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TECTONIC RECONSTRUCTION OF THE ALPINE OROGEN IN THE WESTERN MEDITERRANEAN REGION

BY

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SUMMARY

This thesis focuses on the tectonic evolution of the Alpine orogen in the western Mediterranean region. The aim of the thesis is to better understand orogenic processes and to elucidate the style of tectonism that takes place during the evolution of convergent plate margins. The western Mediterranean region, according to my working hypothesis, has been subjected to complex tectonic interactions governed by plate convergence, subduction rollback, formation of small back-arc basins, independent motions of continental blocks and accretion of allochthonous terranes. In the scope of this thesis, I test the working hypothesis by conducting detailed spatio-temporal analyses and developing a number of reconstruction models.

The relative motions of Africa, Europe, Iberia and Adria are revised based on a compilation of up-to-date data derived from magnetic isochrons in the Atlantic Ocean and palaeomagnetic data. Convergence of Africa relative to Europe commenced during the Cretaceous (120-83 Ma) and was subjected to fluctuations in convergence rates. The data suggest that a collisional episode in the Alpine orogeny at ca. 65 Ma resulted in a dramatic decrease in the relative plate motions, and that a slower motion since ca. 25 Ma promoted extension in the Mediterranean back-arc basins. Distinct changes in plate kinematics are also recognised in the motion of Iberia with respect to Europe. Palaeomagnetic constraints on the motion of Adria suggest that this continental area was attached to Africa since the Permian or the Triassic.

The tectonic evolution of the western Mediterranean since the Oligocene has been modelled using an interactive software package for plate tectonic reconstructions (PlatyPlus). The reconstruction shows that rollback/back-arc extension processes have affected a broad zone of deformation since the Oligocene and involved drifting and rotations of continental terranes, which were eventually accreted to the adjacent continents. In the Tyrrhenian Sea, spatio-temporal analysis provides a lithospheric-scale insight into the effects of such tectonic processes. The reconstruction shows distinct episodes of rollback/back-arc extension, which have been impeded by a number of accretionary events. At 6-5 Ma, the accretion of a large carbonate platform in the central Apennines initiated a slab tear and resulted in the formation of a narrow subducting slab that has consequently undergone higher degrees of subduction rollback and back-arc extension (in excess of 60-100 km/Myr).

According to the proposed model, arcuate orogenic segments in the western Mediterranean evolved during the extensional destruction of an earlier collisional belt. Therefore, the present locus of exposure of high-pressure rocks does not represent sites of continental collisions between major tectonic plates. Since the Oligocene, subduction of Mesozoic oceanic lithosphere, accompanied by rollback of the subducting slab, led to the progressive bending and the episodic tearing of subducting slabs, and resulted in the formation of several slab segments that are presently recognised in tomographic images. These remnant slabs can account for nearly all the volume of oceanic domains that existed in the western Mediterranean during the Oligocene.

Finally, the tectonic evolution of the western Alps is discussed. The orogen consists of ribbon-

like continental terranes that are separated by two or more ophiolitic complexes. High-pressure metamorphism occurred in distinct orogenic episodes and possibly reflects individual accretionary episodes that possibly occurred at the terminations of subduction rollback processes. In particular, widespread extensional tectonism, shortly after a collisional episode at ~35 Ma, possibly occurred as a result of rollback of an east-dipping subduction zone.

STATEMENT

This thesis does not contain material which has been accepted for the award of any other degree or diploma in any university or institution, and to the best of my knowledge contains no material published by any person, except where due reference is made in the text

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Gideon Rosenbaum
September 2003

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CHAPTER 1

INTRODUCTION

Plate tectonics theory is like Venus: very beautiful and born in the sea.

Rudolf Trümpy (1999)

FOREWORD: CHAPTER 1

This chapter provides infromation on the aim and the scope of this thesis, and outlines the working hypothesis that was developed in collaboration with my supervisor Gordon Lister. The chapter points out a number of key questions that will be addressed in the forthcoming chapters. It also includes a historical review on research with regard to orogenic processes. Laurent Aillères, Wouter Schellart and Megan Hough are acknowledged for providing comments on the manuscript. I also wish to thank Homer LeGrand for his interest in the historical review.

1. INTRODUCTION

Abstract

This thesis focuses on the evolution of the Alpine orogen in the western Mediterranean region and aims to elucidate the style of tectonism that takes place in convergent plate margins during orogenesis. I begin with a historical review with regard to orogenic research, showing that during the 18th and 19th centuries, orogeny was commonly co :dered to be associated with cooling and contraction of the whole Earth. In th late 19th century, geologists recognised the role of horizontal motions during orogenesis, and by 1912, Alfred Wegener presented his theory on continental drift, which suggested that mountains formed during the occasional crashing of continents into one another. The advent of plate tectonics, 50 years later, confirmed most of Wegener's ideas, and provided a comprehensive kinematic framework for global tectonics. Nevertheless, the rules of plate tectonics are insufficient to account for complex tectonic interactions that take place in convergent margins during orogenesis. Such interactions, which include rollback of subducting slabs, formation of small back-arc basins and accretion of allochthonous terranes, played a fundamental role during the tectonic evolution of the Alpine-Himalayan orogenic belt. This study focuses on the westernmost part of this belt, located in the western Mediterranean region, and aims to provide insights into the style of tectonic activity that took place in this region. In the thesis, detailed spatio-temporal analysis is performed to reconcile a myriad of geological and geophysical data with kinematically constrained tectonic frameworks. The resulting reconstruction models show the importance of rollback/back-arc-extension processes and interactions between numerous accreted terranes during the evolution of convergent plate margins in the western Mediterranean region.

1.1. Orogeny: an historical outline

1.1.1. Earth's contraction and geosynclines

The term orogeny is derived from the Greek words ορος (oros = Mountain) and γενεση (geneos = creation, origin). It came into use in the 19th century (Gilbert, 1890) to describe the deformation of rocks within mountain chains and the formation of mountainous topography. By present geological usage, orogeny is related only to the formation of structures in mountain chains, including folding and faulting in shallow crustal levels, and plastic folding, metamorphism and magmatism in deep crustal levels (Bates and Jackson, 1980).

From the 17th century (e.g. Descartes, 1664) to the end of the 19th century, it was generally agreed that mountain building processes were related to an ongoing contraction of the Earth. The contraction was believed to occur as a result of a gradual cooling of Earth through time - a process that was considered to be responsible for generating compressive stresses that wrinkled the Earth's crust to accommodate diminishing surface area (e.g. Elie de Beaumont, 1831; Dana, 1873; Suess, 1883). The process was commonly likened to the wrinkling of the skin of a desiccating apple (Oreskes, 1999).

An important benchmark in the history of orogeny research is marked by two works published in the second half of the 19th century, Hall (1859) and Dana (1873). James Hall and James Dana were particularly familiar with the formation of mountain belts located adjacent to continental margins (e.g. the Appalachians). Hall (1859) recognised that deformed sedimentary rocks in mountain belts were considerably thicker than their equivalent correlatives on platforms. He therefore suggested an origin that was linked to the accumulation of sediments along the continental margins. Dana (1873) coined the term "geosynclinal" (or "geosyncline") for elongated basins that were filled with a great thickness of shallow-water sediments and were somehow uplifted to form mountain belts. Dana (1873) suggested that oceans and continents were permanent features on the surface of the Earth, and that continental margins were particularly weak zones in the crust. Therefore, during the ongoing contraction of Earth, only the continental margins with their narrow geosyncline zones were subjected to folding (Dana, 1873).

During the late 19th century and the first half of the 20th century, the concept of the geosyncline was generally embraced by the scientific community. Suess (1883) applied it in Europe in the context of Alpine geology, albeit with a somewhat different meaning compared with the original American counterpart of the Appalachian Mountains. The Alpine mountain belt was not located at continental margins, thus casting doubts on the peculiar weakness of these zones within the Earth's crust. Moreover, the Alps "geosyncline" consisted of deep-sea deposits and ophiolites (Haug, 1900; Steinmann, 1905), which showed that oceans and continents were not permanent features of the Earth's surface. Instead, it was thought that a single "geosyncline" may represent episodic alternations between continental and oceanic environments. Suess (1883) suggested that such modifications in the surface of the Earth occurred across the whole globe as a result of the contraction of Earth. This assumption, however, was soon found to be unfeasible based on Suess's own observations from the Alps.

1.1.2. From nappe tectonics to continental drift

Suess (1875) was one of the earliest geologists to propose that large-scale horizontal movements are expressed in internal structures of the Alps. Heim (1878) drew similar conclusions based on structural observations, and Bertrand (1884) established the nappe hypothesis, which introduced the concept of far-travelled overthrust nappes. These ideas suggested that mountain building processes were accompanied by kilometric-scale lateral displacements, an idea that was somewhat at odds with the theory of the contraction of Earth, which mainly predicted vertical uplift.

The acceptance of Bertrand's nappe hypothesis in the late 19th century and early 20th century resulted in the establishment of a European-based mobilistic school of thought, led predominantly by Argand (1916; 1924) and supported by the theory of continental drift (Wegener, 1912). This theory was based on a synthesis of an enormous amount of data (particularly palaeontological data), which pointed out the possible common origin of the continents. Wegener (1912) concluded that crustal blocks had drifted through the ocean basins during the evolution of the Earth's crust. In a general sense, this concept was consistent with structural observations from the Alps that implied large horizontal displacements related to the "push" of Africa against Europe (Argand, 1924). Nevertheless, at the time, the theory of continental drift was too radical, and was rejected (and even ridiculed) by the majority of the scientific community (particularly in America; Oreskes, 1999), with the notable exceptions of Argand (1924) and Carey (1958).

1.1.3. Plate tectonics

The advent of plate tectonics during the 1960's (e.g. Vine and Matthews, 1963; Wilson, 1965; McKenzie and Parker, 1967; Le Pichon, 1968; Morgan, 1968) revived Wegener's theory of continental drift, and provided a consistent kinematic model involving the motions of rigid lithospheric plates on top of a semiplastic asthenospheric upper mantle. This radical revolution in the Earth sciences was probably not related to the long-standing firmness of the European mobilistic school of thought. Rather, the theory of plate tectonics evolved with the emergence of geophysical data from the ocean floors, which led to the recognition of linear stripes of magnetic anomalies (Mason and Raff, 1961), and to the development of the concept of sea-floor spreading by Hess (1962) and Dietz (1961). The existence of magnetic stripes was subsequently explained by the combination of sea-floor spreading and secular magnetic reversals (Vine and Matthews, 1963)[1]. Soon after the publication of these findings, it was proposed that the surface of Earth consisted of several large and relatively rigid plates, separated by narrow boundaries, which accommodated convergence, divergence or strike-slip (transform) motions (Wilson, 1965; McKenzie and Parker, 1967). According to this theory, the growth of tectonic plates in mid-oceanic ridges was compensated by subduction of oceanic plates at the trenches (Morgan, 1968). Mountain belts, therefore, were now perceived as crustal manifestations of subduction and collisional processes.

During the first few years that followed the advent of plate tectonics, the implication of the theory

[1] Coincidently in 1963, the Canadian geophysicist Lawrence Morley reached similar conclusions. Morley's paper, however, was rejected by *Nature* and subsequently, by the *Journal of Geophysical Research*, on the ground that it was too radical and speculative (Oreskes and Le Grand, 2001).

in natural examples resulted in the publication of works that provided answers to various complexities in global tectonics (e.g. Hamilton, 1969; Atwater, 1970; Dewey and Bird, 1970). Such works were inspiring attempts to apply plate tectonics in continental processes (i.e. processes related to the deformation of the continental crust). Specifically, plate tectonics was sought to explain orogenic processes, and led to an almost euphoric belief that major dilemmas with regard to mountain building processes would quickly be solved (LeGrand, 2000). Nonetheless, geologists soon realised that plate tectonic models require generalisation and oversimplification of a plethora of geological information, and as such, these models provided only a first-order approximations for continental processes (e.g. Isacks et al., 1968). As a result, many geologists have found plate tectonics an inadequate paradigm for addressing questions in continental tectonics (e.g. Molnar, 1988).

In the last 20-30 years, there was a considerable advance in the research on the origin of mountain belts. In the following sections, three aspects related to this research are discussed: (1) the existence of accreted allochthonous terranes in orogens (e.g. Ben-Avraham et al., 1981); (2) the mode of extensional tectonism during orogenesis (e.g. Dewey, 1988); and (3) the recognition of lateral migration of subduction systems (commonly subduction rollback) (e.g. Elsasser, 1971; Garfunkel et al., 1986).

1.1.4. Allochthonous terranes

The concept of allochthonous terranes (also called "accreted" or "exotic" terranes) evolved from field studies in the western part of North America, where various rock units, each characterised by a unique geological history that is different from the history of the contiguous belt, have been recognised (Jones et al., 1983). The origin of such terranes, as inferred from palaeontological and palaeomagnetic studies, has been often attributed to areas that are located great distances from the present position of the allochthonous terranes. In the early 1980's, many allochthonous terranes were recognised throughout the circum-Pacific orogens, as well as in the Alpine-Himalayan (Tethyan) belt (Ben-Avraham et al., 1981; Nur and Ben-Avraham, 1982; Hashimoto and Uyeda, 1983). The existence of such terranes further complicated plate tectonic reconstruction models, because such reconstructions required the incorporation of numerous blocks or "microplates" during orogenesis, accompanied by large degrees of horizontal motions.

1.1.5. Extensional tectonics in orogens

The role of extensional tectonism during and after orogenesis (Dewey, 1988, and many others) attracted considerable research since the early 1980's, and was explained by various kinematic and geodynamic models (e.g. Platt, 1986; England and Houseman, 1988; Buck and Sokoutis, 1994). A particularly interesting mode of extension, first recognised in the Basin and Range region of the western United States (e.g. Hamilton, 1987), is characterised by extension accommodated along

flat-lying detachment faults (Lister and Davis, 1989), and is associated with the exhumation of deep-seated metamorphic core complexes (Lister et al., 1984; Davies and Warren, 1988; Hill et al., 1992; Jolivet et al., 1994). Currently, extensional structures associated with metamorphic core complexes are widely accepted features in orogenic belts. However, their significance and the driving mechanism for extension are matters of controversy. In some models (e.g. Platt, 1986), syn-orogenic extensional tectonics has been attributed to internal rearrangements within the orogenic wedge, following the overthickening of the wedge and its tendency to regain gravitational stability. In other models, the role of syn-orogenic extension has been explained by orogenic-scale alternations between shortening and extensional regimes, which were generated by the geodynamics of a lithospheric slab (e.g. Lister et al., 2001; Collins, 2002). Such processes, related to the destruction of orogenic belts by extensional processes, have usually not been expressed in plate tectonic reconstruction models.

1.1.6. Migration of subduction systems

The theory of plate tectonics, as a geometric model, accounted for global plate kinematics, but did not address geodynamic complexities related to the crust and lithosphere. Research done in the last thirty years has shown that these geodynamic complexities, particularly in subduction systems, may possess profound implications on the evolution of convergent margins. Subduction zones are commonly subjected to backward migration ("rollback") of the subduction hinge due to a gravitational instability generated by the negative buoyancy of the subducting oceanic lithosphere relative to the surrounding asthenosphere (Elsasser, 1971; Molnar and Atwater, 1978; Dewey, 1980; Garfunkel et al., 1986). The process of subduction rollback has been documented by direct GPS measurements that showed a relative motion between volcanic arcs and the hinterland of the subducting plate (e.g. Bevis et al., 1995). Evidence for arc migration associated with subduction rollback has been found in numerous subduction zones, including various arcs in the western Pacific Ocean (Molnar and Atwater, 1978), the New Hebrides-Tonga arcs (Bevis et al., 1995), the Scotia arc (Alvarez, 1982), the Lesser Antilles arc (Alvarez, 1982), the Hellenic arc (Le Pichon and Angelier, 1979), the Calabrian arc (Malinverno and Ryan, 1986) and the Rif-Betic arc (Lonergan and White, 1997).

These observations suggest that the large horizontal displacements that take place during orogenesis are, at least in part, related to the mobilism of subduction zones. Furthermore, it has been speculated that high velocities of subduction rollback may result in widespread extension on the overriding plate (Dewey, 1980; Royden, 1993a), which in turn, may lead to the extensional destruction of existing orogens. Subduction rollback, therefore, may play a fundamental role in switching the tectonic modes of orogens from crustal shortening to extension.

1.1.7. Reflection on plate tectonic reconstructions

Ninety-one years after the promulgation of the theory of continental drift by Wegener (1912), the role of continental mobilism during orogenesis is established as an undebated fact. Yet, plate tectonic reconstructions encounter great difficulties while dealing with the kinematic evolution of orogens because it accounts only for the convergent motions of major plates and not to internal interactions in the convergent zones. Plate tectonic remains the unifying theory of global tectonics but has proven insufficient to account for the complexity of processes that occur during the deformation of the continental crust, particularly in convergent margins (i.e. the role of accreted terranes, syn-orogenic extension and subduction rollback). Therefore, plate tectonic rules must be complemented by additional spatio-temporal constraints.

1.2. Working hypothesis

The working hypothesis for this research suggests that **Alpine-Himalayan orogeny involved complex interactions between retreating subduction systems, back-arc extension and independent motions of numerous continental blocks.** These complexities explain why it is problematic to implement plate tectonic rules in kinematic reconstructions of the Alpine-Himalayan orogen. The working hypothesis is based on the recognition of tectonic processes that take place in convergent plate margins (e.g. in the southwest Pacific; Hall, 2002). In the context of the Alpine-Himalayan orogen, the hypothesis implies that tectonic reconstructions cannot be obtained by utilsing large scale plate kinematics. Rather, additional spatio-temporal constraints are required.

1.3. Aim of this research

The aim of this research is to test the working hypothesis using the example of the Alpine belt in the western Mediterranean region and the western Alps. This necessitates to reconcile a large amount of descriptive observations with kinematic modelling, and to develop a well-constrained reconstruction model for the tectonic evolution of the region. The following key issues are addressed:

- the methodological approach used to a establish kinematic framework for tectonic reconstructions of orogenic belts;
- the role of large-scale convergence associated with the style of tectonism in the orogenic belt;
- the degree of horizontal motions that can take place during convergence of two large-scale plates (Africa and Europe);
- the effect of accretionary events on the geodynamic and kinematic evolution of convergent margins;
- the main processes controlling the formation of curved structures along the Alpine belt in the western Mediterranean region; and
- the type of tectonic activity in the western Alps in the context of the tectonism in the western Mediterranean region.

1.4. Scope of this thesis

The Alpine-Himalayan orogen is a belt of Mesozoic and Cenozoic deformation stretching from Spain to Southeast Asia (Figure 1.1a). In the scope of this research, I focus on the westernmost part of the Alpine-Himalayan belt, in an area bounded by Gibraltar straits to the west, the Maghrebide Mountains to the south, the Adriatic Sea to the east and the central Alps to the north (Figure 1.1b). This area was chosen because of the enormous amount of data that exist from previous research. In addition, the role of back-arc extension (e.g. the western Mediterranean basin) and independent block rotations (e.g. the rotation of Corsica and Sardinia) is particularly well documented in the study area, providing the opportunity to use spatio-temporal constraints to quantify these processes.

Research involved analysis of data derived from multidisciplinary fields, which include structural geology, palaeomagnetism, igneous and metamorphic petrology, geochronology, stratigraphy and seismic tomography. The study area has been subjected to considerable research in the past, and an enormous amount of published data is available in the scientific literature. In the scope of this research, I have extensively used these published data.

The reconstruction procedure is based on developing a hierarchy of rules and assumptions that can help to utilise the available data as spatio-temporal constraints. This methodology provides a logical framework for the development of tectonic reconstructions that are not merely dependent on the opening and closing of existing oceans. The reconstruction involves visualisation of tectonic processes using the PlatyPlus software package, which has been developed by researchers at the Australian Crustal Research Centre (Duboz et al., 1999). In this study, PlatyPlus was used to test kinematic models by developing reconstruction animations, which provided a powerful tool in drawing attention to tectonic complexities and to identify the significance of regional events.

1.5. Layout of this thesis

The thesis starts with a description of global tectonic patterns related to convergent margins. I then discuss large-scale plate motions, and finally evaluate and apply spatio-temporal constraints in specific regions throughout the western Mediterranean and the Alps. This spatio-temporal analysis results in a series of reconstruction models (Chapters 5-9), which provide a test against the working hypothesis and the style of tectonism in convergent margins.

The forthcoming parts of the thesis can be divided into the following sections:

- Reconstruction philosophy (Chapter 2).
- Formulation of a kinematic framework (Chapters 3 and 4).
- Tectonic reconstructions of the Alpine belt in the western Mediterranean region (Chapters 5-7).
- Western Alps reconstructions (Chapters 8-9).
- Conclusions (Chapter 10).

Each chapter between chapter 2 and chapter 9 (inclusive) is written as an independent research article. Therefore, there are some repetitions in the text and in the illustrations (particularly in association

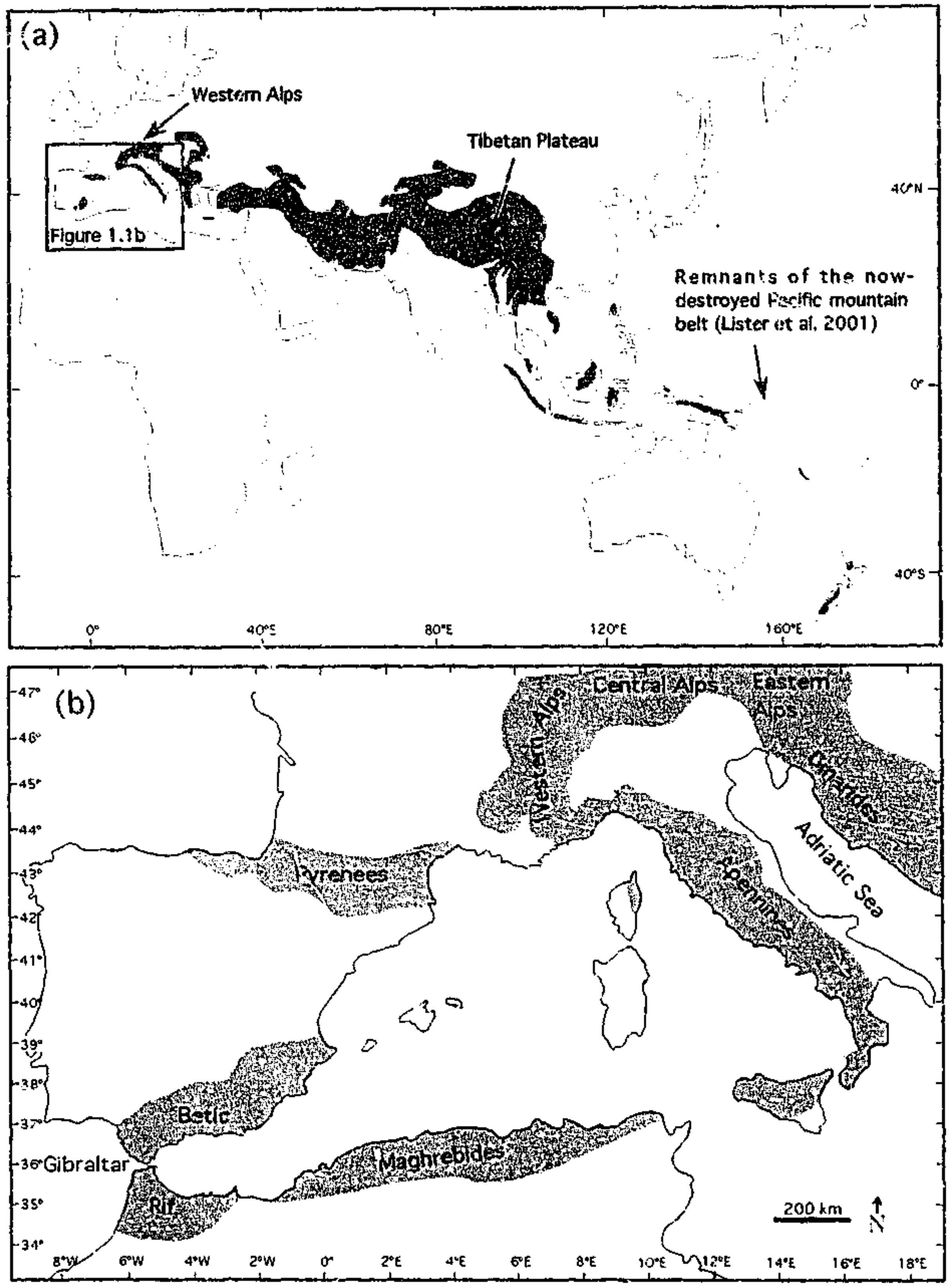


Figure 1.1. (a) Map showing the areal distribution of the Alpine-Himalayan orogenic belt, stretching from Spain to Southeast Asia, and possibly, further southeast towards the Indonesian archipelago, Papua New Guinea and New Zealand (after Lister et al., 2001); (b) Extent of the area covered in this thesis.

with the background material). A brief summary of the content of each chapter is given below.

Chapter 2 draws the attention to characteristic tectonic patterns in convergent plate boundaries and demonstrates how spatio-temporal constraints are utilised in tectonic reconstructions. I outline

major assumptions, evaluate the significance of spatio-temporal constraints and construct "rules" that can be used while reconstructing complex tectonic interactions.

Chapter 3 provides kinematic boundary conditions associated with the motions of Africa and Iberia relative to Europe. The kinematic analysis is based on the history of the opening of the northern Atlantic Ocean since the Jurassic. The implications of these motions with respect to Alpine orogeny are discussed.

Chapter 4 provides constraints on the motion of Adria. The kinematics of this large continental area in the central Mediterranean has been subjected to contradicting interpretations in the literature. The Adriatic problem is revisited based on analysis of palaeomagnetic, stratigraphic and seismic data, which suggest that Adria moved together with Africa since the Permian. The oceanic nature of the Ionian Sea and its age are also discussed in this chapter.

Chapter 5 presents a kinematic reconstruction model of the tectonic evolution of the western Mediterranean since the Oligocene. The reconstruction is accompanied by an animation, which visualises the extent of horizontal motions associated with subduction rollback, back-arc extension and independent motions of crustal blocks.

Chapter 6 presents a more detailed reconstruction of the evolution of the Tyrrhenian Sea and the Apennine-Maghrebide belt. Here, detailed analysis of spatio-temporal constraints provides better resolution on the timing of tectonic processes, showing links between rollback processes and back-arc extension, as well as possible temporal relationships between accretionary processes, tearing of subducting slabs and switches in tectonic modes.

Chapter 7 deals with the formation of arcuate orogenic patterns in the western Mediterranean region. It is shown that the arcuate shape of the orogen was attained during extensional destruction of an earlier high-pressure/low-temperature belt. These extensional processes are predominantly attributed to the rollback of subduction zones in directions that are oblique or orthogonal to the direction of convergence. I therefore propose that the present exposures of high-pressure/low-temperature metamorphic rocks are not necessarily the sites of large-scale continental collisions.

Chapter 8 provides a review on the tectonic evolution of the western Alps from the Jurassic to the Oligocene. Results of this chapter show that tectonic activity in the western Alps involved multiply collisional episodes and possible switches of tectonic modes from crustal shortening to extension.

Chapter 9 focuses on a style of tectonism in the Oligocene Alps, showing evidence for widespread extensional tectonism that immediately postdates a major collisional event at ca. 35 Ma. The kinematic and geodynamic implications of these processes are discussed.

Chapter 10 summarises the conclusions drawn from this thesis, provides brief reflections on relevant key issues and discusses several possible topics for future research.

CHAPTER 2

TECTONIC RECONSTRUCTION OF OROGENIC BELTS USING SPATIO-TEMPORAL CONSTRAINTS

But that is exactly what I like about this science of geology. It is infinite, ambiguous, like all poetry; like all poetry it has secrets, it permeated by them, lives within them, without being destroyed by them. It does not lift the veil, but only moves it, and through tiny holes in the fabric a few rays escape, which dazzle the eye.

Rodolphe Toepffer, Nouvelles Génevoises (1841)

FOREWORD: CHAPTER 2

This chapter discusses how spatio-temporal constraints are utilised in tectonic reconstructions of orogenic belts. The chapter explains the methodological approach used in the forthcoming chapters, and provides several examples from the western Mediterranean region. The ideas presented here benefited from many discussions with Gordon Lister on the 'science of spatio-temporal constraints'. We intend to develop a publication together on this subject. I also wish to thank Jerome Ganne for discussions on this topic and Ivo Vos, Wouter Schellart and Virginia Toy for providing comments on the manuscript.

2. TECTONIC RECONSTRUCTION OF OROGENIC BELTS USING SPATIO-TEMPORAL CONSTRAINTS

Abstract

This chapter provides a methodological framework for developing tectonic reconstructions of convergent margins during orogenesis. Tectonism in these domains is governed by interactions between subduction systems and numerous continental terranes, which result from a number of tectonic processes, e.g. subduction rollback, back-arc extension and accretion of allochthonous terranes. Nevertheless, results of these processes are usually not incorporated in reconstruction models, because these models are predominantly based on data derived from magnetic isochrons, hotspot tracks and palaeomagnetism, which provide only first-order spatio-temporal constraints. In the tectonic setting under consideration, the application of such data is not always possible, or provides an incomplete picture. Therefore, additional lower-order spatio-temporal constraints are required. The evaluation of these constraints is based on a set of rules and assumptions that enable us to extract kinematic criteria from a large amount of geological and geophysical data. In the case study of the western Mediterranean region, I demonstrate that the kinematics of subduction rollback, back-arc extension and the independent motion of allochthonous terranes can be quantified using adequate spatio-temporal analysis. This approach can be used to establish tectonic reconstructions in complex zones, or in pre-Mesozoic reconstructions, for which conventional reconstruction methodologies are not applicable.

2.1. Introduction

In the 35 years since the advent of the plate tectonic theory, the science of global tectonic reconstruction has progressed relatively slowly. In part this is because plate tectonic reconstructions have mainly focused on motions that can be inferred from sea-floor spreading in presently observed ocean basins (e.g. Dewey et al., 1973; Molnar and Tappotnier, 1975; Biju-Duval et al., 1977). These data were complemented by palaeomagnetic observations (e.g. Van der Voo, 1993), and both data sets were taken into account in global reconstructions (e.g. Torsvik et al., 2001). There has been continued progress in terms of data analysis (e.g. Srivastava et al., 2000) leading to refinements of the relative motions of the major plates (e.g. Rosenbaum et al., 2002a). Occasionally, new data in respect to the opening of ocean basins comes to light (e.g. Ross Sea (Cande et al., 2000); Arctic

Ocean (Michael et al., 2003)), resulting in a radical impact on existing paradoxes, dilemmas and controversies.

Although the methodology used for plate reconstructions of this type is well documented (Pitman and Talwani, 1972; Dewey et al., 1973), there is less focus on the types of complexity that occur in convergent plate margins. In these tectonic environments, tectonic processes are associated with complex interactions, as expressed, for instance, in the geology of the western Mediterranean region (Rosenbaum et al., 2002b) and Southeast Asia (Hall, 2002). The methodological approach for this type of tectonic reconstruction has received little attention, which is a shortcoming in the sense that it does not address tectonic patterns, such as subduction rollback, back-arc extension and accretion of allochthonous terranes. All these processes are known to play major roles in global tectonics (e.g. Uyeda and Kanamori, 1979; Ben-Avraham et al., 1981; Jarrard, 1986) and must be incorporated in kinematic tectonic reconstructions.

For example, tectonic models for the Alpine orogen in the western Mediterranean region (Royden, 1993b; Lonergan and White, 1997; Rosenbaum et al., 2002b; Rosenbaum and Lister, in review-a; in review-b) have shown that subduction rollback and formation of extensional back-arc basins have played fundamental roles during orogenesis. This complex tectonic history cannot be reconstructed solely on the basis of magnetic isochrons and palaeomagnetic observations. In the Tyrrhenian Sea, Rosenbaum and Lister (in review-a) have estimated rollback velocities of 60-100 km/Myr during the Miocene. Rollback processes were coeval with lithospheric extension in the back-arc region, and involved large horizontal motions of continental blocks (e.g. Corsica, Sardinia, Calabria and the Kabylies) that have been located on the overriding plate. Such horizontal motions were often oblique or even perpendicular to the direction of overall plate convergence, and continued until the accretion of these continental blocks (as allochthonous terranes) into the continental palaeomargins of the surrounding continents (Rosenbaum et al., 2002b).

Perhaps the best example for this style of tectonism is the Southwest Pacific Ocean (Hall, 2002), where rollback processes are considered to be responsible for the formation of numerous back-arc basins that separate a series of continental ribbons (e.g. Lord Howe Rise, Norfolk Ridge and Three King Rise). In this style of tectonism, plate kinematic models that are based only on the large-scale motions of major plates are insufficient to account for the complex kinematics.

The aim of this chapter is to establish a methodological framework for kinematic reconstructions of convergent margins. I discuss a number of criteria that can be used as spatio-temporal constraints in reconstruction modelling, and demonstrate that such constraints provide a viable tool for reconstructing tectonic complexities in convergent plate margins.

2.2. Tectonic patterns in convergent plate boundaries: subduction rollback, back-arc extension and accretion of allochthonous terranes

The mobility of subduction systems has been discussed by various authors (e.g. Elsasser, 1971; Molnar and Atwater, 1978; Dewey, 1980; Garfunkel et al., 1986; Royden, 1993b) and is currently supported by direct GPS measurements (e.g. Bevis et al., 1995). Most subduction zones have been subjected to oceanward retreat (or "rollback") during the subduction of oceanic lithosphere (Jarrard, 1986). This process has been explained by a gravitational instability resulting from the negative buoyancy of the lithospheric subducting slab in comparison with the surrounding asthenosphere (Elsasser, 1971; Molnar and Atwater, 1978; Dewey, 1980) (Figure 2.1a). The gravitational instability results in sinking of the subducting lithosphere at an angle that is steeper than the slab dip, which in turn, leads to a regressive motion of the hinge of the subducting slab (Dewey, 1980). Tomographic images (e.g. Lucente et al., 1999; Widiyantoro et al., 1999), as well as analogue modelling (Faccenna et al., 2001a; Schellart, 2003), show that at depth of ~670 km, the dip of the subducting slab tends to shallow until it is lying parallel to the 670 km discontinuity (Figure 2.1b,c). Subduction rollback is likely to cease once continental crust arrives at the subduction zone (Figure 2.1c), thus reducing the negative buoyancy of the slab and impeding further rollback (Lonergan and White, 1997; Rosenbaum and Lister, in review-b).

Relationships between subduction rollback and back-arc extension have been discussed by Molnar and Atwater (1978), Dewey (1980) and Royden (1993a). These authors suggested that extension at the overriding plate is triggered by rapid subduction rollback accompanied by relatively slow convergence. If the velocity of subduction rollback exceeds the velocity of convergence, a vacancy in the overriding plate is likely to be accommodated by horizontal extension (Figure 2.1b). Schellart et al. (2002a) and Schellart (2003) have classified the geometry of these rollback/back-arc systems into three groups: unidirectional systems, radial systems and asymmetric systems (Figure 2.2).

In unidirectional rollback/back-arc systems, arc migration occurs in a single direction, as illustrated in the examples of the Scotia arc and the South Shetland arc (Figure 2.2b). In this tectonic configuration, the back-arc region is bounded on both sides by lithospheric-scale discontinuities, along which strike slip motion occurs.

Radial rollback/back-arc systems are characterised by oroclinal bending of the arc during subduction rollback and back-arc extension (Figure 2.2c). Such processed are accompanied by opposite sense of block rotations on both sides of the arc and possible conjugate strike-slip faults (Schellart et al., 2002a). Natural examples for radial rollback/back-arc systems are the Hellenic arc and the Calabrian arc in the Mediterranean Sea (Figure 2.2a,c).

Asymmetric rollback/back-arc systems are indicated by wedge shaped back-arc basin geometry and block rotations that are either clockwise or counterclockwise (but not both). The back-arc basin is bounded on one side by a lithospheric-scale strike-slip fault that accommodates the sense of motion related to arc migration. Schellart et al. (2002b; 2002c; 2003) have recently discussed examples of asymmetric rollback/back-arc systems in the New Hebrides – Tonga arcs (Figure 2.2d) and the Kuril arc, and suggested evolutionary models associated with gradients of rollback velocities along the subduction trenches.

The style of tectonism in rollback/back-arc systems enables large horizontal displacements of continental and oceanic fragments before they are accreted to the orogen as allochthonous terranes (e.g. Nur and Ben-Avraham, 1982). The origin of these terranes is associated with continental ribbons rifted from a larger continent during back-arc extension, or with density anomalies in the ocean floor (e.g. oceanic plateaux, oceanic ridges and hotspot tracks). Such continental blocks or oceanic anomalies consist of relatively buoyant material that is more resistant to subduction and is commonly accreted rather than subducted when arriving at the subduction zone (Figure 2.1c). Furthermore, accretion of allochthonous terranes is likely to impede subduction processes and to slow down or cease subduction rollback. This, in turn, can lead to tearing of subducting slabs, formation of cusps, and segmentation of the subduction system.

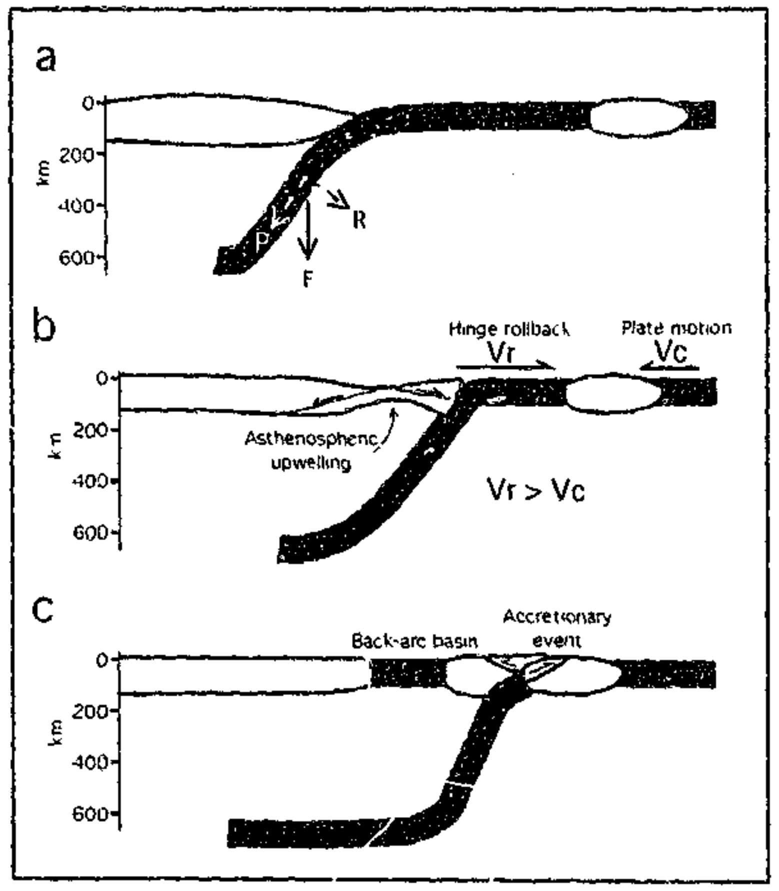


Figure 2.1. (a) Schematic illustration of the evolution of subduction rollback by the vertical negative buoyancy (F) of the subducting slab (P and R are two components of F, parallel and perpendicular to the slab, respectively). The subduction zone is pulled backward if the cold and dense slab is not supported by the mantle asthenosphere (after Lonergan and White, 1997). (b) Formation of back-arc extension when the rate subduction rollback (Vr) exceeds the rate of convergence (Vc) (after Dewey, 1980). (c) Arrival of continental block at the subduction zone. The process impedes rollback processes and results in the accretion of the continental block as an allochthonous terrane.

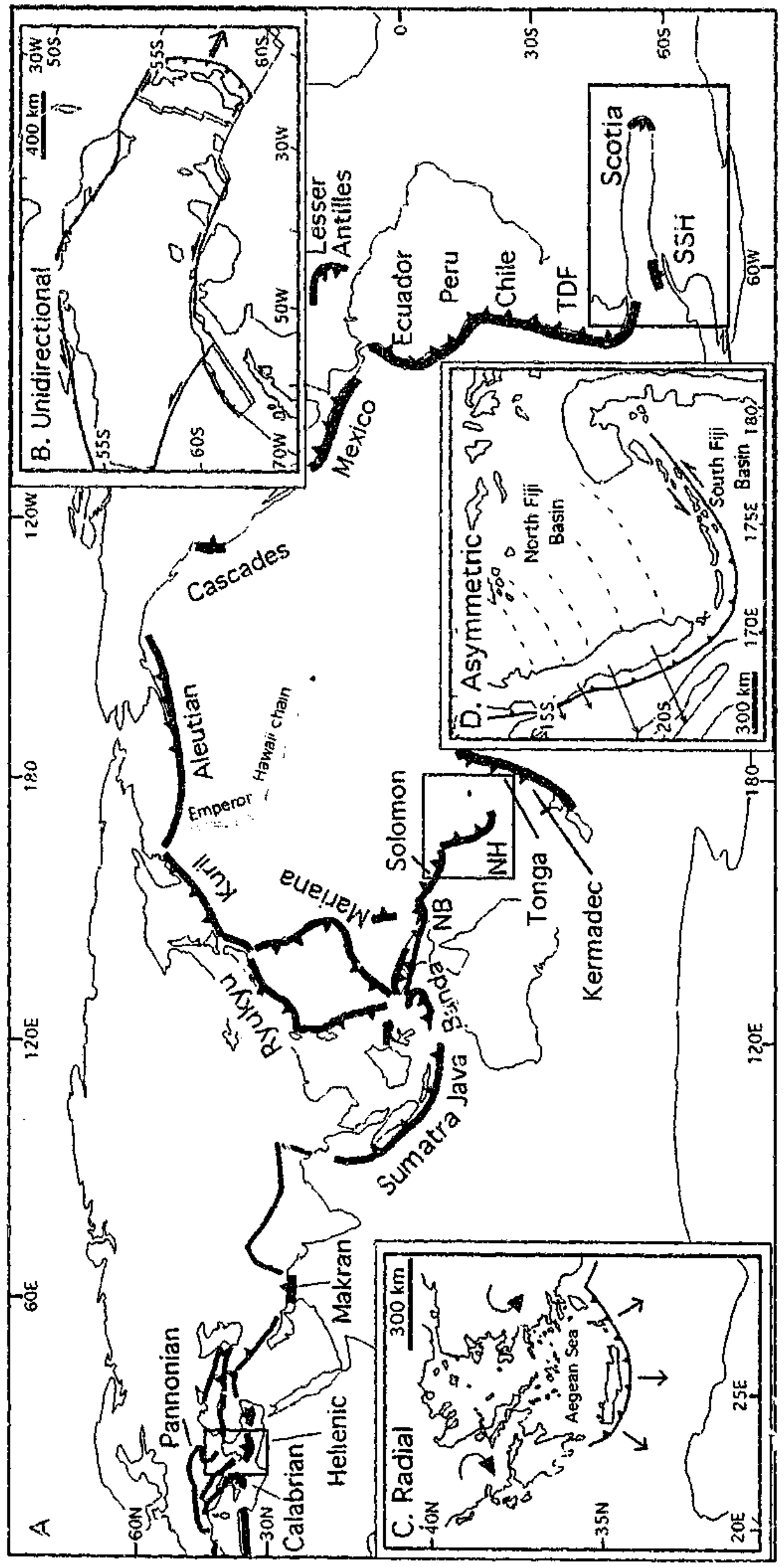


Figure 2.2. (a) Global occurrences of subduction systems and examples of rollback/back-arc systems as classified by Schellart et al. (2002a). (b) Unidirectional rollback in the Scotia arc (after Barker, 2001). (c) Radial subduction rollback in the Hellenic arc and associated block rotations (after Kissel and Laj, 1988). (d) Asymmetric (clockwise) subduction rollback in the New Hebrides arc and in the North Fiji back-arc basin (after Schellart et al., 2002c). NB = New Britain; NH = New Hebrides; SSH = South Shetland; TDF = Tierra del Fuego.

In what follows, I aim to show how these complex kinematic interactions can be incorporated in reconstruction modelling using a systematic framework of spatio-temporal constraints.

2.3. Reconstruction philosophy

The strategy to be employed in building the reconstruction model is based on determining a hierarchy of spatio-temporal constraints ranging from global plate kinematics to regional geological and geophysical criteria. The identification of such constraints provides the ability to translate large amounts of descriptive data into first-order to fourth-order constraints, which in turn, provide logical justifications for the reconstruction model.

The hierarchy of spatio-temporal constraints is based on the relative significance, accuracy and reliability of different datasets when used as kinematic indicators in a regional context. For example, data that are derived from magnetic isochron provide a relatively accurate insight into the motion path of major plates, and are therefore regarded as first-order constraints. In contrast, more local kinematic criteria, such as shear sense directions that are inferred from mineral lineations, are considered here as third- or fourth-order constraints, because they are subjected to large uncertainties and commonly refer to the motions of minor plates or allochthonous terranes.

First-order constraints are associated with the motions of major tectonic plates as indicated by the opening of ocean basins. Reconstructing the tectonic evolution of oceans is relatively straightforward due to the relatively little internal deformation that affected the oceanic lithosphere in comparison with the deformation in the continents (McKenzie, 1969). Furthermore, magnetic stripes (isochrons), formed during sea-floor spreading, document the spreading history of existing oceans and provide an accurate tool for reconstructing large-scale plate motions (see section 2.4.1 and Chapter 3).

Second-order constraints are related to motions of microplates or large continental blocks, such as Iberia and Adria in the Mediterranean region. Constraints on the motion of Adria (see Chapter 4) cannot be inferred from 'conventional' constraints (e.g. magnetic isochrons), and are therefore dependent on less robust criteria, such as palaeomagnetism and the opening history of small ocean basins (the Ionian Sea, in the case of Adria).

Third-order constraints are predominantly associated with the kinematic history of allochthonous terranes and the opening of small basins. These constraints incorporate large amounts of descriptive data that can be used to reconstruct the palaeogeography and the kinematic history of the study area. For example, third-order constraints can include spatio-temporal distributions of syn-rift sediments (see Chapter 6), directions of tectonic transports that are inferred from shear sense criteria (see Chapter 8) and crustal volume calculations.

Fourth-order constraints are based on mapping spatio-temporal variations in sutures and subduction zones. This is done, for example, by mapping fossilized magmatic arcs or by obtaining high-resolution spatio-temporal constraints on the occurrence of high-pressure/low-temperature metamorphism. Based on these data, the locus of subduction and collisional processes can be reconstructed.

During the procedure of reconstruction modelling, the utilisation of each hierarchal step provides boundary conditions for the next step. In other words, additional spatio-temporal constraints are progressively added to an existing framework of larger scale and more robust constraints. The consistency of the kinematic model with regional geological features is constantly verified, resulting in ongoing revisions and refinements of inaccurate or inadequate assumptions.

2.4. First-order constraints: the motion of major plates

2.4.1. Reconstruction of magnetic isochrons

The most robust kinematic criterion used to define plate motions is based on the reconstruction of magnetic isochrons and fracture zones. Pairs of magnetic anomalies of similar ages (isochrons) and fracture zones can be defined on both sides of oceanic spreading ridges, and their fitting back together shows the geometric configuration for the time at which the anomalies were formed (Pitman and Talwani, 1972; Dewey et al., 1973) (Figure 2.3). This procedure results in a series of finite rotations, which define the relative motion of plates on both sides of a spreading ridge. Direct application of this method can only describe the relative motions of diverging plates. In convergent margins, relative (convergence) motion can be calculated by a plate circuit technique using the motion of the converging plates relative to a third plate (e.g. Dewey et al., 1989; Rosenbaum et al., 2002a).

Application of the above method has been useful for determining motions of continents that have a passive margin boundary adjacent to oceanic crust with well-defined magnetic isochrons. For example, reconstruction of magnetic isochrons provides a framework for the implied plate motions around the Atlantic Ocean (e.g. Africa, Iberia, Europe, Greenland, North America; Figure 2.3).

In summary, motions inferred from reconstruction of magnetic isochrons can be used as boundary conditions while reconstructing orogenic belts. However, other criteria are required to constrain the motions of smaller microplates.

2.4.2. Plate motions relative to hotspots

The existence of linear chains of progressively younger volcanics corresponds to the motion of tectonic plates over an underlying mantle plume. Plate motions relative to a hotspot frame of reference can be calculated if hotspots are assumed to be fixed with respect to the deep mantle (Morgan, 1971). Evidence for relative motions between hotpots (Steinberger and O'Connell, 1998) indicates that the fixed-hotspot hypothesis is inaccurate. However, Global plate kinematic models that are based on hotspot track analysis are generally consistent with models that are based on magnetic anomalies (e.g. Morgan, 1983; Duncan and Richards, 1991; Müller et al., 1993), showing that movements of hotspots relative to each other are relatively small.

A major disadvantage using hotspot tracks in kinematic models is the fact that they are best preserved as submarine volcanoes in the oceans whereas tracks on the continents are seldom recognised. Thus, continental volcanism is of a little help in reconstructing plate motions and the few authors attempting to do so had to incorporate into their models additional kinematic constraints from oceanic hotspots (Garfunkel, 1992; Lawver and Müller, 1994). Furthermore, hotspot track analysis usually does not constrain the relative motions of converging plates, because this would require independent constraints for the motion of each plate relative to hotspots. In the case of Africa and Europe, for example, the existence of hotspot tracks on the African plate can be used to determine the relative motion of Africa with respect to hotspots (Morgan, 1983; O'Connor and Duncan, 1990; O'Connor et al., 1999; Rosenbaum et al., in review), but with the absence of hotspot tracks in Europe, the motion of Africa with respect to Europe cannot be independently determined.

In conclusion, kinematic parameters derived from hotspot track analysis can be used in conjunction with other techniques to infer plate motions. In most cases, however, hotspot analysis cannot independently constrain the motion of plates in convergent margins.

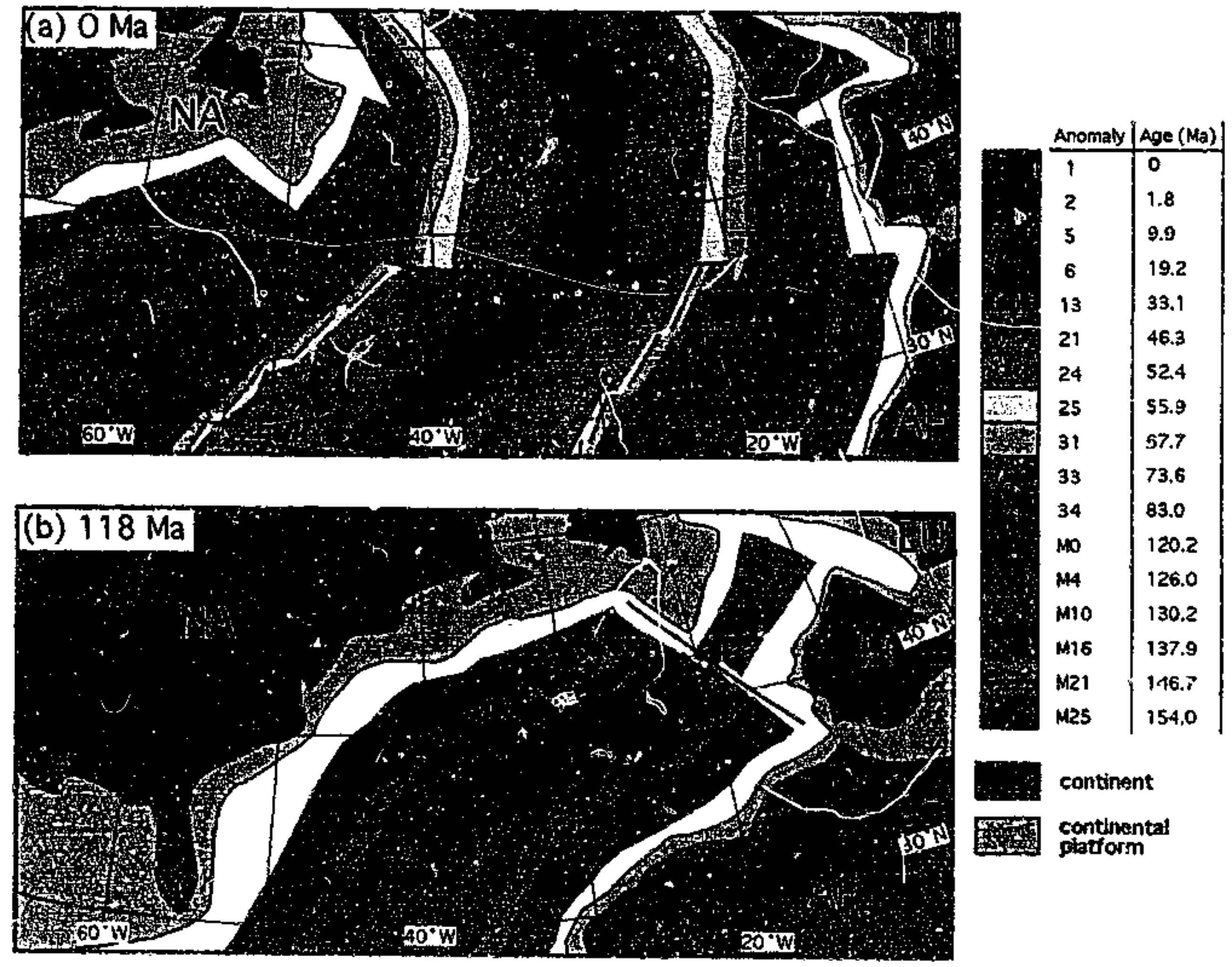
2.4.3. Palaeolatitudes and APWP

The application of palaeomagnetic techniques as first-order kinematic constraints is commonly used in Precambrian and Paleozoic reconstruction models (e.g. Torsvik et al., 1996; Meert, 2003). In such models, magnetic isochrons or oceanic hotspot tracks cannot be taken into account, because these features are only recognised in existing oceans, which are usually younger than ~200 Ma. Therefore, palaeomagnetic data are the most robust constraints for large-scale horizontal motions of the continents prior to the Mesozoic. In Mesozoic and Cenozoic reconstructions, palaeomagnetic constraints are rarely used as first-order constraints because of their large errors and uncertainties in comparison with available data derived from the reconstruction of magnetic isochrons (e.g. Besse and Courtillot, 1991). Another disadvantage of palaeomagnetic data is that they can only provide constraints on latitudinal displacements and cannot account for changes in longitudes.

Plate motions can be inferred from the Apparent Polar Wander Path (APWP), which is a plot of sequential positions of poles from a particular continent, representing the apparent motion of the rotation axis with respect to the continent of observation (Butler, 1992). Based on a similar APWP, for example, it has been suggested that Adria (the continental mass separating Africa and Europe in the central Mediterranean) has been attached to Africa at least since the early Mesozoic (Channell et al., 1979; Muttoni et al., 2001).

The measurement of palaeomagnetic inclinations, when corrected for the effect of tilting, can provide evidence for the latitude in which magnetic properties were acquired in a particular rock (palaeolatitudes). This approach can be used in tectonic reconstructions and is particularly powerful while reconstructing northward motions of Gondwana terranes in Peri-Tethyan domains (e.g. Adria, Anatolia, Iran, Afghanistan, etc). Palaeolatitudes are supposedly related to absolute latitudinal displacements, which can be compared with motions calculated from hotspot track analysis. These

two independent datasets can be used to test the consistency of kinematic models (see Chapter 4). Inferred palaeolatitudes, however, are subjected to large errors (typically of 5-10°), because each degree of uncertainty in the measurement of the inclination results in approximately 0.5° of uncertainty in the palaeolatitude.



(c)	Lat	Long	Angle
AF/NA	66.45	-20.19	-52.95
EU/NA	-69.27	-26.05	22.91
IB/NA	65.87	-18.33	-55.23
AF/EU	43.42	-7.14	-39.51

Figure 2.3. First order constraints on the motions of Africa (AF), North America (NA), Europe (EU) and Iberia (IB) as calculated from the reconstruction of magnetic isochrons in the Atlantic Ocean. (a) Recognised magnetic isochrons in the central part of the Atlantic Ocean (after Cande et al., 1989) (courtesy of R. Sutherland for providing the digital dataset); (b) Plate reconstruction at 118 Ma based on fitting of magnetic isochrons prior to the Cretaceous Normal Superchron. (c) Rotational parameters (latitudes and longitudes of Euler poles and rotational angles) determined from the reconstruction at 118 Ma.

2.5. Second-order constraints: minor plates and ocean basins

As mentioned above, the existence of well-defined magnetic isochrons in ocean basins provides robust constraints for calculating plate motions. This method can occasionally be used also for the motion of minor plates, as in the example of the motion of the Iberian microplate relative to Africa and Europe, which is calculated from magnetic isochrons in the Atlantic Ocean (Rosenbaum et al., 2002a) (Figure 2.3). However, this method cannot provide a framework for the motions of independent microplates throughout the Mediterranean region. There are oceanic domains with no recognised magnetic anomalies (e.g. the Ionian Sea; Rosenbaum et al., in review) or extensional basins that are floored by thinned continental lithosphere. There are also numerous oceanic domains that have been consumed during orogenesis and are presently exposed as ophiolitic complexes in the Alpine-Mediterranean region (Stampfli et al., 1998; Stampfli, 2000). The history of extension and opening of such basins (oceanic or continental) is the basis for the construction of second-order constraints.

An example of second-order constraints is the rotational parameters for the motion of the Corsica-Sardinia microplate with respect to Europe (Figure 2.4). This microplate underwent an independent motion with respect to Europe during the opening of the wedge shaped Ligurian Sea, which separates Corsica and Sardinia from the southern margin of Europe. Therefore, constraints

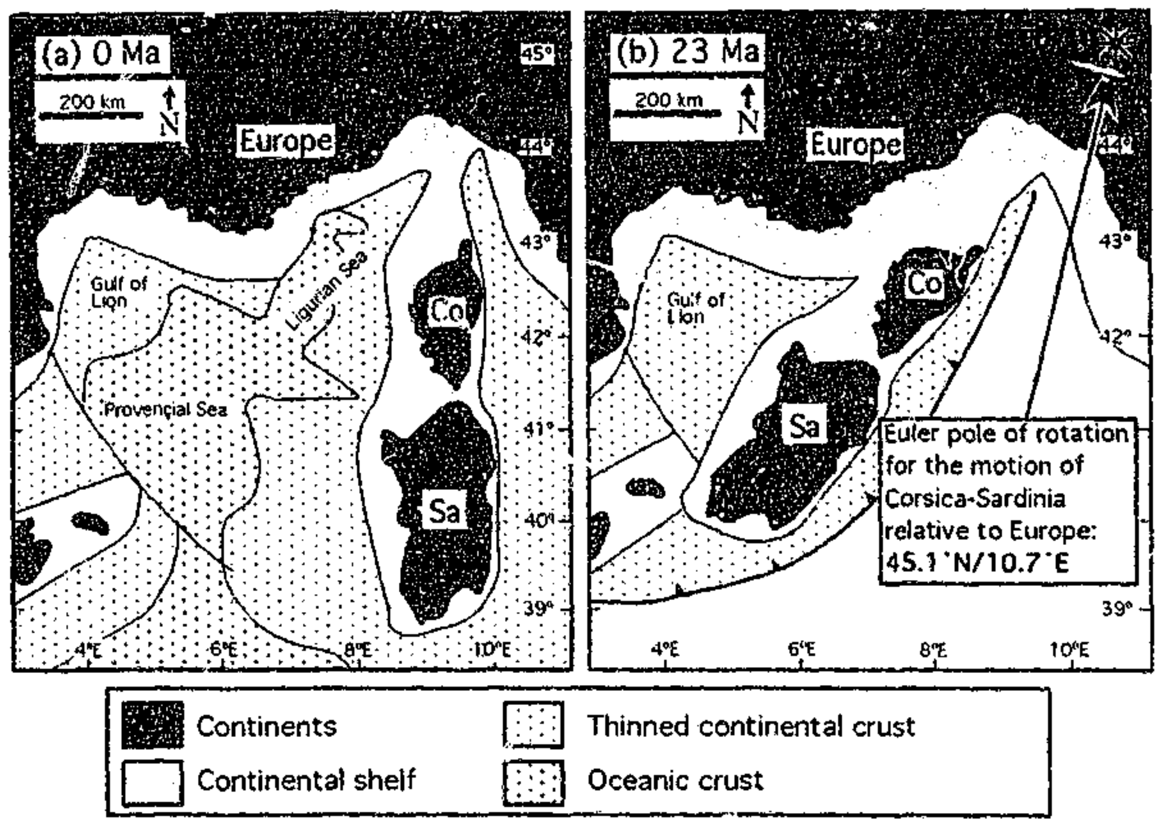


Figure 2.4. The determination of second-order constraints on the basis of the opening history of a small back-arc basin (the Ligurian Sea). (a) The present location of Corsica and Sardinia. (b) Early Miocene (23 Ma) reconstruction. The resulting rotational parameters for the motion of Corsica and Sardinia with respect to Europe (latitude 45.1°N; longitude 10.7°E; angle 30°CCW) are confirmed by palaeomagnetic results (see text).

on the opening of the Ligurian Sea can be rendered into second-order constraints on the motion of the Corsica-Sardinia microplate. There are various data sources that can be used to reconstruct the opening history of the Ligurian Sea; for example: (1) the structure of the passive margins (e.g. Rollet et al., 2002); (2) the spatio-temporal distribution of syn-rift sediments (e.g. Rehault et al., 1985); and (3) the sense of tectonic transport indicated by extensional structures (e.g. Jolivet et al., 1998). Based on these data, it is suggested that extensional tectonism commenced in the Ligurian Sea during the Oligocene (~30 Ma), and was associated with tectonic transport towards the southeast (Jolivet et al., 1998). In addition, palaeomagnetic evidence suggests that Corsica and Sardinia underwent ~30° of counterclockwise rotation in the time span between 22-16 Ma (Speranza et al., 2002). The incorporation of all these data together provides a relatively robust constraint for the position of Corsica and Sardinia during the Oligocene and the kinematics of the opening of the Ligurian Sea (Figure 2.4). The motion of Corsica and Sardinia relative to Europe can accordingly be defined by a Euler pole of rotation located in the Ligurian coast (45.1°N/10.7°E).

Spatio-temporal constraints related to the history of consumed ocean basins are more difficult to obtain and are subjected to large uncertainties. Consumed ocean basins are commonly preserved within orogenic belts, and are marked on the basis of the following features: (1) occurrences of ophiolitic complexes; (2) sedimentological evidence for the existence of deep-sea basins; (3) evidence for extensional structures related to rifting; and (4) the geochemical signature of basalts (e.g. Collins, 2002). In Figure 2.5, I present an example for the opening of the now-consumed Liguride Ocean, which existed in the western Mediterranean region during the Mesozoic and Early Tertiary and is presently exposed in ophiolitic complexes. In the example shown, the opening of this

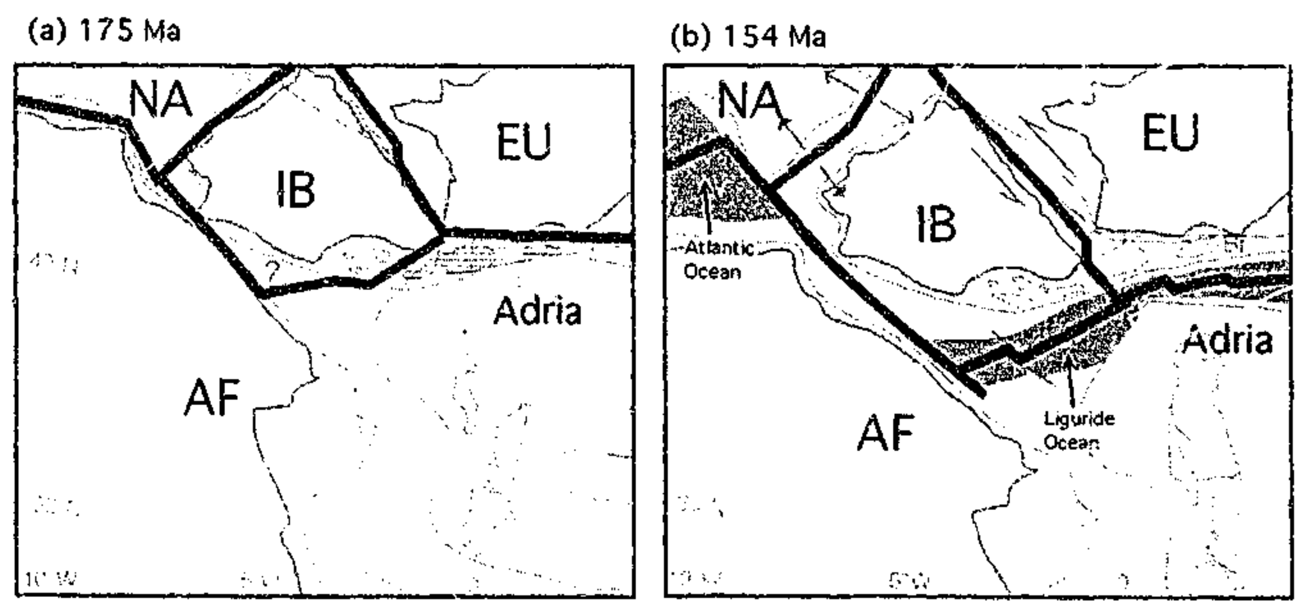


Figure 2.5. Constraints for the opening of the Liguride Ocean during the Jurassic based on the relative motions of Iberia (IB), Africa (AF) (including Adria), Europe (EU) and North America (NA) (after Rosenbaum et al., 2002a). (a) Middle Jurassic (175 Ma) reconstruction; (b) Late Jurassic (154 Ma) reconstruction. The geometry of the Liguride Ocean (shaded) is implied from the new area formed in the boundaries of AF-EU and AF-IB.

Middle Jurassic to Early Cretaceous ocean basin has been implied from the relative motions of North America, Africa (including an Adriatic promontory), Iberia and Europe (Rosenbaum et al., 2002a).

2.6. Third-order constraints: continental blocks and allochthonous terranes

Orogenic belts commonly consist of a collage of continental fragments, fragments of volcanic arcs or fragments of oceanic plateaux that have been accreted during orogenesis (Howell, 1995). In many cases, these fragments underwent large degrees of horizontal displacements prior to their accretion (hence the term *allochthonous terranes*). Thus, the challenge of tectonic reconstructions is to identify correlative terranes and to obtain adequate kinematic constraints with regard to the degree of their displacement.

Accreted terranes are separated from adjoining terranes by faults (or sutures) and are usually identified by a distinct tectonic, sedimentary, magmatic or metamorphic evolution with respect to the surrounding rocks. In the reconstruction process, correlative terranes are reconstructed using trial-and-error experiments with the aim to reconcile the unique geological properties of the accreted terranes with available kinematic evidence from structural geology, palaeomagnetism or paleontology (for latitudinal changes).

An example for accreted terranes in the western Mediterranean region is shown in Figure 2.6. These accreted terranes, which consist of parts of the Rif-Betic belt, the Kabylies massifs, Calabria and northeast Corsica, are now found in discrete localities throughout the western Mediterranean region. However, terrane analysis (e.g. Cohen, 1980; Rosenbaum et al., 2002b) suggests a common origin for all these terranes, which was associated with a contiguous Oligocene NE-SW-striking orogenic belt (Figure 2.6b).

The reconstruction of this Oligocene orogenic belt is consistent with third-order palaeomagnetic constraints (see also Chapter 5). These palaeomagnetic data do not provide sufficient information with regard to absolute displacements, but show significant evidence for post-Oligocene block rotations around vertical axes (Figure 2.6a). These rotations are counterclockwise in Corsica-Sardinia, the Betic cordillera, and the Apennines, and clockwise in Calabria, Sicily, the Balearic Islands and the Rif Mountains of northwest Africa (Rosenbaum et al., 2002b, and references therein).

2.7. Fourth-order constraints: subduction systems

As discussed earlier (in section 2.2), subduction zones are mobile elements that tend to move (or rollback) toward the opposite direction of the dip of the subducting slab (Figure 2.1). Such lateral variations through time in the location of subduction zones can be traced by the spatial and temporal distribution of calc-alkaline magmatism, because subduction processes commonly generate these magmas. Furthermore, the polarity of subduction zones (i.e. the dip of subduction) can be inferred

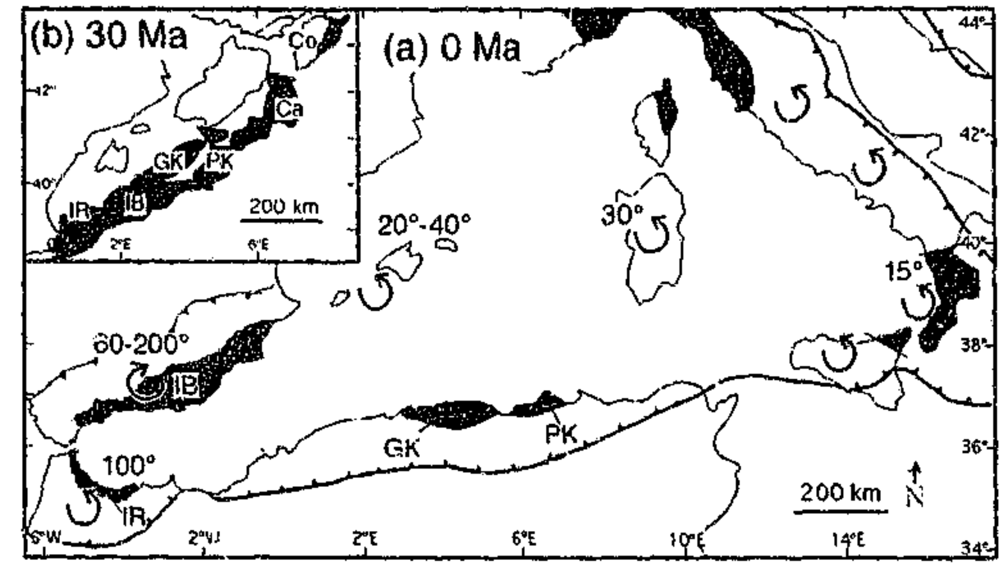


Figure 2.6. (a) Third-order constraints in the western Mediterranean region based on terrane analysis and palaeomagnetic block rotations. Shaded areas indicate allochthonous terranes with evidence for pre-Oligocene Alpine orogeny. (b) Reconstruction of the western Mediterranean allochthonous terranes based on third-order constraints showing an Oligocene NE-SW-striking orogenic belt (after Rosenbaum et al., 2002b). Ca, Calabria; Co, Corsica; IB, Internal Betic; IR, Internal Rif; PK, Petite Kabylie; GK, Petite Kabylie.

from the geochemical zoning of contemporaneous magmatism, with an increasing amount of alkaline composition further away from the subduction system (Keith, 1978).

In the western Mediterranean region, the spatio-temporal distribution of calc-alkaline magmatism shows radial outward younging of arc magmatism since the Oligocene (see Chapter 5). This evidence suggests rollback of a northwest-dipping subduction zone.

Seismic tomography has also been used to locate remnant subduction systems and to test reconstruction models (e.g. Bunge and Grand, 2000; Wortel and Spakman, 2000; Hall and Spakman, 2002; Rosenbaum and Lister, in review-b). The method is based on the assumption that subducted lithospheric slab is colder than the surrounding asthenosphere (McKenzie, 1970), and therefore subject to faster P- and S-wave velocities. These properties are taken into account in 3D tomographic models that show images of active subduction systems and remnant lithospheric slabs. Such images provide an admissibility check to tectonic reconstructions and can be compared with predicted kinematic and geodynamic features (see Chapters 6 and 7).

2.8. Conclusions

In this chapter, I have discussed various spatio-temporal constraints that can be used to reconstruct the tectonic evolution of orogenic belts. These constraints range from first-order constraints that are associated with the motion of major plates to lower-order constraints related to the opening of small back-arc basins, the motions of allochthonous terranes and the mobility of subduction systems. The evaluation of spatio-temporal constraints integrates various data sources and results in relatively realistic reconstruction models that are consistent with geological and geophysical data.

One of the most important aspects of this method is the ability to permit independent horizontal motions of numerous continental terranes in the reconstruction model. The reconstruction is not based only on plate tectonic rules with regard to the motion of major rigid plates. Furthermore, fixed positions of subduction systems are not assumed. This approach provides the means to account for the complex tectonic interactions that take place during the evolution of convergent plate margins.

The chapter illustrated several examples for the utilisation of spatio-temporal constraints and described the hierarchical structure of the reconstruction method that is used in the forthcoming chapters to reconstruct tectonic processes in the western Mediterranean region. Adequate spatio-temporal constraints from multi-disciplinary data sources provide a systematic approach for reconstructing the tectonic evolution of complex zones. This approach increases the resolution of tectonic reconstructions and provides insights into the occurrences and the effects of particular events, such as slab tearing, slab segmentation and formation of cusps in subduction systems.

CHAPTER 3

RELATIVE MOTIONS OF AFRICA, IBERIA AND EUROPE DURING ALPINE OROGENY

Then the Iberian Peninsula moved a little further, one metre, two metres, just to test its strength. The ropes which served as evidence, strung from one side to the other, as firemen do with walls that develop cracks and threaten to cave in, broke like ordinary string, some that were stronger uprooted the trees and posts to which they were tied. Then there was a pause, a great gust of air, like the first deep breathing of someone awakening, and the mass of stone and earth, covered with cities, villages, rivers, woodlands, factories, wild scrub, cultivated fields, with all their inhabitants and livestock, began to move, a ship drawing away from harbour and heading out to an unknown sea one more.

José Saramago, The Stone Raft (1986)
(translated from Portuguese by Giovanni Pontiero)

FOREWORD: CHAPTER 3

This chapter provides constraints for large-scale plate motions during Alpine orogeny, namely the motions of Africa, Iberia and Europe. The approach used here, inspired by the work of Dewey et al. (1989), is based on calculating rotation parameters from the spreading history of the Northern Atlantic Ocean. The work benefited from discussions with Gordon Lister, and the kinematic analysis was facilitated by the PlatyPlus software package developed by Cecile Duboz and Gordon Lister. Earlier versions of this chapter have been reviewed by Rob Van der Voo, Shiri Srivastava, Wouter Schellart, Dave Giles and Mara Pavlidis. This chapter has been slightly modified, based on a paper published in *Tectonophysics*[#].

[#] Rosenbaum, G., Lister, G.S. and Duboz, C., 2002. Relative motions of Africa, Iberia and Europe during Alpine orogeny. *Tectonophysics*, 359: 117-129.

3. RELATIVE MOTIONS OF AFRICA, IBERIA AND EUROPE DURING ALPINE OROGENY

Abstract

A revised kinematic model for the motions of Africa and Iberia relative to Europe since the Middle Jurassic is presented in order to provide boundary conditions for Alpine-Mediterranean reconstructions. These motions were calculated using up-to-date kinematic data predominantly based on magnetic isochrons in the Atlantic Ocean and published by various authors during the last 15 years. It is shown that convergence of Africa with respect to Europe commenced during the Cretaceous Normal Superchron (CNS), between chrons M0 and 34 (120-83 Ma). This motion was subjected to fluctuations in convergence rates characterised by two periods of relatively rapid convergence (during Late Cretaceous and Eocene-Oligocene times) that alternated with periods of slower convergence (during the Paleocene and since the Early Miocene). Distinct changes in plate kinematics are recognised in the motion of Iberia with respect to Europe, indicated by: (1) a Late Jurassic–Early Cretaceous left-lateral strike-slip motion; (2) Late Cretaceous convergence; (3) Paleocene quiescence; (4) a short period of right-lateral strike-slip motion; and (5) final Eocene-Oligocene convergence. Based on these results, it is speculated that a collisional episode in the Alpine orogeny at ca. 65 Ma resulted in a dramatic decrease in the relative plate motions and that a slower motion since the Early Miocene promoted extension in the Mediterranean back-arc basins.

3.1. Introduction

The relative motions of Africa, Iberia and Europe have provided a kinematic framework for numerous tectonic reconstructions of the Alpine-Mediterranean region (e.g. Smith, 1971; Dewey et al., 1973; 1989; Biju-Duval et al., 1977; Dercourt et al., 1986; Savostin et al., 1986; Stampfli et al., 1998; Wortmann et al., 2001). In these reconstructions, the motion of Africa with respect to Europe is usually based on rotation parameters (poles of rotation and rotation angles) as calculated, for example, by Dewey et al. (1973; 1989). In the latter, convergence is considered to commence in the Late Cretaceous and has been accommodated since by subduction processes, continental collision and lithospheric deformation. The accuracy of these data is extremely important in order to set boundary conditions for future reconstructions, and it is the aim of this chapter to provide revised kinematic constraints for these motions based on a comprehensive list of up-to-date rotation parameters.

The opening of the Atlantic Ocean at the expense of the disappearing Tethys Ocean was deduced in early reconstructions based on the least-square fitting of continental platforms across the Atlantic (Carey, 1958; Bullard et al., 1965). These reconstructions, however, only provided constraints for the positions of the plates prior to the opening of the Atlantic Ocean and could not provide a time-dependent incremental path that described the relative motions of the surrounding plates. The latter was made possible in the pioneering work of Pitman and Talwani (1972), who used fracture zones and magnetic anomalies on both side of the Atlantic spreading ridge to reconstruct the kinematic evolution of the Atlantic Ocean. Their work provided sets of a limited number of poles of rotation (Euler poles) and their finite rotation angles for the motions of Africa and Europe relative to North America.

The occurrence of linear magnetic anomalies is limited to divergent plate boundaries where sea floor spreading and formation of new oceanic crust took place. Therefore, the motions of Africa and Iberia with respect to Europe cannot be directly determined but are calculated instead by using published or derived motions of Africa, Europe and Iberia with respect to North America (Dewey et al., 1973; 1989; Savostin et al., 1986; Mazzoli and Helman, 1994). With the emergence of new data that better constrain the evolution of the Atlantic Ocean, these motions have been revised and modified (Srivastava and Roest, 1989; 1996; Lawver et al., 1990; Srivastava et al., 1990a; 2000; Roest and Srivastava, 1991; Roest et al., 1992; Srivastava and Verhoef, 1992; Torsvik et al., 2001), and it is therefore necessary to incorporate these additional data in a revised kinematic model for the relative motions of Africa, Iberia and Europe.

This chapter presents revised rotation parameters for the motions of Africa and Iberia with respect to Europe. In the scope of this chapter, the complex kinematics of microplates within the Alpine orogeny is not discussed. Rather, I aim to provide a framework for the implied motions around the Mediterranean as obtained from plate kinematics of the North Atlantic. These motions can provide kinematic boundary conditions for more detailed reconstructions of the Alpine-Mediterranean region.

3.2. Methodology

Kinematic analysis has been performed using PlatyPlus, a software package developed at the Australian Crustal Research Centre (School of Geosciences) at Monash University. PlatyPlus provides an interactive platform for plate reconstruction by reading and applying motion files (with either absolute or relative motions) and by interpolating rotation parameters at any required stage. PlatyPlus files are found in Appendix 1, and PlatyPlus movies related to the relative motions of the plates in the Northern Atlantic Ocean are found in Appendix 2.

Kinematic data applied in this study have been derived from published rotation parameters predominantly based on geometrical fits of linear magnetic anomalies and fracture zones (Klitgord and Schouten, 1986; Srivastava and Roest, 1989; 1996; Lawver et al., 1990; Müller et al., 1990; Srivastava et al., 1990a; 2000; Roest et al., 1992; Royer et al., 1992; Srivastava and Verhoef, 1992;

Torsvik et al., 2001) (Table 3.1). These data describe the relative motions of Europe, Iberia and Africa with respect to North America. Unfortunately, only a limited number of magnetic anomalies are recognised in the Atlantic Ocean. No information exists for anomalies younger than anomaly 5 (9.9 Ma) and the oldest recognised magnetic anomaly in the central Atlantic is M25 (154 Ma). A significant problem arises from the existence of a quiet period with no magnetic reversals during the Cretaceous Normal Superchron (CNS), between chron 34 (83 Ma) and chron M0 (120.2 Ma). In Table 3.1, the only rotation parameters from this period are derived from extrapolation of the motion of Europe with respect to North America at 92 Ma (the assumed age of the opening of Labrador Sea) based on the rate of spreading between anomalies 31 and 34 (67-83 Ma) (Srivastava and Roest, 1989).

The motion of Africa and Iberia relative to Europe was calculated using a plate circuit analysis with a fixed European coordinate system (Dewey et al., 1973). For example, the rotation of Africa relative to Europe at time t can be expressed as:

$$_{AF}ROT_{EU}(t) = {}_{AF}ROT_{NA}(t) + {}_{NA}ROT_{EU}(t)$$

Where $_{A}ROT_{B}(t)$ is the rotational parameters (latitude, longitude and angle) of plate A relative to B at a given time.

Ages of magnetic anomalies taken here are the youngest ages (initiations) of the geomagnetic chrons after Gradstein et al. (1994), Channell et al. (1995a), Cande and Kent (1995) and Huestis and Acton (1997) (Table 3.2).

3.3. Motion of Africa with respect to Europe

Poles of rotation and rotation angles for the motion of Africa relative to Europe are summarised in Table 3.3 along with rotation parameters derived from the models of Dewey et al. (1989) (for 170-25 Ma) and Mazzoli and Helman (1994) (for 25-0 Ma). This motion is also shown as trajectories and convergence rates of sets of three points moving with Africa as a function of time (Figure 3.1).

The motion of Africa with respect to Europe during the Late Jurassic and Early Cretaceous is characterised by left-lateral strike-slip motion (Figure 3.1a). According to our results, this motion changed to relative convergence between chrons M0 and 34, which is a long period with no magnetic reversals (Cretaceous Normal Superchron). Using these data alone, it is therefore not possible to determine when convergence commenced. Dewey et al. (1989) suggested that convergence commenced at 92 Ma based on Late Cretaceous ages obtained in early geochronological studies of high-pressure rocks from the western Alps (Bocquet et al., 1974). However, this assumption is rather weakened by much younger (Cenozoic) ages of high-pressure metamorphism in the Alps reported in more recent contributions (e.g. Duchêne et al., 1997; Gebauer et al., 1997; Rubatto and Hermann, 2001). Therefore, geological evidence from the western Alps is not considered in this model to support kinematic changes during the CNS.

Table 3.1. Euler poles of rotation and finite rotation angles as inferred from fitting of magnetic isochrons.

Magnetic anomaly	Age (Ma)	Latitude	Longitude	Rotation	Reference
Africa with respect to North America					
5	9.9	80.12	50.80	-2.52	Müller et al. (1990)
6	19.2	81.07	56.51	-5.21	Srivastava et al. (1990a)
13	33.1	75.37	1.12	-10.04	Müller et al. (1990)
21	46.3	75.30	-3.88	-15.25	Müller et al. (1990)
24	52.4	78.33	-2.64	-16.91	Müller et al. (1990)
25	55.9	79.68	-0.46	-18.16	Müller et al. (1990)
30	65.6	82.90	4.94	-20.76	Müller et al. (1990)
31	67.7	82.51	-0.63	-20.96	Klitgord and Schouten (1986)
32	71.1	81.35	-9.15	-22.87	Klitgord and Schouten (1986)
33 (young)	73.6	80.76	-11.76	-23.91	Klitgord and Schouten (1986)
33 (old)	79.1	78.30	-18.35	-27.06	Roest et al. (1992)
34	83.0	76.55	-20.73	-29.60	Klitgord and Schouten (1986)
M0	120.2	66.09	-20.18	-54.45	Srivastava et al. (1990a)
M4	126.0	65.97	-19.43	-56.63	Roest et al. (1992)
M10	130.2	65.95	-18.50	-57.40	Klitgord and Schouten (1986)
M11	131.1	66.14	-18.72	-58.03	Roest et al. (1992)
M16	137.9	66.24	-18.33	-59.71	Roest et al. (1992)
M21	146.7	66.24	-18.33	-62.14	Roest et al. (1992)
M25	154.0	66.70	-15.85	-64.90	Roest et al. (1992)
	170.0	67.02	-13.17	-72.10	Klitgord and Schouten (1986)
	175.0	65.97	-12.76	-76.44	Srivastava et al. (1990a)
Europe with respect to North America					
5	9.9	65.38	133.58	-2.44	Lawver et al. (1990)
6	19.2	68.92	136.74	-4.97	Lawver et al. (1990)
13	33.1	65.64	136.95	-7.51	Lawver et al. (1990)
21	46.3	66.15	135.40	-10.87	Srivastava and Roest (1996)
24	52.4	63.89	139.27	-12.89	Srivastava and Roest (1996)
25	55.9	63.14	141.66	-14.22	Srivastava and Roest (1989)
30	65.6	64.84	143.96	-16.95	Srivastava and Roest (1989)
33 (old)	79.1	66.17	147.74	-19.00	Srivastava and Roest (1989)
34	83.0	66.54	148.91	-19.70	Srivastava and Roest (1989)
Labrador Sea	92	66.67	150.26	-20.37	Srivastava and Roest (1989)
M0	120.2	69.67	154.26	-23.17	Srivastava et al. (2000)
M25	154.0	69.03	155.44	-23.26	Torsvik et al. (2001)
	170	69.1	156.70	-23.54	Royer et al. (1992)
	175	71.61	156.70	-25.27	Torsvik et al. (2001)

Latitudes and longitudes are in degrees with positive values for N and E
Rotation angles are in degrees with positive values for clockwise rotations.

Table 3.1. Continued.

Magnetic anomaly	Age (Ma)	Latitude	Longitude	Rotation	Reference
Iberia with respect to North America					
5	9.9	65.38	133.58	-2.44	Lawver et al. (1990)
6	19.2	68.00	138.20	-4.75	Srivastava et al. (1990a)
13	33.1	76.34	117.33	-7.98	Srivastava et al. (1990a)
21	46.3	74.70	126.96	-11.05	Srivastava et al. (1990a)
24	52.4	72.98	133.28	-12.94	Srivastava et al. (1990a)
25	55.9	73.29	133.88	-14.25	Srivastava et al. (1990a)
31	67.7	74.96	135.34	-17.19	Srivastava et al. (1990a)
33 (old)	79.1	85.49	110.28	-22.41	Srivastava et al. (1990a)
34	83.0	87.18	57.43	-24.67	Srivastava et al. (1990a)
M0	120.2	64.71	-18.94	-58.11	Srivastava et al. (2000)
M25	154.0	66.90	-12.93	-60.45	Srivastava et al. (1990a)
	175	65.72	-12.82	-66.32	Srivastava and Verhoef (1992)

Table 3.2. Ages of magnetic anomalies used in calculations of plate motions in this chapter.

Anomaly	Age (Ma)	Ref.	Anomaly	Age (Ma)	Reference
5	9.9	Huestis and Acton (1997)	33 (old)	79.1	Cande and Kent (1995)
6	19.2	Huestis and Acton (1997)	34	83.0	Cande and Kent (1995)
13	33.1	Huestis and Acton (1997)	M0	120.2	Gradstein et al. (1994)
21	46.3	Cande and Kent (1995)	M4	~126.0	Channell et al. (1995a)
24	52.4	Cande and Kent (1995)	M10	130.2	Gradstein et al. (1994)
25	55.9	Cande and Kent (1995)	M11	131.1	Gradstein et al. (1994)
30	65.6	Cande and Kent (1995)	M16	137.9	Gradstein et al. (1994)
31	67.7	Cande and Kent (1995)	M18	142.5	Gradstein et al. (1994)
32	71.1	Cande and Kent (1995)	M21	146.7	Gradstein et al. (1994)
33 (young)	73.6	Cande and Kent (1995)	M25	154.0	Gradstein et al. (1994)

Ages indicate the initiation of magnetic chrons.

A prominent feature recognised in Figure 3.1b is two periods of rapid convergence: between chrons M0 and 31 (120-67 Ma), and between chrons 24 and 6 (52.4-19.2 Ma). These periods are terminated by rapid decreases in convergence rates, which are particularly evident at 67-65 Ma when convergence basically ceased for a period of 10-15 m.y. Convergence recommenced during chron 25 (~55 Ma) and seemed to reach its maximum rate during Eocene-Oligocene times. Subsequently, convergence has significantly decreased and remained relatively slow at least since chron 6 (19 Ma).

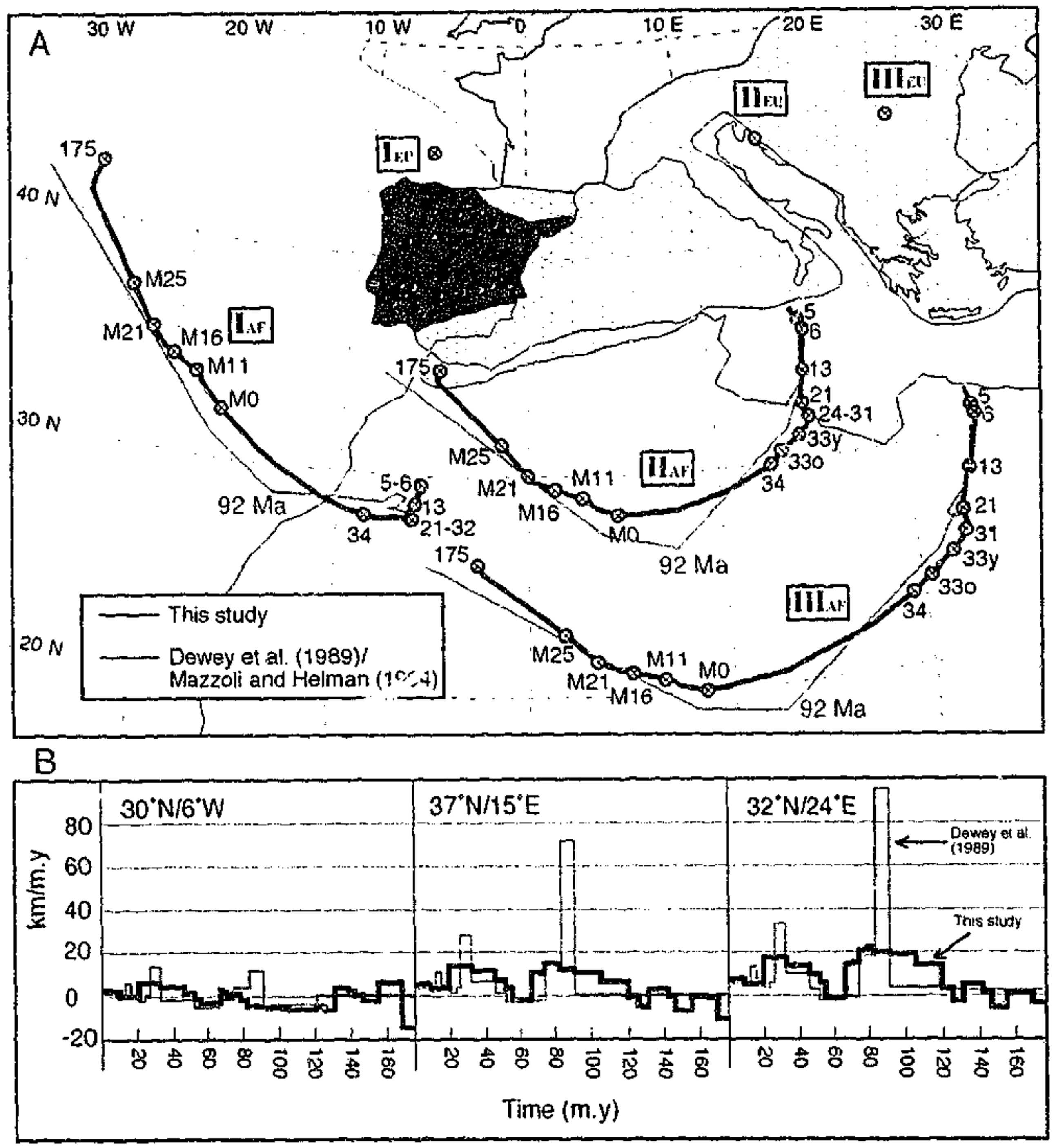


Figure 3.1. (a) Trajectories of three points in Africa relative to fixed points in Europe (I_{AF}=30°N/6°W; I_{EU}=45°N/6°W; II_{AF}=37°N/15°E; II_{EU}=45°N/15°E; III_{AF}=32°N/24°E; III_{EU}=45°N/24°E) plotted as a function of time. Stippled area indicates regions of Mesozoic and Cenozoic deformation. (b) Calculated convergence rates of I, II and III. Compared trajectories and convergence rates after Dewey et al. (1989) and Mazzoli and Helman (1994) are also shown (Grey lines).

Table 3.3. Euler poles for the motion of Africa relative to Europe.

Time (Ma)	This study			Dewey et al. (1989)[1] and Mazzoli and Helman (1994)[2]		
	Latitude	Longitude	Rotation	Latitude	Longitude	Rotation
5	-13.99	158.23	0.55	-8.13	165.07	0.51
10	-13.91	158.94	1.09	-3.15	164.83	1.03
15	-14.77	162.89	1.53	-10.99	161.26	1.67
20	-17.27	164.31	2.13	-21.71	162.41	2.44
25	-24.32	161.90	3.57	-26.84	164.01	3.06
30	-27.24	161.41	5.04	-28.48	161.07	4.70
35	-28.89	160.77	6.41	-29.21	159.45	6.47
40	-30.10	160.57	7.60	-31.87	161.53	7.90
45	-30.96	160.61	8.80	-33.91	163.36	9.31
50	-30.02	160.71	9.61	-34.72	161.63	10.46
55	-28.83	161.90	10.39	-33.24	165.32	10.73
60	-28.73	163.02	10.58	-32.44	166.98	10.97
65	-28.82	163.93	10.64	-31.75	168.71	11.21
70	-29.70	163.75	11.87	-32.43	169.41	11.43
75	-32.66	164.46	13.84	-33.67	169.43	11.91
80	-35.02	165.10	16.46	-34.37	168.77	13.43
85	-36.88	166.30	19.35	-36.21	168.26	16.98
90	-38.68	167.70	22.56	-39.77	168.05	28.77
95	-40.01	168.88	25.77	-41.29	168.60	34.39
100	-41.03	169.89	28.99	-42.51	169.56	35.89
105	-41.85	170.77	32.23	-43.62	170.48	37.41
110	-42.51	171.59	35.12	-44.65	171.36	38.95
115	-43.09	172.35	37.95	-45.59	172.20	40.51
120	-43.60	173.04	40.78	-46.47	173.01	42.07
125	-44.46	173.71	42.65	-47.27	173.79	43.65
130	-44.94	174.58	43.75	-48.02	174.54	45.24
135	-46.03	174.82	45.25	-48.71	175.25	46.83
140	-46.66	175.09	46.50	-49.36	175.95	48.44
145	-47.22	175.28	47.81	-49.96	176.61	50.05
150	-48.06	176.43	49.29	-50.53	177.26	51.67
155	-48.99	177.90	50.98	-51.06	177.88	53.30
160	-49.81	178.93	53.00	-51.56	178.48	54.93
165	-50.56	179.92	55.03	-52.02	179.06	56.57
170	-51.26	180.88	57.07	-52.46	179.62	58.21
175	-50.57	182.00	59.82	-52.88	180.16	59.86

(1) Dewey et al. (1989) for 175-25 Ma

(2)Mazzoli and Helman (1994) for 25-0 Ma.

3.4. Iberia

The motion of the Iberian microplate during the opening of the Atlantic Ocean has been previously discussed by Srivastava et al. (1990a; 1990b) and Roest and Srivastava (1991). Here we present calculated rotation parameters for the motion of Iberia relative to Europe (Table 3.4) and trajectories of the motion of two points in Iberia relative to fixed points in Europe (Figure 3.2). These results are generally in agreement with the kinematic analysis described in Roest and Srivastava (1991).

Table 3.4. Rotation parameters for the motion of Iberia relative to Europe

Anomaly	Age (Ma)	Latitude	Longitude	Rotation
5	9.9	0	0	0
6	19.2	77.93	59.14	0.24
13	33.1	-31.21	166.79	1.73
21	46.3	-23.85	157.12	1.72
24	52.4	-21.60	157.88	2.10
25	55.9	-20.72	162.40	2.61
30	65.6	-12.95	165.77	3.02
31	67.7	-16.45	167.49	3.10
33 (old)	79.1	-37.17	169.00	8.04
34	83.0	-38.86	169.85	10.28
	92.0	-42.64	173.20	16.56
M0	120.2	-43.86	174.17	44.77
M18	142.5	-46.19	177.47	45.91
M25	154.0	-47.12	179.45	46.29
	170.0	-47.55	180.35	50.62
	175.0	-46.80	181.10	50.33

The reconstruction shows that during Middle Jurassic – Early Cretaceous (170-120 Ma) the plate boundary between Iberia and Europe accommodated more than 200 km of left-lateral strike-slip motion (Figure 3.3a-d). This motion commenced during rifting of Iberia from the Grand Banks, prior to the initiation of sea floor spreading in this region (see Srivastava and Verhoef, 1992). A major change in plate kinematics occurred with the onset of sea floor spreading in the Bay of Biscay sometime during the CNS (120-83 Ma), leading to left-lateral strike-slip motion in the Pyrenees and approximately 115 km of convergence in more easterly parts of the 'Greater' Iberian microplate (e.g. Sardinia and Corsica; Figure 3.3e). The relative motion of Iberia and Europe changed to wholesale convergence at chron 34 (83 Ma; Figure 3.3e,f), whereas after chron 31 Iberia stopped moving with respect to Europe for a period of 10-15 m.y (until chron 25; Figure 3.2). An independent motion of Iberia resumed during the Eocene, identified by a right-lateral strike-slip motion with a total displacement of 60-70 km during anomalies 24-21 (55-46 Ma; Figure 3.3g) followed by final convergence until the Oligocene (see also Figure 3.3h; Roest and Srivastava, 1991).

The above kinematic evolution of Iberia with respect to Europe is also supported by geological evidence that implies left-lateral strike-slip displacement along the North Pyrenean Fault at least since 110 Ma (Montigny et al., 1986; Costa and Maluski, 1988) and a change to transpressional regime at ca. 85 Ma (Puigdefabregas and Souquet, 1986; De Jong, 1990). Soula et al. (1986) have reported dextral offsets along oblique mylonites from the Pyrenees, which are found in Late Cretaceous rocks and are unconformably overlain by Eocene rocks. This may suggest that right-lateral strike-slip motion occurred in the time span between the Cretaceous and the Eocene as also implied from my reconstruction. Geological evidence for the final stage of convergence in the Pyrenees is found in successive Eocene to Early Oligocene thrust sheets and piggyback basins (e.g. Puigdefabregas and Souquet, 1986). Crustal balanced cross-sections in the Pyrenees show a decreasing amount of shortening from east to west (Seguret and Daignieres, 1986) with at least 147 km of minimum shortening in the central Pyrenees (Muñoz, 1992; Fitzgerald et al., 1999). This is comparable to an approximately 160-170 km of total convergence between Iberia and Europe since chron 34 calculated for points located in the central Pyrenees (around longitude 1°E; Figure 3.2).

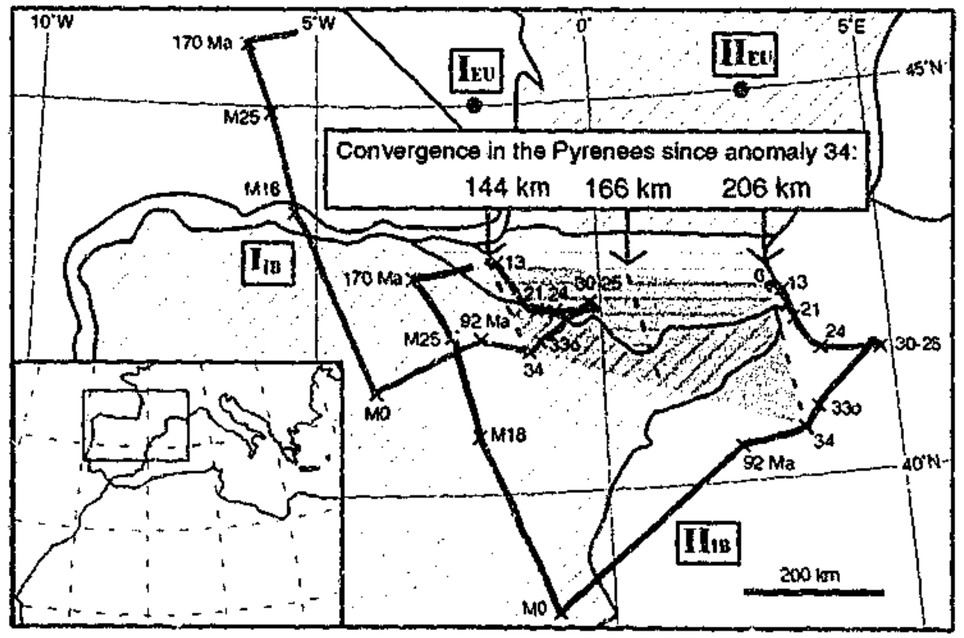


Figure 3.2. Trajectories of two points in Iberia relative to fixed points in Europe plotted as a function of time (I_{IB}=43°N/2°W; I_{EU}=45°N/2°W; II_{IB}=42.5°N/3°E; II_{EU}=45°N/3°E). Shaded area and dashed lines indicates the amount of convergence in the Pyrenees since anomaly 34 (83 Ma).

3.5. Tectonic implications

Figure 3.3 shows a schematic tectonic reconstruction of the western Tethys using the above motions of Africa and Iberia as well as additional information from Stampfli et al. (1998; 2001a). In this reconstruction, Europe is fixed and the Adriatic plate is attached to Africa as implied from palaeomagnetic studies (e.g. Channell, 1996).

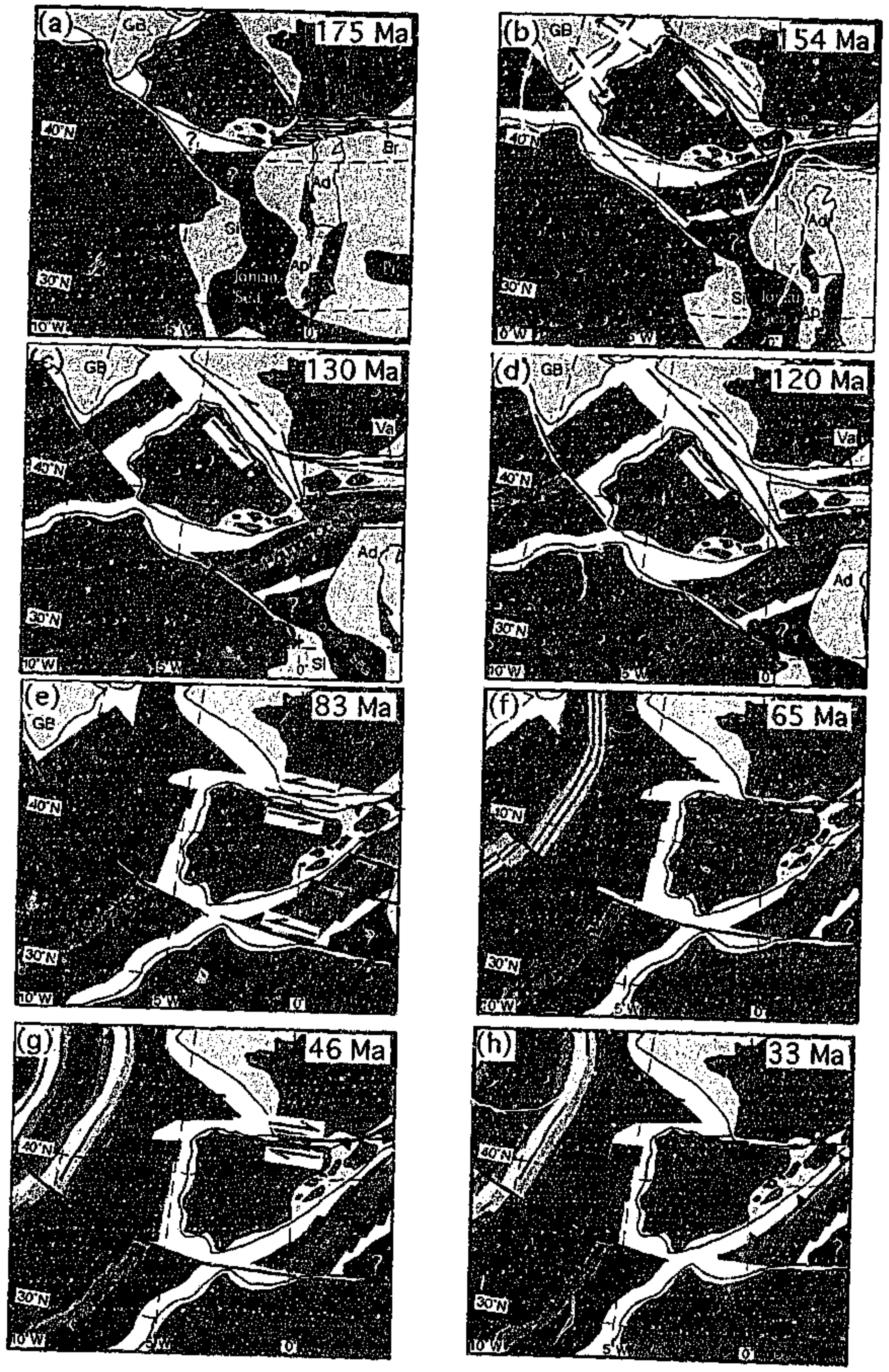


Figure 3.3. Reconstruction of the western Tethys since Middle Jurassic using rotation parameters for Africa, Iberia and Europe. Additional information on Alpine Tethys is partly modified after Stampfli et al. (1998; 2001a). Isochrons in the Atlantic Ocean are after Cande et al. (1989) (courtesy of R. Sutherland for providing the digital dataset) except of M11 off the Iberian and the Grand Banks coasts, which is after Srivastava et al. (2000). Shaded areas mark overlaps of adjacent plates. Abbreviations are: Ad = Adria; Ap = Apulia; Br = Briançonnais; GB = Grand Banks; Pi = Pindos; Si = Sicily; Va = Valais.

The reconstruction of the western Tethys shows a relatively good fit of the continents since the Middle Jurassic using the revised kinematic constraints. This is particularly important for the Adriatic promontory, which overlapped with southern France and eastern Spain in earlier reconstructions (compare figure 2 in Wortmann et al., 2001 with Figure 3.3a). A minor overlap exists between northern Adriatic terranes (Southern Alps and Sesia Zone) and eastern 'Great' Iberian terranes (Corsica, Sardinia, and Kabylies). This geometric problem was possibly a direct result of an assumption (which was not necessarily justified) that these terranes had a rigid connection with Iberia and Adria. The overlaps seen in the western margins of Iberia at 175 Ma (Figure 3.3a) can be explained by the effect of syn-rift stretching prior to spreading in the northern Atlantic (Srivastava and Verhoef, 1992).

The reconstruction shows that the kinematic constraints considered here are sufficient to explain the opening of the Liguride Ocean during the Late Jurassic (Figure 3.3b,c). Nevertheless, using these motions alone it is difficult to resolve the tectonic interactions between different terranes in the Alps (i.e. Briançonnais, Valais, Sesia-Lanzo Zone), which possibly acted as independent microplates during Alpine orogeny.

3.6. Discussion

3.6.1. Commencement of Africa-Europe convergence

A key issue revisited in this chapter, previously discussed by Dewey et al. (1989), considers when convergence between Africa and Europe commenced. The data suggest that this occurred sometime between chrons M0 and 34, when the motion of both Africa and Iberia with respect to Europe changed from overall left-lateral strike-slip to convergence. However, the existence of a quiet magnetic period at this time (the CNS) does not enable precise estimation of the age of the kinematic change. Dewey et al. (1989) suggested a swerve in the plate motion at 92 Ma based on the assumption that a change in the plate kinematics would mark the onset of Alpine deformation in Europe. Figure 3.1b shows that this would have resulted in rapid convergence rates (in excess of 90 km/m.y. in the eastern Mediterranean), but only during a relatively short period between 92-83 Ma.

The issue in respect to Dewey et al's (1989) interpretation concerns the timing of onset of Alpine collision. As indicated, eclogites and blueschists from the western Alps, previously considered to represent a Cenomanian-Turonian (100-90 Ma) metamorphic event, now reveal Cenozoic metamorphic ages using various dating techniques (Duchêne et al., 1997; Gebauer et al., 1997; Rubatto and Hermann, 2001). This evidence does not contradict the possibility that rocks in the western Alps were also subjected to high-pressure metamorphism in earlier times. There is simply no agreement as to when these events took place. However, metamorphism may have occurred as early as ~120 Ma (Paquette et al., 1989; Monié and Chopin, 1991) suggesting that subduction and collisional processes took place earlier than previously thought. This may imply even earlier commencement of convergence between Africa and Europe.

The attempt to correlate orogenic processes with motions of the far-field plates is further complicated if we consider the limits of our current knowledge as to how orogens work. Recent reconstructions of the Tethyan belt (e.g. Hall, 2002; Stampfli and Borel, 2002) show the incorporation of a large number of continental ribbons (or allochthonous terranes) during orogenesis. Many of these terranes functioned as independent microplates detached from their origin plates during back-arc extension and subsequently drifted and collided with adjacent continents (Nur and Ben-Avraham, 1982). With such interactions, shortening in the collisional belt can be fully compensated by extension in the back-arc region, resulting in accretion of allochthonous terranes without convergence of the far-field plates taking place. If this style of tectonism has occurred, it becomes difficult to use metamorphic ages to infer the behaviour of the large-scale plate interactions. It is therefore concluded that timing of the onset of convergence of Africa with respect to Europe remains poorly constrained.

The motion presented here for the period between chrons M0 and 34 has been derived from linear interpolation (except for a single point at 92 Ma that marks the opening of Labrador Sea; Srivastava and Roest, 1989). However, the validity of this interpolation can be doubted if a pirouette in plate kinematics occurred sometime during the CNS. I check this possibility by plotting the temporal distribution of the Africa-Europe Euler pole (Figure 3.4), which shows a continuous migration before and after the CNS. If an abrupt change from overall strike-slip to convergence took place during the CNS, I would expect to see these poles clustered in two groups. Nonetheless, the Euler pole gradually migrated northward. I can therefore assume, despite the lack of data, that ongoing northward migration of the Euler pole occurred during the CNS. It is therefore argued that such pirouette from overall strike-slip to convergence is unlikely and more smoothly varying trajectories are presented.

3.6.2. Fluctuation of plate motions

An interesting aspect of the motions described in this chapter is the fluctuation in convergence rates through time. This is best emphasised between chrons 31 and 24 (67-55 Ma), in which convergence of both Africa and Iberia virtually stopped for a period of 10-15 m.y. (Figure 3.1b and Figure 3.2). In the latest Cretaceous (70-65 Ma), fragments of Gondwana were accreted against the European margin, resulting in obduction of ophiolites over the Arabian margin (Dercourt et al., 1986) and high-pressure metamorphism in the Sesia-Lanzo zone of the western Alps (Duchêne et al., 1997; Rubatto et al., 1999). It is therefore possible that continental collision in Europe led to a temporary quiescence in plate motions followed by plate reorganisation. Similar behaviour involving a dramatic decrease in plate kinematics has also been reported by Molnar and Tapponnier (1975) for the velocity of the Indian plate following the India-Eurasia collision at ca. 50 Ma.

A second period of relatively slow convergence rates is recognised since the Early Miocene (~20 Ma), although the exact chronology of this velocity decrease is not very well resolved. Since the Late Oligocene (~30 Ma), large areas in the interface between Africa and Europe have been subjected to

an extensional regime associated with the formation of back-arc basins in the Mediterranean region (e.g. Dewey et al., 1973; 1989; Biju-Duval et al., 1977; Durand et al., 1999; Jolivet and Faccenna, 2000; Rosenbaum et al., 2002b). In the western Mediterranean, particularly rapid extension occurred between 21-16 Ma (Speranza et al., 2002). Royden (1993b) has shown that back-arc extension can initiate by rollback of subduction hinges combined with slower convergence rates that do not exceed the rates of subduction rollback. Therefore, slower convergence rates can theoretically promote back-arc extension (Northrup et al., 1995; Jolivet and Faccenna, 2000). Accordingly, it is possible that subduction rollback and back-arc extension in the Mediterranean region commenced as a result of slower convergence between Africa and Europe.

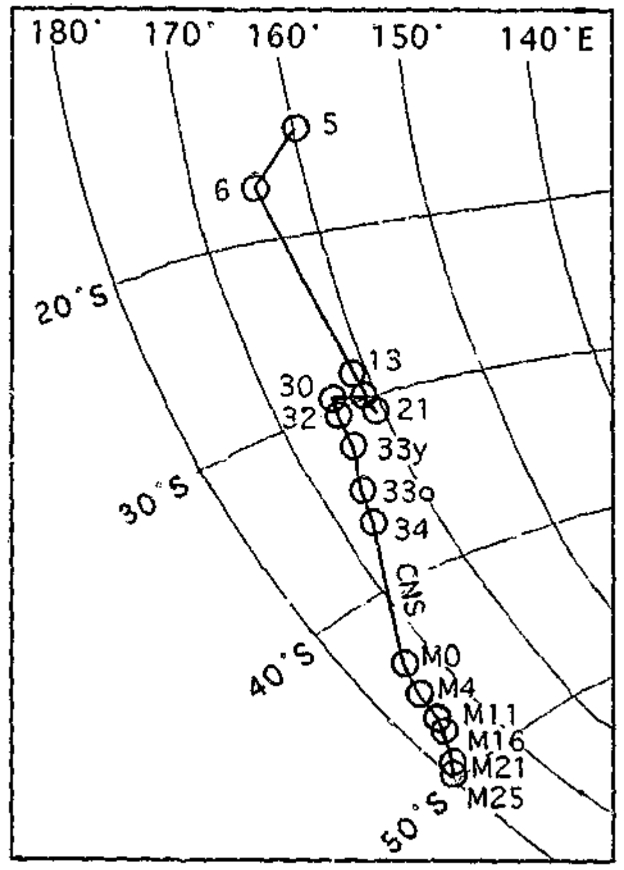


Figure 3.4. Projection of Euler poles for the motion of Africa relative to Europe showing a continuous and gradual migration of the Euler pole through time. CNS, Cretaceous Normal Superchron

3.7. Conclusions

This chapter presented revised kinematic parameters for the motions of Africa and Iberia with respect to Europe in order to provide boundary conditions for Alpine-Mediterranean reconstructions. The following conclusions can be drawn from this study:

1. Convergence of Africa with respect to Europe commenced during the CNS between chrons M0 and 34 (120-83 Ma).
2. Between chrons 31 and 24 (67-55 Ma), Africa and Iberia almost stopped moving relative to Europe, possibly as a consequence of continental collision in the Alps at ca. 65 Ma.
3. The motion of Iberia relative to Europe has been characterised by alternation from left-lateral strike-slip motion, overall convergence and right-lateral strike-slip motion.
4. Since the Early Miocene, a relatively slow convergence has been taking place between Africa and Europe giving rise to wholesale extension in the Mediterranean back-arc basins.

CHAPTER 4

THE MESOZOIC AND CENOZOIC MOTION OF ADRIA: CONSTRAINTS AND LIMITATIONS

Like criminal investigators, historical scientists collect evidence, consider suspects, and follow leads.

Carol E. Cleland (2001)

FOREWORD: CHAPTER 4

This chapter presents kinematic analysis of Adria, which is the main continental mass that separates Africa and Europe in the central Mediterranean. The 'Adriatic problem' – the question whether Adria was an independent microplate or a promontory of Africa - is crucial for reconstructing the tectonic evolution of the Alpine-Mediterranean region. It is therefore revisited here by means of systematic spatio-temporal analysis. I thank Gordon Lister for discussions on this topic, and Cecile Duboz for her role in the development of the PlatyPlus software package, which facilitated the kinematic analysis. The chapter also benefited from discussions with Giovanni Muttoni, Fabio Speranza and Antonio Schettino, and from critical reviews by Mike Raetz, Wouter Schellart and Peter Betts. A slightly modified version of this chapter has been submitted to *Geodinamica Acta* and is currently in review*.

* Rosenbaum, G., Lister, G.S. and Duboz, C., The Mesozoic and Cenozoic motion of Adria (central Mediterranean): constraints and limitations. *Geodinamica Acta*, in review.

4. THE MESOZOIC AND CENOZOIC MOTION OF ADRIA: CONSTRAINTS AND LIMITATIONS

Abstract

This chapter presents kinematic analysis on the motion of Adria, which is the continental mass that bridges Africa and Europe in the central Mediterranean. Palaeomagnetic data show a general coherence between the motion of Adria and Africa since the Late Paleozoic. This mutual motion, for the period between 120 Ma and the present, is verified by comparing inferred palaeolatitudes from relatively stable parts of Adria (Southern Alps, Hyblean Plateau, Apulia, Gargano and Istria), and latitudinal changes that are predicated by the motion of Africa with respect to hotspots. Additional constraints on the motion of Adria are provided from the Late Paleozoic-Early Mesozoic passive margin of Adria in the Ionian Sea. The seismic structure of the floor of the Ionian Sea is similar to the structure of the oceanic crust in marginal back-arc basins, suggesting that it formed as a small ocean basin. Furthermore, the Ionian lithosphere in the Calabrian arc has been subjected to rapid rollback - a process that commonly occurs only when the subducting slab consists of oceanic lithosphere. This ocean marks the Permian-Triassic to Jurassic plate boundary between Adria and Africa, suggesting a small amount of independent motion between Adria and Africa at that time, followed by a coherent motion of Adria and Africa since the Jurassic.

4.1. Introduction

Adria is the name given to a region of continental crust bridging the continental masses of Africa and Europe in the central Mediterranean (Suess, 1883; Channell et al., 1979; Bosellini, 2002) (Figure 4.1). The internal part of Adria, in the area of the present-day Adriatic Sea, consists of a rigid and relatively undeformed continental crust (e.g. Anderson and Jackson, 1987a). Outcrops of the relatively stable and undeformed regions of Adria are exposed in parts of the Southern Alps, Apulia, Gargano, Istria Peninsula and the Hyblean Plateau (Figure 4.1). In contrast, the external parts of Adria underwent intense deformation during Alpine orogeny resulting in the development of three orogenic belts along the Adriatic margins: the Apennines, the Alps and the Dinarides-Albanides (Figure 4.1). In this tectonic configuration, constraints on the motion of Adria are crucial for any attempt to reconstruct the kinematic evolution of the Alpine-Mediterranean region.

During the last few decades the question as to whether Adria was a promontory of Africa, or whether it moved as an independent microplate, have been addressed by numerous authors (Channell et al., 1979; Lowrie, 1986; Anderson, 1987; Platt et al., 1989a; Van der Voo, 1993; Channell, 1996; Stampfli and Mosar, 1999; Mele, 2001; Muttoni et al., 2001; Wortmann et al., 2001; Bosellini, 2002). Palaeomagnetic results were mostly interpreted as indicating a coherent motion between Adria and Africa during the Late Paleozoic and the Mesozoic times (e.g. Channell, 1996; Muttoni et al., 2001). However, there are still considerable uncertainties concerning the motion of Adria, particularly since the Cretaceous, due to the imprecise resolution of palaeomagnetic data (Channell, 1996). Such uncertainties have resulted in contradicting kinematic reconstructions of the Alpine-Mediterranean region, in which Adria has been considered either as an African promontory (e.g. Wortmann et al., 2001; Rosenbaum et al., 2002a), or as an independent microplate (e.g. Biju-Duval et al., 1977; Dercourt et al., 1986; 2000, Stampfli et al., 1998).

The purpose of this chapter is to provide constraints on the motion of Adria based on geological, geophysical and palaeomagnetic data. This is done by (1) considering the significance and limitations of palaeomagnetic data; (2) comparing kinematic interpretations derived from palaeomagnetic data with motions calculated from analysis of hotspot tracks; and (3) constraining the tectonic evolution of the Ionian Sea and its Adriatic passive margin. The analysis supports earlier suggestions that the motion of Adria was roughly similar to the motion of the African plate.

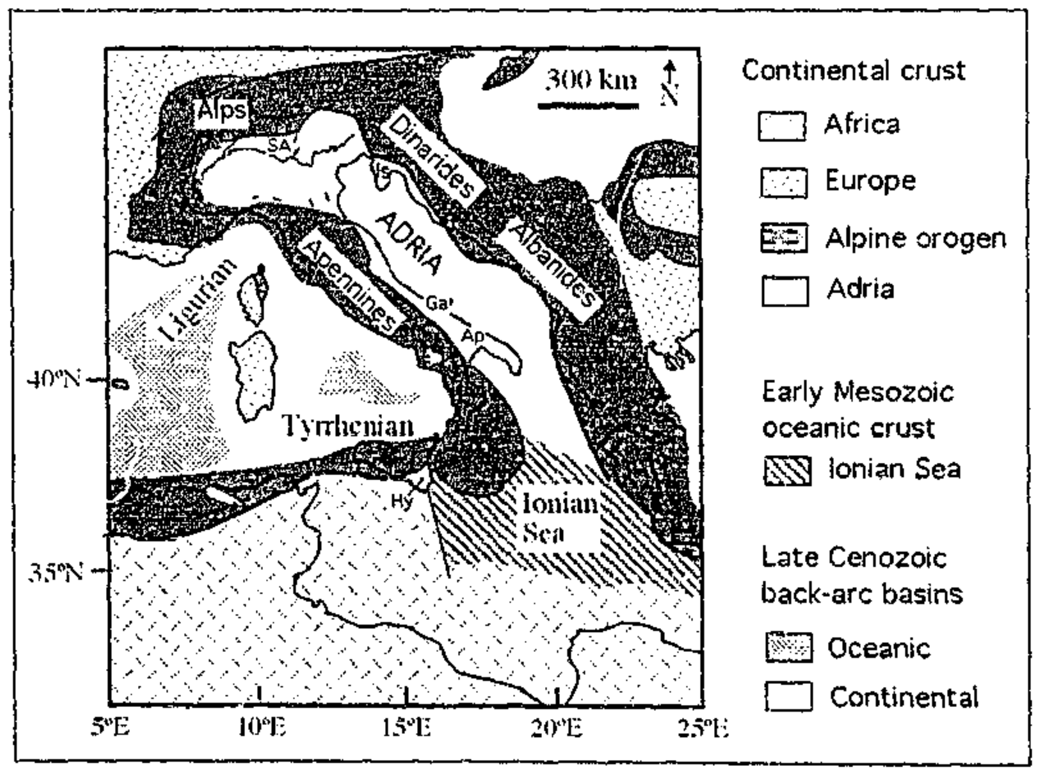


Figure 4.1. Tectonic map of Adria and its margins. Ap, Apulia, Ga, Gargano; Hy, Hyblean Plateau; Is, Istria; SA, Southern Alps.

4.2. Palaeomagnetism of Adria

The lack of robust kinematic criteria (e.g. magnetic isochrons) on the Adriatic margins leaves palaeomagnetism as the major source of information used to provide kinematic constraints on the motion of Adria (e.g. Channell et al., 1979; Channell, 1996; Muttoni et al., 2001). This section discusses relative rotations around vertical axes, the Adriatic Apparent Polar Wander Path (APWP) and palaeolatitudes of Adria as inferred from palaeomagnetic inclinations.

4.2.1. Block rotations

Rotations around vertical axes, as inferred from palaeomagnetic results, are presented in Figure 4.2. These rotations indicate possible angular motions of stable parts of Adria with respect to Africa (Figure 4.2a), as well as the sense of block rotations in the deformed margins of Adria (Figure 4.2b).

In earlier studies (e.g. Lowrie and Alvarez, 1974; 1975; VandenBerg, 1983), it was suggested that the whole Italian peninsula underwent an approximately 30-45° of counterclockwise rotation since the Cretaceous. These studies, however, were based on data derived from deformed nappe structures within the Apennines which were themselves subjected to block rotations during deformation. In subsequent reappraisals of palaeomagnetic data (e.g. Lowrie, 1986; Channell et al., 1992a) it has been agreed that the relatively undeformed parts of Adria have not been significantly rotated with respect to Africa (Figure 4.2). No rotations, or just small amounts of counterclockwise rotations, have been interpreted from data from the Southern Alps and the Hyblean Plateau (Schult, 1973; Barberi et al., 1974; Channell and Tarling, 1975; Gregor et al., 1975). However, results from Istria, Apulia and Gargano indicate either counterclockwise rotations (VandenBerg, 1983; Marton and Nardi, 1994), no rotations (Channell, 1977; Speranza and Kissel, 1993), or clockwise rotations (Tozzi et al., 1988). These contradicting results may be associated with local deformation that nevertheless affected these relatively undeformed parts of Adria (see, for example, Bertotti et al., 1999).

In summary, most palaeomagnetic data suggest little or no rotations of Adria with respect to Africa during the Mesozoic and the Tertiary. In contrast, a pattern of block rotations is recognised in the deformed margins of Adria (Figure 4.2b). These are predominantly clockwise rotations in the Dinarides-Albanides (e.g. Speranza et al., 1995), counterclockwise rotations in the Apennines (e.g. Scheepers et al., 1993; Scheepers and Langereis, 1994; Muttoni et al., 2000), and clockwise rotations in Calabria and Sicily (e.g. Speranza et al., 1999; 2000). These rotational patterns are associated with local deformation in the fold-and-thrust belts, and their sense of motion may represent the direction in which allochthonous terranes were emplaced onto the orogenic edifice at times of subduction rollback and formation of back-arc basins (Lonergan and White, 1997; Rosenbaum et al., 2002b).

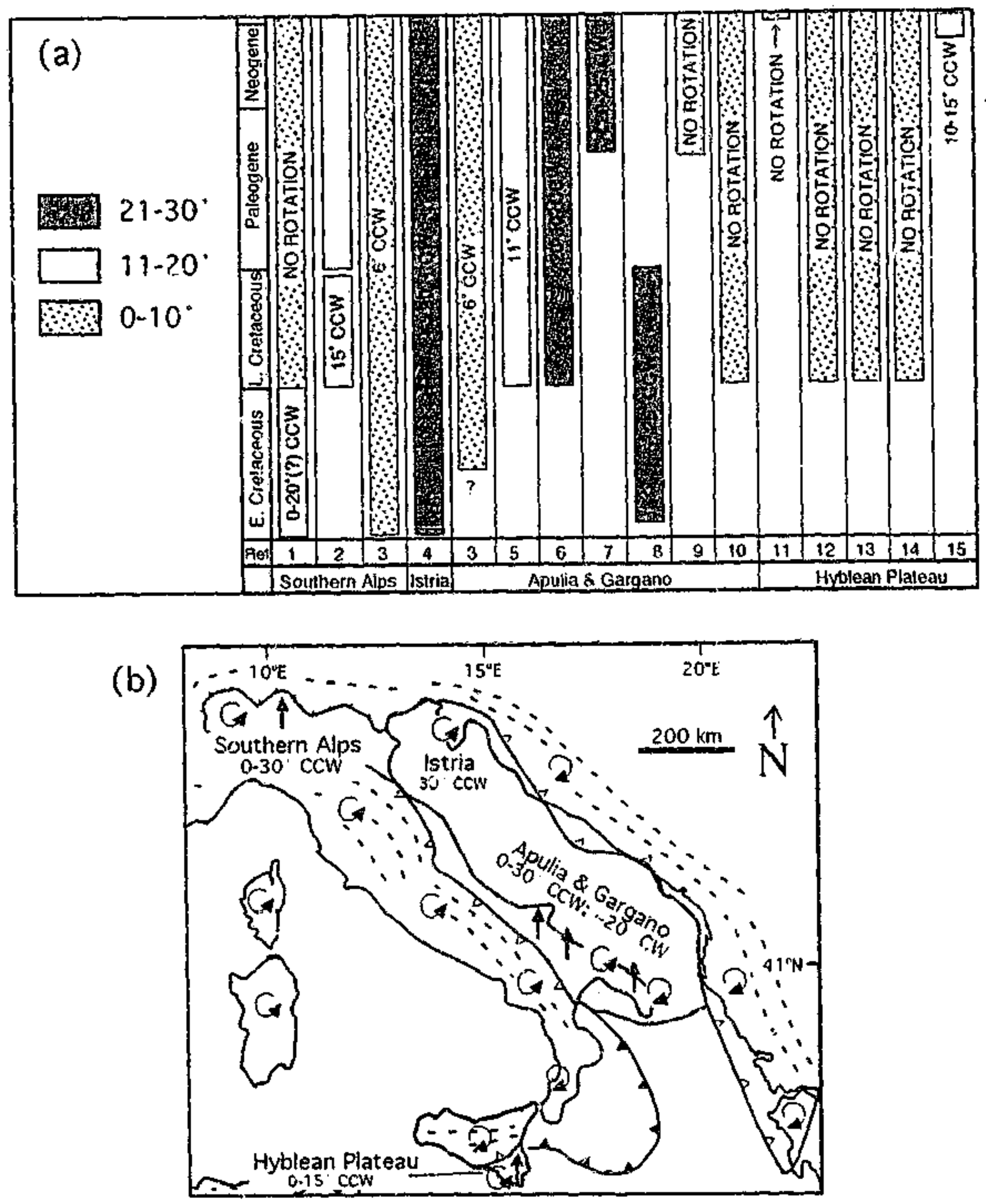


Figure 4.2. (a) Rotations of Adria with respect to Africa as inferred from palaeomagnetic studies. References are (1) Channell et al. (1992a); (2) VandenBerg & Wonders (1976); (3) Channell and Tarling (1975); (4) Marton and Velojvic (1983); (5) Marton and D'Andrea (1992); (6) Marton and Nardi (1994); (7) Tozzi et al. (1988); (8) VandenBerg (1983); (9) Speranza and Kissel (1993); (10) Channell (1977); (11) Scheepers (1992); (12) Gregor et al. (1975); (13) Barberi et al. (1974); (14) Schult (1973); (15) Besse et al. (1984). (b) Map showing the sense of Cretaceous to Present block rotations in Adria and its margins.

4.2.2. Apparent Polar Wander Path (APWP)

The Adriatic APWP is made up of sequential positions of palaeomagnetic poles from the relatively stable parts of Adria (Figure 4.3). The palaeomagnetic poles from Adria are relatively scattered,

resulting in a motion path which is subjected to large errors. Nevertheless, many of these poles still overlap with the 95% confidence circles of the African APWP constructed by Besse and Courtillot (1991) (Figure 4.3). This suggests that the motion path of Adria was not significantly different from the motion path of Africa since the Jurassic. Similarly, in Early Mesozoic data (Muttoni et al., 2001), a close coincidence of palaeomagnetic poles from Adria and Africa is also recognised, suggesting that Adria was an African promontory attached to the African plate at least since the Permian (Channell, 1996; Muttoni et al., 2001).

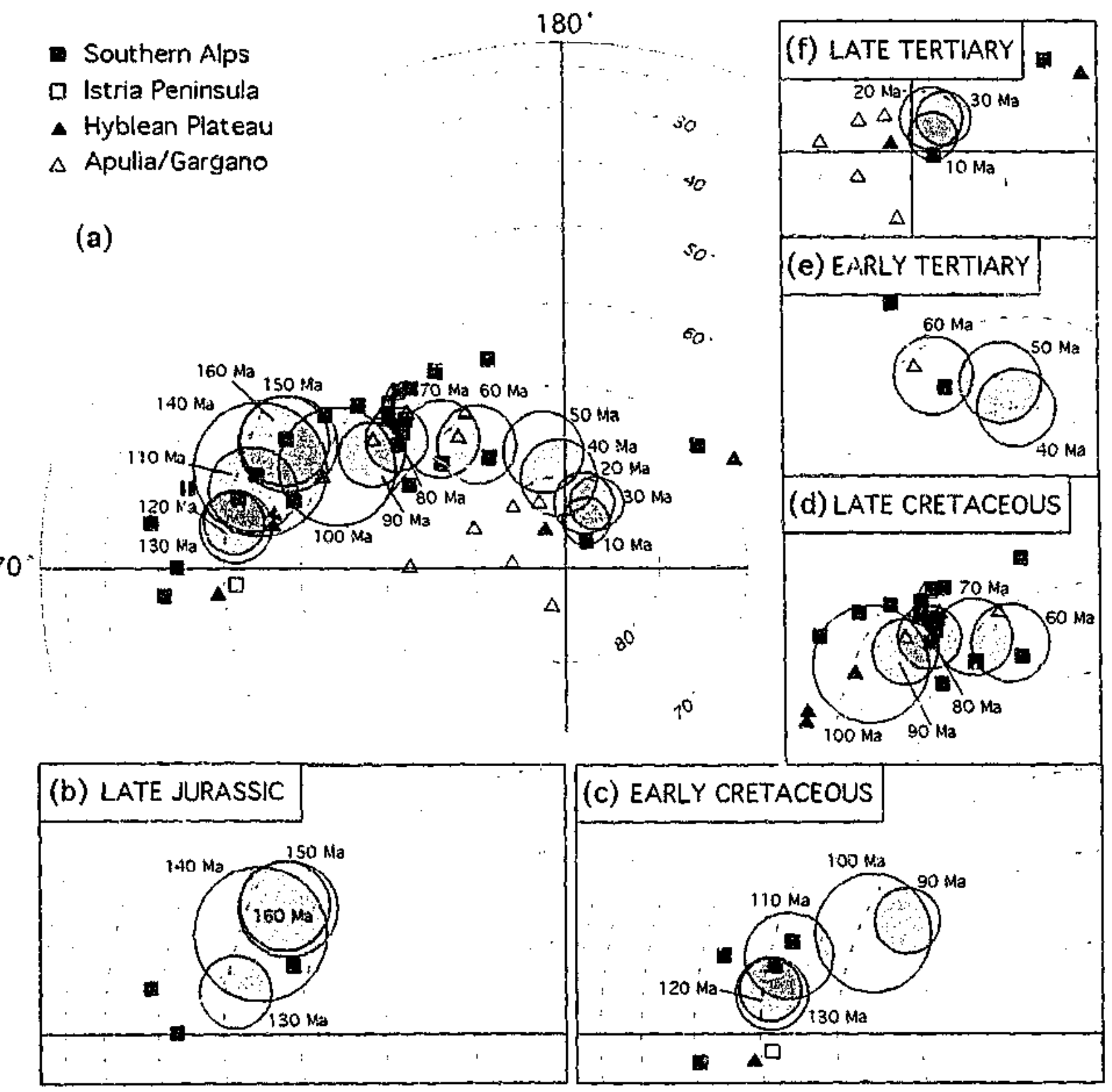


Figure 4.3. (a) Palaeomagnetic poles from Adria (see Table 4.1) and the African APWP (shaded areas) plotted with 95% confidence circles (after Besse and Courtillot, 1991). Clusters of Adriatic poles of different ages along with the relevant circles of the African APWP are shown in insets: (b) Late Jurassic; (c) Early Cretaceous; (d) Late Cretaceous; (e) Early Tertiary; (f) Late Tertiary. 95% confidence circles of Adriatic poles have not been plotted in order to preserve readability of diagrams.

Table 4.1. Palaeomagnetic results from relatively stable parts of Adria

Location	Age	lat /long	B	N	Dec	Inc	KD	α_{95}	Pole	dp	dm	λ_n	λ_n^+	λ_n^-	$\Delta\lambda_n$	Reference
Southern Alps																
Pelagic limestones	97-105	46.0/12.0	8	48	325.8	38.2	0.0	5.4	52.9/252.3	3.7	6.3	21.5	25.5	17.9	24.5	VandenBerg & Wonders (1980)
Pelagic limestones	93-97	46.0/12.0	8	44	336.4	36.2	0.0	4.6	57.6/236.7	3.1	5.3	20.1	23.3	17.1	25.9	VandenBerg & Wonders (1980)
Pelagic limestones	88-90	46.0/12.0	8	50	340.4	37.7	0.0	9.5	60.4/231.4	6.6	11.2	21.1	28.4	15.0	24.9	VandenBerg & Wonders (1980)
Pelagic limestones	88-90	46.0/12.0	11	66	343.8	39.5	0.0	5.5	62.9/226.6	3.9	6.5	22.4	26.6	18.6	23.6	VandenBerg & Wonders (1980)
Pelagic limestones	87-89	46.0/12.0	8	50	345.1	42.9	0.0	7.4	65.7/226.7	5.6	9.1	24.9	31.1	19.6	21.1	VandenBerg & Wonders (1980)
Pelagic limestones	86-88	46.0/12.0	3	16	343.4	45.7	0.0	8.4	67.0/232.6	6.8	10.7	27.1	34.6	20.9	18.9	VandenBerg & Wonders (1980)
Pelagic limestones	83-87	46.0/12.0	3	17	346.7	49.4	0.0	8.3	71.1/230.0	7.3	11.0	30.3	38.[illegible]	23.6	15.7	VandenBerg & Wonders (1980)
Pelagic limestones	79-83	46.0/12.0	2	11	347.6	50.0	0.0	0.0	72.0/228.7	0.0	0.0	30.8	30.8	30.8	15.2	VandenBerg & Wonders (1980)
Vicentinian Alps limestines	78-90	45.8/11.9	6	30	344.2	44.1	139.1	5.7	66.4/229.6	4.5	7.1	25.9	30.6	21.6	19.9	VandenBerg & Wonders (1976)
Vicentinian Alps limestines	86-90	45.8/11.9	2	10	345.5	38.2	0.0	3.3	62.7/222.5	2.3	3.9	21.5	23.9	19.2	24.3	VandenBerg & Wonders (1976)
Vicentinian Alps limestines	85-87	45.8/11.9	2	10	344.2	43.3	0.0	5.3	65.7/228.7	4.1	6.6	25.2	29.6	21.3	20.6	VandenBerg & Wonders (1976)
Vicentinian Alps limestines	78-85	45.8/11.9	2	10	342.7	50.9	0.0	5.3	70.4/240.9	4.8	7.1	31.6	36.8	27.0	14.2	VandenBerg & Wonders (1976)
Vicentinian Alps limestines	152-155	45.8/11.9	2	10	328.5	45.3	0.0	3.0	58.3/255.3	2.4	3.8	26.8	29.3	24.5	19.0	VandenBerg & Wonders (1976)
Dolomites	60-89	45.5/11.5	12	97	343.0	41.4	33.8	7.5	64.1/229.2	5.9	9.2	23.8	29.8	18.6	21.7	Channell & Tarling (1975)
Dolomites	90-97	45.5/11.5	3	18	330.5	36.5	69.5	14.9	54.8/244.6	10.6	17.3	20.3	32.1	11.2	25.2	Channell & Tarling (1975)
Scaglia Rossa limestones	65-89	46.0/12.0	22	496	346.6	38.5	68.7	3.8	63.4/220.2	2.7	4.5	21.7	24.5	19.1	24.3	Channell et al. (1992a)
Upper Maiolica limestones	124-135	46.0/12.0	10	815	322.2	39.5	87.3	5.2	51.5/257.0	3.7	6.2	22.4	26.3	18.8	23.6	Channell et al. (1992a)
Ammonotico Rosso l'stones	146-155	46.0/12.0	11	462	309.7	30.6	89.1	4.9	39.1/263.3	3.0	5.4	16.5	19.6	13.5	29.5	Channell et al. (1992a)
Basalts Priabona, NW Italy	38-50	45.0/11.0		29	353.0	50.0	0.0	0.0	75.0/214.0	0.0	0.0	30.8	30.8	30.8	14.2	De Boer (1965)
Biochemic sediments, Schio	65-146	45.0/11.0		4	356.0	37.0	0.0	0.0	65.0/200.0	0.0	0.0	20.6	20.6	20.6	24.4	De Boer (1965)
Marosnca basalts, NW Italy	25-33	45.0/11.0		16	184.0	-61.0	0.0	0.0	86.0/145.0	0.0	0.0	42.1	42.1	42.1	2.9	De Boer (1965)
Volcanics (Euganei, Lessini)	35-56	45.5/11.7	26		169.9	-36.7	11.8	8.0	63.7/213.1	5.5	9.3	20.4	26.3	15.3	25.1	Soffel (1978)
Volcanics (Euganei, Lessini)	23-35	45.5/11.7	17		200.2	-52.5	28.3	6.4	70.2/133.8	6.1	8.8	33.1	39.7	27.5	12.4	Soffel (1978)
Ammonitico Rosso L'stone	154-161	45.8/11.0	3	163	309.8	40.8	226.6	8.2	43.9/269.4	6.0	9.9	23.3	29.9	17.7	22.5	Channell et al. (1990a)
Maiolica Limestones	118-140	45.0/11.0	5	899	306.2	41.2	48.9	11.0	41.8/273.4	8.2	13.4	23.6	32.8	16.2	21.4	Channell et al. (1995b)
Cimson limestones & shales	120-129	46.0/11.8		343	317.0	31.6	11.8	2.3	44.2/257.3	1.6	2.9	17.1	18.6	15.7	28.9	Channell et al. (2000)

Table 4.1. Continued

Location	Age	lat /long	B	N	Dec	Inc	KD	α_{95}	Pole	dp	dm	λ_n	λ_n^+	λ_n^-	$\Delta\lambda_n$	Reference
Istria peninsula																
Istria limestones	65-141	45.0/14.0	19	119	316.5	46.5	19.5	7.8	51.6/272.5	6.4	10.0	27.8	34.8	21.8	17.2	Marton & Veljovic (1983)
Apulia and Gargano																
Calcari di Castro limestones	23-56	40.0/18.5	13	87	25.7	50.6	41.0	6.0	67.5/123.1	5.4	8.1	31.3	37.2	26.2	8.7	Tozzi et al. (1988)
Plio-Pleistocene Clays	1-4	41.0/16.5	11	180	179.0	-55.7	240.0	3.0	85.2/206.1	3.1	4.3	36.2	39.4	33.3	4.8	Scheepers (1992)
Apulian limestones	88-97	41.0/16.5	12	56	327.0	40.0	42.0	6.7	56.0/261.0	4.7	7.9	22.8	27.9	18.2	18.2	Marton & Nardi (1994)
Scaglia limestone, Gargano	65-90	41.8/16.0	21	123	327.7	38.2	0.0	4.3	56.1/259.2	3.0	5.1	21.5	24.6	18.6	20.3	Vandenberg (1983)
Miolica limestone, Gargano	97-146	41.8/16.6	8	47	312.0	41.0	0.0	8.7	49.0/274.0	5.8	10.0	23.5	30.5	17.5	18.3	Vandenberg (1983)
Limestones, Gargano	86-90	41.9/16.0	17	114	335.0	37.8	31.3	6.5	60.5/249.1	4.5	7.7	21.2	26.0	16.9	20.7	Channell (1977)
Hyblean plateau																
Hyblean plateau volcanics	71-84	36.8/14.7	12	74	173.6	-31.2	36.0	7.4	69.3/212.3	4.6	8.3	16.8	21.8	12.4	20.0	Grasso et al. (1983)
Hyblean plateau volcanics	71-84	36.8/14.7	52	203	167.3	-27.6	22.0	4.3	65.2/225.1	2.6	4.7	14.6	17.3	12.2	22.2	Grasso et al. (1983)
Hyblean plateau volcanics	5-10	37.2/14.9	18	112	359.0	48.3	15.0	9.4	82.1/200.9	8.1	12.3	29.3	38.3	22.0	7.9	Grasso et al. (1983)
Hyblean plateau volcanics	2-4	37.2/14.9	33	249	356.9	59.6	26.0	5.0	85.9/339.3	5.7	7.5	40.4	46.5	35.1	-3.2	Grasso et al. (1983)
Cape Passero volcanics	70-90	36.7/15.1	19	107	167.0	-22.0	22.8	7.1	62.1/223.3	4.0	7.6	11.4	15.6	7.6	25.3	Schult (1973)
Trubi Formation (sediments)	3-5	37.1/14.5	5	28	353.5	54.0	167.0	4.5	84.1/260.5	4.4	6.3	34.5	39.2	30.3	2.6	Besse et al. (1984)
Tellaro Formation (sed.)	8-15	37.1/14.5	3	17	171.0	-49.0	137.0	10.0	79/245	8.7	13.2	29.9	39.8	22.0	7.2	Besse et al. (1984)
Vizzini sxediments	42-50	37.1/14.5	2	33	352.0	35.5	140.0	2.3	71.2/218.5	1.5	2.7	19.6	21.2	18.1	17.5	Besse et al. (1984)
Basalts	1-5	37.2/14.8	10	84	356.0	48.0	24.0	10.0	81/219	9.1	13.9	29.0	38.7	21.3	8.2	Barberi et al. (1974)
Cape Passero basaitic Dykes	72-86	36.7/15.1	8	34	165.0	-26.0	62.0	7.0	63/229	4.1	7.6	13.7	18.0	9.8	23.0	Barberi et al. (1974)
Cape Passero igneous rocks	72-86	36.7/15.1	27	141	167.0	-23.0	28.0	5.3	62.5/224	3.0	5.6	12.0	15.1	9.1	24.7	Barberi et al. (1974)
Mt Iblei Volcanics	0-5	37.2/14.8	19	93	161.0	-50.0	12.8	8.3	73/269	8.8	13.1	30.8	39.0	24.0	6.4	Gregor et al. (1975)
Cape Passero Volcanics	72-86	36.7/15.1	27	141	167.0	-23.0	28.0	5.3	62.5/224	3.0	5.6	12.0	15.1	9.1	24.7	Gregor et al. (1975)
Cape Passero Volcanics	88-90	36.7/15.1	15	88	162.9	-30.1	22.5	9.8	64/236	5.0	9.2	16.2	22.7	10.5	20.5	Gregor et al. (1975)

Comments: Age in Ma; lat = latitude; long = longitude; B = number of sites; N = number of samples; Dec = declination; Inc = Inclination; KD = precis[illegible] parameter; α_{95} = radius of the cone of 95% confidence; Pole = virtual geomagnetic pole; dp,dm = semiaxes of the 95% level of confidence ellipse (in degrees) for the vi[illegible] geomagnetic pole; λ_n = mean palaeolatitude; λ_n^+ = maximum palaeolatitude; λ_n^- = minimum palaeolatitude; $\Delta\lambda_n$ = difference between present-day latitude and palaeolatitude.

4.2.3. Palaeolatitudes

Table 4.1 summarises palaeomagnetic data and calculated palaeolatitudes for rocks from the relatively stable parts of Adria (Southern Alps, Istria Peninsula, Apulia, Gargano and the Hyblean Plateau). The changes between calculated palaeolatitudes and the present-day latitudes of the sampling sites ($\Delta\lambda_0$) are plotted against the age of the rocks (Figure 4.4). This diagram provides a quantified estimate for the latitudinal change of Adria since the Middle Mesozoic. It shows a general northward migration of Adria through time that agrees with the overall northward motion of Africa with respect to Europe. However, there are some anomalously large latitudinal variations (in excess of 20-30°) that correspond to 2000-3000 km of a northward motion. This apparent contradiction may be assessed independently by analysing the hotspot frame of reference for Africa.

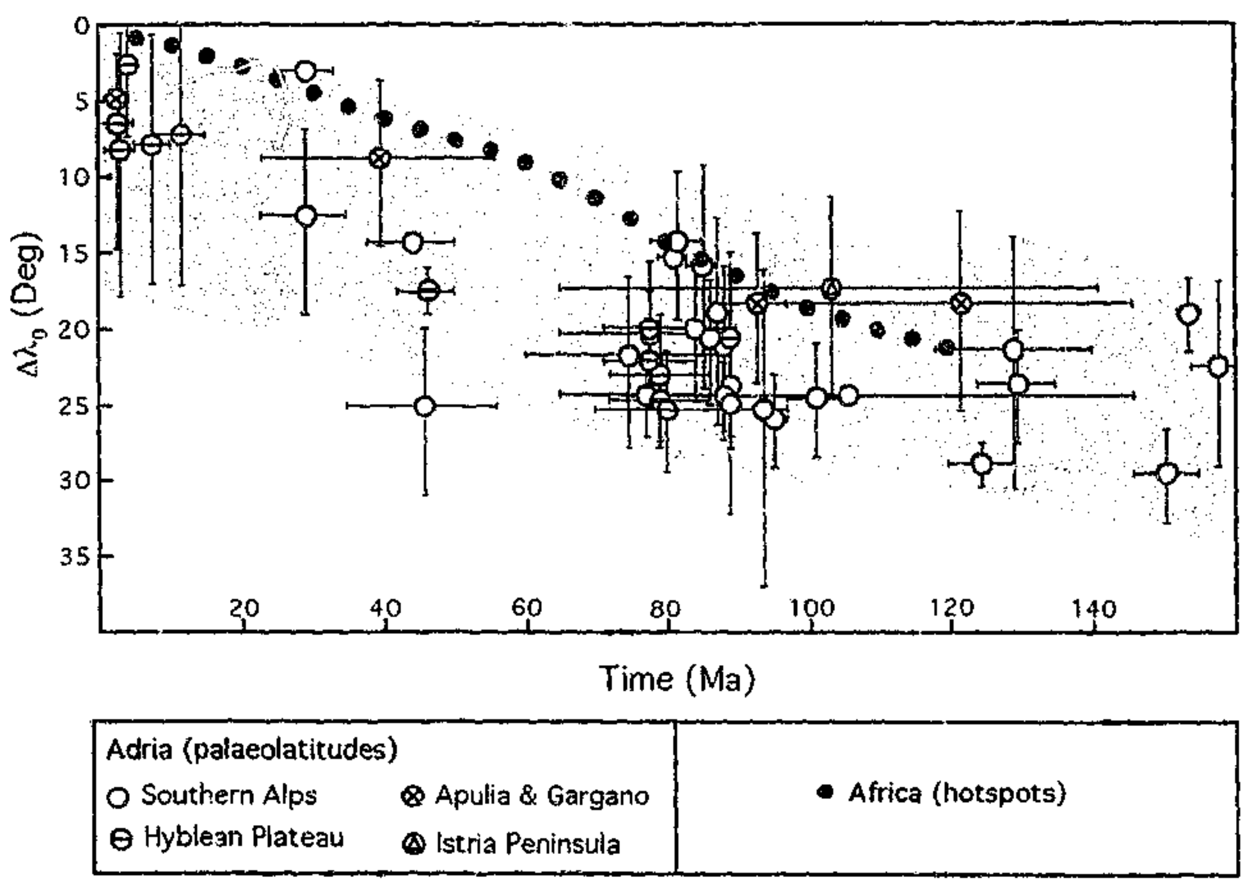


Figure 4.4. Diagram of latitudinal change ($\Delta\lambda_0$) versus time based on palaeolatitudes from the relatively stable parts of Adria. $\Delta\lambda_0$ is the difference between the present-day latitude of a particular site and its palaeolatitude (see Table 4.1). Also shown are latitudinal changes of a point at 37°N/11°E that moves together with Africa in a hotpots frame of reference (see text for discussion).

4.3. Palaeomagnetism and the motion of Africa relative to hotspots

The motion of Africa with respect to hotspots can provide independent constraints on plate motions, which may be compared with data derived from palaeomagnetic studies (i.e. palaeolatitudes). The existence of linear chains of progressively younger volcanics is considered to result from the motion

of lithospheric plates relative to an underlying mantle hotspot. If these hotspots are assumed to be stationary with respect to the mantle, the motion of a plate relative to hotspots may be considered as an 'absolute' motion. Similar assumptions have been used in several plate kinematic models, which are generally consistent with the kinematics known from the reconstruction of magnetic isochrons (Duncan and Richards, 1991; Müller et al., 1993; Wang and Wang, 2001).

4.3.1. The motion of Africa relative to hotspots

Several studies of hotspot tracks on the African plate have been used to determine the motion of Africa in a hotspot frame of reference (Morgan, 1983; O'Connor and Duncan, 1990; Garfunkel, 1992; O'Connor and Le Roex, 1992) (Figure 4.5a). The most prominent linear recognised track is the Tristan da Cunha – Walvis Ridge, which shows ongoing volcanism since 130 Ma (Morgan, 1983; Müller et al., 1993). Hotspot tracks that are related to the motion of Africa are also the St Helena ridge (O'Connor and Le Roex, 1992), the Réunion ridge (Müller et al., 1993), and the continental volcanism of Darfur-Levant in northeastern Africa (Garfunkel, 1992). Other Late Cenozoic volcanic centres throughout Africa (e.g., Cape Verde, Afar) are not characterised by linear tracks, possibly, because of a slower motion of the African plate since the Oligocene (Burke, 1996; O'Connor et al., 1999).

The movement of Africa relative to a hotspot frame of reference (Figure 4.5b) is based on kinematic models of Fleitout et al. (1989), Duncan and Richards (1991), O'Connor and Le Roex (1992), Garfunkel (1992) and Müller et al. (1993). Rotational parameters that represent an average interpolation of the published data are summarised in Table 4.2 (see also movie 'Af_Eu_HS.mov' in Appendix 2). Results show a generally northeastward motion of Africa, which became increasingly faster from 105 Ma to 80 Ma. A marked velocity decrease in the latest Cretaceous and the Paleocene (65-50 Ma) coincides with a particularly slow relative motion of Africa with respect to Europe at that time (Rosenbaum et al., 2002a). The data show that a further velocity decrease occurred since 25-20 Ma (Figure 4.5b).

4.3.2. Palaeolatitudes vs. the motion of Africa with respect to hotspots

Latitudinal changes versus time, as predicted from the motion of Africa with respect to hotspots, are plotted in Figure 4.4. These latitudinal changes were calculated by applying the rotational parameters of Africa with respect to hotspots (Table 4.2) on a point located at 37°N/11°E. The results show that predicted changes in the latitudes of Africa are within the error limits of the Adriatic palaeolatitudes (shaded area in Figure 4.4). Therefore, there is no evidence for an independent motion of Adria with respect to Africa. There are few anomalously low palaeolatitudes derived from the palaeomagnetic data that may be associated with inclination errors or with an original (synsedimentary) shallowing of remnant inclinations.

A similar phenomenon of unusually low palaeolatitudes has been reported in Cenozoic sediments from Central Asia, in which measurements of remnant inclinations have shown values 20-30° shallower than predicted (Chauvin et al., 1996; Gilder et al., 2001; Bazhenov and Mikolaichuk, 2002). These anomalous inclinations may not necessarily reflect true low palaeolatitudes, but were probably acquired during sedimentation (Gilder et al., 2001). The effect of inclination shallowing is considered to be smaller or even negligible in Mesozoic rocks (e.g. Bazhenov and Mikolaichuk, 2002). Indeed, the latitudinal discrepancy, as shown in Figure 4.4, is diminished for ages older than ca. 80 Ma, where predicted latitudinal changes from the hotspot frame of reference are consistent with inferred palaeolatitudes.

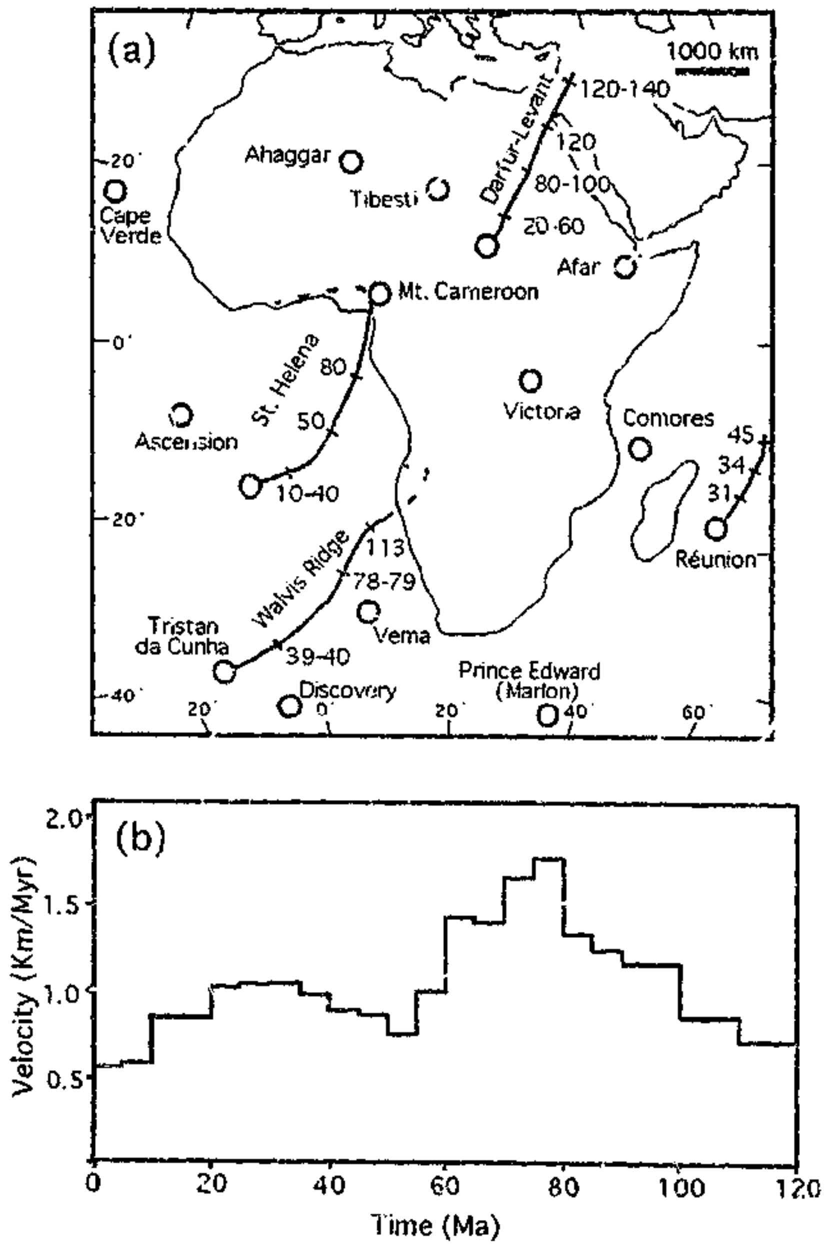


Figure 4.5. (a) Distribution of hotspot-related volcanism (open circles) and hotspot tracks (thick lines) on the African plate (after Burke, 1996). Uncertain tracks are indicated by dashed lines. Numbers indicate the age of volcanism in Ma (after Garfunkel, 1992; O'Connor and Le Roex, 1992; Müller et al., 1993); (b) Motion of Africa in a hotspot frame of reference calculated for a point (at 20°N/10°E) that moves together with Africa. Velocities are calculated relative to a fixed point at 40°N/10°E.

Table 4.2. Poles of rotation for the motion of Africa relative to hotspots.

Age (Ma)	Latitude	Longitude	Rotation	Age (Ma)	Latitude	Longitude	Rotation
5	49.28	-40.32	-0.97	65	34.48	-48.92	-14.23
10	49.20	-40.49	-1.95	70	3[illegible].70	-51.00	-15.42
15	46.87	-44.98	-3.02	75	31.65	-51.55	-16.95
20	45.52	-46.38	-4.11	80	30.60	-52.17	-18.60
25	41.94	-46.61	-5.26	85	29.38	-51.96	-20.04
30	39.57	-46.75	-6.43	90	28.74	-51.60	-21.41
35	38.04	-46.80	-7.65	95	27.55	-50.78	-22.72
40	37.07	-46.45	-8.83	100	26.48	-50.07	-24.04
45	36.46	-46.58	-9.87	105	26.42	-49.50	-25.16
50	35.97	-46.96	-10.83	110	26.37	-48.97	-26.28
55	35.87	-47.14	-11.73	115	26.59	-48.41	-27.33
60	35.31	-47.56	-12.89	120	26.79	-47.89	-28.39

4.4. The Ionian Sea

The southern margin of Adria in the Ionian Sea is the only margin of Adria that was not affected by orogenic processes. Therefore, the geological evolution of this basin can potentially provide pristine evidence on the geodynamic relationships of Adria and Africa. However, due to the depth of the water (>3 km) and the thickness of the sedimentary cover (6-8 km; De Voogd et al., 1992; Catalano et al., 2001; Figure 4.6b), the basement of the Ionian Sea has never been sampled. Therefore, it is still a matter of debate whether the floor of the Ionian Sea consists of an oceanic crust (e.g. Dewey et al., 1973; Biju-Duval et al., 1977; Finetti, 1985; De Voogd et al., 1992; Stampfli and Mosar, 1999; Dercourt et al., 2000; Catalano et al., 2001) or a thinned continental lithosphere (e.g. Morelli, 1978; Farrugia and Panza, 1981; Calcagnile et al., 1982; Boccaletti et al., 1984; Bosellini, 2002).

In the following section, I show that geophysical evidence, as well as the geodynamic context of the Ionian Sea in the central Mediterranean, suggest the existence of oceanic lithosphere in this region. If so, a divergence plate boundary between Adria and Africa existed as long as sea-floor spreading took place. Accordingly, the possible age of this ocean is discussed.

4.4.1. Nature of the Ionian crust

The Ionian crust, with a mean crustal thickness of 17-22 km (Papazachos and Comninakis, 1978; Nicolich et al., 2000), is relatively thick in comparison with typical oceanic crust. However, seismic velocities indicate that approximately 30% of this crustal thickness is contributed by the sedimentary cover (Finetti, 1985; De Voogd et al., 1992; Catalano et al., 2001) (Figure 4.6b). The seismic layers below the sedimentary cover show characteristic features of oceanic crust with seismic velocities of 4.7-7.45 km/s (De Voogd et al., 1992) (Figure 4.6b). The thickness of these layers is approximately 8 km, which is a slightly thicker value relative to comparable seismic layers in a typical oceanic crust (5-7 km; Keary and Vine, 1990). However, the crustal thickness in the Ionian Sea resembles a typical

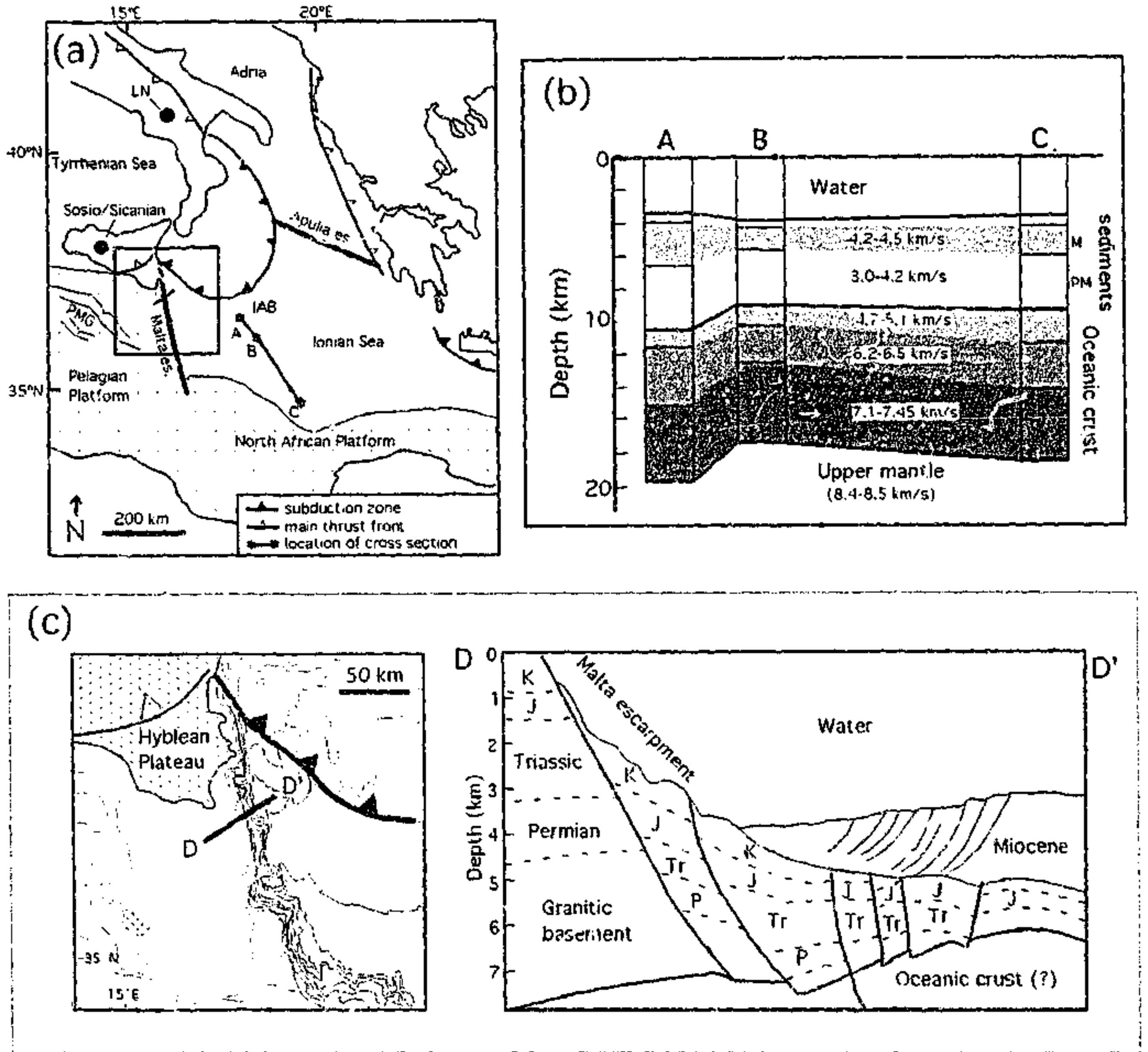


Figure 4.6. Map showing major tectonic features in the Ionian Sea. Shaded areas indicate young (<5 Ma) oceanic crust in the Tyrrhenian Sea and old (>160 Ma) oceanic crust in the Ionian Sea. Apulia es., Apulia escarpment; IAB, Ionian Abyssal Basin; LN, Lagonegro; Malta es., Malta escarpment; PMG, Pantelleria-Malta graben. The map is modified after Boccaletti et al. (1984) and Finetti (1985). (b) cross section across the Ionian Abyssal Basin based on a seismic velocity model. The cross section shows characteristic velocities of oceanic crust below the sedimentary cover (after De Voogd et al., 1992). M, Messinian evaporates; PM, Pre-Messinian sediments. (c) Enlargement of the bathymetry (contoured every 400 m) near southeast Sicily (after De Voogd et al., 1992) and an interpreted seismic reflection line across the Malta escarpment (Line MS 26 in Finetti (1985)). P, Permian; Tr, Triassic; J, Jurassic; K, Cretaceous.

thickness of oceanic crust in marginal back-arc basins (e.g. Japan Sea; Lee et al., 1999), suggesting that it may have formed as a back-arc basin adjacent to the African margin.

Further support for the existence of oceanic crust has been provided from interpreting the transmission of regional waveforms as recorded by the Italian Seismic Network (Mele, 2001). These data show that the regional shear phase (L_g), that typically propagates within continental crust, has not been transmitted in the oceanic crust of the Ionian Sea (Mele, 2001). Extensional structures recognised in seismic reflection profiles through Malta and Apulia escarpments (Figure 4.6a,c), may reflect the conjugate passive margins of the Ionian basin (Catalano et al., 2001).

Subduction of the Ionian lithosphere is currently taking place in the Calabrian arc (Caputo et al., 1972; Cristofolini et al., 1985) and the Hellenic arc (Le Pichon and Angelier, 1979) (Figure 4.6a). In both arcs, the subduction hinges underwent rollback that resulted in the formation of extensional back-arc basins: the southern Tyrrhenian Sea in the Calabrian arc and the Aegean Sea in the Hellenic Arc. The rollback of the Ionian slab is particularly well studied in the Calabrian Arc, where rollback in excess of 400-500 km has occurred since 5-6 Ma (Malinverno and Ryan, 1986; Faccenna et al., 1996; 2001a; 2001b; Carminati et al., 1998; Gvirtzman and Nur, 1999; 2001; Wortel and Spakman, 2000). Such a rapid migration of the subduction hinge is considered by many authors to result from a gravitational collapse of the subducting slab, driven by the negative buoyancy of the slab relative to the asthenosphere (e.g. Elsasser, 1971; Dewey, 1980; Garfunkel et al., 1986). Examples from nature suggest that fast rollback occur when the subducting slab is made up of oceanic lithosphere (e.g. Molnar and Atwater, 1978; Uyeda and Kanamori, 1979; Jarrard, 1986). It is stressed that subduction of continental material is likely to decrease the negative buoyancy, and thus to impede subduction rollback. Therefore, the lithosphere of the Ionian Sea is most likely of an oceanic nature.

4.4.2. Age of the Ionian crust

If the Ionian Sea consists of oceanic lithosphere, when did sea floor spreading take place? Della Vedova et al. (1989) have estimated an age of 180-200 Ma for the Ionian crust based on the relatively low heat flow values (34 mW m^{-2}) measured in the Ionian abyssal plain. This age may correspond to the latest stage of spreading because it postdates a long history of deep-sea sedimentation in the region. Deep-water pelagic sediments were deposited in western Sicily during the Permian and the Triassic (Sosio Valley and Sicanian basin; Catalano et al., 1991) and from Middle Triassic to Oligocene times in the Lagonegro Basin (southern Apennines; Figure 4.6a) (Wood, 1981). The subsidence pattern on the margins of the Ionian Sea infers the existence of a deep basin from at least since the Triassic (Stampfli and Mosar, 1999). Similar subsidence patterns are found in other localities throughout the Neotethyan margins (i.e. Oman, Zagros, Levant etc), suggesting a concomitant opening of a Permian-Triassic Ocean that stretched from central Asia to the Eastern Mediterranean and the Ionian Sea (Stampfli et al., 2001b). This concept is supported by comparable Permian deep-water faunas found in Sicily, Crete and on top of oceanic crust in Oman (Catalano et al., 1991). However, the timing of the transition from rifting to spreading remains poorly constrained. Stampfli and Mosar (1999) and Stampfli et al (2001b) have proposed that spreading occurred during the Permian or the Triassic, suggesting that the Ionian crust may be one of the oldest occurrences of autochthonous oceanic crust.

The opening the Ionian Sea implies the existence of a divergence plate boundary between Adria and Africa at that time. This means that a small amount of independent motion between Adria and Africa must have occurred regardless of their general coherent motion inferred from palaeomagnetic data. Finetti (1985) and Catalano et al. (2001) have suggested that sea-floor spreading may have continued during the Jurassic. No evidence for younger deformation is recognised in the Ionian Sea and it is assumed that spreading ceased sometime during the Jurassic.

4.5. Discussion

Constraints on the motion of Adria imply no significant relative motion between Adria and Africa since the last stage of opening in the Ionian Sea during the Jurassic. Sea-floor spreading of the Ionian Sea took place sometime between the Late Permian and the Jurassic, and resulted in a small relative motion between Adria and Africa that may not be recognised within the error limits of the palaeomagnetic results.

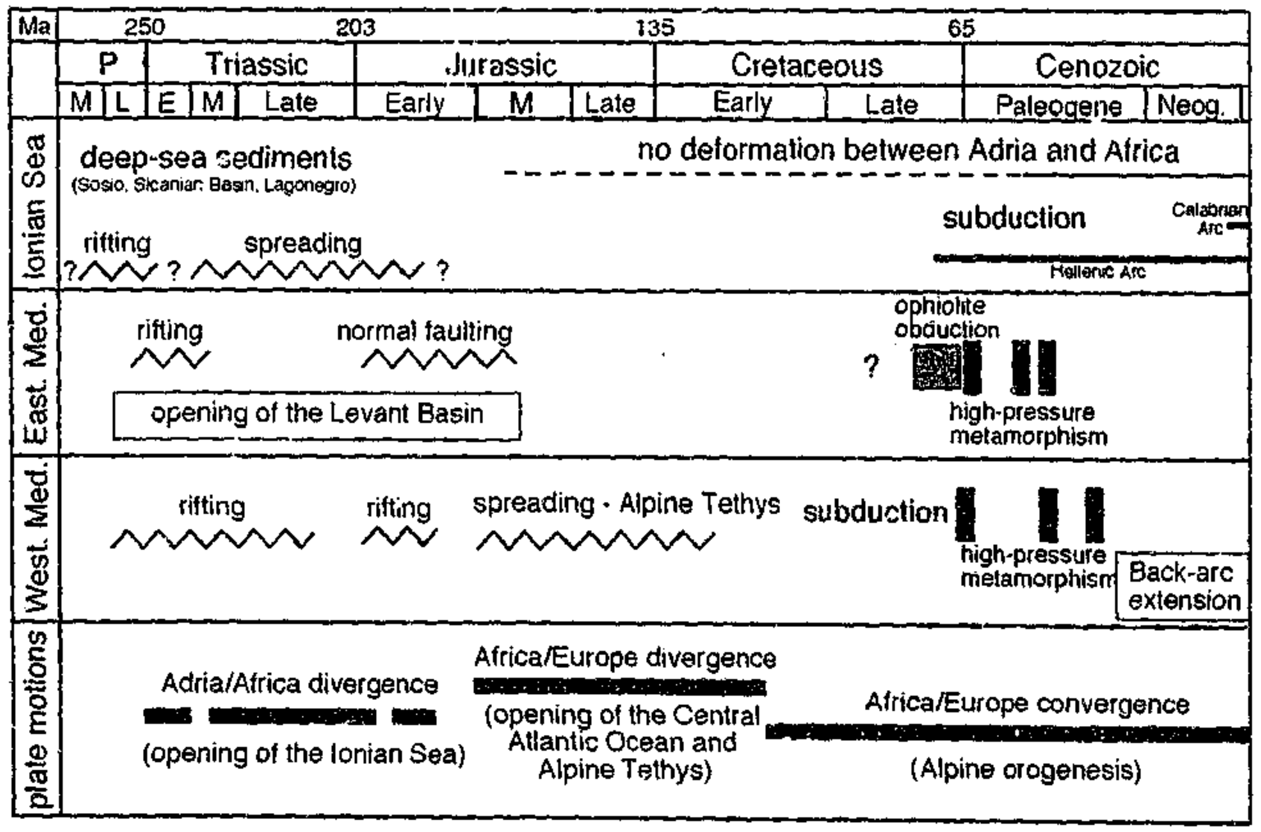


Figure 4.7. Summary of geological events in the Mediterranean region since the Late Permian and their relations with plate motions. Data on the Eastern Mediterranean are from Garfunkel (1998). Time scale is after Remane et al. (2000).

It has been noted that a geometric problem emerges when Adria is rotated back in time from its present position using rotational parameters for the motion of Africa and Iberia with respect to Europe (e.g. Stampfli et al., 1998; Stampfli and Mosar, 1999; Wortmann et al., 2001). The movement of Adria and Africa as a rigid entity results in space problems in the Early-Middle Jurassic configuration, in which Adria overlaps Iberia and southern Europe. In order to resolve this issue, Stampfli et al. (1998) have suggested a lateral northwestward displacement of southern Adria (Apulia) with respect to the northern part of Adria during the Late Cretaceous. This requires the existence of a plate boundary

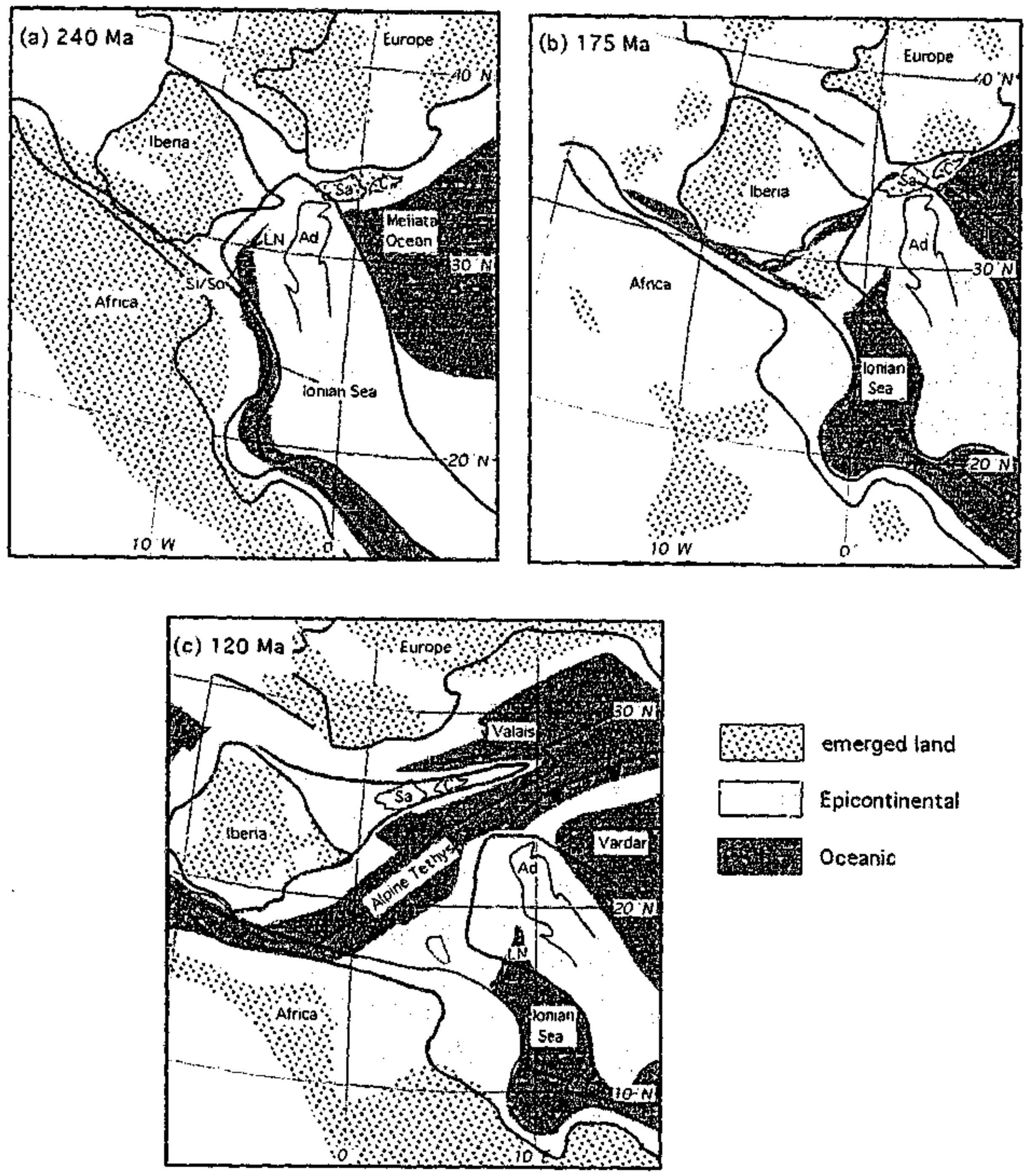


Figure 4.8. Schematic palaeogeographic reconstruction of Adria during the (a) Triassic; (b) Jurassic; and (c) Early Cretaceous. Plate kinematics is after Rosenbaum et al. (2002a) and palaeogeography is after Gaetani et al. (2000), Masse et al. (2000); Thierry et al. (2000), Stampfli et al. (2001a) and Bosellini (2002). Ad, Adria; C, Corsica; LN, Lagonegro; Sa, Sardinia; Si, Sicani; So, Sosio.

within Adria, for which there is no geological evidence. However, Rosenbaum et al. (2002a) have recently shown that the geometric problem of Adria during the Jurassic can be resolved simply by using revised rotational parameters for the motions of Africa, Europe and Iberia, without the necessity to assume a non-rigid behaviour of Adria or the independent motion of an Adriatic plate.

The main geologic events associated with the tectonic evolution of Adria and its surrounding regions are summarised in Figure 4.7, and a schematic palaeogeographic reconstruction of Adria during the Mesozoic is illustrated in Figure 4.8. In this reconstruction, the relative motions of Africa, Europe and Iberia (after Rosenbaum et al., 2002a) are recalculated from a hotspots frame of reference. The reconstruction also shows the existence of a narrow Early Triassic deep-sea basin (Figure 4.8a) in Sicily and in the southern Apennines, represented by the deep-sea sediments of Sosio/Sicani and Lagonegro complexes, respectively. The basin in which these sediments were deposited marks the westernmost termination of the Permian-Triassic Ionian rift and the Neotethys Ocean (Catalano et al., 1991; Stampfli et al., 2001b). At least in this part of the Neotethys Ocean, the seismic structure of the oceanic crust and the relatively large thickness of the oceanic layers suggest that sea-floor spreading took place in a marginal back-arc basin. However, the geodynamic framework associated with the opening of this back-arc basin is relatively unclear because there is no evidence for the existence of a south-dipping subduction zone along the African margin that could be responsible for rifting and back-arc extension in the overriding African plate.

Spreading in the Ionian Sea was completed during the Jurassic (Figure 4.8b). At the same time, Triassic and Jurassic rift basins in Western Europe evolved into ocean basins. The evolution of these new oceans, the Central Atlantic Ocean and Alpine Tethys (also called Liguride Ocean), is relatively well constrained by reconstruction of magnetic isochrons in the Atlantic Ocean (Rosenbaum et al., 2002a).

During the Early Cretaceous (Figure 4.8c), the opening of Alpine Tethys was completed and convergence between Africa and Europe gave rise to Alpine orogenesis. In several Mesozoic reconstructions (e.g. Catalano et al., 2001; Faccenna et al., 2001b), the Ionian Sea has been shown to be connected to the oceanic domains of the western Mediterranean region (i.e. Alpine Tethys/Liguride Ocean) by an oceanic passage. This is not the case in our reconstruction, where the Early Mesozoic Ionian Sea is separated from the Jurassic Alpine Tethys/Liguride Ocean by a narrow continental bridge. This is based on the evidence for contemporaneous occurrences of similar types of dinosaurs in Apulia, Istria and Africa, which suggest migration over a Cretaceous land bridge (Bosellini, 2002); and the occurrence of Numidian flysch of Mid-Tertiary age both in Sicily and the southern Apennines (Patacca et al., 1992). Nevertheless, the existence of a narrow (300-500 km) oceanic passage cannot be ruled out, because the closure of such ocean would not necessarily be recognised with the current resolution of palaeomagnetic data.

4.6. Conclusions

Based on analysis of palaeomagnetic, geophysical and geological data, the following can be concluded:

- Adria and Africa underwent a similar motion path since the Late Paleozoic, suggesting that Adria was an African promontory during most of this period. This, however, does not preclude the possibility that minor rotations occurred between Adria and Africa, as for example during the opening of the Ionian Sea. These relative motions cannot be distinguished within the resolution of the data.
- The opening of the Ionian Sea took place from the Permian-Triassic to the Jurassic times. Its oceanic nature is supported by geophysical evidence, and is consistent with the history of subduction rollback in the Calabrian Arc. Therefore, as long as sea floor spreading took place, the Ionian Sea was a divergent plate boundary between Adria and Africa.
- Geological and geophysical evidence suggests that the Ionian Sea ceased to function as a divergent plate boundary at least since the Jurassic, and since then Adria has again moved as an African promontory attached to the African plate.

CHAPTER 5

RECONSTRUCTION OF THE TECTONIC EVOLUTION OF THE WESTERN MEDITERRANEAN SINCE THE OLIGOCENE

That there was such a Tethys – an open seaway right through to Australia – is one of the most certain facts of palaeogeography. What a complicated business it is to reconstruct it from a tangle of trends and cross-trends all the way from Spain to New Guinea. Yet, if one makes the single induction that the visible bends in the great orogenic belts are impressed strains, undo them, and there is the Tethys, simple and entire!

S. Warren Carey, A Tectonic Approach to Continental Drift (1958)

FOREWORD: CHAPTER 5

This chapter presents a kinematic model of the tectonic evolution of the western Mediterranean since the Oligocene. The model emphasises the role of subduction rollback in the western Mediterranean, and was particularly inspired by the paper of Lonergan and White (1997) on the Rif-Betic orogenic belt. Modelling has been performed using the PlatyPlus software package developed by Cecile Duboz and Gordon Lister at the Australian Crustal Research Centre. Cecile is also acknowledged for providing technical assistance and support. The reconstruction model has benefited from discussions with various people, including Gordon Lister, Laurent Jolivet, Dov Avigad, Zvi Ben-Avraham, Celal Sengor, Maarten Krabbendam, Wouter Schellart, Peter Betts and David Giles. I also wish to thank Robert Hall and Dov Avigad for reviewing the manuscript, Mara Pavlidis for providing grammatical assistance and Megan Hough for technical support. A slightly modified version of this chapter has been published in the *Journal of the Virtual Explorer*[*].

[*] Rosenbaum, G., Lister, G.S. and Duboz, C., 2002. Reconstruction of the tectonic evolution of the western Mediterranean since the Oligocene. *Journal of the Virtual Explorer*, 8: 107-130.

5. RECONSTRUCTION OF THE TECTONIC EVOLUTION OF THE WESTERN MEDITERRANEAN SINCE THE OLIGOCENE

Abstract

This chapter presents a tectonic synthesis of the tectonic evolution of the western Mediterranean since the Oligocene. This work is based on data derived from different geological datasets, such as structural geology, the distribution of metamorphic rocks, magmatic activity, sedimentary patterns, palaeomagnetic data and geophysics. Reconstruction was performed using an interactive software package (PlatyPlus), which enabled me to apply rotational motions to numerous microplates and continental terranes involved in the evolution of the western Mediterranean basins. Boundary conditions are provided by the relative motions of Africa and Iberia with respect to Europe, and the Adriatic plate is considered here as an African promontory. The reconstruction shows that during Alpine orogenesis, a very wide zone in the interface between Africa and Europe underwent extension. Extensional tectonics was governed by rollback of subduction zones triggered by gravitational instability of old and dense oceanic lithosphere. Back-arc extension occurred at the overriding plates as a result of slow convergence rates combined with rapid subduction rollback. This mechanism can account for the evolution of the majority of the post-Oligocene extensional systems in the western Mediterranean. Moreover, extension led to drifting and rotations of continental terranes towards the retreating slabs in excess of 100-800 km. These terranes - Corsica, Sardinia, the Balearic Islands, the Kabylies blocks, Calabria and the Rif-Betic - drifted as long as subduction rollback took place, and were eventually accreted to the adjacent continents. It is concluded that large-scale horizontal motions associated with subduction rollback, back-arc extension and accretion of allochthonous terranes played a fundamental role during Alpine orogenesis.

5.1. Introduction

The geological evolution of the western Mediterranean exhibits complicated interactions between orogenic processes and widespread extensional tectonics. The region is located in a convergent plate margin separating Africa and Europe, and consists of marine basins - the Alboran Sea, the Algerian-Provençal Basin, the Valencia Trough, the Ligurian Sea and the Tyrrhenian Sea (Figure 5.1) - which

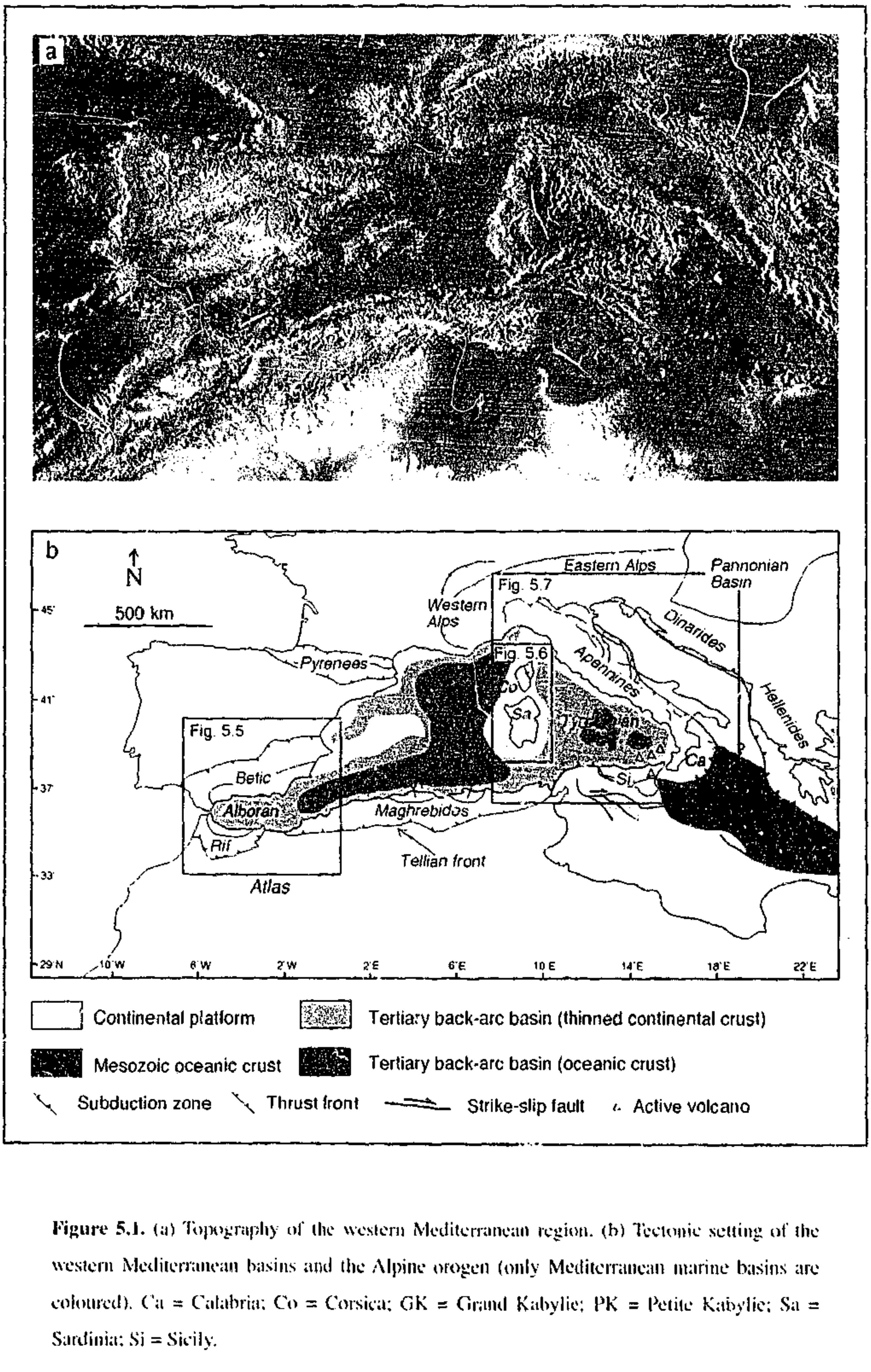


Figure 5.1. (a) Topography of the western Mediterranean region. (b) Tectonic setting of the western Mediterranean basins and the Alpine orogen (only Mediterranean marine basins are coloured). Ca = Calabria; Co = Corsica; GK = Grand Kabylie; PK = Petite Kabylie; Sa = Sardinia; Si = Sicily.

formed as back-arc basins since the Oligocene. The evolution of these basins, simultaneously with ongoing convergence of Africa with respect to Europe, has been the subject of numerous studies (e.g. Stanley and Wezel, 1985; Durand et al., 1999). Widespread extension associated with the formation of these basins led to considerable thinning of the continental crust (i.e., in the Alboran Sea and the northern Tyrrhenian) or to the local initiation of sea floor spreading (i.e., in the southern Tyrrhenian and Provençal Basin). Furthermore, extensional tectonism in the western Mediterranean was coeval with orogenesis in the adjacent mountain chains of the Rif-Betic cordillera, the Maghrebides of northern Africa and Sicily, the Apennines, the Alps and the Dinarides (Malinverno and Ryan, 1986; Crespo-Blanc et al., 1994; Tricart et al., 1994; Cello et al., 1996; Azañón et al., 1997; Frizon de Lamotte et al., 2000; Faccenna et al., 2001b) (Figure 5.1).

The simultaneous formation of extensional basins together with thrusting and folding in adjacent mountain belts has led to several tectonic models that acknowledge the role of large-scale horizontal motions associated with the retreat of the subduction trench (hereafter termed subduction rollback) (Malinverno and Ryan, 1986; Royden, 1993b; Lonergan and White, 1997). These models provide an explanation for the origin of allochthonous terranes, which drifted great distances to their present locations (e.g., Calabria). However, some issues are yet to be resolved and have been the subject of considerable debate. Different models have been proposed to explain the evolution of the Alboran Sea, namely, as a back-arc basin associated with a retreating slab (Lonergan and White, 1997), or as the result of an extensional collapse of thickened lithosphere (Platt and Vissers, 1989; Houseman, 1996). The evolution of the Tyrrhenian Sea is also controversial with some fundamental problems in the current explanations of the evolution of this basin.

In this work, I aim to develop a coherent visual reconstruction that will best explain the large-scale tectonics of the western Mediterranean region. I use a wealth of accumulated knowledge as published in the literature, as well as a new software package that provides the ability to perform an interactive reconstruction. The reconstruction is presented as an animation, which clearly demonstrates some fundamental features seen in convergent plate margins. It shows the complex interactions between subduction processes, horizontal extension, block rotations and accretion events, and it emphasises the roles of subduction rollback and the episodic accretion of allochthonous terranes during orogenesis.

5.2. Summary of previous works

Over the course of the last century, and particularly since the emergence of the modern theory of plate tectonics, numerous studies have aimed to reconstruct the evolution of the Mediterranean basins in the context of the Alpine orogeny. The main concepts used in these reconstructions (as well as in the present study) correspond to continental drift, microplate rotations and migration of subduction zones. They were introduced as early as 80 years ago in the outstanding tectonic synthesis of Argand (1924) (Figure 5.2a). In an early attempt to reconstruct the tectonic evolution of the Mediterranean region. Carey (1958) demonstrated that the unbending of arcuate mountain belts throughout the western Tethys yielded the reassembly of Spain, the Balearic Islands, Corsica, Sardinia, Sicily and Italy in the northwestern margin of the Tethys Ocean (Figure 5.2b). Subsequent reconstruction models have generally followed Carey's ideas concerning the palaeo-position of these terranes (Smith, 1971; Alvarez et al., 1974; Boccaletti and Guazzone, 1974; Biju-Duval et al., 1977; Cohen, 1980) (Figure 5.2c,d), although the significance of the Cenozoic deformation was not always recognised (e.g. Smith, 1971).

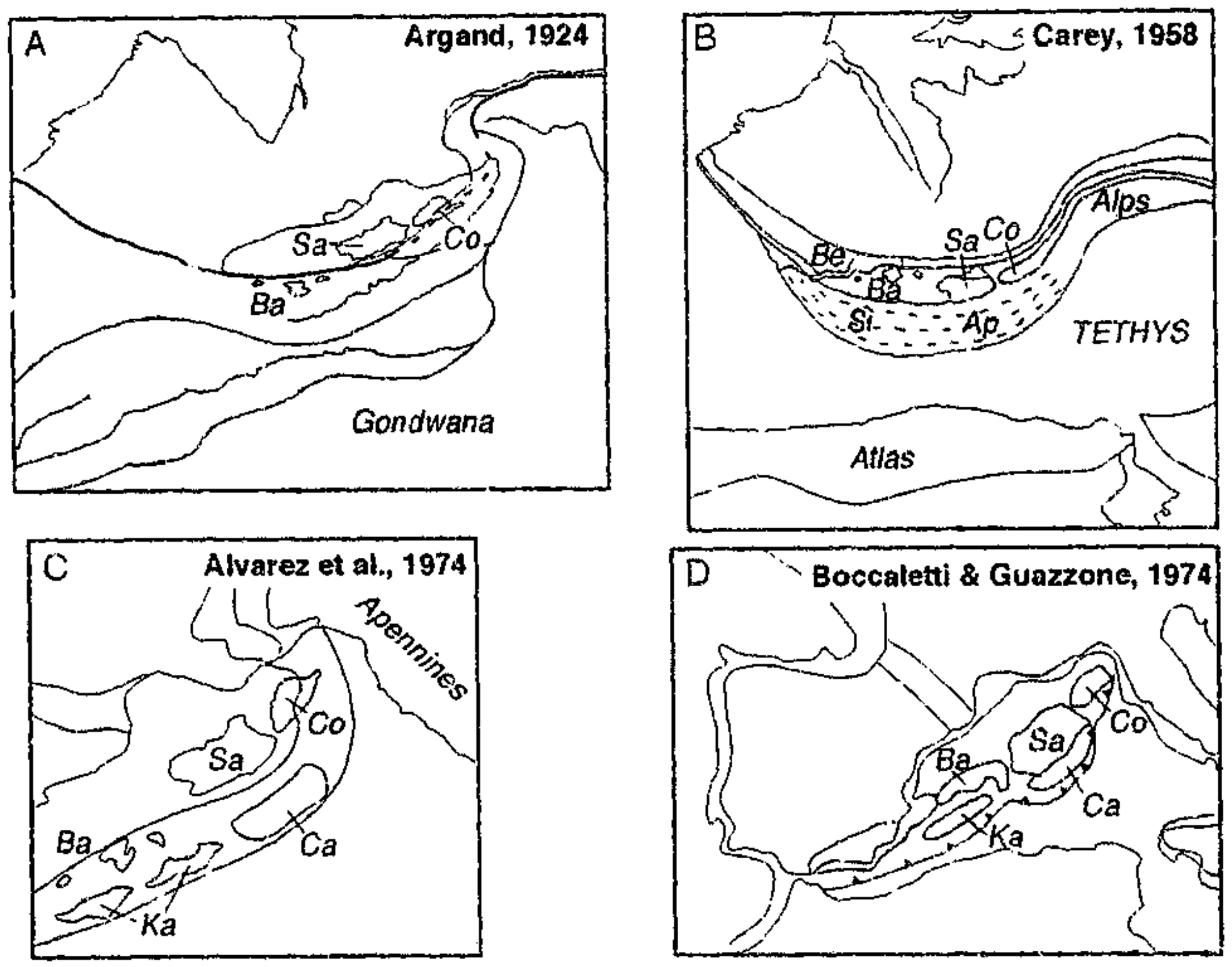


Figure 5.2. Examples of earlier reconstructions showing the western Mediterranean prior to the opening of Late Cenozoic extensional basins. (a) Argand (1924); (b) Carey (1958); (c) Alvarez et al. (1974); (d) Boccaletti and Guazzone (1974). Ap = Apennines; Ba = Balearic Islands; Be = Betic; Ca = Calabria; Co = Corsica; Ka = Kabylies; Sa = Sardinia; Si = Sicily.

During the 1970s, it was suggested that the western Mediterranean basins are relatively late tectonic features formed progressively since the Late Oligocene (Dewey et al., 1973; Alvarez et al., 1974; Biju-Duval et al., 1977). This conclusion was strongly supported by evidence of Oligocene and Miocene extensional deformation and syn-rift deposits on the margins of the western

Mediterranean basins (e.g. Cherchi and Montadert, 1982; Rehault et al., 1985; Bartrina et al., 1992, and many others). Thus, the western Mediterranean basins are essentially different from the Eastern Mediterranean, where the remnants of Mesozoic oceanic crust (Neotethys) are probably preserved below the sediments (De Voogd et al., 1992; Ben-Avraham et al., 2002). Mesozoic oceanic crust is not found on the floor of the western Mediterranean basins. However, the existence of consumed oceanic basins in this region is implied by reconstruction models (e.g. Dercourt et al., 1986; Ricou et al., 1986) and by the occurrence of ophiolitic complexes within the adjacent fold-and-thrust belts (Ricou et al., 1986; Knott, 1987).

The configuration of extensional basins surrounded by continuous mountain belts has been commonly interpreted as the result of back-arc extension (Boccaletti and Guazzone, 1974; Biju-Duval et al., 1977; Cohen, 1980; Ricou et al., 1986). In most reconstructions, it has been stressed that back-arc extension led to drifting of continental blocks and to large-scale block rotations (Dewey et al., 1973; 1989; Alvarez et al., 1974; Biju-Duval et al., 1977). Corsica and Sardinia underwent counterclockwise rotations during the opening of the Ligurian Sea, whereas the opening of the Valencia Trough was accompanied by clockwise rotations of the Balearic Islands (Montigny et al., 1981; Pares et al., 1992). The Kabylies and the Calabrian blocks, which had been deformed and metamorphosed in the Alpine orogen, migrated to their present locations during the opening of the Algero-Provençal Basin and the Tyrrhenian Sea, respectively (Alvarez et al., 1974; Cohen, 1980; Dewey et al., 1989). Boccaletti and Guazzano (1974) have further suggested a southward and eastward migration of subduction arcs during the formation of the western Mediterranean basins. Similar ideas explained the existence of extensional back-arc basins in convergent margins (Dewey, 1980). In the western Mediterranean, it led to the recognition of subduction rollback as an important driving mechanism in the tectonic evolution of the region (Rehault et al., 1985; Malinverno and Ryan, 1986; Royden, 1993b; Lonergan and White, 1997).

The mechanism of subduction rollback has been discussed by Elsasser (1971), Molnar and Atwater (1978), Dewey (1980) and Royden (1993a; 1993b). These authors have suggested that subduction rollback is the result of a negative buoyancy of the subducting slab relative to the asthenosphere (Figure 5.3), obtained when the subducting slab is cold and dense, as in the case of oceanic slabs older than ~50 Ma (Molnar and Atwater, 1978). It results in a vertical sinking of the subducting lithosphere beneath the asthenosphere, which can lead to a regressive motion of the subduction hinge (Lonergan and White, 1997) (Figure 5.3). As rollback occurs, it produces a potential vacant region, which can either be supported by convergence that matches or exceeds the rates of the retreating hinge, or by back-arc extension in the overriding lithosphere (Royden, 1993a). In the works of Rehault et al. (1985), Malinverno and Ryan (1986), Royden (1993b) and Lonergan and White (1997), the evolution of the western Mediterranean basins has been mainly attributed to the rollback of a NNW dipping subduction zone. The reconstruction shown here largely reflects ideas previously presented in these works.

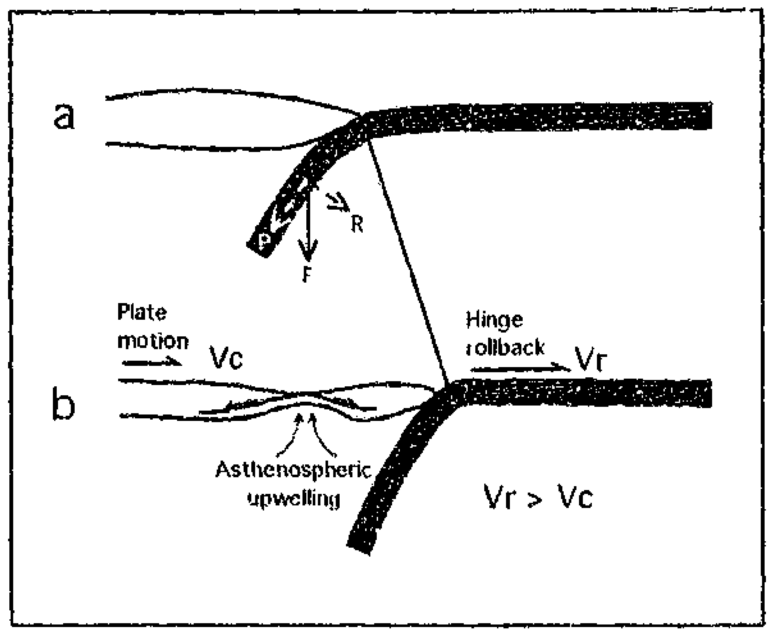


Figure 5.3. Simplified cross section showing the evolution of subduction rollback (modified after Lonergan and White, 1997). (a) P and R are two components of the vertical negative buoyancy (F) of the subducting slab. If the subducting slab is cold and dense, the component R cannot be supported by the mantle asthenosphere, and the subduction zone is pulled backward. (b) Formation of a back-arc extensional basin when the rate of subduction rollback (Vr) exceeds the rate of convergence (Vc).

5.3. The geology of the western Mediterranean region

In this section, I briefly discuss the main characteristics of geological terranes that were incorporated in the post-Oligocene evolution of the western Mediterranean. Time relationships between the different terranes are presented in Figure 5.4.

5.3.1. Rif-Betic cordillera

The mountains of the Betic in southern Spain and the Rif in northern Morocco surround the Alboran Sea to form an arc shaped orogenic belt in the westernmost Mediterranean (Figure 5.5). This belt marks the western terminus of the Alpine orogen. The rocks of the Rif-Betic cordillera are usually divided to three main zones: the Internal Zone, the External Zone and the Flysch Zone. The Internal Zone consists of allochthonous Paleozoic to Early Miocene rocks, which were thrust onto the External Zone during the Miocene (Crespo-Blanc et al., 1994; Crespo-Blanc and Campos, 2001). Alpine deformation and metamorphism affected the Internal Zone during the Cretaceous and the Tertiary, and crustal rocks were buried to great depths and underwent high-pressure metamorphism (De Jong, 1990). The External Zone consists of Mesozoic to Tertiary rocks, which represent the passive margin of Africa and Iberia deforming during Alpine orogeny. The Flysch Zone mainly consists of Early Cretaceous to Early Miocene deep marine clastic deposits (Wildi, 1983).

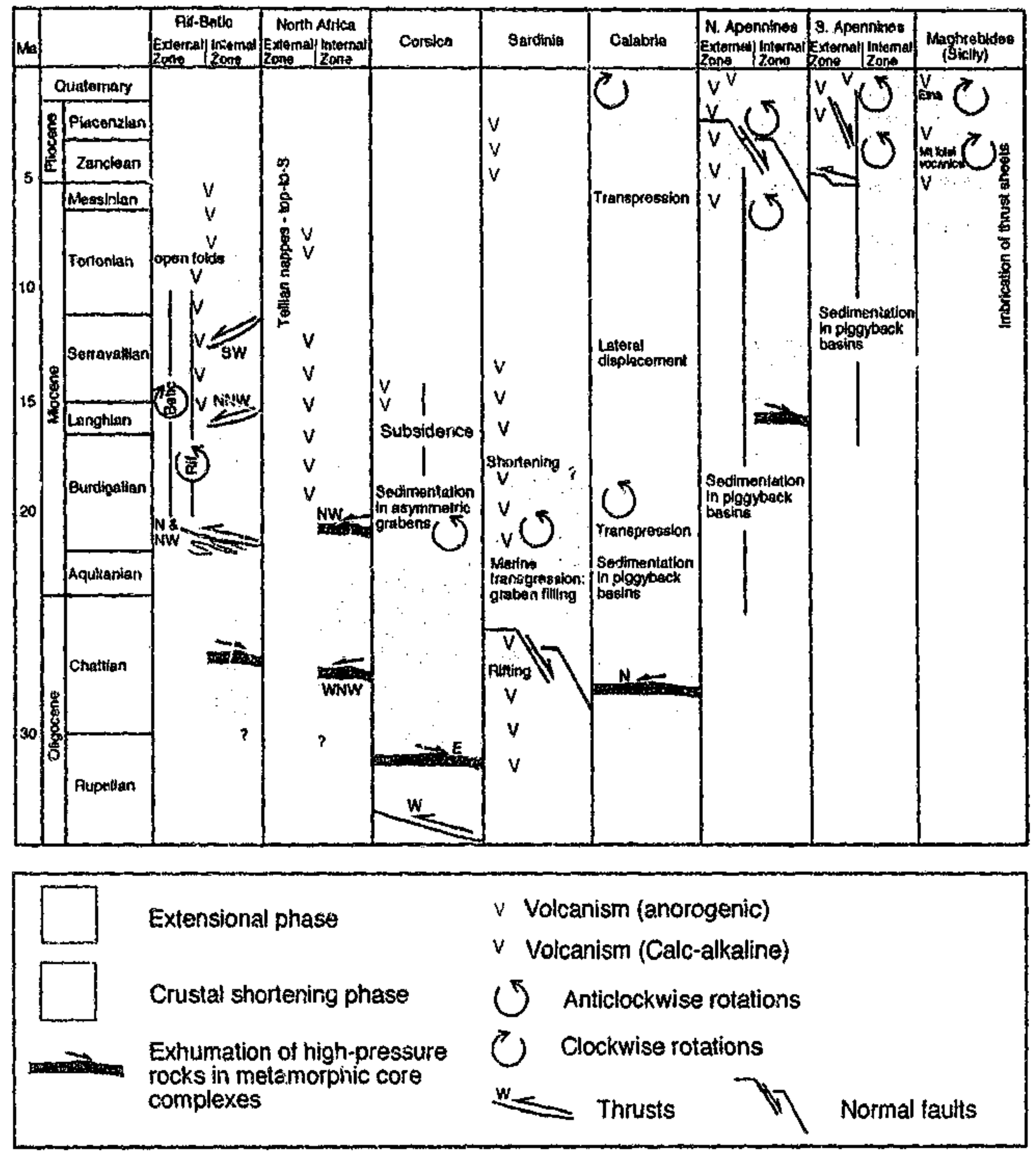


Figure 5.4. Time-space diagram of the tectonic activity in the western Mediterranean (time scale after Palmer, 1983). Data compiled from Boccaletti et al. (1990), Oldow et al. (1990), van Dijk and Okkes (1991), Caby and Hammor (1992), Carmignani et al. (1994), Crespo-Blanc et al. (1994), Ferranti et al. (1996), Saadallah and Caby (1996), Azañon et al. (1997), Sowerbutts (2000) and Crespo-Blanc and Campos (2001).

Structural and metamorphic relationships in the Rif-Betic cordillera, particularly from the Internal Zone, show that several contractional and extensional episodes took place. The earliest extension was probably associated with rapid isothermal exhumation of high-pressure rocks in the Late Oligocene – Early Miocene, and probably commenced at ~30 Ma (Azañón et al., 1997; Platt et al., 1998). A few authors have demonstrated that coeval with crustal shortening in the External Zone, earlier thrust faults in the Internal Zone were rejuvenated as extensional detachments (Azañón et al., 1997; Martínez-Martínez and Azañón, 1997). Extension led to vertical thinning and isothermal exhumation of high-pressure rocks now exposed beneath low-angle normal faults (Platt et al., 1983; Azañón et al., 1997; Balanyá et al., 1997). Furthermore, during the Early Neogene, a slab of diamond-bearing mantle peridotites was exhumed and juxtaposed amidst crustal rocks of the Internal Zone (Van der Val and Vissers, 1993). Radiometric dating of the high-pressure rocks suggests a rapid exhumation during the Late Oligocene - Early Miocene (27-18 Ma) (Monié et al., 1994; Platt et al., 1998).

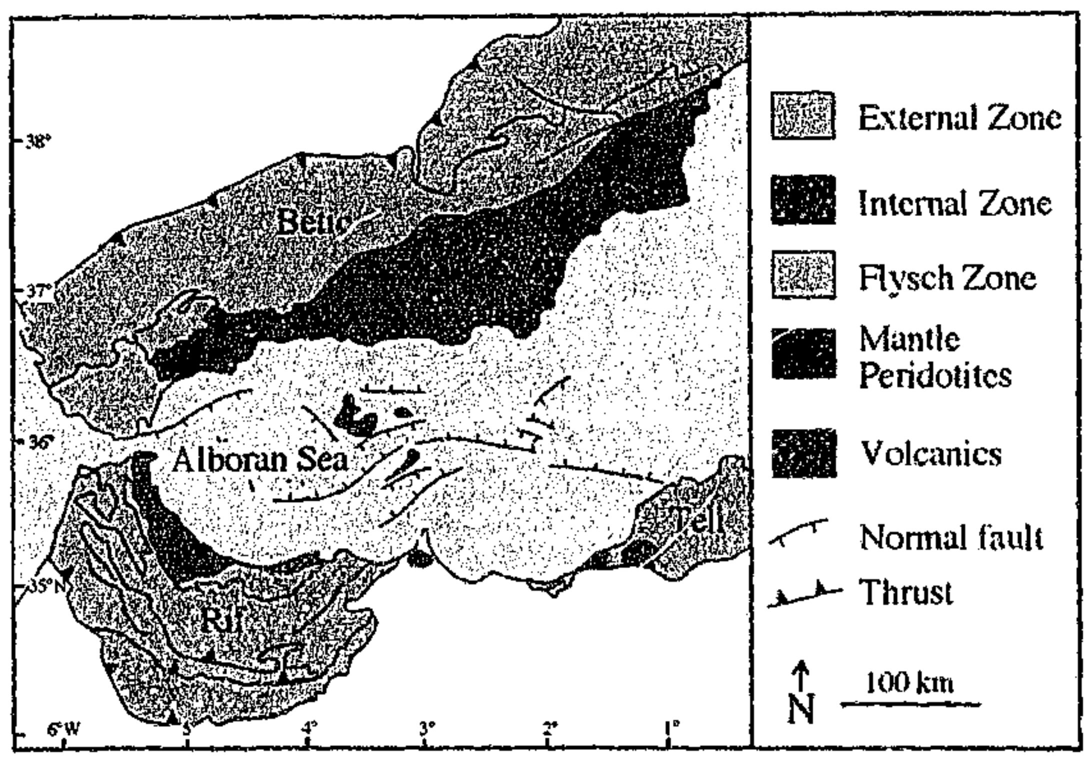


Figure 5.5. Geological map of the Alboran Sea and the Rif-Betic cordillera, after Platt and Vissers (1989).

The floor of the Alboran Sea consists of rocks that are similar to those found in the Rif-Betic cordillera, and covered by Early Miocene syn-rift deposits and post-rift marine sediments (Comas et al., 1992; Platt et al., 1998). It is therefore evident that the region was subjected to widespread extension during the Middle Miocene that led to the formation of the Alboran Sea. However, contemporaneously with extension in the Alboran Sea, thin-skinned thrusting and folding took place in the External Zones of the Rif-Betic cordillera (Platt and Vissers, 1989; Platzman et al., 1993; Crespo-Blanc and Campos, 2001). This deformation was accompanied by a considerable amount of block rotations around vertical axes, with clockwise rotations in the Rif and counterclockwise rotations in the Betic (Allerton et al., 1993; Platzman et al., 1993; Lonergan and White, 1997).

5.3.2 The Maghrebides and Kabylies

The mountain chain of the Maghrebides in Northern Africa and in Sicily consists of a stack of south-verging thrust sheets that bridges the Apennines of Italy with the Rif Mountains of Morocco (Figure 5.1). Most rocks exposed in the Maghrebides are non-metamorphic sedimentary units of Mesozoic to Early Miocene age deposited on the southern margin of the Tethys Ocean (Wildi, 1983). The southern boundary of the Maghrebides is a low-angle thrust fault (the Tellian Front) that delimits the Maghrebide nappes from autochthonous units of the Atlas chain (Wildi, 1983). More internal zones of the orogen are found in northern Algeria and Tunisia (Grand and Petite Kabylies), as well as in a submerged fold-and-thrust belt between Sicily and Tunisia (Compagnoni et al., 1989; Tricart et al., 1994). These internal terranes originated in the Alpine orogen and were overthrust onto the Maghrebides during the opening of the western Mediterranean basins (Cohen, 1980).

The Kabylies consist of a Hercynian basement and Mesozoic sediments, metamorphosed and strongly deformed during Alpine orogeny. Alpine metamorphism took place during the Cretaceous and the Tertiary (Peucat and Bossière, 1981; Monié et al., 1984; 1988; 1992; Cheilletz et al., 1999), and involved metamorphism at high-pressure conditions. High-pressure rocks are presently exposed in metamorphic core complexes, and are directly overlain by rocks that did not suffer Alpine metamorphism (Caby and Hammor, 1992; Saadallah and Caby, 1996). Their exhumation seems to be associated with horizontal extension accommodated along low-angle-normal faults (Caby and Hammor, 1992; Saadallah and Caby, 1996). Deformation along these faults is usually associated with flat lying foliations and mylonitic shear zones with top-to-the-NW sense of shear (Caby and Hammor, 1992; Saadallah and Caby, 1996). Radiometric dating indicates that extensional deformation probably occurred at 25-16 Ma (Monié et al., 1984; 1988; 1992).

Thrusting in the External Maghrebides commenced at the Early/Middle Miocene and was generally directed southward and southeastward (Frizon de Lamotte et al., 2000, and references therein). Since the Tortonian, the locus of major crustal shortening has migrated southward until a ~400 km wide area was structured as a fold-and thrust belt (Tricart et al., 1994).

5.3.3. Corsica and Sardinia

The islands of Corsica and Sardinia consist of a Hercynian basement and a Mesozoic-Tertiary cover which are overthrust (in northern Corsica) by slivers of Alpine origin (Mattauer et al., 1981) (Figure 5.6). The External units probably originated in the southern margin of Europe, and are usually reconstructed to a position adjacent to southern France or northeast Spain (Figure 5.2). The separation of these terranes from Europe is attributed to the post-Oligocene rifting in the Gulf of Lion and the opening of the Ligurian Sea, in which the islands underwent ~30° of counterclockwise rotations (Montigny et al., 1981).

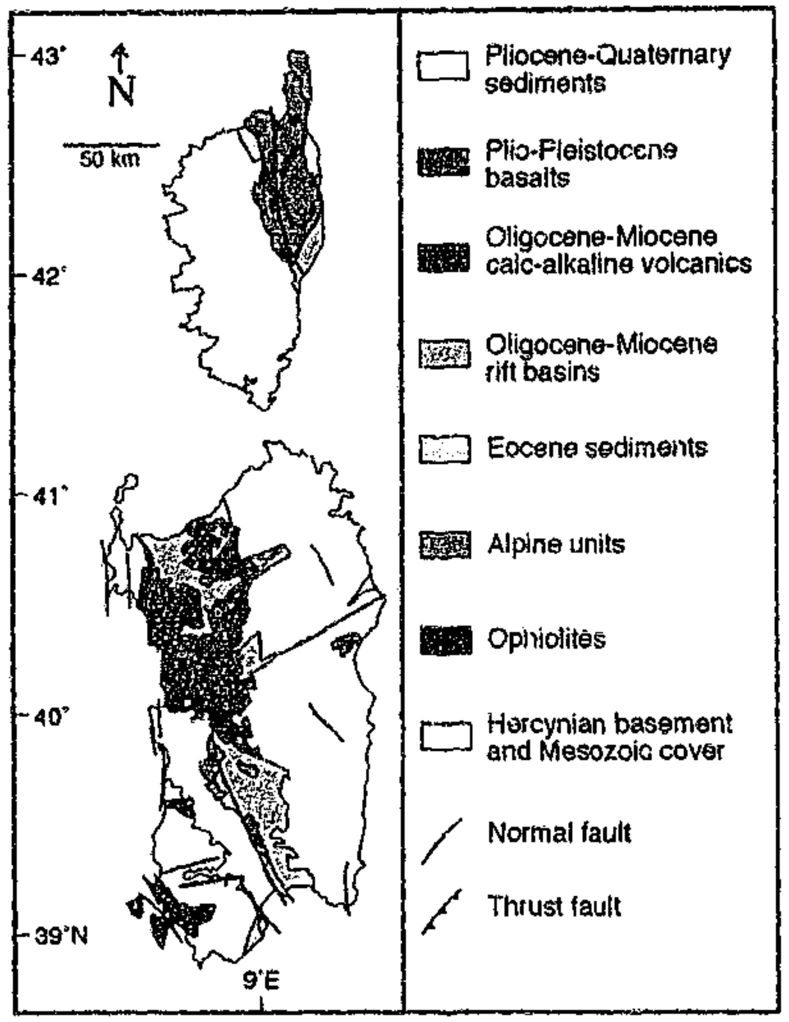


Figure 5.6. Geological map of Corsica and Sardinia, after Cherchi and Montadert (1982) and Jolivet et al. (1990).

In the nappe stack of Alpine Corsica, high-pressure ophiolitic rocks and non-metamorphic rocks are overthrust on top of external (mainly granitic) units. Radiometric ages of the high-pressure rocks in Alpine Corsica indicate that metamorphism took place during the Late Cretaceous and Tertiary (Jolivet et al., 1998; Brunet et al., 2000, and references therein), at the same time when high-pressure metamorphism occurred in the western Alps. Structural evidence shows that Alpine Corsica was thrust over the Hercynian basement with a westward sense of movement and that these structures are overprinted by younger extensional structures with eastward sense of shear (Fournier et al., 1991). Extensional deformation occurred at greenschist-facies conditions and is defined in localised shear zones along extensional detachments (Fournier et al., 1991; Daniel et al., 1996). This style of extension led to isothermal exhumation of the metamorphic core complex now preserved in northeast

Corsica (Jolivet et al., 1990; Fournier et al., 1991). An age of ~32 Ma has been estimated as the commencement of extensional deformation in Corsica (Brunet et al., 2000).

The island of Sardinia consists of a Hercynian basement and a Mesozoic to Eocene sedimentary cover, overlain by syn- and post-rift sediments and volcanics (Cherchi and Montadert, 1982) (Figure 5.6). The island is transected by a N-S-striking rift that forms a succession of tilted blocks filled with Oligocene-Miocene continental to marine sediments (Cherchi and Montadert, 1982). Rifting probably commenced in the middle Oligocene (~30 Ma), and at the end of the Oligocene the trough was deep enough to allow invasion by sea (Cherchi and Montadert, 1982). The latest syn-rift sediments are Aquitanian in age, indicating that rifting ceased at 23-24 Ma (Cherchi and Montadert, 1982). A second phase of extension that postdates the formation of the Sardinian rift took place at mid-Aquitanian to early Burdigalian (23-20 Ma) (Sowerbutts, 2000). Letouzey et al. (1982) have reported Burdigalian (21-17 Ma) NE-SW contractional structures, which seemed to occur simultaneously with a third extensional phase associated with the reactivation of N to NNW trending normal faults (Sowerbutts, 2000).

5.3.4. Calabria

The nappe-structured belt of Calabria, in the southernmost part of the Italian peninsula and the northeastern corner of Sicily, connects the Apennines and the Maghrebides (Figure 5.7). The northern and the southern boundaries of the Calabrian block are major strike-slip faults, the sinistral Sagiento Line and the dextral Taormina Line; their sense of motion indicates that the whole Calabrian block underwent eastern displacement relative to the Apennines and the Maghrebides (Van Dijk and Scheepers, 1995). The rocks in Calabria are remarkably different from those in the Apennine-Maghrebide belts indicating prolonged tectono-metamorphic evolution associated with Alpine orogeny.

The Calabrian nappes are divided into three major tectonic units (Amodio-Morelli et al., 1976). The lowermost unit consists of a Mesozoic carbonate platform that belongs to the margin of Adria. It is overlain by two ophiolitic nappes composed of Mesozoic and Cenozoic sedimentary and ophiolitic rocks partly metamorphosed under high-pressure conditions (Knott, 1987; Cello et al., 1996). The uppermost unit consists of a Paleozoic Hercynian basement and a Mesozoic to Cenozoic sedimentary cover with a strong Alpine signature (Knott, 1987). The latter allochthonous terrane, thereafter referred as the Calabrian block, was progressively emplaced eastward onto the ophiolite-flysch unit and the margin of Adria (Knott, 1987; Dewey et al., 1989; van Dijk and Okkes, 1991).

The age of high-pressure metamorphism in the Calabrian block is ambiguous and varies between 60-35 Ma (Schenk, 1980; Rosseti et al., 2001). During the Oligocene and the Miocene, isothermal exhumation of the high-pressure rocks took place (Rosseti et al., 2001). This led to the emplacement of weakly metamorphosed and nonmetamorphosed rocks on top of high-pressure rocks in tectonic contacts of low-angle extensional detachments (Platt and Compagnoni, 1990; Rosseti et al., 2001).

The commencement of extensional tectonics in Calabria has been inferred as ~30 Ma, based on $^{40}Ar/^{39}Ar$ data associated with ductile extensional structures (Rosseti et al., 2001).

5.3.5. The Apennines

The Apennine belt is a Late Cenozoic fold-and-thrust belt striking parallel to the Italian peninsula from Calabria to the western Alps (Figure 5.7). The nappes of the Apennines predominantly consist of non-metamorphic or weakly metamorphosed Late Triassic to Neogene marine sediments probably deposited on the passive margin of the Adriatic foreland The autochthonous crystalline basement is rarely exposed and is only found in the area of Alpi Apuane tectonic windows (northern Apennines). In these outcrops, low-angle normal faults juxtaposed the allochthonous cover above a Hercynian basement, which was metamorphosed at high-pressure conditions at ~25 Ma (Carmignani and Kligfield, 1990; Carmignani et al., 1994; Jolivet et al., 1998; Brunet et al., 2000). Extensional tectonics during the Oligocene-Miocene has been considered to play an important role in the exhumation of the Hercynian basement (Carmignani et al., 1994; Jolivet et al., 1998).

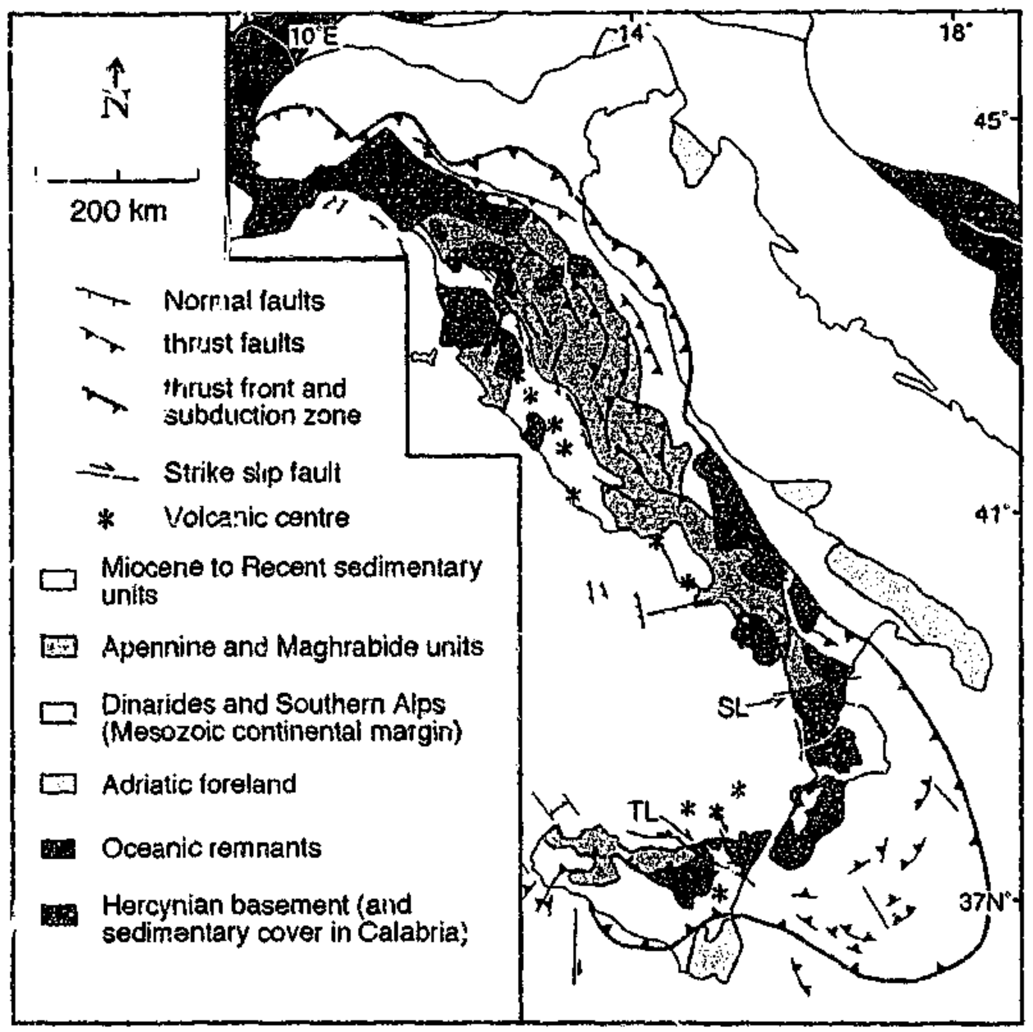


Figure 5.7. Geological map of the Italian peninsula and Sicily, modified after Channell et al. (1979) and Patacca et al. (1993). SL = Sagiento Line; TL = Taormina Line.

Deformation in the Apennines commenced in the Oligocene in the Northern Apennines (Boccaletti et al., 1990a) and has propagated southward since the Late Miocene. During deformation, the front of the Apennine thrust system migrated eastward; it is presently located predominantly in the Adriatic Sea (Jolivet et al., 1998; Brunet et al., 2000). While ongoing thrusting took place in the (eastern) external Apennines, the (western) internal domain was subjected to crustal extension (e.g. Carmignani et al., 1994; Ferranti et al., 1996) forming a series of extensional sedimentary basins that becomes younger towards the east (Patacca et al., 1990).

5.4. Methods and data

5.4.1. PlatyPlus

Reconstruction has been performed using PlatyPlus, a software package developed in the Australian Crustal Research Centre at Monash University. The program provides an interactive platform for tectonic reconstruction on a Unix machine. It enables the user to rotate and translate polygons on a spherical Earth. Polygons are chosen based on geological criteria, and their boundaries are digitised and recorded in ASCII files ('bd' files in Appendix 1). The objects' hierarchy is then defined in 'knox' files (see Appendix 1). Once motions are projected on the PlatyPlus platform, they can be rotated and translated. This can be done either by reading a previously prepared motion file (see Appendix 1) that contains sets of Euler poles of rotations, or by dragging them manually. Thus, known motions are defined in a motion file and applied accordingly, whereas unknown motions are found by trial-and-error experiments using the best available kinematic constraints. This method provides a continuous updating of the motion file during the reconstruction process. Finally, the resulting reconstruction can be viewed in a movie format (see Appendix 2).

5.4.2. Magnetic isochrons

The most robust data used in plate reconstructions are obtained from the existence of magnetic anomalies. In the Atlantic Ocean, anomalies of similar ages (isochrons) are recognised on both sides of the spreading ridge. Thus, fitting of these isochrons can provide the relative motion of the two diverging plates. However, this method cannot be used for convergent plate margins, and is therefore of little importance in the Mediterranean region. In this reconstruction, the only motions constrained by magnetic isochrons are the relative motions of Africa versus Europe and Iberia versus Europe (Table 5.1). These data are based on plate circuit calculations described in Rosenbaum et al. (2002a).

Table 5.1. Motions of Africa and Iberia relative to Europe (after Rosenbaum et al., 2002a).

Time (Ma)	Africa/Europe			Iberia/Europe		
	Lat.	Long.	Rot.	Lat.	Long.	Rot.
5	-13.99	158.23	0.55	0.00	0.00	0.00
10	-13.91	158.94	1.09	78.26	58.40	0.00
15	-14.77	162.89	1.53	78.10	58.72	0.13
20	-17.27	164.31	2.13	63.16	133.18	0.19
25	-24.32	161.90	3.57	-21.55	163.26	0.65
30	-27.24	161.11	5.04	-29.42	165.91	1.31
35	-28.89	160.77	6.41	-30.21	165.32	1.72

5.4.3. Palaeomagnetism

Constraints on the kinematics of deformed allochthonous terranes are obtained from palaeomagnetic studies, which can theoretically provide information about the role of block rotations around vertical axes. However, the regional significance of these rotations in deformed rocks can be ambiguous because of the effect of local deformation. In this reconstruction, I refer only to palaeomagnetic studies that considered the effect of local deformation and provided regional tectonic implications. The amount and timing of block rotations associated with different terranes are presented in Figure 5.8 and are summarised in Table 5.2.

The role of post-Oligocene block rotations in the western Mediterranean is presented in Figure 5.8. The rotation of Corsica and Sardinia is relatively well established by palaeomagnetic constraints that indicate ~30° of counterclockwise rotations after the Late Eocene and prior to the Middle Miocene (De Jong et al., 1973; Montigny et al., 1981; Vigliotti and Langenheim, 1995; Speranza, 1999; Speranza et al., 2002). It has therefore been concluded that Sardinia and Corsica rotated to their present position during the opening of the Ligurian Sea in the Late Oligocene – Early Miocene. Likewise, the Balearic Islands underwent approximately 25° of clockwise rotation during the opening of Valencia Trough (Freeman et al., 1989; Pares et al., 1992).

Post-Oligocene rotations are recognised in allochthonous terranes of the Rif-Betic cordillera and in the deformed External Zones of the African and Iberian margins. In the Betic cordillera, palaeomagnetic studies implied large amount of clockwise rotations (130°-200°) that gradually decrease towards the distal parts of the thrust belt (Platzman, 1992; Allerton et al., 1993; Lonergan and White, 1997). At the same time, rocks in the Rif were subjected to counterclockwise rotations in magnitude of approximately 100° (Platzman et al., 1993). In the following reconstruction, I adopt Lonergan and White's (1997) explanation for the opposite rotational directions on both sides of the arcuate Rif-Betic orogen, which is associated with back-arc extension and a westward subduction rollback.

Most palaeomagnetic results from peninsular Italy show counterclockwise rotations of allochthonous Apennine units (Lowrie and Alvarez, 1974; 1975; Channell, 1992; Dela Pierre et al., 1992; Scheepers and Langereis, 1994; Iorio et al., 1996; Speranza et al., 1998; Muttoni et al., 2000;

Table 5.2. Block rotations inferred from palaeomagnetic data.

Locality	Age of rocks	Inferred rotations	Reference
Betic			
External Betic	Jurassic	20°-68° CW	Osete et al. (1988)
External Betic	L. Jurassic and L. Cretaceous	>60° CW	Platzman (1992)
Eastern Betic (Internal Zone), Southern Spain	L. Jurassic (Miocene folding)	~200° CW (1)	Allerton et al. (1993)
Eastern Subbetic (External Zone), Southern Spain	L. Jurassic (Miocene folding)	68°±22° CW (2)	Allerton (1994)
Western Subbetic (External Zone)	L. Jurassic (Neogene remagnitisation)	~45° CW	Villalain et al. (1996)
Western Internal Zone	Oligocene – E. Miocene	90-140° CW	Calvo et al. (2001)
Rif			
Western External Rif	Jurassic	CCW	Platzman (1992)
Boundary of Internal/External Zone, Rif	Jurassic and Cretaceous	~100° CCW	Platzman et al. (1993)
Balearic Islands			
Mallorca	Jurassic	40° CW	Freeman et al. (1989)
Mallorca and Menorca	Jurassic to Miocene	20° CW (3)	Pares et al. (1992)
Corsica and Sardinia			
Western Sardinia	Middle Tertiary	30° CCW	Montigny et al. (1981)
Calabria			
Messina Straits	Pliocene-Pleistocene	14° and 36° CW	Aifa et al. (1988)
Ionian and Tyrrhenian coasts	Pliocene	10°-20° CW	Scheepers et al. (1994)
Amantea basin, west Calabria	Tortonian-Messinian	19°±11° CW	Speranza et al. (2000)
Italian Peninsula			
Umbria, northern Apennines	L. Cretaceous and Paleogene	43° CCW	Lowrie and Alvarez (1974)
Umbria, northern Apennines	Cretaceous and Eocene	25° CCW (4)	Lowrie and Alvarez (1975)
Gran Sasso, central Apennines	Cretaceous to Pliocene	Up to 90° CCW	Dela Pierre et al. (1992)
Northern Umbria, northern Apennines	Jurassic - Cretaceous	20° CCW (5)	Channell (1992)
Sant'Arcangelo basin, southern Apennines	Pliocene-Pleistocene	22° CCW	Sagnotti (1992)
Apulia	L. Pliocene – E. Pleistocene	No rotation	Scheepers (1992)
Sant'Arcangelo basin, southern Apennines	L. Pliocene – E. Pleistocene	23° CCW	Scheepers et al. (1993)
Central Tyrrhenian coast	Plio-Pleistocene	No rotation	Sagnotti et al. (1994)
Southern Apennines	L. Miocene – Middle Pliocene	39° CCW	Scheepers and Langereis (1994)
Central Apennines	Eocene - Oligocene	35°±5° CCW (6)	Mattei et al. (1995)
Central Apennines	Early-Middle Miocene	12°±13° CW (6)	Mattei et al. (1995)
Tyrrhenian coast, northern Apennines	Late Messinian –Pliocene	No rotation	Mattei et al. (1996)
Monte Matese region, southern Apennines	Miocene	40° CCW (7)	Iorio et al. (1996)
Sibillini-Cingoli, northern Apennines	Messinian	15° CW	Speranza et al. (1997)
Marche-Romagna, northern Apennines	Messinian	~ 20° CCW	Speranza et al. (1997)
Molise region, central Apennines	L. Miocene	35° CCW	Speranza et al. (1998)
Bologna region, northern Apennines	E. Oligocene	52° CCW	Muttoni et al. (2000)
Bologna region, northern Apennines	L. Miocene	28° CCW	Muttoni et al. (2000)
Central northern Apennines	Early Pleistocene	23° CCW	Sagnotti et al. (2000)

Comments:
(1) Late Oligocene and Miocene rotations; (2) Relative rotation of the upper limb of a southeast-vergent recumbent fold
(3) Post Burdigalian rotation; (4) Post-Eocene rotations; (5) Rotation relative to the Southern Alps; (6) Rotation relative to Africa
(7) Rotation relative to Apulia

Sagnotti et al., 2000). However, clockwise rotations have been reported in post-Messinian sediments in the Northern Apennines (Speranza et al., 1997). Palaeomagnetic results from the Apennines are probably biased by the effect of local deformation (Sagnotti, 1992; Scheepers et al., 1993; Mattei et al., 1995), and it is therefore extremely difficult to constrain the kinematics of these terranes. It seems likely, however, that the motion of Apennine nappe thrusts was dominated by counterclockwise rotations during the opening of the Tyrrhenian Sea (ca. 9-5 Ma in the northern Tyrrhenian and 5-0 in the southern Tyrrhenian).

Palaeomagnetic studies from the Calabrian block suggest a counterclockwise rotation in Early – Middle Miocene, followed by ~15-20° clockwise rotations, probably during the Pleistocene (Van Dijk and Scheepers, 1995; Speranza et al., 2000, and references therein). In the Maghrebide belt of Sicily, post-Miocene deformation has been dominated by clockwise rotations, which decrease towards the distal part of the belt (Speranza et al., 1999, and references therein).

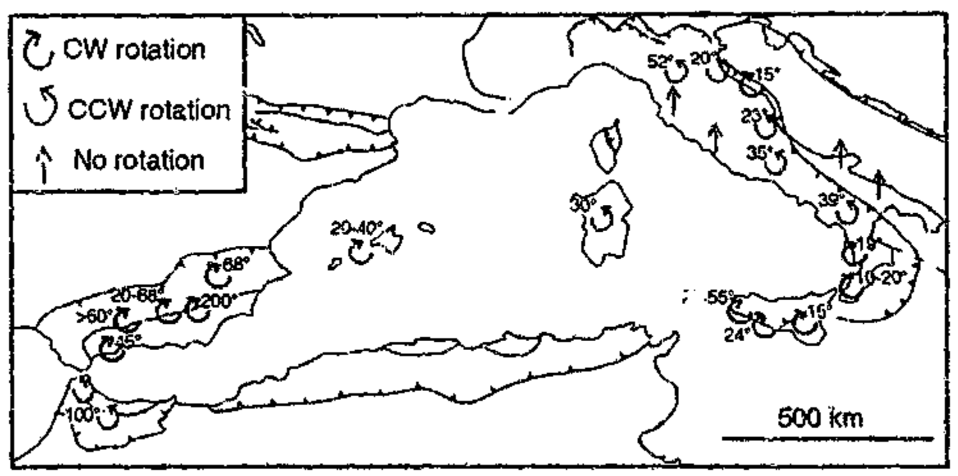


Figure 5.8. Rotations inferred from palaeomagnetic studies throughout the western Mediterranean. References and time constraints are summarised in Table 5.2.

5.4.4. Volcanism

The distribution of calc-alkaline magmatism in the western Mediterranean is summarised in Table 5.3 and presented in Figure 5.9. Calc-alkaline magmatism was probably produced during subduction processes, and the nature of subduction systems can thus be implied from the spatial and temporal distribution of the volcanic rocks (Figure 5.9). Calc-alkaline volcanics show a general younging trend from southern France towards the Apennines, North Africa and the Alboran Sea, suggesting migration of volcanic arcs during subduction rollback. They are locally superimposed by younger alkaline volcanics and oceanic basalts, which were emplaced in the back-arc region.

Evidence for an Oligocene volcanic arc is found in southern France and in Sardinia. In southern France, calc-alkaline magmas erupted between the Oligocene and the Middle Miocene (33-20 Ma) (Bellon, 1981). Based on geochemical criteria, it has been suggested that volcanism was

Table 5.3. Summary of calc alkaline volcanism in the western Mediterranean.

Locality	Type	Age (Ma)	Reference
Ligurian/Valencia region			
Provence	Calc-alkaline	33-20	1
Valencia Trough	Dacites and rheolites	24-18.6	2
North Africa			
Cap Bougaroun, northeast Algeria	High K microgranites and rhyolites	16.4-15.2	4,5
Bejaia-Amizour, north Algeria	Shoshonitic granitoids	16.2-15.3	4
Beni Touffout, northeast Algeria	High K granodiorites and monzogranites	16.2	4,5
Algérios	High K granitoids	16-13.9	4
Cap de Fer Edough, northeast Algeria	High K granitoids	15.9-15.1	4
El Aouana, northeast Algeria	Medium K dacites	15.9-14.5	4
Filfila, northeast Algeria	Shoshonitic syenogranitic	15.3	4,5
La Galite Island, Northern Tunisia	High K granitoids	14.2-10	4
Cherchell, north Algeria	K rich alkaline G	13-11	4
Nefza, Northwest Tunisia	Shoshonitic dacites	12.9-8.2	4
Alboran region			
Alboran	Rhyolites and andesites	18-7	6
Oran, NW Algeria	High-K andesites	11.7-7.5	4
Cabo de Gata, Southern Spain	Pyroxene andesite, dacites, rhyolites	12-10.5	7
Ras Taraf, NE Morocco	Andesites, rhyolites	13.1-12	8
Troi Fourches, NE Morocco	Shoshonitic andesites	6.6-5.4	4
Sardinia			
Sardinia	Andesites and rhyolites	32-13	3
Tyrrhenian region			
Capraia, Northern Tyrrhenian	High-K andesites, dacites	8.5-6.5	3
Montecristo, Northern Tyrrhenian	Cordierite monzogranites	7.3	9
Vercelli seamount, Central Tyrrhenian	Monzogranites	7.2	9
Mt. Capanne, Elba, Northern Tyrrhenian	Granodiorite, monzogranites	6.8-6.2	9
Ponza, central southern Tyrrhenian	Rhyolites	5-2	10
Anchise, central Southern Tyrrhenian	Shoshonitic baslats and dacites	5-2	3,10
Campiglia, Northern Apennines	Monzogranites, syenogranites	?	
Porto Azzurro, Elba, Northern Tyrrhenian	Cordierite monzogranites	5.1	9
Giglio, Northern Tyrrhenian	Cordierite monzogranites	5.1	9
San Vincenzo, Northern Apennines	Cordierite rhyolites	4.7	9
Capraia, Northern Tyrrhenian	High-K andesites, dacites, rhyolites	4.6-3.5	9
Gavorrano, Northern Apennines	Cordierite monzogranites, syenogranites	4.4	9
Castel di Pietra, Northern Apennines	Monzogranites, syenogranites	4.3	9
Monteverdi, Northern Apennines	Cordierite monzogranites	3.8	9
Tolfa	Trachydacites and rhyolites	3.8-2.3	11
Manziana	Rhyolites	3.6	9
Roccastrada	Cordierite rhyolites	2.5-2.2	9
Cerite	Rhyolites	2.4	9
Marsili seamount	High-K calc-alkaline andesites	<1.3	3
Aeolian Islands	Andesites	1-0	3

References are: (1) Bellon (1981); (2) Marti et al. (1992); (3) Savelli (1988) and references therein; (4) Maury et al. (2000) and references therein; (5) Fourcade et al. (2001); (6) Lonergan and White (1997) and references therein; (7) Zeck et al. (2000); (8) El Bakkali et al. (1998); (9) Serri et al. (1993); (10) Argnani and Savelli (1999); (11) Savelli (2000).

associated with a north-dipping subduction system (Rehault et al., 1985). The volcanic arc continued southwestward in Sardinia and Valnecia Trough, although in the latter, volcanism commenced only at 25-24 Ma (Marti et al., 1992).

During the Lower – Middle Miocene, extension took place in the Ligurian Sea and Valencia Trough, and the volcanic arc migrated together with the retreating slab. In Sardinia, ongoing volcanism occurred until 14 Ma (Savelli, 1988). The largest amount of Middle Miocene volcanism is found in the North African coast and in the Alboran Sea (Lonergan and White, 1997; El Bakkali et al., 1998; Maury et al., 2000; Zeck et al., 2000; Fourcade et al., 2001, and references therein). Calc-alkaline volcanism occurred in North Africa between 16.5-8 Ma, following the accretion of the Kabylies blocks to the African margin at ~18-15 Ma (Lonergan and White, 1997). Volcanic activity in the Alboran Sea and the Betic-Rif cordillera has been dated to 15-7 Ma, and was probably associated with an east-dipping subduction zone that retreated westward towards Gibraltar during Middle-Late Miocene (Lonergan and White, 1997).

Since the Late Miocene, the majority of volcanism in the western Mediterranean has been concentrated in the Tyrrhenian Sea (Savelli, 1988; 2000; Serri et al., 1993; Argnani and Savelli, 1999) (Figure 5.9). In this region, calc-alkaline volcanics become younger from west to east, and show a geochemical polarity that resulted from a west dipping subduction (Savelli, 2000). The existence of magmas younger than ~7 Ma eastward of the Sardinia-Corsica axis suggests commencement of eastward slab rollback at 10-7 Ma. It can also be recognised that arc migration first occurred in the northern Tyrrhenian, and only later (~5 Ma) in the southern Tyrrhenian. The youngest calc-alkaline volcanics, found in the Aeolian Islands, mark the present location of the volcanic arc associated with subduction processes beneath Calabria.

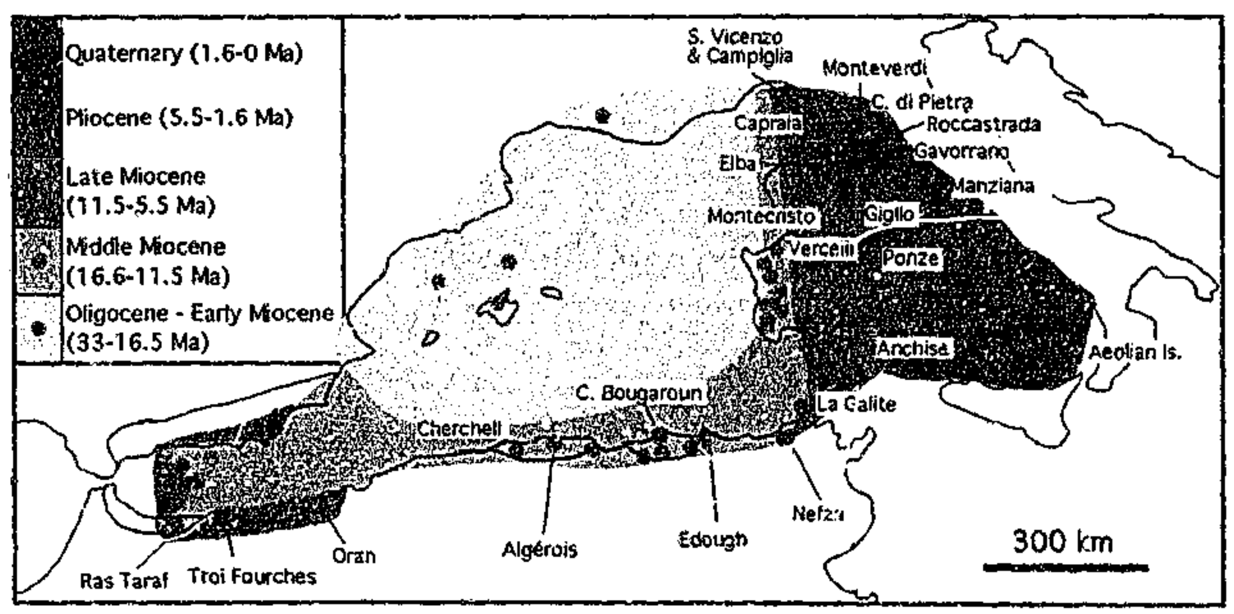


Figure 5.9. The distribution of volcanic rocks in the western Mediterranean (see also Table 5.3).

5.4.5. Seismicity

Deep structures in the Mediterranean region have been studied from the analysis of seismic data, and particularly from seismic tomography images (Wortel and Spakman, 1992; 2000; Spakman et al., 1993; Carminati et al., 1998). Seismic tomography is based on the contrast between seismic velocities produced by the existence of relatively cold subducting lithosphere within the surrounding hot asthenosphere. It thus provides an insight to the three-dimensional structures of subduction systems long after the generation of earthquakes has ceased (Wortel and Spakman, 1992).

Results from seismic tomography strongly support the existence of subduction systems throughout the western Mediterranean. A well-defined Benioff Zone is found below the Calabrian Arc, where active subduction of the Ionian plate is taking place. It is associated with deep (>500 km) earthquakes (Anderson and Jackson, 1987b), and is clearly recognised in tomographic images that show a northwest dipping lithospheric slab subducting beneath Calabria, and flattened in the upper/lower mantle boundary (Lucente et al., 1999). In other localities, such as the Apennines and the Alboran Sea, the seismic tomography images show subducting slabs detached from the lithosphere at the surface (Wortel and Spakman, 2000). Beneath the Apennines, a detached lithospheric slab is recognised at 150-670 km depth. It dips towards the southwest, and indicates the remnant subduction system that existed during the opening of the Tyrrhenian Sea (Lucente et al., 1999). Beneath the Alboran Sea, a detached lithospheric slab at depths of more than 600 km has been deduced from deep seismicity and tomographic images (Buforn et al., 1991a; Seber et al., 1996; Calvert et al., 2000). However, the tectonic significance of the detached slab has been differently interpreted as evidence for east dipping subduction (Lonergan and White, 1997), lithospheric delamination (Seber et al., 1996; Calvert et al., 2000), and a convective removal of thickened lithosphere (Platt and Vissers, 1989). In the North African margin, tomographic anomalies have not been clearly recognised due to a relatively poor spatial resolution (Spakman et al., 1993). However, it has been suggested that existing anomalies below Algeria indicate remnants lithospheric mantle ruptured and segmented from the Tyrrhenian and the Alboran subduction systems (Carminati et al., 1998).

5.5. Reconstruction

A reconstruction movie of the western Mediterranean since the Oligocene is found in the attached CD and can be viewed by using the *QuickTime* application (movie 'W_Med.mov' in Appendix 2). The main tectonic characteristics shown in the movie are discussed below.

5.5.1. Tectonic setting of south Europe in the Oligocene

During the Oligocene, the area between the Iberian peninsula and southern France consisted of several terranes, which are now located hundreds of kilometres away. Among these are the internal zones of the Betic-Rif Cordillera, the Balearic Islands, the Kabylies, Corsica, Sardinia, and Calabria (Ricou et al., 1986; Lonergan and White, 1997) (Figure 5.10). Most of these terranes consist of a Hercynian basement and a Mesozoic cover, which largely underwent deformation and metamorphism during Alpine orogenesis. The origin of these terranes is not entirely clear, but they were possibly attached to the Iberian plate before being incorporated in the Alpine orogeny (Stampfli et al., 1998). Since the Middle Miocene, no rotations occurred in Corsica, Sardinia and the Balearic Islands, and their palaeo-positions can be inferred by applying opposite rotations to those obtained from palaeomagnetic studies. Prior to the opening of the western Mediterranean basins, Calabria was located adjacent to Sardinia (Alvarez et al., 1974; Dewey et al., 1989; Minzoni, 1991). An alternative hypothesis is that during the Oligocene, Sardinia and Calabria overlapped each other, forming the upper (Sardinia) and the lower (Calabria) units of a metamorphic core complex. This hypothesis, however, requires further research. The largest uncertainty in the Oligocene reconstruction is the position of the Internal Zone of the Rif-Betic cordillera, which is here placed southeast to the Balearic Islands after Lonergan and White (1997). This configuration forms a continuous orogenic belt during the Oligocene, from the Rif-Betic to Calabria, Corsica and the western Alps.

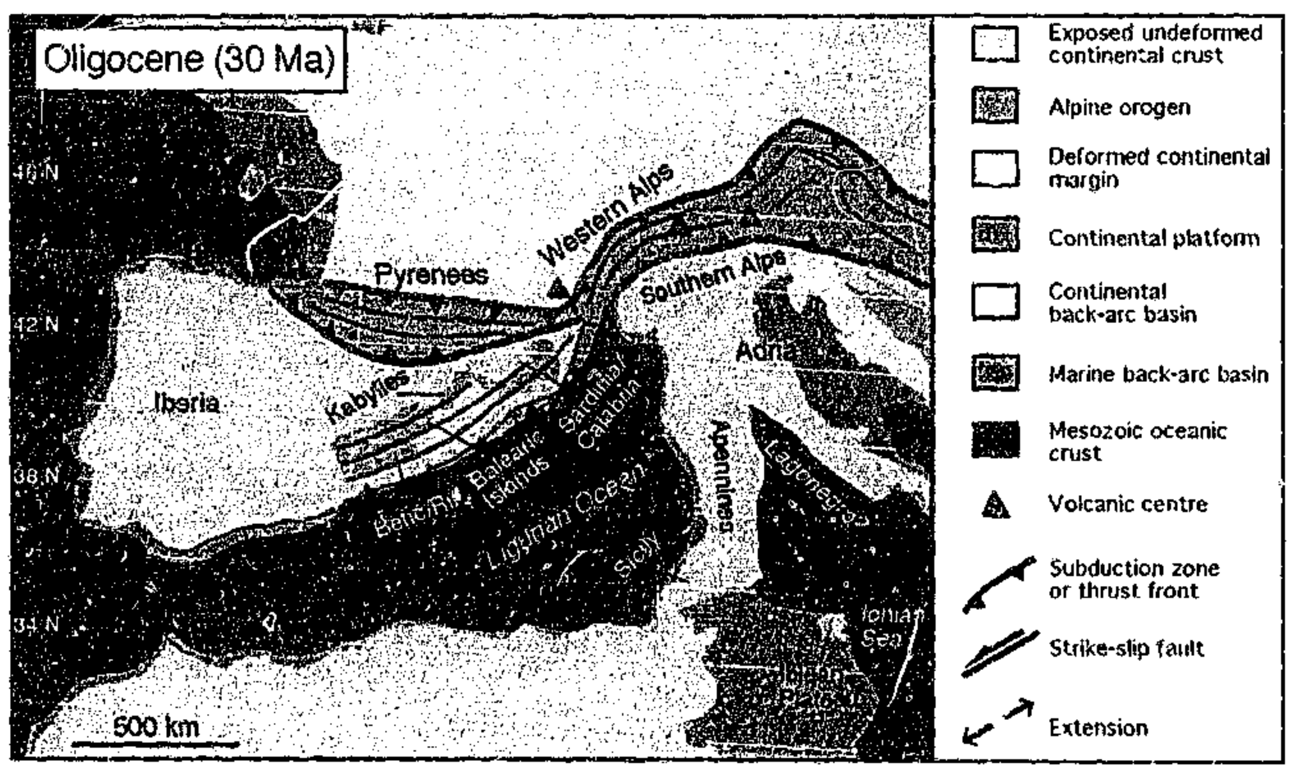


Figure 5.10. Oligocene reconstruction (30 Ma).

The tectonic setting in the Late Oligocene was characterised by a switch in the vergence of subduction systems and by the occurrence of widespread extension in the Alps (termed 'the Oligocene Lull' by Laubscher (1983)) and in the western Mediterranean region. In the Early Oligocene, the Alpine orogen underwent a major orogenic episode, indicated by ~35 Ma ages of high-pressure and ultra-high-pressure rocks exposed in the Internal Crystalline Massifs of the western Alps (Gebauer, 1996; Gebauer et al., 1997; Rubatto and Gebauer, 1999). The present structural configuration of the Alpine sutures in the western Alps and in northeast Corsica suggests that, prior to continental

collision, the area had been controlled by a southeast-dipping subduction system (Figure 5.10). In the Late Oligocene, however, the polarity of the subduction system changed, and a new northwest-dipping subduction system developed in the southern margin of west Europe, producing calc-alkaline volcanism in Provence and Sardinia (Figure 5.10).

The initiation of a new northwest-dipping subduction system was possibly triggered by continental collision in the Alps at 35 Ma. This collision could have blocked the existing subduction system by the arrival of thick crustal material at the subduction zone. Thus, a new subduction system developed in a more southerly location, where relatively old (Jurassic) oceanic lithosphere was found. At 30 Ma the subducting oceanic lithosphere was relatively old (>110 Ma) and cold enough to create a gravitational instability, which would have caused rollback of the subduction hinge towards the SE. In addition to slab rollback, the motion of Africa relative to Europe has been considerably slower since 30 Ma, and particularly since 25-20 Ma (Jolivet and Faccenna, 2000; Rosenbaum et al., 2002a). Thus, with the absence of sufficient convergence to support subduction rollback, extension commenced in the overriding plate, forming the foundations of the western Mediterranean basins.

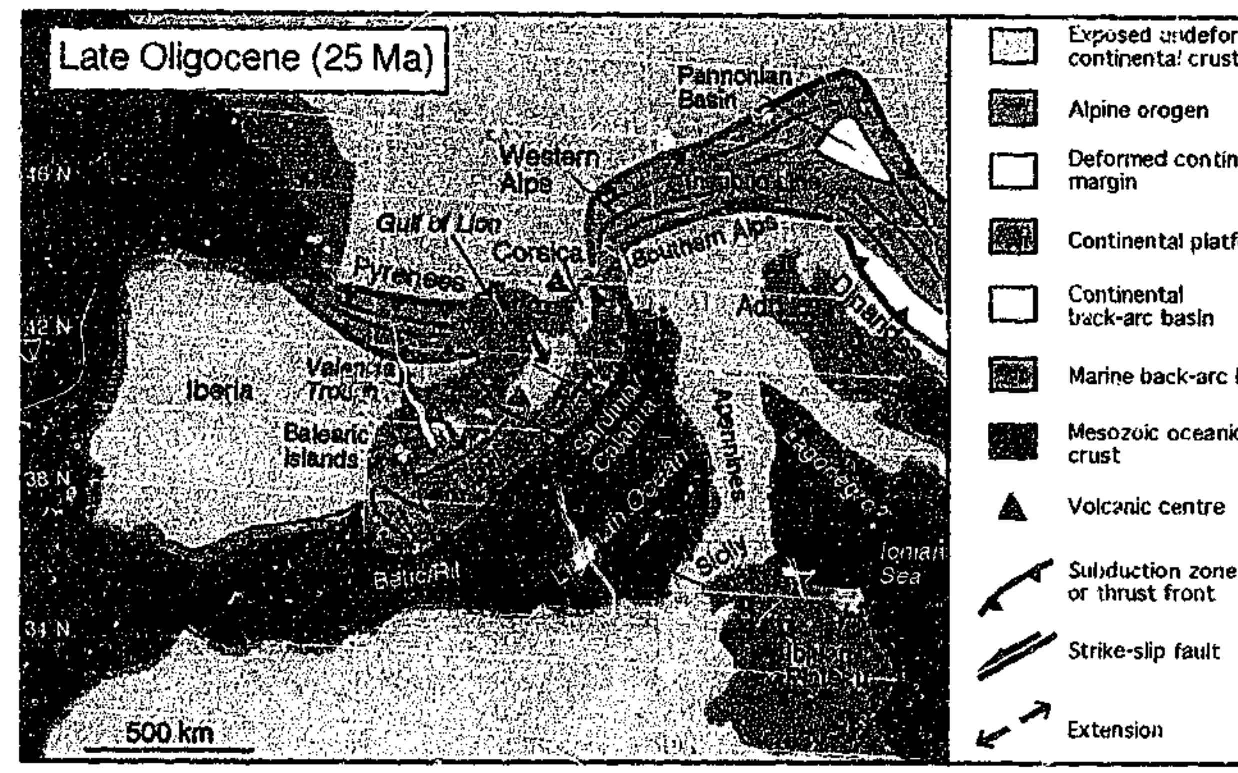


Figure 5.11. Late Oligocene reconstruction (25 Ma).

5.5.2. Valencia Trough, Gulf of Lion and the Ligurian Sea

Earlier rifting is inferred from syn-rift Late Oligocene sediments deposited on Early Oligocene grabens and half grabens in the margins of Valencia Trough, the Gulf of Lion and the Ligurian Sea (Cherchi and Montadert, 1982; Burrus, 1989; Bartrina et al., 1992). Rifting probably commenced in the early Late Oligocene (~30 Ma) in the Gulf of Lion (Séranne, 1999), and in the latest Oligocene

(~25 Ma) in Valencia Trough (Roca et al., 1999). A right lateral strike-slip fault (North Balearic Transfer Zone) separated the Valencia Trough from the Gulf of Lion (Séranne, 1999) (Figure 5.11). Structural observations from the extended margins suggest that horizontal extension was partitioned in different crustal levels, forming rift valleys in the upper crust (e.g., in Sardinia; Cherchi and Montadert, 1982), and low angle extensional detachments in deeper crustal levels (e.g., Corsica and Calabria; Jolivet et al., 1990; Rosseti et al., 2001). Ductile extensional deformation in Corsica and Calabria has been dated at 32-25 Ma (Brunet et al., 2000; Rosseti et al., 2001), that is, before subduction rollback commenced.

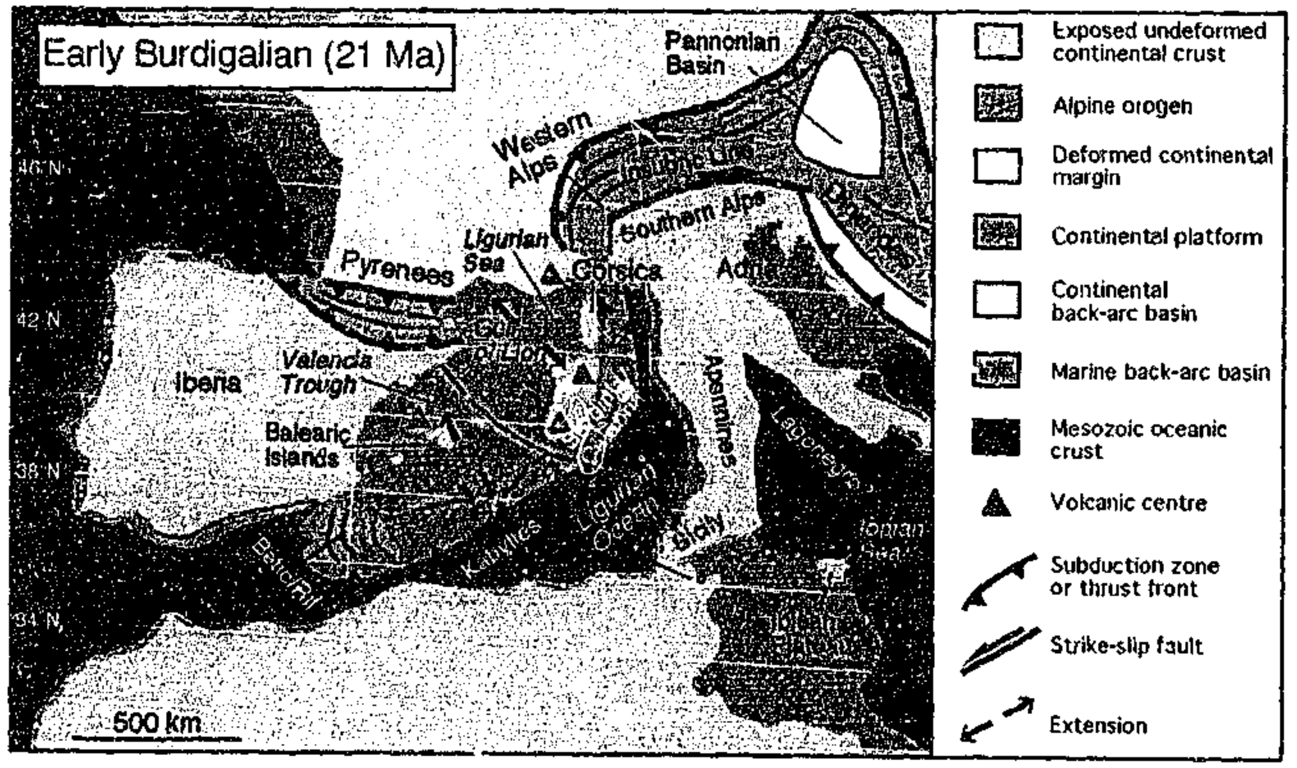


Figure 5.12. Early Burdigalian reconstruction (21 Ma).

As a result of subduction rollback, extension in the Early Miocene led to the breakup and drifting of continental fragments formerly attached to southern France and Iberia. Thus, during the opening of the Ligurian Sea and the Valencia Trough, the Balearic Islands, Corsica, Sardinia and Calabria were subjected to block rotations. Extension in Valencia Trough ceased in the early Burdigalian (21-20 Ma) before it was sufficient to form oceanic crust (Bartrina et al., 1992; Watts and Torné, 1992). However, ongoing southward rollback of the subduction hinge led to the formation of a new rift system between the Balearic Islands and the Kabylies blocks, and further extension resulted in the formation of the Provençal Basin (Séranne, 1999) (Figure 5.12). In the Gulf of Lion, tectonic activity ceased in Aquitanian/early Burdigalian (20-18 Ma) (Cherchi and Montadert, 1982; Burrus, 1989), possibly due to the collision of Corsica, Sardinia and Calabria with the Apennines (Figure 5.13). Following collision, Apennine units arrived at the subduction system and impeded rollback, which in turn, led to the cessation of back-arc extension in the Ligurian Sea.

5.5.3. North Africa

During the Early-Middle Miocene, intense tectonic activity took place in North Africa due to the opening of the Provençal, Algerian and Alboran basins, and the subsequent emplacement of the Kabylies and the Internal Rif onto the African margin. Extension in the Provençal Basin commenced after rifting in Valencia Trough had failed. In early Burdigalian (~21 Ma), continental

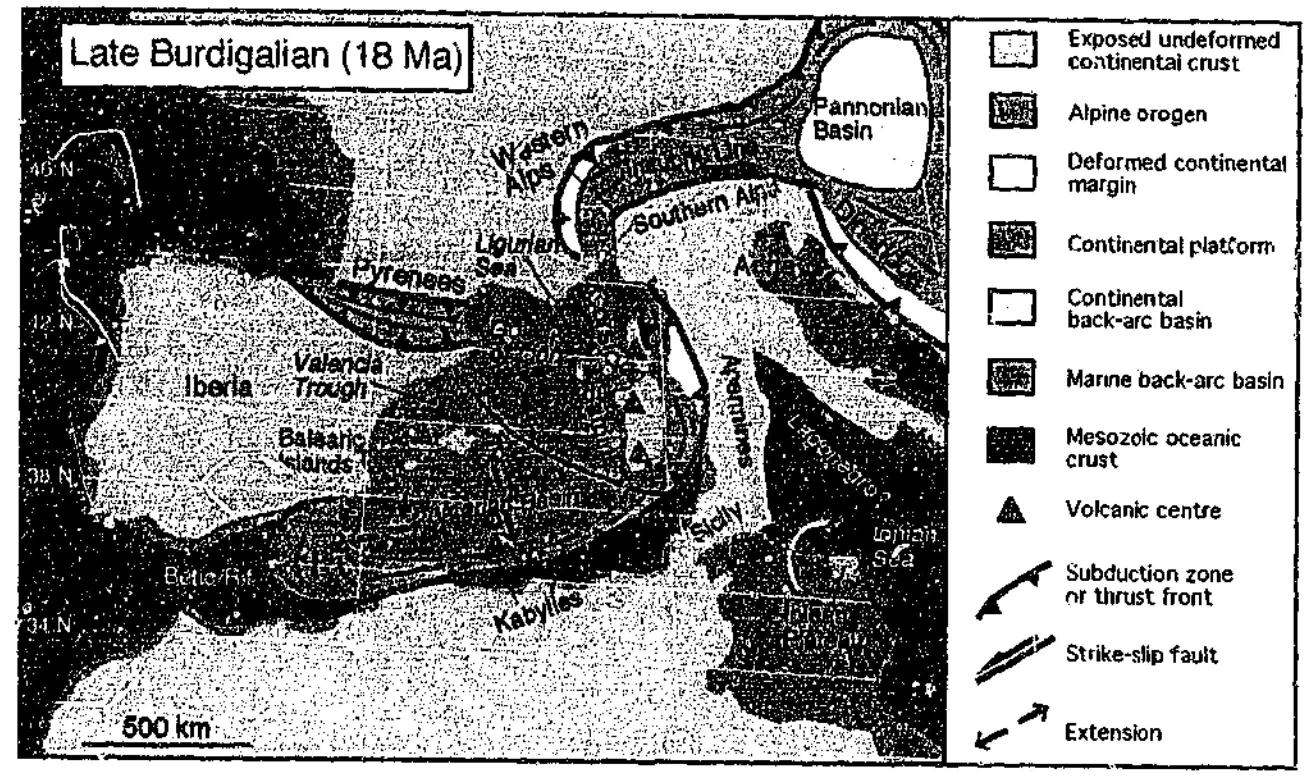


Figure 5.13. Late Burdigalian reconstruction (18 Ma).

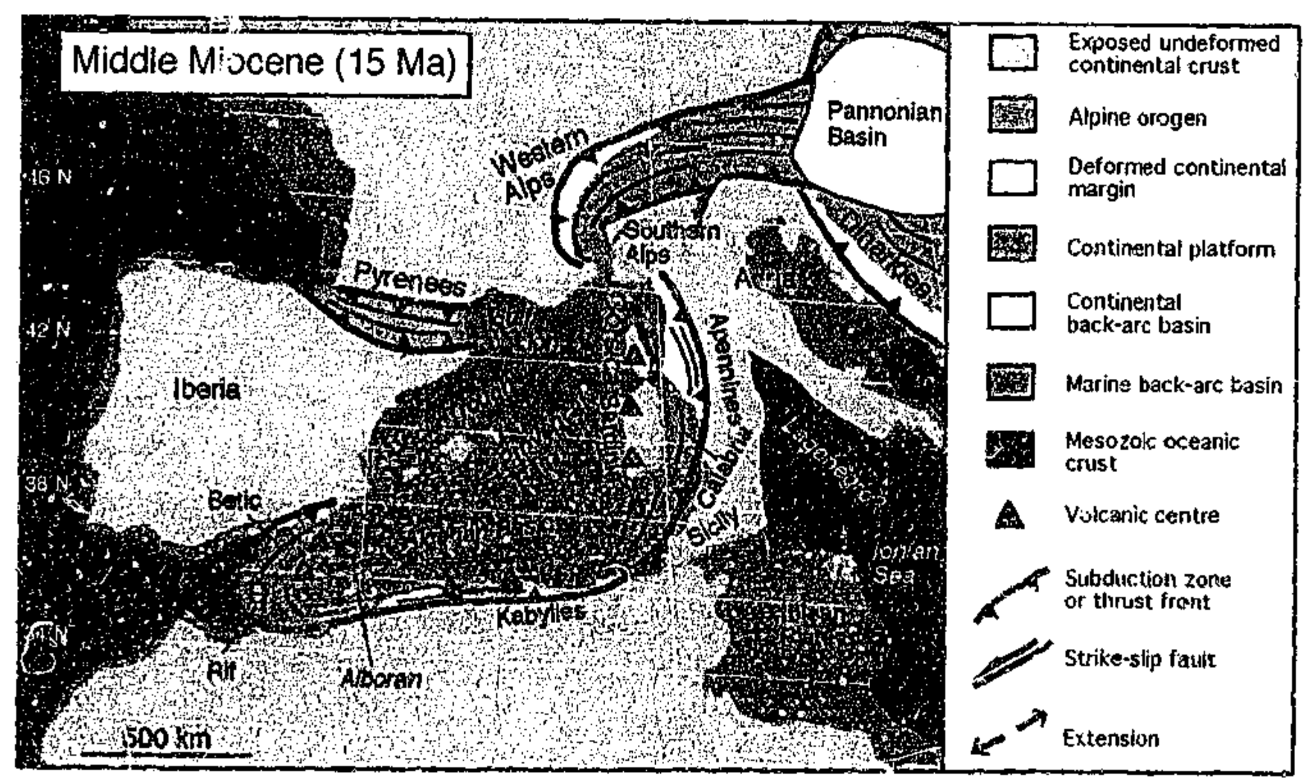


Figure 5.14. Middle Miocene reconstruction (15 Ma).

breakup occurred between the Balearic Islands and the Kabylies blocks, and a new basin developed. Extension was probably governed by a rapid southward rollback and subsequently led to sea floor spreading and formation of new oceanic crust in the Algerian-Provençal Basin (Rehault et al., 1985) (Figure 5.12 and Figure 5.13).

Evidence for extensional fabrics associated with the opening of the Algerian-Provençal Basin is found in the Kabylies metamorphic core complexes, which are now accreted to the African margin. These complexes are characterised by the occurrence of low-angle extensional detachments, which juxtaposed upper crustal rocks on top of high-grade metamorphic rocks (Caby and Hammor, 1992; Tricart et al., 1994; Saadallah and Caby, 1996). The occurrence of core complexes implies that a considerable amount of crustal thinning was made possible by non-coaxial shearing along extensional detachments. Based on $^{39}Ar/^{40}Ar$ dating, it has been suggested that extensional deformation occurred at 25-16 Ma (Monié et al., 1984; 1988; 1992).

The Kabylies blocks drifted southward in response to the southward rollback of the subduction zone until they collided and accreted to the African margin (Cohen, 1980; Tricart et al., 1994) (Figure 5.14). Collision occurred when the Mesozoic oceanic lithosphere, which had previously separated the Kabylies from the African margin, was totally consumed by subduction. The collision occurred between 18-15 Ma, based on the cessation of extensional tectonics in the Algerian-Provençal Basin and the Kabylies core complexes, and the commencement of thrusting in the External Maghrebides (Frizon de Lamotte et al., 2000). The accretion of the Kabylies blocks and the final consumption of old oceanic lithosphere in this region permanently terminated southward subduction rollback. South-directed thrust systems propagated southward after accretion (Frizon de Lamotte et al., 2000), but subduction processes were impeded and eventually halted with the presence of the relatively buoyant continental crust at the subduction zone. The Middle Miocene cessation of subduction processes in North Africa resulted in the segmentation of the western Mediterranean subduction system, with an eastward dipping subduction in the Alboran region, and a westward dipping subduction in the Tyrrhenian region (Lonergan and White, 1997) (Figure 5.14).

5.5.4. The Alboran Domain

The tectonic evolution of the Alboran Sea is a matter of controversy, and several different models have been proposed. This reconstruction follows the ideas of Royden (1993b) and Lonergan and White (1997), who suggested a slab rollback model for the opening of the Alboran Sea. Alternatively, it has been suggested that extension in the Alboran Sea developed as a result of an extensional collapse of a thickened continental crust and its underlying lithospheric mantle (e.g. Platt and Vissers, 1989; Houseman, 1996). These models are not supported by field observations from the Rif-Betic cordillera, which imply episodic alterations from crustal shortening to extension (Azañón et al., 1997; Balanyá et al., 1997; Martínez-Martínez and Azañón, 2002). Neither are such models supported by the block rotations inferred from palaeomagnetic data (see section 5.4.3).

According to the present model, the origin of the Internal Zone of the Rif-Betic cordillera is similar to the origin of Corsica, Sardinia, Calabria and the Kabylies (Figure 5.10). Thus, the Rif-Betic in its Oligocene position formed a contiguous orogenic belt together with the Kabylies, Calabria, Corsica and the western Alps that underwent high-pressure metamorphism during Alpine orogeny (Figure 5.10). As mentioned before, these terranes (excluding the western Alps) were part of an overriding continental slab above a northwest dipping subduction zone, which started to retreat oceanward during the Oligocene. Once rollback of the subduction hinge commenced, rocks of the Rif-Betic Internal Zone were subjected to an extensional regime leading to the formation of metamorphic core complexes and exhumation of high-pressure metamorphic rocks below extensional detachments (Platt et al., 1983; Azañón et al., 1997; Balanyá et al., 1997).

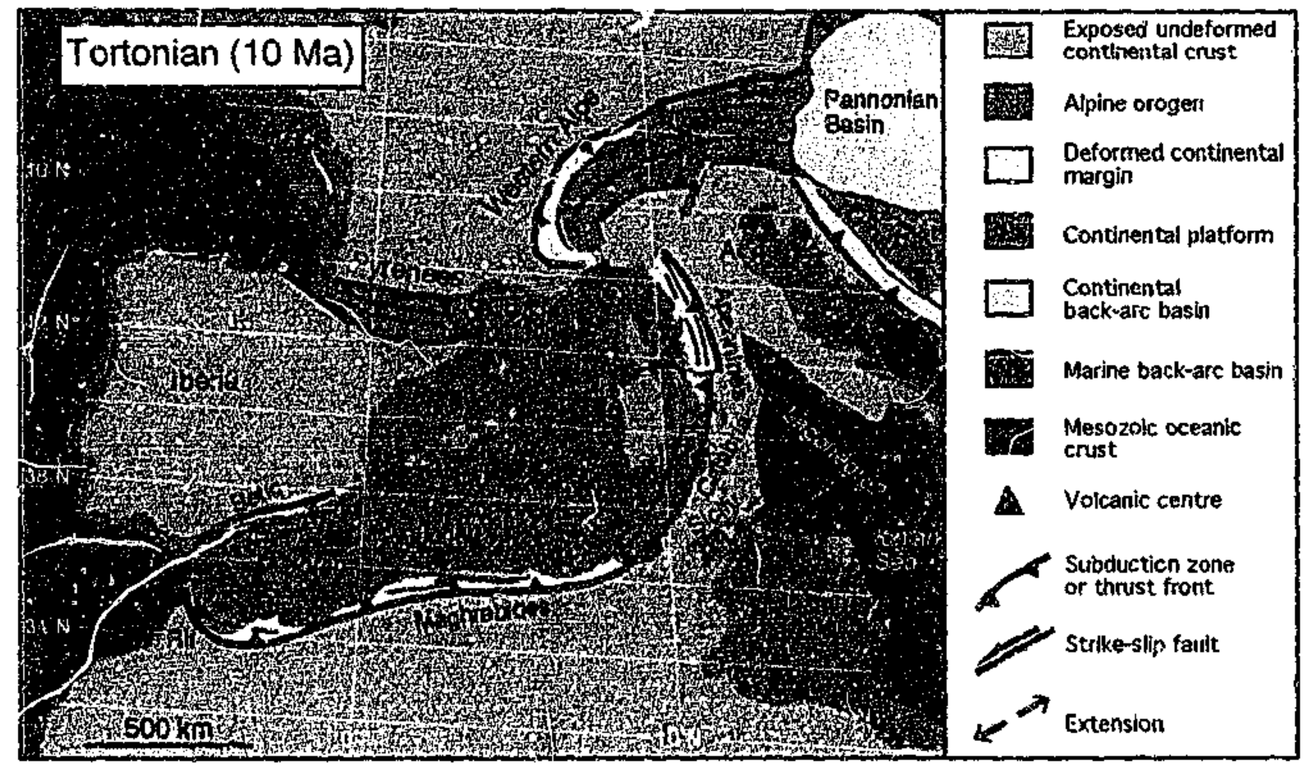


Figure 5.15. Tortonian reconstruction (10 Ma).

Westward rollback of the subduction hinge in the westernmost Mediterranean was probably triggered by an original curvature of the subduction zone in its western terminus. Rollback, therefore, led to the southwestward migration of the subduction hinge accompanied by southwestward drifting of the extended continental fragments of the overriding plate. Southerly migration, however, stopped when the subduction zone reached the passive margin of Africa (see previous section). Rollback continued only in areas where the existence of oceanic lithosphere still permitted oceanward retreat of the subducting lithosphere (Figure 5.14). Thus, in the Middle Miocene (16-15 Ma), the western east-dipping segment of the subduction zone rolled back in the direction of the oceanic lithosphere.

The formation of the Alboran Sea occurred during the westward migration of the subduction hinge. Rapid rollback was compensated by wholesale extension in the overriding continental crust, which was thinned to ~15 km between 23-10 Ma (Lonergan and White, 1997). Contemporaneously, fragments of continental crust were thrust onto the passive margin of Africa and Iberia (i.e. the External Rif-Betic Zones), forming rotation patterns consistent with oblique thrusting derived by the westward rollback of the subduction zone. Final accretion of the Rif-Betic Cordillera occurred at ~10 Ma, when the subduction zone rolled back as far as Gibraltar (Figure 5.15). Subduction rollback then ceased, together with the cessation of back-arc extension in the Alboran Sea (Lonergan and White, 1997).

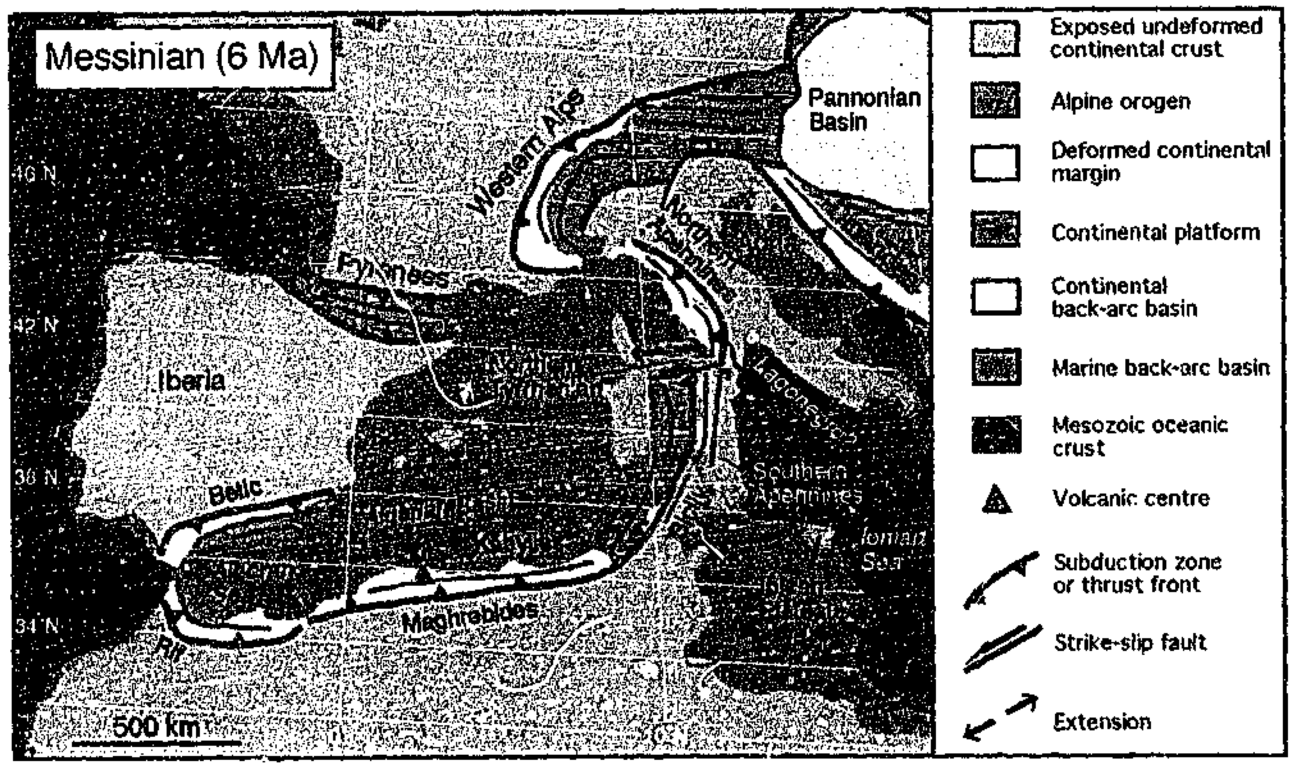


Figure 5.16. Messinian reconstruction (6 Ma).

5.5.5. Tyrrhenian Sea

The Tyrrhenian Sea is the youngest basin in the western Mediterranean, forming since the Tortonian (~9 Ma). It was opened, according to this reconstruction, as a result of a southeastward rollback of subduction systems near the Adriatic margins (Malinverno and Ryan, 1986).

It is suggested here that the collision of Corsica and Sardinia with the Apennines at ~18 Ma led to a relative quiescence in back-arc extension between 18-10 Ma (Figure 5.13 - Figure 5.15). During this period, continental crust of Apennine units incorporated in the subduction zone, and impeded further eastward subduction rollback. Thus, considerable crustal shortening accompanied by thrust systems that propagated eastward occurred in the Apennines.

The opening of the Tyrrhenian Sea occurred in two stages: an early (9-5 Ma) opening of the northern Tyrrhenian Sea (Figure 5.16), and a late (5-0 Ma) opening of the southern Tyrrhenian Sea (Mantovani et al., 1996) (Figure 5.17). It was accompanied by coeval crustal shortening in the Apennines (Malinverno and Ryan, 1986) and counterclockwise rotations of nappe stacks. The reason

for the opening of the northern Tyrrhenian is not entirely clear. It may be associated with subduction of oceanic crust located between the Apennine belt and Adria, which promoted lithospheric gravitational instability during Late Miocene and further eastward subduction rollback. It is possible that deep-sea sediments of the Lagonegro and Molise formations are remnants of these intra-Adriatic basins (Sengör, 1993). At this stage, the subduction zone was oriented ~N-S, that is, roughly parallel to the direction of convergence. Therefore, the rate of convergence at the trench was very low, and consumption of oceanic lithosphere was mostly driven by the negative buoyancy of the subducting slab (Faccenna et al., 2001b).

During the latest Miocene or the Early Pliocene (5 Ma) extension ceased in the northern Tyrrhenian Sea and migrated southward to the southern Tyrrhenian Sea (Figure 5.17). This stage was characterised by considerable extension that culminated during the Pliocene-Pleistocene, when new oceanic crust formed. Contemporaneously, crustal shortening occurred in the Southern Apennines and in Sicily, accompanied by counterclockwise block rotations in the former and clockwise rotations in the latter. These processes have been controlled by rapid rollback of the oceanic Ionian lithosphere beneath the Calabrian arc.

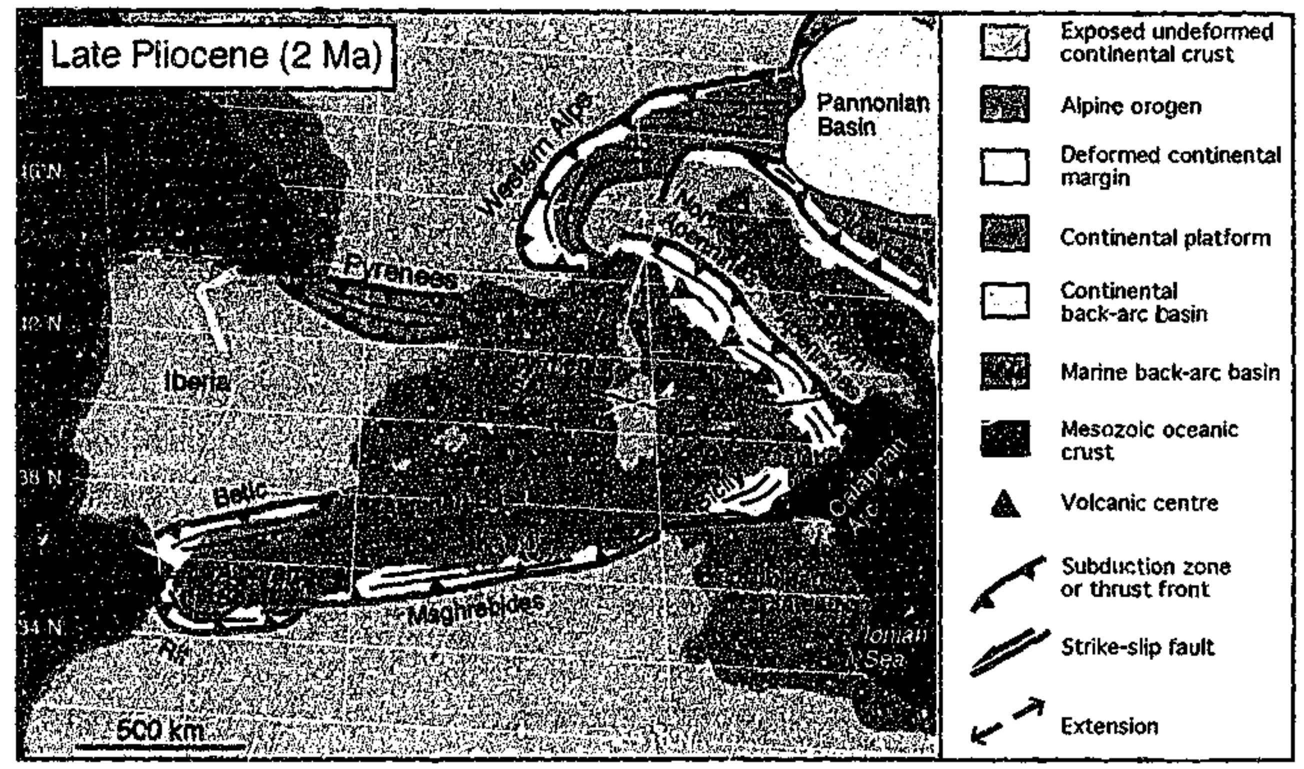


Figure 5.17. Late Pliocene reconstruction (2 Ma).

5.6. Discussion

The reconstruction of the western Mediterranean since the Oligocene emphasises the role of subduction rollback in convergent plate margins. It shows that a widespread extension took place in the convergent interface separating Africa and Europe, forming marine back-arc basins and new spreading centres. Subduction rollback was probably controlled by the gravitational instability produced during subduction of cold and dense oceanic lithosphere. Back arc extension was likely to occur when the velocity of the slab retreat overcame the absolute motion of the overriding plate (Molnar and Atwater, 1978; Dewey, 1980; Royden, 1993a; Lonergan and White, 1997). In the evolution of the western Mediterranean, these processes have played a fundamental role since the Oligocene. Subduction rollback was made possible by the consumption of Mesozoic oceanic lithosphere at the subduction zone. This oceanic lithosphere was probably old and cold enough to be gravitationally unstable relative to the surrounding asthenosphere. The subduction zone has therefore progressively retreated from its Oligocene position near the southern margin of Europe to its final configuration in the Calabrian arc, North Africa and the Alboran arc. In the western Mediterranean, subduction rollback occurred during a period of relatively slow convergence between Africa and Europe (Jolivet and Faccenna, 2000; Rosenbaum et al., 2002a). As convergence alone could not compensate the vacant area formed by subduction retreat, back-arc extension occurred in the overriding plate. Thus, a period of relatively slow convergence was actually characterised by large-scale horizontal motions of smaller microplates and allochthonous terranes.

The dispersion of continental terranes, which drifted and rotated during subduction rollback, is clearly seen in the reconstruction. In the Alpine orogen, this process led to the fragmentation of a continuous belt into continental terranes, which in turn, collided with the passive margins of the surrounding continents. I stress that this mechanism may have profound tectonic implications on the way orogens work. Orogens may be subjected to switches from crustal shortening and extension, controlled by the processes of subduction rollback, rifting in the back-arc region and the subsequent accretion of allochthonous terranes into adjacent passive margins. Thus orogenesis cannot be oversimplified to subduction followed by collision of two continental plates, but includes accretion of numerous continental terranes. Following collisional events, reorganisation of the plate boundaries occurs, associated with termination, jumping or segmentation of subduction zones. A similar style of orogenesis has been proposed by Nur and Ben Avraham (1982) based on numerous examples of allochthonous terranes throughout the Circum-Pacific (excluding the Andes) and the Alpine Himalayan belts. These authors have suggested that continental slivers and microcontinents could actually migrate great distances before colliding with the continents.

In summary, the style of tectonism suggests that fragments of continental crust were subjected to large amounts of horizontal transportation, block rotations on vertical axes, and episodic alterations from crustal shortening to extension.

5.7. Concluding remarks

Extension in the western Mediterranean region commenced at 32-30 Ma and was primarily controlled by subduction rollback. The rapid rollback of the subduction hinge was accompanied by a relatively slow convergence between Africa and Europe. rapid rollback velocities were not supported by convergence, and extension occurred on the overriding plate.

During back-arc extension, marine basins progressively formed from north to south, floored either by thinned continental crust or new oceanic crust. The earliest basins began to form in Late Oligocene in the Gulf of Lion, the Ligurian Sea and Valencia Trough. In Early Miocene, back-arc extension propagated to Provençal, Algerian and Alboran basins, and in the Late Miocene, extension in the Tyrrhenian Sea commenced.

Rifting led to breakup of continental terranes, which drifted and rotated as long as the subduction zone continued to rollback. Subduction rollback temporally or permanently ceased when continental crust arrived at the subduction zone, impeding subduction processes. The continental terranes have then been accreted to the continents and considerable crustal shortening occurred.

CHAPTER 6

THE EVOLUTION OF THE TYRRHENIAN SEA AND THE APENNINE-MAGHREBIDE BELT: INSIGHTS FROM DETAILED SPATIO-TEMPORAL ANALYSIS

Michael said 'lithosphere'. He said 'sandstone', 'chalk bed'. He said 'precambrian', 'cambrian', 'metamorphic rocks', 'igneous rocks', 'tectonics'. ... Those words relate to facts which have meaning for me, for me alone, like a message transmitted in code. Beneath the surface of the earth opposed endogenic and exogenic forces are perpetually at work. The thin sedimentary rocks are in a continuous process of disintegration under the force of pressure. The lithosphere is a crust of hard rocks. Beneath the crust of hard rocks rages the blazing nucleus, the siderosphere.

Amos Oz, My Michael (1968)
(translated from Hebrew by Nicholas de Lange)

FOREWORD: CHAPTER 6

This chapter discusses the Neogene to Quaternary evolution of the Tyrrhenian Sea and the Apennine-Maghrebide belt. The ideas presented in this chapter have benefited from discussions with Gordon Lister, Celal Sengor, Wouter Schellart, Mike Raetz, Fabio Speranza and Carlo Savelli. I also thank Kevin Burke, Darrel Cowan, Wouter Schellart, Mike Raetz, Peter Betts and Ivo Vos for providing comments on the manuscript. A slightly modified version of this achapter has been submitted to *Tectonics* and is currently under review[#].

[#] Rosenbaum, G. and Lister, G.S. The evolution of the Tyrrhenian Sea and the Apennine-Maghrebide belt: insights from detailed spatio-temporal analysis. *Tectonics*, in review.

6. THE EVOLUTION OF THE TYRRHENIAN SEA AND THE APENNINE-MAGHREBIDE BELT: INSIGHTS FROM DETAILED SPATIO-TEMPORAL ANALYSIS

Abstract

In this chapter I present spatio-temporal analysis and a tectonic reconstruction of the evolution of the Tyrrhenian Sea and the Apennine-Maghrebide belt. The earliest rifting in the northern Tyrrhenian Sea took place during the Oligocene-Early Miocene (25-16 Ma). This rifting stage ceased at ca. 16 Ma and remained inactive until the Tortonian (~10 Ma). The major stages of rifting associated with the opening of the Tyrrhenian Sea commenced at ~10 Ma, and involved two episodes of back-arc extension induced by the rollback of a west-dipping subducting slab. The first period of extension (10-6 Ma) was prominent in the northern Tyrrhenian and in the western part of the southern Tyrrhenian. The second period of extension, which began in the latest Messinian (6-5 Ma), mainly affected the southern Tyrrhenian, and was accompanied by extreme rates of subduction rollback, in excess of 60-100 km/Myr. Reconstruction of subduction rollback, combined with additional palaeomagnetic and palaeogeographic constraints, suggests that the accretion of a large carbonate platform in the central Apennines during the latest Messinian (at 6-5 Ma) impeded subduction processes and initiated a slab tear. This in turn, resulted in the formation of a narrow subducting slab in the Ionian Sea that has consequently undergone higher degrees of subduction rollback and back-arc extension.

6.1. Introduction

The Tyrrhenian Sea is a young (< 10 Ma) extensional basin that formed in the complex convergent boundary between Africa and Europe (Figure 6.1a). It is surrounded by an arcuate orogenic belt consisting of the Apennines in the Italian peninsula, the Calabrian arc in southern Italy and the Maghrebides in Sicily. The arcuate belt has been subjected to orogenic processes associated with subduction of a west-dipping lithospheric slab, simultaneously with back-arc extension in the Tyrrhenian Sea (e.g. Malinverno and Ryan, 1986). Both crustal and lithospheric processes are well documented by geological and geophysical data, making the Tyrrhenian-Apennine region an ideal natural laboratory for studying geodynamic interactions between subduction processes, collisional tectonics and back-arc extension (e.g. Faccenna et al., 1996; 2001a; 2001b; Gvirtzman and Nur, 2001).

Many authors regard the Tyrrhenian Sea as an example of a back-arc basin associated with eastward 'rollback' of a west-dipping subduction zone (e.g. Malinverno and Ryan, 1986; Kastens et al., 1988; Patacca and Scandone, 1989, Gvirtzman and Nur, 1999; 2001; Faccenna et al., 2001a ;2001b). The driving mechanism for slab rollback is thought to result from the negative buoyancy of the slab in comparison with the surrounding asthenosphere (e.g. Elsasser, 1971; Molnar and Atwater, 1978; Dewey, 1980; Garfunkel et al., 1986). The gravitational instability may result in a gradual steepening of the dip of the subducting slab and in an oceanward retreat (i.e. rollback) of the subducting hinge. If the rate of hinge rollback exceeds the rate of convergence, then extension will start on the edge of the overriding plate leading to the opening of a back-arc basin (Dewey, 1980; Nur et al., 1993; Royden, 1993a).

In most natural examples, subduction rollback occurs only when the subducting slab is made up of oceanic lithosphere. This is probably because continental crust is relatively light, and is likely to reduce the negative buoyancy of the slab, which in turn would impede subduction rollback. In the southern Tyrrhenian region, geophysical evidence suggests that the slab beneath the Calabrian arc may consist of an Early Mesozoic oceanic lithosphere (e.g. Catalano et al., 2001) Nevertheless, there is no agreement with regard to the nature of a west-dipping subducting slab beneath the Italian peninsula that supposedly rolled back during the opening of the northern and central Tyrrhenian Sea (e.g. Malinverno and Ryan, 1986).

The purpose of this chapter is to evaluate spatial and temporal constraints that may link the geometry of the subducting slab with the history of subduction rollback and back-arc extension in the region. It is shown that consideration of these constraints provides an insight into the tectonic evolution of the orogen, and enables us to better understand the tectonic responses caused by tearing of a retreating slab.

6.2. Tectonic setting

The Tyrrhenian Sea can be divided into a northern domain and a southern domain. The northern Tyrrhenian is a wedge shaped basin, bounded to the south by the major fault zone of 41° Parallel Line (41PL) and floored by thinned continental crust (Figure 6.1a,b). In the southern Tyrrhenian, in addition to stretched fragments of continental crust, there are also deep basins (Vavilov and Marsili basins) that contain Pliocene to Recent (<5 Ma) MORB-type basalts (Kastens et al., 1988). These basins are considered to represent the locus of sea-floor spreading in a back-arc position with respect to a northwest-dipping subduction zone, presently located beneath Calabria (Figure 6.1a). Further evidence for lithospheric stretching in a back-arc environment is based on high values of heat flow (>150 mW m^{-2}) measured in the Tyrrhenian Sea (Della Vedova et al., 1984) (Figure 6.1c).

The Tyrrhenian-Apennine system has been subjected to simultaneous extensional tectonism in the

Tyrrhenian Sea and crustal shortening in the Apennine-Maghrebide orogen since the Late Miocene (e.g. Malinverno and Ryan, 1986; Lavecchia, 1988) (Figure 6.2). Deformation in the Tyrrhenian-Apennine system followed an earlier deformational event that took place during the Late-Oligocene-Early Miocene and involved the opening of the Ligurian-Provençal Basin (Figure 6.1a) as a result of the rollback of a northwest-dipping subduction zone (e.g. Faccenna et al., 1997; Rollet et al.,

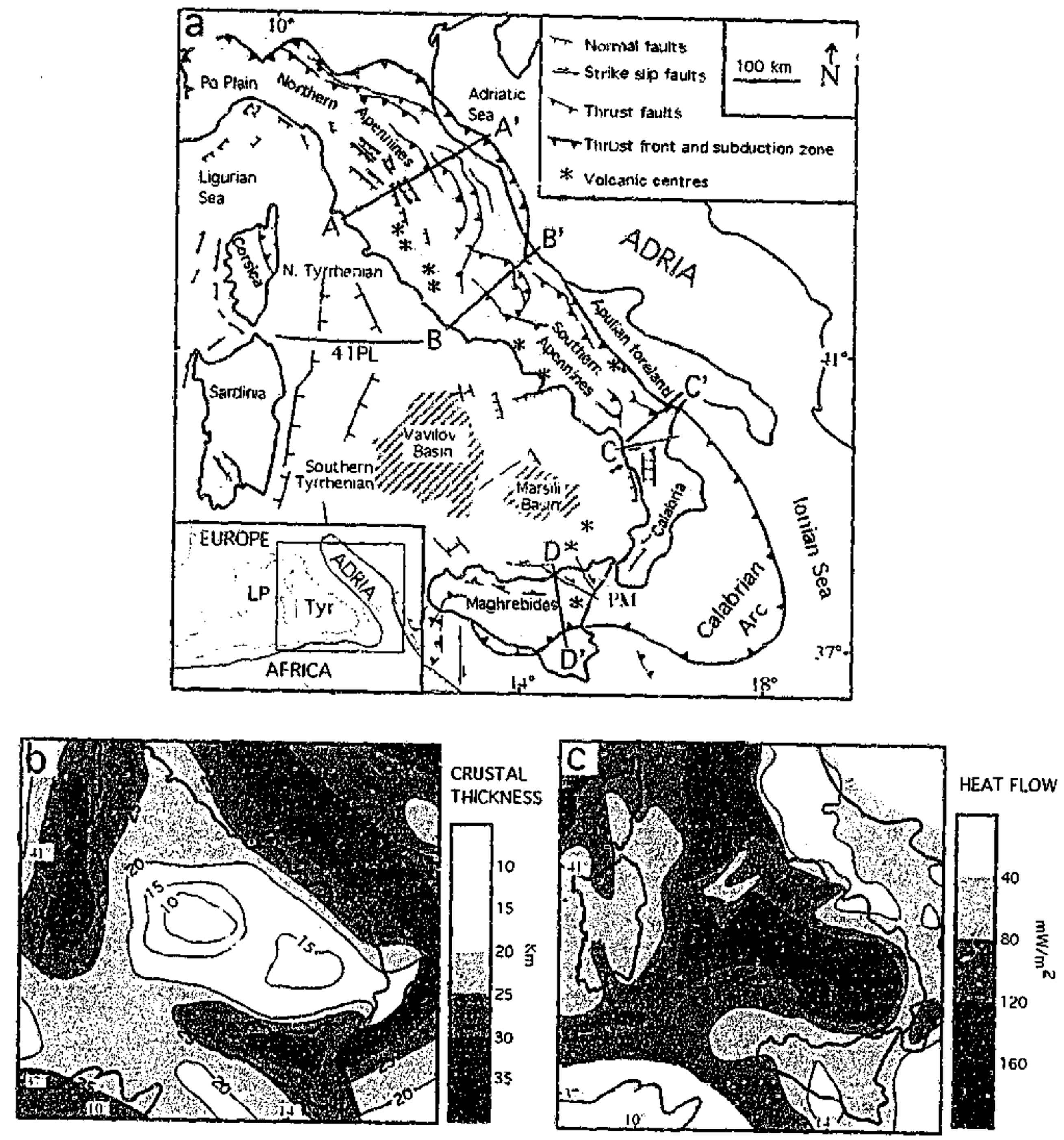


Figure 6.1. Tectonic setting of the Tyrrhenian Sea. (a) Map showing main structural features, modified after Patacca et al. (1993), and locations of cross sections (Figure 6.3). The boundary between the northern domain and the southern domain is marked by the major fault zone of the 41st Parallel Line (41PL). LP, Ligurian-Provençal Basin; PM, Peloritan Mountains; Tyr, Tyrrhenian Sea. (b) Crustal thickness in the Tyrrhenian Sea, after Gvirtzman and Nur (2001). (c) Heat flow in the Tyrrhenian Sea, after Della Vedova et al. (1991).

2002; Rosenbaum et al., 2002b). The opening of the Ligurian-Provençal Basin was accompanied by a counterclockwise rotation of the Corsica-Sardinia microplate, which led to a progressive collision of Corsica-Sardinia with the former western margin of Adria, and gave rise to the formation of a north-south striking orogen in the Apennines (Patacca et al., 1990). Based on palaeomagnetic results, it has been suggested that Corsica and Sardinia stopped rotating at ca. 16 Ma (Speranza et al., 2002, and references therein), which coincides with the time of termination of back-arc extension in the Ligurian-Provençal basin.

Opening of the Tyrrhenian Sea is recognised in the structure of the rifted margins of Corsica and Sardinia, which are associated with Tortonian syn-rift sediments (Mauffret and Contrucci, 1999; Sartori et al., 2001). The extensional front, as well as the front of crustal shortening, migrated eastward through time, leading to the ongoing destruction of internal parts of the orogen by extensional processes. The same process is reflected by the present-day tectonic activity in the Apennines, which is characterised by thrust tectonics in external parts of the orogen and extension in the internal zone (e.g. Mariucci et al., 1999; Montone et al., 1999).

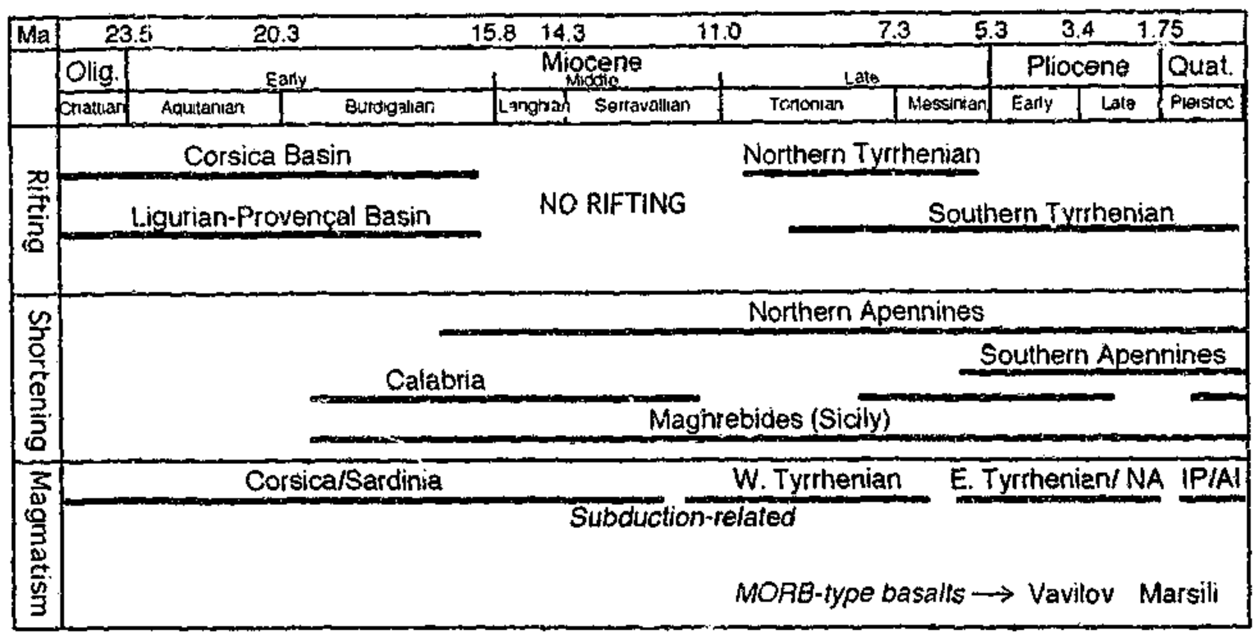


Figure 6.2. Chronology of syn-rift sedimentation, crustal shortening and magmatism in the Tyrrhenian Sea and the Apennine-Maghrebide belt. See text for discussion. Timescale after Remane et al. (2000). AI, Aeolian Islands; IP, Italian Peninsula; NA, Northern Apennines.

6.3. Constraints on crustal deformation

Observations show that deformation in the Tyrrhenian-Apennine region involved simultaneous shortening and extension. During the Neogene and Quaternary, crustal shortening occurred in the front of the Apennine belt, and is currently taking place in the most external parts of the orogen (Figure 6.3). In contrast, more internal parts of the orogen have been subjected to extension associated with the formation of the Tyrrhenian Sea (Figure 6.3 and Figure 6.4).

6.3.1. Migration of the thrust front

The spatio-temporal distribution of crustal shortening is inferred from the ages of thrusting along Apennine-Maghrebide nappe structures and from depositional ages of foreland (molasse) basins (e.g. Boccaletti et al., 1990b). The pattern of crustal shortening in the Tyrrhenian-Apennine system is characterised by migration of thrust faults towards external parts of the orogen (Figure 6.3). In addition, deformation in the southern Tyrrhenian-Apennine system is relative younger compared with the timing of deformation in the northern domain.

Structural evidence from the northern Apennines points to a northeastward migration of crustal shortening (Jolivet et al., 1998) (Figure 6.3). Shortening structures are associated with northeastward-verging folds and thrusts, and deposits in foreland basins become progressively younger towards the external parts of the orogen (Lavecchia, 1988; Boccaletti et al., 1990b). The timing of deformation also becomes younger in the central and southern Apennines. In the southern Apennines, crustal shortening occurred predominantly during the Pliocene and the Pleistocene and was characterised by eastward and southeastward migration of the thrust front (Roure et al., 1991).

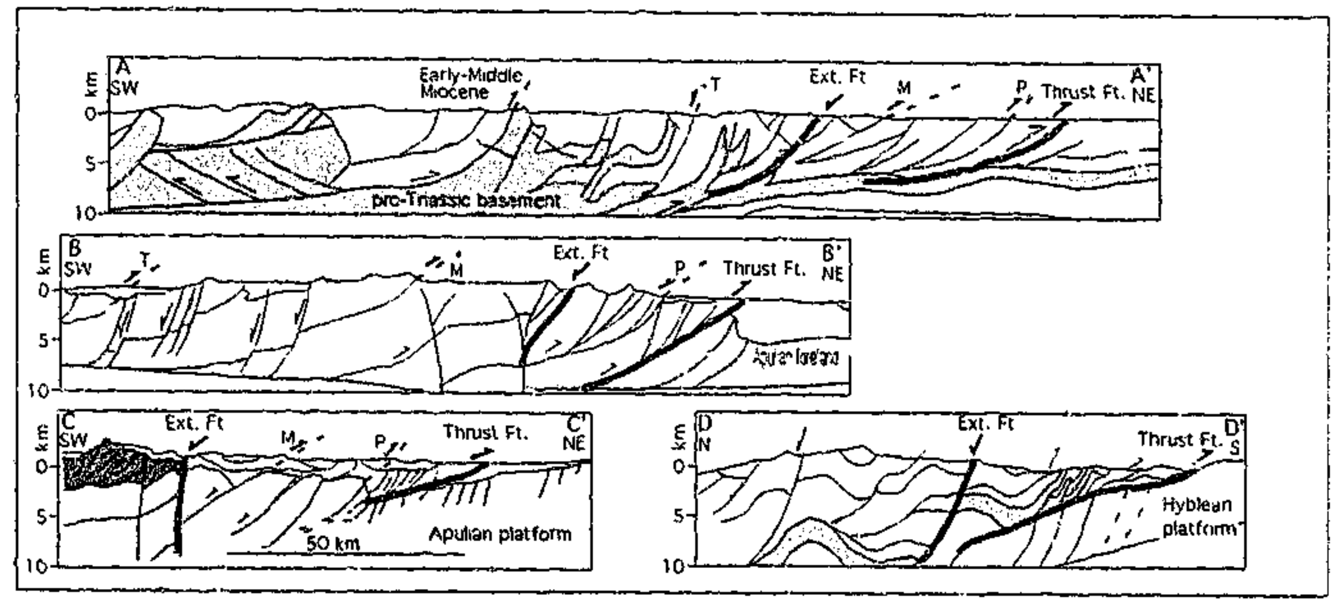


Figure 6.3. Schematic cross sections across the Apennines-Maghrebides (see location map in Figure 6.1a) showing the approximate location of the thrust front through time and the present-day location of the thrust front extensional front (thick lines). Cross sections are modified after Finetti et al. (2001), Mariucci et al. (1999), Cipollari et al. (1999), Cello and Mazzoli (1999) and Roure et al. (1990). M, Messinian (7.5-5.3 Ma); P, Pliocene-Pleistocene (<5.3 Ma); T, Tortonian (11-7.3 Ma).

The pattern of deformation in Calabria is more complex, with evidence for a prolonged tectono-metamorphic history that involved several deformational episodes associated with Alpine orogeny (e.g. Knott and Turco, 1991; van Dijk and Okkes, 1991; Cello et al., 1996; Rosseti et al., 2001). Rock units in Calabria were subjected to high-pressure metamorphism during the Eocene (ca. 44 Ma), and were later exhumed to shallow crustal levels in a period of extensional tectonism (at ca. 30 Ma; Rosseti et al., 2001). In the framework of the Tyrrhenian-Apennine system, the Calabrian block (including the Peloritan Mountains of northeast Sicily; Figure 6.1a) is considered to be an allochthonous terrane that originated approximately 800 km northwest of its present location. It then drifted southeastward since the Oligocene and was accreted into the Adriatic margin during the Neogene (Rosenbaum et al., 2002b, and references therein).

Crustal shortening in Sicily involved progressive underthrusting of rock units during southeastward migration of the thrust front and clockwise block rotations (Oldow et al., 1990; Roure et al., 1990). Crustal shortening in western Sicily began in the Early-Middle Miocene (Oldow et al., 1990), which is somewhat earlier compared with the deformation in the southern Apennines.

6.3.2. Migration of extensional deformation

The spatial distribution of the earliest syn-rift sediments in the Tyrrhenian Sea and its margins suggests that extensional deformation migrated towards the northeast in the northern domain and towards the southeast in the southern domain (Figure 6.4). The earliest syn-rift sediments were deposited during the Late Oligocene-Early Miocene (~25-16 Ma) in the Corsica Basin (Mauffret and Contrucci, 1999) (Figure 6.4). This rifting stage took place before or during the opening of the Ligurian-Provençal Basin (Figure 6.1a), and before Corsica and Sardinia reached their present positions at 16 Ma (Speranza et al., 2002). The Oligocene-Early Miocene extensional deformation in the northern Tyrrhenian may have resulted from subduction rollback of an Oligocene northwest-dipping subduction zone (Rosenbaum et al., 2002b). However, the locus of extension was located in the Ligurian-Provençal rift system, whereas rifting in the northern Tyrrhenian Sea ceased before 16 Ma and was not resumed until ca. 10 Ma (Mauffret and Contrucci, 1999).

The major rifting event associated with the opening of the Tyrrhenian Sea commenced in the Early-Middle Tortonian (~10-9 Ma), based on syn-rift Tortonian sediments in the northern Tyrrhenian and in the western part of the southern Tyrrhenian (Figure 6.4). Tortonian syn-rift sediments are widespread in the northern Tyrrhenian and become progressively younger towards the northeast (Bartole, 1995). In the southern Tyrrhenian, the distribution of Tortonian sediments is restricted to a narrow zone offshore the Sardinian coast (Spadini et al., 1995; Sartori et al., 2001). This spatial distribution may suggest that during the Tortonian, the migration of extensional deformation in the southern domain was slightly slower than the migration in the northern domain. A wedge of syn-rift Late Tortonian sediments has also been reported from the rifted margin of northern Sicily (Pepe et al., 2000), which is in a relatively anomalous location compared with the distribution of Tortonian sediments in the southern Tyrrhenian (Figure 6.4). These Tortonian sediments, as well as older (Middle Miocene) syn-rift sediments found in Calabria (e.g. Argentieri et al., 1998), were probably not deposited in situ, but originated in a more westerly position, from which they have migrated eastward together with Sicilian and Calabrian allochthonous units.

The majority of the southern Tyrrhenian Sea was subjected to extension from the Messinian to the Pleistocene, which culminated in the Pliocene-Pleistocene sea-floor spreading of the Vavilov (5-4 Ma) and Marsili basins (3-2 Ma) (Kastens et al., 1988). The wider distribution of syn-rift sediments during this period suggests that the extensional front migrated faster in the southern Tyrrhenian compared with the northern Tyrrhenian (Figure 6.4). These differential migration rates may have been accommodated by the E-W fault zone around parallel 41° (41PL; Figure 6.4) that was active as a left-lateral strike-slip shear zone during the Messinian and the Pliocene (Spadini and Wezel, 1994; Bruno et al., 2000). Based on the distribution of syn-rift sediments, the rates of migration at 7-2 Ma in the southern Tyrrhenian are estimated as ca. 80 km/Myr.

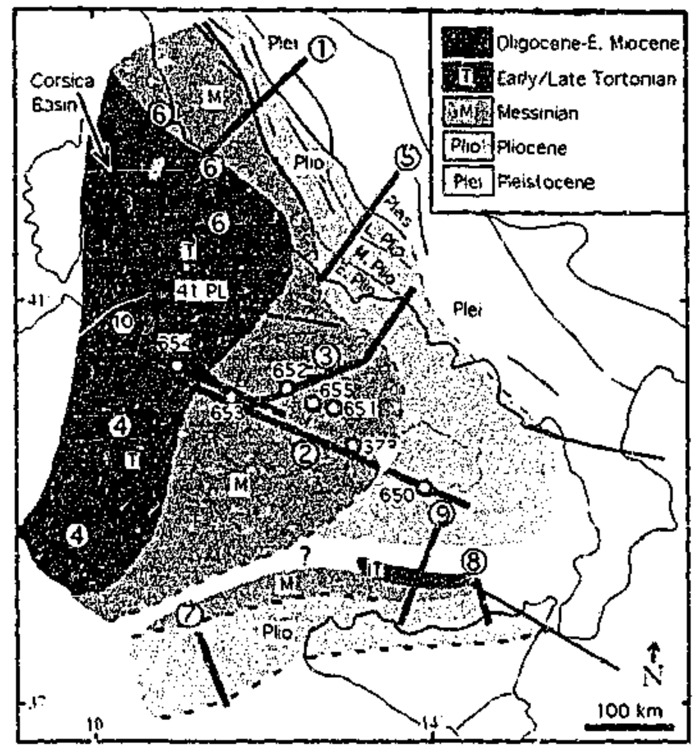


Figure 6.4. Map showing spatial distribution of the earliest syn-rift sediments in the Tyrrhenian Sea and its margins. Thick lines indicate locations of sections from which data have been obtained. Uncertain boundaries are indicated by dashed lines. Syn-rift sediments from the Calabrian block and its margins are not shown because they were probably deposited in a more westerly position. References are: (1) Bartole, (1995); (2) Kastens et al. (1988); Cipollari et al. (1999); (4) Sartori et al. (2001); (5) Cavinato and Celles, (1999); (6) Mauffret and Contrucci (1999); (7) Gamberi and Argnani (1995); (8) Nigro and Sulli (1995); (9) Pepe et al. (2000); (10) Spadini et al. (1995). Locations of ODP sites (650-655) and an OSDP site (373) are also shown. 41PL, 41st Parallel Line.

6.4. Constraints on the role of subduction rollback

The strongest evidence on the role of subduction rollback in the Tyrrhenian-Apennine system is derived from the distribution of subduction-related magmatism, which suggests a pattern of migration of the magmatic arc that is consistent with rollback of the subduction hinge towards the east and the southeast. In the following section, I combine data on magmatic activity with data derived from tomographic cross sections in order to infer the geometry of the subducting slab and to reconstruct the history of rollback at different segments of the Tyrrhenian-Apennine system.

Table 6.1. Magmatic rocks throughout the Tyrrhenian-Apennine region (see Figure 6.5 for location).

No	Location	Rock type	Age (Ma)	References
Mainly calc-alkaline (subduction related)				
1	Sisco, Northeast Corsica	Lamproite	15-14	Civetta et al. (1978)
2	Gallura, west Sardinia	Rhyolitic dacite	15	Savelli (2002)
3	Logudoro, Sardinia	Andesite	13	Savelli (2002)
4	S. Pietro island (SE Sardinia)	Comendite	14-15	Savelli (2002)
5	Montecristo	Cordierite monzogranite	7.3	Serri et al. (1993)
6	Vercelli seamount	Monzogranite	7.5-6.5	Savelli (2002)
7	Cornacya, seamount	Shoshonite	12.9-12.3	Mascale et al. (2001)
8	Capraia	Andesite, dacite, rhyolite	6.9-6	Savelli (2002)
9	Mt. Capanne, Elba	Granodiorite, monzogranite	6.8-6.2	Serri et al. (1993)
10	Porto Azzurro, Elba	Cordierite monzogranite	5.1	Serri et al. (1993)
11	Giglio	Cordierite monzogranites	5.1	Serri et al. (1993)
12	Anchise	Shoshonitic basalts and dacites	5.3-3.5	Savelli (2002)
13	Capraia	High-K andesites, dacites, rhyolites	4.6-3.5	Serri et al. (1993)
14	Montecatini, Tuscany	Lamproite	4.1	Savelli (2002)
15	San Vincenzo	Cordierite rhyolite	4.7	Serri et al. (1993)
16	Castel di Pietra & Gavorrano	Monzogranites, syenogranites	4.4-4.3	Serri et al. (1993)
17	Ponza island	Rhyolite	4.4	Savelli (2002)
18	Monteverdi	Cordierite monzogranites	3.8	Serri et al. (1993)
19	Tolfa	Trachydacite and rhyolite	3.8-2.3	Savelli (2000)
20	Manziana	Rhyolites	3.6	Serri et al. (1993)
21	Roccastrada	Cordierite rhyolite	2.5-2.2	Serri et al. (1993)
22	Cerite	Rhyolite	2.4	Serri et al. (1993)
23	Volturno plain	Basalt, andesite	>2	Savelli (2002)
24	Radicofani	Basalts	1.3	Savelli (2002)
25	Cimini	Trachytic dacite	1.3-0.9	Savelli (2002)
26	Roccamonfina	Basalt, trachytic basalt, latite	1.2-0.3	Savelli (2002)
27	Ponza island	Trachyte	1.1	Savelli (1988)
28	Lower slope	Basalts	1.1	Savelli (1988)
29	Mt Amiata	Trachytic dacite	0.3-0.2	Savelli (2000)
30	S. Venanzo	Melilitite/carbonatite	0.4	Stoppa & Wooley (1997)
31	Torre Alfina	Lamproite	0.8	Savelli (2002)
32	Cupaello	Melilitite/carbonatite	0.4	Stoppa & Wooley (1997)
33	Vulsini	Basanite, shoshonite, latite, trachyte	0.6-0.1	Savelli (2002)
34	Vico, Sabatini	Tephrite, leucitite, phonolite, trachyte	0.6-0.1	Savelli (2002)
35	Albani Hills	Leucitite.	0.6-0.1	Savelli (2002)
36	Ventotene volcanic complex	Trachytic basalts, phonolite	0.8-<0.2	Savelli (2002)
37	Phlegrei fields	Basalt	<0.04	Savelli (2002)
38	Vesuvius	Trachyte, phonolite	<0.03	Savelli (2002)
39	Vulture	Tephrite, phonolite	0.8-0.13	Stoppa & Wooley (1997)
40	Palinuro seamount	Basalt	0.8-0.3	Savelli (2002)
41	Marsili seamount	Andesite	0.8-0	Savelli (2002)
42	Enarete & Eolo seamounts	Basalts, andesites	0.8-0.6	Savelli (1988)
43	Alicudi island	Basalts	<0.2	Savelli (2002)
44	Filicudi island	Basalts	[illegible]	Savelli (2002)
45	Salina island	Basalt, andesite, dacite, rhyolite	0.4-<0.01	Savelli (2002)

Table 6.1. Continued

No	Location	Rock type	Age (Ma)	References
46	Stromboli island	Basalt, andesite	<0.06	Savelli (2002)
47	Panarea volcanic complex	Basalt, andesite	0.8-0.01	Savelli (2002)
48	Lipari island	Andesite, rhyolite	<0.25	Savelli (2002)
49	Vulcano island	Basalt, shoshonite, latite, rhyolite	<0.12	Savelli (2002)
Mainly Tholeiitic/alkaline – interplate				
50	Hyblean Plateau	Tholeiitic and alkaline basalts	8-6	Savelli (2001)
51	Capo Ferrato, Sardinia	Trachytic basalt, trachyte	5	Savelli (2002)
52	Aceste	Trachyte, rhyolite	5	Savelli (1988)
53	Montiferro, Sardinia	Basanite, trachytic basalt, phonolite	3.2-2.3	Savelli (2002)
54	Monte Arci, Sardinia	Basalt, andesite, rhyolite, trachyte	4-2.3	Savelli (2002)
55	Magnaghi seamount	Basalt	3-2.7	Savelli (2002)
56	Hyblean Plateau	Tholeiitic and alkaline basalts	3-1.2	Savelli (2001)
57	Logudoro, Sardinia	Basalt, trachytic basalt	0.9-0.2	Savelli (2002)
58	Vavilov seamount	Basalts	0.4-<0.1	Savelli (1988)
59	Ustica island	Basalt, trachyte, rhyolite	0.8-<0.2	Savelli (2002)
60	Pantelleria island	Alkaline magmas, peralkaline, rhyolites, trachytes	<0.3	Savelli (2001)
61	Etna	Tholeiitic basalts	0.5-0	Savelli (2002)
Oceanic basalts				
62	ODP 655 – Gortani Ridge	Basalts	4.6-4	Feraud (1990)
63	ODP 651	Basalts	3-2.6	Feraud (1990)
64	ODP 650 (Marsili Basin)	Basalts	1.9-1.6	Savelli (2002)

6.4.1. Magmatism

Neogene to Quaternary magmatic activity in the Tyrrhenian-Apennine region involved subduction-related calc-alkaline magmatism, intraplate alkaline magmatism and emplacement of MORB-type basalts (Savelli, 1988; 2000; 2001; 2002; Serri et al., 1993; Argnani and Savelli, 1999; 2001) (Figure 6.5 and Table 6.1). The distribution of subduction-related (mainly calc-alkaline) magmatism defines a geochemical polarity associated with a west-dipping subduction zone (e.g. Savelli, 2000; Argnani and Savelli, 2001). These magmas become younger from west to east, which I believe, is the response to rollback of the subduction hinge during the opening of the Tyrrhenian Sea.

The oldest volcanic rocks in the Tyrrhenian margins are Oligocene-Middle Miocene (32-13 Ma) calc-alkaline rocks found in Corsica and Sardinia (Savelli, 2002, and references therein). The youngest of these magmatic centres (16-13 Ma) were active after Corsica and Sardinia stopped rotating and indicate the location of a N-S-striking volcanic arc that existed at that time (Figure 6.5a). The subsequent migration of this magmatic arc is reconstructed in Figure 6.6.

The earliest stage in the migration of the Tyrrhenian-Apennine magmatic arc is documented by a series of N-S-striking magmatic centres dated at 13-7 Ma (Figure 6.5b). These magmatic rocks are found 100-150 km east of the former position of the magmatic arc, suggesting a rapid episode of slab rollback that probably took place during the Tortonian (11-7 Ma). Subduction rollback continued during the Messinian (7-5 Ma) and led to cessation of calc-alkaline magmatism in the offshore area of Sardinia and southern Corsica, and to the onset of magmatism near the eastern margin of the Tyrrhenian Sea.

A second episode of subduction rollback is inferred from the distribution of post-Messinian (6-0 Ma) calc-alkaline magmatism. During this period, rollback continued in the southern domain, but slowed down or stopped in the northern domain. The episode of rapid rollback in the southern domain is inferred from the observation that the positions of the post-Messinian (<6 Ma) magmatic-arc in the southern Tyrrhenian are relatively spaced (Figure 6.6).

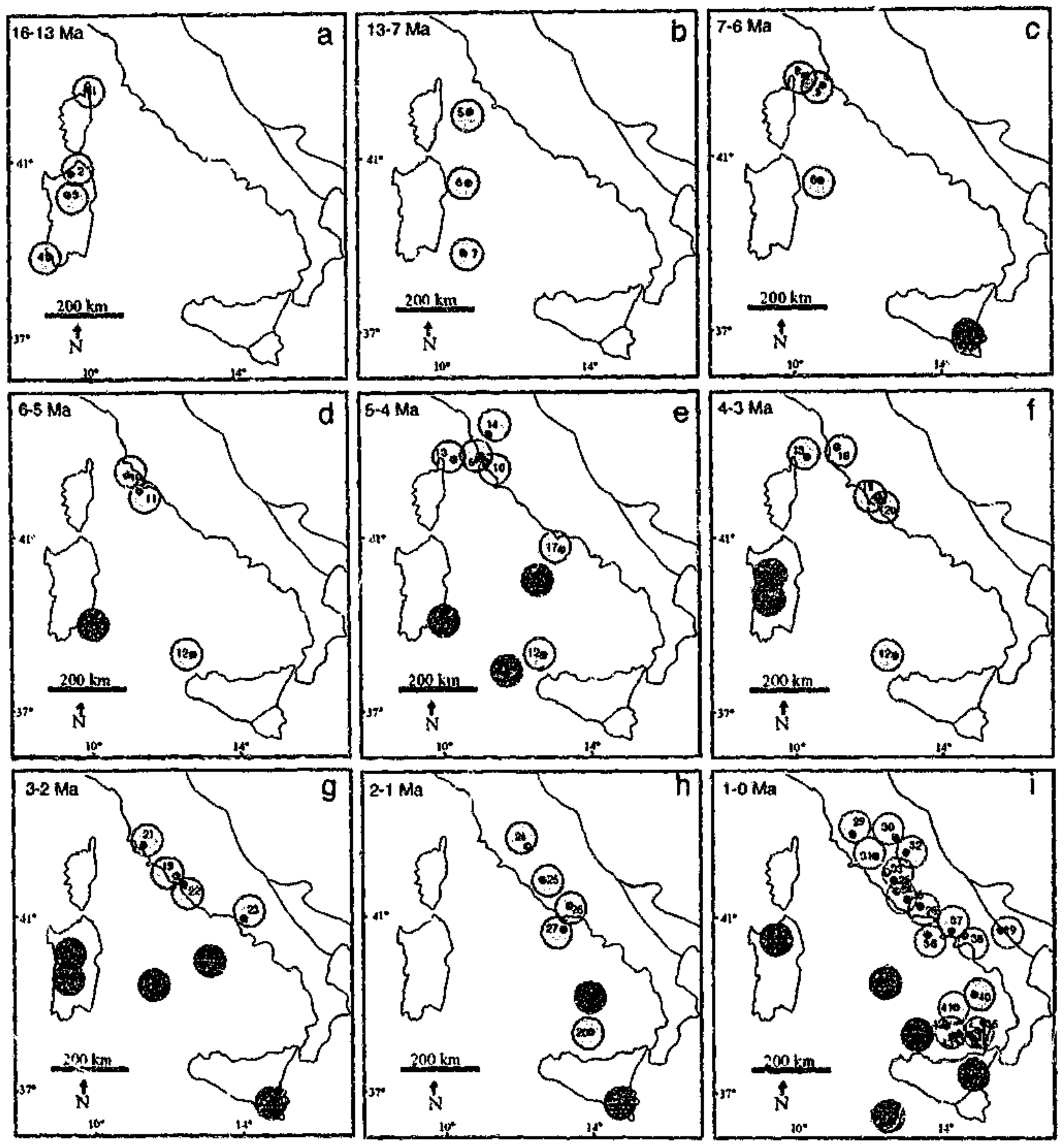


Figure 6.5. Distribution of subduction-related (purple), intraplate (green) and MORB-type (blue) magmatism in the Tyrrhenian region. The numbers correspond to the list of references in Table 6.1.

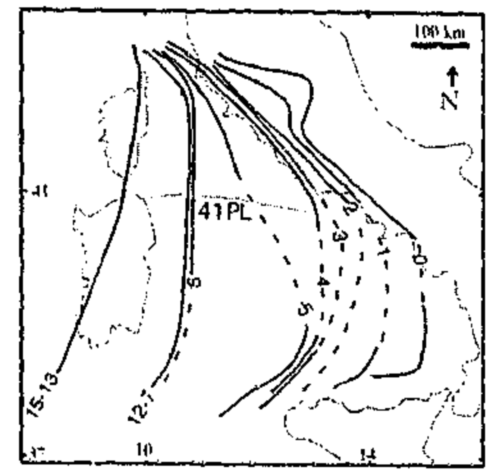


Figure 6.6. Map showing the approximate location of the magmatic arc through time based on the data shown in Figure 6.5. Numbers indicate ages in Ma. The contour map clearly shows an overall eastward migration of the magmatic arc since the Late Miocene. Dashed lines indicate inferred locations where no data exist. 41PL, 41st Parallel Line.

6.4.2. The geometry of the subducting slab

Seismic data from the Apennines and the Calabrian arc support the existence of a segmented, locally detached, west-dipping subduction zone in the region (e.g. Anderson and Jackson, 1987a; Selvaggi and Amato, 1992; Wortel and Spakman, 1992; Lucente et al., 1999). The Tyrrhenian-Apennine slab is illustrated in three segments, the northern Apennine slab, the central-southern Apennine slab and the Ionian slab (Figure 6.7).

The existence of a cold lithospheric slab in the northern Apennine arc is interpreted from few occurrences of deep earthquakes at depths of up to 90 km (Selvaggi and Amato, 1992) and from zones of relatively high seismic velocities recognised in tomographic models (Wortel and Spakman, 1992; Lucente et al., 1999) (Figure 6.7a). Due to the low resolution of the tomographic images, it is debateable whether these seismic anomalies represent a continuous slab that is presently subducting beneath the northern Apennines (Lucente et al., 1999), or a slab that is separated from the overriding plate by a zone of low velocity asthenosphere (e.g. Wortel and Spakman, 1992; 2000). Wortel and Spakman (1992) have explained their model by a detachment of the lithospheric slab, which has been propagated laterally beneath the Apennines. In contrast, the analysis shown here is based on the tomographic model of Lucente et al. (1999), which shows a continuous subducting slab beneath the northern Apennines. Nevertheless, the possibility that the northern Apennine slab has been detached cannot be excluded.

In the central Apennines, evidence for slab detachment has been inferred both from the models of Wortel and Spakman (1992) and Lucente et al. (1999). In both models, there is no high-velocity anomaly at shallow and intermediate depths (up to 250 km) beneath the central Apennines, which means that subduction does not take place at these depths. However, a zone of high seismic velocities at depths of 400-700 km (Figure 6.7b) may point to the existence of a detached slab beneath the central Apennines.

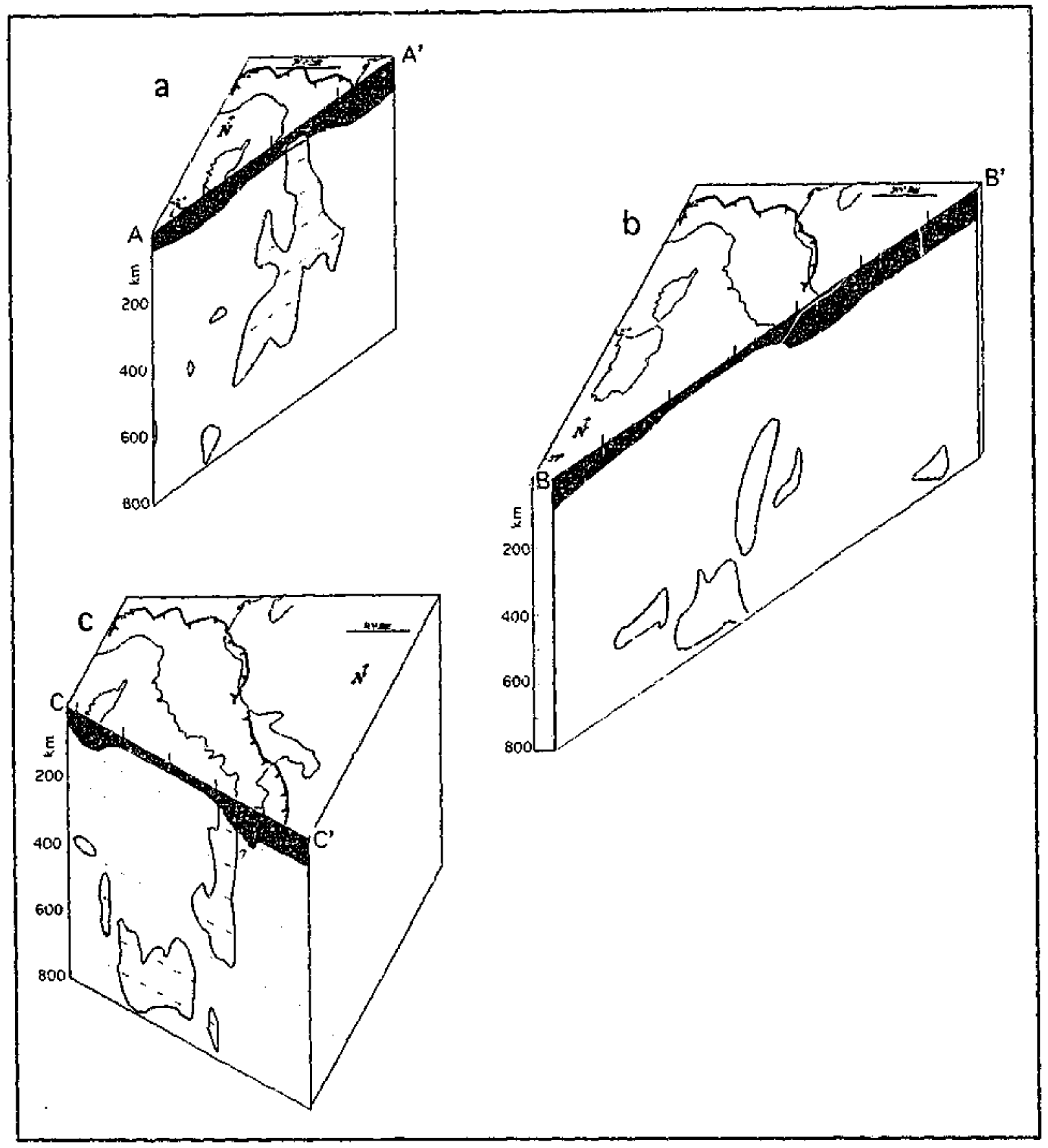


Figure 6.7. Upper mantle cross sections beneath the (a) northern Apennines, (b) central Apennines and (c) Calabria. The cross sections show positive (>2%) seismic velocities of P waves (after Lucente et al., 1999; light grey), and estimated lithospheric thickness (after Della Vedova et al., 1991; dark grey).

The continuous nature of the southernmost segment of the Tyrrhenian-Apennine slab (the Ionian slab) in the Calabrian arc is in contrast to the fragmented central Apennine slab. In this region, there is a good agreement between different tomographic models (Wortel and Spakman, 1992; 2000; Lucente et al., 1999) that show a relatively narrow zone (ca. 300 km) of high seismic velocities. This seismic anomaly indicates a continuous sub-vertically cold slab dipping towards the northwest. The existence of this slab is also supported by occurrences of deep earthquakes (at depths of up to 500 km) defining the Benioff-Wadati zone (Anderson and Jackson, 1987b). The tomographic images (Figure 6.7c) suggest that the subducted slab stagnated at a depth of ca. 700 km, that is, around the

670 km discontinuity. Remnants of the slab can then be traced horizontally parallel to the 670 km discontinuity over a distance of 600-700 km. The zone of active subduction is localised only in the area where the Tyrrhenian-Apennine subduction system meets the oceanic lithosphere of the Ionian Sea (Rosenbaum et al., in review). This may suggest that the Calabrian arc is the only area within the Tyrrhenian-Apennine system where subduction processes have not been impeded by accretion of continental material, as will be discussed later.

6.5. Reconstruction of subduction rollback

Using the above constraints, it is possible to reconstruct the evolution of the subducting slab (Figure 6.8). The reconstruction is based on the assumption that slab segments recognised beneath the Tyrrhenian-Apennine system have been subducted during or before the opening of the Tyrrhenian Sea. The results shown in Figure 6.8 have been obtained by preserving the length of each slab segment, and by applying an incremental reconstruction of the subduction zone. In this process, I assume that the convergence between the downgoing plate and the overriding plate is negligible so all components of horizontal motions are attributed to subduction rollback. This is a sound assumption because the rates of convergence in the last 10 Myr were considerably slower than the rates of subduction rollback (Figure 6.9). In the central Mediterranean, the vector of Africa-Europe convergence since 10 Ma is estimated to be 7-7.4 km/Myr towards the northeast (DeMets et al., 1994; Rosenbaum et al., 2002a). This implies zero convergence in cross sections perpendicular to the Apennine belt, and ca. 6 km/Myr convergence along cross section CC' (Figure 6.9). This value is negligible when compared with the rates of subduction rollback (20-100 km/Myr).

Predictions on the behaviour of the slab during subduction rollback are based on results from analogue experiments (Faccenna et al., 2001a; Schellart, 2003), which simulated subduction by initiating gravitational sinking of a relatively high-viscosity upper layer (simulating the lithosphere) in a model box filled with low viscosity material simulating the sub-lithospheric mantle. Both sets of experiments have shown that under these circumstances, the slab was subjected to rapid rollback, and that stagnation and flattening of the slab occurred when it reached a mechanical boundary simulating the 670 km discontinuity.

The age of each stage in the reconstruction is deduced from the location of the magmatic arc relative to the subducting slab. These time constraints provide the ability to calculate the length of the slab that has been subducted during the opening of the Tyrrhenian Sea. In the northern transect, the total length of the slab subducting since 13 Ma has been ca. 200 km (Figure 6.8a). The deeper parts (>200 km depths) of the northern Apennine slab have therefore been subducted earlier, that is, during the opening of the Ligurian-Provençal basin at the Early Miocene. The rates of rollback in this transect were greater at earliest stages of the Tyrrhenian extension (~10-8 Ma).

The history of the central Apennine slab is relatively poorly constrained (Figure 6.8b). However, based on its 3D position with respect to the northern and southern slab segments (Figure 6.8 and

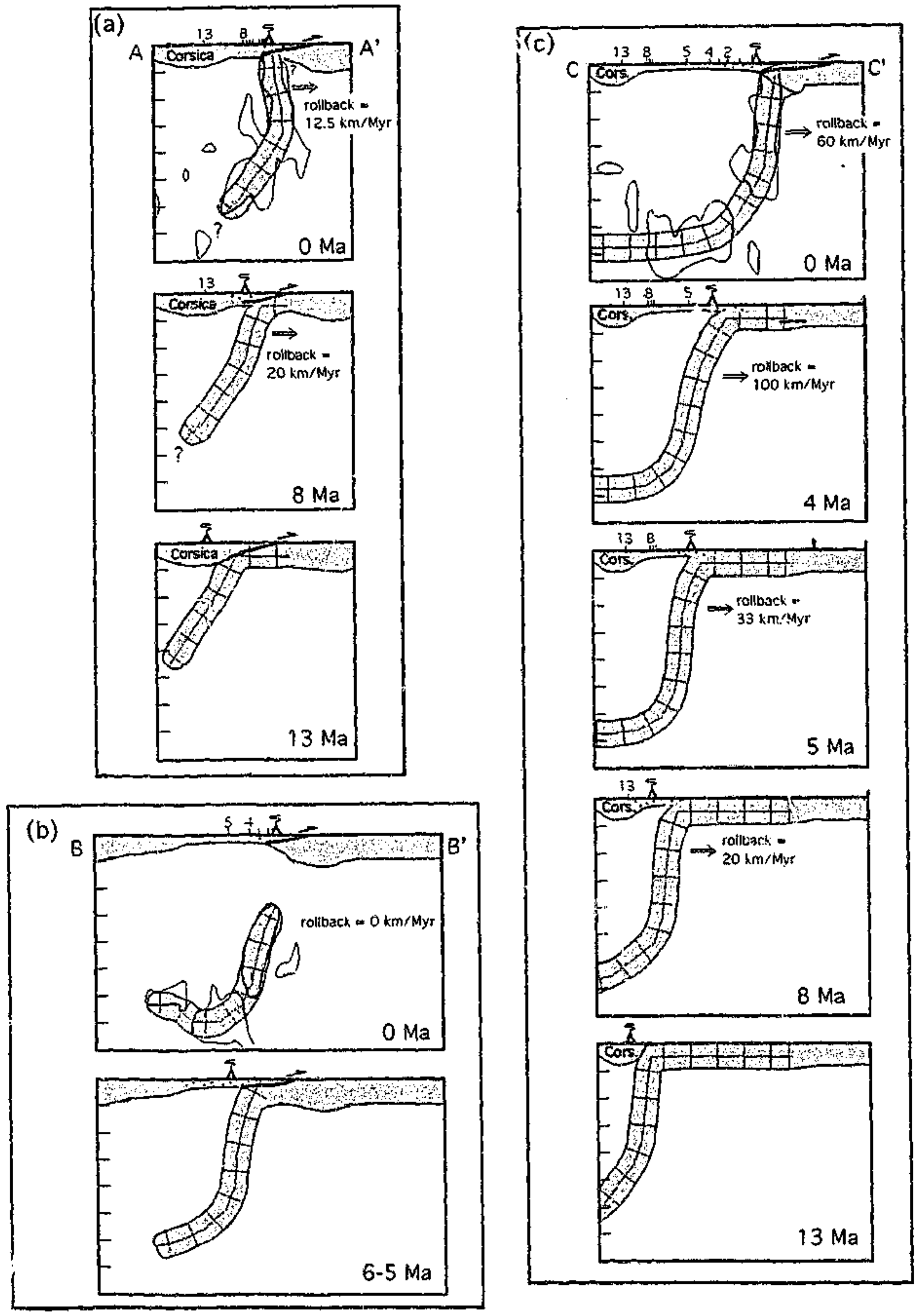


Figure 6.8. Reconstruction of the subducting slab at different stages during the opening of the Tyrrhenian Sea. (a) Northern Apennine transect. (b) Central Apennines transect. (c) Calabrian arc transect. The short vertical thin lines and adjacent numbers indicate locations and ages (in Ma) of the magmatic arc. Thick solid lines indicate the approximate location of the thrust front (unknown for the Calabrian transect) and dots indicate areas of syn-rift sedimentation. Zones of high seismic velocities at 0 Ma are also shown.

Figure 6.10). I suggest an age of 6-5 Ma for the tearing of the slab. Since then, subduction processes have ceased and the slab has been subjected to gravitational foundering. Therefore, it is likely that the origin of younger (3-0 Ma) volcanics in the central Apennines has not been directly derived from subduction processes (e.g. Lavecchia and Stoppa, 1996).

The role of subduction rollback is best illustrated in the cross section across the Calabrian arc (Figure 6.8c). Here, the length of slab subducted during the opening of the Tyrrhenian Sea is estimated as ca. 530 km. The reconstruction shows an increase in the rates of rollback at ca. 5 Ma, with maximum rollback velocities of ca. 100km/Myr.

The evolution of the subducting lithosphere in the Tyrrhenian-Apennine system in a map view is shown in Figure 6.10. I suggest that tearing of the slab occurred at ~6-5 Ma beneath the central Apennines, and was followed by a subsequent period of rapid rollback in the southern Tyrrhenian Sea. I further propose that the tearing of the central Apennine slab may have triggered the episode of rapid rollback.

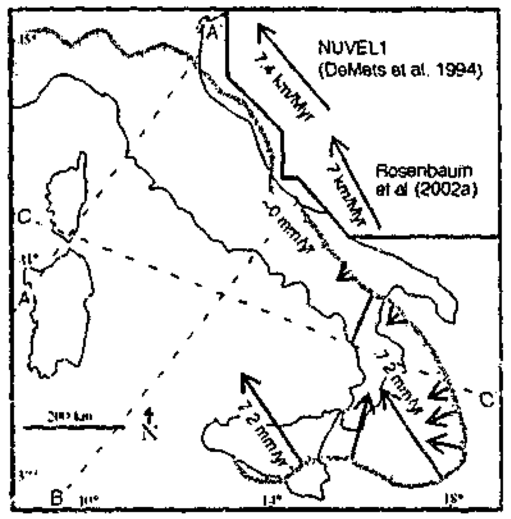


Figure 6.9. Component of convergence perpendicular to the strike of the subduction zone (black arrows) calculated from the relative motion of Africa and Europe (after DeMets et al., 1994; Rosenbaum et al., 2002a). Velocities are indicated by the relative lengths of the arrows. The component of convergence is greatest in the Calabrian arc and in Sicily (7.2 km/Myr) and is equal to zero in the Apennines. Convergence perpendicular to the section CC' is ca 6 km/Myr, which is one magnitude of order smaller than the rates of subduction rollback.

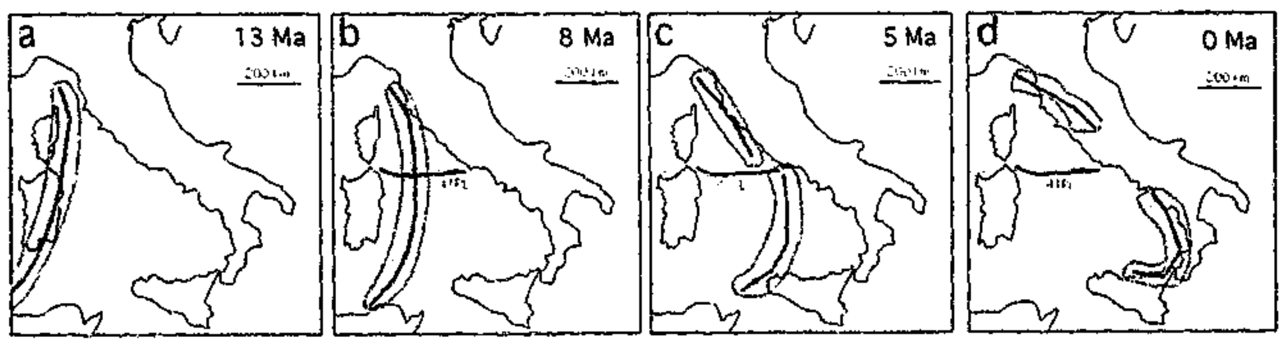


Figure 6.10. Maps showing the evolution of the subducting lithospheric slab beneath Italy since the Late Miocene. Shaded areas are surface projections of the slab at depths of 30-100 km based on the tomographic model of Lucente et al. (1999) and the results of this reconstruction (Figure 6.8). Note that the reconstruction predicts tearing of the slab at the latest Messinian (~6-5 Ma), and a subsequent period of rapid rollback in the southern Tyrrhenian Sea. 41PL, 41st Parallel Line.

6.6. The origin of the Apennines

6.6.1. Palaeomagnetic constraints

In this section I consider the role of Neogene to Quaternary block rotations in the Tyrrhenian-Apennine region as interpreted from palaeomagnetic data (Table 6.2 and Figure 6.11). These rotations indicate the sense of motions during deformation in the Apennine-Calabrian-Maghrebide fold-and-thrust belts. Since deformation took place contemporaneously with the opening of the Tyrrhenian Sea, it is possible to combine these rotations within the framework of subduction rollback and back-arc extension (Figure 6.11b).

Table 6.2. Rotations in the Apennines-Maghrebides since the Tortonian (see Figure 6.11 for location).

No	Inferred rotation	Time span	Reference
Northern Apennines			
1	28° CCW	Since Tortonian	Muttoni et al. (2000)
2	20-60° CCW	Since Messinian	Speranza et al. (1997)
3	20° CCW	Since Messinian	Speranza et al. (1997)
4	15° CW	Since Messinian	Speranza et al. (1997)
5	No rotation	Since Late Messinian	Mattei et al. (1996)
Central Apennines			
6	No rotation	Since Late Pliocene	Sagnotti et al. (1994)
7	28° CCW	Since L. Miocene	Mattei et al. (1995)
8	35° CCW	Since Messinian	Speranza et al. (1998)
Southern Apennines			
9	22-23° CCW	L. Pliocene – E Pleistocene	Scheepers et al (1993)
10	80° CCW	<15 Ma	Gattacceca & Speranza (2002)
11	15° CCW	During Late Pliocene	Scheepers & Langereis (1994)
11	15° CCW	During Middle Pleistocene	Scheepers & Langereis (1994)
11	9° CCW	Since Middle Pleistocene	Scheepers & Langereis (1994)
Calabria-Peloritan			
12	19° CW	Tortonian-Messinian	Speranza et al. (2000)
13	25° CW	During Tortonian	Duermeijer et al. (1998)
14	15-20° CW	Pleistocene	Mattei et al. (1999)
15	10-20° CW	During Middle Pleistocene	Scheepers et al. (1994)
16	14° CW	Plio-Pleistocene	Aifa et al. (1988)
Sicily			
17	25° CW	Since Lower Pliocene	Grasso et al. (1989)
18	34° CW	Since Early Pliocene	Scheepers & Langereis (1993)
19	55° CW	Since Early Pliocene	Speranza et al. (1999)

Clockwise (CW) and counterclockwise (CCW)

The pattern of deformation in the Apennines is predominantly governed by counterclockwise rotations (Figure 6.11a). In general, the degree of rotation increases, and the ages of these rotations become younger, from north to south along the Apennines. There are a few exceptions to this deformational pattern, as for example, in the external part of the Northern Apennines (Figure 6.11a).

where post-Messinian (<5 Ma) clockwise and counterclockwise rotations correspond to the arcuate bending of the orogen (Speranza et al., 1997; Lucente and Speranza, 2001).

Palaeomagnetic evidence from Calabria and Sicily suggest predominantly clockwise rotations since the Late Miocene (Table 6.2 and Figure 6.11a). Speranza et al. (2000) have reported on earlier (Early-Middle Miocene?) counterclockwise rotations of the Calabrian block, which were probably associated with the lateral migration of this block from its Oligocene position in the northwestern Mediterranean (Rosenbaum et al., 2002b). Clockwise rotations in Sicily occur contemporaneously with nappe emplacement and the opening of the southern Tyrrhenian Sea (Oldow et al., 1990). These rotations show an opposite sense of motion with respect to coeval rotations in the southern Apennines, suggesting an arcuate formation of the fold-and-thrust belt during subduction rollback of the Ionian slab (Figure 6.11b).

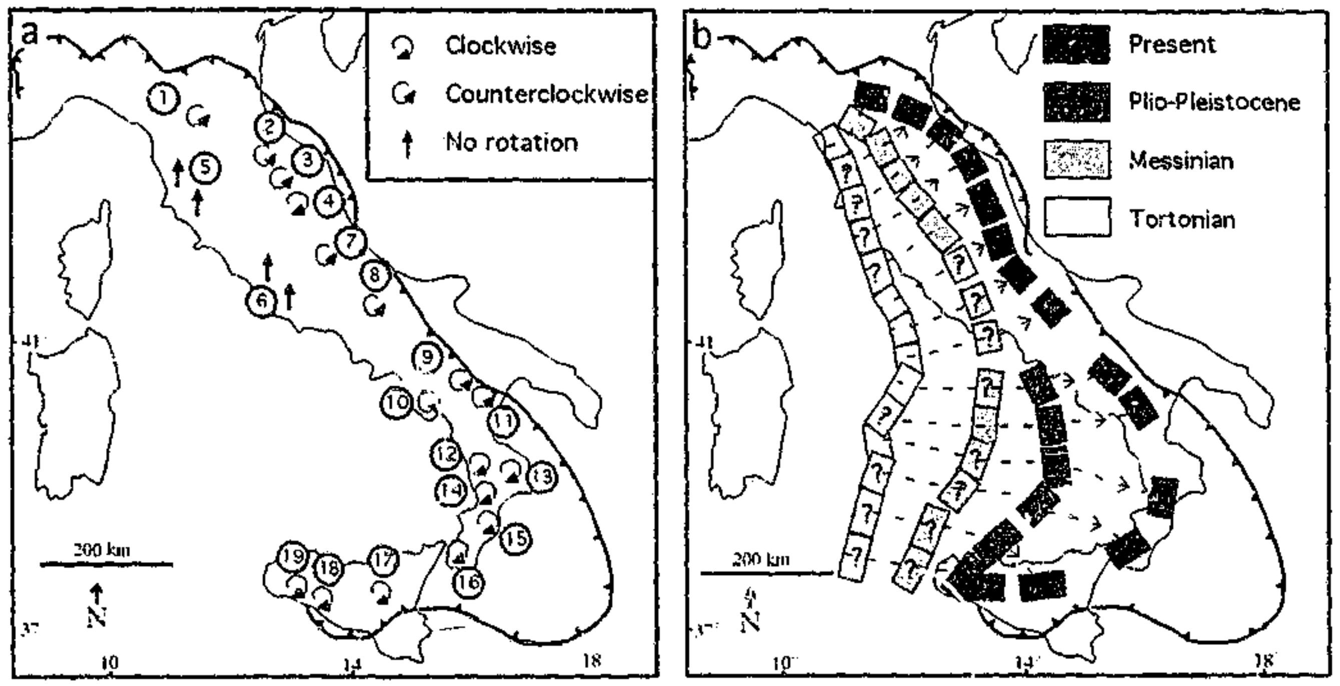


Figure 6.11. (a) Sense of block rotations in the Apennines-Maghrebides as inferred from palaeomagnetic data. Numbers correspond to the list of palaeomagnetic rotations in Table 6.2. (b) Restoration of rigid blocks with respect to the history of extension in the Tyrrhenian Sea (Figure 6.6 and Figure 6.10) and the amount of inferred palaeomagnetic rotations. Dashed lines and arrows indicate the sense of tectonic transport. Rotations that are not based on palaeomagnetic data are indicated by question marks.

6.6.2. Palaeogeography

The evolution of the Apennine-Maghrebide orogen is further investigated on the basis of palaeogeographic interpretations of sedimentary facies (Figure 6.12). Since the Tortonian, the locus of orogenesis occurred in the western margin of Adria, which consisted of a heterogeneous sedimentary cover made of Mesozoic carbonate platforms and basinal pelagic deposits (Figure

6.12b). The Adriatic carbonate platforms were deposited during the Mesozoic in a shallow marine environment (e.g. Bosellini, 2002, and references therein). The most external platform in the central and southern Apennines is the foreland area of the Apulian platform (Figure 6.12a). More internal platforms (hereafter termed 'Internal platform'), such as the Campania-Lucania platform and the Latium-Abruzzi platform, are presently thrust on top of the Apulian platform. The suture between the Apulian and the Internal platform is made of two basinal domains represented by the Lagonegro and the Molise units (Mostardini and Merlini, 1986). These units consist of Triassic to Oligocene clays, cherts, radiolarites and limestones, overlain by Miocene turbiditic sequences (Wood, 1981; Marsella et al., 1995). They were deposited in a deep-marine environment that was located west of the slope of the Apulian carbonate platform (Marsella et al., 1995) (Figure 6.12b).

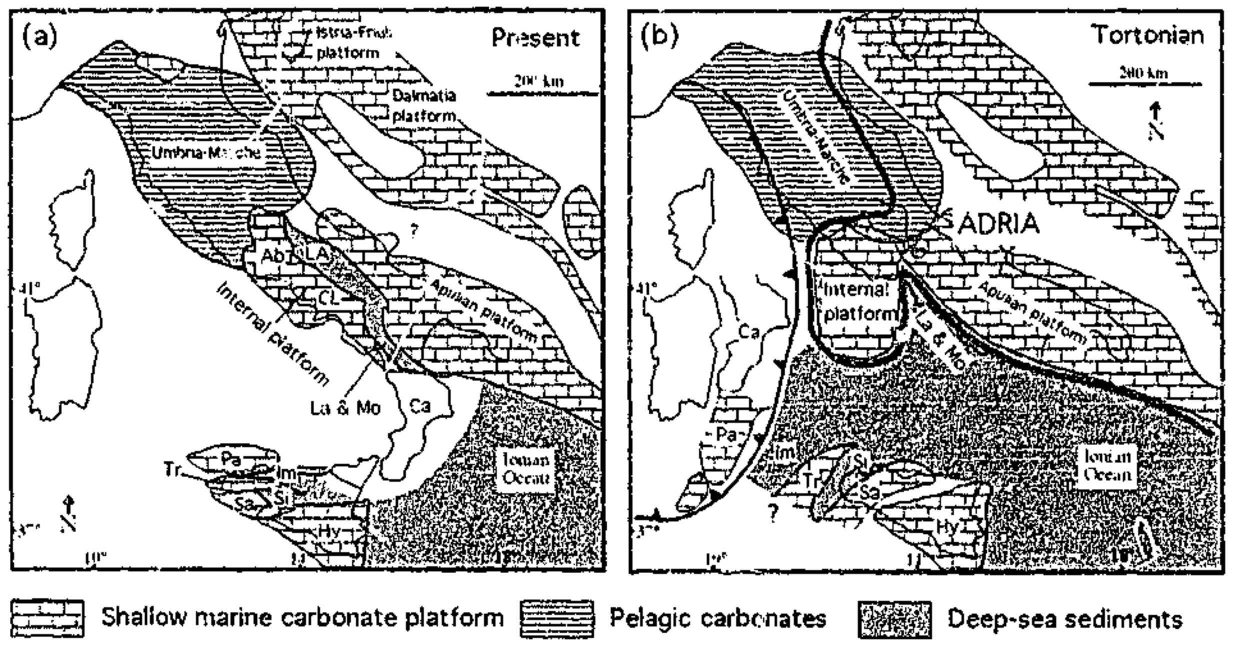


Figure 6.12. (a) The spatial distribution of sedimentary facies in the western margin of Adria (modified after Oldow et al., 1990; Marsella et al., 1995; Bosellini, 2002). (b) Restored positions of Apennine-Maghrebide terranes prior to the opening of the Tyrrhenian Sea. The thick grey line indicates the irregular continental margin of Adria. Ab, Abruzzi; Ca, Calabria; CL, Campania-Lucania platform; Hy, Hyblean platform; Im, Imerese; La, Lagonegro; Mo, Molise; Pa, Panormide platform; Sa, Saccense; Si, Sicanian; Tr, Trapanese.

The palaeogeography of the northern Apennines differs from the palaeogeography of the central and southern Apennines (Figure 6.12a) (see also discussion by Lucente and Speranza, 2001). In the Umbria-Marche region of the northern Apennines, the sedimentary pile predominantly consists of Early Jurassic to Paleogene pelagic limestones deposited on top of Triassic evaporites (Lavecchia et al., 1988, and references therein). Furthermore, the crust and lithosphere beneath the Umbria-Marche region is thinner in comparison with the central and southern Apennines (Calcagnile and Panza, 1980; Geiss, 1987). This may suggest that sediments of the northern Apennines were not deposited on a 'normal' continental crust and that the shape of the continental margin of Adria, when incorporating in Apennine-Maghrebide orogeny, was irregular (Figure 6.12b).

The sedimentary facies in Sicily is characterised by several lithotectonic assemblages of Mesozoic carbonate platforms (Panormide, Trapanese, Saccense and Hybley) and two basinal units with pelagic deposits (Imerese and Sicanian basins) (Catalano and D'Argenio, 1978; Oldow et al., 1990). In the most internal part of the fold-and-thrust belt and at the highest structural position, the Panormide carbonate platform is separated from the underlying platform and seamount of Trapanese by sediments of the Imerese basin (Figure 6.12a). The Sicanian basin, in turn, is juxtaposed between the Trapanese unit and the Saccense platform (Channell et al., 1990b; Oldow et al., 1990). This tectonic configuration suggests that the imbrication of thrust sheets involved closure of pelagic basins that now define local sutures between the carbonate tectonic allochthons (Figure 6.12b).

The suggested palinspastic reconstruction (Figure 6.12b) combines the palaeogeographic data with the constraints on the kinematic evolution of the orogen (i.e. Figure 6.11b). I suggest that the incorporation of heterogeneous crustal material at orogenic processes resulted in an uneven distribution of strain during deformation. Carbonate platforms moved as relatively rigid blocks, whereas basinal units were subjected to intense crustal shortening deformation. I further speculate that the arrival of the Internal platform at the subduction zone impeded subduction processes and led to tearing of the central Apennine slab, and also that the closure of deep-sea basins (Lagonegro, Molise, Imerese and Sicanian basins) in the westernmost Ionian Ocean enabled ongoing rollback in the southern Apennines, Calabria and Sicily.

6.7. Discussion

6.7.1. Tectonic evolution of the Tyrrhenian Basin

Results of the spatio-temporal analysis suggest that the major stages of opening in the Tyrrhenian Sea did not begin before the Late Miocene (ca. 13-10 Ma). It should be stressed that earlier Oligocene-Early Miocene extensional episodes in the Corsica basin (e.g. Mauffret and Contrucci, 1999) and in the northern Apennines (e.g. Carmignani et al., 1994) were probably related to the opening of the Ligurian-Provençal basin, and were accompanied by the counterclockwise rotation of Corsica and Sardinia. Palaeomagnetic evidence (Speranza et al., 2002) suggests that Corsica and Sardinia reached their present positions by 16 Ma. Following that, during the Middle Miocene (16-10 Ma), extension in the Tyrrhenian Sea stopped, as indicated by the absence of Middle Miocene syn-rift sediments in the region (Figure 6.4). Moreover, during this period, the position of the magmatic arc remained relatively fixed (Figure 6.6), suggesting that the subduction zone did not rollback.

Rifting in the Tyrrhenian Sea commenced during the Tortonian (~10-9 Ma) and was localised along the margins of Corsica and Sardinia. This extensional event, which is documented by the occurrence of Tortonian syn-rift sediments, occurred contemporaneously with a rapid eastward migration of the magmatic arc. Therefore, Tortonian extensional tectonics in the Tyrrhenian Sea is interpreted as a direct tectonic response to subduction rollback. These constraints on the early stages of back-arc extension are significant because previous workers have debated whether subduction rollback has actually played a role in the opening of the northern Tyrrhenian Sea (e.g. Faccenna et al., 1996; Mantovani et al., 1996).

A major change in the style of subduction rollback and back-arc extension occurred during the Late Messinian (~6-5 Ma) in the Tyrrhenian-Apennine system. This change is recognised both in the distribution of syn-rift sediments and the distribution of magmatic activity. These independent data sources suggest that at ~6-5 Ma, the rates of subduction rollback and back-arc extension slowed down in the northern Tyrrhenian, but increased considerably in the Southern Tyrrhenian. The differential velocities of extension were accommodated by the deep-seated fault zone of the 41° Parallel Line that was active as a left-lateral strike-slip fault during the Pliocene and the Pleistocene (Lavecchia, 1988; Boccaletti et al., 1990c; Spadini and Wezel, 1994; Bruno et al., 2000). During this extensional episode, the extreme rates of subduction rollback (60-100 km/Myr) induced sufficient lithospheric attenuation in the overriding plate to form new oceanic crust in the Vavilov and Marsilli basins.

6.7.2. Tectonic response to accretion of continental crust

An important aspect of this study is the possible relationships between palaeogeography and the geometry of the subducting slab. Apart from the uncertain role of present-day subduction processes in the northern Tyrrhenian (e.g. Selvaggi and Amato, 1992; Lucente and Speranza, 2001), current subduction in the Tyrrhenian-Apennine system is restricted to a narrow zone in the Calabrian arc. The reason for this geometry is that the Calabrian arc is the only area within the Tyrrhenian-Apennine system that still consumes oceanic lithosphere (i.e. the Ionian Ocean). In contrast, there is no evidence for active subduction in the central Apennines, where a detached remnant slab is found only at depths greater than 250-300 km (Figure 6.7b). In this region, the palaeogeography was dominated by a large carbonate platform (the Internal platform; Figure 6.12b). It is suggested that the arrival of a relatively buoyant continental block at the subduction zone may have impeded the subduction processes, leading to a detachment of the subducting slab from the surface, and to tearing of the slab at the boundary between the continental block and the oceanic lithosphere (Figure 6.13). This mechanism is essentially similar to formation of slab tears as a result of docking of allochthonous terranes in the subduction system (Ben-Avraham et al., 1981; Nur and Ben-Avraham, 1982).

The formation of slab tears is a crucial trigger for changing the history of rollback. In the Tyrrhenian Sea, time constraints on the tearing of the slab and the accretion of the Internal platform indicate that these events took place at approximately 6-5 Ma. These events were followed by a period of rapid rollback in the southern Tyrrhenian (in excess of 100 km/Myr), and the formation of an oceanic back-arc basin. It is suggested that slab tearing was the trigger for the subsequent Plio-Pleistocene episode of increased rapid rollback/back-arc extension in the southern Tyrrhenian. The tearing resulted in the formation of a relatively narrow slab (e.g. Dvorkin et al., 1993) remaining in the southern Tyrrhenian. Consequently, this slab was subjected to reduced hydrodynamic suction

from sideways asthenospheric flow (Figure 6.13), which is the lifting force that prevents sinking of the slab into the asthenosphere (Dvorkin et al., 1993; Nur et al., 1993). It is expected that the decrease of this force would act to accelerate subduction rollback.

From this example, it appears that episodes of back-arc extension related to subduction rollback are strongly influenced by the rheological properties of the lithosphere and the interaction with relatively rigid continental blocks. Subduction rollback is likely to follow the geometry of existing oceanic crust. Thus, while docking of thick continental material in the subduction system may impede subduction processes, the incorporation of deep basins at the subduction zone could provide a relatively free boundary for further rollback. A similar assumption has led Sengör (1993) to suggest that the closure of small, possibly oceanic basins, such as the Lagonegro basin, may have facilitated further subduction rollback of Adria. The present reconstruction shows (Figure 6.12b) that these basins, the Lagonegro, Molise, Imerese and Sicanian basins, probably accommodated subduction rollback, although the main component of subduction rollback is attributed to the closure of the narrow Ionian Ocean.

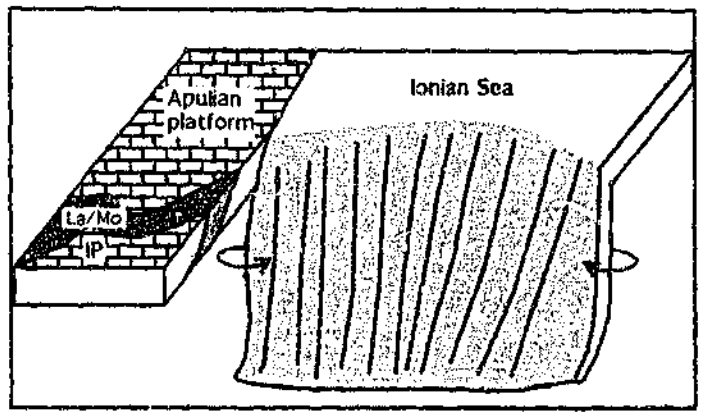


Figure 6.13. Schematic illustration showing possible relationships between the geometry of the subducting slab and the rheological properties of the crust and lithosphere. Tearing of the slab occurred as a result of the arrival of thick continental material (the Internal platform) at the subduction system, combined with ongoing subduction of oceanic lithosphere (the Ionian Sea) in the Calabrian arc. Following tearing, rollback was further accelerated by sideways asthenospheric flow (arrows). IP, Internal platform; La, Lagonegro; Mo, Molise.

6.8. Conclusions

We have used sets of spatio-temporal constraints in order to obtain a better resolution on the occurrence of crustal and lithospheric processes associated with the evolution of the Tyrrhenian Sea and the Apennine-Maghrebide belt. The earliest evidence for extensional tectonism is associated with Oligocene-Early Miocene (~25-16 Ma) syn-rift sediments found in the northern Tyrrhenian. This extensional episode occurred before or during the opening of the Ligurian-Provençal basin, and before Corsica and Sardinia stopped rotating (at ~16 Ma). The major extensional episode associated with the opening of the Tyrrhenian Sea commenced only at ~10-9 Ma, after a period of 6-7 Myr of relative quiescence in extensional processes during the Middle Miocene.

The first stage of opening of the Tyrrhenian Sea, at 10-6 Ma, was characterised by widespread extension in the northern domain, and rifting in the western part of the southern domain. In contrast, the second stage of opening (after ~6-5 Ma) was localised in the southern Tyrrhenian Sea, and involved extreme rates of subduction rollback (in excess of 100 km/Myr) and the formation of new oceanic crust in the back-arc region. The transition between the two stages of opening, at about 6-5 Ma, was triggered by docking of the Internal carbonate platform in the central Apennine subduction system, which I propose, led to the formation of a slab tear. Subsequently, the remaining Ionian slab beneath Calabria was narrower, and was consequently subjected to accelerated rates of subduction rollback.

CHAPTER 7

FORMATION OF ARCUATE OROGENIC BELTS IN THE WESTERN MEDITERRANEAN REGION

The best picture that I can give of the origin of the large mountain ranges consists of imagining that when my hand is scarped by accident, folds of skin become piled up in one direction, while behind them the skin is torn and little blood wells up.

Edward Suess, Die Heilquellen Böhmens (1878)
(translated from German by A.M Celal Sengör)

FOREWORD: CHAPTER 7

This chapter deals with the origin and the evolution of arcuate orogenic belts in the western Mediterranean region. The preparation of the chapter benefited from discussions with Gordon Lister, Roberto Weinberg and Wouter Schellart. Comments on the manuscript by Roberto Weinberg, David Giles and Peter Betts are also acknowledged. A slightly modified version of this chapter has been submitted to *Geological Society of America Special Paper* and is currently in review[a].

[a] Rosenbaum, G., Lister, G.S., Formation of arcuate orogenic belts in the western Mediterranean region. *Geological Society of America Special Paper*, in review.

7. FORMATION OF ARCUATE OROGENIC BELTS IN THE WESTERN MEDITERRANEAN REGION

Abstract

The Alpine orogen in the western Mediterranean region, consisting of the Rif-Betic belt and the Apennine-Calabrian-Maghrebide belt, is a classic example of an arcuate orogen. It contains fragments of Cretaceous to Oligocene high-pressure/low-temperature (HP/LT) rocks, which were exhumed and dispersed during post-Oligocene extensional deformation and are presently exposed in the soles of metamorphic core complexes. In this chapter I illustrate that the arcuate shape of the orogenic belt was attained during extensional destruction of the earlier HP/LT belt, driven by subduction rollback in a direction oblique or orthogonal to the direction of convergence. Since the Oligocene, subduction of Mesozoic oceanic lithosphere, accompanied by rollback of the subducting slab, led to progressive bending and episodic tearing of the slab. This process resulted in the formation of several slab segments presently recognised in tomographic images beneath the Alboran Sea, North Africa and Italy. The remnant slabs can account for nearly all the volume of oceanic domains that existed in the western Mediterranean during the Oligocene. Subduction rollback led to extension in the overriding plate and to the opening of back-arc basins. Extensional tectonism affected the original, relatively non-arcuate HP/LT belt. Allochthonous fragments of the original belt (e.g. Alpine Corsica, Calabria and the Internal Betic) rotated and drifted as independent units until they were accreted in an arcuate fashion into the continental palaeomargins of Africa, Iberia and Adria. Therefore, the present exposures of HP/LT metamorphic rocks in the western Mediterranean region do not represent sites of continental collisions between major large-scale tectonic plates .

7.1. Introduction

Arcuate orogenic belts are common tectonic features defined by along-strike variations in the structural trend of the orogen. Following the works of Carey (1955) and Marshak (1988), arcuate belts have been classified into two groups, namely rotational arcs (oroclines), which are orogens that underwent syn-orogenic bending, and non-rotational arcs that initiated in their present curved form. A further classification of oroclines may be related to the degree to which the crust and lithosphere were involved in the oroclinal bending. Some oroclines, such as the arc of the Jura Mountains in the northwestern Alps (Figure 7.1), are superficial tectonic features that formed by thin-skinned deformation in the fold-and-thrust belt. In other oroclines, however, the arcuate shape of the orogen seems to be rooted deep in the lithosphere. For example, in the northern Apennine arc (Figure 7.1), there is evidence for oroclinal bending at both crustal and lithospheric scales (Lucente and Speranza, 2001). The process of oroclinal bending associated with thin-skinned deformation in fold-and-thrust belts has been subjected to extensive studies (e.g. Marshak, 1988; Ferrill and Groshong, 1993; Hindle et al., 1999; Lickorish et al., 2002), resulting in a relatively good understanding of the development of upper crustal arcuate systems. Nevertheless, apart from a few important contributions (e.g. Royden, 1993b), the lithospheric control on oroclinal bending has received relatively little attention in the literature. In this chapter I address the role of lithospheric processes in the formation of arcuate belts using the western Mediterranean as a natural laboratory.

Arcuate belts are particularly abundant in the west Pacific Ocean and the Mediterranean Sea, which are areas where subduction systems have been subjected to oceanward retreat (subduction rollback) and to widespread back-arc extension in the overriding plate (e.g. Dewey, 1980; Royden, 1993a). The margins of these basins are often marked by arcuate oroclines, which in extreme cases (e.g. the Banda arc) are bent by up to 180°. In contrast, in areas where subduction zones are relatively fixed, such as in the east Pacific Ocean, the mountain belts involved smaller degrees of oroclinal bending. These observations suggest that rollback processes may account for the formation of many curved orogens throughout the world (e.g. Royden, 1993b; Hall, 2002; Schellart et al., 2002a). The history and the nature of such arcs are associated with complexities and irregularities within the subducting plate.

The aim of this chapter is to discuss the formation of the arcuate belts that surround the western Mediterranean region (Figure 7.1). The study is based on a synthesis of large amounts of spatio-temporal constraints (Rosenbaum et al., 2002b; Rosenbaum and Lister, in review-a) that enable me to develop tectonic reconstructions of the evolution of the orogens. Results of this study suggest that the arcuate shape of orogens in this region is attributed to the destruction of an earlier, more linear Alpine belt, and to the subsequent accretion of segments of that belt during rollback and back-arc extension.

7.2. Tectonic setting

The Alpine belt in the western Mediterranean region extends from Gibraltar to the Adriatic Sea and consists of several different segments of the Alpine orogen: the Rif-Betic belt, the Maghrebides-Apennines, the western Alps and the Pyrenees (Figure 7.1). With the exception of the Pyrenees, they are arcuate fold-and-thrust belts, characterised by frontal thrusting in non-metamorphosed external zones and extension in internal zones (Malinverno and Ryan, 1986; Platt and Vissers, 1989; Crespo-Blanc et al., 1994; Tricart et al., 1994; Avigad et al., 1997; Doglioni et al., 1997; Lonergan and White, 1997). Orogenic processes commenced during the Cretaceous and involved several episodes of high-

pressure/low-temperature (HP/LT) metamorphism at ca. 65 Ma, 45 Ma and 35 Ma (Gebauer et al., 1997; Rubatto et al., 1998; 1999). These events may correspond to collisions between continental ribbons and microplates that existed in the area that separated the converging plates of Africa and Europe (Dercourt et al., 1986; Stampfli et al., 1998). HP/LT rocks are presently exposed in the extensional internal parts of the orogens; they are structurally related to ductile deformation along extensional detachments and to the exhumation of metamorphic core complexes (Jolivet et al., 1990; 1998; Carmignani et al., 1994; Saadallah and Caby, 1996; Avigad et al., 1997; Rosseti et al., 2001).

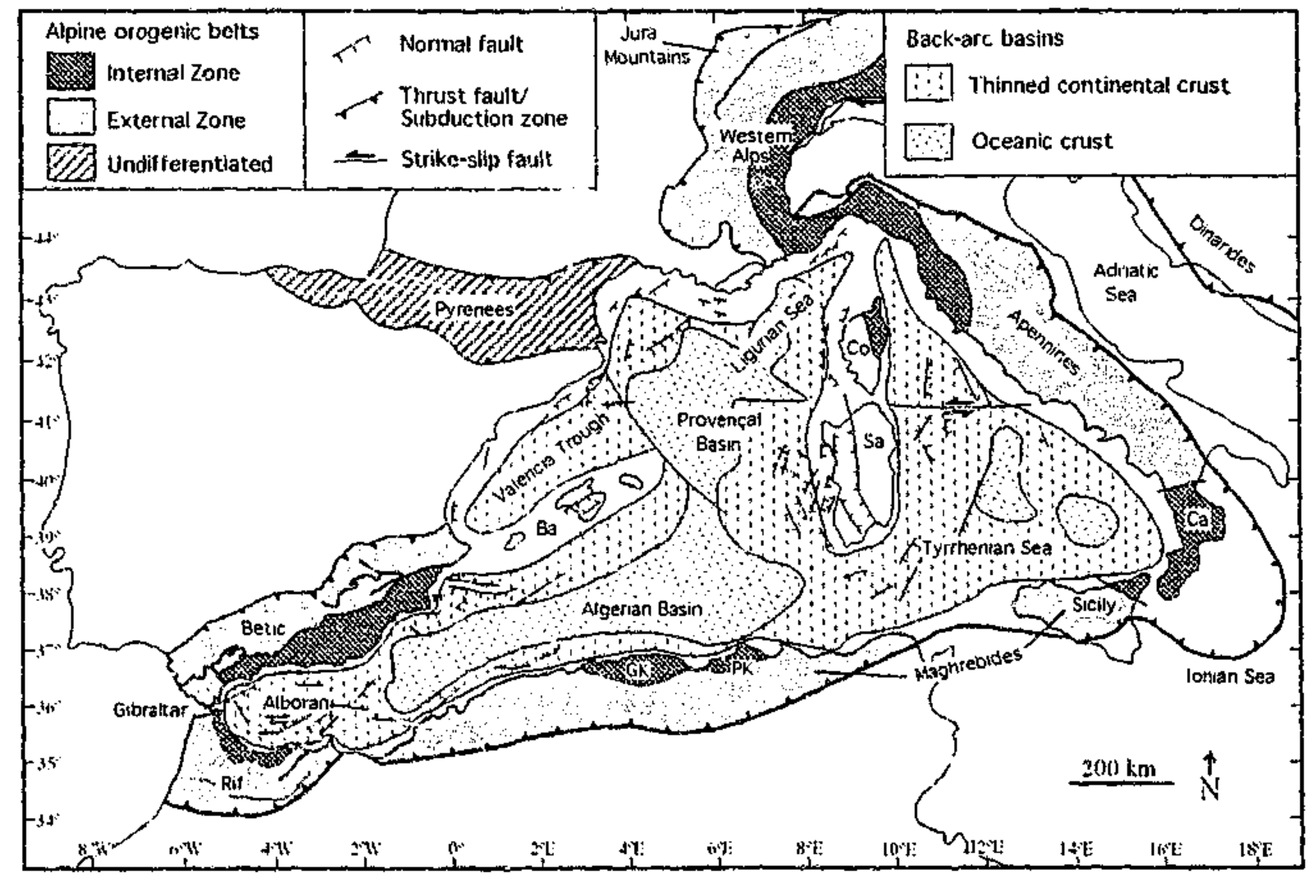


Figure 7.1. Tectonic map of the Alpine orogen in the western Mediterranean region, modified after Carminati et al. (1998) and Coward and Dietrich (1989). Ba, Balearic Islands; Ca, Calabria; Co, Corsica; GK, Grand Kabylie; PK, Petite Kabylie; Sa, Sardinia.

The western Mediterranean region has been subjected to widespread extension since the Oligocene, which occurred simultaneously with the overall convergent motion of Africa with respect to Europe (e.g. Dewey et al., 1989; Jolivet and Faccenna, 2000; Rosenbaum et al., 2002b). Extension gave rise to the formation of several marine basins, some floored by attenuated continental crust (the Valencia Trough, the Alboran Sea and the northern Tyrrhenian Sea) and some floored by Neogene to Recent oceanic crust (the Ligurian-Provençal Basin, the Algerian Basin and the southern Tyrrhenian Sea) (Figure 7.1). The opening of extensional basins in an overall convergent setting has been a matter of numerous studies (e.g. Durand et al., 1999). All the basins have developed since the Oligocene within the overriding European plate, and in the immediate proximity to a northwest-dipping subduction zone. These relationships suggest that the back-arc basins developed as a result of subduction rollback (e.g. Malinverno and Ryan, 1986; Faccenna et al., 1997; Lonergan and White, 1997; Gueguen et al., 1998; Rosenbaum et al., 2002b) combined with relatively slow convergence rates of Africa and Europe since the Oligocene (Jolivet and Faccenna, 2000; Rosenbaum et al., 2002a). In the following sections, I provide further evidence that supports the role of subduction rollback in the tectonic evolution of the arcuate orogenic belts in the western Mediterranean region.

7.3. The Apennines-Maghrebides

The Apennine-Maghrebide belt forms an arcuate orogenic system that follows the margins of the Ligurian-Provençal Basin and the Tyrrhenian Sea (Figure 7.2a). The belt can be traced from the Apennines, through the Calabrian-Peloritan Mountains and the Sicilian Maghrebides, and is linked with the Maghrebides Mountains in Tunisia via a submerged fold-and-thrust belt in the Sardinia Channel (Tricart et al., 1994). In this orogenic system, and particularly in the Apennines, the role of coeval crustal shortening and extensional deformation is reflected by structural, seismic and sedimentological data. Thrust tectonics associated with deformation in the Apenninic fold-and-thrust belt is restricted to the orogenic front, while internal parts of the orogen are continuously destroyed by extensional tectonism.

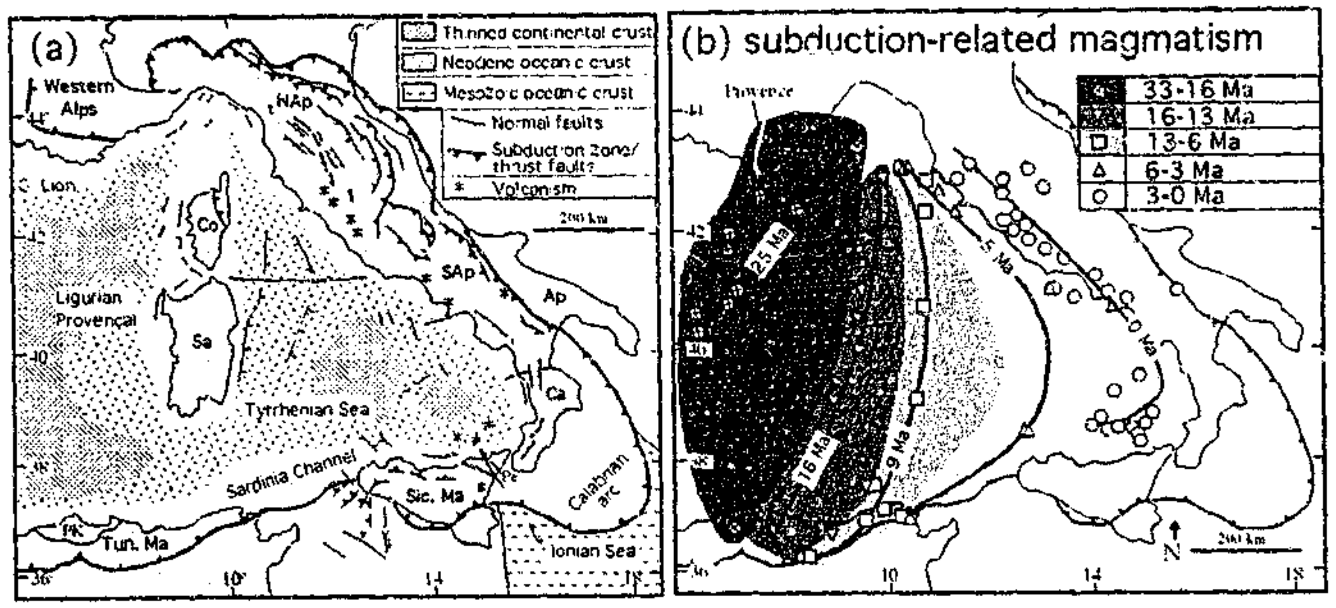


Figure 7.2. (a) Tectonic setting of the Apennine-Maghrebide orogen, modified after Patacca et al. (1993). Ap, Apulian foreland; Ca, Calabria; Co, Corsica; NAp, Northern Apennines; Pe, Peloritan Mountains; PK, Petite Kabylie; Sa, Sardinia; SAp, Southern Apennines; Sic. Ma, Sicilian Maghrebides; Tun. Ma, Tunisian Maghrebides. (b) Spatial distribution of subduction-related magmatism in the Apennine-Tyrrhenian region showing an eastward migration of the subduction zone through time (after Savelli, 2002; Rosenbaum and Lister, in review-a). Locations of magmatic centres that are older than 16 Ma have been rotated 30° counterclockwise with accordance to the rotation of the Corsica-Sardinia microplate (dashed line).

7.3.1. Spatio-temporal constraints

7.3.1.1. Magmatism

There has been intense magmatic activity in the region surrounded by the Apennine-Maghrebide belt since the Late Oligocene, including subduction-related magmatism, intra-plate magmatism and emplacement of MORB-type basalts (Savelli, 2002, and references therein). The spatio-temporal distribution of subduction-related magmatic centres (predominantly calc-alkaline magmas) show progressive younging towards the currently active subduction zones or thrust fronts (Figure 7.2b), associated with increasingly arcuate shape, suggesting that a northwest-dipping subduction zone has rolled-back and has gradually become more arcuate.

Evidence for Oligocene to Middle Miocene (33-16 Ma) subduction processes is documented by occurrences of calc-alkaline magmatism in Provence, the Ligurian Sea, Corsica and Sardinia (Figure 7.2b). Subduction processes took place contemporaneously with the opening of the back-arc Ligurian-Provençal Basin and were accompanied by a counterclockwise rotation of Corsica and Sardinia, which took place between 25-16 Ma. Therefore, magmatic centres in Corsica and Sardinia, which are older than 16 Ma, originated in a more northwesterly position and rotated together with Corsica and Sardinia to their present locations. The 33-16 Ma magmatic arc was probably continuous with contemporaneous calc-alkaline magmatic centres along the present-day Mediterranean coast in France (Figure 7.2b).

Arc-migration as a consequence of subduction rollback can be implied from the spatial distribution of Late Miocene to Recent (13-0 Ma) calc-alkaline magmatic centres. These centres are predominantly found in the Tyrrhenian Sea and the Apennines and show an overall pattern of migration towards the east (Figure 7.2b). The rate of subduction rollback as inferred from the easterly migration of the magmatic arc is in the order of 20-100 km/Myr (assuming that the dip of the subduction zone is constant).

7.3.1.2. Palaeomagnetism

Palaeomagnetic evidence suggests that Corsica and Sardinia underwent approximately 30° of counterclockwise rotation after the Oligocene (<25 Ma) and reached their present position by ~16 Ma (Speranza et al., 2002). In the Apennine-Maghrebide belt, there are numerous palaeomagnetic results that suggest younger (<10 Ma) rotations (e.g. Sagnotti 1992; Scheepers et al., 1993; Scheepers and Langereis, 1994; Muttoni et al., 2000), which are considered to represent the crustal response to the opening of the Tyrrhenian Sea (Rosenbaum and Lister, in review-a).

Palaeomagnetic data suggest that rotations in the Apennines since the Miocene have been dominated by counterclockwise rotations, with the exception of a few clockwise rotations in the northern Apennines (Figure 7.3a). The dominant counterclockwise rotations may correspond to northeastward motion of Apennine units during deformation (Figure 7.3b). The occurrence of clockwise rotations in the northern Apennines seems to be associated with the local arcuate bending of the Umbria-Marche fold-and-thrust belt, which involved both counterclockwise and clockwise rotations at 6-4 Ma (Speranza et al., 1997) (Figure 7.3b).

The pattern of block rotations in the southern part of the Tyrrhenian Sea and the Apennine-Maghrebide belt is characterised by large amounts of Neogene to Quaternary counterclockwise rotations (up to 80°) in the southern Apennines (Scheepers et al., 1993; Scheepers and Langereis, 1994; Gattacceca and Speranza, 2002) and similar amounts of clockwise rotations in Sicily (Speranza et al., 1999; 2003) (Figure 7.3a). In between, the Calabrian block was subjected to small amounts (~15°) of clockwise rotations (Speranza et al., 2000). This symmetry suggests that the southern Apennines and Sicily were subjected to oroclinal bending simultaneously with clockwise rotations in Sicily and counterclockwise rotations in the southern Apennines, with the area in between (the Calabrian block) undergoing smaller degrees of rotations (Figure 7.3c).

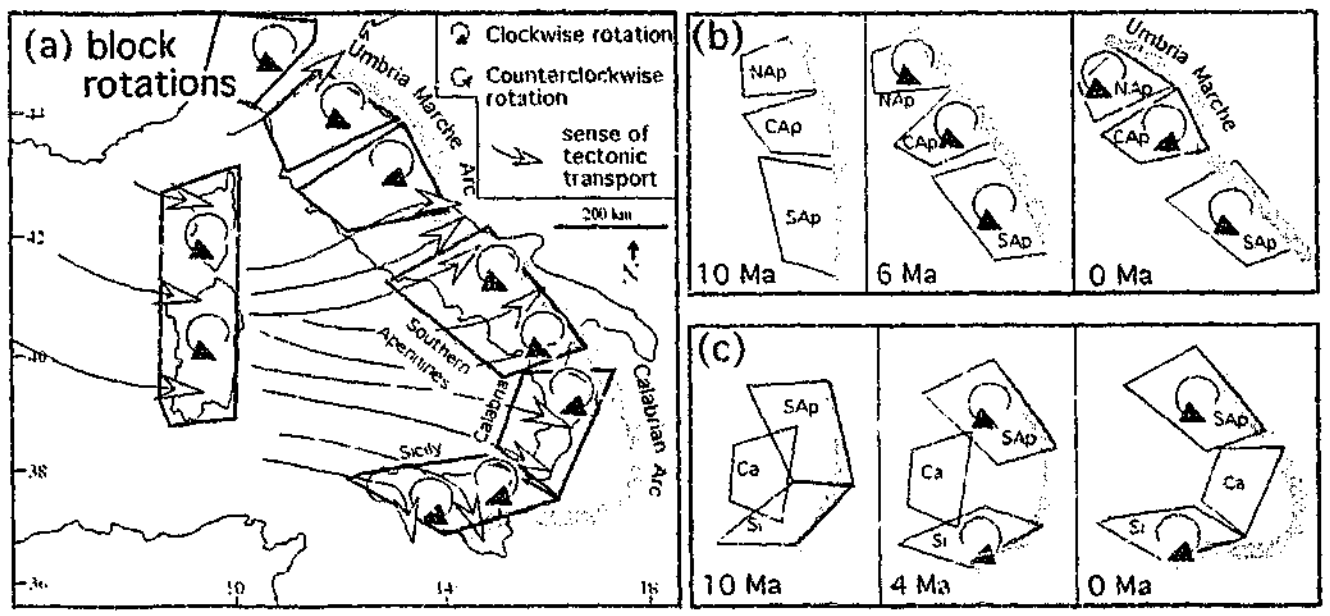


Figure 7.3. (a) Directions of Neogene to Quaternary (25-0 Ma) block rotations as inferred from palaeomagnetic studies (after Rosenbaum and Lister, in review-a, and references therein). (b) Schematic illustration showing the evolutionary pattern of block rotations in the northern Apennines (NAp), central Apennines (CAp) and southern Apennines (SAp). Rotations are predominantly counterclockwise, except of the post-Messinian (~5Ma) clockwise rotations associated with the evolution of the Umbria-Marche arc in the northern Apennines. (c) Schematic illustration showing the evolutionary pattern of block rotations in the southern Apennines (SAp), Calabria (Ca) and Sicily (Si). Note that rotations involved an overall eastward motion accompanied by simultaneous clockwise rotations of Sicily and counterclockwise rotations of the southern Apennines.

7.3.1.3. *Seismicity*

The distribution of deep earthquakes in the southeastern part of the Tyrrhenian Sea defines a narrow Benioff-Wadati Zone (about 200 km long), which is slightly concave upwards and steeply dipping towards the northwest (Anderson and Jackson, 1987b). A well-defined high-velocity anomaly

produced by a cold lithospheric slab is recognised in tomographic images (Lucente et al., 1999; Wortel and Spakman, 2000) (Figure 7.4a,d). These images show that the slab beneath Calabria is subducted down to depth of ~670 km, where it stagnates and lies horizontally parallel to the 670 km discontinuity (Figure 7.4b,d). A relatively cold lithospheric slab, associated with a few occurrences of relatively deep (up to 90 km depth) earthquakes (Selvaggi and Amato, 1992), is also recognised beneath the northern Apennines (Figure 7.4e). Based on the tomographic model of Lucente et al. (1999), it has been suggested that active subduction is currently taking place beneath the northern Apennines (Lucente and Speranza, 2001). In other models, however, tomographic images have been interpreted to indicate a remnant lithospheric slab, which is detached from the surface (Wortel and Spakman, 1992; 2000; van der Meulen et al., 1998). In this respect, the role of active subduction in the northern Apennines remains controversial.

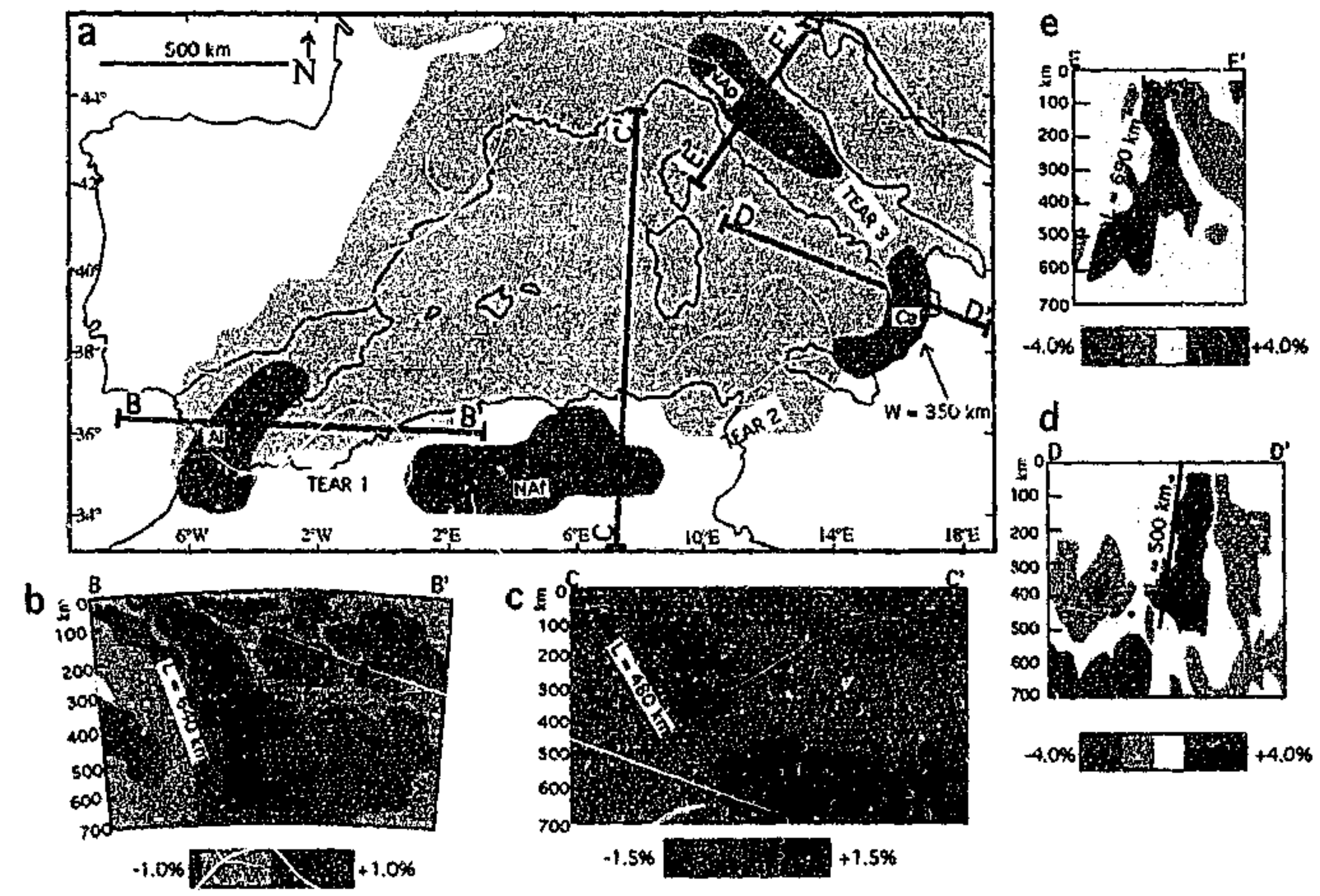


Figure 7.4. Lithospheric structure of the western Mediterranean as inferred from seismic tomography. (a) Map view showing the areas of high-velocity anomalies at depths of 150 km (dark grey) and 600 km (light grey) (modified after Bijwaard and Spakman, 2000; Wortel and Spakman, 2000). Thick lines indicate locations of lithospheric cross sections (b-e). (b) Alboran region, after Gutscher et al. (2002) (original data of Bijwaard and Spakman, 2000); (c) North Africa, after Carminati et al. (1998) (original data of Bijwaard et al., 1998); (d) Calabria, after Lucente et al. (1999) and (e) the northern Apennines, after Lucente et al. (1999) Earthquake hypocentres are indicated by black dots (after Selvaggi and Amato, 1992). Note that different scale is used at different cross sections. Al, Alboran slab; Ca, Calabrian slab; NAf, North African slab; NAp, North Apennine slab.

7.3.1.4. Extensional structures

Structural evidence suggests that both fronts of crustal shortening and extension have migrated through time towards the external parts of the orogen, leading to the simultaneous build up of the fold-and-thrust belt in external zones and its destruction by extension in internal zones (Malinverno and Ryan, 1986). The earliest evidence for extensional tectonics is related to the 32-30 Ma exhumation of metamorphic core complexes in Corsica and Calabria (Jolivet et al., 1990; Brunet et al., 2000; Rosseti et al., 2001). At shallower crustal levels, the onset of rifting is indicated by the occurrence of Late Oligocene (~30-25 Ma) syn-rift deposits in the Gulf of Lion (Séranne, 1999), in the Ligurian Sea (Rollet et al., 2002) and in Corsica Basin (Mauffret and Contrucci, 1999).

Extensional structures become progressively younger towards the east (Jolivet et al., 1998; Brunet et al., 2000; Rosenbaum and Lister, in review-a). Ductile extensional structures related to deformation during the Early-Middle Miocene (25-16 Ma) have been reported from several islands in the northern Tyrrhenian Sea (Rosseti et al., 1999; Brunet et al., 2000) and from the Alpi Apuane metamorphic core complex in the northern Apennines (Carmignani et al., 1994).

The most intense extensional structures in the Tyrrhenian region and the Apennine-Maghrebide belt are related to post-Tortonian (<10 Ma) extensional deformation. Extension is associated with normal faulting and formation of half-grabens, filled by syn-rift sediments that become progressively younger towards the east. Along the Apennines, there is also a southward younging trend in the ages of extensional structures and in the ages of syn-rift deposits (Rosenbaum and Lister, in review-a), suggesting that extension in the southern Tyrrhenian Sea is relatively younger in comparison with the extension in the northern Tyrrhenian Sea.

7.3.2. Reconstruction

A schematic reconstruction of the evolution of the Apennines-Maghrebides based on the above spatio-temporal constraints is presented in Figure 7.5. According to this reconstruction, deformation and progressive curving of the Apennine-Maghrebide belt was directly related to the opening of back-arc basins, i.e., the Ligurian-Provençal Basin and the Tyrrhenian Sea.

The reconstruction shows that during the latest Oligocene (25 Ma), the northern margin of continental Africa and the western margin of Adria were relatively unaffected by orogenic processes (Figure 7.5a). Rather, orogenesis related to a northwest-dipping subduction zone was localised in the European margin, where a NE-SW-striking belt linked internal parts of an earlier (Cretaceous to Oligocene) Alpine orogen: the Alps, Corsica, Calabria, the Kabylies and the Betic-Rif (Figure 7.5a). At the subduction zone, oceanic lithosphere of the Middle-Late Jurassic Liguride Ocean had been subducted, forming a belt of arc-magmatism in Corsica, Sardinia, the Valencia Trough and Provence.

I suggest that sometime between 30-25 Ma, the combination of a gravitational instability at the subduction zone and the relatively slow convergence rates of Africa and Europe (Jolivet and

Faccenna, 2000; Rosenbaum et al., 2002a) triggered the initiation of a southeast-directed subduction rollback. Thus, the overriding plate was subjected to an extensional regime that led to the opening of new back-arc basins (Valencia Trough, the Gulf of Lion and the Ligurian Sea) at the location of the former Alpine orogen. The degree of subduction rollback decreased towards the northeast where it was pinned against the northwest margin of Adria (Figure 7.5a). Therefore, at the same

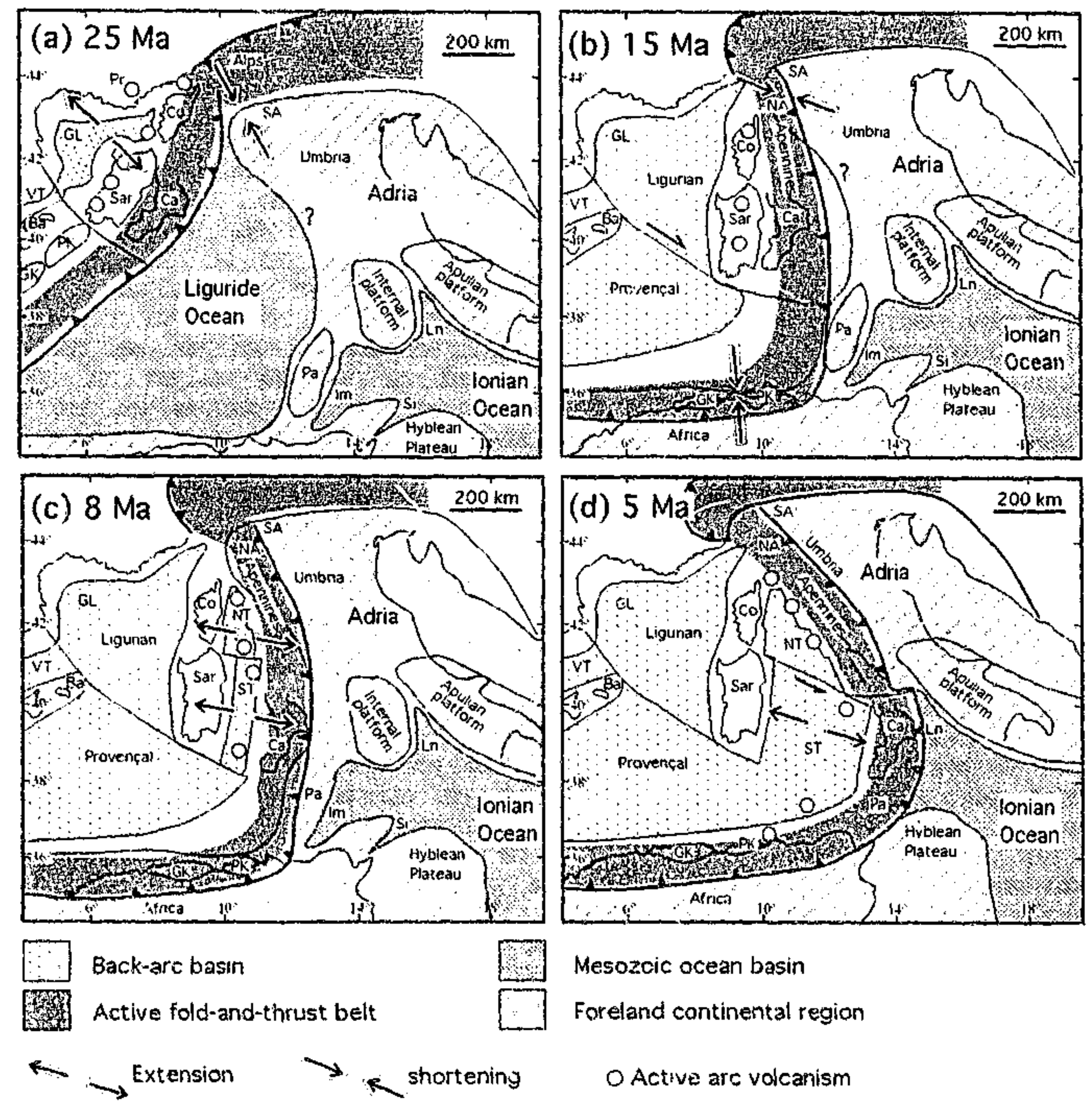


Figure 7.5. Schematic reconstruction of the tectonic evolution of the Apennine-Maghrebide belt (see text for discussion). (a) Early Miocene (25 Ma); (b) Middle Miocene (15 Ma); (c) Late Miocene (8 Ma); (d) Pliocene (5 Ma). Note that the Internal Platform is accreted to the overriding plate between (c) and (d). Ba, Balearic Islands; Ca, Calabria; Co, Corsica; GK, Grand Kabylie; GL, Gulf of Lion; Im, Imerese; Ln, Lagonegro, NA, Northern Apennines; NT, Northern Tyrrhenian; Pa, Panormide platform; PK, Petite Kabylie; Pr, Provence; SA, Southern Alps; Sar, Sardinia; Si, Sicanian; ST, Southern Tyrrhenian, VT, Valencia Trough.

time that extension occurred in the back-arc basins, parts of the northern Apennines were subjected to crustal shortening associated with stacking of nappes in a collisional zone (Figure 7.5a,b). These events mark the initiation of Apennine deformation, which was characterised by crustal shortening associated with the gradual docking of Corsica and Sardinia into the palaeomargins of Adria, and the simultaneous syn-orogenic extension associated with subduction rollback.

During the Middle Miocene (16-15 Ma), Corsica and Sardinia reached their present position and back-arc extension in the Ligurian-Provençal basins ceased. At this stage, most of the oceanic lithosphere of the Liguride Ocean was already consumed (Figure 7.5b). Therefore, continental material of the northern African margin arrived at the subduction zone, causing a disturbance in the system and its rearrangement. This led to the collision of the Kabylies blocks (GP and PK in Figure 7.5) with Northern Africa, and to the emplacement of these allochthonous terranes on top of the palaeomargins of North Africa. This event marks the onset of crustal shortening in North Africa associated with deformation in the Maghrebide belt.

During the period between 16 Ma and 10 Ma, the Apennine-Maghrebide belt was predominantly subjected to crustal shortening as a result of subduction and accretion of continental rocks, whereas extensional processes were relatively quiescent. Back-arc extension resumed during the Tortonian (10-9 Ma), when eastward subduction rollback recommenced, possibly as a result of renewed subduction of dense oceanic crust, leading to the opening of the Tyrrhenian Sea (Figure 7.5c). The reconstruction suggests that the subducting lithosphere consisted of Early Mesozoic ocean basins (e.g. Imerese, Sicanian, Lagonegro) that existed in the westernmost part of the present-day Ionian Sea (Rosenbaum and Lister, in review-a; Rosenbaum et al., in review). During 10-5 Ma, back-arc extension occurred predominantly in the northern Tyrrhenian Sea and in the western part of the southern Tyrrhenian Sea. The latest stage of back-arc extension began at 6-5 Ma and was mainly localised in the southern Tyrrhenian Sea (Figure 7.5d).

7.4. The Rif-Betic

The Betic and the Rif Mountains in southern Spain and northern Morocco form an arcuate orogenic belt that surrounds the Alboran Sea in the westernmost Mediterranean Sea (Figure 7.6). Crustal deformation in the Rif-Betic orogen involved outward migration of thrust nappes, which took place simultaneously with extensional deformation in internal parts of the orogen (e.g. Platt and Vissers, 1989; Crespo-Blanc et al., 1994). The floor of the Alboran Sea is made of internal parts of the orogen, which have been subjected to intense extensional deformation since the Early Miocene (Comas et al., 1992). The Alboran region is often considered as a classic example of a mountain belt that collapsed as a result of post-orogenic extensional tectonism (Dewey, 1988; Platt and Vissers, 1989; Houseman, 1996). The evolution of the Alboran Sea and the Rif-Betic orogen has been explained by several controversial models, which include (1) back-arc extension due to subduction rollback (Royden, 1993b; Lonergan and White, 1997; Rosenbaum et al., 2002b); (2) extension induced by the break-off

of a subducting lithospheric slab (Blanco and Spakman, 1993; Carminati et al., 1998); (3) convective removal of the lithospheric root (Platt and Vissers, 1989; Vissers et al., 1995; Platt et al., 2003); and (4) an asymmetric delamination of a sub-continental lithosphere (Docherty and Banda, 1995).

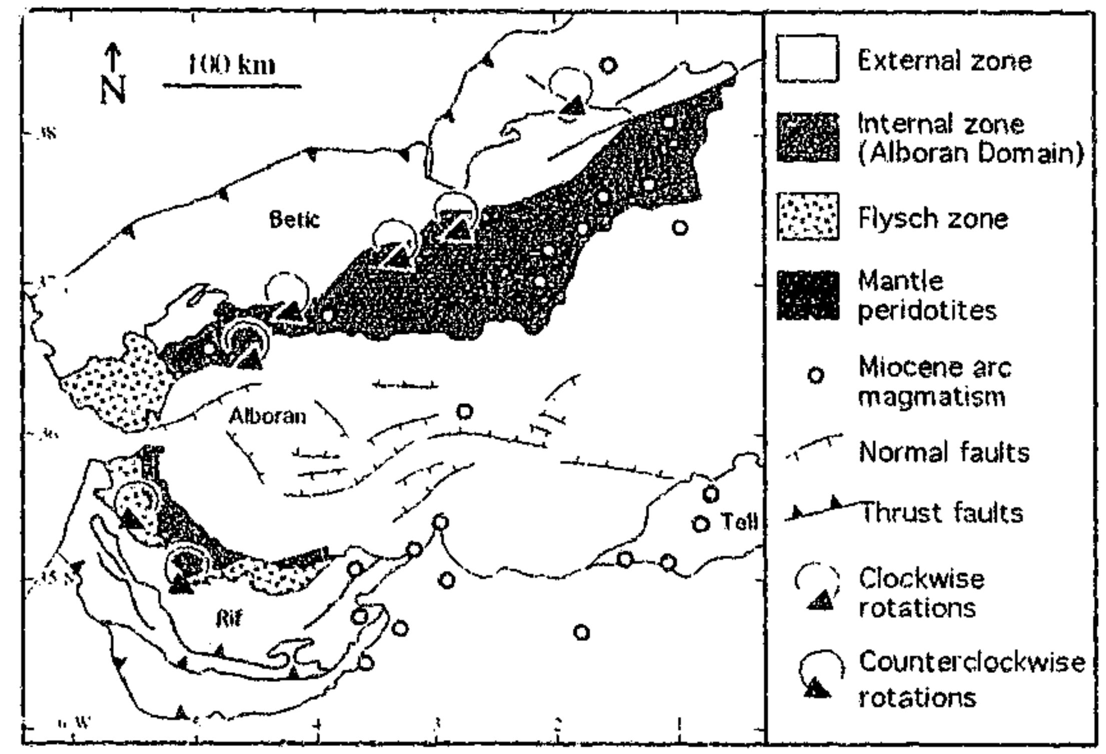


Figure 7.6. Structural setting of the Rif-Betic belt and the Alboran Sea, modified after Platt and Vissers (1989) with locations of Miocene magmatic centres after Savelli (2002). Neogene block rotations around vertical axes as inferred from palaeomagnetic studies are also shown (see text for references).

7.4.1. Spatio-temporal constraints

7.4.1.1. Structural setting

The Rif-Betic orogen consists of three pre-Miocene tectonic domains (Figure 7.6): (1) the palaeomargins of Iberia and Africa; (2) a Flysch Trough underlain by oceanic crust or attenuated continental crust; and (3) the internal 'Alboran Domain' consisting of several strongly deformed and metamorphosed nappe complexes (Crespo-Blanc et al., 1994). The structural and metamorphic signature persevered in the rocks of the Alboran Domain (including the thinned continental basement of the Alboran Sea) suggests a prolonged tectono-metamorphic evolution that included several alternating episodes of crustal shortening and extension (e.g. Crespo-Blanc et al., 1994; Azañón et al., 1997; Balanyá et al., 1997; Azañón and Crespo-Blanc, 2000). These nappe stacks originated in the Alpine belt and have been emplaced on top of the palaeomargins of Africa and Iberia (presently the external zone). The Alboran Domain is interpreted as an allochthonous terrane, which originated in a more easterly position with respect to its present location (Balanyá and García-Dueñas, 1987; Martínez-Martínez and Azañón, 2002).

The boundary between the Alboran Domain and the external parts of the Rif-Betic belt is the

Flysch Trough, which is recognised in the western Betic and in the northern part of Africa (Figure 7.6). It consists of Cretaceous to Miocene deep-water sediments, which were deposited in the basin that separated the allochthonous Alboran Domain from its palaeomargins (Durand-Delga, 1980).

7.4.1.2. Extensional structures

Extension in the Alboran Domain commenced in the Early Miocene and involved activity along flat-lying extensional detachments, which attenuated the thick orogenic pile (e.g. Balanyá and García-Dueñas, 1987; Platt and Vissers, 1989; Jabaloy et al., 1993; Booth-Rea et al., 2002). This extensional episode culminated in a thermal peak at 23-21 Ma (Platt et al., 2003) and led to the exhumation of high-pressure rocks in metamorphic core complexes (e.g. Avigad et al., 1997), as well as to the exhumation of a large volume of diamond-bearing mantle peridotites (Figure 7.6). The orientation of extensional structures related to this deformational stage is predominantly E-W. Early Miocene extensional deformation is also inferred from the occurrence of Aquitanian-Burdigalian (23-16 Ma) syn-rift sediments deposited in half-grabens in the Alboran Sea (Comas et al., 1992). During the Middle Miocene (15-11 Ma), at the same time that back-arc extension temporally stopped in the Apennine-Maghrebide belt, the early extensional structures in the Rif-Betic were superimposed by nearly N-S extensional structures with an overall westward tectonic transport direction (Crespo-Blanc et al., 1994; Booth-Rea et al., 2002; Martínez-Martínez and Azañón, 2002).

7.4.1.3. Palaeomagnetism

Palaeomagnetic data from the Rif-Betic orogen reveal a systematic pattern of block rotations around vertical axes (Figure 7.6). Most of these rotations correspond to crustal deformation during the Miocene, i.e. during the opening of the Alboran Sea (Lonergan and White, 1997). Like the Calabrian arc, the most characteristic kinematic pattern is reflected by opposite directions of Neogene rotations in both sides of the Alboran Sea, with clockwise rotations in the Betic and counterclockwise rotations in the Rif (Platzman, 1992; Allerton et al., 1993; Platzman et al., 1993; 2000). The palaeomagnetic evidence strongly supports the idea that oroclinal bending of the Rif-Betic orogen took place contemporaneously with back-arc extension in the Alboran Sea.

7.4.1.4. Seismicity

Tomographic models in the Alboran region show a pronounced zone of high-seismic velocities at depths of 200-700 km beneath the Alboran Sea and southern Spain (Blanco and Spakman, 1993; Seber et al., 1996; Calvert et al., 2000) (Figure 7.4b). This positive anomaly has been interpreted to indicate an east-dipping subducting slab (Blanco and Spakman, 1993; Gutscher et al., 2002) or a piece of the lithosphere that has been detached and foundered vertically from the lithospheric root (Seber et al., 1996). The existence of lithospheric material at such depths is further supported by few occurrences of deep (ca. 600 km depth) earthquakes beneath southern Spain (Buforn et al., 1991b). In addition, a widespread intermediate-depth seismic activity (60-120 km depth) beneath Gibraltar

(Casado et al., 2001) may correspond to active eastward subduction in this region (Gutscher et al., 2002).

7.4.1.5. Magmatism

Evidence for subduction-related magmatism is widespread in the Alboran region (Figure 7.6). Magmatism took place since the Early Miocene (~22 Ma) and continued until the latest Miocene (~6 Ma) (Lonergan and White, 1997). This magmatic activity suggests that subduction processes took place in the westernmost Mediterranean region during the opening of the Alboran Sea. However, there is no clear evidence with regard to the dip of the subduction zone and the direction of subduction rollback. Moreover, the cessation of subduction-related magmatism at ~6 Ma and the lack of present-day magmatic activity are at odds with evidence for active subduction processes beneath Gibraltar as suggested by Gutscher et al. (2002).

7.4.2. Reconstruction

The proposed reconstruction of the Rif-Betic chain and the Alboran Sea (Figure 7.7) involves westward rollback of an east-dipping subduction zone (Lonergan and White, 1997; Rosenbaum et al., 2002b). I speculate that during the Oligocene (~30 Ma), the internal parts of the Rif-Betic belt (the Alboran Domain) were located near the present-day Balearic Islands (Figure 7.1). During the Oligocene, the Alboran Domain was part of a NE-SW-striking orogenic belt, consisting of the western Alps, Corsica, Calabria and the Kabylies blocks. This belt was located at the overriding plate relative to a northwest-dipping subduction zone (Figure 7.7a). Therefore, the Oligocene-Early Miocene exhumation of metamorphic core complexes in the Rif-Betic may be attributed to a large-scale extensional event associated with earliest stages of rifting in the back-arc region (Figure 7.7b).

The reconstruction shows that extension in the Alboran region could have been produced by the rollback of the Oligocene-Miocene subduction system. As rollback occurred, the hinge of the subduction zone progressively became more arcuate, leading to an overall southwestward direction of tectonic transport in the Alboran region. The earliest stages of subduction rollback were dominated by N-S extension produced as a result of southward and southwestward subduction rollback (Figure 7.7b-d). Extension gave rise to the opening of Valencia Trough and the Algerian Basin, and continued as long as oceanic lithosphere existed between the subduction zone and the continental margin of Africa.

It is speculated that a dramatic change in the direction of rollback occurred at ca. 15 Ma. This change was driven by collision between the subduction zone and the continental margin of Africa, resulting in the cessation of southward rollback and tearing of the subducting slab. However, the existence of an oceanic embayment in the Gibraltar area enabled further subduction rollback towards the west, which led to a change from N-S to E-W extension of the Alboran Domain (Figure 7.7e,f). This extensional phase led to the westward tectonic transport of the allochthonous terranes of the internal

Rif and Betic, accompanied by counterclockwise rotations in the south and clockwise rotations in the north. Subsequently, the Rif and Betic were emplaced and accreted into the palaeomargins of Iberia and Africa, resulting in contemporaneous extension in the Alboran Sea and crustal shortening in the arcuate orogenic belt (Figure 7.7f).

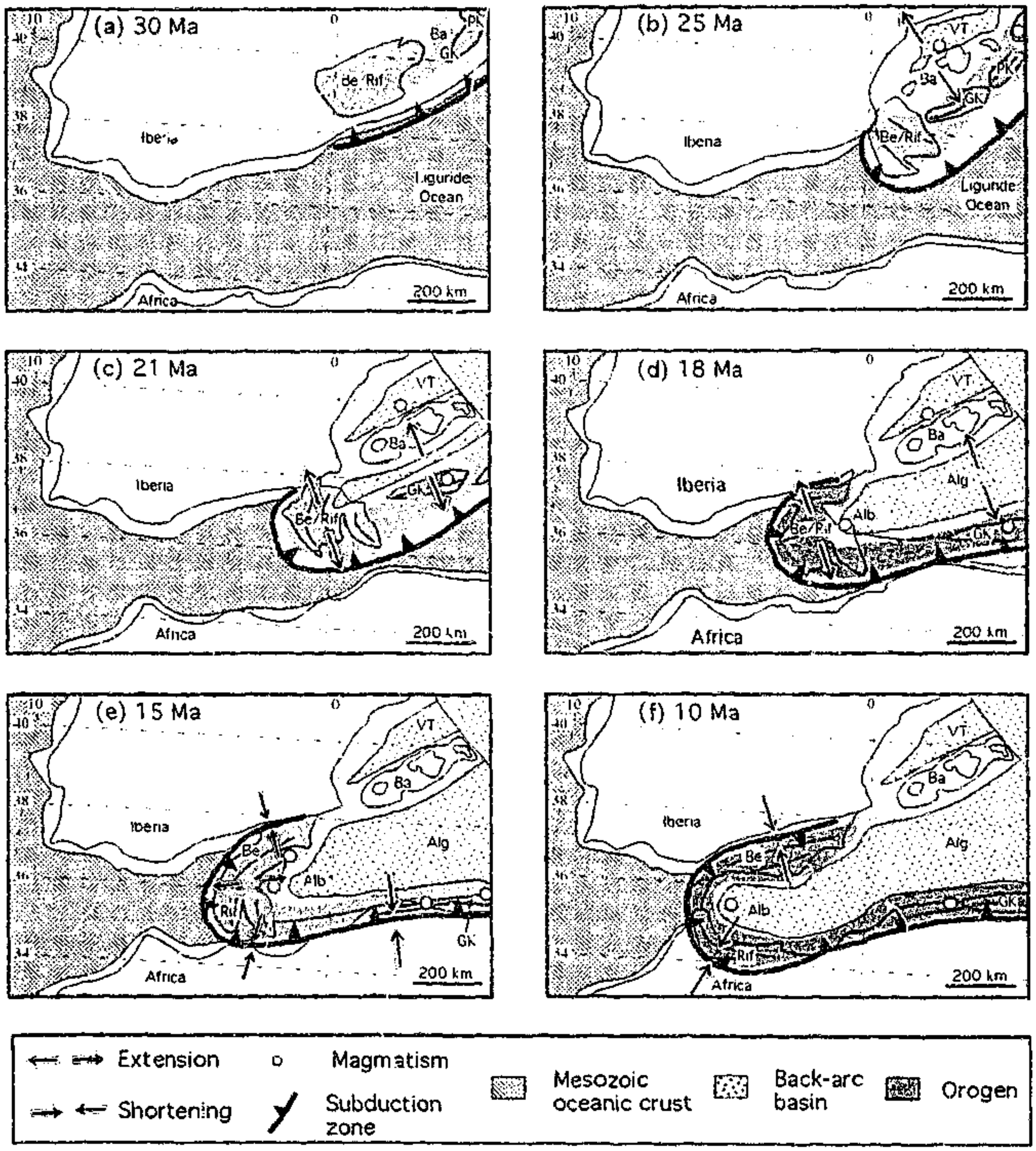


Figure 7.7. Schematic reconstruction of the evolution of the Rif-Betic orogenic belt (see text for discussion). Alb, Alboran Sea; Alg, Algerian Basin; Ba, Balearic Islands; Be, Betic; GK, Grand Kabylie; PK, Petite Kabylie; VT, Valencia Trough.

7.5. Discussion

7.5.1. Lithospheric control on the evolution of Mediterranean-type arcuate belts

The reconstruction of the tectonic evolution of the western Mediterranean since the Oligocene (Figure 7.5 and Figure 7.7) suggests that lithospheric processes associated with rollback of subduction zones were the main driving mechanism controlling the development of arcuate belts in the region. Since the Oligocene, the subducting lithosphere in the western Mediterranean region evolved from a relatively linear NE-SW-striking system located south of the European margin, to several narrow slab segments (separated by tears) presently found beneath the Alboran region, North Africa and Italy (Figure 7.4). In this process, the whole subduction system has progressively migrated backward, presumably, as a result of a gravitational instability induced by the negative buoyancy of the lithospheric slab with respect to the surrounding asthenosphere (Elsasser, 1971; Molnar and Atwater, 1978; Dewey, 1980; Garfunkel et al., 1986). This mechanism, when combined with relatively slow convergence rates, may have led to the opening of back-arc basins due to lithospheric extension at the edge of the overriding plate (Royden, 1993b; Jolivet and Faccenna, 2000).

The existence of remnant slabs that are still subducting in the western Mediterranean has been interpreted from tomographic models, which show regions of relatively fast seismic wave velocities that often coincide with deep seismic activity (e.g. Wortel and Spakman, 1992; 2000; Bijwaard et al., 1998; Lucente et al., 1999; Bijwaard and Spakman, 2000; Figure 7.4). The tomographic images suggest that segments of subducting slabs exist beneath the Apennines, Calabria, North Africa and the Alboran Sea. These slab segments are separated from each other by regions of relatively slow seismic wave velocities where the slab is apparently torn. The deepest slab tears are indicated by the absence of high-velocity anomalies down to depths of ~600 km. Such tears are found between the Alboran slab and the North African slab (Tear 1 in Figure 7.4), and between the North African slab and the Calabrian slab (Tear 2 in Figure 7.4). A shallower tear, which is characterised by the absence of a lithospheric slab down to depths of ~250 km, is found in the central Apennines, between the Calabrian and the northern Apennine slabs (Tear 3 in Figure 7.4) (Lucente and Speranza, 2001).

The remnant slabs that are presently recognised beneath the western Mediterranean are fragments of oceanic lithosphere consumed since the Oligocene (Figure 7.8). A large volume of this oceanic lithosphere was probably derived from the Liguride Ocean, which was a Middle-Late Jurassic oceanic domain separating Iberia and Adria (Rosenbaum et al., 2002a). Based on the reconstruction, I estimate that the area of the 'Liguride-Ocean lithosphere' subducted since the Oligocene was approximately 775000 km^2 (Figure 7.8).

In the eastern parts of the western Mediterranean subduction system (the Apennine and Calabrian arcs) there was also subduction of an oceanic or sub-continental lithospheric domain, known as the 'Adriatic/Ionian lithosphere' (e.g. Gvirtzman and Nur, 2001) (Figure 7.8). This domain supposedly originated in the westernmost parts of the Early Mesozoic Ionian Sea (Rosenbaum and Lister, in review-a; Rosenbaum et al., in review). I estimate that an area of ~370000 km^2 of the Adriatic/Ionian lithosphere has been consumed since the Oligocene (Figure 7.8).

We can now show that the total surface area of the remnant lithospheric slabs is sufficient to account for the whole area consumed since the Oligocene in the western and central Mediterranean (i.e. the Liguride-Ocean lithosphere and the Adriatic/Ionian lithosphere) (Table 7.1). The surface area of a slab is estimated by multiplying the width (W) of the high-velocity anomaly when projected on the earth's surface with the length (L) of the anomaly recognised in tomographic cross sections (Figure 7.4). It is assumed that the currently subducting lithospheric slabs have not been significantly stretched. Note that I have calculated only remnant slabs that are currently subducting (i.e. dipping beneath the overriding plate or sub-vertical). The surface area of these slabs is sufficient to account for nearly 100% of the total area of the Oligocene oceans (Table 7.1). The tomographic images show an additional lithospheric slab lying horizontally parallel to the 670 km discontinuity (Figure 7.4c,d). However, since all the surface area of consumed Oligocene lithosphere is traced in remnant subducting slabs, the stagnated horizontal slab must have been subducted prior to the Oligocene.

The above comparison suggests that mass balance considerations may be used to predict how the lithosphere is deformed during the evolution of subduction rollback. Deep tears occurred where the retreating subduction zone collided with the northern margin of Africa resulting in the incorporation of continental crust in the subduction system. However, in the Gibraltar and the Calabrian arcs, the existence of narrow oceanic passages enabled further rollback in a direction that was roughly perpendicular to the direction of convergence. These changes in rollback directions could have only occurred after tearing of the slabs. During subduction rollback, these tears accommodated strike-slip motions, leading to an overall 'extrusion' of the Alboran to the west and Calabria to the east.

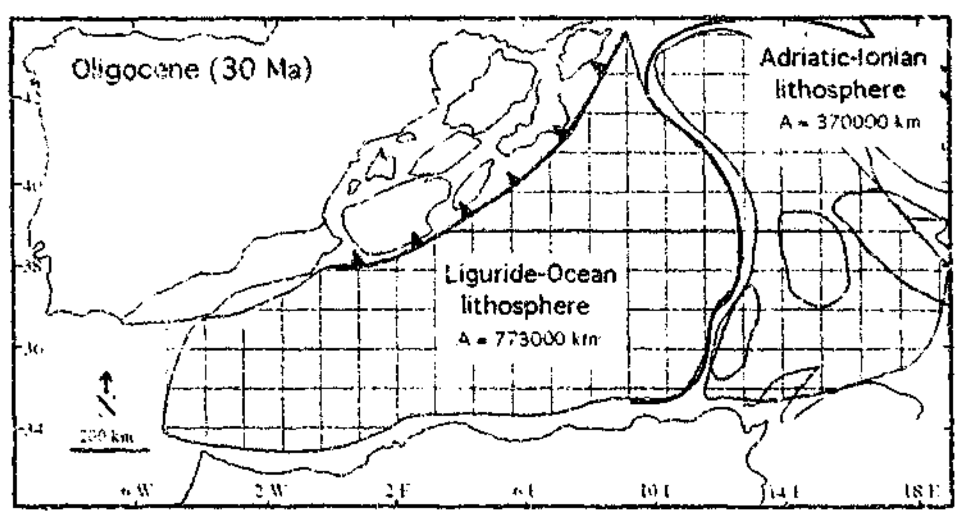


Figure 7.8. Reconstruction of the western Mediterranean region at 30 Ma showing the area covered by Oligocene oceanic domains. This is equal to the surface area which is observed in remnant slab segments (Table 7.1), suggesting that formation of slab tears can be predicted from mass balance considerations. Thick line marks the boundary of the Adriatic domain.

Table 7.1. Comparison between predicted areas that have been consumed since the Oligocene with observed surface areas of remnant slabs.

Reconstruction		**Tomography**			
	Inferred area of subducting lithosphere		Width (km)	Length (km)	Surface areas of slabs (km^2)
Liguride domain	775,000	**Alboran**	450	640	288,000
		North Africa	690	480	331,200
Adriatic-Ionian domain	370,000	**Calabria**	350	500	175,000
		Northern Apennines	500	650	325,000
Total	**1,145,000 km^2**	Total			**1,119,200 km^2**

7.5.2. Crustal response to subduction rollback

All the allochthonous terranes in western Mediterranean (Calabria, Alpine Corsica, Kabylies and the internal Rif-Betic) originated from internal parts of an earlier (Cretaceous to Oligocene) NE-SW-striking orogenic belt that was located at the overriding plate. These allochthonous terranes show different tectonic and stratigraphic evolution relative to the continents onto which they are emplaced, but share a comparable pre-Oligocene tectono-metamorphic evolution (see next section). The early orogenic belt collapsed at the first stage of back-arc extension (at 32-30 Ma), and was subsequently dispersed in an arcuate fashion towards the directions of subduction rollback. Migration of these allochthonous terranes could have proceeded as long as subduction rollback and back-arc extension continued.

Eventually, at different stages during the Miocene, the western Mediterranean subduction systems collided with the palaeomargins of Africa, Iberia and Adria. This collision led to the incorporation of continental crust at the subduction system, which decreased the negative buoyancy of the subducting slab and terminated further subduction rollback and back arc extension. Thus, the allochthonous terranes were accreted into the continental palaeomargins forming several segments of arcuate mountain belts.

From the above discussion it appears that the arcuate belts in the western Mediterranean region are the crustal manifestation of the lithospheric processes associated with subduction rollback and tearing of subducting slabs. The tomographic data show that lithospheric slabs underwent a small degree of lithospheric bending (e.g. in the Calabrian arc and the Alboran arc; Figure 7.4a), which corresponds to the overall curvature of the fold-and-thrust belt. However, the degree of bending is much more prominent in the orogenic belts, as reflected by occurrences of tight arcuate orogens such as the Rif-Betic belt (Figure 7.6). The difference between the geometry of arcuate structures in crustal and lithospheric scales is probably related to the suggestion that bending of lithospheric slabs results in formation of tears, whereas crustal blocks can be progressively emplaced in an arcuate fashion.

In the western Mediterranean region, the rate of subduction rollback was greater than the rate of convergence (Jolivet and Faccenna, 2000; Rosenbaum et al., 2002b); therefore, the edge of the overriding plate was subjected to an overall extensional regime that resulted in the opening of back-arc basins. This style of tectonic activity involved a significant degree of horizontal motions that explain how continental fragments (or 'allochthonous terranes') could have travelled great distances before accreting into the adjacent palaeomargins. Subduction rollback can thus provide a mechanism for transportation and emplacement of allochthonous terranes of the kind discussed by Ben-Avraham et al. (1981) and Nur and Ben-Avraham (1982).

7.5.3. Origin and exhumation of the HP/LT belt

Constraints on the evolution of the western Mediterranean arcuate belts provide interesting outcomes on the history of burial and exhumation of HP/LT rocks. HP/LT assemblages are presently exposed in metamorphic core complexes within internal zones of the orogenic belts (Figure 7.9a). These rocks were subjected to burial at depths that exceeded 40 km, or even 100 km in extreme cases (in the western Alps; e.g. Chopin, 1984), probably as a result of crustal thickening associated with collisional tectonics. The ages of HP/LT metamorphism in the western Mediterranean and the western Alps are predominantly Early-Middle Tertiary, with possibly three groups of ages that may represent large-scale orogenic episodes at ~65 Ma, ~45 Ma and ~35 Ma (Gebauer et al., 1997; Rubatto et al., 1998; 1999). None of the HP/LT assemblages formed during the development of the arcuate belts, that is, since ~25 Ma. I therefore conclude that the post-Oligocene (<25 Ma) evolution of the western Mediterranean involved an extensional destruction of an earlier HP/LT belt. During this period HP/LT terranes have drifted over great distances until reaching their present locations in internal zones of the arcuate orogens.

The above conclusion is of fundamental importance to plate tectonic reconstructions, because it implies that HP/LT belts do not necessarily mark the locus of continental collisions of major tectonic plates. For example, many authors have regarded the Rif-Betic chain as a collisional suture between Iberia and Africa that was later subjected to post-orogenic extensional collapse (e.g. Dewey, 1988; Platt and Vissers, 1989; Houseman, 1996). I suggest, in contrast, that Rif-Betic belt originated in a HP/LT belt that had been located in a more northeasterly position and had formed a contiguous belt with the HP/LT rocks of the western Alps, Corsica and Calabria. It was not until the initiation of subduction rollback and back-arc extension that fragments of the original orogenic belt have started to drift as allochthonous terranes towards the directions of subduction rollback, forming the present-day arcuate geometry of the orogen.

Structural observations from HP/LT terranes within the Alpine belt suggest that the HP/LT rocks have been subjected to intense extensional deformation, which probably played a fundamental role in their exhumation processes (e.g. Jolivet et al., 1990; 1994; Avigad, 1992; Avigad et al., 1997; Azañón et al., 1998; Rosseti et al., 2001). Extension and vertical lithospheric thinning could have commenced when the rocks were still buried at relatively deep crustal levels (Rosenbaum et al., 2002c), shortly after the culmination of HP/LT metamorphism (Figure 7.9b). In the western Mediterranean, the earliest rifting stage took place at 32-30 Ma and led to the exhumation of the Cretaceous to Oligocene HP/LT rocks in the soles of metamorphic core complexes in Calabria, Corsica and the Betic (Jolivet et al., 1990; Azañón et al., 1998; Rosseti et al., 2001). This exhumation stage occurred 5-7 Myr before the commencement of sea-floor spreading in the Ligurian-Provençal basins as indicated by rotations of Corsica, Sardinia and the Balearic Islands (Rosenbaum et al., 2002b). This lag may indicate the transitional period between rifting and sea-floor spreading in the back-arc region.

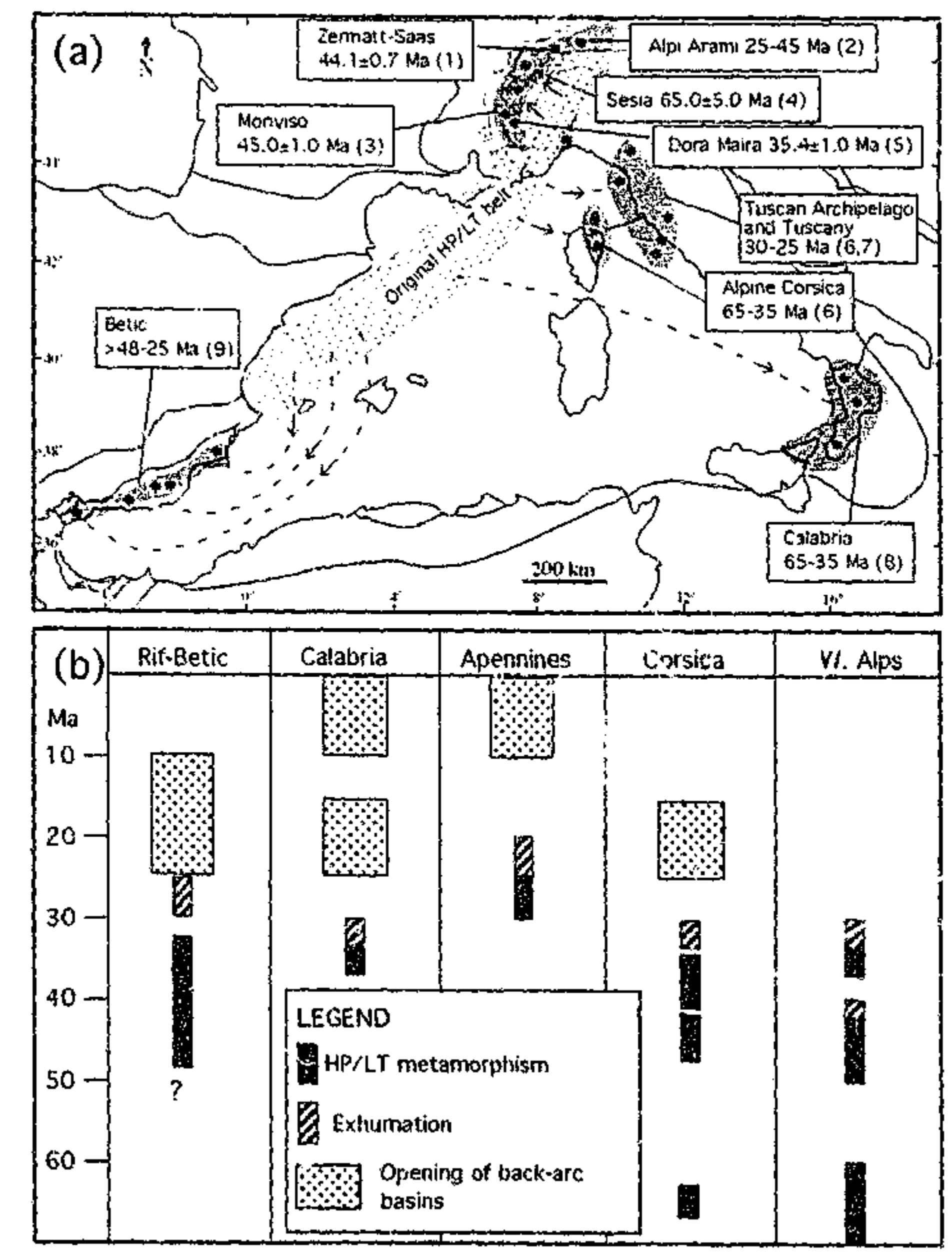


Figure 7.9. (a) The location of the original HP/LT belt in the western Mediterranean and the western Alps (dotted area) and the present-day distribution of HP/LT mineral assemblages (shaded areas and stars). Selected radiometric ages are also shown. Dashed lines and arrows indicate directions of tectonic transport. References are: (1) Rubatto et al. (1998); (2) Gebauer (1996); (3) Rubatto and Hermann (2003); (4) Rubatto et al. (1999); (5) Gebauer et al. (1997); (6) Brunet et al. (2000); (7) Kligfield et al. (1986); (8) Rosseti et al. (2001); (9) Monié et al. (1991). (b) A summary of HP/LT metamorphic ages, exhumation ages and periods of back-arc extension (see text for discussion).

7.6. Conclusions

This chapter emphasised the lithospheric control on the formation of arcuate orogenic belts in the western Mediterranean. I have shown that the geodynamic evolution of the western Mediterranean region involved large degrees of subduction rollback and back-arc extension that were accompanied by lithospheric bending and formations of slab tears. In this tectonic framework, the arcuate belts are the crustal manifestation of these lithospheric processes, but with surface expressions that are associated with 180° arcs and radiuses of ~200 km. In contrast, the degree of bending at depth is much smaller. The orogenic belts consist of fragments of allochthonous terranes that have been accreted into the palaeomargins of adjacent continents during the final stages of subduction rollback. Accordingly, the present exposure of HP/LT rocks does not necessarily correspond to the original locus of HP/LT metamorphism and the collision of major tectonic plates.

CHAPTER 8

TECTONIC EVOLUTION OF THE WESTERN ALPS FROM THE JURASSIC TO THE OLIGOCENE

Getting closer in a driven and harmonious movement, following the meridian, the ancient boundaries of Eurasia and Indu-Africa were melded in the geosynclinal domain, limited by the Tethys. The formation of almost a plastic flow that ended up by breaking into a double train of chains, nappes and folds against the jaws of a vice, therefore designed to behave like a mountain belt.

Emile Argand, Sur l'arc des Alpes Occidentales (1916)
(Translated from French by Jerome Ganne and Cecile Duboz)

FOREWORD: CHAPTER 8

This chapter presents a compilation of published data related to the geology of the western Alps and discusses a number of proposed reconstruction models for the tectonic evolution of the region in the period between the Jurassic and the Oligocene (160-35 Ma). I wish to thank Gordon Lister, Gerard Stampfli, Hans Laubscher, Marco Beltrando, Jerome Ganne, Marnie Forster, David Giles and John Clulow for stimulating discussions on this topic. I also wish to acknowledge Jean-Paul Cadet and Laurent Jolivet for giving me a copy of the "Atlas Peri-Tethys" (Dercourt et al. 2000), which has been a crucial reference in this chapter. The manuscript has benefited from critical comments by Peter Betts, Gordon Lister, Caroline Forbes, Ivo Vos and Mike Hall.

8. TECTONIC EVOLUTION OF THE WESTERN ALPS FROM THE JURASSIC TO THE OLIGOCENE

Abstract

This chapter presents a review of spatio-temporal constraints and of reconstruction models for the tectonic evolution of the western Alps. Despite extensive research in the last 150 years, the regional tectonic reconstruction of this part of the Alpine-Himalayan orogen has remained controversial. The orogen consists of several ribbon-like continental terranes (Sesia/Austroalpine, Internal Crystalline Massifs, Briançonnais), which are separated by two or more ophiolitic sutures (Piedmont, Valais, Antrona?, Lanzo/Canavese?). High-pressure metamorphism of each terrane occurred during distinct orogenic episodes: at ~65 Ma in the Sesia/Austroalpine, at ~45 Ma in the Piedmont zone and at ~35 Ma in the Internal Crystalline Massifs. It is suggested that these events reflect individual accretionary episodes, which together with kinematic indicators and plate motions, provide constraints for the discussed reconstruction models. The models involve a prolonged orogenic history that took place during relative convergence of Europe and Adria (here considered as a promontory of the African plate). The first accretionary event involved the Sesia/Austroalpine terrane. Final closure of the Piedmont Ocean occurred during the Eocene (~45 Ma) and involved ultra-high-pressure (UHP) metamorphism of the Piedmont oceanic crust. The incorporation of the Briançonnais terrane in the accretionary wedge occurred thereafter, possibly during a Late Eocene (45-35 Ma) subduction of the Valais Ocean or the Antrona Ocean. The final closure of these oceans may account for the collisional episode, at ca. 35 Ma, in the Internal Crystalline Massifs.

8.1. Introduction

8.1.1. Scope and objectives

The arcuate belt of the western Alps is one of the most studied orogens in the world and is commonly used as a classic example of a collisional belt (e.g. Argand, 1916; 1924; Trümpy, 1960; Hsü, 1989). Yet its complex structure has never been fully understood in the context of plate kinematics. This shortcoming is largely attributed to the difficulty in applying the plate tectonic theory in continental processes, leading to oversimplified kinematic models or non-kinematic descriptive models.

The aim of this chapter is to provide a review of spatio-temporal constraints (geochronology, distribution of sutures, plate motions and kinematic indicators) and to develop plate-scale reconstructions for the evolution of the western Alps from the Jurassic to the Oligocene. I present alternative reconstruction models for different time frames during this evolution. Alpine research is based on numerous studies, and it is not in the scope of this chapter to provide an exhaustive coverage of the literature. I will focus on recently published results that are applicable for Alpine reconstruction, and I will provide a comparative discussion of reconstruction models, emphasizing the timing of subduction and collisional processes and the number of continental blocks and ocean basins involved in Alpine orogeny. Due to the complexity of the western Alps, it is unlikely that the validity of a specific reconstruction model will ever be proved beyond any doubt. However, I aim to present multiple reconstruction hypotheses, and then to evaluate whether these hypotheses are in agreement with the geological record.

8.1.2. Historical background

Mobilistic concepts with regard to the evolution of the Alps were first published in the second half of the 19th century (Heim, 1878; Bertrand, 1884; Schardt, 1893; see also reviews by Trümpy (1996) and Dal Piaz (2001a)). These ideas came from the interpretation of nappes and large thrust systems, which implied that crustal fragments had been transported over great distances before their final emplacement on the orogen. The recognition of nappe tectonics led to a generation of Alpine geologists that eagerly argued for the role of global-scale horizontal motions, long before the advent of the plate tectonic theory (e.g. Suess, 1875; Wegener, 1912; 1922; Argand, 1916; 1924).

Nonetheless, Alpine research did not play a significant role in the development of the plate tectonic theory (Dal Piaz, 2001a). The foundations for plate tectonics emerged from a plethora of ocean-floor geophysical data, which have been made available since the 1940's (see review by Oreskes and Le Grand, 2001). Plate tectonics has explained sea-floor spreading, subduction processes and continental drift, but it has not provided sufficient information to resolve kinematic complexities that occur in orogenic belts. In this respect, it is perhaps ironic that the kinematics of the Alps – the place where mobilistic ideas were born - has hitherto not been explained by a plate kinematic approach.

There have been numerous attempts to use plate kinematic constraints to reconstruct the evolution of the Alpine orogen (e.g. Smith, 1971; Dewey et al., 1973; Biju-Duval et al., 1977). This approach has been implemented in the western Alps by using the combination of kinematic criteria derived from geological structures together with information concerning the motions of the surrounding plates (e.g. Baird and Dewey, 1986; Platt et al., 1989a). The results of these studies have been inconclusive because of the difficulty in constraining the timing of deformation and the uncertainties concerning the motion of Adria (either a microplate, or an African promontory consisting of the Southern Alps, the Italian peninsula and the Adriatic Sea; e.g. Channell et al., 1979; Dercourt et al., 1986). Such works, together with other notable reconstructions by Laubscher (1988; 1991; 1996), Dercourt et al. (1986; 1993; 2000), Stampfli and Merchant (1997) and Stampfli et al. (1998) provided

a framework for plate tectonic reconstructions of the western Alps, because they incorporated a vast amount of geological information into reconstruction models. However, these models require continual refinements based on new additional spatio-temporal constraints.

Perhaps the most significant progress in Alpine tectonics in recent years has been brought from the field of geochronology, with the recognition of Middle Tertiary high-pressure (HP) and ultra-high-pressure (UHP) metamorphism (Tilton et al., 1989; 1991; Gebauer et al., 1992; 1997; Gebauer, 1996; Duchêne et al., 1997; Rubatto and Gebauer, 1999; Rubatto and Hermann, 2001). These ages dramatically changed the traditional view on the timing of Alpine tectonics, which was earlier considered to be associated with Late Cretaceous HP metamorphism followed by Middle Tertiary Barrovian metamorphism (e.g. Hunziker et al., 1989). Currently, many geologists agree that a major HP metamorphic event occurred in the western Alps during the Eocene/Oligocene (in addition to a Late Cretaceous HP event in the Austroalpine terrane; e.g. Michard et al., 1996; Stampfli et al., 1998; Gebauer, 1999; Dal Piaz, 2001b; Froitzheim, 2001). The controversy with regard to the consequences of these geochronological interpretations are further discussed in this chapter.

8.2. Tectonic setting

The western Alps form an arcuate orogenic belt, striking E-W in central Switzerland, NE-SW to N-S along the Italian-French border and NW-SE in the Maritime Alps (Figure 8.1a). The orogen is divided into three domains, namely the Adriatic domain, the Internal Zone ('Penninic Alps') and the External Zone. A geological cross-section across the western Alps and a space-time chart are presented in Figure 8.1b and Figure 8.2, respectively, and a brief description of each domain is given below.

8.2.1. The Adriatic domain

The Adriatic domain in the western Alps consists of granulite-facies pre-Alpine gneisses of the Ivrea zone, a thick carbonate sequence of the Southern Alps and a clastic sedimentary infill of the Po Basin (Figure 8.1a). These rock units have not been subjected to Alpine metamorphism (Frey et al., 1999) and underwent relatively little Alpine deformation. In the central Alps, the non-metamorphosed Southern Alps are separated from the Penninic Alps by a distinct mid-crustal fault zone, the Insubric Line, which was active as a dextral strike-slip fault during the Oligocene and the Miocene (~32-16 Ma; Laubscher, 1983; Schmid et al., 1989; Steck and Hunziker, 1994). In the northwest Alps, the contact between the Ivrea Zone and the Penninic Alps is marked by the Canavese zone, which is a narrow belt of strongly deformed crystalline basement rocks and Mesozoic sediments with similar sedimentary facies as the rocks of the Southern Alps (Zingg et al., 1976). In the western Alps, the nature of the boundary between the Adriatic domain and the Penninic Alps is not clear because it is covered by Late Tertiary and Quaternary sediments of the Po Basin.

(a)

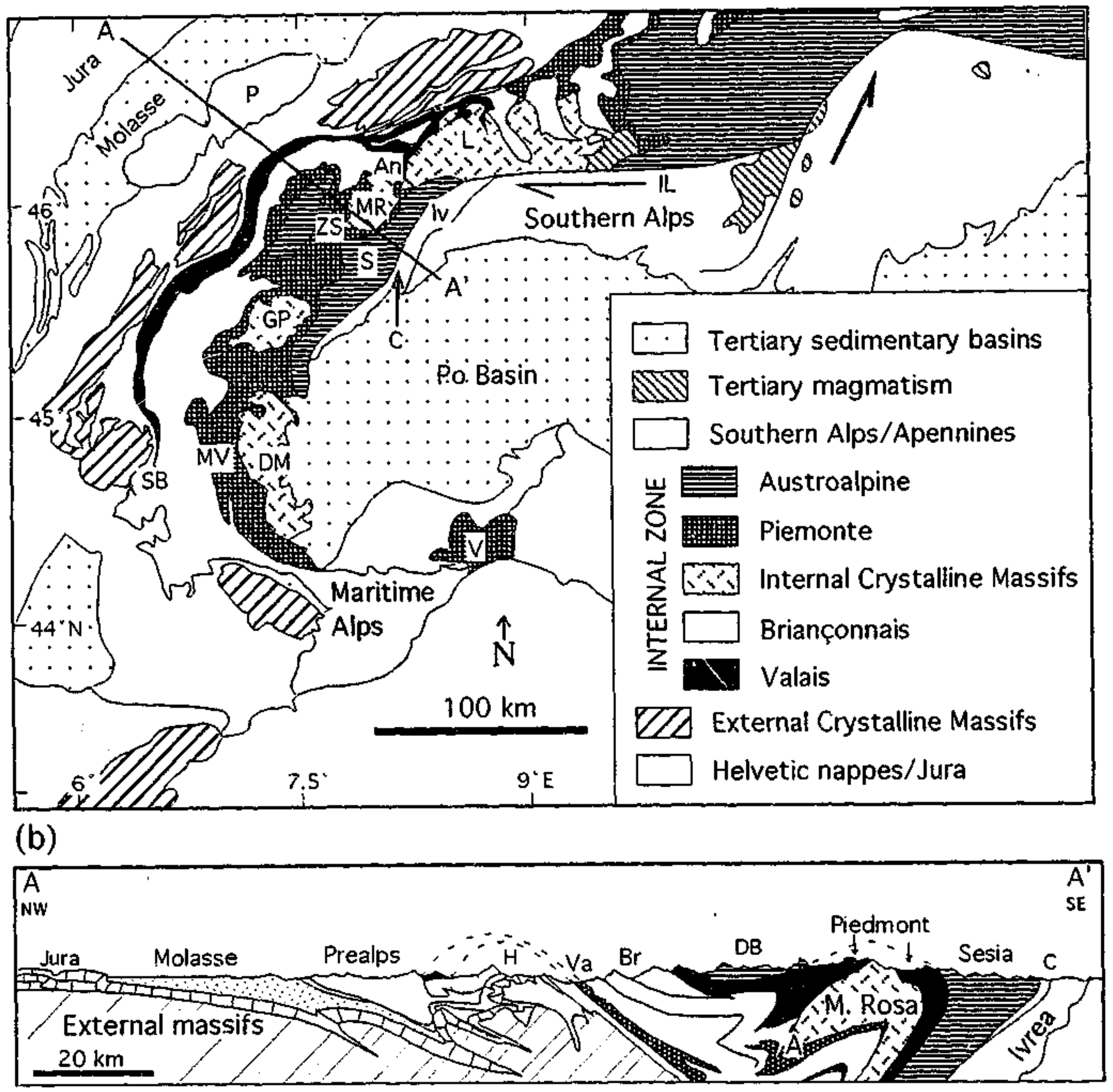


Figure 8.1. (a) Map of the western Alps showing the main tectonic units, modified after Polino et al. (1990), Hsü (1994) and Dal Piaz (1999). An, Antrona; C, Canavese; DM, Dora Maira; GP, Grand Paradiso; IL, Insubric Line; Iv, Ivrea; L, Lepontine; MR, Monte Rosa; MV, Monviso; P, Prealps; S, Sesia; SB, Sub-Briançonnais; V, Voltri; ZS, Zermatt-Saas. (b) Cross section across the western Alps (see Figure 8.1a for location) modified after Escher et al. (1997) and Froitzheim (2001). DB, Dent Blanche; H, Helvetic nappes; Va, Valais.

8.2.2. Internal Zone

The Internal Zone consists of a series of continental terranes and ophiolites, which are aligned as arcuate ribbons parallel to the strike of the orogen (Figure 8.1a). These are, from southeast to northwest, the Austroalpine, the Piedmont zone, the Internal Crystalline Massifs, the Briançonnais zone, and the Valais zone (Figure 8.1b).

The Austroalpine is located in the highest structural position of the Penninic Alps. It is considered to represent the northern margin of Adria because it has a similar sedimentary facies as the rocks of the Southern Alps (Trümpy, 1980). In the western Alps, the Austroalpine is represented by the Sesia-Lanzo zone and by a number of klippen (Dent Blanche, Mt Emilius, Margna; Dal Piaz, 1999). Rocks of the Sesia-Lanzo zone were subjected to HP metamorphism during the latest Cretaceous (70-65 Ma; Duchêne et al., 1997; Rubatto et al., 1999).

The Piedmont zone comprises of metamorphosed ophiolitic rocks derived from a Middle-Late Jurassic ocean (Rubatto et al., 1998; Rubatto and Hermann, 2003), and folded Jurassic to Late Cretaceous calc-schists (Schistes Lustrés) that represent the sedimentary cover of the Piedmont Ocean. The Piedmont Ocean existed on the southeastern flank of the Briançonnais terrane. Coesite bearing eclogites that indicate burial at UHP conditions have been found at Lago di Cignana in northwest Italy (Reinecke, 1991). The timing of HP and UHP metamorphism in these rocks, as well as in HP eclogites from the Monviso ophiolite (Figure 8.1a), have been dated at 49-44 Ma (Duchêne et al., 1997; Rubatto et al., 1998; Rubatto and Hermann, 2003).

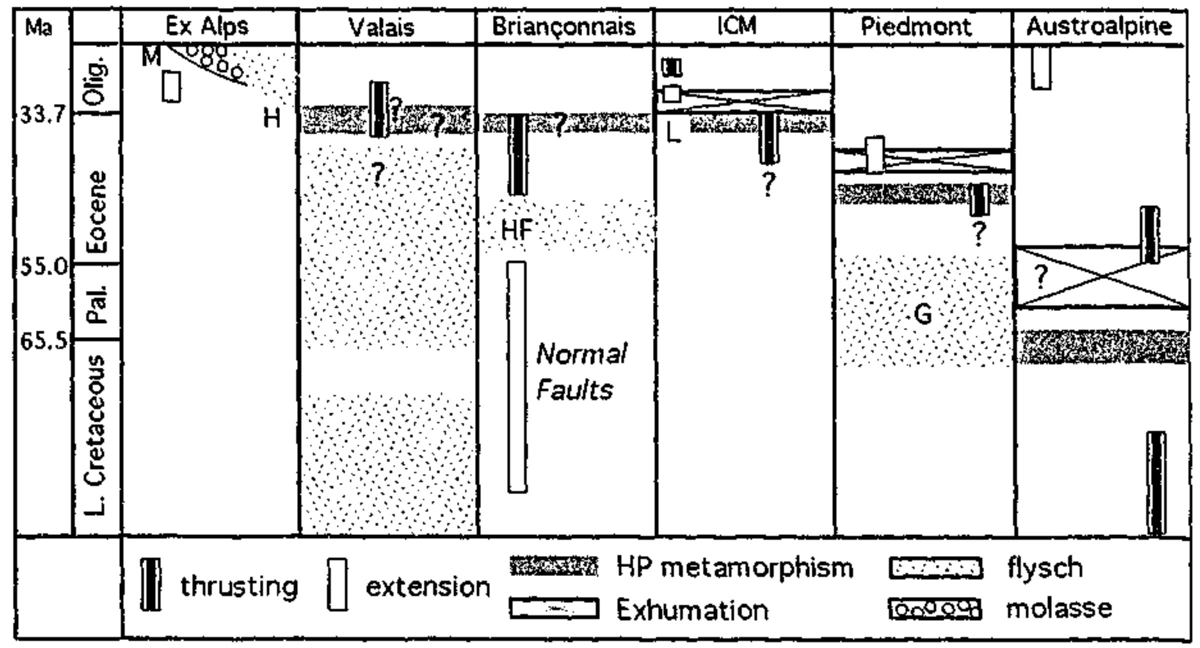


Figure 8.2. Space-time diagram showing important tectonic events in the western Alps during the Cenozoic. Data complied from Trümpy (1980), Schmid et al. (1997); Stampfli et al. (1998); Bistacchi et al. (2001); Michard and Martinotti (2002). For references to the metamorphic ages see Figure 8.3. G, Gurnigel flysch; H, Helvetic nappes; HF, Helmithoid flysch; L, Lepontine; M, Molasse; Time scale is after Remane et al. (2000).

The Internal Crystalline Massifs (Dora Maira, Gran Paradiso and Monte Rosa) consist of pre-Triassic basement and minor Late Palaeozoic – Early Mesozoic cover metasediments. These rocks underwent HP and UHP metamorphism during Alpine orogeny, probably in Late Eocene – Early Oligocene (~35 Ma; Duchêne et al., 1997; Gebauer et al., 1997; Gebauer, 1999; Rubatto and Gebauer, 1999; Rubatto and Hermann, 2001). The rocks are presently exposed in tectonic windows within the Piedmont domain (Figure 8.1). The origin of the Internal Crystalline Massifs is controversial, with a proposed European origin (e.g. Avigad et al., 1993; Froitzheim, 2001), an Adriatic origin (e.g. Polino et al., 1990; Stampfli et al., 1998) or an independent terrane separated both from Europe and Adria (e.g. Platt, 1986) (Figure 8.4). In the Dora Maira massif, the presence of coesite pseudomorphs in metasedimentary rocks suggests metamorphism at depths of more than 100 km (Chopin, 1984).

The Briançonnais zone consists of a pre-Mesozoic crystalline basement overlain by Permo-Carboniferous continental sediments and a Mesozoic cover (Platt et al., 1989b). The origin of this continental domain is a matter of debate; some authors have suggested that the Briançonnais zone originated in the passive margin of Europe (e.g. Lemoine et al., 1986), whereas others (e.g. Stampfli, 1993) have suggested an origin in the Iberian microplate and a later incorporation as an exotic terrane in the accretionary prism of the western Alps.

The most external domain in the Penninic Alps is the Valais zone (Figure 8.1). It consists of flysch units and minor ophiolitic rocks, which underwent HP metamorphism during Alpine orogeny (Oberhänsli, 1994; Cannic et al., 1996; Bousquet et al., 2002). Towards the southern part of the western Alps, the Valais domain is missing, but is possibly represented in the rocks of the Sub-Briançonnais (Bertrand et al., 1996). The Sub-Briançonnais and the Valais domains lie on the northwestern flank of the Briançonnais terrane.

8.2.3. External Zone

The External Alps consist of an autochthonous and parautochthonous crystalline basement that originated in the Hercynian orogen (External Crystalline Massifs) and a Mesozoic to Cenozoic sedimentary cover (Helvetic/Dauphinois nappes) (Trümpy, 1980). The External Alps underwent intense Alpine deformation characterised by folding of basement nappes and thin-skinned thrusting of the Helvetic cover nappes (Escher and Beaumont, 1997; Figure 8.1b). The cover nappes are detached from their underlying basement by weak layers of Triassic evaporites (Hsü, 1994). In the internal parts of the basement massifs, Alpine metamorphism reached greenschist and amphibolite facies conditions (Frey et al., 1999).

Crustal shortening in the External Alps commenced in the Late Oligocene - Early Miocene and has propagated towards the northwest. The latest thrusting occurred in Late Miocene and Pliocene (~10-5 Ma) in the arcuate chain of the Jura Mountains (Trümpy, 1980; Figure 8.2). Since the Oligocene (~30 Ma), detrital formations derived from the rising of the Alps have been deposited in the molasse basins of northwest Switzerland (Figure 8.1). The more internal parts of these basins have undergone Neogene crustal shortening as the orogenic front migrated northward (Trümpy, 1980).

8.3. Spatio-temporal constraints

8.3.1. Age of Alpine high-pressure metamorphism

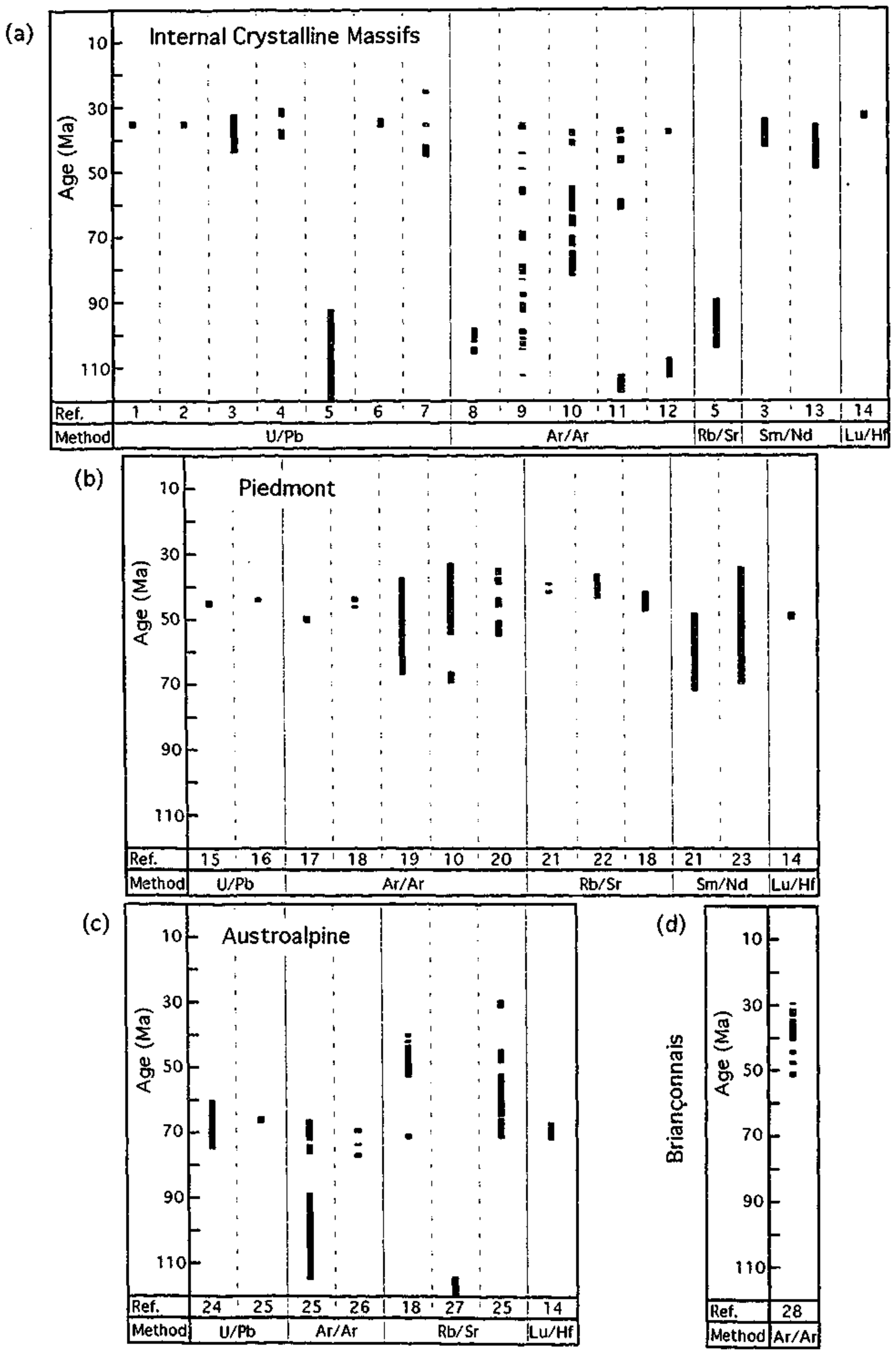
High-pressure metamorphism is considered to take place during subduction or collisional processes and its age may indicate the timing of accretionary events (e.g. Lister et al., 2001). In the western Alps, interpretations of metamorphic ages are a matter of debate, resulting in a number of different reconstruction models (compare, for example, Avigad et al. (1993) with Froitzheim (2001)). One of the major controversies is associated with the significance of Cretaceous (Eoalpine) HP metamorphism - an event that is considered by some authors to be prominent in the whole Internal Alps (e.g. Agard et al., 2002), whereas other authors suggest that it affected only the Austroalpine terrane (Gebauer, 1999; Dal Piaz, 2001b; Froitzheim, 2001). The differences between these contrasting views are significant because they considerably influence reconstruction scenarios.

A compilation of HP metamorphic ages from the western Alps is presented in Figure 8.3 and in Table 8.1. The interpretation of some of these results as peak HP ages remains controversial (see, for example, discussions by Gebauer (1999) and Dal Piaz (2001b)). Nevertheless, the comparison of results derived from five different dating techniques enables us to recognise a number of geochronologically distinct episodes of metamorphism discussed below.

8.3.1.1. Internal Crystalline Massifs

Figure 8.3a shows that the majority of ages obtained from the Internal Crystalline Massifs are Late Eocene –Early Oligocene ages (37-32 Ma). These results have been obtained using four different independent methods (U/Pb, Ar/Ar, Sm/Nd and Lu/Hf), and are therefore considered as relatively robust constraints for the occurrence of HP and UHP metamorphism at ca. 35 Ma.

Significantly older ages (120-40 Ma) have been obtained using ^{40}Ar-^{39}Ar dating techniques (Figure 8.3a), with most studies interpreting their results as Late Cretaceous HP metamorphic ages and younger (Eocene) cooling ages (e.g. Scaillet et al., 1992). More recent studies, however, have shown that some of the dated minerals (particularly from the UHP rocks of the Dora Maira massif) contained excess radiogenic ^{40}Ar products, which could have led to erroneously older ages (Arnaud and Kelley, 1995; Scaillet, 1996). Interestingly, Paquette et al. (1989) obtained Cretaceous ages for the UHP rocks of the Dora Maira massif using U-Pb and Rb-Sr methods. The results of these independent methods indicate that older HP ages cannot be attributed solely to the presence of excess argon in the rocks. Moreover, geological studies in the Internal Crystalline Massifs (e.g. Philippot, 1990) have revealed a long and complex evolution of fabrics and microstructures, which suggest a prolonged tectono-metamorphic evolution of these terranes. Therefore, the debate concerning the occurrence of Cretaceous HP metamorphism in the Internal Crystalline Massifs is as yet inconclusive and cannot be easily refuted.



8.3.1.2. Piedmont

Radiometric dating of HP rocks from the Piedmont domain indicates that metamorphism occurred during the Middle Eocene (50-42 Ma; Figure 8.3b). The most precise ages are obtained by U-Pb SHRIMP zircon dating (Gebauer, 1999), which have yielded ages of 44-45 Ma in both Zermatt-Saas and Monviso ophiolites (Rubatto et al., 1998; Rubatto and Hermann, 2003). Slightly younger ages (42-39 Ma) have been obtained by Rb-Sr techniques (Cliff et al., 1998; Dal Piaz et al., 2001). The latter may indicate the time of recrystallisation in shear zones during exhumation, which is similar to greenschist facies ages obtained by Reddy et al. (1999; 2003), Amato et al. (1999) and Cartwright and Barnicoat (2002) in the Zermatt-Saas area.

8.3.1.3. Austroalpine

The Sesia-Lanzo zone is considered to be an Austroalpine unit (e.g. Dal Piaz, 1999, and references therein) and is the only terrane in the western Alps that shows clear evidence for Late Cretaceous (75-65 Ma) HP metamorphism (Ruffet et al., 1995; Inger et al., 1996; Rubatto et al., 1999; Dal Piaz et al., 2001; Figure 8.3c). Considerably younger ages (48-40 Ma) have been obtained from other lower Austroalpine klippen (Mt Emilius, Glacier Rafray and Etirol Levaz; Dal Piaz et al., 2001), suggesting a possible incorporation of these klippen in the Piedmont accretionary wedge.

8.3.1.4. Briançonnais

The age of HP metamorphism in the Briançonnais zone is poorly constrained because eclogite and blueschist facies rocks are often overprinted by greenschist facies assemblages, which obliterate dateable HP assemblages (Platt et al., 1989b). Greenschist facies assemblages have yielded Oligocene ages (35-27 Ma; Freeman et al., 1997; Freeman et al., 1998). Late Eocene – Early Oligocene ages (40-30 Ma) obtained by Markely et al. (1998) have been interpreted as HP ages (Figure 8.3d).

Figure 8.3. Compilation of geochronological results associated with the age of HP metamorphism in the (a) Internal Crystalline Massifs; (b) the Piedmont zone; (c) the western Austroalpine terrane; and (d) the Briançonnais terrane. References are: (1) Gebauer et al. (1997); (2) Rubatto and Hermann (2001); (3) Tilton et al. (1989); (4) Tilton et al. (1991); (5) Paquette et al. (1989); (6) Rubatto and Gebauer (1999); (7) Gebauer (1996); (8) Monié and Chopin (1991); (9) Scaillet et al. (1992); (10) Chopin and Maluski (1980); (11) Monié (1985); (12) Chopin and Monié (1984); (13) Becker (1993); (14) Duchêne et al. (1997); (15) Rubatto and Hermann (2003); (16) Rubatto et al. (1998); (17) Monié and Phillipot (1989); (18) Dal Piaz et al. (2001); (19) Barnicoat et al. (1995); (20) Agard et al. (2002); (21) Cliff et al. (1998); (22) Amato et al. (1999); (23) Bowtell et al. (1994); (24) Rubatto et al. (1999); (25) Igner et al. (1996); (26) Ruffet et al. (1995); (27) Oberhänsli et al. (1985); (28) Markley et al. (1998).

Table. 8.1. Inferred ages of high-pressure metamorphism in the western Alps

Massif	Rock type	Method	Age (Ma)	Reference
Internal Crystalline Massifs				
Dora Maira	Eclogite	Lu-Hf (whole rock-garnet)	32.8±1.2	Duchêne et al. (1997)
Dora Maira	Pyrop quartzite	U-Pb (SHRIMP-zircon)	35.4±1.0	Gebauer et al. (1997)
Dora Maira	Calc-silicate eclogite	U-Pb (SHRIMP-titanite)	35.1±0.9	Rubatto & Hermann (2001)
Dora Maira	Pyrop quartzite	U-Pb (zircons)	38.0±6.0	Tilton et al. (1989)
Dora Maira	Pyrop quartzite	U-Pb (zircons)	38.0±1.4	Tilton et al. (1991)
Dora Maira	Pyrop quartzite	U-Pb (ellenbergerite)	31.4±1.3	Tilton et al. (1991)
Dora Maira	Pyrop quartzite	U-Pb (zircon)	121+12/-29	Paquette et al. (1989)
Dora Maira	Pyrop quartzite	Sm-Nd (garnet)	38.0±4.5	Tilton et al. (1989)
Dora Maira	Pyrop quartzite	Rb-Sr (phengite)	96.0±4.0	Paquette et al. (1989)
Dora Maira	Pyrop quartzite	Ar-Ar (phengite)	104.5±1.2 100±2.3	Monié & Chopin (1991)
Dora Maira	Fine grained gneiss	Ar-Ar (phengite)	34.5±0.12	Scaillet et al. (1992)
Dora Maira	Leucocratic orthogneiss	Ar-Ar (phengite)	44.2±0.21 35.0±0.16 36.8±0.14	Scaillet et al. (1992)
Dora Maira	coarse grained gneiss	Ar-Ar (phengite)	48.9±0.2	Scaillet et al. (1992)
Dora Maira	Metapelite	Ar-Ar (phengite)	79.0±0.15 68.8±0.12 48.9±0.16	Scaillet et al. (1992)
Dora Maira	Mg-rich micaschist	Ar-Ar (phengite)	104.2±0.24 112.3±0.14	Scaillet et al. (1992)
Dora Maira	Fine grained gneiss	Ar-Ar (phengite)	36.2±0.81	Scaillet et al. (1992)
Dora Maira	Metapelite	Ar-Ar (phengite)	87.5±0.76 80.5±0.96 82.9±0.14 91.5±1.47; 79.0±0.88	Scaillet et al. (1992)
Dora Maira	Mg-rich micaschist	Ar-Ar (phengite)	69.5±1.55 55.7±1.28	Scaillet et al. (1992)
Dora Maira	Fine grained gneiss	Ar-Ar (phengite)	101.0±0.42 102.7±0.56	Scaillet et al. (1992)
Dora Maira	Metapelite	Ar-Ar (phengite)	90.9±0.91 98.9±0.9	Scaillet et al. (1992)
Dora Maira	Mg-rich micaschist	Ar-Ar (phengite)	102.6±0.77	Scaillet et al. (1992)
G. Paradiso (cover)	phengite marble	Ar-Ar (phengite)	57.4±3.2	Chopin & Maluski (1980)
G. Paradiso (basement)	metagranite	Ar-Ar (phengite)	56.4±1.5	Chopin & Maluski (1980)
G. Paradiso (cover)	dolomitic marble	Ar-Ar (phengite)	37.8±1.0	Chopin & Maluski (1980)
G. Paradiso (basement)	garnet-para. glaucophanite	Ar-Ar (phengite)	78.0±4.0	Chopin & Maluski (1980)
G. Paradiso (basement))	gneiss	Ar-Ar (phengite)	41.0±1.1	Chopin & Maluski (1980)
G. Paradiso (basement)	micaschist	Ar-Ar (phengite)	71.1±1.7	Chopin & Maluski (1980)
G. Paradiso (cover)	dolomitic marble	Ar-Ar (phengite)	58.5±3.7	Chopin & Maluski (1980)
G. Paradiso (basement)	micaous glaucophanite	Ar-Ar (pheng. + paragonite)	64.9±1.8	Chopin & Maluski (1980)
G. Paradiso (basement)	q.-pheng. glaucophanite	Ar-Ar (phengite)	60.0±1.9	Chopin & Maluski (1980)
Monte Rosa	phengite metaquartzite	U-Pb (SHRIMP-zircon)	34.9±1.4	Rubatto & Gebauer (1999)
Monte Rosa	HP metapelite	Rb-Sr (whole rock)	102.0±2.0	Paquette et al. (1989)
Monte Rosa	HP metapelite	Rb-Sr (phengite)	91.0±2.0	Paquette et al. (1989)
Monte Rosa - Furgg zone	pelitic micaschist	Ar-Ar (phengite)	114.8±2.9	Monié (1985)
Monte Rosa - Furgg zone	paragneiss	Ar-Ar (phlogopite)	60.0±1.7	Monié (1985)
Monte Rosa - cover	orthogneiss	Ar-Ar (phengite)	40.2±1.0	Monié (1985)
Monte Rosa - basement	granite	Ar-Ar (phengite)	46.1±1.2	Monié (1985)
Monte Rosa - basement	quarzite	Ar-Ar (phengite)	37.3±0.9	Monié (1985)

Table. 8.1. Continued.

Massif	Rock type	Method	Age (Ma)	Reference
Monte Rosa	talc-ph.-chloritoid schist	Ar-Ar (phengite)	110.0±3.0	Chopin & Monie (1984)
Monte Rosa	phengite quartzite	Ar-Ar (phengite)	37.7±0.9	Chopin & Monie (1984)
Lepontine - Alpi Arami	Garnet-pyroxenite	U-Pb (SHRIMP-zircon)	35.4±0.5	Gebauer (1996)
Lepontine –Alpi Arami	Garnet-pyroxenite	U-Pb (SHRIMP-zircon)	43.0±2.0	Gebauer (1996)
Lepontine - Alpi Arami	Garnet-pyroxenite	U-Pb (SHRIMP-zircon)	25.1±0.6	Gebauer (1996)
Lepontine –Alpi Arami	Garnet lherzolite	Sm-Nd (garnet-WR-Cpx)	39.1±3.5	Becker (1993)
Lepontine –Alpi Arami	Garnet lherzolite	Sm-Nd (garnet-WR-Cpx)	39.6±3.8	Becker (1993)
Lepontine –Alpi Arami	Garnet clynopyroxenite	Sm-Nd (garnet-WR-Cpx)	43.9±5.7	Becker (1993)
Lepontine –Alpi Arami	Eclogire	Sm-Nd (garnet-WR-Cpx)	37.5±2.2	Becker (1993)
Lepontine –Alpi Arami	Peridotite	Sm-Nd (garnet-WR-Cpx)	42.3±2.9	Becker (1993)
Lepontine –Alpi Arami	Garnet lherzolite	Sm-Nd (garnet-WR-Cpx)	40.2±4.2	Becker (1993)
Piedmont				
Monviso	Eclogite	Lu-Hf (whole rock-garnet)	49.1±1.2	Duchêne et al. (1997)
Monviso	Eclogite	U-Pb (SHRIMP-zircon)	45.0±1.0	Rubatto & Hermann (2003)
Monviso	Mylonitic eclogite	Sm-Nd (granet-Cpx)	60.0±12	Cliff et al. (1998)
Monviso	Chlorite eclogite	Sm-Nd (granet-Cpx)	62.0±9	Cliff et al. (1998)
Monviso	Mylonitic eclogite	Rb-Sr (phengite-Cpx)	41.6±0.4 39.2±0.4	Cliff et al. (1998)
Monviso	Eclogite	Ar-Ar (phengite)	50.0±1.0	Monié & Phillipot (1989)
Zermatt-Saas	Eclogite	U-Pb (SHRIMP-zircon)	44.1±0.7	Rubatto et al. (1998)
Zermatt-Saas		Sm-Nd (garnet)	52.0±18	Bowtell et al. (1994)
Zermatt-Saas	Coesite bearing eclogite	Sm-Nd (garnet-WR-Cpx)	40.6±2.6	Amato et al. (1999)
Zermatt-Saas	Eclogite	Rb-Sr (5 points isochron)	40.0±3.9	Amato et al. (1999)
Zermatt-Saas	Phengite quartzite	Rb-Sr (WR-pheng. isochron)	45.0±2.8	Dal Piaz et al. (2001)
Zermatt-Saas	Mn quartzite	Rb-Sr (WR-pheng. isochron)	45.0±0.5	Dal Piaz et al. (2001)
Zermatt-Saas	Phengite pyrop quartzite	Rb-Sr (WR-pheng. isochron)	42.0±0.5	Dal Piaz et al. (2001)
Zermatt-Saas	Eclogitic gabbro	Rb-Sr (WR-pheng. isochron)	42.0±0.4	Dal Piaz et al. (2001)
Zermatt-Saas	Phengite quartzite	Ar-Ar (phengite)	46.1±0.4	Dal Piaz et al. (2001)
Zermatt-Saas	Mn quartzite	Ar-Ar (phengite)	44.2±0.4	Dal Piaz et al. (2001)
Zermatt-Saas	Phengite pyrop quartzite	Ar-Ar (phengite)	43.4±0.3	Dal Piaz et al. (2001)
Zermatt-Saas	Eclogitic gabbro	Ar-Ar (phengite)	44.3±0.3	Dal Piaz et al. (2001)
Zermatt-Saas	Eclogite	Ar-Ar (paragonite)	46.0±0.9 44.0±1.7	Barnicoat et al. (1995)
Zermatt-Saas	Glaucophane schist	Ar-Ar (paragonite)	43.0±5.8 56.0±11.0	Barnicoat et al. (1995)
Zermatt-Saas	Retrogressed gabbro	Ar-Ar (paragonite)	44.0±3.1	Barnicoat et al. (1995)
Schistes Lustrés	Phengite quartzite	Ar-Ar (phengite)	38.4±1.1	Chopin & Maluski (1980)
Schistes Lustrés	Piemontite ph. quartzite	Ar-Ar (phengite)	43.6±1.1	Chopin & Maluski (1980)
Schistes Lustrés	Phengite quartzite	Ar-Ar (phengite)	38.6±1.0	Chopin & Maluski (1980)
Schistes Lustrés	Calcschist	Ar-Ar (pheng. + paragonite)	67.8±1.8	Chopin & Maluski (1980)
Schistes Lustrés	Chloritoid q glaucophanite	Ar-Ar (pheng. + paragonite)	43.8±11	Chopin & Maluski (1980)
Schistes Lustrés	Metapelites	Ar-Ar (phengite)	38.2±1.2 35.5±1.1 44.9±1.5 51.9±1.6 53.7±1.7	Agard et al. (2002)

Table. 8.1. Continued.

Massif	Rock type	Method	Age (Ma)	Reference
Austroalpine				
Sesia	Eclogite	Lu-Hf (phengite-garnet)	69.2±2.7	Duchêne et al. (1997)
Sesia	Eclogite	U-Pb (SHRIMP-zircon)	65.0±5.0	Rubatto et al. (1999)
Sesia	Eclogitic micaschist	U-Pb (SHRIMP-zircon)	65.0±3.0	Rubatto et al. (1999)
Sesia	Mafic eclogite	U-Pb (SHRIMP-zircon)	68.0±7.0	Rubatto et al. (1999)
Sesia	Calc-silicate	U-Pb (sphenes)	66.0±1.0	Inger et al. (1996)
Sesia	HP marbles	Rb-Sr (ph. -WR isochron)	71.0±0.8	Dal Piaz et al. (2001)
Sesia	HP metagranitoid	Rb-Sr (omp.-WR isochron)	129.0±15.0	Oberhänsli et al. (1985)
Sesia	HP metagranitoid	Rb-Sr (gar.-omp. isochron)	114.0±1.0	Oberhänsli et al. (1985)
Sesia	Eclogitic micaschist	Rb-Sr (ph-par.-ep. isochron)	53.8±0.7	Inger et al. (1996)
Sesia	Marble	Rb-Sr (ph-calcite isochron)	63.3±0.8 57.0±1.2 63.0±0.8 64.5±0.9 61.0±0.7	Inger et al. (1996)
Sesia	Marble	Rb-Sr (ph-cal.-cpx isochron)	58.5±0.7	Inger et al. (1996)
Sesia	Leucogneiss	Rb-Sr (ph-epidote isochron)	52.3±0.6 59.8±0.7	Inger et al. (1996)
Sesia	Eclogitic micaschist	Rb-Sr (ph-omp. isochron)	46.4±2.1	Inger et al. (1996)
Sesia	Eclogitic micaschist	Rb-Sr (ph-cpx. isochron)	68.8±2.2 30.4±1.3 53.8±1.8	Inger et al. (1996)
Sesia	Metagranite	Rb-Sr (ph-cpx-ep. isochron)	63.0±1.3	Inger et al. (1996)
Sesia	Eclogite	Rb-Sr (ph-cpx isochron)	68.6±3.1	Inger et al. (1996)
Sesia	Marble	Ar-Ar (white mica)	69.0±3.0 70.0±1.0 75.0±1.5 93.7±2.4 71.0±1.5 105.0±10.0 91.0±2.5 92.4±2.8	Inger et al. (1996)
Sesia	Eclogite	Ar-Ar (phengite)	76.9±0.6	Ruffet et al. (1995)
Sesia	Leucocratic gneiss	Ar-Ar (phengite)	73.7±0.2	Ruffet et al. (1995)
Sesia	Micaschist	Ar-Ar (phengite)	73.6±0.3	Ruffet et al. (1995)
Sesia	Paragneiss	Ar-Ar (phengite)	69.4±0.7	Ruffet et al. (1995)
Mt. Emilius	Eclogite	Rb-Sr (WR-ph. isochron)	40.0±0.5	Dal Piaz et al. (2001)
Mt. Emilius	High-pressure gneiss	Rb-Sr (WR-ph. isochron)	48.0±5.0 42.0±0.4	Dal Piaz et al. (2001)
Glacier Rafray	Glaucophane eclogite	Rb-Sr (WR-ph. isochron)	45.0±0.4	Dal Piaz et al. (2001)
Etirol-Levaz	High-pressure gneiss	Rb-Sr (WR-ph. isochron)	45.0±0.7 47.0±0.9	Dal Piaz et al. (2001)

8.3.1.5. Valais

The age of HP metamorphism in the Valais zone has not been determined by geochronological studies because of the limited exposure of these rocks. In the central Alps, the existence of Palaeocene-Eocene fauna in HP rocks of supposedly Valasian origin (Bousquet et al., 2002) suggests that metamorphism occurred during or after the Eocene. This is further supported by recent geochronological constraints from the Chiavenna Unit (central Alps), indicating HP ages of 37-38 Ma (Liati et al., 2003).

Table. 8.1. Continued.

Massif	Rock type	Method	Age (Ma)	Reference
Briançonnais				
Central Siviez-Mischabel	Micaschist	Ar-Ar (white mica)	38.6±0.4 39.5±0.6 36.2±0.4 37.2±0.6 38.3±0.6 33.0±0.6 40.5±0.4 51.2±0.8 36.0±0.4 37.4±0.4	Markley et al. (1998)
Tsaté	Calcschist	Ar-Ar (white mica)	44.5±0.6	Markley et al. (1998)
Mont-Fort	Micaschist	Ar-Ar (white mica)	38.0±0.4 39.4±0.6	Markley et al. (1998)
Eastern Siviez-Mischabel	Micaschist	Ar-Ar (white mica)	34.9±0.6 31.7±0.6 36.5±0.4 31.8±0.4 29.6±0.2	Markley et al. (1998)
Southern Siviez-Mischabel	Micaschist	Ar-Ar (white mica)	35.0±0.4 37.0±0.4	Markley et al. (1998)
Rhone Valley	Micaschist	Ar-Ar (white mica)	40.0±0.6 47.7±0.6	Markley et al. (1998)

8.3.2. Oceanic sutures in the western Alps

Oceanic sutures in the western Alps have been subjected to conflicting interpretations (Figure 8.4). There is a general consensus that the ophiolitic complexes of the western Alps represent at least two Mesozoic ocean basins (the Piedmont and the Valais Oceans). The oceanic origin of the Piedmont domain is indicated by a widespread occurrence of Jurassic metagabbros and metabasalts overlain by Late Jurassic – Early Cretaceous radiolarian cherts, siliceous shales and pelagic limestones (Coward and Dietrich, 1989). The majority of these rocks have been intensely overprinted during the complex poly-deformational evolution of the Alps.

The rocks of the Valais domain are often considered to represent an additional ocean basin (e.g. Stampfli, 1993; Oberhänsli, 1994; Froitzheim et al., 1996; Stampfli et al., 1998; Froitzheim, 2001; Bousquet et al., 2002; Figure 8.4b). Ophiolitic material in the Valais domain is relatively scarce, and evidence for its oceanic origin is inferred from the occurrence of MORB-related volcanic rocks interlayered with Middle Cretaceous to Eocene deep-sea sediments found in the northwestern part of the western Alps (Marthaler, 1984; Oberhänsli, 1994; Bousquet et al., 2002). These rocks underwent Alpine HP metamorphism and do not appear to be connected with the Piedmont ophiolites (Figure 8.1).

Contrasting tectonic interpretations have been proposed for the origin of the Antrona ophiolite, which is located adjacent to the Monte Rosa massif (Figure 8.1a). Platt (1986) has proposed that these rocks mark the suture between the Internal Crystalline Massifs (here an independent ribbon) and the Briançonnais terrane, forming a third oceanic suture in the western Alps together with the Piedmont and the Valais Oceans (Figure 8.4c). In other works, the same rocks have been considered to represent the Piedmont suture (Escher and Beaumont, 1997) or the Valasian suture (Froitzheim, 2001).

An additional oceanic suture, separating the Sesia block from Adria, has been proposed by

Mattauer et al. (1987) based on the occurrence of ultramafic rocks in the southern tip of the Sesia Zone (Lanzo peridotites). The geochemistry of the Lanzo peridotites resembles the upper mantle residual of a mid-ocean-ridge (Bodinier, 1988), suggesting an origin from oceanic lithosphere. Evidence for this suture also exists in the rare serpentinised peridotite and metabasalt lenses of the Canavese zone (Escher and Beaumont, 1997). However, the contact between the Sesia rocks and the Lanzo peridotites is intensely deformed, and it is therefore possible that these oceanic rocks are in fact folded remnants of the Piedmont Ocean (e.g. Avigad et al., 1994; Michard et al., 1996).

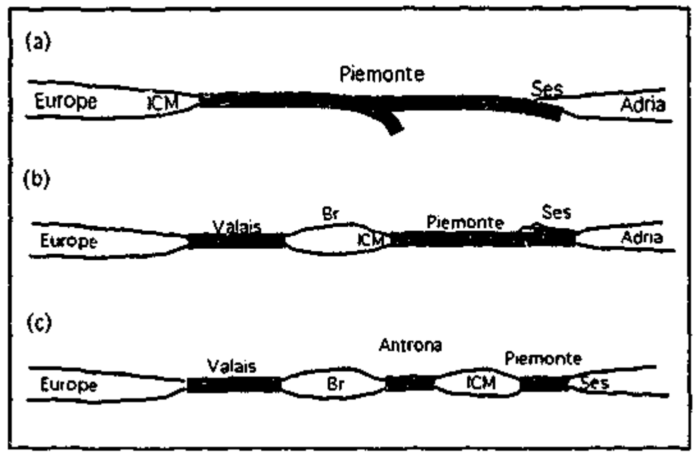


Figure 8.4. Cartoons showing conflicting Early Cretaceous reconstructions of ocean basins in the western Alps. (a) A single ocean basin, the Piedmont Ocean, separating Europe (Internal Crystalline Massifs) and Adria (Sesia) (Avigad et al., 1993). (b) Two ocean basins, the Piedmont and Valais Oceans. Note the location of the Internal Crystalline Massifs in the margin of the Briançonnais terrane (Michard and Martinotti, 2002). (c) Three oceans separating continental ribbons of the Briançonnais and the Internal Crystalline Massifs from both Europe and the Adria (Platt, 1986). Br, Briançonnais; ICM, Internal Crystalline Massifs; Ses, Sesia.

8.3.3. Plate motions and kinematic indicators

A major problem in Alpine reconstruction concerns with the difficulty to provide adequate kinematic constraints for the different terranes and microplates involved in the orogenic processes. At the largest scale, the relative convergence of Africa and Europe, as calculated from magnetic isochrons in the Northern Atlantic Ocean, can provide a first order approximation of kinematic boundary conditions (e.g. Dewey et al., 1973; 1989; Rosenbaum et al., 2002a). Motion paths of Africa with respect to Europe as calculated from the models of Dewey et al. (1989) and Rosenbaum et al. (2002a) are shown in Figure 8.5, as well the extrapolated motion based on NUVEL-1 plate velocities (DeMets et al., 1994). These boundary conditions, however, are applicable in western Alps reconstructions only if Adria has moved as an African promontory.

The controversy whether Adria behaved as an African promontory or as an independent microplate has been extensively discussed in the literature (e.g. Channell et al., 1979; Anderson, 1987; Wortmann et al., 2001; Rosenbaum et al., in review). Palaeomagnetic results suggest a

coherent motion of Africa and Adria since the Early Mesozoic (Channell, 1996; Muttoni et al., 2001). However, there is a considerable uncertainty about the motion of Adria during the Tertiary, due to the scarcity of palaeomagnetic data and an apparently independent motion inferred from active tectonics (Anderson, 1987). Moreover, the relatively poor resolution of palaeomagnetic data permits an independent motion of Adria in excess of hundreds of kilometres.

A further complication to describe the evolution of the Alps in terms of plate kinematics comes from the possibility that certain terranes (i.e. the Briançonnais or the Austroalpine terranes) moved independently in the area between Adria and Europe. This style of tectonism, which involves independent motions of allochthonous terranes, has been shown to take place in the western Mediterranean since the Oligocene (Rosenbaum et al., 2002b), and it is possible that analogous independent motions also took place in the convergence interface between Adria and Europe.

Several workers have attempted to analyse the relative motions of Adria and Europe on the basis of kinematic indicators found in rock structures in the western Alps (Baird and Dewey, 1986; Platt et al., 1989a; Figure 8.6a). The kinematic indicators are considered to represent the longest dimension of the strain ellipsoid, which may correspond to the direction of the tectonic transport. As seen in Figure 8.6a, the kinematic indicators show a radial pattern, which is orthogonal to the arc of the western Alps. Platt et al. (1989a) have suggested that this arcuate pattern could have been formed by body forces induced by an overall northwestward motion of Adria (Figure 8.6b). However, the existence of a SW-NE structural grain in the southwestern Alps may cast doubt on the validity of this model (Fry, 1989). It is possible that the arcuate pattern reflects a more consistent direction, which was reorientated during the formation of the arc of the western Alps (Figure 8.6c).

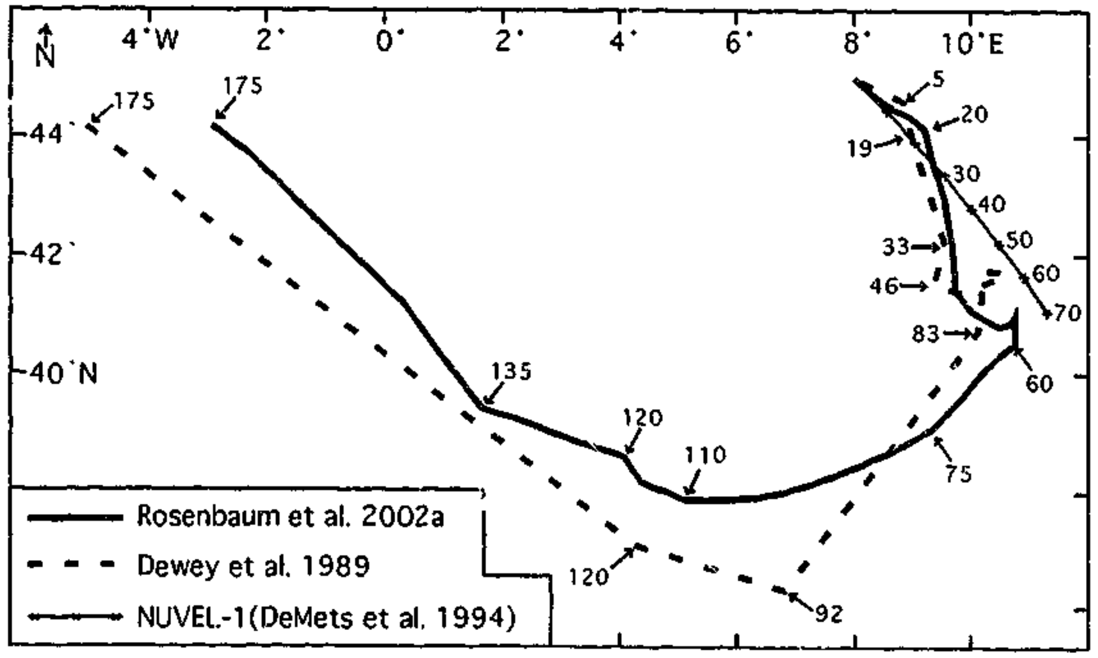


Figure 8.5. Comparison of the motion path of Africa relative to Europe shown by one point (45°N/8°E) that moves together with Africa (after Dewey et al., 1989; DeMets et al., 1994; Rosenbaum et al., 2002a). The motion path of NUVEL-1 is calculated by extrapolating the current angular velocity of Africa-Europe (DeMets et al., 1994) back in time. Numbers indicate time in million years.

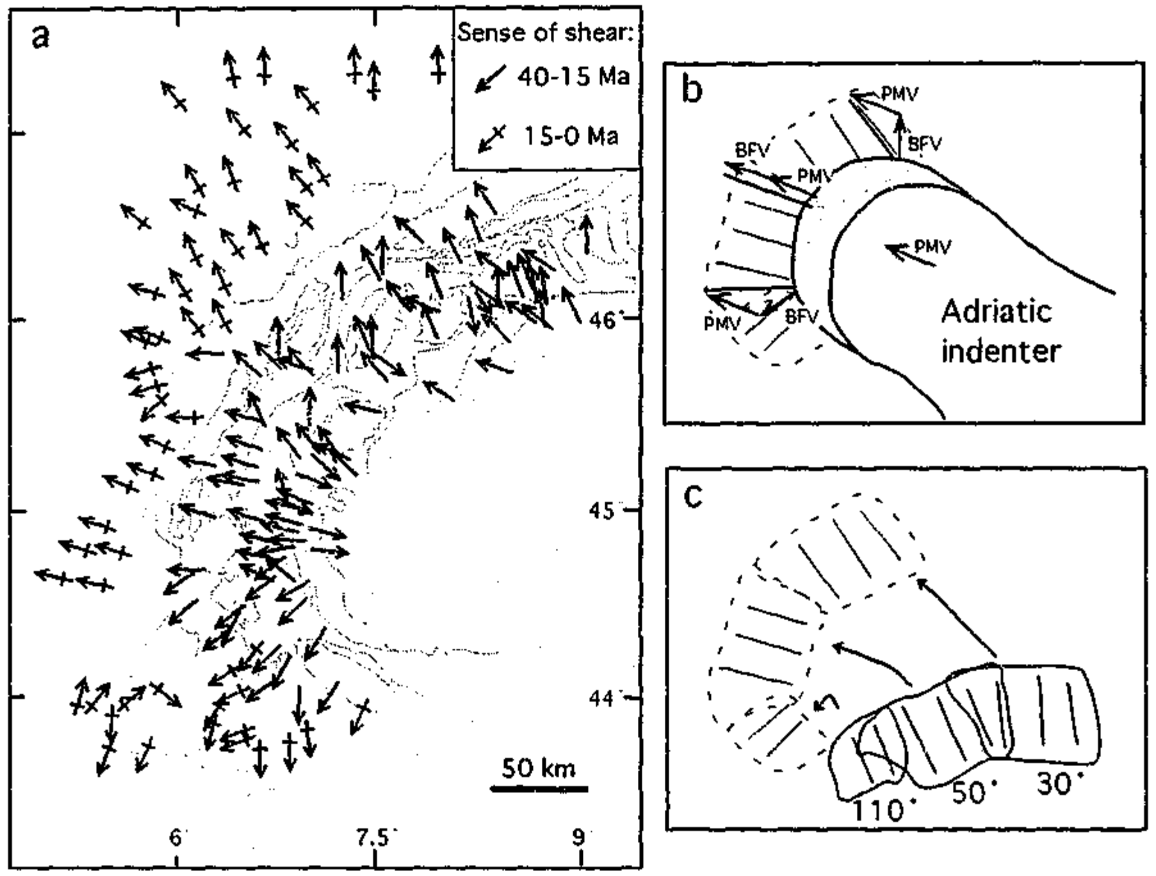


Figure 8.6. (a) Compilation of post-Eocene kinematic data from the western Alps inferred from linear structures (i.e. stretching lineations, crystal fibres on fault planes, wear grooves and stylolite axes). Arrows show sense of shear in rocks that are located at the head of arrows. Data are taken from Platt et al. (1989a, and references therein) and Lickorish et al. (2002, and references therein). (b) Formation of arcuate kinematic indicators as a result of a NW-ward indentation of Adria (after Platt et al., 1989a). BFV, body force vector perpendicular to the strike of the orogen; PMV, plate-motion vector. (c) An alternative explanation for the arcuate pattern of kinematic indicators, which involves reorientations of the lineations as a result of block rotations. Counterclockwise block rotations are applied according to palaeomagnetic results: 110° in the southern western Alps (Collombet et al., 2002), 50° in the central western Alps (Thomas et al., 1999) and 30° in the northern western Alps (Heller et al., 1989).

8.4. Tectonic reconstructions

In the following section, I discuss alternative reconstructions that have been proposed for different stages in the evolution of the western Alps. In most cases, differences between these reconstructions result from alternative interpretations of geological features. These controversies are discussed in accordance with the available spatio-temporal constraints.

8.4.1. Early/Middle Mesozoic

During the Early Mesozoic, the western Tethys was subjected to continental extension associated

with the breakup of Pangea (Coward and Dietrich, 1989). In northern Africa and in northwestern Europe, extensional basins, bounded by NW-SE faults, formed during the Permian and the Triassic (Coward and Dietrich, 1989). These faults were probably inherited from earlier Variscan structures (Coward and Dietrich, 1989). Pre-oceanic rifting in the Alps is recognised in Early Jurassic and early Middle Jurassic extensional structures (Lemoine and Trümpy, 1987).

Sea-floor spreading in the Alps commenced in the Middle Jurassic, contemporaneously with the opening of the Central Atlantic Ocean. It has therefore been suggested that the opening of these two ocean basins was linked by a major transform fault associated with a sinistral strike-slip motion (e.g. Laubscher and Bernoulli, 1977; Dercourt et al., 1986; Le Pichon et al., 1988). In the Alps, evidence for oceanic crust is found in the Piedmont ophiolite, for which crystallisation ages of 166-160 Ma have been obtained by U-Pb (SHRIMP) dating (Rubatto et al., 1998; Rubatto and Hermann, 2003).

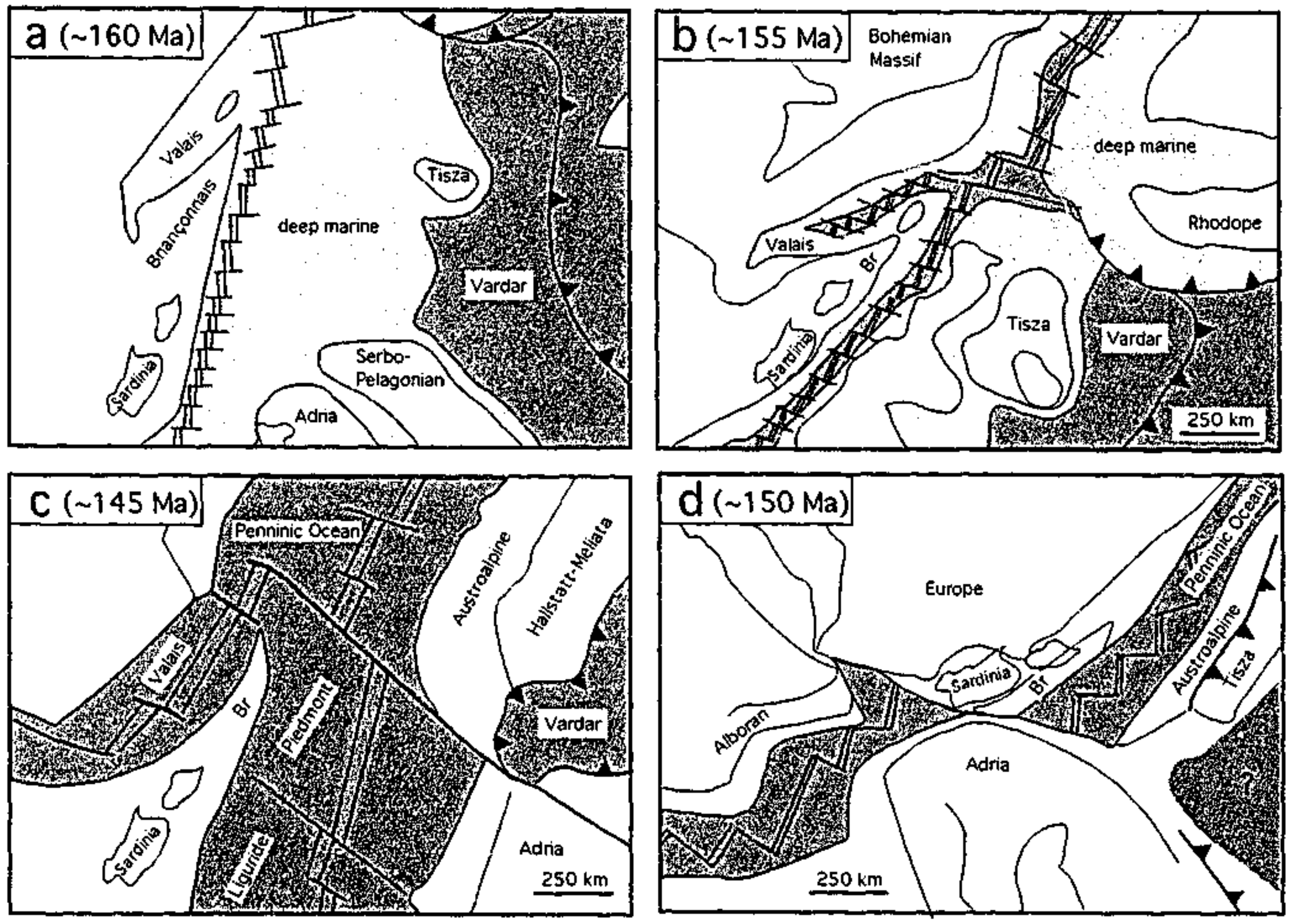


Figure 8.7. Middle-Late Jurassic reconstructions (160-145 Ma); (a) Dercourt et al. (1993); (b) Dercourt et al. (2000); (c) Stampfli and Marchant (1997); (d) Wortmann et al. (2001). Br, Briançonnais.

Middle-Late Jurassic reconstructions are shown in Figure 8.7. These models represent sea-floor spreading of an Alpine Ocean and subduction of the Vardar Ocean east of the Austroalpine terrane. The reconstruction of Dercourt et al. (1993) shows a single branch of sea-floor spreading related to Alpine Tethys (Figure 8.7a). This has been modified in the palaeogeographic map of Dercourt et al. (2000), which shows a relatively narrow (200-250 km) ocean basin separating Adria and Iberia, and an additional Jurassic oceanic embayment of similar width north of the Briançonnais terrane (Valais Ocean; Figure 8.7b). Alpine Tethys widens northeast of the eastern extension of the Briançonnais terrane, showing a deep marine connection with the eastern part of the Tethys Ocean (Dercourt et al., 2000). Similar tectonic features are shown in the Late Jurassic reconstruction of Stampfli and Marchant (1997) (Figure 8.7c), with an Iberian promontory of the Briançonnais zone bounded by the Piedmont and the Valais Oceans. Stampfli and Marchant's (1997) reconstruction, however, considers a much larger ocean (up to 700 km wide) for Alpine Tethys. The reconstruction of Wortmann et al. (2001) shows a complete separation between the northern and the southern branches of Alpine Tethys (Figure 8.7d). According to these authors, the Jurassic Briançonnais zone (including Corsica and Sardinia) was already detached from the Iberian microplate, and no oceanic domain existed in the Valais domain.

The plate tectonic configuration in the western Tethys prior to the opening of the Alpine Tethys has been discussed by Wortmann et al. (2001). These authors have argued that Early-Middle Mesozoic reconstructions of Adria using African rotational parameters result in a considerable overlap of Adria with southern Europe and Iberia. The degree of this space problem has been considerably diminished in the work of Rosenbaum et al. (2002a), using revised rotational parameters for the motion of Africa (and Adria) with respect to Europe. A relatively good fit of the continents in the Middle Jurassic was thus obtained.

A major difference between the alternative Late Jurassic reconstructions shown in Figure 8.7 is the width of the Alpine Tethys Ocean. It varies in width from ~250 km in the model of Dercourt et al. (2000) and Wortmann et al. (2001) to ~700 km in the model of Stampfli and Marchant (1997). The width of Alpine Tethys can be obtained by reconstructing the motion of Africa and Europe assuming that Adria moved together with Africa. Using this technique, Coward and Dietrich (1989) have estimated a width of ~500 km (based on data from Dewey et al., 1989). Revised rotational parameters by Rosenbaum et al. (2002a) resulted in ~400 km width of the Jurassic Alpine Tethys.

8.4.2. Cretaceous

One of the major aspects with regard to Cretaceous reconstructions of the western Alps is related to the tectonic significance of the Valais domain. In some reconstruction models, Valaisian rocks have been interpreted to originate from a small wedge shaped marine basin, underlain by continental crust (e.g. Polino et al., 1990; Dercourt et al., 1993; Figure 8.7a and Figure 8.8a). In other reconstructions, however, the existence of an ocean basin that separated the Briançonnais terrane from the European margin has been considered (e.g. Stampfli, 1993; Schmid et al., 1997; Stampfli and Marchant, 1997; Stampfli et al., 1998; Bousquet et al., 2002; Figure 8.8b).

Stampfli et al. (1998) have linked the opening of the Valais Ocean with the opening of the Northern Atlantic Ocean during the Late Jurassic – Early Cretaceous (ca. 155-120 Ma). Based on this study, the two spreading centres were connected by a transform fault through the Bay of Biscay and

the Pyrenees and underwent similar amounts of opening. At the same time that sea-floor spreading occurred in the Valais Ocean, a southeast-dipping subduction zone formed in the Alpine Tethys, leading to Early and Late Cretaceous orogenesis in the eastern Alps (Figure 8.8b). In the model of Stampfli et al. (1998), the closure of the Valais Ocean has been attributed to the rotation of Iberia and the opening of the Bay of Biscay at 120-83 Ma.

The timing of sea-floor spreading of the Valais Ocean has been recently addressed by radiometric dating of metabasic rocks in an ophiolitic unit within the central Alps (Chiavenna unit) that is attributed to the Valais Ocean (Liati et al., 2003). Oceanic-related magmatism in this locality appears to have occurred at ca. 93 Ma, suggesting that the Valais Ocean was still opening during the Late Cretaceous.

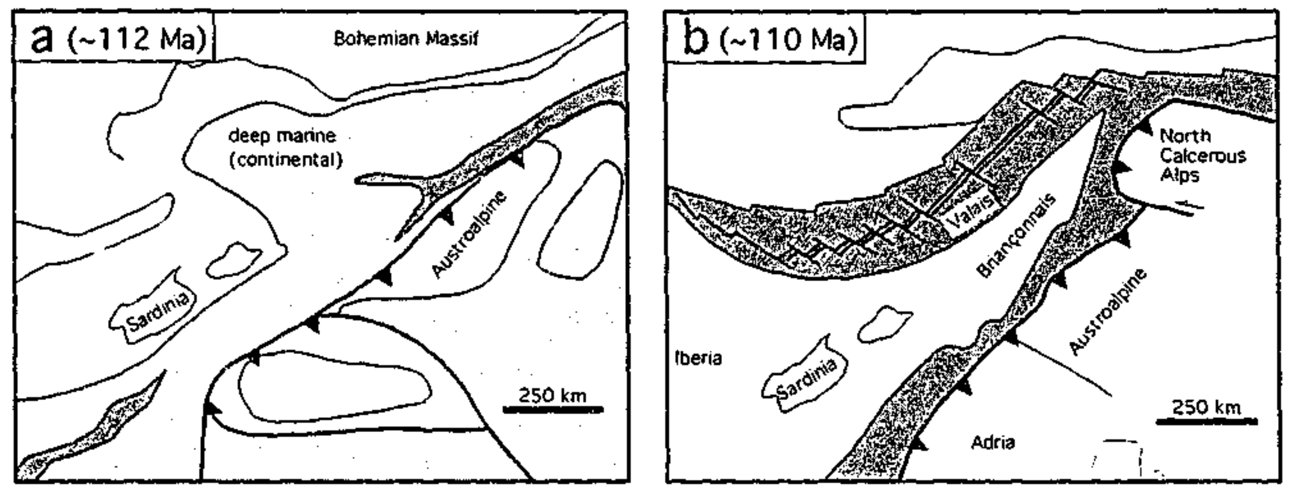


Figure 8.8. Early Cretaceous reconstructions of the western Alps: (a) continental marine basin in the Valais domain, after Dercourt et al. (2000); (b) sea floor spreading in the Valais Ocean, after Schmid et al. (1997) and Stampfli et al. (1998).

Orogenic processes during the Cretaceous are documented by major unconformities and HP metamorphism in the eastern Alps (e.g. Thöni and Jagoutz, 1993). However, in the Internal Crystalline Massifs and the Piedmont domain, the role of Cretaceous HP metamorphism (e.g. Hunziker et al., 1989) is now debated (e.g. Michard et al., 1996; Gebauer, 1999). Nevertheless, it is widely accepted, based on geochronolgical evidence, that subduction and collisional processes were responsible for HP metamorphism in the Sesia/Austroalpine terrane during the Late Cretaceous.

According to the reconstruction of Stampfli and Marchant (1997) (Figure 8.9a), Late Cretaceous orogenesis was dominated by two systems of subduction-accretionary complexes, striking roughly NE-SW. Northwest of the Briançonnais terrane, the Valais Ocean was consumed in a southeast-dipping subduction zone, which continued westward into the Pyrenees (Figure 8.9a). At the same time, the Briançonnais terrane was subjected to extensional faulting and tilting, and was not incorporated in the accretionary prism before the Middle-Late Eocene (45-35 Ma; Stampfli et al., 1998; Michard and Martinotti, 2002). In this reconstruction, the Hellminthoid Flysch separated the southeast side of the Briançonnais terrane from the Austroalpine terrane. Towards the east, a second active margin existed, associated with the collisional suture of the Austroalpine terrane and the Dinaro-Pelagonian microplate (Figure 8.9a).

A considerably different Late Cretaceous reconstruction by Dercourt et al (2000) shows occurrences of orogenic processes further to the south, predominantly in an E-W-striking accretionary belt (Figure 8.9b). The Austroalpine terrane was bounded by an active margin in the north, which was associated with the subduction of a ca. 300 km wide ocean basin separating the southern margin of Europe from the front of the Alpine orogen. The overall E-W orientation of the Austroalpine terrane is similar to its present orientation in the eastern Alps. This is inconsistent with palaeomagnetic data, which indicate at least 30° of clockwise rotation of the eastern Alps relative to Europe since the Jurassic (Channell et al., 1992b).

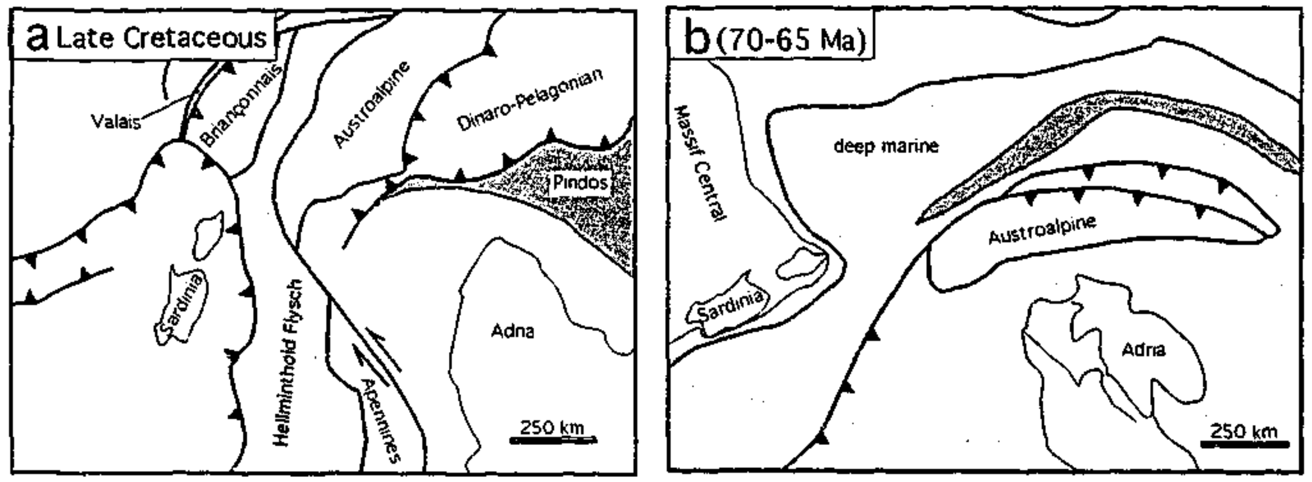


Figure 8.9. Alternative Late Cretaceous reconstructions of the Alps; (a) Stampfli and Marchant (1997); (b) Dercourt et al. (2000).

8.4.3. Early Tertiary

The period between 65 Ma and 50 Ma was characterised by relatively small convergence rates of Africa with respect to Europe (Dewey et al., 1989; Rosenbaum et al., 2002a). During the Palaeocene, convergence actually stopped for a period of ~10 Myr (at 65-55 Ma; Figure 8.5). Eocene (~45 Ma) ages of HP and UHP metamorphism of the Piedmont ophiolite (Figure 8.3b) have been interpreted to represent the closure of the Piedmont Ocean during southeast-dipping subduction (e.g. Michard et al., 1996; Stampfli et al., 1998; Froitzheim, 2001). The subduction zone is shown in the Eocene reconstructions of Figure 8.10.

The Middle Eocene reconstruction of Dercourt et al. (2000) shows a southeast-dipping subduction of the deep marine 'Alpine Flysch Basin' beneath Adria (Figure 8.10a). In contrast, according to Schmid et al. (1997), all oceanic material had already been consumed in the Middle Eocene, and the Briançonnais terrane was incorporated in the subduction/collisional suture (Figure 8.10b). An additional collisional suture is related to the southeast-dipping subduction of the Valais Ocean beneath the Briançonnais (Schmid et al., 1997). Froitzheim (2001) proposed that subduction of the Piedmont Ocean led to the collision of the Briançonnais terrane with the Adriatic margin in the

Middle Eocene, whereas the consumption of the southeast-dipping Valais Ocean occurred only in the latest Eocene. This reconstruction seems to agree with the timing of collisional deformation in the Briançonnais terrane, which shows no evidence for orogenesis prior to the Middle Eocene (Stampfli et al., 1998).

An important question is whether the age of HP metamorphism (~45 Ma) represents subduction of the Piedmont Ocean or obduction of the ophiolitic complex during collisional processes. The geochronological data suggest that a short-lived HP event occurred at ~45 Ma (Rubatto et al., 1998; Rubatto and Hermann, 2003), followed by a rapid exhumation to mid-crustal levels at 42-38 Ma (Amato et al., 1999; Reddy et al., 1999; 2003; Cartwright and Barnicoat, 2002). Smilar ages of HP metamorphism has been obtained from two different branches of the Piedmont Ocean, which are located more than 100 km apart (Zermatt Saas and Monviso ophiolites; Figure 8.1). In addition, HP metamorphism during subduction processes should result in a wide range of ages. I therefore suggest that at 45 Ma, a short-lived orogenic episode affected the whole area and possibly involved obduction and collisional processes, which are commonly overlooked in existing reconstructions. These events may have been followed by a period of intense stretching of the overriding plate.

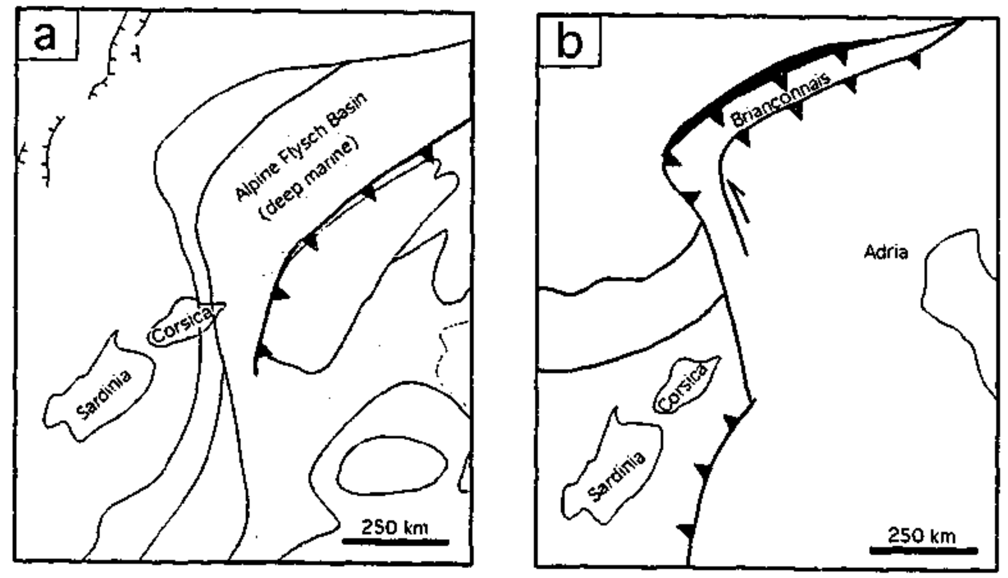


Figure 8.10. Eocene (~45 Ma) reconstructions; (a) Dercourt et al. (2000); (b) Schmid et al. (1997).

8.4.4. The Eocene-Oligocene collision in the western Alps

A major collisional event that involved HP and UHP metamorphism of the Internal Crystalline Massifs occurred at ~35 Ma. This is supported by precise radiometric dating of HP and UHP rocks that yielded Oligocene ages in the Dora Maira (Gebauer et al., 1997; Rubatto and Hermann, 2001), Monte Rosa (Rubatto and Gebauer, 1999), the Lepontine Dome (Gebauer, 1996) and the Voltri Massif (34±1 Ma; Rubatto and Scambelluri, 2002) (Figure 8.3a).

Froitzheim (2001) has interpreted the Oligocene HP metamorphic ages as the final stage of subduction of the Valais Ocean. In this interpretation, the Internal Crystalline Massifs originated in the European margin and were brought into the Valasian subduction trench after the incorporation of the Briançonnais terrane in the accretionary wedge and the subduction of the Valais Ocean. This model is consistent with the timing of deformation and metamorphism in the Briançonnais, which is estimated as 45-34 Ma (Freeman et al., 1997; Markley et al., 1998; Stampfli et al., 1998; Michard and Martinotti, 2002). The post-Eocene subduction of the Valais Ocean is supported by the occurrence of Palaeocene-Eocene fauna within the HP rocks of the Valais (Bousquet et al., 2002). However, the model of Froitzheim (2001) implies that orogenic processes did not affect the Internal Crystalline Massifs prior to the Oligocene. Therefore, this model would have to be refined if the Internal Crystalline Massifs underwent earlier (Cretaceous) HP metamorphism (Figure 8.3a).

8.5. Discussion

An important issue arising from this chapter is the episodic temporal distribution of HP metamorphism in the western Alps. Different terranes within the orogen underwent HP metamorphism at different stages: 70-65 Ma in the Sesia zone, 45-44 Ma in the Piedmont, ~40-35 Ma in the Briançonnais and ~35 Ma in the ICM. Lister et al. (2001) interpreted this episodic pattern of HP ages as evidence for switches in tectonic modes, from crustal shortening to extension, following accretionary events. These authors have shown that orogenic processes (e.g. in the Southwest Pacific Ocean) are commonly governed by subduction of small ocean basins, rollback of subduction zone and accretion of continental ribbons (Lister et al., 2001). In this tectonic configuration, each accretionary event is followed by resumption of subduction and slab rollback in an adjacent ocean basin, which in turn, would switch the overriding plate into an extensional mode. The latter could then lead to the formation of metamorphic core complexes in the overriding plate and to rapid exhumation of HP rocks.

The existence of distinct HP ages for different ribbons within the western Alps suggests that similar processes, related to episodic accretionary events and switches to extension, may have occurred in this orogen. This working hypothesis is considered in the possible evolutionary model for the western Alps, from Cretaceous to Oligocene, shown in Figure 8.11.

The pre-Alpine configuration is presented in Figure 8.11a. It shows that the area between Adria and Europe consisted of continental ribbons (e.g. Sesia zone, Briançonnais) that were separated by ocean basins (e.g. Piedmont, Valais, Lanzo Ocean?). The kinematic boundary conditions for the motion of Adria and Europe are based on the motion of Africa with respect to Europe (after Rosenbaum et al., 2002a). This pre-Alpine configuration shows spatial relationships between Alpine terranes and their HP metamorphic ages, with older HP ages located further from the European margin. Based on the geological evidence, it seems that both shortening and extensional deformation migrated through time towards Europe (Figure 8.2). This suggests that the orogenic history was controlled by a retreating southwest-dipping subduction system, which was active during the progressive consumption of a sequence of ocean basins and the accretion of continental ribbons into each other (Figure 8.11).

Accretion of the Sesia zone occurred in the Late Cretaceous (70-65 Ma), possibly after the

consumption of an ocean basin (Lanzo Ocean?) that separated this terrane from Adria (Figure 8.11b,c). Following accretion, subduction recommenced in the Piedmont Ocean, and the rollback of its subduction hinge led to an overall extension regime in the overriding plate resulting in the exhumation of HP units (Figure 8.11d). This process continued as long as subduction of the Piedmont Ocean occurred. Note that during the subduction of the Piedmont Ocean, in the Paleocene and Early Eocene, very little or no convergence occurred between Africa and Europe (Figure 8.5; Rosenbaum et al., 2002a). The horizontal motions associated with the subduction of the Piedmont Ocean were probably accommodated by subduction rollback.

Subduction of the Piedmont Ocean ceased at ~45 Ma and probably involved the accretion of the Briançonnais terrane into the orogenic wedge (Figure 8.11e). The switch to extensional regime was once again controlled by resumption of subduction rollback, this time in the Valais ocean, leading to extensional deformation and exhumation of the Piedmont oceanic lithosphere (Figure 8.11f). Subduction of the Valais Ocean eventually led to accretion and UHP metamorphism of the Internal Crystalline Massifs (at ~35 Ma; Figure 8.11g).

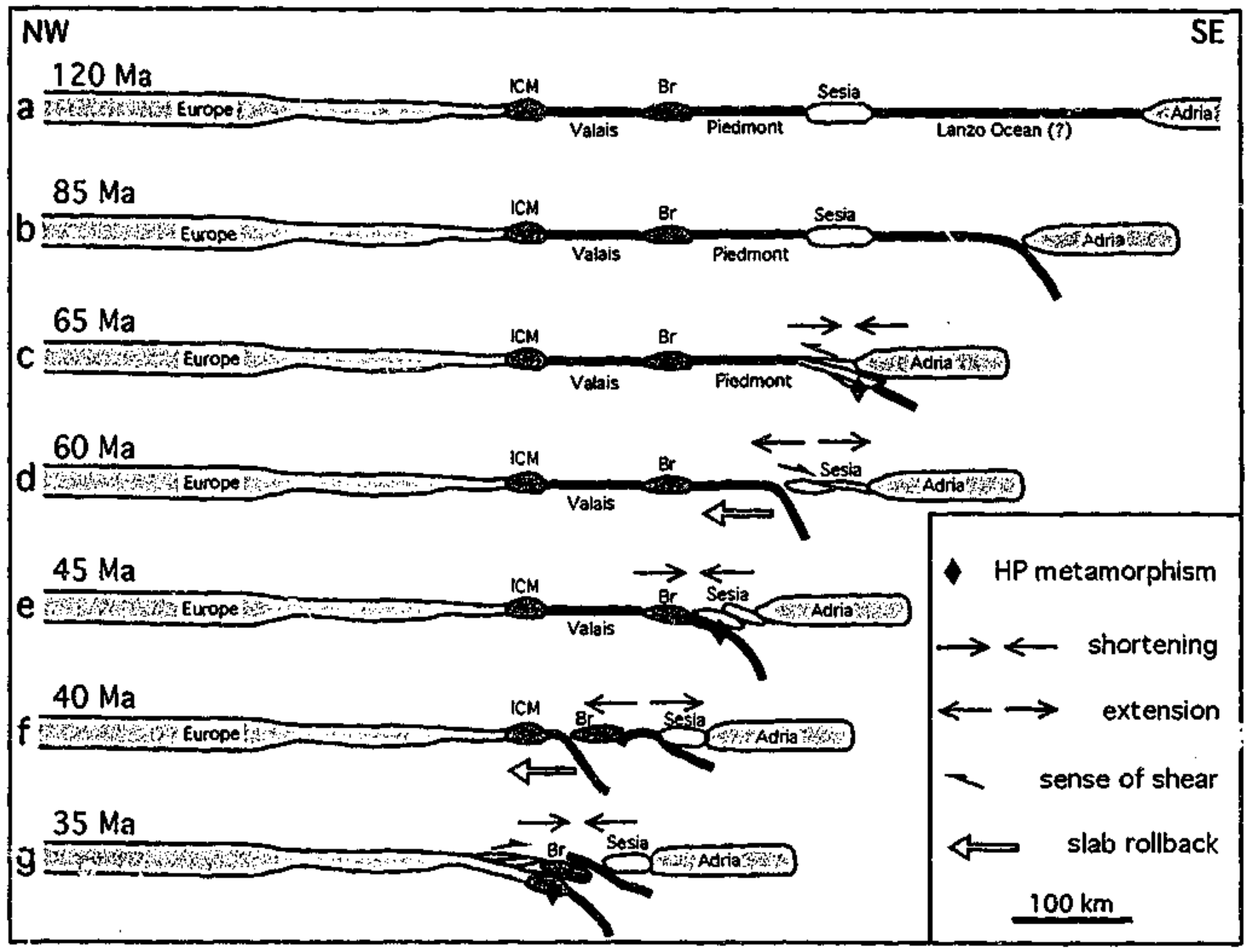


Figure 8.11. Possible tectonic model for the evolution of the western Alps from the Early Cretaceous to the Oligocene. Note that switches from crustal shortening to extension correspond to transitions from accretionary events to subduction rollback. Br, Briançonnais; ICM, Internal Crystalline Massifs.

The above model has profound implications. It implies that Africa has never collided with Europe. Rather, continental ribbons were accreted to the orogen in a tectonic environment that resembles the modern Southwest Pacific region.

8.6. Concluding remarks

In this chapter, I have presented spatio-temporal constraints and alternative reconstruction models for the tectonic evolution of the western Alps. The ages of HP metamorphism in the western Alps suggest that orogenesis did not occur during a progressive and continuous process of subduction and collision (e.g. Hsü, 1989). Rather, the process appears to have incorporated internal episodic interactions between continental allochthonous terranes and relatively small ocean basins. The orogen was subjected to several relatively short episodes of HP metamorphism (Lister et al., 2001). Three such episodes are identified in the western Alps: (1) at ~65 Ma in the Austroalpine terrane; (2) at ~45 Ma in the Piedmont zone; and (3) at ~35 Ma in the Internal Crystalline Massifs. The age clusters may further suggest that dated metamorphic signatures were acquired during collisional events and not during progressive subduction, because the latter would lead to a wider range of metamorphic ages.

Each of the accretion episodes may have been followed by a period of intense horizontal extension. In the example of the Latest Eocene episode (~35 Ma), a period of widespread extension followed, and was associated with the rapid exhumation of UHP rocks (see Chapter 9). Likewise, extensional tectonism occurred in the Piedmont zone at 45-36 Ma, exhuming UHP rocks metamorphosed at ~45 Ma (Reddy et al., 2003), and possibly also in the Sesia zone following HP metamorphism at 65 Ma (Avigad, 1996). The geodynamics of such switches from shortening to extension may be associated with interactions between subduction rollback and accretionary events, but is nevertheless a matter of further research.

CHAPTER 9

WESTWARD SLAB ROLLBACK AS A MODEL FOR OLIGOCENE EXTENSIONAL TECTONISM IN THE WESTERN ALPS

.....reexamine all you have been told at school or church, or in any book, dismiss what insults your soul, and your very flesh shall be a great poem.

Walt Whitman, Leaves and Grass (1855)

FOREWORD: CHAPTER 9

This chapter discusses enigmatic observations related to the role of extensional tectonics in the western Alps during the Oligocene. Concepts presented in this chapter have been influenced by the paper of Lister et al. (2001) on episodicity during orogenesis, and have been extensively discussed with Gordon Lister. The chapter have benefited from discussions with Hans Laubscher, Roberto Compagnioni, Jean-Michel Bertrand, Marnie Forster, Marco Beltrando, Jerome Ganne, Wouter Schellart, Maarten Krabbendam, Norman Fry and David Giles. Comments on the manuscript by Wouter Schellart, Gordon Lister, Ivo Vos and Quinton Hills are also acknowledged.

9. WESTWARD SLAB ROLLBACK AS A MODEL FOR OLIGOCENE EXTENSIONAL TECTONISM IN THE WESTERN ALPS

Abstract

The arcuate orogen of the western Alps was subjected to widespread extensional tectonism shortly after a major collisional episode at the Eocene-Oligocene boundary (i.e. from ~35 Ma). Extension was accommodated by movements along gently inclined extensional shear zones, which facilitated rapid exhumation of high-pressure and ultra-high-pressure rocks. The extension in the western Alps was coeval with widespread extension in the western Mediterranean, which was associated with the exhumation of metamorphic core complexes in Corsica and Calabria at 32-30 Ma, and the subsequent opening of the western Mediterranean back-arc basins (Ligurian, Provençal and the Gulf of Lion). Three mechanisms for this extensional episode are discussed: (1) internal readjustments within the orogenic wedge; (2) collapse of an overthickened lithosphere; or (3) rollback of a subducting slab. The latter mechanism is favoured because it can account for the large horizontal motions predicted by shearing along gently inclined extensional shear zones. However, the validity of a rollback model implies that an oceanic domain (Valais Ocean) must have existed between the Internal Massifs and Europe, and was not entirely consumed before ca. 30 Ma. Therefore, the ophiolitic suture associated with the collision at ~35 Ma may indicate another oceanic domain (Antrona Ocean), which is usually overlooked in reconstruction models.

9.1. Introduction

The arcuate orogenic belt of the western Alps is located at the convergent boundary between Europe and the African promontory of Adria (Figure 9.1a). The orogen is commonly regarded as a classic collisional belt representing the closure of one or more ocean basins from the Late Cretaceous to Late Eocene (~90-35 Ma) and the subsequent progressive indentation of Adria into Europe (e.g. Platt et al., 1989a). The internal part of the orogen (the 'Penninic Alps') consists of a series of continental terranes (the Austroalpine-Sesia zone, the Briançonnais zone and the Internal Massifs), which are aligned as arcuate ribbons parallel to the strike of the orogen (Figure 9.1b). These terranes are separated from each other and from the stable foreland by several ophiolitic sutures (Piedmont, Antrona(?), Valais), which mark the locations of consumed Jurassic to Cretaceous ocean basins (Stampfli et al., 1998).

Occurrences of high-pressure (HP) and ultra-high-pressure (UHP) rocks in the internal western Alps indicate that during orogenesis rock units were subjected to burial at depths of 40-100 km. HP and/or UHP metamorphism affected each zone within the Penninic Alps at different stages of Alpine orogenesis (e.g. 70-65 Ma in the Sesia Zone, ~44 Ma in the Piedmont zone, 45-36 Ma in the Briançonnais, and ~35 Ma in the Internal Massifs; Duchêne et al., 1997; Gebauer et al., 1997; Markley et al., 1998; Rubatto et al., 1998; 1999; Rubatto and Gebauer, 1999; Rubatto and Hermann, 2003). These ages suggest that Alpine orogeny may have evolved during episodic collisional events associated with accretion of a number of small continental terranes (Lister et al., 2001; Figure 9.1c). In this tectonic configuration, it is possible that HP or UHP rocks formed as the result of each individual accretionary event, and were subsequently rapidly exhumed due to renewed rollback of an adjacent subduction zone (Lister et al., 2001).

This chapter focuses on the tectonism in the western Alps following the collisional event at the Eocene-Oligocene boundary (~35 Ma). I refer to geochronological evidence (e.g. Rubatto and Hermann, 2001) that implies extreme rates of exhumation shortly after the culmination of UHP metamorphism in the Internal Massifs (Dora Maira, Gran Paradiso and Monte Rosa; Figure 9.1b). It is then discussed that during a short period in the Oligocene, the Internal Massifs were predominantly subjected to extensional tectonism. These results challenge the current tectonic paradigm of the western Alps, which implies progressive crustal shortening due to the indentation of Adria into Europe (e.g. Platt et al., 1989a), and thus call for a re-examination of evolutionary tectonic models of the western Alps.

9.2. Evidence supporting extensional tectonism during the Oligocene

The idea that the Oligocene Alps were subjected to widespread extensional tectonism was first suggested by Laubscher (1983) based on the occurrence of Oligocene (32-25 Ma) magmatism in the central Alps and in the northern part of the western Alps (Figure 9.1b). Oligocene magmatism is associated with calc-alkaline intrusions and mafic dyke swarms, which outcrop in a narrow zone close to the Insubric Line (the present-day boundary between Adria and Europe; Figure 9.1b). This evidence alone does not provide sufficient support for extensional tectonism. However, it shows that an acute lithospheric-scale process, which involved mantle melting (indicated by mafic dykes), took place during the Oligocene. In what follows, it is demonstrated that magmatism along the Insubric Line was coeval with extensional tectonism that affected the Internal Massifs in the western Alps.

Recent geochronological data from the HP and UHP rocks of the internal massifs have revealed consistent metamorphic ages at ca. 35 Ma (Gebauer et al., 1997; Rubatto and Gebauer, 1999; Rubatto and Hermann, 2001). Most of these ages were obtained by U-Pb SHRIMP dating of coesite-bearing UHP rocks in the Dora Maira massif, indicating that at ~35 Ma, rocks were buried to depths of more than 100 km. Additional in situ SHRIMP dating of retrogressed minerals within the UHP rocks has shown that at 33-32 Ma, the UHP unit was already exhumed to intermediate depths of 20-40 km

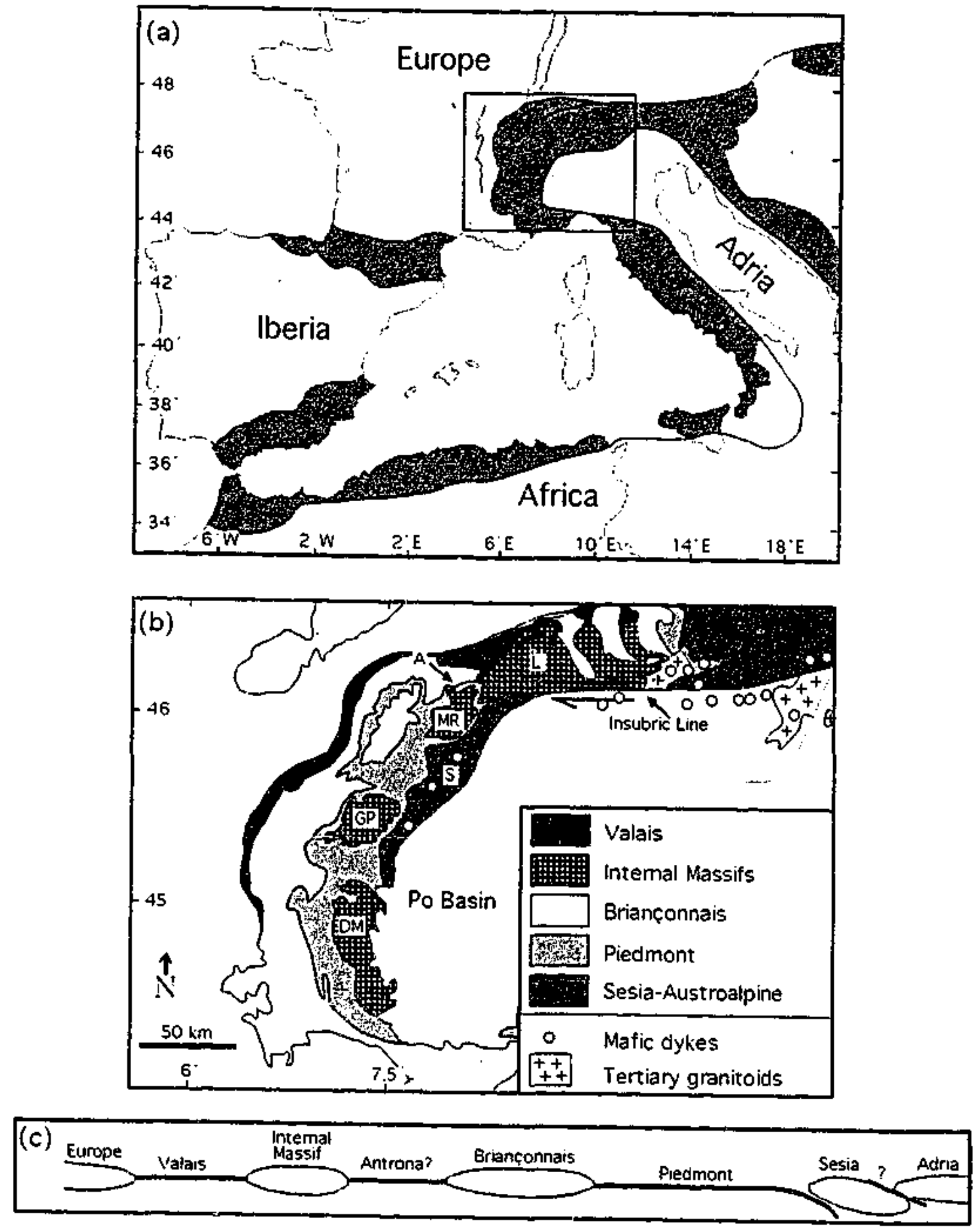


Figure 9.1. (a) The arcuate shape of the Alpine belt (dark grey) in the western Mediterranean region. (b) Tectonic map of the Internal western Alps (modified after Polino et al., 1990) and distribution of Eocene-Oligocene magmatic centres along the Adria-Europe plate boundary (after von Blanckenburg and Davies, 1995). A, Antrona; DM, Dora Maira; GP, Gran Paradiso; L, Lepontine; MR, Monte Rosa; S, Sesia. (c) Hypothetical tectonic configuration of Alpine terranes and ocean basins at ~65 Ma.

(Rubatto and Hermann, 2001; Figure 9.2). These results imply extreme rates of exhumation, of 1.6-3.4 cm/yr, immediately after the culmination of UHP metamorphism at 35 Ma (Rubatto and Hermann, 2001). Gebauer et al (1997) have provided an additional support for the role of rapid exhumation by using zircon fission track dating of a sample that was subjected to UHP metamorphism at 35 Ma and was cooled to temperatures of about 290° at ~30 Ma. Their results suggest possible exhumation rates of 2.0-2.4 cm/yr. These exhumation rates are considerably higher than predicted exhumation rates from erosion in a mountainous, tectonically active, wet environment (e.g. 0.5-1.3 cm/yr in the Southern Alps of New Zealand and the Taiwan Alps; Ring et al., 1999). Therefore, exhumation was possibly facilitated by tectonic activity such as extension (e.g. Hill et al., 1992).

In the Dora Maira massif, there is evidence for the existence of extensional structures associated with low-angle tectonic contacts juxtaposing relatively low-pressure rocks on top of an UHP unit (Avigad, 1992; Avigad et al., 2003). The existence of such contacts implies that a thick crustal section above the UHP rocks was excised before or during the movement along the contacts. The contacts are characterised by normal sense, ductile shear zones, which were active during retrograde greenschist facies metamorphism (Philippot, 1990; Avigad, 1992; Avigad et al., 2003).

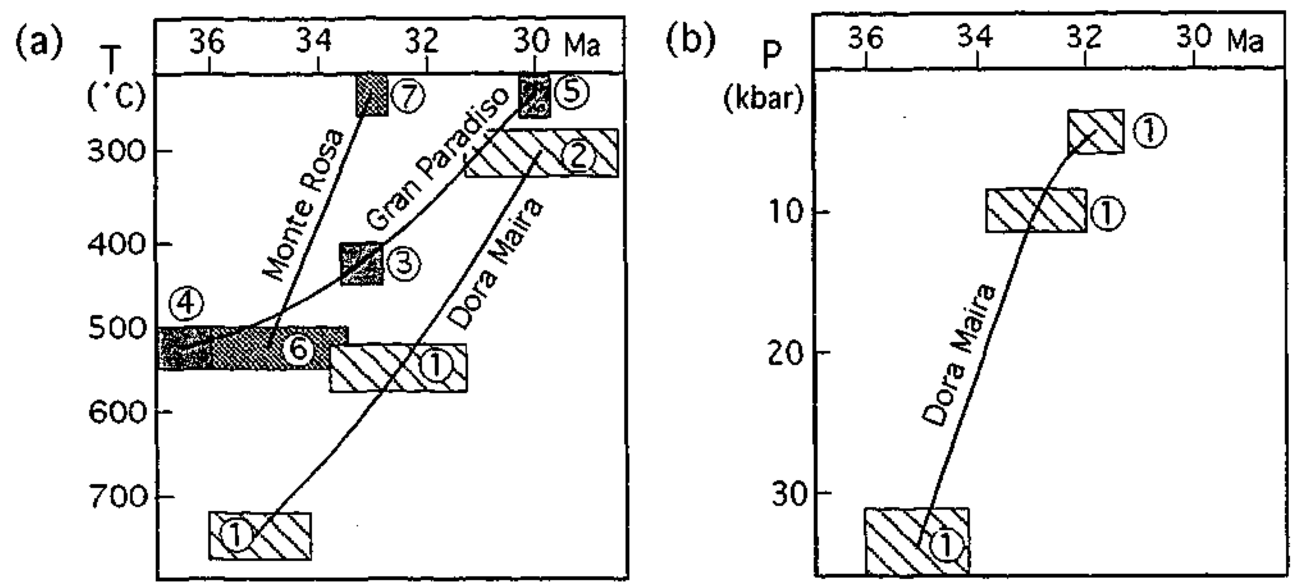


Figure 9.2. Temperature-Time (a) and Pressure-Time (b) paths in the Internal Massifs showing extreme rates of cooling and exhumation during the Early Oligocene. Note that robust constraints on the timing of exhumation in the Gran Paradiso and Monte Rosa are lacking; however, the cooling paths in these massifs resemble the history of cooling during exhumation in Dora Maira. References are: (1) Rubatto and Hermann (2001); (2) Gebauer et al. (1997); (3) Freeman et al. (1997); (4) Brouwer et al. (2002); (5) Hurford and Hunziker (1989); (6) Rubatto and Gebauer (1999); (7) Hurford et al. (1991).

The exact timing of extensional deformation in the Dora Maira massif is yet to be constrained by geochronological methods. However, an age of 34-30 Ma is suggested for the extensional phase, because extension postdates UHP metamorphism at ~35 Ma and was possibly related to the subsequent rapid exhumation (Figure 9.2). Further research is required in order to determine whether such an extensional event was also responsible for the exhumation of the Gran Paradiso and Monte Rosa massifs. In these internal massifs, there is a lack of reliable constraints on the age of HP metamorphism, and existing ages are highly controversial (e.g. see review by Gebauer, 1999). An age of 34.9±1.4 Ma has been proposed for the time of HP metamorphism in the Monte Rosa massif (Rubatto and Gebauer, 1999), and possible HP ages of 43-36 Ma are documented from the

Gran Paradiso massif (references in Brouwer et al., 2002). Zircon fission track ages of ca. 30 Ma in Gran Paradiso and ca. 33 Ma in Monte Rosa (Hurford and Hunziker, 1989; Hurford et al., 1991) indicate that cooling to ~225°C took place during the Oligocene, that is, contemporaneously with the exhumation of UHP rocks in the Dora Maira massif (Figure 9.2). Extensional structures related to a short-lived extensional deformation during the Oligocene (ca. 31 Ma) have also been reported from the area between the Monte Rosa and the Gran Paradiso massifs (Bistacchi et al., 2001). These observations suggest that extensional deformation between 34-30 Ma may have affected all the Internal Massifs of the western Alps.

9.3. The western Alps in context of the Oligocene tectonism in the western Mediterranean

The Oligocene extensional phase in the western Alps occurred at the same time that extensional tectonism commenced in the western Mediterranean region, leading to the exhumation of metamorphic core complexes in Alpine Corsica and in Calabria, and the subsequent back-arc extension in the Ligurian-Provençal Basins and the Gulf of Lion (Fournier et al., 1991; Rosseti et al., 2001; Rosenbaum et al., 2002b, and references therein). Extension was responsible for the destruction of an earlier HP belt that consisted of Alpine Corsica, Calabria, the Kabylies blocks and the internal parts of the Rif-Betic chain (Rosenbaum et al., 2002b; Figure 9.3). Prior to the Oligocene, these terranes were subjected to a prolonged tectono-metamorphic evolution that involved Cretaceous to Oligocene HP metamorphism and obduction of Jurassic ophiolites similar to those found in the internal zone of the western Alps (Fournier et al., 1991; Rosseti et al., 2001). It is therefore suggested that during the Oligocene, the SW-NE striking orogenic belt of the western Mediterranean was linked to the HP belt of the western Alps (Figure 9.3).

During the Oligocene (32-30 Ma), HP rocks in Corsica and Calabria were exhumed below flat-lying extensional detachments (Fournier et al., 1991; Rosseti et al., 2001). These detachments accommodated extensional strain during exhumation, and involved large horizontal motions sufficient to account for such exhumation processes. Horizontal extension was probably generated by rollback of a northwest-dipping slab, which induced lithospheric extension in the overriding plate and led to the destruction of the Oligocene western Mediterranean HP belt (Rosenbaum et al., 2002b) (Figure 9.3). Fragments of this belt now form an arcuate pattern of orogenic segments that encircles the western Mediterranean basins (Figure 9.1a and Figure 9.3).

Currently, the arcuate belt of the western Alps has an analogous shape compared to the western Mediterranean arcuate belts (Figure 9.3), suggesting comparable geodynamic evolution during rollback processes. It is speculated that the arc of the western Alps formed due to westward rollback of a southeast-dipping slab (Figure 9.4a). This hypothesis explains the role of Oligocene extensional tectonism in the western Alps and the arcuate shape of the orogen. However, the concept of rollback is somewhat problematic in the tectonic framework of the western Alps because it requires that the internal massifs were located at the overriding plate above a southeast dipping oceanic slab, and that subduction and rollback of oceanic lithosphere was still taking place during the Oligocene (Figure 9.3 and Figure 9.4a). I will further discuss these issues in the next section.

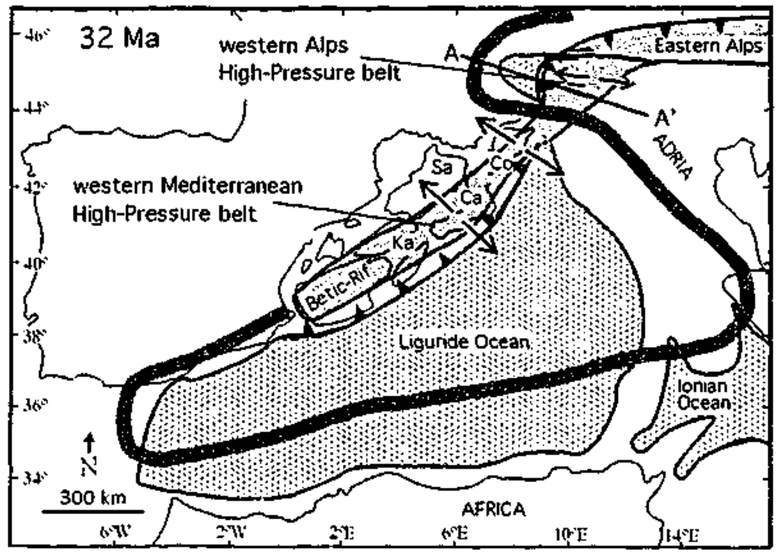


Figure 9.3. Schematic reconstruction of the western Mediterranean region during the Oligocene. Shaded areas indicate the location of the Alpine orogen, which was a contiguous HP belt stretching from the Rif-Betic to the western Alps. Note that the internal western Alps were located further to the southeast relative to their present position and were separated from the European foreland by a small oceanic embayment (see discussion in text). Thick curved line indicates the present-day geometry and position of the Alpine belt. Ca, Calabria; Co, Corsica; Ka, Kabylies; Sa, Sardinia. Line A-A' indicates cross-section plotted in Figure 9.4a.

9.4. Discussion

The role of extensional tectonism during the Oligocene has hitherto not been considered in geodynamic and kinematic evolutionary models of the western Alps. Three possible explanations are discussed here: (1) extension associated with internal readjustments within the accretionary wedge; (2) extension induced by a slab breakoff following lithospheric overthickening; and (3) extension induced by slab rollback.

Chemenda et al. (1995) have proposed a model for syn-orogenic extension and exhumation of HP rocks associated with internal readjustments within the orogenic wedge. Their model is based on a buoyancy-driven extrusion of orogenic wedges within the accretionary prism, involving the reactivation of frontal thrust faults as extensional detachments during an overall compressional regime. Nevertheless, evidence from the western Alps shows extensional contacts that have opposite dip directions with respect to frontal thrusts, indicating that these contacts are not reactivated frontal thrusts (Avigad et al., 2003). Therefore, Oligocene extensional tectonics during exhumation in the western Alps cannot be solely attributed to an extrusion model of the kind proposed by Chemenda et al. (1995).

An alternative explanation for the role of Oligocene extension may be linked to the overthickening of the crust and lithosphere, which led to HP metamorphism at the culmination of orogenesis at ~35 Ma (e.g. Lister et al., 2001). Von Blanckenburg and Davies (1995) have explained the occurrence of magmatism along the Insubric Line (Figure 9.1b) by the breakoff of a lithospheric slab beneath the central Alps. This model, however, did not consider the contemporaneous extensional tectonism in the western Alps. As discussed in this chapter, extension in the western Alps was accommodated along gently inclined extensional shear zones, which were active during rapid exhumation of UHP rocks. The operation of these shear zones during exhumation implies large amounts of horizontal motions, which were less likely to result solely from breakoff and vertical sinking of the slab. It is therefore possible that slab breakoff (involving a horizontal tear) or slab tearing (involving a vertical tear) was accompanied by horizontal migration of the subduction hinge (Figure 9.4b).

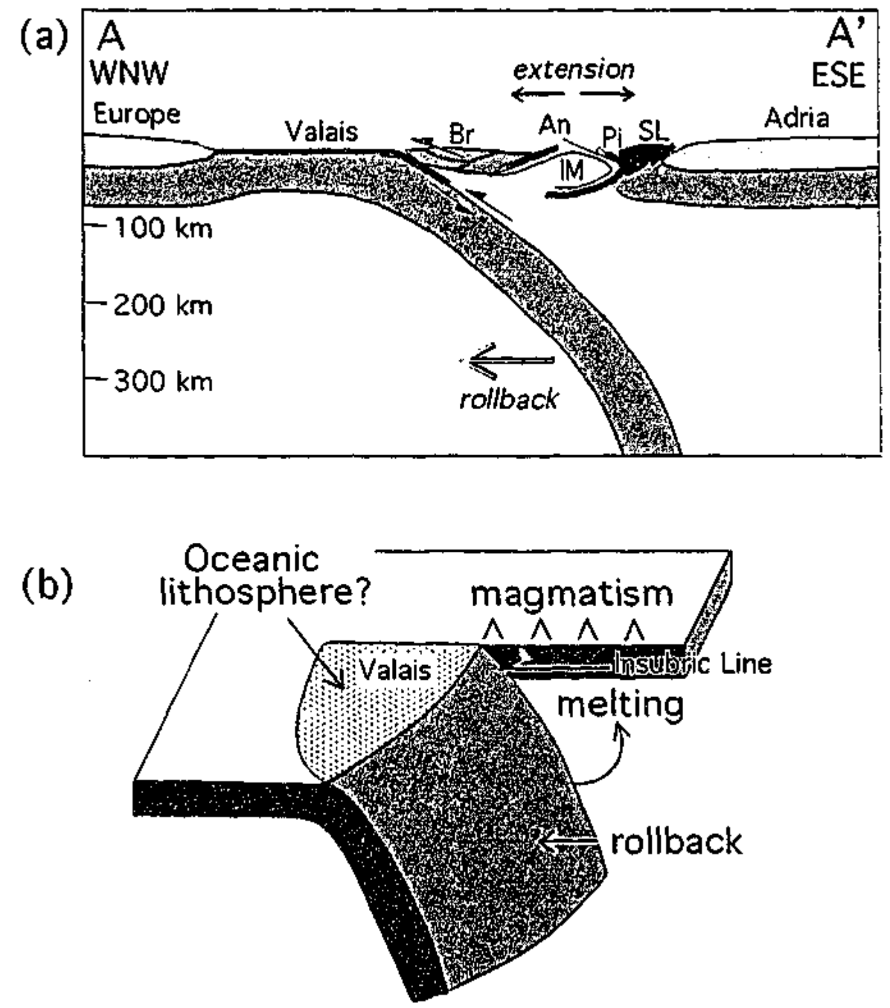


Figure 9.4. (a) Cross section through the Oligocene orogenic wedge in the western Alps (see Figure 9.3 for location) showing extension and exhumation of the Internal Massifs simultaneously with slab rollback. An, Antrona; Br, Briançonnais; IM, Internal Massif; Pi, Piedmont; SL, Sesia-Lanzo. (b) The geometry of the Oligocene slab showing tearing of the slab in the central Alps and rollback in the western Alps.

Extension in the orogenic belt is here linked to westward slab rollback (Figure 9.4a). This model requires that an oceanic subducting slab existed between the Oligocene orogenic wedge and the European margin. This oceanic embayment still existed in the Western Alps at 35 Ma (Figure 9.3) and was consumed during the Oligocene. This hypothesis contradicts the general agreement that the last Alpine ocean, the Valais Ocean, was entirely consumed by 35 Ma (e.g. Stampfli et al., 1998; Froitzheim, 2001). However, there is no direct evidence showing that the final closure of the Valais Ocean (sensu stricto) occurred before 35 Ma (Bousquet et al., 2002). It is accordingly proposed that the final closure of the Valais Ocean occurred later in the Oligocene, at ca. 30 Ma. The final closure of the Valais Ocean is also indicated by a transition from flysch to molasse sedimentation at ~30 Ma (Trümpy, 1980). If this scenario is true, then the collision at ~35 Ma may have occurred from the closure of an additional Alpine ocean, the Antrona Ocean (Platt, 1986), which separated the Internal Massifs from the Briançonnais Zone (Figure 9.1c).

The proposed model combines tearing of the slab in the central Alps and slab rollback in the western Alps (Figure 9.4b). Tearing of the slab beneath the central Alps triggered crustal and mantle melting along the plate boundary and resulted in a gravitational instability of the slab beneath the western Alps, which was now a narrow slab segment dipping towards the southeast or the east. This slab rolled back and led to the destruction of the internal orogenic belt by extensional processes. In this tectonic configuration, the Insubric Line in its early stages (35-30 Ma; Steck and Hunziker, 1994) was actually a transform fault with a dextral sense of motion, which accommodated the overall lateral migration of the arc of the western Alps.

9.5. Conclusion

Structural, metamorphic and geochronological evidence suggests that extensional tectonism affected the Internal Massifs of the western Alps during the Oligocene (34-30 Ma), immediately after the culmination of HP and UHP metamorphism at ca. 35 Ma. These observations are explained by the existence of an Oligocene east- or southeast-dipping slab, which was subjected to tearing beneath the central Alps and to westward rollback in the western Alps. The geodynamic model is consistent with the onset of dextral movement along the Insubric Line during the Oligocene, and the occurrence of widespread magmatic activity as a result of a slab tear. In the western Alps, however, the model requires reappraisal of earlier tectonic reconstructions, with final closure of the Valais Ocean at ca. 30 Ma, and the incorporation of an additional oceanic domain (the Antrona Ocean) in Tertiary subduction and collisional processes.

CHAPTER 10

CONCLUSIONS

If god had consulted me when he created the world, I would have recommended greater simplicity.

Alphonso "The Wise" (1221-1284)

10. CONCLUSIONS

10.1. General

The thesis aimed to provide better spatio-temporal resolution on tectonic processes that occurred during the tectonic evolution of the Alpine orogen in the western Mediterranean region. Such spatio-temporal constraints have been used to develop well-constrained reconstruction models that demonstrate that simple plate tectonic rules are not sufficient to account for the complex geodynamic and kinematic interactions that take place during orogenesis. It has been shown that convergent margins are subjected to deformation within a broad area, where large horizontal motions occur. These motions are primarily controlled by subduction rollback and the opening of back-arc basins, and by independent motions of small continental blocks. All these processes are interconnected; for example, subduction rollback would induce extension in the back-arc region, whereas accretion of allochthonous units could play a fundamental role in switching the tectonic mode from extension to crustal shortening.

It has also been shown that orogenic processes must be analysed in 3D, and in both crustal and lithospheric scales. The 3D structure of subducting lithospheric slabs plays a major role in the regional geodynamics, and can lead to acute changes within the orogenic wedge. In the Tyrrhenian Sea, for instance, the coincidence of tectonic processes at 6-5 Ma suggests that slab tearing was subsequent to a major accretionary event, probably because of the arrival of buoyant continental material at the subduction system. Slab tear, in turn, triggered rapid subduction rollback that resulted in the opening of the southern Tyrrhenian Sea.

An interesting outcome of this thesis is the possibility that the style of tectonism in the western Mediterranean is analogues to tectonic processes that occurred during the evolution of the arc of the western Alps. This segment of the Alpine orogen underwent a prolonged tectono-metamorphic history that involved several stages of deformation and distinct metamorphic events. Using the analogue example from the western Mediterranean region, I speculated that such complex interactions possibly involved rollback processes and episodic accretionary events.

10.2. Reflection on key issues

In this section, I will briefly respond to key issues that have been outlined at the beginning of this thesis (see Chapter 1).

10.2.1. Approach used to establish the kinematic framework for tectonic reconstructions

The methodological approach is based on constructing a hierarchy of spatio-temporal constraints ranging from global plate kinematics to regional geological and geophysical criteria. First order constraints are related to motions of major tectonic plates and are derived from relatively robust kinematic data (e.g. magnetic isochrons, hotspot tracks and palaeomagnetic data), whereas lower-order constraints are related to the motions of minor plates and allochthonous terranes or to the migration of subduction zones through time. Using these spatio-temporal constraints, I was able to establish well-constrained kinematic reconstructions, which adequately demonstrated tectonic complexities in the Alpine-Mediterranean region.

10.2.2. Role of large-scale convergent motions

Alpine orogeny has been controlled by convergence between Africa and Europe, which commenced during the Cretaceous Normal Superchron (120-83 Ma). Nevertheless, results of this thesis show that major tectonic processes within the orogen were not necessarily related to the relative convergence motion between the large plates. Rather, orogenic episodes often corresponded to large horizontal motions that occurred in the broad zone that separated the large converging plates. Moreover, the combination of subduction processes together with relatively slow convergence rates may have led to faster rates of subduction rollback, which in turn, resulted in the opening of back-arc extensional basins. In the western Mediterranean, the termination of rollback/back-arc extension processes was usually marked by an accretionary event, which was associated with tectonic transport directions that were oblique or even perpendicular to the direction of convergence (e.g. in the Apennine belt). It is therefore concluded that large-scale convergence motions are not sufficient to account for the history of deformation in the orogen.

10.2.3. Degree of horizontal motions

The spatio-temporal analysis shows that hundreds of kilometres of horizontal motions have occurred in the western Mediterranean since the Oligocene. The Calabrian block, for instance, has been displaced approximately 800 km from a position adjacent to the European margin (offshore southern France) to its present position in the southernmost part of the Italian peninsula. The primary mechanism for these large displacements is attributed to subduction rollback, which was estimated to occur at rates of 60-100 km/Myr during the opening of the Tyrrhenian Sea.

10.2.4. Effect of accretionary events

It appears that there is a possible link between accretionary events and episodic transitions from shortening to extension within the orogenic wedge. The occurrence of accretionary processes is likely to occur when relatively buoyant crustal material arrives at the subduction zone and impedes rollback processes. This process would result in the cessation of extensional processes (induced by subduction rollback) and to transition to an overall compressional regime. It has been further speculated that a major accretionary event can eventually lead to tearing of the subducting slab.

10.2.5. Formation of curved orogenic structures

A major outcome of this thesis is related to the observation that arcuate orogenic belts in the western Mediterranean region formed during the extensional destruction of an earlier high-pressure/low-temperature orogenic belt. The original belt was contiguous with the western Alps before the Oligocene, and was then subjected to oroclinal bending during subduction rollback. It has also been demonstrated that this process must have involved tearing of the subducting slab in several places. The formation of such tears may have further accelerated subduction rollback because they enabled sideways asthenosphereic flows, which acted against the negative buoyancy of the slab.

10.2.6. The western Alps in the context of the western Mediterranean tectonics

The last two chapters of this thesis pointed out on major problems with regard to the current paradigm on the tectonic evolution of the western Alps. These problems are predominantly concerned with evidence for the occurrence of several distinct high-pressure events in the internal zone of the western Alps and for seemingly episodic switches between shortening and extension in the orogenic wedge. I therefore propose to test the evolution of the western Alps by means of western Mediterranean-type of tectonism. In this respect, the geometrical analogy between the arcuate shape of the western Alps and the arcuate western Mediterranean belts is truly a striking feature. However, reconstruction of the evolution of the western Alps requires further studies and high-quality spatio-temporal constraints.

10.3. Suggestions for further research

The outcomes of this thesis can lead to a wide range of research proposals. Several examples are briefly discussed below.

10.3.1. Refinements of existing reconstruction models

The reconstruction approach used in this work was based on trial-and-error experiments using the best available constraints. The resulting reconstruction models must be subjected to ongoing scrutiny in order to expose existing flaws in the reconstruction process or to refine numerous compexities by utilising additional spatio-temporal constraints. This process is crucial in order to maintain strong relationships between models and nature.

10.3.2. Obtaining high-quality spatio-temporal constraints on specific problems

The reconstruction process exposed numerous problems that can be tested by implementing adequate techniques (e.g. geochronology, palaeomagnetism, seismic tomography etc). Two examples from the field of palaeomagnetism are given below.

Recent palaeomagnetic results from the western Alps (Collombet et al., 2002) show increasing amounts of Neogene counterclockwise rotations from north to south along the strike of the orogen. This pattern of rotations may correspond to asymmetric oroclinal bending during late-orogenic (and probably extensional) stages in the evolution of the arc of the western Alps. This view is in contrast with the current paradigm on the formation of the arc of the western Alps, and can be further supported by systematic palaeomagnetic testing in key localities.

Another example is related to the motion of Adria. As discussed earlier, the Adriatic problem has received considerable attention in the literature. However, the structural history of one of the key areas, the Scutari-Pec Fault Zone in Albania, has remained relatively unconstrained. A combined structural and palaeomagnetic study can lead to profound implications on the kinematics of the central Mediterranean and the Alpine orogeny. Based on my existing reconstruction, I predict that the Scutari-Pec Fault Zone was active as a major lithospheric discontinuity, accommodated by dextral strike-slip motion during the Neogene, clockwise block rotations around vertical axes south of the fault zone, and no significant rotations north of the fault zone (e.g. see movie 'Adria.mov' in Appendix 2). Testing this hypothesis can be addressed by palaeomagnetic sampling from different localities north and south of the fault zone, combined with structural mapping of the area to elucidate the role of strike-slip activity and its relative timing.

In-situ dating of deformation (e.g. Freeman et al., 1997) can be used to test specific hypotheses, particularly with regard to the timing of deformational events in the western Alps. For example, dating extensional shear zones in the Internal Massifs can verify, or refute, the significance of widespread extensional tectonism in the western Alps at 34-32 Ma. The timing of deformation in the Valais and the Antrona ophiolites are also crucial for testing the working hypothesis presented in chapter 9.

10.3.3. Extending the reconstruction model to the eastern Mediterranean

The Alpine belt in the eastern Mediterranean (e.g. Turkey and Greece) provides a perfect example for an episodic collisional orogen associated with rollback/back-arc extension processes. The preservation of ophiolitic sutures and the amount of published works from this region provides an opportunity to extend the scope of the existing models. An eastern Mediterranean reconstruction model can incorporate large amounts of published data, complemented by additional structural, geochronological and palaeomagnetic data from key localities. The combination of such project with the existing western Mediterranean reconstructions will result in a visual reconstruction model of the whole Mediterranean region.

10.3.4. Applying a similar reconstruction philosophy in Proterozoic reconstructions

The reconstruction philosophy discussed in this thesis can be used to develop well-constrained reconstruction models in other tectonic environments, especially where robust constraints from the present-day oceans are lacking (i.e. pre-Mesozoic reconstructions). Proterozoic reconstructions, for example, can involve implementation of palaeomagnetic data, which will be used as kinematic boundary conditions, and will be tested against additional spatio-temporal constraints. These constraints can include: (1) correlations between different parts of the orogenic belts; (2) time and space constraints on subduction processes; (3) distribution of extensional back-arc basins; and (4) spatio-temporal distribution of mineral deposits. The reconstruction can be developed progressively by testing the admissibility of alternative scenarios, and by using reconstruction software packages for interactive modelling. The reconstruction can lead to a breakthrough in continental reconstruction, as its scope will take into account various tectonic processes, such as subduction rollback, back-arc extension and accretion of allochthonous terranes. All these processes have not been considered hitherto in Proterozoic reconstructions.

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APPENDICES

All appendices are found in the attached CD-ROM.

Appendix 1 - PlatyPlus files

This directory includes the following sub-directories:

- Af_Eu
- Atlantic
- Model1
- Model2
- Model3
- Model4
- Model5
- Model6
- Model7
- Model8
- Model10
- Model11
- Model12
- Model13
- Model14
- Model15
- Model16
- Model19
- Model20
- Model21
- Model22
- Model23
- Model24
- Model25
- Model26
- Model27
- Model28
- Model29
- Model30
- Model31

The directories 'Model1' to 'Model 31' are all slight modifications of the tectonic model for the western Mediterranean since the Oligocene (see Chapter 5). The results of Chapter 5 are predominantly reflected in 'Model 26'.

The directory 'Af_Eu' includes information on the motion of Africa, Iberia and Europe (Chapter 3)

The directory 'Atlantic' includes information on magnetic isochrons in the Atlantic Ocean (Chapter 3).

All PlatyPlus files are text (ASCII) files. They are divided into three categories:

1) object (.bd) files;

2) knox files; and

3) motion files

Object files

Object (.bd) files consist of lists of latitudes and longitudes, which outline the boundary of PlatyPlus objects (i.e. continents, continental platforms, terranes, magnetic isochrons, etc).

An example of an object file ('Southern_Alps.bd') is shown below:

READ	05	close	0		
31	Southern Alps				
4000.00	0.00	0.00	200.00	1001	0.00
45.567	7.913				
45.702	8.162				
45.789	8.347				
45.822	8.558				
45.853	8.842				
45.851	9.113				
45.799	9.371				
45.709	9.746				
45.641	10.095				
45.607	10.246				
45.636	10.322				
45.713	10.395				
45.884	10.533				
46.079	10.655				
46.275	10.783				
46.437	10.893				
46.380	10.712				
46.299	10.406				
46.249	10.077				
46.247	9.665				
46.237	9.365				
46.239	8.963				
46.227	8.652				
46.078	8.353				
45.956	8.162				
45.915	8.088				
45.761	8.038				
45.567	7.913				

The first line includes: "READ" which is a compulsory comment ; "NUMBER" which identifies a family of objects; "close" or "open" which define whether the object is close (e.g. a continent) or open (e.g. a magnetic isochron); and a compulsory "zero".

The second line includes a "NUMBER" which indicates the number of pairs of numbers in the following lines (including the third line). The rest of the line is an "optional comment"

The first two numbers in the third line are "4000.00" and "0.00". They indicate that this object file is time windowed; the third "NUMBER" indicates the time (in Ma) of the last appearance of the object ; the fourth "NUMBER" indicates the time (in Ma) of the first appearance of the object (object files that are not time windowed do not include the first four numbers of the third line); the fifth "NUMBER" indicates the colour of the object; the sixth number is a compulsory "zero".

The following numbers include pairs of latitudes and longitudes.

Knox files

A knox file consists of a summary of all the object files to be plotted in a PlatyPlus model. Below is an example of a knox file ('Af_Eu.knox'):

File /home/gawler/users/gideon/PLATY270900/Af_Eu/Af_Eu.knox
Objects-Directory /home/gawler/users/gideon/PLATY270900/Af_Eu

OBJECT_NAME	PARENT_NAME	BOUNDARY_FILE_NAME
Gearth	None	GEarth.bd
All	Earth	GEarth.bd
Africa	All	Africa.bd
G05Africa	Africa	Africa.bd
G05Southern_Alps	Africa	Southern_Alps.bd
Eurasia	All	Eurasia.bd
G11Eurasia	Eurasia	Eurasia.bd
G11Britain	Eurasia	Britain.bd
G11Ireland	Eurasia	Ireland.bd
END		

The first line indicates the exact location of the file.

The second line indicates the location of the directory that includes the object files.

Each of the following lines includes: (1) the object name; (2) the hierarchical order (for example, "Southern Alps" is an object in a family of objects titled "Africa"); and (3) the name of the matching object file.

The last line is the word "END"

Motion files

Motions files consist of rotational parameters (i.e. latitude and longitude of Euler poles of rotations and rotational angles).

An example of a motion file with absolute motions ('Af_Eu_HS.motion') is shown below.

Total No_object

5	5	49.28	-40.32	-0.97	{This study
5	10	49.20	-40.49	-1.95	{This study
5	15	46.87	-44.98	-3.02	{This study
5	20	45.52	-46.38	-4.11	{This study
5	25	41.94	-46.61	-5.26	{This study
5	30	39.57	-46.75	-6.43	{This study
5	35	38.04	-46.80	-7.65	{This study
5	40	37.07	-46.45	-8.83	{This study
5	45	36.46	-46.58	-9.87	{This study
5	50	35.97	-46.96	-10.83	{This study
5	55	35.87	-47.14	-11.73	{This study
5	60	35.31	-47.16	-12.89	{This study
5	65	34.48	-48.92	-14.23	{This study
5	70	33.70	-51.00	-15.42	{This study
5	75	31.65	-51.55	-16.95	{This study
5	80	30.00	-52.17	-18.60	{This study
5	85	29.38	-51.96	-20.04	{This study
5	90	28.74	-51.60	-21.41	{This study
5	95	27.55	-50.78	-22.72	{This study
5	100	26.48	-50.07	-24.04	{This study
5	105	26.42	-49.50	-25.16	{This study
5	110	26.37	-48.97	-26.28	{This study
5	115	26.59	-48.41	-27.33	{This study
5	120	26.79	-47.89	-28.39	{This study
11	5	-70.8393	86.982	0.638903	{This study
11	10	-59.7205	127.327	1.55797	{This study
11	15	-61.2313	92.7561	2.08261	{This study
11	20	-57.25	91.3484	2.76655	{This study
11	25	-49.0803	77.0025	2.77502	{This study
11	30	-39.9015	67.4035	2.92727	{This study
11	35	-32.7976	63.5127	3.23281	{This study
11	40	-28.4496	63.1605	3.54898	{This study
11	45	-23.8599	60.5817	3.89346	{This study
11	50	-24.8273	60.2341	4.38741	{This study
11	55	-26.2186	59.3677	5.0313	{This study
11	60	-28.7402	68.3955	5.7861	{This study
11	65	-29.3464	76.4509	7.01532	{This study
11	70	-25.2892	72.7482	7.77549	{This study
11	75	-15.4551	70.9739	8.58108	{This study
11	80	-5.97795	66.6457	9.74361	{This study
11	85	1.53274	61.5431	11.0518	{This study
11	90	8.66003	56.9394	12.7874	{This study
11	95	14.7814	53.8777	14.7327	{This study
11	100	19.1788	51.4252	16.8763	{This study
11	105	21.9671	48.2229	18.9455	{This study
11	110	23.7572	46.0922	20.9549	{This study
11	115	25.0509	44.0608	22.8958	{This study
11	120	26.062	42.3364	24.8754	{This study

Each line includes five numbers. The first number is the identification number of the family of objects to be rotated; the second number indicates the time (in Ma) at which rotation is applied; the third and the fourth numbers indicate the latitude and the longitude of the Euler pole of rotation, respectively; the fifth number indicates the rotational angle (relative to the present position of the object). The rest of the line is a comment.

Below is an example of a motion file with relative motions ('Af_Eu.motion'):

Total Eurasia 11

05	11	5	-13.9873	158.228	0.548543	{This study
05	11	10	-13.9089	158.039	1.09475	{This study
05	11	15	-14.7748	162.887	1.53213	{This study
05	11	20	-17.2716	164.307	2.12614	{This study
05	11	25	-24.3237	161.899	3.57059	{This study
05	11	30	-27.2359	161.108	5.04152	{This study
05	11	35	-28.8937	160.766	6.40983	{This study
05	11	40	-30.102	160.568	7.60205	{This study
05	11	45	-30.9609	160.614	8.79721	{This study
05	11	50	-30.0189	160.714	9.61363	{This study
05	11	55	-28.8313	161.897	10.3875	{This study
05	11	60	-28.7309	163.016	10.5849	{This study
05	11	65	-28.8156	163.927	10.6366	{This study
05	11	70	-29.7049	163.752	11.8735	{This study
05	11	75	-32.6604	164.459	13.8412	{This study
05	11	80	-35.0188	165.096	16.4589	{This study
05	11	85	-36.8819	166.302	19.3539	{This study
05	11	90	-38.6826	167.702	22.5568	{This study
05	11	95	-40.0061	168.876	25.7721	{This study
05	11	100	-41.0257	169.888	28.9946	{This study
05	11	105	-41.8549	170.771	32.225	{This study
05	11	110	-42.5148	171.591	35.1185	{This study
05	11	115	-43.0863	172.345	37.9504	{This study
05	11	120	-43.5956	173.037	40.7849	{This study

The first line includes the name and the identification number of the object, to which motions are relative to.

Each of the following lines includes six numbers. The first number is the identification number of the family of objects to be rotated; the second number indicates the identification number of the object, to which motions are relative to; the third number indicates the time (in Ma) at which rotation is applied; the fourth and the fifth numbers indicate the latitude and the longitude of the Euler pole of rotation, respectively; the sixth number indicates the rotational angle (relative to the present position of the object). The rest of the line is a comment.

Appendix 2 - Movies

Movies are available in QuickTime formats. The application can be obtained from: <http://www.apple.com/quicktime/download/>

Western Mediterranean movies

The following movies are available:

1) 'W_Med.mov'

The movie shows the tectonic evolution of the western Mediterranean since the Oligocene. See detailed discussion in Chapter 5.

2) 'Ligurian.mov'

The movie is a close-up of the western Mediterranean movie in the Ligurian Sea (see Chapter 5).

3) 'Alboran.mov'

The movie is a close-up of the western Mediterranean movie in the Alboran Sea (see Chapter 5).

Additional movies

The following movies are available:

1) 'Adria.mov'

The movie shows internal deformation at the margin of Adria associated with the opening of the Tyrrhenian Sea and the Aegean Sea (e.g., see Chapter 6). Note that a considerable movement along the Scutari-Pec Fault Zone (separating the Dinarides-Albanides and the Hellenides) is predicted (this issue is not discussed in this thesis but is suggested for further research in Chapter 10).

2) 'Af_Eu.mov'

The movie shows the motion of Africa with respect to Europe for the last 150 Myr based on data of Dewey et al. (1989) (see Chapter 3).

3) 'Af_Eu_HS.mov'

This movie shows the motions of Africa and Europe in a hotspot frame of reference (see Chapter 4).

4) 'Af_Eu_Ib_Ad.mov'

The Movie shows relative motions of Africa (with Adria attached to it), Europe and Iberia as calculated in Chapter 3. The red outline of Africa indicates the motion of Africa with respect to

Europe based on data from Dewey et al. (1989).

5) 'Biscay.mov'

The movie shows the opening of the Bay of Biscay at the tectonic framework of the Northern Atlantic Ocean and the western Mediterranean region (see Chapters 3 and 5).

6) 'N_Atlantic.mov'

This movie shows a plate tectonic reconstruction of the Northern Atlantic Ocean since the Jurassic (see Chapter 3).

7) 'N_Atlantic_isochrons.mov'

This movie shows the opening of the Northern Atlantic with respect to the age of magnetic isochrons (see further discussion in Chapter 3). Data are after Cande et al. (1989).