Coeval extension, sedimentation and arc-volcanism along the Oligo-Miocene Sardinian Rift.

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Declaration

All the data, their interpretation and the conclusions drawn from them are my own work unless otherwise stated.

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Abstract

The Oligo-Miocene Sardinian Rift is an intra-arc basin which formed in response to multiphase extension and transtension on several orientations of normal and strike-slip faults. Rifting occurred during and after the separation and rotation of the Corsica-Sardinia microplate from Eurasia, coeval with northwestward dipping subduction of Neotethyan oceanic crust beneath the islands. The Sardinian rift comprises many semi-independent sub-basins which are filled with complex arrangements of non-marine and marine, siliciclastic, carbonate and marlstone sediments plus subduction-derived extrusive and pyroclastic volcanic rocks. Exposures in the Sardinian Rift provide a rare opportunity to study the evolution of, and processes active within intra-arc and back-arc basins. Field observations from along the Oligo-Miocene Sardinian Rift are presented and placed within a new chronostratigraphic framework. A tectono-stratigraphic synthesis of the Sardinian Rift has implications for the Oligo-Miocene tectonic development of the Western Mediterranean and for extensional settings in general. Geochemical analysis of volcanic-arc rocks provides clues as to what happens at depth when continental arc magmatism and extension are combined.

Rifting commenced in the mid-late Oligocene, coeval with the eruption of the first volcanic-arc rocks, whilst the Sardinia-Corsica microplate was attached to Eurasia. The resultant proto-Sardinian rift formed with considerable along-strike variability. It consisted of a N-S segment (post-rotation orientation) in northern Sardinia which intersected with NE-SW trending, elongate transtensional sub-basins. In southern Sardinia, the main rift segment was oriented NW-SE with separate E-W trending grabens dissecting the southernmost pre-rift basement. Geometries within continental clastic sediments shed from local topographic highs and basinward lacustrine limestones suggest that the first phase of extension was short-lived (<few Ma) and that tectonic subsidence far exceeded sedimentation rates. Strike-slip faults crossing the eastern Sardinian basement on inherited late Hercynian trends may have moved before and during the first mid-late Oligocene rifting phase. These strike-slip faults are thought to have facilitated the tectonic escape of continental crust from the Northern Apennines compressional zone towards the extending areas of western Sardinia and the southern Eurasian plate. Extension may have been driven by the steepening or roll-back of the subducting slab.

From the latest Oligocene until the early Burdigalian (early Miocene), extension was focused to the west of the Corsica-Sardinia microplate by the opening of the Western Mediterranean back-arc basin, which resulted in the separation and rotation of the Corsica-Sardinia microplate. Over this period, the arc-magmas which appear to have been derived from a constant mantle wedge source, may have been relatively depleted in an independently varying subduction-component in comparison to the oldest and youngest arc-volcanics. The subduction-component may have increased slightly to the north. The dominant signature was passive infilling of the Sardinian proto-rift structure by voluminous andesites, ignimbrites, marginal and shallow marine clastic and epiclastic sediments. Volcanic rocks blanketed
the accommodation space created by the older faulting event whilst clastic sediments were deposited in fan-deltas, localised by the degrading tectonic topography and reworked by marine currents. Complex intercalations and mixtures of non-marine and shallow marine volcanic-epiclastic-siliciclastic basin filling units occurred. In northern Sardinia, strike-slip fault movement recording possible tectonic escape continued until the latest Aquitanian. Late Aquitanian-early Burdigalian E-W normal faults with impressive syn-rift geometries in northernmost Sardinia may have formed due to the separation of the Corsican and Sardinian fragments as the microplate rotated away from the Eurasian plate.

In the mid Burdigalian, as back-arc spreading and the majority of arc volcanism ended, a second major extension phase occurred along the length of the Sardinian Rift. In northern Sardinia, complex sub-basins were caused by the interference of older rift structures and new N-S and E-W trending normal faults which were active towards the basin centre. Regional evidence suggests that this extension phase, and occasional, outcrop scale, late Burdigalian-Langhian syn-rift deposits associated with N-S faults, may be related to successive subduction zone roll-back to the south and east of Sardinia. From the late Burdigalian, the Sardinian Rift was dominated by shallow marine sedimentation following a late Burdigalian transgression over the faulted and degraded topography. The mixed siliciclastic-carbonate-marlstone basin fill was controlled by the relative sea level and sediment supply in response to the existing tectonic relief. Platform carbonates were deposited on fault-block highs whilst clastic material was focused by intersecting fault trends or along the elongate transtensional sub-basins. Calcarenites and marlstones accumulated in deeper waters in sub-basin centres.

In the Sardinian Rift, syn-rift deposits are localised and voluminously minor. The majority of basin fill records post-rift deposition partly controlled by pre-formed tectonic relief. Thus, extension and transtension were pulsed and short lived. A complex basin fill sequence results from the interaction of active tectonism, degrading tectonic relief, sea level change, active volcanism and clastic sediment supply. The relative sea level controlled the broad types of basin filling lithofacies. The evolving rift structure and active volcanism controlled the provenance, supply and accumulation of the basin fill. Arc-volcanism played a major role in basin filling by forming topographic edifices, blanketing tectonic topography and by affecting the composition of sediments. Individual elements of the Sardinian intra-arc rift history show similarities to other extensional settings. However, multiphase, variably oriented extension and transtension with a mixed siliciclastic-carbonate-volcanic basin fill makes the Oligo-Miocene Sardinian Rift a unique area, providing evidence for and enabling interpretation of these processes in an intra-arc setting.
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'In nature there are no models. Each environment and rock sequence is unique'

Reading and Levell 1996.
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Volume A: Text
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Chapter 1 - Introduction

1.1 Justification of research

The Oligo-Miocene Sardinian Rift (Fig. 1.1) is a rare example of a well exposed, extensional intra-arc basin formed on continental crust. Intra-arc basins contain thick volcanic, pyroclastic and sedimentary successions located within an arc (Ingersoll and Busby 1995; Smith and Landis 1995). The origin of and controls on sedimentation in such basins are poorly understood (e.g. White and Robinson 1992; Ingersoll and Busby 1995; Smith and Landis 1995) largely because few well-exposed, relatively undeformed ancient field examples exist. Modern examples are often submerged beneath the sea and there is a 'paucity of studies that integrate volcanology, sedimentology and basin analysis' (Ingersoll 1988). Intra-arc basins can be considered as incipient back-arc basins (Ingersoll and Busby 1995). The Sardinian Rift's evolution is associated with the opening of the Western Mediterranean back-arc basin. The Sardinian rift formed by multiphase extension on normal and oblique-slip faults and was created from numerous 'sub-basins' of different ages and structural styles. It was filled with non-marine to marine clastic, carbonate and marly sediments plus subduction-derived calc-alkaline volcaniclastic and volcanic rocks. Today, the rift is dissected and exposed at elevations up to 700m. The Sardinian Rift thus provides a rare opportunity to study a complex basin fill sequence resulting from coeval extension, sedimentation and volcanism with implications for specific processes active in intra- and back-arc basins and for Western Mediterranean geology in general. This thesis describes an integrated field-based study of the Sardinian Rift which provides detailed syntheses of basin evolution from well-exposed areas incorporating sedimentological, stratigraphical and structural data.

This chapter summarises the aims of the research and the main methods by which it was undertaken.

1.2 Aims of research

A diverse yet integrated study of the Sardinian Rift was facilitated by a multidisciplinary approach which utilised field-based observations combined with dating and geochemical work. The aims of the research can be broadly considered under four groups:

A) Structural evolution

A description of the structural geometries and fault kinematics which accommodated extension and the timing of extension. Key issues include:

- What is the role of inherited basement structures in rift development?
- Are fault styles and fault linkage similar to those in other extensional basins?
- Did the rift propagate through time?
• Does faulting young into the basin centre?
• Over what sort of timescales was faulting active?
• How much extension occurred?
• Was extension continuous or episodic?
• Did extension occur in all places at the same time?
• Are volcanism and faulting spatially and temporally related?

B) Nature of and controls on basin filling
A description of the sedimentary and volcanic facies which fill the rift, their provenance, dispersal and accumulation, basin filling geometries and the recognition of 'syn- and post-rift' basin filling. Key issues include:
• What are the roles of tectonics, relative sea level change and volcanism in controlling basin filling?
• How does the nature of the basin fill compare to other extensional settings where volcanism may not be as important?
• Are the sedimentary geometries and responses observed similar to those in other extensional basins?

C) Geochemistry of volcanic-arc rocks
An assessment of the geochemical evolution of parental arc magmas through time. Are there any systematic changes in magma geochemistry that may be linked to the observed phases of crustal extension in Sardinia?

D) Rift evolution
A tectono-stratigraphic synthesis placed within the framework of Western Mediterranean tectonics and back-arc basin opening. Questions posed include:
• How does the Sardinian Rift evolution fit into the regional tectonic framework and what are the implications?
• What was the relative elevation and subsidence history of the rift?
• What are the implications for the way in which volcanic arcs split and back-arc, intra-arc basins open?
• What might have been the geodynamic causes of the Sardinian intra-arc rift and the related back-arc basin?
• Why is the Sardinian Rift an abandoned intra-arc basin rather than a back-arc basin?
1.3 Previous work

1.3.1 Sardinian Rift

Accessible work published in international journals describes in detail specific aspects of the Oligo-Miocene Sardinian Rift. For example, the age and nature of arc volcanics (Coulon et al. 1973; Coulon and Dupuy 1975; Dostal et al. 1976; Bellon et al.; 1977, Savelli et al. 1979; Assorgia et al. 1984; Beccaluva et al. 1985; Morra et al. 1994, Lecca et al. 1997) and biostratigraphy/stratigraphy of logged sections (Pecorini and Pomesano Cherchi 1969; Pomesano Cherchi 1971b; Cherchi 1974). Palaeomagnetic studies and regional hypotheses concerning the tectonic history and rotation of the Corsica-Sardinia microplate abound (Auzende et al. 1973; Tapponnier 1977; Montigny et al. 1981; Cohen 1980; Burrus 1984; Rehault et al. 1984; Hill and Hayward 1988; Dewey et al. 1989; Carmignani et al. 1995; Vigliotti and Langhenheim 1995). Rift structure and geophysics are the subject of a few publications (Cherchi and Tremolieres 1984 see Appendix 1A and section 2.2.3; Pecorini et al. 1988; Balia et al. 1991) Geological maps (1:100,000) produced by the Servizio Geologico d’Italia in the 1950-70’s make some subdivisions of Oligo-Miocene stratigraphy and are generally accurate. A limited number of papers discuss the rift stratigraphy and evolution of specific areas (Thomas and Gennesseaux 1984; Assorgia et al. 1986a; Martini et al. 1992). Perhaps the seminal works on the Sardinian Rift are the related publications of Cherchi and Montadert (1982a,b) which provide a short synthesis of rift stratigraphy, structure and evolution placed in the Western Mediterranean tectonic framework (summary chart: Fig 1.2).

Less accessible are the numerous local studies concerning the stratigraphy or biostratigraphy of specific areas published in the Sardinian literature or in conference volumes (Cherchi 1971a; Porcu 1983; Leone et al. 1984; Spano and Asunis 1984; Cherchi 1985; Assorgia et al. 1988; Mazzei and Oggiano 1990; Assorgia et al. 1995a) and 1.25000, 1:50000 scale geological maps (Assorgia et al. 1986b; Assorgia et al. 1992c; Assorgia et al. 1993; Oggiano 1987; Assorgia et al. 1995b) These publications were acquired from contacts in Sardinia in the second and third field seasons.

The majority of previous work can therefore be broadly divided into three groups; local stratigraphic and biostratigraphic studies, palaeomagnetic and geochemical analyses of volcanic rocks and regional tectonic studies. Such work provides an invaluable basis for this study. However, present stratigraphy and stratigraphic nomenclature is confused (section 2.1.4) and apart from the brief descriptions in Cherchi and Montadert (1982a,b), only unpublished field guides (BP, ENSPM) utilise a tectono-stratigraphic approach. Local studies often take a geological leap from their findings to a regional tectonic scheme (Assorgia et al. 1984, 1986a versus Assorgia et al. 1995a). This approach is flawed because regional models are often simplified hypotheses and because the evolution of the rift is more complex than previously thought.
Whilst this study was in progress, other groups working on the Sardinian Rift were also active (Cagliari, Sassari, Roma Tre and Urbino Universities). Of particular value was the 'Fossa Sarda' conference held at Villanovaforru in June 1997 which gathered together all those working on the Sardinian Rift. As well as new biostratigraphic zonations (e.g. Serrano et al. 1997), radiometric dates (e.g. Dieno et al. 1997) and descriptions of local stratigraphy, some work at this conference provided the first real attempts at stratigraphic correlation along the whole rift (e.g. Assorgia et al. 1997c). The ideas of Carmignani et al. 1992b, 1994, 1995 (see 2.1.2.5, 2.1.5) concerning regional tectonic evolution seem to be gathering wide support and there exists an appreciation that the rift is not a simple graben created by one extensional event (as in Cherchi and Montadert 1982a,b). Whilst this thesis was being assembled in November 1997, Lecca et al. (1997) published an independent review of the Sardinian volcanic sequences and rifting stages. The publication proposes a new volcanic lithostratigraphy applicable to the whole rift basin and considers the rift evolution in three stages. It is quoted and incorporated into the thesis wherever applicable, though the ideas presented here were formulated entirely independently.

There are therefore, a dearth of integrated tectono-stratigraphic studies correlated along the rift basin and process-based studies recording the general significance of this unique area. Particularly lacking are publications which illustrate, with data, the nature and timing of faulting in the rift basin, the response of the basin fill to faulting. Published work in general does not consider the Sardinian Rift as a natural laboratory to study extension, sedimentation and volcanism but rather concentrates on describing the geology of Sardinia. The research presented here aims to rectify this situation.

1.3.2 Intra-arc basins and arc volcanism

In the last decade, numerous studies of modern and ancient volcanic and volcaniclastic facies have greatly increased our knowledge of the nature of and controls on volcanic basin filling in extensional settings (e.g. Fisher and Schminke 1984; Geol. Soc. Spec. Publ. 25 (eds) Frostick et al. 1986; Cas and Wright 1988; Fisher and Smith 1991; R. Smith 1991, Busby and Ingersoll 1995; Orton 1996). Detailed basin analyses of intra-arc and back-arc rifts have started to rationalise the evolution of individual areas (e.g. Sato and Amano 1991; White and Robinson 1992) though the origins of such basins are still poorly understood. In continental arcs, links between extension and subduction-derived volcanism are commonly cited (Cole 1984: Pe-Piper et al. 1994: Petford and Atherton 1994). One possible explanation evaluated here, is the proposition that rifting events preceding back-arc basin opening are linked at depth to lithospheric thinning, as suggested for the Lau basin (Clift and Dixon 1994).
1.3.3 Extensional basins and basin filling

Inherent in this study was an appreciation of the published literature and current thinking on the formation of extensional and back-arc basins (e.g. Mckenzie 1978; Karig 1974), the style and rates of normal faulting observed in such settings (e.g. Jackson et al. 1988; Jackson and White 1989; Morley et al. 1990; Gawthorpe and Hurst 1993 etc.) and of sedimentation in extensional settings (e.g. Rosendahl et al. 1986; Leeder and Gawthorpe 1987; Burchette 1988; Prosser 1991, 1993; Lambiase and Bosworth 1993; Collier and Gawthorpe 1995; Gawthorpe et al. 1997; Sharp et al. in press).

1.4 Logistics and methods

1.4.1 Study areas

The complexity of the Sardinian Rift basin meant that the approach taken was to subdivide the region into 'study areas' based on geographical location and geological characteristics such as structural style and basin fill sequence (Fig 1.3). This thesis describes 6 'study areas' from along the Sardinian Rift and from Corsica (Fig 1.3) from south to north. Some of the study areas or parts of them are termed 'sub-basins' where a subsided region can be described by a coherent geological evolution. The Sarcidano, Anglona, Funtanazza and Logudoro areas were studied in detail, including mapping at 1:25000 scale whilst the strike-slip systems of eastern Sardinia and Miocene basins of Corsica were studied at a reconnaissance level.

1.4.2 Logistics

Fieldwork totalling 28 weeks was split into three field seasons; August-September 1994, April-July 1995 and September-October 1996. Field studies utilised the Istituto Geografico Militare 1:25000 topographic maps. More information was gained from field samples via petrographic microscopy, nannofossil dating (section 2.2.1), Sr isotope stratigraphy (section 2.2.2), Ar-Ar dating (2.2.3), X-ray diffraction (Appendix 5B) and X-ray fluorescence (Appendix 9A,B). In particular, follow-up work was used to provide a high resolution temporal framework necessary for interpreting field data.

1.4.3 Fieldwork Methodology

The Oligo-Miocene Sardinian rift is moderately exposed. Though natural exposures occur in volcanic, limestone or basement rocks and along coastal sections, the best outcrops are often man-made (quarries or road cuts). The level of exposure does not generally allow the tracing of horizons for any considerable lateral extent (more than a few tens of metres). The basin fill sequence is complex with rapid lateral and vertical facies changes. Therefore the approach taken in the field was to:

- deduce a lithostratigraphy for each studied area, constrained, where available, by biostratigraphy and radiometric dates.
• carry out facies analysis to build up a picture of the laterally varying and temporally evolving palaeoenvironments using traditional, accepted criteria (e.g. in Leeder 1988, Reading 1986, 1996).
• record data concerning the supply, provenance, dispersal and accumulation of the basin fill.
• describe and measure sedimentary and volcanic geometries and basement/basin fill geometries with particular emphasis on syn- and post-rift deposits
• describe and measure faults, folds etc.
• use all this information to build tectono-stratigraphic hypotheses for each area. The ideas were developed and tested in the field.
• undertake focused sampling programs e.g. for geochemical analysis of basalts and Sr-isotope stratigraphy of carbonates.

Standard techniques such as mapping at various scales, logging, structural measurement, outcrop sketches, clast counting, and measurement of palaeocurrents were employed. Appendix 1A describes the techniques in more detail. Figure 1.4 and Appendix 1B provide an explanation of the symbols and definitions used throughout the thesis.

1.4.4 Limitations

Though the aims of this study were broad and there were a huge number of tasks that would ideally be attempted, work was often limited by several factors. The main limitation was the level of exposure both in terms of the amount of outcrop and the burial of the Oligo-Miocene sequence. Some potential questions (for example, ‘was there uplift before rifting commenced?’) could not be addressed because of the type of facies deposited. In addition, poor temporal control on some stratigraphy and contradictions between different dating methods (section 3.4.2) limit interpretations in some areas.

1.5 Thesis structure

This first chapter provides an introduction to the research whilst Chapter 2 summarises in a literature review the regional geological setting and geology of Sardinia. Chapter 3 sets out the results of Sr isotope stratigraphy and Ar-Ar dating and defines the stratigraphy observed in each study area. The main body of this thesis, Chapters 4-8, describes the sedimentology and geometries of the basin fill, the structure and tectono-stratigraphic development of each study area in turn. This approach has been used because of the complexity and individuality of the studied areas. Chapter 9 describes the results of a focused geochemical analysis of Sardinian basalts and basaltic andesites. Chapters 10 and 11 form a synthesis of this study, by describing the tectono-stratigraphic development of the Sardinian rift, the relationships to Western Mediterranean regional tectonic evolution and by examining processes and features observed in this intra-arc basin. So that they may be viewed together, the thesis is split into a volume of text and appendices(A) and a volume of figures, tables and enclosures (B).
Chapter 2
Chapter 2 - Geological setting

This chapter summarises the regional tectonics of the Western Mediterranean, includes a brief description of the geological history of Sardinia and the structural framework defining the studied sub-basins.

2.1 Cenozoic Tectonics of the Western Mediterranean

The purpose of this literature review is to summarise the regional geological history of the Western Mediterranean in the Cenozoic with particular emphasis on and implications for Sardinia. Such a summary is necessary since the evolution of the Sardinian rift and geodynamic causes of extension are intimately related to the regional tectonic setting. Chapter 10 discusses the detailed evolution of the Sardinian Rift and how this may relate to the regional tectonic history. A great deal of information on the Western Mediterranean exists and a number of contrasting models have been suggested. Therefore, care has been taken to consider the geological data as much as possible. Figure 2.1 illustrates the localities mentioned in the text.

2.1.1 Cretaceous-Palaeocene-Eocene

It is widely accepted that before the opening of the Western Mediterranean basin in the early Miocene, the Corsica-Sardinia microplate formed the southern edge of the Eurasian plate (e.g. Cohen 1980; Dewey et al. 1989; Carmignani et al. 1995; Robertson and Grasso 1995; Figs 2.2-2.6). Convincing evidence includes the similarity in the pre-Oligocene geological history of Corsica, Sardinia, northern Spain and southern France (Le Pichon et al. 1971; Montigny et al. 1981; Cherchi and Montadert 1982a b; Le Douran et al. 1984; Sartori et al. 1987) and the proposed good fit of the microplate to the southern Eurasian plate (Le Pichon et al. 1971; Auzende et al. 1973). The Calabria and the Kayblie continental fragments are thought to have been somewhere out with this region and the Balearics can be restored to a position before the opening of the Valencia Trough (e.g. Cohen 1980; Rehault et al. 1984; Hill and Hayward 1988; Dewey et al. 1989).

2.1.1.1 Africa-Eurasia collision and related compressional events

The collision of the ‘African’ and ‘Eurasian’ plates from the Cretaceous to the present day has dominated the geological history of the circum-Mediterranean region (e.g. Hill and Hayward 1988; Dewey et al. 1989; Robertson and Grasso 1995). The collision commenced when the relative movement of the African plate changed to a northeasterly direction in the Cretaceous at ~92 Ma (Dewey et al. 1989). In the Western Mediterranean, the most conspicuous result of this plate convergence was the development of the Alpine, Pyrenean and Betic compressional belts and the closure of Mesozoic Neotethys (Dewey et al. 1989; Robertson and Grasso 1995). Associated
In southern France, E-W trending thrusts were developed during the late Cretaceous-late Eocene 'Pyreneo-Provencal' tectonism (Guieu and Roussel 1990; Hippolyte et al. 1993; Benedicto et al. 1996; Mauffret and Gorini 1996). The compression led to a regional uplift (Guieu and Roussel 1990).

In Corsica, two compressional phases are distinguished, a late Cretaceous ophiolite stacking and a late Eocene-Oligocene continental collision (Egal 1992) which ended in late Oligocene time (Jolivet et al. 1991). The Corsican 'Alpine' tectonics (i.e. Eocene age) have been interpreted both as continuation of the Alps collisional belt (e.g. Jolivet et al. 1991) or as part of a doubly-vergent accretionary wedge extending to the Northern Apennines, and associated with the subduction of Neotethys beneath the Corsica-Sardinia block from the Cretaceous (Carmignani et al. 1994, 1995).

In Sardinia, Cherchi and Tremolieres (1984) recognised an "Alpine Austrian" compressional phase of mid-late Cretaceous age oriented at 100° (pre-Miocene rotation orientation) within the Mesozoic strata of northern Sardinia. A 'Pyrenean', early/mid Eocene, compressional phase is also recorded with a NW-SE (post-rotation) shortening direction in the Monte Albo area of eastern Sardinia (Fig. 1.1) and far northwestern Sardinia (Cherchi & Montadert 1982a,b). Due to this compressional phase, early Eocene marine facies were unconformably covered by Lutetian (mid Eocene) - early middle Oligocene continental clastic sediments of the Cixerri formation which are believed to have been sourced from northeastern Iberia based on palaeocurrent information, petrologic and faunal constraints (Pittau Demelia 1979; Cherchi and Schroeder 1976, Cherchi and Montadert 1982a,b). Sartori et al. (1987) state that late Eocene deformation in Sardinia increased eastwards, resulting in gentle folding and block faulting of the Mesozoic rocks though they do not give any details. The tectonic structures and deformation resulting from Cretaceous-Eocene compression in Sardinia are poorly defined and it is therefore difficult to assess their role in the subsequent geological history.

2.1.1.2 Initiation of subduction beneath the Corsica-Sardinia microplate
Convergence of Africa and Eurasia led to the northward subduction of Neotethyan oceanic crust (e.g. Beccaluva et al. 1987; Hill and Hayward 1988; Robertson and Grasso 1995; Figs 2.2-2.5). The calc-alkaline, subduction derived, volcanic rocks found in Sardinia, the Gulf of Valencia, Provence and offshore Corsica (the volcanic arc) dated at ~33-15 Ma (Beccaluva et al. 1985, 1987, 1994; Girod and Girod 1977; Maillard and Mauffret 1993; Fig. 2.2) are the evidence that northward/northwestward directed subduction of Neotethys oceanic crust occurred from the mid-Oligocene-mid Miocene. Some authors argue that subduction commenced in the late Cretaceous (Hill and Hayward 1988; Carmignani et al. 1994, 1995; Robertson and Grasso 1995) whereas Beccaluva et al. (1987) and Tapponnier
(1977) favour a late Eocene age due to consideration of the timing of volcanism and regional tectonic considerations respectively. Accretionary wedges which were later deformed and emplaced onto mainland Italy (e.g. Liguride Complex) developed in response to the subduction system (Knott 1987; Hill and Hayward 1988; Carmignani et al. 1994, 1995; Monaco and Tortorici 1995; Fig. 2.3)

2.1.2 Late Eocene-Burdigalian

2.1.2.1 Extension of the southern Eurasian plate and initiation of the Western Mediterranean Basin

Some extensional systems of Eocene age are observed. In southern Sardinia for example, lower and upper Eocene sediments accumulated in east-west (post-rotation orientation) trending grabens (Cherchi and Montadert 1982ab; Assorgia et al. 1992bce; Barca and Costamagna 1997; Fig. 1.1). The larger, N-S oriented, late Eocene-late Oligocene ‘Western European’ rift systems of the Bas-Rhone, Rhine and Bresse grabens were initiated by a change in Africa-Eurasia plate motions at the end of the Eocene (Hippolyte et al. 1993). The rifting propagated southwards down the Rhone valley (Banda and Santenach 1992) and was cross-cut by mid/late Oligocene-early Miocene, NE-SW trending extensional faults of the ‘Liguro-Provençal’ rift system (Hippolyte et al. 1993).

From the mid-late Oligocene, an array of NE-SW- NNE-SSW trending extensional faults developed in southern France, western Sardinia and the Gulf of Lion, at the southern margin of the Eurasian plate (e.g. Cherchi and Montadert 1982ab; Le Douran et al. 1984; Rehault et al. 1984; Dewey et al. 1989; Hill and Hayward 1988; Robertson and Grasso 1995; Figs. 2.2-2.5). As a result of this extensional faulting, the Corsica-Sardinia-Balearic-Kayblie-Calabria system and part of the volcanic arc began to separate or rift from the southern European plate (Figs. 2.3, 2.4).

2.1.2.2. The Western Mediterranean basin rifting period.

Onland exposures in Provence and western Sardinia together with offshore data from the Gulfs of Lion, France and Valencia, Spain, facilitate a description of the timing and geometries of active extension at the southern margin of the Eurasian plate.

In southern mainland France and in the Gulf of Lion, there was a strong control by earlier structures formed by the Pyreneo-Provençal compressional events (Gorini et al. 1993; Benedicto et al. 1996; Mauffret and Gorini 1996). The structures were inverted such that in places, extension occurred on low-angle detachments (Gorini et al. 1993; Benedicto et al. 1996; Mauffret and Gorini 1996). Rehault et al. (1984) believe that in southern France and the Gulf of Lion most extension and crustal thinning took place in the late Oligocene, since bounding faults show clear syn-sedimentary activity until the late Oligocene, perhaps up to the early Miocene. By the Aquitanian, the rifting phase had ended from observations of sediments in the Gulf of Lion (Cravatte et al. 1974, Biju -Duval et al. 1987). The Gulf of Valencia is another aborted branch of the ~NE-SW trending extensional system and is thought to
have been active in late Oligocene-early Miocene times (Cohen 1980; Banda and Santenach 1992; Maillard and Mauffret 1993). Seismic studies show that the basin is formed on thinned continental crust (Banda and Santenach 1992). The Sardinian Rift would also have been oriented approximately NNE-SSW before the rotation of the Corsica-Sardinia microplate (section 2.1.2.3). In Sardinia, published data constrains the rifting period between post 29.9 Ma (mid-late Oligocene) until mid Aquitanian times (23-24 Ma, Cherchi and Montadert 1982ab). The evidence for syn-rift activity is defined by Cherchi and Montadert (1982ab) as the disposition of Aquitanian sediments in half-graben and the faulted offsets of shallow marine lower Aquitanian limestones. The end of the syn-rift phase was dated by the sealing of faulted blocks by later Aquitanian sediments (Cherchi and Montadert 1982ab, see section 5.5). Utilising a similar database, Carmignani et al. (1994, 1995) believe that rifting in Sardinia and the surrounding Balearic basin (Western Mediterranean basin) occurred from the Late Aquitanian and mainly in the Burdigalian. As discussed in Chapter 10, two main, mid-late Oligocene and mid Burdigalian, extensional phases can be detected in the Sardinian Rift.

Thus, the phase of active rifting that preceded the opening of the Western Mediterranean basin in Sardinia, southern France and the Gulf of Valencia is documented as mid-late Oligocene - early Miocene in age. The end of active rifting which was marked by the change from syn-rift to post-rift deposition and ‘thermal subsidence’ was associated with the separation and rotation of the Corsica-Sardinia microplate, the volcanic arc and the formation of the oceanic Western Mediterranean back-arc basin (Cherchi and Montadert 1982ab; Rehault et al. 1984; Figs 2.3-2.5).

2.1.2.3 The Western Mediterranean back-arc basin and coeval rotation of the Corsica-Sardinia microplate.

The drift and/or rotation about a pole in the Gulf of Genoa of the Corsica-Sardinia microplate contemporaneous with oceanic accretion was first considered in detail by Le Pichon et al. (1971) and Auzende et al. (1973) based on magnetic anomalies, lineations and structural features observed in the basin. The Western Mediterranean Basin is commonly considered a marginal or back-arc basin which formed as a continental fragment (Corsica-Sardinia), volcanic arc (on Sardinia) and trench-subduction system migrated to the south and east (e.g. Rehault et al. 1984; Beccaluva et al. 1987; Robertson and Grasso 1995; Figs 2.3-2.5). The timing and kinematics of the back-arc basin formation and microplate rotation are reasonably well constrained.

Palaeomagnetic studies on the rotation of the Corsica-Sardinia microplate, placed within a time framework by radiometric or biostratigraphic dating, constrain Western Mediterranean Basin opening. Montigny et al. (1981) provide a widely accepted dataset which combines palaeomagnetic analysis with K-Ar dating of volcanic mineral separates from outcrops in Sardinia. A counterclockwise rotation of 30° between 20.5 and 19 Ma is deduced, in agreement with earlier work (e.g. 30°, Bellon
et al. 1977; 30° 21.7-17.6 Ma, Edel 1980) but over a more accurately defined timescale. After a re-
assessment of existing palaeomagnetic data by Todesco and Viglotti (1993) suggesting that the end of
rotation was unconstrained, Vigliotti and Langhenheim (1995) and Speranza et al. (1997) used new
samples to propose that most rotation had occurred by ~18Ma but did not end until about 15 Ma.
Aeromagnetic studies also illustrate deep-rooted structures (?Miocene volcanic bodies) with a 30°
westward remnant magnetism thought to result from the rotation of Sardinia (Galdeano and Ciminale
1987). A palaeomagnetic study with samples from either side of the Straits of Bonifacio (Vigliotti et
al. 1990) confirms the theory of no relative rotation between Corsica and Sardinia as suggested by the
geological similarities (Arthaud and Matte 1977) and geophysics (Egger et al. 1988; Blundell et al.
1992). Thus Corsica and Sardinia formed a coherent rotating microplate (Burrus et al. 1984; Vigliotti
et al. 1990).

The magnetic anomaly pattern and structures in the Western Mediterranean Basin can be used to
deduce the geometry of basin opening and possible correlations to the magnetic anomaly timescale
(Burrus 1984; Rehault et al. 1984; Vigliotti and Langhenheim 1995; Fig. 2.7). Three main spreading
axes are identified; NE-SW trending in the Western Mediterranean basin, NW-SE in the south
Balearic basin and E-W in the northern Algerian Basin (Rehault et al. 1984; Fig. 2.7). Spreading
segments are bounded by fracture zones (Rehault et al. 1984; Fig. 2.7). The magnetic anomaly pattern
was interpreted by Burrus (1984) such that oceanic accretion occurred at magnetic anomaly 6B-6 time
(23.5-19 Ma) whereas Vigliotti and Langhenheim (1995) reinterpret the end of accretion at magnetic
anomaly 5A (~15 Ma). Estimates of spreading rates thus vary from 16cm/yr (Cohen 1980) to
4.5cm/yr (Vigliotti and Langhenheim 1995). The spreading age of the Western Mediterranean Basin
is also constrained by K-Ar radiometric ages of 18±0.5 Ma and 19 Ma (Rehault et al. 1984; Cherchi
and Montadert 1982a,b) from ‘Tristanite’ dredged from the axis of the Ligurian sea.

Seismic studies provide evidence of the geometry of the back-arc basin. Stretched continental crust,
only with tilted fault block/half graben geometries forms the ‘passive margins’ to the basin and
oceanic crust 5-7km thick is present in the basin centre (Burrus 1984; Le Douran et al. 1984; Rehault
et al. 1984). The nature of the ‘passive margins’ surrounding the ocean basin is variable. The Gulf of
Lion is a wide (100km) region of thinned continental crust whereas the NW Sardinian, Ligurian and
Corsican margins are narrow and steep (~25km, Fig. 2.2, Le Douran et al. 1984). The cause of this
difference is thought to be the inherited Mesozoic and Eocene structures present in the Gulf of Lion
(Mauffret and Gorini 1996).

Burrus (1984) used many of these criteria, combined with field evidence of active rifting to constrain
oceanic spreading in the Western Mediterranean Basin and the rotation of the Corsica-Sardinia
microplate from 21-19 Ma, after an earlier ‘drifting’ phase of microplate translation from 23-21 Ma.
Rehault et al. (1984) used similar information to suggest oceanic spreading from 21-18 Ma, the timing which is used in this thesis. Thus, by the early-mid Miocene the Corsica-Sardinia microplate was oriented approximately north-south as it is today and the Western Mediterranean back-arc basin had formed (Figs 2.2-2.5). The migrating subduction zone is likely to have been somewhere to the east of the Corsica-Sardinia microplate. Where oceanic crust no longer remained, for example at the northeastern end of the subduction zone, continental collision occurred (section 2.1.2.5, Figs. 2.3-2.5).

2.1.2.4 Oligo-Miocene calc-alkaline volcanic arc

Calc-alkaline volcanic rocks derived from the subduction of oceanic crust beneath the Corsica-Sardinia microplate, are best exposed on Sardinia and found in southern France, offshore western Corsica and Sardinia, the Gulf of Valencia and the Kabylies (Cann and Hsü 1973; Dupuy, et al. 1974; Coulon and Dupuy 1975; Girod and Girod 1977; Dewey et al. 1989; Maillard and Mauffret 1993; Fig. 2.2) An Andean-type volcanic arc can be envisaged at the southern margin of the European plate such that arc-volcanism occurred coeval with extension (Macciotta et al. 1978; Robertson and Grasso 1995; Mauffret et al. 1996; Fig. 2.5) The arc supplied material into the trench-accretionary wedge system of the subduction zone to the southeast, now present within compressional belts in northern Africa, Sicily and the Apennines (Girod and Girod 1977; Sartori et al. 1987; Dewey et al. 1989). Volcanism began at ~34 Ma in Provence and at ~33 Ma in Sardinia (Beccaluva et al. 1985, 1987, K-Ar dates). Andesites and tuffs found in the Kabylies are dated at 19.1±1 Ma (in Cohen 1980) and volcaniclastic layers in the Northern Apennines are dated at 21-20 Ma (Odin et al. 1994). The migration of the subduction zone to the east during opening of the Western Mediterranean Basin results in volcanism ending at ~20 Ma in Provence but continuing in Sardinia until ~13 Ma (Beccaluva et al. 1987) or ~18Ma (this study, chapter 3).

In Sardinia, the early phase of volcanism (33-26 Ma) shows characteristics of an immature arc, has calc-alkaline affinity (some tholeiites in southern Sardinia) with spiderdiagrams typical of a subduction signature (Beccaluva et al. 1994). From about 26 Ma, volcanism was more voluminous (Beccaluva et al. 1987) with the first ignimbrites at 23 Ma (Coulon 1977; Savelli et al. 1979). The maximum volcanic intensity occurred contemporaneous with rifting and oceanic spreading (Beccaluva 1987, 1994). A significant north-south zonation of magmatism occurred from 21-18 Ma (Beccaluva et al. 1987). The N-S tholeiitic to calc-alkaline character (Coulon and Dupuy 1975) and an increase in K, Rb, Li, Sr and Ba and light rare earth elements (LREE) northward (Coulon and Dupuy 1975, Coulon 1977) are cited as evidence for a northward dipping subduction slab (Coulon and Dupuy 1975; Cohen 1980; Beccaluva et al. 1987). Beccaluva et al. (1994) link back-arc spreading, microplate rotation and geochemical zonation of volcanic rocks together to imply a deepening of the subducting slab at 21-18 Ma. After ~18 Ma (this study), western Sardinia became a remnant volcanic arc.
2.1.2.5 Compression, transpression, transtension and extension in belts surrounding the Western Mediterranean basin.

Contemporaneous or subsequent compression occurred in areas surrounding the extensional region of the Western Mediterranean Basin, South Balearic basin and Alboran Sea in an arcuate belt from the Betics to the Northern Apennines (Figs. 2.3-2.5). Compression occurred diachronously when Neotethyan oceanic crust had been consumed and/or extensional basins opened and the migrating subduction zone, accretionary wedge and fragment of continental crust collided with part of the Africa-Apulia plate (Robertson and Grasso 1995; Lonergan and White 1997).

In the Northern Apennines, continental collision and emplacement of oceanic crust is recorded in the late Oligocene-Aquitanian (Kligfield et al. 1986 and Carmignani and Kligfield 1990 in Carmignani et al. 1994, 1995) The start of deformation has been dated at 27 Ma (late Oligocene, Kligfield et al. 1986 in Dewey et al. 1989) Emplacement is commonly bracketed between the late Oligocene-early Miocene (Robertson and Grasso 1995) and compression which migrates to the northeast is thought to have ended in the western Northern Apennines by the Langhian or Tortonian (Sartori 1990; Robertson and Grasso 1995). A younger 'intra-Burdigalian' age for with the first strong Adria-vertgent compression and also for north African deformation was preferred by Cherchi and Montadert (1982a,b), Rehault et al. (1984) and Sartori (1990) who relate this convergence directly to the end of rotation of the Corsica-Sardinia microplate.

In Corsica, compression which commenced in the Eocene (section 2.1.1.1) may have occurred until the late Oligocene (Jolivet et al. 1990, 1991; Carmignani et al. 1994, 1995). Carmignani et al. (1994, 1995) argue that this compressional phase records Corsica-Apulia collision in a system with opposite vergence but equivalent of the northern Apennines event (Fig. 2.8a), though the published data suggests that the Corsican compression was slightly older than the Northern Apennines. In Calabria, Sicily, the southern Apennines and north Africa, compression occurred from the late Oligocene-late Miocene, peaking in the early-mid Miocene (Knott 1987; Channell and Mareschal 1989; Sartori 1990; Monaco and Tortorici 1995; Robertson and Grasso 1995; Fig. 2.5). Neogene, outward directed thrusting in the Betic and Rif-Tell areas occurred contemporaneously with the late Oligocene-late Miocene extension of the Alboran Sea (Platt and Vissers 1989), whilst at the SE margin of the Valencia Trough and in the Balearics, compression occurred in the late Oligocene-mid Miocene as a continuation of the Betic belt (Banda and Santenach 1992; Torres et al. 1993; Fig. 2.5, 2.6). In Sardinia, Cherchi and Montadert (1982a,b) and Cherchi and Tremolieres (1984) record a phase of Burdigalian compression oriented at 40° which they relate to the end of Corsica-Sardinia microplate rotation and collision with the Apulian block. No evidence for this phase was recorded during this research.
Transpression was caused by movement on NE-SW and N-S trending sinistral strike-slip faults in eastern Sardinia and southwestern Corsica (Carmignani et al. 1992b, 1994, 1995; Figs 2.5, 2.8). In the M. Albo area of Sardinia (Fig. 1.1), this phase occurred sometime after the post Palaeocene and before the Burdigalian, probably in the late Oligocene-Aquitanian according to Carmignani et al. (1992b, 1994, 1995) rather than in the mid-Eocene (Cherchi and Montadert 1982a; Cherchi and Tremolieres 1984). Activity on the sinistral strike-slip faults also caused late Oligocene-Aquitanian age pull-apart basins to be formed on ‘releasing bends’ in central Sardinia (Oggiano et al. 1995). The seismo-stratigraphic studies of Thomas et al. (1988) suggest the presence of Oligo-Miocene age transcurrent flower structures within a dominantly extensional regime offshore central western Sardinia. Carmignani et al. (1994, 1995) see transpression in Sardinia and SW Corsica as a hinterland to the contemporaneous compression in the Northern Apennines and Corsican collisional belts (Fig. 2.8a,b) in the late Oligocene-Aquitanian.

Ductile and brittle extension is described in ‘Alpine’ Corsica by Jolivet et al. (1990, 1991) commencing in the late Oligocene (no temporal data provided) and continuing until the late Miocene (deformation of dated sediments). The initial ductile extension reactivated the older compressive thrust contacts and later, low to high angle brittle normal faulting cut lower Miocene sediments (Jolivet et al. 1990, 1991). Carmignani et al. (1994, 1995) propose that the extension in Corsica and the Northern Apennines, as well as over Sardinia, the Western Mediterranean Basin and the northern Tyrrenian sea commenced in the Burdigalian.

2.1.3 Langhian - Serravalian -Tortonian

From the mid-Miocene onwards, compression and outward thrusting at the orogenic front of the Apennines took place with extension in the orogenic hinterland (Malinverno and Ryan 1988; Robertson and Grasso 1995). The coupled extensional-compressional belt migrated progressively towards the northeast and east in the northern and central Apennines respectively (Dewey et al. 1989; Robertson and Grasso 1995; Carmignani et al. 1995). In the late Tortonian, the locus of extension migrated to the east of Sardinia and resulted in the initiation of the Tyrrenian Sea and the eastern Sardinia passive margin (Kastens et al. 1988). The migrating compressional belt became more ‘arcuate’ with time and the Apulian plate rotated anticlockwise (Hill and Hayward 1988; Dewey et al. 1989; Lonergan and White 1997).
2.1.4 Late Miocene-Recent

The extension which started in the late Tortonian continued to the present day with the locus of extension moving progressively to the southeast to form the Tyrrenian Sea (Kastens et al. 1988; Fig. 2.3, 2.6). One idea is that extension was caused by the roll back of the subduction zone which was under the Corsica-Sardinia microplate, to its present position underneath the Aeolian arc (Kastens et al. 1988; Malinverno and Ryan 1986; Lonergan and White 1997; Fig. 2.6) through a strand of oceanic crust in the Apulian platform (Malinverno and Ryan 1986). Robertson and Grasso (1995) suggest that a new subduction system became active from the early Miocene onwards related to removal of oceanic crust to the southeast (in the vicinity of the Ionian Sea) and that this new system drove supra-subduction zone extension in the Tyrrenian Sea. Another theory is that extension in the Tyrrenian Sea occurred as a response to the lithospheric delamination of overthickened continental crust (Channell and Mareschal 1989; Carmignani et al. 1995; Fig. 2.8c) resulting from earlier continental collision.

In Sardinia, extension is evident in the formation of the Pliocene Campidano graben (Cherchi and Montadert 1982ab; Sartori et al. 1987) which re-uses older Oligo-Miocene extensional faults (Thomas et al. 1988) and occurred in the same time frame as the eruption of 5.3 - 0.9 Ma age alkaline basalts (Beccaluva et al. 1985; Macciotta et al. 1978, Fig. 1.1, 2.10). Dyke swarms of <3.4 Ma in Sardinia are oriented N-S, implying that the extension direction from this time was east-west (Feraud and Campredon 1983) consistent with Plio-Quaternary N-S trending extensional faults (Carmignani et al. 1994).

2.1.5 Validity of plate tectonic reconstructions

Plate tectonic reconstructions regarding the opening of Western Mediterranean basin can be subdivided into two groups:

1) Models which propose that the early Miocene rotation of the Corsica-Sardinia microplate was the cause of compression in the northern Apennines due to resultant space problems (e.g. Rehault et al. 1984; Dewey et al. 1989; in Channell and Mareschal 1989; Robertson and Grasso 1995). In this scenario, 'Alpine' compression in Corsica occurred previous to the Late Oligocene as a separate tectonic event.

2) The model of Carmignani et al. (1994, 1995) in which Oligocene compression in Corsica and the Northern Apennines occurred contemporaneously. Subsequently, Burdigalian extension in Sardinia, the Western Mediterranean Basin and the northern Tyrrenian Sea took place. Thus, extension and
Corsica-Sardinia microplate rotation happened at the same time as extension in Corsica, the Northern Apennines and what was to become the northern Tyrrhenian Sea (Fig. 2.8b).

In the first model, late Oligocene-early Miocene extension in Sardinia occurred as the easternmost part of a rift system affecting the Eurasian plate (Cherchi and Montadert 1982a,b). Space problems resulting from microplate rotation and back-arc basin formation caused compression in areas to the east, the end of extension in Sardinia and a possible compressional phase in Sardinia from the Burdigalian (e.g. Cherchi and Montadert 1982ab; Rehault et al. 1984). In the second model, Sardinia acted as a hinterland to Corsica-Apennines collision in the Oligocene-lowermost Miocene and extension commenced only after the late Aquitanian (Carmignani et al. 1994, 1995).

Crucial to distinguishing the true scenario is the timing of events and a detailed history of the Sardinian Rift evolution. In the published literature there seems to be clear evidence from age data that, contrary to both models, Eocene-Oligocene extension in the western areas of southern France, southern Sardinia, Valencia Trough, etc. occurred contemporaneously with compression in eastern areas such as northern Corsica and the Northern Apennines. Chapter 10 discusses a model for the Oligo-Miocene evolution of Sardinia and implications for the regional tectonic development of the Western Mediterranean which reconciles both datasets.

2.1.7 Geodynamic causes of back-arc extension contemporaneous with compression

Several processes have been suggested as the cause of extension in the Western Mediterranean Basin contemporaneous with compression in surrounding belts.

Tapponnier (1977) considered that the opening of the Western Mediterranean basin was a direct consequence of collision where 'Apulia' acted as a rigid indentor, causing the Alps, and extension occurred as a response to 'lateral expulsion'.

More common is the notion that extension in the Western Mediterranean Basin and Tyrrhenian Sea occurred as a response to subduction (Rehault et al. 1984; Hill and Hayward 1988; Beccaluva et al. 1987; Kastens et al. 1988; Malinverno and Ryan 1988; Channell and Mareschal 1989; Lonergan and White 1997) and in particular to the steepening and migration of the subduction zone i.e. 'roll-back' (Beccaluva et al. 1987, Kastens et al. 1988; Malinverno and Ryan 1988; Channell and Mareschal 1989; Lonergan and White 1997). Certainly, this simple idea seems to most easily account for the regional history described above. Based on seismic tomography, Wortel and Spakman (1992) extend
the roll-back idea to one where the subduction system progressively rolled back, collided, and may have undergone 'slab detachment' resulting in the uplift of the compressional zones.

'Delamination' or 'convective removal' of overthickened continental lithosphere caused by collision is another mechanism proposed for generating extension (extensional collapse) in the Alboran Sea, Tyrrenian Sea and Western Mediterranean Basin contemporaneous with shortening and uplift in surrounding external belts (Carmignani et al. 1995; Platt and Vissers 1989; Channell and Mareschal 1989). Lonergan and White (1997) discard the convective removal hypothesis because the predicted 'radial' thrusting is not observed and because the palaeomagnetically determined rotations detected cannot be accommodated. Some 'asymmetric delamination' models (Channell and Mareschal 1989; Carmignani et al. 1994, 1995) seem to involve a compromise between overthickened lithosphere with a subducted oceanic slab underneath and sinking, rather like 'roll-back' or slab-detachment, of the slab and thickened lithosphere resulting in uplift and extension (e.g. Fig. 2.8c).

2.1.8 Summary
Within the overall framework of Africa-Eurasia collision, a NE-SW trending extensional system developed at the southern margin of the Eurasian plate in the late Oligocene (Fig. 2.5). Extension occurred contemporaneously with a calc-alkaline volcanic arc derived from subduction of oceanic crust, north to northwestwards beneath the Eurasian plate (Fig. 2.5). Rifting went to completion (i.e. β=∞) to the northwest of Sardinia in the early Miocene, with the opening of the Western Mediterranean back-arc basin and the separation and rotation of the Corsica-Sardinia microplate, which contained part of the volcanic arc (Fig. 2.5). The Sardinian Rift thus became an aborted intra-arc basin. The locus of extension moved to the east of Sardinia from the late Miocene with the opening of the Tyrrenian Sea active until the present day.

2.2 Geological history of Sardinia
This literature review briefly describes the Phanerozoic geology and geological history of Sardinia. Of particular importance is an appreciation of inherited structures which were re-used in the Oligo-Miocene episode, the late Eocene-early Oligocene 'pre-rift' topography, the nature and location of the types of 'pre-rift' rocks, needed for provenance analysis, and the post-Miocene tectonism which cuts the succession studied here. Figure 1.1 summarises the Sardinian geology.

2.2.1 Palaeozoic
Palaeozoic rocks from Cambrian to early Carboniferous age crop out in three regions, the Iglesiente-Sulcis area of southwest Sardinia, eastern Sardinia and the Nurra area of northwestern Sardinia (Cherchi 1985; Fig. 1.1). The rocks, strongly deformed by the Hercynian orogeny, were laid down on
a ?Precambrian-Ordovician passive margin, middle Cambrian-early Ordovician continental arc (southern side) and late Ordovician-late Devonian rifted margin (northern side) on opposing sides of the 'south Armorican Ocean' (Carmignani et al. 1992a). Continental collision between Armorica and Gondwana began in the late Devonian-early Carboniferous and remnants of oceanic crust are preserved in the Posada-Asinara suture zone (Carmignani et al. 1992a). This Hercynian orogeny resulted in intense deformation, with thrusts oriented NW-SE to NNW-SSE, SW vergent nappes in the northwestern internal zone and folding in the southwestern external zone (Fig. 2.9). Metamorphic grade decreased from granulite/eclogite and amphibolite facies in the northwest to greenschist/zeolite facies in the southwest (Cocozza and Jacobacci 1975; Sartori et al. 1987; Carmignani et al. 1992a; Fig. 2.9). Thus rocks originally deposited as conglomerates, sandstones, mudstones, dolomites and volcanics became metamorphosed to migmatites (rare), metaconglomerates, psammites, pelites, marbles and metavolcanics (e.g. Cocozza and Jacobacci 1975; Cherchi 1985; Sartori et al. 1987; Carmignani et al. 1992a). Clasts of these rocks are commonly found within the Oligo-Miocene succession.

In the middle and late Carboniferous, the orogenic wedge collapsed causing ductile extension at mid-lower crustal levels accompanied by the emplacement of peraluminous anatectic granites (~300 Ma, Carmignani et al. 1992a). The metamorphic rocks were unroofed by reversal of the earlier thrusts and in the Sarcidano area of central Sardinia a major antiform (Flumendosa antiform, NW-SE trend) was unroofed as a "metamorphic core complex" on low angle detachments to the NE and SW (Fig. 2.9). Post-orogenic calc-alkaline plutonism which formed the Corsica-Sardinia batholith ended by ~275 Ma (Carmignani et al. 1992). Thus much of the Palaeozoic basement is composed of pelites and psammites cross-cut by granitoid batholiths. Late Carboniferous-Permian 'molasse basins' were filled with clastic material and rhyolite-rhydacitic volcanics (Cocozza and Jacobacci 1975; Carmignani et al. 1992a).

The main structural trend of the Hercynian orogeny was NW-SE (Carmignani et al. 1987; Sartori 1987; Carmignani et al. 1992a; Fig. 2.9). However, Chabrier and Chorowicz (1982) described structures trending 25-80° as 'late Hercynian' and Oggiano et al. (1995) and Assorgia et al. (1995) also considered that large NE-SW trending faults (Fig. 1.1) are inherited 'late Hercynian' structures. In the Monte Albo area of northeastern Sardinia, Carmignani et al. (1992b) described an east-west Hercynian structural trend. Thus a pervasive NW-SE Hercynian compressional fabric seems to have been cut by localised and spaced NE-SW to E-W trending, 'late Hercynian' brittle and ductile structures.
2.2.2 Mesozoic

Throughout the Mesozoic, Sardinia was part of a relatively stable platform (Cocozza & Jacobacci 1975). The Triassic of Sardinia comprises continental and lagoonal quartzose conglomerates, marls and dolomites of the 'Germanic facies' (Cocozza & Jacobacci 1975). From the late Triassic - late Cretaceous, carbonate platform facies dominate in western Sardinia, whilst in central and eastern Sardinia, continental-deltaic sediments were overlain by marine deposits following a Bathonian transgression (Cocozza and Jacobacci 1975; Sartori et al. 1987). Dolomite and limestone which crop out discontinuously in the Nurra region of northwestern Sardinia, eastern central Sardinia and the Sarcidano region are the main rock types (Fig. 1.1). The succession can reach thicknesses of 900m and in central and eastern Sardinia the deposits rest discordantly on peneplaned Palaeozoic units (Cocozza & Jacobacci 1975). Clasts of these rocks are also found within the Oligo-Miocene succession. The offset and tilting of the carbonates allows estimation of Oligo-Miocene extensional fault throws.

2.2.3 Cenozoic

Small lenses of Eocene sediments are unconformable on the older rocks (Cocozza and Jacobacci 1975). Lower Eocene calcareous, clastic and ligniferous sediments of continental to marine facies (Ferrara et al. 1992) are unconformably overlain by the continental clastic, Upper Eocene-?Oligocene deposits of the Cixerri Formation (Cocozza and Jacobacci 1975; Cherchi and Montadert 1982ab; Assorgia et al. 1992b). Compressional deformation during the Eocene was discussed in section 2.1.1.1 and it is unclear what effect the deformation had on the paleotopography. In the southwest of Sardinia, Eocene sediments accumulated in the east-west trending Sulcis and Cixerri grabens (Fig. 1.1). The nature of the pre-rift topography over the rest of Sardinia is not known prior to the Oligo-Miocene extensional phase, but the environment of deposition was demonstrably subaerial.

The Oligo-Miocene rift which dominates the geology and topography of western Sardinia was filled with continental to marine sediments and calc-alkaline volcanic rocks from the late Oligocene until the late Miocene. The youngest rocks related to this phase consist of small Messinian-age outcrops of lagoonal facies including evaporites (Cherchi 1985). Cherchi and Tremolieres (1984); Cherchi and Montadert (1982a) provided a description of the structural evolution of Sardinia in the Oligo-Miocene based on 'microtectonic' measurements. They identify Messinian NW-SE compression, Burdigalian NE-SW compression, Aquitanian compression (80°) and Oligocene-Aquitainian extension. However, the amount of data used in this hypothesis is very small and consists of 3-20 readings from 17 'microtectonic stations'. It is unclear how the timing of fault movement was constrained. Based on these publications, it now widely stated in the literature that a Burdigalian compressional phase is present all over Sardinia (e.g. Cherchi and Montadert 1982ab; Rehault et al. 1984; Sartori et al. 1987).
for which no supportive evidence was found in this study. The approach is not thought to be useful when considering such a complex rift basin and the results should be viewed with caution (Appendix 1A also). Lecca et al. (1997) also state that there are no Burdigalian age compressive structures along the rift basin.

The Oligo-Miocene basin fill of the Sardinian rift exposed today was most probably only buried under its own weight. Subtle folding in some localities is thought to result from sediment and volcanic compaction. Much of the basin fill is transitional between sediments (i.e. sands, ashes) to rocks (sandstones, tuffs) and has high porosity.

In the Pliocene, the southern portion of the Oligo-Miocene rift was re-activated by the 20km wide, NW-SE trending Campidano graben (Cherchi and Montadert 1982ab; Fig. 1.1, 2.10). Up to 500m of continental clastic sediments (Samassi Formation) were deposited in this graben (Cherchi 1985). Large volumes of alkali basalts were erupted as extensive flows and cinder cones from 5.3 - 0.9 Ma over the Oligo-Miocene structure and parts of eastern Sardinia (Beccaluva et al. 1985; Fig. 1.1). The basalt flows form hard-caps to much of the peneplaned Miocene succession. At some time between the Messinian and Pliocene, significant uplift of Sardinia must have occurred such that the Miocene marine sequences are exposed today at elevations up to 700m. This may have been caused by thermal rift-flank uplift at the margins of the Tyrrenian Sea back-arc basin. In central southern Sardinia, localised uplift also occurred in the footwall of the Pliocene Campidano graben.

Plio-Pleistocene N-S normal faulting is common throughout Sardinia (Carmignani et al. 1994) Tyrrenian age (Pleistocene) marine deposits are exposed discontinuously around the coast of Sardinia in terraces at elevations up to 12m (Cocozza and Jacobacci 1975), though other areas of the coast are subsiding (e.g. Roman ruins beneath the sea ~30km SW of Cagliari).

2.3 Structural framework of the Oligo-Miocene Sardinian rift

This section outlines the large scale structure of the Sardinian Rift. An appreciation of the structural framework is necessary to place in context the studied areas (Chapters 4-8, Fig. 1.3) which present the evidence for the assertions made below.
Outside the main Sardinian Rift basin, the southernmost part of Sardinia is cross-cut by a number of sub-basins. The Sulcis area (Fig. 1.1) with east-west and NW-SE normal faults and the Cixerri E-W graben both contain Eocene and Oligo-Miocene age rocks (Cherchi and Montadert 1982ab; Assorgia et al. 1992bce; Barca and Costamagna 1997; Fig. 1.1). The Funtanazza sub-basin (Fig. 1.1, 1.3, Chapter 4) is an asymmetric graben trending east-west. It is bounded by high angle normal faults, the northerly one thought to have the greater throw. The Funtanazza and Cixerri graben are intersected by the Pliocene Campidano graben in the east.

In southern Sardinia, the NW-SE trending western bounding fault of the Sardinian rift and other NW-SE faults were re-activated by the narrower Campidano Graben in the Pliocene (Fig. 1.1, Fig. 2.10). The eastern margin of the rift was defined by major ~N-S faults and other ~E-W faults, north of Cagliari and by NW-SE trending normal faults in the Sarcidano sub-basin of central Sardinia (Chapter 5, Fig 1.1, 1.3). Two fault sets define the basin margin of the Sarcidano area (Fig. 1.1). NW-SE high-angle faults define a tilted block and half graben geometry. Slightly younger NE-SW to NNE-SSW high-angle faults define discrete, fault-bounded basement blocks. The dominant NW-SE trend of the basin margin and the location of basement blocks seems to be controlled by the Hercynian ‘Flumendosa antiform’ structure (section 2.2.1).

To the north, the NW-SE Sarcidano trend is replaced by N-S trending normal faults which define the eastern rift margin in central northern Sardinia and intersect with a series of NE-SW trending sinistral strike-slip faults and transtensional sub-basins (e.g. Ottana and Oschiri, Chapter 6, Fig 1.1, 1.3). Within the basin centre, the Logudoro study area of central Sardinia (Chapter 7) is divided into a series of volcanic fault blocks and sub-basins defined by E-W faults and N-S to NNW-SSE faults (Fig. 1.1). East of Sassari, the Ploaghe fault trends NW-SE, bounding a deep sedimentary sub-basin. The northernmost Anglona study area has a complex structure (Chapter 8). NE-SW faults in the eastern basement are cross-cut by NNW-SSE trending faults bounding a 46km wide, tilted fault block and half-graben sub-basin structure (Castelsardo and Portotorres sub-basins). NE-SW, E-W to NW-SE trending normal faults define a high volcanic platform (Tergu Platform) and synclinal sub-basin (Perfugas) to the south of the half-graben (Fig. 1.1). The poorly exposed western margin of the northern Sardinian rift consists of N-S trending faults bounding pre-rift rocks (Nurra area) or N-S faults with little associated topography (central western Sardinia and just offshore, Fig. 1.1 after Assorgia et al. 1995a). These western faults have variable, but generally smaller throws than the eastern margin bounding faults giving the northern Sardinian rift an asymmetric, half-graben structure.
Chapter 3
Chapter Three - Stratigraphy

The purpose of this chapter is to describe the dating techniques used and to define the stratigraphy of each of the study areas along the Sardinian Rift.

The sedimentary and volcanic basin fill sequence of the Sardinian rift is complex, with abrupt vertical and lateral facies changes. No simple stratigraphic scheme can be used to describe the entire basin fill adequately, though some correlations and general trends can be observed within a chronostratigraphic framework. Existing work does not provide or define any coherent rift stratigraphy and nomenclature is often confused. The approach of this study was to identify and define a lithostratigraphy for each studied sub-basin based on field outcrops. The lithostratigraphy was then placed in a chronostratigraphic framework constrained by existing biostratigraphic zonations, radiometric ages and by new temporal data.

This chapter firstly summarises the nature of and problems with published stratigraphy thus justifying the need for this new work. Secondly, the techniques and results of dating of stratigraphically important sedimentary or volcanic rocks are described. The variety of the basin fill rocks meant that no unified dating technique was applicable. The successful methods utilised were nannofossil zonations of marlstones, $^{87}$Sr/$^{86}$Sr isotope stratigraphy of shallow marine carbonates and calcarenites and single crystal $^{40}$Ar/$^{39}$Ar dating of volcanic plagioclases and biotites, correlated to the timescale of Berggren et al. (1995, Fig. 3.1). A formal stratigraphy for each study area is then proposed which incorporates the new and existing age constraints.

3.1 Existing Stratigraphy

All published and located stratigraphic criteria were considered and recalibrated to the timescale of Berggren et al. (1995, Fig. 3.1). The use of stage names whose range can differ widely dependent on the timescale (e.g. Aquitanian, 21.5-23.8 Ma; Berggren et al. 1995; 22.5-24 Ma, Vail and Hardenbohl 1979; 19.5-24.5 Ma, Edel 1980) was avoided.

3.1.1 Existing Biostratigraphy

In depth biostratigraphic studies, particularly ones utilising microfossils to 'zone' level, provide an invaluable basis for this research (e.g. Pomesano Cherchi 1971ab; Spano and Asunis 1984; Cherchi 1985; Mazzei and Oggiano 1990; Serrano et al. 1997). The original data (nannofossil, foraminifera lists or zonations) were used wherever possible to correlate to the timescale of Berggren et al. (1995). However, several problems occur with this biostratigraphic data:
a) Over the entire Western Mediterranean region, difficulties exist in the early Miocene such that calcareous nannofossils of the NN1 zone (Early Aquitanian), occur with the NS planktonic foraminifera \(G.\ aliaperturus-C.dissimilis\) sub-zone (early Burdigalian; Cherchi 1985). Cherchi (1985) and Cherchi and Montadert (1982a,b) utilise the ‘older’ the nannofossil zones in preference to foraminifera zonations as stratigraphic constraints.

b) In general, the zonation/correlation schemes used in the literature are those of Martini (1971, nannofossils) and Blow (1969, planktonic foraminifera) but several others are also common (e.g. Iaccarino 1985 in Porcu et al. 1997). The different zonation schemes utilise different ‘marker’ microfossils and correlations to a number of timescales leading to a complex and confusing picture. Some authors do not even specify which zonation scheme they used (e.g. Maxia and Pecorini 1969) and, in addition, this oldest literature uses different stage names for the Miocene (Aquitanian - Langhian - Helvetian = ?Aquitanian - Burdigalian - Langhian -Serravalian). For a non specialist, it therefore proved difficult to evaluate some sample zonations with respect to contemporary schemes. Where the zonation is tentative a question mark is placed in front of it.

c) Diachronous facies, faulting and abrupt lateral facies changes within the basin fill of the Sardinian Rift mean that to high resolution, simple temporal correlation of outcrops over kilometre scales is not a realistic approach. For example, marlstones basinward of the Isili fault block containing N4 zone microfossils (Fig. 5.1, Sarcidano sub-basin), equated to shallow marine carbonates in this area, are not necessarily the same age as topographically higher, shallow marine carbonates north of the fault block (e.g. as in Cherchi and Montadert 1982ab, see also Appendix 3D.2).

3.1.2 Existing \(^{40}\text{K}-^{40}\text{Ar}\) and \(^{40}\text{Ar}-^{39}\text{Ar}\) radiometric dates and volcanic stratigraphy

A sizeable number of whole rock \(^{40}\text{K}-^{40}\text{Ar}\) dates exist (e.g. Coulon et al. 1974; Savelli et al. 1979; Assorgia et al. 1984; summarised in Beccaluva et al. 1985) plus fewer \(^{40}\text{Ar}^{39}\text{Ar}\) and \(^{40}\text{K}-^{40}\text{Ar}\) determinations on mineral separates (Montigny et al. 1981; Odin et al. 1994; Assorgia et al. 1995a and in press; Balogh et al. 1997; Deino et al. 1997; Mameli and Oggiano 1997). Combined with lithostratigraphy, the dates have been used to define 5 volcanic series (Table 3.1). Lecca et al. (1997) provide an alternative lithostratigraphic nomenclature along the Sardinian Rift. Some contradiction exists in the published data. In part, this may arise because the timing and nature of volcanism changes from south to north over the island and one simple scheme may not be applicable for the entire area. However, considering the data from similar locations, some \(^{40}\text{K}-^{40}\text{Ar}\) dates cannot be recording the true timing of volcanism (e.g. SA1 Montresta 18.7±0.7 Ma versus SA2 Bosa 19.4± 0.7 Ma; SI2 Banari 19.6± 0.5 Ma versus SA2 Padria 18± 0.8 Ma, in Beccaluva et al. 1985, Fig. 7.1 for locations). Also, when compared to biostratigraphic data from overlying sediments, \(^{40}\text{K}-^{40}\text{Ar}\) ages can appear too
'young' e.g. SA2 Paulilatino (Fig. 6.1) 14.5±0.6 Ma (Serravalian, Beccaluva et al. 1985) versus N7 zone (early Burdigalian) for overlying marine sedimentation in Odin et al. (1994). Whole rock or bulk separate ages may be in error because of alteration (e.g. strong rock-water interactions) and xenocryst contamination (Montigny et al. 1981, 3.2.3) and must be treated with caution. Vigliotti and Langhenheim (1995) consider that 'most of the radiometric ages calculated by whole rock dating in Sardinia are too young' e.g. 16.8-17.5 Ma from the M. Traessu area (Fig. 7.1, Coulon et al. 1974) are 'more adequately' the 19.9 Ma of Montigny et al. (1981) on separates from this area.

### 3.1.3 Existing Stratigraphic Schemes
Published studies often provide a confusing picture of rift stratigraphy due to the lack of properly defined stratigraphic nomenclature, resultant misquoting and misinterpretation, and because of over simplification of facies variability. Stratigraphy and stratigraphic variations are complex even within sub-basins and one stratigraphic scheme is not easily applicable to the entire rift. Published stratigraphy generally consists of informal and often undefined local terms (e.g. Arenarie di Gesturi, Calcare di Sassari, Cherchi 1985) As a result, published work sometimes uses the same terms differently (below). Some examples are given below and solutions are described in section 3.3.

#### a) Southern Sardinia
Pecorini and Pomesano Cherchi (1969) and Cherchi (1974) provide some ‘formal’ stratigraphic definitions based on outcrops in the Sarcidano, Campidano and Cagliari areas of southern Sardinia (e.g. Ussana Formation, Marne di Ales, Formazione della Marmilla, Marne di Gesturi). These terms are used but not always quoted in later publications. Cherchi and Montadert (1982ab) not only change the meaning of some of the ‘formations’ (e.g. Marmilla Fm.; Ussana Fm. from continental/lagoonal conglomerates and muds in Pecorini and Pomesano Cherchi (1969) to then include shallow marine sandstones and carbonates in 1982ab), but they imply that the whole of the Oligo-Miocene stratigraphy of Sardinia can be described by their stratigraphic column (Fig. 1.2) which fieldwork shows is based on outcrops from only southern Sardinia. Particular problems occur with the definition of the ‘Marmilla Formation’ first defined by Cherchi (1974) as sands, marls, hyaloclastites and pillow lavas of Aquitanian age in the Marmilla region, north of the Campidano plain. In contrast, Cherchi and Montadert (1982ab) describe the Marmilla Formation as ‘marls rich in planktonic foraminifera’ of Aquitanian-late Burdigalian age (for all Sardinia) and Leone et al. (1984) use the term for Aquitanian age conglomerates and tuffaceous sandstones around the Genoni fault block (Fig. 5.1).
b) Northern Sardinia

Less work has been published on northern Sardinia and local informal names are used (e.g. Calcare di Sassari, Lacustre in Cherchi 1985; Tufo a Vaginella, Molassa a Vaginella in Spano and Asunis 1984). The terms are poorly defined with respect to the nature, age and aerial extent of the basin filling unit. Thomas and Gennesseaux (1986) provide a correlation between their seismic-stratigraphy, published and unpublished outcrop data from northern Sardinia to attempt a simple 'informal' stratigraphic scheme. They utilise the terms Ussana and Marmilla Formations by facies correlation to the southern Sardinian outcrops. This stratigraphic scheme is not used here because of the confusion that exists over this nomenclature and because in detail, the facies types, age and tectonic context from the two areas are different. Martini et al. (1992) present the clearest record of stratigraphic relationships in northern Sardinia through a facies-based sequence stratigraphic analysis. However, this study lacks a thorough systematic stratigraphic description.

3.1.4 Use of existing stratigraphy

A large and valuable stratigraphic database on the Oligo-Miocene rocks of Sardinia exists though at present stratigraphic terms are often confused. This study aims to resolve many of the problems detailed above by defining a formal stratigraphy for each sub-basin based on lithostratigraphic field observations, new age constraints and reliable ages of located samples, using the guidelines of Whittaker et al. (1991). In each study area, the existing stratigraphy and temporal constraints have been evaluated and incorporated wherever possible (section 3.3, Appendix 3D). Where unambiguous terms have been used previously, existing nomenclature has been utilised. Though it introduces a great deal of new terminology, such a detailed and high resolution approach is necessary to constrain the complex facies variability of the basin fill, basin events and thus rift evolution.

3.2 Dating Techniques

This section describes the methodology and discusses the results of dating performed in this study.

3.2.1 Biostratigraphic Analyses

Several types of biostratigraphic analyses were evaluated;

a) Macrofossil and macroforaminifera assemblages abundant in some sediments (e.g. corals such as Porites sp. and Favites Neglecta, abundant Lithothamnium sp. and Lithophyllum sp. coralline red algae and macroforaminifera such as Heterostegina sp., Amphistegina sp., Miogypsina sp. are diagnostic of the Miocene. However, no detailed analysis of such samples was attempted because it was unlikely to give results to a high enough resolution and would require specialist knowledge beyond the scope of this research work.
b) Planktonic foraminifera are rare, poorly preserved and thought to be unsuitable for dating (D. Kroon pers. comm. 1995) in stratigraphically important sandstones, carbonates and calcarenites. Thus zonations on such samples was not attempted.

c) Nine samples consisting of continental muds and organic rich horizons were sent to Geochem Ltd. for pollen analysis in the hope that they would provide key information on the start of continental sedimentation in the Sardinian rift. No results were forthcoming, most probably because the samples were oxidised and the pollen poorly preserved.

d) Nannofossil Zonations

Ten marlstone samples were analysed for nannofossils by E. Gervais of J & G Consultants, organised by Dr. Dick Kroon of Edinburgh University. The results are presented in Appendix 3A and summarised in Table 3.2. Six of the samples have a diagnostic nannofossil assemblage which enables accurate dating, and one sample was more tentatively assigned a nannofossil zonation. The results are in agreement with existing dates from central Sardinia which were tentatively assigned to the NN1-NN2 zones (see section 3.1.1; Cherchi 1985) and provide reasonable ages for marlstone sedimentation in Sarcidano where abrupt facies variations exist (Appendix 3D.2). The tentative NN1-NN2 zonation from Castelsardo is in agreement with the stratigraphy of Spano and Asunis (1984, N4 zone) and Francolini and Mazzei (1992, NN1-NN2)

3.2.2. $^{87}$Sr/$^{86}$Sr Isotope Stratigraphy

3.2.2.1. Rationale

Much of the basin fill of the Sardinian rift consists of shallow marine platform carbonates, calcarenites and calcirudites which contain a variably preserved macrofauna and rarely contain a diagnostic microfossil assemblage suitable for biostratigraphic dating. However, dating of these rocks is particularly important in reconstructing the timing of tectonic events and the sedimentary palaeogeographies which define the evolution of the rift. Where biostratigraphic correlation fails, an alternative approach is to use $^{87}$Sr/$^{86}$Sr isotope stratigraphy. Here, a $^{87}$Sr/$^{86}$Sr isotope study was used to provide additional temporal constraints for samples from along the Sardinian rift and from southern Corsica (Fig 3.2). The techniques of $^{87}$Sr/$^{86}$Sr isotope stratigraphy are still being developed and emphasis was placed on analysing the validity of the results. The study was performed by myself under the supervision of Dr R. Ellam at SURRC, East Kilbride.
3.2.2.2. Theory

The Sr isotopic ratio and Sr concentration in the oceans can be considered constant at any given time since the residence time of Sr is long (2.5-5 Ma) compared to the ocean mixing time (500-1000 years, Beets 1992). On average, the Sr isotope ratio has been increasing since the Jurassic (Beets, 1992) and several recent studies have constrained at high temporal resolution, Miocene marine $^{87}\text{Sr}/^{86}\text{Sr}$ isotope changes by combined biostratigraphic, magnetostratigraphic and isotopic studies on sediments and planktonic foraminifera from ODP/DSDP drilling sites (Hodell and Woodruff 1994; Oslick et al. 1994). Marine organisms which precipitate tests from seawater appear to do so without discriminating between the different Sr isotopes (Palmer and Elderfield 1985; DePaolo and Ingram 1985). Providing that the amount of $^{87}\text{Sr}$ produced by the decay of $^{87}\text{Rb}$ is negligible and that no diagenetic alteration has occurred since deposition, the organisms skeletal material thus records the ambient $^{87}\text{Sr}/^{86}\text{Sr}$ ratio during life. Thus, published studies which correlate $^{87}\text{Sr}/^{86}\text{Sr}$ ratios to biostratigraphic and magnetostratigraphic time constraints define curves which can be used as a 'reference' for $^{87}\text{Sr}/^{86}\text{Sr}$ seawater evolution in the Miocene (Figure 3.3).

The implication is therefore that the $^{87}\text{Sr}/^{86}\text{Sr}$ of seawater recorded in biogenic carbonates can then be used as a geochronometer (Beets 1992) to give a Sr isotope 'age' for that sample. The technique is particularly useful in the Miocene where the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of seawater increases rapidly and is particularly well constrained (Hodell 1994), facilitating results of high precision.

3.2.2.3. Sample selection and preparation

Skeletal material of two types was chosen for $^{87}\text{Sr}/^{86}\text{Sr}$ isotope analysis in this study. Oysters (Ostrea sp.) and red algae (Lithothamnium sp.) are the most widespread fossils in the Miocene marine rocks of Sardinia, thus sample collection from a wide range of sections and sites was possible. Samples were chosen to constrain rift events such as transgression or fault movement and to constrain the rift stratigraphy. The study was restricted to the two fossil types for consistency and because the skeletal material of these organisms most often appears to be diagenetically unaltered carrying a strong likelihood that the $^{87}\text{Sr}/^{86}\text{Sr}$ signature is undisturbed.

Oysters precipitate tests of stable, low Mg calcite or low Mg calcite and aragonite (Milliman 1974; Flugel 1982; Scoffin 1987) and are thus relatively resistant to diagenetic alteration. In thin section, oyster tests from Sardinia look fresh and X-ray diffraction shows them to be composed of low Mg calcite (Appendix 3B). In addition, X-ray fluorescence on two samples shows a Rb/Sr ratio of <0.006, such that the amount of $^{87}\text{Sr}$ derived from the decay of $^{87}\text{Rb}$ is negligible. Thus, the oyster tests should be accurate recorders of the original $^{87}\text{Sr}/^{86}\text{Sr}$ seawater composition in the Miocene.

Original red algae skeletal material is composed of relatively unstable, high Mg calcite (4-30%) (Milliman 1974; Flugel 1982; Scoffin 1987). X-ray diffraction shows that red algae samples from
Sardinia are presently composed of low Mg calcite (Appendix 3B). In thin section, the intricate test structure is often well preserved, though sometimes micritized on a small scale (Fig. 3.4). It is possible that this micritization happened just after the time of life and therefore accurately records seawater composition. In any case, the process of transformation from high Mg to low Mg calcite is thought to occur on 'scales small enough that the grains slowly alter to calcite without disturbing the original character' (Milliman 1974). In addition, there is evidence that marine cements and shell material can preserve their original isotopic compositions despite the presence of diagenetic fluids and alteration (Carpenter et al. 1991).

Although transformed from high to low Mg calcite, red algae appear to record the original $^{87}Sr/^{86}Sr$ seawater ratio, though where possible, oyster samples were used in preference to red algae. In addition, X-ray florescence shows that in red algae samples (n=2), Rb/Sr <0.014. Calculation shows that in this case, the maximum amount of $^{87}Sr$ derived from $^{87}Rb$ will be less than the analytical error.

Oysters were sampled such that if possible, one individual lamina was used for analysis. Each lamina represents one growth stage (i.e. one year ?) and the use of one lamina means that interlamina sediment or cement of a different $^{87}Sr/^{86}Sr$ ratio was not be included in the analysis. However, about half the oyster samples contained 2 or 3 laminae, but with no visible interlaminae sediment or cement. Red algae samples were derived by drilling (~0.5mm drill head) along ~10-20 rhodolith laminae to give a fine powder.

The powdered samples were cleaned in pure ethanol in a ultrasound bath for 40 minutes, dissolved in 0.25M HCl and the Sr extracted by conventional techniques as described in Appendix 3B. The samples were analysed in the VG Sector 54-30 multicollector mass-spectrometer at East Kilbride.

3.2.2.4. Calculation

Two $^{87}Sr/^{86}Sr$ seawater evolution curves for the Miocene were used as reference curves for this study (Hodell and Woodruff 1994; Oslick et al. 1994, Fig 3.3). The curves were chosen because they provided a large number of data points (n=169, n=88 respectively) constrained by magneto- and biostratigraphy using up-to-date techniques.

The curves were first recalibrated to the recently published timescale of Berggren et al. (1995) by utilising a linear correlation between the timing of magnetic chron as defined in Cande and Kent (timescale x, 1992) and Berggren et al. (timescale y, 1995) such that $y=0.9863x+0.2415$, $R^2=0.994$ ($R^2$=regression coefficient). The data of Hodell and Woodruff (1994) and Oslick et al. (1994) was also corrected for the interlaboratory bias using the comparative measurements of the NBS 987 Sr standard for which the SURRC VG Sector 54 gave 0.710237 ± 0.000020 (2 standard deviations, n=15).
A line-fitting software package was used to give a best-fit sixth order polynomial equation for the data of Hodell and Woodruff (1994) and Oslick et al. (1994) such that the $^{87}\text{Sr}^{86}\text{Sr}$ seawater evolution curve through time can be described:

Hodell and Woodruff (1994)

$$^{87}\text{Sr}^{86}\text{Sr} = -7.99513418 \times 10^{-1} x^6 + 2.38527872 \times 10^{-8} x^5 - 2.50761539 \times 10^{-6} x^4 + 1.59696423 \times 10^{-3} x^3 - 1.34434978 \times 10^{-2} x + 0.752935164$$

where $x$ = time in Ma $R^2 = 0.88$

Oslick et al. (1994)

$$^{87}\text{Sr}^{86}\text{Sr} = -2.52081009 \times 10^{-6} x^6 + 1.88164021 \times 10^{-8} x^5 - 4.16892737 \times 10^{-7} x^4 - 6.11184222 \times 10^{-7} x^3 + 1.39879780 \times 10^{-4} x^2 - 1.80766023 \times 10^{-3} x + 7.16174535 \times 10^{-1}$$

where $x$ = time in Ma $R^2 = 0.80$

Farrell et al. (1995) show that for the past 7 Ma, 97% of $^{87}\text{Sr}^{86}\text{Sr}$ data fitted to a seawater evolution curve does so within a 'confidence interval' of 0.000019, that is, the error involved in fitting a seawater evolution curve is less than the analytical sample error ($\pm 0.000020$). For the early Miocene, the curve fitting error is likely to be less since the curve is steeper and well defined.

The $^{87}\text{Sr}^{86}\text{Sr}$ isotope 'age' of samples according to the two curves was then found by iteration on an 'Excel' spreadsheet (Table 3.3). The $^{87}\text{Sr}^{86}\text{Sr}$ isotope 'age' is not like a traditional radiometric date but relies upon correlation of a $^{87}\text{Sr}^{86}\text{Sr}$ sample ratio to a measured reference curve. Although the internal precision calculated on the 150 measurements of the isotope ratio in the mass spectrometer is smaller (e.g. $\pm 0.000017$ 2$\sigma$ error), the error used was the external reproducibility or precision measured from repeated standards of NBS 987 ($^{87}\text{Sr}^{86}\text{Sr}$ $\pm 0.000020$, 2$\sigma$ error). This value was used to calculate the appropriate 2$\sigma$ error in the 'age' of each sample (Table 3.3).

3.2.2.5. Results and Discussion

The detailed results are tabulated in Appendix 3B and are summarised in Table 3.3. The results show that the $^{87}\text{Sr}^{86}\text{Sr}$ isotope 'ages' calculated from both datasets (Hodell and Woodruff 1994 and Oslick et al. 1994) are generally identical within error though they may be slightly different for the oldest or youngest samples (Fig 3.5). For the purposes of clarity, only the dataset correlated to Oslick et al. (1994), believed to be the most complete and best constrained dataset (R. Ellam pers. comm. 1997) has been plotted on Figs 3.10-3.18 which illustrate the results compared to other stratigraphic criteria.

The $^{87}\text{Sr}^{86}\text{Sr}$ isotope dating technique on 27 samples produced 22 high resolution results which were not obviously diagenetically altered. The technique was successful in that samples from stratigraphic
sections gave ‘dates’ in the correct order, over a reasonable timescale (e.g. Is Paras Mbr. base-top, Fig. 5.1 for location, 18.4-16.9± 0.3 Ma; M. Santo, Fig. 7.1 for location, 16.2-14.8± 0.7 Ma; Scala Giocca, Fig. 7.1 for location, 15-12.7± 1 Ma) and on a broad scale agree with the majority of biostratigraphic, radiometric and lithostratigraphic criteria (section 3.3, Appendix 3D). However, some $^{87}\text{Sr}/^{86}\text{Sr}$ isotope dates need further discussion.

a) Five samples (23-27, Table 3.3) gave much younger ages (<8.6 Ma, $^{87}\text{Sr}/^{86}\text{Sr}>0.708894$) than are possible from stratigraphic constraints. These samples are considered to be diagenetically altered. The alteration process would occur if rainwater dissolved basement rocks (e.g. Hercynian metamorphics $^{87}\text{Sr}/^{86}\text{Sr}$ typically 0.7170, Beccaluva et al. 1985), ran off onto Miocene carbonates and reacted with the oyster or red algae tests to give a higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratio or ‘younger’ values than the original carbonate.

b) The lowermost carbonate southwest of Florinas (sample 13, Fig. 7.1, 19.5±0.3 Ma), possibly the transgressive carbonates at S.M. Iscalas, north of Coissone (sample 8, Fig. 7.1, 18.2±0.3 Ma) and Castelsardo (sample 20, Fig. 8.1, 18.88±0.3 Ma; sample 21, 19.4±0.3 Ma:) are perhaps ‘too old’ when compared to radiometric dates on underlying volcanics and biostratigraphic data on laterally equivalent sediments or overlying sediments (e.g. N7-N8 zone, ~16-17 Ma, east of Coissone; see Appendix 3D). $^{87}\text{Sr}/^{86}\text{Sr}$ isotope stratigraphy would give erroneously older ages if the sample did not record the true seawater signal because it contained interlaminae volcanic-derived sediment. Cenozoic volcanic rocks from Sardinia have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.704-0.706 (Dupuy et al. 1974), 0.706-0.707 (Mona et al. 1994) and some oysters visibly incorporate sediment between test lamina. Any future analysis should attempt to utilise only single laminae.

Work in progress (R. Ellam pers. comm. 1997) suggests that at high resolution, the errors involved in $^{87}\text{Sr}/^{86}\text{Sr}$ isotope dating due to diagenetic alteration and sediment incorporation could be approximately three times more than the analytical error. In this case, the $^{87}\text{Sr}/^{86}\text{Sr}$ isotope results from Sardinia would all agree well.

$^{87}\text{Sr}/^{86}\text{Sr}$ isotope stratigraphy is a useful technique for dating marine carbonates and calcarenites containing no diagnostic fauna within error limits of a few million years. The implications of the results are discussed in detail in section 3.3 and Appendix 3D, sub-basin stratigraphy. Because of limitations due to sample alteration and possible sediment incorporation, most reliability can be placed on the high resolution results from samples which constrain stratigraphic sections and/or where some external stratigraphic constraints may be applied. In this respect, the problems with the technique make it difficult to evaluate to very high resolution the reliability of any one result. This could be solved by further testing of the technique, for example by using some $^{87}\text{Sr}/^{86}\text{Sr}$ isotope samples which could be
accurately biostratigraphically dated and by looking in more detail at the nature and amount of sample alteration.

### 3.2.3 $^{40}$Ar/$^{39}$Ar radiometric dating

#### 3.2.3.1. Rationale

The $^{40}$Ar/$^{39}$Ar technique provides a means of dating the crystallisation age of igneous minerals based on the decay of $^{40}$K to $^{40}$Ar. In this study, the technique was used to date the timing of extrusive volcanism in Sardinia. The original aims were to constrain events and stratigraphy in the basin fill and changes in parental magma geochemistry through time. However, due to the results of the geochemical sampling and analysis (Chapter 9), the second aim was replaced by a collaborative project with the University of Cagliari investigating the timing of two phases of ignimbritic volcanism, where samples also constrained important basinal events. The study was funded by NIGL grant no. IP/453/0995 awarded to Dr J. Dixon and A. Sowerbutts and was carried out by myself under the supervision of Dr M. Pringle at SURRC, East Kilbride.

#### 3.2.3.2. Theory (modified after Faure 1986)

$^{40}$K/ $^{40}$Ar geochronology is based on the decay of radioactive $^{40}$K to $^{40}$Ar where the age of the rock or mineral can be calculated by measuring the amounts of two the isotopes and using equation 1

$$t = \frac{1}{\lambda} \ln \left[ c \left( \frac{40 \text{Ar}}{40 \text{K}} \right) + 1 \right]$$

where $\lambda$ is the total decay constant of $^{40}$K, $c$ is a constant (the total decay constant of $^{40}$K divided by the decay constant of $^{40}$K to $^{40}$Ar) and by assuming that no $^{40}$Ar or $^{40}$K has been lost or added since the system closed through its closure temperature, that no $^{40}$Ar was incorporated into the mineral at the time of its formation (a good assumption since Ar is inert) and that corrections are made for atmospheric $^{40}$Ar.

$^{40}$Ar/$^{39}$Ar geochronology relies on the same decay process and the same assumptions, but has major advantages over $^{40}$K/ $^{40}$Ar dating. For example, with the $^{40}$Ar/$^{39}$Ar technique one measures the ratio of $^{40}$Ar/$^{39}$Ar in a mass spectrometer which is more accurate than measuring the amounts of $^{40}$K and $^{39}$Ar separately. Also, $^{40}$Ar/$^{39}$Ar dating can be performed on single crystals rather than on a bulk samples. This has particular advantages in volcanic rocks erupted through and onto older basement where there may be incorporation of xenocrysts, since a number of individual crystals are dated and xenocrysts can be identified. Such rocks are common in Sardinia. In addition, single crystal step-heating can be used to check for and allow for the effects of alteration. In Sardinia, a reasonable number of published $^{40}$K/$^{39}$Ar dates exist, however some doubt exists as to their validity due to the effects of alteration and
the possibility of xenocryst incorporation. The $^{40}$Ar/$^{39}$Ar dating technique was employed in this study to overcome these problems.

The method of $^{40}$Ar/$^{39}$Ar geochronology is to bombard the sample with fast neutrons in a nuclear reactor such that $^{39}$K (n,p) -> $^{39}$Ar. The ratio of $^{39}$K to $^{40}$K is a constant such that equation (1) can be modified

$$t = \frac{1}{\lambda} \ln \left(J(\frac{^{40}Ar_{rad}}{^{39}Ar}) + 1\right)$$

(2)

$^{40}$Ar$_{rad}$ = radiogenic $^{40}$Ar, $^{39}$Ar$_K$ = $^{39}$Ar derived from $^{39}$K

where J is a parameter that relates to the neutron flux density and can be calculated along the length of the irradiated vial by including a number of standard crystals of known age. $^{40}$Ar$_{rad}$/$^{39}$Ar$_K$ is calculated from the total $^{40}$Ar/$^{39}$Ar measured in the mass spectrometer by making corrections for air, K and Ca.

3.2.3.3. Methodology

Samples were carefully chosen for their stratigraphic significance and their location is illustrated on Fig 3.6. The majority of the samples that were finally analysed were from ignimbrites and crystal tuffs, though two samples were from reworked tuffs (9638 and B36) As described in detail in Appendix 3C, plagioclase and biotite crystals were separated, cleaned, irradiated and single crystals were analysed by total laser fusion, laser step-heating or laser degassing plus laser fusion in a static vacuum collector mass-spectrometer. $^{40}$Ar/$^{39}$Ar ages and errors were calculated using standard techniques and corrections after Dalrymple et al. (1981).

3.2.3.4. Results

The results of the $^{40}$Ar/$^{39}$Ar dating are tabulated in Appendix 3C, summarised in Table 3.4 and Figs. 3.7 and 3.8.

A closely grouped set of single crystal analyses from each sample would be indicative of the crystallisation age of that sample. However, the results were more diverse than this simple scenario. Single crystal total fusion on plagioclases gave closely grouped age results with the exception of sample 3 (B36) which contained one crystal thought to be altered (18.05±0.66 Ma, Fig. 3.7). Single crystal biotites analysed by step-heating and by total fusion after degassing generally gave a wider scatter of age results (Fig. 3.7, samples 4, 5, 7). However, sample 6 gave a very good result from total fusion of single biotite crystals (Fig. 3.7).
Experiments 9710090 and 9710092 produced step-heating profiles with a consistent plateau age (Fig. 3.8). Experiments 9710091, 9710086, 9710088 gave consistent plateau ages for the bulk of the sample but gave younger ages for the first (and second 9710091, 9710089) heating steps (Fig. 3.8) interpreted as a consequence of minor amounts of crystal alteration. The consistent plateau ages indicate the igneous crystallisation age, which varied from crystal to crystal. Experiment 9710089 has a profile which does not form a plateau but increases with the amount of $^{39}\text{Ar}$ released. The first two steps may result from alteration whilst an increasing ‘pseudoplateau’ at ~23 Ma may be the consequence of partial resetting of an older biotite crystal.

Step-heating on single crystal biotites shows that the majority of the crystals have undergone minor amounts of alteration, the effects of which can be removed by a ‘degassing step’ (Appendix 3C). Simple total fusion experiments on such slightly altered crystals would give ages a little too young. Odin et al. (1994) also describe low temperature alteration of biotites from their $^{40}\text{Ar}/^{39}\text{Ar}$ step-heating experiments on bulk biotite separates. Regardless of the alteration, the bulk of the biotite crystals gave good plateau ages indicative of an igneous origin, but at varying times.

**Analysis of results**

Some explanation must be given to explain why consistent, single crystal plagioclase ages and older, scattered, single crystal biotite ages are gained from the same ignimbrite sample (samples 4 and 7, Fig. 3.8, Table 3.4). Three working hypothesis were considered:

**Hypothesis 1:** The favoured hypothesis is that fresh, euhedral, clear plagioclases record an explosive igneous eruption phase of the ignimbrites at ~20.6-20.9 Ma, whereas the variably older biotites are xenocrysts incorporated into the erosive pyroclastic flows.

The erosive nature of ignimbrites and the incorporation of xenocrysts seems to be a fairly common phenomenon (e.g. LoBello et al. 1987; Sparks et al., in press). In samples 4 and 7, the presence of entrained metamorphic basement clasts (Fig. 5.68) is consistent with the erosive and incorporating nature of these flows. Some basement derived, deformed biotite flakes are rarely present within these samples but can be distinguished from the euhedral ‘igneous’ biotites that were used for single crystal dating. Step-heating experiments showed that the majority of biotite crystals analysed were indeed Oligo-Miocene igneous phases, but with a variety of eruptive ages. Such biotites could be derived from andesitic volcanism which underlies the ignimbrites and commenced at ~33-30 Ma in southern Sardinia (Savelli et al. 1979; Assorgia et al. 1984; Beccaluva et al. 1985). However, in thin section, there is no evidence that the igneous biotites came from a different source. For example, they are not contained within volcanic ‘clasts’ and look quite content (Figs. 3.9, 4.4).
The hypothesis that the plagioclases record the age of igneous eruption is supported by the closely grouped results of the other plagioclase samples analysed, in particular that of sample 1 (B92) which occurs within the same volcanic succession as sample 4 (9690, Fig. 3.7) at the same time (within error). In addition, Odin et al. (1994) analysed bulk plagioclase separates using the step-heating technique on an ignimbrite sample from Chiaramonti in northern Sardinia (Fig. 6.1) and gained ‘an excellent plateau age’ which ‘actually represents the crystallisation age’ whereas biotites from the same sample showed small amounts of alteration.

This hypothesis may be tested by step-heating of individual plagioclase crystals when a new and more sensitive machine is installed at SURRC in late 1997.

**Hypothesis 2:** All crystals are variably altered to younger ages, with the oldest biotite crystals indicating most closely the age of volcanism.

This hypothesis is apparently supported for sample 4 (9693, oldest biotite 23.32±0.26 Ma) by published data from the Funtanazza coastal section (Fig. 4.1). Assorgia et al. (1995a) date the ignimbrite at 23.8± 2.4 Ma using $^{40}$K – $^{40}$Ar on plagioclase separates, Barbieri et al. (1997) date underlying fossils at 23 ±0.3 Ma using the $^{87}$Sr/$^{86}$Sr technique and the fossils are believed to be N4 zone (Cherchi 1982 in Assorgia et al. 1997c). However, since other N4 zonations by A. Cherchi are now believed to be ~N6 zone (Appendix 3D.2), an old, simplified curve was used in the $^{87}$Sr/$^{86}$Sr analysis (Hodell et al. 1991) with no methodology specified, and bulk plagioclase data has a large error, the published data is not conclusive. If this hypothesis were correct it seems an unlikely coincidence that all the plagioclases within the lowermost ignimbrite samples (1,4,7,8) all consistently alter to restricted age groups from 20.6-20.9 Ma.

**Hypothesis 3:** Plagioclases are variably altered to younger ages of 20.6-20.9 Ma, the oldest biotites are incorporated xenocrysts and biotites of ~21-22 Ma (9693) or 23-24 Ma (9690) populations record the timing of igneous eruption.

From the samples analysed in this study it seems most likely that in the stratigraphically lowest ignimbrites, plagioclases record the crystallisation age whilst variably older biotites are xenocrysts incorporated into the pyroclastic flow (hypothesis 1). This scenario is assumed to be the case throughout the thesis (Table 3.4). Samples which lie stratigraphically above the lowest ignimbrites (5:9691, 6:9692) do not contain entrained basement clasts and give more consistent biotite ages (5:9691, Fig. 3.7) suggesting that that xenocryst contamination is not such a problem higher up in the ignimbrite succession.
The results show that single crystal dating which identifies xenocrystic contamination and step-heating which identifies sample alteration are essential for constraining the timing of extrusive volcanism along the Sardinian Rift. Existing whole rock $^{40}$K-$^{40}$Ar dates and $^{40}$K-$^{40}$Ar or $^{40}$Ar/$^{39}$Ar dates on bulk separates where xenocryst contamination and alteration are possible must be treated with caution. The implications of the $^{40}$Ar/$^{39}$Ar dating on the basin stratigraphy and evolution are discussed in sections 3.3, 3.4 and in the following chapters.

3.3 Formal stratigraphy of the study areas

This section summarises the stratigraphy of each studied area (Fig. 1.3) by integrating and evaluating all available lithostratigraphic, biostratigraphic, $^{87}$Sr/$^{86}$Sr isotope and radiometric data and by defining new formal terminology following the guidelines of Whittaker et al. (1991). Appendix 3D gives a detailed discussion of the data available from each area which is needed because of the problems outlined in sections 3.1 and 3.2. The stratigraphy and temporal constraints are illustrated on chronostratigraphic diagrams (Fig 3.10 key, Figs 3.11-3.18), in tables (3.5-3.8) and on geological maps/enclosures (Encl. 1-5, Figs. 4.1, 5.1, 6.1, 7.1, 8.1). Ar-Ar and $^{87}$Sr/$^{86}$Sr isotope dates are given with 2σ errors (95% confidence level).

3.3.1 Funtanazza sub-basin (Fig. 3.11, Table 3.5, Chapter 4, Fig. 4.1)

Lithostratigraphic relationships in this area have been described by Assorgia et al. (1984, 1986ab, 1992) though no formal stratigraphy has previously been documented. In the east of the sub-basin, the succession is totally dominated by the andesites, andesitic breccias, basalts and rhyolitic ignimbrites of the mid Oligocene-late Burdigalian Arcentu Group. The Arcentu Group was split into four formations (A-D) by Assorgia et al. (1984, 1986ab). In the west, the basin fill of the coastal section commenced with the late Chattian-late Aquitanian (this study), non-marine to marine Campu Sali Formation and continued with a pyroclastic flow of the volcanic Arcentu Formation (late Aquitanian, section 3.2.3). Sedimentation on the coastal section continued with shallow marine calcarenites of the latest Aquitanian-mid Burdigalian (this study) Sartori Formation and was finally cross-cut by mid-late Burdigalian basaltic dykes and intrusions (Assorgia et al. 1984, 1986ab) of the volcanic Arcentu Group. Further inland, a second rhyolitic pyroclastic deposit and ?early Burdigalian (this study) lacustrine limestone unit (Sa Tellura Formation) are lateral equivalents of the middle part of the Sartori Formation. Figure 3.11 provides two alternative chronostratigraphies for the area, one based on published data and one based on new Ar-Ar and $^{87}$Sr/$^{86}$Sr isotope results (sections 3.2.2, 3.2.3 and Appendix 3D). Both versions of the stratigraphy are considered in chapter 4 since neither result is conclusive.
3.3.2 Sarcidano sub-basin (Figs 3.12, 3.13, Table 3.6, Chapter 5, Fig. 5.1)

Sedimentation in Sarcidano commenced with the coarse clastic deposits of the Nureci Formation. The basal ?late Oligocene- late Aquitanian Villanovatulo Member continental conglomerates and breccias passed diachronously upwards into early Aquitanian - late Burdigalian shallow marine sediments of the Duidduru Member. Pyroclastic volcanic rocks of the late Aquitanian-early Burdigalian Araxigi Formation occurred as a partial lateral equivalent of the Nureci Formation in the north-west of the sub-basin. The lower/mid-upper Burdigalian Isili Formation comprises shallow marine siliciclastic-carbonate alternations, thick shallow marine carbonates (Is Paras Member) and calcarenites/calcirudites (Serra Longa Member) partly equivalent to and overlying the Duidduru Member. The above units comprise the ?late Oligocene- late Burdigalian Sarcidano Group. Basin filling was completed with deposition of the partly lateral equivalent and overlying marlstones of the late Burdigalian -Langhian Giara Group.

3.3.3 Strike-slip systems of eastern Sardinia (Fig. 3.14, 3.15, Table 3.7, Chapter 6, Fig. 6.1)

Though this 'study area' covers a large part of eastern Sardinia (Fig. 1.3) and much the area was studied only at reconnaissance level, it is possible to describe a lithostratigraphy common to these zones which is most complete in the 'pull-apart basins' of Oschiri and Ottana (Fig. 1.3). The succession comprises ?late Oligocene-earliest Burdigalian basal continental conglomerates or breccias overlain and intercalated with ignimbrites, tuffs and lacustrine limestones-cherts; this unit is here termed the Oschiri Formation. The mid-late Burdigalian Chilvani Formation consists of alluvial to marginal marine clastic sediments unconformable on the Oschiri Formation.

3.3.4 Logudoro study area (Fig 3.14, 3.15, Table 3.7, Chapter 7, Fig. 7.1)

The lowest exposed basin fill in the Logudoro area comprises a thick, mid Oligocene-mid/late Burdigalian succession of basalts, andesites, ignimbrites and tuffs of the Logudoro Group which can be subdivided into five lithostratigraphic series (Table 3.1). The volcanic rocks are overlain by and intercalated with lacustrine sediments and tuffs correlated to the Oschiri Formation as defined above (3.3.3). The continental to marginal marine clastic sediments of the mid-late Burdigalian Chilvani Formation pass upwards and westwards into a thick marine succession termed the Florinas Group which is unconformable on the Logudoro Group and Oschiri Formation. The mid/late Burdigalian-Serravalian marine sediments of the Florinas Group comprise a complex arrangement of sandstones, carbonates and marlstones which were not studied at high enough resolution to describe a formal 'formation and member' stratigraphy. This group could be formally subdivided into formations if more detailed studies were undertaken.
3.3.5 Anglona study area (Fig 3.17, 3.18, Table 3.8, Chapter 8, Fig. 8.1)
The complex stratigraphy of the Anglona sub-basin is here defined by a lower transgressive-regressive, volcano-sedimentary, ?late Oligocene-early Burdigalian Elefante Formation, laterally equivalent volcanic Tergu Formation, an upper, early-mid Burdigalian, lacustrine Perfugas Formation and an uppermost ?early Burdigalian-early Serravalian, marine carbonate-marlstone Laerru Formation.

Sedimentation of the Elefante Formation in the Anglona sub-basin commenced with continental clastic deposition adjacent to the sub-basin's eastern margin (?late Oligocene, Casteldoria Member) and fluvio-lacustrine-lagoonal deposition of limestones, marlstones and cherts in the sub-basin centre (?late Oligocene-earliest Aquitanian Valledoria Member). The succession passes gradationally upwards into a thick, ?late Oligocene-early Burdigalian succession of pumice and crystal lapilli tuffs which are marine at the top (Vaginella Member) and lateral equivalents of the early/mid Aquitanian-early Burdigalian Castelsardo Member offshore to shallow marine clastic sediments. The laterally equivalent, ?mid Oligocene-mid Burdigalian Tergu Formation comprises andesites, volcanic breccias and ignimbrites in the west of the sub-basin.

Aerally extensive ignimbrite and tuff deposition towards the top of the Tergu Formation is overlain by and intercalated with the early-mid Burdigalian lacustrine limestones of the Perfugas Formation in the south of the study area. The non-marine basin fill passes diachronously upwards into early Burdigalian-early Serravalian marine carbonates, calcarenites and marlstones of the Campulandru, Sedini, Martis and Sennori Members (Laerru Formation).

3.3.6. Capo Testa and Corsica (Fig 3.18)
The Capo Testa succession of northernmost Sardinia (Fig. 1.3) has been dated as late Burdigalian to Langhian (Cherchi 1974; Assorgia et al. 1997a). In the Miocene basins of Corsica (Fig. 1.3), sedimentation is believed to be of Burdigalian-Langhian age (St Florent limestones; Bonifacio calcarenites) and Langhian (N8) - Messinian age (N16-N17, Plaine Orientale, Orzag-Sperber and Pilot 1976). A ⁸⁷Sr/⁸⁶Sr isotope sample from the Bonifacio sub-basin gave an age of 17.4± 0.3 Ma (sample 22, mid-late Burdigalian) in agreement with the published data.
3.4 Stratigraphy of the Sardinian Rift

3.4.1 Problems with Sardinian Stratigraphy

The ages gained from the various dating techniques combined with lithostratigraphy mean that a high resolution stratigraphic analysis is possible. However, consideration of sub-basin stratigraphies (Appendix 3D) highlights some general problems encountered with 'matching' the different dating techniques:

- Biostratigraphic, new radiometric and isotopic data may be inconsistent with respect to local lithostratigraphy (e.g. Sarcidano, Appendix 3D.2). This may be because older biostratigraphy needs to be re-evaluated and correlated using the same zonation scheme and timescale, and because simple temporal facies correlations are not viable within such a complex structural regime.

- $^{40}$K - $^{40}$Ar and $^{87}$Sr/$^{86}$Sr isotope dates from different stratigraphic units may overlap in time (Logudoro Appendix 3D.4; Anglona Appendix 3D.5) but since whole rock $^{40}$K - $^{40}$Ar samples may be altered (section 3.1.2) and $^{87}$Sr/$^{86}$Sr isotope samples may be altered or incorporate material with a different $^{87}$Sr/$^{86}$Sr isotope signature (section 3.2.2) it is difficult to make a reasoned assessment of whether the volcanic samples or carbonate samples are altered or have incorporated material.

3.4.2 Trends and Correlations

Correlation of sub-basin stratigraphy's along the whole of the Sardinian Rift illustrates a general trend of initial continental clastic and lacustrine sedimentation in the ?Rupelian-Aquitian (Villanovatulo, Casteldoria, Valledoria Mbrs., Campu Sali Fm.) contemporaneous with andesitic and ignimbritic volcanism (Logudoro, Arcentu Groups; Tergu, Araxigi Fm.). The deposition of marine siliciclastic and epiclastic sediments adjacent to the rift margin and volcanic centres occurred from the latest Oligocene-Aquitian (Duidduru, Vaginella Mbrs.) coeval with volcanism (Logudoro, Arcentu Groups, Tergu, Araxigi Fm.) intercalated continental clastics and lacustrine sediments (Oschiri Fm.). From the mid Burdigalian, shallow marine siliciclastic, carbonate and marlstone sediments dominated the fill of the Sardinian rift and Corsican basins (Florinas, Giara Groups, Laerru, Isili Fms.). The trends and correlations along the Sardinian Rift are discussed further in chapters 10 and 11 (see Fig. 10.8, 10.9).

3.4.3 Timing of Volcanic Series

Previous workers have identified a volcanic lithostratigraphy of interbedded 'andesitic' and 'ignimbritic' series constrained to some extent by radiometric dates (Coulon 1974; Savelli et al. 1979; Assorgia et al. 1995a and in press, Table 3.1). However, some confusion exists because of problems with the $^{40}$K - $^{40}$Ar dating (section 3.1.2). Of particular interest is whether the series represent discrete temporal events with possible geodynamic consequences. For example, are there two distinct phases of
ignimbritic volcanism all over the island? The $^{40}$Ar/$^{39}$Ar dating performed in this study along with other recent radiometric dates on mineral separates helps to answer this question (Table 3.9, Fig. 10.8 also). The results show that the lithostratigraphically ‘lower ignimbrites’ span 23.8-19.1 Ma whilst the ‘upper ignimbrites’ span 19.08-16.6 Ma and thus the two series are continuous within analytical error. However, there do appear to have been peaks of large volume ignimbrite eruption at ~20.9-20.6 Ma and ~18 Ma with deposits found in several localities across the island.

3.5 Summary

- The complex basin fill succession of the Sardinian Rift requires high resolution stratigraphic schemes for each studied sub-basin. Such stratigraphy is possible by incorporating and evaluating lithostratigraphic information and temporal constraints.

- $^{87}$Sr/$^{86}$Sr isotope stratigraphy is a useful tool for dating marine sediments containing no diagnostic fauna, but for total reliability to be placed on the results the technique may need to be further tested.

- Single crystal $^{40}$Ar/$^{39}$Ar dating which identifies xenocrystic contamination and single crystal step-heating experiments which identify sample alteration are essential for constraining the timing of extrusive volcanism along the Sardinian Rift. Existing whole rock $^{40}$K-$^{40}$Ar dates and $^{40}$K-$^{40}$Ar or $^{40}$Ar/$^{39}$Ar dates on bulk separates where xenocryst contamination and alteration are possible must be treated with caution. From the samples analysed in this study it seems most likely that in the stratigraphically lowest ignimbrites, plagioclases record the crystallisation age whilst variably older biotites are xenocrysts incorporated into the pyroclastic flow. Major phases of ignimbritic volcanism occurred in the Sardinian Rift at ~20.9-20.6 Ma and at ~18 Ma.
Chapter 4
Chapter Four - The Funtanazza sub-basin

The interaction of contemporaneously deposited sedimentary, volcaniclastic and extrusive volcanic rocks can be studied within the mid Oligocene-lower Miocene fill of the Funtanazza sub-basin. The sub-basin is a 2-9 km wide, east-west trending graben which formed to the west of the main Sardinian Rift in southwestern Sardinia. (Fig. 1.3, 4.1). This chapter builds on existing work which describes and dates the rock types present (Assorgia et al. 1984 1986ab, 1992a, 1997c; Assorgia and Gimeno 1994; Barbieri et al. 1997). Also presented is new data that enables the response of sedimentation to active volcanism and relative sea level change to be evaluated and a tectono-stratigraphic model for extension within this part of the southern Sardinian Rift basin.

The wider, eastern part of the Funtanazza sub-basin was filled with mid Oligocene-Burdigalian basaltic and andesitic rocks of the Arcentu Group. In the western part of the sub-basin, intercalated lacustrine, shallow marine sediments and rhyolitic pyroclastic flows of ?late Oligocene-mid Burdigalian age crop out (Campu Sali Formation, Sa Tellura Formation, Sartori Formation, section 3.3.1, Appendix 3D.1; Fig. 3.11). The stunning coastal section provides the majority of data on the Funtanazza sub-basin since inland the sedimentary series is poorly exposed.

4.1 Structural geometry

The Funtanazza sub-basin is a 2-9 km wide graben with an overall east-west trend (Fig. 4.1, Assorgia et al. 1986b; Assorgia and Gimeno 1994; Barca et al. 1996). Normal faults are thought to have formed this significant depression within Palaeozoic basement rocks (this study; Barca et al. 1996). The E-W strike of the Funtanazza sub-basin is unusual compared to the main Oligo-Miocene depocentres (Fig. 1.3). It is possible that the E-W orientation results from reactivation of western continuations of 'late Hercynian' structures which crop out in eastern Sardinia (section 2.2.2). The E-W to NE-SW trending structures were used as sinistral strike-slip faults until the late Aquitanian and passed at their western ends into transtensional basins (Chapters 6 and 7). From the present day outcrop pattern, it is apparent that ENE-WSW to E-W trending normal fault segments were linked by NW-SE to WNW-ESE trending transfer faults (e.g. northwest of Montevecchio and at Sa Tellura [577867], Fig. 4.1). The eastern end of the Funtanazza sub-basin was bounded by the NNW-SSE trending Campidano normal fault which defined the main Oligo-Miocene Sardinian rift trend as well as the Pliocene Campidano graben (Fig. 1.1; Barca et al. 1996; Fais et al. 1996). Offshore seismic reflection profiles reveal that the Campidano fault displays mainly extensional activity with limited evidence for sinistral strike-slip displacement (Thomas et al. 1988). The Arcentu Group volcanic centre formed at the intersection zone between the Funtanazza sub-basin and Oligo-Miocene proto-Campidano graben most probably because the extensional fault systems provided a route through the crust for magma.
At the coast, and traceable for a few kilometres inland, a high angle contact (>70°) between the surrounding Palaeozoic basement rocks and sub-basin fill can be observed (Fig. 4.1 A-A'). This feature strongly suggests that both the northern and southern margins of the Funtanazza sub-basin were bounded by -ENE-WSW trending normal faults (Fig 4.1). At the northern margin of the sub-basin, a late Burdigalian age basalt dyke is aligned in a E-W orientation in the vicinity of the fault zone [543855]. The dyke was most probably intruded along an extensional structure associated with the graben-bounding fault. Further inland (e.g. [576825]) Miocene sedimentary and volcanic rocks overstep the degraded fault scarp.

Structural data from the well-exposed coastal section and from inland exposures implies that the northern bounding fault has the larger throw, thus making the present day Funtanazza sub-basin an asymmetric graben (Fig. 4.1, A-A'). The Oligo-Miocene graben most probably had a similar structural geometry. Basin asymmetry would have been enhanced by uplift and tilting associated with late Burdigalian basaltic intrusion and/or late Burdigalian faulting (Fig. 4.1, A-A'). There is no evidence for post-Burdigalian, ENE-WSW normal fault movement.

Several NE-SW normal faults are found in the southern part of the sub-basin (this study, Assorgia et al. 1986ab). These structures post-depositionally offset a late Aquitanian (this study) rhyolitic lapilli tuff and basal calcarenites of the Sartori Formation (latest Aquitanian-earliest Burdigalian, Fig. 4.1) and progressively downfault the units to the northwest. The main post-early Burdigalian sedimentary depocentre which is exposed in the coastal section was apparently defined by a ~3 km long NE-SW normal fault (e.g. fault zone at [566851], Fig. 4.1). Sediments lying to the northwest of this structure generally dip to the NNW whilst those to the southeast are flat lying or dip to the southeast. Small, post-depositional normal faults with offsets of tens of centimetres (e.g. stereogram E, Fig. 4.1) and late Burdigalian basaltic dykes are also oriented NE-SW. In common with other areas of the Sardinian Rift (2.2.3, chapters 5-8), post depositional, N-S trending normal faults cut across the sedimentary and volcanic fill of the Funtanazza sub-basin (Assorgia et al. 1986b; Barca et al. 1996).

4.2 Basin filling geometries

4.2.1 Geometries within the Arcentu Group

The Arcentu Group consists of basaltic-andesitic lava domes, lava flows, andesitic breccia cones and pyroclastic flows cut by basaltic intrusions (e.g. Fig. 4.2; Assorgia et al. 1984, 1986ab). The volcanic rocks record a complex internal geometry with abrupt variations in thickness, local angular unconformities and breccia cones with dips of up to 30°. No conclusive, systematic dip variations or unconformities were observed. Although local tectonic movements may have occurred during the eruption of the Arcentu Group, the complex internal geometries can be attributed to eruptive and depositional processes active in the volcanic terrain (stratovolcanoes, Cas and Wright 1988). Dips of
andesitic breccia beds suggest that the main volcanic centres were located at M. Arcentu (Fig. 4.1), M. Maiori (2 km west of M. Arcentu) and to the east of the Campidano fault.

In the west of the Funtanazza sub-basin, breccia beds within epiclastic cones commonly dip at greater angles than the underlying volcano-sedimentary succession (e.g. M. Perdosu [575865] Fig 4.1, 20° as opposed to 12° of underlying Sa Tellura Formation).

Mid-late Burdigalian dykes trend systematically from NNW-SSE in the far east of the outcropping Arcentu Group to ENE-WSW/E-W in the western Funtanazza sub-basin (Fig. 4.1). The focal point of these dykes would lie to the northeast of the Funtanazza sub-basin in the present day Campidano plain (Fig. 4.1). It is curious that the dykes align with the NNW trending Campidano fault and the ENE-WSW trending Funtanazza sub-basin faults in the vicinity of the structures. A possible explanation would be that at the time of dyke intrusion, extension was occurring on both fault sets and that the systematic change in dyke orientation represented the changing stress field across the area.

4.2.2 Geometries within the sedimentary basin fill

Poor inland exposures mean that geometries within the sedimentary basin fill cannot be assessed. The majority of the rocks cropping out along the coastal section are conformable, parallel bedded and appear to passively infill the graben topography. One exposure at [543847] (Fig. 4.3) may show subtle bed thickening and divergence to the NNW, but this cannot be confirmed without examination of the section from a perpendicular viewpoint (i.e. offshore).

4.3 Sub-basin fill

This section briefly discusses the rock types and depositional environments of the Funtanazza sub-basin fill. The lithofacies used and justification of interpreted depositional environments are based on observations from subsequent study areas which were studied in more detail (Chapters 5 and 8).

4.3.1 Arcentu Group (Mid Oligocene-late Burdigalian)

The volcanic rocks which make up the Arcentu Group today form an impressive topography reaching 800m in height. Subaerially erupted, basalt and basaltic andesite flows and domes form the basal units (mid-late Oligocene, Formation A, Assorgia et al. 1984; Table 3.5). They are capped by a ~25m thick, areally extensive, rhyolitic pyroclastic flow (Assorgia et al. 1986b; Fig. 4.4). At the coast [547838, 545844] the flow contains centimetre-sized cognate andesite, Palaeozoic basement and pumice clasts, quartz, K-feldspar and biotite crystals (Assorgia et al. 1984, 1992a). It may have been deposited in a submarine environment according to Assorgia et al. (1992a). The flow forms an important marker horizon within the western sub-basin stratigraphy. Subaerial and submarine andesitic lava flows, pillows and breccias form the majority of the Arcentu Group (Aquitanian-Mid Burdigalian,
Formations B and C, Assorgia et al. 1984, 1986ab; Fig. 4.2). Rhyolitic-rhydacitic tuffs crop out in lenses within the Arcentu Group and sedimentary succession (Assorgia et al. 1984, 1986ab). Some of the tuffs have fine parallel lamination and are thought to have been deposited in a subaqueous environment (Assorgia et al. 1986a). Mid-Late Burdigalian basaltic dykes and intrusions (Formation D, Assorgia et al. 1984, 1986ab, Fig. 4.5) cut the older volcanic rocks and western sedimentary succession (Fig. 4.1, 4.2, 4.6).

4.3.2 Campu Sali Formation (late Oligocene, Assorgia et al. 1986, late Oligocene-late Aquitanian, this study)

The Campu Sali Formation is defined as fluvio-lacustrine with uppermost shallow marine sediments (Table 3.5). The rocks crop out under the rhyolitic flow marker bed and are best exposed on the Funtanazza coastal section [545844] (Lithofacies lh, 1g, 1e, 5k, 3d, Tables 5.6, 5.8, 8.4, 8.5; Fig. 4.7, log B).

The fluvial conglomerates (Assorgia et al. 1986a) consist of red, matrix-supported conglomerates (lithofacies 1e) with rounded Palaeozoic basement clasts. The unit is cross-cut by nodular caliche (lithofacies 1g) at [545844]. Lacustrine limestones (Assorgia et al. 1986a, 1992a) with a nodular texture evolve upwards into parallel laminated limestones containing rootlet horizons and organic layers (Lithofacies 1h, 5k; Figs. 4.7 log B, 4.8). The limestones are micrites with millimetre-sized, curved shells thought to be disarticulated ostracods. Marine transgression is recorded by the presence of oysters, Turritellid gastropods and bivalves in a shell hash bed (Lithofacies 3d) which also contains rare rip-up clasts of the underlying lacustrine limestones (Fig. 4.7 log B).

4.3.3 Sartori Formation (Aquitanian-mid Burdigalian, published work, latest Aquitanian-mid Burdigalian, this study)

The Sartori Formation consists of shallow marine calcarenites with minor calcirudites, cross-bedded calcarenites and grainstones (Lithofacies 2e, 2g, 3b, 3d, 3e, 3g, 4a, Tables 5.6, 5.8) and reaches up to ~150m in thickness. It overlies the rhyolitic flow marker bed and is well exposed in the Funtanazza coastal section (Fig. 4.7 logs A and C, type section).

The palaeontology and water depth of deposition of the coastal exposures were examined by Assorgia et al. (1992a). They found a wide marine fauna with numerous species of bivalves, Turritellid gastropods, bryozoans, echinoids, corals and bioturbation. Water depths immediately overlying the rhyolitic pyroclastic flow were 0-10m, passing upwards into a two transgressive (50-80m water depth) and regressive (10-50m water depth) cycles (Assorgia et al. 1992a).
The findings of Assorgia et al. (1992a) are in agreement with the trends and facies variations observed (Fig. 4.7 logs A and C). At Marina di Arbus [547838] (Fig. 4.1), the transgression over the rhyolitic pyroclastic flow occurs via a slightly irregular hiatal surface and is recorded by the deposition of shell hash beds (coquina) deposited in very shallow waters (Lithofacies 3d; Figs. 4.7 log A, 4.9). The overlying bioturbated muddy calcarenites containing whole echinoids and transported shell and bryozoan fragments (Lithofacies 3b, 4a) indicate further marine transgression (Fig. 4.7 log A).

At the base of log C, the Sartori Formation (Fig. 4.7) consists of bioturbated calcarenites (Lithofacies 4a, Table 5.8). The calcarenites commonly contain both broken, transported marine fauna and in-situ bivalves and echinoids (e.g. Fig. 4.10). Further up the succession, the calcarenites become coarser, with abundant volcanic clasts and are intercalated with volcanic horizons and calcirudite beds (Fig. 4.3, 4.11, 4.7 log C 22-48m; Lithofacies 2g, 3g). In agreement with Assorgia et al. (1992a), these sediments apparently record relative sea level fall. They also record the influx of volcanic and carbonate material from non-marine or very shallow marine settings and possible volcanic airfall deposition. Further relative sea level changes are indicated by deposition of fine bioturbated and faintly cross-bedded calcarenites (Lithofacies 3b, 4a; transgression; Fig. 4.3, 4.7 Log C 40-52m) followed by grainstones, calcarenites rich in wood pieces and coarse, shelly calcarenites (Lithofacies 3d, 3g; regression; Fig. 4.7 log C, 53-65m). The interaction of volcanism and marine sedimentation is discussed in more detail below.

Calcarenite and calcirudite outcrops exposed inland commonly contain coarse sand-pebble grade volcanic and Palaeozoic basement clasts, red algae, oysters and broken shell fragments suggesting a slightly shallower marine environment than the muddy calcarenites of the coastal section. The top surface of the calcarenite bed underlying the Sa Tellura Formation [579865] is a calcirudite breccia formed from the underlying calcarenite and is interpreted as 'palaeokarst' (after descriptions in Tucker and Wright 1990).

### 4.3.4 Sa Tellura Formation (Late Aquitanian, Assorgia et al. 1986, ?early Burdigalian, this study)

The ~30m thick lacustrine limestone succession (Assorgia et al. 1986a) of the Sa Tellura Formation crop out on top of the Sartori Formation and a tuff horizon in the hangingwall of the northern sub-basin bounding fault and transfer fault [575867] (Fig. 4.1). The sediments are thought to have been deposited during a period of relative sea level fall and can be correlated to shallowing in marine sediments exposed in the coastal section (Assorgia et al. 1986a, 1992a). The lacustrine limestones have rare quartz, andesite and Palaeozoic basement clasts, are sometimes silicified and contain algal structures, ostracods and non-marine gastropods (Assorgia et al. 1986a, Lithofacies 1h, 1j, Table 8.4).
In thin section, they are micritic with micrite intraclasts, micrite pellets, abundant thin shell fragments and ?ostracods (Fig. 4.12).

4.4 Mixed volcaniclastic-carbonate-siliciclastic sedimentation

This section briefly examines the interaction of volcanism and volcaniclastic sedimentation in a shallow marine shelf setting as exemplified by the Funtanazza sub-basin. This theme is discussed in more detail in chapter 11.

Basic volcanism in the Funtanazza sub-basin was concentrated in volcanic centres in the east of the area. Erosion of the volcanic centres resulted in the supply of basaltic and andesitic clasts from the high volcanic topography to the topographically depressed western sedimentary depocentre. Acidic pyroclastic volcanism, presumably from the same volcanic centres, supplied rhyolitic material in the form of flows and most probably as ash fallout.

Mixed volcaniclastic-carbonate sedimentation on a marine shelf is represented by crystal and pumice rich calcarenites, volcanic calcirudites (Fig. 4.11) and fine tuffaceous calcarenites to tuffs (Fig. 4.3, 4.7 log Q). The volcanic component is mixed into the sediment in a manner similar to either lime mud or clasts in a mixed carbonate-siliciclastic setting. That is, the volcanic crystals, pumice and glassy shards may be mixed into the sediment matrix or the volcanic component may take the form of discrete clasts. Apart from in fine tuffaceous calcarenites which may represent admixed airfall deposits, a diverse marine fauna exists within the volcaniclastic-carbonate sediments and is rapidly re-established after a volcanic event (e.g. base Sartori Formation). Volcanic calcirudites (Fig. 4.11) are interpreted as the transported products of volcanic fan deltas which coexisted at the same time as coral, red algae and bryozoan reefs. The coexistence of coarse ‘clastic’ material and carbonate build-ups is observed in modern settings. For example, in the Red Sea, carbonate reefs rim alluvial fans (Purser et al. 1986, Roberts and Murray 1988).

An unusual volcanic horizon consisting of purple volcanic mud spattered with centimetre sized ?pumice clasts and rare andesitic volcanic clasts crops out near Calada Bianca [543846] (Fig. 4.13). This bed may have formed due to the fallout of ash and pumice supplied directly from volcanic eruption.

The emplacement of Late Burdigalian basaltic magma formed dykes and other intrusions in the Funtanazza sub-basin (Assorgia et al. 1984, 1986ab). At the northern end of the Funtanazza coastal section [543855], the basaltic magma was intruded into shallow marine sediments and seawater forming impressive pillow lavas and peperites (Assorgia and Gimeno 1994). The pillow lavas are highly vesicular, have a glassy rind and can be up to 1.2m in diameter (this study, Assorgia and
Gimeno 1994). Peperites consists of pillows with entrained marine sediments and areas where pillows have disrupted the semi-lithified sediment (Assorgia and Gimeno 1994; Fig. 4.14). Further down the sedimentary succession, basaltic dykes cut across the marine sediments with little or no associated deformation (Fig. 4.6).

### 4.5 Timing of extension

It is generally accepted that the Funtanazza sub-basin was a late Oligocene-early Miocene graben associated with the formation of the Sardinian Rift (Assorgia et al. 1984, 1986; Assorgia and Gimeno 1994). However, no conclusive evidence constraining the timing of active extension in the Funtanazza sub-basin exists. Poorly exposed structural geometries do not provide any further information. However, a proto-graben structure must have formed before or contemporaneous with initial sedimentation and volcanism in the mid-late Oligocene in order to have created the accommodation space. A phase of early Burdigalian NW-SE normal faulting is tentatively identified after initial, widespread calcarenite deposition (?latest Aquitanian-earliest Burdigalian) and before this unit was covered by ?mid-late Burdigalian age andesitic flows and breccias of the Arcentu Group. Finally, an extensional phase may have facilitated the emplacement of late Burdigalian dykes and other intrusions.

### 4.6 Tectono-stratigraphic development

#### 4.6.1 Late Oligocene (published temporal data) or late Oligocene-late Aquitanian (this study) (Fig. 4.15a)

The E-W to ENE-WSW oriented Funtanazza structure apparently existed before the majority of sub-basin filling. Oligocene volcanism in the Funtanazza sub-basin consisted of localised basaltic domes and flows exposed in the east of the study area (Assorgia et al. 1984). In the west of the sub-basin, matrix supported conglomerates, calcrete palaeosols and lacustrine limestones with rootlet and organic layers were deposited in non-marine conditions (Campu Sali Formation). At the present day coast, the non-marine sediments were overlain by marine shell beds before the eruption of a rhyolitic pyroclastic flow (20.87± 0.3 Ma) which covered much of the Funtanazza sub-basin.

#### 4.6.2 Aquitanian (published temporal data) or latest Aquitanian-earliest Burdigalian (this study) (Fig. 4.15b, 4.16)

Widespread marine sedimentation is observed on top of the rhyolitic pyroclastic flow. Shallow marine shell hash beds fine upwards into bioturbated calcarenites deposited on a mixed volcanioclastic-carbonate-siliciclastic shelf in response to continued marine transgression. Subaerial and submarine andesitic volcanism resulting in lava flows, pillows and epiclastic breccia cones was active in the east of the sub-basin. Some volcanic material was supplied westwards into the sedimentary depocentre.
4.6.3 Late Aquitanian (published temporal data) or early Burdigalian (this study) (Fig. 4.14c)

After initial calcarenite deposition, a phase of NE-SW trending normal faulting can be tentatively identified. Relative sea level fall and contemporaneous volcanism recognised in the coastal section by the deposition of calcirudites and volcanic rich sediments can be correlated to rhyolitic tuff bed and lacustrine limestones (Sa Tellura Formation).

4.6.4 Mid-late Burdigalian (Fig. 4.15d)

Shallow marine, calcarenite dominated sedimentation continued in the westernmost part of the Funtanazza sub-basin until the mid Burdigalian with a transgressive to regressive trend. The final phase of late Burdigalian volcanism resulted in the emplacement of basaltic dykes and other intrusions, possibly in response to renewed extensional tectonism. Pillow lavas and peperites were formed where the basaltic lavas were erupted at shallow levels or into the sea.

4.7 Summary

- The Funtanazza sub-basin is a 2-9 km wide graben thought to have formed by extension on E-W to NE-SW and NW-SE trending normal faults. The sub-basin lies to the west of the main Sardinian rift trend as defined by the Campidano fault. At the intersection zone between the two extensional systems the Arcentu Group volcanic complex developed.

- It is unclear exactly when the extensional faults were active but the Funtanazza graben must have existed before the mid-late Oligocene to accommodate the sub-basin fill. A NW-SE phase of faulting is tentatively identified in the early Burdigalian and extension may have accompanied dyke emplacement in the late Burdigalian.

- The wider, eastern part of the Funtanazza sub-basin was filled with more than 800 metres of basaltic and andesitic volcanic rocks (Arcentu Group). At the same time, the western sedimentary depocentre was filled with non-marine conglomerates and lacustrine limestones (Campu Sali and Sa Tellura Formations), shallow marine carbonates and calcarenites (Sartori Formation) intercalated with rhyolitic tuffs and lapilli tuffs.

- Volcanic epiclastic material and ash fallout affected the composition of shallow marine sediments but did not prevent colonisation by a diverse fossil fauna. Peperites were caused by intrusion of basaltic magma into marine sediments just beneath the sediment-water interface (Assorgia and Gimeno 1994).
Chapter 5
Chapter Five - The Sarcidano sub-basin.

This chapter describes the structure, sedimentology and depositional architecture of the Sarcidano sub-basin (Fig. 1.3). A model for the tectono-stratigraphic development of the area is proposed. The Sarcidano sub-basin was studied in detail because the complex sedimentary and volcanic basin fill exhibits rapid lateral and vertical facies changes which show a close relationship to the basin structure. Sedimentation is interpreted to be a response to three phases of normal faulting contemporaneous with relative sea level change. This study is useful for evaluating the main controls on sediment dispersal and accumulation in a complex intra-arc setting, spatial and temporal models of syn- and post rift sedimentation (e.g. Leeder and Gawthorpe 1987; Prosser 1993) and the relative importance of local post-rift as opposed to syn-rift sedimentation (chapter 11).

The stratigraphy of this area was described in section 3.3.2, Figures 3.12, 3.13 and Appendix 3D. The succession comprises basal clastic sediments of the Nureci Formation (continental Villanovatulo and shallow marine Duidduru Members), the volcanic succession of the Araxigi Formation, the mixed siliciclastic-carbonate rocks of the Isili Formation (Is Paras and Serra Longa Members) and finally the calcarenites and marlstones of the Giara Group. The basin fill was deposited within a complex structural framework of tilted, uplifted fault blocks and half-graben defined by two main sets of high angle normal faults in a rift-margin position (Fig. 5.1). An examination of basin filling geometries (section 5.3) indicates that this rift-margin structure had formed before the majority of basin filling occurred. Facies variability at the basin margin developed partly because sedimentary and volcanic rocks accumulated within quasi-independent depocentres defined by the normal fault structures. Enclosures 1 and 2 accompany this chapter.

5.1 Previous Work

To date, published work on the sedimentary rocks of the Sarcidano sub-basin has concerned small areas or sections with a palaeontological and biostratigraphic bias (Leone et al. 1984, Genoni area; Cherchi 1985, Giara di Gesturi section; Cherchi and Montadert 1982ab). Cherchi and Montadert (1982ab) and Tremolieres et al. (1988) discuss briefly the facies variability and rift margin structure around Monte Trempu (the Isili fault block, Fig. 5.1). The volcanic succession termed here the Araxigi Formation was described by Assorgia et al. (1995ab, in press). Unpublished ENSPM and British Petroleum field guides provide some interesting ideas concerning the evolution of the Sarcidano sub-basin. Unfortunately, a number of Institut Francais du Petrole theses on parts of the Sarcidano sub-basin were not obtainable.
5.2 Structure of the Sarcidano rift margin

The Sarcidano sub-basin is an approximately 45 km long, NW-SE trending zone bounded at its southeast and northwest margins by N-S to NNE-SSW trending zones of normal faults which follow the main Sardinian rift trend (Figs. 1.1, 1.3, 5.1). This section describes the structural geometries which accommodated extension at the rift margin. At present, Oligo-Miocene palaeofault scarps and topography are degraded and/or buried by sedimentary and volcanic rocks. Thus a variety of techniques were utilised to define the Oligo-Miocene rift margin structure (see Appendix 1A). After a description of the general characteristics of the Sarcidano sub-basin, structural field data is tabulated for four separate study areas.

5.2.1 Previous Work

Although there exists a general appreciation in the literature that the Sarcidano area represents an Oligo-Miocene rift margin which was bounded by normal faults, the location, geometry and kinematics of such structures are largely undefined. Cherchi and Montadert (1982ab) provide the exception to this in their description of '5-20° tilted basement blocks', bounded by a 'master fault' of throw ~500 m, resulting in a half-graben structure at the basin margin (Figure 5.2). In addition, Cherchi and Montadert (1982ab) mention, but do not locate, a series of 130-150° oriented faults, dipping 50-80°, with throws of 20-100 m, and a second intersecting normal fault set, trending 170-190°, dipping at ~80° and defining a horst, graben and half-graben topography.

5.2.2 General structural characteristics

5.2.2.1 Structural style

Although not directly exposed, several lines of evidence can be used to delimit the large, high-angle, planar normal faults which accommodated extension at the Sarcidano basin margin. Observations include the nature of pre-rift topography, pre-rift and basin filling geometries, offsets and tilting of pre-rift and basin fill units, small faults sub-paralleling larger structures and the geometry of degraded fault scarps. It must therefore be noted that much of the large fault data has been estimated rather than directly measured.

Synthesis of field data (below) shows that extension occurred on large, roughly NW-SE trending (~130-150°), and cross-cutting NNE-SSW/NE-SW trending (~020-050°), 5-10 km long, segmented normal faults with throws up to hundreds of metres (Figs. 5.1, 5.3). Fault planes commonly dip at high angles between 45 and 90°, often 60-70°, and are planar at the surface. The two fault sets defined a rift-margin composed of a line of uplifted, tilted, 3.5-7 km wide basement blocks and 3-5 km wide half-graben (Figs. 5.1, 5.3, 5.4). Low areas existed between fault blocks along the length of the sub-basin (Figs. 5.1, 5.3). Smaller, planar, generally high angle normal faults with throws up to tens of metres are observed sub-parallel to the larger structures. They complicated the pre-rift basement
topography and disrupted the basin fill. The NNE-SSW/NE-SW trending cross-faults occurred at high angles (±perpendicular) to the rift-margin, demonstrably cutting across NW-SE trending normal faults (e.g. Isili front [050975], Villanovatulo [184013], Fig. 5.1). These faults tip out from throws of a few hundred metres to tens of metres or zero over a few hundred metres (section 5.2.2.3). The throw on the Laconi fault (Fig. 5.1) may decrease towards its NW and SE ends. The maximum throw would have been north of Nurallao if the present day topography and dolomite offsets record an Oligo-Miocene feature. The process of fault scarp degradation was active from the Oligo-Miocene (section 5.3.3) and continued until present day. The consequence of erosion is the retreat of the fault scarp and the lowering of the scarp angle (Figs. 5.1, 5.4). If tilted fault blocks are restored to an uneroded state, the amount of rock removed by fault scarp degradation on the block crest may be up to a few hundred metres.

5.2.2.2 Basement inheritance

The change in orientation of rift-bounding faults from N-S to NW-SE in the Sarcidano sub-basin mirrors those of Hercynian thrust-fronts and shear-zones in the pre-rift basement (Carmignani et al. 1987; Fig. 5.1). In addition, the NW-SE trending line of the topographically high, fault bounded Isili, Genoni, San Antonio Ruinas and Grighini tilted blocks is on the culmination of the Hercynian 'Flumendosa Antiform' (Carmignani et al. 1992a; Fig. 2.9). In common with other extensional settings (Chadwick, 1986; Powell and Williams 1989; Williams et al. 1989), some thrust lineaments and late Hercynian detachment faults could be re-activated at depth. In addition, it may be that NNE-SSW/NE-SW trending cross-faults followed a 'late Hercynian' structural fabric (section 2.2.1).

5.2.2.3 Transfer zones and cross faults

Overlapping, segmented normal faults in extensional rifts are linked by transfer zones which exhibit geometries from relay ramps between fault tip lines, through a complex arrangement of small faults to one distinct transfer fault (Larsen 1988; Morley et al. 1990; Nelson et al. 1992; Gawthorpe and Hurst 1993; Fig. 5.5). In the Sarcidano sub-basin, normal fault linkage occurs with morphologies similar to the two end members. The NW-SE trending, synthetic, en-echelon fault segments of the Laconi, Nurallao and Asuni faults transfer their displacement across a low area without evidence of linking faults and the Laconi fault may tip out towards this area, consistent with a relay ramp morphology (sensu Larsen 1988; Gawthorpe and Hurst 1993; 4 km WSW of Laconi [000100]).

Two hypothesis can be proposed for the origin of the cross faults;
a) they are similar to transfer-faults observed in other extensional settings
b) they represent a change in the way extension was accommodated along this section of the basin margin and a change in the orientation of the extensional stress field.
In Sarcidano, high angle cross-faults have a morphology similar to transfer faults (Morley et al. 1990; Gawthorpe and Hurst 1993) in that they are at high angles to major basin bounding faults and tip out over tens of metres. Transfer-faults are characterised by oblique and strike-slip movement, they transfer basin bounding structures (Nelson et al. 1992; Gawthorpe and Hurst 1993; Fig. 5.5). In the Sarcidano sub-basin, cross-faults occur where large NW-SE trending structures jump basinward (e.g. Isili cross-fault to Nurallao fault, Genoni cross fault to Asuni fault, Fig. 5.1), but although data is limited, it appears that cross-faulting occurred by dip-slip, normal fault movement (section 5.2.2.4). Transfer faults may move synchronously with the fault segments they join, or at a late stage between propagating en-echelon segments (Nelson et al. 1992). In the Sarcidano sub-basin, cross-faults cut across NW-SE structures (e.g. west Isili block [050975], Villanovatulo [184013]). The second hypothesis for the origin of cross-faults is favoured here because the primarily dip-slip cross-faults cut and post-date the NW-SE structures rather than appearing to be related to them (Fig. 5.5). Also, sediment geometries in the basin fill indicate that extension occurred on outcrop scale NE-SW to N-S trending normal faults from time to time after the main phase of cross-faulting (section 5.3.2), indicating that the extensional stress field indeed changed. Thus, it is proposed that in the Sarcidano sub-basin there are two phases of major extension; NW-SE faulting followed by NE-SW to NNE-SSW cross-faulting.

5.2.2.4 Small faults and fault slip data
Rare slickenside data on small faults parallel to the large extensional structures, along with the geometries of syn-sedimentary faults illustrate that the Oligo-Miocene extension occurred dominantly by dip-slip fault movement on NW-SE and NE-SW/NNE-SSW faults (e.g. Fig. 5.6 st (stereogram) 1-5, 7, 8). Minor components of dextral and sinistral oblique slip are recorded by some slickensides but the limited data set shows no consistent sense of oblique slip. Of particular interest is the predominant dip-slip sense of fault slip from NE-SW cross-faults (e.g. Fig 5.6 st3), though slickensides from the Isili block cross-fault also show a sinistral component (Fig. 5.6 st3).

FIELD OBSERVATIONS

5.2.3 Villanovatulo-Serri area and the surrounding pre-rift basement
Table 5.1 summarises the post mid Oligocene observed structures for this area. Large normal faults which step the basement down towards the rift are observed in pre-rift rocks bounding the southeast Sarcidano sub-basin. The basinward dip of these rift-bounding normal faults is opposed to the dip of Hercynian thrusts, yet the change in strike of the rift-bounding structures from NW-SE (e.g. Ortuabis fault) to ~N-S (e.g. Serri-Mandas fault) mirrors that of Hercynian thrust fronts on Carmignani et al. (1987; Fig. 5.1) suggesting some control by the older fabric. Around Villanovatulo, a complex pattern of NNW-SSE trending, rift-bounding and antithetic faults were cut by the Villanovatulo cross-fault which forms the accommodation space for the deposition of the Villanovatulo Member conglomerates.
Syn-sedimentary faults (section 5.3.2) within the lowest exposed levels of the Villanovatulo Member sub-parallel the NW-SE trend of the basin margin (Fig. 5.6 st1).

### 5.2.4 Isili fault block to Laconi

Table 5.2 summarises the post mid Oligocene observed structures for this area. The Isili fault block is defined by NW-SE and NE-SW trending, high angle normal faults with throws of hundreds of metres. Together with the Laconi and Nurallao NW-SE trending faults, a tilted, uplifted basement block and half-graben geometry is delimited (e.g. Fig. 5.1 C-C’). If Mesozoic dolomite was originally flat lying, the Isili fault block has been tilted by 15°. The Isili block cross-fault and Gergei fault both tip-out to the northeast over 100’s of metres. Smaller faults with throws from a few centimetres to 10’s of metres complicate the block and half-graben topography with the formation of smaller terraces (Gemmuri fault; Fig. 5.1).

### 5.2.5 Genoni fault block to Asuni

Table 5.3 summarises the post mid Oligocene observed structures for this area. The Genoni fault block and Asuni fault define a similar structural style as in the Isili area but the Genoni block fault was cross-cut by several, small NE-SW to N-S trending normal faults at its southeast margin (Figs. 5.8, 5.9).

### 5.2.6 Grighini-Samugheo area

Although not studied in great detail, the present day topography and that beneath the sediments and volcanic rocks of the Nureci and Araxigi formations indicates that there must have been a series of NW-SE and NNE-SSW trending normal faults with throws of hundreds of metres and lengths up to 7 km (Fig. 5.1). They defined the uplifted Grighini fault block, subsided half graben geometry and adjacent low areas, in a style similar to that observed to the southeast. For example, south of Samugheo [939159], exposures on a hillside reveal Hercynian pelites bounded by a high-angle, planar boundary, believed to be the scarp of the Asuni fault, which Araxigi Formation ignimbrites outcrop adjacent to and cover. Other abrupt thickness variations in the Araxigi Formation (e.g. between [930180 and 947174]) over hundred of metres, related to linear pre-rift morphologies are strong indicators normal faults must have been present. West of Allai, metre scale, sealed fault bounded basement blocks on the dip-slope of the Grighini fault block, trend NNW-SSW to NNE-SSW (Fig. 5.6 st9, 5.11) in general alignment with the NNE-SSW orientation of the rift-bounding Busachi fault to the north.
5.3 Basin filling geometries

This section summarises and illustrates the basin filling geometries identified at the Sarcidano rift-margin in a series of tables, sketches and photos. The results are particularly useful in constraining the timing of sub-basin events.

5.3.1 Pre-rift -basin fill unconformity

Table 5.4 summarises the nature of the pre-rift - basin fill unconformity in the Sarcidano sub-basin. At outcrop scale, the pre-rift-basin fill unconformity consists of covered and onlapped irregular basement topography (Figs. 5.11, 5.12), fault planes (Figs. 5.10), degraded fault planes (Figs. 5.3, 5.8) or fault related topography (Fig. 5.13) such as tilted fault-block dip slopes (Fig. 5.14). The larger scale nature of the unconformity, which is not directly exposed, is discussed below (section 5.3.3).

5.3.2 Evidence for syn-depositional fault movement

Table 5.5 summarises the evidence for syn-depositional fault movement within the Sarcidano sub-basin. Although the Sarcidano sub-basin margin is defined by normal faults with throws of least 500m, Table 5.5 summarises all the evidence observed for syn-depositional fault movement in this region. Common syn-depositional features include bed divergence, angular unconformities associated with sealed normal faults and fold structures. It is noticeable that syn-depositional features are observed only rarely and are of outcrop scale (<50m).

5.3.3 Large scale geometries - post rift basin infilling

This section describes the large scale sediment geometries identified by correlating bed dips and outcrop geometries between exposures.

5.3.3.1 Villanovatulo Member

Although it is difficult to constrain the geometry of the discontinuous Villanovatulo Member outcrops across the study area, it appears that the sedimentary rocks occur as 'wedges' which thin away from the sediment source point, controlled by the local fault topography. For example, an alluvial fan (section 5.4.1.4) sourced from the dip-slopes of the Grighini fault block is intercalated with and laterally equivalent to ignimbrites of the Araxigi Formation (Fig. 5.3 J-J'). The internal geometry of these coarse clastic deposits deposited by alluvial processes (section 5.4.1) is complex. Bedding is not easily discernible, surfaces are often cross-cutting (e.g. channels) and the sediments may have a depositional dip. Syn-depositional faulting is identified only at the base of the Villanovatulo Member at Villanovatulo ([186017] Fig. 5.15) Isili reservoir ([083004] Fig. 5.16) and west of Allai ([875239] Fig. 5.11). In addition, the dispersal paths and sedimentary supply points of the Villanovatulo Member suggest that NW-SE faults (and N-S to E-W faults at the Sarcidano sub-basins periphery) had formed a
block and graben topography before the deposition of the sediments. Thus much of the Villanovatulo Member may represent a phase of topographic degradation which occurred after the first phase of basin margin faulting. The sediments are classified as syn-rift to early post-rift deposits.

5.3.3.2 Duidduru Member

The large scale geometry of the Duidduru Member is that of passive infill and onlap onto fault created topography. North and west of the Isili and Genoni cross-fault hangingwalls, the Duidduru Member comprises an overall wedge shaped geometry (Figs. 5.1, 5.3). Although bedding is rare, the sediments appear to passively infill the cross-fault topography rather than be associated with coeval fault growth. Onlap on to the degraded cross-fault scarp is also observed e.g.[056980]. Perpendicular to the block and half-graben topography, the Duidduru Member also infills fault topography and onlaps onto degraded fault planes (Figs. 5.1, 5.3, 5.8). Along the front of the Genoni block, the onlap of the Duidduru member onto the degraded Genoni block fault is particularly well exposed (Fig. 5.3, 5.8).

On the dip-slope of the Isili fault block, Duidduru Member clastic sediments transported in fan deltas form a thin layer (Fig. 5.14). Limited syn-depositional fault movement within the Duidduru Member is indicated at Isili Reservoir where faults and bed thickening (Figs. 5.3, 5.17) indicate movement on local NW-SE trending faults, and east of Senis on local N-S to NNE-SSW trending faults (Fig. 5.18). The bulk of the Duidduru Member which was deposited after the major phases of normal faulting had created the topography can be classified as a post-rift deposit.

5.3.3.3 Araxigi Formation

The majority of the Araxigi Formation is parallel-bedded, infills, onlaps and covers fault-created accommodation space in the northwest of the Sarcidano sub-basin (Figs. 5.1, 5.3). The geometry of the pre-rift to Araxigi Formation contact was controlled by the underlying fault-topography, for example south Samugheo [939159] and the relay-ramp geometry WSW of Laconi [000100]. Although individual ignimbrite eruption events up to 40-80m thick (Assorgia \textit{et al.} 1995a), instantaneous on a geological timescale, would not record syn-depositional fault movement, one would observe dip variations and unconformities between volcanic units if the Araxigi Formation were syn-depositional with major fault displacements. West of Allai, the basal ignimbrite flow of the Araxigi Formation, intercalated with Villanovatulo Member conglomerates, crops out locally with a dip of 22° whereas the rest of the overlying Araxigi Formation dips at 10°. Because of the level of exposure, it is difficult to deduce whether the dip variation could be a result of Grighini Block NW-SE normal fault movement and block rotation or local, late faulting as suggested by the abrupt end to the outcrop. In any case, the majority of the Araxigi Formation was erupted after major fault displacements had occurred i.e. post-rift deposition.
5.3.3.4 Is Paras Member

The majority of the Is Paras Member consists of parallel beds which passively infill and onlap fault formed topography with an aggradational then retrogradational geometry (Figs. 5.1, 5.3). Particular areas of the Is Paras Member were affected by syn-depositional normal faulting. For example, north-west of the Isili fault block, the tip splay of the Isili block cross-fault was active within the lower Is Paras Member ([067014-075006] Fig. 5.19). At similar depositional levels in Isili Reservoir, local dip variations and slumping imply coeval, local normal faulting (Fig. 5.3) on a NW-SE trending structure. Localised syn-sedimentary NNE-SSW normal faulting affected certain levels of the upper Is Paras member (Figs. 5.20, 5.21). Cherchi (1985) observed a ‘remarkable fan-shaped variation between the base and the top of the carbonatic series’ (i.e. Is Paras Member) implying coeval fault-block rotation. Although there are dip-variations within the carbonates (above, Fig. 5.3), they appear to be local and a result of short-lived, local normal faulting as opposed to through the entire sedimentary unit due to long-lived fault-block rotation. Thus, the Is Paras member was deposited after the formation of the block and half-graben topography (post-rift), but was affected by local faulting episodes in several orientations (local syn-rift deposits).

5.3.3.5 Giara Group

The vast majority of the thick Giara Group sequence consists of parallel-bedded units which infilled and onlapped any remaining accommodation space (Fig. 5.1). At the base of the group, where time-equivalent to the Isili Formation, ~N-S oriented normal faulting caused disruption and an angular unconformity to develop west of the Isili fault block ([037992, 960023] Figs. 5.22, 5.23). Along the front of the Isili fault block [080950 and 084959], local structures such as dips variations, open folding, unconformities (Fig. 5.24) and disturbed beds are believed to be the result of fault propagation folding (sensu Withjack et al. 1990; Gawthorpe et al. 1997; Sharp et al. in press) related to small ~Burdigalian displacements of the Isili block fault or fault drag, perhaps enhanced by early sediment compaction. However, the majority of the Giara Group at the Sarcidano sub-basin margin was a post-rift megasequence with the characteristic geometries of onlap and infill.

5.3.3.6 Summary

A syn-rift megasequence (sensu Prosser 1991, Appendix 1B) can be identified from the late Rupelian - late Burdigalian (~30Ma-17Ma, Villanovatulo Member-basal Giara group). However, within the megasequence, the majority of the succession passively infills and onlaps onto a earlier fault-formed topography. Only localised syn-rift deposits are recognised.
5.3.4 Post-depositional deformation

High angle, post-depositional normal faults and fractures commonly cut across and offset the Oligo-Miocene basin fill. The faults are high angle (70°-90°), with thin, anastomosing fault planes, offsets of a few millimetres to tens of metres and often trend N-S (Fig. 5.25).

Gentle, post-depositional, open synclinal folding and tilting of the entire Oligo-Miocene succession is common at the Sarcidano sub-basin margin (Figs. 5.1, 5.3). Adjacent to fault-scarps, dips can reach up to 54° (front Isili block [056970]), and at a longer wavelength, dips are commonly between 5° and 10°. These structures are interpreted to result from fault propagation folding (sensu Withjack et al. 1990; Gawthorpe et al. 1997; Sharp et al. in press) of the Oligo-Miocene succession, which occurred due to post-Langhian reactivation of the Oligo-Miocene normal faults, and from compaction of the basin fill sequence over basement blocks. The timing of the late fault movement is probably early Pliocene, contemporaneous with the formation of the NW-SE trending Campidano Graben but before the eruption of the Giara di Gesturi alkaline plateau basalt (2.67±0.2 Ma, Beccaluva et al. 1985) which unconformably covers slightly tilted Miocene sediments (Cherchi 1985).

5.3.5 Evidence for a Burdigalian phase of compression

Cherchi and Montadert (1982ab) and Cherchi and Tremolieres (1984) proposed that a NE-SW directed Burdigalian phase of compression affected Sardinia (section 2.2.3). In the Sarcidano sub-basin, one outcrop of the Villanovatulo Member (Isili factory section [100029]) shows a folded conglomerate bed associated with a post-depositional thrust fault oriented NW-SE (Fig. 5.26). It is unclear when or how this occurred. Enhanced dips, unconformities and disturbed beds within the Giara Group, which may have been interpreted as compressional features, are here thought to relate to fault propagation folding and/or early sediment compaction (above) because of their location and lack of other compressional features. The scenario in Sarcidano is perhaps similar to the Oligo-Miocene of the Gulf of Suez rift where it has recently been argued that structural geometries previously related to a phase of compression (Knott et al. 1995) are in fact the result of the extensional phase (Gawthorpe et al. 1997; Sharp et al. in press). In general, localised compressional deformation may occur within an extensional terrain due to local space problems, for example, in between faults of different orientations.
5.4 Facies descriptions and interpreted depositional environments

This section describes in detail the nature of sedimentary and volcanic rocks found within the Sarcidano sub-basin fill. The basin fill was deposited within a complex structural framework (section 5.2). Here, the complex variety of sedimentary and volcanic rocks which developed between and around the emergent Isili, Genoni and Grighini tilted fault blocks (Fig. 5.1) and the basin margin is discussed in stratigraphic succession. The section is structured such that for each stratigraphic unit a description of lithofacies is followed by a description of the textures, compositions, sedimentary structures, lithofacies architecture and interpreted depositional environment of the rocks for different areas of the sub-basin. This description is necessary to present all the information needed to constrain the basin fill variability. The data represented in bar charts and rose diagrams is listed in Appendix 5A.

5.4.1 Nureci Formation, Villanovatulo Member (?late Oligocene - late Aquitanian).

The Villanovatulo Member is defined here as a succession of continental conglomerates and breccias which were deposited by alluvial fan to braided fluvial systems. No simple type section exists due to incomplete exposure, rather some typical outcrops are defined around [179029-178028] (Table 3.6). The formation is the stratigraphically lowest unit and overlies a faulted pre-rift basement topography. Outcrops occur discontinuously throughout the study area (Fig. 5.1, Encls. 1, 2). Previous work has described these sediments as "megabreccias at the foot of fault scarps" and "heterometric conglomerates in a red-violine sandy-muddy matrix" deposited in a continental environment, with structures typical of alluvial fans or torrential deposits (Cherchi and Montadert 1982a). In the vicinity of Allai, Assorgia et al. (in press) describe continental, polygenic conglomerates deposited by 'massive gravitative flows' and 'channelled flows' in an alluvial fan system under semi-arid conditions.

5.4.1.1 Lithofacies

The sediments of the Villanovatulo Member can be divided into seven lithofacies (Table 5.6).

- **Lithofacies 1a** is a massive, disorganised clast supported breccia, with a chaotic arrangement of very poorly sorted and angular clasts (Fig. 5.27). It is locally sourced, found in proximity to basement topography, and is interpreted as the deposits of scree or talus gravity fall (sensu Blair and McPherson 1994) to very proximal, cohesionless debris flows (sensu Postma 1990).

- **Lithofacies 1b** is a massive, disorganised clast supported conglomerate with sub-rounded, poorly sorted clasts (Fig. 5.28) which are occasionally massively stratified or imbricated. These features are indicative of deposition from high concentration cohesionless debris flows (Nemec and Steel 1984; Postma 1990; thicker, coarser beds or clast-rich debris flows, Miall 1996) to fluidal sheetflow (Bull 1972) or streamflood events (Nemec and Steel 1984, thinner and better sorted beds).
• **Lithofacies 1c** is a crudely reverse-graded, clast-supported conglomerate with sub-rounded and moderately sorted clasts (Fig. 5.29). It is interpreted as the product of a surging debris flow (Nemec and Steel 1984) or a high strength, clast rich debris flow (Miall 1996).

• **Lithofacies 1d** comprises clast-supported conglomerates and gravelstones with sub-rounded and moderately sorted clasts which are channelised, cross-bedded, imbricated or normally graded (Figs. 5.30, 5.31). The sedimentary structures and discontinuous nature of this lithofacies indicate that they were deposited by sheetflow (Bull 1972; Blair and McPherson 1994; or fluidal sediment flow, Nemec and Steel 1984) to braided stream channel processes (Steel and Wilson 1976) such as longitudinal bars and channels (Miall 1996).

• **Lithofacies 1e** is a thinly bedded (< 0.5m) ungraded, matrix-supported conglomerate with poorly sorted, angular to rounded clasts. Such sediments are interpreted to have been deposited by viscous, cohesive debris flows (Bull 1972; Postma 1990; Miall 1996).

• **Lithofacies 1f** consists of moderately sorted, coarse to fine sandstones with faint parallel lamination or poorly defined troughs and channels (Fig. 5.32). These structures are indicative of stream channel reworking (Bull 1972; Steel and Wilson 1976) by bars, dunes and scour fill (Miall 1996).

• **Lithofacies 1g** consists of mm- to cm-scale planar, crack filling and nodular micrite generally within reddened, oxidised iron rich sandstones and siltstones (Fig. 5.32). It is interpreted to represent the early stages of caliche or calcrite development, formed within subaerially exposed sediments (after Collinson 1986; Miall 1996).

5.4.1.2 Villanovatulo Town

The Villanovatulo Member is exposed in road cuttings over an area of ~10 km² SSW of Villanovatulo town (Fig. 5.1, enclosure 1) where it consists of up to 175 m of clast supported conglomerates (Lithofacies 1b, 1c, 1d; Table 5.6) with minor calcretes, coarse sandstones and matrix-supported conglomerates (Lithofacies 1e, 1f, 1g; Table 5.6). The clastic sediments lie at the eastern edge of the Sarcidano sub-basin within a fault-bounded depocentre created by the intersection of a NE-SW to E-W trending cross fault with a slightly older NNW-SSE trending basin-bounding fault (Fig. 5.1, Enclosure 1). NW-SE trending faults antithetic to the basin-margin faults complicated the depositional topography further (Fig. 5.1, enclosure 1).

5.4.1.2a Grain size, shape and sorting.

The variability of clast size and sorting with location is illustrated on Fig. 5.33. Organised conglomerates (Lithofacies 1d) consist of moderately sorted clasts less than 10 cm in size (locality 1). Disorganised conglomerates (Lithofacies 1b) exhibit poor sorting with the majority of clasts less than 10 cm in size but with a percentage ranging up to 80 cm. Within Lithofacies 1b, the most poorly sorted deposits with the largest outsize clasts occur in proximity to the basement source (localities 5 and 8) at the top of the succession.
Clast composition influences the clast shape. Dolomite clasts tend to be sub- to well-rounded and ellipsoidal in shape, whilst metamorphic clasts are angular to sub-rounded. Many of the dolomite clasts are pitted and bored, with the borings exhibiting no preferential orientation. It is difficult to envisage how this boring occurred in the Oligocene continental environment. It seems likely that some pebbles are recycled clasts, initially eroded and bored in the Mesozoic or lower Eocene, at a time when other shallow marine deposits are found in central Sardinia (Cocozza and Jacobacci 1975; Ferrara et al. 1992).

5.4.1.2b Composition and provenance

Graphs of clast composition from 10 different localities show that within the clast-supported conglomerates (Lithofacies 1b and 1d) the stratigraphically highest and most proximal conglomerates are dominated by dolomite clasts (locality 2, dolomite 90%, Fig. 5.33). The stratigraphically lowest conglomerates have a greater proportion of metamorphic to dolomite clasts (locality 7, metamorphic 43%, dolomite 49%). Overall, the percentage of dolomite generally increases up the succession. It seems likely that this is because the dolomite is harder and more resistant to erosion through sediment recycling, whereas the metamorphic, pelitic rock is softer and has planes of weakness defined by clay mineral schistosity. The composition of minor, matrix supported conglomerates is distinctly different, with higher percentages of metamorphic clasts than the surrounding clast-supported conglomerates (e.g. locality 2, metamorphic 37%, dolomite 57%), implying that such flows sampled a different source area. The matrix fill of conglomerates (Appendix 1A) is mainly composed of very poorly sorted, angular metamorphic clasts, rare dolomite clasts, and red sand - siltstone grade material. In places (e.g. locality 6), the matrix contains a fresh volcanic mica, volcanic feldspar and micrite (ripped-up caliche) component.

Clast composition suggests that the majority of clasts were eroded from Mesozoic dolomite and Hercynian metamorphic lithologies which are found surrounding the Villanovatulo Member in this area (Fig. 5.1). Minor clast components include soft, yellow siltstone-sandstone which shows affinities to upper Eocene Cixerri Formation or lower Eocene sediments found in southern Sardinia, rare Eocene clasts packed with Nummulites sp., intraclasts of micrite (caliche rip-ups), and occasional red trachyte clasts with fresh plagioclase, pumice and glass relicts. The source of the trachyte, which shows many textural and compositional similarities to Oligo-Miocene volcanics, is unknown.

5.4.1.2c Sedimentary structures and facies architecture

Disorganised clast supported conglomerates (Lithofacies 1b) outcrop as massive banks or as laterally extensive layers greater than 1 m thick. Their basal contact is often cross-cutting and erosive with little evidence of multiple flows other than faint massive stratification and diffuse changes in grain sizes. Organised conglomerates (Lithofacies 1d) exhibit cross-bedding (Fig. 5.31), clast imbrication, normal grading and may be channelised (Fig. 5.30). Channels are concave, usually about metre deep, up to ~7
m in width, and often occur in cross-cutting, stacked systems. Coarse to fine sandstones (Lithofacies 1e) occur as channel-fill and also as poorly exposed zones surrounding channels. In the latter location they sometimes contain micrite caliche and are reddened (Lithofacies 1g).

Figures 5.34 and 5.35 illustrate the palaeocurrents from 8 localities in the Villanovatulo area. Apart from two channel orientations (Locality 4) and two cross-bedding readings (Locality 2, 7), the data come from clast imbrications in organised conglomerates (Lithofacies 1d) and rarely in disorganised conglomerates (Lithofacies 1b). The rose diagrams indicate that the overall palaeoflow trend was one of supply from the northeast and north and dispersal to the southwest and south, basinwards. However, in detail, local fault topography is likely to have influenced local drainage patterns, for example, the variability at localities 2-4 and the dominant southwards palaeocurrents in the vicinity of the cross-fault from localities 5-8.

Facies distributions indicate that proximal areas (north and northeast) are dominated by massive, erosively based, disorganised conglomerates which cross-cut channelised, cross-bedded and palaeosol horizons. In more distal areas, channels and imbricate beds plus patches of finer disorganised conglomerates are interbedded with poorly exposed coarse sandstones (Fig. 5.35).

5.4.1.2d Interpreted depositional environment

Proximal facies are thought to be characterised by deposition of cohesionless debris flows and sheetflows with occasional stream channel reworking and palaeosol development. Distal facies indicate deposition by streamflow processes with less common and voluminously smaller cohesionless debris flows. Studies have shown that debris flows and coarse grain sizes occur on the proximal parts of alluvial fans, whilst there is an increase in channel processes and a decrease in grain size to lower or distal parts of fans (Hooke 1967; Bull 1972; McGowen and Groat 1971; Collinson 1986; Miall 1996). Thus, proximal facies at Villanovatulo represent the upper parts of an alluvial fan system whilst distal facies are interpreted to have been deposited on the mid-part of an alluvial fan transitional to a braided fluvial system (Fig. 5.35). The area shows characteristics of both the debris-dominated and sheetflow dominated fans described by Blair and McPherson (1994). The supply of sediments to the alluvial fans, the sediment architecture and dispersal are intimately related to the fault-defined palaeotopography (Fig. 5.35).

5.4.1.2 Isili Area

Less than 10 m of the Villanovatulo Member is exposed directly over pre-rift rocks in two localities, on the dip-slope of the Isili fault block, cropping out at abnormally low water levels of Isili reservoir west of the town [803004] and to the north of the town in a road cutting south of a NW-SE trending normal fault [100029] (factory section). It consists of disorganised conglomerates, graded
conglomerates, cross-bedded channel filling sandstones, palaeosols and thin pebble beds within massive coarse sandstones and gravelstones (Lithofacies 1b, 1c, 1d, 1e, 1g; Table 5.6). The continental conglomerates and sandstones of the Villanovatulo Member pass conformably upwards into similar lithofacies of the Duidduru Member which contain a marine fauna (e.g. Fig. 5.97 below).

5.4.1.2a Grain size, shape and sorting.
Disorganised and organised clast-supported conglomerates (Lithofacies 1b, 1c, 1d) are moderately sorted, have moderate sphericity, are sub-angular to sub-rounded with clasts commonly 5-10 cm in size, rarely up to 25 cm. The percentage of matrix can be quite high such that some outcrops are only just clast supported.

5.4.1.2b Composition and provenance.
Four graphs (Fig. 5.36) illustrate the compositional variation present in the Villanovatulo Member. The sediments were sourced from the dip-slope of the Isili fault block (below) which is composed of Palaeozoic pelites, psammites and metavolcanics (Carmignani et al. 1987) intruded by numerous granites, and originally capped by Mesozoic dolomite of which only outliers remain today (Fig. 5.1). A diverse clast assemblage could be produced by erosion of this source and slight variations in supply systems would sample a different source area. Clast variations up-succession in the Villanovatulo Member north of Isili record the local unroofing of metamorphic basement (factory section [098027], Fig. 5.36). The basal exposure contains 95% dolomite (locality A), whereas at 7.8 m (see Fig. 5.82) the first major input of metamorphic clasts is observed (72% dolomite, 14% metamorphic Fig. 5.36). The trend continues up into the lower Is Paras Member, where clast composition is 48% dolomite, 30% metamorphic (Locality C, Fig. 5.36). Gravelstones to fine sandstones comprise angular fragments of pre-rift material, green or red clay minerals plus occasional fresh, volcanic derived crystals (feldspar and mica; north Isili) and reworked micrite caliche rip-up clasts (Isili reservoir).

5.4.1.2c Sedimentary structures and facies architecture
Disorganised clast-supported conglomerates (Lithofacies 1b) and inversely graded conglomerates (Lithofacies 1c) outcrop as sharp-based layers from 0.5 m to 1.5 m thick. They commonly truncate gravelstone and sandstone beds. Gravelstones and sandstones occur as massive, metre-thick horizons interbedded with thin pebble layers, cross-bedded channel fill, and as thin parallel laminated beds (Fig. 5.32). Finer, reddened sandstones and siltstones are in places disrupted by centimetre-scale nodular caliche. The small exposures do not allow a detailed description of facies architecture, though in general, laterally extensive conglomerate horizons truncate and are interbedded with lenses of finer gravelstone and sandstones.
Figure 5.37 illustrates palaeocurrent directions in the Villanovatulo Member from Isili reservoir. Rose diagrams A and B consist of data measured from clast imbrications and do not give a clear indication of the palaeocurrent direction. Rose B may indicate southwestward palaeoflow. Palaeocurrents are often difficult to obtain in sediments dominated by debris flow deposits (Collinson 1986). If the debris flow is concentrated and sufficiently cohesive, a non-turbulent or viscous plug flow will develop (Leeder 1982; Miall 1996) thus prohibiting the formation of systematic imbrication which occurs by grain-grain contact (Miall 1996). Rose diagram C consists of data measured from cross-bedding and clearly shows a northwards directed palaeocurrent direction (Fig. 5.37) implying that drainage was from the dip-slope of the Isili Block into the half-graben. North of Isili town (factory section), the Villanovatulo Member exposure has rare, southerly palaeocurrent indicators reflecting the local fault-created topography.

5.4.1.2d Interpreted depositional environment

The sediments appear to have been deposited in a system where small, cohesionless debris flows alternated with waning sheetflow, stream channels and palaeosols. These processes are indicative of deposition on the mid-parts of alluvial fans (e.g. Hooke 1967; Miall 1996) in a semi-arid climate. The sediments were locally sourced from fault-created topography (Fig. 5.47 below).

5.4.1.3 Allai area

South and west of Allai village [865234], the Villanovatulo Member is exposed by a series of road cuttings on the dip-slope of the Grighini fault block. The dip-slope was cross-cut by smaller NNW-SSE to NNE-SSW trending normal faults (Figs. 5.6, 5.11) thought to be at the tip-line of the Busachi fault on the main Sardinian Rift trend (Fig. 5.1). The sediments, at least 50 m thick, are striking because of their intense red colour and the dominance of very coarse, poorly sorted boulder-pebble breccias and conglomerates (Lithofacies la, lb, minor Lithofacies ld, if) outcropping very close to (10 m) or on top of pre-rift basement rocks (Fig. 5.11). Towards its top and distally to the north, the Villanovatulo Member becomes intercalated with volcanic flows of the Araxigi Formation (e.g. at [875237])

5.4.1.3a Grain size, shape and sorting.

Breccias and conglomerates (Lithofacies la, lb, ld) are very poorly sorted with larger clasts commonly between 0.5 and 1m in size (Fig. 5.38). Clasts range from angular in the breccia which resembles a tightly packed scree, to moderate - sub-rounded in the conglomerates. The proximal, coarsest conglomerates have, in general, a very small percentage of matrix, whereas distal and finer microconglomerates are in places almost matrix supported. Gravelstones to fine sandstones (Lithofacies lf) are poorly to moderately sorted and comprise angular clasts. There is a notable lateral
decrease in clast sizes from sediments in proximity to or directly overlying the basement [865234, average size ~20 cm] to the east [884227, average size ~2 cm] and north.

5.4.1.3b Composition and provenance.
Monomict breccia derived from and directly overlying faulted metamorphic basement passes stratigraphically upwards and distally into sediments of a more diverse composition sourced from a wider area (Fig. 5.38). Pre-rift clasts sourced locally from the Grighini block are dominant. In the distal locality [856244] which is stratigraphically higher in the succession, Oligo-Miocene volcanic clasts are also present.

5.4.1.3c Sedimentary structures and facies architecture.
Coarse breccias and conglomerates most often exhibit a chaotic clast fabric and occur as massive beds up to 6m thick (Fig. 5.27). Gravelstone and coarse sandstone beds occur within wide depressions (possibly 10'sm wide channels) over faulted basement highs and occasionally contain poorly defined, metre-scale troughs. Coarse to fine sandstones also occur as metre-scale lenses or channels within massive conglomerates. Close to the pre-rift basement, coarse breccias and conglomerates form the majority of outcrops, whereas laterally, on a scale of 2.5 km, gravelstones, sandstones and finer, thinner bedded conglomerates represent distal deposition. Although tilted by ~10° to the NE, the sediments in this area also record a depositional slope from proximal to distal outcrops of 2.3 degrees over 2.5 km. Such a slope is equivalent to that observed on other alluvial fans (1.5-25°, Blair and MacPherson 1994). Rare clast imbrication implies that palaeocurrents were to the east and north-east in proximal areas near to the topographically high sediment supply point and to the south-east (i.e. axial, oriented parallel to the Grighini fault block) in distal localities (Fig. 5.1, 5.38).

5.4.1.3d Interpreted depositional environment
A dominance of lithofacies indicative of gravity-dominated depositional processes, such as cohesionless debris flows in proximal areas and debris flows to braided stream deposits in distal areas, suggest that the Villanovatulo Member in the Allai area represents a small alluvial fan deposit (Fig. 5.38). The fan was dominated by debris flow processes (after Miall 1996) and sourced from the topographically high Grighini fault block.

5.4.1.4 Other Areas
Small outcrops of the Villanovatulo Member are exposed in the immediate hangingwall of fault scarps at Asuni [964142], west Gergei [062958] and west Isili [079996]. South of Serri [129935], the sediments outcrop directly over an irregular basement topography (section 5.4.1). The sediments are poorly sorted, often red continental breccias containing locally derived angular clasts arranged in a
chaotic fabric (Lithofacies 1a). They represent gravity fall (talus) to subaerial debris flow deposits sourced from a degrading fault scarp topography.

5.4.1.5 Villanovatulo Member - summary
The Villanovatulo Member consists of polymict continental conglomerates, sandstones and palaeosols deposited in alluvial fan to braided fluvial systems. No fossils are found within the rocks, which are often red in colour. The sediments show considerable lateral variability in palaeocurrents, clast size, composition and sorting dependent on the local fault defined topography and pre-rift lithologies. The alluvial fan conglomerates were sourced from local topographic highs such as fault-block dip-slopes and adjacent to fault scarps. In the Villanovatulo area, a large volume of conglomerates accumulated in a topographic low caused by intersecting fault trends and possibly enhanced by fluvial erosion (Fig. 5.1, E-E').

5.4.2 Nureci Formation, Duidduru Member (early Aquitanian-late Burdigalian)
The Duidduru Member comprises marginal to shallow marine clastic sediments sourced from the erosion of pre-rift basement and Oligo-Miocene volcanic rocks (Araxigi Formation) together with the eruption of contemporaneous volcanic rocks. The sedimentary succession records shallow marine transgression over the faulted topography of the basin margin, resulting in diachronous timing of the Duidduru Member across the basin margin and along strike (Fig. 3.12, 3.13). Considerable spatial variability in sediment thickness and lithofacies architecture occurs over the Sarcidano sub-basin and makes it difficult to define a widely applicable type section, though the area in front of the Genoni fault block [990055-922049] exhibits many typical characteristics (Table 3.6). The sediment variability makes it necessary to subdivide the Sarcidano sub-basin into three areas to describe the Duidduru Member fully.

Previous work has described the fluvio-deltaic to littoral sediments known locally as the ‘Arenarie di Gesturi’ as homogeneous sandstones, debris flow megabreccias, conglomerates, cross-bedded sandstones and silty fine sandstones towards the basin (Cherchi and Montadert 1982a; Cherchi 1985). The sediments, with up to 95% quartz and 10% limey matrix, contain local accumulations of marine fauna, tool marks, tool casts and the lower levels were thought to have been deposited in a ‘palaeocanyon’ near the Isili fault block (Cherchi 1985). Around Genoni, Leone et al. (1984) describe conglomerates and tuffaceous sandstones with oysters, echinoids and Turritellid gastropods deposited by a deltaic system.
5.4.2.1 Lithofacies

The sediments of the Nureci Formation are subdivided into seven lithofacies, summarised in Table 5.6

- **Lithofacies 2a** consists of moderately to well sorted, disorganised, clast supported conglomerates containing whole and reworked *Ostrea sp.*, *Pecten sp.* and other bivalve shells. Postma (1990) describes fan delta systems sourced by alluvial fans where, on the delta plain, cohesionless debris flows and sheetflows (unconfined stream flow) fed by the fan would represent delta fan lobes. Such a depositional environment appears appropriate to Lithofacies 2a of the Nureci Formation.

- **Lithofacies 2b** comprises channel filling clast-supported conglomerates with whole and reworked, broken *Ostrea*, *Pecten* plus rare imbrication and cross bedding. Similar facies have been described from the Neogene fan deltas of southern Spain where they can be interpreted as contained debris flows in submarine or fan delta channels (Dabrio 1990).

- **Lithofacies 2c** are thin, well sorted conglomerate layers, often with imbrication and surrounded by massive gravelstones and sandstones of Lithofacies 2d. Dabrio (1990) interprets similar facies as a wave produced lag after fan-derived conglomerates or reworking of shallow marine shoreface sediments.

- **Lithofacies 2d** consists of massive bedded, matrix supported conglomerates, gravelstones and coarse sandstones. occasionally with slumps, convolute bedding and water escape structures. These sediments are interpreted to be the products of subaqueous, cohesive debris flows (Nemec and Steel 1984; Dabrio 1990; Postma 1990, coarser facies) to rapid suspension fallout (finer facies, Postma 1990) on the delta plain to delta front (Postma 1990).

- **Lithofacies 2e** comprises often complex, stacked arrangements of high to low angle, cross-bedded and channelised sandstones with a rare marine fauna. Such deposits are indicative of shallow marine foreshore-shoreface deposition with reworking by tidal and wave currents (Johnson and Baldwin 1986). Dabrio (1990) also outlines similar lower foreshore sedimentation in association with fan delta systems.

- **Lithofacies 2f** comprises 1-40 cm thick, sharp based, parallel laminated and/or normally graded sandstone to siltstones which are interpreted as the products of dilute density (turbidity) currents to suspension fallout and hemipelagic sedimentation on the prodelta (after Postma 1990; Dabrio 1990)

- **Lithofacies 2g** consists of parallel-bedded or graded tuffs and ashes, plus andesitic breccias. These deposits are the products of the rapid re-deposition of volcanic rocks (i.e. epiclastic) as dilute turbidity flows or debris flows, local extrusive eruptions and pyroclastic flows and fallout.
5.4.2.2 Around the Isili fault block

Significant outcrops of the Duidduru Member are exposed in three structural positions around the Isili fault block. In the half-graben zone between the emergent fault block and basin margin [Isili reservoir, 8030041], up to 20m of disorganised and organised conglomerates and massive sandstones (Lithofacies 2a, 2b, 2d; Table 5.6) conformably overlie the Villanovatulo Member. To the northwest and west of the Isili fault block up to 100m of matrix-supported gravel-pebble conglomerates (Lithofacies 2d) and interbedded sandstones-conglomerates (Lithofacies 2c-2e) were deposited in the hangingwall of the cross-fault [058007], whilst cross-bedded gravelstones (Lithofacies 2e) were dispersed along the block front [054973]. To the north of the block, towards the basin margin in the vicinity of the Lacuni fault hangingwall [087047], small exposures consist of <10m of massive conglomerates and sandstones (Lithofacies 2a, 2d). Thin (<5m thick) successions of the Nureci Formation are present over wide areas around the Isili block (P. Mannu [115020], Crastu [107977], S. Isili [085997]) often between the pre-rift basement and overlying carbonate (Fig. 5.83 below).

5.4.2.2a Grain size, shape and sorting

The majority of conglomerates (Lithofacies 2a, 2b, 2c, 2d) are moderately to well sorted with sub-rounded to rounded, moderate to high sphericity clasts (Fig. 5.39) with clast size averages from a few to twenty centimetres. However, the coarsest boulder-pebble conglomerate beds are poorly sorted and have large angular clasts up to 1.3m. Massive bedded gravelstones and sandstones (2d) are angular-sub-rounded and are moderately to well sorted (Fig. 5.40). Cross-bedded sandstones (2e) are well sorted with sub angular to sub-rounded clasts.

5.4.2.2b Composition, provenance and fossil content

Counts of conglomerate clast composition from Isili reservoir and west of the Isili block (Fig. 5.41) show a wide compositional range, sourced primarily from the pre-rift metamorphic, granitoid and dolomite lithologies of the Isili block, but with a variable percentage of reworked intraclasts. Soft micrite rip-up clasts which must have been derived very locally from caliche horizons make up ~50% of the clasts in one exceptional outcrop [082004]. Whole in situ encrusting and reworked oysters are common in the basal outcrop of this formation at Isili reservoir, though are not present throughout. Towards the top of the section, broken oysters, Pecten, other bivalve and bryozoan pieces are found within some conglomerate beds indicating that a marine environment was present periodically, with marine influence increasing upwards. East of Nurallao [087047], clast counts up section record the inverse basement stratigraphy from dolomite to an increasing metamorphic component due to erosional unroofing (Fig. 5.42).

Gravelstones and coarse sandstones consist primarily of fresh and altered metamorphic, granitic and dolomitic clasts, metamorphic quartz, quartz and feldspar with occasional muddy intraclasts, often
surrounded by a fine brown-green matrix of clay minerals (Fig. 5.40). XRD analysis has identified montmorillonite which is a common alteration product of volcanic rocks such as tuffs (Deer, Howie and Zussman 1992; Appendix 5B). Thin sections confirm that the majority of clasts were derived from the pre-rift basement (altered metamorphic clasts, metamorphic quartz, altered potassium feldspar, abraded albitic plagioclase, micas), though euhedral andesine plagioclase could be derived from the reworking of Oligo-Miocene volcanics (Fig. 5.40). The gravelstones and sandstones have a high primary porosity (up to 40%) which is partially occluded by pore filling clay minerals and/or a patchy calcite cement. In outcrop, patchy calcite cementation gives rise to irregular and ellipsoidal well indurated patches (Fig. 5.43). Gravelstones and sandstones generally contain rare and scattered broken shell (?bivalve) fragments, foraminifera pieces and carbonised wood pieces. However a few localities contain well preserved, shallow marine fossil fauna. For example the basal outcrop at [057995], west of the Isili Block, contains Pecten, oyster, bivalve, coral, bryozoa and Turritellid gastropod fragments plus a sharks tooth. North of the Isili block at [098028], whole in situ and broken Clypeaster sp. echinoids colonise one level of a tuffaceous sandstone bed (Fig. 5.82 below). Along the front of the Isili block, cross-bedded coarse sandstones and gravelstones are clean arkoses with open porosity and very little matrix (Fig. 5.44).

5.4.2.2c Sedimentary structures and facies architecture

Around Isili reservoir, laterally extensive, sharp-based and channelised clast-supported conglomerates grade into or are interbedded with massive, low-angle cross-stratified or channelised brown/green sandstones (Fig. 5.45). Fine red siltstone or black mudstone interbeds occur occasionally as small lenses less than 20 cm in thickness. Coarse (>10 cm clasts), laterally extensive conglomerates (Lithofacies 2a), probably deposited by cohesionless debris flows and sheetflows, are rarely imbricated, and measured clast imbrications (Fig. 5.46ab) do not provide a consistent sense of palaeoflow, possibly for reasons discussed in section 5.4.1.2c. However, channelised conglomerates (Lithofacies 2b) indicate palaeoflow to the north (Fig. 5.46c), from the dip slope of the Isili Block. Thus deposition at Isili reservoir was dominated by sheet and channelised cohesionless debris flows interspersed with massive sandstones reworked by channels and into low-aspect dunes in a marginal marine setting.

To the NW of the Isili block, facies variability and variations in sediment thickness are observed from north to south. In the north at [965025] and in the vicinity of the Laconi fault [087047], 0.5m scale bedded, clast-supported conglomerates, conglomerate layers, massive, channelised and cross bedded brown/green sandstones (Fig. 5.43), up to 40m thick, represent deposition by debris flows interspersed with times of wave and shallow marine shoreface reworking. These sediments pass south and basinward into at least 100m thick, massive, matrix-supported pebble-gravel conglomerates and gravelstones which are areally extensive for 9 km² adjacent to the fault block side. Outcrops of these
massive conglomerates and gravelstones can be up to 30m in quarried faces near Gesturi, yet no bedding is discernible and outsize intra- and pre-rift clasts, up to 3m, float within the conglomerate. The sediments are interpreted to be deposits of amalgamated cohesive debris flows and/or rapid suspension fallout. Some variety is observed at the top of the succession where metre-scale bidirectional cross-bedding is observed in one locality [038981]. Immediately adjacent to the cross-fault scarp defining the west Isili block at [057995], the basal sediments are clast-supported conglomerates and sandstones with palaeocurrents to the south and SSW (Fig. 5.46). They are overlain by a 12m thick, disorganised boulder bed which is interpreted as gravity fall to debris flow deposition sourced locally from the adjacent cross-fault scarp. Along the front of the Isili block [056970], matrix supported conglomerates were reworked into longitudinal, cross-bedded bars by currents parallel to the Isili block fault (i.e. axial to the fault-block). Irregular, knobly calcite cementation (Fig. 5.44) may reflect the preferential cementation of burrows through the bar.

5.4.2.2d Interpreted depositional environment
The facies and facies architecture in the Isili reservoir area suggest that a punctuated shallow marine transgression occurred over the Villanovatulo Member alluvial fan to produce a low-angle fan delta system (Hjulstrom Type A shoal water profile sensu Postma 1990) sourced from the dip-slope of the emergent Isili fault block. The Isili reservoir section represents the delta plain where sheet and channelled debris flows were interspersed with marginal to shallow marine reworking (Fig. 5.47). North and NNW of the Isili fault block, fan deltas supplied clastic material southwards and shallow marine reworking of fan delta lobe conglomerates dominated deposition. To the northwest of the block, clastic sediments were transported by submarine debris flows and/or rapid suspension fallout into the hangingwall depocentre of the Isili block cross-fault (Fig. 5.47).

5.4.2.3 Around the Genoni fault block
This section sub-divides the area around the Genoni fault block into three zones. The area in ‘front’, i.e. SW of, the degraded NW-SE trending fault scarp of the Genoni block is characterised by a >22m thick fining upwards sequence of conglomerates and sandstones (Lithofacies 2a, 2c, 2d, rare 2e; Fig. 5.48) and is exposed in road cuttings roughly parallel and perpendicular to the block bounding fault (Fig. 5.8). The area to the north and west of the Genoni fault block, located in the hangingwalls of the NW-SE Asuni fault and NE-SW Genoni block cross-fault, exhibits notable lateral facies variations (Lithofacies 2a-2f) in a sedimentary sequence up to 55m thick, exposed by a series of road cuts and small quarries. The area to the north, on the footwall of the Genoni fault block is not well exposed (Lithofacies 2a, 2d) though sediments overlying the Araxigi Formation are estimated to be up to 30m thick.
5.4.2.3a Grain size, shape and sorting

Along the front of the Genoni block, the sediments immediately adjacent to the pre-rift basement are boulder-pebble clast-supported conglomerates consisting of poorly sorted, moderate to well rounded, ellipsoidal to spherical clasts up to 2m in size (Fig. 5.50). It seems unusual that such large, locally derived, disorganised conglomerates are so rounded. An explanation is that the palaeoshoreline was at this level during conglomerate deposition, as indicated by Lithophaga borings (Fig. 5.51), and that wave action may have caused clast rounding. The majority of conglomerates around the Genoni block are moderately to fairly poorly sorted, with clast sizes up to 40 cm. The matrix of the conglomerates is generally poorly sorted, often with angular fragments down to silt and mud grade. Gravelstones and coarse sandstones tend to be fairly poorly to moderately sorted, often with a significant mud or silt fraction. Lithic fragments, reworked quartz and feldspar clasts are angular to sub-rounded whilst crystals of plagioclase and biotite are euhedral and angular. Finer sandstones and siltstones are moderately to well sorted, with sub-angular to sub-rounded clasts.

5.4.2.3b Composition, provenance and fossil content

Clast counts of conglomerates along the front of the Genoni block show that in the most proximal areas, closest to the degraded fault scarp basement, the conglomerate composition is of granite and dolomite (Fig. 5.49) - the two lithologies that make up the Genoni block (Fig. 5.1). However, in stratigraphically higher conglomerates, a variable percentage of metamorphic clasts which must have been sourced from a local outcrop (now eroded away) or from outside the immediate vicinity are present. Also notable is the lack of volcanic clasts which are common in other conglomerates of the Duidduru Member to the northwest. The upper surfaces of boulders at the base of the succession and oyster pieces have Lithophaga borings (Fig. 5.51). Frey (1975) states that such rock-borers are most common in the intertidal and shallow subtidal zone. In areas of active extension such as Greece, Lithophaga borings are taken to position the palaeoshoreline. Clast counts to the northwest of the Genoni block (Fig. 5.52) record a variable percentage of pre-rift basement, volcanic and intra-clasts.

Volcanic clasts would have been eroded from the Araxigi Formation which outcrops to the north and northwest. Gravelstones and sandstones of the Duidduru member are composed of material derived from the erosion of the pre-rift basement, the erosion of volcanic rocks plus presumably from coeval volcanic eruption. Sediment composition is variable dependent on the importance of each of these sources, ranging from lithic rudites to tuffaceous sandstones. A typical gravelstone or sandstone might consist of metamorphic, granite, dolomite, quartz, altered K-feldspar and micas as a basement component and volcanic clasts, euhedral plagioclase, quartz, mica and glassy shards as a volcanic component. The green/brown matrix which gives the sediments their characteristic colour is composed, at least in part, of the clay minerals clinoptilolite and analcime (plus possible montmorillonite) characteristic of
volcanic alteration (Tucker 1991; Deer, Howie and Zussman 1992, Appendix 5B). Gravelstones and sandstones often contain broken, ribbed bivalve fragments and carbonised wood pieces. Along the front of the Genoni block ([990055], Fig. 5.53), Turritellid gastropods, Scutella echinoids, oysters, Pecten, bivalves (Leone et al. 1984) rare red algae, ?Nuculidae sp. bivalves and Arenicolites trace fossils are found. North of the Genoni Block [e.g. 018064], the uppermost, massive bedded tuffaceous sandstone-gravelstone was colonised by a rich in situ shallow marine fauna of echinoids, pectenids, several species of ribbed bivalves, large (8 cm) Turritellid gastropods and serpulid casts indicative of a pause in sediment deposition plus broken bryozoan and carbonised wood fragments (Fig. 5.53).

5.4.2.3c Sedimentary structures and facies architecture

Front Genoni Block

The road cutting along the front of the Genoni fault block occurs along the degraded fault scarp and exposes an embayment along the degraded NW-SE fault plane, caused by cross-cutting faults (Fig. 5.8). The sediments that fill this embayment range from massive-boulder conglomerates and a large conglomerate olistolith adjacent to the degrading scarp, fining laterally and stratigraphically upwards into better sorted, parallel-bedded clast- and matrix- supported conglomerates and massive and rarely cross-bedded coarse sandstones (Fig. 5.8). Thus, whilst talus fall to debris flow processes operated adjacent to the scarp, fan delta lobe material supplied by debris flows was reworked by foreshore/shoreface processes tens of metres away.

Fumaroles

The penultimate bed of the Duidduru Member along the front of the Genoni fault block consists of a massive tuffaceous sandstone cross-cut by numerous vertical pipes up to 20 cm wide and 2m high (Figs. 5.48, 5.54). These structures are post-depositional, such that the matrix of the sandstone has been altered and is better indurated. They consist of an outer white rind which is cemented by calcite and lacks brown matrix clays (Fig. 5.55), plus an inner brown portion where clasts are surrounded by a dirty, clay mineral matrix and where fossil preservation is good. The clay mineral clinoptilolite, a common alteration mineral of volcanic rocks (Tucker 1991) was identified using XRD in both the white and brown portions of the pipe (Appendix 5B). The structures are planar, aligned and utilise fractures sub-parallel to the block bounding fault. They could be water circulation features utilising joints and fractures or 'fumaroles' (after Cas and Wright 1988, not gas escape pipes). Lines of hot springs parallel to normal faults, linear fumaroles and carbonate precipitation are found in extensional areas with contemporaneous volcanic activity (e.g. Valley of Ten Thousand Smokes; Cas and Wright 1988; East African Rift J.J. Tierclin pers. comm. 1996). This water circulation may in addition cause the variable shell dissolution and reprecipitated calcite cementation visible at [989049].
NW of the Genoni fault block outcrops expose lateral facies variations. SE of Asuni [966115], channelised and sheet conglomerates with rare imbrication, plus massive tuffaceous sandstones (Lithofacies 2a, 2b, 2c; Fig. 5.56) with no marine fauna and indicative of subaerial debris flow and sheetflow deposition are present. Palaeocurrents in channelised gravelstones and conglomerates were directed SE along the trend of the basin bounding fault [969108] (Fig. 5.57). These sediments pass laterally basinwards (south to southeast) into a zone of sheet conglomerates, channelised conglomerates, conglomerate lags (lenses and layers), cross-bedded and channelised coarse sandstone gravelstones (Lithofacies 2a, 2b, 2c, 2d; [978096]; Figs. 5.57, 5.58) with a fauna of whole and broken bivalves, echinoids and brachiopods (Fig. 5.23). Such facies are indicative of conglomerates supplied by debris flows in fan delta lobes and channels plus shallow marine reworking. Palaeocurrents measured in these conglomerates indicate a southward palaeoflow (Fig. 5.27). Laterally equivalent sediments to the northwest, away from the major sediment supply point, are cross-bedded, channelised and decimetre-0.5m parallel-bedded brown sandstones. They are rich in carbonised wood with rare, Turritellid sp. gastropods (2 species) and a small, thin bivalve fauna. Such sediments may be indicative of marginal marine, tidal deposition (after Elliott 1986; Leeder 1988; Fig. 5.57). Further basinwards, finer, parallel-bedded tuffaceous sandstones and graded sandstones-siltstones (Fig. 5.59) are occasionally cut by coarse, erosively based, channelised conglomerates containing broken marine shell fragments [969072] (Lithofacies 2b, 2e, 2f; Fig. 5.57).

5.4.2.3d Interpreted depositional environment

Along the front of the Genoni fault block, the boulder conglomerates represent rock fall to debris flow deposits sourced from the degrading fault scarp. Laterally equivalent and stratigraphically higher, extensive conglomerate sheets, massive and cross-bedded gravelstones plus tuffaceous sandstones indicate shallow marine shoreface reworking of material supplied by small fan deltas. The tuffaceous nature of the sediments, lack of lithified volcanic clasts and Ar-Ar dating (section 3.2.3) indicate that the shallow marine tuffaceous succession occurred at the same time as ignimbrite deposition further towards the basin margin. Tuffaceous beds such as that containing the fumaroles probably represent material supplied by volcanic eruption and reworked very soon afterwards.

North and west of the Genoni block, lithofacies arrangements reflect the basinward change from probable alluvial fan/fan delta marginal marine sedimentation at clastic supply points in intersecting fault trends and marginal marine channels/bars away from supply points to the delta plain of a low angle fan delta system (sensu Postma 1990). The delta plain was represented by massive and channelised conglomerates which occurred as fan-delta lobes and channels (Fig. 5.57). These pass basinward into a zone of shoreface reworking indicated by conglomerates lags and bars. Further basinward, finer, graded sandstones and siltstones of the prodelta (after Postma 1990) were cut by submarine fan delta channels (Fig. 5.57). To the north of the Genoni fault block, a poorly exposed
succession is capped by a bed with in situ colonisation of parallel-bedded tuffaceous sandstones (5.28). The colonised bed indicates a shallow marine environment with limited sediment supply.

5.4.2.4 Around the Grighini fault block

In this area, the Duidduru Member is exposed by 4.5 km of new road cuttings from Mogarella - Villaurbana [872117-830127] along the front of the Grighini Block and between Mogarella and west Senis (Fig. 5.1). The outcrops along the Mogarella-Villaurbana road section expose over 150m of cross-bedded, channelised sandstones, laterally extensive clast-supported conglomerate and sandstone sheets, along with minor graded sandstones, siltstones, tuffs, lapilli tuffs and an andesitic breccia (Lithofacies 2a, 2c-2g). Outcrops of parallel-bedded tuffaceous sandstones and andesitic breccias (Lithofacies 2d, 2g) are present east and south of Fordongianus but have not been studied in any detail. The Grighini area is located 6 km north of the Monte Arci volcanic centre (Assorgia et al. 1976) and in proximity to volcanics of the Araxigi Formation. Thus the rocks of the Duidduru Member are of dominant volcanic composition, resulting from erosion of volcanic and basement topography, from local submarine extrusion of andesitic magmas and presumably also from fallout of explosive eruptions.

5.4.2.4a Grain size, shape and sorting

Conglomerates exhibit moderate to poor sorting with sub-rounded to angular clasts commonly 5-40 cm in size, but with rare outsize clasts of up to 1.5m (Fig. 5.60). Massive and cross-bedded sandstones, tuffites, ashes and siltstones are moderately to well sorted with angular to sub-rounded clasts.

5.4.2.4b Composition, provenance and fossil content

Conglomerates consist of volcanic, metamorphic and granite clasts in variable proportions (e.g. v 71%, m 26%, qz 3%, n=131 at [858119]). Coarse sandstones-siltstones have a dominantly volcanic composition. In thin section, they are composed of plagioclase, quartz and biotite crystals, rare glauconite and lithic clasts, shell, bryozoan and globeriginid foraminifera fragments in a glassy and ashy matrix (e.g. Fig. 5.61). Primary and epiclastic volcanic products occur as massive, brown, matrix-rich lapilli tuffs with altered glass and clay minerals or as grey pumice-ash-crystal (quartz, plagioclase, biotite) tuffs. Wood and plant fragments are common (Fig. 5.61).

Compositional segregation of the volcanic rich sediments often outlines cross-bedding (Fig. 5.62). Two sorts of compositional segregation occur. Pumice-ash-crystal tuffs separate into a basal dark grey layer rich in biotite and quartz crystals, often calcite cemented, and an upper light grey part rich in pumice fragments. Muddy volcanic arenites separate into a basal, light brown layer of poorly sorted basement and volcanic clasts, brown altered volcanic glass and crystals, and an upper darker brown,
better sorted layer rich in mud and wood fragments. The compositional segregation is interpreted to be the result of a drop in flow velocity, perhaps during a single tide or flow event (see below). In this case the basal white/grey layers would represent the time of greatest flow when the high energy of currents sorted and cleaned the sediments. The relative thickness of cross-bed foresets may relate to the duration of the flow competence for each grain size. Carbonised wood fragments are fairly common throughout the succession and occasionally occur as concentrated layers or lags (Fig. 5.63). Transported pieces of oysters, bryozoan, red algae, echinoids, bivalves and gastropods are particularly concentrated at the base of conglomerate or massive sandstone layers. Broken shells and rare in situ bivalves and gastropods are occasionally found within massive sandstones or channelised/cross-bedded sandstone. Trace fossils such as Ophiomorpha, Planolites and ?Skolithos are infrequently found towards the tops of massive sandstone beds, which taken together, are indicative of a shallow littoral sea (Collinson and Thomson 1982; Fig. 5.63).

5.4.2.4c Sedimentary structures and facies architecture

The eastern part of the Mogarella-Villaurbana road section is dominated by cross-bedded and channelised sandstones interbedded with laterally extensive, coarse conglomerate sheets with transported shallow marine fossils and tuffaceous sandstone sheets (Fig. 5.64). The western part of the road section is dominated by massive sandstones, parallel-laminated to graded sandstones, siltstones and ashes with minor channels and cross-beds (Figs. 5.65, 5.66). North-east of the road section, an exposure of andesitic breccia crops out [873118] and may be the source for some of the volcanic clasts observed in the road section.

Conglomerates occur as either clast or matrix-supported sheet-like layers interpreted as debris flow deposits (Fig. 5.64) or as discontinuous single pebble layers interpreted as current reworked debris flow deposits. One notable outcrop exposes a highly erosive submarine conglomerate channel within massive bedded sandstones.

Massive sandstones are commonly bedded on the 0.5-2m scale and are generally ungraded, though they may pass upwards into finer, parallel-laminated sandstones (Fig. 5.65). In the western part of the Mogarella - Villaurbana road section, contorted and slumped bedding plus occasional water escape structures indicate fast depositional rates and an unstable slope (Fig. 5.63). The massive sandstones are often interbedded with finer, parallel-bedded and graded sandstones, siltstones and ashes, some with flame structures (Fig. 5.66). These sediments are interpreted as the products of debris flows, turbidity currents and suspension fallout respectively.

Cross-bedding with sets 0.5-1m in size is very common, though sets can reach up to 2m in height (Fig. 5.64). Foresets and toesets are preserved with angles of dip from 32° on the foresets to 18° on
the toesets giving the cross-beds an asymptotic or swept out appearance. Individual beds of cross-sets show no evidence for reversed flow conditions though a few outcrops expose bidirectional cross-bedding in different beds. Some high and low-angle areas of cross-bedding show evidence of depositional cyclicity, since the thickness of compositionally different foresets are grouped or bundled (after Allen and Homewood 1984) Such features are found in tidally formed sandwaves due to variability in flow energies between spring and neap tides (Homewood and Allen 1981, Allen and Homewood 1984). Figure 5.64 illustrates the grouping of white and brown bundles. In this case brown bundles, rich in a mud matrix, may represent the lower flow energies of neaps and the white thicker bedded bundles the higher energies of spring tides, as discussed above. In addition, a few localities exhibit this cyclicity such that there are 14 beds making up one brown-white package (Fig. 5.64). This may be interpreted as the astronomic control of the tidal flow i.e. 14 day diurnal spring-neap tides (after Homewood and Allen 1981).

The cross-bedded units often have erosive and cross-cutting basal surfaces. They are truncated at their top surfaces by a further set of prograding cross-beds (Fig. 5.64). Although cross-bedding is common, it proved very difficult to measure accurately the true-dip directions of the sets because the new road cuttings provided only a two dimensional section. However, it is obvious from measurements of the 2D sections that the dominant palaeocurrents were to the NW, axial to the Grighini fault block. The few measured examples (n=8) from 3D exposures show palaeocurrents to the west and northwest.

Shallow channel structures are also abundant in outcrops along the Villaurbana-Mogarella road (Fig. 5.60). These features are commonly less than a metre in thickness, greater than 2m wide, are only slightly concave and often occur in stacked and superimposed groups (Fig. 5.60) Generally, outcrops trending NE-SW or N-S expose these structures such that they may be preferentially oriented NW-SE to E-W. They may represent tidal channels oriented approximately axial to the Grighini fault block or be structures formed due to erosion by rip-currents set up by a wave-induced circulation system (after Elliott 1986). As well as cross-bedding and channels, sigmoidal shaped dune-forms and a range of truncating and onlapping low-angle structures interpreted to be oblique or perpendicular sections through sandwaves and channels are exposed (Fig. 5.60). The sedimentary structures are similar to those observed from tidally influenced shallow marine seas (Johnson and Baldwin 1986, Allen and Homewood 1984) and represent a series of shifting, prograding sandwaves and channels with dominant NW currents, axial to the Grighini fault block.

5.4.2.4.d Interpreted depositional environment

Lithofacies architecture indicates that to the south-east of the Grighini fault block, fan deltas supplied from the erosion of local basement and volcanic rocks supplied coarse material as sheet like debris flows on delta-plain fan lobes (after Postma 1990; Dabrio 1990). Material was probably also supplied
by coeval volcanic eruption (Fig. 5.67). The sediments became reworked by tidal currents into
sandwaves and channels with dominant palaeocurrents to the northwest. To the south-west of the
Grighini fault block, where reworking into bars and channels is less common, slumped and contorted
debris flow material interbedded with the products of dilute turbidity currents and suspension fallout
represent sediments of the delta front (after Postma 1990; Dabrio 1990; Fig. 5.67).

5.4.2.5 Duidduru Member - Summary
The Duidduru Member represents a transgressive shallow marine shoreface system at the basin
margin with high, but spatially variable, sediment supply. Facies architecture implies that the
palaeoshoreline consisted in part of fan deltas localised by fault topography, transporting large
volumes of material eroded from high, fault-bounded or volcanic topography as debris flows (Figs.
5.47, 5.57). Sediment reworking was most intense in the form of tidal currents axial to fault blocks
which resulted in a system of migrating tidal bars and channels (Figs. 5.47, 5.57, 5.67). Further
basinward, finer grained turbidites and suspension fallout deposits pass laterally into the marlstones of
the Giara Group (Figs. 5.47, 5.57).

5.4.3 The Araxigi Formation (late Aquitanian-early Burdigalian)
The Araxigi Formation consists of subaerially deposited rhydacitic - dacitic volcanioclastic products
with minor epiclastic intercalations (after Assorgia et al. 1995a). Up to 300m in thickness, it
dominates the half-graben fill in the north-western part of the Sarcidano sub-basin (Fig. 5.1). This
study utilised and complemented ongoing work (Assorgia et al. 1995a, in press, Lecca et al. 1997) by
examining the nature of interbedded sediments and large scale field relationships relating to the
timing of explosive volcanism and extensional faulting (section 5.3.3.3). Table 5.7 summarises the
stratigraphic divisions and volcanological characteristics of the Araxigi Formation after Assorgia et
al. (1995a, in press).

The Araxigi Formation consists of a lower part, often white in colour and composed of poorly sorted
plagioclase, biotite, quartz, alkali feldspar phenocrysts, lithic clasts, pumice and glassy cusped shards
in a glassy, ashy matrix (Fig. 5.68). These dacitic tuffs are variably welded. Particularly well lithified
outcrops contain aligned vesicles and fiamme textures. Bedding is usually sub-horizontal and massive,
though at the base of the Allai unit metre-scale beds have low angle truncation, onlap and scour
structures (Fig. 5.69). These structures possibly represent the dilute, turbulent base surge of the
ignimbrite, though the scale is much larger than documented base surge deposits (Cas and Wright
1988; Fisher and Schminke 1984). Upper horizons of the Araxigi Formation are generally red,
welded and jointed in outcrop. In thin section, they consist of euhedral plagioclase in a linedated or
fiamme glass matrix (Fig. 5.70). Small red hematite patches overprint the glass and are responsible for
its red colour. Southeast of Asuni, the uppermost flow which overlies marginal marine sediments,
exhibits interesting gas pipes and contains large (10's cm) pieces of carbonised wood (Fig. 5.71). The relationship may imply that a pyroclastic flow traversed a coastal zone colonised by trees and that large volumes of steam escaped through the pyroclastic flow.

No obvious volcanic centre can be identified within the Sarcidano sub-basin. Volcanic rocks of the Araxigi Formation do not appear to be related to local fault movement (i.e. fissural eruptions, see section 5.3.3.3 also). The Allai unit ignimbrite exhibits trends in the size of lithic fragments and in lithofacies relations which imply a source of volcanism from the Ottana graben to the north of the Sarcidano sub-basin (Assorgia et al. 1995a; Fig. 1.3).

5.4.3.1 Inter-ignimbrite sediments

5.4.3.1a San Antonio Ruinas [924123]

Green sediments situated in the hangingwall of the San Antonio Ruinas normal fault (Fig. 5.1, Enclosure 1) are matrix-supported volcaniclastic conglomerates, gravelstones, sandstones and siltstones with a minor quartz and lithic basement component. In thin section they consist of poorly sorted and angular plagioclase with minor clinopyroxene, hornblende, quartz, oxide crystals and rare metamorphic clasts in an ash and glassy matrix. Assorgia et al. (in press) name them the terrigeneous-tuffitic unit (TT). Clasts in the matrix-supported conglomerates reach up to 25 cm with a composition primarily of volcanic rocks (83%) plus minor pre-rift lithologies (schist 6%, marble 2.5%) and sandstone intraclasts (8.5%, n=115). XRD analysis suggests that the matrix clay minerals are clinoptilolite and analcime, which are both considered to be alteration products of volcanic material (Tucker 1991; Appendix 5B). Thus the sediments were sourced mainly from the surrounding ignimbrites but also from basement fault scarps which were present to the north. Beds are massive or normally graded on the metre scale (Fig. 5.72). Parallel lamination and trough cross bedding occurs rarely at the top of fining up units. Erosively based units and amalgamated sandstones are occasionally present (Fig. 5.72). No fossils were found within the sediments. They are interpreted to have been deposited by subaerial debris flows and sheetflows which at a waning flow stage became dominated by streamflow deposition (sensu Nemec and Steel 1984).

5.4.3.1b South Asuni [948124]

Green sediments also occur in road cuttings to the south of Asuni at approximately the same stratigraphic level as the terrigeneous-tuffaceous unit described above. Here, they consist of moderate to well sorted, angular quartz gravelstones with a green, volcanic derived silty matrix, and poorly sorted, angular, matrix-supported conglomerates with pre-rift basement and volcanic clasts. No fossils were found. The conglomerate units are often erosively based, less than 1m in thickness and are laterally discontinuous (Fig. 5.73) They are interpreted to be debris flow deposits. Gravelstones exhibit high to low-angle cross-bedding and trough cross-bedding within indistinct channels.
Structures are eroded and superimposed by numerous re-activation surfaces (Fig. 5.73). This organisation may be indicative of braided fluvial systems where small channels, transverse and longitudinal bars, dunes and scour pits are superimposed on one another (Collinson 1986; Miall 1996). The outcrops are taken to represent deposition in a gravelly fluvial system dominated by streamflow processes, but with rare debris flows, sourced from basement fault scarps to the north and from the erosion of local volcanic material.

5.4.3.1c Southeast Asuni [965115]
Road cuttings southeast of Asuni expose ~10m of brown sandstones, gravelstones and microconglomerates which were deposited beneath the uppermost pyroclastic flow of the Araxigi Formation. The sediments comprise tens of centimetre thick, parallel-bedded, normally graded and laterally extensive volcanic sandstones which contain rare *Ostrea* sp. and shell fragments plus two types of *in situ* ?Turritellid* sp. gastropod, up to 5mm in size. The microconglomerate bed is derived from basement and volcanic lithologies. A 1.5m thick, cross-bedded gravelstone horizon suggests palaeoflow to the south-east. The sediments are interpreted to be debris flow deposits occasionally reworked into bars in a marginal marine environment.

5.4.4 Isili Formation (early/mid Burdigalian - late Burdigalian)
The Isili Formation consists of shallow marine, mixed siliciclastic-carbonate facies (Lithofacies 2a-2e, 3a-3g; Table 5.6, 5.8). It is subdivided into the Is Paras Member comprising of a lower part with siliciclastic-palaeosol-carbonate alternations (Lithofacies 3a, 3b, 3d, 2a-2e; Tables 5.6, 5.8), an upper part representative of a shallow marine carbonate platform (Lithofacies 3c-3f; Table 5.8) and partially equivalent calcarenites and calcirudites of the Serra Longa Member (Lithofacies 3b, 3g; Table 5.8). The Isili Formation crops out around the town of Isili and the lithofacies are correlated to small outcrops north of Mogarella [894150] and south of Serri [129935]. Around Isili, the Is Paras Member passes laterally into the Serra Longa Member which thins westwards to a metre or so thick near Genoni and Nureci.

5.4.4.1 Previous work
The palaeontology and sedimentology of the 'Calcare d'Isili' (Isili Formation) platform carbonates has been described by Cherchi and Montadert (1982a) and Cherchi (1985). A diverse fossil fauna including *Amphistegina* sp., *Miogypsina* sp. (e.g. *M. complanata*, *M. gunteri*, *M. bantamensis*), *Heterostegina* sp., *Lepidocyclina* sp., abundant *Lithothamnium*, bryozoans, corals, serpulids and oysters is characteristic of a *Lithothamnium-Gypsina* community deposited in water depths of less than 60m and probably 5-20m (Cherchi and Montadert 1982a, Cherchi 1985). Frequent micritization and *Lithophaga* borings also indicate that the platform was within the photic zone (Cherchi and Montadert 1982a; Cherchi 1985). Geopetal birds-eye structures are thought to denote subaerial exposure and
current agitation is thought to have occurred due to the large size of oncolites (Cherchi and Montadert 1982a, Cherchi 1985). Platform instability is indicated by slumps with a transport direction to the east. The carbonates are classified as well cemented biosparites containing large algal oncolites and frequently micritized bioclasts (Cherchi and Montadert 1982a; Cherchi 1985). The Serra Longa Member is briefly described by Leone et al. (1984) as a polygenic, 'heterometric' (i.e. poorly sorted) conglomerate in the Genoni area.

5.4.4.2 Lithofacies

Seven additional lithofacies are needed to describe the Isili Formation:

- **Lithofacies 3a** comprises of rootlet beds overlain by a loamy soil horizon (Fig. 5.74) and reddened, mottled sandstone-siltstone horizons with possible caliche filled veins. These features are characteristic of palaeosol development and plant/tree colonisation (e.g. Collinson 1986).

- **Lithofacies 3b** consists of moderately to well sorted, poorly cemented calcirudites containing pre-rift, red algae, oyster and broken shell clasts within a lime mud - patchy spar matrix. Deposition must have occurred in a zone with both clastic and carbonate sediment supply such as found at the margins of a carbonate platform, an area where alluvial fan/fan delta systems were rimmed by carbonates (e.g. Red Sea, Purser et al. 1986; Friedman 1988; Roberts and Murray 1988) or on an open platform with clastic supply and limited wave reworking (Tucker and Wright 1990).

- **Lithofacies 3c** consists of wackestones, often transitional to packstones (Dunham 1962 scheme) or sparry biomicrites (Folk's 1959 scheme) where a variety of broken bioclastic fragments, often micritized, are enclosed within a patchy lime mud and micraspar matrix, with small areas of blocky carbonate spar (Fig. 5.75). This microfacies is indicative of deposition in shallow waters with open circulation, close to wavebase and of areas where clasts from higher energy environments have moved down local slopes to lower energy settings (after Tucker and Wright 1990).

- **Lithofacies 3d** comprises packstones, grainstones and rudstones (or packed, poorly washed biosparites, Fig. 3.4) containing a variety of reworked bioclasts with rare, small patches of algal bindstone, oyster reef mounds and oyster beds. This microfacies is indicative of areas with constant wave action and with small patch reefs (Tucker and Wright 1990). Oyster build-ups are indicative of shallow waters (e.g. ~<30m, Milliman 1973).

- **Lithofacies 3e** consists of large scale (5-20m), cross-bedded wackestones, packstones, grainstones and rudstones (Fig. 5.77) with similar microfacies to 3c and 3d but reworked and deposited by the progradation of large carbonate bars.

- **Lithofacies 3f** comprises slumped wackestones of similar microfacies to 3c but re-deposited when semi-lithified by slumping and sliding (Fig. 5.78).

- **Lithofacies 3g** consists of well-cemented calcirudites with pre-rift clasts, red algae, oysters and other bioclasts in a dominantly sparry matrix (Fig. 5.79). The sediments were deposited at the
margin of an open carbonate platform in a fairly high energy environment (Tucker and Wright 1990), as a transgressive lag (section 5.4.4.5c) or debris flow (section 5.4.5.2).

5.4.4.3 Lower Is Paras Member (mid-Burdigalian)

5.4.4.3a Grain types, fossil composition and classification
The composition of conglomerates and sandstones (Lithofacies 2a-2e) within the lower Is Paras Member are similar to that of the underlying Duidduru Member in their dominant pre-rift basement component (e.g. Fig. 5.41). However, the sediments are better sorted, more rounded and of smaller average grain sizes than the Duidduru Member. Lithic clasts in calcarenites and calcirudites are dominantly pre-rift basement (e.g. pelites, metamorphic and igneous quartz, dolomite, micas) whilst the matrix ranges from lime mud to equigranular spar. Calcarenites and calcirudites were first colonised by small red algae (Fig. 5.80) and oysters and pass gradationally upwards into packstones with a full marine fauna of bryozoa, bivalves, benthic foraminifera, echinoids and barnacles. In the lower part of the Is Paras Member, a bed consisting only of broken bryozoan fragments forms a particularly useful marker horizon. Trace fossils such as Skolithos, Thalassanoides and Planolites indicative of shallow marine deposition (Reineck and Singh 1980; Collinson and Thomson 1982; Johnson and Baldwin 1986; Tucker 1991) are occasionally observed in the calcarenites and carbonates.

5.4.4.3b Sedimentary structures and facies arrangements
Northwest of the Isili fault block, conglomerates and sandstones occasionally exhibit cross-bedding or imbrication caused by northwesterly directed palaeocurrents. At Isili reservoir, north of Isili (factory section) [098028] and east of Nuragus [071007], the vertical succession of the Is Paras Member consists of cycles such that gradual colonisation of coarse clastic rocks evolved upwards into to pure shallow marine carbonates over scales from a few centimetres to a few metres (Figs. 5.19, 5.81, 5.82). Clastic sediments cover carbonate beds with a sharp contact indicative of a sudden influx of material (Figs. 5.19, 5.82). Palaeosols were developed in sandstones and calcarenites (Figs. 5.74, 5.19). In topographically higher locations, on the dip-slope of the Isili fault block, the lower Is Paras member is condensed into a succession of calcirudites and calcarenites less than 10m thick (Fig. 5.83). The lower Is Paras Member also exhibits abrupt lateral facies and thickness variations on passing from the half graben behind the Isili block over the zone of the west Isili block cross fault (Fig. 5.19 [067014-075006]). Faulting and relative sea level controlled lateral facies variations such that carbonates and palaeosols were formed on footwall highs (Logs B and C; Fig. 5.19), and wedge out laterally towards clastic/carbonate-siliciclastic rocks which were deposited in hangingwall lows (Log A; Fig. 5.19). Passing up the succession, the widespread platform carbonates of the upper Is Paras Member seal syn-depositional faults (Fig. 5.19).
South of Serri [129935], calcirudites and packstones occur over an irregular pre-rift topography and as a lateral equivalent to basinward Giara Group marlstones. These carbonates are of similar lithofacies to the Is Paras Member but are probably considerably older because of their lower topographic position (Fig. 3.13).

5.4.4.3c Interpreted depositional environment
The alternations of conglomerates, packstones and palaeosols are indicative of pulsed supply of clastic material transported through fan deltas, shallow marine reworking, periodic emergence, and gradual biogenic colonisation to form a shallow marine carbonate platform (e.g. Log C: Fig. 5.19, 5.82, 5.84). On the dip-slope of the Isili fault block and south of Serri, calcarenites and calcirudites record diachronous transgression onto the palaeotopography (Fig. 5.84).

5.4.4.4 Upper Is Paras member (late Burdigalian)
5.4.4.4a Grain types, fossil composition and classification
In the Isili area, abraded, coated and micritized bioclastic fragments are the most common grain types. Whole and in-situ organisms, some forming monospecific build-ups, plus rare peloids are also present. The types of bioclasts observed are coralline red algae (Lithothamnium sp. rhodoliths, Lithophyllum sp.), which often coat other bioclastic fragments, Ostrea sp., various bivalves including large Pecten, bryozoa, benthic foraminifera (see 5.2.4.1, Cherchi and Montadert 1982a), echinoderms, serpulid worm casts, and corals (Figs. 5.75, 5.85). Some outcrops are totally dominated by rhodoliths up to 10 cm in size (Fig. 5.86). A dominance of red-algal rhodoliths is found in other Miocene shallow marine sediments from Corsica and Turkey (Orzag-Sperber et al. 1977). Occasionally, Planolites and Skolithos trace fossils and scattered gravel sized pre-rift clasts are found. The bioclasts are surrounded by a patchy lime mud and microspar matrix with small areas of diagenetic, blocky, equigranular cement. The Is Paras Member carbonates are classified as unsorted-sorted sparry biomicrites (after Folk 1959) or sparry wackestones, packstones and rudstones to muddy grainstones with additional rare algal bindstones and oyster boundstones (after Dunham 1962). North of Mogarella [894150], an Ostrea sp. and bivalve boundstone (Fig. 5.85) outcrops in the same stratigraphic position as the Is Paras Member.

5.4.4.4b Sedimentary structures and facies arrangements
Prograding, cross-bedded, grainstone-wackestone units up to 20m in height are particularly conspicuous towards the top of the Is Paras Member north and northwest of Isili (Fig. 5.21, 5.77, 5.87). Cross-beds up to ~80-100 cm thick, dip at angles up to 26°, with a bidirectional nature in different outcrops. The base of individual cross-beds are often concentrated in large rhodoliths and other bioclastic fragments which must have been moved by the high energy event that initiated cross bedding. Some cross-beds have an asymptotic geometry of swept out tosets. In one road cutting...
[098010], reactivation surfaces dividing zones of cross bed progradation can be identified (Fig. 5.87). Palaeocurrents taken perpendicular to cross-bedding indicate that flow would have been northwest-southeast (i.e. axial to the fault block) and to the northeast [0880141 (Fig. 5.1, Enclosure 1). Slumped horizons of semi-lithified wackestones have a chaotic internal fabric, erosive basal surface and an overall wedge shaped geometry (Fig. 5.78).

Parallel-bedded wackestones-rudstones with isolated in-situ Ostrea sp. and Lithothamnium sp. build-ups dominate the Is Paras Member (Fig. 5.88). This study examined many outcrops and found no obvious systematic arrangement of microfacies. Towards the top of the section north of Isili, the carbonate platform became disrupted by localised, outcrop scale, NNE-SSW oriented normal faults (section 5.3.2). At similar stratigraphic levels but in different locations, chaotic slumps and 15-20m cross-bedded units crop out. To the south of Isili, continued sedimentation and aggradation to retrogradation of parallel-bedded wackestones and packstones up the dip-slope of the Isili fault block [080980] occurred coeval with Giara Group marlstones in the half-graben centre. (Fig. 5.1, section C-C').

5.4.4.4c Interpreted depositional environment

The production and deposition of thick, parallel-bedded wackestones, packstones and grainstones with rare algal and oyster patch reefs occurred on a stable, open, shallow marine carbonate platform, with 5-20m water depth (Cherchi & Montadert 1982a) and reasonable water movement (open platform, Tucker and Wright 1990; Fig. 5.84). Normal faulting disrupted the carbonate platform and slumps may have been generated by associated seismic events [e.g. 102013]. At the same stratigraphic level, prograding carbonate bars may have been formed by tidal currents axial to the Isili fault block (i.e. sandwaves). In the Jurassic of England, bidirectional oolite sandwaves are thought to have formed due to tidal currents and occurred as a sand belt which paralleled the shelf margin (Sellwood 1986). The Burdigalian age, shallow marine facies of the Apt-Fourcalquier basin of southern France (Jones 1988) and molasse of the Swiss Basin (Allen et al. 1985) contain sandwaves of similar or greater dimensions (30-40m) and similar characteristics, interpreted to have formed in narrow, tidal seaways. The uppermost beds of the Is Paras Member were deposited as a carbonate rim up the dip slope of the Isili fault block as relative sea level rose and whilst deeper marine marlstones were deposited in the half-graben centre. North of Mogarella, the boundstone represents an isolated patch reef which was formed on a footwall high where sedimentary supply would have been restricted.

Discussion

The Is Paras Member platform carbonates around Isili were isolated in a zone behind and to the NNW of the Isili fault block because this area was restricted from the clastic material which was transported into Isili block cross-fault hangingwall [e.g.052987]. In addition, little clastic material was eroded
from the Mesozoic dolomites exposed in fault scarps to the north or was supplied from the emergent Isili fault block dip-slope which may have been rimmed by a carbonate reef.

### 5.4.4.5 Serra Longa Member (early/mid Burdigalian - mid/late Burdigalian)

The Serra Longa Member occurs in two positions and with two modes of genesis. The maximum thickness is observed northeast and west of Isili as a lateral equivalent to the Is Paras member where there was both clastic and carbonate supply (Lithofacies 3b, 3g; Fig. 5.89). Further to the northwest, for example around the Genoni fault block, the Serra Longa Member consists of a 0.5-2m, sharp based calcirudite bed (Lithofacies 3g; Figs. 5.90, 5.91). In this region it is particularly noticeable because it is much better indurated than the surrounding rocks (Fig. 5.91) and often crops out forming a ridge in the present day topography.

#### 5.4.4.5a Grain types and fossil composition.

Rounded to angular, gravel to cobble size clasts dominantly of pre-rift and occasionally of volcanic composition (Fig. 5.49) occur in the calcirudites. Calcarenites vary from medium to coarse sand grade with angular to moderately rounded, moderately sorted clasts of pre-rift basement (Fig. 5.79). Fossils present include whole and broken red algae, oysters and bivalve shells. The matrix varies from soft lime mud (Lithofacies 3b) to sparry carbonate and calcarenite (Lithofacies 3g).

#### 5.4.4.5b Sedimentary structures and facies arrangements

Low angle cross bedding is occasionally observed within the thickest, well cemented units (3g) of the Serra Longa Member. Northwest of the Isili fault block [058016-045995], the Serra Longa Member consists of 5m of poorly cemented calcarenite (Lithofacies 3b) overlain by up to 10m of well cemented calcirudite (Lithofacies 3g). The calcirudite thins to the northwest until it forms a 0.5 to 2m thick bed which fines upwards and marks a distinct change from clastic rocks of the Duidduru Member to the marlstones and calcarenites of the Giara Group. In the Genoni region, there is some evidence that the Serra Longa Member may lie over an omission or hiatal surface at the end of a period of clastic sediment supply. For example, the uppermost bed of the Duidduru Member has, in places, an extensive in situ shallow marine fauna (Fig. 5.53). Also, in one place [003066], northwest of Genoni, the sharp and erosive boundary of the calcirudite bed with the underlying tuffaceous clastics is iron-stained. Moving further to the northwest, the distinct Serra Long Member band disappears towards Mogarella where Duidduru Member sedimentation continued until the late Burdigalian (Fig. 5.1, 3.12)

#### 5.4.4.5c Interpreted depositional environment

Adjacent to the Is Paras Member carbonate platform, the sediments of the Serra Longa Member were deposited by reworking of clastic and carbonate material in a shallow marine setting.
In other ancient successions, erosion surfaces and disconformities overlain by a thin sheet of extensively bioturbated sandstones or surf-winnowed gravel lag reflect the passage of non-depositional transgressions across the area, involving the landward migration of the shoreline under conditions of moderately rapid sea level rise (Elliott 1986). Transgressive sequences with greater amounts of deposition record the passage of the shoreface or wavebase over an unconformity with a conglomerate lag (Elliott 1986). In the Genoni area, the Serra Longa Member is thought to represent such a marine transgressive lag where sediment supply was limited. New temporal constraints from this area (chapter 3) combined with lithostratigraphy imply such a transgression associated with Serra Longa member deposition can be dated at ~19Ma (early/mid Burdigalian).

5.4.5 Giara Group (Burdigalian-Langhian)
At the basin margin, the Giara Group consists of up to 275m of parallel-bedded, planktonic foraminifera-rich marlstones and fine calcarenites with rare calcirudite debris flows, marl and calcarenite turbidites (Lithofacies 3g, 4a-4c; Table 5.8).

5.4.5.1 Previous Work
Cherchi (1985) describes the Tuili-Giara di Gesturi succession, 6 km to the south of the Genoni fault block, as consisting of marlstones with rare arenaceous and tuffaceous horizons capped by calcarenites. Cherchi and Montadert (1982ab) analysed Aquitanian benthic and planktonic foraminifera from marlstones to estimate the palaeobathymetry at the foot of rift-margin faults as 1000-1300m, compared to 200-300m in unspecified ‘other places’ at the same time. In the vicinity of Genoni, Leone et al. (1984) describe yellow, well cemented sandstones above the calcirudite (Serra Longa Member) which pass gradually upwards into marlstones.

5.4.5.2 Lithofacies
Three additional lithofacies describe the Giara Group sediments.

- **Lithofacies 4a** consists of well cemented, well sorted, fine calcarenites composed of pre-rift and volcanic clasts, planktonic and benthic foraminifera, calcareous nanofossils, broken red algae and shell fragments, whole bivalves, brachiopods (?Terabratulid sp.) and echinoids (Schizaster sp., Clypeaster sp.) within a lime mud, clay or fine spar matrix (Fig. 5.93). Thalassanoides, Planolites, Ophiomorpha and Rhizocorallium trace fossils are also common (Fig. 5.92). Such hemipelagic sediments are indicative of deposition on an oxygenated marine shelf, probably below wavebase, with limited clastic sediment supply and firmground colonisation (after Coniglio and Dix 1992).

- **Lithofacies 4b** are marlstones composed of planktonic foraminifera, calcareous nanofossils, clay and lime mud with very rare silt to fine sand size quartz clasts (Figs. 5.94, 5.95). Hemipelagic to pelagic sediments such as this characteristic of an offshore marine environment, below wavebase,
with very little clastic sediment supply (after Jenkyns 1986) such as in an ocean basin adjacent to a carbonate platform (Sellwood 1986; Tucker 1991, Coniglio and Dix 1992).

- **Lithofacies 4c** are marlstone and calcarenitic turbidites with a sharp, erosional base, normal grading to parallel lamination and background sedimentation (Fig. 5.94; Bouma T$_{ak}$). Marlstone turbidites contain centimetre-sized marlstone rip-up clasts. These sediments are indicative of gravity flow deposits supplied locally downslope (after Congilio and Dix 1992).

The Giara Group also includes clastic conglomerates and calcirudites near the basin margin with similar textural and compositional characteristics to Lithofacies 2a and 3g. However, the 0.5-2m scale beds are sharp-based, normally graded, interbedded with Lithofacies 4a, 4b, and are thought result debris flows supplied from the basin margin.

**5.4.5.3 Facies Arrangements**

The vertical succession at the basin margin is such that basal interbedded calcarenites and marlstones (Fig. 5.94, 5.96) with rare calcirudite debris flows (front Isili Block at [053962] and [112933]) and turbidites pass upwards into a thick, monotonous succession of marlstones capped by calcirudite, calcarenite and marlstones alternations (e.g. Giara di Gesturi -950000, Serri-Nurri road [149951]). Further basinward the succession is dominated by marlstones (e.g. southwest of Isili fault block) and by tuffaceous marlstones with pillow lavas and hyaloclastites (Marmilla region south of the Grighini fault block; Cherchi 1985).

**5.4.5.4 Interpreted depositional environment**

The rocks of the Giara Group represent a deeper marine environment with water depths estimated from tens of metres to 1000m (Cherchi and Montadert 1982ab) and a greatly decreased coarse clastic supply. With initial deepening, hemipelagic sediments were intercalated with material transported by gravity flows supplied from shallower environments at the basin margin. Increased water depths are evident from the proceeding, thick pelagic marlstone succession (Fig. 5.96). Finally, the uppermost Giara Group calcarenites record a renewed shallowing of the marine environment.

**5.4.6 Summary and relationships between basin filling units**

Initial continental clastic sedimentation of the Villanovatulo Member was conformably followed by deposition of the shallow marine clastic sediments of the Duidduru Member with transgression onto the Sarcidano sub-basin margin. This clastic Nureci Formation was transported from eroding, fault-created and volcanic topographic highs by alluvial and debris flow processes. In marginal and shallow marine units, sediment reworking by axial tidal currents occurred. The Araxigi Formation volcanic succession was partly contemporaneous with this clastic sedimentation. The volcanic rocks were deposited from voluminous pyroclastic flows (Assorgia et al. 1995a) and interbedded with non-marine
to marginal marine sediments dominated by debris flow and fluvial processes. Concomitant with renewed transgression in the early/mid Burdigalian, further facies variability developed such that a mixed carbonate (Isili Formation) and clastic (Duidduru Member) system existed around the Isili fault block whilst marlstones of the Giara Group were deposited in distal areas. The Isili Formation comprises shallow marine siliciclastic sediments and palaeosols intercalated with dominant platform carbonates. The carbonates were deposited in a shallow sea and sometimes reworked into large scale cross-bedded bars by marine currents. From the mid-late Burdigalian, clastic sedimentation (Duidduru Member) ended with further transgression. Carbonate-marlstone deposition of the upper Is Paras member and Giara Group continued, filling the remaining accommodation space in response to gradual relative sea level rise.

The basin filling succession of the Sarcidano sub-basin is thus both vertically and laterally variable in terms of facies, composition, depositional environments and thickness. Four composite logs illustrate how the basin filling succession changes from northwest to southeast across the study area (Fig. 5.97). The basin filling sequence is predominantly conformable with some minor pauses in sedimentation (e.g. top Duidduru Member-Serra Longa Member) and reaches a maximum thickness of around 300m at the Sarcidano rift-margin. The controls on basin filling are discussed in chapter 11.

5.5 Timing of extension

5.5.1. Previous Work

Cherchi and Montadert (1982ab) appear to use sedimentary geometries from the Sarcidano sub-basin to constrain the timing of the Sardinian rifting phase from mid Oligocene (~30 Ma) to 23-24 Ma (mid-late Aquitanian on their timescale of Vail and Hardenbohl 1979). Their criteria for syn-rift deposits are '1) the typical arrangements of Aquitanian sediments in half-graben 2) the vertical offset of hundreds of metres of shallow marine lower Aquitanian limestones along the master fault'. They identify the end of rifting due to 'the sealing of faulted blocks by Aquitanian sediments'. Several comments must be made here. Firstly, as discussed in chapter 3, the dating of the sediments by A. Cherchi in this area may be in error such that they are in fact younger than the Aquitanian N4 zone. Secondly, it is unclear what sorts of arrangements in half graben are proposed, though Cherchi (1985) mentions dip-variations as discussed above (section 5.3.3.4). Thirdly, this study did not observe 100's metres of offset of lower Aquitanian or any age limestones along a fault, but found that limestones were deposited at different topographic levels, probably at different times (section 3.3.2, Appendix 3D). Cherchi and Montadert (1982ab) and Cherchi (1985) imply that extension occurred continuously from 30-23 Ma and that the sediments here classified as the Nureci and Isili Formations formed a syn-rift sequence with large-scale, divergent sediment wedges. Assorgia et al. (1995a) favour a different hypothesis for the evolution of the Sarcidano and Ottana sub-basins (Fig. 1.3) such that the 'Ussana Formation' (i.e. Villanovatulo Member) conglomerates and lower parts of the Araxigi Formation were
deposited over Palaeozoic 'troughs and shoulders' yet in an unspecified 'wide transtensional left lateral regime'. Eruption of upper parts of the Araxigi Formation were triggered by extensional events post-20 Ma resulting in a 'horst-graben' system and 'increasing rifting' occurred after the marine transgression in Aquitanian-Burdigalian times (Assorgia et al. 1995a). No evidence is presented to support this hypothesis which essentially seems to suggest only post-early Burdigalian extension.

5.5.2 Timing of extension in the Sarcidano sub-basin

The syn- and post-depositional geometries, palaeocurrents and dispersal paths of the basin fill of the Sarcidano sub-basin, constrained by new biostratigraphic, radiometric and Sr isotope dates can be used to deduce the timing of extension.

First phase of extension. Extension initially occurred on ?late Rupelian - ?mid Chattian (sometime between 30-26 Ma) NW-SE trending normal faults with throws of hundreds of metres and on NE-SW/NNE-SSW faults at the sub-basin margin, largely before deposition of the Villanovatulo Member. The timing of this extension phase is poorly constrained since no dates exist for the continental Villanovatulo Member in this region.

Second phase of extension. The main phase of NE-SW trending cross-faulting cut, and therefore post-dated the NW-SE faulting, and occurred before the deposition of the Villanovatulo Member at Villanovatulo, the Duidduru Member at Isili and Genoni and the Araxigi Formation around the Grighini fault block. The ?late Chattian initiation of cross-faulting is thus poorly constrained but the majority of fault displacement had accumulated by the early Aquitanian.

Minor re-activation of rift-margin and cross-faults seems to have occurred from time to time through the deposition of the basin fill sequence until the mid-late Burdigalian (Giara Group). For example, the Isili block cross-fault was active around the mid Burdigalian (~18.5 Ma, lower Is Paras Formation Fig. 5.19).

Third phase of extension. Small, distributed, outcrop scale NNE-SSW to N-S trending normal faults were active at times from the mid-late Burdigalian (~17-18.5 Ma) within the basin fill sequence (Duidduru, Is Paras Members and basal Giara Group)

Reactivation of the NW-SE trending faults occurred and new N-S oriented faults became active sometime after the Langhian, probably in the Pliocene, resulted in the gentle folding and tilting of the entire succession.
5.5.3 Amount of syn and post-rift basin filling.

In the Sarcidano sub-basin, stratal geometries indicative of syn-depositional normal faulting are observed only rarely and are of outcrop scale (Table 5.5) even within the syn-rift megasequence (section 5.3.36). Significantly, the exposed basin fill does not contain large-scale, divergent, rotated bedding with major angular unconformities, which would be indicative of prolonged movement on large half-graben bounding faults (as suggested in Cherchi and Montadert 1982ab). It is possible syn-rift deposits may be present, buried beneath the level of present day exposure, but these will be volumetrically minor. However, the basin fill generally consists of parallel-bedded units of varying lithofacies (post-rift deposits; Figs. 5.1, 5.3) which infill earlier fault-formed topography. The implication is that for both fault sets, the majority of extensional fault activity which formed the block-and-graben topography mainly occurred before basin filling (i.e. the rate of tectonic subsidence far exceeded sedimentation rates during the phase of active extension), that the accommodation space created by faulting was initially underfilled, and that the extensional phases must have been short-lived (<few Ma).

5.6 Kinematic evolution of the Sarcidano sub-basin

In isotropic rocks, principal stress directions responsible for faulting can be deduced from the orientation of fault planes using simple models of brittle failure (e.g. Anderson 1951). However most faults occur in anisotropic rocks and fault-slip data must be used to define movement on fault planes that may have been oriented at varying angles to the principal stresses. In the Sarcidano sub-basin, rare slickenside data and fault offsets indicate that the majority of extension occurred on dip-slip faults. Therefore in the simplest terms one can assume that the extension direction (or minimum stress \( \sigma_3 \), horizontal) was oriented perpendicular to the trend of fault planes similar to the model of Anderson (1951, Fig. 5.98). The lack of fault slip data means that a thorough analysis of Sarcidano sub-basin kinematics is not possible. However, broadly speaking the three identified phases of extension can be related to changes in the overall stress field.

This study identified a set of ?Oligocene NW-SE trending (present day orientation) normal faults which underwent dip-slip extension with a style similar to the ‘tilted dominoes’ of Jackson et al. (1988). According to the simple model, the extension direction associated with these faults would have been NE-SW. The second set of ?latest Oligocene-early Aquitanian cross-faults oriented NE-SW to NNE-SSW would have been the result of NW-SE to WNW-ESE directed extension. Syn-sedimentary Burdigalian normal faults trend NNE-SSW to N-S implying an ENE-WSW to E-W extension direction. The possible causes and implications of these differing stress fields are discussed in the regional context, within the framework or Sardinia-Corsica microplate separation and rotation (Chapter 10).
5.7 Tectono-stratigraphic development

5.7.1 Late Rupelian - mid Chattian (Oligocene; Fig. 5.99a): initial extension

Initial extension in the Sarcidano sub-basin occurred in a continental environment on NW-SE trending, segmented normal faults and on faults with NNE-SSW/NE-SW orientations at the sub-basin margins. The newly formed tilted-block and half-graben topography controlled clastic sediment provenance, dispersal and accumulation of the basal Villanovatulo Member. Volumetrically small, coarse talus and alluvial fan deposits were shed off local fault-bounded footwall highs and fault block dip slopes, leaving the majority of the accommodation space underfilled. Only the lowest exposed continental clastic rocks exhibit geometries directly related to syn-depositional faulting, implying initial rates of tectonic subsidence far exceeded sedimentation rates and subsequent underfilling of fault-created topography.

5.7.2 Late Chattian - early/mid Burdigalian (Fig. 5.99b): Second extension phase, transgression and sedimentary response to fault topography

A pre-early Aquitanian cross-faulting phase resulted in the segmentation of the basin margin into fault blocks separated by a series of quasi-independent depocentres. Geometries observed in the subsequently deposited Araxigi Formation and Duidduru Member are those of onlap onto and over degraded fault scarps and up the dip-slopes of the fault blocks. In the north-west of the Sarcidano sub-basin, basin filling was controlled by a voluminous supply of ignimbritic volcanoclastics (Araxigi Formation) which effectively blanketed and rapidly filled the accommodation space created by faulting. In topographically lower areas to the southeast, the complex basin-margin structure between the emergent fault blocks and basin margin controlled the supply and accumulation of shallow marine clastic facies (Duidduru Member) concomitant with marine transgression. Material was supplied from the erosion of fault scarps and topographic highs and transported basinward into the underfilled, fault-derived topography. For example, the variable topographic gradients around the fault blocks led to the coarsest and thickest clastic sediments being supplied to, and accumulating in, the depressed areas associated with the hangingwalls of the NW-SE and NE-SW intersecting faults. Such areas were supplied by fan deltas sourced from the degradation of fault scarps plus volcanic material in proximity to volcanic centres. Laterally, the sediments were reworked by shallow marine processes and tidal currents focused axial to fault blocks. Basinward, finer grained turbidites and suspension fallout were cut by conglomerate-filled submarine channels which were deposited as the coarse clastic material accumulated at the rift basin margin.
5.7.3 Mid - late Burdigalian (Fig. 5.99c): Reduced tectonism and continued filling of fault topography with rising sea level.

Renewed transgression in the early-mid Burdigalian (~19Ma) resulted in intense lateral facies variations at the Sarcidano sub-basin margin. Clastic material continued to be supplied into the Isili block cross-fault depocentre (Duidduru Member), whilst a largely parallel-bedded, infilling succession of mixed carbonate-siliciclastic, disrupted by localised NNE-SSW outcrop scale normal faulting, was deposited behind the Isili fault block (lower Is Paras Member). The aggradational-retrogradational upper Is Paras Member platform carbonates, again disrupted by localised, outcrop scale NNE-SSW syn-sedimentary faulting occurred contemporaneous with continued transgression. Adjacent to the Is Paras carbonate platform, calcarenites and calcirudites of the Serra Longa Member passed laterally into calcarenites and marlstones of the Giara Group. Around the Genoni fault block, the Serra Longa Member was deposited as a transgressive lag which marked the change from clastic sedimentation to drowning of sediment source areas and Giara Group marlstone sedimentation. Duidduru Member volcanic-rich sediments supplied from local, topographically high source areas (Araxigi Formation, Monte Arci) continued to be deposited along the front of the Grighini block, with axial tidal dispersal of material transported by fan deltas.

5.7.4 Late Burdigalian - Langhian (Fig. 5.99d): Post-rift deposition

The parallel-bedded, planktonic foraminifera-rich marlstones of the Giara Group which fill and onlap further onto the basin margin palaeotopography, can largely be considered as a post-rift megasequence. Continued subsidence/relative sea level rise resulted in a progressive deepening of the shallow and marginal marine environments. Drowning of clastic sediment source areas and of the Is Paras Member carbonate platform occurred. In proximity to the basin margin, debris flows and turbidites were deposited.
5.8 Summary

- The structure of the Sarcidano basin margin was defined by three phases of normal faulting; 1) extension on large NW-SE (~130-150°) trending normal faults and NE-SW/NNE-SSW faults at the sub-basin margin during the ?late Rupelian-?Mid Chattian (30-26 Ma), 2) extension on large NE-SW to NNE-SSW (~020-050°) normal faults sometime within the ?late Chattian -early Aquitanian (25-23 Ma), 3) small, widely distributed outcrop scale NNE-SSW to N-S (~360-030°) trending normal faults active during the mid-late Burdigalian (17-18.5 Ma). At the surface, faults were high-angle and planar. Rare slickenside data indicates dominant dip-slip movement. Large normal faults with throws of tens to 500m defined a line of tilted fault blocks, half graben with low areas adjacent to fault blocks. Smaller faults with throws of tens of metres disrupted the basin fill.

- Sedimentary geometries indicate that basin filling occurred mainly after extensional faulting had created the rift margin topography (post-rift deposition). Syn-rift sedimentation occurred only rarely and was localised at outcrop scale. The rate of tectonic subsidence thus exceeded the rate of sediment supply. The underfilled fault topography was progressively filled with the complex volcano-sedimentary succession.

- Near volcanic centres, a rapid and voluminous supply of ignimbritic volcanic material led to complete filling of the rift topography (Araxigi Formation) and the provision of a new sediment source, whilst adjacent areas remained underfilled.

- The Sarcidano sub-basin fill is complex with abrupt lateral and vertical facies changes within a predominantly conformable succession. Basin filling commenced with deposition of the continental clastic sediments of the Villanovatulo Member which were deposited by alluvial fans sourced from local, fault-bounded topographic highs. After ?early Aquitanian transgression onto the basin margin, deposition of the Duidduru Member was characterised by erosion of fault scarps which supplied clastic material to a shallow marine coastal system of fan deltas, tidal bars and channels. Further basinward deeper water marlstones and turbidites cut by submarine channels accumulated. At the same time, the Araxigi Formation volcanic succession was deposited in fault-formed accommodation space to the north-west of the Sarcidano sub-basin and consisted mainly of dacitic pyroclastic flow deposits (e.g. large volume ignimbrites) supplied from outside the study area. Renewed transgression into the Sarcidano sub-basin led to the deposition of the Is Paras Member platform carbonates behind the Isili fault block, Serra Longa Member transgressive lag and Giara Group marlstones and calcarenites.
Chapter 6
Chapter 6 - Strike-slip systems of eastern Sardinia and the eastern rift margin.

The basement rocks of central and northern eastern Sardinia are cross-cut by a number of 10-70 km long, sinistral strike-slip faults (Carmignani et al. 1987, 1992b, 1994; Assorgia et al. 1995; Oggiano et al. 1995; Barca et al. 1996) trending from NNE-SSW to E-W. Fault movement occurring between the post-Palaeocene and Burdigalian (Dieni et al. 1987; Carmignani et al. 1992b; Oggiano et al. 1995) caused transpressional duplexes to develop in eastern areas (Monte Albo, Fig. 6.1; Carmignani et al. 1992b) and transtensional basins to form on releasing bends towards the margins of the Sardinian Rift (Oschiri and Ottana sub-basins; Oggiano et al. 1995; Assorgia et al. 1995). The precise timing of movement on the strike-slip structures is poorly documented and will be addressed here. This chapter examines the evidence for Oligo-Miocene strike-slip fault movement, the nature of the ‘transtensional’ sub-basin fill, and the character of the zones where the strike-slip systems intersect the main ~N-S Sardinian Rift trend. The results presented come from a focused, reconnaissance scale study which utilised the existing database.

The basin fill within the Oschiri, Ottana sub-basins and along the Sardinian Rift margin is considered within the framework of a basal Oschiri Formation comprising intercalated continental conglomerates, lacustrine sediments, tuffs and ignimbrites, a volcanic Logudoro Group (chapter 7) and an upper continental-marginal marine, clastic Chilvani Formation (chapter 7, see sections 3.3.3 and 3.3.4). The lithofacies and depositional environments of the basin fill are considered in more detail in chapters 7 and 8, where the better exposed successions are discussed in more detail.

6.1 Overall structural framework

6.1.1 Structural geometries

Eastern Sardinia is composed of a metamorphic basement that was deformed and intruded by granitoid batholiths during the Hercynian Orogeny. This basement is cross-cut by Cenozoic strike-slip faults (Fig. 6.1) that are believed to have reactivated late Hercynian fabrics (Chabrier and Chorowicz 1981, Carmignani et al. 1992b, Oggiano et al. 1995). For example, faults and fault bends follow the same trends as late Hercynian dykes within the basement (Oggiano et al. 1995, Barca et al. 1996). Whilst the Nuoro fault has long been recognised as a left-lateral strike-slip structure (Alvarez and Cocozza 1974), the other left-lateral E-W to NNE-SSW trending structures were identified by Carmignani et al. (1987, 1992b) and Barca et al. (1996; Fig. 6.1). The NE-SW to NNE-SSW trending faults are linear to curvilinear (e.g. Olbia, Tavrolara, Nuoro, Capo Comino faults) with spacings of 5-20 km (Fig. 6.1). Together with the E-W structures (Cedrino fault, Posada fault) wedge-shaped geometries reminiscent of 'strike-slip duplexes' were formed (sensu Woodcock and Fisher 1986 in Carmignani et al. 1992b;
In the field, the large, strike-slip faults can be recognised from linear or curvilinear depressions within a mountainous topography. Where exposed, the rocks within these zones tend to be heavily sheared and brittlely fractured with multiple, but mostly sinistral strike-slip senses of offset (Fig. 6.2). The faults tectonically juxtapose different basement units or form a sharp basement topography having tens to hundreds of metres of relief next to which Cenozoic units were deposited (Fig 6.3). For example, southwest of Ottana, the Nuoro fault bounds a relatively high granite paleotopography, adjacent to and onlapped by flat lying ignimbrites and other units filling the subsided basin (Fig. 6.1). The granite exposed at or near the fault is heavily sheared with an almost vertical fabric striking at $220^\circ$, sub-parallel to the Nuoro fault. New road cuttings along the Capo Comino fault (Fig. 6.1), show a variety of brittle fault fabrics with cataclastic fault zones from a few millimeters to several centimetres wide.

Similar observations have been used to define the faults mapped by Carmignani et al. (1992b, 1994), Oggiano et al. (1995) and Barca et al. (1996; Fig. 6.1). For example, along the Nuoro fault, granites are cut by vertical or steeply dipping cataclastic fault zones that occur up to tens of metres away from the main fault plane (Carmignani et al. 1992b). Fault zones commonly show near vertical shear planes, left lateral movement indicators and place a number of tectonic basement slices in contact (Carmignani et al. 1992b). On geological maps (sheets 194, 297, Barca et al. 1996), the sinistral strike-slip faults clearly offset the boundaries of basement units (leucogranites, monzogranites, metamorphic rocks) by up to 4 km.

6.1.1.2 Transpressional structures
At the eastern end of the Nuoro fault, northwest striking, high-angle overthrusts place Hercynian basement rocks over Mesozoic carbonates and occur within the Mesozoic carbonates, forming the high mountain of Monte Albo, a 'half-flower' structure (Carmignani et al. 1992b; Fig. 6.4). The orientation of structures observed within this area are consistent with a sinistral transpressional origin with a NNE shortening direction (Carmignani et al. 1992; Fig. 6.4B(2)). This transpressional duplex is thought to have developed on a restraining bend and in area of convergence between the E-W Cedrino and NE-SW Nuoro faults (Carmignani et al. 1992b, Fig 6.4). Carmignani et al. (1992b) suggest that a similar transpressional duplex may have developed between the Posada and Tavrolara faults (Fig. 6.4).
6.1.1.3 Transtensional structures

Topographically depressed sub-basins with fault patterns similar to those predicted for sinistral transtensional duplexes are observed at the western ends of the exposed strike-slip faults (e.g. Ottana sub-basin, Assorgia et al. 1995; Oschiri sub-basin, Oggiano et al. 1995; Figs. 6.1, 6.4B3). Both sub-basins are thought to have formed in response to sinistral strike-slip movement at a releasing bend on the southern bounding fault (Nuoro, Olbia) and by oblique slip movement with an important normal component on the northern bounding fault (Bolotana, Berchidda/Tula, Assorgia et al. 1995, Oggiano et al. 1995). Seismic data from the Oschiri sub-basin confirms that the northern fault has a dominant normal movement sense, such that the northern Berchidda/Tula faults bound a half-graben structure cut by high angle fault strands (Barca et al. 1997). The Oschiri sub-basin shows geometries typical of a transtensional basin whereas the evolution of the Ottana area may also have been controlled by coeval faulting on the -N-S Sardinian Rift trend (below).

6.1.1.4 The eastern margin of the Sardinian Rift

The eastern margin of the Sardinian Rift is complicated due to the intersection of the NE-SW trending, strike-slip fault family described above with normal, N-S trending rift bounding faults (Fig. 6.1). Basement-cover relationships in this area are poorly exposed because volcanic rocks of the basin fill conceal the underlying structure and because of forestation. North-south trending normal faults define the basin margin between some strike-slip faults (e.g. south of Ozeiri, south of Lago Omodeo, Fig. 6.1) and similar structures probably exist under the volcanic rocks along the central eastern rift margin in order to form the necessary accommodation space. The onlap of volcanic rocks suggests that a palaeotopography formed by the strike-slip faults and normal faults existed before Oligo-Miocene eruption. For example, basement promontories bounded by the NE-SW structures protruded into the basin (e.g. Bolotana area) and the basin fill onlaps the slopes of such features (e.g. Foresta Burgos area, Fig. 6.1)

6.1.2 Measured fault geometries and kinematics

Fault planes were examined at outcrop scale wherever possible with the aim of deducing the nature and movement histories of the large faults (Fig. 6.1), and to compare this data with that presented in the literature. Accessible outcrops with exposed fault planes are fairly rare. Figure 6.5 shows the location and nature of the measured small faults which had throws ranging from a few centimetres to a few metres.

The small faults often parallel or sub-parallel the large faults as expected, and the majority of the faults dip at greater than 55° (Fig. 6.5 intra-basement st1, 3, basement-cover st4, intra-basin fill st5, 6, 7). Syn-sedimentary normal and oblique-normal faults within the Oschiri sub-basin are sub-parallel to the Oschiri fault (Fig. 6.5 st5, 7), and whilst some post-depositional faults follow an E-W trend, others are
oriented NW-SE (Fig. 6.5 st8) and their genesis is uncertain (section 6.4.3). Stereoplot 2 (Fig. 6.5) shows the NNW-SSE to N-S orientation of faults bounding a small fault block. This outcrop is very close to the Nuoro fault but the observed structures are at high angles to it. The faults here were probably related to other ~N-S structures observed in this area (e.g. Oddoni fault, Fig. 6.9).

Normal faults defining the main Sardinian rift margin have dip-slip slickensides (Fig. 6.5 st6). Slickensides measured along the Capo Comino fault together with fault offsets show a mixture of normal (Fig. 6.2) and reverse, oblique sinistral and sinistral strike-slip movement (Fig. 6.5. st3). No cross-cutting relationships between the different fault types were observed. It is possible that the Capo Comino fault had a polyphase movement history under a number of different stress regimes but lack of cross-cutting relationships may suggest that they formed in one complex deformation phase.

### 6.1.3 Summary

Field observations and published data show that linear and curvilinear features cutting across the basement of eastern Sardinia are strike-slip faults. However, it is unclear exactly when the strike-slip faults were active and how they relate to the formation of the Oligo-Miocene Sardinian Rift. An examination of the Cenozoic rock sequences and geometries associated with the strike-slip faults and along the eastern Sardinian rift margin is the key to unravelling this relationship. The sections below describe the results of such a study.

### 6.2 The Ottana sub-basin

#### 6.2.1 Structural style

The Ottana sub-basin is the topographically depressed area between the releasing bend of the sinistral Nuoro fault and the Bolotana-Bono oblique slip faults (Fig. 6.1). The eastern side of the basin is bounded by the ~N-S trending Oddini normal fault, which is visible in one road cutting (Fig. 6.1). Other ~N-S structures are observed to the east of the Ottana sub-basin (Fig. 6.1, 6.5 st2, 6.9) and may represent the interference of the Sardinian Rift trend with the strike-slip fault systems. Within the basement to the northeast of the Ottana sub-basin, E-W trending structures (Orotelli, Aurora faults) form topographic depressions. These depressions are partly filled by pumice-ash ignimbrite which shows similar characteristics to the Miocene ignimbrites of the Ottana sub-basin. To the south of the Ottana sub-basin, the Busachi NNE-SSW normal fault defines the rift margin towards the Sarcidano area (Chapter 5). Porcu (1983) shows a set of N-S to NNW-SSE trending normal faults, which cut the strike-slip structures and volcanic rocks at the Ottana sub-basin margins with offsets of a few tens of metres. The area was not examined in enough detail to detect all of these structures, but it is clear that a pre-depositional and post-depositional ~N-S fault family exists over the Ottana sub-basin (e.g. Oddini fault, Figs. 6.1, 6.5, 6.8, 6.9 below).
6.2.2 Basin filling units

6.2.2.2 Oschiri Formation (late Aquitanian-early Burdigalian)

The unit termed here the ‘Oschiri Formation’ comprises a basal succession of rhyolitic-rhydacitic ignimbrites, tuffs, and inter-eruption coarse channelised conglomerates containing volcanic and basement clasts (Porcu 1983, Assorgia et al. 1995). In the field, poorly exposed, pumiceous white ignimbrites with basement xenocrysts (Allai unit, Assorgia et al. 1995) and tuffs alternate with red, jointed ignimbrites with fiamme structures (Porcu 1983), aligned flow textures and quartz, plagioclase ± clinopyroxene and hornblende crystals. The volcanic succession reaches up to 180m in thickness (Porcu 1983). Four kilometres north of Ottana, thinly bedded (mm-20cm) tuffs and pumice tuff/lapilli tuffs exhibit low angle cross-bedding and trough structures (Fig 6.6). These rocks may represent the deposits of pyroclastic surges (after Fisher and Schminke 1984, Cas and Wright 1988), or ignimbrite and tuffs reworked at the margins of a shallow lake or sea.

In the west of the sub-basin, the volcanic rocks are overlain by up to 60m of parallel and cross-bedded coarse sandstones composed of red feldspar, quartz and metamorphic clasts, thought to have been deposited in a fluvial environment (Sedilo sands, Porcu 1983). The ‘fossil forest of Zuri’ is found within this unit and comprises the remains of palm and other trees (Porcu 1983; Porcu et al. 1997). The 60-70m thick Sa Manenzia pumice-ash pyroclastic flow overlies the fossil forest (Porcu 1983; Porcu et al. 1997).

6.2.2.3 Chilvani Formation (?mid/late Burdigalian)

The ‘Dualchi sands’ (Porcu 1983), here correlated to the Chilvani Formation (section 3.3, Appendix 3D) are 30-40m of conglomerates, sandstones and mudstones derived from weathering of the Palaeozoic basement and underlying volcanic rocks deposited in a fluvial (at the base) to marine environment (at the top), with oysters, Pecten and other bivalves (Porcu 1983). Near Dualchi, the coarse sandstones crop out as a massive, 4m thick unit derived from the weathering of granite with a silt grade matrix, yielding very little evidence for deduction of depositional processes. To the west of the Ottana sub-basin this clastic unit passes upwards into deeper littoral-epibathyal marlstones (Tadasuni-Sorradile marls, Porcu 1983, Assorgia et al. in press)

6.2.3 Basin filling geometries

On a large scale, flat lying volcanic rocks fill and cover the fault-controlled basement palaeotopography. The rocks are found in the linear fault depressions of the Bono, Nuoro, Orotelli and Aurora structures, the Ottana sub-basin depocentre, between fault strands on the releasing bend of the Nuoro fault and capping the fault ‘footwalls’ (Nuoro and Bolotana faults). In areas where a few tens of metres of the volcanic succession is exposed, for example between the fault strands on the releasing bend of the Nuoro fault east of Sedilo (Fig. 6.1), the rocks form a conformable, parallel bedded unit.
In the footwall of the Oddini fault, ignimbrite horizons gently onlap onto the basement slope of a few degrees. These geometries indicate that the Ottana sub-basin had formed by movements on the strike-slip faults and on some N-S normal faults before this phase of volcanic eruption, dated at 20.71± 0.22 Ma (section 3.2.3) or 21.1± 1.4 Ma (late Aquitanian, Assorgia et al. 1995). Dip variations up to −15° occur along the length of the Nuoro fault (Fig. 6.1) and could have been caused by tilting associated with the cross-cutting NNW-SSE faults identified by Porcu (1983) which moved after volcanic eruption.

South of Ottana, the onlap of flat-lying tuffs and welded ignimbrites onto the Nuoro fault is observed (Fig. 6.7), confirming that the Nuoro fault paleotopography existed before the eruption of the volcanic rocks exposed at the surface today. Stratigraphically beneath the flat lying volcanic rocks, an angular unconformity of 13°, indicates syn-depositional fault movement occurred within fluvial volcanic sandstones and channelised conglomerates (Fig 6.8). The outcrop is sub-parallel to the Nuoro fault and the angular unconformity is probably related either to local ~N-S fault movement or to a splay off the Nuoro fault, rather than movement on the Nuoro fault (Fig. 6.1, 6.5 Stl). Ten kilometres further northeast along the Nuoro fault, breccias and conglomerates crop out adjacent to and cover a block bounded by a ~N-S fault set. These are in turn overlain by volcanic rocks at the top of the outcrop (Fig. 6.9). This geometry also shows that a ~N-S faulted basement topography existed, and was degraded, before the volcanic succession was erupted from the late Aquitanian-early Burdigalian.

The eruption of the volcanic rocks outcropping at the surface was not therefore contemporaneous with fault movement or related to an ‘important extensional phase’ as suggested by Assorgia et al. (1995).

6.2.4 Constraints on the nature and timing of movement on the Nuoro fault

Fault geometries and kinematics show that the Nuoro fault was a sinistral strike-slip structure (this study; Alvarez and Cocozza 1974; Carmignani et al. 1992; Barca et al. 1996). Basin filling geometries show that Nuoro fault movement occurred before volcanic rocks dated at 20.71± 0.22 Ma (latest Aquitanian) were erupted. Some ~N-S normal faulting also occurred before this time. The presence of N-S structures is not incompatible with the sinistral strike-slip system (Fig. 6.4B (3)) and they may reflect extension on the main Sardinian Rift trend. The ~N-S normal fault family also moved after the eruption of the volcanic succession according to the map of Porcu (1983). The present day topography of this area is dominated by the steep Bolotana scarp and flat-lying Ottana plain, a result of the upper-mid Pliocene tectonic reactivation on the Bolotana fault reactivation, suggested by Porcu (1983). By examining the offset of basin filling units, the Nuoro fault has not undergone a reactivation of more than a few tens of metres since the Oschiri Formation volcanic rocks were erupted.
6.3 Sediments at the eastern end of the Nuoro-Cedrino fault system

Syn-tectonic breccias and conglomerates derived from erosion of the Palaeozoic basement and Mesozoic carbonates are found within thrust slices within the Monte Albo transpressional duplex (Cuccuru e Flores conglomerate, Carmignani et al. 1992b; Fig 6.10). In places, the conglomerate lies in tectonic contact with the metamorphic and limestone rocks and is itself deformed and crushed to a clayey cataclastic zone, 1-1.5m wide, containing sinistral kinematic indicators (Carmignani et al. 1992b). East of Siniscola, the breccia has a carbonate matrix with ooids, chert horizons, is faintly stratified and unconformably overlies the metamorphic basement (Carmignani et al. 1992b). No marine indicators were found within the rocks, which are therefore interpreted as scarp breccias entering a lacustrine environment.

Dating these sediments is of particular importance for constraining the timing of transpression (6.1.1.2). At present, the only constraint comes from a reworked clast indicating a post Palaeocene age (Dieni et al. 1987). Dieni and Massari (1965) correlated the sediments of Monte Albo to calcirudites of Cuisian age near Orosei (14 km to the southeast, Fig. 6.1), which contain a rich macroforaminifera fauna (Nummulites sp. and ?Discocyclina sp. identified here). This correlation is now widely used so that the Monte Albo conglomerates are said to be post-early Lutetian in age (Carmignani et al. 1992b, 1994, 1995; Oggiano et al. 1995). Given the structural complexity of this area, the facies diachroneity present in the Sardinian rift and the implications that the age of Monte Albo transpression has for models of Western Mediterranean tectonic evolution, this simple correlation of ?non marine to marine sedimentary facies must be treated with caution. Using the above criteria and their preference for a regional model, Carmignani et al. (1994) state that ‘we relate these deposits (the Cuccuru e Flores conglomerate) to a Late Eocene or more probably Oligocene tectonic (transpressional) event’.

In the present study, sediments at the eastern end of the Nuoro-Cedrino fault system were examined in order to verify their structural position and to find material suitable for dating. It was hoped that some marine carbonate skeletal material suitable for $^{87}$Sr/$^{86}$Sr isotope dating (Chapter 3) would be found, but apart from organisms within the Jurassic dolomites, no fossils were discovered.
6.4 The Oschiri sub-basin

6.4.1 Structural style

The Oschiri sub-basin was formed along the releasing bend of the sinistral strike-slip Olbia fault and in the hangingwall of the oblique slip Berchidda and Tula faults (Oggiano et al. 1995; Fig. 6.11, 6.12). On the releasing bend, the Olbia fault changes strike from 035 to 060° and a transtensional fault splay developed, with the 075° trending Oschiri fault forming the other major structure (Oggiano et al. 1995; Fig. 6.1, 6.12). A number of reverse and normal faults trending at 120° were mapped by Oggiano et al. (1995), though only one normal fault on this trend was identified in this study (Fig. 6.11). Two and a half kilometres southwest of Oschiri, three 020-030° normal faults juxtapose granite and basin filling units (Fig 6.11). The pattern of faults compares well with those predicted for a sinistral strike-slip system (Figs. 6.4113, 6.12; Oggiano et al. 1995). Post-depositional, N-S normal faults east and southwest of Tula (Fig. 6.11) belong to a different fault family related to events identified within the main Sardinian Rift (also in Oggiano et al. 1995).

6.4.2 Basin filling units

The Oschiri sub-basin is poorly exposed. The age and stratigraphic location of continental conglomerate units, thought to occur in the Oligocene-Aquitanian, Burdigalian (Oggiano et al. 1995) and from the Pliocene onwards, are very difficult to ascertain.

6.4.2.1 Oschiri Formation (?Oligocene-early Burdigalian)

The Oschiri Formation is defined from a few type outcrops within the Oschiri sub-basin. It is comprised of finely interbedded lacustrine limestones and tuffs ([098079], Lithofacies lh, lj, 5g, Tables 8.4, 8.5) and intercalated basement derived conglomerates, tuffs, lapilli tuffs and ignimbrites [079071] (Lithofacies la, lb, le, 5e, 5h, 5a, 5j, Tables 8.4, 8.5). The thickness exposed over the Palaeozoic basement ranges from a few metres to a few tens of metres.

Adjacent to the Berchidda fault, a massive boulder breccia composed only of basement clasts crops out (Breccia di Codinattu, Oggiano et al. 1995). These sediments, which have a haematite cement, were deposited in an alluvial fan at the edge of the sub-basin (Oggiano et al. 1995). Over approximately 20m, the boulder breccia fines away from the basin edge into massively stratified conglomerates and coarse sandstones with pebble strings, diffuse matrix, clast supported patches, and rare, reddened medium sand grade horizons (Lithofacies la, 1b, 1e, Table 5.6). These clastic sediments have a granitic composition with a tuffaceous matrix, they are poorly sorted with angular clasts and may represent deposition by debris flows in a more distal part of the alluvial fan (after Hooke 1967, Bull 1972, McGowen and Groat 1971, Collinson 1986, Miall 1996). Within the sub-basin centre, lacustrine limestones, cherts and tuffs containing abundant silicified plant pieces and mammal remains were deposited (Oggiano et al. 1995). The lacustrine deposits are finely laminated (Fig. 6.13), preserve
silicified plant stems in hand specimen and in thin section (Fig. 6.14), and are similar to the rocks of the Perugias Formation (Lithofacies 1h, 1j, Table 8.4) to the north of the Tula fault. The conglomerates and lacustrine sediments are intercalated with, and overlain by volcanic rocks. These include red, welded dacitic-andesitic pyroclastic units with flow textures, dirty green-brown epiclastic tuffs, thin plagioclase-hornblende-opaque tuffs and white-green pumice lapilli tuffs containing plagioclase, rare orthopyroxene and hornblende (Lithofacies 5e, 5h, 5a, 5j, Table 8.5; Figs. 6.15, 6.16).

6.4.2.2 Chilvani Fm. (mid-late Burdigalian)

In the eastern end of the Oschiri sub-basin, the lacustrine sediments of the Oschiri Formation are unconformably overlain by a conglomerate (Oggiano et al. 1995; Lecca and Tilocca 1997) deposited by alluvial fan and fluvial systems thought to be Burdigalian in age (Oggiano et al. 1995). Further to the west, the conglomerates sit on the volcanic rocks of the Oschiri Formation (Oggiano et al. 1995). Whilst these continental clastics could not be clearly distinguished from Pliocene to modern conglomerates, similar rocks of the Chilvani Formation outcrop to the west, beneath the middle Miocene marine succession (Chapter 7).

6.4.3 Basin filling geometries

Within the Oschiri sub-basin, two outcrops show clear evidence of syn-depositional faulting and folding. Listric normal faults occur in finely laminated lacustrine sediments of the Oschiri Formation, a few hundred metres WSW of Oschiri town (Fig 6.13, 6.17). The sediments show progressive bed rotation, and thickening into the normal fault, onlap and pinch out away from the fault (Figs. 6.13, 6.16). The syn-sedimentary fault set is overlain by onlapping, parallel bedded sediments (Figs. 6.13, 6.16). Whilst the faults could have developed in response to an unstable slope, the alignment with the Oschiri fault (Fig. 6.5 st5) located a few hundred metres to the south, suggests they are directly associated with movement on the releasing bend of the strike-slip fault.

A few kilometres to the northwest of Oschiri, road cuts on the new 'superstrada' expose a gentle syncline, with an axis trending 060°, overlain by flat lying beds (Fig. 6.11, 6.17). The fold developed within dirty brown-green epiclastic tuffs and a prominent white bed. The white bed is a pumice-plagioclase-glass tuff and is also syn-depositionally faulted by 10-20cm at its base. The syn-sedimentary fault planes have a strike of 083-090° sub-parallel with the Oschiri fault. The synclinal axis is aligned with the trend of the Oschiri sub-basin and therefore may have developed in response to differential movements on faults within the sub-basin. In any case, the white bed records the end of active tectonism within the transtensional basin and plagioclase crystals from it are dated at 20.60±0.24 Ma (latest Aquitanian) by the 40Ar/39Ar technique (section 3.2.3).
Post-depositional normal and oblique-sinistral faults form a series of tilted blocks within tuffs and lapilli tuffs just west of the syn-depositionally folded outcrop (Fig 6.18). The faults strike from E-W to NW-SE (Fig. 6.5 st8). Since block rotations are commonly observed in strike-slip zones (Christie-Blick and Biddle 1985; Woodcock and Schubert 1994), the orientation of this fault set could be a result of anticlockwise block rotation.

On a large scale, the red, jointed volcanic rocks that form the uppermost unit of the Oschiri Formation are flat lying or dip at less than 10° to the east. They cap the lacustrine succession or sit directly upon the granitic basement. The geometries within the Oschiri sub-basin fill indicate that whilst sinistral strike-slip fault movement was active within the lower Oschiri Formation, it had ended by the time the uppermost volcanic units were erupted. Eastward tilting of the Oschiri sub-basin fill can be directly related to post-depositional –N-S faulting relating to extension on the Sardinian rift trend (e.g. southwest of Tula on Fig. 6.11).

6.4.4. Constraints on the timing and nature of Olbia fault movement

Fault geometries and kinematic indicators show that the Olbia fault is a sinistral strike-slip structure (Carmignani et al. 1994, 1995; Oggiano et al. 1995; Barca et al. 1996). Oggiano et al. (1995) recognise normal, reverse and oblique-slip faults, folds and unconformities within the Oschiri sub-basin. Transtensional basin formation in the late Oligocene-Aquitanian was constrained by the lower Aquitanian age of Redini (1940) on lacustrine sediments and a pyroclastic flow dated at 22± 0.8 Ma (unpublished data, Oggiano et al. 1995). Lecca and Tilocca (1997) recognise upper Oligocene-Langhian transtension but strangely, for this non-marine setting, use correlation to the ‘global’ sea level curve of Haq et al. (1988) as the temporal constraint.

By combining the observed sub-basin filling geometries with a single crystal $^{40}\text{Ar}/^{39}\text{Ar}$ date, the main phase of transtension in the Oschiri sub-basin can be constrained as up to 20.60± 0.2 Ma. Post-depositional –N-S faulting occurred after this time. The northern Berchidda and Tula faults were reactivated sometime after the Miocene (Oggiano et al. 1995) resulting in the abrupt topography present today (Fig 6.3).
6.5 The central part of the eastern margin of the Sardinian Rift

6.5.1 Basin fill (late Oligocene-early Miocene)

The volcanic rocks which cover the eastern margin of the Sardinian Rift are a direct continuation of the Logudoro Group (section 3.3.4, Appendix 3D) described in the Logudoro study area to the west (Chapter 7). The most prominent rocks outcropping along the eastern margin of the Sardinian rift are red, jointed ignimbrites, a few to tens of metres thick with aligned vesicle and crystal textures. Tuffs, white pumice-lithic lapilli tuffs with non-juvenile clasts up to a few centimetres in size, white pumice-lithic ignimbrites with aligned vesicles and stratified pumice-ash deposits with low angle cross bedding, dune forms, lensoidal and wavy, wedge shaped beds (Fig. 6.19) are intercalated with the jointed flows. The latter unit has depositional structures similar to those found in pyroclastic surges (Fisher and Schminke 1984, Cas and Wright 1988). The volcanic succession of dominant rhyolitic-dacitic composition is interpreted as the products of pyroclastic flows, surges and fallout resulting from explosive volcanic eruption in subaerial conditions (after Fisher and Schminke 1984, Cas and Wright 1988).

Immediately overlying the basement unconformity and beneath the volcanic succession, red continental, basement breccias, conglomerates, sandstones and caliche less than 10m thick is found in one locality, a kilometre northwest of Ozeiri (Fig. 6.1).

6.5.2 Basin fill geometries

The volcanic rocks crop out as a conformable, parallel bedded succession which is flat lying or dips at less than ~10° westward into the main Sardinian Rift. By correlating between exposures, the rocks appear to onlap onto a basement topography caused by existing NE-SW and N-S faults (6.1.1.4).

6.5.3 Relative timing of structures

The volcanic succession exposed at the surface effectively blanketed the existing faulted structure and did not record evidence on the relative timing of the NE-SW strike-slip and N-S normal fault structures. In addition, no area was found where the structures cross-cut each other. Therefore it is impossible to accurately constrain the events forming the central part of the eastern rift margin. Evidence from surrounding areas (e.g. Ottana, Oschiri sub-basins) suggests that NE-SW strike-slip faults may have occurred with and were cross-cut by a N-S fault family on the Sardinian rift trend. This scenario consistent with the observed outcrop patterns at the eastern rift margin.
6.6 Timing of strike-slip movement and extension

Recent regional tectonic models for the evolution of the northern part of the Western Mediterranean (section 2.1) describe a change from compression and transpression during the Late Oligocene - Aquitanian to a phase of extension starting in the Burdigalian (Carmignani et al. 1994, 1995). Previous models describe extension during the mid Oligocene - Aquitanian and compression in the Burdigalian (Cherchi and Montadert 1982ab; Rehault et al. 1984). It is thus crucial to examine the timing and relationship between the movement on sinistral strike-slip faults and extensional faults forming the Sardinian rift (see also section 7.4).

The main observations from this chapter are that:

- a system of sinistral strike slip faults dissects the central and northern eastern Sardinian basement. As displacement accrued, transpressional fault duplexes developed at restraining bends and transtensional basins formed at releasing bends.
- basin fill geometries in the Ottana sub-basin show that Nuoro fault movement and some N-S faults had formed before 20.71±0.22 Ma (latest Aquitanian).
- basin fill geometries in the Oschiri sub-basin show that a transtensional phase with movement on the Olbia, Berchidda and Tula faults occurred until 20.60±0.24 Ma (latest Aquitanian). Post-depositional N-S faults formed sometime after this.

Though the last transtensional activity has been identified in the latest Aquitanian, the start of sinistral strike-slip fault movement or the whether the faults moved in several phases has not been deduced due to lack of datable exposures. However, from the geometry of the basin fill and the identification of hundreds of metres of Oschiri sub-basin fill on seismic reflection profiles (Barca et al. 1997), it is most probable that fault movement had commenced at, or before, the late Oligocene. In addition, the field and seismic observations suggest that the majority of the fault displacement had accumulated before the late Aquitanian.

Whilst the transpressional and transtensional duplexes could have formed coevally on restraining and releasing bends respectively, the literature states that M. Albo was formed under a transpressional stress regime (Carmignani et al. 1992b), whilst the Oschiri and Ottana sub-basins were a response to a transtensional stress regime (Oggiano et al. 1995, Assorgia et al. 1995). Since the syn-transpressional sediments can only be constrained as post-Palaeocene and are dissimilar to any Oligo-Miocene units studied by the author, it seems most probable that if there were two phases of strike-slip fault movement, the phase of transpression would have preceded the phase of transtension.
Dated basin filling geometries show that in detail, movement on the Nuoro fault (Ottana sub-basin) ended before movement on the Olbia fault (Oschiri sub-basin). Lack of suitable exposures meant that the timing of movement on other strike-slip faults cannot be determined, though in places they are cut by late Oligocene normal faults and fault depressions were filled with Miocene sedimentary and volcanic rocks (Anglona study area, chapter 8). It therefore seemed the best assumption that in their pre-late Aquitanian history, the family of similarly oriented strike-slip faults in eastern Sardinia and tentatively those in southern Corsica moved at the same times within the post-Palaeocene - ?mid Aquitanian.

Carmignani et al. (1994, 1995) argued that transpressional tectonics on the strike-slip faults crossing the eastern Sardinian basement and southern Corsica occurred within the hinterland of the Northern Apennines collisional belt (Fig. 2.8), but do not consider the transtensional basins at the western ends of these faults. To the south of the sinistral strike-slip system, ~N-S arcuate faults with dextral movement sense are interpreted as a zone of tectonic escape from the collisional area (Carmignani et al. 1994, Fig. 2.8). Data presented here has shown that the sinistral strike-slip faults and associated transtensional sub-basins within the eastern Sardinian basement were active starting sometime after the Palaeocene, certainly in the late Aquitanian and ending by the latest Aquitanian. Of great importance is that some ~N-S faults with normal offsets had also formed at the western margins of the strike-slip faults before the latest Aquitanian. The implications of this and other recent studies on the regional tectonic evolution are discussed in Chapter 10.
Chapter 7
Chapter 7 - The Logudoro study area

The Logudoro study area exposes a normally faulted succession of late Oligocene-mid Burdigalian volcanic rocks covered by mid Burdigalian-Serravallian marginal marine and marine sediments. This chapter examines the style and timing of faulting which created the sediment depocentres and the large scale facies architecture of the siliciclastic-carbonate basin fill in response to the fault-created topography, sediment supply, active tectonics and relative sea level change. This complements existing studies which have described the rock types and proposed depositional models (e.g. Coulon 1977; Mazzei and Oggiano 1990; Martini et al. 1992).

The stratigraphy of the Logudoro study area was defined in section 3.3.4 and consists of a >500m thick, andesitic-dacitic volcanic succession (Logudoro Group) intercalated at its top with fluvio-lacustrine sedimentary and pyroclastic rocks (Oschiri Formation). These units are unconformably overlain by the fluvio-deltaic Chilvani Formation and marginal to fully marine sandstones, limestones and marlstones of the Florinas Group (Figs. 3.14-3.16). Figure 7.1 and Enclosures 3 and 4 are geological maps of the area and show the localities discussed in the text.

7.1 Structural geometries observed within the Logudoro study area

On a large scale, the Logudoro study area represents the main N-S trending part of the Sardinian Rift in central northern Sardinia and its intersection with the transtensional basins (chapter 6) towards the east (Figs. 7.1,1.3). The submarine eastern, and exposed western margins of the rift basin are thought to be bounded by -N-S trending normal faults (e.g. section 6.1.1.4) forming a graben structure which dominated western Sardinia. (Pecorini et al. 1988; Fais et al. 1996). Today, the graben is buried beneath the thick Oligocene-lowermost Miocene volcanic succession (Logudoro Group) and subsurface imaging is required to characterise it. Sedimentary depocentres which formed on top of the Logudoro Group were not so simply delimited. A number of sub-basins were created by extension on several different fault trends (Fig. 7.1). It is the evidence for the nature and timing of these structures which is described below.

7.1.1 Previous Work

Various authors have recognised E-W to NE-SW and NNW-SSE fault sets within the Logudoro study area in addition to 'late' (post Miocene) N-S normal faults (Assorgia et al. 1988; Martini et al. 1992; Barca et al. 1996; Funedda et al. 1997). The interference of NE-SW trending transtensional basins with the NNW-SSE or N-S fault sets defining the main Sardinian Rift trend was described by Oggiano et al. (1995, Fig. 7.2) and Funedda et al. (1997).
A gravity, resistivity and magnetic study by Pecorini et al. (1988) was used to try and define the deeper structure of the ‘sub-basins’ within the Logudoro area of the Sardinian Rift. Whilst the geophysical data was complicated by the presence of large volcanic bodies (Bouguer gravity anomalies and magnetic highs), negative Bouguer anomalies and magnetic measurements indicate a N-S to ENE-WSW trending, ~2000m deep sub-basin, west and south of Sassari and a depocentre in the Ardara-Chilvani region (compare with Fig. 7.1). As a synthesis, Pecorini et al. (1988) suggest a progressive deepening of the Oligo-Miocene basin to the west (Fig. 7.3). This is in contrast to seismic reflection profiles taken ~40km to the north which show progressive eastward subsidence (chapter 8, Fig. 8.4, Enclosures 6 and 7). The geophysical data supports the hypothesis of a large, ~N-S trending (100’s m throw) normal fault bounding the main Sardinian Rift in the vicinity of Ardara as suggested by Oggiano et al. (1995, Figs. 7.2, 7.3 from Pecorini et al. 1988).

7.1.2 Large scale geometry

The sedimentary depocentres overlying the Logudoro Group which formed by extension on several fault trends can be split into three groups (Figs. 7.1, 7.4).

- The **Sassari sub-basin** trends ~N-S west of Sassari and ~E-W south of Sassari and appears to have a graben geometry (Fig. 7.4). The eastern margin of the sub-basin was bounded by the proposed, N-S trending Calancui fault and WNW-ESE trending Ploaghe fault. The western margin of the sub-basin was bounded by N-S extensional faults (Molafa, Ozzastru) in the west and a number of northward dipping, E-W and NNW-SSE normal faults in the south (Cuga, Brundette, Rocca Bianca, Ittiri, Banari; Fig. 7.4).

- The triangular shaped **Oschiri-Sassari sub-basin intersection zone** in the central part of the Logudoro study area was delimited by the NE-SW trend of the Olbia and Berchidda strike-slip faults (chapter 6) and by the WNW-ESE trending Ploaghe fault (Fig. 7.4). The sedimentary depocentre was most probably created by now buried N-S to NNW trending normal fault or faults as suggested by Pecorini et al. (1998) and Oggiano et al. (1995) in the vicinity of Mores and Ardara. NNW-SSE trending structures in the west of the area and NE-SW structures in the east were formed or reactivated in the Plio-Quaternary (e.g. Thiesi, Santo, ?Iscalas; Fig. 7.4).

- In the **southern part of the Logudoro study area**, the depocentres around Romana were created by NW-SE oriented normal fault movement (Fig. 7.4). West of Padria, sediments appear to have accumulated adjacent to a NNE-SSW trending normal fault (Fig. 7.4). However, it is unclear how Miocene sediment depocentres formed in south central Logudoro or whether, such as in the case at S.M. Iscalas and Mara (Fig. 7.1), sediments merely infilled a remnant volcanic topography (7.2.6.8).
The Logudoro study area passes northwards into the Anglona study area. The structural style of the eastern margin of the Sassari 'graben' and Portotornes half-graben (section 8.2) sub-basins is similar. They were formed by westward dipping, N-S to NNE-SSW/NNW-SSE trending normal faults (Calancui, Sennori and Tramontana respectively) with throws of hundred of metres.

Without extensive seismic reflection profiles or other high resolution geophysical techniques, the precise structural geometries defining the Logudoro study area remain unclear. However, sedimentary depocentres above the Logudoro Group appear to have been created by the interaction of NE-SW to E-W, NNW-SSE and N-S fault sets. It may be that the NE-SW to E-W fault family were related to the similarly oriented strike-slip fault systems which are exposed to the east, and reflect the inherited basement fabric. The NNW-SSE to N-S faults apparently represent a new fault family defining the main Sardinian Rift.

The section below presents the evidence for normal faulting within the Logudoro study area. Basin filling geometries constrain the timing of faulting and are used to help evaluate the relationships between the two main fault sets.

### 7.1.3 Field evidence defining normal faulting

#### 7.1.3.1 Fault geometries in the Sassari sub-basin

Table 7.1 summarises the evidence used to define post mid Oligocene faults in the Sassari sub-basin and the proposed character of the faults. In combination with normal offsets of the Logudoro Group, the planar, high angle contacts between volcanic rocks and the Florinas Group are recognised as extensional faults. Very few actual fault planes are exposed, but those that are, are high angle, planar and sub-parallel the proposed larger structures. No slickensides were found to deduce whether the faults with visible normal offset moved by dip- or oblique-slip movement. The normal faults bounding the Logudoro Group are thought to have throws of at least hundreds of metres since the sedimentary succession deposited in adjacent accommodation space reaches a few hundred metres thick.

#### 7.1.3.2 Timing of fault movement in the Sassari sub-basin

Within the Sassari sub-basin, E-W trending faults formed before late Burdigalian marine transgression. The evidence is a sealed E-W oriented fault plane, east of Uri ([591994], Fig. 7.6), sediments onlapping onto fault planes at outcrop scale (Fig.7.5f) and the large scale arrangement of thick marine sediments within normal fault hangingwalls which appear to onlap the fault planes rather than cut by them (all east-west faults, e.g. Fig. 7.5de). The last two criteria also apply to the WNW-SEE trending Ploaghe fault, NNW-SSE trending normal faults and the Ozzastru N-S fault which also formed before marine transgression (e.g. Fig. 7.5b). The majority of movement on the Molafa and Calancui N-S faults is thought to have occurred before marine transgression because the sediments which thicken
over the fault topography, are not cross-cut by faults, and because of facies variations across these structures (below). Some syn-depositional fault movement is recorded as bed divergence within the basal, mid Langhian age marlstones and overlying late Langhian limestones. Bed divergence was caused by ‘forward’ rotations of up to 30° (i.e. dipping in the same direction as the fault) in the immediate fault hangingwall (Figs. 7.5ac). Such ‘forward’ rotations are common in other areas of the Sardinian Rift and could be attributed to fault propagation folding where the normal fault propagates up through the basin fill but does not necessarily cut through it (sensu Withjack et al. 1990; Gawthorpe et al. 1997; Sharp et al. in press). Sealed N-S normal faults of late Burdigalian-Langhian age are also found north of Ittiri (Fig. 7.1 B-B’). In the Sassari sub-basin, the majority of the Florinas Group can be classified as post-rift deposits, though localised syn-rift deposits are found. Sediment geometries adjacent to outcrop scale fault planes sometimes show subtle bed divergence (e.g. Fig. 7.5b,f) and this could be a product of syn-depositional tectonic movements (fault propagation folding, fault drag) or syn-depositional sediment compaction.

The Sassari sub-basin thus existed before late Burdigalian marine transgression, though N-S faults also moved in the Langhian. The southern margin of the Sassari sub-basin was defined by E-W and N-S or NNW-SSE normal faults. No cross-cutting relationships were observed and it is unclear whether the two fault sets moved at the same time, or consecutively, in response to two tectonic phases. The faulted volcanic structure formed a template for sediment deposition; its role on facies architecture and further evidence for faulting within the basin fill is discussed below.

7.1.3.3 Oschiri-Sassari sub-basin intersection and south Logudoro areas

Table 7.2 summarises the evidence used to define post mid Oligocene faults in the Sassari-Oschiri sub-basin intersection and south Logudoro areas and the proposed character of the faults. The triangular geometry of the Oschiri-Sassari sub-basin intersection zone was formed by subsidence on the strike-slip Olbia and Tula faults (chapter 6) and normal Ploaghe fault, which have throws of hundreds of metres. The strike-slip faults were active before transgression and movement ended in the latest Aquitanian (chapter 6). Proposed N-S trending faults, which are thought to form the eastern edge of the main Sardinian Rift trend (Pecorini 1988; Ogiano et al. 1995) in the Ardara - Mores region, are not exposed at the surface. North-south trending normal faults were tentatively identified at the eastern margin of the Sardinian Rift (chapter 6, Fig. 7.1). The western margin of the sediment depocentre is defined by the onlap of the Florinas Group onto the Logudoro Group and by backtilting in the footwall of the post-depositional Thiesi fault.
In southern Logudoro, some fault defined sedimentary depocentres were formed by normal offsets on NW-SE trending structures (south of Romana) and a NE-SW oriented structure (west of Monteleone Roccadoria). Fault movement on both fault sets occurred before late Burdigalian marine transgression. Evidence includes small faults within the Logudoro Group on the Monteleone fault-trend which are sealed by the basal carbonate of the Florinas Group (Figs. 7.4st1, 7.7), the degradation of the NE-SW fault before transgression (Fig. 7.5k) and the large scale geometry of flat-lying, undeformed sediments within the depocentre defined by the Monteleone fault. No observations were made to elucidate how accommodation space was formed above the Logudoro Group in south central Logudoro. It is possible that this area was underfilled at the end of Logudoro Group volcanism.

In the Pliocene, alkali basalts were erupted onto a dissected topography and crop out at different topographic levels. The Oschiri-Sassari intersection zone and southern Logudoro were cross-cut and tilted by Plio-Quaternary N-S and NNW-SSE faults and dykes (this study; Feraud and Campredon 1983; Assorgia et al. 1988; Carmignani et al. 1994; Funedda et al. 1997). Some published maps also show NE-SW structures cutting across the Pliocene basalt plateaus (Assorgia et al. 1995) and Oligo-Miocene basin fill (Barca et al. 1996, north of Bonorva, Fig. 7.1) suggesting a post-Pliocene phase of movement on the 'strike-slip' fault family. In this study, a ?post-depositional NE-SW fault north of Mores was identified (Fig. 7.5h).

7.2 Basin fill architecture

This section describes the nature, geometries and arrangements of basin filling units within the Logudoro study area with a view to determining overall depositional environments and palaeogeographies. Emphasis is placed on the large scale lithofacies relationships over this extensive study area since in detail, the lithofacies present are similar to those defined in chapters 5 and 8 and were described by Mazzei and Oggiano (1990) and Martini et al. (1992).

7.2.1 Logudoro Group (mid Oligocene-mid Burdigalian)

The Logudoro Group volcanic rocks are exposed at the western, eastern and northern margins of the Logudoro study area (Fig 7.1). These topographically high areas are positioned in the footwalls of mid-late Burdigalian, post-Logudoro Group (below) and pre-marine transgression normal faults (section 7.1, Fig 7.1). The succession exposed at the surface is at least 500m thick.
7.2.1.1 Rock Types

C. Coulon and his co-workers formulated a volcanic stratigraphy for the Logudoro area in the seventies (Table 3.1; Coulon et al. 1974; Coulon 1977). The succession consists of subaerially erupted, intercalated basaltic to andesitic domes, lava flows, dykes and breccias plus voluminous dacitic to rhyolitic pyroclastic rocks, domes and flows (Coulon et al. 1974; Coulon 1977; Assorgia et al. 1988; Table 3.1). The succession of pyroclastic rocks in the Monte Traessu area (Fig. 7.1) was examined by Assorgia et al. (1988) who document a basal ash-pumice flow, ash-pumice fall, ground surge and block and ash flow deposits. Within the volcaniclastic succession, Coulon (1977) recognised welded ignimbrites intercalated with partially welded pumice tuffs, cross-bedded ash deposits and pyroclastic breccias. The rock types are similar to those described in chapter 8 (section 8.4.5) for the Tergu Formation which represents the northward continuation of the Logudoro Group. Figure 7.8 is an example of a typical ‘block and ash’ pyroclastic deposit containing additional, large white pumice clasts. Lava flows and welded pyroclastic flows weather to form ridges whilst andesitic volcanic centres and dacite-rhyolite domes tend to have a rounded geometry (Fig. 7.9).

The volcanic rocks in the Logudoro study area have traditionally been viewed as a calc-alkaline suite derived from a high-alumina basalt parental magma (e.g. Coulon 1977; Coulon et al. 1974; Dostal et al. 1982) though primary high magnesia basalts have recently been found near Montresta (Morra et al. 1997). The porphyritic volcanic rocks commonly contain zoned plagioclase and clinopyroxene with olivine in basic rocks and orthopyroxene in acid rocks (Dostal et al. 1982). Amphibole and biotite are found sporadically and quartz is scarce (Dostal et al. 1982). Figure 7.10 shows a typical porphyritic andesite with plagioclase, altered hornblende and clinopyroxene phenocrysts.

7.2.1.2 Basin filling geometries

Within the Logudoro Group, andesitic volcanic centres crop out over scales of 10-20km (e.g. Seda Oro, Fig. 7.1, Coulon 1977; Osilo area, Servizio Geologico, sheet 180) whilst voluminous rhyolitic-dacitic pyroclastic products are much more widespread (Fig. 7.1). The source of the ignimbrites is uncertain (Lecca et al. 1997 also), for example no large caldera structure has been found. Coulon (1977) suggests that they may result from fissural eruptions. There is no clear evidence for fissural eruptions and since the andesitic volcanic centres represent the ‘roots of an ancient volcanoes’ (Coulon 1977) it is possible they erupted both andesitic and dacite-rhyolitic products such as are observed at other continental arcs (Cas and Wright 1988; Wilson 1993).

Published cross-sections show that whilst volcanic constructions have ‘domal’ geometries and some pyroclastic units thin from volcanic centres, the Logudoro Group forms an overall conformable, parallel-bedded succession which does not contain major, tectonically-derived angular unconformities (Coulon 1977; Servizio Geologico, sheet 193). This indicates that the Logudoro Group infilled the accommodation space created by earlier faulting and that the pre-late Burdigalian normal faulting
(7.1.3.2), which defined sedimentary depocentres, occurred after the eruption of the Logudoro Group (i.e. <~18 Ma). Andesite domes associated with the latest stages of volcanism are clearly aligned on an ENE-WSW trend, northwest of Coissione (Servizio Geologico sheet 193, Lecca et al. 1997). The implication is that NE-SW faults may have moved during the latest stages of volcanism and facilitated a path to the surface for andesitic magmas (Lecca et al. 1997 also). At the present day, the entire volcanic succession dips gently in a variety of orientations as a result of post-depositional faulting (Servizio Geologico, sheet 193) and/or compaction.

The rock types and rock arrangements suggest that from the mid-Oligocene-mid Burdigalian, the Logudoro study area was a volcanic terrain consisting of stratovolcanoes (sensu Cas and Wright 1988) built dominantly from andesitic flows, domes and breccias and surrounded by rhyolitic-dacitic pyroclastic fall, flow and surge deposits. The Logudoro Group accumulated within the ~N-S trend of the ‘main Sardinian Rift’ and continued northwards into the Anglona study area (chapter 8, Tergu Formation). The location and thickness of the volcanic rocks indicate that a ~N-S graben had formed by the mid Oligocene. The volcanic centres were most probably significant topographic constructions and voluminous pyroclastic material may have filled the majority of the tectonically derived accommodation space (as in Sarcidano sub-basin, chapter 5). This meant that the ?late Oligocene-Aquitanian transgression recorded in other parts of the Sardinian Rift (Sarcidano, Funtanazza, Anglona) did not reach this area.

7.2.2 Oschiri Formation (?Late Oligocene-earliest Burdigalian)

In the Logudoro study area, the Oschiri Formation crops out in only a few localities, stratigraphically above the Logudoro Group and, where exposed, is less than twenty metres thick. The finely laminated lacustrine marls, limestones with pisolithes and silicified plant fragments, cherts and tuffs are thought to have been deposited in lakes formed in ‘sags between stratovolcanoes’ (quote from Martini et al. 1992; Pomesano Cherchi 1971a; Cherchi 1974; Servizio Geologico Sheet 180 and 193).

Four kilometres NNW of Florinas [700033], the Oschiri Formation consists of variably chertified, finely laminated, grey tuffs (Fig. 7.11), tens of centimetres thick chertified white ash with occasional biotite and carbonised wood fragments, and grey marly beds with desiccation cracks (Fig. 7.12; Lithofacies 1h, 1j 5g, Tables 8.4, 8.5). Six kilometres SSW of Florinas [704964] and [680965], finely laminated, volcanic derived sediments with rare low angle cross-bedding and metre thick chert beds crop out (Lithofacies 1j, 5e). These sediments are indicative of lacustrine sedimentation where much of the material supplied came from erosion of the surrounding volcanic massifs and/or directly from fall out of volcanic ash. Drying up of the lakes occurred periodically.
7.2.3 Chilvani Formation (mid-late Burdigalian)

The Chilvani Formation is poorly exposed in the Oschiri-Sassari sub-basin intersection zone (Fig 7.4) where it lies conformably beneath the sediments of the Florinas Group. No continuous type section crops out, so that 'type outcrops' comprising non-marine sediments [878884], massive sandstones [872913] and marine sediments at the transition to the Florinas Group [873887] have been designated. The maximum exposed thickness of the Chilvani Formation is ~60m.

Fluvial-marginal marine sandstones and conglomerates classified here as the Chilvani Formation are thought to have been deposited in environments from alluvial fans and braided fluvial systems (in the northeast, Oggiano et al. 1995; Funedda et al. 1997) to deltas (marginal marine facies, Servizio Geologico sheet 193; Pomesano Cherchi 1971a; Martini et al. 1992) and barred shorelines (Martini et al. 1992). Non-marine, clast and matrix supported conglomerates, gravelstones (e.g. Fig 7.13) and coarse sandstones are massively or parallel-bedded (Lithofacies similar to If, Table 8.4). The sediments are moderately sorted with sub-angular to rounded clasts. Rare clast imbrication [878884] and cross-bedding [837876] indicates westward to northwestward palaeocurrents (Figs. 7.1, 7.4). Pomesano Cherchi (1971a) recognised that the clastic material had a provenance from granitic rocks present to the northeast. Palaeozoic metamorphic plus Oligo-Miocene volcanic clasts are also found. North of Ardara [824983], parallel stratified coarse sandstones contain a discontinuous patch, ~1m thick, containing a high percentage of organic material, indicative of local vegetation cover.

Outcrops along the Mores-Ardara road e.g. [872913] are characterised by massively bedded, coarse, granite derived sandstones and gravelstones with occasional pebble strings (Fig. 7.13) and poorly defined low angle cross-bedding. It is unclear exactly what processes, in what environment, the massive units formed. If these sediments are the products of debris flows, considerable reworking occurred before redeposition. Passing upwards, the percentage of marly/lime mud matrix increases and a bioturbated calcarenite to calcirudite containing shell fragments and whole echinoids (Clypeaster sp.) marks the contact with the Florinas Group (Fig. 7.14).

In summary, the basal, eastern outcrops of the Chilvani Formation were deposited by fluvial/alluvial fan systems which supplied clastic material westwards. The genesis of overlying massive sandstones is unclear, but since marine transgression is observed in western, upper parts of the Chilvani Formation it is likely that they represent fluvio-deltaic sediments, in agreement with published work (Servizio Geologico sheet 193; Pomesano Cherchi 1971a; Martini et al. 1992).
7.2.4 Florinas Group (late Burdigalian-Serravalian)

It has long been recognised that the marginal marine and marine sediments termed here the Florinas Group consist of 'heterogeneous banks' of carbonates, marlstones, marly sandstones, massive and cross-bedded sandstones (Pomesano Cherchi 1971a). This section attempts to rationalise the complex lithofacies arrangements by describing the nature and geometries of the Florinas Group in different zones of the Logudoro study area (defined on Fig 7.14). A tectono-stratigraphic synthesis for the whole area is then presented in section 7.4 using the temporal data discussed in Appendix 3D. A Florinas Group type section up the slopes of Monte Santo [796936] to [810920] has been chosen since it contains many of the typical lithofacies. The Florinas Group is partly equivalent to the Laerru Formation found in the Anglona study area to the north. For example, the Langhian carbonates and marlstones around Sassari (section 7.2.4.5) form a continuous exposure with the topographically lower, upper Burdigalian succession of the Sennori Member (section 8.4.7.6).

7.2.4.1 Previous Work

Several attempts have been made to describe and rationalise the Florinas Group. Pomesano Cherchi (1971a) produced five logs from the hillsides of Monte Santo and Monte Pelao (Fig. 7.1). Two carbonate levels were correlated using planktonic foraminifera. Between the carbonate levels, thick sandstones at M. Santo become replaced by marlstones and sandy marlstones towards the west (M. Pelao, Pomesano Cherchi 1971a). The variability in the sedimentary succession between Giave and Coissone was recognised by Assorgia et al. (1988). Three logs show the transgressive, fining upwards succession which passes through a conglomerate to biogenic limestones, thickest in the shallower, western part of the basin and passing upwards into fine calcarenites and marlstones, thickest in the east (Assorgia et al. 1988). Mazzei and Oggiano (1990) studied the complex area around Florinas (Fig. 7.1) where they recognised a succession comprising lower sandstones, lower limestones, many sandstones, unconformable upper sandstones and unconformable upper limestones. Transgression over the lower fluvio-deltaic sandstones resulted in the deposition of the lower, shallow marine platform carbonates which in turn were replaced by marly sedimentation as the basin deepened (Mazzei and Oggiano 1990). The upper sandstones overlie an unconformity resulting from "a general uplift" and the upper limestones mark a new marine transgression (Mazzei and Oggiano 1990). Whilst the publication provides lithofacies descriptions and age data, some stratigraphic relationships are not in agreement with field observations made in this study (below).

The most detailed work on the basin fill of the Sardinian Rift by Martini et al. (1992) describes the siliciclastic-carbonate rocks of the northern Logudoro study area and utilises the results of the Mazzei and Oggiano (1990) study. The sediments are divided into lithofacies (Table 7.3) and lithofacies architecture is considered within a simple 'sequence stratigraphic' framework. The publication describes a succession at the northern Ploaghe-Florinas basin margin where sandy carbonates pass
upwards into an offshore algal carbonate platform developed in response to rotation on a hypothesised, N-S trending 'main boundary fault' (Fig. 7.15). In the deeper parts of the Ploaghe-Florinas basin, laterally equivalent marly limestones were deposited (Fig. 7.15). After rapid regression, siliciclastic sandstones ("lowstand wedges") prograded over an erosional unconformity and a carbonate platform was once again established on the fault bounded high (Fig. 7.15). Within the Oschiri-Sassari sub-basin intersection zone, a mixed siliciclastic-carbonate succession developed in response to three transgressive-regressive cycles and to the supply of elastic material along the Oschiri sub-basin (Fig. 7.15; Martini et al. 1992). Siliciclastic rocks are replaced by basinward marlstones passing westwards (Fig. 7.15; Martini et al. 1992).

Whilst field studies corroborate the majority of direct observations made within Martini et al. (1992), the structural framework and hypothesised fault movement (Fig. 7.15) are not justified, no data is presented on the significant unconformities illustrated within the basin fill (e.g. on Fig. 7.15). The approach of correlating poorly dated, transgressive-regressive cycles to a 'eustatic' curve is not scientifically rigorous. The proceeding sections describe key field observations of the Florinas Group with the aims of building on the previous work and deducing a more rigorous tectono-stratigraphic development for the area (7.4).

### 7.2.4.2 Lithofacies

Table 7.3 summarises the main lithofacies which make up the Florinas Group (from Martini et al. 1992) and their equivalence to similar lithofacies found within the Sarcidano and Anglona study areas (chapters 5 and 8). The lithofacies of Martini et al. (1992) provide an accurate description of the rocks of the Florinas Group and, with the addition of calcirudites (Lithofacies 3g, Table 5.8), are used in the descriptions below (Figures 7.16 to 7.30 also).

In brief, marginal and shallow marine sandstones were deposited in fan delta, coastal plain and beach-bar type environments with clastic material supplied from the erosion of basement rocks. A range of calcarenites and marly limestones occurred in shallow marine settings, some directly adjacent to carbonate platforms and marlstones were deposited in deeper marine regions with water depths of a few hundred metres. Large scale (10-20m high) cross-bedded carbonates are common within the Logudoro study area. The cross-beds within these units dip at ~10-25°. In other areas of the Sardinian Rift (e.g. Is Paras Mbr, chapter 5) similar structures are interpreted as current-formed bars or carbonate sandwaves since the high angle cross-beds (up to ~30°) and swept-out, asymptotic toesets indicate strong palaeoflow. Whilst smaller scale (0.5-5m high) cross-bedding in the Logudoro study area were current formed, the larger, lower dipping structures apparently record the progradation of a carbonate unit into the deeper marine parts of the depocentre (Fig. 7.1).
In close proximity to the WNW-ESE trending Ploaghe fault, a late Burdigalian, shallow marine, mixed carbonate-siliciclastic system passes basinward and into a limestone-marlstone succession and is overlain by ?Langhian age carbonate-siliciclastic deposits.

Three kilometres south of the Ploaghe fault at Ardara e.g. [820920], less than ten metres of packstones, grainstones and reef framestones with *in situ* corals crop out (Lithofacies Ca, Cr, Table 7.3). These nodular weathering platform carbonates developed with a sharp basal contact over the calcarenites and calcirudites of the uppermost Chilvani Formation.

Cyclic siliciclastic-carbonate sedimentary rocks are exposed at a similar topographic level to the Ardara carbonates, 3km to the WSW of Ploaghe [750005] (Figs. 7.26, 7.31, 7.32; Lithofacies Gs, Sf, Cm, Table 7.3). The succession here consists of two, 6-10m thick, coarsening upwards sequences composed of marly sandstones with *Ophiomorpha*, passing upwards to cross-bedded and massive, coarse, quartzose sandstones with *Thalassanoides* and a well indurated, calcarenite bed capped with a horizon of whole *Clypeaster sp.* echinoids (Figs. 7.31, 7.32). The latter horizon is interpreted as marine flooding surface which caps the top of a regressive cycle. The cross-beds at this locality have asymptotic, swept out toesets (Fig. 7.31), progradational topsets (Fig. 7.26) and show palaeocurrents to the west and northwest, towards the Sassari sub-basin. The regressive cycles are formed from offshore marine sandstones and marlstones passing up to prograding shallow marine bar forms and shelfal sandstones. With renewed relative sea level rise, clastic sediment supply most probably decreased and calcarenites were deposited with a 'flooding surface' identified at their top before renewed marlstone deposition. These units have the characteristics of 'parasequences' *sensu* Van Wagoner *et al.* (1988). To the west and northwest of this outcrop, the succession consists of a basal, metre-thick, shell hash bed (Fig. 7.18, [733017]) overlain by ~3m of red algal and oyster grainstone and passing upwards to a marlstone succession with occasional calcarenite and wackestone beds (Lithofacies Ca, 3b, Mf). The sediments record marine transgression to an offshore marine setting and form the base of the sequence analysed at Florinas (below).

Marly lime sandstones (Sm) containing scattered metamorphic, granite, quartz and volcanic clasts and shell fragments are found immediately adjacent to the Ploaghe fault, 4km west of Ploaghe (e.g. [738030], Figs. 7.1 A-A', 7.27). The sedimentary rocks contain low angle laminations, plus trough and mounded bedding (Fig. 7.27) in cliff exposures perpendicular to high angle (20-30°) cross-beds. Whilst the low angle structures may be a record of storm events on the marine shelf (after Dott and Bourgeois 1982; Duke *et al.* 1991; Johnson and Baldwin 1997; in Martini *et al.* 1992) they are more probably sections perpendicular to bars or sandwaves which prograded into the basin from the Ploaghe fault topography. Thus, initial transgression in the Ploaghe-Ardara-Codrongianus area resulted in the development of a shallow marine carbonate platform around Ardara and mixed siliciclastic-carbonate
shelf deposits in areas of clastic supply. The shelf prograded to the west and northwest, towards the
deep marlstone dominated basin exposed around Codrongianus. (see Fig. 7.45c).

South of Ploaghe town and immediately adjacent to the Ploaghe fault [e.g. 785013], ~60m of
massively- and cross-bedded, coarse sand grade calcarenites and calcirudites (Fig. 7.16) are overlain
by ~10m of red algal, ripple bedded limestones with scattered granite clasts (Lithofacies Sqd, Gs, 3b.
3g Table 7.3, Fig. 7.1 B-B'). These rocks crop out at a topographically higher level than those
discussed above and may record a different phase of basin evolution (see Fig. 7.45d). The sediments
comprise poorly to moderately sorted quartz and granite clasts, red algae, bryozoan, oyster, coral and
shell fragments with a sparse lime mud matrix. In places, red algal bindstone patches a few tens of
centimetres across are exposed. Cross-bedding in the lower unit reaches up to 15m in thickness and
palaeocurrents taken perpendicular to the dip of the cross-beds indicate westerly to southwesterly
progradation directions. Smaller, asymmetrical, catenary out-of-phase ripples (after Collinson and
Thompson 1982) are observed on the exposed surfaces of the cross-bedded bars (Fig 7.33) and
indicate the same palaeoflow direction as the larger structures. The sediments south of Ploaghe may
represent the progradation of a number of ‘Gilbert’ type deltas into a shallow sea where small
carbonate build-ups formed on areas away from the active delta front or rimmed to the fan deltas. A
similar coexistence of coarse clastic and carbonate sedimentation can be observed in ancient settings
(e.g. Cretaceous, southern Spain, Garcia-Mondejar 1990) and modern day settings (Red Sea, Roberts
and Murray 1988, Purser et al. 1986). The progradation directions of the Gilbert deltas suggests that
they may have formed in response to the Ploaghe fault topography (i.e. footwall-derived sensu
Gawthorpe and Colella 1990) and to westward basin deepening. However, sediments were not derived
from erosion of the Ploaghe fault scarp but from reworking of pre-rift basement clasts and carbonate
material (formed in situ?).

7.2.4.4 Florinas area (Fig. 7.14)
Sediment architecture in the Florinas area is complex, with abrupt lateral and vertical facies changes
and discontinuous angular unconformities within the basin fill. The proposed tectono-stratigraphic
development (Fig. 7.34) accompanies this discussion.

Stage 1- late Burdigalian (NN4 zone, Fig. 7.34i)
The base of the succession comprises a series of marlstones with occasional wackestone beds between
Florinas and Codrongianus [722006] and occasional calcarenites south and east of Monte Mannu
[678974] (Lithofacies Mf, Cm). A typical log from south of M. Mannu (Fig. 7.35a) shows an
intercalation of marlstones with fine calcarenites containing transported shell fragments, in situ
echinoids and bivalves plus frequent bioturbation. Marlstones west of Florinas town [708006] contain
centimetre-sized, transported red algae and shell pieces which must have formed on an adjacent
carbonate platform. Apart from south of Monte Mannu where bed rotation and divergence are
observed in stratigraphically higher marlstones (Fig. 7.38 below), this offshore marine succession is a parallel-bedded unit, up to 100m thick (Fig. 7.36).

Stage 2 - earliest Langhian (Fig. 7.34ii)
Massive, parallel- and cross-bedded, well sorted, quartz rich sandstones-gravelstones with a lime mud matrix overlie the marlstone succession (Fig 7.25, 7.37, Lithofacies Sqd). The sediments were most probably deposited in marginal marine to shoreface environments with some prograding Gilbert type deltas. The thickness of this siliciclastic unit is extremely variable from ~10m south and southwest of Florinas town [720003, 720988, 725994], to ~100m north of Monte Mannu [682985], and to zero west of Monte Mannu (Figs. 7.1 B-B', 7.35). The basal contact of the unit, deemed to be a large scale, erosive, angular unconformity in Martini et al. (1992, see Fig 7.15), was not observed. The crucial stratigraphic relationship proposed by Mazzei and Oggiano (1990) was that the sandstones overlie thick ‘lower limestones’ northeast of Monte Mannu. However, the poor exposures in this area suggest that the so-called ‘lower limestones’ in fact overlie the sandstones and are continuous with the ‘upper limestones’ such as is clearly observed in the continuous limestone cliff section from west of Florinas [703004], to Carheghe [675017]. The western boundary of the thick sandstone is sharp (Fig. 7.38, Fig. 7.1 B-B') and to the west, marlstone deposition continued. The contact is thought to be the southward continuation of the Calancui N-S trending normal fault due the alignment of the structures and the syn-depositional, tectonic rotation of marlstone beds in the proposed fault hangingwall (Fig. 7.38). These divergent marlstone beds indicate a phase of normal fault movement not long before the transition to carbonate sedimentation in the early-mid Langhian, after sandstone deposition.

The cause of the abrupt thickness variations within the sandstone unit remain unclear. It is possible that a large angular unconformity developed on top of the marlstone succession due to subaerial exposure in the footwall of the Calancui fault, similar to the hypothesis of Martini et al. (1992). Alternatively, the sandstones may have accumulated in a tectonically formed accommodation space due normal faulting on a westward dipping ~ N-S structure and the ~E-W trending Mantedda fault (Figs. 7.34ii, 7.1 B-B').

Stage 3 - early/mid Langhian-Serravalian (Fig. 7.34iii)
The contact of the earliest Langhian sandstones with transgressive shallow marine carbonates is both conformable and an angular unconformity, dependent on location. A striking angular unconformity, downcutting to the northeast was developed at [698992], 2km southwest of Florinas (Fig. 7.39), but the cause of the unconformity is unclear. It was affected by a NW-SE striking normal fault (Fig. 7.39). The unconformity may have occurred in response to local, fault-associated tectonic uplift, either in the footwall of the same structure which perhaps caused the accommodation space for the sandstones, or a NE-SW trending fault identified tens of metres away from the unconformity outcrop (Fig. 7.1). At all other localities within Florinas area, the contact of the sandstones with the overlying platform
carbonates is conformable [698009, 720002, 708996, 718987] (Lithofacies 3b, Ca). It is gradational, with progressive red algae, oyster and echinoid colonisation, a corresponding decrease in clastic material and appears as a sharp change in induration (Figs. 7.35bcd, 7.37).

A thick carbonate platform unit characterised by wackestones, packstones, grainstones and red algal bindstones developed over the top of the sandstones in the north and west of the Florinas area (Lithofacies Ca). The Florinas platform carbonates are continuous with the slightly younger limestones observed in the Sassari area (below). Cross bedding, 10-20m high, which prograded southwards in the footwall of the Calancui fault and westwards in the fault hangingwall is observed (Fig. 7.1). To the west of the Calancui fault, south and west of Florinas and Monte Mannu, the sedimentary succession consists of deeper marine marlstones and calcarenites which surrounded the carbonate platform (e.g. Figs. 7.1 B-B', 7.35c).

7.2.4.5 Sassari area (Fig. 7.14)

An exceptional cliff face exposure south of Sassari (Fig 7.40) and road cuttings through the same succession just to the east (Scala Giocca, Fig. 7.1) provide insights into the aggradation and progradation of a mid Langhian-Serravalian carbonate platform (Lithofacies Ca).

The carbonates south and east of Sassari are thought to have accumulated in accommodation space created by the westward dipping, N-S trending Calancui normal fault. In the proposed footwall of the fault, tens of metres of shallow marine, red algae dominated carbonates were deposited on top of the Logudoro Group. Over the fault zone, angular unconformities developed within the basin fill (Fig. 7.5a) and the thickness of marine sediments increased to ~200m. The cliff face section lies within the hangingwall, 2km from the normal fault. Further evidence for the Calancui fault is the exposure of lacustrine sediments (Oschiri Formation) in the fault footwall to the south e.g. [684035] (Fig. 7.1, A-A').

The cliff face succession dips gently to the west. At the base of the cliff face, parallel-bedded marlstones pass upwards into a carbonate unit ~60m thick which contains syn-depositional normal faults, wide concave channels and a number of chaotic, slumped and slide zones. Coherent blocks are common within the slumped zones. Listric, syn-depositional faults crop out in cliff exposures to the south side of the same valley such that limestone beds in the hangingwall form a rollover anticline and diverge into the normal fault ('growth anticline', Fig. 7.41). A listric fault is probably also responsible for the growth anticline observed at the western end of the main cliff face (Figs. 7.40, 7.41). Listric faults and growth anticlines, slumps and slides such as these are commonly found within deltaic successions deposited on a slope (e.g. upper Triassic Svalbard, Edwards 1976; Gulf of Mexico, Roberts and Yielding 1994; Johnson and Baldwin 1996). Thus although the listric faults are oriented in a similar direction to the active N-S Calancui fault (7.1.3.2), they are not necessarily related to
extension. The top ~50m of the cliff face is formed by parallel-bedded units with subtle, low angle
downlap surfaces indicating platform progradation in combination with the dominant aggradational
signature.

The cliff-face exposure is interpreted as a regressive succession from mid Langhian marlstones
(Cherchi 1974) through slumped and syn-depositionally faulted limestones interpreted as platform
slope deposits to parallel-bedded, prograding packages on the platform top (mid Serravalian, 3.2.2).
The cliff face carbonate succession thins westward such that 5km from the proposed N-S fault
[617050] only a few tens of metres of limestones crop out, the majority of the succession being marly
limestone and marlstones (Fig. 7.1 B-B', Lithofacies Mf and Cm). The sedimentology of the adjacent
Scala Giocca section confirms the overall regressive trend. The basal limestone outcrops comprise
muddy lime sandstones with scattered broken shells, rhodoliths and bioturbation (Fig. 7.20). Further
up the Scala Giocca [658064], slumps are observed within muddy, bioturbated lime sandstones and
represent failure of an unstable platform slope (Fig. 7.22, 7.40). A regressive succession of lime
sandstones through parallel-bedded, rhodolith dominated carbonates and a massive reefal body is
exposed in outcrops on the mid-upper parts of the Scala Giocca ([657064] Fig 7.42). Finally, at the top
of the Scala Giocca [657066] parallel- (Fig. 7.21), cross-bedded and channelised rhodolith
conglomerates (each rhodolith 2-6cm diameter) represent a high energy depositional environment on
the carbonate platform top.

7.2.4.6 Southern and western margins of the Sassari sub-basin (Fig. 7.14)
In the footwalls of the Molafa and Ittiri faults (Fig. 7.4), the western Sassari sub-basin was filled by
bioclastic carbonates, calcarenites and lime sandstones. The late Burdigalian (3.2.2) transgression over
E-W and NW-SE faulted volcanic rocks is marked by a sharp contact over a lignite layer at [591994]
(Fig. 7.6) and [628945], suggesting vegetation cover before submergence.

At Molafa [595048], cross-bedded carbonates (Fig. 7.17) in the footwall of the Molafa fault occur
adjacent to a dominantly marlstone succession in the fault hangingwall (Figs. 7.24, 7.1 B-B') and
calcirudites in the fault zone (Fig 7.30, Lithofacies Ca, Mf, 3g). This trend is mirrored at the faulted
southern margin of the Sassari sub-basin between Ittiri and Banari. Here carbonates occur on normal
fault footwalls. In the hangingwalls they pass laterally basinward into marlstones (Fig. 7.1 B-B'). The
onlapping geometries and undeformed nature of the sediments indicate that the fault topography
existed before marine transgression (7.1.3.2). This topography obviously controlled the location
of shallow marine carbonates at the basin margin whilst marlstones were deposited in the Sassari sub-
basin centre. In this situation, it may be expected that carbonate platforms and reefs would have
formed on footwall highs and shed debris material basinward (e.g. Leeder and Gawthorpe 1987).
However bioclastic packstones were ubiquitous in all the areas examined (e.g. across the Brundette
fault [653954]). Large scale cross-bedding within the carbonates indicates eastward progradation
directions, towards the main Sassari sub-basin depocentre. Marlstones to the north of the southern basin margin are very similar to those described in the Florinas area (e.g. Figs. 7.23, 7.35a) though fine limey sandstones are common north of Ittiri [620000] (Fig. 7.28).

7.2.4.7 Oschiri-Sassari sub-basin intersection zone (Fig. 7.14)
The Miocene succession in this area is preserved underneath Pliocene alkali basalt caps such as Monte Santo and Monte Pelao and reaches 400m in thickness (Figs. 7.1 C-C', D-D', 7.44c). Late Burdigalian transgression (Cherchi 1974) over the volcanic Logudoro Group or Chilvani Formation is recorded as a calcarenite or carbonate bed, a few to tens of metres in thickness, which passes upwards into a marlstone succession (Lithofacies Ca, Cm, Mf; Figs. 7.43, 7.44bc, 7.1 C-C' D-D'). In the eastern part of the zone, the basal units are overlain by tens of metres of ?early Langhian, massive, granite-derived sandstone and gravelstones with rare granite pebbles (Sqd; Figs. 7.43, 7.44c). The clastic sediments are probably correlatives of, and were deposited in, a similar marginal marine environment to the thick sandstones found west of Florinas (7.2.4.4). To the west, for example around Thiesi and Monte Pelao, marlstones with occasional calcarenite bands are laterally equivalent to the clastic sediments (Lithofacies Mf, 3b; Pomesano Cherchi 1971a also). They were deposited in deeper waters away from the clastic sediment supplied along the Oschiri sub-basin (Martini et al. 1992 also). In the mid Langhian, renewed relative sea level rise resulted in the deposition of a second carbonate, ~10m thick, and marlstone succession in the east of the area (Ca, 3b, Mf; Fig. 7.43; top N8 zone, Pomesano Cherchi 1971a). At Mores [865903] and Monte Santo [810923], late Langhian-early Serravalian (section 3.2.2, Appendix 3D) carbonates and calcarenites form a thicker, third carbonate, calcarenite and marly limestone unit at the hill tops (Lithofacies Ca, Cr, 3b, Cm; Figs. 7.19, 7.1 C-C', D-D').

Sediments within the Oschiri-Sassari intersection zone indicate that from the late Burdigalian-early Serravalian a shoreface existed in the east of the area and that the depocentre deepened to the west. Siliciclastic-carbonate and marlstone cycles resulted from variations in relative sea level and clastic sediment supply which was dispersed along the existing transtensional sub-basin topography. The conformable sedimentary succession is flat lying or gently dipping with lateral facies changes. No evidence was observed for syn-depositional tectonism. Post depositional NNW-SSE, N-S and ENE-SSW faulting offsets and tilts the sediments.
7.2.6.8 South central Logudoro

A simple, late Burdigalian-?mid Langhian transgressive sequence crops out in south central Logudoro. It comprises an occasional 20-40cm thick basal conglomerate, bioclastic carbonate succession of variable thickness and deeper marine marlstones (Lithofacies Ca, 3b, Mf; Figs. 7.44d-g, 7.1 E-E’). Calcarenites evolving to packstones and wackestones commonly make up the prominent weathering bioclastic carbonates (Fig. 7.44dg; Lithofacies 3b, Ca, Cm). These carbonates are sometimes cross-bedded on a metre scale (e.g. northwest of Bonorva [761756]) or on a 10-15m scale (e.g. south of Romana [647796]).

The influence of a remnant volcanic topography on sedimentation can be observed in south central Logudoro. For example, north of Mara [680756], parallel-bedded, flat lying marlstones (Mf) surround andesitic cones related to the last phase of Logudoro Group volcanism (Figs. 7.9, 7.1 E-E’). At S.M. Iscalas [732776], a similar geometry is observed but here a succession of grainstones, packed calcirudites and packstones surrounds the andesitic cone (Lithofacies Ca, 3g; Figs. 7.44e, 7.1 E-E’). These Iscalas limestones were deposited in a shallow marine, high energy environment. As they currently crop out at 600m, they may either have been post-depositionally elevated in the footwall of the Iscalas fault or, if the present day topography reflects the Miocene palaeotopography, they represent equivalents of marlstones deposited in deeper parts of the basin (e.g. Giave [777788], Fig. 7.44d; Assorgia et al. 1988).

In the Romana area, sediments clearly lie within a fault defined depocentre. In the hangingwall of the Monteleone fault, ~60m of marlstones overlie a metre to 10m thick oyster to bioclastic carbonate bed (Lithofacies Ca, Mf; Fig 7.44g, 7.7, 7.1 C-C’), whilst in the fault footwall, ~60m of parallel- and cross-bedded carbonates are found (Ca, Monteleone Roccadoria, [627808]). Calcarenites and calcirudites with volcanic lithic clasts, thick shelled oysters (Fig. 7.29, Lithofacies 3b, 3g) and intercalated tuff beds occur in the small half graben defined by the Temo and NE-SW trending fault at [605806] (Fig. 7.44f). The sediments show clear onlap to the Logudoro Group to the west and north (Fig. 7.1 C-C’).
7.3 Timing of extension

7.3.1 Previous Work

The presentation of evidence for extensional events in the Logudoro study area is rare. Oggiano et al. (1995) and Carmignani et al. (1994, 1995) believe that the main N-S Sardinian Rift formed wholly on Burdigalian age normal faults which cut across the older Oligo-Aquitanian transtensional basins (Fig. 7.2). No data is given to show the actual fault locations or timing. Activity on NE-SW faults during marine Miocene sedimentation is mentioned by Assorgia et al. (1988) in the Giave area. Funedda et al. (1997) give the most complete account, describing E-W and NNW faults which formed a 'pre-Burdigalian' topography which was filled with 'post-mid Burdigalian' sediments. Minor Langhian movements were related to the last phases of extension (Funedda et al. 1997).

7.3.2 Timing and nature of extension in the Logudoro study area

The mid Oligocene-mid Burdigalian Logudoro Group accumulated within the main N-S trending part of the Sardinian Rift basin. The volcanic rocks exposed at the surface onlap the faulted, eastern rift margin after N-S and NW-SE faulting had occurred and do not exhibit evidence for syn-eruptive tectonism within their geometries. Similar volcanic products are found along the strike-slip basins where they are intercalated with sediments (chapter 6, Oschiri Fm).

Thus, although the main N-S Sardinian rift structure is poorly defined, it clearly existed in some way before the mid-Oligocene and had formed by the time the majority of Late Oligocene-early Burdigalian volcanic rocks were erupted. The transtensional basins which intersect with the eastern rift margin also existed at this time though strike-slip fault movement also occurred in the latest Aquitanian (chapter 6). Based on field evidence, it is unclear how the mid-late Oligocene NE-SW to E-W trending 'strike-slip' faults and ~N-S trending 'main Sardinian rift' fault sets interacted but they may well have been contemporaneous.

Basin filling geometries show that extension on E-W and NW-SE to N-S oriented faults formed sedimentary depocentres overlying the Logudoro Group after the end of volcanic eruption (mid Burdigalian) and before late Burdigalian marine transgression (i.e. at 17-18 Ma). It is again unclear whether the two fault sets moved independently, in separate events, or synchronously. This major extension phase occurred over a short (<1 Ma) time period, just after the end of subduction-related volcanism and the rotation of the Corsica-Sardinia microplate (section 2.1). The possible tectonic regimes responsible for the two fault sets and significance of the extension phase are discussed in chapter 10.

Localised syn-depositional faulting within the basin fill occurred on some ~N-S trending normal faults in the early/mid Langhian (e.g. Calancui fault, M. Mannu, Fig. 7.38), mid-late Langhian (Calancui
fault, east of Sassari, Fig. 7.5a) and sometime within the late Burdigalian-Langhian (Molafa fault and north of Ittiri, Figs. 7.5c, 7.1 B-B'). This minor phase of fault activity, after Corsica-Sardinia microplate rotation, probably occurred in response to an east-west oriented extensional stress field. There is no evidence for syn-depositional fault activity on east-west trending structures.

7.4 Tectono-stratigraphic development

Tectono-stratigraphic models for the development of the Logudoro study area, central eastern rift margin and the Oschiri transtensional sub-basin (chapter 6) are given below (Figs. 7.45, 7.46).

7.4.1 Mid Oligocene-early Burdigalian (Fig. 7.45a) Volcanism in the main Sardinian Rift contemporaneous with strike-slip fault movement.

Basic volcanism commenced at ~28 Ma in the Logudoro study area (Beccaluva et al. 1985). A >500 m thick succession of basaltic, basaltic andesite and andesitic lava flows, dome and breccias plus voluminous rhyolitic-dacitic pyroclastic deposits (Logudoro Group) was erupted until the mid Burdigalian (~18 Ma, section 3.2.3 and Odin et al. 1994). Towards the top of the volcanic succession and in the Oschiri transtensional sub-basin, volcanic rocks were intercalated with fluvio-lacustrine sediments (Oschiri Formation). The Logudoro Group rocks infilled the main Sardinian Rift basin which had already been defined by ~N-S trending normal faults. In the Oschiri sub-basin, the late Aquitanian Oschiri Formation was deposited contemporaneously with a phase of movement on the bounding strike-slip faults (chapter 6). Younger volcanic rocks covered topography created by movement on NE-SW trending strike-slip faults (chapter 6). The Logudoro Group volcanic constructions were probably topographic highs. The Late Oligocene-Aquitanian marine transgression observed in other areas of the Sardinian Rift did not reach the Logudoro study area.

7.4.2 Mid Burdigalian extension (Fig. 7.45b)

Between the end of Logudoro Group volcanism at ~18 Ma and late Burdigalian marine transgression at ~17 Ma a phase or phases of extension created a number sub-basins on top of the Logudoro Group. Extension occurred on E-W and NW-SE to N-S trending normal faults though the relationship between the fault sets is not clear.
7.4.3 Late Burdigalian; marine transgression and mixed carbonate-siliciclastic sedimentation (Fig. 7.45c)

In the late Burdigalian, marine transgression occurred over the mid Burdigalian faulted topography. Non-marine clastic sediments (Chilvani Formation) derived from the erosion of the Palaeozoic basement and Oligo-Miocene volcanic rocks were supplied along the Oschiri sub-basin and passed laterally into marginal marine and marine sediments (Florinas Group) in the Oschiri-Sassari sub-basin intersection zone. The transgressive Florinas Group typically comprises a basal, shallow water calcarenite/carbonate bed rich in red algae and oysters passing upwards to deeper marine marlstones. The sediments progressively onlapped the faulted and remnant volcanic topography and began to fill the fault-formed depocentres. Between Ploaghe, Ardara and Codrongianus lateral facies variability can be observed. A carbonate platform developed in shallow waters away from clastic supply at Ardara whilst mixed siliciclastic-carbonate sediments were deposited on a shallow marine shelf west of Ploaghe. The shelfal sediments consists of cyclic, cross-bedded and massive sandstones capped by calcarenites and marly limestones related to marine flooding plus cross-, trough- and mounded-bedded marly lime sandstones. Further towards the Sassari sub-basin (Codrongianus-Florinas), marlstones with rare wackestones and calcarenites were deposited in deeper marine conditions.

7.4.4 Lower Langhian; siliciclastic sedimentation (Fig. 7.45d, 7.46)

Within the eastern part of the Oschiri-Sassari intersection zone and west of Florinas, 10-100m of marginal marine coarse quartzose sandstones and gravelstones were deposited over the transgressive carbonate-marlstone series. The influx of clastic material and change to marginal marine conditions in some parts of the Logudoro study area can most easily be related to relative sea level fall (e.g. Mazzei and Oggiano 1990; Martini et al. 1992). The clastic material within the Oschiri-Sassari intersection zone was clearly supplied along the Oschiri sub-basin. Based on their composition, a similar source seems likely for the Florinas sandstones. In the Florinas area, it is unclear what caused the rapid thickness variations observed within the sandstone unit. The N-S trending Calancui fault, to the west of which no sandstones are found and which exhibits early/mid Langhian movement (Fig. 7.38), clearly played a major role. Whilst the sandstones may overlie an erosive angular unconformity, stratigraphic and structural relationships are not consistent with the location of the fault-block model of Martini et al. (1992; Fig. 7.15). Marlstone and carbonate sedimentation continued in other areas of the Logudoro study area. Shallow marine carbonates developed on fault-created topographic highs at the margins of sedimentary depocentres and prograded into deeper waters dominated by marlstone deposition, dependent on the local palaeotopography.
7.4.5 Mid Langhian-Serravalian; renewed transgression, carbonate-marlstone sedimentation and regression (Fig. 7.45e)

The lower Langhian sandstone units were capped by marine calcarenites, carbonates and marlstones resulting from mid Langhian (N8, NN5 zone) transgression. Movement on the Calancui fault is identified in the mid-upper Langhian due to angular unconformities developed with the basin fill (Fig. 7.5a). In the fault footwall west of Florinas, a thick mid-?upper Langhian carbonate platform developed. In the fault hangingwall, mid Langhian marlstones pass upwards to late Langhian-Serravalian carbonate slope and platform top deposits resulting from carbonate platform progradation and/or relative sea level fall. Away from the palaeotopographic high associated with the Calancui fault and carbonate platform, limestones pass laterally to deeper marine marlstones. At Ploaghe, large scale prograding calcarenite/calcirudite bars formed in a shallow sea with coarse clastic material and carbonates produced \textit{in situ}. In the Oschiri-Sassari intersection zone (at Monte Santo, Mores), mid Langhian carbonates and marlstones pass upwards into a late Langhian - early Serravalian carbonate dominated succession.

7.5 Summary

- Mid Oligocene-early Burdigalian volcanic rocks were erupted into the N-S trending Sardinian rift and NE-SW transtensional basins at the eastern rift margin (chapter 6). Late Burdigalian-Serravalian sediments filled mid Burdigalian depocentres created by E-W and NW-SE to N-S oriented normal faulting of the volcanic rocks. N-S trending normal faults were active in the Sassari sub-basin at times between the late Burdigalian-late Langhian.

- Sedimentation over the faulted topography began with the non-marine-marginal marine, clastic, mid-late Burdigalian Chilvani Formation which was derived from the weathering of the Palaeozoic basement and dispersed along the Oschiri sub-basin. The late Burdigalian-Serravalian Florinas Group comprises laterally variable sandstones, carbonates and marlstones which developed in response to clastic sediment supply, local tectonism and relative sea level change superimposed on the existing fault palaeotopography. Initial late Burdigalian transgression resulted in a widespread calcarenite-carbonate-marlstone succession. Lower Langhian regression was accompanied by the deposition of thick sandstones in the Oschiri-Sassari intersection zone and Florinas area. Carbonate and marlstone sedimentation was re-established by mid Langhian transgression and continued until the Serravalian.
Chapter 8
Chapter 8 - The Anglona study area and Miocene basins of northernmost Sardinia and Corsica

The Anglona study area (Fig. 1.3) formed by three phases of extension during the Oligo-Miocene and was affected by two marine transgressions. Coupled with voluminous andesitic-dacitic eruptions, this lead to the deposition of a complex non-marine to marine, volcanic, clastic and carbonate basin filling succession. The Anglona study area provides particular insights into syn- and post-rift basin filling where a large proportion of the basin fill was volcanic material and contains some of the most impressive syn-rift deposits in Sardinia. This chapter describes the nature of structures defining the Anglona study area, the geometry of basin filling units linked to the timing of extension and the sedimentology and volcanology of the basin filling units defined in chapter 3. This data is used to reconstruct a tectono-stratigraphic evolution for the area. A brief description of the basin filling units, structures and possible tectono-stratigraphic evolution of related Miocene basins of northern Sardinia and Corsica is given in this chapter since it is necessary to incorporate these areas in any regional model (chapter 10).

Enclosure 5 is a 1:25000 map of the Anglona study area, summarised on Figure 8.1. The stratigraphy of the Anglona study area is divided into the basal, transgressive-regressive clastic and volcanic Elefante Formation (Casteldoria, Valledoria, Vaginella and Castelsardo Members) and laterally equivalent volcanic Tergu Formation. The Elefante Formation is capped by an areally extensive ignimbrite of the Tergu Formation (τ2). The τ2 marker is overlain by the lacustrine Perfugas Formation and transgressive Laerru Formation which consists of shallow marine carbonates, calcarenites and marlstones overlying a basal conglomerate (Sedini, Sennori, Martis and Campulandru Members, see sections 3.3.5, Appendix 3D, Figs. 3.17, 3.18).

8.1 Previous Work

Three logged sections in the vicinity of Castelsardo provide the basis for the majority of published work on this area (Maxia and Pecorini 1969; Spano and Asunis 1984; Francolini and Mazzei 1991). The focus of these papers was biostratigraphic analysis, though there are limited descriptions of the basin fill, a geological map (Spano and Asunis 1984) and an appreciation of lateral facies variability (Maxia and Pecorini 1969). Seismic reflection profiles shot offshore the Anglona study area were interpreted by Thomas and Gennesseaux (1986) and provide a useful suggestions for the large scale structure and basin filling geometries. Their seismo-stratigraphic to onland correlations and temporal constraints must be treated with caution since no well data exist. Seismic reflection lines interpreted by Thomas and Gennesseaux (1986) show that, just offshore from the Anglona study area, a ~23 km wide N-S/NNW-SSE trending half-graben with a fill of up to 2.5s two-way travel time (TWTT, ~1.9 km
assuming seismic velocity 1.5 km/s) exists, and that extension occurred at two times (Oligocene-mid Aquitanian, Seq. 1,2 and Burdigalian Seq. 4 suggested; Fig. 8.2). A French thesis provides valuable sedimentological descriptions of the eastern part of the Anglona study area (Quesney-Forest and Quesney-Forest 1984). The far west of the Anglona study area was mapped by Oggiano (1987) and is further constrained by the biostratigraphic study of Francolini (1994). An integrated, tectono-stratigraphic model for the evolution of the Anglona study area has not before been deduced.

8.2 Structural geometries observed within the Anglona study area

Field observations and seismic reflection data from just offshore show that Anglona study area has a complex structural history of extension (Fig. 8.1, 8.3). The area can be divided into three ‘sub-basins’ defined by different fault styles and consequent basin filling histories. The three sub-basins can be defined as (Fig 8.3):

- **the Castelsardo sub-basin**, a 23 km wide, NNW-SSE trending half-graben lying to the north of the San Giovanni and San Maria Coghinas faults. Smaller, ~0.5 km wide half-grabens and tilted fault blocks were defined by E-W to NW-SE normal faults.

- **the Perfugas sub-basin**; a synclinal basin of triangular shape defined by the WNW-ESE Ortigu, NE-SW Perfugas and Tula (chapter 7) faults. The sub-basin was developed after the eruption of the t2 ignimbrite, and was bordered in the west by the ‘Tergu platform’ in the footwall of the San Giovanni fault and the Sedini north-south fault.

- **the Portotorres sub-basin**; a ~21 km wide half-graben bounded by the NNW-SSE trending Tramontana and NE-SW Sennori faults, extending to the west of the Anglona study area.

After using the offshore seismic reflection data to define the large scale structure of the Castelsardo and Portotorres sub-basins, this section describes the field evidence for, and the geometry of, the normal and oblique-slip faults defining the three sub-basins which make up the Anglona study area.

8.2.1 Seismic reflection profile interpretation - structures

British Petroleum kindly lent seismic reflection lines shot in the Gulf of Asinara for AGIP in 1968. Two lines (125 and 123) which run sub-parallel to the coast (at 7 km and 20 km respectively, Fig. 8.1) provide useful information on the structures and basin filling geometries with the Castelsardo and Portotorres sub-basins (Fig. 8.4, Encls. 6,7). The results are in general agreement with Thomas and Gennesseaux’s (1984) interpretation of these seismic reflection lines (Fig. 8.2) though not all their basin-filling megasequences are visible.
The data shows that, just offshore, the Castelsardo sub-basin is a ~23 km wide, tilted half-graben structure, 2-2.5 seconds (~1.9 km) deep, bounded by the Viddalba (eastern) and Tramontana (western) faults. The hangingwall dip slope of the half-graben structure is complicated by smaller 1-4 km, scale fault bounded blocks with throws of 0.1-0.4 seconds (~75-300m). The Portotorres sub-basin also has a half-graben structure, 2.5s (TWTT, ~1.9 km) deep and bounded in the east by the Tramontana fault. The normal faults bounding the pre-rift basement are high angle and planar though the dip-angles change slightly with depth (e.g. Viddalba fault on line 125, shallower with depth, Tramontana fault on line 125, steeper with depth) and high angle faults splay off into the basin fill.

FIELD OBSERVATIONS
8.2.2 Previous work
In contrast to Sardinian workers, Quesney-Forest and Quesney-Forest (1984) were not 'fault shy' and their map of the Perfugas and eastern Castelsardo sub-basins contains abundant normal, strike-slip and thrust faults. Some of the faults clearly exist (e.g. Sedini, Bulzi and Concatile post-Miocene normal faults) whereas there is no evidence for some of the others, or they may exist but have small throws (<10m). In particular, there is little data to support a 'Messinian' compressional phase producing 8 km long, NE-SW thrust faults or a series of kilometre long, north-south strike-slip faults. 'Microtectonic' studies from one or two localities which do show compressional structures (a fold and reverse slip slickensides on centimetre scale faults) cannot be used to simply define the structural history of such a complex sub-basin (as in Quesney-Forest and Quesney-Forest 1984). Spano and Asunis (1984) indicate some 'presumed' faults on their geological map of the Castelsardo town region but make no comment on them. Many of the 'presumed' structures are clearly normal faults (below), though there is no evidence for some others. The margin of the Portotorres sub-basin was mapped in detail by Oggiano (1987) and he identified to high resolution, several fault strands relating to the Tramontana and Sennori fault systems.

8.2.3 Castelsardo sub-basin
Table 8.1 summarises the evidence used to define larger scale, ?post mid-Oligocene faults in the Castelsardo sub-basin.

8.2.3.1 Eastern margin of the Castelsardo sub-basin
The eastern margin of the Castelsardo sub-basin is clearly defined by the NNW-SSE trending Viddalba normal fault which bounds high, pre-rift basement mountains and the flat Coghinas plain. The fault cuts sinistral strike-slip faults identified in the eastern basement rocks (Barca et al. 1996; Fig. 8.1, 8.3). The throw on the NNW-SSE trending Viddalba fault decreases in the vicinity of Casteldoria, where several high angle fault strands with throws of tens of metres can be identified (Therme Fig 8.5 st1, Fig. 8.6; Ruiu, Casteldoria faults).
The Casteldoria granitic and metamorphic rocks crop out as an E-W trending horst (the Casteldoria horst), bounded by the Ortigiu fault in the south (below) and by the proposed San Maria Coghinas fault in the north (Fig. 8.3). The San Maria Coghinas fault may represent the offset tip of the large NE-SW trending, post Hercynian sinistral strike-slip fault identified by Barca et al. (1996, Fig. 8.1) in the Palaeozoic basement to the east. On a smaller scale, the Casteldoria horst block is cut by faults with throws of ~10-20m (e.g. Scalitta and 2 almost vertical faults at [905274]) which trend NE-SW and WNW-ENE. Along with the Casteldoria fault, these smaller structures clearly formed the accommodation space and controlled the extent and dispersal of the Casteldoria member sediments which were shed northwards (section 8.4.1) from the horst (Fig. 8.39 below). Measurements of small faults and fractures within the granitic and metamorphic rocks in the Casteldoria area generally gave a random scatter. However, NW-SE trending faults had a dextral strike slip sense where slickensides were found (2 readings) and could have been related to the larger, NE-SW trending sinistral strike-slip systems (1 outcrop scale reading).

8.2.3.2 The San Giovanni and Ortigiu faults within the Castelsardo sub-basin

The San Giovanni fault, although not directly observed, is thought to run east-west, dipping to the north and defining the southern margin of the Castelsardo sub-basin (Fig. 8.1 A-A’). It is important to characterise this structure since it divides the largely unfauloted Tergu Platform to the south and the complexly faulted Castelsardo town area to the north. Several observations point towards the existence of a ~15 km long, high angle normal fault: the offset of the τ2 ignimbrite by ~50m, the offset of ignimbrites (Lithofacies 5j) within the Vaginella Member at [858270], the presence of numerous outcrop scale E-W to NW-SE (western areas) trending normal faults (Fig. 8.5 St2-5) including one ~20m wide deformed fault zone at [755273] (Fig 8.8).

Though evidence was sought, the structure of the zone bounding the western end of the Casteldoria horst is unclear. The 110° trending, Ortigiu normal fault (below) appears to bound the southern horst e.g. [895273]. Based on evidence such as syn-sedimentary faulting, post-depositional faulting and abrupt dip variations within the Valledoria and Castelsardo Members, the Ortigiu fault may continue for ~4 km into the Casteldoria sub-basin. The oppositely dipping E-W San Giovanni fault bounding the platform to the south appears to 'tip out' towards this zone giving a geometry similar to an 'overlapping, antithetic, interference zone' (sensu Gawthorpe and Hurst 1993; Fig. 8.12).
8.2.3.3 Castelsardo town area

The area between Castelsardo and the San Giovanni fault was cut by normal faults of several orientations (Fig. 8.3). The dominant structures are a set of curved faults trending NW-SE to E-W defining a series of tilted normal fault blocks and half-graben, 0.5-2 km wide and offsetting τ2 with throws of 40-100m (Castelsardo, Lu Pozzu, Fragiagu, Oschiri, Bagliaoglìa, Paligheddu faults; Figs. 8.1 A-A', B-B', 8.5 St6-9, 8.9). Tilting of the fault blocks increases to the north (e.g. 6-7° at [775271] to 15° at [758296]) and the small fault blocks seem to ‘roll-over’, perhaps into a larger structure lying just offshore (Fig. 8.1 A-A', B-B'). The exposed Castelsardo fault plane [756294], is made of a series of high angle, planar fault strands containing deformed Castelsardo Member sediment slices (Fig. 8.5 St6). The present day topography exhibits a typical degraded fault block geometry (Fig 8.1 B-B'). In the hangingwall of the Castelsardo fault, the τ2 ignimbrite is folded into an asymmetric syncline (Fig 8.1 A-A', B-B') thought to develop as a result of fault propagation folding (sensu Withjack et al. 1990; Gawthorpe et al. 1997; Sharp et al. in press) through the ignimbrite layer and/or later drag along the fault.

NNW-SSE to NNE-SSW trending, high angle normal faults form the other main fault family in this area (N-S fault family). These faults probably reflect the same extensional events which caused the NNW-SSE Castelsardo half graben and may be equated to the surface expression the smaller faults identified on seismic reflection profiles (section 8.3.2). Whilst many of the N-S fault family have small offsets and are post-depositional (?Pliocene), the Castelsardo east fault can be traced for 5 km and may show subtle syn-depositional movement (Castelsardo Member, 8.4.3.1 below). The geometry of the fault is shown on Fig 8.10, the fault is almost vertical (appears slightly curved from this viewpoint) with a ~2m wide breccia zone exposed on wave cut platform. Whilst the footwall is largely undeformed, the hangingwall is cut into a series of blocks by smaller faults (Fig. 8.10). Fault and fracture relationships on the wave cut platform indicate that the last phase of extension occurred on N-S rather than E-W trending faults. Syn-and post-depositional outcrop scale faults on a ~N-S orientation are found within the Valledoria Member in the west of the sub-basin (Fig 8.5 St4). Post-depositional ~N-S faults are common within the Vaginella Member (Fig. 8.5 St5) and Castelsardo Member (Fig. 8.5 St7,8). Very high angle normal fault planes (70-90° at [785266] (Fig. 8.5 St5) with throws of 40cm-1m are characteristic of the post-depositional structures.

A set of NE-SW trending normal faults can also be identified within the Castelsardo town area. To the south-east of the town, such faults post-depositionally offset the τ2 ignimbrite with throws of 20-40m (e.g. Pt. Molino area, Fig. 8.5 St10). At Tergu junction [738272], a 230° trending fault drags a red ignimbrite (Lithofacies 5j) within the Vaginella Member to almost vertical (Fig 8.13). The fault has oblique sinistral slickensides and causes local folding within a footwall fault set. The NE-SW trending
faults are in the correct location to represent the extension within the Castelsardo sub-basin of a buried strike-slip fault (Vignola fault; Fig. 6.1) identified in the basement (Barca et al. 1996).

The τ2 ignimbrite is also offset by a number of smaller faults which sub-parallel the larger structures and have throws of 10-20m (Fig 8.1, 8.11). Fault slip data deduced from slickensides and from offset geometries on well exposed faults indicate that the majority of extension occurred on high angle, planar normal faults. Scattered slickenside data show that dextral oblique or strike slip faults are oriented NW-SE to E-W whilst sinistral oblique or strike-slip faults are oriented NE-SW.

8.2.4 Perfugas sub-basin

Table 8.2 summarises the evidence for ?post mid-Oligocene structures in the Perfugas sub-basin. The Ortigiu fault clearly defines a topographic change between high basement mountains and the low-lying Perfugas sub-basin. The 70-80° dipping fault plane is exposed at the base of the Casteldoria type section. Undeformed, pink granite crops out a few metres away from the fault zone which comprises a well indurated, foliated and fractured granite mylonite with quartz veining (Fig. 8.14). The Ortigiu fault plane is not offset by N-S faults (e.g. Ruiu fault) present to the north suggesting that the ~1.9 km throw on the Viddalba fault (identified on seismic profile ~20 km to the north) had tipped out to zero at this point and/or that the Ortigiu fault post-dated this structure (see 8.3.2). Since similar geometries are observed within the Hercynian basement (e.g. on Barca et al. 1996), the Ortigiu fault may represent a reactivated transverse structure associated with the large sinistral strike slip-faults crossing the eastern Sardinian basement.

The southeastern margin of the Perfugas sub-basin is bounded by a gently northwestwards dipping slope in the footwall of the Tula fault (chapter 7) some 13 km to the south, defining a triangular sub-basin shape. Approximately 10m of τ2 ignimbrite offset, plus disruption and tilting of the basin fill on a NE-SW trend occurs along the Perfugas fault. The nature and extent of this fault is poorly constrained but it may represent a reactivation of a NE-SW trending, late Hercynian basement discontinuity which moved after Miocene basin filling (Funedda et al. 1997 also).

The Perfugas sub-basin is cross-cut by N-S to NNW-SSE trending normal faults (Sedini, Bulzi, Concatile) and passes westwards into the high and largely unfaulted Tergu platform in the footwall of the San Giovanni fault. Linear, NE-SW orientated structures, believed to be normal faults with throws of 10-50m on the basis of the large scale geometries, cut the τ2 scarp in the footwall of the San Giovanni fault. To the southwest of the Perfugas sub-basin, the Martis Member was deposited at various levels over a highly irregular volcanic topography. An E-W trending normal fault may cause the topographic change obvious south of the Sedini Member platform carbonates [845177] (Fig. 8.1, Encl 5., see Fig 8.72) but otherwise no structures were clearly identified.
New cuttings around Bulzi football pitch, within Perfugas Formation sediments, expose a syn-depositional anticline, 10m in diameter, with a SSW-NNE oriented, southward plunging axis. In this area, Quesney-Forest and Quesney-Forest (1984) identified an anticline with a NW-SE hinge which they related to dextral strike-slip movement on a N-S fault, plus reverse offset slickensides related to NW-SE compression. The poor exposures in this area did not allow an evaluation of these structures. However, they are minor and of limited areal extent. The triangular shaped Perfugas sub-basin suffered a complex history of extension superimposed on inherited basement structures. The presence of compressional structures may be indicative of local space problems rather than of a ‘compressional phase’ documented in Quesney-Forest and Quesney-Forest (1984).

8.2.5 Portotorres sub-basin margin

Table 8.3 summarises the evidence for post-mid Oligocene faults at the Portotorres sub-basin margin and proposed character of the faults. The Tramontana fault, identified on seismic reflection profiles and causing the present day topography, bounds the western Castelsardo sub-basin and eastern Portotorres sub-basin. The degraded fault scarp becomes less noticeable towards the south where, in the Monte Uri area a transfer zone with several smaller faults (Molasa, Fiori Fig. 8.5 st13) accommodates the change to the NE-SE oriented Sennori/San Giusta faults (Fig. 8.15).

8.2.6 Summary

The Anglona study area has a very complex structure where ‘new’ Oligo-Miocene extensional faults were superimposed on a pre-rift basement with major NW-SE lineaments. The half-graben geometries visible on seismic reflection profiles and on land are similar to those commonly found in extensional basins (e.g. in East Africa, Rosendahl 1987; Greece, Gawthorpe et al. 1994; Collier & Gawthorpe 1995; North Sea, Underhill 1991a,b; Rattey & Hayward 1993; Thomson & Underhill 1993, Underhill et al. 1997; Gulf of Suez; Angelier 1985; Patton et al. 1995, and in Leeder & Gawthorpe 1987; Gawthorpe & Hurst 1993) but the overall structure is much more complex. Two main fault families can be observed; E-W to NW-SE and NNE-SSW to NNW-SSE. ‘Large’ faults which defined the sub-basins structure were high, angle, planar features with dominantly dip-slip movement and throws from ~2 km to a few tens of metres. Smaller syn and post-depositional faults sub-parallelled the larger structures.
8.3 Basin filling geometries

The basin filling geometries observed within the Anglona study area, summarised here, are of particular use in constraining the timing of structural events. In the Castelsardo sub-basin, syn-rift sedimentary geometries are well exposed and the sediment response to active extension is examined.

8.3.1 Seismic reflection profiles-interpretation

In common with Thomas and Gennesseaux (1986), basin filling geometries on seismic reflection lines 123 and 125 (Fig 8.4, Encls. 6,7) are interpreted to have resulted from two phases of extension. Seismic ‘megasequences’ defined by seismic character and onlap/truncation relationships can be identified.

- The seismic ‘basement’ is characterised by irregular reflections and the contact with the overlying basin fill is often unclear.
- Megasequence 1 (yellow) consists of tilted, parallel-bedded or poorly imaged parallel-bedded units, thought to be pre-rift rocks or undefined, early basin fill.
- Megasequence 2 (orange) consists of a lower unit part where reflectors gently diverge into a normal fault and an upper aggradational, mounded unit which thickens into a normal fault and downlaps onto the lower part. This unit shows similarities to the ‘rift initiation’ and ‘rift climax’ syn-rift deposits of Prosser (1993) and represents a phase of active extension.
- Megasequence 3 (green) consists of a ‘post-rift’ tilted, faulted, parallel-bedded and irregular bedded package which may onlap and infill the accommodation space adjacent to top of Megasequence 2 and is widespread over the seismic lines.
- Megasequence 4 (blue) on line 125 consists of a mounded, aggradational package adjacent to the fault plane which downlaps onto megasequence 4 and thickens into the fault plane. This package records the second phase of active extension where tectonic subsidence > basin filling and accommodation space for megasequence 5 was formed.
- Megasequence 5 (lilac) consists of a thick, post-rift parallel-bedded, infilling package which onlaps onto lower megasequences/the basement. It is variably tilted and folded in response to post-depositional movement on the Tramontana fault and compaction. The top surface of the megasequence is an unconformity at 0.2-0.3s TWTT.
Field Observations

8.3.2 The Casteldoria horst area

In the Casteldoria horst area (Fig. 8.3), the basal, continental Casteldoria Member outcrops directly over the pre-rift basement in four different ways:

a) It onlaps onto low angle basement topography (Fig. 8.16) within a gentle depression at [910271]. The sediments are derived directly from the weathering of the basement.

b) It onlaps directly onto an existing fault plane at [905275] (Fig. 8.7). Synclinal folding, very local angular unconformities and bed thickness variations within the sediments are thought to result from early sediment compaction. Note that the high angle, planar fault is not degraded and it is likely that sedimentation occurred soon after faulting.

c) It lies adjacent to the degraded Therme fault scarp and exhibits syn-depositional bed divergence towards the basin at [915283] (Fig. 8.17). This geometry could result from filling of fault defined accommodation space, but with dips this divergent is likely to result from syn-depositional fault movement with fault drag (Fig. 8.17).

d) It is rotated and outcrops over the Ortigiu fault plane [905205] (Fig. 8.18). The variably dipping 'syn-rift' sediments and the t2 ignimbrite are thought to have been rotated during the growth of the Ortigiu fault in the style of a 'fault-propagation fold' (sensu Withjack et al. 1990; Gawthorpe et al. 1997; Sharp et al. in press), maybe later enhanced by drag on the fault. Figure 8.19 provides the interpreted scenario for the development of the Ortigiu fault and Ruiu faults based on sedimentological (8.4.1.2 below) and structural evidence.

Thus, in the vicinity of the faulted pre-rift basement, the basal Casteldoria Member exhibits a variety of sedimentary geometries with variable thickness dependent on local fault configurations and movement histories. The t2 ignimbrite and Casteldoria Member outcrop with high basinward dips (~40°) along the length of the Ortigiu fault (Fig. 8.1). In contrast, the flat lying Perfugas Formation sediments infill and onlap onto this topography (Fig. 8.1, F-F') indicating that movement on the Ortigiu fault had ended before Perfugas Formation deposition.

8.3.3 The Castelsardo sub-basin

8.3.3.1. Geometries associated with syn-depositional faulting

The sediments of the Valledoria Member contain rare syn-depositional faults on ~N-S and NW-SE trends at [725625]. The largest structure is the NW-SE trending sealed fault visible on Fig. 8.21. Numerous centimetre scale, north-south trending, sealed syn-sedimentary faults occur at the base of gastropodal marlstone beds.
The Castelsardo Member exhibits abundant evidence for syn-depositional normal faulting. The Strada Bianca section contains sealed, syn-sedimentary normal faults at the base of the section (Fig. 8.22) and at ~51m on Fig 8.55 [875279] (Fig. 8.23). The stratigraphically higher outcrop contains a basal, faulted and fractured sandy unit where bedding was rotated into a small anticline, with laterally adjacent, diverging conglomerate units and overlying coarsening upwards conglomerates draping over the fault-block (Fig. 8.23). The syn-sedimentary faults trend at ~120° and are located between the Ortigiu and San Giovanni faults within the inferred 'antithetic interference zone' (Fig. 8.12). They are interpreted as small faults which moved at the same time as the larger faults. Movement on the larger faults would be linked to uplift, degradation and a local influx of clastic material (8.4.4.5). This transfer zone may have been an entry point for clastic material found within the Castelsardo sub-basin (8.4.4.7).

To the west of Castelsardo [755287], a quarried exposure shows a tilted normal fault block within Castelsardo Member sediments and the response of the sediments to movement on this strand of the Fragiagu fault. The proto-fault block is thought to have been exposed at the sea-surface since it has a faintly bioturbated top (Fig. 8.24a). It was onlapped by overlying tuffaceous sandstones (debris flows, Lithofacies 6c) which thickened gently into the hangingwall depocentre. Several faulting events occurred before the deposition of the green conglomerate bed such that the fault strands moved progressively to the southwest, the fault block increased in size and subtle bed divergence continued (Fig. 8.24a). Synthetic and antithetic normal faults were also active (Fig. 8.24a). On the fault block dip-slope, exposed ~15m to the north (Fig. 8.24b), a series of cross-bedded sediment ‘wedges’ thickened and prograded away from the fault block high (Lithofacies 6d). Thus the sediment response to normal faulting within a supply dominated, shallow marine system (8.4 below) was for debris flows to thicken and pond in the hangingwall depocentre and sediment bars/fans to prograde and thicken down the dip slope. The fault and sediment geometries observed in these exposures are typical of syn-rift settings observed at a much larger scale on seismic profiles (e.g. Prosser 1991, 1993; Partington et al. 1993; Rattey and Hayward 1993).

The Castelsardo west beach section [756295-757298], in the footwall of the Castelsardo fault, shows very subtle bed divergence to the north in three localities and a sealed, faulted horizon (Fig. 8.25) within massive tuffaceous sandstones, tuffs and granite-volcanic conglomerates (Lithofacies 5b, 5g, 5h, 6e).

East of Castelsardo, the exposed section sits on the dip-slope of the Castelsardo fault and in the hangingwall depocentre of the fault proposed just offshore (Fig. 8.1, A-A’, B-B’). Very subtle bed divergence and thickening is observed within an E-W oriented section, away from the footwall of the N-S trending Castelsardo east fault [770293]. Clear syn-depositional onlap is observed in the hangingwall, NW-SE oriented cliff section (Fig. 8.10). Whilst the N-S fault may have been active, the
dominant fault activity occurred on E-W faults (below) and the above thickening may reflect variable
along strike throw on these latter faults. Bed thickening and divergence to the north is obvious from
the base of the Castelsardo east section at [770293] (Fig. 8.26). Further up the section [774295], the
Punta Viuledda cliffs show abundant evidence for thickening into a hangingwall depocentre (Fig.
8.27). The basal part of the cliffs exhibit subtle thickening to the northeast (Fig 8.28). Bed dips
truncated by, and onlapping onto an erosion surface are the same indicating a non-tectonic origin for
the surface (Fig 8.28). The surface is thought to have developed due to relative sea level fall, erosion
and later relative sea level rise with deposition of a transgressive pebble and fossil lag. The upper part
of the Punta Viuledda cliffs [774295] begins with a subtly diverging sequence of bioturbated
sandstones, truncated by a higher angle erosion surface, and is overlain by an orange, sediment wedge,
thickening to the north (Fig. 8.29). The anatomy of the sediment wedge consists of three sets of
stacked, cross-bedded units composed of granite gravelstones-fine conglomerates, passing northwards
into ?toesets (Lithofacies 6d, Fig 8.29). The sediments are envisaged as small 'Gilbert' type fan deltas
supplied down the tilted fault block dip-slope into the hangingwall depocentre to the north (8.4.8).
Non cross-bedded units exhibit approximately the same dips above and below the erosion surface,
again indicative of a non-tectonic origin for the higher angle unconformity. Although no discrete
boundary exists across the erosion surface the simplest explanation is that it resulted from relative sea
level fall since clear, bioturbated marine facies outcrop beneath the surface (Lithofacies 6e) and
marginal to non-marine facies outcrop on the top (Lithofacies 6d, 6c, 5h; 8.48 below).

Thus, apart from the top ~20m, the Castelsardo Member can clearly be identified as a unit deposited
contemporaneously with movement on E-W to NW-SE trending normal faults. The abundant supply of
reworked volcanic and clastic material records the syn-depositional activity as a series of divergent,
thickening wedges similar to those expected in an idealised active half-graben model (Appendix 1B.1).

8.3.3.2 Geometries associated with post depositional faulting
The τ2 ignimbrite records an instantaneous geological event and so could not record episodes of syn-
depositional faulting (section 8.4.5.2). However, the unit is post-depositionally faulted. Some faulted
outcrops of the τ2 ignimbrite do appear slightly thicken into the hangingwall and thin onto the
footwall. This could be a record of a subtle pre-eruptive faulted topography which the τ2 pyroclastic
flow covered but could equally be a product of later erosion of the footwall highs. The τ2 ignimbrite is
conformable on the Castelsardo Member. Extrapolating the dips of the Vaginella Member and Tergu
Formation through the zones of no exposure indicates that a small angular unconformity, thought to
result from movement on the Tramontana fault, exists between these units and the τ2 ignimbrite.

The τ2 ignimbrite was post-depositionally faulted on E-W-NW-SE and N-S trends before the
deposition of the Campulandru Member with a maximum throw of ~ 100m (Castelsardo fault),
resulting also in the tilting of the underlying Castelsardo Member. A clear angular unconformity exists between the faulted, tilted τ2 ignimbrite/Castelsardo Member and the approximately flat lying carbonates of the Campulandru Member deposited after mid Burdigalian marine transgression (Fig 8.1 A-A’, B-B’, 8.67 below).

The remaining NE-SW trending and small faults within basin fill appear to be post-depositional but their timing cannot be constrained (e.g. Fig. 8.5 St2-5, 8, 10, 11). Since the faults occur as related 'groups' of similar nature and orientation it is likely that they occurred at the same time as those faults which are constrained or represent post-Miocene reactivation of these trends.

8.3.3.3 Large scale geometries
The parallel-bedded sediments and volcanics of the Valledoria and Vaginella Members which infilled the Castelsardo half-graben now outcrop in a gentle, asymmetric syncline (N-S axis, Fig. 8.1 C-C’) probably as a result of back tilting in the footwall of the Tramontana fault and sediment compaction. The Tergu Formation volcanic rocks pass laterally into and are occasionally overlain by the Vaginella Member (Fig. 8.1, A-A’, C-C’). The Castelsardo Member outcrops as a series of wedges of variable thickness, laterally equivalent to the uppermost Vaginella Member (Fig. 8.1. A-A’, B-B’).

8.3.4 Perfugas sub-basin
In the eastern Perfugas sub-basin, the τ2 ignimbrite was post-depositionally folded by movement on the Ortigiu, Tula and San Giovanni faults causing a triangular synclinal geometry in which the Perfugas Formation lake formed (Fig. 8.1 F-F’). Towards the west, the τ2 ignimbrite was faulted/folded along the N-S Sedini fault before the deposition of the Sedini Member, since at 838184, flat lying carbonates onlap the faulted and folded τ2 horizon (Fig 8.30). The Sedini Member platform carbonates do not contain any evidence for syn-depositional faulting. Rather the thickness of the Sedini Member varied dependent on the underlying τ2 topography. For example, the carbonates thicken westwards towards the Sedini fault (Fig. 8.1 D-D’). They onlap the southward dipping τ2 topography to the east of the fault, thicken, prograde and pass laterally into the Martis member marlstones southwards into the deepest parts of the sub-basin (Fig. 8.72 below).

Basin fill geometries in the west of the Perfugas sub-basin and towards the Portotorres basin margin consist of variably tilted andesite flows capped by prevalently flat lying τ2 ignimbrite and carbonates of the Martis and Sennori Members. Whilst some dips within the Tergu Formation may be depositional (8.4.5.3), tilting of the entire succession is thought to result from of a phase of fault movement on the Tramontana-Sennori system before τ2 eruption. The present day Perfugas sub-basin geometries result from a phase of post-depositional faulting on the Sedini, Bulzi and Concatile faults.

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and gentle tilting of the southern basin fill to the northwest in the footwall of the Tula and Perfugas faults.

8.3.5 Portotteros sub-basin margin

Sediment geometries and thicknesses within the Sennori Member reflect depositional onlap onto and filling of a faulted Tergu formation topography (e.g. Sennori fault). Two outcrops may relate to syn-depositional fault movement: an outcrop of slumped marlstone adjacent to the Tramontana fault at [708213] and carbonate bed dip variations in the hangingwall of the Molaza fault [724192]. Post-depositional faults offset the carbonate units (e.g. San Giusta fault). Marlstones within the Portotteros sub-basin dip gently in the same direction as the bounding faults most likely as a consequence of post-depositional movement on the sub-basin bounding faults and sediment compaction.

8.3.6 Correlation of seismic and onland geometries

Tentative correlations can be made between the basin filling 'megasequences' observed on seismic profiles and the large scale basin filling geometries observed in the Anglona study area by assuming that the half-graben basins observed on seismic profiles are of Oligo-Miocene age. The aggradational, mounded part of megasequence 2 (Fig. 8.4), relating to the first phase of extension on ~N-S half-graben bounding faults may be equated with the basal clastic deposits shed from fault scarp topography (Casteldoria Member). The infilling nature of megasequence 3, capped by a strong reflector is comparable to the half-graben filling of the Valledoria and Vaginella Members, capped by the τ2 ignimbrite (also in Thomas and Gennesseaux 1986). The faulting event (megasequence 4) recorded before post-rift basin infilling (megasequence 5) is likely to be the same event recorded before the deposition of the Campulandru, Sedini and Sennori Members on the Castelsardo east, Sedini, Tramontana and Sennori faults respectively. Finally, the dips to the west of megasequence 5 on line 125, west of the Tramontana fault and to the east in the footwall of the fault are similar to those observed onland.
8.4 Facies descriptions and environments of deposition

Lateral facies variations require that the complex basin fill of the Anglona study area be subdivided into a number of lithostratigraphically defined members (chapter 3). This section outlines the sedimentology and volcanology of these basin filling units. The basin fill was deposited within the developing structural framework described above, contemporaneous with normal fault movement (e.g. basal Casteldoria member, Castelsardo Member) and as a response to a complex fault-defined palaeotopography (e.g. Vaginella Member, Laerru Formation).

8.4.1 Elefante Formation, Casteldoria Member (?late Oligocene)

The Casteldoria Member comprises continental conglomerates, sandstones and siltstones deposited by alluvial fans adjacent to fault topography (Lithofacies la, lb, lc, le) passing laterally into fluvial (ld, lf) and lacustrine (lf, lh) environments. Rare outcrops of the sediments, which overlie the pre-rift basement in a variety of structural situations are exposed around Casteldoria [913276]. Outcrops of similar lithofacies (1b, ld, lf) at [959238] in the NE corner of the Perfugas sub-basin are correlated to the Casteldoria Member.

8.4.1.1 Lithofacies

The Casteldoria Member can be described by the seven lithofacies (1a-1f, 1h) shown in Table 8.4. Although the Casteldoria Member has its own distinct characteristics described below, Lithofacies 1a-1f are interpreted to have been deposited by the same processes and in similar environments to those discussed in chapter 5.

- As well as related to stream channel reworking, coarse to fine sands (Lithofacies 1f), composed of graded, amalgamated beds with rip-up clasts which are intercalated with Lithofacies 1h, represent deposition from sediment gravity flows, possibly turbidites, which occurred as a fluvial system entered a lake over topographic relief (as examples in Tucker and Wright 1992, Talbot and Allen 1996).

- Lithofacies 1h consists of bedded white fine marlstones to limestones composed of lime mud and sand-silt size quartz clasts, biotite, white mica, quartz and plagioclase crystals. The sediments are characteristically bedded on a millimetre to ~40cm scale with grading from fine sandstones to siltstone occasionally observed. In some areas the sediments contain a monospecific, millimetre-sized, high spired gastropod assemblage [725625], in other areas, ostracods or thin shelled bivalves are present [860285]. Round, pinkish, <1mm sized carbonate spheres are occasionally observed, scattered within marlstones and are thought to be lacustrine oogones (Charophyte reproductive structures, after Tucker and Wright 1992). Carbonised or silicified organic fragments are reasonably common. Such lithofacies and fossil fauna have been described in lacustrine environments (Talbot and Allen 1996, Tucker and Wright 1992).
The following sections describe the differing composition, sedimentary structures and lithofacies arrangements of the Casteldoria Member in four localities

8.4.1.2 Casteldoria road section [905205]
The Casteldoria type section crops out directly over the 120° trending Ortigiu fault plane bounding the Monte Ruiu granite and underlies the τ2 ignimbrite. The beds have been post-depositionally rotated (Fig. 8.1, 8.3a, see 8.3.2). The section consists of sheet-like and channelised pebble conglomerates (Lithofacies 1b, 1c, 1d), massive gravelstones and sandstones (1f, Fig 8.18).

The sorting of clast-supported conglomerate pebbles is fairly poor with clast sizes up to 27cm (Fig. 8.31). Outcrop F exhibits a bimodal clast population (Fig 8.31). Clasts in beds D-F are sub-rounded to rounded (Fig. 8.18) whilst those in beds A-C are sub-rounded to sub-angular.

The pre-rift basement lithologies to the west and north of the Casteldoria area are Silurian rocks metamorphosed in the Hercynian orogeny (gneisses, schists, intermediate and basic intrusives, post Hercynian granites and Permian quartz porphyry; Servizio Geologico sheet 180, Fig. 8.1, Enclosure 5). These rock types form the clasts found within the Casteldoria road section (Fig. 8.32) which were therefore supplied from the local area rather than the immediate vicinity. It is curious that the clast types exhibit a systematic change from a diverse range of granitic, metamorphic and porphyry at the base of the section to dominant porphyry clasts in beds C-F (Fig 8.32). It is difficult to envisage how the upper units sampled only a (now eroded) porphyry source area. Perhaps the quartz porphyry was the most resistant to sediment recycling?

Intercalated matrix supported conglomerates, gravelstones and sandstones are composed of poorly to moderately sorted basement clasts, quartz and feldspars plus a possible volcanic component of euhedral plagioclase and mafic (?biotite, hornblende) crystals and a fine green, possibly clay mineral fraction.

Palaeocurrents from re-oriented clast imbrications show dominant westward palaeoflow for the lower beds A-D and an abrupt change to northeast palaeoflow in the conglomerate bed F (Fig 8.33). Beds A-D were transported westwards from the ~N-S trending Ruiu fault high whereas bed F may have been deposited contemporaneous with the growth of the Ortigiu fault (8.3.2). Quesney-Forest and Quesney-Forest (1984) used average clast orientations to suggest a similar change in sediment supply linked to the formation of the Ortigiu fault.

The sheetlike conglomerate beds B and C (Fig 8.18) have slightly gradational contacts and they are thought to represent cohesionless debris flows to sheetflows (after Bull 1972; Nemec and Steel 1984; Miall 1996) deposited in ‘flash flood’ type events. Bed D exhibits reverse grading over ~40cm,
indicative of a clast-rich debris flow (after Miall 1996). It has a sharp top (Fig 8.18), where the uppermost clasts have a shiny red iron coating thought to represent a ‘desert varnish’ surface. Other conglomerates outcrop as matrix-supported patches and as discontinuous ‘strings’, one to two pebbles thick. The facies and facies arrangements are representative of alluvial fan deposition (sensu Bull 1972, Miall 1996).

8.4.1.3 Casteldoria car park [910274]

In the vicinity of Casteldoria, the sediments of the Casteldoria Member were deposited in fault-related topographic depressions directly overlying the granitic and metamorphic pre-rift basement (Fig 8.1, Enclosure 5). Fluvially deposited conglomerates, gravelstones and coarse sandstones [910274] (Lithofacies Id, If) pass northwest, westward and basinward into intercalated sandstones and marlstones [905275] thought to represent proximal lacustrine deposits (Lithofacies If, Ih, Fig. 8.7).

The voluminously minor conglomerates comprise moderately sorted, sub-angular to sub-rounded, dominantly metamorphic clasts in a granite derived matrix. Gravelstones and coarse sandstones consist of moderately well sorted, sub-rounded quartz and feldspar with granite and minor metamorphic clasts. They are ‘granite wash’ sourced locally by erosion of the Monte Ruiu granite. Average clast sizes decrease laterally from a few centimetres-0.5cm [910274] to a few millimetres at [905275] though centimetre sized rip-up marlstone clasts are found in the latter locality. Marlstones are composed of lime mud, volcanic ash with biotite crystals and plant fragments.

Cross-bedding, characteristically ~50cm in height and demonstrating N to NNE directed palaeocurrents (Fig 8.34a), is the dominant sedimentary structure observed at Casteldoria car park [910274]. Individual, centimetre scale cross-beds exhibit coarsening up to gravel grade, indicative of avalanche face, grain flow processes (Bagnold 1954; Leeder 1982). Toesets show lateral bed thinning and have a ‘swept out’ geometry. Trough-filling cross-bedding is observed both perpendicular (Fig. 8.34b) and parallel to the palaeoflow direction. This is a characteristic fluvial signature (Leeder 1982, Collinson 1996, Miall 1996). Cross-bedded units sometimes occur within metre scale beds which fine up from a conglomerate base and are intercalated with thin marlstone beds (Fig. 8.34c) perhaps recording material supplied in debris flows, reworked by a fluvial system and times of overbank/lacustrine deposition. The lithofacies and sedimentary structures at Casteldoria car park represent deposition from a dominantly fluvial system with periodic high energy flow towards the basin. Lacustrine sedimentation may have been active from time to time, recorded between flow events.

Graded and amalgamated-graded gravelstone and coarse sandstone beds intercalated with fine white marlstones are common in the more ‘distal’ localities such as [905275] (Fig 8.7). The amalgamated graded unit shown in Fig. 8.7c has a rippled top surface. These sediments are thought to have been
deposited in a proximal lacustrine environment where material supplied by fluvial systems entered a lake, was deposited as a turbid gravity flow which ripped up underlying marlstone clasts and was then reworked by waves at the lake edge.

8.4.1.4 Viddalba-Therme di Casteldoria
Sediments lying adjacent to the NNW-SSE trending Viddalba fault and overlying a pumiceous, welded, biotite-hornblende ignimbrite (Tergu Formation) consist mainly of coarse breccias, conglomerates and gravelstones (Lithofacies la, lb, ld, le, lf) sourced locally from the fault topography. The outcrops are thought to be of Oligo-Miocene age because of their lithification and facies, though it is possible they could be more recent since they are not constrained by any overlying units.

Clasts are poorly sorted, angular to sub-rounded and composed of locally supplied metamorphic, granite and quartz porphyry rocks. Conglomerates and breccias range from clast to just matrix-supported and crop out as massive banks to discrete beds. Immediately adjacent to the metamorphic basement at [918292], monomict metamorphic boulder breccia fines laterally over 20m and passes into a polymict conglomerate cut by rare gravelstone beds. The sediments are interpreted as the products of gravity fall to subaerial debris flow processes active on small sediment cones or fans adjacent to fault topography.

8.4.1.5 Cantionera Coghinas [959238]
Quarried exposures correlated to the Casteldoria Member crop out in the area of the intersection of the Ortigiu and Perfugas faults, under the τ2 ignimbrite. Although not studied in any detail, the massive banks (20m high) of coarse red sandstones and gravelstones cut by lenses and channels of conglomerates are characteristic of braided fluvial systems (Collinson 1996, Miall 1996).

8.4.2 Elefante Formation, Valledoria Member (?late Oligocene -earliest Aquitanian)
The fluvio-lacustrine Valledoria Member consists of parallel-beded white marlstones, cherts and tuffs (Lithofacies 1h, 1j, 5a, 5b, 5e, 5h, Table 8.4) with rare, massive, granite-derived gravelstones (Lithofacies 1d). The sediments are best exposed in cuttings along the new coast road in the west of the Castelsardo sub-basin, south of Monte Oschiri [725625] and in the east [858292], south of Valledoria (type section). The gradational contact with the overlying Vaginella Member occurs where volcanic facies begin to dominate (Encl. 5, Fig. 8.1). The sediments are classified on the geological map (Sheet 180, Servizio Geologico) as m1, ‘lacustré’ composed of sandstones, limestones, marly limestones, intercalated cherts, lignites and tuffs containing a fauna of the continental molluscs Helix, Planorbis and Limnaea.
8.4.2.1 Lithofacies

Six additional lithofacies are needed to describe the Valledoria Member (Table 8.4)

- **Lithofacies 1j** consists of parallel-bedded, brown and black cherts and partly chertified white marlstones with similarities to Lithofacies 1h. The lithofacies is thought to have been deposited in a lacustrine environment. Cherts are commonly found in lake sediments and represent post-depositional modification of diatom tests to opaline silica supplied after algal blooms (Talbot and Allen 1996). In lakes associated with faults and hydrothermal circulation, alkaline hot springs can be an additional source of siliceous deposits (e.g. Baker 1986; Renaut et al. 1986; Talbot and Allen 1996). Hydrothermal circulation occurs at the present day (Therme di Casteldoria hot springs) and is likely to have been more active in the Oligo-Miocene (active faulting and volcanism) so it is quite possible that both mechanisms contributed to the chertification of the marlstones.

- Tuffs, pumice lapilli tuffs, crystal tuffs and epiclastic tuffaceous sandstones (Lithofacies 5g, 5h, 5a, 5b, 5e) which are described in section 8.4.3 are also interbedded with the Valledoria Member.

8.4.2.2 Valledoria area

Road cuttings in the Valledoria area e.g. [850292] and [875295] provide a horizontal section through laterally extensive, finely parallel-bedded lacustrine marlstones to limestones, chertified marlstones, cherts and tuffs (Lithofacies 1h, 1j, 5g, 5h, Fig 8.35). In thin section, marlstones comprise lime mud with minor quartz clasts and rare articulated and disarticulated bivalves or ostracods (Fig 8.36). Southsouthwest of Valledoria [853281], low rounded hills have been formed by the weathering of soft marlstone beds beneath a hard cap of indurated, well sorted, granite derived gravelstone (Lithofacies 1d). This massive, metre-scale bed is interpreted to have been transported by a debris flow from fault-bounded, topographically high granite basement into a calm, carbonate dominated lacustrine environment.

8.4.2.3 South of Monte Oschiri

Road cuttings south of Monte Oschiri [725625-728267] expose marlstones, chertified marlstone, cherts (Fig. 8.37) and tuffs/lapilli tuffs of the Valledoria Member (Lithofacies 1h, 1j, 5a, 5b, 5e, 5h, Fig. 8.38). Towards the base of the exposure, a large 'channel-like' body, filled by sandy marlstones with faint ripple and trough traces was eroded into underlying tuffs outcrops (Fig 8.21). The base of the 'channel' contains three curious ~50cm depressions (Fig 8.21) one of which has morphology similar to groove structures common at the base of turbidites. The structures are oriented just to the west of north and could have been small, subaerial channels formed before the larger, overlying channel. Alternatively, the outcrop could represent a submarine lacustrine channel with groove marks at its base, though the energy of the depositional environment does not seem sufficient to support this hypothesis. Above the crest of a syn-sedimentary fault (Fig. 8.21), the succession continues with tens
of centimetre thick, parallel-bedded alternations of cherts (Fig. 8.37), tuffs and partly chertified marlstones (Figs. 8.38, 8.21). A monospecific gastropod assemblage indicative of non-marine or stressed lagoonal conditions totally dominates some partially chertified marlstone beds and scattered, spherical oogones are occasionally found.

8.4.2.4 Depositional environment

Parallel-bedded alternations of marlstones to limestones, cherts and tuffs represent deposition in a quiet, dominantly carbonate lacustrine environment where material was sometimes supplied from the erosion of the surrounding basement or from volcanic eruption. Fine siliciclastic-carbonate alternations are observed in other ancient lake sediments such as those from the Orcadian basin of Scotland (in Talbot and Allen 1996) as are volcanic tuffs and ashes are mixed into lake sediments (East Africa Renaut et al. 1986; Watkins 1986; Wilkins Peak Member, USA in Tucker and Wright 1992).

The western outcrops of the Valledoria Member which were deposited in proximity to the Tergu Formation volcanics contain a much higher proportion of volcanic material and chert beds than those in the east. Along with the channels and gastropodal mud banks, the western exposures may represent a shallower fluvo-lacustrine environment than those in the east, which were a distal equivalent of the Casteldoria Member within the half-graben basin. Figure 8.39 synthesises the possible palaeogeography of the eastern Castelsardo sub-basin and shows the variable dispersal paths and basinward fining of the Casteldoria Member.

8.4.3 Elefante Formation, Vaginella Member (?late Oligocene -early Burdigalian)

The Vaginella Member comprises a volcano-sedimentary succession of tuffs and lapilli tuffs with minor cherts and lignites (Lithofacies 5a-5j, Table 8.5). The succession is here interpreted to have been deposited as a result of subaerial pyroclastic flows and fallout from explosive ?phreatoplinian eruptions entering a shallow lacustrine/lagoonal (at the base) to marine environment (top). The 250m thick Vaginella Member is well exposed in numerous road cuttings in the north, central part of the Anglona study area. It is differentiated from the overlying Castelsardo Member which contains intercalated clastic sediments and basaltic-andesite lava flows.

8.4.3.1 Previous work

First described as the ‘Molassa a Vaginella depressa’ (Parona 1887) or ‘Tufo a Vaginella’ (Moretti 1942) because of the occurrence of the Vaginella sp. pteropods, the Vaginella Member is simply described as ‘silty tuffs and sands’ by Spano and Asunis (1984). The variably silicified sandstones consist of volcanic glass, pyroxenes, micas, foraminifera, radiolarian, sponge-spicules and sometimes pumice clasts (Spano and Asunis 1984). Carbonised and silicified wood, gastropods, bivalves, scaphopods, echinoids, calcareous nanofossils and pteropods are found within the well stratified
tuffaceous horizons (Servizio Geologico, sheet 153, 1953, Spano and Asunis 1984, Francolini and Mazzei 1991). No-one has described the interesting depositional structures found within the Vaginella Member or the sorts of processes which might have operated during the deposition of the 'volcano-sedimentary' succession.

8.4.3.2 Lithofacies

Thirteen distinct lithofacies can be identified within the Vaginella Member (Table 8.5). The environments of deposition are considered in 8.4.3.4.

- **Lithofacies Sa** consists of pumice lapilli tuffs of rhyolitic composition. They are pale grey in colour with abundant, millimetre to 5cm sized white pumices, rarer non-juvenile volcanic clasts and show massive or faint parallel stratification within 5cm-1m thick beds (Fig. 8.40). The 'matrix' surrounding the pumices consists of blocky, cuspate and platy glass shards, plagioclase, quartz, clinopyroxene, orthopyroxene and opaque (?magnetite) crystals in a glassy groundmass.

- **Lithofacies Sb** are crystal lapilli tuffs and crystal tuffs of dacitic-andesitic composition. The rocks are dark grey in colour, of fine to coarse sand grade, with up to 40%, black crystals, with small (few mm) rare pumices and variable proportions of non-juvenile volcanic clasts. The crystal lapilli tuffs are generally 5cm-2m scale bedded with no internal stratification or grading but are rarely channelised [e.g. 790271] or normally graded. Opaques (?magnetite), plagioclase, clinopyroxene, orthopyroxene crystals and blocky glassy shards which form the bulk of the rock are surrounded by a glassy matrix (Fig. 8.41). Towards the top of the Vaginella Member, small, thin shelled bivalves and gastropods are found [753236] and [790271] within this lithofacies.

- **Lithofacies Sc** comprises elaborately interbedded pumice lapilli tuffs, crystal lapilli tuffs and tuffs (Fig. 8.42) with beds up to a few metres thick. The structures seem to have formed as a series of small troughs, channels and climbing ripple forms with complex pumice-crystal segregation and reactivation histories. The most complex structures where pumice lenses pinch in and out, beds truncate and split into two cannot be easily explained. Crystal tuff-tuff structures such as low angle, sometimes bidirectional cross-bedding, climbing ripples and low angle trough cross-bedding are found in of marine settings with high sediment fallout or in pyroclastic surge deposits (Allen 1984; Fisher and Schminke 1984).

- **Lithofacies Sd** consists of laterally extensive, fine grained, dark green beds, a few to ~20cm thick (Fig. 8.43). XRD analysis suggests that the green beds are of quartz, augite, clinoptilolite and illite composition and they are thought to be partly chertified, altered volcanic tuffs.

- **Lithofacies Se** consists of epiclastic tuffs, lapilli tuffs and agglomerates. The reworked volcanic rocks outcrop as massive, faintly stratified, cross-bedded and channelised units (Fig. 8.43, 8.44). Polymict, non-juvenile volcanic clasts occur within a matrix containing quartz, plagioclase, mafic minerals and glassy shards. Whilst cross-bedding and channelisation result from turbulent fluid
flow, massive beds, beds with diffuse matrix-supported volcanic clast patches and with 'strings' of lithic clasts are indicative of mass flow processes (Nemec and Steel 1984; Postma 1990).

- Lithofacies 5f consists of very finely laminated tuffs and lapilli tuffs (Fig. 8.45). The rocks comprise millimetre-centimetre scale interbedded layers of pumice (1mm sized) tuffs, crystal tuffs and tuffs. Soft sediment deformation structures are common (Fig. 8.45). This lithofacies represent deposition in a low energy, subaqueous environment with limited volcanic supply.

- Lithofacies 5g consists of 5-50cm bedded white, well sorted tuffs with faint parallel lamination.

- Lithofacies 5h comprises massive pale green to brown tuffs with scattered crystals, lithic clasts and pumice, commonly 0.5-3m scale bedded.

- Lithofacies 5j consists of hard, red-purple or white, 1-3m thick units, with a glassy groundmass, aligned vesicles, rare aligned crystals and lithic clasts.

- Lithofacies 5k consists of lignite beds which outcrop as 5-15cm thick layers containing abundant carbonised and silicified plant remains.

- At the base of the succession gastropodal marlstones and cherts (Lithofacies 1h, 1j) are interbedded. Marlstones contain a slightly more diverse, three fauna of three gastropod types (Fig. 8.38).

8.4.3.3 Facies architecture

At the outcrop scale, the Vaginella Member is characterised by parallel-beded units of the lithofacies described above, generally between 5cm and 2 metres in thickness (Fig. 8.46). Interbedded pumice and crystal lapilli tuffs dominate. The thinner intercalations sometimes contain water escape and soft sediment deformation structures. On a larger scale, interbedded cherts, gastropodal marlstones and lignites (Lithofacies 1h, 1j, 1k) indicative of a lacustrine/lagoonal and subaerial environments are found towards the base of the section [728267] (Fig. 8.38, 8.46a) whilst marine fossils are found towards the top of the succession (e.g. at [790271 and 753236]).

8.4.3.4 A possible depositional environment

The vesicular (pumice), non-vesicular (glassy shards) and non-juvenile volcanic clast types found within the rhyolitic-andesitic, volcanic-derived Vaginella Member, combined with the areal extent of the deposits suggest that the material was supplied from phreatoplinian eruptions driven by the vesiculation of magma and by magma-water interaction (after Self and Sparks 1978). It is thought that the volcanoes were located in the west and southwest of the Anglona study area (Tergu Fm., below). However, trying to unravel how the Vaginella Member relates to subaerial eruption is difficult. Non-volcanic lithofacies and fossils imply that the Vaginella Member was deposited in a fluvio-lacustrine environment at its base and a marine environment towards its top. Although some of the Vaginella Member may have been deposited subaerially or in very shallow water, most of it must have been deposited in a subaqueous environment. The marine transgression is not recorded by typical features
such as shoreface reworking, a transgressive lag, etc. and indeed the majority of the volcanic-derived succession has not undergone any obvious reworking.

The Vaginella Member exhibits some features and structures common to subaerial, pumiceous pyroclastic flow deposits. For example, lithofacies 5a comprises massive or crudely stratified pumice units similar to those found in some ignimbrites but lacks crystal-pumice grading and associated basal surge or upper ash deposits. Lithofacies 5c with its complex pumice-crystal tuff interbeds contains structures which may be similar to those observed from pyroclastic flows and surges, in particular those from 'wet surges' (i.e. free water within the flow, Allen 1984, Fisher and Schminke 1984; Cas and Wright 1992; Fig. 8.42). The hard, white and red units with aligned crystal and vesicle textures (Lithofacies 5j) may be the products of a hot, gas rich, pyroclastic flow with high temperature welding (ignimbrites).

A point of debate is whether 'submarine' ignimbrites resulting from the passage of a hot, gas rich pyroclastic flow into a body of standing water and seemingly present in some ancient successions can really exist. Cas and Wright (1991) examined the problem and concluded that the only unequivocal 'submarine', hot gas rich pyroclastic flow deposits occurred in the shallow water, nearshore area, that when pyroclastic flows enter water they are disrupted explosively or ingest water and transform into water-supported mass flows and that there is no evidence to suggest that welding in subaqueous environments is common. The general consensus seems to be that when a pyroclastic flow enters a body of water a variety of hot, water supported debris flows, low and high concentration turbidites and secondary mass flows develop (Fig. 8.47; Fisher 1984; Cas and Wright 1988; Kano 1991; Orton 1996)

For the majority of the Vaginella Member an environment of deposition where subaerial pyroclastic flows and surges enter a shallow body of water and become transformed into water supported mass flows seems quite appropriate. Massively bedded tuffs and lapilli tuffs (Lithofacies 5a, 5b, 5h) could represent primary and secondary debris flow deposits. Complex interbeds of pumice and crystal tuffs (Lithofacies 5c) might result from water-lain, highly turbulent, low to high concentration flows with rapid rates of fallout (Fig. 8.47). Finer, laminated tuffs (Lithofacies 5d, 5f, 5g) would be the record of fallout deposits (Fig 8.47). Soft sediment deformation and water escape structures would be expected in such a succession and could be triggered by eruptions. Similar problems to those in other ancient settings exist within the Anglona study area such that hard 'ignimbrites' with eutaxitic textures (Lithofacies 5j) exist in a 'subaqueous' succession. Since no thin section was taken, it is impossible to say whether these truly represent the 'welded' products of hot pyroclastic flows. The scenario might have been similar to that described by Cas and Wright (1991) with an emergent to shallow water welded ignimbrite deposited across the shoreline (Fig 8.47) and mass flows supplied basinward. Reworking of the Vaginella Member occurred as channels supplied material from the Tergu Formation volcanic centres and marine currents may have caused cross-bedding (Lithofacies 5e). A planktonic
fauna (foraminifera, pteropods) and a small-sized and restricted benthic fauna lived within the
treacherous waters of such a sea.

The Vaginella Member facies appear similar to those 'subaqueous' deposits derived from subaerial
flows recorded by Fiske (1963 in Cas and Wright 1992) in a marginal freshwater-marine basin and by
Kokelaar et al. (1984) in the Ordovician of Wales. Although there is evidence for a marine
environment, the Vaginella Member does not have clear shoreface deposits such as are observed other
settings where subaerial flows are thought to have entered the sea (e.g. Mweelrea Fm. Ireland Dewey
1963; Capel Curig Formation, Wales, Orton 1987). In summary, parts of the Vaginella Member may
represent the products of subaerial pyroclastic flows, surges and fallout but the majority is interpreted
as the products of subaerial pyroclastic flows, surges and fallout entering a shallow lake, lagoon or sea.

8.4.8 Elefante Formation, Castelsardo Member (early/mid Aquitanian-early
Burdigalian)

The Castelsardo Member is a succession of interbedded clastic breccias, conglomerates, sandstones
and siltstones sourced from the pre-rift basement and from underlying volcanic rocks, lapilli tuffs, tuffs
and basaltic andesite lava flows. The rocks were deposited in a variety of environments from offshore
marine to very shallow marine/subaerial conditions (Lithofacies 6a-g, 5b, 5g, 5h; Table 8.6). Lying
stratigraphically above the volcano-sedimentary Vaginella Member and beneath the t2 ignimbrite, the
Castelsardo Member is well exposed in coastal sections to the west [756295-757298] and east
([770293-775297] Punta Viuledda type section) of Castelsardo and in a road outcrop southwest of San
Maria Coghinas ([873278] to [879270] Strada Bianca section). Scattered exposures outcrop inland for
example at Punto Molino[790273] and San Giovanni [800260 and 796257].

8.4.4.1 Previous work

The prominent cliffs of Punta Viuledda [774295] (Fig. 8.27) comprising 'green-yellow, granite,
porphyry and quartz conglomerates-sandstones' are included within the logged sections of Maxia and
Pecorini (1969) and Spano and Asunis (1984). Some graded and cross-bedding is noted and a
provenance from Gallura (eastern Palaeozoic basement) is suggested. The authors note that there
exists 'facies heterotropy' between Punta Viuledda and their other logged sections ([763294] and
[790270]) and that the clastic sediments which contain Pinus sp. pine cones, foraminifera, pteropods,
turritellids, bivalves, oysters, pecten, echinoids and rare corals were deposited in the sea. The Strada
Bianca section was studied by Quesney-Forest and Quesney-Forest (1984). They describe a succession
commencing with fluvialite subangular, 'heterometric' breccias and conglomerates sourced from the
Palaeozoic basement, fining upwards into interbedded sandstones and conglomerates with a littoral
marine fauna, notably bivalves and gastropods, to planar and parallel-bedded, well sorted sandstones
with interbedded tuffs.
8.4.4.2 Lithofacies

The Castelsardo Member consists of a diverse range of marine and marginal/non-marine clastic and volcanic units which can be classified by ten lithofacies (Table 8.6)

• **Lithofacies 6a** consists of poorly sorted, massively bedded, clast-supported boulder-pebble breccias with a disorganised fabric (Fig. 8.48). The units have irregular, erosive bases, are channelised in one locality and were sourced locally from the pre-rift basement. The sediments are interpreted to be submarine (from surrounding sediments), cohesionless debris flow deposits (after Nemec and Steel 1984; Postma 1990) deposited close to an exposed fault scarp.

• **Lithofacies 6b** consists of matrix supported conglomerates which occur rarely, either as discrete beds or as beds with conglomerate ‘strings’. They are interpreted to be the products of submarine (from surrounding sediments) cohesive debris flows (after Nemec and Steel 1984; Dabrio 1990, Postma 1990).

• **Lithofacies 6c** comprises 10cm-5m parallel-bedded, clast-supported pebble-gravel conglomerates and sandstones with little internal structure and moderate sorting (Fig 8.49, 8.25). Clast compositions range from dominantly pre-rift basement to dominantly volcanic rocks. Four types of high-spired *Turritellid* gastropods [790273], some with serpulid encrustations (Fig. 8.49), whole and broken bivalves are found occasionally. The sediments are thought to represent submarine, cohesionless debris flows (after Nemec and Steel 1984; Dabrio 1990; Postma 1990) which sourced an area of sediment reworking.

• **Lithofacies 6d** consists of 0.5-3m scale, cross-bedded conglomerates, gravelstones and sandstones (Fig. 8.50). Cross-bed sets and individual, high angle (e.g. 30°) cross-beds commonly exhibit fining up and have ‘swept-out’ toesets (Figs. 8.50, 8.24, 8.29). At Punta Viuledda, topsets, foresets and possibly toesets are preserved (Fig. 8.29) Clasts are moderately to well sorted, sub-angular to rounded and sourced from the erosion of pre-rift basement (dominantly granitic). The cross-beds tend to have a very open framework, though where matrix is present, it is of a volcanic composition (e.g. Fig. 8.50 [7552871]). The cross-bedded units are interpreted as prograding sand/gravel bars existing as small Gilbert type fan deltas in a shoreface-marginal marine setting (after Postma 1990, Kazanci 1990, Johnson and Baldwin 1996).

• **Lithofacies 6e** consists of 20cm-2m bedded, bioturbated gravelstones-sandstones, often graded and composed of pre-rift, often granite derived clasts and reworked volcanic rocks (Fig 8.23, 8.28). On the Strada Bianca [875276], finely planar laminated sandstones and symmetric climbing ripples with straight, bifurcating crests and superimposed bundled lenses (Bouma units C, D Fig. 8.52) occur at the top of graded beds. Trace fossils include *Thalassanoides*, *Skilithos*, *Diplocraterion*, *Ophiomorpha*, centimetre diameter vertical burrows with a gravel-sized clast lining and areas of a complex, superimposed ichnofabric. Whole and broken marine fossils such as echinoids (*Scutella* sp. Spano and Asunis 1984), gastropods, bivalves including pectenids and *Ostrea* are found. These
rocks represent the colonisation of mass flows (debris flows to high density, coarse turbidites), below wavebase in oxygenated waters (after Nemec and Steel 1984).

- **Lithofacies 6f** comprises 20cm-1m graded, bioturbated grey volcanic sandstones-siltstones. The normally graded beds sometimes have an irregular erosional base, planar gravelly strings, concentrations of pumice and bioturbation at the top (Fig. 8.51). Carbonised wood fragments are common and one particular layer contains fossilised pine cones (Fig. 8.53, *Pinus sp.* Maxia and Pecorini 1969) Trace fossils include *Asterasoma* (Fig. 8.54), *Rhizocorallium*, *Diplocraterion*, ?*Skilithos*, *Planolites* (Fig. 8.54), *Chondrites* and colonies with a complex, cross-cutting nested ichnofabric. The rocks are composed of silt-sand size, euhedral, broken and abraded plagioclase, clinopyroxene and opaque crystals, broken bryozoa, shells, whole planktonic foraminifera and nannofossils (3.2.1) within a dirty, altered glassy matrix. They represent turbidites (Bouma units A, B, D, E) supplied from the reworking of underlying volcanic rocks and colonised in the marine realm below wavebase.

- **Lithofacies 6g** are subaerially erupted, holocrystalline, basaltic andesite lava flows, composed of plagioclase, orthopyroxene, clinopyroxene and iddingsite after olivine with interstitial opaques.

- **Lithofacies 5b, 5g, 5h** (above), massive and crystal tuffs, outcrop as parallel-bedded units (Fig. 8.25) within the Castelsardo Member. There are few clues to the depositional environment of these volcanic rocks, and they may have been the products of either subaerial or subaqueous mass flows or fallout from proximal eruptions.

### 8.4.4.3 Type section, east of Castelsardo

East of Castelsardo, a regressive marine to non-marine sequence commencing with offshore marine volcanic sandstones-siltstones and capped by the subaerially erupted τ2 ignimbrite (Tergu Formation) outcrops along the coast from [770293] to [775297] (Figs. 8.51, 8.52, 8.28, 8.29, 8.55). The basin filling units were deposited in a small, E-W trending half graben defined by the Castelsardo fault and a proposed fault just offshore (Fig. 8.1).

The lower units (Lithofacies 6f) represent turbiditic deposition of volcanic material (?reworked Vaginella Member), with reworked marine fossils, carbonised wood fragments, pine cones and intense bioturbation. The rocks represent water depths below wavebase, probably greater than a few tens of metres. Passing up the succession into well sorted, massive, green sandstones and bioturbated sandstones with abundant shell fragments (Lithofacies 6c, 6e, basal Punta Viuledda cliffs, Fig. 8.28), the sediments record the influx of quartz clasts, thought to have been sourced dominantly from the Palaeozoic granitic basement. Today the granitic basement outcrops to the east, but was perhaps exposed in a fault scarp to the south in Oligo-Miocene times (8.4.4.5). Beds at the base of Punta Viuledda cliff contain iron-stained *Diplocraterion* and *Ophiomorpha* trace fossils. The base of the cliff is cross-cut by a higher angle (15°) erosion surface, lined with granitic pebbles and large *Pecten*
sp. (Fig. 8.28). Above the erosion surface prominent, indurated beds (Fig. 8.28) consist of *Thalassanoides* and other centimetre-sized, superimposed, cross-cutting ichnofabrics deposited below wavebase. A change to orange coloured, cross-bedded gravelstones (Lithofacies 6d) over a truncation surface (Fig. 8.29) records the influx of rounded, granitic pebbles. The stacked, prograding cross-bedded units with no marine fauna record the build-out of small Gilbert fan deltas in a probable marginal marine setting. Overlying this, conglomerates and cross-bedded conglomerates containing rounded and polished granitic and porphyry clasts with no marine fauna are interbedded with lapilli tuffs (Lithofacies 6c, 6d, 5b, 5h) and pass upwards into parallel-bedded tuffs (Lithofacies 5b, 5h) and the τ2 ignimbrite. These upper beds were probably deposited in a subaerial environment. Cross-bedding from Punta Viuledda indicates northwest to northeast palaeoflow orientations, off the fault block dip slope into the hangingwall depocentre.

**8.4.4.4 West of Castelsardo**

The coastal section west of Castelsardo exposes a similar series of offshore tuffaceous, bioturbated turbidites (lithofacies 6f), passing up into massive granitic sandstones (lithofacies 6c) and interbedded lapilli tuffs (lithofacies 5b, 5g, 5h) to that exposed in the east. As Figure 8.55 shows, the western succession which occurs in a similar structural situation, is less dominated by basement-derived clastic sediments and more dominated by tuffaceous lithologies.

The Castelsardo west quarry outcrop [755287] exposes a succession of massively bedded sandstones and conglomerates, cross-bedded conglomerates (Fig. 8.50) and matrix-supported conglomerates (Lithofacies 6b-d) associated with syn-sedimentary normal faulting on the Fragiagu fault, located ~30m beneath the τ2 ignimbrite (Fig. 8.24). Whilst the facies types are similar to those observed elsewhere, the composition of the rocks - well sorted pre-rift basement and poorly sorted volcanic clasts in a tuffaceous matrix - suggests a diverse source area. Centimetre-sized basement clasts within a green, upper conglomerate bed are commonly rounded, polished and faceted with a triangular morphology indicative of subaerial, aeolian reworking (ventifacts, Kocurek 1996) before incorporation of the larger volcanic clasts and debris flow redeposition.

**8.4.4.5 Strada Bianca section**

Cuttings up the Strada Bianca (white road) provide a vertical succession through the Castelsardo Member (Fig. 8.55) in the hangingwall of the San Giovanni fault and in the transfer zone between it and the Ortigiu fault (Fig. 8.12). Sedimentation commenced with a fining upwards sequence of massive boulder-pebble breccias with cross-bedded and massive conglomerates and gravelstones (lithofacies 6a-d) to graded, bioturbated sandstones (Lithofacies 6e). Rare, indistinct bioturbation traces in the basal part of the exposure and a clear marine influence at the top of the exposure (*Thalassanoides* and gravel lined burrows) indicate that this basal part of the Castelsardo Member was deposited in a marine environment rather than the fluviatile environment proposed by Quesney-Forest.
and Quesney-Forest (1984). The sediments are envisaged to have formed as a shallow marine debris cone sourced from an exposed fault-scarp or scarps where the fining-up signature records the degradation of the fault scarp. Continuing up the road, breccia and debris flow units occur directly associated with syn-sedimentary fault activity (Fig. 8.23) and again exhibit a fining-upwards trend to a unit of normally graded, gravelstones-siltstones with climbing ripples and bioturbated tops (Lithofacies 6a-c to 6e). Climbing ripples and cross-beds indicate N-NNE palaeocurrent directions, towards the half-graben basin. A third fining upwards series can be identified at ~131m [875273], where the erosively based, coarsest conglomerates and overlying conglomerates contain a high percentage of disarticulated and broken oysters plus other shell fragments suggesting a proximal zone of colonisation.

8.4.4.6 Other Exposures
Scattered outcrops expose sandstones and gravelstones in the footwall and hangingwall of the San Giovanni fault (e.g. Punto Molino [790273] Fig. 8.49, south of San Giovanni [796257]). Basaltic andesite lava flows and breccias are found at, and southwest of, Pt Molino [782258]. The location and extent of the volcanic rocks (Fig. 8.1, Enclosure 5) may result from fissural eruption along the proposed San Giovanni fault.

8.4.4.7 Correlations and trends
The Castelsardo Member outcrops as discontinuous wedges within the Castelsardo sub-basin (Figure 8.1, Encl 5), laterally equivalent to the upper Vaginella Member. Figure 8.31 shows logs from the most extensive outcrops of the Castelsardo Member. Logs 1-3 show a regressive trend where offshore, volcanic rich marine sediments pass upwards into marginal marine clastic sediments, tuffs and a lava flow. The provenance of the clastic sediments indicates a source in the pre-rift basement lithologies, particularly granite, exposed today to the southeast. It is unclear how the sediments were transported to this area and when aeolian processes operated. Log 4 from the Strada Bianca section is dominated by clastic sediments in three main fining upwards cycles. The clasts were supplied locally, perhaps after fault movement, scarp exposure and degradation. The thickness and lateral variability of the Castelsardo Member wedges can in part be correlated to structural position. For example, the thickest, clastic dominated succession of Log 4 was deposited within a transfer zone adjacent to pre-rift basement (see 8.12). Log 1, being furthest west and closest to the volcanic source is most dominated by reworked volcanic material and log 3 containing a lava flow is located near a fault along which fissural eruption may have occurred.
8.4.5 Tergu Formation (mid Oligocene-mid/late Burdigalian)

Volcanic rocks, commonly andesite domes, flows and breccias, dacitic ignimbrites and pumice-lithic tuffs are here classified as the Tergu Formation with a characteristic type area west of Tergu [around 735225]. The rocks crop out as an andesitic complex, capped by ignimbrites in the west and south-west of the Anglona study area and form the northerly continuation of the Logudoro Group (chapter 7). The Tergu Formation is a lateral equivalent of the Elefante and Perfugas Formations where eruption took place from 31 Ma to 18.4 Ma (chapter 3). The t2, red, dacitic ignimbrite of the Tergu Formation which is extremely widespread over the Anglona study area is of particular importance for stratigraphic correlation within the basin fill.

8.4.5.1 Previous work

Hypersthene-augite andesites, andesite breccias and ignimbrites (called trachyandesites in some publications) have long been recognised in the west of the Anglona study area (Servizio Geologico Sheet 180 1953; Coulon and Dupuy 1975; Dostal et al. 1976; Savelli et al. 1979; Spano and Asunis 1984; Oggiano 1987). Geochemically, the volcanic rocks can be classified as calcalkaline, high-K calc-alkaline and shoshonitic (Beccaluva et al. 1994; Mameli and Oggiano 1997). Spano and Asunis (1984) describe the red-brown colour, fluidal, fiamme and vitrophyric textures and the incorporated lithic and ignimbrite clasts within the t2 ‘upper’ ignimbrite. Mameli and Oggiano (1997) outline a volcanic succession consisting of a basal andesitic complex with domes and intercalated breccias, a volcano-sedimentary continental (termed here Valledoria Member) and marine (termed here Vaginella Member) unit with local intercalations of eruptive rocks and quenching autobreccias, an upper andesite/basaltic andesite flow and an uppermost pyroclastic flow of acidic composition characterised by basal obsidian, a fiamme rich middle (t2) and an upper, less welded top.

8.4.5.2 Rock Types

The Tergu Formation was not examined in great detail though five main rock types can be easily distinguished.

- **Lava flows**, 5-20m in thickness are commonly massive or jointed, porphyritic with millimetre sized plagioclase and/or pyroxene phenocrysts. They occasionally contain lithic fragments and have vesicular or brecciated tops (Fig. 8.56). Multiple flows are exposed as a series of ridges on hillsides. Lava flows are commonly clinopyroxene-orthopyroxene-plagioclase andesites, sometimes with a glassy matrix (Fig. 8.57).

- **Lava domes/volcanic constructions** are suggested by the nature of outcrops on present day 'mounded' topography and are described by Mameli and Oggiano (1997). They represent ancient volcanoes which would have supplied andesitic and dacitic material to the surrounding area.
• **Volcanic breccias and agglomerates** are common. Monomict autobreccias at the top or leading edge of lava flows probably formed due to mechanical friction (Fig 8.56). Polymict epiclastic units, some with a tuffaceous matrix were the products of rock falls and debris flows (Fig 8.58). Some agglomerate outcrops contain oyster fragments [750250] and small bivalves and gastropods [786265] within the matrix indicating deposition in a marine setting.

• **Ignimbrites**

The stratigraphically lowest ignimbrite beneath the Casteldoria Member near Viddalba [903284] is composed of pumice and ash with plagioclase, quartz, biotite and amphibole crystals, lithic (volcanic and basement) fragments. The pyroclastic flow has a texture of aligned crystals and aligned elongate vesicles, is well indurated, jointed and may be welded.

The τ2 'upper' dacitic ignimbrite is composed of plagioclase, rare quartz, rare unidentified mafic crystals and volcanic clasts enclosed with a laminated, welded, red-brown glassy matrix (Fig. 8.59). Vesicles, obsidian and glassy shards are common, aligned and elongate (Fig. 8.60). The ignimbrite is ~20m thick on average, has a typically red-brown, jointed appearance and forms extensive natural outcrops and topographic ridges in the Anglona study area (e.g. Fig. 8.11).

The ignimbrites were deposited from dacitic pyroclastic flows which are commonly derived from the collapse of eruption columns or lava domes (Fisher and Schminke 1984, Cas and Wright 1992). The large areal extent and thickness (implies >50 km$^3$ volume) of the τ2 ignimbrite must have resulted from a huge eruption, but ignimbrites of this and of a much greater size have been observed in other parts of the world (e.g. Upper Bandelier Tuff, USA 200 km$^3$ Cas and Wright 1992).

• **Pumice-lithic lapilli tuffs** are commonly found overlying the τ2 ignimbrite in the southwest of the Anglona study area (Fig. 8.61). They outcrop as massive, ungraded beds and could represent the proximal airfall products of volcanic eruption or unwelded block, ash pumice flow deposits (after Fisher and Schminke 1984, Cas and Wright 1992).

8.4.5.3 **Lithofacies arrangements and interpreted palaeoenvironment**

The detailed structures and lithofacies arrangements within the volcanic complex are difficult to define because of the amount of exposure. The stratigraphically lowest areas west and southwest of Tergu comprise volcanic constructions, breccias and variably dipping lava flows, some to very high angles (70°, Enclosure 5). The dips of the lava flows do not occur on any simple or consistent trend. Some of the dips could well be 'depositional' though others are probably related to later fault movement, for example, on the Tramontana half-graben bounding fault. Stratigraphically higher lava flows, ignimbrites and tuffs are much flatter lying (zero to ten degrees e.g. [760175]) and were erupted after major fault disruption. The alignment of the Tergu Formation volcanic rocks with the Tramontana fault suggests that the fault had some control on the location of andesitic volcanic centres. The Tergu Formation is interpreted to represent a chain of topographically high, largely subaerial, andesitic
volcanoes which erupted andesitic lavas and were reworked by epiclastic processes (Fig. 8.47). The volcanoes also erupted explosively with the production of large volumes of expanded pumiceous material, the fragmentation of lithified volcanic rocks, ignimbrite flows and volcanic fallout. The products of explosive eruptions were supplied into the more basinal zones of the Anglona study area where the volcano-sedimentary Vaginella Member was deposited (Fig. 8.47).

8.4.6 Perfugas Formation (early-mid Burdigalian)

The Perfugas Formation consists of finely bedded lacustrine sediments and tuffs (Lithofacies 1h, 1j, 1k 5a, 5b, 5e, 5g, Table 8.4, 8.5) which unconformably overlie the ‘t2’ ignimbrite of the Tergu Formation within the synclinal Perfugas sub-basin. The sediments are exposed in road cuttings in the south-east of the Anglona study area e.g.[909217] and are up to ~30m thick.

8.4.6.1 Previous work

Known locally as the ‘lacustre’, the Perfugas Formation has not before been considered as a different unit than the stratigraphically lower Valledoria Member (beneath τ2) since the two exhibit similar characteristics (e.g. Servizio Geologico 1953). In their detailed examination, Quesney-Forest and Quesney-Forest (1984) describe interbedded grey ‘tuffs’ and ‘ashes’, laminated lacustrine carbonates with silica and fluviatile conglomerates. Interbedded tuffs and ashes, common around Laerru and Martis are associated with banks of calcareous ‘oolites’ and concretions which formed around the roots of ancient trees found within this region (petrified forest, Quesney-Forest and Quesney-Forest 1984). Laterally equivalent, laminated limestones, marlstones, ashes and fine yellow sandstones rich in organic material show affinities to varved sedimentation (Quesney-Forest and Quesney-Forest 1984). Lagoonal ostracods indicate a ‘paralic’ environment near the top of the succession (Quesney-Forest and Quesney-Forest 1984).

8.4.6.2 Lithofacies

The Perfugas Formation can be described by seven lithofacies (Tables 8.4, 8.5).

- Finely laminated limestone beds (few mm- 20cm, Lithofacies 1h) composed of lime mud with pellets, radial and concentric ooids/oogones are characteristic (Fig. 8.62). Larger concentric algal structures up to 40cm size are found around silicified plant/tree stems within the lacustrine limestones (e.g. Bulzi area [861215], Fig. 8.63).
- Partly chertified limestones and cherts (Lithofacies 1j) are composed of concentric crystalline carbonate (oolites, oncocolites ?) and textureless carbonate ‘blobs’ within a chert matrix (Fig. 8.64).
- Lithofacies 1k is a fine yellow sandstone, bedded from a few to tens of centimetres and with rare normal grading (Fig. 8.62). The lithofacies is thought to represent debris flow to high concentration turbidite deposits. No distinct clasts were identified within the rock to hint at its provenance.
• Parallel-bedded pumice and crystal lapilli tuffs and tuffs (Lithofacies 5a, 5b, 5e, 5g) are also found within the Perfugas Formation

8.4.6.3. Facies arrangements and depositional environment
In the south-central part of the Anglona study area, closer to the Tergu Formation volcanic source area, the Perfugas Formation consists mainly of interbedded lapilli tuffs and tuffs (Lithofacies 5a, 5b, 5e, 5g) with intercalations of lacustrine limestones. The 'petrifed forest' comprising silicified tree trunks surrounded by a concentric irregular algal coat attests to the presence of a continental or lagoonal environment in this area (Fig. 8.65). In the southeastern part of the Anglona study area, the Perfugas Formation comprises finely laminated lacustrine limestones and cherts (Fig. 8.62) with abundant 'oogones', algal oncolites, rare, monospecific gastropods (lithofacies 1h, 1j) and occasional, mass flow deposits (lithofacies 1k). These rocks were deposited in a quiet, carbonate producing, lacustrine environment where plants such as Chara (Charophyte green algae producing oogones and lime mud, Tucker and Wright 1992) must have been ubiquitous.

8.4.7 Laerru Formation
Shallow marine carbonates, calcirudites, calcarenites and marlstones (Lithofacies 2a, 3b-3h, 4b, Table 8.7) which unconformably overlie the Tergu Formation and ?conformably overlie the Perfugas Formation are classified as the Laerru Formation. The Laerru Formation is split into 4 members which exhibit differing lithofacies architecture, locations, structural situations and are likely to be of slightly varying ages (chapter 3). The Sedini Member, outcropping in the south-central part of the Anglona study area is characterised by shallow marine platform carbonates (lithofacies 3b-3e), up to ~50m in thickness, overlying a thin transgressive conglomerate (lithofacies 3g) and passes southwards into the slightly deeper marine Martis Member calcarenites, carbonates and marlstones (lithofacies 3b, 3c, 4b). The Campulandru Member shallow marine calcirudites, calcarenites, grainstones and framestones (Lithofacies 3b, 3d, 3g, 3h) outcrop around Castelsardo overlying the faulted τ2 ignimbrite. The Sennori Member conglomerates, calcirudites, platform carbonates and marlstones (Lithofacies 2a, 3b-3g, 4b) exposed in the far west of the Anglona study area represent the northward continuation of the Florinas Group (chapter 7) and were deposited over a complex faulted topography. With the exception of the Campulandru Member which was studied in detail, the Laerru Formation sediments were examined with an emphasis on large scale facies architecture and sediment geometries rather than on detailed lithofacies variations.
8.4.7.1 Previous Work

Around Castelsardo (Campulandru Member), carbonates rich in corals (*Thegiosaura miocenica*, *Helioastrea* *sp.* and *Porites* *sp.*), bryozoa, *Lithothamnium*, oysters, pecten, other bivalves (e.g. *Chlamys*) and echinoids (*Clypeaster* *sp.*), overlying a basal transgressive conglomerate and with 'biohermal' (reefal) patches are documented by the Servizio Geologico (1953), Maxia and Pecorini (1969) and Spano and Asunis (1984). Oggiano (1987) and Francolini (1994) mapped and dated a basal sandy-conglomeratic unit, carbonate unit, marly and marly arenaceous units in the west of the Anglona study area (Sennori Member). Oggiano (1987) notes that the basal conglomerates exhibit a provenance from Gallura, the carbonates are rich in corals, *Lithothamnium*, *Clypeaster* and bivalve pieces supplied from reef reworking and that marlstones are bioturbated and *Spangatoida* echinoid bearing. Quesney-Forest and Quesney-Forest (1984) examine the 'Sedini limestones' (Sedini Member) in the south-central part of the Anglona study area, noting the progressive upward change from conglomerate to carbonate rocks with a wide marine fauna (bivalves, foraminifera, *Lithothamnium*, bryozoa) deposited in an open, littoral, agitated shallow marine carbonate platform. The rocks are classified as packstones and grainstones with sedimentary structures including 8m scale cross-bedding composed of 7-8cm diameter rhodoliths. The Sedini limestone thickens to the south reflecting the Burdigalian palaeotopography (Quesney-Forest and Quesney-Forest 1984).

8.4.7.2 Lithofacies

The Laerru Formation can be described by the seven Lithofacies 2a, 3b, 3c, 3d, 3e, 3g and 4b which were introduced in chapter 5 and lithofacies 3h which is introduced here (Table 8.7).

- **Lithofacies 2a** consists of disorganised, clast-supported conglomerates sourced from the eastern, Gallura area pre-rift basement, with scattered shell fragments.
- **Lithofacies 3b** are calcarenites, characteristically containing coarse sand grade quartz, volcanic and basement clasts, whole and broken bivalve shells, red algae and echinoids with some *Thalassanoides* bioturbation.
- **Lithofacies 3c** consists of wackestones with broken shell fragments and lime mud pellets in a lime mud and patchy microspar matrix.
- **Lithofacies 3d** are massively bedded grainstones and packstones with common bryozoan, rhodoliths, corals, bivalve shells, benthic forams, micritized pellets and partly micritized shell fragments within a patchy lime mud, microspar and spar matrix.
- **Lithofacies 3e** comprise cross-bedded packstones and grainstones with sets up to ~20m high, a metre or so thick and with angles of 10-15°.
- **Lithofacies 3g** is a calcirudite composed of boulder-gravel sized basement, volcanic and quartz clasts within a carbonate matrix containing broken shells, corals, and red algae.
• **Lithofacies 3h** consists of reef framestones-bindstones (after Embry and Klovan 1971 and James 1984 in Tucker and Wright 1992) with in situ corals and algal buildups. *Porites sp.* and *Favites Neglecta*, common Miocene reefbuilders (B. Rosen pers. comm. 1996), are found in the Campulandru Member along with other poorly preserved coral types. Coral pieces are commonly 50cm in diameter and have a hemispherical morphology where the space is infilled by lime mud and encrusting algae (e.g. Fig 8.66). Such reefstones would have built up in a shallow marine environment within the photic zone, with strong currents and away from clastic sediment supply (e.g. Tucker and Wright 1992; Wright and Burchette 1996). Rudstones derived from the framestones (i.e. reef talus) are occasionally found as small, associated patches e.g. [756294].

• **Lithofacies 4b** consists of marlstones and bioturbated marlstones with occasional syn-depositional slumping.

8.4.7.3 *Campulandru Member (?mid-late Burdigalian)*

The Campulandru Member exhibits lateral facies variations in response to marine transgression onto a complexly faulted topography (Fig. 8.67). On the scarp of the Castelsardo fault at [756294] and [762293], boulders of the τ2 ignimbrite with an oyster grainstone matrix (Fig. 8.68) fine upwards over a few metres through an oyster-red algae calcirudite (Lithofacies 3g) into a *Porites sp.* rich coral-algal rudstone then framestone (Lithofacies 3h, Fig. 8.66). The outcrops are interpreted as a fringing reef which developed on the degraded Castelsardo fault scarp (Fig. 8.67). Large coral boulders are found in small patches on the footwall high of the Lu Pozzu [760289] fault scarp and on topographic ‘benches’ on Campulandru hillside [785296]. Although not *in situ*, it seems likely that these large pieces reflect the presence of coral framestones derived from the unexposed, underlying bedrock and that small patch reefs developed on footwall crests and fault block dip slopes (Fig. 8.67). South of Castelsardo, the τ2 ignimbrite is folded into a syncline adjacent to the Castelsardo fault (Fig 8.1 A-A', B-B'). In the core of the syncline and in the hangingwall of the Castelsardo east fault [762292], calcarenites and calcirudites (Lithofacies 3b, 3g) were deposited, whilst on the hangingwall dip-slope of the Lu Pozzu fault [762289], grainstones with coral, oysters, *Pecten*, echinoids and bivalve fragments, benthic macroforaminifera (*?Amphistegina* and two other types, Fig. 8.69) and *Thalassanoides* traces overlie a basal calcirudite containing τ2 ignimbrite clasts. Thus, clastic material was supplied into hangingwall lows whilst reefs fringed footwall highs and fault scarps producing material which was reworked on fault block dip-slopes (Fig. 8.67).
8.4.7.4 Sedini Member (?mid/late Burdigalian - early Serravalian)
The Sedini Member platform carbonates thicken from a few to 50m, to the south (Quesney-Forest and Quesney-Forest 1984) into the Perfugas sub-basin syncline, and west into the hangingwall depocentre of the N-S trending Sedini Fault (Fig. 8.72).

North of Sedini [846248], the thinnest units expose the transition from a basement and volcanic (τ2 and tuff) conglomerate, through a calcirudite with biogenic colonisation by red algal sticks and rhodoliths (Lithophyllum and Lithothamnium, Fig. 8.70), to red algal grainstone over a few metres (Lithofacies 2a, 3g, 3b, 3d, Fig. 8.71). This succession records marine transgression. Laterally, in the hangingwall of the Sedini fault [840234], wackestones, calcarenites and marlstones with in-situ, smooth shelled infaunal bivalves (2 types ?Nuculoidea sp.), ?Schizaster echinoids, Ophiomorpha and Planolites traces (Lithofacies 3b, 3c, 4b) were deposited in a calmer, deeper environment reflecting the palaeotopography (Fig 8.72).

To the south of Sedini [845220], the basal calcirudites (Lithofacies 3g) commonly contain angular basement clasts, oyster, bivalve and red algae fragments with ?Thalassanoides trace fossils. The overlying platform carbonates (grainstones and packstones, Lithofacies 3d) also contain coral pieces and large red algal rhodoliths. Cross-bedded units within the carbonate platform (Fig. 8.73) show south to south-south-east progradation directions. On a large scale, the platform carbonates exhibit a ramp-like geometry, dipping at 7° and prograding to the south (Fig. 8.74). At around [850197], the thick carbonate platform units are replaced by calcarenites and deeper marine marlstones (Lithofacies 3b, 4b) of the Martis Member.

The Sedini Member is interpreted as an open, high energy carbonate platform (as Quesney-Forest and Quesney-Forest 1984) which progressively onlapped, covered and filled topography in response to marine transgression. A series of carbonate sand bars developed on the platform which had a ramp-like geometry which deepened to the south. Facies architecture and carbonate thickness was partly controlled by the N-S trending Sedini fault which is interpreted to have been active before and after Sedini Member deposition (8.3.4, Fig. 8.72).

8.4.7.5 Martis Member (?mid/late Burdigalian-early Serravalian)
The Martis Member consists of a basal calcarenite, calcirudite or grainstone (Lithofacies 3b, 3g, 3d) overlain by a series of poorly exposed marlstones (Lithofacies 4b; Fig. 8.75) with intercalated 'patches' of prominent weathering calcarenite, tuffs and lapilli tuffs (Fig 8.1, Encl. 5) and an uppermost calcarenite/calcirudite cap (e.g. Chiaramonti [847113]). The sediments occur in a ENE-SSW oriented strip in the core of the Perfugas sub-basin syncline. The basal carbonates which outcrop at many different topographic levels (e.g. around Nulvi) clearly show a marine transgression over an
irregular topography, yet it is unclear to what extent faulting controlled this topography (8.2.4). Calcarenites and calcirudites commonly consist of pre-rift basement clasts, dominantly granite, which was probably supplied 'axially' along the Perfugas sub-basin from the northeast. At Chiaramonti [853106], the basal calcarenite which overlies the Tergu Formation pumice-crystal lapilli tuff is rich in volcanic clasts and biotite crystals (Fig. 8.76). It also contains a bed of in-situ *Scutella sp.* and *Amphiope transversifora* echinoids (Odin et al. 1994).

8.4.7.6 Sennori Member (Late Burdigalian-early Langhian)
The Sennori Member outcrops on top of and adjacent to faulted Tergu Formation volcanic rocks at the margin of the Portotorres sub-basin. The generalised succession is a basal pre-rift basement derived conglomerate (Lithofacies 2a; Fig. 8.77) passing upwards into calcirudites and calcarenites (Lithofacies 3g, 3b; Fig. 8.78), platform carbonates (Lithofacies 3c, 3d, 3e) and basinward into marlstones (Lithofacies 4b; Figs. 8.79, 8.1, Encl. 5). A faulted topography seems to have existed before deposition of the Sennori Member (see 8.4.5) and, with marine transgression, exerted a control on lithofacies architecture. For example, the Monte Uri succession is located within a 'transfer zone' between the NNW-SSE trending Tramontana fault and the NE-SW trending Sennori, San Giusta faults (Fig. 8.15). Within the transfer zone, the basal units comprise a thick siliciclastic-carbonate succession. In other topographically higher areas, platform carbonates were deposited directly on the Tergu Formation (Fig. 8.15). At Sennori [6631521, road cuttings with massive banks and faint channels filled by large rhodoliths (5cm) and thick, broken bivalve shells are interpreted as the margin of a high energy carbonate platform on the Sennori fault footwall crest. In the fault hangingwall, deeper marine marlstones (Fig. 8.79) with whole *?Nuculoidea sp.* bivalves, 10cm sized high spired gastropods, *Ophiomorpha, Thalassanoides* and other, unidentified trace fossils are indicative of deeper marine sedimentation, below wavebase. The marlstones commonly exhibit parallel layers of irregular carbonate cementation (Fig 8.79) where some of the cementation occurs preferentially around *Thalassanoides* burrows. Thus the sediment architecture of the Sennori Member shows lateral facies changes relating to sediment supply and marine bathymetry.

8.4.8 Summary and relationships between basin filling units
The Elefante Formation exposed in the Anglona study area records basal alluvial fans and fluvial deposition, shed from high fault scarps into lacustrine, half-graben depocentres (Casteldoria and Valledoria Members). Large amounts of material supplied from explosive volcanic eruption is believed to have been deposited in a lacustrine and then marine subaqueous environment (Vaginella Member). Finally, a regressive sequence of offshore marine to marginal marine volcanic and clastic rocks were deposited (Castelsardo Member). Contemporaneous with the deposition of the Elefante Formation, volcanic rocks of the Tergu Formation were erupted from andesitic volcanoes in the west and south-west of the study area. Overlying the areally extensive t2 ignimbrite, the lacustrine
sediments of the Perfugas Formation and shallow marine carbonates, calcarenites and marlstones of the Laerru Formation were deposited on an irregular faulted topography. In the west of the study area, volcanic and volcanioclastic rocks are common whilst in the east, clastic sedimentary rocks are exposed.

8.5 Miocene basins of northernmost Sardinia and Corsica

8.5.1 Castelsardo to Capo Testa

A linear depression filled with patches of Miocene dacitic lapilli tuff and more recent continental conglomerates runs along the line of the strike-slip fault identified by Carmignani et al. (1987), Barca et al. (1996) from the northeastern margin of the Castelsardo sub-basin towards the northeast. The fault bends to trend northwards and intersects Capo Testa in the far north of Sardinia. At Capo Testa, a small basin (few km$^2$), filled with late Burdigalian-Langhian sediments (Cherchi 1974; Assorgia et al. 1997) was formed by the N-S trending fault. A discrete, high angle fault plane crops out within granite and bounds the eastern side of the basin which is filled with nodular, shallow marine carbonates and marly sandstones. Thus the sinistral strike-slip fault (Barca et al. 1996) must have been active sometime prior to the mid-Miocene with a component of extension.

8.5.2 Corsica

Three Miocene basins of Corsica were examined at reconnaissance scale.

The St Florent basin in northern Corsica is an intermontane basin resulting from ductile and brittle east-west extension following late Cretaceous-Oligocene crustal thickening of 'Alpine Corsica' (Jolivet et al. 1991; Egal 1992). It was filled by Burdigalian age sediments (Orzag-Sperber and Pilot 1976). The basin fill comprises a basal, basement-derived conglomerate, calcirudites with rhodoliths coating basement clasts, calcarenites, rhodolith grainstones, graded calcirudite debris flows and bidirectional cross-bedded units. The abundant rhodoliths and marine fauna are similar to those observed in Sardinia. The Francardo basin, north of Corte, is another Miocene intermontane basin within 'Alpine Corsica' with fluvio-lacustrine sediments and Burdigalian age marine marlstones (Orzag-Sperber and Pilot 1976).

Western and southern 'Hercynian' Corsica consists of a Palaeozoic granitoid basement cut by extensive NE-SW to N-S trending faults which sinistrally offset units in the granitoid basement (Carte Geologique de la France, Corse, 1980). These faults controlled the formation of Miocene basins in southern Corsica. The Plaine Orientale basin (Orzag-Sperber and Pilot 1976) of central eastern Corsica was bounded by the continuation of a NNE-SSW trending fault which runs through the granitoid basement. The fault defines at the present day, a high, basement topography and low, poorly
exposed sedimentary basin. No outcrops of the fault were found to deduce the fault kinematics. In the Plaine Orientale basin, conglomerates, sandstones and marlstones of Langhian-Messinian age are exposed (Orzag-Sperber and Pilot 1976).

The Bonifacio basin of southernmost Corsica (Fig 8.80) was studied in the most detail because of its vicinity to Sardinia. The basin was defined by a NE-SW to NNE-SSW trending, main bounding fault in the northwest. The fault is a part of a larger structure running to the NNE through the granitoid basement which shows sinistral offset (Carte Geologique de la France, Corse, 1980). Fractures within the granitic basement close to the fault zone align with the fault (Fig 8.80) but no slickenside data were found to determine whether the basin was formed due to oblique sinistral strike-slip or due to dip-slip fault movements. Smaller faults on various trends are tentatively identified at the contact between the granitic basement and sediment cover (e.g. SE and NE of Maora, Fig. 8.80). The basin is envisaged as a 'half-graben' type structure with onlap to the SE over the basement such as is observed east of the Phare di Pertusato (Fig 8.80). Sediment geometries are indicative of basin filling after a phase of basin formation (post-rift sedimentation).

The basal basin fill comprises a few metres of conglomerate-calccrudite, carbonates with a diverse marine fauna (e.g. bryozaos, bivalves, macroforaminifera), a palaeosol and boulder horizon and/or granitic sandstones. In the NE corner of the basin, the unconformity over fractured granite in the hangingwall of the bounding fault is observed, whilst a reef rudstone outcrops the same structural situation in the far SW (Fig. 8.80). Cross-bedded calcarenites composed of granite, granite derived quartz, feldspars, broken shells and red algae within a lime mud or sparry matrix form the majority of the Bonifacio basin fill. Sedimentary structures within the cross-bedded lithofacies include complex superimposed arrangements of bidirectional cross beds with 'swept out' geometries forming sigmoidal sets (Fig 8.81, 8.82), smaller ripples superimposed on metre scale sandwaves and channels filled with cross-bedded units (Fig. 8.82). The structures are indicative of a shallow marine, tidal environment of deposition (after Homewood and Allen 1981; Johnson and Baldwin 1996). Cross-bedding indicates that palaeocurrents were oriented N-S to NNW-SSE around Bonifacio and westwards in the Pertusato area (Fig. 8.80) suggesting a rather peculiar system of tidal currents.

Thus, the Miocene basins of southern Corsica were formed either by sinistral strike slip fault movement or normal fault movement on these existing structures and were filled by shallow marine sediments from the Late Burdigalian.
8.6 Timing and kinematics of extension

Sediment and structural geometries are here used to suggest when, and how, extension occurred in the Anglona study area, and what relationship this extension may have to other basins of northern Sardinia and Corsica.

8.6.1. Previous Work

Based on their seismostratigraphic study, Thomas and Gennesseaux (1986) suggest Oligocene to mid Aquitanian and lower Burdigalian rifting phases within the Castelsardo and Portotorres half graben sub-basins. The second phase was thought to be contemporaneous with a short ‘distensive’ phase between Corsica and Sardinia, the tilting of the Castelsardo sub-basin and the creation of the continental slope. From their field based study of the eastern Anglona study area, Quesney-Forest and Quesney-Forest (1984) suggest NE-SW Plio-Quaternary extension (Concatile, Bulzi, Sedini faults), Messinian NW-SE trending compression and movement on N-S strike-slip faults, NE-SW oriented Burdigalian compression and Oligocene-Aquitanian NNW-SSE oriented extension. As discussed above (8.2.4) there is little evidence for anything other than local compressional or strike-slip movements.

8.6.2 Timing of extension in the Anglona study area

The first phase of extension cuts across sinistral strike-slip faults within the Hercynian basement which were identified by Carmignani et al. (1987), Barca et al. (1996; Fig. 8.83). The strike-slip faults of northernmost Sardinia and Corsica may have moved in the late Oligocene or earlier, in common with areas to the south (chapter 6), resulting in the accommodation space which was filled by ?Miocene age volcanic rocks and Miocene sediments. Strike-slip faults were followed by ?late Oligocene-mid Burdigalian extensional events. Three main phases of Oligo-Miocene extensional faulting can be identified in the Anglona study area. The development discussed below is consistent with, but not dependent on, the seismic reflection data. The phases can be identified as:

Phase 1 - ?Late Oligocene movement on NNW-SSW normal faults creating the Castelsardo and Portotorres half graben sub-basins (Fig. 8.83). The timing of this phase is poorly constrained by the age of marine sediments higher up in the succession. The faults appear to be new structures which cut across the existing NE-SW lineaments.

EVIDENCE: The basal, continental Casteldoria member was deposited contemporaneous with fault movement and shed westward and basinward. The basal Valledoria Member contains syn-sedimentary N-S normal faults.
Phase 2 - mid Aquitanian-early Burdigalian extension on E-W to NW-SE trending normal faults (Fig. 8.83).

EVIDENCE: The Castelsardo Member was deposited contemporaneously with movement on the Castelsardo, Lu Pozzu, Fragiagu and Ortigiu faults. Faulting in this orientation also occurred after the eruption of the τ2 ignimbrite but ended before marine transgression (Laerru Fm.) and lacustrine deposition (Perfugas Formation).

These faults are thought to have formed as an extensional system related to the opening of the Bonifacio Straits between Sardinia and Corsica as the microplates rotated in from 21-18 Ma. In this scenario the overall extension direction would be -NNE-SSW (Fig. 8.83). This extension phase may also have caused normal reactivation of the fault bounding the Bonifacio sub-basin in southern Corsica.

Phase 3 - early/mid Burdigalian extension on NNW-SSE to N-S faults (Fig. 8.83).

EVIDENCE Movement on the Tramontana/Sennori faults is thought to have caused tilting of lithologies in the footwall before and after the eruption of the τ2 ignimbrite, but before the deposition of the Laerru Formation. The Sedini fault was active after τ2 ignimbrite, but before the deposition of the Martis Member.

Thus, in the early/mid Burdigalian, basin filling geometries indicate that just before and after the eruption of the τ2 ignimbrite both E-W and N-S fault families were active. This is compatible with the observations from offshore made by Thomas and Gennesseaux (1986). In the mid Burdigalian, the extensional stress field changed such that only -N-S structures were active.

Faulting episodes within the Anglona study area lasted for only a few million years and were accommodated on different orientations of faults, presumably in response to a rapidly changing overall stress field. Over the short-lived periods of extension, tectonic subsidence was generally greater than basin filling, resulting in an underfilled tectonic relief at the end of each extension phase. Fault throws on the largest structures may have been up to ~500m in each extension phase. The exception to this was the E-W faulting episode which was recorded by a voluminous clastic-epiclastic sediment supply (Castelsardo Member). Intervening periods of tectonic quiescence and sub-basin filling span timescales of ~4-5 Ma.

In common with the whole of Sardinia, post-depositional, Plio-Quaternary N-S trending normal faults dissect and tilt the Oligo-Miocene succession.
8.7 Tectono-stratigraphic development

This section synthesises all the information on the Anglona study area to reconstruct the Oligo-Miocene evolution of the zone. Figure 8.84 illustrates the tectonostratigraphic development, whilst Figs 8.39, 8.47, 8.67, 8.72 have also given an idea of the facies variations and palaeotopography.

8.7.1 Late Oligocene - first extensional phase and the start of half-graben filling, Fig.8.84A.

Sinistral strike-slip movement on extensive NE-SW faults within the basement occurred sometime before the ?Late Oligocene. Basin formation commenced in the ?Late Oligocene with extension on two large NNE-SSW trending normal faults which formed a tilted fault-blocks and half-graben structure. Continental clastic sediments of the Casteldoria Member were shed westwards from the eastern, Viddalba fault scarp, partly contemporaneous with fault movement and northwards from the Casteldoria horst. The horst may have been a remnant east-west high, probably related to the large strike-slip fault in the adjacent basement. Fine grained lacustrine sediments of the Valledoria Member were deposited in calm waters in the Castelsardo half-graben centre. In the west and southwest of the Anglona study area, andesitic volcanism may already have been active (dubious K-Ar date). This volcanism might have supplied the basal ignimbrite at Viddalba and a limited amount of volcanic ash onto the footwall high of the Castelsardo half-graben where shallower lacustrine sediments such as gastropodal marlstones, ?fluvial channels and cherts were intercalated with tuffs.

8.7.2 Latest Oligocene-early Aquitanian - volcanism and half graben filling, Fig. 8.84B.

A phase of half-graben filling in the latest Oligocene-early Aquitanian is recorded by the rhyolitic-dacitic pyroclastic and/or epiclastic volcanic rocks of the Vaginella Member. The volcanic eruptions most probably took place along a line of active andesitic volcanoes in the vicinity of the Tramontana fault and to the south-west of the Anglona study area. The volcanic centres are now exposed as a unit with variably dipping volcanic cones, lava flows and epiclastic breccias - the Tergu Formation. Acidic pyroclastic flows, surges and ash clouds resulting from explosive eruption, on passing into the Castelsardo sub-basin, entered a lacustrine-lagoonal to shallow marine (from the mid Aquitanian) environment. The resultant basin fill consists of interbedded pumice and crystal lapilli tuffs, tuffs, complexly interbedded tuffs and lapilli tuffs with unusual sedimentary structures and rare epiclastic channels which do not obviously record marine transgression. The basin fill was dominated by the supply of volcanic material.
8.7.3 Mid Aquitanian-early/mid Burdigalian - syn-depositional faulting associated with the second extensional phase, Fig. 8.84C.

A mid Aquitanian-early/mid Burdigalian regressive, coarsening upwards, epiclastic and clastic succession records NW-SE to W-E oriented syn-sedimentary normal fault activity around Castelsardo. Syn-rift sedimentation consists of divergent sediment wedges which thicken into hangingwall depocentres and prograde off tilted fault-block dip-slopes. Erosion surfaces related to relative sea level fall, cut across the clastic sediment wedges. Normal faulting on this trend continued until after the subaerial τ2 ignimbrite at the top of the regressive succession. Some N-S faulting may also have occurred prior to the eruption of the τ2 ignimbrite.

In the area of the proposed ‘antithetic interference zone’ between the San Giovanni and Ortigiu faults, NW-SE trending syn-sedimentary normal fault activity is related to the sudden influx of coarse, basement derived clastic material and then gradual fining upwards to a background sedimentation of graded, marine sandstones and siltstones. Clastic material was probably derived from locally exposed pre-rift basement scarps and was supplied basinward through the topographically depressed transfer zone. Fault movement, which caused a fault propagation fold to develop along the Ortigiu fault, and back-tilting on the San Giovanni fault, ended sometime after the eruption of the τ2 ignimbrite.

8.7.4 Early/mid Burdigalian - third faulting phase and the start of marine transgression, Fig. 8.84D.

After the eruption of the τ2 ignimbrite and before the deposition of the Perfugas and Laerru Formations, NNW-SSE and NE-SW trending normal faults defined the Portotorres sub-basin margin exposed today. Movement on the Tramontana fault, probably caused eastward tilting of Castelsardo sub-basin. A number of N-S to NNW-SSE trending normal faults cut and offset the τ2 ignimbrite within the Anglona study area. Therefore, in the early/mid Burdigalian, before marine transgression, a complicated, faulted and folded topography existed on the τ2 ignimbrite surface. In the footwall of the San Giovanni fault and adjacent to the Casteldoria horst, the Perfugas sub-basin was ‘perched’. Whilst the lacustrine-lagoonal sediments of the Perfugas Formation were deposited in this perched lake, marine transgression occurred in the topographically lowest zone around Castelsardo. The Perfugas Formation is composed mainly of lapilli tuffs closest to volcanic centres and pumice-lithic lapilli tuffs of the Tergu Formation are laterally equivalent. The control of the underlying faulted topography on sedimentation contemporaneous with marine transgression is shown by the Campulandru Member. Coral reefs were established on footwall highs, grainstones and patch reefs developed on fault block dip slopes, whilst reef rudstones and calcarenites accumulated on degraded fault scarps and in hangingwall depocentres.
8.7.5 Mid Burdigalian-early Serravalian - post rift sedimentation, Fig. 8.84E.

Marine transgression resulted in the progressive invasion of the Anglona study area by the sea and the continued onlap of marine sediments onto an existing faulted and folded topography. Shallow marine platform carbonates overlying a transgressive conglomerate occurred at the margins of the sea on footwall highs (Sedini member, Sennori carbonates). Marlstones and calcarenites were deposited in the deeper waters in the Perfugas sub-basin centre (Martis Member) and hangingwall depocentre of the Tramontana fault (marlstones of Sennori Member).

8.8 Summary

- The Anglona study area exposes a complex, laterally variable basin fill succession that accumulated in response to three phases of active extension and two transgressive marine cycles. The area provides a unique example of how clastic, carbonate and volcanic rocks filled an evolving accommodation space.

- The complex structure of the Anglona area is thought to have resulted from three extension phases superimposed on an inherited basement structure of NE-SW trending, sinistral strike slip faults. The first ?late Oligocene extension occurred on NNW-SSE trending normal faults to form a half-graben and tilted fault-block topography. The second, mid Aquitanian-early Burdigalian phase of E-W to NW-SE extensional faulting is thought to have resulted from the opening of the Bonifacio Straits between Sardinia and Corsica as the microplate rotated. The final, early/mid Burdigalian phase of ~N-S trending normal faulting occurred on the same half-graben bounding normal faults as before and other, distributed structures.

- Filling of the Anglona sub-basin commenced with deposition of clastic material adjacent to fault-bounded topographic highs transported in alluvial fan/fluvial systems (Casteldoria Member) and distal lacustrine sedimentation (Valledoria Member). Volcanic centres in the west and southwest of the area, present today as a series of volcanic domes, flows and epiclastic breccias (Tergu Formation), supplied dacitic-rhyolitic pyroclastic rocks which were deposited subaqueously in the half-graben centre (Vaginella Member). A regressive volcaniclastic-clastic succession then recorded syn-rift sedimentation (Castelsardo Member). After the eruption of a widespread ignimbrite and the formation of a perched lake (Perfugas Formation), marine transgression resulted in the deposition of shallow marine platform carbonates, calcarenites and deeper marine marlstones (Laerru Formation). Tidal calcarenites of a similar age to the Laerru Formation outcrop in the Bonifacio basin of southern Corsica.
Chapter 9
Chapter Nine - Geochemistry of Oligo-Miocene arc-volcanics.

The Oligo-Miocene volcanic series of Sardinia has long been recognised as a calc-alkaline-tholeiitic sequence with geochemical signatures typical of subduction-related magmatism (e.g. Dupuy et al. 1974; Coulon and Dupuy 1975; Dostal et al. 1976; Coulon 1977; Dostal et al. 1982; Rutter 1985; Beccaluva et al. 1987). The geochemistry of this volcanic arc succession can provide clues to mantle behaviour during Sardinian rifting events (Chapters 4-8, 10) and the opening of the Western Mediterranean back-arc basin.

Oligo-Miocene volcanic-arc rocks crop out in the extensional sub-basins of southern France, the Gulfs of Lion and Valencia, at the southern margin of the Eurasian plate, and are particularly well exposed in the Sardinian Rift. In common with other continental arc settings such as New Zealand, Greece, the Peruvian Andes (Cole 1984; Pe-Piper et al. 1994; Petford and Atherton 1994), the nature of the association between extension and subduction-related volcanism poses an interesting problem as to the connection between upper-crustal extension and deeper mantle processes. Recent studies from the oceanic-arc setting of the Lau Basin, S.W. Pacific have suggested that asthenospheric upwelling resulting from lithospheric stretching may be an important, additional melt generation process preceding back-arc basin opening (Clift and Dixon 1994; Fig. 9.1a). It was thus of interest to see if any evidence of changes in the extent of melting could be detected through the rifting period, which might be consistent with progressive upwelling of mantle.

This chapter aims to use the major and trace element geochemistry of datable suites of Sardinian basalts and basaltic andesites to deduce whether lithospheric thinning may have been an active process concomitant with upper crustal rifting (Chapters 4-8, 10) or whether a subduction signature was dominant and unchanged. The aim was to understand what happens at depth when arc-magmatism and extension are combined. Appendix 9A describes the X-ray fluorescence techniques used to obtain the major and trace element data, Appendix 9B tabulates the results of the sample analysis, sample locations and normalising factors for spider diagrams.
9.1 Previous Work

9.1.1 Arc-magmatism at active continental margins

Magma generation at active continental margins occurs within one of the most complex geodynamic settings on earth (Fig. 9.1a). This section provides a brief summary of the current thinking on the sources of continental-arc magmas and the processes which modify the magmas within this 'multistage-multisource phenomenon' (Wilson 1993, Tatsumi and Eggins 1995).

The current consensus is that most arc rocks crystallised from parental magmas generated in the mantle wedge overlying the subducting slab (Fig. 9.1a; Hawkesworth et al. 1993; Wilson 1993). Melting results from the lowering of the peridotite solidus due to the infiltration of slab-derived water-bearing fluids or partial melts - the 'subduction-component' (Fig. 9.1; Hawkesworth et al. 1993; Wilson 1993). In oceanic arcs, it is thought to be the asthenospheric mantle wedge which melts, but in the case of continental arcs, the subcontinental lithosphere may also cross the wet peridotite solidus (Pearce 1983; Wilson 1993). By modelling trace element behaviour, Pearce and Parkinson (1993) suggest that in continental arcs, the magma sources vary from fertile MORB source mantle to enriched mantle with a subcontinental lithosphere component.

Many processes may affect a magma before it is erupted at the surface. Fractional crystallisation may occur at many levels as the melt rises. Crustal contamination, magma mixing and volatile loss may also modify the magma composition (Fig. 9.1a). Specific geochemical indicators and systematic element trends can be used to help distinguish between possible magma sources, the amount of partial melting, the extent of fractional crystallisation and alteration history of the magma (e.g. Pearce 1983; Pearce and Parkinson 1993; Rollinson 1993; Wilson 1993, etc.)

9.1.1.1 Typical geochemical characteristics of continental arc-magmas

Continental arcs such as the Sardinian example are commonly composed of voluminous calc-alkaline andesites and dacitic-rhyolitic pyroclastic products (e.g. ignimbrites) with relatively minor basalts and basaltic andesites (Chapters 6-8; Wilson 1993). The rocks often span a range from low-K tholeiites through calc-alkaline and high-K calc-alkaline to shoshonitic compositions (Wilson 1993).

Continental arc-rocks are generally enriched in the incompatible, large ion lithophile elements (LILE; Rb, K, Ba, Sr, Th) and light rare earth elements (LREE; e.g. La) relative to mid-ocean ridge basalt (MORB). The high field strength elements (HFSE; Ti, Zr, Hf, Nb, Ta, Y) are not as enriched, and in some cases may be depleted, producing high LREE/HFSE or LILE/HFSE ratios relative to MORB. The enrichments are believed to result from metasomatism of the mantle source of arc-basalt by fluids released from the subducted slab (Pearce 1982; Wilson 1993). Relative depletions may represent the retention of the HFSE in a stable mantle or slab phase such as rutile or sphene coupled with high
degrees of partial melting (Pearce 1982; Wilson 1993), though this process is not fully understood (e.g. Tatsumi and Eggins 1995).

9.1.1.2 Importance of continental arc rock geochemistry for models of back-arc basin formation
Four models of mantle and crustal dynamics relating to back-arc basin formation have been proposed (e.g. Karig 1974, see section 10.2.1) but the evolution of these often submarine and hence relatively inaccessible zones still remains poorly understood. Observations from the continental upper crust suggest that many back-arc basins formed in response to extensional or transtensional stresses applied to the lithosphere i.e. 'passive' back-arc spreading rather than 'active' back-arc spreading in response to mantle diapirism (Molnar and Tapponier 1975; Dewey 1980; Lallemand and Jolivet 1985; Malinverno and Ryan 1986; Sibuet et al. 1987; Kastens et al. 1988; Faure and Charvet 1990; Charvet and Ogawa 1994; Windley 1995; section 10.2.1). By using geochemistry to test the hypothesis that lithospheric stretching may promote volcanism in intra-arc settings one may be able to assess a fundamental connection between upper crustal and deep earth processes.

9.1.2 Geochemical characteristics and petrogenesis of Oligo-Miocene volcanic rocks of Sardinia.
Major and trace element studies have suggested that the Oligo-Miocene Sardinian volcanic rocks resulted from a north-northwest dipping subduction zone beneath the southern Eurasian plate (e.g. Coulon et al. 1973; Coulon et al. 1975; Dostal et al. 1976; Beccaluva et al. 1987, 1994). Calc-alkaline and tholeiitic rocks are found within southern Sardinia whilst calc-alkaline to shoshonitic rocks crop out in northern Sardinia (Coulon et al. 1975; Dostal et al. 1976; Beccaluva et al. 1987, 1994). The progressive increase in the concentrations of K, Li, Rb, Sr, LREE at a given silica content and the LREE/Y ratio northwards is thought to reflect the depth to the northwards dipping subduction zone (Coulon et al. 1973; Dupuy et al. 1974; Coulon et al. 1975; Dostal et al. 1976; Beccaluva et al. 1987, 1994).

Geochemical and isotopic studies have led to proposals that the Oligo-Miocene Sardinian rocks were produced by partial melting of a MORB source mantle modified by enrichments and depletions related to the presence of subducted oceanic crust (Dupuy et al. 1974; Coulon et al. 1975; Dostal et al. 1976; Rutter 1985; Morra et al. 1997). Melting is thought to have occurred in the garnet lherzolite stability field, from chondrite-normalised REE patterns (Rutter 1985). Morra et al. (1997) mention that the extensional regime in which volcanism occurred could have resulted in the upwelling of lithospheric mantle contaminated by subduction fluids, causing intersection of the hydrous peridotite solidus, but do not quantify the statement or present any specific argument in support of upwelling as a process.
There is abundant geochemical and isotopic evidence that fractional crystallisation controlled the evolution of the Sardinian magma series from basalts to rhyolites (Coulon et al. 1973; Dupuy et al. 1975; Dostal et al. 1976; Savelli et al. 1979; Rutter 1985; Morra et al. 1994). Major and trace element modelling by Rutter (1985) showed that the majority of the element trends could be explained by fractional crystallisation of an anhydrous gabbro assemblage of olivine, clinopyroxene, plagioclase and titanomagnetite. Geochemical and isotopic studies by Morra et al. (1994, 1997) show that crustal contamination occurred along with fractional crystallisation in the differentiation from basalts to andesites but that changes from andesitic to rhyolitic compositions were solely a result of fractional crystallisation. The general consensus appears to be that crustal contamination played a minor and/or local role in modifying magma compositions and that fractional crystallisation was the major control (Coulon et al. 1973; Dostal et al. 1982; Coulon et al. 1978; Savelli et al. 1979; Rutter 1985; Beccaluva et al. 1987; Morra et al. 1994, 1997).

9.2 Sample selection and temporal constraints

9.2.1 Sampling policy

In order to elucidate the source and the melting characteristics of the volcanic arc rocks it is necessary to filter out the effects of fractional crystallisation and crustal contamination. The simplest first step in this process is to collect a suite of the most basic volcanic rocks within a particular terrain. Thus, the sampling policy for the Sardinian volcanic arc was to try and collect a series of basalts from each volcanic centre along the rift basin which were well constrained in absolute or relative age. Volcanic sequences containing andesites to basalts crop out in six main localities along the Sardinian Rift (Fig. 9.2). It was important to collect a time-constrained series from each of these areas since previous work has shown significant spatial zonations in geochemical character (Coulon et al. 1973; Dupuy et al. 1974; Coulon et al. 1975; Dostal et al. 1976; Beccaluva et al. 1987, 1994).

However, collection of basaltic rocks representing the temporal extent of magmatism from each of these areas proved problematic for two reasons. Firstly, the Sardinian volcanic-arc rocks are dominated by voluminous andesites and ignimbrites. The minor basalts and basaltic andesites proved difficult to find. Secondly, levels of dissection and exposure of the volcanic terrains meant that, apart from in the Arcentu Group (Funtanazza sub-basin Fig. 9.2), it was not possible to collect a stratigraphically well-constrained basaltic series. The analysed samples tend to come from exposures kilometres apart and the stratigraphic relationships between the exposures are often not completely clear.
9.2.2 Sample analysis and selection

85 samples were analysed for major and minor trace elements using standard XRF techniques at the University of Edinburgh (Appendix 9A). Petrographic examination and inspection of the major element compositions allowed altered samples and a large number of andesites and more evolved rocks with SiO₂ > 55% or MgO < 3.5% to be discarded. Alteration in the Sardinian volcanic rocks occurred by water infiltration (Assorgia et al. 1994ab) and is particularly prominent in areas adjacent to large normal faults where the rocks were buried and subsequently exhumed (e.g. central and north areas, Fig. 9.2). Since the data analysis required evaluation of mobile element concentrations (Rb, Sr etc.) any samples which were, for example, calcite cemented, or had obviously had large quantities of fluid passing through them were not used. This left only 30 samples which were fresh and of basaltic to basaltic andesite composition. The spatial and temporal changes already documented dictated that it was not appropriate to process all the analysed samples together. However, this meant that there were not enough samples for a rigorous analysis from each of the six basaltic-andesitic volcanic centres.

9.2.3 Temporal constraints

A considerable number of K-Ar dates and stratigraphic series constrain the Oligo-Miocene volcanic succession (Table 3.1, section 3.1.3) though many of the K-Ar dates are considered unreliable (sections 3.1.3, 3.2.3.4). Within extended error limits, the published temporal and stratigraphic information was used to help constrain many of the collected samples which were not taken from a discrete stratigraphic section. The published K-Ar data does show that the oldest basalts and basaltic andesites (~30 Ma, coeval with initial rifting) crop out in southern Sardinia whilst younger products are located in northern Sardinia (~24-18 Ma, coeval with passive subsidence in Sardinia and back-arc basin formation). This indicates another obstacle to the analysis of basalt series over the entire period of Sardinian rifting and back-arc basin formation - the fact that no single suite spans the whole Oligo-Miocene Rift evolution. A programme of Ar-Ar single crystal dating was proposed to provide high resolution temporal constraints on any systematic geochemical trends. However, since the only area with a reasonable number of samples was already fairly well documented (Arcentu Group, Assorgia et al. 1984; Assorgia et al. 1986) the prepared samples were not analysed.

9.2.4 Data used in the geochemical analysis

The Arcentu Group is represented by ten basaltic or basaltic andesite samples, nine of which were taken in stratigraphic succession over a time span believed to be ~30-18 Ma (after Assorgia et al. 1984, 1986ab). It is worth noting here, that on the plots below, one Arcentu sample often plots off the main trend. However, this is not the same sample on each plot, and so the scatter cannot be attributed to a single rogue sample. Selected geochemistry on 10 samples from the Logudoro and Anglona (northern) study areas (Logudoro Group, Tergu Formation, chapters 7, 8) is also presented. Five analyses of high magnesia basalts from Morra et al. (1997) have been added to selected plots since
these are thought to represent the most primary magmas in this area and were possibly erupted at the same time as the late dykes in Arcentu (~18 Ma, Morra et al. 1997, e.g. Fig. 9.6). The Nb concentrations for the Morra et al. (1997) samples, which are all less than 5 ppm, are not used here since the errors are unknown and may be significant for such small absolute concentrations. The temporal constraints on samples from north central Sardinia is much poorer than in the Arcentu area since they come from widely differing locations. Existing radiometric dates and relative stratigraphic locations have been used to suggest possible ages for the rocks between a range from ~25-18 Ma.

9.3 Major element trends

A graph of SiO₂ vs. K₂O (Fig. 9.3) shows that the majority of the analysed samples are calc-alkaline according to the criteria of Le Maitre et al. (1989). A number of samples from the Logudoro area and one from northern Sardinia are high-K calc-alkaline. The samples show a general fractionation trend of increasing K₂O with increasing silica. Graphs of SiO₂ vs. MgO and Fe₂O₃ (Fig. 9.3) show a general decrease in MgO and Fe₂O₃ content with increasing SiO₂. The Fe₂O₃ variation is of characteristic calc-alkaline rock series and results from the early crystallisation of Fe-Ti oxides (e.g. titanomagnetite) in contrast to the Fe and Ti enrichment characteristic of tholeiitic suites (Wilson 1993). The variations in these and other major elements (e.g. CaO, Na₂O) are consistent with the fractional crystallisation of a gabbroic assemblage (as in Rutter 1985).

The SiO₂ vs. MgO plot also discriminates between two magma series in the Arcentu area, a high MgO basalt series, with 8-10 wt% (Group 1), and a moderate MgO series with 3-6 wt% (Group 2). Other major and trace element variations (see below) further characterise these two groups; the first having distinctly lower TiO₂, Al₂O₃, and higher MnO than basalts of Group 2 which can be described as high-Al. The high MgO set (Group 1) came from the stratigraphically lowest basaltic flows exposed, which are thought to have been erupted at the onset of Sardinian Rifting (30-24Ma, Assorgia et al. 1984, 1986ab) as well as from the late dykes which cut across the Arcentu Group (~18Ma, Assorgia et al. 1984, 1986ab) and which may have been emplaced at the same time as a mid Burdigalian phase of extension (chapter 4). The lower MgO set (Group 2) were sampled from the middle and voluminously major part of the Arcentu succession (~24-18Ma, Assorgia et al. 1984). Morra et al. (1997) suggested that high Mg basalts from the Logudoro area were near to primary melts of hydrated mantle, whereas the high Al basalts were derived from them by fractional crystallisation of olivine and clinopyroxene. Derivation of the Group 2 Arcentu suite from Group 1 by olivine-cpx extraction appears difficult since the SiO₂ contents are similar. It is therefore important to deduce whether the two magma series in the Arcentu area were sourced from a different parental melt which varied through time or whether they reflect different fractional crystallisation and/or crustal contamination histories of the same parental melt erupted under different tectonic conditions.
The Al$_2$O$_3$ vs SiO$_2$ plot (Fig. 9.3) also distinguishes between the high- and low-Al Arcentu suites. In addition, the onset of calcic plagioclase extraction probably explains the negative correlation in the Logudoro and northern samples after ~49.5% SiO$_2$.

### 9.4 Systematic geochemical analysis

#### 9.4.1 Testing for fractional crystallisation and sample alteration

Major element variations (above) and published work (e.g. Rutter 1985, Morra et al. 1997) suggest that fractional crystallisation is an important process controlling the composition of Sardinian volcanic arc rocks. Before attempting to answer questions relating to the source of the magmas, one must first filter out the effects of fractional crystallisation and assess the effect of alteration on trace element concentrations and trace element ratios.

The positive correlation on the Nb vs. SiO$_2$ plot (Fig. 9.4) is the clearest demonstration of the behaviour of Nb as an incompatible element. That is, with the extraction of phases such as ol-cpx-opx-titanomagnetite and resultant increase in SiO$_2$, the concentration of Nb in the melt increases. This fractional crystallisation trend is mirrored by the positive Zr vs SiO$_2$ correlation (Fig. 9.4). The Nb and Zr vs SiO$_2$ plots appear to suggest that the Logudoro samples may have come from a parental magma relatively richer in Nb at a given SiO$_2$ content. Alternatively they may have undergone different fractional crystallisation histories (see below).

A plot of log Nb vs. log Zr shows that, within error limits of Nb±0.5 ppm (Appendix 9A), the Arcentu, Logudoro and northern samples lie on a positive correlation with constant Nb/Zr ratio (Fig. 9.4). The samples behave in a consistent way, suggesting that they have not been subject to alteration or to analytical error. Particularly noticeable on this plot is a low Nb concentration Arcentu sample (4 ppm) which occurs as an outlier on all plots and ratio plots involving Nb.

A plot of MgO vs. Zr (Fig. 9.4) exhibits a weak, curved negative correlation if the high MgO Arcentu Group 1 samples are excluded. The shape of the correlation may reflect the initial extraction of olivine with decreasing MgO content and then the extraction of plagioclase, such that the incompatible Zr continues to become concentrated in the melt whilst MgO content stays the same.

Mobile elements were plotted to see if they were affected by alteration. The assumption is that random scatter (particularly extreme depletion) is more likely to be the result of low T weathering or alteration than a source characteristic. Plots such as Sr vs. Rb, Sr vs. Ba, Rb vs. K$_2$O (Fig. 9.5) gave reasonable correlations suggesting that alteration of the samples was relatively minor. The Sr concentrations from Logudoro may have been variably enriched by alteration or some other process. In addition, plots of these elements against SiO$_2$ (e.g. Rb vs. SiO$_2$, Fig. 9.5) gave a sensible correlation suggesting a control
on element concentration by fractional crystallisation and/or crustal contamination rather than alteration (Fig. 9.5). The correlations and spread of data on these plots can be explained by extraction of calcic plagioclase, such that Sr and Ba are removed into the plagioclase whilst Rb and K become enriched in the melt. Since Rb increases and Sr decreases with plagioclase extraction, Rb/Sr should be a good fractional crystallisation index. This is confirmed by the broad positive correlation of Rb/Sr with SiO₂ (Fig. 9.5) and is important because the Rb/Sr ratio cannot therefore be used to detect mobile element variations in the source, in the absence of sufficient samples of constant Sr content.

A plot of two LREE elements such as La vs. Ce (Fig. 9.5) shows a broad positive correlation, suggesting that whilst scatter may be caused by analytical uncertainty, the elements behave in a coherent way rather than being affected by later losses.

Although the Nb concentration varies systematically as a result of fractional crystallisation (above), the Nb/Zr ratio is independent of fractionation, since a plot of Nb/Zr vs SiO₂ or Nb gives constant Nb/Zr values with respect to the fractionation index within the analytical error limits (Fig. 9.6; also Fig. 9.4, Log Nb vs Log Zr).

In contrast, other incompatible element pairs do show systematic variations with a fractionation index. In most cases, the Zr/Y ratio might be expected to be constant because both elements are moderately incompatible in silicate phases. However, a plot of Zr/Y vs Y shows that the ratio increases with decreasing Y concentrations (Fig. 9.6). The positive correlation between TiO₂ and Y (Fig. 9.6) strongly suggests that Y is behaving as a compatible element and becomes incorporated into a fractionating titanomagnetite phase. This explains the positive Zr/Y vs. SiO₂ correlation (Fig. 9.6). A plot of La against SiO₂ shows a weak positive correlation for Arcentu Group 2 and scatter for the remaining samples which have an inherent variability in LREE contents. The Arcentu Group 2 trend is also reflected in variations in the La/Y ratio, suggesting a control by the La concentration.

Thus a great majority of the trace element concentrations and trace element ratios within the Sardinian samples were affected by fractional crystallisation of an ol-opx-cpx-titanomagnetite assemblage (also in Rutter 1985, Morra et al. 1997). Ratios such as La/Y and Rb/Sr are affected by fractionation and cannot therefore be used to detect source processes.
9.4.2 General characteristics of spider diagrams normalised to MORB

Samples from along the length of the Sardinian Rift produce spider diagrams normalised to MORB (Saunders and Tarney 1984, Sun 1980 given in Rollinson 1993) which show a typical pattern for volcanic arc rocks (Fig. 9.7). The pattern is characterised by enriched LILE and LREE e.g. K, Rb, La and much less enriched or even depleted HFSE e.g. Nb, Zr (Fig. 9.7). The pattern is comparable to samples taken from the medium-K and high-K series of continental arcs where involvement of the sub-continental lithosphere can be inferred to have resulted in the enrichment of all the elements relative to oceanic-arc samples (e.g. Pearce 1983, Wilson 1993, Tatsumi and Eggins 1995). Alternatively, the pattern could represent lower degrees of melting of a MORB source (given mild enrichment) as a result of a thicker lithosphere than in the oceans. The typical spider diagram pattern together with the data discussed below and published data (e.g. Rutter 1985, Morra et al. 1997), suggests that the Oligo-Miocene volcanics were derived from melting of a mantle source modified by a subduction component. Spider diagrams plotted for all the samples show no obvious variations, though the absolute concentrations of elements may have been controlled by fractional crystallisation and/or crustal contamination (above). The fact that the spider diagrams do not show any noticeable spatial or temporal changes suggests, by itself, that the source of the magmas could have been constant.

9.4.3 Testing for variations in the source of the volcanic-arc rocks

9.4.3.1 Theory

In the early stages of crystal fractionation, elements incompatible in mantle phases are likely to remain strongly incompatible in the olivines and pyroxenes crystallising at lower pressure. For as long as the extract phases do not change the modal proportion significantly, many incompatible ratios are not likely to change. They may be a good indicator of source characteristics (Latin et al. 1990; e.g. LREE/HREE, Ce/Y or Nb/Zr). In the case of rift-related basalts, a decrease in the LREE/HREE ratio might then indicate either an increase in the extent of melting of a constant source or an increasing level of depletion in a changing source melted to the same degree. In the subduction-modified MORB system however, an increase in the LILE/HFSE or LREE/HREE ratio may also be indicative of an enrichment in the subduction-component of the source area. It is thus important to examine separately the immobile trace element ratios unaffected by fractionation and those mobile incompatible elements reflecting potential slab influence.

9.4.3.2 Hypothesis

Initial examination of trace element ratios indicative of source processes indicates that the Group I high-Mg basalts from Arcentu have higher La/Nb (~3) values than the Group 2 high-Al basalts (La/Nb ~1.6). However, the two groups have similar Nb/Zr ratios (0.04-0.07). Since Nb and Zr do not appear to be involved in slab-fluid enrichment processes and the Nb/Zr ratio is unaffected by low to moderate levels of crustal fractionation, as a simplest working hypothesis, it seems that the Group 1 samples
might have been derived from the same mantle source with similar extent of melting, but that the Group 1 source region had been preferentially enriched in a LREE bearing slab component.

Similarly, if one compares La/Nb (Fig. 9.8), Rb/Zr (Fig. 9.8) ratios and K, La, Nb, Zr (Figs. 9.3, 9.4, 9.6) concentrations at a given level of SiO₂, then the samples from southern Sardinia generally have lower concentrations or ratios than those in the north. Yet the northern and southern areas have Nb/Zr ratios that are the same within error, again suggesting that the samples may have been derived from the same source mantle with similar extent of melting, but that the northern source region had been preferentially slightly enriched in LREE and mobile elements. An alternative explanation, such as differing fractionation histories is needed to accommodate changes in the Nb and Zr concentrations.

9.4.3.3 Analysis and implications
The constancy of the Nb/Zr ratio indicates that variations in the extent of melting or degree of enrichment in the mantle wedge prior to slab-fluid enrichment cannot be detected. No conclusions can be drawn as to the variation with time, space or any geochemical parameter of the Nb/Zr ratio. In answer to the original question, changes in the degree of melting which might relate to increased upwelling are not detectable over the observed intra-arc and back-arc basin rifting periods. The sensitivity of the approach has not however, been modelled quantitatively. However, there may be other variations worth testing, such as the possible variations in subduction-component enrichment.

Subduction-component enrichment can be detected by fractionation-independent variations in mobile elements thought to be supplied in the hydrous flux (e.g. Rb, Ba, Sr, La). The La/Y, Rb/Ba, Rb/Sr ratios could not be used because it was shown above that they are affected by fractional crystallisation. The pairs Rb/Zr and La/Nb were chosen since they both appear to be fractionation independent (Fig. 9.8). Each pair is made of a mobile/immobile trace elements of comparable incompatibility to minimise any fractionation effects. A positive correlation would indicate a possible variation in the subduction-component, with high ratios indicating a high slab-flux. The plot of Rb/Zr vs La/Nb (Fig. 9.8) gives a wide positive scatter outwith the proposed error limits, suggesting that there may be some variation in a slab-flux component independent of the constant mantle wedge source.

Broadly speaking, the oldest and youngest Group 1 Arcentu samples have systematically higher ratios than the Group 2 samples from the middle, voluminous part of the succession, independent of the proposed analytical error (Fig. 9.8). This suggests that the Group 1 samples were preferentially enriched in a subduction-component in the source. Similarly, plots of Rb/Zr and La/Nb against time appear to show a progressive decrease and then increase in the values for the Arcentu samples (Fig. 9.8). No systematic temporal relationships can be identified for the Logudoro and northern areas. When linked to Sardinian rifting and regional events, this suggests that samples erupted over the period of Sardinian thermal subsidence and back-arc basin opening were less enriched in a subduction component. Samples which may have been erupted at the same time as the Sardinian Rifting phases
(chapter 10) were relatively enriched in a subduction-component. No simple geodynamic explanation for this is yet apparent.

Spatial variability between the time-equivalent Arcentu Group 2 rocks and Logudoro plus northern samples suggests that the northern samples may have been slightly more enriched in the subduction component since they have generally higher Rb/Zr, La/Nb ratios (Fig. 9.8). The fractionation history of the Logudoro plus northern samples with relatively higher Zr, Nb, and lower Y, Ti concentrations than the Arcentu Group (Fig. 9.4, 9.6) was also probably different. It seems that the samples from the north were more affected by the fractionation of titanomagnetite than those in the south (Fig. 9.6, Y, Ti contents). This is consistent with the inference that an increased water component (i.e. slab flux) promotes the crystallisation of oxides such as titanomagnetite (Wilson 1993, Tatsumi and Eggins 1995) and that a slightly greater slab-flux may be present in the north.

9.5 Summary

- Due to the fractionated nature of the Sardinian volcanic-arc sequence and the level of exposure, problems were encountered collecting a basaltic series spanning the Oligo-Miocene from differing locations along the rift.
- Systematic variations in the major and trace element ratios, plus some incompatible trace element ratios are consistent with fractional crystallisation of an olivine-clinopyroxene-orthopyroxene-titanomagnetite assemblage as postulated by Rutter (1985).
- The Nb/Zr ratio was constant, within error, throughout the period of Sardinian rifting events, back-arc basin opening and along the length of the rift. Thus, there was no detectable change in either the primary mantle wedge source enrichment or degree of partial melting. It does not appear that active lithospheric thinning and associated, increased asthenospheric upwelling accompanied the extensional events observed in the Sardinian Rift to a detectable extent.
- Subtle and tentative systematic changes in fractionation-independent mobile/immobile trace element ratios suggest that the subduction-component may have varied independently of a constant source. Variations from the most complete temporal succession in southern Sardinian suggest that the subduction-flux may have decreased over the period of back-arc basin opening relative to the first and last extension-related volcanism. Spatial variations within rocks of the same age indicate that magmas erupted in northern Sardinia may have sampled a source richer in a subduction component than those in the south.
- Different fractional crystallisation histories may have resulted in the observed variations in trace element concentrations for northern and southern samples at a given silica level. The Y and Ti variations suggest that in the northern magmas, titanomagnetite may have been relatively favoured as a crystallising phase, possibly due to an increased water content or slab-flux component in these samples.
Chapter 10
The Oligo-Miocene Sardinian Rift evolution records information regarding the nature and timing of back-arc basin formation in the Western Mediterranean (Cherchi and Montadert 1982; Rehault et al. 1984). The rift has been recognised as a complex extensional structure since the works of Cocozza and Jacobacci (1973), but until recently, published accounts failed to provide a specific model for the formation of this basin. Cherchi and Montadert (1982ab) and Rehault et al. (1984) regarded the Sardinian Rift simply as a basin formed by Late Oligocene-mid Aquitanian (30-23 Ma) extension, but they did not quantify this assertion over the length of the structure. Subsequent work by Thomas and Genesseaux (1986) and Martini et al. (1992) on areas of the northern Sardinian Rift recognised that the rift was formed by more than one phase of extension which occurred between mid/late Oligocene to mid Miocene times, but their models still lacked a detailed kinematic account of basin formation. However, the model of Carmignani et al. (1994, 1995), which considered the approximately N-S trending Sardinian Rift to be a Burdigalian age extensional structure which was superimposed upon Late Oligocene-Aquitanian transpressional sinistral strike-slip fault systems (section 2.1.5), was the first to specify a detailed kinematic evolution for the entire Sardinian Rift. The Carmignani et al. (1994, 1995) model has been quoted subsequently, but with modifications, such that the late Oligocene-early Miocene was a time of a ‘wide, sinistral transtensive regime’ according to Assorgia et al. (1995a), Oggiano et al. (1995) and Morra et al. (1997).

From observations made in the present study, it is apparent that whilst strike-slip faulting was active in the Late Oligocene-Aquitanian (chapter 6), the Carmignani et al. (1994, 1995) model misinterprets and fails to recognise a great deal of published data and field evidence. In particular, the formation of late Oligocene-Aquitanian extensional sub-basins to the west and south of the strike-slip structures are not considered. This chapter uses field observations (chapters 4-8) together with temporal constraints (chapter 3) and reliable published data in order to produce a detailed tectono-stratigraphic model for the evolution of the Oligo-Miocene Sardinian Rift. The resultant evolutionary model has significant implications for Western Mediterranean tectonic development.
10.1 Tectono-stratigraphic model of Sardinian Rift evolution

Seven time slices from the mid Oligocene to the late middle Miocene illustrate the main events in the tectono-stratigraphic evolution of the Sardinian Rift (Figs. 10.1-10.7). Figures 10.8 and 10.9 highlight stratigraphic correlations along the rift. All orientations refer to the present day trend of features rather than the trend at their time of formation. The significance of the re-oriented trends will be discussed in section 10.3.

10.1.1. Late Rupelian-mid/late Chattian (30-25 Ma, mid-late Oligocene; Fig. 10.1)

Although little temporal data exists (Appendix 3D), the late Rupelian-mid/late Chattian appears to have been the time when Oligo-Miocene extension of the Sardinian continental crust commenced. The majority of this 'first-phase' extension is thought to have occurred over a short time period (<few Ma, section 11.2.4), creating the tectonic relief which was subsequently infilled (chapters 5, 6, 8). This major phase of extension may have occurred within the late Rupelian-mid/late Chattian period based on the radiometric dates of volcanics rocks within the sub-basins, and the age and arrangement of overlying basin filling units.

In eastern Sardinia, sinistral strike-slip faults whose movement is constrained at some time(s) between the post-Palaeocene (Carmignani et al. 1992b) and late Aquitanian (chapter 6), may also have been active within the late Rupelian-mid/late Chattian. The strike-slip faults in the eastern Sardinian basement are cut by mid-late Oligocene NNW-SSE faults in the Anglona study area and, although no cross-cutting relationships have not been observed in central northern Sardinia, the resultant geometries are consistent with the majority of strike-slip faulting occurring prior to ~N-S rift formation (Chapters 6 and 7).

Basaltic-andesitic volcanic rocks, commonly considered to have a continental tholeiitic character (Dostal 1976; Coulon 1975), formed small lava domes, dykes and subvolcanic bodies from ~32 Ma (Cherchi and Montadert 1982ab; Beccaluva et al. 1985). Many of these Oligocene rocks are found in southern Sardinia, some within the re-activated late Eocene, east-west trending grabens of Sulcis and Cixerri (Cherchi and Montadert 1982ab; Assorgia et al. 1992bce; Barca and Costamagna 1997) and the east-west Funtanazza sub-basin (Assorgia et al. 1984). The volcanic rocks are located only within the sub-basins, suggesting that in southern Sardinia, E-W and NW-SE trending normal faults which facilitated magma emplacement had formed and/or were active. For the same reasons, ~N-S trending normal faults may have been active in northern Sardinia. It is possible that the unusually oriented E-W grabens of southern Sardinia may have formed in response to the reactivation of 'late Hercynian' E-W to NE-SW trending structures which crop out in the eastern Sardinian basement and may extend westwards.
At the eastern margin of the Sardinian Rift (Sarcidano sub-basin), the age of basin initiation is poorly constrained because of poor exposure and paucity of temporal constraints on the basal continental clastic sedimentary rocks. However, coarse clastic rocks (Villanovatulo, Casteldoria Members) whose composition and dispersal was controlled by tectonic relief created by NW-SE to N-S trending normal faults are thought to be Late Rupelian-Aquitanian in age (post 29.9 Ma, Cherchi and Montadert 1982ab), suggesting that extension initiated in the late Rupelian-mid/late Chattian. In the Sarcidano area, and possibly in the Campidano region of southern central Sardinia, these earliest NW-SE faults developed on a trend inherited from the Hercynian basement, whilst slightly younger NE-SW cross-faults followed more closely the main Sardinian Rift trend.

10.1.2 Late Chattian (25-23.8 Ma, latest Oligocene; Fig. 10.2)

By the latest Chattian, the N-S trending Sardinian proto-rift and transtensional basins at the eastern proto-rift margin had formed. The exact timing of fault movement is not constrained but syn-rift deposits are localised and voluminously minor (chapters 5, 8), suggesting that the rates of tectonic subsidence far exceeded sedimentation rates.

The proto-rift consisted of a complex system of sub-basins created by extensional and transtensional faulting on a variety of fault trends. In southern Sardinia, the E-W trending Sulcis, Cixerri and Funtanazza sub-basins existed whilst in south-central Sardinia, NW-SE and N-S oriented normal faults defined the Campidano depocentre and the eastern rift margin. In northern Sardinia, N-S to NNW-SSE normal faults defined the proto-rift. NE-SW trending transtensional sub-basins may have already existed or may have formed in the late Chattian. The area of central Sardinia west of the Ottana sub-basin (Fig. 1.3) may have been situated in a 'wide transtensive regime', as suggested by Assorgia et al. (1995a) since seismic profiles from just offshore show flower structures which developed in response to sinistral strike-slip displacement (Thomas et al. 1988). Though the exact relationship between the NE-SW sinistral strike-slip faults and N-S normal faults is not apparent, it is clear that the tectonic relief caused by both mechanisms was present, since Late Oligocene-early Burdigalian sedimentary and volcanic rocks infill both sub-basin styles. This is in direct contrast to the Carmignani et al. (1994, 1995) model which considers that transpressional strike-slip faulting was active in the late Chattian and that the extensional Sardinian rift basin did not form until the Burdigalian.

Radiometric dates indicate that andesitic volcanism occurred along the length of the N-S trending Sardinian rift in the Late Chattian (e.g. Beccaluva et al. 1985). Sedimentation was dominantly fluviolacustrine with coarse clastic material dispersed from local, fault-bounded topographic highs by alluvial fan systems (Villanovatulo, Casteldoria Members). Finely laminated lacustrine limestones accumulated in fault-created accommodation space away from clastic supply (e.g. Valledoria Member,
Campu Sali, Oschiri Formations). In southern Sardinia, the first marine marlstones were deposited in the basin centre (Campidano area, Cherchi and Montadert 1982ab) in response to regional or eustatic sea level rise (see 10.1.9 below).

10.1.3 Early-mid Aquitanian (23.8-21 Ma, early Miocene; Fig. 10.3)
A phase of activity on sinistral strike-slip faults within the transtensional Oschiri sub-basin which ended in latest Aquitanian times (20.6±0.2 Ma, section 2.3.3, chapter 6) may have commenced in the early-mid Aquitanian. As discussed in chapter 6, all the NE-SW structures may have been active at this time, but the amount of movement is thought to have been relatively small. No active extension was detected within the Sardinian proto-rift, rather degradation and filling of the older tectonic relief dominated. This phase of passive infilling may relate to a phase of thermal subsidence (sensu McKenzie 1978).

The first rhyolitic-dacitic ignimbrites plus andesitic volcanic rocks were erupted along the length of the proto-Sardinian rift in the early-mid Aquitanian (e.g. in Beccaluva et al 1985). Fluvio-lacustrine sedimentation intercalated with volcanic products continued in the Oschiri transtensional basin (Oschiri Formation) whilst latest Oligocene-early Aquitanian marine transgression proceeded over low-lying areas of the Sardinian proto-rift. Subaerial volcanic centres in central and northern Sardinia must have formed a topographic high which transgression did not overstep. The character of sedimentation associated with the 'first' marine transgression varied, and was dependent on the supply of clastic or volcanic material. For example, in Sarcidano, clastic material supplied from the erosion of tectonic relief was localised into fault-created depocentres by fan-deltas and reworked by tidal currents (Duidduru Member). At the western margin of the Funtanazza sub-basin, shell hash beds were deposited (top Campu Sali Formation) whilst in the Anglona study area a volcano-sedimentary succession accumulated in the half-graben adjacent to andesitic volcanic centres (Vaginella Member).

10.1.4 Late Aquitanian -early Burdigalian (21-18 Ma, early Miocene; Fig. 10.4)
The late Aquitanian-early Burdigalian was a time dominated by sub-basin filling after sinistral strike-slip fault movement had ended (20.6±0.2 Ma, Oschiri sub-basin). This phase of passive infilling may relate to a continued phase of thermal subsidence (sensu McKenzie 1978). Four features suggest that an extensional stress field (i.e. \( \sigma_3 \)) oriented \( \sim \)N-S may have acted on the Sardinia-Corsica microplate at this time. In northernmost Sardinia, E-W normal faults with associated late Aquitanian-early Burdigalian syn-rift deposits are well exposed (Castelsardo sub-basin). Extension on NE-SW faults may have occurred in the Funtanazza sub-basin (chapter 4), Bonifacio sub-basin (chapter 8), and in Logudoro where the last andesitic cones were aligned on a NE-SW trend (chapter 7).
Voluminous pyroclastic products including pumice, block and ash ignimbrites and welded lava-like ignimbrites filled the accommodation space, onlapped and covered the eastern rift margin of the central and northern Sardinian proto-rift between 21 and 18 Ma. Basalts and andesites show a distinct geochemical zonation from tholeiitic in the south, through calc-alkaline in Logudoro, to shoshonites (high-K) in northernmost Sardinia (Coulon and Dupuy 1975; Beccaluva et al. 1987). The andesitic volcanic centres are thought to have formed stratovolcanoes which also erupted explosively and supplied pyroclastic material over a wide area. Basin filling geometries indicate that this phase of intense volcanism between 21-18 Ma does not appear to be related to active extension. The most reliable radiometric data indicates that volcanism in the Sardinian Rift largely ended at ~18 Ma (section 3.2.3, Appendix 3D; Odin et al. 1994).

In the transtensional basins of the eastern rift margin, fluvio-lacustrine sediments were intercalated with pyroclastic volcanic rocks (Oschiri Formation). Marine sedimentation continued in areas away from the volcanic centres and records gradual transgression until ~late Aquitanian times (e.g. onlap and filling by clastic Duidduru Member), early Burdigalian regression at ~19 Ma (e.g. Sa Tellura Formation; Castelsardo Member) followed by renewed transgression in the mid-Burdigalian (e.g. Serra Longa Member; Campulandru Member; upper Sartori Formation). Submarine volcanism occurred in the Campidano depocentre with formation of pillow lavas and hyaloclastites (Cherchi 1974; Assorgia et al. 1995a).

10.1.5 Mid Burdigalian (18-17 Ma, early Miocene; Fig. 10.5)

The mid Burdigalian was a time of active extension which defined the Oligo-Miocene Sardinian Rift structure as exposed today. Normal faulting occurred largely after volcanism had ended at ~18 Ma, and created numerous complex depocentres before the sizeable late Burdigalian transgression which covered the majority of the rift basin.

In northern Sardinia, extension occurred via two fault-families, those with a NW-SE to N-S trend, and those with a E-W to NE-SW trend which downthrow to the north. Before the mid- Burdigalian, the E-W to NE-SW fault family were active (section 10.1.4), whilst after the mid Burdigalian, syn-sedimentary N-S trending normal faults moved (section 10.1.6). This suggests that the large scale extensional stress field changed from ~N-S (late Aquitanian-mid Burdigalian) to E-W (mid Burdigalian-Langhian). Apart from near Castelsardo, where N-S faults cut E-W faults, the relationships between the fault sets is unclear and they apparently moved together in the mid Burdigalian. In the Sardinian Rift, extension on the two fault sets may have been facilitated by the presence of inherited fault trends (e.g. NE-SW, late Hercynian, N-S Oligocene rift-forming structures). In southern Sardinia, the emplacement pattern of mid/late Burdigalian age dykes within the Funtanazza...
sub-basin suggests that E-W and NW-SE extension may have occurred at this time (chapter 4). In the Sarcidano sub-basin, localised NE-SW to N-S faulting is recorded within the basin fill succession.

In the mid/late Burdigalian, clastic sediments derived from weathering of the Palaeozoic basement (Chilvani Formation) were transported by fluvial/alluvial fan systems to the southwest along the Oschiri and Ottana transtensional sub-basins, towards fault created depocentres on the main Sardinian Rift trend. Marine transgression onto the lowest depocentres occurred, resulting in the deposition of calcarenites and carbonates (Sartori Formation, Campulandru Member, Isili Formation). At the Sarcidano sub-basin margin, intense facies variability developed within a mixed carbonate-siliciclastic-marlstone system controlled by the fault-created topography and clastic sediment supply. Areas of central northern Sardinia which formed paleotopographic highs were covered by non-marine sediments, such as the lacustrine Perfugas Formation in a perched basin (Perfugas sub-basin), and conglomerates, breccias and a discontinuous organic layer in the Logudoro study area.

10.1.6 Late Burdigalian-early/mid Langhian (17-16 Ma, early-mid Miocene; Fig. 10.6)

From the Late Burdigalian to mid Serravalian, marine sedimentation was the dominant process active over the Sardinian Rift. Following a major late Burdigalian transgression, correlation of marine outcrops implies that a marine seaway existed along the length of the Sardinian Rift basin. Occasional movements on N-S trending normal faults can be deduced from outcrop scale syn-rift deposits in both northern and southern Sardinia. However, gradual post-rift thermal subsidence (sensu McKenzie 1978) is likely to have been the dominant mechanism by which hundreds of metres of shallow marine sediments were able to accumulate. Within the shallow marine succession, relative sea level changes can be detected.

Late Burdigalian and earliest Langhian sedimentation records marine transgression over the mid Burdigalian tectonic relief and degraded late Oligocene-lowermost Miocene topography. Platform carbonates were commonly deposited in shallow water areas at the sub-basin margins, on fault bounded highs which were restricted from clastic supply (Florinas Group, Isili Formation, Sedini, Sennori Members). Marlstones and muddy calcarenites accumulated in deeper water depths of a few hundred metres in the sub-basin centres (Florinas Group, Giara Group, Martis Member). Clastic material supplied along the NE-SW trending transtensional sub-basins (Chilvani Formation) passed westwards and basinwards through marginal marine to mixed carbonate-siliciclastic marine shelfal environments (Florinas Group). Calcarenites, calcirudites and carbonates accumulated in shallow marine, tidally-dominated conditions in the Bonifacio and St. Florent basins of Corsica. A minor regressive period in the early-mid Langhian led to the influx of siliciclastic material in the Logudoro area and shallowing of the depositional environment from offshore marlstones to shallow marine
calcarenites in Sarcidano. A calc-alkaline to peralkaline (high Na, K) suite of fractionation-derived ignimbrites erupted in the Sulcis sub-basin of southernmost Sardinia represents the last phase of the subduction-derived volcanism from 17.6± 0.8-13.8± 0.7 Ma (Morra et al. 1994).

10.1.7 Mid Langhian-Serravalian (16-12 Ma, mid Miocene; Fig. 10.7)
In the mid Miocene, marine sedimentation continued to fill the earlier fault-created accommodation space with evidence for limited amounts of localised N-S normal faulting.

After the minor early/mid Langhian regression (10.1.6), renewed transgression resulted in the widespread deposition of shallow marine platform carbonates whilst marlstones continued to fill deeper depocentres. In the areas studied here the record of marine sedimentation ends in the mid Serravalian (section 3.2.2) with carbonate and calcarenite deposition capping marlstone sequences and indicating another phase of relative sea level fall. The final part of the Oligo-Miocene cycle was Tortonian to Lower Messinian age carbonates, calcarenites and lagoonal sediments documented in the area around Cagliari, the Campidano plain (Fig. 1.1), and on the Sinis Peninsula of western central Sardinia (NN6-NN11 zone, Pecorini and Pomesano Cherchi 1969; Cherchi 1974, 1985).

10.1.8 Summary of tectonic events
The main tectonic events which formed the Sardinian intra-arc rift can be summarised as:

- Transpressional and transtensional movement on NE-SW trending sinistral strike-slip faults which crossed the eastern Sardinian basement. Fault movement occurred sometime between the post-Palaeocene and late Aquitanian, most probably in several phases. The majority of strike-slip movement probably accumulated before ~N-S Sardinian proto-rift formation. The two fault sets may have moved at the same time in the Oligocene. Strike-slip fault activity occurred after ~N-S Sardinian proto-rift formation in the late Aquitanian, though the amount of movement in this phase may have been relatively minor.

- Mid-late Oligocene formation of the Sardinian 'proto-rift' in western Sardinia. The 'proto-rift' consisted of E-W grabens in southernmost Sardinia, a NW-SE trending graben structure in south central Sardinia and N-S trending grabens and half-grabens in northern Sardinia. The faults bounding the sub-basins had throws of ~500-1000m. Passive filling of these fault-created depocentres occurred from the latest Oligocene to the mid Burdigalian.

- Late Aquitanian-early Burdigalian extension on E-W to NE-SW trending normal faults with throws of 10-100's metres.

- Mid-Burdigalian extension on E-W to NE-SW and NW-SE to N-S trending faults with throws of hundreds of metres.
occasional activity on Late Burdigalian-Langhian –N-S trending normal faults with throws of tens of metres.

The cause of the different tectonic phases is discussed after the Sardinia-Corsica microplate has been reconstructed to its pre- and syn-rotation positions, in a regional context (10.3).

10.1.8.1 Comparisons with published work

The tectonic framework deduced in this study provides an coherent explanation for the complex structure of the Sardinian Rift basin and the abundance of basin fill ages and facies types. It is clear that the Sardinian Rift did not simply form as a Late Oligocene-Aquitanian basin with one, relatively long-lived extension phase as suggested by Cherchi and Montadert (1982), Rehault et al. (1984) and quoted in many subsequent publications. However, a ‘proto-rift’ structure consisting of several orientations of sub-basin did form an overall north-south trending series of depressions in western Sardinia sometime in the mid-late Oligocene. The presence of mid-late Oligocene extensional basins that formed a –N-S proto-rift structure in western Sardinia is in direct contrast to the Carmignani et al. (1994, 1995) model which suggests that at this time, the Sardinian continental crust was only affected by transpression (Carmignani et al. 1995), or transpression and transtension (Oggiano et al. 1995), on sinistral strike slip faults. This research shows that movement on strike-slip fault systems in eastern Sardinia could have occurred either before and after, or coeval with and after the formation of the extensional ‘proto-rift’ in western Sardinia. Lecca et al. (1997) suggest that the Sardinian Rift evolution comprised a late Oligocene-Burdigalian transtensional and extensional phase followed by a second, late Burdigalian-Langhian ‘collapse event creating wide structural lows’. This evolution is more compatible with the observations made in this study.

Published work from northern Sardinia which suggests a phase of early Burdigalian extension between Corsica and Sardinia (Thomas and Gennesseaux 1986), and Burdigalian –N-S faulting which cut across older strike-slip faults (Martini et al. 1992; Carmignani et al. 1994, 1995; Oggiano et al. 1995) is in general agreement with the field observations of late Aquitanian -early Burdigalian E-W faulting and mid Burdigalian E-W and N-S normal faulting.

The second phase of mid-Burdigalian extension within the main Sardinian Rift basin occurs within the same time span as the suggested phase of regional NE-SW directed compression (Cherchi and Montadert 1982ab; Cherchi and Tremolieres 1984). No evidence for such a compressional phase was found in this study. Instead it is possible that some structures, such as folds or unconformities immediately adjacent to fault planes, which may have been interpreted as compressional features by Cherchi and Montadert (1982ab) and Cherchi and Tremolieres (1984), resulted from renewed activity on extensional faults which created fault propagation folding (sensu Withjack et al. 1990; Gawthorpe
et al. 1997; Sharp et al. in press) or fault drag. The rarity of compressional structures within the Oligo-Miocene basin fill and lack of consistent structural trends suggests that the few that are present may have formed in response to local space problems within an overall extensional regime.

10.1.9 Stratigraphic correlations, rift subsidence and comparisons to a ‘eustatic’ curve

This section describes the trends which can be deduced from a correlation of sub-basin stratigraphy’s which are illustrated on Figures 10.8 and 10.9. Comparisons between the interpreted relative sea level along the rift basin, the ‘eustatic’ curve of Haq et al. (1988, Fig. 10.10) and regionally recognised sea level changes, allows the control of global or regional sea level variations as opposed to local tectonism to be deduced. The Haq et al. (1988) sea level curve for the Oligo-Miocene was based on outcrops from Europe, the USA, New Zealand and Australia and although the short term curve clearly appears to be biased to the Mediterranean in the late Miocene (Messinian) it is cautiously taken here to represent ‘eustatic’ sea level change.

10.1.9.1 Initial extension and transtension

Sedimentary basin filling commenced with the deposition of coarse, continental clastic sediments sometime after the Eocene and probably in the Oligocene (?Rupelian), post 29.9±1.5 Ma (Villanovatulo Member, Sarcidano; basal Campu Sali Formation, Funtanazza; basal Oschiri Formation; Casteldoria Member, Anglona). Lacustrine limestones and cherts, sometimes rich in organic matter, are either basal equivalents, or else they overlie the alluvial conglomerates and breccias (e.g. mid Campu Sali Formation, Funtanazza). In northern Sardinia, these lacustrine deposits are intercalated with volcanic tuffs, lapilli tuffs and ignimbrites (e.g. Valledoria Member, Oschiri Formation) and are as young as latest Aquitanian in the Oschiri sub-basin, which was restricted from marine transgression (20.60±0.24 Ma, $^{40}$Ar/$^{39}$Ar date, section 3.2.3). Similar continental conglomeratic, lacustrine and lagoonal marlstones of this age are found within extensional sub-basins of southern France and the Gulf of Lion (Gorini et al. 1993).

Volcanic basin filling commenced with the eruption of basalt and basaltic andesite flows in discrete volcanic centres sometime between 32-29 Ma in southern Sardinia (Beccaluva et al. 1985; Balogh et al. 1997) and at ±31.2± 1.1 Ma or ~28 Ma in northern Sardinia (Beccaluva et al. 1985; Balogh et al. 1997). The majority of volcanism had ended by 18 Ma (mid Burdigalian). Volcanism resulted in the construction of topographically high volcanic areas (e.g. Logudoro, Funtanazza) dominantly composed of andesite flows and breccias which supplied large volume rhyolitic-dacitic pyroclastic flows and ashes over a greater areal extent. Mid-late Oligocene subsidence was caused by active extensional and transtensional tectonism which occurred in continental conditions.
10.1.9.2. First marine transgression

In the topographically lowest areas away from the volcanic centres, the first marine transgression occurred in the latest Oligocene (marls, Campidano region, Cherchi and Montadert 1982ab), and Aquitanian (shell bed, top Campu Sali Formation, Funtanazza; clastic sediments, Duidduru Member, Sarcidano, Serrano et al. 1997; within volcano-sedimentary Vaginella Member, Anglona, Spano and Asunis 1984; Francolini and Mazzei 1991). Similarly, in the Gulf of Lion, Aquitanian to mid-Burdigalian shallow marine clastic sediments, calcarenites and carbonates were deposited (Gorini et al. 1993).

Over the whole of Sardinia, there is evidence for a minor regression in the late Aquitanian-early Burdigalian (regressive clastic sediments, Castelsardo Member; shallowing of Sartori Formation and inland Sa Tellura Formation, Funtanazza; ‘Sedilo’ alluvial unit/Sa Manenzia pyroclastic flow, Ottana sub-basin, Porcu 1983) and transgression after ~19 Ma (early-mid Burdigalian; Campulandru Member carbonates; Serra Longa transgressive lag, Is Paras Member carbonates, Sarcidano; calcarenites, upper Sartori Formation, Funtanazza).

The first, latest Oligocene-early/mid Burdigalian transgressive-regressive-transgressive cycle which was deduced from lithofacies within the Sardinian Rift, shows general similarities to the 'eustatic' curve of Haq et al. (1987; Fig 10.10) suggesting that worldwide fluctuating sea level may have been responsible for the trends rather than extension in the western Mediterranean and back-arc basin opening. Relative subsidence of the rift basin was therefore a result of an independently fluctuating sea level possibly superimposed upon a phase of thermal subsidence.

10.1.9.3 Second marine transgression

Following a mid Burdigalian extensional phase, late Burdigalian (~17 Ma, N7-N8 zones), marine transgression reached most parts of the Sardinian Rift. At high stratigraphic resolution, the timing of transgression was controlled by the local tectonic relief (e.g. Perfugas Formation, Campulandru Member to Sedini Member, Anglona; Florinas Group). Those areas previously dominated by shallow marine sedimentation were further inundated (e.g. Giara Group, Sarcidano). In the Gulf of Lion, sedimentation was also dominated by marlstones and turbidites from the late Burdigalian-Tortonian (Gorini et al. 1993). A late Burdigalian-lower Langhian transgression is recognised in other areas of the Mediterranean (Cyprus, Robertson et al. 1991; Turkey, Flecker 1995; Robertson and Grasso 1995) and shown on the Haq et al. (1988) eustatic curve (Fig. 10.10). Thus basin subsidence was a result of the mid Burdigalian extensional tectonism, independent sea level change and possibly continued thermal subsidence.
10.1.9.4 Fluctuating sea level

Carbonate-marlstone sedimentation continued until the mid Serravalian (Florinas Group, Laerru Formation) with a period of relative sea level fall identified by the influx of clastic sediments within the Florinas Group and calcarenite deposition at the top of the Giara Group in the early/mid Langhian (~16 Ma). The short-term Haq et al. (1988) curve has two abrupt regressions in the earliest and mid Langhian, one of which could represent the same event recorded by the observed early/mid Langhian regression in Sardinia (Fig. 10.10). In this stage it appears that an "independently fluctuating sea level was superimposed on continued, gradual thermal subsidence of the rift basin.

In study areas described during the course of this research, the end of marine sedimentation in the Sardinian Rift occurred in the Serravalian (Florinas Group), after a third, mid Langhian transgression. Marine sedimentation may have continued until the Tortonian or early Messinian (Leone et al. 1992, Cherchi 1974), but with deposits subsequently eroded away. Alternatively, marine sedimentation may have ended contemporaneous with ‘eustatic’ regression from the mid Serravalian (Haq et al. 1988; Fig. 10.10), or may have ended because regional uplift commenced in response to ‘rift flank-uplift’ on the margins of the proto-Tyrrhenian Sea.

10.1.9.5 Alternative rift correlation

Independent work by Assorgia et al. (1997c) also recognises three transgressive-regressive cycles across the Sardinian Rift. They are 1) Aquitanian-mid/upper Burdigalian followed by a widespread hiatus at ~18 Ma, 2) upper Burdigalian-mid/upper Serravalian, 3) ?upper Serravalian- Tortonian/early Messinian. The first two cycles are of slightly different ages than the three identified in this study because of the different temporal constraints and timescales used. The Assorgia et al. (1997c) ‘third cycle’ is based on hypothesised ages of shallow marine carbonates from northern Sardinia (e.g. M. Santo, ?Tortonian, Pomesano Cherchi 1971a, see Appendix 3D), from mid-late Miocene sediments found around Cagliari and on the Sinis Peninsula of western Sardinia (N17, NN11 zone Cherchi 1974, 1985).
10.2 Back-arc basin formation and continental deformation

Comparisons of the Western Mediterranean back-arc and intra-arc basin evolution with other tectonically similar areas is useful in determining the feasibility of the Western Mediterranean tectonic evolution proposed in section 10.3. General implications from the Sardinian Rift study for the development intra- and back-arc basins can also be highlighted.

Destructive plate margins which involve subduction of oceanic lithosphere beneath continental lithosphere are commonly complex zones of deformation. Such zones cannot be viewed in a classical '2D perspective' (i.e. solely a trench-arc-back-arc system) because subduction-related forces interplay with intra-continental deformation forces (Jolivet et al. 1989; Windley 1995) such that these areas are rarely long continuing, steady state systems (Hamilton 1994). Compressional stress fields in the vicinity of the plate suture often occur contemporaneously with strike-slip and extensional stresses in the overlying plate, at progressive distances from the suture zone (i.e. 'extensional' arcs Dewey 1980; New Zealand, Cole 1984; S.E.Asia, Fitch 1972, Nakamura and Uyeda 1980, Malod and Kemal 1996; Woodcock and Schubert 1994). Strike-slip fault movements are often associated with oblique plate convergence or areas of intra-continental deformation (e.g. Fitch 1972; Sibuet et al. 1987; Jolivet et al. 1989; Woodcock and Schubert 1994; Ingersoll and Busby 1995), and extension with forces related to sinking of the oceanic lithosphere such as subduction zone roll-back (e.g. Dewey 1980).

The Oligo-Miocene evolution of the Western Mediterranean also occurred in a complex tectonic setting where subduction, intra-arc and back-arc basin formation was accompanied by continental collision and strike-slip fault movement. The tectonic development shows similarities both to 'indentor' tectonics (McKenzie 1969; Molnar and Tapponnier 1975; Tapponnier 1977; Tapponnier et al. 1986) and to typical examples and models of back-arc basin formation. The proceeding sections review simple '2D' models of back-arc and intra-arc basin formation, the 'indentor' model, and then discuss implications from the Western Mediterranean.

10.2.1 Back-arc and intra-arc basin formation

Back-arc basins form when an existing island arc is rifted into two halves, or rifted along the front or rear of the arc-axis, followed by oceanic spreading between the two arc fragments (Karig 1970). Intra-arc basins can be viewed as incipient back-arc basins (where extension has not gone to completion) or aborted back-arc basins (on the remnant arcs, Smith and Landis 1995), and as such may retain a record of processes directly related to back-arc basin formation.

The Western Mediterranean back-arc basin and Sardinian intra-arc rift system has many general characteristics similar to other such settings. For example, the faulted morphology of the rifted margins, the location and size of the spreading centre, the basin size and bathymetry, and the presence

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of rotated continental microplates are similar to areas such as the Okinawa trough (Fig. 10.11, Sibuet et al. 1987; Letouzey and Kimura 1985), south China Sea and the Japan Sea (in Lonergan and White 1997; Marsaglia 1995)

10.2.1.1 Models of back-arc basin formation

Hypotheses for the formation of back-arc basins have been split into four models (Karig 1974). They may have formed by:

1) 'active spreading', where spreading is induced by forceful upwards asthenospheric flow or 'mantle diapirism' (Karig 1971).

2) 'passive spreading', where spreading is induced by passive upwelling of asthenosphere as a response to extensional forces acting on the lithospheric plate. The extensional forces are thought to result from several different mechanisms, one of which is subduction zone roll-back (Forsyth and Uyeda 1975; Dewey 1980; Carlson and Melia 1984; Malinverno and Ryan 1986; Viallon et al. 1986; Sibuet et al. 1987; Kastens et al. 1988; Charvet and Ogawa 1994; Windley 1995). On continental crust, marginal seas (i.e. back-arc basins) may be generated by transtensional strike-slip fault movements derived from stress propagation through the continent after continent-continent collision (e.g. indentor model, south China Sea, Molnar and Tapponnier 1975; Sea of Japan, Lallemand and Jolivet 1985, Jolivet et al. 1989). However, strike-slip extension may also be caused by oblique convergence (Okinawa trough, Sibuet et al. 1987) but in this case, rifting is oblique to the arc-trend. Often, several different mechanisms may operate to generate extension (e.g. Okinawa trough, oblique convergence plus roll-back, Sibuet et al. 1987; Charvet and Ogawa 1994). Finally, post-collisional extensional collapse has also been suggested as a possible mechanism creating continental back-arc basins (e.g. Faure and Charvet 1990; Carmignani et al. 1994, 1995).

3) extension induced by secondary convection cells in the mantle wedge overlying the subduction zone (e.g. Hsui and Toskoz 1981)

4) step-back in the location of oceanic lithosphere underthrusting (Karig 1974, Kearey and Vine 1996).

In Sardinia there is no evidence for mantle diapirism, such as domal uplift proceeding back-arc basin spreading. Admittedly, domal uplift would be difficult to detect because of the poor exposure and continental nature of the initial basin filling facies. However, continental facies are dispersed in response to local tectonic topography rather than radially outwards. Arc volcanism was localised by extensional sub-basins over a region of 200 km. Extension facilitated a route to the surface for volcanism (section 11.3.1) rather than volcanism facilitating extension. In addition, the passive margins of the Western Mediterranean basin have classic graben/half graben geometries suggesting that extension occurred in response to lithospheric stretching by pure shear (sensu McKenzie 1978) or simple shear (proposed by Mauffret et al. 1995).
The dominant force generating the extension is thought to have been subduction zone roll-back (this study; Malinverno and Ryan 1986; Lonergan and White 1997) which commenced contemporaneous with transtension before the opening of the back-arc basin (described below). The reasoning behind this comes from comparisons and similarities with other better studied areas such as the western Pacific where subduction zone roll-back is commonly accepted as a driving mechanism behind arc and subduction zone migration (e.g. Okinawa: Viallon et al. 1986; Sibuet et al. 1987; general: Forsyth and Uyeda 1975; Dewey 1980; Carlson and Melia 1984; Charvet and Ogawa 1994; Windley 1995).

The reasons are:
i) The aligned volcanic arc and extensional basins developed sub-parallel to the proposed subduction zone.

ii) The geometry of the back-arc basin can be explained by lateral variations in the amount of roll-back. That is, the most spreading occurred to the west of Sardinia rather than west of Corsica or the Balearics (Fig. 2.7) because the subduction system was pinned at its ends. This may have resulted from greater amounts of roll-back in the area east of Sardinia. This would also explain the increasingly arcuate nature of the subduction zone (see iv). Similar geometric criterion combined with a wealth of other data has been used to suggest extension created by roll-back in the Okinawa Trough (Viallon et al. 1986; Sibuet et al. 1987).

iii) In Sardinia, the timing of extension before and after back-arc basin opening and the timing of volcanism can be easily explained within a framework of progressive subduction zone roll-back. Regional compressional events such as the development of the Apennine chain and emplacement of the Kabylies (Fig. 2.1) also fit in well with a model of progressive subduction zone roll-back.

iv) The progressive roll-back of the subduction zone is thought to have continued from the late Miocene to the present and resulted in the formation of the Tyrrhenian Sea (Malinverno and Ryan 1986; Kastens et al. 1988). Steeply-dipping zones have been imaged to depths of 600 km using seismic tomography along the length of Italy and between Spain and north Africa and interpreted to be subducted oceanic lithosphere (Wortel and Spakman 1992). In some of these areas, beneath continental collision zones, the subducted slab appears to be ‘detached’ (Wortel and Spakman 1992). This detachment process would cause rapid uplift and extension in the overlying continental crust (Wortel and Spakman 1992), and such a model forms an alternative mechanism for the observed post-collisional extension to ‘asymmetric lithospheric delamination’, ‘asthenospheric intrusion’ (Channell and Mareschal 1986; Carmignani et al. 1994, 1995) or the convective removal of a lithospheric root (e.g. Platt and Vissers 1989, section 2.1.7).

Other models have advocated that extension in the Western Mediterranean back-arc basin, Sardinian Rift, northern Tyrrhenian Sea, Corsica and northern Apennines was caused by post-collisional extensional collapse in the Burdigalian, due to lithospheric delamination and asthenospheric intrusion after continental collision (Carmignani et al. 1994, 1995). As shown above, the timing of observed events is inconsistent with such a mechanism, space problems are created by the model (Robertson and
Grasso 1995), and a simpler, more logical mechanism is possible (below). Similar arguments to those discussed by Lonergan and White (1997) for the Betic-Rif orogen, such as the lack of radial extension/compression, the observation that potential energy considerations do not explain why extension should continue after the continental crust has thinned to a normal thickness, and the presence of rotated continental fragments, negate the use of a 'convective removal' model for extension in the Western Mediterranean back-arc and Sardinian intra-arc basins.

10.2.1.2 Characteristics associated with back-arc and intra-arc basins
Back-arc basins often initiate along the line of the volcanic arc because of the relative weakness of the lithosphere in the hot volcanic zone (Dewey 1980; Tamaki 1985) and because the topographic effect of the volcanic arc induces the highest vertical stress in this region (Charvet and Ogawa 1994). This is not always the case, for example, spreading in the Okinawa Trough is thought to have initiated behind the volcanic arc (Sibuet et al. 1987; Fig. 10.11). In the Western Mediterranean, extension in Sardinia was clearly accompanied by a volcanic arc before back-arc basin formation. The volcanic arc continued southwestwards into the Valencia Trough (Mauffret et al. 1995) Definitive information on nature and amount of volcanism in the submerged rift system to the northwest is lacking, but it seems that, although calc-alkaline volcanism did occur (e.g. Girod and Girod 1977), a clear volcanic arc did not exist in this area (Mauffret et al. 1995). Thus the opening of the western Mediterranean back-arc basin occurred to the west of the main volcanic arc and approximately in the centre of a diffuse system of extensional sub-basins. Rifting on the volcanic arc did not go to completion (i.e. the formation of oceanic crust) most likely because the Sardinian continental crust was subject to both strike-slip fault movement and extension (section 10.1.1), whereas extension dominated slightly further to the northwest.

Back-arc basins apparently often show an initial phase of diffuse, widespread extension which is succeeded by a phase of passive thermal subsidence once spreading is focused in the basin centre (e.g. Japan Sea, Sato and Amano 1991; Okinawa Trough, Letouzey and Kimura 1985; Lau Basin, Clift and Dixon 1994; Tyrrenhenian Sea, Kastens et al. 1988). After back-arc basin opening, the rifting may recommence on the fragment on the fore-arc side of the back-arc basin (Carey and Sigurdsson 1984). This pattern is observed in the Western Mediterranean. The first phases of extension in the Western Mediterranean occurred over a wide area, with numerous diffuse sub-basins. After initial extension, probably lasting less than a few million years (section 11.2.4), extensional stresses must have been concentrated in the area of the proto-back-arc basin, since passive basin filling is observed on seismic lines from the Gulf of Lion (Cravatte et al. 1976) and in Sardinia, after initial extension and before oceanic spreading (latest Oligocene-mid/late Aquitanian). Passive infilling of the earlier tectonic relief continued in Sardinia over the time of back-arc basin opening, but extension occurred again on the forearc side (Sardinia) of the back-arc basin after spreading had ended.
Periods of spreading in an associated back-arc basin often correspond to hiatal periods in volcanic activity in the intra-arc basin (western Pacific basins, Smith and Landis 1995; Sibuet et al. 1987, Okinawa trough). However, Sato and Amano (1991) presented a model for the Japan Sea in which back-arc basin opening was accompanied by rapid subsidence and voluminous bimodal volcanism in an intra-arc setting on the fore-arc side of the spreading centre. Volcanism in the Sardinian intra-arc basin was also basaltic-rhyolitic and most voluminous over the period of Western Mediterranean back-arc spreading. Volcanism in the ‘remnant arc’ (i.e. southern France) stopped approximately when back-arc spreading starts (Beccaluva et al. 1987), suggesting roll-back of the subduction zone linked to back-arc basin formation and similar to the model illustrated by Carey and Sigurdsson (1984) for the timing of volcanism in arc basins.

10.2.2 ‘Indentor’ models of continental deformation

Subduction in the Western Mediterranean was related to the closure of Neotethys and the convergence of Africa-Apulia and Eurasia. Similar processes were active along the entire Tethyan chain and subduction of oceanic crust was frequently interrupted by the collision of continental lithospheres. Continental collision resulted in complex patterns of deformation that are explained by an ‘indentor’ model (McKenzie 1969; Molnar and Tapponnier 1975; Tapponnier et al. 1986). The model consists of a rigid indentor of continental crust surrounded by oceanic crust colliding with fixed continental crust (Fig 10.12a). Compression in the immediate zone of collision passes laterally to areas of strike-slip faulting where ‘tectonic escape’ to areas away from the compressional zone occurs (Fig. 10.12a). The model is most commonly applied to the Himalayan collision (Molnar and Tapponnier 1975; Tapponnier et al. 1986; Fig 10.12b).

The zones of strike-slip faulting facilitate tectonic escape away from the indentor and towards the areas of oceanic crust subduction where transtension and extension occur (Fig 10.12a-c). For example, in the eastern part of the Himalayan collision, ‘Sundaland’ underwent tectonic escape to the south-east on sinistral strike-slip faults, towards the zones of extension, such as south China Sea and the Burman subduction zone (Tapponnier et al. 1986; Fig. 10.12b). Similarly, the indentation of the Arabian plate into Eurasia caused the westwards tectonic escape of Turkey on dextral strike-slip faults towards the zone of extension behind the Hellenic subduction zone (Taymaz et al. 1991; Fig 10.12c). In both these cases, extension is partly controlled by subduction zone roll-back (S. China Sea, Hall 1996; Greece, Taymaz et al. 1991; Fig 10.12bc).

Thus, in zones of continental collision and subduction, back-arc basin formation and regional tectonic development commonly result from the interaction of subduction-related forces and intracontinental deformation.
10.3 Implications for Western Mediterranean tectonics

The detailed model of Sardinian Rift evolution has implications for the Oligo-Miocene tectonics of the Western Mediterranean when the Corsica-Sardinia microplate is restored to its palaeogeographic position and linked to other regional events (published data). The deduction of the Western Mediterranean regional tectonic history also allows possible geodynamic models which may have driven intra-arc and back-arc basin formation to be evaluated and proposed. A modified ‘indentor’ model of continental collision (Adria-Eurasia) and tectonic escape, combined with subduction zone roll-back driven back-arc basin extension, can accommodate many of the events recorded in Sardinia and the Western Mediterranean (Sowerbutts 1997; Fig 10.12e). Figures 10.13 to 10.17 illustrate the following discussion.

10.3.1 Phase one - Paired compression- tectonic escape- extension (Mid-Late Oligocene-Late Aquitanian, 30-21 Ma; Fig 10.13, 10.14)

10.3.1.1 Tectonic events

Until the late Oligocene, the Sardinia-Corsica microplate was still attached to the southern Eurasian plate (section 2.1.1). A mid Oligocene-lower Miocene volcanic arc formed at the margin of the continental plate due to westwards-northwestwards dipping subduction of Neotethyan oceanic crust, which may have commenced in the Eocene (section 2.1.1.2). The subduction zone is thought to have extended from the Betic-Rif areas in the west to the Corsica/northern Apennines zone in the east (e.g. Beccaluva et al. 1987; Hill and Hayward 1988; Robertson and Grasso 1995; Lonergan and White 1997).

Prior to the mid-late Oligocene, the history of the Sardinian continental crust is poorly constrained. However, at the northeastern end of the proposed subduction system, continental collision occurred in northern Corsica in the Late Eocene-Oligocene (Egal 1992; Jolivet et al. 1990, 1991; Fig. 10.13) when no Neotethyan oceanic crust remained between Eurasia and Adria (Carmignani et al. 1995). The post-Palaeocene transpression on sinistral strike-slip faults in eastern Sardinia (Carmignani et al. 1992b) and southern Corsica may have occurred at this time in a manner similar to the model proposed by Carmignani et al. (1994, 1995; Figs. 2.8, 10.13). At the same time, considerable shortening occurred in the Alpine collision. In France, late Eocene extensional basins were formed on N-S trending normal faults and NNE-SSW to NE-SW trending faults (Hippolyte et al. 1993). The latter basins had a similar trend to the Eocene grabens of southern Sardinia (Fig. 10.13).

By the late Oligocene-earliest Miocene, Jolivet et al. (1990, 1991) show that ductile extension had commenced in the Corsican continental collision zone. The zone of compression had migrated to the Northern Apennines area to the east (Fig. 10.14, Kligfield et al. 1986; Carmignani and Kligfield 1990; Carmignani et al. 1994, 1995). The Alpine collisional belt was separated from the Northern Apennines....
by a transpressional deformation zone, active from the Late Oligocene-Burdigalian (Piana and Polino 1995). In southern France and the Gulf of Lion, a series of NE-SW trending extensional sub-basins were forming (Hippolyte et al. 1993, Fig. 10.14). The proto-Sardinian Rift also formed at this time, with the overall trend of the basin on a similar NNE-SSW trend. In detail, the initial sub-basins in southern Sardinia formed on different trends, most probably as a result of inheritance on older structures. For example, in south central Sardinia initial extensional faults were oriented NW-SE, parallel to Hercynian trends, and were cross-cut by slightly younger NE-SW to NNE-SSW more closely paralleling the main Sardinian Rift trend. The majority of active extension in the Sardinian Rift (this study) and over southern France/Gulf of Lion (Cravatte et al. 1987; Biju-Duval et al. 1987) occurred sometime within the mid-late Oligocene. The extensional system comprised numerous sub-basins over a wide zone with arc-volcanism, most prominent in Sardinia, which extended to sub-basins in southern France and the Gulf of Valencia (Cann and Hsü 1973; Girod and Girod 1977; Dewey et al. 1989; Maillard and Mauffret 1993; Fig. 10.14). This wide zone of extensional deformation and volcanism (multirift type, Tamaki 1988) may have been related to the presence of a shallowly-dipping slab of oceanic crust beneath the region (after Tamaki 1988).

In the Aquitanian, the locus of extension must have transferred to a narrower region in the centre of the extensional system (the proto-back-arc basin) since in Sardinia and the Gulf of Lion (Cravatte et al. 1974) passive infilling of the Oligocene tectonic reliefs occurred. The Sardinian Rift became an aborted intra-arc basin on the forearc side of the proto-back-arc basin. Arc-volcanism continued in France until ~20 Ma (Beccaluva et al. 1987) and ~18 Ma in Sardinia. Movement on sinistral strike-slip faults in eastern Sardinia and southern Corsica may have occurred before and during proto-rift formation since an extensional-transtensional topography was infilled by late Oligocene-lower Miocene basin fill. The last activity on the strike-slip faults is identified at 20.6±0.2 Ma (section 3.2.3), after proto-rift formation and concomitant with the start of back-arc basin opening.

Thus within the mid-late Oligocene-late Aquitanian, compression in the Northern Apennines occurred at the same time as extension at the margin of the Eurasian plate. The recognition of paired extension and compression contrasts with a development which considers that only extension (e.g. Cherchi and Montadert 1982ab; Hill and Hayward 1988; Dewey et al. 1989) or compression and transpression (Carmignani et al. 1994, 1995) occurred at this time.

10.3.1.2 A modified ‘indentor’ model combined with subduction zone roll-back
Tapponnier (1977) suggested that the Western Mediterranean basin and N-S fault systems in Europe formed in a ‘expulsion latérale mixte’ of continental crust resulting from Alpine collision (Fig. 10.12d). Whilst this model may be applicable for the Eocene (Fig 10.13), it does not account for the observed orientation of Oligo-Miocene extensional basins. Lecca et al. (1997) favour an indentor
model where back-arc extension occurred as a counterpart to shortening in the Northern Apennines rather than in response to subduction zone roll-back.

In the model presented here, the large scale 'indentor' was the Adria microcontinent which collided with the Eurasian plate to form the Alps (Tapponnier 1977). The convergence between Africa and Eurasia set-up the N-NW dipping subduction system of Neotethyan oceanic crust extending from Corsica to the Betics. At the eastern margin of this zone, where the Neotethyan oceanic crust ran out, continental collision occurred (Fig. 10.12e). In the late Eocene, this occurred in northern Corsica, but by the Late Oligocene, progressive oceanic crust subduction and/or roll-back meant that continental collision propagated westwards and southwards to form the N-S trending Northern Apennines (Fig. 10.12e, 10.14). It is the northern Apennines collisional 'indentor' which led to the re-activation of strike-slip faults in southern Corsica and eastern Sardinia and which accommodate tectonic escape (Fig. 10.12, 10.14). The tectonic escape was directed towards zones which were in extension to the west/southwest (Fig. 10.12, 10.14). Transtension occurred at the boundary of the extension zone (Fig. 10.12, 10.14). Extension is thought to have been driven by subduction zone roll-back (10.2.1.1).

Continental deformation in the Western Mediterranean thus exhibits similar patterns to other Tethyan zones, though the geometries are slightly more complex (Fig. 10.12; Molnar and Tapponnier 1975; Tapponnier 1977; Tapponnier et al. 1986; Taymaz et al. 1991; Hall 1996). The geodynamics of the regions are complex, involving forces related to continental collision and paired with those caused by subduction zone roll-back.

10.3.2 Phase two - Opening of the Western Mediterranean back-arc basin and rotation of the Corsica-Sardinia microplate (Late Aquitanian-early Burdigalian, 21-18 Ma; Fig. 10.15)

Concomitant with the opening of the Western Mediterranean back-arc basin, thrust fronts in the Northern Apennines migrated to the east, extension in the western Northern Apennines commenced, and tectonic escape recorded by movement on sinistral strike-slip faults in the Sardinian continental crust ended (Fig. 10.15). These events could be linked to a peak in subduction zone roll-back, such that the proposed shallowly-dipping slab underneath Sardinia/southern Eurasia steepened, possibly resulting in the observed, geochemically-zoned magmatism (Beccaluva et al. 1987). Slab roll-back, steepening, and/or detachment under the continental collision zone of the Northern Apennines would create the migrating compressional belt paired with extension.

In the Late Aquitanian-early Burdigalian, extension created by east/south eastwards roll-back was accommodated by the opening of the back-arc basin (Figs. 10.15, 10.16). A set of E-W trending faults and separation between Corsica and Sardinia may relate to an additional component of N-S
extension created by southwards directed roll-back to the south of the microplate. Back-arc basin spreading and microplate rotation may have largely ended by ~18 Ma because the main phase of roll-back ended and/or the subduction zone had migrated too far away for a major influence to be felt. On a larger scale, the size of the extensional system was controlled by the compressional zones present to the west and east.

10.3.3 Phase three - Post rotation extension and migration of compressional belts (mid Burdigalian, ~17-18 Ma; Fig. 10.16)

After back-arc basin opening, progressive subduction zone roll-back away from the Corsica-Sardinia microplate dominated the evolution of the eastern Western Mediterranean. This meant that calc-alkaline volcanism in the Sardinian volcanic arc ended. In the back-arc and intra-arc basins, marine sedimentation kept pace with thermal subsidence and infilled earlier fault-formed topography. A mid Burdigalian phase of ~E-W and ~N-S extension in Sardinia may have resulted from an extensional stress-field created by eastwards and southwards roll-back of the now arcuate subduction zone. Compression continued to migrate outwards in zones surrounding the subduction zone.

10.3.4 Phase four- Relative tectonic quiescence and intra-arc basin filling (Late Burdigalian-mid Serravalian, 17-12 Ma; Fig. 10.17)

Though compression continued in areas surrounding the Western Mediterranean (Apennines, N. Africa etc.), the mid-Miocene appears to have been a time of tectonic quiescence in the central Western Mediterranean with only small amounts of east-west extension on syn-sedimentary N-S normal faults observed. Again, extension may be a possible consequence of further subduction zone roll-back, but in an area some distance to the west of the Sardinian continental fragment in the region of the proto-Tyrrhenian Sea. Intra-arc basin sedimentation was the main event recorded over this time period.

The Tyrrhenian Sea is believed by some to be have formed due to continued and progressive roll-back of the same subduction zone from the Late Tortonian to the present day (Malinverno and Ryan 1986; Kastens et al. 1988), which resulted in Plio-Pleistocene N-S normal faulting in Sardinia and reactivation of NW-SE faults in the Campidano area.

10.3.5 Geodynamic summary

The complex Sardinian Rift evolution is thought to result from the combination of tectonic escape resulting from continental collision and pulsed extensional stresses approximately perpendicular and parallel to the main rift trend which were a consequence of subduction zone roll-back (Fig. 10.12c).
10.4 Comparisons with other intra-arc and back-arc settings

The Sardinian Rift and Western Mediterranean back-arc basin have features in common with other intensively studied modern and ancient arc, back-arc and subduction systems. These can be summarised:


- Rates of back-arc basin opening and microplate rotation are rapid. For example, the entire sea of Japan opened in from 28-18 Ma resulting in the rotation of Japan (Tamaki et al. 1992). Intra-arc rifting in northern Japan lasted for only 3 Ma with high rates of initial subsidence (1 km/Ma, Yamaji 1990). Other arc-basins commonly have rates of tectonic subsidence >200 m/Ma (Smith and Landis 1995). In the Western Mediterranean, the back-arc basin opened over a 3 Ma period with an associated microplate rotation of 30° (Rehault et al. 1984). Pulsed phases of rapid subsidence and volcanism are common (e.g. Kobayashi 1983, northwestern Pacific periods of volcanism and subsidence of a few Ma). Periods of tectonic subsidence in the Sardinian intra-arc rift last for only a few Ma with rates of 50-250 m/Ma.

- Similar geometries and dimensions. For example, the common 10's-100 km width of back-arc basins, block-faulted margins, the location and size of the spreading centres, the basin bathymetry, the 100's metres-10 km thick fill of intra-arc basins and the arcuate nature of compressional belts and subduction zones (Okinawa, Sibuet et al. 1987; Letouzey and Kimura 1985; south China Sea and Japan Sea, Tamaki et al. 1992; Marsaglia 1995, Smith and Landis 1995; Lonergan and White 1997)

- Diffuse extension focused on the volcanic arc before back-arc basin opening, thermal subsidence in the intra-arc over the period of opening and continued extension on the volcanic arc after back-arc basin formation (e.g. general, Carey and Sigurdsson 1984; Japan Sea, Sato and Amano 1991; Okinawa Trough, Letouzey and Kimura 1985; Lau Basin, Clift and Dixon 1994; Tyrrhenian Sea, Kastens et al. 1988, Chapter 10, 11).

Chapter 11
Chapter 11- Significance of the Sardinian Rift for processes active in extensional settings

Synthesis of detailed field observations made along the Sardinian Rift enable the styles and timing of extension, basin filling and the controls on basin filling to be determined within this intra-arc basin. The results enable each of these processes to be compared to features typical of other extensional settings with the aim of elucidating common characteristics. The overall theme of this chapter is to determine whether processes active in the intra-arc basin can be rationalised. Any resultant, predictive models or general characteristics would be particular use in determining the character of submerged intra-arc basins which may be suitable for hydrocarbon exploration.

11.1 Structural geometry of the intra-arc basin

The complex structural geometry of the Sardinian Rift results from multiphase extension and transtension on normal and strike-slip faults of numerous orientations.

11.1.1 Extensional geometries

Normal faults within the Sardinian Rift occur on a variety of different orientations and sizes with throws from a few millimetres to ~2 km (Fig. 8.4). 'Large' normal faults with throws of 50-2000m accommodate the majority of extension, and define a number of semi-independent depocentres or sub-basins, each with its own characteristics. Within the depocentres, smaller faults with throws of a few to tens of metres cause local topographic irregularities.

11.1.1.1 Fault plane geometries

Though large Oligo-Miocene normal fault planes are virtually always either degraded or covered, the abrupt variations in basin fill thickness over planar surfaces (e.g. Sassari sub-basin; section 7.1.3), bed offsets (e.g. Anglona study area; section 8.2.3), basement-basin fill relationships (e.g. Sarcidano sub-basin; section 5.2) and basin filling geometries (chapters 4-8) suggest that large normal faults accommodating extension were planar and had steep dips, often 60-90°. This is consistent with structures commonly observed regions of active continental extension (Jackson and White 1989) though the fault-plane dip is commonly greater than the average value of 50-60° (Jackson and White 1989), perhaps because of the small amount of overall extension in Sardinia (section 11.2.6). Normal faults with a listric geometry and associated, divergent rotated beds with rollover anticlinal geometries in the hangingwall, are found only within the sedimentary basin fill (Sassari sub-basin 7.2.4.5, Oschiri sub-basin 6.4.3). They are not commonplace and do not accommodate the majority of extension.
Small fault planes which can be measured directly in outcrop are often sub-parallel to the proposed larger structures. Most are high angle, planar and occasionally have dip-slip or slightly oblique-slip slickensides. Within individual sub-basins, extensional faults with differing throws form fault sets or populations with similar orientations. Fault populations are common in extensional settings (e.g. Davison 1994). Most study areas contain more than one fault set, where each population moved at different times (e.g. Sarcidano, section 5.5) or at the same time (e.g. Logudoro, section 7.3; Anglona, section 8.6).

11.1.1.2 Fault linkage

Extensional faults within the Sardinian rift are segmented. Faults with cumulative throws of hundreds to thousands of metres are commonly 5-20 km long, with approximately linear surface traces. Normal --E-W faults in the Castelsardo sub-basin have curvilinear traces (Fig. 8.1). Fault segments in other extensional settings have similar characteristics (e.g. Jackson and White 1989; East Africa, Rosendahl et al. 1986; North Sea, Rattey and Hayward 1993, Thomson and Underhill 1993, Underhill 1994; Gulf of Suez, Colletta et al. 1988, Patton et al. 1995; Greece, Gawthorpe et al. 1994).

Segment linkage occurred in a variety of ways. Faults on similar orientations were linked by synthetic relay ramp geometries (Sarcidano, section 5.2, Fig. 5.5) or synthetic interference zones (Portotorres basin margin, section 8.2.5, Fig. 8.15; terminology and examples in Larsen 1988; Morley et al. 1990; Gawthorpe and Hurst 1993). Faults on different orientations were linked by antithetic interference zones (sensu Gawthorpe and Hurst 1993, Castelsardo-Perfugas sub-basins, section 8.2, Fig. 8.12) or appear to join or cross-cut each other. Lack of exposure means that the relationship between apparently joining or cross-cutting faults cannot often be deduced. It is possible that faults on different orientations moved at the same time, as has been documented from other places (sections 7.3, 10.1.8; Arran, Woodcock and Underhill 1987; North Sea, Cherry 1993; Angelier 1994).

11.1.1.3 Fault Growth

In Sardinia, evidence suggesting incremental vertical growth of extensional faults is recorded by occasional syn-rift deposits in 2D sections roughly perpendicular to the fault plane (section 11.2.1). The effects of lateral fault growth and propagation, now recognised as a major control on the development of syn-rift sequences (e.g. Leeder and Jackson 1993; Gawthorpe et al. 1997; Sharp et al., in press), were not recognised within the moderately exposed Sardinian Rift.

11.1.1.4 Geometries associated with normal faulting

The geometries of uplifted, tilted footwall blocks and subsided hangingwall depocentres around Sardinian normal faults are similar to those observed in present day or ancient extensional settings (e.g. Borah Peak, Stein et al. 1988; Greece, Gawthorpe et al. 1994 and personal observations; Gulf of
Suez, Angelier 1984, Jackson et al. 1988; North Sea, Yielding 1990, Rattey and Hayward 1993, Fig. 11.1). Such behaviour results in the formation of the half-graben geometry which is recognised as the fundamental structural style in many extensional settings (e.g. Rosendahl et al. 1986; Leeder and Gawthorpe 1987; Prosser 1993). Often, an extensional basin may be formed from a series of half-graben depocentres separated by uplifted, rotated fault blocks (e.g. Gulf of Suez, Angelier 1984, Jackson et al. 1988; North Sea, Yielding 1990, Glennie 1990, Rattey and Hayward 1993, Fig. 11.1). Individual elements of the Sardinian Rift structure regularly have half-graben geometries (e.g. Logudoro area, Fig. 7.1), some with tilted fault blocks (e.g. Sarcidano sub-basin, Fig. 5.1; Anglona study area, Figs. 8.1, 8.4). Tilted fault blocks in the Sardinian Rift are commonly degraded in a style similar to those observed in the North Sea (e.g. Yielding 1990; Macleod 1995; Underhill et al. 1997; Fig. 11.1).

Within individual sub-basins or study areas, the fault-created topography shows characteristics similar to that of other extensional terrains, with segmented normal faults bounding tilted, uplifted blocks and half-graben, bounded by transfer zones (e.g. in East Africa, Rosendahl 1987, Morley 1990; Greece, Gawthorpe et al. 1994, Collier and Gawthorpe 1995; North Sea, Underhill 1991, Rattey and Hayward 1993, Thomson and Underhill 1993, Underhill et al. 1997; Gulf of Suez; Angelier 1984, Patton et al. 1995, and in Leeder and Gawthorpe 1987; Gawthorpe and Hurst 1993). For example, uplifted, tilted blocks and subsided half-graben in Sarcidano and the Anglona study areas show affinities to the 'tilted domino' scenario proposed by Jackson et al. (1988, e.g. Fig. 11.1). In these areas, the structural style which formed a complex template for the deposition of sediments is one which allows comparison with, and has applications to, other rift basins. In other areas, the combination of differently oriented normal faults and strike-slip faults, leads to sub-basins and an overall rift structure with unique palaeogeographies.

### 11.1.2 Strike-slip geometries

Strike-slip faults present in the eastern Sardinian basement crop out as narrow zones of intense brittle deformation containing high-angle fault planes with left-lateral, oblique, normal and reverse slickensides (this study; Alvarez and Cocozza 1974; Carmignani et al. 1992b). Sinistral offsets of up to 4 km are visible in basement lithologies adjacent to the faults (on Barca et al. 1996). Strike-slip duplexes sensu Woodcock and Fisher (1986) formed in areas of E-W and NE-SW inherited fault trends in eastern Sardinia (Carmignani et al. 1992b). Transpressional duplexes were located at restraining bends (M. Albo, Carmignani et al. 1992b) whilst transtensional sub-basins formed at releasing bends (Oschiri, Oggiano et al. 1995). The geometries within these sub-basins are similar to theoretically predicted and observed structural styles in strike-slip settings (e.g. Christie-Blick and Biddle 1985; Woodcock and Fisher 1986; Woodcock and Schubert 1994). The transtensional basins formed elongate depressions within the Palaeozoic basement which linked to the main Sardinian rift.
Where the strike-slip fault systems meet the extensional fault systems of the rift, an irregular tectonic relief was formed by the intersecting fault trends (section 6.5).

11.1.3 Importance of inherited basement structures

The southern part of the Oligo-Miocene Sardinian Rift was formed by a number of different sub-basins with differing orientations. This is due to inherited structural trends. In southernmost Sardinia, the Sardinian Rift is composed of a number of -E-W trending sub-basins. Two of the sub-basins were present during the Eocene (Sulcis, Cixerri, Fig. 1.1, Cherchi and Montadert 1982ab; Assorgia et al. 1992bce; Barca and Costamagna 1997) and were reactivated in the Oligo-Miocene. In central Sardinia, the overall NW-SE trend of the Campidano and Sarcidano areas reflects the underlying orientation of Hercynian thrust fronts and shear zones (e.g. on Carmignani et al. 1987). In the Sarcidano study area, the orientation of the basin margin changes in an area where thrust fronts trend from -N-S to -NW-SE (Fig. 5.1). In addition, the line of tilted fault blocks within the Sarcidano sub-basin occurs along the line of a Hercynian antiformal culmination which was intruded by late Hercynian granites (Carmignani et al. 1992a). These features suggest that inherited structures controlled the orientation of brittle, extensional faults at depth (i.e. negative inversion sensu Williams et al. 1989), in a manner similar to that observed in the American Rockies (Powell and Williams, 1989) and Wessex basin (Chadwick 1986). It is worth noting that in the Sarcidano sub-basin, normal faults on the inherited trend moved first, whilst structures aligned on the main Sardinian rift trend accommodated extension at a later stage. Lecca et al. (1997) also recognise that NW-SE trending, inherited Hercynian structures influenced the formation of the Oligo-Miocene rift.

Strike-slip faults trending NE-SW to E-W in the eastern Sardinian basement and southern Corsica are thought to have re-activated late Hercynian structures (Chabrier and Chorowicz 1981; Carmignani et al. 1992b; Oggiano et al. 1995). These inherited weaknesses were in the correct orientation to facilitate tectonic escape from the Northern Apennines compressional zone (section 10.2, 10.3). It is unclear whether the faults continued to the west, underneath the Sardinian Rift. However, the orientation of the -E-W Funtanazza, Sulcis and Cixerri grabens suggests that such structures may continue south and westwards. In addition, the latter graben occurs adjacent to a ~30km long, -E-W trending Hercynian fault mapped in eastern Sardinia (Barca et al. 1996).

The sub-basin variability and Oligo-Miocene history of Sardinia can therefore be partly explained by the re-activation of inherited structures. Inherited fault trends are often important in determining the structural styles of normal faulting and transfer zones in other extensional basins (e.g. North Sea Glennie 1990, Cherry 1993; Gulf of Suez, Patton et al. 1995).
11.2 Basin filling geometries and structural development

In this study, careful examination of basin filling geometries combined with temporal constraints have enabled the structural development of a complex rift basin to be established. The approach used is to define strict criteria for the identification of periods of fault growth (syn-rift deposits, Appendix 1B) on a particular fault or set of faults as opposed to periods of passive infilling of earlier formed fault topography (post-rift deposits, Appendix 1B).

11.2.1. Syn-rift geometries

In previous field-based and seismic studies of extensional settings, a variety of criteria have been used to identify 'syn-rift sequences'. Examples which may occur, but are not valid criterion for defining coeval fault movement, include the presence of coarse clastic sediments, abrupt facies changes, abrupt thickness changes and the filling of half-graben geometries (e.g. in Spec. Publ. Int. Ass. Sedm. 20, 1993; Sardinia, Cherchi and Montadert 1982ab; Gulf of Lion, Gorini et al. 1993). The approach of Prosser (1991, 1993), who examined the classification of syn-rift sequences and specifically defined criteria which relate to fault movement (Appendix 1B), is used here. Because there are several phases of short-lived fault movement in the Sardinian Rift, individual outcrop scale syn-rift deposits are very different than the syn-rift megasequence (11.2.4 below, Fig. 11.2).

Typical syn-rift and syn-transtensional geometries observed in the Oligo-Miocene Sardinian Rift are:

- angular unconformities related to differentially rotated beds which diverged into a hangingwall depocentre (e.g. Castelsardo sub-basin, Figs. 8.23, 8.26, 8.27, 8.29; Sarcidano sub-basin, Figs. 5.20, 5.21)
- bed thickening and thinning over an active fault, sealed by overlying sediments (e.g. Castelsardo sub-basin, Figs. 8.22, 8.24; Sassari sub-basin, Fig. 7.1 B-B').
- localised variations in bed dips in the vicinity of faults, sealed by overlying sediments (e.g. Sassari sub-basin, Fig. 7.5a; Sarcidano sub-basin Figs. 5.15, 5.22). In these cases, an unconformity and the increase in dips occurs in the fault hangingwall, towards the hangingwall depocentre. Dip variations are thought to be caused by fault-propagation folding (sensu Withjack et al. 1990; Gawthorpe et al. 1997; Sharp et al. in press) and/or fault drag.
- bed divergence and rollover anticlines associated with listric faults (Oschiri sub-basin, Figs. 6.13, 6.17; Sassari, Figs. 7.40, 7.41).
- syn-depositional folding and associated angular unconformities (e.g. Oschiri, Fig. 6.16; Sarcidano, Figs. 5.23, 5.24).

Other possible syn-rift geometries were localised angular unconformities which could not be directly related to a fault (e.g. Figs. 5.16, 5.17), and may have been caused by erosive sedimentary processes.
Syn-rift geometries occur at a variety of scales from a few centimetres up to 50m. It is very noticeable that syn-rift deposits are rarely found, only recognised in close vicinity to faults and form a voluminously minor part of the basin fill (Fig. 11.2). The sorts of syn-rift geometries observed are similar to those found in other extensional settings at a much larger scale on seismic sections (e.g. Prosser 1991, 1993; Glennie 1990; Rattey and Hayward 1993; Thomson and Underhill 1993). At an outcrop scale, stacked progradational fan-delta deposits similar to those in the Castelsardo sub-basin are found within syn-rift sequences in Greece (Gawthorpe et al. 1994), fault-propagation folds are found in the Gulf of Suez (Gawthorpe et al. 1997; Sharp et al. in press) and similar, sealed, syn-rift fault geometries are observed in the south of France (Hippolyte et al. 1993). The seismic reflection line from offshore the Castelsardo and Portotorez sub-basins (Fig. 8.4, Enclosures 6 and 7) shows that syn-rift megasequences formed only a small proportion of the half-graben fill and consisted of beds which diverged slightly into the hangingwall depocentre and mounded, aggradational beds adjacent to the fault scarp. These geometries are similar to the rift-initiation and rift climax arrangements classified by Prosser (1991, 1993, Fig. 11.6).

Divergent, differentially rotated syn-rift deposits are recorded most frequently within rapidly deposited clastic sediments (e.g. Castelsardo Member) or in platform carbonates (e.g. Isili Formation) where the rate of sedimentation was high enough to record contemporaneous fault movements. In areas with lower sedimentation rates, rapid fault growth resulted in the creation of an underfilled tectonic relief which was passively infilled with post-rift deposits (below). The lowest exposed, coarse continental clastics (Villanovatulo Member, Casteldoria Member) are syn-rift deposits. Immediately overlying continental clastics show no evidence for syn-depositional tectonism, though this may be difficult to detect because of the nature of the sediment. Combined with their variable dispersal paths, one can consider that these sediments are a response to the topography created as a result of tectonism. It was suggested by Blair (1987), and is inherent in the model of Prosser (1991, 1993), that the influx of coarse clastic material just after active tectonism ended is due to the establishment of drainage systems. Whilst this may be generally true in Sardinia, the influx of coarse clastic material is strongly dependent on the spatial location around the evolving tectonic relief, as observed in other areas (see sections 11.4, 11.5; Gawthorpe et al. 1994, 1997). Voluminously supplied volcanic rocks would be likely to record syn-eruptive fault movements. The Sardinian volcanic rocks exposed at the surface today passively infill rift topography, the interpretation being that the majority were erupted after active extension.
11.2.2 Post-rift geometries

The majority of the basin fill within the Sardinian Rift forms a largely conformable, parallel-bedded succession which passively onlaps and infills earlier fault-created topography. These are interpreted as post-rift deposits (Fig. 11.2). Many examples have been given of the way in which post-rift deposits onlap and cover degraded fault planes, rotated fault blocks and basement topography (e.g. Figs. 5.1, 5.2, 5.10, 5.12, 5.13, 5.14, 6.7, 6.9, 7.5d-fjk, 7.6, 8.7, 8.16, 8.30). Within the post-rift deposits, lateral facies changes and abrupt facies thickness variations are common, as a result of the influence of the existing tectonic relief. The infilling of tectonic relief by Sardinian post-rift deposits which form the majority of the basin fill sequence is similar to those described from other extensional basins by Prosser (1991, 1993). The seismic reflection lines from offshore the Anglona study area (Fig. 8.4, Encl. 6, 7) show a similar signature, such that the bulk of the basin fill records parallel-bedded, post-rift deposition.

At the present day, the post-rift deposits are flat lying or dip in a variety of orientations at less than 10°. Post-depositional dips and gentle folding are thought to have been caused by sediment compaction and by post-depositional, high angle brittle faulting, often on a ~N-S trend. Steep dips immediately adjacent to fault planes (e.g. Sarcidano sub-basin, Isili block, 5.3.4) are interpreted as being caused by post-depositional fault-propagation folding (sensu Withjack et al. 1990; Gawthorpe et al. 1997; Sharp et al. in press).

11.2.3 Unconformities within the basin fill sequence

Contemporaneous fault activity and sea level change in many extensional terrains results in a basin fill which is cut by numerous unconformities and/or flooding surfaces which form the basis for sequence stratigraphic interpretations (e.g. Gawthorpe et al. 1994, 1997). In the Sardinian Rift, unconformities related to syn-depositional fault movements occur only rarely and locally, within the voluminously minor syn-rift deposits. Unconformities which relate to relative sea level change, if present, are not exposed, apart from a hiatal surface in the Genoni area of the Sarcidano sub-basin (Serra Longa Member) and erosion surfaces in the Castelsardo Member syn-rift deposits (Figs. 8.28, 8.29). In particular, the extensive mid Miocene unconformity surfaces shown by Martini et al. (1992) in the Logudoro study area were not recognised. A major unconformity was developed in northern Sardinia due to phases of E-W and N-S faulting in the mid Burdigalian. Apart from this, correlation between outcrops indicates that the majority of the basin fill sequence appears to be conformable and parallel-bedded.
11.2.4. Implications for the timing of fault movement

The voluminously minor syn-rift deposits which formed in response to pulsed extensional events, and dominant post-rift basin filling geometries observed in the Sardinian Rift, suggest that phases of active extension were short-lived and that within these, the rate of tectonic subsidence far exceeded rates of basin filling. At the end of a phase of active extension, an underfilled tectonic topography existed which was gradually infilled by post-rift deposits. Temporal constraints applied to the basin fill geometries suggest that the phases of active extension occurred for periods of less than a few million years and therefore that rates of tectonic subsidence were ~50-250m/My.

Prosser (1991, unpublished) defines a syn-rift megasequence as the period over which active extension occurs on any fault within an extensional system. That is, the majority of the syn-rift megasequence can be post-rift deposits, as is the case in Sardinia (Fig. 11.2). In the Sardinian Rift, syn-rift megasequences can be identified in the mid-late Oligocene (first extension phase) and from the late Aquitanian-mid Burdigalian (second and third extension phases) though minor outcrop scale extensional faults occur outside these time periods (Fig. 11.2).

11.2.5 Comparison with other extensional settings

The lack of large scale syn-rift sedimentation with the Oligo-Miocene Sardinian Rift contrasts with other extensional settings such as the syn-rift sequence of Upper Jurassic of the North Sea (e.g. Brae depositional system of the South Viking Graben, Rattey and Hayward 1993; Partington et al. 1993), the Helmsdale Boulder Beds of the Inner Moray Firth, (Underhill 1991ab; Thomson and Underhill 1993), and examples from the Barents Sea (Prosser 1993), where large-scale, divergent, rotated wedges span timescales of millions to tens of millions of years and reflect large cumulative throws on extensional faults.

Back-arc and intra-arc basins on continental crust, driven by subduction zone roll back, often exhibit episodic extension with very fast subsidence rates and shorter durations of extension (e.g. 1 km/My for 3 Ma, Japan, Yamaji 1990; Smith and Landis 1995; Orton 1996) in comparison to 'normal' extensional basins e.g. locally 100m/My maximum tectonic subsidence in the North Sea (Turner 1995) for periods of 10Ma (Underhill 1991ab; Rattey and Hayward 1993).

The implication of the above on the basin filling succession of arc-basins would be to accentuate the initial underfilling and subdued syn-rift deposition such as is observed within the Sardinian Rift. Short-lived phases of extension associated with high strain rates in the Sardinian Rift may be due to softening of crustal rocks as a result of increased heat flow and systems of dykes, sills and faults which dissipated variations in strain with depth in the lithosphere (after England et al. 1993), driven by pulsed subduction zone roll back (chapter 10).
11.2.6 Patterns of fault movement

The synthesis of this outcrop scale study (chapter 10) allows some general points to be made about the patterns of fault movement within the Sardinian Rift. Within the temporal resolution of this study (i.e. ~1 Ma) all faults within a fault set or population in a particular sub-basin moved at the same time. In northern Sardinia, two, differently oriented fault families may have moved at the same time, in the mid-late Oligocene and the mid Burdigalian. In truth, the faults may have been active at different, but over a very short timescale (<1 Ma). Examples from other settings indicate that, in the absence of a consistent cross-cutting relationship, faults of different orientations may have been active in the same stress field (Woodcock and Underhill 1987, Angelier 1994). Such a scenario is consistent with the observations from northern Sardinia.

Many extensional basins exhibit propagation of the rift structure through time (e.g. Banda and Santenach 1992; Sharp et al., in press). In Sardinia, extensional faulting began in southernmost areas in the Eocene (Cherchi and Montadert 1982ab; Assorgia et al. 1992bce; Barca and Costamagna 1997). By the latest Oligocene, the overall ~N-S trending Sardinian Rift had formed along the entire length of western Sardinia (chapter 10), though it is unclear from the poor exposures and temporal constraints whether the structure propagated northwards or occurred coevally.

Extension within the rift basin was pulsed (section 10.1.8). Whilst the older Oligocene faults defined the rift margins, mid Burdigalian normal faulting occurred towards the centre of the rift (Portotorres sub-basin, Logudoro study area, Fig. 11.2). Fault activity in extensional rifts often appears to young towards the centre of the basin (e.g. Gulf of Corinth area, Greece, Collier and Gawthorpe 1995; East Africa, Baker 1986; North Sea, J. Underhill pers. comm. 1997), although the mechanisms for this do not appear to be understood. In the case of the Sardinian Rift, younger faults may have occurred preferentially in the zones of hot, weak crust dominated by Oligocene-lowermost Miocene subduction-related volcanism (after ideas in England et al. 1993).

11.2.6 Quantifying extension

Detailed, quantitative work on the amount of upper crustal extension was not undertaken because in each sub-basin centre there is no way to accurately estimate the fault throws and no complete or representative section was available for backstripping and subsidence analysis (see 11.4 also). In a very crude way, fault throws on the seismic section from just offshore northern Sardinia were summed by using a conversion velocity of 1.5 km/s. The vertical throw on all the faults would result in ~2.1 km of horizontal extension of the crust from the ?Oligo-Miocene to the present day (assuming an average fault plane dip of 60°). This is likely to be an underestimate since not all faults are visible on the seismic reflection profile. With the present day rift width of 46 km, this gives $\beta \approx 1.05$ ($\beta$=extended
length/original length). Geophysical studies show that, at the present day, there is no thinning of the Sardinian crust over the area of the Oligo-Miocene or Plio-Pleistocene rift basins (Egger et al. 1988). This suggests that the crust was thinned in the Oligo-Miocene but underplated by Oligo-Miocene or Plio-Pleistocene volcanism.

Features one would expect in continental crust extended by a factor of 1.05 are noted in the Sardinian Rift e.g. the high angle of the normal faults which have not been rotated by continued extension (as in the domino model of Jackson et al. 1988) and the ~2 km maximum thickness of sediment (seismic section) or 1500m estimated by (Cherchi 1985). In comparison, the North Sea which is reported to have β=1.6-2 (Schlatter and Christie 1980; Barton and Wood 1984; Klemperer 1988) has a sediment thickness up to 4 km (Glennie 1990, but over 100’s Ma) and greater tilting of fault blocks (e.g. Yielding 1990; Rattey and Hayward 1993).

11.3 The interaction of extension, transtension and arc-volcanism

11.3.1 Spatial relationships between basin structures and arc-volcanism

At the Western Mediterranean, southern margin of the Eurasian plate, Oligocene-early Miocene arc-volcanism is found only within extensional sub-basins of the rift system (Sardinia, France, Gulf of Lion, Valencia). In Sardinia, andesitic volcanic centres are found within the main part of the rift basin and within smaller E-W trending graben in southern Sardinia. Several of the andesitic volcanic centres are located at the intersection points of differently oriented major extensional faults. For example, the Monte Arci volcanic centre occurs where the Sarcidano sub-basin meets with the main N-S rift trend, the Arcentu Group at the intersection between the NW-SE Campidano system and Funtanazza sub-basin, and the Tergu Formation, between the Ploaghe, Sassari, Sennori and Portotorres fault systems (Figs. 1.1, 4.1, 7.1, 8.1). Final andesitic cones in the Logudoro study area are aligned on a NE-SW trend indicating possible control by inherited basement structures at depth. These observations suggest that, in common with other settings (e.g. East Africa, Baker 1986; Greece, Pe-Piper et al. 1994; Orton 1996), upper-crustal extensional faulting facilitated magma emplacement.

11.3.2 Temporal relationships between extension, transtension and arc-volcanism

Since the location of volcanic centres is apparently related to extensional faults, it would appear logical that periods of intense ‘subduction-derived’ magmatism would be related to phases of upper crustal extension. Such a link can be observed in other back-arc extensional settings, even where the amount of extension may be relatively small (e.g. Greece, Pe-Piper et al. 1994; Peru, Petford and Atherton 1994; Japan, Sato and Amano 1991).
In Sardinia, the first phase of mid-late Oligocene extension is associated with volcanism. The Oligocene eruptive products were basaltic-andesitic in composition and record the initial pulse of subduction-modified magmas, where a direct path to the surface was provided by extensional and transtensional faults. However, intense volcanism continued after the first extension phase had ended, during back-arc basin formation (Aquitanian-early Burdigalian). In contrast, Lecca et al. (1997) propose that in the Sardinian Rift, ignimbritic volcanism was associated with a transtensional and extensional Aquitanian-Burdigalian rift phase, but give no actual evidence for syn-sedimentary faulting. Sato and Amano (1991) also document continued volcanism on the fore-arc side of the spreading centre contemporaneous with back-arc basin formation in Japan.

The mantle source and degree of melting appears to have been constant over the periods of Sardinian rifting, back-arc basin formation and along the length of the rift. Fractional crystallisation often controlled major and trace element concentrations and some trace element ratios. However, the most basaltic magmas record an possible decrease in subduction-enrichment over the period of back-arc opening, acting independently on the constant source (chapter 9). Zoned magmatism with increasing northward concentrations of K, Rb, La/Nb, Rb/Zr etc. (this study; Coulon and Dupuy 1975; Coulon 1977) tentatively suggests a northward increase in the slab-flux component combined with different fractional crystallisation histories in the north and south (chapter 9). Over this time, a range of voluminous andesitic-rhyolitic magmas which formed by crustal level fractional crystallisation and contamination (Rutter 1985; Morra et al. 1997) were erupted.

The emplacement of late basaltic dykes in the Funtanazza area and the eruption of a widespread ignimbrite horizon (t2) in northern Sardinia correlates with the start of Burdigalian extensional phases identified in the Sardinian Rift (section 10.1.8). The youngest basaltic magmas record a possible increase in the subduction-component enrichment of the source area (chapter 9). Final andesites and dacites were also erupted. The majority of volcanic activity ended before the mid Burdigalian faulting phase (section 10.1.8), a factor attributed to the roll-back of the subduction zone, away from the Corsica-Sardinia microplate (section 10.3).

**11.3.3 Volcanism and uplift**

Modern volcanic eruptions are often preceded by inflation of the cone topography due to emplacement of molten magma in a chamber (Dzurisin et al. 1984). Records from Italy show that preceding a caldera eruption, an uplift of 7m occurred several kilometres from the eruptive centre (Campi Flegri, Orton 1996). Within the geological record, Watkins (1986) detects inflations of the land surface which affect lake levels preceding volcanic eruption in the East African Rift. Orton (1996) gives examples of uplift and incision prior to pyroclastic eruptions. It is therefore clear that volcanic uplift may be an important factor in controlling sub-basin subsidence patterns.
Due to the nature of the Sardinian exposures and the other controlling factors on deposition (section 11.5), the signature of volcanic uplift can only be tentatively recognised in the Castelsardo sub-basin. The Castelsardo Member succession shows an overall regression from offshore marine deposition to marginal marine fan-deltas capped by a non-marine ignimbrite (τ2, section 8.4.8). The regression which is more marked than in other areas of Sardinia (section 10.1.9) may have been enhanced by local volcanic uplift preceding ignimbrite eruption. Eruption may have been triggered by extensional tectonism. Renewed transgression occurred over the top of the faulted ignimbrite, perhaps partly as a response to land surface depression after eruption as well as eustatic sea level rise (section 10.1.9).

11.4 Nature of intra-arc basin filling

In some areas, the Sardinian depocentres which have geometries similar to those from other extensional settings (section 11.1) form a framework for a fairly predictable basin fill or ‘rift-related depositional system’ of either siliciclastic or carbonate facies when compared to published observations and models (e.g. Leeder and Gawthorpe 1987; Prosser 1993). However, many zones of the Sardinian Rift also show examples of the interaction between mixed carbonate-siliciclastic and mixed carbonate-siliciclastic-volcaniclastic sedimentation. Many areas are characterised by abrupt lateral and vertical facies variations on a scale of tens of metres. A general appreciation of these basin filling lithofacies and their architecture will be useful for the forward modelling of such basin filling sequences in other extensional terrains.

The lithofacies of the basin fill and interpreted depositional environments have been detailed in chapters 4-8 and will not be individually discussed here. Briefly, the basin fill of the Sardinian Rift comprises coarse continental clastic sediments deposited by alluvial fan-fluvial systems, lacustrine cherts, limestones and palaeosols and a variety of andesitic-rhyolitic lava flows, domes breccias and pyroclastic products. Marine sediments include clastics deposited by fan-deltas and littoral processes, shallow marine platform carbonates, deeper marine marlstones and a range of mixed carbonate-siliciclastic-volcanic sediments deposited in the littoral zone or marine shelf.

11.4.1 Nature of the basin fill and the subsidence history of the rift basin

Correlation of stratigraphic information and basin filling geometries allowed the generalised subsidence history of Sardinian the rift basin to be evaluated (section 10.1.9). Relative subsidence was created by two main phases of active tectonism separated by phases of thermal subsidence and an independently fluctuating sea level (Fig. 11.3; ‘eustatic’ sea level, Haq et al. 1988; regional, Robertson et al. 1991, Flecker 1995).
In simplified terms, the Sardinian Rift basin fill sequence shows a fining-upward, transgressive trend (Fig. 11.3). This general theme is one common to many rift basins and passive margins though such sequences often span longer timescales (e.g. North Sea Jurassic, Rattey and Hayward 1993; Miocene Gulf of Suez, Sellwood and Netherwood 1984; Jackson et al. 1988; Jurassic/Cretaceous of east Greenland, Surlyk 1984). The pattern occurs in response to initial, active tectonic subsidence and degradation of tectonic reliefs followed by thermal subsidence and the drowning of clastic source areas. In detail, the Sardinian Rift subsidence history is more complex because the relatively small amounts of multiphase extension (section 11.2.6), in addition to contemporaneous volcanism, resulted in a fairly shallow rift structure which was particularly affected by sea level fluctuations and local tectonic events.

Detailed quantitative analysis to produce subsidence curves (backstripping) could be attempted by utilising sections with good temporal constraints. There are two reasons why it was not thought worthwhile performing such an analysis here. Firstly, the exposed sections within the Sardinian Rift reflect only part of the basin’s ~18Ma history. If one correlated between sections from different sub-basins to gain a composite Sardinian section, a confusing and unrealistic analysis would result. This is because the subsidence history of each individual sub-basin is different in response to local tectonics, etc.

11.4.2 Models and features of sedimentation in extensional settings common to the Sardinian Rift

Spatial models for the variable sedimentation patterns around an active half-graben were synthesised for a number of depositional environments by Leeder and Gawthorpe (1987; Fig 11.4). Some patterns of Oligo-Miocene Sardinian basin filling show similarities to the models even though the Sardinian examples are mainly early post-rift deposits. For example, the ‘continental half-graben with interior drainage model’ (Fig 11.4) which is common to many of the east African lakes (Williams and Chapman 1986; Cohen et al. 1986), is similar to the initial continental clastic-lacustrine deposition in the Anglona study area (Casteldoria and Valledoria Members, compare Fig. 11.4 with 8.39). In this area, coarse clastic material also enters the basin via drainage from an area of intersecting fault trends, a feature common to many extensional settings (e.g. Leeder and Gawthorpe 1987, model B Fig. 11.4; Morley et al. 1990; Leeder and Jackson 1993; Gawthorpe et al. 1994; Collier and Gawthorpe 1995). Particularly noticeable is the local dispersal of clastic material in small alluvial fans sourced from the adjacent tectonic topography.

Environments of deposition proposed for the marginal-shallow marine, clastic Duidduru Member in the Sarcidano sub-basin (Figs. 5.47, 5.57, 5.67) are similar to the coastal/marine gulf half-graben of Leeder and Gawthorpe (1987; Fig. 11.4 model C). In particular, the response of fan delta lobes to
migrate to the area of maximum subsidence, whether that is in zones of cross-fault intersection (Figs. 5.47, 5.57) or along the axis of the half-graben (Fig. 5.67) is prevalent. Alluvial fans/fan deltas were supplied both as small cones from degrading fault scarps (Fig. 5.57) and from the dip-slopes of tilted fault blocks (Fig. 5.47). Within the Duidduru Member, an additional factor was the existence of tidal currents axial to the fault blocks which caused reworking of the material supplied by fan-deltas.

Carbonate platforms which formed on existing fault block highs or on fault-block dip slopes and passed basinward into deeper marine marlstones (Isili Formation, Florinas Group, Laerru Formation) show some similarities to the carbonate/coastal shelf half-graben of Leeder and Gawthorpe (1987, Fig 11.5). However, in Sardinia, wackestone and packstones were generally supplied basinward rather than coherent blocks of reef talus (Fig 11.5).

Prosser (1991, 1993) developed a spatial and temporal model for clastic half-graben sedimentation based on seismic and field observations. The model predicts a rift climax phase where tectonic subsidence outpaces sedimentation, resulting in an underfilled half-graben geometry. Sediments are supplied by voluminously small, footwall talus and hangingwall-derived sediment fans with deep marine conditions in the half-graben centre (Fig 11.6a). Though marine inundation did not accompany the Oligocene rift climax phase along the Sardinian rift, the nature of extension, sediment dispersal paths and volume of sedimentation in Prosser’s model are similar to those observed in Sardinia (e.g. Figs. 5.47, 8.39).

Prosser’s (1991, 1993) subsequent ‘immediate post rift systems tract’ records the influx of coarse clastic material into the half-graben resulting from establishment of drainage systems and degradation (Fig 11.6b). The marginal-shallow marine Duidduru Formation in the Sarcidano sub-basin (section 5.4.2 and above, Figs. 5.47, 5.57, 5.67) shows similarities to this systems tract. The ‘late post-rift systems tract’ is dominated by fine grained sedimentation which fills the remaining accommodation space and onlaps the degraded topography (Prosser 1993; Fig. 11.6c). The marlstones, carbonates and calcarenites of the Giara, Florinas Groups, Isili, Laerru Formations have similar facies characteristics and geometries. Though the sea-level variations are different, the ‘rift-related depositional systems’ model of Prosser (1991, 1993) happens to be appropriate to the non-volcanic area of the Sarcidano sub-basin which exhibits initial syn-rift basin underfilling, clastic influx after the main phase of fault movement and later marlstone filling.

11.4.2.1 Clastic sediment dispersal

Clastic sediment dispersal paths within the Sardinian Rift fill are highly variable since they reflect the response of the palaeo-transport systems towards the lowest local topography created by a variety of fault configurations. Intersecting fault zones and transfer zones characteristically channel the clastic material, in a manner similar to many other extensional systems (e.g. Leeder and Gawthorpe 1987;
Morley *et al.* 1990; Gawthorpe and Hurst 1993; Baker 1986; Cherry 1993; Collier and Gawthorpe 1995). NE-SW trending, elongate strike-slip basins surrounded by basement footwall highs directed clastic sediments basinwards to shallow marine zones on the main Sardinian Rift trend.

Siliciclastic and carbonate sediment dispersal by marine tidal currents is also prominent in the Miocene of Sardinia and southern Corsica. These deposits may have accumulated in tectonically formed tidal seaways as is found in other areas of the Miocene in southern Europe (southern France, Jones 1988; Alpine Molasse, Allen and Homewood 1984)

11.4.2.2 Mixed carbonate-siliciclastic facies

No 'typical' rift-related depositional system in the style of Leeder and Gawthorpe (1987) or Prosser (1993) exists for mixed carbonate-siliciclastic facies. General similarities between lithofacies arrangements and responses observed in other areas and the Sardinian Rift can be observed.

In the Eocene of NE Spain (Lopez-Blanco 1993) and modern day environments along the Red Sea and Gulf of Suez coarse clastic fan deltas are rimmed by carbonate reefs (Purser *et al.* 1986; Friedmann 1988). In such environments, episodic supply of coarse clastic material means that carbonate production can become established in a semi-arid climate (Purser *et al.* 1986; Friedmann 1988; Roberts and Murray 1988). Many possible examples of such a setting can be observed in the Sardinian Rift, since the carbonate platform deposits often become established over, or are intercalated with coarse clastic sediments (e.g. lower Is Paras Member, chapter 5; basal Florinas Group, chapter 7). The transition is normally gradational and occurs via colonisation by oysters and red-algae in a calcarenite or calcirudite, evolving upwards into a carbonate dominated sediment with a more diverse marine fauna (Figs. 5.19, 5.80, 5.83, 7.43, 7.44, 8.68, 8.70, 8.71). In the rock record, carbonate-clastic transitions also occur abruptly over unconformity surfaces related to sea level change (e.g. Yemen, Bosence *et al.* 1996). In Sardinia, siliciclastic-carbonate alternations within the Florinas Group may be partially unconformity bounded (chapter 7; Martini *et al.* 1992) but lack of exposure prevents full investigation.

Clastic bypass is thought to have been responsible for the establishment of some Sardinian carbonate platforms (e.g. Isili area chapter 5; Ploaghe-Ardara chapter 7). In the Gulf of Suez and Red Sea, the spatial arrangement of carbonate platforms on fault block highs and marginal half-grabens, with muddy bioclastic, silicate sands, slope aprons and mudstones in central troughs (El Haddad 1983; Purser *et al.* 1986; Burchette 1988; Fig 11.7) is similar to sedimentation patterns seen in Sardinia (e.g. Isili, Laerru Formations, Florinas Group, Figs. 5.84, 7.46, 8.67, 8.72 11.5).
Idealised models of carbonate systems (Reading 1986b) show that segregated and mixed siliciclastic-carbonate sediments sometimes occur on a carbonate shelf. Many of the Sardinian calcarenites, which occur adjacent to carbonate platforms, in deeper marine areas with some terrigeneous influx (e.g. Isili, Laerru Formations Florinas Group, Figs. 5.84, 7.46, 8.67, 8.72, 11.5), may have been deposited on a ‘carbonate’ shelf.

11.4.3 Features of volcanic and volcaniclastic facies.

In the last decade, numerous studies of modern and ancient volcanic and volcaniclastic facies have greatly increased our knowledge of processes active in such settings (e.g. Fisher and Schminke 1984; Cas and Wright 1988; Fisher and Smith 1991; Ingersoll and Busby 1995; Orton 1996). Particular emphasis has been placed on the nature of gravity-driven processes in areas of volcanic topography (e.g. mass flows, lahars, in Fisher and Smith 1991; White and Robinson 1992), the sorts and processes of pyroclastic deposits (Fisher and Schminke 1984; Cas and Wright 1988), the importance of syn- and inter-eruption deposits in areas surrounding volcanic centres (G. Smith 1991; R. Smith 1991; Buesch 1991) and the possibilities of subaqueous pyroclastic flows (Kokelaar et al. 1984; Orton 1987, 1996). Such studies were of particular use in interpreting many of the volcaniclastic and pyroclastic facies found within the basin fill of the Sardinian Rift.

The volcanic and volcaniclastic deposits of the Sardinian Rift basin fill resulted from andesitic-dacitic stratovolcanoes with abundant breccias and mass flows derived from degradation of the volcanic edifices (Logudoro, Arcentu Groups, Tergu Formation) and were intercalated with continental and shallow marine sedimentary rocks. Whilst a number of studies have examined extension, volcanism and sedimentation in continental settings (e.g. Fig 11.8; Baker 1986; Williams and Chapman 1986; Watkins 1986; Mathisen and McPherson 1991), relatively little is known about the interaction of extension, volcanism and sedimentation in shallow marine areas not dominated by clastic slope aprons (e.g. Palaeozoic of North Wales, Kokelaar et al. 1984; Orton 1987; Devonian, Australia, Stratford and Aitcheson 1996; Costa Rican beach ridges, Orton 1996). This is simply because the marine sediments are presently submerged. The Sardinian Rift provides an excellent opportunity to study such deposits. The Sardinian Rift outcrops show that in all these settings, rapid facies changes occur, in common with other volcano-tectonically active areas (Sato and Amano 1991; Orton 1996; Reading and Levell 1996).

11.4.3.1 Non-marine basin filling

The Oschiri Formation consists of intercalated basement and epiclastic conglomerates, lacustrine limestones and cherts (inter-eruption deposits sensu G. Smith 1991) with welded and unwelded ignimbrites and tuffs (syn-eruption deposits, chapters 6 and 7, Fig. 6.15). The Oschiri sub-basin contained these deposits in a ‘fault-bounded basin’ (sensu Smith and Landis 1995) whilst where
volcanism dominated to the west (Logudoro Group) the Oschiri Formation was deposited in 'volcano-bounded basins' (volcanic sags, Martini et al. 1992). A 3D facies model for the Logudoro Group and Oschiri Formation (Fig 11.8a) shows rapid facies changes and depositional environments similar to those predicted in other settings (e.g. Watkins 1986; Japan Kano 1991; Fig. 11.8b; Mathisen and McPherson 1991, Fig. 11.8c). Sardinian lacustrine limestones (Oschiri Formation, Valledoria Member) are often partly chertified suggesting that the lakes were weakly alkaline, possibly in response to contemporaneous volcanism, similar to, but not as extreme as the East African lakes (Baker 1986; Renaut et al. 1986).

11.4.3.2 Mixed siliciclastic-carbonate-volcaniclastic basin filling in a shallow marine environment

Different responses to volcanism are seen within the shallow marine Sardinian basin filling sequence. These are dependent on the volume and nature of volcanic material supplied.

The marginal marine to offshore marine Vaginella Member of the Anglona study area was totally dominated by the supply of acidic pyroclastic material, most probably from line of andesitic volcanic centres lying to the west (Fig. 8.47). Rock types include massive pumice lapilli tuffs, crystal tuffs, finely laminated tuffs, strangely-interbedded lapilli and crystal tuffs, epiclastic conglomerates and welded ignimbrites. The depositional processes are thought to have been such that subaerially supplied pyroclastic flows and surges entered a shallow marine sea, transformed into secondary turbulent flows, with reworking by secondary mass flows and additional ash-falls deposits (Fig. 8.47). The 3D facies model (Fig. 8.47) shows some similarities to the scenario proposed by Kano (1991; Fig 11.8b) for the Miocene of part of Japan where secondary surges enter a lake rather than a shallow sea. The Castelsardo Member, which lies above the volcanic-dominant succession firstly records reworking of the volcanic material by turbid and mass flows and then intercalation and mixing of tuffs/lapilli tuffs into a clastic-dominated syn-rift succession (Fig. 8.55).

Other shallow marine basin filling lithologies in which active volcanism or the erosion of volcanic reliefs played a role tend to contain mixtures of volcanic-carbonate or volcanic-siliciclastic sediments as well as discrete volcanic beds (ash falls, ignimbrites). The Duidduru Member of the Sarcidano sub-basin which was supplied both from the erosion of the pre-rift basement, eroding volcanic reliefs (Araxigi Formation) and from contemporaneous eruption, consists of sandwaves, mass flow, tidal channel and shoreface deposits similar to typical siliciclastic systems. However, the rocks can be made almost entirely of crystals, volcanic lithics, glass shards and altered glass (Fig. 5.55). Rare in-situ fossils tend to be small and infaunal.

In the Funtanazza sub-basin (chapter 4), offshore-littoral calcarenites and calcirudites were intercalated with volcanic rich calcarenites and bizarre volcanic-rich horizons (Figs. 4.3, 4.7, 4.13, 4.11). Volcanic lithic fragments within the calcarenites were derived from the weathering of the coeval
Arcentu Group, supplied basinward. Airfall and pyroclastic flows, supplied from the active Arcentu Group volcanism, mixed within the sediments on the marine shelf, are likely to have produced the beds rich in a crystal and tuff component. Again, in this setting, facies types are similar to 'normal' shelfal settings but the composition can vary. The fossil fauna within volcanic-rich beds is restricted, but is much more diverse in immediately overlying sediments. In the Funtanazza sub-basin, one can also observe contemporaneous basaltic intrusion into the shallow marine sediments and the formation of peperites and pillow lava (Fig. 4.14; Assorgia and Gimeno 1994).

In many areas, calcirudites which developed as a transgressive lag over volcanic topography contain abundant volcanic clasts and tuff-rich layers. Some fossils such as oysters and red algae (e.g. Fig. 7.29, 7.44) seem quite able to live an environment where the waters must have been regularly clouded by epiclastic and ?pyroclastic volcanic material.

In summary, andesitic fragments eroded from volcanic centres tended to be mixed into sediments as cognate clasts. Pyroclastic deposits supplied over greater areal extent were intercalated with, or mixed into crystal, glass or volcanic clast rich sediments which were affected by normal sedimentary processes.

11.4.4 Use of sequence stratigraphy

'Sequence stratigraphy is the analysis of genetically related depositional units within a chronostratigraphic framework' (Reading and Levell 1996). This approach has been inherent throughout this research. The use of a chronostratigraphic framework to a resolution of ~1Ma has enabled the complexity of contemporaneous depositional systems within the intra-arc basin to be understood.

However, the level of exposure within the Sardinian Rift means that one is not able to apply sequence stratigraphic models (sensu Exxon e.g. Haq et al. 1988; Van Wagoner et al. 1988 or Galloway 1989) and terminology in any useful way. For example, the key elements bounding stratigraphic sequences either unconformities (sensu Exxon) or maximum flooding surfaces (sensu Galloway 1989) do not obviously subdivide the Sardinian basin fill, or where they may, are not exposed. In any case, the majority of sequence stratigraphic models for which sea level fluctuations are the main control on sedimentation are not applicable for the tectonically and volcanically active Sardinian rift basin. Recent studies have concerned the application of sequence stratigraphic models to clastic systems in tectonically active extensional areas (Gawthorpe et al. 1994). However, a predictive model for tectonically active areas with mixed siliciclastic-carbonate-volcanic basin fills does not exist and is unlikely to be formulated (see discussion 11.7).
11.5 Controls on intra-arc basin filling

11.5.1 Introduction

In ‘normal’ extensional basin, tectonics, eustatic sea level change and climate are the major controls on basin filling (Leeder and Gawthorpe 1987; Frostick and Steel 1993; Prosser 1991, 1993; Gawthorpe et al. 1994; Collier and Gawthorpe 1995; Lambiase and Bosworth 1995; Fig. 11.9). The Sardinian intra-arc rift has an additional and unpredictable factor which controls basin filling - volcanism. Climatic controls are recognised as being most important in continental settings (e.g. Frostick and Reid 1993). These factors interact to control the relative sea level change, location and amount of sedimentary supply and basin fill architecture (Fig 11.9). Isolating and predicting the controls can help to rationalise the nature of sedimentary successions.

Extrusive volcanism exerts a strong control on basin filling. It affects the filling of a depocentre in several ways;

i) it results in the construction of topographic edifices with complex uplift and subsidence patterns i.e. volcanoes (e.g. Reading 1986a; Reading and Levell 1996; R. Smith 1991; Buesch 1991)
ii) the edifices are major sources of intrabasinal coarse clastic material and influence the drainage within the depocentre (e.g. Reading 1986a; Reading and Levell 1996)
iii) laterally extensive lava flows and ignimbrites often cause damming and drainage diversions within depocentre, thus creating a number of smaller sub-basins (Reading 1986a; Baker 1986; Cohen et al. 1986. They fill the tectonic relief relatively quickly and depocentres are rarely underfilled (e.g. East Africa, Gregory Rift, Williams and Chapman 1986; western rift, Baker 1986)
iv) local uplift may precede eruption (Orton 1996; Watkins 1986).

With the recent increase in our knowledge of volcaniclastic processes (11.4.3), facies models allow the nature of products away from a volcanic centre to be evaluated providing the composition of the eruptive products are known (e.g. Mathisen and McPherson 1991). However, the timing of volcanism, in particular that of voluminous pyroclastic products which may be dispersed for large areas around the volcanic centre, is largely unpredictable.

11.5.2 Observations from the Sardinian Rift

In the Sardinian rift, the main controls on basin filling can be isolated (Fig 11.9):

active extensional and strike-slip tectonics
thermal subsidence
degrading tectonic relief
eustatic sea level change
arc volcanism
sediment supply
?climate?
The influence of each of these factors in controlling the observed basin fill sequence can be assessed by careful consideration of the lithofacies types and architecture in relation to the structural development of the rift (chapter 10), 'eustatic' sea level (Haq et al. 1988) or regional sea level and climate variation curves (Calvo et al. 1993).

On a large scale, several phases of active and thermal subsidence formed the accommodation space for the deposition of the basin fill (Figs. 11.9, 11.10). The tectonically-formed depocentres consisted of a number of differently oriented and timed sub-basins of varying sizes. Relative sea level along the rift basin was a result of 'eustatic' or a least regional fluctuations superimposed on a palaeotopography created by extension, transtension and volcanic edifices and gradual, passive subsidence (Fig. 11.10). The relative sea level controlled the general types of facies which developed within the basin fill i.e. continental, marginal marine, shallow marine or offshore marine.

11.5.2.1 Continental basin filling
In continental conditions, the active and degrading topography created by faulting and volcanism controlled the dispersal paths and accumulation of clastic sediments (Villanovatulo, Casteldoria Members, Chilvani Formation). For example, areas of coarse clastic sediment supply were located close to the tectonic reliefs whereas lacustrine or volcano-lacustrine sediments accumulated away from the zone of sediment supply (Villanovatulo, Casteldoria, Valledoria Members, Oschiri Formation; Fig. 11.8a).

11.5.2.2 Role of volcanism
In areas close to volcanic centres, the erupted basalts-dacites and associated breccias dominated the basin fill succession over the time period of active volcanism (e.g. Logudoro, Arcentu Groups, Tergu Formation; Figs. 11.10, 11.11). In other areas of the Sardinian Rift, pyroclastic volcanism intermittently dominated the basin fill by covering the entire area and/or swamping and filling tectonic relief (e.g. Tergu Formation τ2 ignimbrite, Araxigi, Oschiri Formations; Fig. 11.11). The pyroclastic rocks themselves were a control on basin filling since they rapidly filled the accommodation space, prohibiting the development of any other basin fill and becoming a source for erosion of epiclastic sediments (Fig 11.11). Local uplift preceding voluminous volcanic eruptions may have occurred (Castelsardo Member, section 11.3.3) and resulted in local fluctuations in lake or sea level.
11.5.2.3 Marginal and shallow marine basin filling

In marginal and shallow marine conditions, the nature and geometry of the basin fill was controlled by the palaeobathymetry, the routes of clastic sediment supply, the amount of volcanic material supplied, the location of carbonate platforms and dispersal by marine currents (Fig 11.12). Routes of clastic sediment supply were strongly influenced by the local, often degrading tectonic configuration. For example, in the Sarcidano sub-basin, the Duidduru Member clastic sediments were supplied into the lowest depocentres created by the intersection of fault trends, in a manner similar to cross-faults in other areas (e.g. Morley et al. 1990; Cherry 1993; Gawthorpe and Hurst 1993; Fig. 11.12). Away from these focused conduits of clastic sediment supply, shallow marine carbonates developed (Isili Formation, Fig. 11.12). An additional control on the location of carbonate sedimentation here was the composition and induration of the local Mesozoic basement lithology. Basement lithology plays a role in the amount of clastic sediment provided in other localities (e.g. Greece, Leeder et al. 1991; Leeder and Jackson 1993). Shallow marine sediments in the Sarcidano sub-basin were dispersed axially to fault blocks by tidal currents (Fig. 11.12).

The interaction of active tectonism, relative sea level change and clastic sediment supply can be seen northwest of the Isili fault block in the Sarcidano sub-basin and around Castelsardo in the Castelsardo sub-basin. In Sarcidano, normal faulting controlled lateral facies variations such that carbonates and palaeosols were formed on footwall highs (Logs B and C; Fig. 5.19) and clastic/carbonate-siliciclastic rocks were deposited in hangingwall lows (Log A; Fig. 5.19). Vertical facies alternations resulted from relative sea level change and pulsed clastic sediment supply. In the Castelsardo Member, it is possible to assess the controls of active tectonism which produced divergent bed geometries and thickening across normal faults and relative sea level change which caused the development of erosion surfaces (Figs. 8.28, 8.29) plus the gradual regression observed in the lithofacies types.

The amount of volcanic material supplied had a profound influence on marginal marine-marine basin filling in some areas. In the Sarcidano sub-basin, proximity to pre-rift and volcanic source areas, the amount of coeval, erupted volcanic material and the amount of marine reworking controlled the composition of the Duidduru Member.

In the Anglona study area, the basin fill is dominated by volcanic or volcanic-derived rocks rather than clastic material supplied from the degradation of the pre-rift basement. Clastic sediments are only found immediately adjacent to pre-rift topography, within transfer zones and as syn-rift deposits. Before and after voluminous volcanism, limestones were deposited (Valledoria Member, Perfugas, Laerru Formations) but during active volcanism, the supply of volcanic material adjacent to volcanic centres 'swamped' the basin and any marine reworking (Vaginella Member). This situation occurred
because there were no major drainage paths into the basin from the Viddalba fault footwall and because of the proximity to volcanic centres.

11.5.3 The possible role of climate

Studies which have isolated climatic controls at high resolution in a shallow marine setting indicate that sediment supply and the type of sediment transport are controlled by climatic processes whilst sedimentary geometries result from the interplay of faulting and eustacy in accommodation space created by tectonics (Collier et al. 1997). Analysis at high temporal or geometric resolution is not possible in the Sardinian Rift basin.

It is clear that tectonics, sea level and volcanism interacted to produce a complex basin fill succession, but it is difficult to assess to what extent long term climate variability may have affected the Sardinian basin fill. From direct field observations, it is only apparent that the Oligo-Miocene climate was warm and semi-arid (caliche, fauna of carbonates, alluvial fan and fluvial deposits). Calvo et al. (1993) produced a relative climate curve based on the studies of the continental Neogene deposits of Spain (Fig. 10.10). The Spanish climate variability noted by Calvo et al. (1993) may well have been present over the adjacent Sardinian region. The curve shows that the Late Oligocene-mid Burdigalian were relatively moist (Fig. 10.10), consistent with the presence of clastic material supplied from degrading tectonic topography in the Sardinian Rift. The mid/late Burdigalian-Serravalian was relatively hot and dry (Fig. 10.10), consistent with the formation of carbonate platforms along the Sardinian Rift. Thus the climate signal may enhance the effects caused by tectonism and relative sea level change. In addition, since volcanic eruptions cause climatic effects (Rampino 1991), it is likely that arc-volcanism and climate change were interrelated.

11.5.4 Time-varying controls on basin filling

From the Oligocene-mid Burdigalian, active tectonism, degrading topography, volcanism and sea-level change all interacted to determine the nature of the basin fill (Fig. 11.10, 11.11). In the parts of the basin dominated by volcanic rocks, the volcanic supply controlled basin filling, but in other areas, rift structure and relative sea level were the most important controls on sedimentation, a feature common to other extensional basins (e.g. Leeder and Gawthorpe 1987; Prosser 1993; Lambiase and Bosworth 1995). Whilst tectonism had formed the template for marine deposition from the mid-Burdigalian to the Serravalian, the relative sea level was the dominant control on basin filling in the latter stages of rift evolution (e.g. Is Paras Member, Giara and Florinas Groups).
11.5.5 The resultant intra-arc basin complexity

In the Sardinian Rift, complex spatial and temporal variations in Oligo-Miocene lithofacies occur on the sub-kilometre scale within a vertically and laterally variable basin fill succession. The complexity of the basin fill results from the interaction of many factors. In particular, arc-volcanism played a major role in basin filling by creating substantial volcanic edifices, blanketing tectonic topography and becoming mixed or intercalated with sediments of the basin fill. From detailed studies, the contribution of each control at a particular time and place can be evaluated.

However, in such a region of multiphase extension/transtension with unpredictable active volcanism it would be difficult to predict some of the complex arrangements from, for example, a well or 2D seismic line. Whilst this study presents general themes about the behaviour of basin filling in intra-arc basins, it is difficult to comprehend that any predictive model would be able to begin to rationalise such a unique setting.

11.6 Implications for hydrocarbon prospectivity

Due to a lack of a mature source rock, there does not appear to be any likelihood of finding hydrocarbons suitable for exploration within the Sardinian Rift itself. However, the field observations made do have potential implications for petroleum provinces in general because they give insights into reservoir distribution, tectonic styles and hence structural and stratigraphic trapping potential in intra-arc settings.

There are a number of different possibilities for analogies to reservoir rocks within the Oligo-Miocene succession. Most obvious are the continental and shallow marine conglomerates and sandstones of the Nureci Formation, Sarcidano sub-basin. These clastic sedimentary rocks form thick accumulations in the hangingwalls of intersecting fault trends, similar to the control of transfer faults or zones in other extensional settings (e.g. Morley et al. 1990; Gawthorpe and Hurst 1993; Cherry 1993). The scale of fault blocks and clastic deposits are similar to those seen in a block of the North Sea or the Gulf of Suez. However, the rocks porosity and permeability characteristics, which are already fairly poor (Fig. 5.40), would probably decrease even further if buried to the necessary depth for hydrocarbon maturation. Pore space is occluded by clay minerals in areas of the 'cleanest' basement derived sediments (Fig. 5.40) and almost totally occluded by altered volcanic glass in areas close to volcanic centres (Fig. 5.55). Mathisen and McPherson (1991) suggest that the poroperm characteristics of such rocks may be enhanced by grain dissolution on burial. Similar pore filling of volcanic derivation or by lime mud occur within some parts of the Florinas Group and Castelsardo Member clastic sediments. However, other poorly consolidated beds within these units have excellent poroperm characteristics.
Secondly, Miocene shallow marine carbonates sometimes show intra-granular or inter-granular porosity, though lime mud, microspar and spar fills the majority of pore space (Isili Formation, Florinas Group, Laerru Formation, Fig. 3.4). If the reservoir quality of carbonates were enhanced, for example by fracturing, they would form excellent reservoir rocks which were located on structural highs within the basin (fault-blocks). Other oil fields situated in volcanic provinces have reservoirs within fractured volcanic rocks (Nevada, Japan, Georgia, BP field guide). If some post-depositional tectonic event were to cause lithification and fracturing of the Sardinian volcanics, they would form an extensive reservoir rock.

Possible structural traps are caused by normal faulting and drape anticlines covered by marlstones (e.g. Fig. 5.1). Stratigraphic traps might include the onlap and pinch-out of clastic sediments onto degraded fault planes and lateral pinch out into marlstones (e.g. Fig. 5.1). Other stratigraphic traps might include the pinchout of shallow marine carbonates on structural highs into basinward marlstones (Figs. 5.1, 7.1).

Particular additional risks to exploration exists in intra-arc settings such as the rapid lateral and vertical changes which are apparently characteristic (this study; Smith and Landis 1995; Orton 1996), the complex structural styles resulting from multiphase extension, the intercalation of volcanic material and the unpredictable effect of magmatism on a basin's thermal history and maturation.

11.7 Discussion

The Sardinian Rift intra-arc basin shows similarities to other volcanic and non-volcanic extensional settings. Field observations have been used to describe and help rationalise the structural development of an intra-arc basin in relation to back-arc basin formation, and the nature of and controls on a mixed siliciclastic-carbonate-volcanic basin fill succession. General points and simple models can be made from different aspects of the observations. However, the Oligo-Miocene Sardinian Rift history is unique and no model could easily predict the succession of events which took place. The uniqueness of the basin comes down to several factors, the use of inherited versus new structural trends, a changing overall stress field resulting in multiphase extension and along with the location and timing of arc-volcanism.
11.8 Summary

- Structural elements of the Sardinian Rift basin comprise planar normal faults, tilted fault blocks and half-graben and transfer zones which show similarities to structures commonly observed in other extensional settings. However, relationships between some fault sets are not clear and the overall structure of the rift basin is complex, resulting in many independent or semi-independent depocentres. Sinistral strike-slip faults in the eastern Sardinian basement formed transpressional duplexes at restraining bend and transtensional basins at releasing bends with geometries similar to those predicted for idealised models. Inherited basement structures were important in determining the Oligo-Miocene rift development.

- Syn-rift deposits within the Sardinian intra-arc basin fill are voluminously minor in contrast to other extensional settings. The majority of sedimentary and volcanic rocks passively infill the topography created by extensional or transtensional tectonism and thermal subsidence (post-rift deposits). Along with temporal constraints, geometries and the response of the basin fill suggest that active extension occurred for relatively short time periods (<few Ma) where tectonic subsidence far exceeded the rates of sediment supply. Several phases of such extension occurred on differently oriented structures and faulting youngs into the basin centre.

- There were links between the location of andesitic volcanic centres and extensional faults suggesting that extensional faulting facilitated initial magma emplacement. Volcanism commenced with rift formation and continued through a phase of basin infilling. Volcanism may have caused local uplift which affected sub-basin filling.

- The basin fill of the Sardinian rift exhibits rapid lateral and vertical facies variations in response to the interaction of passive and active tectonism, eustatic sea level change and arc-volcanism. The thickness, dispersal and accumulation of sediments were largely controlled by the rift structure and the presence of volcanic centres, whilst the types of facies which developed were controlled by relative sea level.

- The basin filling commenced with continental clastic and lacustrine deposition contemporaneous with andesitic volcanism. Mixed shallow marine sedimentation and andesitic-rhyolitic volcanism followed. In the final stages of basin filling, marine carbonate-marlstone sedimentation dominated.

- Whilst some parts of the basin fill exhibit predictable facies arrangements and geometries, the presence of volcanism within the intra-arc basin and the complex, multiphase structure result in a rapidly varying and unique basin fill succession.
Chapter 12
Chapter 12 - Conclusions and suggestions for future work

12.1 Conclusions

The Sardinian Rift is an intra-arc basin which formed in response to multiphase extension and transtension on several orientations of normal and strike-slip faults. The variable rift structure resulted in the formation of many semi-independent depocentres or sub-basins. The sub-basins were filled with continental to shallow marine siliciclastic sediments, lacustrine sediments, andesitic-rhyolitic volcanic rocks, shallow marine platform carbonates and deeper marine marlstones. Complex lateral and vertical facies variations within the basin fill resulted from the interaction of active and passive tectonism, eustatic sea level change and arc-volcanism.

The main focus of this research has been to utilise detailed field observations placed within a new chronostratigraphic framework to determine a tectono-stratigraphic synthesis of the complex intra-arc basin. A geochemical analysis of the volcanic-arc rocks has investigated links between the observed extension and possible variations in the source of the magmas. This study, which has produced a high resolution evolution of the Sardinian intra-arc rift, has implications both for the regional tectonic development of the Western Mediterranean and for the development of and processes active in intra-arc, back-arc and extensional settings in general.

Chapters 3-9 formed a detailed description of the individual study areas and the geochemical analysis. These observations are integrated into the main conclusions of this research which are summarised below.

12.1.1 Tectono-stratigraphic development of the Oligo-Miocene Sardinian intra-arc rift.

Extension of the Sardinian continental crust on high angle, planar normal faults commenced sometime in the mid-late Oligocene, resulting in a complex proto-rift structure with significant along-strike variability. The main −N-S proto-rift trend in northern Sardinia was interrupted by NE-SW oriented, elongate strike-slip basins formed on the releasing bends of sinistral strike-slip faults which crossed the eastern Sardinian basement. The strike-slip faults may have moved before and coeval with the −N-S rift forming faults. In south-central Sardinia, the proto-rift was oriented NW-SE, most probably as a result of the influence of Hercynian structural trends. The rift margin consisted of a line of tilted normal fault blocks cut by NE-SW trending cross-faults and half-grabens. Small E-W trending graben off the main Sardinian Rift in southernmost Sardinia are considered to be reactivated late Eocene sub-basins.
Basin filling in the mid-late Oligocene commenced with the eruption of basalts and andesites within
the Sardinian proto-rift and the deposition of coarse continental clastics transported by small alluvial
fans off the newly formed tectonic reliefs. The locally supplied clastic sediments passed laterally into
finely laminated lacustrine limestones, cherts and tuffs deposited in half-graben or sub-basin centres.
Geometries within the basin fill imply the first mid or late Oligocene rifting phase lasted for less than a
few million years, that tectonic subsidence far exceeded sedimentation rates and that the proto-rift was
significantly underfilled.

From the latest Oligocene until the early-mid Burdigalian, the dominant signature along the Sardinian
Rift was one of filling of the proto-rift topography. However, localised extension on ~E-W trending
normal faults resulted in impressive late Aquitanian-early Burdigalian syn-rift geometries in the
regressive clastic sediments of the Castelsardo area of northern Sardinia. A final, late Aquitanian phase
of movement on the sinistral strike-slip faults crossing the eastern Sardinian basement was recorded
within the siliciclastic-volcanic-lacustrine transtensional sub-basin fill.

Voluminous andesitic lava domes, flows and breccias plus dacitic and rhyolitic pyroclastic flows,
surges and falls dominated the latest Oligocene - early/mid Burdigalian proto-rift fill in the vicinity of
volcanic centres. Away from the volcanic edifices, a marine transgression covered lower parts of the
proto-rift. A marine volcano-sedimentary succession filled the Castelsardo-sub basin adjacent to
andesitic volcanic centres. At the eastern rift margin in central Sardinia, marginal and shallow marine
siliciclastic-epiclastic sediments eroded from fault-bounded basement and volcanic reliefs were
transported by fan-deltas into the lowest palaeotopography created by intersecting fault trends. The
conglomerates and sandstones were reworked and dispersed by marine currents axial to the tilted fault
blocks and passed laterally basinwards into deeper marine marlstones.

In the early-mid Burdigalian, a second phase of extension occurred along the length of the Sardinian
Rift, after arc-volcanism had largely ended. Normal faulting of the underlying volcanic succession is
particularly prominent in northern Sardinia where ~E-W and ~N-S trending fault sets may have moved
at the same time. Again, no major associated syn-rift deposits are found, suggesting a short pulse of
extension where tectonic subsidence far exceeded rates of basin filling. In southern Sardinia, dyke
emplacement seems to have accompanied this extension phase and NNE-SSW trending faults were
active.

In the mid-late Burdigalian, renewed marine transgression proceeded over both the mid-Burdigalian
fault-scarps and the degraded Oligocene fault topography. Mixed carbonate-siliciclastic-marlstone
deposition with abrupt vertical and lateral facies changes formed in response to the localised supply of
clastic material and the fault-created palaeobathymetry. Clastics continued to be supplied into the
hangingwalls of intersecting fault trends and along the elongate transtensional sub-basins. Shallow
marine carbonate platforms developed on fault-block highs whilst calcarenites and marlstones were deposited in half-graben or sub-basin centres.

From the late Burdigalian-mid Serravalian, two trangressive-regressive cycles can be identified within the carbonate-siliciclastic-marlstone sedimentary succession which largely infilled the earlier formed tectonic relief. Localised ~N-S trending normal faulting affected parts of the basin fill from time to time.

The generalised subsidence history of the Sardinian Rift therefore reflects two short-lived phases of active extension and tectonic subsidence which were separated by longer periods of passive, possibly 'thermal' subsidence and basin filling. Superimposed on the tectonic events were eustatic, or at least regional, sea level fluctuations. Volcanic uplift may have affected basin subsidence locally.

12.1.2 Structural styles and timing of extension in the Sardinian Rift

The vast majority of structures which accommodated extension were high angle, planar normal faults on a variety of different orientations. High-angle strike-slip faults formed transpressional and transtensional duplexes at restraining and releasing bends respectively. Normal and strike-slip faults defined simple graben, half-graben linked by relay ramps and cross-faults and typical transtensional basin geometries. However, an overall unique set of depocentres and sub-basins were formed by the superposition of these individual structures. The re-use of inherited Hercynian and late Hercynian structures in the Sardinian basement seems to have caused much of the rift complexity.

Basin filling geometries generally indicate that several 'pulsed' periods of short-lived active extension and transtension were responsible for the formation of the Sardinian Rift. Active extension and strike-slip fault movement occurred before back-arc basin opening. Active extension concentrated in the centre of the rift structure occurred after back-arc basin opening. Whilst the majority of the basin filling sequence may be considered a syn-rift megasequence (sensu Prosser 1991, 1993), syn-rift deposits are rare and voluminously minor, with the majority of the basin fill covering an existing topography (i.e. post-rift deposition). Within the temporal and spatial resolution of this study, the rift and faults within it do not appear to propagate and grow with time. Faults within a particular population appear to have moved at the same time. The amount of extension in the Sardinian Rift is small (β~1.05) as shown by the high-angle faults and maximum thickness of ~2km of sediments.
12.1.3 Arc volcanism and extension

Spatial relationships between the location of andesitic volcanic centres and large normal faults implies that these upper crustal discontinuities facilitated magma emplacement. Arc-volcanism commenced at the same time as mid-late Oligocene rifting and was most voluminous over the period of back-arc basin opening and passive subsidence in Sardinia. Volcanism had largely ended before the mid-Burdigalian phase of extension, which was associated with late dyke emplacement. Geochemical analysis of the most primary Sardinian volcanic rocks suggests that the majority of the systematic variations in the major and trace element ratios, plus some incompatible trace element ratios are consistent with fractional crystallisation of an olivine-clinoptyroxene-orthopyroxene-titanomagnetite assemblage as postulated by Rutter (1985). The Nb/Zr ratio was constant, within error, throughout the period of Sardinian rifting events, back-arc basin opening and along the length of the rift. Thus, there was no detectable change in either the primary mantle wedge source-enrichment or degree of partial melting. It does not appear that active lithospheric thinning and associated, increased asthenospheric upwelling accompanied the extensional events observed in the Sardinian Rift to a detectable extent. Subtle and tentative systematic changes in fractionation-independent mobile/immobile trace element ratios suggest that the subduction-component may have varied independently of a constant source. Variations from the most complete temporal succession in southern Sardinian suggest that the subduction-flux may have decreased over the period of back-arc basin opening relative to the first and last extension-related volcanism. Spatial variations within rocks of the same age indicate that magmas erupted in northern Sardinia may have sampled a source richer in a subduction component than those in the south.

12.1.4 Nature of and controls on basin filling in the Sardinian Rift

The mixed volcanic-siliciclastic-carbonate fill of the Sardinian rift exhibits complex lithofacies architecture in response to the interaction of active extension and transtension, existing tectonic topography, sea level change, clastic supply and the nature and location of arc-volcanism. The relative sea level controlled the broad types of basin filling lithofacies. The evolving rift structure and active volcanism controlled the provenance, supply and accumulation of the basin fill. Arc-volcanism played a major role in basin filling by creating substantial volcanic edifices, blanketing tectonic topography and becoming mixed or intercalated with sediments of the basin fill. Andesitic fragments were mixed as cognate clasts whilst pyroclastic flows supplied over greater areal extent were mixed into the sediments. Individual systems associated with typical fault-block and half-graben geometries have characteristics similar to published models (e.g. Leeder and Gawthorpe 1987, Prosser 1993). The superposition of mixed basin filling systems in various sub-basin morphologies provides insights into the behaviour of these more complex zones within in an intra-arc basin.
12.1.5 Implications for Western Mediterranean tectonics

A combination of the Sardinian Rift tectono-stratigraphic development with published data on the Western Mediterranean has allowed a regional model to be proposed. In the mid-late Oligocene, the Sardinia-Corsica microplate was in a crucial position between the northern Apennines compressional zone and the extensional systems at the southern margins of the Eurasian plate. The microplate records a phase of tectonic escape from the northern Apennines towards the extensional region, rather like the 'indentor' type model of continental deformation. Tectonic escape occurred on large sinistral strike-slip faults which crossed the eastern Sardinian and southern Corsican basement. In contrast, extension occurred in a variety of orientations in western Sardinia. The extension in the south Eurasian plate is thought to have been driven by subduction zone steepening and roll-back. Thus the mid-late Oligocene in Sardinia was a time of paired tectonic escape and extension. After initial Sardinian rifting, extension appears to have been focused by the opening of the Western Mediterranean back-arc basin to the west of Sardinia in the Aquitanian-early Burdigalian. Tectonic escape ended just before back-arc spreading as the compressional fronts migrated outwards (e.g. Carmignani et al. 1994) Back-arc basin opening and Sardinia-Corsica separation and rotation may have been driven by continued subduction zone roll-back. After back-arc basin spreading ended, Mid Burdigalian extension was again focused on Sardinia, possibly in response to the continued roll-back of the subduction zone, which also resulted in the arcuate outwards migration of compressional zones.

12.2 Suggestions for future work

Many of the initial aims of this research have been satisfied. Some questions still remain unanswered (e.g. the role of pre-rift topography in controlling the early basin fill) but this is largely because the amount and style of outcrops do not allow such considerations to be evaluated. This section suggests how hypothesis formed in this study could be tested or better constrained and what additional research may prove fruitful.

12.2.1 Detailed field mapping

Outcrop by outcrop scale mapping of some features of the Logudoro study area and strike-slip systems of Eastern Sardinia may facilitate a clearer understanding of the tectono-stratigraphic development of these areas. However, it is felt that the majority of crucial outcrops were examined in this study.

Detailed work on the Miocene basins of Corsica and the timing of strike-slip fault movement would be useful in further constraining the regional tectonic evolution. W. Cavazza from Bologna University has students working in these basins.
12.2.2 Improved temporal constraints

More biostratigraphic, radiometric and Sr-isotope dates would obviously allow better temporal correlation and constraints of basin forming and filling events. Further testing and improvements of the Sr-isotope technique would allow more emphasis to be placed on the results to the high precision of the analytical resolution (often ±0.3 Ma, section 3.2.2). Single crystal Ar-Ar step-heating experiments on plagioclases would allow the cause of discrepancies between biotite and plagioclases to be deduced (section 3.2.3). Work in progress is active on the latter two suggestions.

12.2.3 Further basin analysis

Quantitative subsidence analysis could be performed on some well dated sections, β factors and strain rates calculated. However, the problems and constraints outlined in section 11.4.1 must be appreciated. Seismic reflection profiles would be of great use in determining the complex structure of the Sardinian Rift at depth.

12.2.4 Rationalisation of the sources and evolution of the volcanic-arc rocks

The knowledge of subduction zone processes and arc-magma petrogenesis has increased greatly since the seminal works of Dupuy et al. 1974; Dostal et al. 1976; Coulon 1977 etc. on the Sardinian volcanic rocks. Modern criteria and analytical methods could be applied to determine the formation and evolution of these magmas. Detailed geochemical and isotopic work with links to Sardinian Rift formation is being performed by F. Secchi (Università di Cagliari) and V. Morra (Università di Napoli, see Morra et al. 1994, 1997).

12.2.5 Late Miocene-present day geology

A regional scale study of the late Miocene-present day geological history of Sardinia would provide further insights into the history of a remnant arc and stranded continental fragment which underwent renewed extension contemporaneous with the formation of a second back-arc basin - the Tyrrenian Sea.

An active research group at the Università di Cagliari headed by Profs. A. Assorgia, S. Barca, C. Spano and A. Porcu continues detailed research into the development of the Sardinian Rift.
References
References


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BP Field work to examine syn-rift sequences in Sardinia - notes and map. Unpublished field guide.


A240


Foglio 192, Alghero, 1955
Foglio 226, Mandas, 1959
Foglio 193, Bonorva, 1959
Foglio 297, Nuoro, 1976
Foglio 194, Ozeiri, 1962
Foglio 180, Sassari, 1953
Foglio 224-225, Guspini, 1962


Sparks, R.S.J., Gardeweg, M.C., Calder, E.S. and Matthews S.J. in press. Erosion by pyroclastic flows on Lascar Volcano, Chile. *Bulletin of Volcanology.*


Appendices
Appendix 1A - Field Techniques

Geological Maps
Geological mapping was the basic technique used to record much of the data collected. Whilst the most complicated areas were mapped by examination of every outcrop, other zones were mapped at a 'reconnaissance' scale. In areas of poor or no exposure, vegetation and topographic changes gave hints on the underlying geology. Thus some parts of the Enclosures 1-5 which illustrate this dataset, modified after published maps (e.g. Servizio Geologico, Carta Geologica), are inferred.

Palaeocurrents
Palaeocurrents were measured using a variety of paleoflow indicators:

a) clast on clast imbrication - measurement of aligned clast axes where 2 or more clasts are aligned on top of one another (Collinson and Thompson 1982). 180° was added to the dip-direction measurement to a give paleoflow direction (after Collinson and Thompson 1982, listed in Appendices 5A, 8A). The ideal situation would be to have measurements from a statistically relevant number of readings. This was not always possible since conglomerates are generally very poorly imbricated, partly because pebbles are often quite spherical. The measurements taken represent all the possible measurements that could be taken from a particular bed or outcrop.

b) cross bedding - palaeocurrent flow implied perpendicular to cross beds in down dip direction

c) channels - where the 3D channel orientation is observable, palaeoflow parallel to the channel axis.

Clast counting for size and composition
A representative part of the outcrop was picked and the composition and/or size of the longest axis of all clasts greater than 1cm in size was noted. For statistical relevance, as many clasts as possible (up to ~150) were counted, though in some cases the number was limited by outcrop size.

Techniques used to define rift structures and kinematics
Fault, fracture and bedding planes were measured such that the dip direction was always to the right of the strike direction. They are represented on lower hemisphere stereoplots as planes and poles to planes. Slickensides were measured as an angle on a plane from the maximum dip-direction of the fault plane and are represented on stereoplots by arrows.

Field relationships demonstrate the importance of fault scarp degradation from the Oligo-Miocene period to the present day and the burial of fault planes beneath younger sediments, making the direct measurement of 'large' fault planes (throws 10's -100's m) impossible. In addition, younger episodes of tectonism are superimposed on the Oligo-Miocene rift. Thus a variety of techniques must be used to identify Oligo-Miocene structural geometries and kinematics.
The nature of large fault planes has been deduced from the present day outcrop topography of pre-rift rocks, for example the degraded fault scarps of tilted fault blocks (Chp. 5). From such observations one can deduce the strike of the faults and get some idea of the minimum fault offset and dip. The Oligo-Miocene palaeotopography can be further reconstructed by examining pre-rift outcrop geometries under or adjacent to the basin fill. The tilting and/or offset of originally flat-lying Mesozoic platform carbonates deposited on a peneplaned surface (Cocozza and Jacobacci 1975, e.g. Villanovatulo area, Sarcidano sub-basin) or of older basin fill rocks (e.g. volcanic rocks, Logudoro study area) may verify fault throws and the amount of fault block rotation.

Measurement of smaller, outcrop-scale faults (cm to 10'sm) and other tectonically related structures is possible in many places. The faults may be intra-basement (e.g. spaced brittle structures through Hercynian ductile deformation or metre-scale, brecciated fault zones in Mesozoic dolomites), intra-basin fill (e.g. spaced normal faults and fractures) or between basement and basin fill. Great care must be taken to distinguish pre-, syn- and post- Oligo-Miocene faults and deformation.

Scale invariance relationships amongst populations of faults imply that in simple systems of faults (away from transfer zones) the fault displacement-length relationship, fault strike and kinematics of smaller structures behave in the same way as larger ones (e.g. Marrett and Allemendinger 1990; Cowie and Scholz 1992; N. Dawyers pers. comm. 1997). In a general sense, this is true in Sardinia, such that locally, the nature of measured smaller faults parallels that of the larger, degraded basin forming structures. Thus it is a viable approach to treat the measured geometries and kinematics on small faults as indicative of the history of larger faults. However, it is important to note that this technique is only suitable over small areas where the relationship between small and large faults is visible. Because of the complexity of the Sardinian rift it is not appropriate to use very few measurements of small faults from one or a few scattered locations to define a history of extension (or compression) as proposed in Cherchi and Montadert (1982a,b) and Cherchi and Tremolieres (1984). The results from measuring small faults generally consisted of 1-3 values of the individual fault plane (illustrated on the Enclosures) or more than 6 measurements of spaced fault planes and fractures with rare slickenside data (illustrated on lower hemisphere, equal area stereoplots, planes and poles to planes plotted). All possible measurements were taken from each outcrop. Measurements of faults and fractures, particularly within pre-rift basement rocks often gave a wide, almost random scatter of data. This data has not been included in the thesis.
Appendix 1B - Definitions

This appendix lists the meaning taken here of terms often used ambiguously and explains the abbreviations used.

Syn-rift deposits were deposited contemporaneously with normal or transtensional faulting. They are recognised by individual bed thickening and progressive rotation into a hangingwall depocentre, thickening across an active fault or are bounded by angular unconformities caused by syn-sedimentary rotation (sensu Prosser 1993, Fig. 1B.1).

Post-rift deposits were deposited after normal or transtensional faulting had occurred. They are recognised by their lack of angular unconformities and are parallel bedded, passively infilling of topography, often with strong onlap patterns. Note, however, that this may still result in post-rift deposits having an overall wedge shape (sensu Prosser 1993, Fig. 1B.1).

A syn-rift megasequence is defined as consisting of the rocks deposited over the period that extension occurred on any active fault throughout the sub-basin. Thus periods of post-rift deposition away from area of the active faulting can occur within the syn-rift megasequence (sensu Prosser 1993).

Particular emphasis was placed on the recognition of syn- and post- rift deposits in conjunction with structural measurements. The technique was used so that the timing of differently oriented phases of extension and transtension could be identified.

Proximal deposition is a term used to here to describe sedimentary or volcanic rocks which were deposited close to their source area. Distal deposition describes sedimentary or volcanic rocks which were deposited some distance from the source area, in many cases this is in the ‘basinward’ direction.

The matrix of a sedimentary or volcanic rock is a term used to describe the population of smaller material inbetween much larger clasts. In very poorly sorted breccias and conglomerates which contain clasts from boulder to clay size, matrix is taken to be clasts <1cm in diameter. Clast supported is used here for conglomerates, breccias and gravels where >50% of clasts are in contact with one another. Matrix supported is a term used where >50% of the population of larger clasts are not in contact with one another.

Ignimbrite is a term used for the deposits of pumaceous pyroclastic flows. These sometimes exhibit welded, fiamme and elongate vesicle textures but can also consist of massive and stratified pumice flows and block and ash flows. Pyroclastic deposits are those resulting directly from primary volcanic flows, surges or fallout whereas epiclastic deposits are derived from the erosion and reworking of volcanic rocks. Volcaniclastic is a general term used for pyroclastic and epiclastic deposits.
Calcarenite and calcirudite are terms used for sediments with non-carbonate clasts (i.e. pre-rift basement or volcanic or siliciclastic sediment) within a carbonate matrix. ‘Gravel’ or gravelstone is a term used for sediments with clast of sizes >2mm and <4cm (i.e. intermediate between a coarse sandstone and conglomerate).

A sandwave is a flow transverse bedform coupled to oscillatory currents of tidal origin (Allen 1980).

pl=plagioclase, cpx=clinopyroxene, opx=orthopyroxene, hbl=hornblende, Ti-mt=titanomagnetite, bt=biotite, qz-=quartz.

The symbol – is used as an abbreviation for ‘approximately’. For example, when discussing a fault set which may trend from 130-150°, the abbreviation –NW-SE is used.

Italian Words
marne= marl, calcare= carbonate, arenarie and sabbie= sand, lacustre= lacustrine, di= of, Formazione=Formation
IMMEDIATE POST-RIFT:
Discontinuous parallel reflectors, with possible progradational and aggradational reflectors close to the footwall. Compaction syncline over the basement footwall cut-off point.

LATE POST-RIFT:
Continuous parallel reflectors, less compaction induced deformation. Strong onlap and burial.

RIFT INITIATION:
Perfect wedge shapes to reflector packages, minor onlap onto the hangingwall, discontinuous hummocky internally. Possible progradation (real or apparent), no evidence of important footwall derived sediments.

RIFT CLIMAX:
Chaotic zone close to the footwall scarp, aggradation and downlap if resolution is good enough. Divergence of basinal equivalents, Lozenge shapes or low angle downlaps on the hangingwall dip-slope if preserved. Minor onlap at top of hangingwall slope.

1B.1 Sketch of syn- and post-rift geometries within a half-graben from Prosser (1993).

1B.1 Syn-rift deposition contemporaneous with half-graben bounding normal faulting.

1B.1 Post-rift infill of half-graben topography.
Appendix 3A - Nannofossil Zonations

Results of sample analysis by E. Gervais, J & G Consultants. Refer to Table 3.2 for summary and sample locations. The global zones are after Martini (1971).

Sample ASB40
Zone: upper part NN2 to NN3 (except uppermost part)
Stratigraphic remarks: The presence of Helicosphaera ampliaperta, whereas Sphenolithus heteromorphus is absent indicate that this sample belongs to the upper part of zone NN2 or to zone NN3 (except the uppermost part of zone NN3, where Sphenolithus belemnos and Sphenolithus heteromorphus overlap according to Perch-Nielsen, 1985). Unfortunately Sphenolithus belemnos, which is rarely found in the Mediterranean area, was not found in all the samples studied.
Preservation: moderate

<table>
<thead>
<tr>
<th>Coccolithus miopelagicus</th>
<th>common</th>
<th>in situ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coccolithus pelagicus</td>
<td>abundant</td>
<td>in situ</td>
</tr>
<tr>
<td>Coronocyclus nitescens</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Cyclicargolithus abisectus</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Cyclicargolithus floridanus</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster deflandrei</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Hayella aperta</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Helicosphaera ampliaperta</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Helicosphaera carteri</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Micula decussata</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Pontosphaera multipora</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Reticulofenestra dictyoda</td>
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<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra spp (medium)</td>
<td>abundant</td>
<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra spp (small)</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Sphenolithus conicus</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Sphenolithus moriformis</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Thoracosphaera spp</td>
<td>rare</td>
<td>in situ</td>
</tr>
</tbody>
</table>

Sample ASB53
Zone: uppermost part NN3 to NN4
Stratigraphic remarks: The co-occurrence of Helicosphaera ampliaperta and Sphenolithus heteromorphus indicates that this sample belongs to the uppermost part of zone NN3 to zone NN4.
Preservation: good

<table>
<thead>
<tr>
<th>Coswolithus miopelagicus</th>
<th>common</th>
<th>in situ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coccolithus pelagicus</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Coronocyclus nitescens</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Cyclicargolithus abisectus</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Cyclicargolithus floridanus</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster deflandrei</td>
<td>rare</td>
<td>in situ</td>
</tr>
</tbody>
</table>
**Sample ASB58**
barren

**Sample ASB71**
Zone: uppermost part NN3 to NN4
Stratigraphic remarks: Co-occurrence *H. ampliaperta* and *S. heteromorphus*.
Preservation: good

<table>
<thead>
<tr>
<th>Coccolithus miopelagicus</th>
<th>common</th>
<th>in situ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coccolithus pelagicus</td>
<td>abundant</td>
<td>in situ</td>
</tr>
<tr>
<td>Cyclicargolithus floridanus</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster adamanteus</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Discoaster deflandrei</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster signus</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Effellithus spp</td>
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<td>in situ</td>
</tr>
<tr>
<td>Hayella aperta</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Helicosphaera ampliaperta</td>
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<td>in situ</td>
</tr>
<tr>
<td>Helicosphaera carteri</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Helicosphaera palaeocarteri</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Prediscosphaera cretacea</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Rectapontis compactus</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reticulofenestra spp (medium)</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra spp (small)</td>
<td>abundant</td>
<td>in situ</td>
</tr>
<tr>
<td>Sphenolithus heteromorphus</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Sphenolithus moriformis</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Thoracosphaera spp</td>
<td>few</td>
<td>in situ</td>
</tr>
</tbody>
</table>
Sample ASB88  
Zone: NN1 to NN2  
Stratigraphic remarks: The presence of Triquetrorhabdulus carinatus and Helisphaera carteri assigns the sample to NN1 to NN2 zones.  
Preservation: moderate

<table>
<thead>
<tr>
<th>Coccolithus miopelagicus</th>
<th>few</th>
<th>in situ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coccolithus pelagicus</td>
<td>abundant</td>
<td>in situ</td>
</tr>
<tr>
<td>Cyclicargolithus abisectus</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Cyclicargolithus floridanus</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster adamanteus</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster deflandrei</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Hayella aperta</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Helicosphaera carteri</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Helicosphaera euphratis</td>
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<td>in situ</td>
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<tr>
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</tr>
<tr>
<td>Pontosphaera multipora</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra spp (medium)</td>
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<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra spp (small)</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Sphenolithus conicus</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Sphenolithus moriformis</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Triquetrorhabdulus carinatus</td>
<td>rare</td>
<td>in situ</td>
</tr>
</tbody>
</table>

Sample ASB90  
Zone: NN1 to NN2  
Stratigraphic remarks: The presence of Helicosphaera carteri, whereas Helicosphaera ampliaperta is absent, may indicate that this sample belongs to the NN1 to NN2 zones. Triquetrorhabdulus carinatus, known to be very rare in the Mediterranean area, was not found in this sample.  
Preservation: moderate

| Chiastozygus litterarius          | rare | reworked |
| Coccolithus miopelagicus         | few  | in situ  |
| Coccolithus pelagicus            | abundant | in situ |
| Cyclicargolithus floridanus      | common | in situ |
| Discoaster adamanteus           | few  | in situ  |
| Discoaster deflandrei            | few  | in situ  |
| Helicosphaera carteri            | few  | in situ  |
| Helicosphaera palaeocarteri      | rare | in situ  |
| Helicosphaera scissura           | rare | in situ  |
| Micula decussata                 | rare | reworked |
| Pontosphaera multipora           | few  | in situ  |
| Pontosphaera ribosa              | few  | in situ  |
| Reticulofenestra spp (medium)    | abundant | in situ |
| Reticulofenestra spp (small)     | abundant | in situ |
| Sphenolithus conicus             | few  | in situ  |
| Thoracosphaera spp               | rare | in situ  |
Sample ASC24
Zone: NN1 to NN2?
Stratigraphic remarks: The few nannofossils are not diagnostic.
Preservation: very poor

<table>
<thead>
<tr>
<th>Nannofossil Type</th>
<th>Abundance</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coccolithus miopelagicus</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Coccolithus pelagicus</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster deflandrei</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster spp</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra spp (medium)</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Triquetorhabdulus carinatus</td>
<td>few</td>
<td>questionable</td>
</tr>
</tbody>
</table>

Sample AS54
Zone: uppermost part NN3 to NN4
Stratigraphic remarks:
Co-occurrence of H. ampliaperta and S. heteromorphus.
Preservation: good

<table>
<thead>
<tr>
<th>Nannofossil Type</th>
<th>Abundance</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coccolithus miopelagicus</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Coccolithus pelagicus</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Cribrosphaerella ehrenbergii</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Cyclicargolithus abisectus</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Cyclicargolithus floridanus</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster cf Discoaster exilis</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Discoaster deflandrei</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Hayella aperta</td>
<td>few</td>
<td>in situ</td>
</tr>
<tr>
<td>Helicosphaera ampliaperta</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Helicosphaera carteri</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Helicosphaera palaeocarteri</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Micula decussata</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Prediscosphaera cretacea</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Reticulofenestra dicyoda</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Reticulofenestra spp (medium)</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra spp (small)</td>
<td>abundant</td>
<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra umbilicus</td>
<td>rare</td>
<td>reworked</td>
</tr>
<tr>
<td>Sphenolithus heteromorphus</td>
<td>common</td>
<td>in situ</td>
</tr>
<tr>
<td>Sphenolithus mortiformis</td>
<td>few</td>
<td>in situ</td>
</tr>
</tbody>
</table>

Sample AS74
Zone:?
Stratigraphic remarks: This sample is almost barren. The very rare nannofossils are not diagnostic.
Preservation: very poor

<table>
<thead>
<tr>
<th>Nannofossil Type</th>
<th>Abundance</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>Discoaster spp</td>
<td>rare</td>
<td>in situ</td>
</tr>
<tr>
<td>Reticulofenestra spp (medium)</td>
<td>rare</td>
<td>in situ</td>
</tr>
</tbody>
</table>

Sample AS81 barren
Appendix 3B- $^{87}$Sr/$^{86}$Sr isotope stratigraphy

**Chemistry**

The powdered sample was weighed to be within the range of 0.05g and 0.1g and placed in labelled, clean Savillex (Teflon) beakers with screw tops. 2 mls of pure 0.25M HCl was added to each sample to dissolve the calcium carbonate (leaching step) and left for 1 hour. A residue remained in the base of the beaker at this stage and the sample was centrifuged and the liquid (leachate) pipetted off into the beaker. The residue was washed in approximately 2mls clean water, centrifuged and the water placed in the beaker along with the leachate. This was then dried down under lamps. Initial experiments used different strengths of acid (0.25M vs 2.5M HCl) for the first, sample dissolution step to see if this had any effect on the results (see below).

The leached samples were then redissolved in 2mls of 2.5M HCl and loaded onto cation exchange columns containing Bio-Rad AG50W-X8, 200-400 mesh resin. The column tops were washed with 2 times 1ml 2.5M HCl, then 60mls 2.5M HCl was eluted. The Sr was collected into a clean teflon beaker by a further elution of 12mls of 2.5M HCl and then evaporated under a lamp for several hours. Finally, a drop of concentrated HNO₃ was added and evaporated off. This acid converts Sr chlorides to Sr nitrates which are preferable for the mass spectrometry ionisation. The total procedure blank for this Sr chemistry is typically 0.5ng which is negligible for the samples studied here.

The samples were loaded onto outgassed, oxidised Ta filaments using 1 microlitre of 1M phosphoric acid and dried by passing a current of up to 2.5A through the filament. The samples were then analysed in the VG SECTOR 54-30 mass spectrometer under dynamic multicollection mode along with NBS 987 strontium standards. Isotope ratios were measured as 15 blocks of 10 cycles through axial collector masses 84.5 (baseline), 86, 87, 88, (i.e. 150 measurements). A $^{88}$Sr peak intensity of 1V ($1 \times 10^{-11}$A) ± 10% was maintained and the $^{87}$Sr/$^{86}$Sr ration was corrected for mass fractionation using $^{87}$Sr/$^{86}$Sr =0.1194 and an exponential law.
<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>[GR]</th>
<th>Type</th>
<th>XRD analysis</th>
<th>Measured 87Sr/86Sr</th>
<th>Age Ma (H &amp; W) 1994</th>
<th>Error +Ma</th>
<th>Error -Ma</th>
<th>Age Ma (Oslick et al, 94)</th>
<th>Error +Ma</th>
<th>Error -Ma</th>
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</thead>
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<tr>
<td>1</td>
<td>9635</td>
<td>566851</td>
<td>Oy</td>
<td>L Mg calcite, trace quartz</td>
<td>0.708413</td>
<td>20.04</td>
<td>0.28</td>
<td>0.3</td>
<td>20.66</td>
<td>0.28</td>
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<td>17.1</td>
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<td>16.68</td>
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<td>861121</td>
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<td>17.29</td>
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<td>733775</td>
<td>Oy</td>
<td>L Mg calcite</td>
<td>0.708594 av</td>
<td>17.84</td>
<td>0.23</td>
<td>0.22</td>
<td>18.22</td>
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<td>9</td>
<td>9656 Romana south, base Florinas Group</td>
<td>642805</td>
<td>Oy</td>
<td>L Mg calcite, trace quartz</td>
<td>0.708698 av</td>
<td>16.55</td>
<td>0.29</td>
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<td>0.34</td>
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<td>9649 Ittir - Uri road, base Fl. Grp.</td>
<td>592994</td>
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<td>16.30</td>
<td>0.32</td>
<td>0.29</td>
<td>16.41</td>
<td>0.36</td>
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<td>810919</td>
<td>Oy</td>
<td>L Mg calcite</td>
<td>0.708791 av</td>
<td>14.58</td>
<td>2.58</td>
<td>0.67</td>
<td>14.76</td>
<td>0.66</td>
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<td>12</td>
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<td>808918</td>
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<td>16.13</td>
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<td>16.18</td>
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<td>0.53</td>
<td>0.41</td>
<td>15.44</td>
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<td>16</td>
<td>9674 Scala Giocca basal, Fl. Grp.</td>
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<td>14.95</td>
<td>1.05</td>
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<td>11.32</td>
<td>0.26</td>
<td>0.38</td>
<td>12.7</td>
<td>0.98</td>
<td>1.0</td>
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<td>9668 Sassari east, Fl. Grp.</td>
<td>685090</td>
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<td>0.708831</td>
<td>11.47</td>
<td>0.30</td>
<td>0.51</td>
<td>13.2</td>
<td>1.05</td>
<td>0.90</td>
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</table>

Table 3B1. H&W 94 = correlated to the curve of Hodell and Woodruff 1994. SURRC NBS987 standard is 0.710237 ±0.000020. Oy=oyster, RA=red algae. L=leaching step in 2.5M HCl. Fl. Grp= Florinas Group
<table>
<thead>
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<th>Error -Ma</th>
<th>Error +Ma</th>
<th>Error -Ma</th>
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<tbody>
<tr>
<td>19</td>
<td>Sedini south, mid Sedini Member</td>
<td>848215</td>
<td>RA</td>
<td>LMg calcite</td>
<td>0.708737av 0.708732,2.5M 0.708737,0.25M</td>
<td>15.94</td>
<td>0.38</td>
<td>0.33</td>
<td>16.01</td>
<td>0.41</td>
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<td>20</td>
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<td>764286</td>
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<td>759290</td>
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<td>0.24</td>
<td>19.4</td>
<td>0.26</td>
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<td>22</td>
<td>Corsica, Bonifacio</td>
<td>842635</td>
<td>RA</td>
<td>LMg calcite, trace qz, albite</td>
<td>0.708656</td>
<td>17.1</td>
<td>0.25</td>
<td>0.25</td>
<td>17.35</td>
<td>0.3</td>
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</table>

Samples lying outwith data of published curves, younger overprint.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>[GR]</th>
<th>Type</th>
<th>XRD analysis</th>
<th>Measured 87Sr/86Sr ±0.000020</th>
<th>Age Ma (H &amp; W) 1994</th>
<th>Error +Ma</th>
<th>Error -Ma</th>
<th>Error +Ma</th>
<th>Error -Ma</th>
</tr>
</thead>
<tbody>
<tr>
<td>23</td>
<td>Isili, base Is Paras Member</td>
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<td>Oy</td>
<td>LMg calcite</td>
<td>0.709120</td>
<td>&lt;2.5Ma</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>24</td>
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<td>784958</td>
<td>RA</td>
<td>LMg calcite, quartz, albite, orthoclase</td>
<td>0.709027</td>
<td>4.5-5.5Ma</td>
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<tr>
<td>25</td>
<td>M. Uri north, Sennori Mbr.</td>
<td>736195</td>
<td>Oy</td>
<td>LMg calcite</td>
<td>0.709052</td>
<td>2.5-4.5Ma</td>
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<td></td>
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<tr>
<td>26</td>
<td>Giave west, base Fl.Grp.</td>
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<td>Oy</td>
<td>LMg calcite, trace qz</td>
<td>0.708894</td>
<td>~8.6Ma</td>
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<tr>
<td>27</td>
<td>M. Santo mid., Fl. Grp.</td>
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<td>LMg calcite</td>
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<td>~8.4Ma</td>
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<td></td>
</tr>
</tbody>
</table>

Table 3B1. H&W 94 = correlated to the curve of Hodell and Woodruff 1994. SURRC NBS987 standard is 0.710237±0.000020. Oy= oyster, RA= red algae. L = leaching step in 2.5M HCl. Fl. Grp= Florinas Group.
APPENDIX 3C- Ar-Ar Methodology

Sample Preparation
The samples were disaggregated to approximately crystal size in a jaw crusher and a tungsten carbide
tema such that sieved fractions of 1.4-0.75mm, 0.75-0.25mm and <0.25mm were collected. The
crushed samples were washed and dried. A combination of hand picking, ‘Frantz’ magnetic separation
and heavy liquids were used to separate plagioclase (andesine composition) and biotite crystals from
the surrounding rock. Plagioclase crystals were cleaned by placing in an ultrasound bath in 5-10% HF
for between 5-15 minutes (dependent on size) to remove any volcanic glass and in warm 3-6M HCl for
up to 20 minutes to remove calcite and other alteration phases. Biotite crystals were cleaned in acetone.
The cleanest, euhedral crystals were picked for irradiation since these were most likely to be primary
igneous phases and placed in foil packets in EK26 irradiation, vial D. The irradiation took place at the
CLICIT facility in the Oregon State TRIGA reactor for 16 hours in December 1996. In preparation for
analysis, individual crystals were placed in individual depressions in a numbered copper sample pan
which was placed under vacuum.

Sample analysis
The individual crystals were vaporised under vacuum to release the Ar gas either by simple total laser
fusion (plagioclases), by laser step heating (biotites), or by a laser degassing step (to remove alteration)
followed by total laser fusion (biotites, after step heating had revealed what temperature degassing was
appropriate). The sample gas was held in the extraction line for 10 minutes in the presence of 2 Zr-Al
‘getters’ of a 10A° molecular zeolite sieve held at 400°C and 800mg to absorb non inert gases (O, water,
hydrocarbons), leaving rare gases present. The sample was then introduced into the single collector
mass spectrometer which measured peaks of mass 41-35 in nine cycles. Most samples were measured
using an electron multiplier with a sensitivity of 2e-19, but larger biotites which gave greater amounts
of Ar gas were measured using a Faraday couple.

The extraction line and mass spectometer were pumped to vacuum for at least 10 minutes in between
analyses. Blank analyses which measure the very small amount of gas in the mass spectrometer and
extraction line were performed first thing in the morning, after the first sample and after every 4 or 5
consequently. Average or increasing blanks could then be taken from each sample analysis.

Processing Data
The amount of each isotope and its associated error was calculated by extrapolating the nine mass
spectrometer cycles back to the time when the sample was introduced into the machine. Ages and errors
were calculated after the blank correction had been deducted using standard Ar age equations after
Dalrymple et al. 1981. An age was produced for each total fusion experiment or for each step of the
incremental heating experiments.
For the total fusion experiments, the age of the sample was gained by taking a weighted mean of the results whilst excluding significant outliers resulting from sample alteration (e.g. closely grouped plagioclases). This can be compared with the total gas age calculated from a simple average of all the results and representing the age gained if bulk crystal separates had been used. Where total fusion experiments gave more scattered results (e.g. biotites), more emphasis was placed on explaining the variety of ages than on the weighted mean age.

For the step-heating experiments, a date was gained for each crystal by calculating the weighted mean plateau age, excluding initial or final steps which were not on the 'plateau'.

The mean square weighted deviates (MSWD) value given in Table 3.4 is the ratio of the total scatter about the mean and the scatter expected from the analytical error. In an ideal situation the MSWD would be one. For 6-10 values, a statistically relevant result was produced when the MSWD <2.5. In this dataset, the MSWD was <2.5 except for the scattered results of total biotite fusions, where each individual crystal was considered, rather than taking a weighted mean of the results.
+ 9'OZ

Z10

-+ SSOZ

Et'O

-+ Pot
+ 95'OZ
-+ 15'OZ
'+ 89'OZ

8t'O
ZZ'O
ZE'O
1.rO

-+ 69061

VIVO

+ L4561
-'I' 91, 1'61
+ EW61
-4' 11.0'91
-+ 160*61

160*0
61.1'!

90*0
59E'O
9EP'O

+ 658'8!

tco

'+ 68E'61

E9Z'O

-+ 8P0Z

+ 5L961

8950
9L0

-+ 885'61
+ E9L'61
-+ It'OZ

6110*0
P190
L5O

•+ zcosi

990

+ 19'Ot

80'0

-+ Lilt
+ S'Ot

£Z'O
81'0

+ 69'OZ
-+ SP•OZ
-+ E6OZ

Zi'O
WO
P1'O

Slot
-+ PCOZ

ZVO
WO

(I)
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6'PI
LL.1
E'II
0,91
£8
9'01

TWO
60'0
980'0
Z60'O
8800
Z800
LLO'O

66E9
ES
6L8
V9k
LV58
PLZOI
ZE'OL.

ago sag IIJ
VIZ
COt
5•I1
P'PI
9S1
L'Ol

1900
9900
Z90'0
150'O
iSOO
1500

ZEOO
EEOO
ISOO
LPOO
Lt'O'O

6'iZ
I'OZ
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151

167L

1.061.

Ito
9E'O
8C'O
9E'O
LEO

P50

co
ESO
59'0
I'90
P90

101
860
119'0
CO
690
90'1

WO
t'O
6E'O
9E0

Ito

tO•O
ZOO
ZOO
ZO'O
Z00
ZOO
ZO'O

coc

LE6
6'L6
516
696
V86
£0OI
56

PI-30VP
PI-301E
101309'E
0I-30E't
1,1-301t
101'30L1
101-30Z'Z

ELL
61SZ
60'PE
996t
8ZPE
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Z0'O
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ZOO
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t00
ZOO
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U.PV6E

LtP
9'68
9'P6
StE
9E6

171-30t1
01'3001
51-30cc
S1-30VL
SI -3091
SI1301'S

ZL'O
191
t'S'Lt
t8'OS
EL'!
ES'SS

CL.8
5'06
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S'PL
1'16
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9t6
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0I30CI

LZ'OZ
t'S'ZZ
9P0Z
899
tESt
L8'ZL

PVPI
90'S!
1.Z'IE
89'LOI
IE'ZP

CIO
ZO
Z'O
VO
CIO

to

(low)
U"VOP

Z•O
tO
Z'O
ZO
Z0
Z0
Z0

to
Z0
tO
to
ZO
t0

VO
tO
Z0
VO
to

ro

Z0
Z0
tO

E'O
P'O

Z'O

P JVLL

ro
VU
tO
Z0
V0
Z'O
Z•O

96PZ00O
611.b00'0
90L1000
E8100'O
EZLIOO'O
9Z91000
I8EZOO'O

SO'9
POP'S
6L9'S
6Z'S
9955
8E6'S
1P9

LZt'P
681'11
501'P
SIP
581'1'

Li)
9J1

U)
999

ti)
LI)
tI)
Li)

9E1'P

180'9

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511
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EPZP
Eti'!'
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Zil

968E
ESO'!'
Sold

SIE'9

Sil

6801
89E'0
PZZ'P
1,80'!'

888'S
PELS
66Z'S

Ii)
El)
ZJ)
LI)
0696
9J)

198t

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ti)
El)

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0196

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LLVI'

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90'6

9E11
Si)

SOt'!'

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88E'OI
PZEOI
ZZWSI

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596100'O

Z'O

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Z6ZEO()'O

Z'O
EO
E'O

56EZ00
IEPEOOO

Z0
Z'O

25 1d

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9)1.5000

VO

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SZLEOO'O
LSILOO'O

VO
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Z'O

zszcoo'o
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586ZO0'O

tO

E6PZOO'O
E61Z00'O
105100*0

Z0
VO
tO

2d

=f

-+ ZESZOO'O

Z69

(q)
(q)
('1)
(16)
'Ps JYOP JY6E/V9E JV6E1V1.E JV6EIVOP

('16)
P' "V6E

(%)

('16)
'P.S V9E

t06E00'O

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**Note:** The table above contains data related to gas age and concentration, with values expressed in scientific notation. The values are derived from a series of samples, each with its own concentration and error. The total gas age is calculated and provided at the end of each section. The data is organized in a tabular format for easy reading and analysis.
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(a) Steps labeled either as temperature (deg C), laser power (W), or individual total fusion analyses (TF#).
(b) Corrected for 37Ar and 39Ar decay, half-lives 35.1 days and 259 years, respectively.
(c) Radiogenic (R), calcium-derived (Ca), and potassium-derived (K) argon, respectively (percent).
(d) Ages calculated relative to 85G003 TCR Sanidine at 27.92 Ma with lambda e = 0.581E-10/yr and lambda b = 4.692E-10/yr.
Appendix 3D - Detailed stratigraphic considerations for each study area.

3D.1 Funtanazza sub-basin (Fig. 3.11, Table 3.5, Chapter 4)

Previous Nomenclature

The Arcentu Group has been separated by Assorgia et al. (1984) into four parts such that Formation A consists of basalt domes and flows overlain by a rhyolitic ignimbrite, Formations B and C consist of voluminous andesite flows with pillows, breccias and Formation D consists of basaltic dykes and intrusions.

Age Constraints

Whole rock $^{40}$K - $^{40}$Ar dates from the volcanic succession of the Arcentu Group define the four formations such that A= 30-24 Ma, B= 24-21 Ma, C= 21-18 Ma, D= 18-16 Ma (Assorgia et al. 1984). Radiometric dates from the pyroclastic flow overlying the Campu Sali Fm. and underlying the Sartori Fm. give ages of 20.87± 0.3 Ma (late Aquitanian, single crystal plagioclases, section 3.2.3, biotites 21.4-23.3 Ma), 23.8± 2.4 Ma ( Chattian-Aquitanian boundary, $^{40}$K - $^{40}$Ar plagioclase separates, Assorgia et al. 1995a).

The dated pyroclastic flow sits directly on top of the Pereiraia gervaisi marine bed at the top of the Campu Sali Fm. and is attributed to the N4 zone (early Aquitanian, in Assorgia et al. 1997c after Cherchi 1982). This fossiliferous horizon been dated by the $^{87}$Sr/$^{86}$Sr isotope technique at 23.9±0.3 Ma (latest Chattian, Barbieri et al. 1997 using Hodell et al. 1991 curve), roughly equivalent to the biostratigraphic data and in good agreement with the published radiometric age on the overlying pyroclastic flow. Published data are thus consistent in suggesting the ‘first’ marine transgression in the latest Oligocene-earliest Aquitanian.

According to the $^{87}$Sr/$^{86}$Sr isotope technique, marine sedimentation was re-established after the pyroclastic volcanism at 22.3±0.3 Ma (base of the Sartori Formation) and continued until 20.3±0.2 Ma (early Burdigalian) with an increase in sedimentation rate at 20.7±0.4 Ma (Barbieri et al. 1997). The $^{87}$Sr/$^{86}$Sr results of Barbieri et al. (1997) from the coastal section are older than the $^{87}$Sr/$^{86}$Sr isotope sample taken inland at the base of the Sartori Formation (sample 1, 20.7±0.3 Ma, latest Aquitanian), immediately overlying the pyroclastic flow. Whilst it may be expected that transgression occurred diachronously it is unlikely that it would take ~1.5 Ma. The Sr-isotope sample and Ar-Ar date from this study are consistent and suggest marine transgression overlying the pyroclastic flow in the late Aquitanian.
Microfossil analysis suggests that marine sedimentation (Sartori Formation) continued until the N6-N7 zones (~17.5 Ma, in Assorgia et al. 1997c). The Sa Tellura Fm. is dated at mid-late Aquitanian age (no data presented, Assorgia et al. 1986), though stratigraphic constraints from this study would mean that it was ?early Burdigalian in age.

Discussion

The map of Assorgia et al. (1986b) shows volcanic rocks of Formation B (21-24 Ma, late Oligocene-Aquitanian) stratigraphically overlying marine sediments of the Sartori Formation (until ~17.5 Ma Assorgia et al. 1997c) and lacustrine limestones of the Sa Tellura Formation (late Aquitanian 21 Ma). The published temporal data thus contains inconsistencies which are even more apparent if the data from this study are used. No simple explanations are obvious.

3D.2 Sarcidano sub-basin (Figs 3.12, 3.13, Table 3.6, Chapter 5, Fig. 5.1).

Relation to previous nomenclature

The sediments of the 'Ussana Formation' (Cherchi and Montadert 1982a,b: mid-Oligocene - mid/late Aquitanian continental clastics, marine sandstones and carbonates) and the 'volcanic succession' (Assorgia et al. 1995ab) are here redefined as the Sarcidano Group. Previous nomenclature for the Sarcidano Group includes the terms 'Gesturi sands' (west Isili fault block, Cherchi 1985), Marmilla Formation, (conglomerates and tuffaceous sandstones, Genoni fault block, Leone et al. 1984), Marmilla Formation (sils, tuffs and volcanic breccias, south of the Grighini fault block, Cherchi 1974; Assorgia et al. 1976; Cherchi 1985) for the shallow marine deposits of the Duidduru Member, Nureci Fm. The term 'Isili carbonates' (Cherchi 1985) has been used for the Is Paras Member and possibly the Serra Longa member of the Isili Formation. The marlstones of the Giara Group have been termed the 'Ales Marls' or Ales Formation', ‘Gesturi Marls or Formation’ (Cherchi and Montadert 1982ab; Cherchi 1985) and Marmilla Formation (Cherchi and Montadert 1982ab).

Age constraints

The age of the Villanovatulo Member is very poorly defined due to the lack of material suitable for dating within the continental conglomerates. Several sand-silt grade samples from developing paleosols were sent for pollen analysis but no results were forthcoming. Cherchi and Montadert (1982a) have correlated what is here defined as the Villanovatulo Member conglomerates of the Sarcidano sub-basin with the basal part of the Ussana Formation (Pecorini and Pomesano Cherchi 1969, Campidano region) due to facies similarities. Near Ussana, the basal part of the Ussana Formation has been dated as younger than 29.9Ma ± 1.5 Ma by a 40K-40Ar radiometric age of an andesite which cross-cuts the underlying Cixerri formation. In addition, rounded, well lithified clasts containing Nummulites sp. have been found as conglomerate clasts near Villanovatulo, indicating a
post-Eocene age. By inference, and in accordance with other workers in the area, the Villanovatulo Member is considered to be post-mid Oligocene in age.

The Araxigi Formation is intercalated at its base with continental clastics of the Villanovatulo Member and is partially equivalent to, plus is covered by and intercalated with, shallow marine sediments of the basinward Duidduru Member. $^{40}\text{K}-^{40}\text{Ar}$ dating of mineral separates (Assorgia et al. 1995a & in press) and single crystal $^{40}\text{Ar}/^{39}\text{Ar}$ dating (section 3.2.3) implies that the Araxigi Formation ranges from 20.64± 0.24 Ma (late Aquitanian, basal flow within Villanovatulo Member), to 19.07± 0.46 Ma (early Burdigalian, north of Nureci). Marine sedimentation (Duidduru Member) was contemporaneous with volcanism further basinward (e.g. front Genoni fault block, reworked tuff, 19.85± 0.52 Ma; $^{40}\text{K}-^{40}\text{Ar}$, Villanovaforru (30km to south) 20.5± 0.8 Ma Assorgia et al. 1995a). Plant remains within the sediments of the ‘Asuni unit’ towards the top of the volcanic succession (Assorgia et al. in press) are believed to be of Aquitanian age (C. Spano, pers. comm. 1995).

The sediments here defined as the Duidduru member are considered to be late Oligocene-Aquitanian age in the Isili area (no actual data presented in Cherchi and Montadert, 1982ab; Cherchi 1985), N4 microfossil zone in the Genoni area (Fig. 5.1, mid-Aquitanian-early Burdigalian, Leone et al. 1984) and early Burdigalian in the south Grighini area (Cherchi 1985). The most recent work and discussion with Sardinian workers gives biostratigraphic dates for this formation commencing at the uppermost Chattian-Aquitanian (N4 zone, ~50m in front Genoni fault block) and continuing until the N6-N7 zones (front Genoni fault block, mid-late Burdigalian, Serrano et al. 1997). Radiometric dating, stratigraphic relations and $^{87}\text{Sr}/^{86}\text{Sr}$ isotope dates from the Duidduru Mbr. support the younger N6-N7 zonation and confirm that the member youngs to the northwest. Oyster samples analysed by $^{87}\text{Sr}/^{86}\text{Sr}$ isotope dating gave results of 19.2± 0.3 Ma, west Nureci [974083] (sample 5, basinward of intercalated ignimbrite, early-Burdigalian), 19.4± 0.3 Ma, uppermost bed, Laconi-Genoni road [017064] (sample 6, early Burdigalian) and 17.6±0.3 Ma. southwest of Mogarella [961121] (sample 7, late Burdigalian). $^{40}\text{Ar}/^{39}\text{Ar}$ dating of plagioclase crystals from a shallow marine reworked tuff, south of the Genoni block ([994052] Duidduru member type section) were dated as 19.85± 0.52 Ma (early Burdigalian). Reworking is interpreted to have occurred almost ‘instantaneously’ on a geological timescale (5.4.2.3).

Although Cherchi and Montadert (1982a,b) date basinward carbonates as N4 zone and describe benthic foraminiferal from the Isili area, there does not seem to be any direct biostratigraphic analysis of the Isili Formation at Isili to give the ‘late Oligocene-Aquitanian’ age of Cherchi and Montadert (1982ab) since ‘microforaminifera are not particularly abundant’ (Cherchi 1985). $^{87}\text{Sr}/^{86}\text{Sr}$ isotope stratigraphy from a section through the Is Paras Member around Isili gives results from 18.4-16.9± 0.3 Ma (samples 2-4, mid-late Burdigalian), distinctly younger than the published ages described above.
Recent discussions have illustrated, firstly, confusion and errors caused by the use of different zonation schemes/timescales (C. Spano pers. comm. 1997), and an acceptance that the marine transgression in this area may be latest Aquitanian (A. Cherchi pers. comm. 1997) to Burdigalian in age (L. Simone 1997 pers. comm.). In consideration with stratigraphic relationships and radiometric dates on the underlying volcanics, the younger $^{87}\text{Sr}/^{86}\text{Sr}$ isotope dates are taken to be 'correct' in this locality.

Leone et al. (1984) use planktonic foraminifera to assign an late Aquitanian age (?top N4 zone) to the Serra Longa Member in the vicinity of Genoni though Serrano et al. (1997) date underlying sediments as N6-?N7 zones as discussed above.

At the basin margin, planktonic foraminifera and nannofossil zonations in marlstones and calcarenites of the Giara Group commence at ?N5 (early Burdigalian in Leone et al. 1984), NN3 (mid-Burdigalian, Cherchi 1985) and end at N7-N8 (Langhian, Cherchi 1974), N8 (late Langhian, Iaccarino et al. 1985) or N6-?N7 (late Burdigalian, Assorgia et al. in press). Calcarenites overlying the clastic Duidduru Member, west of the Genoni block, are dated by nannofossils as mid-Burdigalian age (upper NN2-NN3). Marlstones overlying the Serra Longa Member west and basinward of the Isili block marlstones are of upper NN3-NN4 (late Burdigalian) age.

In the basin centre, marine marlstone sedimentation commenced in the late Oligocene (P22-NP25, Cherchi and Montadert 1982ab, Ales marls) or latest Oligocene-early Aquitanian (NN1) or early Burdigalian (N5; see section 3.1.1, Cherchi 1985) and ended at the NN3/N7 zone (mid-late Burdigalian Villanovaforru section (30km to south of Allai), mid-Burdigalian, Cherchi 1985 and Serrano et al. 1997), ?Serravalian (G. Menardii zone, Oristano well, Pomesano Cherchi 1971). New nannofossil zonations confirm basin centre marlstones to the NNI-NN2 nannofossil zones (Aquitanian-early Burdigalian).

**Discussion**

The high resolution dating and stratigraphic relations in this area highlight the presence of abrupt facies changes and the diachroniety of lithostratigraphic units (e.g. Duidduru Mbr.). In addition, recent dating (radiometric, $^{87}\text{Sr}/^{86}\text{Sr}$ isotope and the biostratigraphy of Serrano et al. 1997) has shown that earlier publications (e.g. Cherchi and Montadert 1982ab; Cherchi 1985; Leone et al. 1984, dating by A. Cherchi) which assign all the sediments of the Duidduru Member and Isili Formation to the N4 foraminifera zone (Aquitanian) must be in error. Rather, at least the upper parts of these sediments are of Burdigalian age. This problem may be ubiquitous within published Sardinian biostratigraphy where the majority of zonations have been assessed by A. Cherchi.
In Sarcidano, Serrano et al. (1997) still attribute the start of marine sedimentation to the latest Oligocene-earliest Aquitanian, they present no data to confirm this. In the north-west of the Sarcidano area (Allai) non-marine basin filling occurs at a similar or lower topographic level until the early Burdigalian (constrained by Ar date on intercalated ignimbrite, 19.07± 0.46 Ma) and it is considered here, that transgression onto the parts of the basin margin exposed today may not have occurred until late Aquitanian/early Burdigalian times.

3D.3 Strike-slip systems of eastern Sardinia (Fig. 3.14, 3.15, Table 3.7, Chapter 6, Fig. 6.1)

Previous Nomenclature

Local lithostratigraphic terminology prevails in some areas. In the Oschiri sub-basin, the ‘Brecce di Codinattu’ fanglomerate, ‘conglomerato di Piano Ladu’, the ‘Lacustre auct.’ and acidic volcanics (Oggiano et al. 1995) are constituents of the Oschiri Formation.

In the Ottana sub-basin, Porcu (1983) and Porcu et al. (1997) define a stratigraphy comprising basal ignimbrites and tuffs, passing upwards and westwards into alluvial conglomerates and sandstones of the Sedilo Unit and the Sa Manenzia Unit pyroclastic flow. Assorgia et al. (1995a) subdivide the basal volcanic succession in this area into the ‘Allai Unit’ large volume ignimbrite overlain by the ‘Samugheo unit’ dacitic lava-like ignimbrites intercalated and underlain by the continental clastics of the ‘Ussana Fm’. These units are here correlated with the Oschiri Formation. They are unconformably overlain by fluvo-deltaic conglomerates, sandstones and mudstones of the Dualchi Unit, here correlated to the Chilvani Formation, and marlstones and calcarenites of the Tadasuni-Sorradile Unit (here correlated to the Florinas Group, see 3.3.4). On the eastern segment of the Nuoro fault, syn-transpressional continental conglomerates and breccias are known as the ‘Conglomerato di Cuccur’e Flores’ (in Carmignani et al. 1992).

Age Constraints

In the Oschiri sub-basin, radiometric dates from the Oschiri Fm give late Aquitanian ages of 20.60±0.24 Ma (single crystal ⁴⁰Ar/³⁹Ar, section 3.2.3) and 22±0.8 Ma (pyroclastic flow, unpublished data in Oggiano et al. 1995). These dates agree well with the mid Aquitanian-early Burdigalian age of mammal remains described by Bruijn and Rumke (1974) in Martini et al. (1992). The ‘upper conglomerates’ of the Chilvani Formation are lateral equivalents of late Burdigalian marine sediments to the west (Oggiano et al. 1995).

The stratigraphy is well constrained in the Ottana-Lago Omodeo area. The lowest exposed volcanism of the Oschiri Fm. is dated at 20.71±0.22 Ma (pyroclastic flow, latest Aquitanian, 3.2.3) and 21.1±1.4 Ma (late Aquitanian, ⁴⁰K-⁴⁰Ar biotite separates, Assorgia et al. 1995a) whilst the upper ‘Sa Manenzia’ unit was erupted at 19.08±0.54 Ma (early-mid Burdigalian, section 3.2.3). The Tadasuni-Sorradile
marlstones (Florinas Group) contain a foraminifera fauna diagnostic of the latest Burdigalian-early Langhian (G. Bisphericus, top N7 zone, Porcu et al. 1997). Eastwards and basinwards of the Ottana area, marine marlstones of Aquitanian age are found intercalated with andesites at Pauliliatino (Odin et al. 1994).

The ‘Conglomerato di Cuccuru’e Flores’ found at the eastern end of the Nuoro fault area is thought to be post early Lutetian (Eocene) since correlative conglomerates nearby contain reworked Cuisian-earliest Lutetian faunas (in Carmignani et al. 1992b) and pre-Burdigalian in age (Carmignani et al. 1994). No age constraints are available for the breccias, conglomerates and volcanics actually associated with strike-slip fault systems which can only be constrained as post-Paleocene (Dieni et al. 1987).

3D.4 Logudoro study area (Fig 3.14, 3.15, Table 3.6, Chapter 7, Fig. 7.1)

Previous Nomenclature

The volcanic stratigraphy of the Logudoro Group in this area has been defined into five broad series summarised in Table 3.1 and illustrated on the geological maps, Sheets 193 (Bonorva) and 180 (Sassari). The ‘Formation Lacustre’ (Cherchi 1985) and ‘tufi lacustri’ (Pomesano Cherchi 1971) are classified as the Oschiri Formation. Parts of the Florinas Group are known locally as the ‘Calcare di Sassari’ (Cherchi 1985) and ‘lower and upper sands/carbonates’ (Mazzei and Oggiano 1990, Florinas area).

Age Constraints

Volcanic rocks of the Logudoro Group are dated from -28 Ma (mid Oligocene, Beccaluva et al. 1985) until -18 Ma (i.e. mid-Burdigalian, uppermost pyroclastic flows, this study and Odin et al. 1994) and possibly until 17 Ma (final andesites and dacites, \(^{40}\)K-\(^{40}\)Ar dates in Beccaluva et al. 1985). Some of the whole rock \(^{40}\)K-\(^{40}\)Ar dates on the final phases of volcanism (13-16 Ma, Beccaluva et al. 1985) must be in error (presumably through alteration), since marine sediments which unconformably cover them are dated as late Burdigalian (N7 zone, ~16.7 Ma, below) by microfossils and by \(^{87}\)Sr/\(^{86}\)Sr isotope stratigraphy (e.g. Romana, 3.2.2).

Age constraints on the Florinas Group are described by geographical area, see Fig. 7.1:

In the south of the Logudoro area, Assorgia et al. (1987) describe marlstones near Giave containing foraminifera diagnostic of the N7-N8 zones (G. Bisphericus to Praeorbulina glomerosa zones, late Burdigalian-Langhian). \(^{87}\)Sr/\(^{86}\)Sr isotope stratigraphy from a shallow marine carbonate sample nearby (?laterally equivalent, S.M. Iscalas, Coissone, sample 8) gives an ‘age’ of 18.22±0.3 Ma (mid-Burdigalian), which could be considered in error (too old) when compared to the microfossil and radiometric constraints as discussed in section 3.2.2.
Moving northwards, Pomesano Cherchi (1971) provides a detailed analysis of clastic-carbonate-marlstone sedimentation and of planktonic foraminifera from logged sections of Monte Santo and Monte Pelao in central Logudoro. Marine sedimentation commenced at the top N7 - base N8 zone (G. Bisphericus, latest Burdigalian - earliest Langhian) and continued at least until the N9 zone (G. Altispira-G. Miozea or Orbulina universa (Cherchi 1974, zonation scheme from Iaccarino 1985, latest Langhian - earliest Serravalian). Pomesano Cherchi (1971) suggests that the overlying, uppermost carbonates which do not contain a diagnostic fauna are Tortonian in age. Martini, Oggiano and Mazzei (1992) suggest that these carbonates could be as young as early Messinian. 87Sr/86Sr isotope samples (11, 12) which came from the base and top of these uppermost carbonates and gave ages of 16.2±0.4 and 14.8±0.7 Ma respectively (early Langhian-latest Langhian/earliest Serravalian), not inconsistent with the microfossil zonations in underlying marlstones but distinctly older than the speculated ages.

Mazzei and Oggiano (1990) document biostratigraphic zonations from the Florinas area such that marlstones are NN4, upper N7-N8 zones (Spenolithus heteromorphus-helicosphaera ampliaperta, late Burdigalian-early Langhian) in age and the top of their ‘upper sands’ are upper NN5 zone (Spenolithus heteromorphus, mid-late Langhian). They suggest that the ‘upper carbonates’ which unconformably overlie the ‘upper sands’ may be as young as early Messinian but have no data to support this. An 87Sr/86Sr isotope sample taken from the lowermost calcarenite bed just beneath the thick sandstones which gave an age of 19.5±0.3 Ma (sample 13, lower Burdigalian) is obviously in error when compared to the timing of underlying volcanic rocks and NN4 zone marlstones (3.2.2). A sample taken at a conformable boundary between the thick sandstones and overlying carbonates gave an 87Sr/86Sr age of 16.3±0.4 Ma (sample 14, Burdigalian-Langhian boundary) in general agreement with the NN5 zonation taken at a similar stratigraphic location further to the south (Mazzei and Oggiano 1990).

Cherchi (1974) dates marlstones at the base of the Scala Giocca section (SE Sassari) as the O. Suturalis-Praeorbulina glomerosa, top N8 zone (mid Langhian). It is therefore unclear why Cherchi (1985) reports that the carbonates overlying these marlstones, which do not contain a diagnostic fauna, are late Burdigalian to Serravalian in age. 87Sr/86Sr isotope samples (16-18) taken from a continuous section at and west of the ‘Scala Giocca’ and give ages from 15±0.6 Ma to 12.7±1 Ma (latest Langhian-mid Serravalian) in good agreement with the biostratigraphic data.

Other 87Sr/86Sr isotope samples from Ploaghe (sample 15, calcarenite, 15.4±0.5 Ma, Langhian), Ittiri northwest (sample 10, transgressive carbonates, topographic high, 16.4±0.4 Ma, Burdigalian-Langhian boundary) and Romana south (sample 9, transgressive carbonate, 16.7±0.4 Ma, latest Langhian).
Burdigalian) record the late Burdigalian-Langhian transgression in the Logudoro area dependent on the local topography.

Discussion
In the Logudoro study area, some $^{87}$Sr/$^{86}$Sr isotopic and $^{40}$K-$^{40}$Ar radiometric dating techniques produce overlapping results between 19 Ma -15 Ma in a region where marine carbonates unconformably overlie already deposited ignimbritic, andesitic and dacitic volcanic rocks. This highlights a general problem which results from alteration of the volcanic rocks, giving an date which is too young and/or possible incorporation of volcanic sediment into some $^{87}$Sr/$^{86}$Sr isotope samples which gives a 'date' which is too old.

3D.5 Anglona study area (Fig 3.17, 3.18, Table 3.9, Chapter 8, Fig. 8.1)

Previous Nomenclature
Spano and Asunis (1984) and the geological map (Sheet 180, Sassari) provide a summary of the lithostratigraphy and informal terms used in this area. The ‘Tufo a Vaginella’ of Moretti (1942) and ‘Molassa a Vaginella’ of Parona (1887), M1t, are here defined as the Vaginella Member, ‘conglomerates and sands’ as the Castelsardo Member, ‘ignimbrites ($\tau_1$, $\tau_2$), tuffs, andesites (a), andesitic breccias (aX) and ash falls’ as the Tergu Formation and ‘carbonates and marls’ (M2c) as the Campulandru, Martis, Sennori and Sedini Members.

Thomas and Gennesseaux (1984) provide the most complete ‘stratigraphy’ for this area. It consists of the ‘Ussana Formation’ continental clastics (Valledoria member) which pass upwards into a transgressive ‘Marmilla Formation’ (Vaginella Member, Castelsardo Member), ‘lacustrine limestones’ (Perfugas Formation) and ‘Sedini limestones’ (Sedini Member). For the reasons discussed in section 3.1.3 the stratigraphic scheme of Thomas and Gennesseaux (1984) is not used here.

Age Constraints
As with other areas, continental facies at the base of the succession (Casteldoria and Valledoria Members) are temporally unconstrained, but are considered to be late Oligocene to early Aquitanian in age by utilising basin-wide correlation and based on the biostratigraphy of overlying sediments (e.g. Thomas and Gennesseaux 1984, Quesney-forest and Quesney-forest 1984).

The Vaginella and Castelsardo Members are dated as Aquitanian-early Burdigalian, (N4 zone, G. Dissimilis, Maxia and Pecorini 1969; N4-N6 zone, Spano and Asunis 1984; NN1-NN2 discoaster deflandrei-discoaster drugii zone, Francolini and Mazzei 1992; Assorgia et al. 1997a; ?NN1-NN2 nannofossil zones this work, Table 3.2).
Several studies have performed radiometric dating on the volcanic rocks of the Tergu Formation. Odin et al. (1994) use the $^{40}$Ar/$^{39}$Ar step-heating technique to date the uppermost pyroclastic flow at Chiaramonti (Fig. 6.1) using bulk plagioclase and biotite separates at $18.37 \pm 0.14$ Ma. This is the most reliable radiometric date for the area from the stratigraphically highest volcanic unit. Montigny et al. (1981) used $^{40}$K-$^{40}$Ar bulk separates to constrain 'lower ignimbrites, $t_1$' to 20.3-22.6 Ma and 'upper ignimbrites, $t_2$' to 18.3-19.2 Ma. Mameli and Oggiano (1997) do not give details of a $^{40}$K-$^{40}$Ar date of $18.1 \pm 0.3$ Ma on the ' $t_2$' ignimbrite which is stratigraphically beneath the flow sampled by Odin et al. (1994) and thus is considered to be erroneous. Synthesis of whole rock $^{40}$K-$^{40}$Ar data summarised in Beccaluva et al. (1985), is in general agreement with these trends such that 'lower ignimbrites' range from $21.7 \pm 0.9$ Ma - $20.3 \pm 0.8$ Ma, 'upper ignimbrites' from $19.5 \pm 0.5$ Ma to $18.9 \pm 0.5$ Ma and andesites span $31.2 \pm 1.1$ Ma (Osilo, to the south), or $23.5 \pm 0.9$ Ma to $17.7 \pm 0.8$ Ma.

The Campulandru Member carbonates are considered to be of Serravalian age in Spano and Asunis (1984, macrofaunal correlation), late Burdigalian (no data given, Francolini and Mazzei 1991) or latest Burdigalian to Serravalian (Assorgia et al. 1997a, no data given). $^{87}$Sr/$^{86}$Sr isotope dating provides a mid-Burdigalian age for the Campulandru member (samples 20, 21, $18.9-19.4 \pm 0.3$ Ma). Radiometric dates on the volcanic horizon ($t_2$) which the Campulandru member unconformably overlies are in the range $18.3-19.5$ Ma, highlighting the problems discussed at the end of section 3D.4. Since no other definitive data exists, on the chronostratigraphic diagram it has been assumed that the majority of $^{40}$K-$^{40}$Ar samples were altered, the ' $t_2$' horizon was erupted at $-19.5$ Ma and the $^{87}$Sr/$^{86}$Sr isotope ages are correct.

The Sedini and Martis Members are thought to be Langhian (from $-16$ Ma) to Tortonian age in Thomas and Gennessaux (1984), base N5 zone (mid-Aquitanian, Quesney- Forest and Quesney-Forest, by A. Cherchi, 1984) and uppermost Burdigalian -Serravalian in Odin et al. (1994, distinct Pecten sp. assemblage). The Langhian $^{87}$Sr/$^{86}$Sr isotope for the middle part of the Sedini Member (sample 19, $16 \pm 0.4$ Ma) is in agreement with Odin et al. (1994) and indicates that, as expected, carbonate sedimentation in this area occurred at a later time than topographically lower carbonates cropping out to the north (Campulandru Member).

The Sennori Member marlstones are latest Burdigalian-early Langhian in age (uppermost NN4 - base NN5 nannofossil zones, Spenolithus heteromorphus-Helicosphaera ampliaperta to S. heteromorphus, Francolini 1994).
Appendix 4A- Structural and palaeocurrent data
Funtanazza sub-basin

Funtanazza
Fig. 4.1

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### Appendix 5A - Structural, palaeocurrent and composition data, Sarcidano sub-basin

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APPENDIX 5B

X-ray diffraction methodology
Sample preparation involved crushing a small amount of sediment, oyster shell or drilled red algae powder using a mortar and pestle and loading the sample onto a glass disc using acetone. Analysis was performed using standard techniques on the Philips PW 1800 X-ray diffractometer at the University of Edinburgh. Peaks were identified and matched by utilising the standard peak matching program and with the expertise of Mr Geoff Angell.

Results of X-ray diffraction analysis
XRD analysis was used to test the high Mg calcite, low Mg calcite or aragonite carbonate composition of oyster shells and red algae samples used for Sr isotope dating. The results are listed in the table in Appendix 3B, Sr dating. The remaining XRD analyses were performed with the aim of determining the types of clay mineral present in the matrix of tuffaceous sandstones, tuffs/ashes and pelite rich sandstones.

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cc=calcite, or=orthoclase, qz=quartz, ab=albite, san-sanidine, an=anorthite

Thus although the XRD analyses identify the main minerals which occur as clasts within the rocks (e.g. quartz, plagioclases), in some cases the clay minerals clinoptilolite, analcime and ?montmorillonite are also recognised. These are common alteration phases of volcanic rocks (Tucker 1991).
Appendix 6A - Eastern Sardinia structural data

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### Appendix 8A - structural, paleocurrent and compositional data for the Anglona study area

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| gr                  | 42                     |
| im                  | 27                     |
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| py                  | 22                     |
| v                   | 0                      |
| gi                  | 21                     |
| gw                  | 9                      |
| o                   | 10                     |

| A2                  | n=155                  |
| 2<4cm               | 35                     |
| 4<6cm               | 43                     |
| 6<8cm               | 21                     |
| 8<10cm              | 15                     |
| 10<15cm             | 21                     |
| 15<20cm             | 15                     |
| 20<25cm             | 4                      |
| >25cm               | 1                      |

| B                   | n=77                    |
| gn                  | 7                       |
| sch                 | 7                       |
| gr                  | 42                      |
| im                  | 5                       |
| qz                  | 5                       |
| py                  | 44                      |
| v                   | 0                       |
| gi                  | 5                       |
| gw                  | 4                       |
| o                   |                         |

| B                   | n=77                    |
| 2<4cm               | 20                     |
| 4<6cm               | 24                     |
| 6<8cm               | 14                     |
| 8<10cm              | 7                      |
| 10<15cm             | 9                      |
| 15<20cm             | 1                      |
| 20<25cm             | 1                      |
| 25<30cm             | 1                      |

| C                   | n=77                    |
| gn                  | 2                       |
| sch                 | 2                       |
| gr                  | 5                       |
| im                  |                         |
| qz                  | 1                       |
| py                  | 58                      |
| v                   |                         |
| gi                  | 4                       |
| gw                  | 2                       |
| o                   | 3                       |

<p>| C                   | n=77                    |
| 2&lt;4cm               | 38                     |
| 4&lt;6cm               | 23                     |
| 6&lt;8cm               | 5                      |
| 8&lt;10cm              | 8                      |
| 10&lt;15cm             | 4                      |
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clasts >2cm counted

Fig. 8.32 Casteldoria type section

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Standard X-ray fluorescence techniques (XRF) were used to analyse major and trace element data.  

Rock preparation
Approximately 50g of material was cut from each rock sample to be analysed. The material selected was chosen to be as free from alteration and veins as possible. The cut sample was crushed and the chips ground to a fine powder in a tungsten carbide barrel for 3 minutes.

XRF preparation - major elements
Glass discs formed from the rock-powder were used in the analysis of the major elements. Preparation involved firstly drying the rock powders in an oven at 1100°C overnight. Approximately 1g of each powder was placed into a Pt-5% Au crucible. The samples were ignited for 20 minutes at 1100°C and a value for the loss on ignition (H₂O+CO₂-O₂) calculated. The ignited powder was then fused for 20 minutes at 1100°C using a lithium borate flux (Johnson Matthey Spectroflux 105) with a 5:1, flux:sample dilution. The molten material was placed onto a graphite plate and pressed into a disc by lowering an aluminium plunger onto the globule. The casting operation was carried out on a hotplate at 220°C and the glass disc allowed to anneal at this temperature for 10 minutes before cooling.

Trace elements
Pressed pellets formed from the rock powder were used in the analysis of the trace elements. The pellets were made by mixing ~6g of rock powder with 4 drops of binding agent (2% PVA in distilled water). The mixture was placed in a steel mould, surrounded and backed by boric acid powder and compressed at 8 tons using a hydraulic press to form a 40mm diameter pellet.

XRF Analysis
Samples were analysed using standard procedures on the Edinburgh University Philips PW 1480 wavelength-dispersive, sequential X-ray fluorescence spectrometer which is regularly calibrated and monitored for drift using standard samples. Major elements were screened by combining the total of the measured oxides with the loss on ignition. Where the total was outside the range 99.4-100.4 the sample was reanalysed using a new glass disc.

Errors in XRF analysis
Repeat analysis and regular calibration of standards on the Edinburgh machine shows that major element data has typical errors, incorporating the instrument and sample reproducibility, of for example of SiO₂ ± 0.44wt%, MgO± 0.16wt%, K₂O ± 0.04wt%, Nb ±1.2ppm (but G. Fitton pers.comm. ±0.2ppm), Zr ± 2ppm, Rb±0.8ppm (all 2σ errors, from James 1995). Some of the errors are significant when compared to the concentrations observed here, thus errors are plotted on selected graphs. Repeat analysis of some trace elements in some samples showed that Nb±0.5ppm is realistic.
Appendix 9B- Sample locations, normalising factors and XRF data

The locations of the samples from Logudoro, northern and Arcentu areas discussed in detail in chapter 9 are given below in Table 9B1. The locations for the samples collected in year one are approximate since no 1:25000 scale maps of the area were available at that time. Table 9B1 also gives the possible ages for the samples which were constrained using local stratigraphy and published K-Ar dates.

Table 9B2 gives the major element data of all the samples analysed with Si<55 which were not altered.

Table 9B3 gives the trace element data of all the samples analysed with Si<55 which were not altered.

Table 9B3 shows the re-analysed trace element data.

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Table 9B1 Location and estimated age of samples used in chapter 9. *after Assorgia et al. 1984, 1986

Spider diagrams were normalised to average N-type MORB using the values of Saunders and Tarney (1984), Sun (1980) given in Rollinson (1993). These are Sr/136, K2O%/0.15, Rb/I, Ba/12, Th/0.2, Nb/2.5, La/3, Ce/10, P%/0.12, Nd/8, Zr/88, Ti%/1.5 and Cr/290 (Cr from Pearce 1983).
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