slides was probably tectonic, and the development of the facies solely at the top of the formation either indicates a period of pronounced tectonic activity in the Late Jurassic or, more plausibly, that depositional slopes had built up and become oversteepened.

7.4. Discussion and depositional model for the formations of the Monte Foraporta Unit

The basic pattern of sedimentation throughout the Monte Foraporta Unit is of deposition of fine-grained carbonate, probably derived in part from the adjacent Campania-Lucania Platform. This was achieved either by gravitational settling from the water-column or from very low-density turbid flows. Deposition on slopes is evidenced by the abundance of slumps and, at the base and top of the succession, by intraformational mass flow deposits; the

orientation of the slope or slopes is unknown. Euxinic conditions pertained throughout the early history of this minor basin due to restricted water circulation which prevented mixing of basin waters with the well-oxygenated ocean water of the Lagonegro Basin.

The nature, dimensions and morphology of this basin are difficult to establish. The unit outcrops along the contact between the Lagonegro Units and the Alburno-Cervati carbonate platform Unit and structurally overlies either the Triassic white dolomites of the Monti della Maddalena Unit or the basinal Lagonegro Unit II (See Chapter 1). This tectonic sequence of stacking has led previous workers to conceive of the Monte Foraporta Unit as the remains of a basin within the Campania-Lucania Platform, separated from the Lagonegro Basin by the shallow-water carbonate build-ups of the Monti della Maddalena Unit (Boni et al., 1974)(Fig.7.1a).. An alternative model, and one that leans as heavily upon analogy and structural considerations as upon the sedimentology of the formations themselves, proposes that the Monte Foraporta and Monti della Maddalena Units, though now separated by a thrust, were originally in stratigraphic contact. It is suggested that listric downfaulting of a large segment of the Campania-Lucania Platform margin produced a basin bounded on one side by a fault scarp and on the other by a tilted surface of the subsided block (Fig.7.1b); filling of this basin produced an onlap at the base of the Monte Foraporta Unit. There are several lines of evidence for this model:

a) No time overlap exists between the Monti della Maddalena and Monte Foraporta Units in this area. The age of the dolomites in the former is given as Upper Triassic, whereas the La Calda Formation has been attributed to the Upper Triassic-Lower Liassic, although no

biostratigraphic data are available (Scandone, 1972; Boni et al., 1974). The abundant evidence of redeposition in the latter suggests that rates of sedimentation were high, and backtracking from the Middle Liassic age for the Lower Limestone Member of the Serra del Palo Formation does not necessarily imply that the unit is as old as Late Triassic (c.f. Boni et al., 1974).

b) Downfaulting of the margins of the Campania-Lucania Platform is also manifested in other units, particularly the Bulgheria-Verbicaro Unit, which once constituted the internal margin of the platform (See Chapter 1). Here, Late Triassic white dolomites are succeeded by Jurassic and Cretaceous limestones of fore-reef transitional to basinal environments (d'Argenio & Scandone, 1970; Ippolito et al., 1975). Moreover, in other parts of the Monti della Maddalena Unit itself, which formed the external margin of the platform, similar facies of Liassic and younger age, unconformably overlying Triassic and Infraliassic dolomites, may also be found (Scandone & Bonardi, 1968; d'Argenio & Scandone, 1970; Scandone, 1971; Ippolito et al., 1975). Liassic synsedimentary tectonics have been held responsible for the differentiation of these platform margin units from the Alburno-Cervati platform carbonate (s.s) Unit (Scandone & Bonardi, 1968; d'Argenio & Sgrosso, 1974; Ippolito et al., 1975).

c) The extremely angular (about 30°) relationship between bedding of the Monte della Maddalena and Monte Foraporta Units need not necessarily imply major thrusting (Pl.3.17). If the downfaulted block had tipped towards the platform, as a listric fault model would suggest, onlapping of horizontally-bedded basinal deposits could equally well have produced such a relationship between the two units (Fig. 7.2).

The suggested sequence of events is therefore as follows. Foundering of the margin of the Campania-Lucania Platform in the Late Triassic or Early Jurassic along listric faults produced a small basin bounded to the west by a fault scarp and to the east by the uptilted surface of the faulted block (Fig.7.2). Euxinic conditions prevailed in the basin during its early history as a result of limited circulation probably induced by density stratification; the summit of the upturned block acted as a 'threshold' between the Monte Foraporta and Lagonegro Basins and inhibited water mixing between the two. As the basin filled up during the Jurassic, and as the entire area subsided, anoxic conditions were gradually eliminated and a normal platform marginal sequence developed in the Late Liassic, Dogger and Malm.

A comparable model has been proposed for the Late Jurassic siliciclastic sediments of the Wollaston Forland Group in east Greenland (Surlyck, 1978) (Fig.7.3). Submarine rock-fall breccias pass laterally into conglomerates, sandstones and mudstones in quick succession, and form a clastic wedge that thins rapidly in a distal direction. These sediments, which occur in fining-upwards megacycles, are considered to be the deposits of short-lived, coalescing submarine fans that accumulated at the foot of fault scarps in elongate troughs overlying tilted fault blocks. The tectonic regime under which these basins developed was different to that of the Tethyan region since faulting was accompanied by uplift. This caused emergence and erosion of the summits of the faulted blocks, which were thus able to contribute large volumes of clastic sediment to the basin. All units of the southern Apennines, on the other hand, had been subsiding gradually since the

Middle Triassic and at no time did the platforms become wholly emergent prior to the Cretaceous (Ippolito et al., 1975). It can be seen, therefore, that although listric faulting can be invoked in each case, the contrasting structural regimes in Greenland and southern Italy fostered somewhat different patterns of sedimentation.

An idealised cross-section across the Lagonegro Basin during the Late Jurassic from northwest to southeast shows how the stratigraphies of the Monte Foraporta and Lagonegro Units relate to the Liassic rifting patterns (Fig.7.4). Although the relative positions of the units have been deduced on the basis of structural considerations, evidence for block faulting in both the Monte Foraporta and Lagonegro II units lends further weight to the contention that only the western parts of the basin experienced the full weight of the Liassic tectonics.

CHAPTER EIGHT

GEOCHEMISTRY, DIAGENESIS AND DOLOMITISATION OF THE CARBONATE ROCKS OF THE LAGONEGRO ZONE

8.1. Introduction

A combination of petrographic and geochemical techniques have been employed in order to elucidate the late diagenetic history of the basinal carbonate lithologies of the Lagonegro Zone. In the hemipelagic lime-mudstones of the Sirino Formation, aggrading neomorphism has obliterated the primary textures of the sediment, while solution and re-precipitation of biogenous silica has fostered the development of nodular cherts. Such is the degree of diagenetic modification that little evidence remains of the constitution of the primary lime-mud. Original fabrics are further obscured by dolomitisation, which has affected limestones of both the Sirino and Lagonegro Formations in Unit II north of Marsico Nuovo. In this chapter, these various diagenetic trends are followed, culminating in a description and interpretation of the cause of dolomitisation.

The methods employed are briefly reviewed in Section 8.2, and the significance of the results regarding the diagenesis of the limestones and cherts of the Monte Facito and Sirino Formations is discussed in Section 8.3. This is followed by a consideration of aspects relating to dolomitisation and a review of the implications of diagenesis and dolomitisation for the development of the basin in Sections 8.4 and 8.5 respectively.

8.2. Methods

8.2.1. Scanning electron microscopy

Polished surfaces of limestones were etched in dilute HCL and

examined at magnification of 200X to 5000X on a Cambridge Stereoscan instrument; cherts were prepared similarly, but were etched in dilute HF.

8.2.2. X-ray fluorescence

Samples were powdered using a tungsten carbide swing mill, and an agate planetary ball mill, and pressed pellets were analysed for Sr^{++} using conventional techniques. Results were compared with a range of U.S.G.S. standards. Replicate analyses of one sample gave a standard deviation of 7.9 ppm, and a least squares regression calculated from the measured standards gave a regression coefficient of 0.999 and a standard deviation of 11.25 ppm. The results are given in Appendix 2.

8.2.3. Electron microprobe

Selected samples were analysed for Si, Ca, Mg, Fe and Sr by conventional electron microprobe techniques using a Cambridge Microscan 9 instrument. Z.A.F. corrections were made for carbonates. The results are tabulated in Appendix 3.

8.2.4. Stable-isotope mass spectrometry

Samples were crushed using a steel percussion mortar, and powdered with an agate pestle and mortar. All trace of organic material was removed by leaving the sample for three days in a 3-6% NaOCL solution; they were subsequently rinsed several times in de-ionised water, and dried. Calcite:dolomite ratios were determined semi-quantitatively by X-ray diffraction. 10-25 mg of powder were reacted with 100% phosphoric acid at 25.0^{o}C according to the method outlined by McCrea (1950). CO_2 gas was extracted after 12 and 72 hours in the case of pure calcite and dolomite samples respectively, and after 1, 3 and 72

hours for mixed calcite/dolomite samples; in the latter case, the 1 and 72 hour extractions are assumed to represent the gas derived from the calcite and dolomite respectively. The gas was analysed with a Micromass 903 triple-collector mass spectrometer with reference to CO_2 prepared from a calcite standard in the same way as outlined above. δO^{18} values for dolomite were therefore recalculated to account for the different fractionation factors for acid decomposition of dolomite and calcite (Sharma & Clayton, 1965). Results are expressed as per mil deviations from the PDB standard by the following expression:

$$\delta = ((R_{sample}/R_{standard})-1)1000$$

where R is C^{13}/C^{12} or O^{18}/O^{16} as appropriate

The analytical precision of a single determination, expressed as 2X the standard error of the mean of seven readings, is about $\pm$ 0.019 per mil for δO^{18} and $\pm$ 0.013 for δC^{13}. The repeatability of the measurements, as the standard deviation of the mean calculated from replicate analyses, is about 0.09 and 0.03 per mil for δO^{18} and δC^{13} respectively. A table of results is given in Appendix 4.

8.3. Diagenesis of limestones and cherts of the Sirino and Monte Facito Formations

8.3.1. Description and results

8.3.1.1. Petrography

The petrography of the limestones of the Sirino and Monte Facito Formations has been described in chapters 4 and 5 respectively, but description of their ultrastructure has so far been excluded. The radiolarian biomicrites of the latter comprise an interlocking mosaic of calcite grains chiefly between 2 and 10μ across (Pl.8.1). The grains are of irregular shape and show no internal structure;

recognisable calcareous nannoplankton have not been found. Insoluble grains, probably of quartz, of similar size to the calcite grains are also present; they, too, are of irregular form and may coalesce to form stringers. Larger amoeboid calcite crystals greater than 10μ across appear to have formed by cannabalisation of adjacent micrite grains, and small particles can be observed that are only partially incorporated into these clots; one such grain preserves some features that are possibly relict organic structures, and may be an overgrown and recrystallised coccolith. These patches of microspar may accrete grains over a wide area, and measure as much as 100μ across. Radiolarian moulds are filled by sparry calcite, and preserve no trace of primary structure.

A detailed description of the chert nodules that abound in these lime mudstones has also been so far omitted. They are white, yellow, grey or black in colour, are elongate and arranged in layers parallel to bedding, and may join up to form continuous bands. Although generally spherical or ellipsoidal, they may also be of irregular shape (Pl.5.7 and 5.8). Rows of nodules show no common relationship to bedding planes. In some laminated, silty beds, however, nodules appear to have been exhumed and reworked. The cherts invariably possess a carbonate component, either predominantly calcitic or dolomitic, which is either concentrated around the margins or occurs as concentric layers, giving the nodule a banded appearance. On the basis of these features, it is possible to distinguish four types of nodules; calcitic nodules can be found in both units, but the dolomitic varieties are peculiar to Unit 1.

a) Unbanded calcitic chert nodules

The aspect of these nodules, which are the most common type, is

shown in Pl.8.2, and a typical sequence of zones can be observed between the core and the margin. The inner zone A, or cortex, comprises the bulk of the nodule and is composed of clear microcrystalline quartz. Scattered rhombs of calcite may be present, particularly along the thin, bed-normal veins which pervade the chert. These rhombs show evidence of corrosion and replacement by fine-grained chalcedony, which also fills the veins; commonly, the cores of the rhombs are preferentially replaced. Radiolaria are preserved as moulds filled by brown, sheaf-like aggregates of chalcedony, but primary geopetal fills are represented by finer-grained granular chalcedony

Zone B contains more abundant calcite rhombs which form an interlocking mosaic of euhedral crystals up to 3mm across. They are a dirty grey colour and contain phantoms of Radiolaria as clear, spar-filled spherical moulds (Pl.8.4). The rhombs are commonly zoned and certain zones may show preferential silicification (Pl.8.5). The calcite mosaic and the brown microcrystalline quartz groundmass contain an abundance of predominantly opaque rhombs that are an order of magnitude smaller than the calcite varieties and are commonly zoned; staining reveals that the translucent crystals are composed of dolomite, whereas the opaque ones contain calcite and, probably, iron oxides. The groundmass is darker than in the cortex, and contains abundant small, brown flecks, probably of pyrite; the impurities commonly preserve the outlines of rhombic ghosts. Large calcite rhombs, in the process of corrosion and replacement by chert, are abundant; in contrast, the smaller, opaque rhombs show no sign of alteration. As in the cortex, the calcite rhombs are preferentially situated, and show least sign of corrosion, along bed-normal veinlets.

Zone C is composed entirely of a calcite mosaic and forms a rim up to 5 mm thick around the nodule. Zone D comprises a secondary cortical layer of clear chalcedony, which passes through a transition zone (E) of mixed chalcedony and micrite, into the pure radiolarian biomicrite of the host rock (F). In the latter, radiolarian moulds are invariably filled by sparry calcite, but at some distance from the nodule, they may contain chalcedony.

Under the scanning-electron microscope, the replacement of rhombic calcite by micro-crystalline quartz can be clearly observed (Pl.8.6). The euhedral crystals contain patches of quartz, and their indented margins show evidence of corrosion. The zoning of the dolomite rhombs is also clearly discernable, and it appears that the iron oxide impurities have developed preferentially in specific zones. In Zone E, etching with HF has removed the silica and left a pattern of closely-spaced spherical pits about 50μ in diameter that are lined by micrite and may contain large rhombs of calcite.

b) Banded calcitic chert nodules

In these nodules, Zone B comprises concentric zones of rhombic calcite and microcrystalline quartz that are parallel to the margins of the nodule, and become progressively thinner and close together near the rim (Pl.8.7); there they merge to form a calcite mosaic similar to Zone C (Pl.8.9). As well as being concentrated into bands, the calcite rhombs may also be arranged along veinlets normal to the banding, and have undergone partial replacement by chalcedony, the cores generally being preferentially corroded (Pl.8.10). Smaller opaque crystals are scattered amongst the chalcedony and the calcite mosaic, apparently resisting corrosion, but they are almost absent from

the quartz bands. The siliceous layers and the cortex of the nodule are composed of featureless microcrystalline quartz, but rare chalcedonic pseudomorphs after calcite rhombs can also be distinguished. Zone D, the second cortical layer, comprises very fine grained chalcedony that becomes progressively coarser grained through Zone E, in which radiolarian moulds are filled by single quartz crystals that are commonly overgrown to form euhedral, cubic grains (Pl.8.9). These crystals thin out away from the margin as calcite-filled radiolarian moulds become more common.

c) Unbanded dolomitic chert nodules

These nodules comprise a cortex of pure chert surrounded by a dolomitic chert zone that weathers to a characteristic yellow/brown colour, and commonly possesses an outer siliceous crust (Pl.8.11). The cortex is similar to that of the calcitic nodules; the dolomitic zone, however, is pervaded by rhombs of dolomite 50-100μ in diameter. These form a dense interlocking mosaic, and microcrystalline quartz is only present in the interstices (Pl.8.12). Staining reveals that the dolomite is ferroan and commonly zoned, the cores being more ferroan than the rims. The cores are in some cases composed of calcite shot through with dark brown opaque material (Pl.8.13).

d) Banded dolomitic chert nodules

Some of the dolomitic chert nodules contain relatively minor dolomite which is scattered thinly through the microcrystalline quartz groundmass and is concentrated into bands parallel to the nodule margin (Pl.8.8); in general, these bands become closer together away from the core. The nodules are lined by a rim of rhombic calcite that is also distributed along veins in the chert (Pl.8.14 & 8.15).

The veins may continue into the host rock where they are filled by sparry calcite. One chert band was sampled in which the laminae appear to have been deformed and the quartzose rim ruptured, forming a teepee-like structure that resembles the cross-section of a volcano (Pl.8.8). In the overlying micrite, patches of opaque rhombs form 'mushroom clouds' up to 2 cm above the 'crater'.

8.3.1.2. X-ray fluorescence and microprobe analyses

Selected samples were analysed for Sr^{++}, and the results are given in Tables 8.1 and 8.2. The ordinary radiolarian biomicrite contains lower Sr^{++} concentrations than either the micrite intraclast, that shows evidence of early lithfication, or the calcarenites. The highest values in the limestones are from the intrabiosparite; microprobe analysis of the cements and clasts show little variation, although the intrabiosparite contains appreciably more than the intrabiomicrite. Rhombic calcite from the chert nodules contain variable amounts of Sr^{++}, but they are significantly higher than the host biomicrite; the smaller, and commonly opaque, rhombs contain negligible amounts, however. Dolomite rhombs from dolomitic chert nodules contain approximately half as much Sr as the rhombic calcite; whole rock analyses contain less Sr^{++}, but these samples include considerable amounts of quartz.

Table 8.1 Results of X-ray fluorescence analysis of limestones and cherts of the Sirino Formation - Sr contents.

Sample No.	Sample description	Sr (ppm)
4022	Radiolarian biomicrite	282
5220	Radiolarian biomicrite intraclast	699
5229	Intrabiosparite	1077
5231	Intrabiomicrite	608
5217	Dolomitic chert nodule	271
5218	Dolomitic chert nodule	268
5233	Chert nodule with de-dolomite	67

Table 8.2 Results of electron microprobe analyses of limestones and cherts of the Sirino Formation - Sr contents.

Sample No.	Sample description	Sr (ppm)*
4022	Micrite grains from radiolarian biomicrite	430±86
5229	Micrite intraclasts from intrabiosparite	875±96
5229	Sparry calcite cement from intrabiosparite	867±665
4022, 5215b	Rhombic calcite from chert nodules	1400±642
4022	Small calcite/iron oxide rhombs from chert nodules	65±58
5215b	Dolomite rhombs from dolomitic chert nodules	528±95

* Values quoted represent the mean and standard deviation of several analyses from within the sample(s); the latter, therefore, is a measure of their range rather than of the accuracy of the determinations.

The compositions of the carbonate phases in terms of Ca, Mg and Fe are illustrated diagramatically in Figs. 8.1, 8.2 and 8.3. It can be seen that all calcites, either from the radiolarian biomicrites, intrabiosparite and chert nodules, contain less than 2 mol% Mg and 0.5 mol % Fe; the rhombic calcites tend to contain slightly less Mg than the other varieties, however. The smaller calcite rhombs on the other hand, are markedly more ferroan and may contain up to 16 mol % Mg and 11 mol % Fe. Dolomites from the dolomitic chert nodules are also ferroan, as well as being slightly calcian, having an average composition of $(Ca_{52.6}:Mg_{45.6}:Fe_{1.8})$; they thus contain more Fe than rhombic calcites, even when they coexist within the same nodule.

8.3.1.3. Stable-Isotope Analyses

Samples of various lithologies of the Sirino and Monte Facito Formations were analysed, including rhombic calcite and dolomite from chert nodules, and the results are depicted graphically in Figs. 8.4 and 8.5. Irrespective of rock type, the calcites of the Sirino Formation all lie within a narrow range of δO^{18} and δC^{13} values, from -2.85 to -4.85 per mil and +2.01 to +2.85 mer mil respectively (Fig.8.4). Mean δO^{18} for samples collected from the top of the formation in Units I and II are -4.12 and -3.42 per mil respectively, and a Student t-test indicates that they are significantly different at the 95% confidence level. Biomicrites and intrabiosparites show no significant differences. Two samples from the Gianni Griecu Member fall in the lower part of the δO^{18} range, the radiolarian biomicrite showing slight enrichment with respect to the marly limestone, but a δC^{13} depletion of 0.76 per mil. A comparative sample of radiolarian biomicrite from the Scillato

Formation of the Madonie Mountains, Sicily, gives similar values, although it is slightly enriched in O^{18} and depleted in C^{13} relative to the Lagonegro limestones.

Rhombic calcite and dolomite from chert nodules also fall within the same δO^{18} range; however, the dolomites are both slightly enriched in C^{13} by about 1 per mil in comparison to the host biomicrite, whereas the rhombic calcite is depleted by a similar amount.

Calcites from the Monte Facito Formation contain lighter oxygen than those of the Sirino Formation, and show a much greater scatter (Fig.8.5). Once more, however, the δC^{13} values show a far smaller variation than δO^{18}, in this case of little over 1 per mil. The nodular limestones and the pink micrite from the neptunian dyke show the most O^{18} depletion, but the dolomite in the matrix of the Valle del Pesce nodular limestone is markedly less depleted than the co-existing calcite and the calcite of the nodules. In contrast, the nodular limestones collected from the Lercara Formation at Palazzo Adriano, Sicily, shows negligible depletion.

8.3.2. Interpretation

8.3.2.1. The limestones of the Sirino Formation

The interlocking texture of the micrites is strongly reminiscent of the amoeboid mosaics observed in many fine-grained limetones (Fischer et al., 1967). Such limestones, that have almost zero porosity, clearly possess a very low diagenetic potential, and have probably reached the penultimate stage in a diagenetic sequence that culminates in the production of a clean pavement mosaic of almost equigranular crystals with planar intercrystalline boundaries (Fischer et al., 1967; Bathurst, 1975;

Schlanger & Douglas, 1974). Production of these final pavement mosaics is though to involve thermal and stress effects, but the texture of the Sirino Formation limestones do not evince such dramatic conditions. Documentation of the developmental stages in the production of amoeboid mosaics of this kind provides evidence that both cementation and recrystallisation are involved (Bathurst, 1975; Fischer et al., 1967; Wachs & Hein, 1974); the progressive development of overgrowths on coccoliths, and their gradual loss of structure, reflect the interplay of these processes. Studies made of the ultra-structural changes associated with the down-hole transition from calcareous ooze to chalk and finally limestone in deep-sea sediment cores have often noted the development of overgrowths on the larger nanno-organisms at the expense of smaller calcite particles (Wise & Hsü, 1971 ; Andelseck et al., 1973; Schlanger & Douglas, 1974; Matter et al., 1975; van der Lingen & Packham, 1975; McKenzie et al., 1978); smaller coccoliths tend to disaggregate into small crystals due to dissolution along sutures. The driving force behind this process must be the reduction in free energy associated with a lowering of surface area as the average crystal size increases (Schlanger & Douglas, 1974). It has been argued that these processes alone could have produced the characteristic fabric of the Solenhofen Limestone, for which a nannofossil origin has been argued (Laffite & Noël, 1967; Wise & Hsü, 1971). However, over-growths on coccoliths and other calcite grains are usually euhedral, and a process that simply involved dissolution and reprecipitation of calcite cements should produce grains with planar intercrystalline boundaries. The development of amoeboid mosaics, therefore, must also involve an element of recrystallisation; this has been elegantly demonstrated by the marked similarity of the artificially produced

micrite textures of Hathaway & Robertson (1961) to those of the Solenhofen limestone.

It has also been suggested that the formation of amoeboid mosaics involves pressure solution, even though the commonly observed preservation of fragile shells attests to a lack of major compaction (Bathurst, 1975). The experiments of Shinn et al. (1977), however, indicates that significant compaction can occur without fracture of bioclasts. But Fischer et al. (1967) figure amoeboid mosaics, that comprise strongly interlocking grains, in which circular outlines, probably of coccoliths, are only mildly indented; welding of grains inside and outside the sphere, however, has obliterated the margins in several places. It does not appear, therefore, that significant dissolution has occurred between grains, but that they have passively coalesced. In the case of the Sirino Formation limestones, this process has gone one step further and clots of microspar, equally irregular in form to the micrite grains, have developed at the expense of the smaller particles. This process, termed 'coalescive neomorphism', is an important feature in the development of neomorphic spar, and is probably achieved by wet boundary migration after dissolution and reprecipitation of calcite had almost eliminated primary porosity (Folk, 1965; Bathurst, 1975).

It appears, therefore, that these limestones have undergone a complex transformation from the original lime mud, probably involving phases both of solution/reprecipitation and wet recrystallisation. These processes produced an amoeboid mosaic of micrite grains that subsequently coalesced to form microspar. The dominance of amoeboid, rather than euhedral pavement mosaics suggests that the temperatures and pressures at which the transformation took place were not abnormally high.

Sr^{++} concentrations provide a valuable tool in the interpretation of diagenetic mechanisms in limestones, and in determining their initial composition (Kinsman, 1969; Veizer & Demovič, 1974). Knowledge of the distribution coefficients of strontium for the various carbonate phases involved can beused to determine the Sr^{++}/Ca^{++} ratio of the precipitating solutions. Aragonite, which can substitute Sr^{++} into its lattice relatively easily, has a much higher distribution coefficient than calcite (Kinsman & Holland, 1969). Using these data, and assuming an appropriate Sr^{++}:Ca^{++} ratio, it is possible to predict the Sr^{++} content or aragonite and calcite precipitated in equilibrium with seawater; the respective values are 8200 and 1200 ppm respectively. It is clear from the wide variation in the Sr^{++} compositions of skeletal calcites and aragonites that considerable biochemical fractionation occurs during their formation. Diagenetic modification of Sr^{++} contents will vary depending on whether the recrystallisation system is open or closed, the former commonly leading to depletion whereas, in the latter case, initial compositions may be preserved (Kinsman, 1969). Thus recrystallisation of an aragonitic mud in an open system may produce a calcite that contains less Sr^{++} than a limestone that originally comprised only calcite, since the latter may not have passed through a solution-reprecipitation phase (Veizer & Demovič, 1974); low Sr^{++} concentrations, therefore, do not necessarily exclude the presence of aragonite in the primary sediment.

The generally low Sr^{++} contents of the limestones analysed suggest that an open system of crystallisation pertained during diagenesis. The value determined by X-ray fluorescence for normal radiolarian biomicrite is similar to the average reported for coeval cherty limestones of the central western Carpathians (Veizer & Demovič, 1974)(Fig.8.6)

The Triassic limestones have Sr^{++} contents that are low in comparison to the range observed in post-Tithonian pelagic limestones which were originally composed of calcareous nannoplankton tests. This has been interpreted as a reflection of a higher content of metastable carbonate minerals in the Triassic sediments, whose Sr^{++} was largely lost during solution and reprecipitation (Veizer & Demovič, 1974); on the basis of a single determination, it is not possible to dwell on the validity of this argument.

In contrast, the biomicrite intraclast, that was evidently lithified at an early stage, together with the intrabiomicrite and intrabiosparite, are all enriched in Sr^{++}, the latter showing a value approaching that predicted for inorganically precipitated calcite of 1200 ppm. This suggests that early cements were composed of low-magnesian calcite, rather than either high-magnesian calcite or aragonite, as are those currently formining in the peri-platform oozes of Tongue of the Ocean (Schlager & James, 1978).

It might be expected that the carbon and oxygen isotope results would reflect this trend towards early lithfication (c.f. Hudson & Coleman, 1978; Schlager & James, 1978). Surprisingly, however, the radiolarian biomicrites and intrabiosparites all fall within a similar range, including the biomicrite intraclasts, and are depleted in oxygen relative to modern marine cements and lithified crusts (Hudson, 1977; Milliman & Müller, 1977). It seems, therefore, that diagenetic processes have modified the isotopic compositions even of those sediments that were already lithified. Accepting that such modification has occured, it is possible to put some constraints upon the mode of diagenesis and the nature of the primary sediment.

The narrow range of δC^{13} values suggests that they have withstood any changes during diagenesis; δC^{13} is relatively insensitive to temperature, and variations are generally related to organic processes, which can influence the isotopic composition of carbonates dramatically (Russel et al., 1967; Emrich et al., 1970; Hudson, 1977; Irwin et al., 1977). Although high for a calcite precipitated in direct equilibrium from seawater, as well as in comparison to the average Triassic value for marine limestones quoted by Keith & Weber (1964), these values are typical of both shallow and deep-water sediments forming today (Hudson, 1977; Milliman & Müller, 1977 (Fig.8.7). Numerous documentations of stable-isotope trends in deep-sea sediment cores have noted that similar δC^{13} values persist to considerable depths and show little down-hole variation, whereas δO^{18} shows a steady depletion (Douglas & Savin, 1971; Savin & Douglas, 1973; Anderson & Schneidermann, 1973; Coplen & Schlanger, 1973; Matter et al., 1975; McKenzie et al., 1978). It is suggested, therefore, that the δC^{13} values reflect the original isotopic composition of the sediment and imply that organic reactions have not played a significant part in diagenesis.

The δO^{18} values, on the other hand, are all depleted relative to normal marine values, even though they fall within the range of most pelagic limestones and chalks (Hudson, 1977; Milliman & Müller, 1977)(Fig.8.7). This could reflect either the influence of isotopically lighter meteoric water during diagenesis, or re-equilibration at higher temperatures. Meteoric water containing atmospheric CO_2 is in equilibrium with a calcite of δC^{13} between +1 and +2 per mil, but lighter, soil-derived CO_2 commonly produces waters that are significantly lighter; the relatively constant δC^{13} values of about 2.75 per mil mitigate against involvement of meteoric

water, therefore. Oxygen fractionation, however, shows a pronounced sensitivity to temperature, and it seems more likely that the depletion observed in these limestones relates to diagenetic processes at higher temperatures (Craig, 1965). The isotopic composition of calcite is related to the temperature and isotopic composition of the precipitating fluid by the following expression:

$$T = 16.9 - 4.38\ (\delta_c - \delta_w) + 0.10\ (\delta_c - \delta_w)^2$$

where T is the precipitation temperature, δ_c is the measured δO^{18} of CO_2 derived from the calcite and δ_w is the measured δO^{18} of CO_2 in equilibrium with the precipitation fluid (Shackleton & Kennett, 1975). Assuming a pre-glacial δ_w of -1.2 per mil. it is possible to calculate the temperature at which the calcites finally equilibrated.

The limestones fall within a temperature range of 24.4-32.3^{o}C and, as one might expect, the highest temperatures are recorded by the oldest samples, those from the Gianni Griecu Member of the Sirino Formation. However, samples from one section may show a range of temperatures that do not correlate with depth; in the Monte Nicola and Pignola-Abriola Sections, the respective ranges are 27.5-31.6^{o}C and 24.4-31.4^{o}C. Since the samples were collected over stratigraphic distances of less than 100 m, it is clear that temperature was not the only control upon δO^{18}, particularly since the highest temperature in the Pignola-Abriola Section II is recorded in the youngest sample. One must assume, therefore, that beds have reached different stages of diagenesis which do not relate to depth, or that the pore-water compositions were not constant, or that the calcites were not always in equilibrium with them. It is possible that all of these mechanisms played a part in modification of the δO^{18}

values; the variable development of microspar suggests that late stage neomorphism occurred after the rock was finally lithified, and considerable modification of the isotopic composition of the pore waters is likely to have occurred during diagenesis. The progressive devlopment of lighter diagenetic calcites at higher temperatures probably caused a corresponding enrichment in the pore fluids, which would have effectively buffered the composition of subsequent precipitates. This effect has been observed in deep-sea sediment cores, and explains why O^{18} depletion in calcites with depth is not always as great as would be expected on the basis of calculation of the thermal gradient (McDuff et al., 1978; McKenzie et al., 1978). It is likely, therefore, that the palaeotemperatures implied by the δO^{18} values of these limestones are anomalously low, but they confirm the textural evidence that diagenesis did not occur at abnormally elevated temperatures. Choquette (1968) documents a similar δO^{18} depletion with the development of microspar, with negligible change in the δC^{13} value; this author advocates an evolution of isotopically lighter formation waters, however, rather than an increase in temperature.

Summarising the evidence presented, it appears that some of the limestones of the Sirino Formation may have experienced some early lithification by low-magnesian calcite; this is reflected only in the Sr^{++} concentration data, however, the isotopic compositions documenting a progressive re-equilibration with pore waters at higher temperatures during diagenesis. Temperatures and pressures during this process were insufficient to produce an equidimensional pavement mosaic, but coalescing neomorphism has fostered the development of an amoeboid mosaic of irregular micrite and microspar. The variable diagenetic maturity and inhomogeneity of these limestones are probably reflected in the poor correlation of δO^{18} with depth.

8.3.2.2. Cherts of the Sirino Formation

The preservation of primary sedimentary textures in the cherts, and their nodular development, implies that they are of secondary replacement origin, and the plethora of radiolarian moulds in the limestones points to a biogenous origin for the silica (c.f. Wise & Weaver, 1974). The homogeneous microcrystalline quartz probably developed by a single process of replacement by opal-CT lepispheres, forming a featureless procellanite that subsequently inverted to quartz (c.f. Heath & Moberly, 1971; Wise & Weaver, 1974; Hein et al., 1978; Robertson, 1977). Where ghosts of radiolarian are preserved, an early quartz cementation phase is probably indicated which developed while pore-waters were still undersaturated with respect to opal-CT. The horizontal-planar distribution of the nodules suggests that conditions were more conducive to silicification at certain levels in the sediment pile. This may be a reflection of primary sediment composition or of pore-water control, but there is no evidence that cherts are any more plentiful in beds that contain abundant Radiolaria. Reworking of some nodules attests to their early development, and dissolution of unstable carbonate components near the sediment water interface may have stimulated their formation by releasing Mg^{++} and OH^{-} ions upon which opal-CT is thought to nucleate (Kastner et al., 1977). Drilling in the peri-platform oozes of Northeast Providence Channel, however, has revealed no such relationship, the youngest cherts found being of early Eocene age, whilst no high-magnesian calcite was encountered in sediments below the Pleistocene (Hollister, Ewing, et al., 1972). It is alternatively possible that nodule development may relate to lithology in a more subtle way, as a function of permeability, since abundant migration of pore fluids is necessary.

The zonal nature of the nodules implies accretionary growth, silicification proceeding from the core along reaction 'fronts' (Fig.8.8.) In the cortex, dissolution of $CaCO_3$ and precipitation of SiO_2 occur sympathetically, as nucleation of opal-CT on hydroxyl ions reduces alkalinity that in turn fosters micrite dissolution and production of new nuclei. Expulsion of bicarbonate, calcium and magnesium ions encourages the precipitation of calcite and dolomite in the perifery of the nodule (c.f. Wise & Weaver, 1974); forming principally in radiolarian moulds, the cements gradually develop into rhombs in which the outlines of the Radiolaria are commonly preserved (c.f. Mišík, 1973). Where pore-waters are undersaturated with respect to calcite, dolomite probably precipitates preferentially. These ferroan dolomites may later be dedolomitised, expelling Fe^{++} in the process which is preserved as iron oxides. The low Sr^{++} contents of these opaque calcite grains is typically low (c.f. Shearman & Shirmohammadi, 1969). Migration of the silicification front through the calcite mosaic causes corrosion of the rhombs, but preservation of dolomite suggests that solutions were still oversaturated with respect to that mineral. The zonal replacement fabrics of calcite rhombs indicates that zones of certain compositions were less stable, perhaps those in which substitution of Mg^{++} for Ca^{++} ions had produced a less-ordered lattice structure.

The vertical veins that commonly pervade chert nodules are probably a diagenetic feature that relate to shrinkage during solution and reprecipitation of $CaCO_3$ and SiO_2 (c.f. Wise & Weaver, 1974). The development of calcite rhombs along these veins shows that they acted as pathways for dissolved $CaCO_3$, the solutions that they contained being sufficiently supersaturated at times to allow precipitation of calcite. During subsequent phases of silicification

however, they became filled or replaced by chalcedony, causing partial corrosion of the rhombs.

In banded cherts, migration of the silicification front was periodically checked, development of the calcitic mosaic preventing migration of SiO_2 in solution into the nodule core. Re-nucleation on the rim of the nodule produces a secondary cortical layer, and further silicification in turn generates another peripheral zone of rhombic calcite. The process may have been repeated several times in the case of banded cherts until the local source of biogenous opal had been exhausted. Episodic re-precipitation of dissolved carbonate species is also evinced in the banded dolomitic cherts; pore-water saturation with respect to dolomite periodically increased until such times as precipitation of dolomite caused it to drop again rapidly. Rupturing of some of the cherts during diagenesis allowed supersaturated fluids to escape into the overlying micrite where they precipitated as 'clouds' of dolomite rhombs,

Dolomite rhombs are commonly found in replacement cherts, even when the host rock is formed uniquely of limestone (Dietrich et al., 1963; Lloyd & Hsü, 1972; Mišík, 1973; Jacka,1974; Wachs & Hein, 1974; Price, 1977b). Dolomitic cherts have also been encountered in deep-sea sediment cores from the Atlantic and Pacific Oceans, where they occur as nodules of porcellanite or quartzose chert containing scattered rhombs of dolomite (Maxwell et al., 1970; Lloyd & Hsü, 1972; Keene, 1975). Since the rhombs are seldom corroded, it must be assumed either that chertification stimulates dolomite formation or else that diagenetic conditions favourable to silicification are equally agreeable for dolomite. However, it is not altogether clear at what time during chert formation the dolomite developed; Keene (1975)

considered that the presence of rhombs in the chalk surrounding nodules as well as within them, implied that silicification occurred after dolomitisation whereas Kastner et al. (1977) suggested that the rhombs formed as a result of expulsion of magnesium from opal-CT during its conversion to quartz. The recognition that chert nodules 'grow' however, implies that the presence of dolomite both within and outside the nodules is not at variance with a possible genetic relationship between dolomitisation and silicification, and rhombs could be precipitated ahead of the silica front while the nodule was still forming (c.f. Wise & Weaver, 1974).

Where rhombic calcite and dolomite coexist, as is generally the case, the latter may have formed simply to accommodate any magnesium released during the replacement of micrite. Although both minerals are ferroan, the dolomite lattice appears able to accommodate more Fe^{++} than calcite; explusion of iron during dedolomitisation promotes the formation of iron oxides, which give the rhombs their characterisitic opaqueness. Commonly, however, dolomite occurs alone, suggesting that pore fluids are undersaturated with respect of calcite. Furthermore, the amount of dolomite in many nodules suggests that far more Mg^{++} ions were available than could have been derived simply from the replacement of micrite. Although in the case of these peri-platform sediments, inversion of a possible high-magnesian calcite component can be invoked, the presence of dolomitic chert nodules in nannoplanktonic pelagic oozes implies that the source of magnesian must lie in the marine pore-waters. How, then, does chert formation produce conditions favourable for dolomitisation?

The isotopic composition of cherts commonly reveals a prominent meteoric water influence, and it has been suggested that mixing of

meteoric water with seawater can generate conditions simultaneously favourable to chert formation and dolomitisation (Knauth & Epstein, 1976; Knauth, 1979)(See Section 8.4.3). In the words of Knauth (1979):

> 'The included dolomite should show isotopic evidence of having precipitated from meteoric waters low in O^{18} content.'

The δO^{18} of the samples analysed do indeed show considerable depletion, particularly if a +4 to +7 per mil fractionation relative to calcite is assumed (See Section 8.4.3.); their non-stochiometric composition, however, suggests that they may have precipitated as proto-dolomites, in which case the fractionation may have been only +3 to +4 per mil (Fritz & Smith, 1970). Similar δO^{18} values have been reported for dolomitic cherts recovered from deep-sea sediment cores recovered off the west coast of North Africa and from the Lower Cretaceous Tamaulipas Formation of Central Mexico (Lloyd & Hsü, 1972). But the rhombic calcite has a similar composition to the dolomites, and both calcite and dolomite are within the range of the host sediments, although the δO^{18} values of the latter have probably been modified during late-stage diagenesis (See Section 8.3.2.2). Despite the O^{18} depletion of the dolomite, there are several arguments against a meteoric water influence during chert development. Although secondary dolomites are present in the formation in parts of Unit II, cherts occur in almost equal abundance throughout the formation, and the dolomitic cherts occur only in Unit I, which has escaped secondary dolomitisation, Furthermore, if meteoric water was available to stimulate chert formation, one might expect to see a greater depletion and range of δO^{18} values in the host limestones, whereas in fact the isotopic compositions of biomicrites associated with dolomitic chert are no different from the average. Also, the Sr^{++}

compositions of both calcite and dolomite are comparable to the expected values for inorganic precipitation from seawater, and are significantly higher than those recorded from secondary dolomites for which a meteoric water influence has been invoked (Kinsman,1969;Behrens & Land,1972;Land et al,1975; Randazzo & Hickey, 1978)(Table 8.4). Finally, it is difficult to conceive of a mechanism whereby meteoric water could be introduced into the sediment pile at an early stage throughout the basin, since there is no evidence that the adjacent carbonate platforms were exposed during the Triassic (d'Argenio, 1976) (See section 8.4.4.); in the case of dolomitic chert nodules in deep-sea sediments far removed from land and which have never been subaerially exposed, meteoric water diagenesis is even less credible, particularly when data are available on the marine nature of the coexisting pore-waters (Manheim et al., 1970; Lloyd & Hsü, 1972: Keene, 1975).

It is suggested, therefore, that undersaturation with respect to calcite, coupled with the development of high concentrations of bicarbonate, Ca^{++} and Mg^{++} ions in solution due to the replacement of micrite, and in the latter case to the inversion of high-magnesian calcite, causes precipitation of dolomite. The evidence that opal-CT nucleation causes a reduction in alkalinity may explain the reluctance of calcite to precipitate. The O^{18} depletion of the dolomite is difficut to explain, but similar depletion in the rhombic calcite suggests that no fractionation occurred, and it is probable that both precipitated out of equilibrium.

8.3.2.3. Limestones of the Monte Facito Formation

The isotopic compositions of the Monte Facito Formation limestones show a similar trend towards O^{18} depletion to those of the Sirino Formation, even in the case of the nodular varieties , which might have been expected to preserve some indication of early lithification (c.f. Hudson & Coleman, 1978). It appears, therefore, that diagenetic modification at this level was even more intense, possibly a reflection of the proximity of the basal thrust unit rather than simply depth of burial. The isotopic composition of the Sicilian nodular limestone, on the other hand, is more compatible with early submarine cementation (c.f. Hudson & Coleman, 1978; Schlager & James, 1978). In the case of the neritic limestones, evidence for palaeokarstification implies that the δO^{18} depletion may have been caused by meteoric water diagenesis (See Chapter 4).

The O^{18} enrichment of the dolomitic matrix in relation to the calcite nodules of the Valle del Pesce Member may be the result of different fractionations between calcite and water, and proto-dolomite and water (Fritz & Smith, 1970). Alternatively, the dolomite may have formed at a relatively early stage and has resisted subsequent modification (c.f. Choquette, 1968). Whatever fractionation factor is accepted, the measured δO^{18} is clearly low in comparison to recent supra- and sub-tidal dolomites and may betray the influence of meteoric water (Fig.8.11); the significance of dolomitisation is discussed more fully in Section 8.4.

8.4. Dolomitisation of basinal limestones in the Lagonegro Zone

In the northern part of Unit II, limestones of the Sirino and Lagonegro Formations are extensively dolomitised. Stratigraphic, petrographic and geochemical data are discussed and interpreted in terms of a mixed meteoric/sea water model.

8.4.1. Field relations and petrography

8.4.1.1. The Sirino Formation

The thinly bedded calcilutites, calcarenites and calcirudites of the Sirino Foramtion in Unit II north of Marsico Nuovo have undergone pervasive dolomitisation; in the central part of the Zone, dolomites are only present at the top of the formation, but in the north, the entire exposed sequence is involved. The contact between the dark-grey lime-mudstones and cream-coloured dolomites is razor sharp but highly irregular and hence rarely conformable with the bedding; in the Pignola-Abriola Section II, the vertical contact follows a zig-zag course up the section (Pl.8.16 & 8.17). Isolated patches of limestone may be entirely surrounded by dolomite. Bedding planes and rows of chert nodules are continuous across the contacts, the latter being of identical appearance in both lithologies. The dolomites are highly fractured in comparison to the limestone, and are pervaded by an anastomosing network of closely spaced horizontal veins of coarse, white dolomite up to 5 mm thick (Pl.8.18 & 8.19): The rocks have a coarse, saccharoidal texture and are coarser-grained than the adjacent limestones. The extraformational carbonate breccias at the top of the formation are similarly dolomitised but the outlines of clasts can be recognised (Pl.5.24 & 5.25); these also manifest a saccharoidal texture.

In the La Ralla Section, the Sirino Limestones are completely dolomitised and bedding can be distinguished only with difficulty. The nature of the original sediments is revealed, however, by the presence of beds and nodules of chert. Field relations are obscured by faulting and the succession is better exposed on the west face of Monte Pierno, 7 km southeast of San Fele.

In thin section, the dolomites are seen to comprise a mosaic of subhedral, yellow/brown crystals about 250μ across that are of almost unitform size; their colour is imparted by abundant dark inclusions, which are particularly numerous at the centre of the crystals (Pl.8.20); staining reveals that they are non-ferroan. A lighter-coloured dolomite band 50μ wide separates the dolomite mosaic from the radiolarian biomicrite (Pl.8.21). This band is almost straight and invariably normal to bedding, but may pass into a more diffuse contact zone 250μ wide in which dolomite and biomicrite are separated by patchy calcite microspar. A set of identical bands a few millimetres apart and parallel to the contact can be observed in the dolomite (Pl.8.21). Within the biomicrite, scattered irregular cavities are developed, which are lined by sparry calcite and filled by dolomite. A network of calcite-filled veins is developed normal to bedding that are also parallel to the contact.

The horizontally veined dolomites are composed of a similar mosaic of smaller crystals about 50μ across (Pl.8.22). The dolomite in the veins shows a typical void-filling texture, crystals increasing in size towards the centre; progressive growth stages are preserved by zones of inclusions but the last precipitated dolomite is almost clear. The veins either merge imperceptibly into the mosaic or are bordered by stylotites.

Chert nodules in the limestone and dolomite are identical, and ghosts of Radiolaria and pelagic bivalves, filled or underlain by chalcedonic cements, can be recognised in each case. Outlines of relict dolomite rhombs have not been observed in the microcrystalline quartz groundmass. Cherts from the extraformational carbonate breccias show similar fabrics.

8.4.1.2. The Lagonegro Formation

Dolomitisation of the calciturbidites of the Lagonegro Formation is abundant in the lower part of the La Ralla Section, and has affected almost all beds to some degree. In contrast to the Sirino Formation, however, it is pervasive, and although beds may be only partially dolomitic, sharp limestone/dolomite contacts have not been observed. Instead, entirely, partially and non-dolomitic beds alternate almost at random through the section, although at the base the majority are completely dolomitised. The dolomitic beds may be distinguished by their paler colour and saccharoidal tecture. Chert nodules are present in both lithologies.

The distribution of dolomite in mixed calcite/dolomite beds can be highlighted by staining with alizarin-red, which reveals that the clasts are predominantly calcitic whilst dolomite is concentrated in the interstices (Pl.8.23). In thin section, it can be seen that clear, euhedral rhombs of dolomite between 100 and 500μ across are scattered randomly within the sparry calcite that cements the grains (Pl.8.24); they are excluded, however, from the intraclastic cements. The rhombs may entirely obliterate the contacts between grains and cement, and in some cases are also present within the clasts. Some of the youngest beds, which contain only small amounts of interstitial dolomite, may comprise clasts that are composed entirely of

fine-grained dolomite. Staining indicates that all rhombs are non-ferroan and that some are calcitic; the latter either comprise single crystals or a fine-grained mosaic.

In the more fine-grained calcarenites, rhombs may occur equally within grains or cement (Pl.8.25); in the most dolomitised beds, they make up almost all of the rock, pelletal limestone being preserved only as small patches within the mosaic (Pl.8.26).

8.4.1.3. Platform dolomites

No study was made of the field relations in the dolomitic platform lithologies. Dolomites, however, predominate in the Triassic parts of the succession in both the Campania-Lucania and Abruzzi-Campania Platforms (Ippolito et al., 1975); this relationship is common to the majority of peri-Adriatic carbonate platforms (d'Argenio, 1976). In the Triassic, the dolomites may be associated with evaporites at the base of the platforms and, particularly in the Norian, with supra-tidal cyclic carbonate facies or Lofer Cyclothems (Fischer, 1964; d'Argenio, 1976); these dolomites are commonly fine-grained and bedded, and occur primarily in the central parts of the platforms. Massive dolomites, however, characterise the platform margins and may occur within a wide variety of limestone facies (d'Argenio, 1976). In some cases it is possible to recognise rudite fabrics, but primary texture is generally poorly preserved (Scandone & Bonardi, 1968). In thin section, these white dolomites comprise crystal mosaics similar to those of the Sirino Formation. Such rocks dominate the Triassic part of the Monti della Maddalena Unit, and samples were collected from this unit west of Brienza.

8.4.2. Results of geochemical analysis

8.4.2.1. Electron microprobe

Dolomites from the Sirino and Lagonegro Formations were analysed for comparison; the samples selected were a horizontally veined rock from the former and an incipiently dolomitised calciturbidite of the latter, collected from the Pignola-Abriola II and La Ralla Formations respectively. Ca, Mg, Fe, and Sr were analysed in order to determine the composition of the carbonate phases, and Si was measured to ensure their purity. The results of the Ca, Mg and Fe analyses are represented diagramatically in Fig.8.9, and it can be seen that the compositionsof the two dolomites are quite different; as was suggested by staining, neither are ferroan. The dolomites from the Sirino Formation are almost stochiometric and have an average Ca:Mg ratio of 51.3:48.7. Rhombs from the calciturbidite, however, are much more calcian, ranging from 55 to over 60 mol percent calcium, with an average of 56.9. Comparison of the means of the two groups using a Student t-test reveals that they are significantly different at the 99.5% confidence level.

The Sr contents ranged from 0 to 300 ppm and 0 to 100 ppm for the Sirino and Lagonegro samples respectively,with averages of 65 and 70 ppm; they are not therefore significantly different (Table 8.3). Undolomitised clasts and cements of the calciturbidite, however, showed Sr contents of 90 to 570ppm, with an average of 240 ppm.

Table 8.3. Range and mean of strontium contents of dolomites and calcites of the Sirino and Lagonegro Formations

Sample	Sr ppm High	Sr ppm Low	Sr ppm Mean
Sirino Formation dolomite - 21 analyses	290	0	65
Lagonegro Formation - dolomite - 14 analyses	110	0	70
Lagonegro Formation - calcite - 10 analyses	570	90	240

8.4.2.2. Stable isotopes

The results of oxygen and carbon isotope analyses are displayed in Fig.8.10. It can be seen that some samples are markedly depleted in oxygen, but carbon values, ranging from +1 to +3 per mil, show only minor variation. The δO^{18} values of samples collected from the Pignola-Abriola Section II and in the vicinity of San Fele fall into two distinct groups, the latter being significantly lighter; the averages for the two groups are -7.97 and -0.60 per mil respectively, which are significantly different at the 99.5% confidence level when compared using a Student t test. Of the Pignola-Abriola samples, the most depleted in O^{18} was that collected from the horizontal veins and is 1 per mil lighter than the enclosing rock; the breccia is the most enriched sample. Dolomites from the La Ralla group show a similar spread of values and an average O^{18} enrichment of 0.4 per mil relative to co-existing calcites in partially dolomitised beds. Sirino Formation dolomites contain even heavier oxygen, although the breccia has a δO^{18} of -1.7, lighter than any of the calciturbidites. 2 dolomites from the carbonate platform are slightly heavier still, δO^{18} being positive in each case.

δC^{13} values in each group cover a similar range and show a slight positive correlation with δO^{18} values. In all cases, co-existing calcites are relatively depleted in C^{13}, and dolomites and calcites from the calciturbidites have δC^{13} values of +2.1 and +1.3 per mil respectively.

Dolomites from the Scillato and Fanusi Formations of the Madonie Mountains, Sicily, were analysed for comparison. δC^{13} values are comparable with those of the Sirino Formation, but δO^{18} ranges from -5 to 0 per mil. All these samples were collected from the same type succession on Monte Fanusi, but the stratigraphically higher Fanusi Formation dolomites show the greatest depletion (See Chapter 9 and Schmidt di Friedberg et al., 1960, for details of the section).

8.4.3. Interpretation

The transgressive nature of the limestone- dolomite contact, the patchy development of dolomite, the continuity of bedding and bands of chert nodules across contacts and the preservation of relict textures of Radiolaria and pelagic bivalves in cherts surr-ounded by dolomite, all imply that dolomitisation of the Sirino Formation was of secondary, replacement origin. Similarly, the documentation of progressive replacement of primary textures, and their preservation in chert nodules, supports the contention that dolomitisation of the La Ralla calciturbidites was also secondary. The dolomite and biomicrite textures adjacent to contacts in the Pignola-Abriola Section II give the impression that replacement occurred episodically and that vertical calcite veins acted as barriers to migrating dolomitising fluids; since the veins can only have developed after considerable induration of the calcareous ooze,

it is clear that dolomitisation took place long after deposition. The development of dolomite-filled cavities ahead of the dolomitisation front attests to some dissolution of the lime mudstone, and fabrics along non-veined contacts suggest that micrite is replaced by microspar prior to dolomitisation. Judging by the irregular, step-wise nature of the contact, and the evidence that vertical calcite veins acted as barriers, it appears that fluids migrated horizontally along the bedding planes.

The horizontal veinlets that are commonly developed in the dolomites have no counterparts in the limestones, and their anastomosing tecture rules out any relationship to primary bedding. Their void-filling texture and relatively constant thickness suggest that they developed as tensional cracks rather than by solution; since they merge with the mosaic of the surrounding rock, rather than nucleate on planar substrates, it seems probable that they formed after dolomitisation of the surrounding rock, and that the two processes are genetically related. Their consistent horizontal orientation implies removal of the vertical stress component and it is suggested that this was achieved as a result of a volume decrease during dolomitisation. The volume of the veins in proportion to the total volume of the rock, however, is greater than can be explained by the 10-13% molar difference between calcite and dolomite; the veins are clearly not simply the product of local shrinkage, therefore. The patchy nature of dolomitisation in the section where these veins are found suggests the possibility that masses of 'spongy' dolomite, with high secondary porosity and which were surrounded by relatively cohesive lime mudstones, subsided under their own weight, causing the formation of tensional cracks in the upper part of the brittle 'sponge'; vertical stress was taken up by

the overlying mudstones until such time as the veins were finally cemented up by dolomite. Filling of the cracks, and the elimination of secondary porosity in the mosaic, imply an extraneous source of carbonate that was rich in Mg as well as Ca.

In the Lagonegro Formation, dolomitisation of calciturbidites followed a different pattern. Vertical limestone/dolomite contacts are absent, and beds contain varying proportions of calcite and dolomite. The preferential growth of rhombs between rather than within clasts is interpreted as a reflection of porosity contrasts rather than a greater propensity for dolomitisation on the part of the cements, particularly since intraclastic cements are preserved. The pervasive distribution of the rhombs throughout the beds also supports the contention that dolomitising fluids enjoyed unimpeded movement through the rock; dolomitisation may thus have proceeded more rapidly than in the Sirino Formation. It is likely that submarine cementation of the calciturbidites would have been sluggish, in view of the sub- CCD depositional depths (See Chapter 6); on the other hand,pore waters would have become rapidly supersaturated with respect to calcite if the sediments had originally contained a high proportion of metastable carbonate components (c.f. Scholle, 1971b). Some early cementation is evinced by the lack of evidence for compaction or pressure solving of grains. In summary, therefore, it appears that the different dolomite textures developed in the two formations are principally a reflection of porosity control upon rates of pore fluid migration.

The timing of dolomitisation is difficult to determine precisely. In the Lagonegro Formation, dolomite may occur in beds as young as Late Jurassic in age, and even in the youngest dolomitised beds, rhombs

are almost entirely absent from the chert nodules, which preserve primary textures. In all cases, therefore, it appears that dolomitisation postdated silicification. Geochemical and field data suggest, however, that chert nodule formation can occur relatively rapidly in carbonate sediments and may, in some cases, even predate soft sediment deformation (Bernoulli, 1972; Kastner et al., 1977) (See Chapter 6). The preservation of primary fabrics in cherts may not, therefore be a significant constraint upon the timing of dolomitisation.

The presence of dolomite clasts in some of the younger calciturbidites suggests that parts of the platform from which they were derived were already dolomitised by the Late Jurassic. There is no record of dolomite in the calciturbidites of the Brusco Formation in either of the Lagonegro Units, or in any of the younger formations of the zone (Ippolito et al., 1975). In the Pignola-Abriola Section II, the oldest dolomites are Late Carnian or later in age (de Capoa Bonardi, 1970). It may be inferred, therefore, that dolomitisation occurred at some time, or several times, between the Late Triassic and the Late Jurassic; dolomites in different areas, however, did not necessarily form at the same time or even during the same dolomitising event.

An indication of the nature of the dolomitising fluids can be obtained from the geochemistry of the dolomite. Using the appropriate partition coefficient, strontium contents can be used to determine the strontium:calcium ratios of the precipitating fluids (Kinsman, 1969) (See Section 8.3.2.1). The partition coefficient for dolomite is unknown, but it has been predicted that dolomite should contain about half as much strontium as a coexisting calcite; thus a dolomite precipitated in equilibrium from seawater should contain about

600 ppm (Behrens & Land, 1972). Modern supratidal dolomites of the Persian Gulf indeed contain Sr concentrations of this order, and Holocene subtidal examples from Texas are even higher (Behrens & Land, 1972)(Table 8.4). It is clear, therefore, that the averages of 65 and 70 ppm for the Sirino and Lagonegro Formation dolomites are too low for them to have been precipitated from seawater. These values are more comparable with those of dolomites for which a meteoric water influence has been proposed (Behrens & Land, 1972; Badiozamani, 1973; Land, 1973a, 1973b; Land et al., 1975; Dunham & Ohlson, 1977; Randazzo & Hickey, 1978)(Table 8.4).

Table 8.4. Mean strontium concentrations of recent and ancient dolomites

Description and reference	Strontium (ppm)
Supratidal dolomite; Holocene: Florida and Bahamas (Behrens & Land, 1972)	630
Supratidal dolomite; Holocene: Persian Gulf (Behrens & Land, 1972)	640
Subtidal dolomite; Holocene: Baffin Bay, Texas (Behrens & Land, 1972)	916
Dolomite from meteoric water, Ordovician, Wisconsin (Badiozamani, 1973)	37
Dolomite from meteoric water; Eocene; Egypt (Land, Salem & Morrow, 1975)	90
Dolomite from meteoric water, Ordovician, Canada (Land, Salem & Morrow, 1975)	81
Dolomite from meteoric water; Ordovician/Silurian, Nevada (Dunham & Ohlson,1977)	55
Stochiometric dolomite from meteoric water; Eocene; Florida (Randazzo & Hickey, 1978)	88
Non-stochiometric dolomite from meteoric water, Eocene; Florida (Randazzo & Hickey, 1978)	190
Dolomite; Triassic; Sirino Formation, Lagonegro Unit II (This study)	65
Dolomite; Jurassic; Lagonegro Formation, Lagonegro Unit II	70

The possibility that groundwaters might play a major part in dolomitisation was first suggested by Hanshaw et al., (1971) who

noted that Mg:Ca ratios in the aquifers of the Floridan and Yucatan carbonate platforms approach unity, which represents the point at which calcite, dolomite and water are probably in equilibrium. In the zones of brackish water that underlie these fresh-water lenses, Mg:Ca ratios are almost certainly in excess of 1:1, and it has been suggested that circulation of groundwaters may be responsible for effectuating dolomitisation on a massive scale. A rationale for this process was provided by Folk & Land, (1975), who proposed that, given the necessary time, dolomite may form from solutions with Mg:Ca ratios as low as 1:1 provided that the concentration of foreign ions, which tend to hinder the precise Ca/Mg ordering required, is low; dilution of seawater by meteoric water in these mixing zones has the effect of reducing the absolute concentrations, whilst leaving the Mg:Ca ratio unchanged. Because of the low rate of supply of Mg^{++} ions, dolomitisation is necessarily slow, encouraging the development of large, limpid crystals of stochiometric composition.

The dolomites of the Sirino Formation in the Pignola-Abriola Section II contain almost equal proportions of calcium and magnesium, whereas those of the Lagonegro formation at La Ralla are clearly non-stochiometric. Ordering in the latter may have been hindered by high concentrations of foreign ions, suggesting that the precipitating fluids were more saline; it is surprising, therefore, that there is no significant difference in the strontium concentrations of the two types, since they might be expected to be higher in the non-stochiometric dolomites (c.f. Randazzo & Hickey, 1978). It may be, however, that the resolution of the microprobe is not sufficiently fine to discriminate at these levels.

The stable-isotope analyses may also indicate a meteoric water influence. Comparison with values for modern supra- and sub-tidal

dolomites, shows that all the Lagonegro samples are relatively depleted in O^{18} (Fig.8.11). It is possible, therefore, that dolomitisation by waters of less than normal marine salinity, rather than by hypersaline brines, may have been involved (c.f. Adams & Rhodes, 1960; Gross & Tracey, 1966; Choquette, 1968; Fischer & Rodda, 1969; Veizer, 1970; Badiozamani, 1973; Land, 1973a, 1973b; Behrens & Land, 1972; Land et al., 1975; Dunham & Ohlson, 1977). Alternatively, low δO^{18} values may simply reflect the temperature of diagenesis. Interpretation of δO^{18} for dolomite is confounded by the lack of agreement concerning its possible fractionation with respect to calcite. Extrapolation of high-temperature experimental data suggests that dolomite should be enriched by 4-7 per mil relative to calcite precipitated from the same water (Epstein, 1959; Clayton & Epstein, 1958; Northrop & Clayton, 1966; O'Neil & Epstein, 1966). Study of recent coexisting calcites and dolomites, however, have shown that such a relationship rarely pertains in reality (Friedman & Hall, 1963; Degens & Epstein, 1964; Northrop & Clayton, 1966); this has been taken as evidence that modern dolomites do not represent primary precipitates but replace pre-existing carbonates (Degens & Epstein, 1964). Modern supra- and sub-tidal dolomites, however, generally precipitate as poorly-ordered proto-dolomites, due probably to the influence of foreign cations outlined above (Goldsmith & Graf, 1958); it has been suggested that fractionation relative to calcite in this case may be lower (Northrop & Clayton, 1966). Experiments of Fritz & Smith (1970) seem to confirm the suspicions of Northrop & Clayton: proto-dolomite was enriched in O^{18} by only 2-4 per mil relative to calcite. This value was also determined empirically from the measurement of the isotopic compositions of sub-tidal Holocene dolomites from Texas (Behrens & Land, 1972).

Compared with co-existing calcites, dolomite from the calciturbidites of the Lagonegro Formation show an average enrichment in O^{18} of 0.4 per mil. The oxygen and carbon isotope values for the undolomitised limestones are almost within the range of recent shallow and deep-water carbonate sediments and negligible meteoric water diagenesis is evinced (Milliman, 1974; Hudson, 1977; Milliman & Müller, 1977; c.f. Choquette, 1968)(Fig.8.7.), Microprobe data indicate that the co-existing dolomites are highly calcian and it is likely that they precipitated as protodolomite; accepting a 3-4 per mil fractionation, it appears either that meteoric water has affected the dolomite to a greater degree than the calcite or else that the dolomite precipitated at a higher temperature. In the case of the Sirino Formation dolomites of the Pignola-Abriola Section II, relative depletion of O^{18} relative to the radiolarian biomicrites and intrabiosparites is apparent, even without taking account of any possible fractionation effects; their more nearly stochiometric compositions suggests that these dolomites actually precipitated as ordered crystals, in which case the fractionation would have been even greater than in the case of the Lagonegro Formation dolomites at La Ralla. Furthermore, the δO^{18} of the limestones themselves in the Pignola-Abriola Section II may have become more negative as a result of diagenetic re-equilibration at high temperatures (See Section 7.3.2.1.). A considerable meteoric water influence in this area may be implied by these values, therefore. The most depleted sample is from the late-stage horizontal veins, suggesting that the dolomitising fluids possibly become progressively less 'marine' during the process.

The contrast between the δO^{18} values of dolomite from the two localities shows no relationship to primary rock type, since samples from both the Sirino and Lagonegro Formations in the San Fele area fall

within the same range. There are two possible explanations for this regional discrepancy; the pore fluids at San Fele, which appear to have been isotopically heavier , either contained a higher proportion of sea-water or else their δO^{18} had been modified by water-rock interactions in some way. The calcian composition of the calciturbidite dolomites favour a more saline water, but this is at variance with the Sr data. The alternative is that leaching of metastable carbonate components, which may have been abundant in the calciturbidites, fostered the maintenance of 'marine' δO^{18} values, even though the fluids were not especially saline. Since the Sirino Formation dolomites show a similar depletion, the latter hypothesis seems untenable. It is therefore suggested that meteoric water exerted a greater influence upon dolomitisation in the Pignola-Abriola region than at San Fele.

The δC^{13} values show only minor variation in comparison to δO^{18} and are within the range of most modern carbonate sediments (Hudson, 1977; Milliman & Müller, 1977)(Fig.7.10). Atmospheric CO_2 is in equilibrium with dissolved bicarbonate with a δO^{13} between +1 and +2 per mil (Emrich, et al., 1970); meteoric water would not be expected, therefore, to have influenced the isotopic compositions of carbonates significantly, unless any additional CO_2 had been derived from the soil; the latter is organically derived and thus highly negative in carbon. However, δC^{13} should show an enrichment of 2.5 per mil relative to calcite although, as in the case of oxygen, fractionation may be less in the case of protodolomite (c.f. Fritz & Smith, 1970; Hoefs, 1973); this may be reflected in the slight enrichment of calciturbidite dolomites relative to co-existing calcites.

8.4.4. Discussion and model for dolomitisation in the Lagonegro Basin

Recognition for the potential of mixed saline and meteoric waters for producing chemical conditions favourable to dolomitisation, coupled with the knowledge that zones of brackish water underlie the fresh-water lenses of modern carbonate platforms, has inspired several recent models for sub-surface dolomitisation (Hanshaw et al., 1971; Badiozamani, 1973; Land, 1973a, 1973b; Land et al., 1975; Dunham & Ohlson, 1977, 1978; Randazzo & Hickey, 1978; Miller, 1979). The dynamic hydrological regimes of these platforms ensures that dolomitisation is effected on a massive scale; furthermore, the position of the mixing zone may fluctuate in response to changing climate and subaerial relief (Kohout, 1967; Back & Hanshaw, 1970; Hanshaw et al., 1971). Subterranean geology can also exercise a significant control on the distribution of the brackish zone.

The hitherto favoured 'seepage reflux' model involved sinking of dense, hypersaline brines, produced by evaporation in lagoons, into the underlying rocks, displacing the less-saline marine pore water and effecting dolomitisation (Adams & Rhodes, 1960; Deffeyes et al., 1965; Fisher & Rodda, 1969; Veizer, 1970). However, this model requires that large quantities of gypsum be precipitated in the lagoon in order to raise the Mg:Ca ratio to a level that will allow dolomitisation; the absence of associated gypsum or anhydrite deposits in the majority of cases therefore detracts from its credibility (Badiozamani, 1973; Folk & Land, 1975; Land et al., 1975; Dunham & Ohlson, 1978). The geochemistry of many dolomites is also commonly at variance with this model, although in some cases elevated salinities are indicated by high Sr^{++} and δO^{18} values (Behrens & Land, 1972).

A host of geochemical, petrographic and stratigraphic evidence has been martialled, however, in support of the 'mixing model'. Ancient dolomites commonly show Sr^{++}, O^{18} and slight C^{13} depletion relative to those currently forming under conditions of elevated salinity (Fig.8.11)(Badiozamani, 1973; Land, 1973a, 1973b; Land et al., 1975; Dunham & Ohlson, 1977; Randazzo & Hickey, 1978). The large size and euhedral habit of the crystal are thought to evince slow rates of precipitation from solutions with low Mg:Ca ratios, but relatively free of foreign ions (Folk & Land, 1975). On a larger scale, the geometry of several dolomite bodies has been shown to relate to features of elevated palaeo-relief, such as structural highs and shelf edges; moreover, migrations of the limestone/dolomite contact in response to regional transgression and regressions have been recognised (Badiozamani, 1973; Dunham & Ohlson, 1978; Miller, 1979). Land et al. (1975) have even identified palaeoflow directions on the basis of regional Sr^{++} and δO^{18} patterns which mirror the postulated flow regimes of modern platforms (Kohout, 1967).

The majority of these models for groundwater dolomitisation have involved the replacement of shallow-water carbonate lithologies in shelf areas at relatively moderate depths below sea level. If a similar model is to be applied to the limestones of the Lagonegro Basin, a far less conservative hydrological regime must be invoked to those envisaged hitherto. The freshwater lens of the Floridan Peninsula extends to depths of about 700m and is countered by a 15m 'head' of water above sea level (Back & Hanshaw, 1970). It is clear, however, that this zone extends a considerable distance out to sea, since freshwater submarine springs have been encountered on the continental slope at depths of up to 1 km and as far as 120 km from the coast (Kohout, 1967; Manheim, 1967); Manheim predicts that

with a piezometric head of 150m, discharges could occur as much as 2,000 m deeper. Back & Hanshaw (1970) have shown that the presence of a relatively impermeable layer high in the platform aids the development of a head of freshwater greater than that supported by a normally karstified surface; a similarly impermeable layer at the base of the aquifer would encourage lateral flow (c.f. Land et al., 1975). During the Jurassic, the Lagonegro Basin deepened from about 2 km to more than 4 km, so a considerable head of water would have been required to introduce freshwater into the basinal limestones (See Chapter 6); applying the Ghyben-Herzberg principle that every 40 m of freshwater below sea level, overlying saline formation water, requires a head of 1 m of fresh-water above sea level, implies that at least 50m of the neighbouring platforms must have been exposed. There is abundant evidence of stratigraphic hiatuses in both of the neighbouring platforms during the Jurassic, particularly in the Monti della Maddalena and Monte Croce Units which bordered directly onto the basin on each side (Scandone & Bonardi, 1968; Ippolito et al., 1975; d'Argenio, 1976); in parts of the former, Triassic and 'Infraliassic' dolomites are unconformably overlain by Maastrichtian shallow water limestones, implying protracted emersion during the Jurassic and Cretaceous, whereas in the latter, the entire Lias and Dogger are missing. Subaerial exposure of these areas is attributed to Liassic synsedimentary tectonics (Scandone & Bonardi, 1968). Furthermore, palaeo-continental reconstructions suggest that the platforms occupied latitudes, and enjoyed a climate, not unlike that of the modern Carribean (Smith & Briden, 1977). It is not unlikely, therefore, that a fresh-water aquifer developed in the carbonate platforms which extended to considerable depths; the impermeable shales of the underlying Monte Facito Formation may have

promoted extension of the aquifer some way out beneath the basin floor. The progressive reduction of platform relief and deepening of the basin, both in response to regional subsidence, may have caused the gradual cessation of dolomitisation by elevating the mixing zone above the level of the basin floor.

Geochemical, stratigraphic and petrographic data, all point towards a dolomitisation model involving mixing of sea water and fresh water recharged from the adjacent platforms (Fig.8.12). Tectonic elevation of the platform margins led to establishment of a fresh water lens and mixing zone that extended out beneath the basin floor for several tens of kilometres; dolomitisation occurred in the mixing zone, where pore waters were mildly super-saturated with respect to dolomite, and salinities were sufficiently low to allow ordering of the Mg and Ca ions. Regional variations in δO^{18} are thought to reflect contrasting basinal depths in the different localities following Early Liassic faulting (See Chapter 6); in the La Ralla area, where depths were greatest, only the base of the mixing zone reached the level of the basin floor, whereas in the shallower area around Pignola, basinal sediments impinged upon the higher, less-saline parts of the zone. The δO^{18} variations in the Fanusi and Scillato Formations of Sicily may similarly reflect palaeosalinities at different levels in the mixing zone. Reduction of platform relief and deepening of the basin in response to regional subsidence caused upward migration of the mixing zone such that dolomitisation of basinal sediments ceased in the Late Jurassic.

8.5. Implications of diagenesis and dolomitisation for the development of the Lagonegro Basin

It had been hoped that geochemical data on the limestones of the

Sirino Formation might provide constraints upon the nature of the parent ooze. It is evident that dissolution/reprecipitation and recrystallisation mechanisms are effective in destroying textural evidence; this is clearly demonstrated by the diverse geneses suggested for the Jurassic Solenhofen limestone, for which an origin as either an aragonitic mud or a nannofossil ooze have been argued (Hathaway & Robertson, 1961; Wise & Hsü, 1971). The paucity of traces of coccoliths cannot therefore be taken as *prima facie* evidence against a nannoplanktonic origin, since coccoliths may simply have been completely recrystallised (Bramlette, 1958; Fischer et al., 1967; Wachs & Hein, 1974; di Nocera & Scandone, 1977); judging by published studies, transmission electron microscopy seems to be a more powerful tool in the detection of nanno-organisms than use of the scanning electron microscope.

The Sr^{++} data are too few to provide much assistance, but the generally high values recorded from the rhombic calcites may possibly attest to an initially high content of aragonite, whose Sr^{++} became 'locked' in the nodules; they are certainly high in comparison to inorganically precipitated calcites from normal seawater. The isotopic data paint an equally dismal picture, the oxygen values showing a strong diagenetic overprint while the C^{13} contents simply reflect a normal marine origin, equally applicable for a nannoplanktonic ooze or an aragonitic mud. It is impossible, therefore, to make any valid deductions concerning the origin of these limestones on the basis of these geochemical data.

The evidence presented in support of a mixing model for dolomitisation suggests that those parts of Unit II affected were relatively close to an exposed platform margin. This conclusion is borne out by the presence also of extraformational carbonate breccias in these

areas. Since block faulting must have been responsible for uplifting the platform margin sufficiently to allow fresh water to flow out beneath the basin floor, restriction of dolomitisation to Unit II provides a further indication that only the western side of the Lagonegro Basin was significantly affected by the Liassic tectonics.

CHAPTER NINE

EVOLUTIONARY SYNTHESIS OF THE LAGONEGRO ZONE AND IMPLICATIONS FOR THE MESOZOIC AND CENOZOIC DEVELOPMENT OF THE ALPINE-MEDITERRANEAN REGION

9.1 Introduction

In the preceding chapters, the stratigraphy, structure and sedimentology of the Lagonegro Zone have been examined in order to determine the manner in which the basin originated, developed, was destroyed and finally incorporated into the nappes of the southern Apennines. To put the Zone into perspective, an attempt is made in Section 9.2 to synthesise the models presented into a palaeogeographic frame, and to relate the basin's history to that of the other structural-stratigraphic units of the region. It emerges that the network of platforms and basins from which these units were derived owed their origin to regional subsidence, and evidence is presented in Section 9.3 in support of the contention that this subsidence was brought about by extension and attenuation of continental basement; a comparison is made with the parallel evolution of the Bahamas. Finally, the significance of the Lagonegro Basin, and other 'pre-oceanic' basins of Sicily, Yugoslavia and Greece, is assessed in terms of the overall development of the Alpine-Mediterranean region during the Mesozoic and Cenozoic.

9.2 Synthesis of the Lagonegro Zone in relation to the other structural-stratigraphic units of the southern Apennines.

9.2.1 Middle and Late Triassic - origins of the basin

A sedimentological model for the Monte Facito Formation has been presented in Chapter 4; this relates the origin of the Lagonegro Basin

to the destruction of an embryonic carbonate platform under the influence of extensional tectonics. With the sole exception of the Alburno-Cervati Unit in Calabria, which comprises alternations of shallow-marine limestones and phyllites of Anisian and Ladinian age at the base, rocks of Middle Triassic age are not represented in the other structural-stratigraphic units of the southern Apennines (Bousquet and Dubois, 1967; Ippolito et al., 1975; Amodio-Morelli et al., 1979; Dietrich, 1979). It is not possible, therefore, to present a palaeo-geographic reconstruction for this period, but the similarity of the basal lithologies of the Lagonegro and Alburno-Cervati units suggest that this period saw the beginnings of widespread shallow-water carbonate sedimentation as the area subsided below sea-level. A phase of minor extensional tectonics during the Anisian and Ladinian caused periodic emergence and collapse of extensive tracts of the young platform which, as a result, ceased to accumulate neritic carbonate sediments. These areas formed the template upon which the enfolding pattern of basins and platforms subsequently developed.

By the Late Triassic, the area had acquired considerable topographic relief. While hemi-pelagic lime mudstones accumulated in the Lagonegro Basin at depths which approached the CCD towards the end of the period, a continuation of shallow-marine carbonate sedimentation is recorded in the majority of other units (d'Argenio et al., 1973; Ippolito et al., 1975; d'Argenio, 1976). The rapid rates of deposition maifested by these platforms (≤100-150 m/MY) are not only typical of much of the peri-Adriatic region, but of the Alpine Triassic carbonate platform facies as a whole, and they bear witness to dramatic basement subsidence over a wide area (Bosellini and Hsü , 1973; Bernoulli and Jenkyns, 1974; d'Argenio, 1976). Despite this regional

subsidence, the Late Triassic appears to have been tectonically quiescent in comparison to the preceding epoch. Since rates of deposition in the basin (13-26 m/MY) were so much lower than on the platforms, however, the basin grew progressively deeper at a substantial rate during that period.

The palaeogeography of the region in the Late Triassic was relatively simple, therefore, comprising a single basin, surrounded by carbonate platforms, that probably terminated to the north but which may have extended southwards into Sicily (Fig.9.1). Data relating to basin morphology are scarce in the cherty limestones of the Sirino Formation, but the presence of extraformational carbonate breccias in sections from the northern outcrop area of Unit II suggest the proximity of the platform margin; dolomitisation in these sections, though probably not achieved until the Jurassic, provides an additional testimony to a nearby platform. It was at this stage that the palaeogeographic configuration of the southern Apennine domains most resembled that of the modern Bahamas (Bernoulli 1972; d'Argenio, de Castro, et al., 1975).

9.2.2 Jurassic - renewed block faulting and further deepening of the basin

In the Early Jurassic, the palaeogeography became more complex as renewed extensional tectonic activity further subdivided the existing platforms and basins. In the Lagonegro Basin itself, some areas escaped these disturbances (Lagonegro Unit I) while others experienced varying degrees of block faulting and rapid subsidence (Lagonegro Unit II) (Fig.9.2). In addition, new basins developed both within and on the margins of the carbonate platforms, probably founded upon

subsided fault blocks; these second-order basins, represented by the Monte Bulgheria-Verbicaro, Monte Foraporta and Frosolone Units, began to accumulate both redeposited shallow-marine calciclastic and pelagic sediments. Parts of the Abruzzi-Campania platform, after foundering, may even have developed a seamount topography, for the Triassic shallow-marine carbonate succession of the Monte Croce Unit is succeeded by red nodular pelagic limestones comparable to those of western Sicily (Turco, 1976; c.f. Jenkyns and Torrens, 1971). On the other hand, block faulting caused parts of the Campania-Lucania Platform to become emergent for long periods, the post-Triassic succession of the Monti della Maddalena Unit being notably incomplete (Scandone and Bonardi, 1968). It is additionally possible that the Lagonegro Basin was extended northwards during this period, separating the Campania-Lucania and Abruzzi-Campania platforms, but there is no trace of any basinal units between the Alburno-Cervati and Matese-Monte Maggiore Units north of Naples to justify this contention. All the southern Apennine platforms and basins must have terminated along the Anzio-Ancona Line, which developed as a major fault scarp in the Liassic. To the northwest of this line, the carbonate platforms of the Triassic and Early Liassic were downfaulted en masse to form the Umbria-Marche Basin (Castellerin, Colacicchi and Praturlon, 1978).

There is evidence in the internal units and Calabride Complex for the opening of the Tethys Ocean to the west during the Jurassic. Ophiolite units, comprising pillow basalts overlain by Upper Jurassic radiolarites, Lower Cretaceous Calpionella limestones and black shales, are present at the base of the Calabrian nappes, and as olistoliths in the internal units, that are directly comparable to the Tethyan

oceanic sequences of the northern Apennines (Dietrich and Scandone, 1972; Amodio-Morelli et al., 1979; c.f. Decandia and Elter, 1969; Abbate and Saggri, 1970; Bernoulli and Jenkyns, 1974).

Sedimentation in the Lagonegro Basin occurred below the CCD throughout the Jurassic, and rates of deposition were therefore very low (approximately 1 m/MY and 6 m/MY in Units I and II respectively). Most parts of the surrounding platforms continued to accumulate shallow-marine carbonate sediments as regional subsidence, though not as rapid as in the Triassic (30-40 m/MY) continued. The basin continued to deepen, therefore, particularly in the block-faulted areas. By the end of the Period, the platforms and basins manifested a relative relief in excess of 4km (See Section 9.3).

9.2.3 Cretaceous - an end to regional subsidence

The thicknesses of the Cretaceous successions in the carbonate platform units indicate that rates of subsidence and sedimentation continued to fall during that period (10-20 m/MY). The Abruzzi-Campania Platform became temporarily emergent in the Middle Cretaceous, and the Monti della Maddalena Unit remained above sea-level until fore-reef to basinal carbonate sedimentation was renewed in the Maastrichtian (Fig. 9.3) (d'Argenio, 1970b; Scandone and Bonardi, 1968). A tectonic phase of major importance at the end of the Cretaceous was responsible for bringing deposition to a halt in many parts of the platforms.

During the Early Cretaceous, the sediment-water interface in the Lagonegro Basin remained below the CCD, suggesting that water depths must have been greater than in the adjoining Tethys, parts of which

began to accumulate pelagic limestones at that time (Dietrich and Scandone, 1972; Amodio-Morelli et al., 1979). Deposition of the black shales of the Brusco Formation, however, must have outstripped the drastically reduced subsidence rate because accumulation of pelagic limestones and marls was renewed during the Late Cretaceous accumulation of the 'flysch rosso'. The source of terrigenous mud for the Brusco Formation must have lain to the south since the Umbria-Marche and Molise Basins were free of such material. The Cretaceous did, however, see the deposition of thick black shale sequences in the internal units, for which comparisons can be found in the Liguride units of the northern Apennines (Selli, 1962; Ietto et al., 1965; Vezzani, 1969; Bernoulli and Jenkyns, 1974). The coeval deposition of black shales in both the internal and Lagonegro units suggests that the Lagonegro Basin might have been connected with the Tethyan Ocean somewhere to the south.

9.2.4 Latest Cretaceous and Palaeogene - the onset of regional compression

The onset of flysch depostion in the internal units at the end of the Cretaceous probably indicates that the Tethyan Ocean had begun to close (c.f. Amodio-Morelli et al., 1979). Large parts of the carbonate platforms became emergent and remained so for the remainder of the period. There is little record of deposition in the Lagonegro Basin, but scattered outcrops of the 'flysch rosso' indicate that the basin floor was once more above CCD. These conditions must have prevailed until the Lower Mioene, when a marine transgression in the Campania-Lucania Platform heralded the onset of the thrusting and superposition of the southern Apennines structural units described in Chapter 3.

9.3 Mesozoic regional subsidence in the southern Apennines: an appraisal and comparison with the Bahamas

The underlying philosophy behind the foregoing account is that early block-faulting and regional subsidence, rather than localised rifting, were the principle agents in determining the origin and development of the Lagonegro Basin. There remain, however, the fundamental questions concerning the nature of the original basement of the southern Apennine units and the mechanism by which it subsided on such a massive scale; clearly, it is necessary to solve the first problem before the latter can be addressed. In the following sections, two lines of enquiry are followed: first, the evidence from the southern appenines themselves is evaluated, and then a comparison is made with postulated mechanisms for the origin and development of the analogous modern Bahamas platform.

9.3.1 Mesozoic regional subsidence of the southern Apennines

Subsidence rates in the Lagonegro Basin can only be monitored in relation to the 'CCD, which during the Mesozoic was in a state of flux (Garrison and Fischer, 1969; Bosellini and Winterer, 1975). This lack of constancy in the depths of the dissolution facies is not entirely disadvantageous, however, since it has resulted in a greater number of facies changes in many Tethyan Mesozoic successions than would have been achieved under more static conditions. Using a number of typical sequences of the Alpine-Mediterranean region, and by making comparisons with the relationships of the various dissolution facies to depth in the modern oceans, Bosellini and Winterer (1975) have attempted to plot the depression of the CCD and

related dissolution surfaces in the Late Jurassic, and to monitor the subsidence of the basins themselves (Fig.9.4). Although the smooth subsidence curves that these authors assume are probably unrealistic, their model elegantly accounts for the different facies successions observed in different places. The typical pre-Cretaceous facies of these successions reflect low rates of sediment-supply, but following the dramatic development of calcareous nannoplankton in the Late Jurassic, deposition above the CCD was dominated by calcareous pelagic oozes, now represented by cherty limestones of the Maiolica type. In the Lagonegro Basin, on the other hand, the Triassic and Early Jurassic carbonate facies are likewise stratigraphically expanded due to the input of lime mud from the neighbouring carbonate banks. Thus the sequence of facies changes recorded in the Lagonegro successions do not follow the same pattern as those considered by Bosellini and Winterer; furthermore, as suggested in Chapter 5, the continual supply of calcareous material to the basin in the Early Mesozoic is likely to have caused a progressive deepening of the CCD throughout that period. It is more difficult, therefore, to plot the subsidence of the Lagonegro Basin, but a gradual transition from calcareous to siliceous pelagic deposition at the end of the Triassic provides one fix on the curve (Fig.9.4). The basin floor is assumed to have been at sea level in the Anisian, and at the end of the Jurassic must have been deeper than the CCD since, unlike the majority of the Tethyan basins, calcareous pelagic sedimentation was not resumed until the Late Cretaceous. If it is assumed that the CCD in Tethys during the Cretaceous was at a similar depth to that determined by deep-sea drilling for other oceans at that time, a minimum water depth of 3.6-4 km is implied (van Andel, 1975).

The curve plotted in Fig. 9.4 is at best an approximation, therefore, and only refers to the depth of the sediment-water interface. In order to arrive at a true basement subsidence curve, this must be corrected for the cumulative thickness of sediment at any particular time (Bosellini and Winterer, 1975). The new curve thus generated is, necessarily, tentative, since it takes no account of possible compaction. Although little faith can be placed upon the actual depths predicted, therefore it is evident that the rate of subsidence was initially rapid but decreased gradually during the Mesozoic; average values of 100, 40 and 10 m/MY are implied for the Triassic, Jurassic and Cretaceous respectively.

In general, the tops of the carbonate platforms remained at sea-level throughout the Mesozoic, so their stratigraphic thickness at any particular time can also be used to determine the depth to the basement. Although complete sequences through the platform units are rare, the younger parts of the succession in the autochthonous Apulian Platform are well known from borehole data (Parotto and Praturlon, 1975). The thickness of the Triassic in this unit, which is chiefly constituted by dolomites and evaporites, may reach 2,500m, and 1,500 m of Triassic dolomites are present at the base of the allochthonous Alburno-Cervati Unit in Calabria (Ippolito et al., 1975; Vezzani, 1975). Rates of subsidence in the autochthon were apparently more uniform than in the Lagonegro Basin, being lower in the early stages but greater in the Jurassic and Cretaceous (Fig.9.4). During the Middle Triassic, however, subsidence in the basin was probably accentuated by block faulting, whereas sediment

loading may subsequently have maintained higher rates beneath the platforms as they thickened. Nonetheless, the reasonable agreement between the subsidence curves generated for the basin and the platform supports the notion that their relative relief developed as an expression of differential rates of sediment accumulation rather than in response to any 'rifting' mechanism. The dramatic reduction in rates of sedimentation during the Upper Cretaceous and Tertiary in all of the platforms indicate that regional subsidence tailed off towards the end of the Mesozoic.

The platform and basin of the southern Apennines apparently subsided from sea-level in the Middle Triassic almost in sympathy. It seems unlikely, therefore, that either could have been founded upon Tethyan oceanic crust, particularly in view of the paucity of Middle Triassic volcanism in either domain. Besides, the Middle Jurassic age of the ophiolites in Liguria and Calabria militates against the existence of Tethys during the Triassic; moreover Liassic block-faulting associated with the opening of that ocean has been extensively documented throughout the Alpine-Mediterranean region (Dietrich and Scandone, 1972; Bernoulli and Jenkyns, 1974; Laubscher and Bernoulli, 1977; Amodio-Morelli et al., 1979). It might alternatively be argued that the southern Apennine units were founded upon oceanic crust belonging to the now-vanished Palaeotethys (c.f. Laubscher and Bernoulli, 1977). This, however, requires the existence of a thick clastic wedge, for oceanic crust adjacent to a well-established continental margin would have been several kilometres below sea-level. There is no evidence of any such wedge, and the likely presence of continental basement beneath the autochthonous carbonate platforms of the Adriatic argues against such a

model (Schütte, 1978). Palaeomagnetic data from these foreland domains suggest that the Adriatic region as a whole formed part of the African continent throughout the Mesozoic and Tertiary (Channell, 1976; Channell and Horváth, 1976; Horváth and Channell, 1977).

It is considered, therefore, that the missing basement was continental and that responsibility for the rapid regional subsidence, and associated Middle Triassic block-faulting, must lie with some process of crustal thinning. Subsidence of this nature and magnitude is typical of passive continental margins, and in the case of the Alpine-Mediterranean Mesozoic, is commonly linked with the opening of Tethys (Smith, 1971; Bosellini and Hsü, 1973; Dewey et al., 1973; Bernoulli and Jenkyns, 1974; Scandone, 1975; Channell and Horváth, 1976; Biju-Duval et al., 1977; Bourbon et al., 1977; Celet, 1977; Laubscher and Bernoulli, 1977). The majority of recent models for continental margin basin formation, however, can only account for regional subsidence some time after the inception of the adjacent ocean (Bott, 1971, 1976; Sleep, 1971; Kinsman, 1975); furthermore, the thermal-based hypothesis of Sleep and Kinsman associate rifting with early regional uplift. There is no evidence for either in the southern Apennines, where subsidence and block-faulting began simultaneously at least 60 MY before the formation of the ocean crust preserved in Calabria and Liguria. Computations of the relative motions of Africa and Europe at the beginning of the Jurassic from the magnetic anomalies of the Atlantic ocean suggest a major strike-slip component (Smith, 1971; Dewey et al., 1973); transcurrent margins, however, rarely exhibit major subsidence (Kinsman, 1975).

Two alternative hypotheses, both of which involve crustal extension, have been advanced recently which overcome many of the problems inherent in thermal-based models. In the first, brittle failure in the upper part of the continental crust and ductile flow at the base causes attenuation of the lithosphere during an early phase of extension (McKenzie, 1978). This is followed by a period of passive subsidence due to thickening of the crust by heat conduction to the surface. Alternatively, Sclater et al.(in press) have proposed that extension may cause large-scale intrusion of vertical dykes, the replacement of light crustal rocks by denser mafic or ultramafic material causing subsidence. Both models obviate the need for regional uplift prior to subsidence, and the rate of sinkage in each case is directly related to the amount of stretching. The two models can be distinguished by the fact that initial subsidence in the case of dyke intrustion is much more rapid.

Data on rates of basement subsidence beneath both the platforms and the basin of the southern Apennines are insufficient to allow discrimination between the two models. A plot of cumulative sediment thickness against time elapsed since the Middle Triassic for the autochthonous platform sequence, however, gives good agreement with McKenzie's model for a stretching factor of between 2 and 3, assuming a sediment density of 2.5 gm/cm^3 (Fig. 9.5.). The lack of Middle Triassic volcanism in the southern Apennines also favours this model; however, this is atypical of the Tethyan region as a whole, where volcanics of this age are commonly held to be a testimony to early continental rifting (Dietrich & Scandone, 1972; Bernoulli and Jenkyns, 1974; Hynes, 1974; Scandone, 1975; Biju-Duval et al., 1977;

Bosellini et al., 1977; Celet et al., 1977; Laubscher and Bernoulli, 1977; Bechstädt, 1978; Castellerin, Rossi et al., 1978). Some authors, on the other hand, attribute much of the volcanism in the Dinarides at that time to subduction of the Paleotethyan ocean (Blanchet, 1977; Celet et al., 1977; Bébien , et al., 1978). Although dyke intrustion may well, therefore, have played a part in causing regional subsidence, the abundant evidence of extensional tectonics in the Middle Triassic lends considerable credence to early stretching of the lithosphere as the dominant mechanism. This crustal attenuation was probably initiated as the continent of Pangea began to split apart, but prior to true rifting and opening of the Tethys.

9.3.2 Comparative evolution of the Bahama Platform

The origin of the Bahama Platform has long been a subject of controversy. Polemic has focused chiefly upon the nature of the underlying basement, and upon the mode of formation of the deep channels that dissect the platform. The sedimentary successions, both of the basins and the banks, are well known through drilling, but such is the thickness of the shallow-water carbonates that no wells have penetrated to their base (Fig.9.6)(Paulus, 1972; Mullins and Lynts, 1977). Early workers favoured a continental basement, but the majority of recent hypotheses, to account for the overlap of the Bahamas platform and the continental margin of Africa on pre-drift palaeo-continental reconstructions, have assumed that the underlying crust is oceanic (Fig.9.7)(e.g. Sheridan, 1974; for review, see Mullins and Lynts, 1977). The available geophysical data indicate that the basement is of intermediate density, seismic velocity and thickness, but taken in isolation, the data are ambiguous

and cannot provide a unique solution. Recently, following Meyerhoff and Hatten (1974), Mullins and Lynts (1977) have once more advanced the notion that the Bahamas are underlain by continental basement, and have invoked a 25° rotation of the platform to the northeast to account for the overlap, and corresponding underlap to the south, in palaeocontinental reconstructions;they attribute this rotation to impingement by the Carribean Plate during the Cretaceous and Early Tertiary.

The origin of the deep channels has been variously attributed to tectonic control, submarine erosion and differential sediment accumulation upon a regionally subsiding basement. Site 98 of the Deep Dea Drilling Project, located in Northeast Providence Channel, penetrated 357 m of Upper Campanian to recent pelagic and hemipelagic calcreous oozes and chalks, plus platform-derived calciturbidites, which indicate that the basins have been in existence for at least 70 MY (Hollister,Ewing et al.,1972; Paulus, 1972). By comparing rates of sedimentation in the basins and on the banks, Paulus (1972) suggested that the former are unlikely to have developed prior to the Early Cretaceous. However, recognition by Tator and Hatfield, (1975) of Upper Triassic arkoses near the base of the Great Isaac Well, drilled on the northwestern extremity of the Great Bahama Bank, has led Mullins and Lynts to suggest that the troughs may be as old as Triassic. According to their model, the basins are thought to have developed initially as grabens, but higher rates of sedimentation on the banks during regional subsidence are held responsible for creating most of the platform relief (Fig. 9.8).

The lack of consensus concerning the nature of the Bahamian base-

ment prevents the possibility of making an actualistic model for the southern Apennines. Furthermore, even if the 'continental' hypothesis of Mullins and Lynts (1977) is accepted, their envisaged mechanism for basin formation cannot be applied to the Lagonegro Basin. Although directly comparable at many levels, the evolution of the central Atlantic and Tethyan regions show marked differences in the early stages (c.f. d'Argenio, 1970a; Bernoulli, 1972; Bernoulli & Jenkyns, 1974; d'Argenio, de Castro et al.,1975). The eastern continental margin of the United States underwent a phase of tensional rifting during the Late Triassic, with concomitant volcanism and red-bed deposition, but regional subsidence probably did not begin until the inception of sea-floor spreading in the Early Jurassic (Bernoulli , 1972; Bernoulli and Jenkyns, 1974; Sheridan, 1974). This contrasts with the sequence of events in the Alpine-Mediterranean region where minor block faulting, volcanism and regional subsidence began simultaneously in the Middle Triassic, several tens of million of years before the opening of Tethys (Dewey et al., 1973; Bosellini and Hsü, 1973; Bernoulli and Jenkyns, 1974; Scandone, 1975; Bourbon et al., 1977; Laubscher and Bernoulli, 1977; Bechstädt et al., 1978). Borehole and geophysical data from the Bahamas suggest that here too subsidence began in the Early to Middle Jurassic, and the Triassic is thought to have been a period of uplift and graben formation (Meyerhoff and Hatten, 1974; Mullins and Lynts, 1977). The network of Upper Triassic to Lower Jurassic dolerite dykes in North America, West Africa and South America, when restored to their relative pre-drift positions, appears to converge on the Bahamas, suggesting that it may have been related to a domal uplift centred on that region (May, 1971; Dietz and Holden, 1973; Glockhoff, 1973;

Mullins and Lynts, 1977). The orientation of the deep channels is consistent with their having developed as radial grabens such as might be produced by a tumescent stress system (Mullins and Lynts, 1977). Thus although the Bahamian and southern Apennine carbonate platforms may equally have developed upon subsiding continental basement, their evolutions showing a resultant similarity, the conditions that governed their birth may have been quite different.

9.4 An appraisal of other deep-water basins of the Alpine-Mediterranean Triassic and implications for the early evolution of the Tethyan region

9.4.1 Deep-water basins of the Alpine Mediterranean Triassic - a review

The Lagonegro Basin was only one of a number of similar basins that evolved amongst the carbonate platforms bordering the Palaeo-tethyan Ocean during the Triassic. Some of these troughs persisted throughout the Mesozoic, and preserve a history of continuous pelagic and hemipelagic sedimentation. Many, however, survived only a short time before they filled up and shallow-water carbonate sedimentation was resumed in the Late Triassic and Liassic; the Hallstatt and associated basins of the Eastern and Southern Alps are notable examples (Schlager, 1969; Zankl, 1971; Bernoulli and Jenkyns, 1974; Bosellini and Rossi, 1974; Bechstädt, 1978, Bechstädt et al., 1978). The zones whose stratigraphies show the most striking resemblance to the Lagonegro successions are the Imerese and Sicani Zones of Sicily, and the Budva, Cukali and Pindos Zones of Yugoslavia and Greece (Fig.9.9)(Broquet et al., 1966; Aubouin, Blanchet et al., 1970; Aubouin, Bonneau et al., 1970; Smith and Moores, 1974; Celet, 1977; Catalano and d'Argenio, 1978). Parallels may also be drawn

with the Triassic successions of several other Tethyan basins which are of more oceanic affinity, such as the Sub-Pelagonian or Othris Zone of Greece, the Mamonia Complex of Cyprus, the Antalya Complex of Turkey and the Baër-Bassit Complex of Syria (Lapierre and Parrot, 1972; Lapierre, 1975; Marcoux, 1976; Delaune-Mayere et al., 1977; Robertson and Woodcock, 1979). Comparable Triassic stratigraphies are preserved still further east in the Neyriz region and Hawasina Complex of Iran and Oman respectively (Glennie et al., 1973; Hallam, 1976). In view of these similarities, it is suggested that many of the 'oceanic' basins of the Tethyan Mesozoic may have developed in the Triassic along similar lines to the intra-platform basins of southern Italy.

9.4.1.1. The Imerese and Sicani Zones of Sicily

The Imerese and Sicani Zones, which outcrop in the pile of décollement nappes forming the mountains of Western Sicily, each comprise several structural units that are thought to have originated in two, distinct, intra-platform basins similar to the Lagonegro Basin (Fig.9.10) (Catalano et al., 1976; Catalano and d'Argenio, 1978). Indeed, it has been suggested that these Basins may have merged to the east and eventually connected with the Lagonegro Basin (Catalano et al., 1976; Catalano and d'Argenio, 1978; Amodio-Morelli et al., 1979). There are marked similarities in the stratigraphies of these zones, and an evolutionary model comparable to that outlined in this thesis has recently been proposed by Catalano and d'Argenio (1978).

The units of the Imerese Zone comprise Upper Triassic Halobia-bearing cherty limestones, Liassic dolomite breccias and calcarenites, the Upper Liassic to Lower Cretaceous siliceous mudstones and cherts,

all of which have their counterparts in the Lagonegro Zone (Fig.9.11) (Montanari, 1966; Catalano and d'Argenio, 1978). The Cretaceous and Eocene, however, are constituted by pelagic limestones and marls of the 'Scaglia' facies, with further intercalations of platform-derived calcarenites and calcirudites, although in the Western Madonie Mountains, cherts and siliceous mudstones predominate (Schmidt di Friedberg et al., 1960; Broquet et al., 1966). The redeposited platform carbonates, which are notably abundant in the Jurassic and Cretaceous parts of the succession, may form packets up to 50 m thick (Scandone et al., 1972). Basaltic pillow lavas, probably of Middle Jurassic age, are also present in this zone.

The stratigraphic sequence in the Sicani Zone commences in the Middle Triassic with shales, marls and tuffs, surrounding blocks of Permian reefoidal limestone, that are overlain by Carnian cherty limestones and shales (Fig.9.11)(Mascle, 1967; Broquet et al., 1966). These are succeeded by Upper Triassic to Lower Liassic cherty limestones, Middle Liassic crinoidal and oolitic redeposited calcarenites, and Upper Liassic to Malm radiolarites (Daina, 1965; Broquet et al., 1967). Calpionellaid-bearing cherty limestones of Late Jurassic and Early Cretaceous age give way to Upper Cretaceous and Palaeogene 'scaglia' marly limestones, Eocene redeposited numulitic breccias and Oliocene to Upper Miocene calcarenites, meggabreccias and marls. The principle contrast between the Imerese and Sicani successions is the virtual absence of Jurassic and Cretaceous redeposited carbonates from the latter. This reflects the widespread submergence of carbonate platforms surrounding the Sicani Basin in the Liassic, at which time they developed as pelagic seamounts (Jenkyns and Torrens, 1971; Catalano and d'Argenio, 1978).

The basal parts of the successions are, therefore, closely comparable to those of the Lagonegro Zone, but the predominance of pelagic limestones and marls in the Cretaceous suggests that the Sicilian basins were generally shallower, only attaining depths below CCD in the Western Madonie Mountains. The absence of black shales equivalent to those of the Brusco Formation suggests that their deposition was relatively localised. Although the terrigenous clastics and limestone blocks at the base of the Sicani succession evoke a block-faulting mode of basin origin similar to that suggested for the Monte Facito Formation, the ages of the blocks imply that regional subsidence must have begun much earlier in this area.

9.4.1.2 The Pindos, Budva and Cukali Zones of Greece and Yugoslavia

The Pindos Zone outcrops in a broad swathe through Albania, western Greece and the Peloponnese,apparently terminating northwards in the Budva and Cukali Zones of southwestern Yugoslavia (Fig. 9.9 & 9.10)(Aubouin, Blanchet et al., 1970; Aubouin, Bonneau, et al., 1970; d'Argenio et al., 1971; Smith and Moores, 1974; Celet, 1977). Possible continuations to the south have been identified in the Ethia Zone of Crete and the Xindothio-Profitis Ilias Zone of Karpathos and Rhodes (Orombelli and Pozzi, 1967; Aubouin and Dercourt, 1970; Aubouin, Bonneau and Davidson, 1976; Aubouin, Bonneau, Davidson et al., 1976). Structurally, the Pindos Zone is made up of a stack of north-south-striking imbricate thrust slices with a westerly vergence. The sedimentary facies vary considerably, both along and across the outcrop of the zone, but the succession in central Greece embodies many typical features (Fig.9.11)(Fleury, 1977). At the base is an unknown thickness of flyschoid sandstones and shales

of supposedly Carnian age (Fleury, 1977). In the Budva Zone, similar deposits, though of Anisian age, are underlain by Werfenian (Lower Triassic) red beds, and are succeeded by a series of tuffs and radiolarian cherts (Aubouin, Blanchet et al., 1970; Celet, 1977; Celet et al., 1977). The Upper Triassic throughout these zones is characterised by up to 300 m of cherty limestones with Halobia (Orombelli and Pozzi, 1967; Aubouin, Blanchet et al., 1970; Celet, 1977; Fleury, 1977). These are succeeded by Liassic shales and siliceous mudstones, Middle and Upper Jurassic radiolarian cherts, Upper Jurassic to Lower Cretaceous calpionelled-bearing limestones and marls, and Middle to Upper Cretaceous pink pelagic limestones (Fleury, 1977). In the Budva Zone, the Cretaceous succession is punctuated by abundant redeposited limestones (Aubouin, Blanchet et al., 1970; Celet, 1977). The Late Cretaceous witnessed a transition throughout from pelagic to flysch deposition (Aubouin, Blanchet et al., 1970; Celet, 1977; Fleury, 1977).

9.4.2 Deep-water basins of the Alpine Mediterranean Triassic - significance and implications

The Pindos and associated successions also, therefore, bear a marked resemblance to those of the Lagonegro Zone, although the resumption of calcareous pelagic sedimentation in the Cretaceous is more reminiscent of the Sicilian equivalents. A unique feature of the Pindos Zone, however, is that for much of its length it is structurally overlain by ophiolites (Fig.9.10). Only in Yugoslavia and southern Greece can internal carbonate platforms, represented by the High Karst and Parnassos Zones, be recognised. To the east of the Pindos ophiolites lies the Pelagonian (s.l)

Ophiolite Zone which, in addition to scattered ophiolites, comprises metamorphic basement, Mesozoic platform carbonates and a variety of pelagic and clastic basinal lithologies of Early Mesozoic age (Smith and Moores, 1974; Smith et al., 1975; Smith et al., 1979). A further ophiolite belt, the Vardar Zone, separates the Pelagonian Zone from the Serbo-Macedonian and Rhodope internal massif zones (Smith and Moores, 1974).

It has been argued that the presence of redeposited platform carbonates in the easternmost Pindos units implies that the internal platform continues unexposed beneath the Pelagonian Zone (Scandone and Radoičić, 1975). This presupposes, however, that the units of the latter, in common with those of the remainder of Greece, have a southwesterly vergence and are structurally overlying the external domains (c.f. Bernoulli and Laubscher, 1972); according to this model, the ophiolites of the Pindos and Pelagonian Zones are all considered to have originated in the Vardar Zone. Recent work in the Othris and Olympos districts of eastern Central Greece, however, has indicated that the Pelagonian Zone has a northeasterly vergence, and it has been re-interpreted as the deformed remnants of a continental margin that lay to the northeast of a small ocean basin during the Jurassic (Hynes et al., 1972; Smith and Moores, 1974; Barton 1975, 1976; Smith et al., 1975; Smith and Woodcock, 1976; Price, 1976; Smith et al., 1979). The ophiolites of the Pindos and Pelagonian Zones, as distinct from those of the Vardar Zone, are thought to be the relics of that ocean, which has been interpreted as part of the central Tethys. It follows, therefore, that during the Jurassic, the Pindos basin must have lain adjacent to, or even formed part of the Tethyan Ocean for much of its length.

The apparent juxtaposition of the Pindos Basin with oceanic Tethys during the Jurassic, and the imbricate thrust structure of the zone itself, has led Smith (1976) and Smith et al. (1979) to suggest that the zone represents an accretionary prism stripped from Tethyan oceanic crust subducted beneath the Pelagonian continental margin. The onset of subduction is thought to have caused emplacement of a marginal strip of this ocean on to the continental margin to form the Pindos and Pelagonian ophiolites in the Late Jurassic or Early Cretaceous (Smith and Woodcock, 1976). Similarities in the early Mesozoic stratigraphies of the Pindos Zone and the more distal units of the sub-pelagonian Zone indeed suggest the possibility that they may have originated in the same basin (Smith et al., 1975; Price, 1976). In southern Greece, however, the Pindos Basin was demonstrably separated from the ocean by the Parnassos Platform, and in Yugoslavia, the Budva Basin was similarly bordered on the east by the High Karst Platform (Celet, 1977; Johns, 1977). It is difficult to envisage how such platforms could also have been founded upon oceanic crust, particularly since they were formed prior to sea-floor spreading in the Jurassic. It is suggested, therefore, that the Pindos Basin initially evolved in a similar manner to the basins of southern Italy, through regional subsidence and relative accretion of the surrounding platforms, but that parts of the basin were also sites of Late Triassic and Liassic volcanism and rifting that led to the formation of an ocean basin (Fig.9.12). Subsequent subduction of this ocean, and the 'missing' attenuated continental basement of the Pindos Zone, could have produced the observed imbrication of the Pindos thrust sheets and led to their subsequent emplacement onto the external platform. A non-oceanic mechanism of basin formation, such as evinced by the Lagonegro one, should perhaps be given serious

consideration when attempting to elucidate the early histories of other 'oceanic' basins of the Tethyan region.

In order to examine the spatial relationships between the pre-oceanic deep basins of the Alpine-Mediterranean Triassic, a palinspastic/palaeogeographic restoration of the Mediterranean region prior to the breakup of Pangea has been devised (Fig. 9.13). Several such reconstructions have previously been attempted, the majority of which have been based upon the restoration of the European and African continents to their positions relative to America prior to the opening of the Atlantic Ocean (Smith, 1971; Bosellini and Hsü, 1973; Dewey et al., 1973; Scandone, 1975; Channell and Horváth, 1976; Biju-Duval et al., 1977; Laubscher and Bernoulli, 1977). All of these reconstructions have located the region at the apex of the large, triangular embayment in the continent of Pangea occupied by the now-vanished Palaeotethys. There is little agreement, however, concerning the location of the relics of this Palaeozoic ocean and its continental margins in the present structural framework of Europe and the Middle East, or the significance of the numerous ophiolite belts of the Eastern Mediterranean.

The premise for the reconstruction in Fig.9.13 is that the Apulian/Adriatic foreland has constituted a rigid promontory of the African continent throughout the Mesozoic and Tertiary, and is indicated by palaeomagnetic data from the autochthonous platform sequences of southern Italy and Sicily (Channell and Tarling, 1975; Catalano et al., 1976; Channell, 1976; Channell and Horváth, 1976; Horváth and Channell, 1977). Since the décollement nappes of Sicily, the southern Apennines, Yugoslavia and western Greece are all

stacked onto this stable foreland, by removing the effects of thrusting, it is possible to determine their original positions. Palaeomagnetic data from the nappes of southern Italy and Sicily indicate that they have been rotated anticlockwise and clockwise respectively during their emplacement, and the Lagonegro-Sicani-Imerese basin may therefore have been almost linear and oriented approximately north-south (Catalano et al., 1976). There is no evidence, on the other hand, for rotation of the Pindos thrust sheets, so they have simply been unstacked back towards the east.

In contrast to the majority of previous reconstructions, it is not considered that the deep basins of southern Italy and Pindos were interconnected. There is no compelling evidence that they continued north to join up in the area now represented by the Eastern or Southern Alps, or that they opened out onto the Palaeotethys to the east (c.f. Scandone, 1975; Laubscher and Bernoulli, 1977). It is perhaps surprising, therefore, that prior to the opening of Tethys in the Jurassic, the facies of these basins were not more euxinic. On the other hand, without any mechanism for producing density stratification, such as fosters anoxic conditions in the Modern Black Sea, there is no reason for believing that the deep basins were not well-oxygenated, even though they may have been entirely surrounded by shallow marine banks.

The Lagonegro-Sicani-Imerese and Budva-Pindos Basins are both oriented almost normal to the direction of relative movement of Africa and Europe during the Liassic; it is thus unlikely that the block faulting in these troughs was related to strike-slip tectonics, as may have been the case in the western Alps (c.f. Bourbon et al., 1977). It is considered more likely that the

platforms and basins owe their origin to extension and attenuation of continental crust during the first phases of motion between the two continents prior to the opening of Tethys. The subsequent histories of the two basins differ, however; whereas the Budva-Pindos Basin was also the site of subsequent oceanic rifting, the Liassic faulting episodes left the southern Italian basins virtually intact. The juxtaposition of the Pindos Zone with Tethyan ophiolites is thus largely coincidental, and is of little significance in relation to the origin of the basin during the Triassic. Although the Triassic Lagonegro, Sicani and Imerese basinal zones are almost unique in their isolation from the ophiolite belt, basins of their type may have been one of the dominant features of the Alpine-Mediterranean region prior to the opening of the Tethys Ocean.

9.5 Epilogue

The stratigraphy, sedimentology and structure of the Lagonegro Zone provide a unique insight into the origin and history of one of the pre-oceanic basins of the Alpine-Mediterranean region. It is clear that the extreme depths that these basins achieved in the Late Mesozoic and Tertiary represent a gross magnification of the relatively minor block-faulting that was responsible for their genesis in the Middle Triassic. Nevertheless, pre-occupation with the more dramatic events related to the opening of Tethys may have prevented recognition of similar stages in the early development of several of the 'oceanic' basins of the Tethyan belt. It may be necessary, therefore, to view the basal stratigraphies and hence early histories of these 'oceanic' domains in a slightly different light.

Appendix 1

Stratigraphic sections

A total of 25 sections have been measured at various localities throughout the zone; a stratigraphic correlation chart summarising the data from these sections, together with location maps, can be found in the enclosures (Chart I). For each structural unit, the sections are presented in an approximately lateral sequence from north to south, enabling recognition of facies and thickness variations. In addition, a detailed lithological log, together with a brief description of features of stratigraphic and lithological importance, is given for each section; detailed discussions of the sedimentology of the formations is given in Chapters 4, 5, 6 and 7. All formations of the Lagonegro Units up to and including the Brusco Formation are represented, although there are no complete sections through the latter or the Monte Facito Formation at the base; coverage of the Monte Foraporta Unit is similarly incomplete.

A. Lagonegro Unit I

LI-1 Sasso di Castalda I-I.G.M. Sheet 199-III NE (Brienza) (Pl.A1)

The base of the section is located 500m north of the village of Sasso di Castalda at the confluence of two streams, marked .914 on the map, where they are crossed by a NE-SW fault. From this point the section passes north along the stream bed for 100m and thence up the north-west side of the gorge; it cannot be followed up to the crest due to recent afforestation.

90m of thinly bedded micrites of the Sirino Formation, with shale intercalations becoming more common towards the top, are succeeded by 20m of the Lagonegro Formation. A peculiar feature of this section is the interval of green mudstones and claystones with planar- and

cross-laminated calcarenites 32m above the base. The transition between the two formations differs from that in the type section at Lagonegro; the thinly bedded silicified limestones of the latter are absent, but several beds of green marly limestone and laminated calcarenite occur. Although no section can be measured through the remainder of the Lagonegro Formation, exposure beside the path on the other side of the hill reveals the presence of a number of pink, green or grey fine-grained carbonate breccias and calcarenites interbedded with the siliceous mudstones and cherts; this is in contrast to the solitary breccia in the Lagonegro Section.

L1-2 Sasso di Castalda II - I.G.M. Sheet 199-III NE (Brienza)

A short section beside the road from Sasso di Castalda to S. Michele exposes the contact between the Lagonegro and Brusco Formations. It is situated on the south side of the new road in a recent cutting 500m east of the village. Superficially, the cherts of the Lagonegro Formation differ markedly from those of other sections, commonly being red in the centre with green margins, more reminiscent of the cherts at the top of the Pignola-Abriola Section. Several thick, graded extraformational carbonate breccias occur in the overlying Brusco Formation.

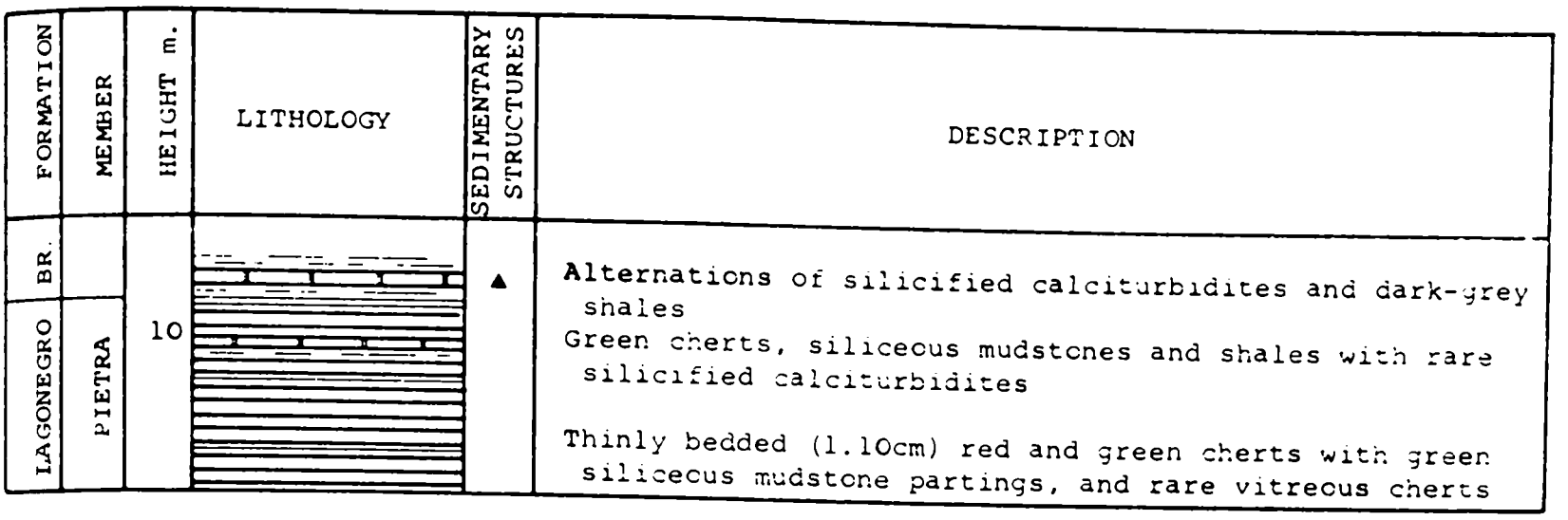


Fig. A1 Section LI1 Sasso di Castalda I

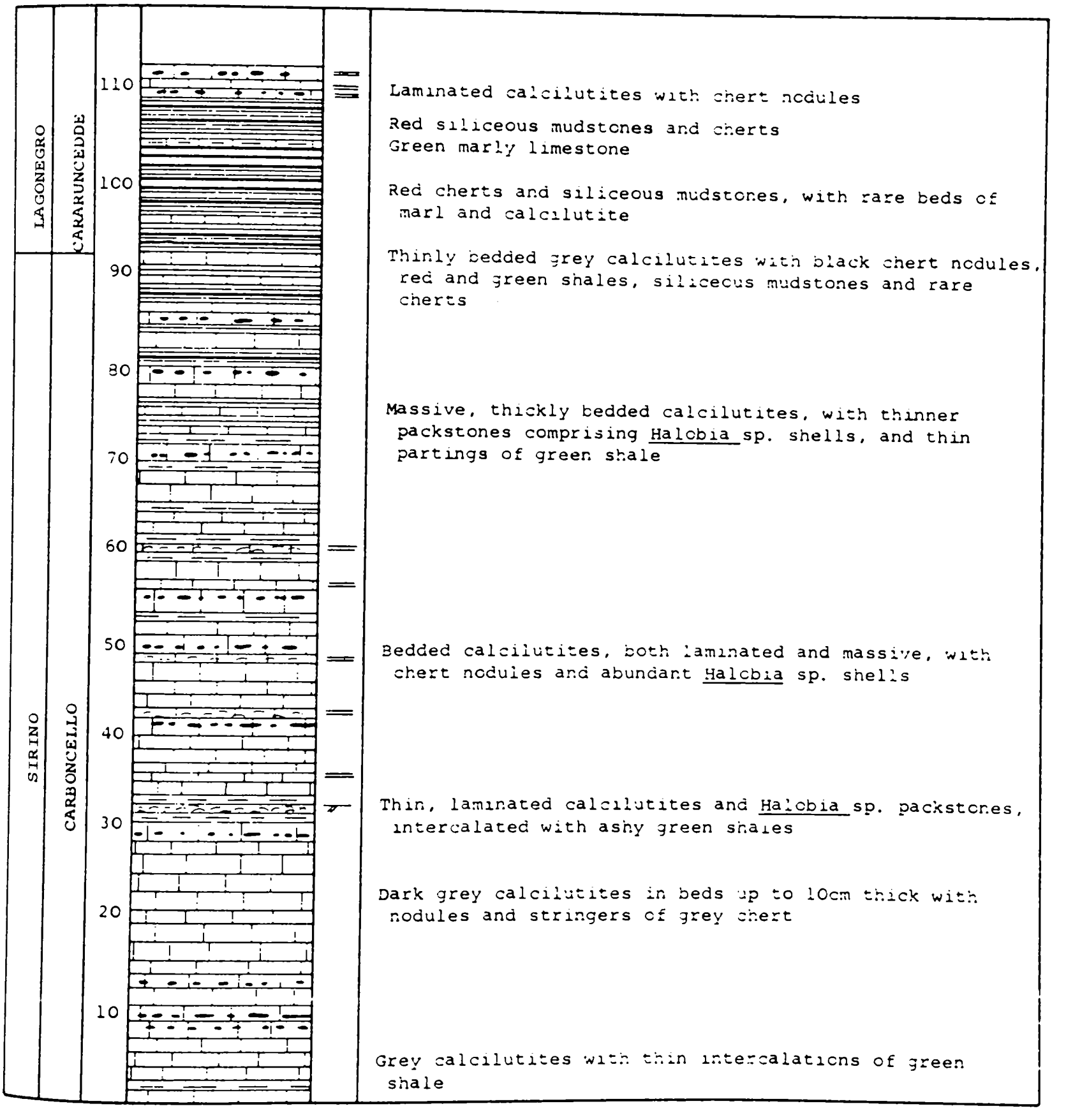


Fig. A2 Section LI2 Sasso di Castalda II

LI-3 Monte Lama - I.G.M. Sheet 199-II NO (Marisco Nuovo)

The section at M. Lama is exposed in the western limb of a northerly-plunging anticline. Situated at the base of the southern flank of the mountain, it extends from the top of the Lagonegro Formation, at the western end of the hillside, east for about 500m. The section is finally obscured by vegetation and tectonic repetition near the core of the anticline. It represents the longest measurable section in the unit but does not encompass any single formation; although the top is bounded by a fault, it is probably very close to the contact with the Brusco Formation.

The transition between the Sirino and Lagonegro Formations is similar to that at Lagonegro, although laminated, thinly bedded and unsilicified calcarenites and calcilutites, that can be found at the base of the latter, are absent; so too are the coarser grained carbonate breccias seen at Sasso di Castalda, even though the sections are only 8km apart. The 200 succession of the Sirino Formation includes the entire Carboncello Member and 130m of the Monte Sirino Member. The limestones and shales of the first are no different to those at Sasso di Castalda, although the distinctive green claystone/laminated calcarenite interval of that section is not seen.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LAGONEGRO	PIETRA	260			Banded green and grey radiolarian cherts, and milk-white vitreous cherts containing abundant Radiolaria
		250			
		240			
	CARAMUNCEDDE				Red siliceous mudstones with occasional beds of red radiolarian chert that may be laminated
					Red and green siliceous mudstones
		230			Thinly bedded, and rarely silicified, laminated calcilutites, with green shale intercalations
		220			Red siliceous mudstones with rare beds of radiolarian chert
		210			Red and green siliceous mudstones, cherts and light-coloured calcilutites
SIRINO	CARBONCELLO	200			Thinly bedded grey calcilutites with nodules and bands of predominantly black chert, and intercalations of red and green paper-shales
		190			
		180			Red and green shales and marls
		170			Medium-bedded grey calcilutites up to 0.5m thick, with nodules and bands of chert, and intercalations of red and green shales up to 10cm thick
		160			Thinly bedded, laminated calcilutites and Halobia sp. packstones
		150			
	MONTE SIRINO	140			Grey calcilutites in beds up to 0.5m thick, with bed-parallel nodules and bands of grey or black chert
		130			
		10			

Fig. A3 Section LI3 Monte Lama

LI-4 Monte Farno - I.G.M. Sheet 210-I SO (Rocca Rossa)

This short section in the Brusco Formation is located in a gulley a short distance east of the Lagonegro-Moliterno road, 600m north by east of M. Farno. The exposure of the formation in this area is separated from the outcrop of the underlying formations, which form the mountain itself, by a fault, and thus it is not possible to judge at what stratigraphic level the section occurs. The alternation of silicified calcilutites, calcareous breccias and grey/black shales is typical of the formation throughout the Sirino Window.

LI-5 Mangoso - I.G.M. Sheet 210-I SO (Rocca Rossa)

A prominent ENE-WSW trending cliff face north of the area called Mangoso, and close to the sharp left-hand bend in the Lagonegro-Moliterno road 6km north of Lagonegro, exposes a 50m section through the upper part of the Lagonegro Formation. Dipping gently to the north-west, these rocks outcrop directly beneath the basal thrust of Unit II. The presence of calcarenites and carbonate microbreccias in the Cararuncedde Member resembles the successions in the Vulturino Window more than that at Lagonegro. The Pietra Member measures 33m and is identical to the type section; although the contact with the Brusco Formation is not exposed, the thrust marking the top section follows the boundary between this and the Lagonegro Formation.

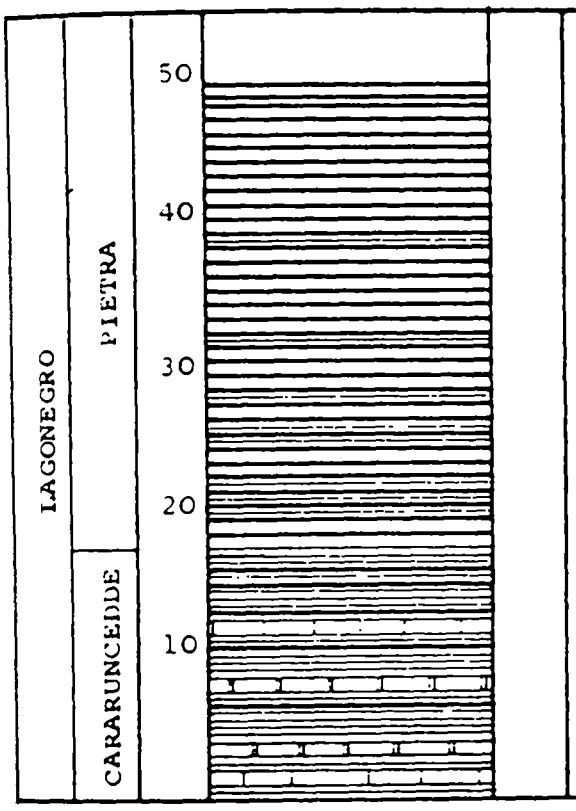
LI-6 Ponte della Pietra - I.G.M. Sheet 210-I SO (Rocca Rossa) (Pl.A2)

This section, only 700m distant from the Mangoso section, was measured in a similar ENE-WSW trending cliff face occupying an analogous structural position on the northern flank of the Gianni Griecu anticline. It extends from the topmost hairpin bend in the Lagonegro-Moliterno road, about 100m north of the bridge over the

F.Pietra, down to the foot of the cliff. Embracing the entire Pietra Member, it incorporates the best exposure of the contact between the Lagonegro and Brusco Formations, which is notably sharp and is marked by a dramatic reduction in the degree of silicification, and a transition from radiolarian chert to black shales.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
BRUSCO		10		● ▲	Silicified calciturbidite with flute marks (30cm thick)
					Alternations of silicified calcilutites up to 30 cm thick and dark grey or black paper shales
				●	Dark- coloured calciturbidite with sole structures Grey and black paper-shales, and rare calcilutites

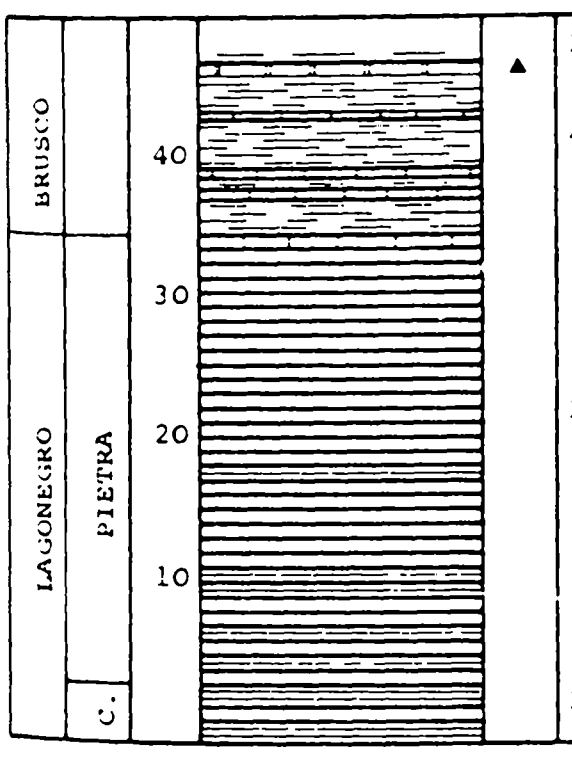
Fig. A4 Section LI4 Monte Farno



Green and grey banded radiolarian cherts, and bands of light-grey or milk-white vitreous chert; rare brown cherts

Red or purple siliceous mudstones with Fe/Mn patina
Red siliceous mudstones and cherts
Massive calcisiltite with chert bands
Green siliceous shales and cherts
30 cm thick calciturbidite
Green siliceous shales and cherts
De-calcified calciturbidite; red siliceous mudstones

Fig. A5 Section LI5 Mangoso



▲ Dark-grey paper-shales
2 graded calciturbidites with erosive bases

Alternations of silicified calcilutite and dark-grey or black paper-shales, commonly bearing an Fe/Mn patina

Banded geen and grey radiolarian cherts, and bands of light-grey, milk-white and brown vitreous chert

Red siliceous mudstones and rare cherts

Fig. A6 Section LI6 Ponte della Pietra

LI-7 Gianni Griecu - I.G.M. Sheet 210-II NO (Lagonegro)(Pl.A3)

Representing the type succession for the lower part of the Sirino Formation, this section extends from the base in the Fiume della Pietra stream bed up the cliff on the south side of Gianni Griecu, 400m south-south-west of the summit. Here it is possible to observe 80 metres of section before it becomes completely inaccessible. Occupying the core of the Gianni Griecu anticline, it is not possible to see what lies beneath. Several mesoscopic asymmetrical folds enable the section to be followed horizontally beside the stream near the base. 12.5m of fine-grained limestones, marly limestones and shales of the Gianni Griecu Member are overlain by 67.5m of thinly bedded limestones with chert bands and nodules. The cherts are especially well developed at the top of the section, where the beds are commonly entirely silicified; this may possibly correspond to the abnormally silicified interval immediately beneath the H. styriaca level identified by Scandone (1967) at Sorgente Acero, and which he considered to be underlying rather than overlying the limestones, marls and shales of the Gianni Griecu member (H. superba level).

LI-8 Costa del Alto - I.G.M. Sheet 210-I SO (Rocca Rossa)

Along the gorge of the Fiume Pietra it is possible to follow the entire sequence through the Sirino and Lagonegro Formations, from the Gianni Griecu Member in the core across either the eastern or western limb of the Costa del Alto-M. Nicola-Mizzo Milego anticline. The stream, however, follows the line of several east-west trending faults, and it has only been possible to measure one discrete section through the uppermost part of the Sirino Member. This is situated on the west-dipping limb, extending eastwards from the sharp, right-angle bend in the stream west-north-west of M. Nicola, where the gorge

is crossed by a fault breccia, for about 50m before the section is faulted out again at the base. A large proportion of the limestone beds are planar- or, more rarely, convolutely laminated, with common shale intercalations towards the top of the section; otherwise, the lithology is identical to that on M. Nicola.

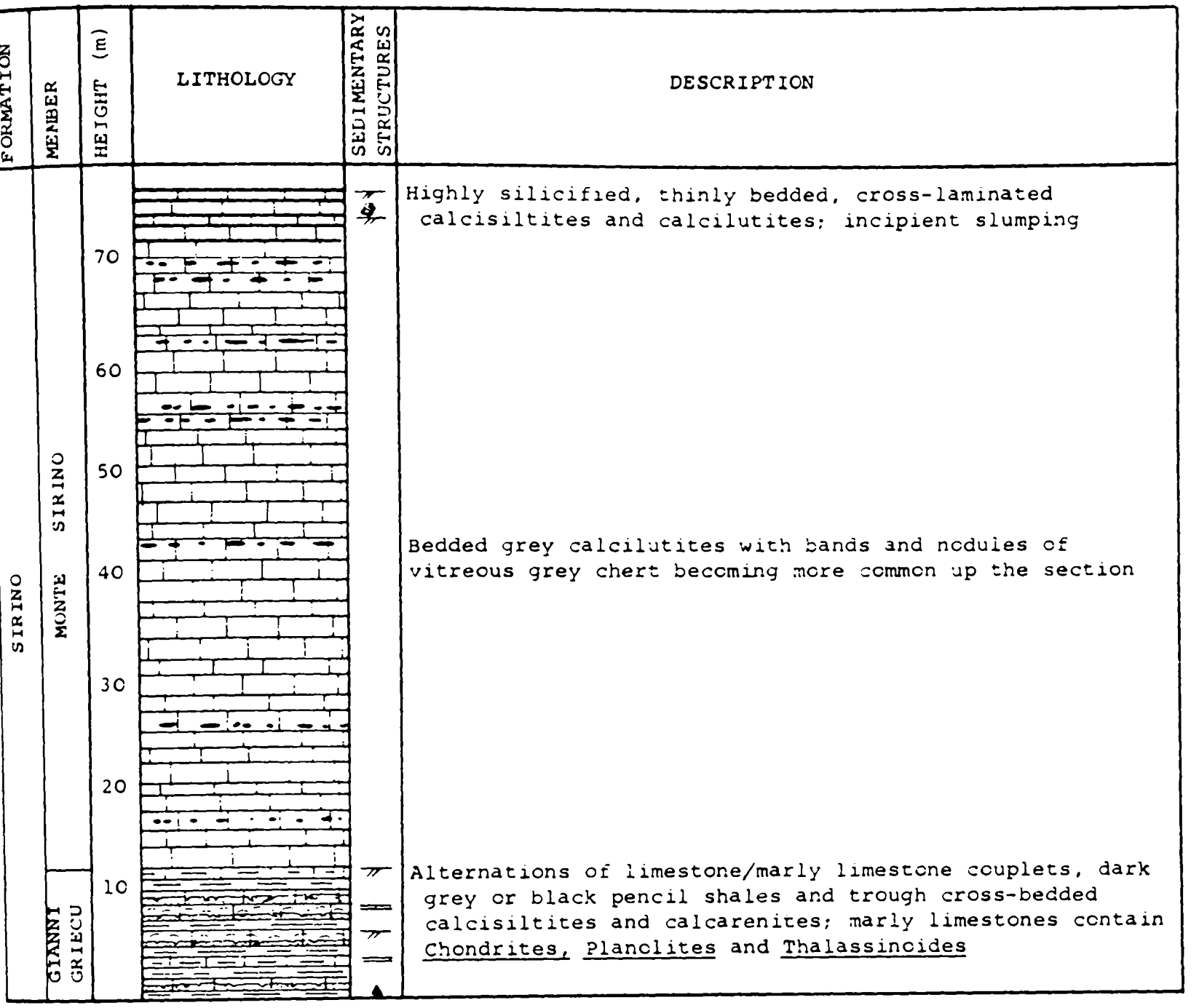


Fig A7 Section LI7 Gianni Griecu

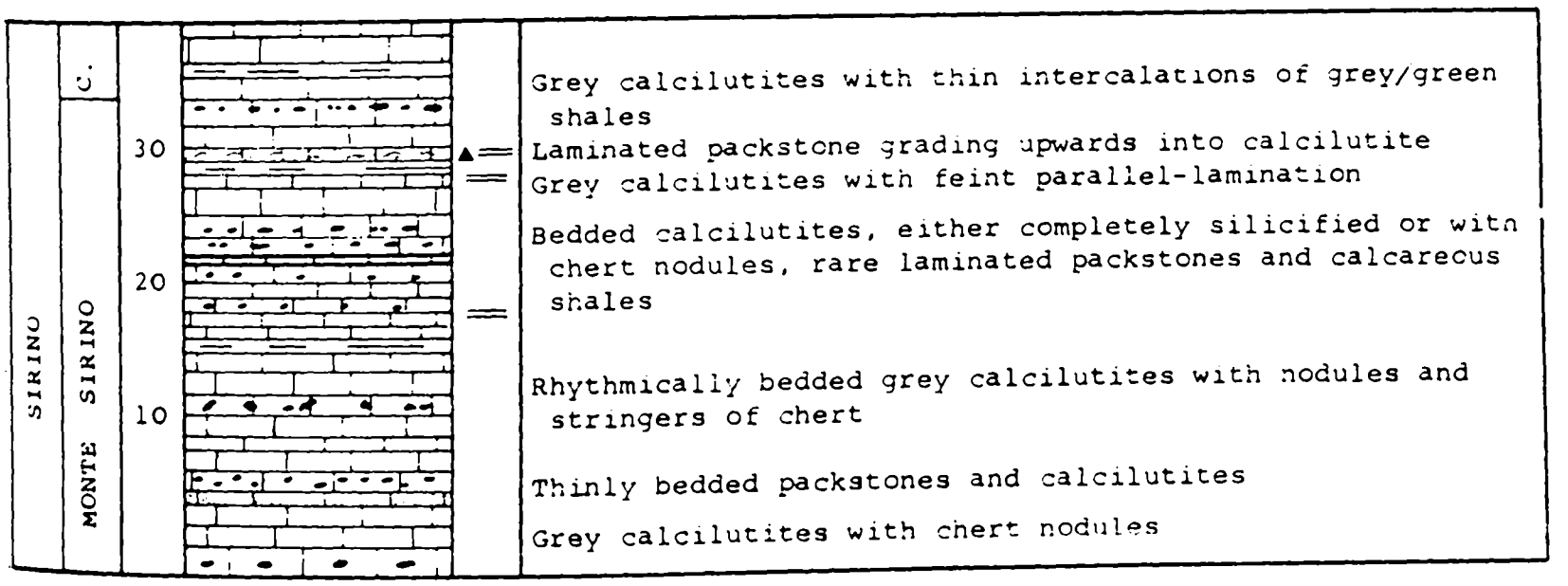


Fig. A8 Section LI8 Costa del Alto

LI-9 Monte Nicola - I.G.M. Sheet 210-I SO (Rocca Rossa)(Pl.A4)

A 52m section through the Monte Sirino Member of the Sirino Formation can be found north of an east-west trending fault on the south side of M. Nicola. This is an excellent section for observing facies variations in the sequence, both laterally and vertically. It is notable for the presence of slumped horizons, several grainstone beds and an abundance of dolomitic chert nodules; these features are described in detail in Chapters 5&6. It is not possible to determine the precise stratigraphic level at which this section occurs, but in view of its position within the anticline, it must be close to the top of the Monte Sirino Member, and possibly corresponds approximately to the lower part of the Costa del Alto Section.

LI-10 Monte Sirino - I.G.M. Sheet 210-II NE (Monte Sirino)(Pl.A5)

The section on M. Sirino, although twice as long as that on M. Nicola, is of comparable lithology and, probably, stratigraphic position . It runs up the east side of Serra del Spalla d'Imperatrice, from the limit of exposure at the base as far as the ridge crest about 550m north east of the summit. Although no slumped beds are exposed, the section shows features identical to those at M. Nicola.

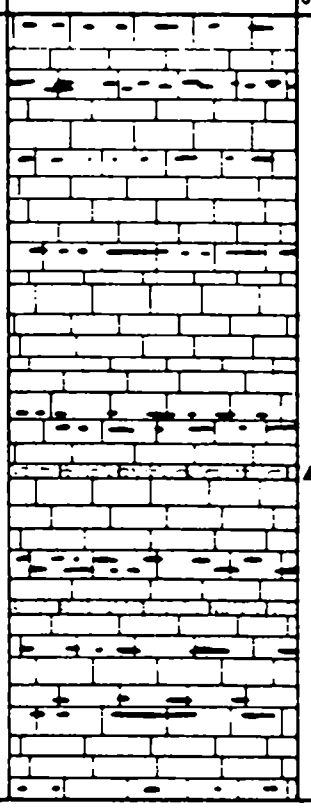
FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
SIRINO	MONTE SIRINO	50			Thinly bedded grey calcilutites with bands and nodules of dolomitic chert
		40			
		30			
				▲ —	Laminated and graded packstone (filling small depression)
		20			Bedded calcilutites with dolomitic chert nodules
					Packstone containing abundant pelagic bivalve shells; fills neptunian dykes in the underlying calcilutite bed
		10			
				§	Slumped calcilutites

Fig. A9 Section LI9 Monte Nicola

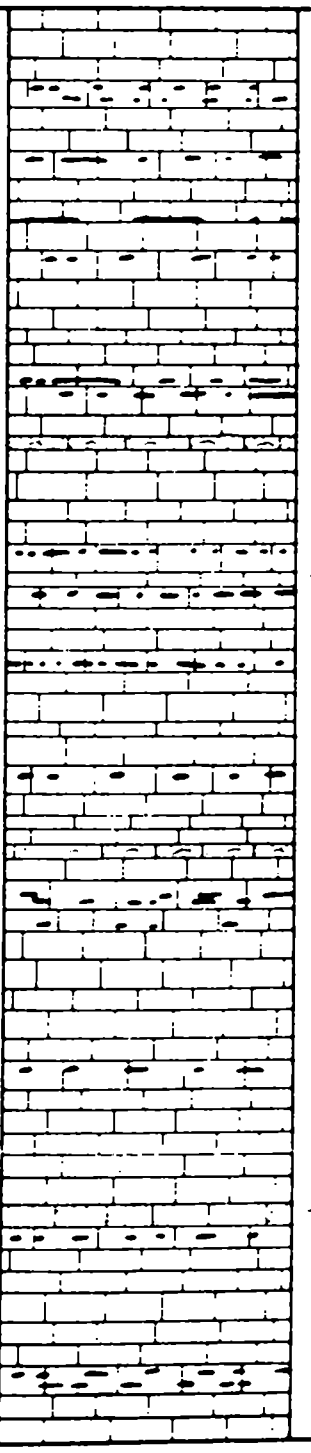
FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
SIRINO	MONTE SIRINO	90			Grey calcilutites with nodules and stringers of dolomitic chert
		80			
		70		=	Laminated packstone with abundant pelagic bivalve shell fragments
					Medium-bedded grey calcilutites
		60		=	Fine-grained laminated packstone
		50			Grey calcilutites with nodules of dolomitic chert
		40			Packstone with abundant pelagic bivalve shells
		30			Grey calcilutites with dolomitic chert nodules and stringers
		20			
				//	Cross-laminated packstone
		10			Well-bedded grey calcilutites with beds and nodules of dolomitic chert

Fig. A10 Section LI10 Monte Sirino

LI-11 Lagonegro - I.G.M. Sheet 210-II NO (Lagonegro)(Pl.A6-A8)

The section at Lagonegro is probably the best exposed in the entire zone. It is located in the gorge of the Fiume Carboncello, beneath the viaduct on the S.S.19 and immediately to the east of the town. It is in two parts; the lower 40m, up to the first red siliceous mudstones, although well exposed on the east side of the gorge, are only accessible on the west. (The succession can also be followed downwards, but not measured, by following the stream bed northwards). The red siliceous mudstones can be correlated across to a terrace on the other side of the gorge that can be reached from the viaduct. From here the section extends continuously up to and along the road without interruption for about 100m.

Up to the point at which it joins the road, the section passes through the Carboncello Member of the Sirino Formation, from near the base to the contact with the Lagonegro Formation. Above this point the latter formation is exposed almost in its entirety, although the contact with the Brusco Formation has been eroded and is not visible. The greeny grey silicified limestones near the base of the formation are unique to this section which includes only one silicified carbonate breccia. The transition from the Sirino to the Lagonegro Formations is thus more clearly defined than at M. Lama and Sasso di Castalda, where laminated calcilutites and calcisiltites are also found above the contact. The chert nodules in the limestones near the top of the Sirino Formations are usually black, and the intervening mudstones appear more siliceous than in other sections, commonly including cherts. There is an overall increase in the degree of silicification upwards through the Lagonegro Formation.

LI-12 Bersaglio - I.G.M. Sheet 210-II NO (Lagonegro)

Although not actually a part of the Lagonegro Section, this exposure of the Brusco Formation lies directly above the contact with the cherts of the Pietra Member 600m east-north-east along strike. It outcrops beside the path leading from Bersaglio towards Bonfilio beneath the autostrada viaduct. It is constituted by grey and black shales, silicified calcilutites and carbonate breccias identical to those at Ponte della Pietra and Monte Farno.

LI-13 Fiume Torbido - I.G.M. Sheet 210-II NO (Lagonegro)

A short section through the Brusco Formation at a higher but unknown stratigraphic level is to be found beside the S.S.19 near KM 124-VI. In contrast to other sections described in the formation, carbonate breccias entirely predominate over silicified calcilutites and show more or less complete Bouma sequences and sole structures, giving the aspect of a typical flysch succession.

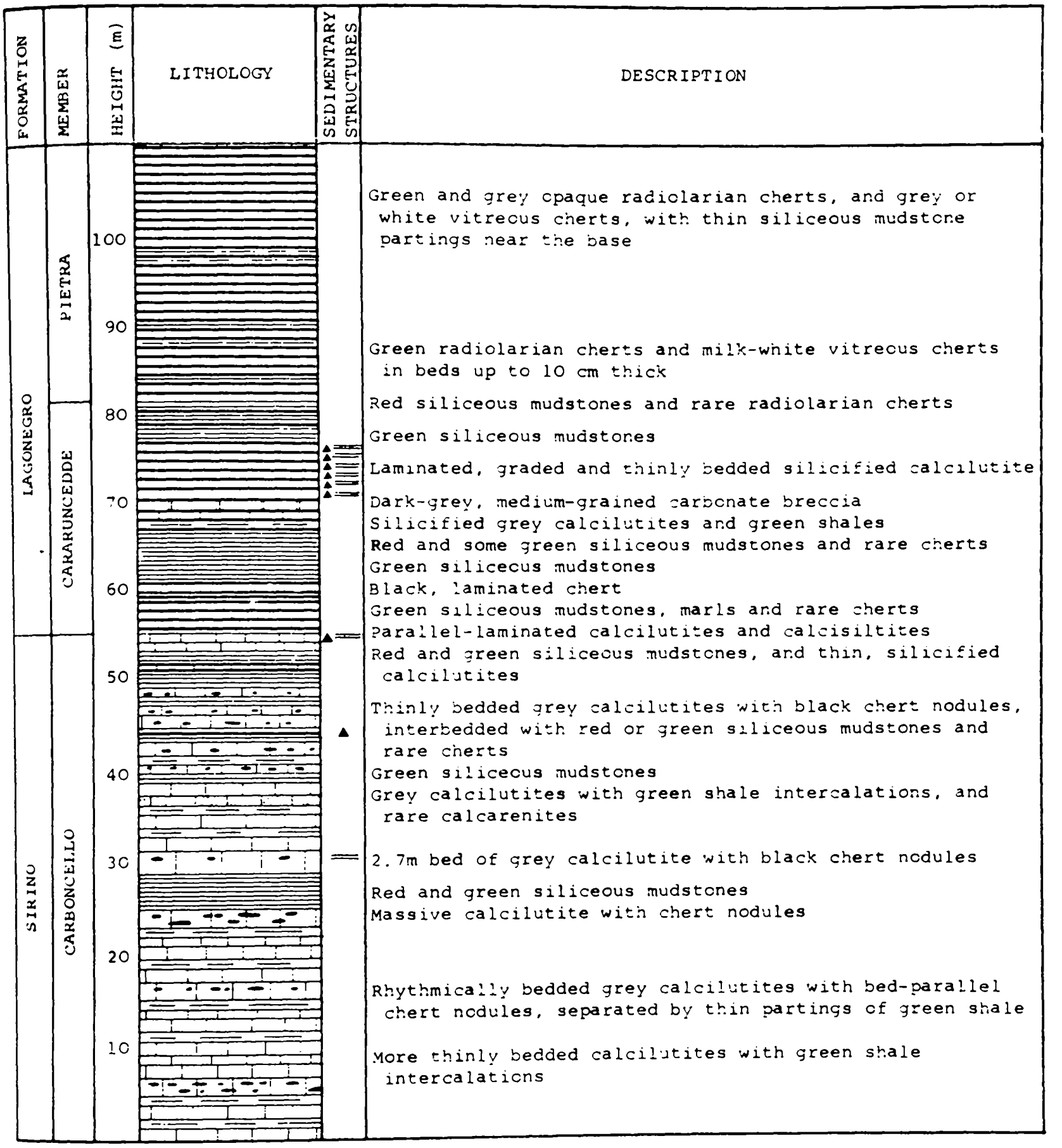


Fig. A11 Section LI11 Lagonegro

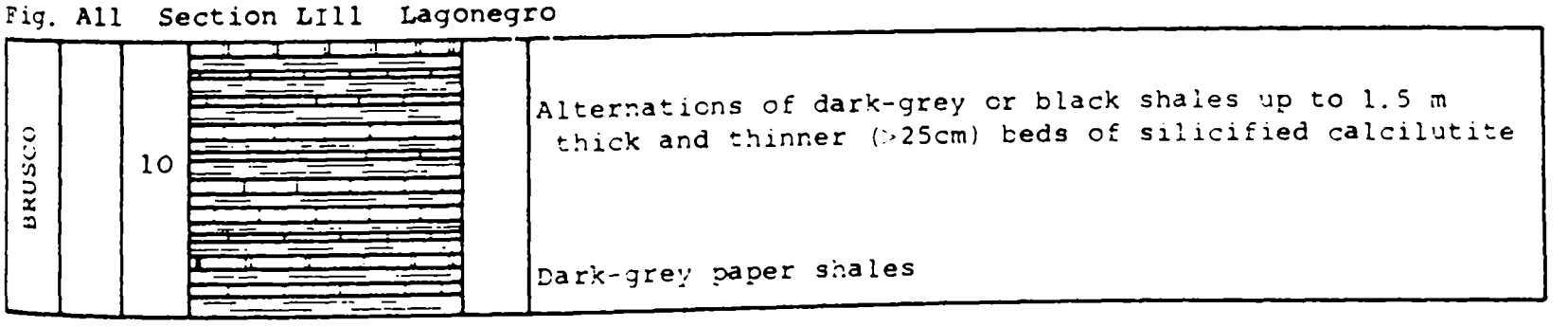


Fig. A12 Section LI12 Bersaglio

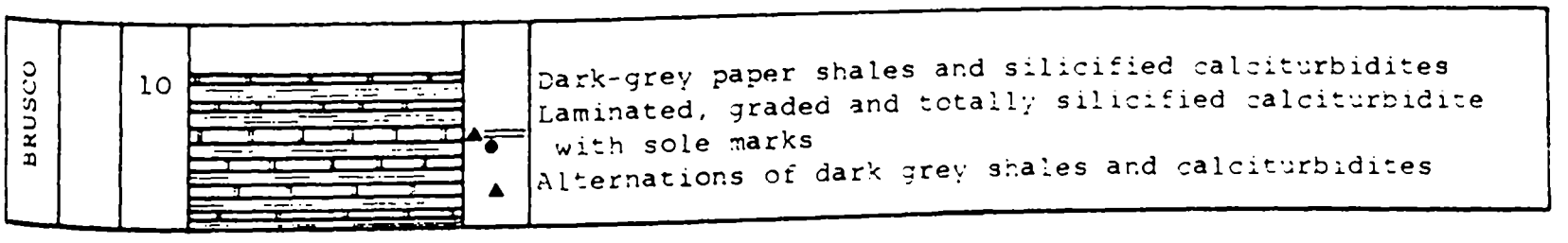


Fig. A13 Section LI13 Fiume Torbido

B. Lagonegro Unit II

LII-1 La Ralla - I.G.M. Sheet 187-III NO (Muro Lucano)(Pl.A9)

This 300m section near S. Fele extends along the crest of La Ralla, from the base 500m north-west of M. Castello as far as the road on the north side of the bridge across the Fiume Bradano. The succession dips at about 60° to the north-east and is bounded by faults to the north-east and south-west. It provides a complete section through the top of the Sirino Formation and the Lagonegro Formation.

The section is dominated by an abundance of calciturbidites that occur throughout the entire Lagonegro Formation, becoming decreasingly abundant towards the top. Details of the Sirino Formation are obscured by pervasive dolomitisation and disruption of bedding along the plane of the fault crossing the valley at the base of the section. However, it is possible to observe that the top few metres comprise coarse angular breccias containing fractured cherts, similar in aspect to the dolomitised breccias at the base of the Pignola-Abriola Section I; these are also exposed on the north side of M. Pierno, 6.5km to the south-east, where they occupy a similar stratigraphic position. Partial or complete dolomitisation of the carbonate breccias can be observed up to 125m from the base of the Lagonegro Formation, throughout which the carbonate beds contain chert nodules or are entirely chertified. The interbedded siliceous deposits show an overall increase in degree of silicification.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LAGONEGRO					Silicified calciturbidites and green siliceous mudstones
					More or less silicified calciturbidites with rare shales
		150			Partially dolomitised calciturbidite with ooids
					Partially or totally dolomitised calciturbidites with chert nodules
		140			Shales and dolomitised calciturbidites
					Dolomitised carbonate breccias with chert nodules
					Dolomitised or silicified calciturbidites
		130		▲	Amalgamated, graded and rarely laminated calciturbidites with dolomitised matrix
				▲ ▲	Thinly bedded fine-grained limestones, calciturbidites and shales
		120		▲	Thick calciturbidites, either completely silicified or with chert nodules
		110			Thinly-bedded, laminated and silicified fine-grained limestones, siliceous mudstones and cherts.
		100		▲ ▲	Thick, graded calciturbidites and thin shales
				▲	Thick beds of dolomitised and rarely graded calciturbidite
		90		▲	Thinly bedded calciturbidites
					Dolomitised, coarse sand and granule-grade calciturbidites
					Laminated limestone
				▲	Calciturbidites with chert nodules, and dark grey shales
		80		▲	Silicified and graded, coarse-grained calciturbidite
					Thick calciturbidites with nodules and bands of chert
		70		▲	Mainly green shales with rare beds of dolomitised calciturbidite
		60			Thin, dolomitised calciturbidites, with intercalations of green shale
		50		▲	Dolomitised and rarely graded dolomitised calciturbidites and poorly exposed siliceous mudstones
		40		▲	Very coarse-grained, graded calciturbidite
					Thinly bedded laminated dolomites, and siliceous mudstones
		30		▲ ▲	Graded, dolomitised calciturbidites and shales
SIRINO		20			Brecciated dolomite with fractured grey or white chert: indistinct bedding
		10			Massive, poorly bedded white dolomite with rare chert nodules

Fig. A14 Section LIII La Ralla - lower part

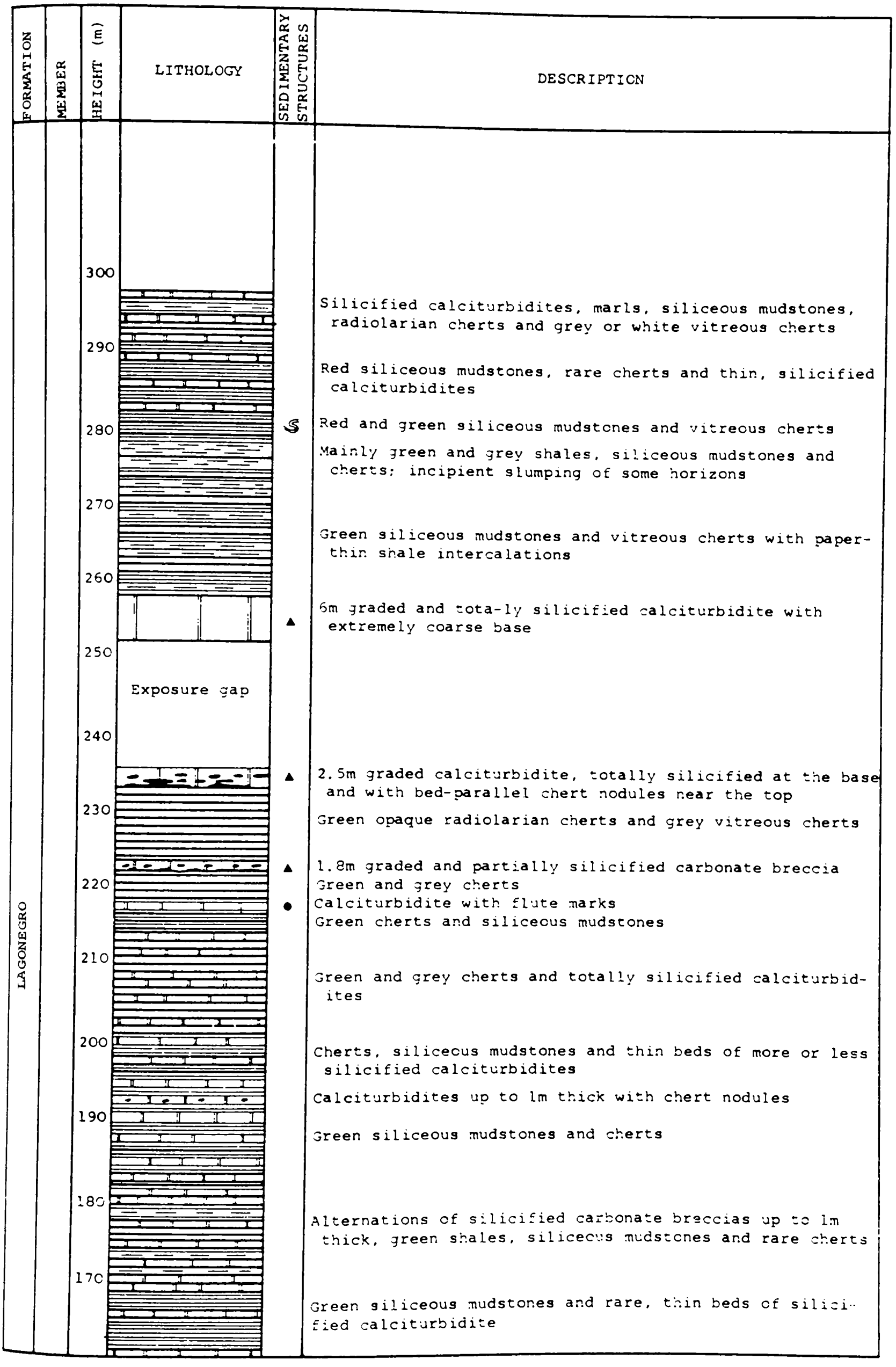


Fig. A14 (cont.) Section LIIl La Ralla - upper part

LII-2 Pignola-Potenza - I.G.M. Sheet 199-I NO (Potenza)

95m of red and green cherts interbedded with silcified carbonate breccias are exposed beside the Pignola to Potenza road between Km5 and Km4, up to andincorporating the contact with the Brusco Formation. The section contrasts with that at La Ralla by virtue of the greatly reduced thickness of the calciturbidites,which seldom exceed 20cm and are entirely silicified. The siliceous mudstones and cherts which constitute the dominant lithology are rarely slumped over vertical intervals not exceeding 2m.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LAGONEGRO		90			Red and green radiolarian cherts
		80			
		70			Thinly bedded silicified calciturbidites Red and green radiolarian cherts and siliceous mudstones Thinly bedded silicified calciturbidites Red and green siliceous mudstones and cherts
		60			Silicified calciturbidites and vitreous cherts
		50			Red and green banded radiolarian cherts and siliceous mudstones
		40			
		30		S	Red and green cherts and silicified calciturbidites Green cherts with slumps
		20			Red and green radiolarian cherts and grey or white vitreous cherts
		10		S	30 cm slumped chert horizon
					Green radiolarian cherts with thin silicified calciturbidites and grey or white vitreous cherts

Fig. A15 Section LII2 Pignola-Potenza

LII-3 Pignola-Abriola I - I.G.M. Sheet 199-I SO (Pignola)

Although grouped together, five individual sequences separated by exposure gaps of unknown thickness comprise this section; it is clear from the regional structure, however, that these outcrops form part of a complete vertical succession. They are located in ascending stratigraphic order along the road from Abriola to Pignola between Km10.8, on the north side of the quarry, and Km8.7, a short distance after the sharp right hand bend around the south-west extremity of Le Coste.

Coarse carbonate breccias of the top of the Sirino Formation, that are entirely dolomitised, comprise the first 6m section, which is in faulted contact with the ensuing 63m sequence. This embraces the contact with the Lagonegro Formation, marked here by the loss of dolomitisation above the last of the breccias. Several cream-coloured, laminated calcilutites, and thinly bedded calciturbidites interbedded with green, yellow and red siliceous mudstones and some pink or white marly limestones characterise the remainder of the formation; the latter are unique to this section, red marls being found only rarely in the Torrente Bitonto section. The red and green cherts of the highest section in the sequence are identical with those of the Pignola-Potenza section, and may be laterally equivalent.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LAGONEGRO		10			Rhythmically bedded red and green radiolarian cherts with thin shaley, and thin silicified calciturbidites and red, green, white or grey vitreous cherts with planar, convolute and small-scale cross-lamination
		40			Red and green shales and rare marls
					Red marls and white marly limestones in beds ≤ 10 cm thick, with rare chert nodules
		30			
		20			Rhythmically bedded red and green cherts and siliceous mudstones
		10			Red and some green siliceous mudstones
					Fine-grained marly limestone
		10			Red and green shales, marls and thin calciturbidites
				▲	Graded calciturbidite Red and green shales, with rey marl at the base
		60			Green shales and silicified calcilutites
					Thinly bedded silicified calcilutite
					Green shales, thin calcilutites and calciturbidites
		50			
					Silicified calcilutites and green shales
		40			
					Cream-coloured, parallel-laminated calcilutites
		30			
					Green shales with thinly bedded calcilutites and fine-grained calciturbidites
		20			Thin, laminated, cream-coloured calcilutites, with rare chert nodules, and thin green shales
					Thinly bedded dolomites with chert nodules and thin intercalations of green shale
SIRINO		10			Beds of dolomitic breccia up to 0.5m thick, with fragmented chert nodules, and thin intercalations of green shales and rare cherts.
					Green and red dolomitic marls and green shales
		5			Thickly-bedded chaotic dolomitic breccias with fragmented chert nodules

Fig. A16 Section LII3 Pignola-Abricla I

LII-4 Pignola-Abriola II Section - I.G.M. Sheet 199-I SO (Pignola)

The section follows the disusedrailway track from the bridge due west of Km12 on the Pignola to Abriola road for 200m and thence up onto the road, along which it extends for a further 600m as far as the quarry. The base is located only a short distance from the contact with the Monte Facito Formation. The thinly bedded calcilutites are characteristic of the Sirino Formation of Unit II, but a higher proportion of laminated calcarenite beds are present than at Torrente Bitonto. Also, a 4m succession of green claystones, siltstones, marls and thin limestones, not found elsewhere, occurs at the point at which the section joins the road. A short distance above this level the limestones are dolomitised, and the bedding-transgressive contact between limestone and dolomite is clearly visible. Exposure of the gap between the two Pignola-Abriola sections is obscured by a quarry , and crossed by several faults.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
SIRINO		110			Thinly bedded dolomites with chert nodules and bands; sub-horizontal dolomite veinlets abundantly developed
		100			Bedded calcilutites and dolomitised calcilutites with bed-parallel nodules and bands of chert; contacts between limestone and dolomite may be normal to bedding
		90			
		80			Rhythmically bedded calcilutites with chert nodules
					Grey nodular limestone
				▲	Laminated and graded calcarenite
		70			Green tuffaceous siltstones and marls with thin limestones containing Halobia superba (de Capoa Bonardi, 1970)
					Laminated calcarenite with H. austriaca (Scandone, 1968)
		60			
					Bedded calcilutites, with bed-parallel bands and nodules of chert, and rare laminated calcarenites with abundant pelagic bivalve shells
		50			
		40			
		30			Thinly bedded grey calcilutites with cherts, and thin shaley partings
		20			
		10			Thinly-bedded, laminated calcarenites separated by thin chert bands
					Beds of laminated calcarenite up to 30 cm thick, and green shales
					Calcarenites, with parallel-laminated bases, and shales

Fig. A17 Section LII4 Pignola-Abriola II

LII-5 Rupe del Corvo - I.G.M. Sheet 199-II SE (Viggiano)

A 170m section can be measured on Rupe del Corvo, a prominent mountain 3.5km north-by-west of Viggiano. It is situated up the eastern side of the south-facing cliff and along the ridge to the north-east as far as the mule track. The section includes a large part of the Sirino Formation and, although poorly exposed, the contact with the Lagonegro Formation. The coarse extraformational breccias seen further north at the top of the Sirino Formation are absent, but several calciturbidites are present in the Lagonegro Formation above; dolomitisation is absent. At several levels in the upper 65m of the Sirino Formation, intraformational calcirudites, comprising rounded calcilutite fragments in a more marly limestone matrix, are interbedded with the thinly bedded calcilutites. In common with the section at Torrente Bitonto, bedding in the calcilutites becomes progressively thicker towards the base, a feature also of the central part of the M. Armizzone section.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LAGONEGRO		160			Green, yellow and grey shales, with beds of parallel-laminated and more or less silicified fine-grained calciturbidites
			Exposure gap		
					Graded and laminated calciturbidites
		150			Laminated calcarenites and calcilutites, with minor shales
		140			Calcilutites and calcarenites with minor shale intercalations
		130			Bedded, grey calcilutites in beds up to 60 cm thick, with bands and nodules of chert, and thin calcarenites with green shale intercalations
		120			
		110			
		100			Thickly bedded calcilutites, with beds and nodules of chert, and thinner interbeds of intraformational calcirudite, both fine- and coarse-grained, 10-20 cm thick
		90			
		80			
SIRINO		70			Alternations of packets of thickly and thinly bedded grey calcilutite, with bands and nodules of chert, and thin beds of calcarenite containing abundant pelagic bivalve shells.
		60			
		50			Thickly bedded grey calcilutites with bands and nodules of chert, and a few thinner calcilutite beds
		40			
		30			
		20			very thickly bedded grey calcilutites with bands and nodules of chert
		10			

Fig. A18 Section LII5 Rupe del Corvo

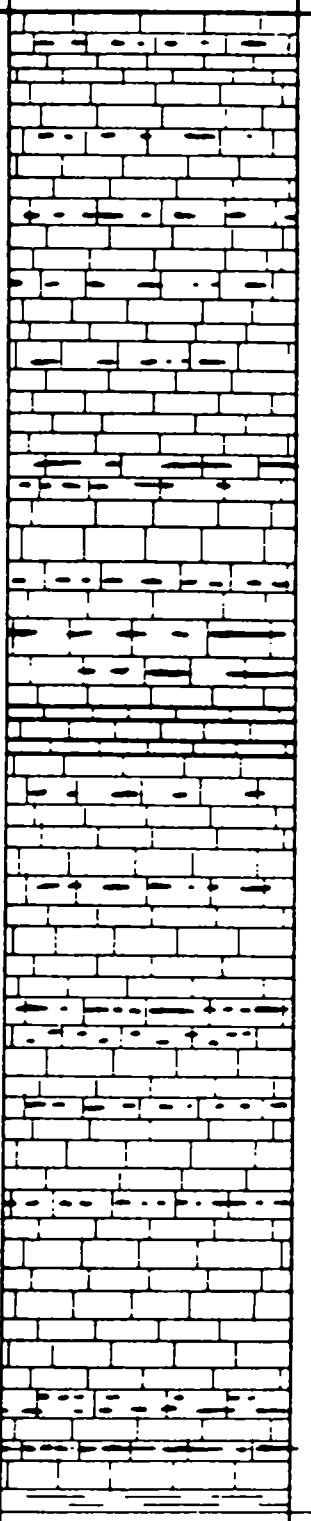
LII-6 Monte Armizzone - I.G.M. Sheet 210-III NO (Latronico)

The section at M. Armizzone, 2.5km south-south-west of Castelsaraceno, is in two parts, correlation between the two being based upon the recognition of a clayey layer containing Halobia superba (Scandone, 1967); however, the stratigraphic significance of this species has been shown to be ambiguous and the correlation is thus a tenuous one (c.f. de Capoa Bonardi, 1970). Furthermore, the top of the section is crossed by a fault which cuts out an unknown thickness of the Sirino Formation. Nonetheless, by virtue of the absence of dolomitisation, and the well exposed upper and lower contacts, it provides the most suitable type of section for the formation.

The first and lower part is located on the east side of the mountain, stretching from its northern extremity south for about 200m; it measures 45m and covers the contact between the Monte Facito and Sirino Formations. The other part follows the western ridge from the H. superba level of Scandone (1967) as far as the summit, although a fault crosses the section at the point where the cliff is offset to the north; a 50m section through the Sirino - Lagonegro Formation contact occurs beyond the fault.

The top of the Monte Facito Formation seen in the lower part is not as well exposed as in the Valle del Pesce; it is lithologically similar, however, except for the presence of some rare sandstone beds. The lower part of the Sirino Formation is characterised by calcilutites with chert nodules and thin shale intercalations. These latter die out upwards in the upper section, where the rocks are similar to those at Rupe del Corvo; the central part is more thickly bedded and contain rare beds composed almost entirely of chert. Across the fault, shales

are once more intercalated with the limestones, and intraformational calcirudites identical to those at Rupe del Corvo and Torrente Bitonto are present. The Lagonegro Formation lacks extraformational carbonate breccias, but contains several laminated calcarenite beds.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
SIRINO		100			Thinly bedded (≤30 cm) grey calcilutites with cherts
		90			Bedded, grey calcilutites with bed-parallel nodules and bands of chert
		80			
		70			Thickly-bedded grey calcilutites with cherts
		60			
					Bands of replacement chert, and thin grey calcilutites with chert bands
		50			
		40			
					Well-bedded calcilutites with bed-parallel nodules and bands of chert
		30			
		20			
		10			
		*			Grey-green shales and thin limestones containing Halobia superba (Scandone, 1968)
					SECTION ON NORTHEAST SIDE OF MONTE ARMIZZONE
		*			Green-grey shales and thin limestones with H. superba (Scandone, 1968)
		40			Bedded calcilutites with chert nodules
		30			Thinly bedded calcilutite with bands of chert, and intercalations of yellow shale
		20			Beds of grey calcilutite up to 1m thick with nodules and bands of black chert
					Thinly bedded calcilutites, with chert nodules, and green shale intercalations
M. FAC	V.d.P.	10			Red and green siliceous mudstones, marls, nodular limestones and replacement cherts

*Suggested correlation

Fig. A19 Section LII6 Monte Armizzone - lower part

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LAGONEGRO		60			Red and green rhythmically bedded radiolarian cherts
		50			Laminated, fine-grained limestones
					Red siliceous mudstones
					Laminated fine-grained limestones with intercalations of siliceous mudstones and cherts
		40			Red and green siliceous mudstones and cherts
					Alternations of thinly bedded and laminated fine-grained limestones, siliceous mudstones and cherts
		30			Grey calcilutites with chert nodules and intercalations of red and green shales
SIRINO		20			Bedded calcilutites and calcirudites with green shale intercalations
		10			

SECTION ON SOUTH SIDE OF MONTE ARMIZZONE

Fig. A19 (cont.) Section LII6 Monte Armizzone - upper part

LII-7 Valle del Pesce - I.G.M. Sheet 210-II NE (Monte Sirino)(Pl.A10)

The 56m type section for the transition between the Monte Facito and Sirino Formations is situated on the northwest side of a small gulley immediately to the west of Tempa di Roccarossa. It passes from red and green claystones at the base, through the nodular limestones and marls of the Valle del Pesce Member, up into the thinly bedded grey calcilutites of the Sirino Formation.

LII-8 Tempa di Roccarossa - I.G.M. Sheet 210-II NE (Monte Sirino)

This very short section beside the path leading from Giardini dei Tuori round the south side of Tempa di Roccarossa exposes a succession of red and green micaceous sandstones and siliceous mudstones in depositional contact with the limestone mass of the mountain itself. Its stratigraphic position within the Monte Facito Formation is not known, but being so close to the Valle del Pesce section, it may be relatively high in the sequence; the sandstones are similar to those found at the base of the Monte Armizzone Section.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
SIRINO		55			Thickly bedded calcilutites with bed-parallel chert nodules and stringers
		50			
		45			
		40			Grey calcilutite beds with some thinner beds of more nodular aspect, and paper-thin shale intercalations and beds and nodules of chert
		35			
MONTE FACITO	VALLE DEL PESCE	30			Grey and green shales, nodular limestones and dolomitic marls with nodules of chert Red and green shales and dolomitic marls with beds and nodules of chert Grey and pink nodular limestones, dolomitic marls, red siliceous mudstones and red or green chert bands
		25			
		20			Red and green shales Grey nodular calcilutites with red and yellow siliceous mudstone intercalations, and red and green chert bands
	PIETRA MAURA	15			Greeny/yellow and red mudstones becoming siliceous towards the top
		10			
		5			Greeny/yellow mudstones

Fig. A20 Section LII7 Valle del Pesce

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
MONTE FACITO	PIETRA MAURA	10			Green, grey and red, structureless micaceous sandstones in beds up to 10 cm thick
		5			Red, green and yellow mudstones

Fig. A21 Section LII88 Tempa di Roccarossa

LII-9 Torrente Bitonto - I.G.M. Sheet 210-II NO (Lagonegro)(Pl.All)

Interrupted only by two short exposure gaps, this section, located 1km northeast of Vignale, affords 300m of continuous exposure from the top of the Lagonegro Formation well down into the Sirino Formation. It is situated on the southern flank of the ridge leading west-south-west from the .787 spot height down to the confluence of the Torrente Bitonto and a tributary,marked .606. This hillside is being afforested, however, and exposure is no longer complete.

Thickly bedded calcilutites at the base pass upwards into more thinly bedded limestones with green shale intercalations and abundant intraformational calcirudites. Following the stream bed for a short distance, the contact with the Lagonegro Formation is crossed; this is broadly similar to that at M. Armizzone, although there are fewer laminated calcarenites. The remainder of the succession is constituted by red, green and yellow siliceous mudstones and cherts, the latter becoming more common towards the top of the section; also present are rare marls and several silicified calciturbidites up to 20cm thick that are particularly common in the central part of the formation. A 1m bed of pink calcarenite is situated 5m below the top of the section.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LAGONEGRO		150	Exposure gap		
		140			Laminated fine-grained limestones with chert nodules and green shale intercalations
		130			Laminated calcarenites and calcisiltites with chert bands, intercalated with green, more or less siliceous mudstones and dark calcareous shales
		120			Red and green siliceous mudstones, cherts, laminated calcarenites and white calcilutites
SIRINO		110			Red and some green siliceous mudstones with beds of intraformational calcirudite
		100			Intraformational calcirudites up to 1 m thick, with intercalations of red or green siliceous mudstone
					Bedded calcilutites with chert nodules
		90			Green shales and calcarenites with pelagic bivalve shell fragments
					Grey calcirudites and calcilutites; bases of some calcirudites are erosive
		80			Grey calcilutites, with cherts, and calcarenites, intraformational calcirudites and green shales
		70			
		60			
		50			Grey calcilutites with cherts and rare thin beds of laminated calcareinte that contain abundant pelagic bivalve shells
		40			
		30			
		20			
		10			

Fig. A22 Section LII9 Torrente Bitonto - lower part

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
BRUSCO					
					Dark-grey and black paper shales
		300			Green cherts 1m bed of multiply-graded pink calcarenite
		290			Red and some green, rhythmically bedded cherts with paper-shale partings, and rare red marls
		280			
		270			Green siliceous mudstones, cherts and many silicified, fine-grained calciturbidites and vitreous cherts
		260			
		250			Red siliceous mudstones, some cherts and many thin, silicified calciturbidites and vitreous cherts
		240			Green/yellow cherts, siliceous mudstones and thin, silicified calciturbidites and vitreous cherts Silicified, cross-laminated and graded calciturbidite
		230			Red siliceous mudstones and radiolarian cherts
					Yellow shales
		220			Red and some yellow finely bedded siliceous mudstones 20 cm silicified calciturbidite
LAGONEGRO					Green siliceous mudstones and radiolarian cherts
		210			Red siliceous mudstones and rare marls
			Exposure gap		
		200			Green cherts and siliceous mudstones
					Green siliceous mudstones and cherts
		190			Red, and some green, siliceous mudstones and rare cherts
		180			Red and yellow siliceous mudstones and cherts
					Red and yellow cherts and siliceous mudstones
		170			Red and green siliceous mudstones and cherts
					Red siliceous mudstones

Fig. A22 (cont.) Section LII9 Torrente Bitonto - upper part

C. Monte Foraporta Unit

MF-1 Malamugliera - I.G.M. Sheet 210-II NO (Lagonero)

Situated in a road cutting a short distance east of the Malamugliera Viaduct on the Superstrada del Noce (S.S.18), this 20m section includes part of the Lower Dolomite Member of the La Calda Formation at an unknown stratigraphic level. It comprises thinly bedded and laminated fine-grained dolomites interbedded with rare thin sapropelic horizons and thicker, coarser grained, graded dolomitic breccias.

MF-2 Marea - I.G.M. Sheet 210-II NO (Lagonegro)

51m of more or less dolomitised grey to black calcarenites, calcisiltites and calcilutites, intercalated with graded dolomitic breccias, are exposed in the road cutting beside the S.S.18 Superstrada del Noce between the Marea Viaduct and the exit for Lagonegro. This section, which occurs at an unknown level in the Limestone/Dolomite Member of the La Calda Formation, is notable for the large number of slumps present.

MF-3 Serra Luceta - I.G.M. Sheet 210-II NO (Lagonegro)

A 40m section through laminated or structureless dark grey or black calcarenites and calcisiltites of the Serra del Palo Formation can be observed on the west face of Serra Luceta. They occur at a high stratigraphic level in the formation but their precise position in relation to the other sections in the unit is not known.

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LA CALDA	DOLOMITE	20			Finely bedded dark-grey or brown laminated dolomites with rare concentrically banded chert nodules and thin sapropelic horizons
		10			Finely bedded, dark-grey, laminated and commonly graded fine- and medium grained dolomites, brown oxidised sapropelic horizons and rare bands of replacement chert

Fig. A23 Section MF1 Malamugliera

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
LA CALDA	LIMESTONE/DOLOMITE	5			
		50			Alternations of grey, laminated, fine-grained limestones and dark-grey dolomites in beds ≤ 10 cm thick, and graded dolomitic breccias
		40			Slumped laminated limestones and dolomites
		30			Graded dolomitic breeccias with planar-laminated tops Thinly bedded, graded and laminated limestones and dolomites
		20			Graded dolomitic breccias Slumped calcarenites and dolomites Dolomitised graded breccias Slumped calcarenites and dolomites
		10			Chaotically slumped calcarenites and calcilutites

Fig. A24 Section MF2 Marea

FORMATION	MEMBER	HEIGHT (m)	LITHOLOGY	SEDIMENTARY STRUCTURES	DESCRIPTION
SERRA DEL PALO	UPPER LIMESTONE	40			Well-bedded, dark-grey calcilutites, calcisiltites and calcarenites, commonly with parallel-laminations
		30			
		20			
		10			

Fig. A25 Section MF3 Serra Luceta

Appendix 2

Results of X-ray flourescence analyses

Sample No.	Description	Sr (ppm)
4022	Radiolarian biomicrite	282 ± 13
5217	Dolomitic chert nodule	271 ± 13
5218	Dolomitic chert nodule	268 ± 13
5220	Radiolarian biomicrite intraclast	699 ± 15
5229	Intrabiosparite	1007 ± 17
5231	Intrabiomicrite	608 ± 14
5233	Chert nodule with dedolomite	67 ± 12

Appendix 3

Results of electron microrobe analyses

Sample No.	Description	$CaCO_3$ Mol %	$MgCO_3$ Mol %	$FeCO_3$ Mol %	Sr (ppm)
4022	Radiolarian biomicrite	98.58	1.11	0.31	500
		98.10	1.47	0.40	300
		98.37	1.39	0.19	500
4022	Rhombic calcite from chert nodule	99.05	0.63	0.25	800
		97.51	2.19	0.15	1500
		98.86	0.65	0.38	1100
		97.97	1.36	0.49	2200
		98.66	0.85	0.30	1900
		98.51	0.86	0.43	2000
		98.96	0.66	0.27	1100
		98.70	0.73	0.42	1500
		99.42	0.34	0.15	900
		98.58	0.80	0.44	1900
		98.29	0.99	0.43	2900
		98.28	1.01	0.44	2700
4022	Dedolomite from chert nodule	83.58	4.85	11.55	200
		96.38	1.46	2.16	0
		93.09	1.46	4.57	0
5215b	Dolomite from dolomitic chert nodule	52.57	44.82	2.54	600
		52.53	45.51	1.90	600
		53.27	44.99	1.60	600
		52.92	45.58	1.47	400
		52.62	46.14	1.19	400
		52.86	45.80	1.28	600
		52.89	45.36	1.67	800
		51.70	46.37	1.87	600
5215b	Rhombic calcite from dolomitic chert nodule	99.01	0.79	0.08	1200
		98.94	0.84	0.12	900
		99.04	0.75	0.13	800
		99.14	0.71	0.06	900
5229	Intrabiosparite - cements	98.60	1.11	0.17	1200
		98.14	1.68	0.18	300
		98.19	1.63	0.17	100
5229	- intraclasts	98.35	1.64	0.01	1000
		97.95	1.92	0.05	800
		98.46	1.45	0.01	800
		97.98	1.92	0.01	900

Appendix 3 (cont.)

Sample No.	Description	$CaCO_3$ Mol %	$MgCO_3$ Mol %	$FeCO_3$ Mol %	Sr (ppm)
5139	Dolomitised limestone of the Sirino Formation	51.59	48.37	0.04	0
		51.56	48.40	0.01	290
		50.99	48.98	0.01	100
		51.43	48.56	0.00	100
		51.14	48.86	0.00	0
		50.59	49.41	0.00	0
		52.09	47.85	0.06	0
		51.17	48.74	0.08	100
5139	Horizontal dolomite veins in dolomitised limestone of the Sirino Formation	52.93	47.06	0.01	0
		51.19	48.80	0.00	100
		50.53	49.45	0.00	80
		50.83	49.16	0.01	0
		50.83	49.15	0.01	100
		50.88	49.08	0.04	0
		51.91	48.07	0.01	100
		51.30	48.68	0.01	100
		50.98	49.01	0.01	0
		51.97	47.97	0.06	0
		51.59	48.41	0.00	0
		50.78	49.21	0.00	100
		50.99	48.99	0.01	100
5344	Dolomitised calciturbidite of the Lagonegro Formation - dolomite rhombs	56.76	43.22	0.01	90
		56.72	43.28	0.00	0
		55.12	44.87	0.00	90
		56.22	43.76	0.01	90
		56.59	43.39	0.01	90
		56.61	42.98	0.01	100
		57.11	42.87	0.01	90
		57.53	42.40	0.12	0
		55.98	43.98	0.04	0
		60.76	39.15	0.08	100
		57.53	42.66	0.12	110
		56.29	43.70	0.01	0
		56.36	43.63	0.00	90
5344	- micritic grains	99.73	0.25	0.00	90
		95.77	4.21	0.00	270
		98.86	1.13	0.00	90
		98.61	1.38	0.00	90
		98.67	1.30	0.00	280
		98.67	1.29	0.01	280
5344	- biosparite grains	99.89	0.05	0.01	560
		99.56	1.27	0.01	90
		99.33	0.60	0.01	570
		98.48	1.50	0.01	90

Appendix 4 - Results of stable isotope analyses

Sample No.			$\delta^{18}O$ PDB	$\delta^{13}C$ PDB
4022	Rhombic calcite from chert nodule	Sasso di Castalda I	-4.01	+2.10
5215(i)	Radiolarian biomicrite	Monte Nicola	-3.85	+2.83
5215(ii)	Dolomite from chert nodule	Monte Nicola	-3.50	+3.97
5217(i)	Radiolarian biomicrite	Monte Nicola	-3.74	+2.85
5217(ii)	Dolomite from dolomitic chert nodule	Monte Nicola	-3.63	+3.72
5220	Radiolarian biomicrite showing evidence of early lithification	Monte Nicola	-3.87	+2.85
5221	Top of intraformational calcarenite	Monte Nicola	-4.34	+2.66
5222	Base of intraformational calcarenite	Monte Nicola	-4.15	+2.16
5229	Top of intraformational calcarenite	Monte Sirino	-4.76	+2.48
5459	Marl from limestone/marly limestone couplet	Gianni Griecu	-4.84	+2.73
5462	Limestone from limestone/marly limestone couplet	Gianni Griecu	-4.48	+2.07
5321	Radiolarian biomicrite	Pignola-Abriola II	-2.87	+2.75
5310	Intraformational bioclastic calcarenite	Pignola-Abriola II	-3.41	+2.57
5134	Intraformational bioclastic calcarenite	Pignola-Abriola II	-2.85	+2.41
5135	Base of intraformational intraclastic calcarenite	Pignola-Abriola II	-3.67	+2.30
5137	Radiolarian biomicrite	Pignola-Abriola II	-4.30	+2.81
5138	Dolomitised biomicrite	Pignola-Abriola II	-7.34	+2.63
5139a	Dolomitised biomicrite	Pignola-Abriola II	-9.53	+0.90
5139b	" " - horizontal vein	Pignola-Abriola II	-8.49	+1.27
5140	Dolomitised extraformational calcirudite	Pignola-Abriola II	-6.54	+2.05
5335(i)	Slightly dolomitised calciturbidite - calcite	La Ralla	-1.72	+1.15
5335(ii)	" " " - dolomite	La Ralla	-0.35	+1.94
5341	Oosparite from calciturbidite	La Ralla	-1.80	+1.65
5342(i)	Slightly dolomitised calciturbidite - calcite	La Ralla	-0.77	+1.30
5342(ii)	" " " - dolomite	La Ralla	-0.54	+2.07
5343(i)	Partially dolomitised calciturbidite -calcite	La Ralla	-1.34	+0.71
5343(ii)	" " " -dolomite	La Ralla	-0.18	+2.31
5345(i)	" " " -calcite	La Ralla	-0.35	+1.76
5345(ii)	" " " -dolomite	La Ralla	-1.42	+1.95
5352	Dolomitised limestone of the Sirino Formation	Monte Pierno	+0.29	+3.08
5353	Dolomitised extraformational calcirudite	Monte Pierno	-1.68	+2.39
5469	Dolomitised limestone of the Sirino Formation	S. Fele	-0.32	+2.94
4017	Dolomitised platform carbonate	Brienza	+1.32	+2.39
6058	Triassic platform carbonate	T. Pergola, Brienza	-1.38	+1.88
6059(i)	Dolomitised platform carbonate - calcite	T. Pergola, Brienza	-0.92	+2.14
6059(ii)	" " " - dolomite	T. Pergola, Brienza	+0.03	+2.50
6061	Triassic platform carbonate	T. Pergola, Brienza	-2.50	+1.63
6129	Dolomitised radiolarian biomicrite	Fanusi Fmt., Sicily	-4.83	+2.42
6130	Dolomitic breccia	Fanusi Fmt., Sicily	-5.26	+2.25
6131	Dolomitised biomicrite	Scillato Fmt., Sicily	+0.06	+3.02
6132	Radiolarian biomicrite	Scillato Fmt., Sicily	-2.59	+1.93
5057a	Nodular limestone - nodules	Monte Armizzone	-7.12	+1.17
5057b	" " - matrix - calcite	Monte Armizzone	-5.44	+0.79
5057b	" " " - dolomite	Monte Armizzone	-4.37	+2.42
5425	Nodular pelagic limestone	Valle del Pesce	-7.12	+0.92
6147	Middle Triassic pink nodular limestone	Palazzo Adriano, Sicily	-0.69	+2.37
6050	Tabular limestone unit	La Cerchiara	-5.55	+1.65
6055	Karstified tabular limestone unit	La Cerchiara	-5.89	+1.54
6007	Nodular pelagic limestone from olistolith	Fontana d'Eboli	-7.25	+0.86
6009	" " " " "	Fontana d'Eboli	-7.25	+0.86
5442	Pink micrite from neptunian dyke	Murge del Principe	-8.77	+1.80
6064	? Upper Cretaceous radiolarian biomicrite	T. Pergola, Brienza	-2.74	+0.91

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