Global kinematic constraints to the tectonic history of the Mediterranean region and surrounding areas during the Jurassic and Cretaceous

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Abstract: The formation of small fragments of continental lithosphere, which rift away from a passive margin and are carried toward a trench together with the surrounding oceanic crust, is a characteristic of many collisional settings, in particular of the northern margin of Gondwana during the Mesozoic. This motion, though chaotic in appearance, can be described rigorously in terms of plate kinematics driven by local temporal variations in the relative velocity field between the main colliding plates. In this model proposed here, the rifted continental fragments approach the trench with the same stationary velocity as the oceanic lithosphere in which they are embedded, while the spreading centers that separate these microplates from the rifted continental margin, either speed up or slow down in order to compensate for small variations of the velocity field between the major colliding plates. Hence, the oceanic leading edge of a subducting plate may separate from its continental part and move independently to ensure a constant convergence rate at the trench. The application of this principle to the complex tectonic history of the Mediterranean region during Jurassic and Cretaceous times is performed starting from a revised global plate motion model. A set of maps illustrating the regional velocity and acceleration fields is presented for nine major phases from the Bajocian through the Maastrichtian. These maps provide new constraints that may be helpful for the construction of plate tectonic models of the Tethyan realm. New insights into some of the major tectonic events that occurred during the Jurassic and the Cretaceous in the Mediterranean region are gained from the correlation between kinematic events and geologic evidence.

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Introduction

The Mediterranean region traditionally represents a difficult but intriguing study area for plate tectonic modeling. During the last three decades many workers from almost all disciplines of geology and geophysics have tried to illustrate the geologic evolution of this region from different perspectives and using a variety of geological or geophysical data. It is meaningful that over 700 Jurassic, Cretaceous and Cenozoic paleopoles have been determined for the plates and microplates of the Mediterranean region, whereas only 126 paleopoles are listed in the Global Paleomagnetic Database (GPMDB) for the North American craton [McElhinny and Lock, 1990].

The main problem that researchers have faced has been to identify and map the dozens of microplates involved in the Alpine-Himalayan collision. It is often difficult to identify the ancient boundaries between these tectonic elements (e.g., Pelagonian, the Austroalpine). However, even when the boundaries are well established, a high degree of uncertainty remains with regard to the original undeformed shape of these small tectonic blocks. Additionally, for some of the identified elements, it is still not clear whether they are continental or oceanic terranes. For instance, it is still uncertain if the Panormide carbonate platform was built on a rifted continental margin or if it developed on thickened oceanic crust.

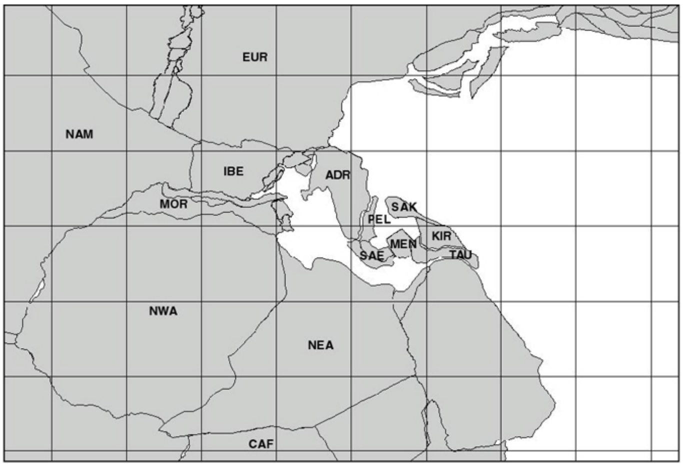
A second problem arises from the absence of direct kinematic constraints for most of the Mediterranean microplates. The best way to accurately assess the rotation parameters of a plate or smaller tectonic element is through the identification of marine magnetic anomalies. Unfortunately, the few and sparse remnants of Mesozoic and Cenozoic oceanic crust of the Mediterranean basin do not provide sufficient information for this purpose. This lack of data was partially overcome by the relatively high number of paleomagnetic studies, which were followed by several quantitative reconstructions of the Mesozoic and Cenozoic paleogeography of the Tethyan realm [e.g., Van den Berg and Zijderveld, 1982 ; Lauer, 1984 ; Morris and Tarling, 1996 ; Channel, 1996 ; Muttoni et al., 1996].

Previous investigations of the tectonic history of the Mediterranean region have followed either qualitative or quantitative approaches. Studies that fall in the first category are based prevalently on geologic data. Examples of this kind of reconstructions can be found in Robertson and Dixon [1984], Şengör et al. [1984], Dercourt et al. [1986], Stampfli et al. [1991], Dercourt et al. [1993], Robertson et al. [1996], Robertson [1998], and Stampfli et al. [1998]. The second group comprises works that highlight the kinematic and structural aspects of the problem [e.g., Dewey et al., 1973 ; Dewey et al., 1989 ; Alvarez, 1991], or those studies that are based on interpretations of paleomagnetic data.

In this paper a new quantitative model of the tectonic history of the Mediterranean region is proposed, which is based on the analysis of variations of the global plate motions and plate circuits, and the identification of corresponding constraints to the local kinematics of small microplates at the interface between Africa and Eurasia during the Jurassic and the Cretaceous. Unlike other classic quantitative studies [e.g., Dewey et al., 1989], theoretical arguments and data are presented to support the idea that the local velocity of intervening microplates is constrained by global variations in plate velocities at stage boundaries (and the corresponding stress fields) and not, as proposed by previous authors, by the path of relative motion of Africa with respect to Eurasia. Therefore, we propose that the relationship between global plate motions and local microplate kinematics is, at best, indirect, whereas a direct connection exists between variations of relative velocity fields of the major plates and the local deformation pattern as expressed by microplate formation and subsequent motion.

The starting point of this study is a revised global plate tectonic model for the Jurassic, Cretaceous and Cenozoic times [Schettino and Scotese, 2000 ; Schettino and Scotese, 2001 ; Schettino and Scotese, 2002, in prep.], and some basic assumptions about the paleogeographic configuration of the western Tethyan realm at the beginning of Jurassic. These assumptions are based on the geological and geophysical evidence for the existence of a Late Permian – Triassic oceanic seaway that separated Adria from the Northern Gondwana margin (Fig. 1). The Imerese and Lagonegro basins in the western Mediterranean [Zappaterra, 1994 ; Pescatore et al., 1999], and the Levant basin in the easternmost Mediterranean [Garfunkel, 1998 ; Robertson, 1998] are well documented domains characterized by pelagic sedimentation during the Triassic. Stampfli et al. [1991] pointed out that the formation of this basin could have been related to the Cimmerian rifting during the Late Permian. However, the Cimmerian rift model requires that an eastward widening of the ocean formed as a result of the counterclockwise rotation of the Cimmerian blocks as they were subducted towards the Eurasian margin. Conversely, we propose an alternative scenario, which results in a westward-widening ocean that opened by clockwise rotation of the Mediterranean terranes with respect to Northern Gondwana. This rifting event may be associated with the extensive Late Triassic rifts of Central Pangea [Veevers, 1989], which culminated at the beginning of Jurassic (about 200 Ma) in the eruption of an enormous quantity of basalts in a geologically short interval and the formation of the Central Atlantic Magmatic Province (CAMP) [Schlische, 1993 ; Olsen, 1997 ; Withjack et al., 1998 ; Hames et al., 2000]. In many aspects, our assumed starting configuration for the Eastern Mediterranean region is similar to the one proposed by Robertson [1998].

Figure 1. Reconstruction of the Tethyan realm at 200.0 Ma (lower Sinemurian)



Labeled tectonic elements are essential protagonists of the Jurassic and Cretaceous history of the Mediterranean region. NWA = Northwest Africa; NEA = Northeast Africa; EUR = Eurasia; MOR = Morocco; IBE = Iberia; ADR = Adria; PEL = Pelagonian; SAE = Southern Aegean; MEN = Menderes; SAK = Sakarya; KIR = Kirsehir; TAU = Taurides; CAF = Central Africa.

The next two sections of this paper illustrate data and methods that were used to construct the proposed model. This is followed by a discussion of the correlation between the kinematic behavior and the geologic events that characterized the geologic history of the Tethyan realm during the Jurassic and the Cretaceous.

The basic kinematic model

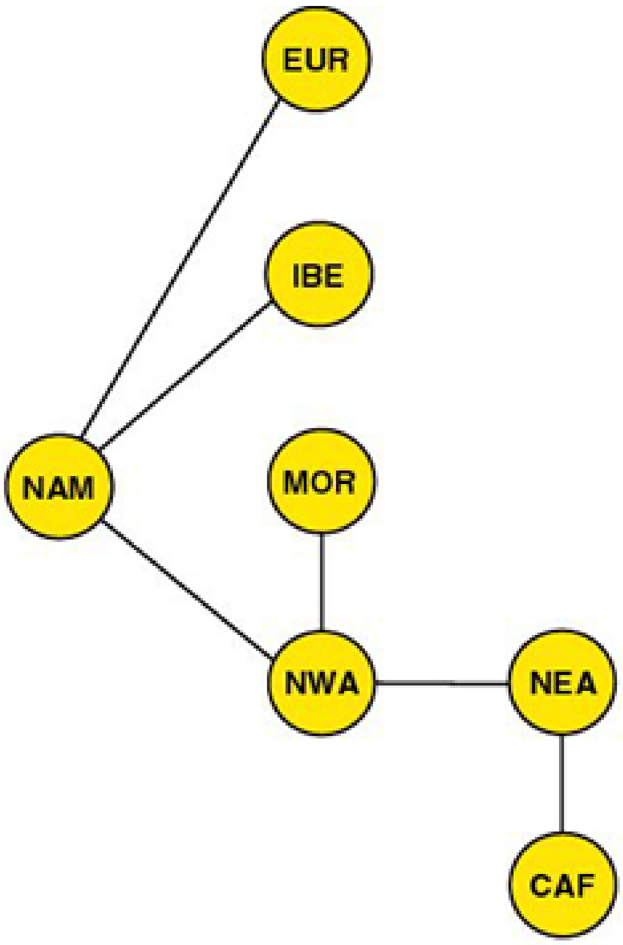
In this paper a simplified set of tectonic elements is used to illustrate the kinematic evolution of the Mediterranean region during the Jurassic and the Cretaceous. In particular, attention is focused on the formation of new plate boundaries in the Tethys realm in response to temporal variations of the velocity field between the large surrounding plates. Therefore, a complete and detailed classification of the microplates involved is not necessary. As the strongest phase of deformation occurred in the Cenozoic, four simple or composite blocks can be used to illustrate the main paleogeographic features. These tectonic elements are: Adria, the Pelagonian terrane, Southern Turkey, and Northern Turkey.

We consider Adria (Fig. 1) as the undeformed assemblage resulting from Permian and Triassic rifting episodes, and comprising the present South Alpine and Northern Apenninic domains, the Apulian promontory and the undeformed Panormide platform (that is, the Southern Apenninic domain of Dewey et. al. [1989] or Turco and Zuppetta [1998]). The Pelagonian terrane is defined by the crystalline and sedimentary formations of Northwest Macedonia and Central Greece that are bounded by the Sub-Pelagonian zone and the Pindos ophiolites to the West, and by the basal thrust of the Vardar zone to the East [Mountrakis, 1984]. Southern Turkey is considered here as the undeformed assemblage comprising the Menderes Massif, the Cycladic units, Crete and the Taurides belt. Finally, Northern Turkey is viewed as a composite microplate comprising the Sakarya and Kirsehir units. It is assumed that all these microplates behaved as rigid elements during the Jurassic and the Cretaceous.

A kinematic model for the major plates surrounding the Mediterranean region is shown in table 1.

The initial fit of North America to Northwest Africa is based on the Blake Spur Magnetic Anomaly of Eastern North America and a pre-stretching match of the continental margins [Klitgord and Schouten, 1986], whereas the North Atlantic fit is based on the reconstructions of Rowley and Lottes [1988] for the North Atlantic and Arctic regions. The starting configuration of the continents around the Tethys at the beginning of the Jurassic (200 Ma, Lower Sinemurian) is illustrated in Figure 1. This fit also takes into account of the pattern of deformation in Africa during the Cretaceous [Wilson and Guiraud, 1992 ; Genik, 1992 ; Guiraud and Maurin, 1992], by restoring the North African blocks to their initial orientations with respect to Central Africa. This includes the closure of the West African Rift System (WARS), and the restoration of transcurrent motion along the Central African Shear Zone (CASZ).

Figure 2. Jurassic and Cretaceous plate motions



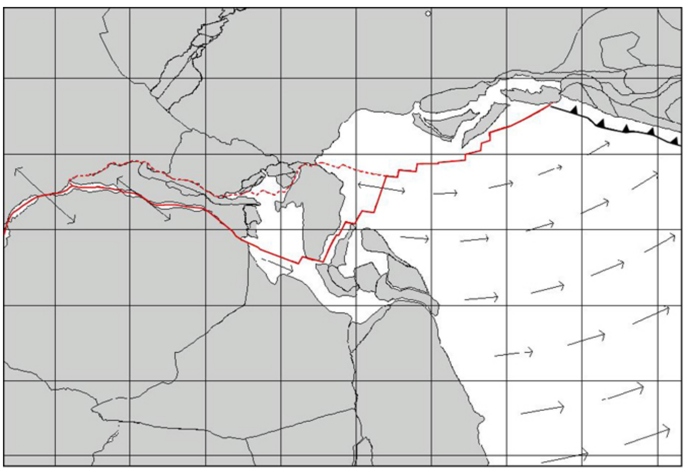
The global circuit used to describe the Jurassic and Cretaceous plate motions around the Atlantic and Tethys oceans. The reference plate is Central Africa, which has been oriented with respect to a paleomagnetic reference frame using the method of Schettino and Scotese [2001]. According to this method, a synthetic and smoothed Apparent Polar Wander Path (APWP) for Central Africa was generated by restoring paleopoles from other continents to the Central African reference frame via the global rotation model [e.g., Besse and Curtillot, 1991 ; Bocharova and Scotese, 1993 ; Van der Voo, 1993]. Further details about this method will be published elsewhere [Schettino and Scotese, in prep.].

Finite rotations that describe the pattern of seafloor spreading in the Central Atlantic have been calculated according to the digital compilation of isochrons of Royer et al. [1992] and the ocean floor age grid [Müller et al., 1997]. The finite rotations of Rowley and Lottes [1988] were adopted for the North Atlantic region because these rotations give the best results when paleopoles from Eurasia are transferred to Central Africa via the global circuit.

Figure 2. Global circuit for the plates surrounding the Atlantic and Tethys oceans.

An important feature of the model proposed in this paper is the definition of the Atlas and Rif-Tell regions as a composite tectonic element which rifted from the Northwest African plate during the Late Triassic – Early Jurassic. In fact, the inclusion of this terrane helps to better explain the connection between the early stages of rifting in the Central Atlantic and the tectonic activity in the Mediterranean region during the Late Triassic and the Early Jurassic. A summary of the main geologic and structural features of the Atlas region can be found in Beauchamp et al. [1996]. Here attention is focused on the Late Triassic - Early Jurassic phase of extension and rifting, which produced an aulacogen in the High Atlas region and transtensional features in the Eastern Mediterranean. These events are correlated with the formation of the Newark Basin in the eastern United States [e.g. Olsen, 1997] and the Pindos Basin in the eastern Mediterranean [Degnan and Robertson, 1998]. These three extensional systems were part of a single divergent boundary that crossed Central Pangea during the Late Triassic and the Early Jurassic (Fig. 3). This boundary existed for about 50 Myrs until an abrupt change in the relative velocity of Africa with respect to North America resulted in the cessation spreading along the Pindos spreading center at 172 Ma, as evidenced by youngest radiometric ages of the Pindos ophiolites [Spray et al., 1984]. This event also coincides with extinction of the Atlas rift system as a major plate boundary and the northward jump of the shear zones between Iberia and Morocco.

Figure 3. Reconstruction at 172.0 Ma (Upper Bajocian)



Vectors represent direction and magnitude of the relative velocity field between Northern Gondwana and Laurasia. Dashed line represents an incipient spreading center. Red line marks the assumed trend of the Pindos ridge and the eastern North America and Atlas rift systems.

Velocity fields and acceleration fields

The plate reconstructions proposed in this paper were made using an interactive software tool for plate tectonic modeling designed by the first author [PCME, Schettino, 1998]. This program also allows to plot relative or absolute velocity and acceleration fields for a given rotation model. If A and B are two conjugate plates and we indicate the instantaneous Euler vector (or angular velocity) of A relatively to B at time t as $\Omega_{AB} = \Omega_{AB}(t)$, then the relative velocity field between these plates at an arbitrary location r is given by:

$$\mathbf{v}_{AB}(\mathbf{r}) = \Omega_{AB} \times \mathbf{r}$$

Equation 1

Hence, in order to calculate a velocity field, we need an estimate of the angular velocity for the considered time t. This estimate is accomplished as follows. A finite rotation pole is a representation of the orthogonal transformation matrix $R_{AB}(t_k)$ that is necessary to move a plate A from its present location to a position with respect to plate B at time t_k [e.g., Cox and Hart, 1986]. Typically, in the case of Mesozoic and Cenozoic models, times t_k correspond to chrons associated to identified magnetic anomalies. Two successive times t_k and t_{k+1} ($t_k > t_{k+1}$) are respectively the lower and upper age limits of a stage. If we consider the reference plate B as fixed to its present day location, the relative motion of A during a stage $[t_k,t_{k+1}]$ can be described through a stage pole matrix $S_{AB}(t_k,t_{k+1})$, given by:

$$\boldsymbol{S}_{AB}(t_k, t_{k+1}) = \boldsymbol{R}_{AB}(t_{k+1})\boldsymbol{R}_{AB}^{-1}(t_k)$$

Equation 2

This quantity represents the rotation that is necessary to carry a plate A from the position assumed at time t_k to the position assumed at time t_{k+1} relatively to a reference plate B which is kept fixed in its present day location.

It must be pointed out that a stage pole matrix does not represent an instantaneous axis of rotation for the motion of A relatively to B between times t_k and t_{k+1}. In fact, by definition the plate A is considered at rest in the present geographic reference frame, whereas any instantaneous rotation of B with respect to A at time $t \in [t_k,t_{k+1}]$ must be calculated relatively to the "correct" location of A at that time. An interesting property of the stage poles is that by Equation (2) we can calculate an unknown finite rotation of A relative to B at time t_{k+1}, provided that the finite rotation at time t_k is known, and an estimation of the stage pole matrix between t_k and t_{k+1} is available. In fact,

$$\boldsymbol{R}_{AB}(t_{k+1}) = \boldsymbol{S}_{AB}(t_k, t_{k+1})\boldsymbol{R}_{AB}(t_k)$$

Equation 3

We will see that this Equation plays a key role in the description of the kinematics of the Mediterranean region. An estimation of the instantaneous angular velocity Ω_{AB} at time t can be obtained as follows. We first calculate the stage pole vector $e_{AB}(t_k,t_{k+1})$ and the stage angle associated to the matrix $S_{AB}(t_k,t_{k+1})$. This pole represents the point of intersection between the rotation axis corresponding to $S_{AB}(t_k,t_{k+1})$ and the Earth's surface, whereas the stage angle represents the total rotation angle about this axis. If we now rotate the vector $e_{AB}(t_k,t_{k+1})$ using the total reconstruction matrix for the reference plate B at time t, the resulting vector represents the instantaneous Euler pole $\omega_{AB}(t)$ at time t. Finally, in order to obtain the angular velocity rate Ω_{AB} we divide the total stage angle by t_k-t_{k+1}.

The method described above allows to calculate velocity fields for any time in the geologic past, provided that a rotation model is available for the considered time period. It also allows to detect discrete variations of the velocity fields at stage boundaries. These variations are representative of acceleration fields and corresponding stress fields that act on the lithosphere at the transition times between two successive stages. The acceleration field between two plates A and B at the transition time t is defined as:

$$\boldsymbol{a}_{AB}(\boldsymbol{r}, t) = \boldsymbol{v}_{AB}(\boldsymbol{r}, t + \delta t) - \boldsymbol{v}_{AB}(\boldsymbol{r}, t - \delta t)$$

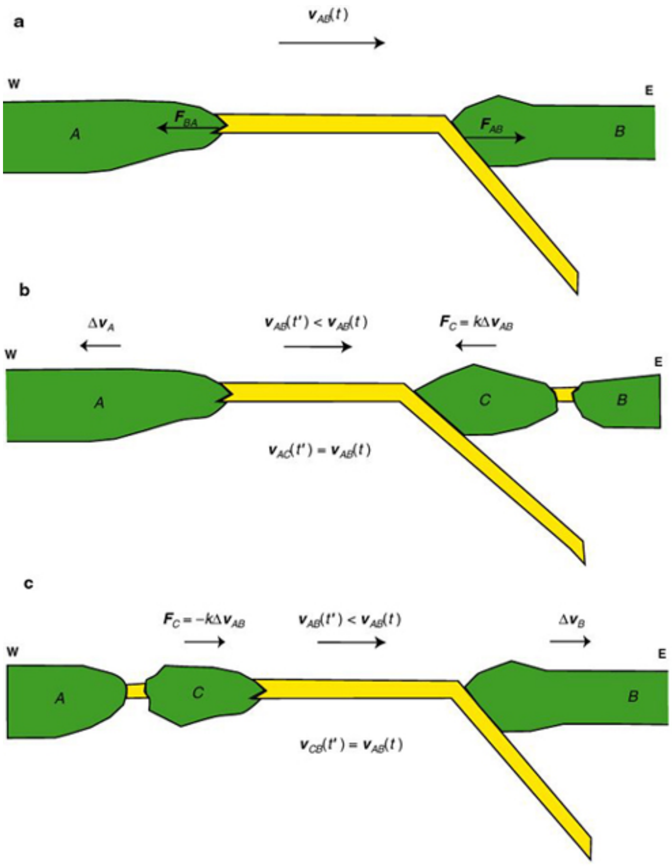
Equation 4

where δt represents a short time interval about the considered stage boundary (e.g. 0.1 Myrs). It should be noted that the term "transition time" refers here to any stage boundary in the rotation model that modifies one or more relative velocities along a circuit path. For instance, if plates A and B are related in the model through a third plate C, an acceleration between A and B may result either from a stage boundary between A and C or a transition between C and B.

We can now illustrate the theoretical framework that was adopted for the reconstruction of the tectonic history of the Mediterranean region proposed in this paper. In the following discussion we assume that the motion of intervening slivers at the interface between two main colliding plates is subject to precise physical laws that require periodic changes in the boundary geometry and formation of broad deformation zones. These changes have the primary role in restoring a state of equilibrium at the trench systems, in response to temporal variations of the main velocity field.

Consider an oceanic plate A that is subducting beneath a plate B. At the equilibrium the net torque exerted on the plate boundary is zero, because all the forces applied at the trench zone are balanced (Fig. 4a). In this instance, the subduction rate is determined by the relative velocity of A with respect to B. Suppose now that a sudden change of the absolute velocity of A (the velocity of A with respect to the deep mantle) occurs that determines a decrease of its velocity relatively to B (Fig. 4b). An analogue situation could be represented by two persons that are pushing one another, and by a subsequent sudden release from one of the two opponents. The resulting predictable conclusion would be an unbalancing of the other one and thrusting toward the retreating person. In the case of two plates the unbalancing would generate a tensional stress field at the margin of B which may activate a subsequent process of back-arc deformation.

Figure 4. Sketch illustrating two different scenarios for the formation of slivers



A: Starting configuration. The trench system is in a state of equilibrium at time t. B: At time t' plate A accelerates away from the trench. The resulting tensional stress field at the trench causes back-arc spreading in marginal areas of B. C: In a different scenario an eastward acceleration of plate B determines rifting at the continental margin of the conjugate plate and the formation of the proxy sliver C.

A somewhat different situation would occur if the retreating block coincides with the upper-plate. In this instance, a tensional stress field would be applied to the subducting plate (Fig. 4c) and transferred to the deformable continental margin of A. Therefore, the oceanic part of this plate would be decoupled from the undeformed continent and a small fragment of continental lithosphere would travel toward the trench along with the surrounding oceanic crust. In both scenarios, a force $F = k\Omega$ is applied to the continental fragment with a magnitude proportional to the variation of relative velocity.

This model may explain several kinematic features of the subduction process in the Mediterranean region, in particular the rifting of continental terranes from the northern margin of Gondwana during the Mesozoic and the formation of extensional basins during the Cenozoic (Balearics, Tyrrhenian, Aegean, etc.). For instance, when Africa accelerated eastwards at chron 7 (25.2 Ma) about a pole located at (78.81N,301.25E), this retreat from the former Iberian trench may have resulted in the rifting of Sardinia, Corsica, Calabria and the Balearics from the Iberian margin between chrons 7 and 6. However, in this paper attention is focused on the kinematics of the Mediterranean region during the Mesozoic, which was dominated by the fragmentation of the northern Gondwana margin by successive retreat of the overriding plate (Fig. 4c) and by intermediate stages of trench compression.

A simple method for calculating finite rotations of continental fragments that traveled from a passive margin towards a trench is based upon the concept of decoupling of the oceanic part of a plate by fragmentation of the adjoining continental margin. As illustrated in Figure 4c, the new oceanic plate retains the subduction rate as an invariant of the process, and the whole mechanism allows the trench system to restore the state of equilibrium when the transition to the new stage is completed. Hence, the stage pole of the new plate with respect to the upper plate for the successive time period must coincide with the former stage pole of the subducting and still undeformed lower plate. Moreover, conservation of the subduction rate is required for any subsequent stage. This observation implies in general that successive small variations of the relative velocity between A and B will be compensated by different spreading rates at the ridge that separates the additional plate C from A. However, a strong acceleration of A in the direction of B could not be compensated by a lower spreading rate. In this instance, the ridge would collapse determining the formation of ophiolites [Robertson and Dixon, 1984].

In order to calculate finite rotations for the Mediterranean microplates, Equations (2) and (3) will be applied iteratively, starting from the initial stage associated to the early formation of oceanic crust in the Central Atlantic and from the assumed initial assemblage at the Northern Gondwana margin. It will be shown that the stage pole associated to the subduction of oceanic lithosphere beneath the Eurasian margin between 175.0 Ma and 170.0 Ma was conserved until chron M0 (120.4 Ma), when an abrupt change in the relative motion of Africa with respect to Eurasia determined the onset of the Alpine collision. During this long time period a unique stable system comprising a stationary trench in the Northern Tethys and a buffering spreading center to the South existed. According to the theoretical model discussed above, this process can be viewed as the driving mechanism for the drifting of Northern Turkey towards the Eurasian margin.

In the following sections the fourteen stages that characterized the tectonic history of the Mediterranean region during the Jurassic and the Cretaceous will be grouped in nine major phases. For each phase boundary, reconstruction maps will be discussed that show the velocity and acceleration fields, the modeled pattern of seafloor spreading [Schettino, 1999] and the distribution of the continental lithosphere. The geomagnetic time scales of Cande and Kent [1995] and Gradstein et al. [1994] are used respectively for anomalies younger than chron 34 (83.5 Ma) and for older times. A full animation of the paleogeographic evolution of the Tethyan realm is shown in the Appendix and is also accessible through the Internet at: http://www.serg.unicam.it/Geo.html [Schettino and Scotese, 2001].

Phase 1: The onset of seafloor spreading in the Central Atlantic

This phase comprises a single stage that falls in the Bajocian, between 175.0 Ma and 170.0 Ma. The younger limit is associated with the fit of the Blake Spur Magnetic Anomaly of Eastern North America [Klitgord and Schouten, 1986] with a corresponding anomaly in Northwest Africa, whereas the older limit of this stage coincides with the onset of an accelerated phase of rifting which culminated in the formation of the first oceanic crust in the Central Atlantic. As discussed in a previous section, during the first few million years the left-lateral motion of Gondwana with respect to Laurasia continued to be transferred from the Atlantic region to the Western Tethys through the Atlas rift system. This model requires that Morocco and Adria remained attached to Laurasia during this initial period (Fig. 3). Similarly, spreading in the Pindos Basin persisted until 172.0 Ma, when the ridge jumped northward into the Liguride Basin initiating sea floor spreading in the Alpine Tethys, and transtensional motion between Iberia and Morocco.

Table 2. Rotation model for the major Mediterranean microplates. Only Jurassic and Cretaceous Euler poles are

shown. The fit of Arabia to Africa was proposed by LePichon and Gaullier [1988]

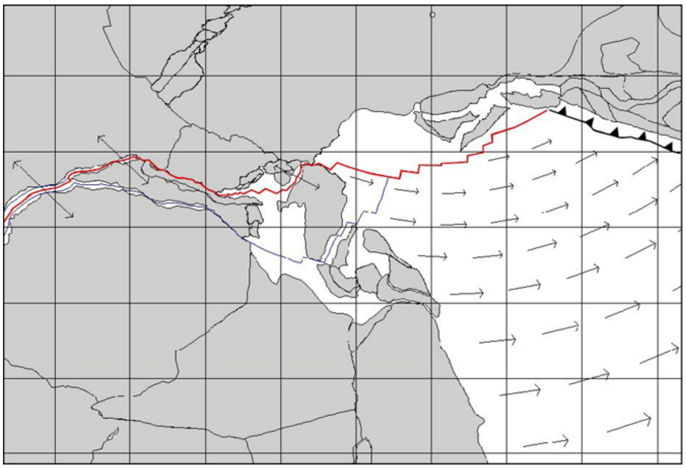
Table 2. Rotation model for the major Mediterranean microplates

Age	Lat	Lon	Angle	Reference Plate
Adria				
67.7	-35.54	194.61	18.33	Northeast Africa
72.5	-35.54	194.61	18.33	Northeast Africa
72.5	-33.2	185.02	24.74	Iberia
80.2	-33.2	185.02	24.74	Iberia
80.2	-35.3	194.5	18.3	Northeast Africa
172	-35.3	194.5	18.3	Northeast Africa
175	-31.29	195.84	16.62	Northeast Africa
200	-31.29	195.84	16.62	Northeast Africa
Pelagonian				
67.7	68.34	331.04	2.37	Adria
172	68.34	331.04	2.37	Adria
175	62.07	225.26	0.36	Adria
200	62.07	225.26	0.36	Adria
Southern Turkey				
67.7	-33.71	220.83	13.86	Arabia
72.5	-33.71	220.83	13.86	Arabia
72.5	-32.02	201.13	25.44	Iberia
74.3	-32.02	201.13	25.44	Iberia
74.3	-33.22	220.6	13.83	Arabia
200	-33.22	220.6	13.83	Arabia
Vardar Ocean				
67.7	67.26	1.69	17.46	Northeast Africa
120.4	67.26	1.69	17.46	Northeast Africa
126.7	65.61	5.16	14.71	Northeast Africa
130	64.3	7.93	12.81	Northeast Africa
131.9	63.63	8.42	11.84	Northeast Africa
139.6	62.44	4.46	8.89	Northeast Africa
147.7	61.69	359.48	5.4	Northeast Africa
154.3	69.34	3.49	3.51	Northeast Africa
170	0	0	0	Northeast Africa

Age	Lat	Lon	Angle	Reference Plate
200	0	0	0	Northeast Africa
Arabia				
67.7	32.13	22.58	-6.36	Northeast Africa
200	32.13	22.58	-6.36	Northeast Africa
Northern Turkey				
67.7	-26.58	205.76	35.47	Eurasia
72.5	-25.16	205.2	35.37	Eurasia
74.3	-25.16	205.2	35.37	Eurasia
80.2	-22.69	202.44	30.78	Eurasia
83.5	-19.07	197.98	25.62	Eurasia
87	-18.29	197.47	26.65	Eurasia
87	8.94	239.36	14.36	Northeast Africa
120.4	8.94	239.36	14.36	Northeast Africa
120.4	-34.32	215.58	23.24	Vardar
170	-34.32	215.58	23.24	Vardar
170	-34.32	215.58	23.24	Northeast Africa

A rotation model for the Mediterranean microplates during the Jurassic and the Cretaceous is shown in Table 2. Figure 5 illustrates the paleogeographic configuration at 170.0 Ma and the corresponding velocity field. Sea floor spreading rates during this stage range from the 65 mm/yr of the southern Central Atlantic to the 33 mm/yr of the Liguride Ocean and the 23 mm/yr of the Northeastern Tethys. A subduction zone at the southern margin of Eurasia was fully established from Sanandaj to Armenia and the Eastern Pontides [Robertson and Dixon, 1984]. A description of the main geologic and structural features of the assemblage of slivers that developed above the subduction zone by oblique convergence can be found in McCall [1996].

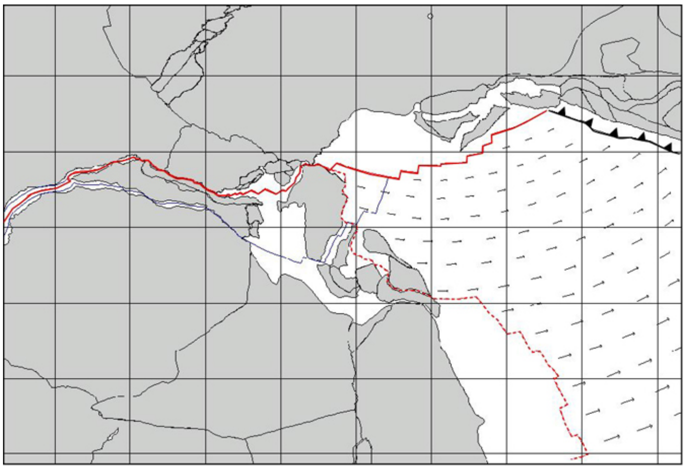
Figure 5. Reconstruction at 170.0 Ma (Upper Bajocian)



Vectors represent direction and magnitude of the relative velocity field between Northern Gondwana and Laurasia. Blue line represents the modeled 172.0 Ma isochron. The instantaneous Euler pole location is marked by a small empty circle.

An abrupt change of the absolute velocity of Laurasia at 170.0 Ma occurred at the end of this initial stage. The acceleration field shown in Figure 6 indicates that the Central Atlantic and Liguride ridges were only affected in the spreading rate, whereas transform azimuths remained unchanged. However, a tensional stress field in the Tethyan realm was generated that caused decoupling of the oceanic lithosphere from the Northern Gondwana margin and rifting of microcontinents, the largest being represented by the Northern Turkey. Evidence of this event can be found in the Sakarya zone [Okay, 1984 ; Yilmaz et al., 1995], where basal conglomerates of the Lower Lias are overlain by sandstones, then by neritic carbonates (until the Late Jurassic), finally by hemipelagic cherty limestones during the Early Cretaceous. The oceanic plate that formed at 170.0 Ma, corresponding to the Vardar Ocean of Şengör et al. [1984], existed as an independent plate until chron M0 (120.4 Ma). Following Şengör and Yilmaz [1981], we will use the term “Inner Tauride Ocean” to indicate the oceanic crust that formed to the South of Sakarya and Kirsehir and that progressively separated these blocks from the Menderes-Tauride platform.

Figure 6. Acceleration field at 170.0 Ma (Upper Bajocian)

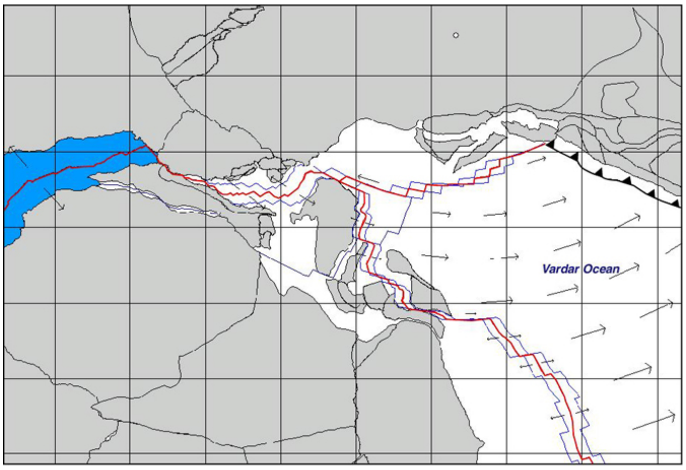


Vectors represent direction and magnitude of the relative acceleration field Laurasia with respect to Northern Gondwana. Dashed line represents a modelled trend of the incipient spreading center associated to the birth of the Vardar Ocean. The instantaneous Euler pole of acceleration is marked by a small empty circle.

Phase 2: Seafloor spreading in the Southern Tethys

This episode comprises, as the previous one, a single stage. The time interval is located between 170.0 Ma (Uppermost Bajocian) and chron M25 (154.3 Ma, Oxfordian-Kimmeridgian boundary) and is characterized by slower spreading rates, ranging from the 45 mm/yr of the southern Central Atlantic to the 23 mm/yr of the Liguride Ocean, where 380 Km of new oceanic crust were formed between the passive margins of Eurasia and Adria. The modeled isochrons and the velocity field at the upper limit of this stage are illustrated in Figure 7. This map also shows the amount of new Tethyan crust that was created. During this time interval Iberia remained attached to Laurasia, while a wide shear zone accommodated left-lateral motion with respect to Morocco.

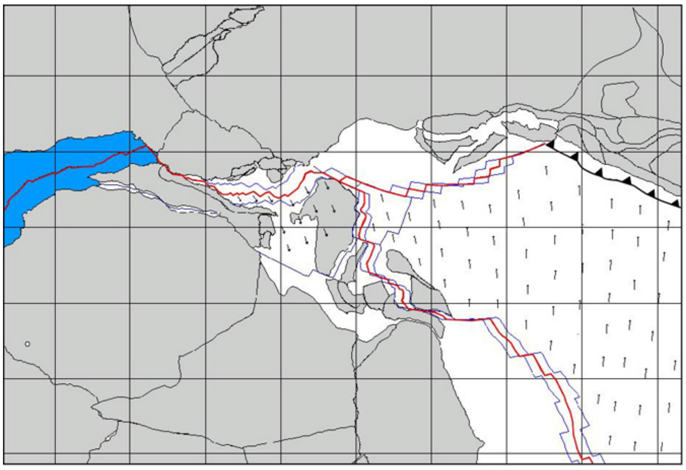
Figure 7. Reconstruction at 154.3 Ma (Upper Oxfordian)



Vectors represent direction and magnitude of the relative velocity field between conjugate pairs of plates. Blue line represents the modeled 172.0 Ma and 170.0 Ma isochrons. The location of the instantaneous Euler pole of relative convergence between Vardar and Eurasia is marked by a small empty circle.

The acceleration field that determined a stage transition at chron M25 is illustrated in Figure 8. It can be described as a northward acceleration of Eurasia that was balanced by a reorientation of the transform faults in the Inner Tauride Ocean and in the Liguride Basin.

Figure 8. Acceleration field at 154.3 Ma (Upper Oxfordian)



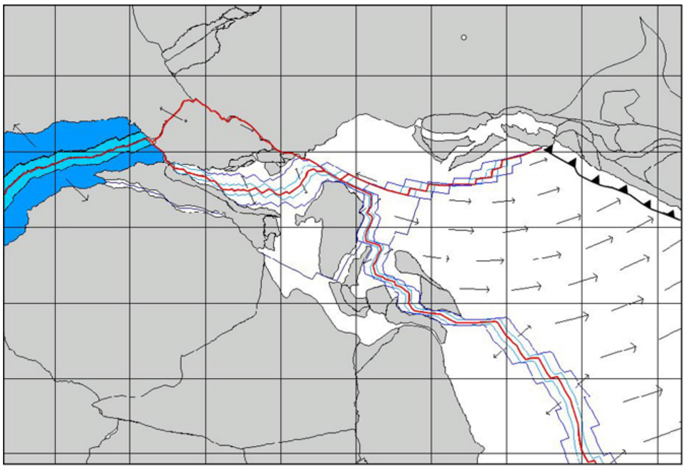
Vectors represent direction and magnitude of the relative acceleration field of Africawith respect to Iberia and Eurasia with respect to Africa. The instantaneous Euler pole of acceleration of Eurasia with respect to Africa is marked by an empty circle.

Phase 3: Iberia separates from North America

The third phase comprises a single short stage between chrons M25 (154.3 Ma, Uppermost Oxfordian) and M21 (147.7 Ma, Middle Tithonian), during which Iberia began to separate from North America (150.0 Ma, Lower Tithonian). This event caused the formation of a new transform plate boundary to the North, between Iberia and Eurasia, and the progressive transfer of left-lateral motion from the southern boundary (the former shear zone between Iberia and Morocco) to this newly constituted fault system (Fig. 9). By the end of the stage, any motion between Iberia and Africa ceased, and the existing spreading center of the Liguride Ocean became extinct.

The northern boundary of Iberia during the Jurassic could be represented by the North Pyrenean Fault (NPF), a major vertical structure that several authors assume to be associated with strike-slip motions during the Middle Cretaceous [Choukroune, 1976 ; Teixell, 1998 ; Beaumont et al., 2000]. However, the global kinematic pattern requires that this fault system was active well before the Albian-Cenomanian, because no other major structure has been recognized that could account for the Late Jurassic-Early Cretaceous motion of Iberia with respect to Eurasia. An indirect confirmation of this hypothesis comes from a recent study of calcareous nannofossil assemblages from the Iberia Abyssal Plain [Concheryo and Wise, 2001]. This study shows that the oldest assemblages of calcareous nannofossils in this area are early to mid-Tithonian in age and formed in a restricted interior basin, though open marine conditions were not established until the major Berriasian rifting episode.

Figure 9. Reconstruction at 147.7 Ma (Middle Tithonian)



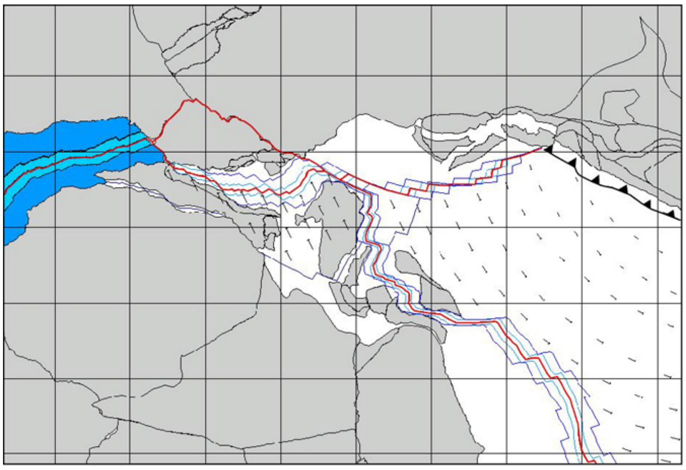
Vectors represent direction and magnitude of the relative velocity field between conjugate pairs of plates. Blue line represents the modeled 172.0 Ma, 170.0 Ma and M25 isochrons. The location of the instantaneous Euler pole of relative convergence between Vardar and Eurasia is marked by a small empty circle.

The plate velocity field at anomaly M21 is shown in Figure 9. At this time the Liguride Ocean reached its maximum extension, about 490 Km in the northern part and 640 Km in the southern zones. The spreading rates did not change significantly with respect to the previous stage, ranging from 46 mm/yr in the southern Central Atlantic to 16 mm/yr in the Liguride Ocean.

The acceleration field illustrated in Figure 10 shows the pulse that extinguished the Liguride spreading center in the Western Tethys and that forced the Vardar ridge to re-orient the system of transform faults that connected spreading segments. It is important to note that the model proposed in the previous sections requires that the ridge that existed in the Northern Tethys, between the area to the North of Adria (Alpine Tethys) and the Eastern Pontides, remained stable during the whole sequence of phases that were characterized by a constant subduction rate of the Vardar Ocean. In fact, this complex system of small spreading segments and large transform faults separated Eurasia from

Vardar in a way that was kinematically consistent with the invariant stage pole of subduction of the decoupled oceanic plate. Hence, new oceanic crust in the Alpine Tethys continued to be formed until the Lower Aptian, well beyond the time of extinction of the spreading centers that existed in the Western Mediterranean. This persistence of tectonic activity is confirmed by the presence of resedimented materials in the pelagic facies of the South Alpine domain [Winterer and Bosellini, 1981; Santantonio, 1994].

Figure 10. Acceleration field at 147.7 Ma (Middle Tithonian)

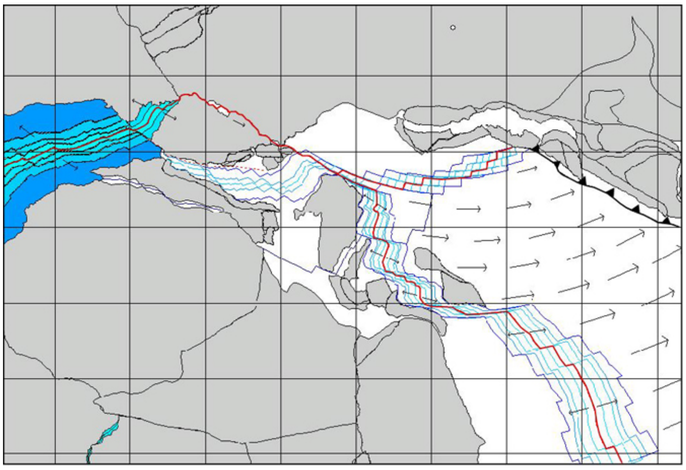


Vectors represent direction and magnitude of the relative acceleration field of Africa with respect to Iberia and Eurasia with respect to Africa.

Phase 4: Transcurrent motion along the Northern Pyrenean Fault

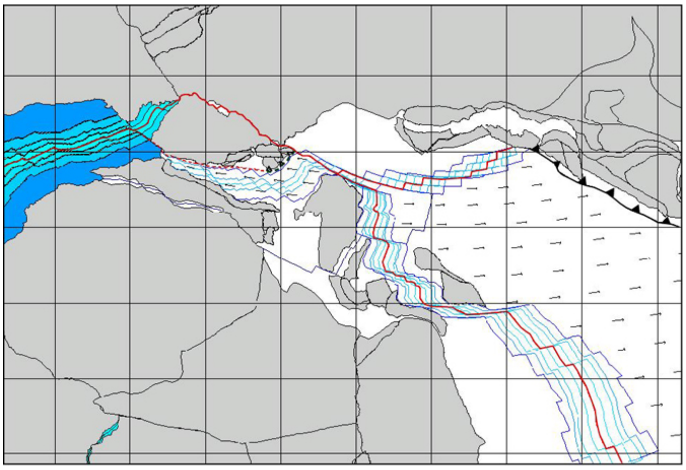
This phase comprises two stages, the first being associated to the time interval from chron M21 (147.7 Ma, Middle Tithonian) to chron M16 (139.6 Ma, Upper Berriasian), while the second one ranges between 139.6 Ma and anomaly M10 (131.9 Ma, Valanginian-Hauterivian boundary). The regional velocity field is illustrated in Figure 11. As anticipated in the previous section, seafloor spreading in the Central Atlantic is now linked to the kinematic pattern of the Tethyan realm via the NPF. No other major tectonic events occurred in the Tethyan realm during these two stages. However, the spreading rate in the southern Central Atlantic diminished significantly with respect to the previous phase, from about 46 mm/yr to 30 mm/yr. The onset of an early compressional setting in the Liguride Ocean and the renewal of tectonic activity at the southern boundary of Iberia marked the end of this phase at anomaly M10. The acceleration field that was responsible of this event is shown in Figure 12. In the Vardar Ocean, an eastward directed variation of velocity of Eurasia determined a faster spreading rate at the beginning of the successive phase.

Figure 11. Reconstruction at 131.9 Ma (Lower Hauterivian)



Vectors represent direction and magnitude of the relative velocity field between conjugate pairs of plates. Blue lines represents the modeled 172.0 Ma, 170.0 Ma, M25, M21 and M16 isochrons. The location of the Euler pole of relative convergence between Vardar and Eurasia is marked by a small empty circle.

Figure 12. Acceleration field at 131.9 Ma (Lower Hauterivian)



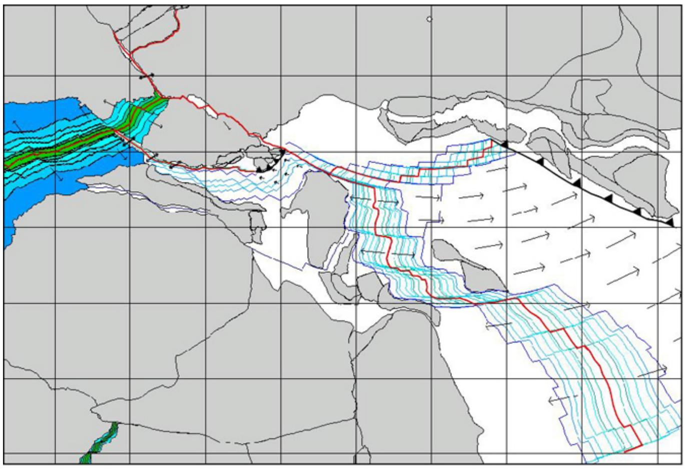
Vectors represent direction and magnitude of the relative acceleration field of Africa with respect to Iberia and Eurasia with respect to Africa.

Phase 5: Last spreading stages in the Vardar Ocean and Alpine Tethys

The general tectonic setting around the Tethys during this important time interval appears as an Early Cretaceous prolongation of the kinematic pattern that characterized the Jurassic. However, its upper limit marks the epilogue of the tectonic history that started in the Middle Jurassic and lasted about 55 Myrs. Three stages are comprised in this phase: from chron M10 (131.9 Ma, Valanginian-Hauterivian boundary) to chron M9 (130.0 Ma, Lower Hauterivian-Upper Hauterivian boundary), from 130.0 Ma to chron M4 (126.7 Ma, Hauterivian-Barremian boundary), and from 126.7 Ma to chron M0 (120.4 Ma, Barremian-Aptian boundary).

A first important event that occurred during this phase is represented by the onset of rifting and seafloor spreading in the North Atlantic since 130.0 Ma [Rowley and Lottes, 1988]. However, in spite of the relevance of this event at global scale and the apparent synchronism, it can be shown that it did not trigger the mechanism of Alpine subduction during the Lower Aptian. This aspect will be discussed further in this section. Another feature of this time period is associated with the weak compressional setting that established in the Liguride Basin (Fig. 13), though the predicted total amount of shortening (about 67 Km) could be too small for a field evidence of this event. Similarly, the very small amount of extension predicted by the model for the southern margin of Iberia can be hardly observed through geological studies. In general, most of the tectonic activity around Iberia continued to occur at the northern transtensional boundary with the Eurasian plate. In this region two important basins formed that are well documented in the geologic literature, the Valais trough and the Asturian basin. To the East, the Valais oceanic trough opened by separation of the Sardinia-Corsica-Briançonnais domains from the southern French margin [Stampfli et al., 1998]. The model proposed in this paper predicts that about 200 Km of new oceanic crust formed between the northeastern margin of Iberia and Eurasia during this phase. At the western side, geologic and structural data from the Asturian region indicate that small rift basins formed at the northern margin of Iberia after the Late Jurassic [Lepvrier and Martínez-García, 1990], several million years before the onset of seafloor spreading in the Biscay Bay [Srivastava et al., 1990].

Figure 13. Reconstruction at 120.4 Ma (Lower Aptian)

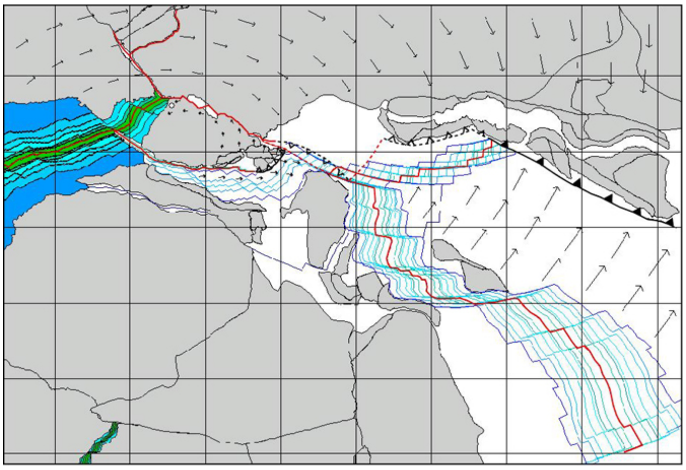


Vectors represent direction and magnitude of the relative velocity field between conjugate pairs of plates. Blue lines represent the modeled 172.0 Ma, 170.0 Ma, M25, M21, M16 and M10 isochrons. The 130.0 Ma and M4 isochrons are indicated in green. The location of the Euler pole of relative convergence between Vardar and Eurasia is marked by a small empty circle.

During the first two stages, between anomalies M10 (131.9 Ma) and M4 (126.7 Ma), the spreading rate in the southern Central Atlantic slowed again, only 16.80 mm/yr with respect to the 30 mm/yr of the previous phase. However, it returned to be 30 mm/yr during the third stage. At the end of the phase (120.4 Ma, Barremian-Aptian boundary) the Alpine Tethys and the Inner Tauride Ocean reached their maximum extensions, respectively 1400 and 1100 Km.

The acceleration field at 120.4 Ma (Barremian-Aptian boundary) is illustrated in Figure 14. The absolute variations of velocity (that is, relative to the spin axis) of Eurasia and North America are shown in the upper part of this map. They are almost identical, though Eurasia was moving with respect to N. America at that time. The anti-pole of this absolute acceleration field was located in southern Morocco (24.03N,13.48E), with a magnitude of clockwise rotation equal to 0.36°/Myr. Therefore, the ongoing separation of Eurasia from the North American craton did not affect the establishment of a strong compressional field in the Tethyan realm, as the two plates accelerated with the same pole and magnitude. This event generated at the same time a strong pulse in the spreading rates of the Central Atlantic and the onset of the Alpine collision. The acceleration of Africa with respect to Eurasia is shown in Figure 14 it is apparent that such pulse of convergence could not be compensated by a slower spreading rate of the Vardar ridge. Hence, the entire system of spreading segments collapsed. According to the model of Robertson and Dixon [1984], a long slice of ophiolites should have formed as a consequence of this Neothetyan ridge collapse, consistently with models that predict at least two distinct root-zones for the Turkish ophiolites [e.g., Çengör and Yilmaz, 1981 ; Robertson and Dixon, 1984]. Therefore, our model confirms that the Neotethyan oceanic seaway that formed between Sakarya-Kirsehir and the Menderes-Tauride platform during the Jurassic and the Early Cretaceous, herein referred to as the "Inner Tauride Ocean", was the source zone for the ophiolites of the Central Anatolia, which were emplaced onto the suture between these blocks during the Late-Cretaceous-Eocene collision.

Figure 14. Acceleration field at 120.4 Ma (Lower Aptian)

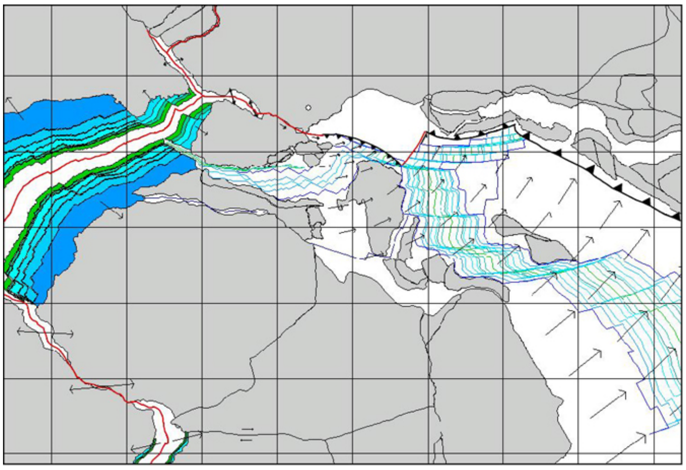


Vectors on N. America and Eurasia represent direction and magnitude of the absolute acceleration field of these plates. Vectors in the Eastern Tethys represent the acceleration field of Northeast Africa with respect to Eurasia. Vectors on Iberia are relative to Eurasia. Finally, the acceleration field in Western Mediterranean is relative to Iberia. A small circle represents the Euler pole of acceleration of N. Africa woth respect to Eurasia.

Phase 6: The early Alpine subduction

This time interval comprises a single stage between anomaly M0 (120.4 Ma, Barremian-Aptian boundary) and 110.0 Ma (Lower Albian). The reconstruction of the velocity and acceleration fields at 110.0 Ma is illustrated in Figures 15-16. As shown in Figure 16, the younger boundary of this stage is not associated to important variations of the relative velocity of Africa with respect to Eurasia. However, the reconstruction shows some important features of the Alpine tectonics during the Cretaceous Quiet Zone.

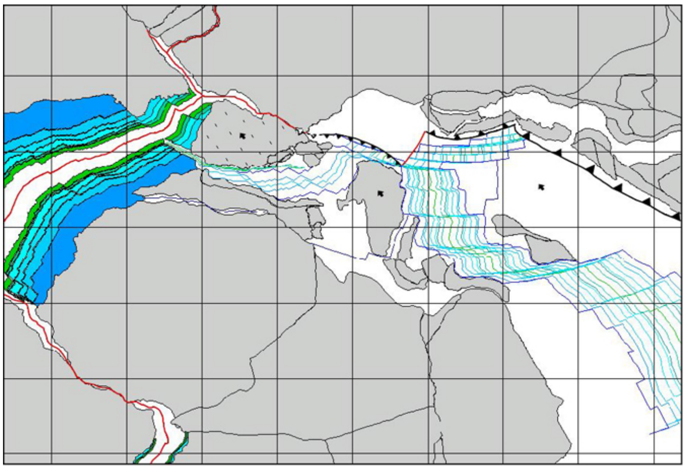
Figure 15. Reconstruction at 110.0 Ma (Lower Albian)



Vectors represent direction and magnitude of the relative velocity field between conjugate pairs of plates. Blue lines represent the modeled 172.0 Ma, 170.0 Ma, M25, M21, M16 and M10 isochrons. The 130.0 Ma, M4 and M0 isochrons are indicated in green. The location of the Euler pole of relative convergence between Northeast Africa and Eurasia is marked by a small empty circle.

A first important feature of this time period is represented by the onset of seafloor spreading in the North and South Atlantic, and by the opening of the Biscay Bay [Srivastava et al., 1990]. In particular, spreading rates were high in the South Atlantic (~43 mm/yr at the Romanche Fracture Zone and ~63 mm/yr at the Agulhas-Falkland Shear Zone) and the Central Atlantic (55 mm/yr in the southernmost region). Simultaneously, the African plate experienced internal deformation with the onset of right-lateral motion along the CASZ and rifting episodes in the Benue Trough and Termit Basins [Wilson and Guiraud, 1992 ; Genik, 1992 ; Guiraud and Maurin, 1992].

Figure 16. Acceleration field of Africa withrespect to Eurasia at 110.0 Ma (Lower Albian).



Acceleration field of Africa withrespect to Eurasia at 110.0 Ma (Lower Albian).

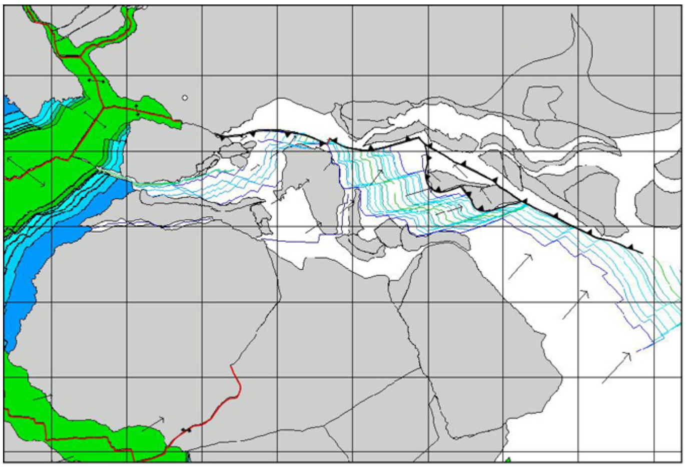
Regarding the Tethyan realm, we note the establishment of two different trench systems, which were presumably separated by a transform fault (Fig. 15). In the Western Mediterranean, old oceanic crust of the Pennine Ocean (Eurasian plate) was subducting southwards under the active margin of Adria. In the Eastern regions the old Jurassic trench of Sanandaj-Armenia-East Pontides extended further west to reach the West Pontides and Rodophe, where Jurassic oceanic crust was subducting northwards. Convergence rates and styles of subduction indicate a strong differentiation between these two trench systems. In general, we observe an eastward increase of the subduction rate, starting from the ~5.2 mm/yr at the highly oblique trench of the Valais Trough through the ~18.0 mm/yr of the Eastern Alpine trench. To the East, North-dipping oblique subduction under the Rodophe-West Pontides zone occurred at rates ranging between ~18.3 mm/yr and ~31 mm/yr. Finally, the highest magnitudes were reached in the easternmost regions, where pre-Jurassic oceanic crust was subducted at a rate of ~60 mm/yr.

From the point of view of the kinematic model presented in the previous sections, the overall result of the plate boundary reorganization at chron M0 (120.4 Ma) was the formation of a system of converging plates characterized with a new stage pole. From the Middle Jurassic through the Early Cretaceous, a large oceanic plate subducted under Eurasia at a constant rate of 0.698°/Myr about a stage pole located at (55.21N,31.76E). This stage pole was calculated assuming that the kinematic pattern established during the initial stage of seafloor spreading in the Central Atlantic, between 175.0 Ma and 170.0 Ma, constrained the rate of convergence in the Eastern Tethys during the subsequent stages. Then, finite rotations of the Vardar plate with respect to Northeast Africa (Table 2) were calculated using Equation 3 and the assumption that small variations of the relative velocity between Africa and Eurasia were compensated by variations of the spreading rate at the ridge that separated Vardar from Northern Gondwana. As discussed in the previous section, the events that occurred at anomaly M0 caused the end of this mechanism of compensation, and a new system of subduction zones established that were characterized by a different rate and pole of convergence. As for the previous phases, we could assume that the stage pole for the motion of Africa with respect to Eurasia during this initial stage (between 120.4 Ma and 110.0 Ma) constrained the rate of convergence of all the subsequent stages. However, the field associated to variations of the relative velocity between Africa and Eurasia illustrated in Figure 16 shows a weak component of acceleration towards the trench. Moreover, we will show that a state of trench compression characterized the Alpine and Tethyan subduction systems until the Lower Campanian. Hence, a state of equilibrium in the mechanism of convergence between the major plates was not established until the Late Cretaceous. This means that mechanisms for the conservation of the equilibrium were not activated until that time.

Phase 7: First collisions in the Eastern Tethys

This phase comprises a single stage between 110.0 Ma (Lower Albian) and chron 34 (83.5 Ma, Santonian-Campanian boundary). A reconstruction of the velocity field at chron 34 is illustrated in Figure 17. Perhaps the most important event of this time period is represented by the establishment of a subduction zone south of Sakarya-Kirsehir, once the leading tip of this composite microplate reached the active margin of Eurasia between Armenia and Eastern Pontides. The model proposed in this paper predicts that Northern Turkey started to collide at about 87 Ma (Upper Coniacian). In order to calculate the motion of this block for subsequent times, we combined the stage pole of the surrounding oceanic plate (which was part of Africa) with an additional pole that removed the trench-normal component of velocity at the contact zone. In this way, no overlap between Northern Turkey and the Eurasian margin was allowed, and the resulting motion was simply a right-lateral translation along the trench of the leading edge and an additional clockwise rotation of the other points. Of course this mechanism required the establishment of a new zone of convergence at the southern margin of Sakarya-Kirsehir, where the Inner Tauride Ocean initiated to subduct. Geologic data from the Tuzgölü Basin of Central Turkey, Sakarya and the Pontides substantially confirm the model proposed here [Görür et al., 1984, and references therein ; Yilmaz et al., 1995]. In fact, Görür et al. [1984] describe a scenario in which north-dipping subduction of Tethyan oceanic crust under Rodophe-Pontide may have initiated during the Aptian-Albian, whereas subduction under Sakarya and Kirsehir occurred since the Cenomanian-Turonian, which is in agreement with the model proposed here.

Figure 17. Reconstruction at 83.5 Ma (Santonian-Campanian boundary)



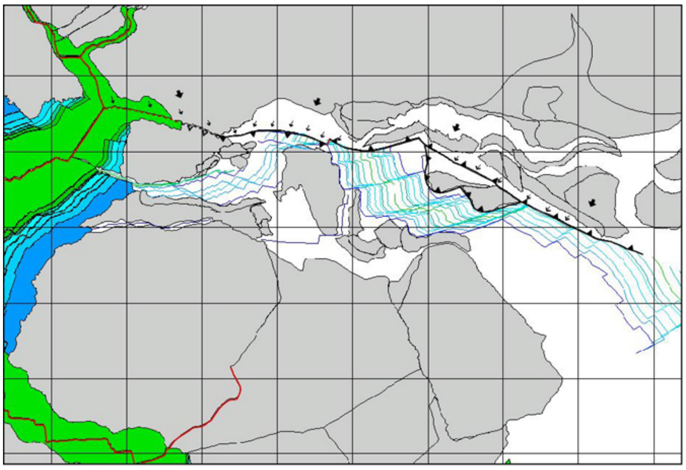
Vectors represent direction and magnitude of the relative velocity field between conjugate pairs of plates. Blue lines represent the modeled 172.0 Ma, 170.0 Ma, M25, M21, M16 and M10 isochrons. The 130.0 Ma, M4 and M0 isochrons are indicated in green. The location of the Euler pole of relative convergence between Africa and Eurasia is marked by a small empty circle.

An interesting feature of the reconstruction at chron 34 is associated with the progressive approach of the south-dipping Alpine subduction zone to the Rodophe-Pontide trench (Fig. 17). At the Santonian-Campanian boundary the length of the transform fault that connected these two systems was strongly reduced, giving rise to a complex subduction zone characterized by reversal of subduction polarity, similar to some present day zones of subduction of Southeast Asia and Western Pacific (e.g., the Solomon-San Cristobal trench system).

Flysch sedimentation became widespread in the South Alpine region from the Turonian [Castellarin, 1976 ; Bichsel and Häring, 1981 ; Bernoulli et al., 1981 ; Bernoulli and Winkler, 1990]. The direction of transport of these terrigenous turbidites suggests a northern provenance, possibly from a trench-forearc system comprising Austroalpine units. At the western end of the Alpine trench, oblique subduction of the Valais Ocean under Corsica continued at a rate of ~8 mm/yr for the whole stage. This process, as well as further spreading in the Biscay Bay, was clearly related to the continuation of counterclockwise rotation of Iberia with respect to Eurasia.

The pattern of the acceleration fields at 83.5 Ma is illustrated in Figure 18. It indicates ridge compression in the Biscay Bay, hence the extinction of the corresponding spreading center. In the Mediterranean region the whole system of subduction zones was subject to compression. Therefore, the next stage began with a plate tectonic configuration of the Tethyan realm characterized by instability of the trench systems.

Figure 18. Acceleration field Eurasia with respect to Iberia and Africa at 83.5 Ma (Santonian-Campanian boundary)



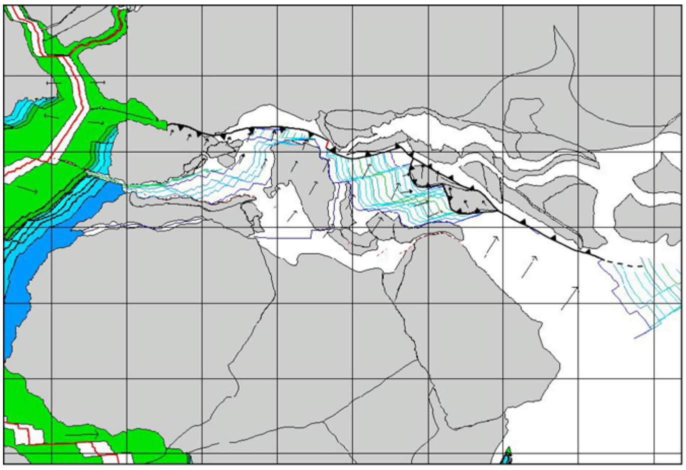
Empty barbs represent incipient subduction zone. Dotted line indicates ridge extinction.

Phase 8: The Early Pyrenean Orogeny

This phase encompasses two stages, the first of which being represented by a short time interval between anomaly 34 (83.5 Ma, Santonian-Campanian boundary) and 80.2 Ma (Middle Campanian), whereas the second one lasted at

74.3 Ma (C33n, Upper Campanian). A uniform velocity field (Fig. 19) characterized the western regions, from the northern margin of Iberia to Western Pontides, because no relative motion occurred between the Mediterranean microplates (including Iberia) and Africa during this phase. To the East, subduction of the Inner Tauride Ocean beneath the Sakarya-Kirsehir block continued to be constrained by the ENE-directed field of velocity of Africa relative to this block. An interesting feature of the reconstruction of Figure 19 is represented by the westward prolongation of the Alpine trench, in the region presently occupied by the Pyrenean orogen. The observation of recent seismic reflection profiles interpretations [Teixell, 1998 ; Beaumont et al., 2000] suggests that this early stage of compression was accommodated by thrusting of continental crust of Iberia onto Eurasia along the North Pyrenean Frontal Thrust (NPFT). If this scenario is correct, the south-dipping Alpine subduction zone can be traced further West without inversion of polarity, as indicated in Figure 19. However, independently from these structural details, we observe that the onset of the Pyrenean orogeny was not related to the opening of the Biscay Bay and the rotation of Iberia, but occurred after the completion of this process.

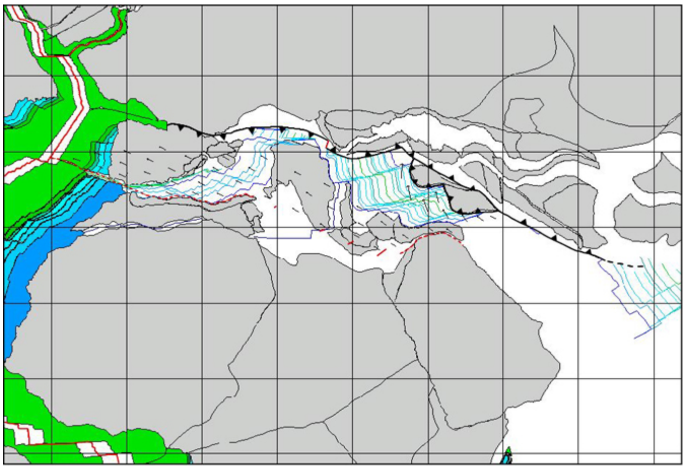
Figure 19. Reconstruction at 74.3 Ma (Upper Campanian)



Vectors represent direction and magnitude of the relative velocity fieldof Africa relative to Eurasia. Blue lines represent the modeled 172.0 Ma, 170.0 Ma, M25, M21, M16 and M10 isochrons. The 130.0 Ma, M4 and M0 isochrons are indicated in green. Dashed red lines represent areas of incipient extension.

The acceleration field at the Upper Campanian boundary is shown in Figure 20. We note the absence of variations of relative velocity at the convergent boundaries of the Northern Mediterranean region. This indicates that the main trench systems reached a state of equilibrium at the end of the phase. Conversely, the Inner Tauride Trench was still in a highly rotational compression field, as illustrated in Figure 20. The principal feature of this map is the weak northwest-directed variation of velocity of Eurasia and Iberia with respect to Africa, which implies either the onset of an extensional stress field in the Southern Mediterranean region (if Adria and Southern Turkey are kept fixed to Iberia) or extension at the western margin of Adria. Geologic data from the Eastern Mediterranean (Antalya Complex and Kyrenia Range) seems to indicate spreading episodes during the Late Cretaceous [Robertson, 1998], whereas no evidence of extensional phenomena comes from the Apenninic domain. Hence, we will assume that Upper Campanian rifting and possibly spreading affected the Southern Mediterranean region, immediately prior of the onset of subduction in the easternmost areas.

Figure 20. Acceleration field of Iberia, Adria and Southern Turkey with respect to Africa at 74.3 Ma (Upper Campanian)



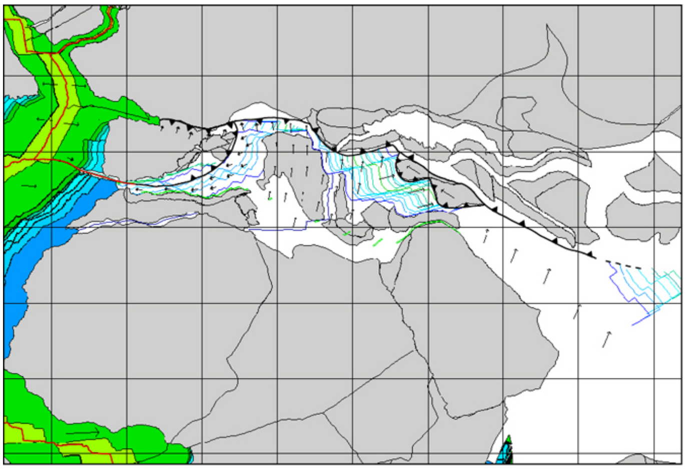
Arrows South of Sakatya and Kirsehir indicate acceleration of subduction rate of the Inner Tauride Ocean. Dashed lines indicate zones of extension.

Phase 9: Closure of the Eastern Oceanic Connection

The last phase that characterized the tectonic history of the Mediterranean region during the Mesozoic comprises, as the previous one, two stages: from 74.3 Ma (C33n, Upper Campanian) to 72.5 Ma (C32n.2n, Upper Campanian), and from 72.5 Ma to 67.7 Ma (C31n, Upper Maastrichtian).

In the eastern regions, this time period marks the completion of the process of subduction of the Vardar Ocean (Northeast Neo-Tethys) under Eurasia and the first direct contact between Kirsehir and the Tauride Platform, with the subsequent closure of the former oceanic connection with the Indian Ocean (Fig. 21). This event also triggered compression in the southern branch of Neotethys, between the Tauride Platform and Arabia [e.g., Robertson, 1998]. At the end of the second stage, in the Upper Maastrichtian, only a remnant of the oceanic crust that formed during the Jurassic and Early Cretaceous survived in the eastern region: the Inner Tauride Ocean, which was still subducting under the Western Pontides, Sakarya and Kirsehir.

Figure 21. Reconstruction at 67.7 Ma (Upper Maastrichtian)



Vectors represent direction and magnitude of the relative velocity fields in the Mediterranean region. Blue lines represent the modeled 172.0 Ma, 170.0 Ma, M25, M21, M16 and M10 isochrons. The 130.0 Ma, M4, M0 and 74.3 Ma isochrons are indicated in green.

Regarding the Western region, the velocity field shown in Figure 21 indicates the onset of weak compressive tectonics in the Liguride Ocean, associated to the eastward motion of Iberia with respect to Africa since 72.5 Ma (C32n.2n, Upper Campanian), tough an earlier phase of compression is likely to have occurred in conjunction with relative motion between Northwest Africa and Northeast Africa. Evidence of compression in this region during the Late Cretaceous is well documented by both geological and geophysical data. A recent paper of Faccenna et al. [2001] shows that initiation of slow subduction under the eastern margin of Iberia (Corsica-Sardinia-Balearics) occurred between 80 Ma and 70 Ma. These authors estimate an average velocity of convergence of 8 mm/yr from the Late Cretaceous to the Oligocene, which agrees well with the value predicted by our model for this initial stage: 6 mm/yr offshore Corsica. However, radiometric ages of the blueschist metamorphic facies of Alpine Corsica indicate an earlier initiation of compressive motions in the Liguride Ocean, about 105 Ma for the emplacement of the Schistes Lustrés nappes [Cohen et al., 1981]. We assume that this event, which predates the onset of subduction of oceanic crust in this region, is associated to large-scale tectonic motions that deformed the African continent between anomalies M0 and C34, when Northwest Africa and Iberia were subject to a small clockwise rotation with respect to Northeast Africa.

Conclusion

A new model for the formation of microplates at convergent boundaries has been discussed and applied to the tectonic history of the Mediterranean region for Jurassic and Cretaceous times. Dewey et al. [1989] already recognized that the motion of large surrounding plates is not always directly related to the regional tectonics. In this paper we show that the kinematics of small intervening terranes is constrained by variations of relative velocity between the major plates. In particular, we have shown that deformation of passive margins and rifting of continental slivers is directly related to the conservation of equilibrium at trench systems. A discussion about the major tectonic events that characterized the Mediterranean region demonstrates that the quantitative approach proposed here accounts for timing and style of processes that are documented in the geologic literature. Hence, an application of this method to other complex tectonic histories (e.g., Southeast Asia, West Pacific, Cenozoic Mediterranean) could give a better understanding of the geologic processes in these areas.

References

Alvarez, W., Tectonic evolution of the Corsica-Apennines-Alps region studied by the method of successive approximations, Tectonics, 10(5), 936-947, 1991.

Beauchamp, W., M. Barazangi, A. Demnati, and M. El Alji, Intracontinental rifting and inversion: Missour Basin and Atlas Mountains, Morocco, AAPG Bull., 80(9), 1459-1482, 1996.

Beaumont, C., J. A. Muñoz, J. Hamilton, and P. Fullsack, Factors controlling the Alpine evolution of the central Pyrenees inferred from a comparison of observations and geodynamical models, J. Geophys. Res., 105(B4), 8121-8145, 2000.

Bernoulli, D., M. Bichsel, H. M. Bolli, M. O. Häring, P. A. Hochuli, and P. Kleboth, The Missaglia Megabed, a catastrophic deposit in the Upper Cretaceous Bergamo Flysch, northern Italy, Eclogae Geol. Helv., 74(2), 421-442, 1981.

Bernoulli, D., and W. Winkler, Heavy mineral assemblages from Upper Cretaceous South- and Austroalpine flysch sequences (Northern Italy and Southern Switzerland): source terranes and paleotectonic implications, Eclogae Geol. Helv., 83(2), 287-310, 1990.

Besse, J., and V. Courtillot, Revised and synthetic apparent polar wander paths of the African, Eurasian, North American and Indian plates, and true polar wander since 200 Ma, J. Geophys. Res., 96(B3), 4029_4050, 1991.

Bichsel, M., and M. O. Häring, Facies evolution of Late Cretaceous flysch in Lombardy (northern Italy), Eclogae Geol. Helv., 74(2), 383-420, 1981.

Bocharova, N. Y., and C. R. Scotese, Revised global apparent polar wander paths and global mean poles, Paleomap Project progress report no. 56, Department of Geology, University of Texas at Arlington, Texas, 20 pp., 1993.

Cande, S.C., and D.V. Kent, Revised calibration of the geomagnetic time scale for the late Cretaceous and Cenozoic, J. Geophys. Res., 100(B4), 6093-6095, 1995.

Castellarin, A., Ipotesi paleogeografica sul bacino del flysch sudalpino Cretacico, Boll. Soc. Geol. It., 95, 501-511, 1976.

Channel, J. E. T., Paleomagnetism and paleogeography of Adria, in Palaeomagnetism and Tectonics of the Mediterranean Region, edited by A. Morris and D. H. Tarling, pp. 119-132, Geol. Soc. Spec. Pub. 105, London, 1996.

Choukroune, P., Structure et evolution tectonique de la zone Nord Pyreneenne: analyse de la deformation dans une partie de chaine à schistosité sub-verticale, Mem. Soc. Geol. Fr., LV(127), 116 pp., 1976.

Cohen, C. R., R. A. Schweickert, and A. L. Odom, Age of emplacement of the Schistes Lustres nappe, Alpine Corsica, Tectonophysics, 73, 267-283, 1981.

Concheryo, A., and S. W. Wise Jr., Jurassic calcareous nannofossils from prerift sediments drilled during ODP Leg 173, Iberia Abyssal Plain, and their implications for rift tectonics, Proc. ODP, Sci. Results, 173, 1-24, 2001.

Cox, A., and R. B. Hart, Plate Tectonics: How It Works, 392 pp., Blackwell Scientific Publications, Palo Alto, Calif., 1986.

Degnan, P. J., and A. H. F. Robertson, Mesozoic-early Tertiary passive margin evolution of the Pindos ocean (NW Peloponnese, Greece), Sedimentary Geology, 117, 33-70, 1998.

Dercourt, J., L. P. Zonenshain, L.-E. Ricou, V. G. Kazmin, X. LePichon, A. L. Knipper, C. Grandjacquet, I. M. Sbortshikov, J. Geyssant, C. Lepvrier, D. H. Pechersky, J. Boulin, J.-C. Sibuet, L. A. Savostin, O. Sorokhtin, M. Westphal, M.L. Bazhenov, J. P. Lauer and B. Biju-Duval, Geological evolution of the Tethys belt from the Atlantic to the Pamirs since the Lias, Tectonophysics, 123, 241-315, 1986.

Dercourt, J., L. E. Ricou, and B. Vrielinck, Atlas Tethys, Paleoenvironmental Maps, Gauthier-Villars, 307 pp., Paris, 1993.

Dewey, J. F., W. C. Pitman, III, W. B. F. Ryan, and J. Bonnin, Plate tectonics and the evolution of the Alpine System, GSA Bull., 84(10), 3137-3180, 1973.

Dewey, J. F., M. L. Helman, E. Turco, D. H. W. Hutton, and S. D. Knott, Kinematics of the Western Mediterranean, in Alpine Tectonics, edited by M. P. Coward, D. Dietrich and R. G. Park, pp. 265-283, Geol. Soc. Spec. Publ. 45, London, 1989.

Faccenna, C., T. W. Becker, F. P. Lucente, L. Jolivet, and F. Rossetti, History of subduction and back-arc extension in the Central Mediterranean, Geophys. J. Int., 145, 809-820, 2001.

Garfunkel, Z., Constrains on the origin and history of the Eastern Mediterranean basin, Tectonophysics, 298(1-3), 5-35, 1998.

Genik, G. J., Regional framework, structural and petroleum aspects of rift basins in Niger, Chad and the Central African Republic (C.A.R.), Tectonophysics, 213, 169-185, 1992.

Görür, N., F. Y. Oktay, I. Seymen, and A. M. C. Şengör, Paleotectonic evolution of the Tuzgölü basin complex, Central Turkey: sedimentary record of a Neo-Tethyan closure, in The geological evolution of the Eastern Mediterranean, edited by J. E. Dixon and A. H. F. Robertson, pp. 467-482, Geol. Soc. Spec. Pub. 17, London, 1984.

Gradstein, F. M., F. P. Agterberg, J. G. Ogg, J. Hardenbol, P. van Veen, J. Thierry, and Z. Huang, A Mesozoic time scale, J. Geophys. Res., 99(B12), 24,051-24,074, 1994.

Guiraud, R., and J. C. Maurin, Early Cretaceous rifts of western and central Africa, Tectonophysics, 213, 153-168, 1992

Hames, W. E., P. R. Renne, and C. Ruppel, New evidence for geologically-instantaneous emplacement of earliest Jurassic Central Atlantic magmatic province basalts on the North American margin, Geology, 28(9), 859-862, 2000.

Klitgord, K. D., and H. Schouten, Plate kinematics of the central Atlantic, in The Geology of North America, vol. M, The Western North Atlantic Region, edited by P. R. Vogt and B. E. Tucholke, pp. 351-378, Geol. Soc. of America, Boulder, Colorado, 1986.

Lauer, J. P., Geodynamic evolution of Turkey and Cyprus based on paleomagnetic data, in The geological evolution of the Eastern Mediterranean, edited by J. E. Dixon and A. H. F. Robertson, pp. 483-491, Geol. Soc. Spec. Pub. 17, London, 1984.

Lepvrier, C., and E. Martínez-García, Fault development and stress evolution of the post-Hercynian Asturian Basin (Asturias and Cantabria, northwestern Spain), Tectonophysics, 184, 345-356, 1990.

McCall, G. J. H., The inner Mesozoic to Eocene ocean of South and Central Iran and the associated microcontinents, Geotectonics, 29(6), 490-497, 1996.

McElhinny, M. W., and J. Lock, Global palaeomagnetic data base project, Phys. Earth and Planet. Int., 63, 1-6, 1990.

Morris, A., and D. H. Tarling, Paleomagnetism and tectonics of the Mediterranean region: an introduction, in Palaeomagnetism and Tectonics of the Mediterranean Region, edited by A. Morris and D. H. Tarling, pp. 1-18, Geol. Soc. Spec. Pub. 105, London, 1996.

Mountrakis, D., Structural evolution of the Pelagonian zone in northwestern Macedonia, Greece, in The geological evolution of the Eastern Mediterranean, edited by J. E. Dixon and A. H. F. Robertson, pp. 581-590, Geol. Soc. Spec. Pub. 17, London, 1984.

Müller, R.D., W. R. Roest, J.-Y. Royer, L. M. Gahagan, and J. G. Sclater, Digital isochrons of the world's ocean floor, J. Geophys. Res., 102(B2), 3211-3214, 1997.

Muttoni, G., D. V. Kent, and J. E. T. Channell, Evolution of Pangea: paleomagnetic constraints from the Southern Alps, Italy, Earth Planet. Sci. Lett., 140(1-4), 97-112, 1996.

Okay, A. I., Distribution and characteristic of the north-west Turkish blueschists, in The geological evolution of the Eastern Mediterranean, edited by J. E. Dixon and A. H. F. Robertson, pp. 455-466, Geol. Soc. Spec. Pub. 17, London, 1984.

Olsen, P. E., Stratigraphic record of the early Mesozoic breakup of Pangea in the Laurasia-Gondwana rift system, Annual Rev. of Earth and Planet. Sci., 25, 337-401, 1997.

Pescatore, T., P. Renda, M. Schiattarella, and M. Tramutoli, Stratigraphic and structural relationships between Meso-Cenozoic Lagonegro basin and coeval carbonate platforms in southern Apennines, Italy, Tectonophysics, 315(1-4), 269-286, 1999.

Robertson, A. H. F., and J. E. Dixon, Introduction: aspects of the geological evolution of the Eastern Mediterranean, in The geological evolution of the Eastern Mediterranean, edited by J. E. Dixon and A. H. F. Robertson, pp. 1-74, Geol. Soc. Spec. Pub. 17, London, 1984.

Robertson, A. H. F., J. E. Dixon, S. Brown, A. Collins, A. Morris, E. A. Pickett, I. Sharp, and T. Ustaömer, Alternative tectonic models for the Late-Palaeozoic-Early Tertiary development of Tethys in the Eastern Mediterranean region, in Palaeomagnetism and Tectonics of the Mediterranean Region, edited by A. Morris and D. H. Tarling, pp. 239-263, Geol. Soc. Spec. Pub. 105, London, 1996.

Robertson, A. H. F., Mesozoic-Tertiary tectonic evolution of the easternmost Mediterranean area: Integration of marine and land evidence, Proc. ODP, Sci. Results, 160, 723-782, 1998.

Rowley, D. B., and A. L. Lottes, Plate-kinematic reconstructions of the North Atlantic and Arctic: Late Jurassic to Present, Tectonophysics, 155, 73-120, 1988.

Royer, J. Y., R. D. Müller, L. M. Gahagan, L. A. Lawver, C. L. Mayes, D. Nürnberg, and J. G. Sclater, A global isochron chart, Technical Rep. 117, 38 pp., University of Texas Institute for Geophysics, Austin, Texas, 1992.

Santantonio, M., Pelagic carbonate platforms in the geologic record: Their classification, and sedimentary and paleotectonic evolution, AAPG Bull., 78(1), 122-141, 1994.

Schettino, A., Computer aided paleogeographic reconstructions, Computers & Geosciences, 24(3), 259-267, 1998.

Schettino, A., Polygon intersections in spherical topology: Application to Plate Tectonics, Computers & Geosciences, 25(1), 61-69, 1999.

Schettino, A., and C. R. Scotese, A Synthetic APWP for Africa (Jurassic-Present) and Global Plate Tectonic Reconstructions, AGU Spring Meeting, Washington D.C., 2000.

Schettino, A., and C. R. Scotese, New Internet Software Aids Paleomagnetic Analysis and Plate Tectonic Reconstructions, Eos Trans. AGU, 82(45), 2001.

Schlische, R. W., Anatomy and evolution of the Triassic-Jurassic continental rift system, Eastern North America, Tectonics, 12(4), 1026-1042, 1993.

Şengör, A. M. C., and Y. Yilmaz, Tethyan evolution of Turkey: a plate tectonic approach, Tectonophysics, 75, 181-241, 1981.

Şengör, A. M. C., Y. Yilmaz, and O. Sungurlu, Tectonics of the Mediterranean Cimmerides: nature and evolution of the western termination of Palaeo-Tethys, in The geological evolution of the Eastern Mediterranean, edited by J. E. Dixon and A. H. F. Robertson, pp. 77-112, Geol. Soc. Spec. Pub. 17, London, 1984.

Spray, J. G., J. Bébien, D. C. Rex, and J. C. Roddick, Age constraints on the igneous and metamorphic evolution of the Hellenic-Dinaric ophiolites, in The geological evolution of the Eastern Mediterranean, edited by J. E. Dixon and A. H. F. Robertson, pp. 619-627, Geol. Soc. Spec. Pub. 17, London, 1984.

Srivastava, S. P., W. R. Roest, L. C. Kovacs, G. Oakey, S. Lévesque, J. Verhoef, and R. Macnab, Motion of Iberia since the Late Jurassic: Results from detailed aeromagnetic measurements in the Newfoundland Basin, Tectonophysics, 184, 229-260, 1990.

Stampfli, G. M., J. Marcoux, and A. Baud, Tethyan margins in space and time, Palaeogeogr., Palaeoclimatol., Palaeoecol., 87, 373-410, 1991.

Stampfli, G. M., J. Mosar , D. Marquer, R. Marchant, T. Baudin, and G. Borel, Subduction and obduction processes in the Swiss Alps, Tectonophysics, 296(1-2), 159-204, 1998.

Teixell, A., Crustal structure and orogenic material budget in the west central Pyrenees, Tectonics, 17(3), 395-406, 1998.

Turco, E., and A. Zuppetta, A kinematic model for the Plio-Quaternary evolution of the Tyrrhenian-Apenninic system: implications for rifting processes and volcanism, J. Volc. Geotherm. Res., 82, 1-18, 1998.

Van den Berg, J., and H. Zijderveld, Paleomagnetism in the Mediterranean area, in Alpine-Mediterranean Geodynamics, edited by H. Berckhemer and K. Hsü, pp. 83-112, AGU Geodynamics Series Vol. 7, Washington D.C., 1982.

Van der Voo, R., Paleomagnetism of the Atlantic, Tethys and Iapetus Oceans, 411 pp., Cambridge University Press, Cambridge, UK, 1993.

Veevers, J. J., Middle/Late Triasssic (230 ± 5 Ma) singularity in the stratigraphic and magmatic history of the Pangean heat anomaly, Geology, 17(9), 784-787, 1989.

Wilson, M., and R. Guiraud, Magmatism and rifting in Western and Central Africa, from Late Jurassic to Recent times, Tectonophysics, 213, 203-225, 1992.

Winterer, E. L., and A. Bosellini, Subsidence and sedimentation on Jurassic passive continental margins, Southern Alps, Italy, AAPG Bull., 65(1), 394-421, 1981.

Withjack, M. O., R. W. Schlische, and P. E. Olsen, Diachronous rifting, drifting, and inversion on the passive margin of central eastern North America: An analog for other passive margins, AAPG Bulletin, 82, 817-835, 1998.

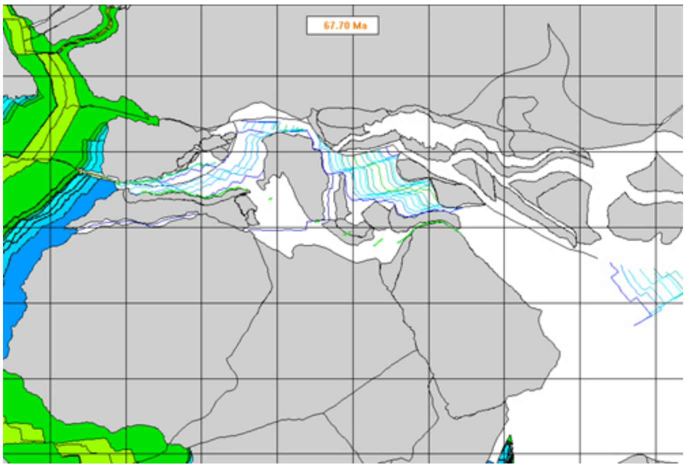
Yilmaz, Y., Ş. C. Genç;, E. Yiğitbaş, M. Bozcu, and K. Yilmaz, Geological evolution of the late Mesozoic continental margin of Northwestern Anatolia, Tectonophysics, 243, 155-171, 1995.

Zappaterra, E., Source-rock distribution model of the Periadriatic region, AAPG Bull., 78(3), 333-354, 1994.

A. Animation

Figure 22 is a movie illustrating the model proposed in this paper for the tectonic evolution of the Mediterranean region during the Jurassic and the Cretaceous. This animation can give a better understanding of the mechanism discussed in the previous sections regarding the formation of new oceanic crust during the Jurassic and the subsequent subduction during the Cretaceous. The process is represented in the movie by the appearance of modeled isochrons at stage boundaries and their subsequent motion as new crust forms at spreading centers. Hence, the total amount of oceanic crust that formed during the Jurassic is clearly bounded by the oldest isochrons. Similarly, the destruction of this crust at trench systems during the Cretaceous is easily identified by disappearance of isochrons as they approach the subduction complex.

Figure 22. Computer animation



Computer animation representing model proposed in this paper for the tectonic evolution of the Mediterranean region during the Jurassic and the Cretaceous