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Breakup of Pangaea and plate kinematics of the central Atlantic and Atlas regions

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SUMMARY

A new central Pangaea fit (type A) is proposed for the late Ladinian (230 Ma), together with a plate motions model for the subsequent phases of rifting, continental breakup and initial spreading in the central Atlantic. This model is based on: (1) a reinterpretation of the process of formation of the East Coast Magnetic Anomaly along the eastern margin of North America and the corresponding magnetic anomalies at the conjugate margins of northwest Africa and the Moroccan Meseta; (2) an analysis of major rifting events in the central Atlantic, Atlas and central Mediterranean and (3) a crustal balancing of the stretched margins of North America, Moroccan Meseta and northwest Africa. The process of fragmentation of central Pangaea can be described by three major phases spanning the time interval from the late Ladinian (230 Ma) to the Tithonian (147.7 Ma). During the first phase, from the late Ladinian (230 Ma) to the latest Rhaetian (200 Ma), rifting proceeded along the eastern margin of North America, the northwest African margin and the High, Saharan and Tunisian Atlas, determining the formation of a separate Moroccan microplate at the interface between Gondwana and Laurasia. During the second phase, from the latest Rhaetian (200 Ma) to the late Pliensbachian (185 Ma), oceanic crust started forming between the East Coast and Blake Spur magnetic anomalies, whereas the Morrocan Meseta simply continued to rift away from North America. During this time interval, the Atlas rift reached its maximum extent. Finally, the third phase, encompassing the time interval from the late Pliensbachian (185 Ma) to chron M21 (147.7 Ma), was triggered by the northward jump of the main plate boundary connecting the central Atlantic with the Tethys area. Therefore, as soon as rifting in the Atlas zone ceased, plate motion started along complex fault systems between Morocco and Iberia, whereas a rift/drift transition occurred in the northern segment of the central Atlantic, between Morocco and the conjugate margin of Nova Scotia. The inversion of the Atlas rift and the subsequent formation of the Atlas mountain belt occurred during the Oligocene–early Miocene time interval. In the central Atlantic, this event was associated with higher spreading rates of the ridge segments north of the Atlantis FZ. An estimate of 170 km of dextral offset of Morocco relative to northwest Africa, in the central Atlantic, is required by an analysis of marine magnetic anomalies. Five plate tectonic reconstructions and a computer animation are proposed to illustrate the late Triassic and Jurassic process of breakup of Pangaea in the central Atlantic and Atlas regions.

Key words: Magnetic anomalies: modelling and interpretation; Marine magnetics and palaeomagnetics; Mid-ocean ridge processes; Oceanic transform and fracture zone processes; Kinematics of crustal and mantle deformation.

1 INTRODUCTION

The breakup of Pangaea, and the subsequent origin of the modern continents, occurred in two main steps, made up, in turn, by several secondary tectonic phases. During the first, relatively short, interval encompassing the late Triassic–early Jurassic time interval, both Laurasia and southeast Gondwana separated from central Pangaea, a large continental block comprising Africa and South America (e.g. Schettino & Scotese 2005). The second, much longer, time interval spanned the entire Cretaceous and was characterized by further fragmentation of the three continental masses formed previously. In this paper, we initially focus on the process of separation of Laurasia from Gondwana and the subsequent formation of the central Atlantic during the early Jurassic. We will show that this time interval can be described by a succession of three distinct phases, each being characterized by different tectonic style, plate boundary system and kinematic parameters. Next, we will review how this initial mechanism of rifting affected the subsequent tectonic evolution of the central Atlantic and Atlas regions until recent times.

The reconstruction of the initial phase of breakup of Pangaea, and the subsequent opening of the central Atlantic, has been the subject of detailed studies since the late 1960s. Bullard *et al.* (1965) were the first authors to use rigorous computational methods for determining the relative positions of the continents around the North and South Atlantic before the late Triassic–early Jurassic phase of breakup. This pioneering work was followed during the 1970s and 1980s by a series of studies focused on the details of the seafloor spreading (SFS) process that took place after the rifting stage (Heirtzler *et al.* 1968; Le Pichon & Fox 1971; Pitman & Talwani 1972; Franchetau 1973; Sclater *et al.* 1977; Olivet *et al.* 1984; Klitgord & Schouten 1986; Sundvik & Larson 1988; Roest *et al.* 1992). In particular, a better knowledge of the seismic structure of the continental margins of eastern North America and northwest Africa, along with the availability of new marine magnetic anomaly data, allowed the development of comprehensive models for the early stages of Pangaea breakup as well as for the subsequent tectonic evolution of the central Atlantic ocean. More recently, Sahabi *et al.* (2004) proposed a new fit of North America and northwest Africa, which, for the first time, considers Morocco as an independent plate and predates by 20 Myr, the formation of the first oceanic crust in the central Atlantic with respect to the previous models.

The most widely used kinematic model is undoubtedly the one proposed by Klitgord & Schouten (1986), where the North American and northwest African pre-breakup relative positions are estimated using post-breakup SFS data. These authors criticized both the geometrical best fit of Bullard *et al.* (1965) and the reconstructions based on palaeomagnetic data (e.g. Morel & Irving 1981; Van der Voo 1983); whereas the latter used too sparse data that did not necessarily account for the pre-breakup geometry, the reconstruction of Bullard *et al.* (1965) was simply obtained by matching selected isobaths from both sides of the Atlantic Ocean. This approach seemed rather arbitrary, because there is not a simple relationship between bathymetric features and continent–ocean boundaries (COBs).

Bullard *et al.* (1965) applied the fitting algorithm to the 100, 500, 1000 and 2000 fm isobaths (1 fathom = 6 ft), then chose the best fitting set of curves (500 fm = 914.4 m) as most representative of the conjugate COBs. New geophysical data from the two conjugate margins, obtained during the 1980s, allowed a more accurate definition of the COBs. We note, nevertheless, that Van der Voo (1990) carried out a comparison of seven reconstructions of the pre-breakup configuration of the North Atlantic region, six of them being based on SFS data, and showed that the best match between the North American and European apparent polar wander (APW) paths for the Middle Ordovician through Early Jurassic interval was obtained exactly using the tight North Atlantic closure pole of Bullard *et al.* (1965). This surprising result can be easily explained considering that SFS data alone do not include all the information necessary to estimate the amount of continental stretching during the rifting stage; accurate reconstructions of the pre-rift geometry always require analysis and balancing of crustal profiles at both continental margins.

In this regard, the central Atlantic fit proposed by Klitgord & Schouten (1986) considers minimum and maximum Euler rotations for the closure of the central Atlantic. Starting from anomaly M25, the stage pole associated with these rotations was calculated using fracture zone trends. The minimum closure angle simply removes the oceanic crust between the East Coast magnetic anomaly (ECMA) on the North American margin and a chain of salt diapirs offshore northwest Africa, because these features were considered markers of the COBs. The maximum closure angle was estimated by assuming that basement hinge zones must not overlap because they represent the limits of unstretched continental crust or, in other words, the landward edges of marginal rift basins.

We will solve the problem of finding a reliable pre-breakup configuration of central Pangaea on the basis of the following simple observations: the southern segment of the ECMA and the Blake Spur magnetic anomaly (BSMA; Fig. 1) are conjugate oceanic anomalies that were generated during the initial pulse of SFS between North America and northwest Africa. This concept, first proposed by Vogt (1973), implies that the BSMA does not have a conjugate on the African side because it formed at the African margin, just while the ECMA was forming at the North American margin. Therefore, a ridge jump occurred later that left both anomalies close to the North American margin. If this scenario is correct, we can apply the following steps for determining quantitatively the fit parameters: (1) a geometrical best-fit algorithm, improved version of the classic Bullard's *et al.* (1965) procedure, is used to find the best match of the ECMA against the BSMA and determine the closure reconstruction pole of the proto-Atlantic ocean, which is the region bounded by these two anomalies; (2) an algorithm for finding the location and geometry of the African margin, conjugate to the BSMA at chron M25, is then used to determine the reconstruction pole of North America with respect to northwest Africa at the ridge jump time, and (3) a crustal balancing procedure is finally applied to the North American margin to estimate the pre-rift reconstruction parameters.

The northern segment of the ECMA, which is associated in part with a non-volcanic margin (Keen & Potter 1995a; Keen & Potter 1995b), can be considered as the conjugate of the Moroccan margin S1 anomaly (Roeser 1982; Verhoef *et al.* 1991; Roeser *et al.* 2002; Figs 1 and 2). We will not assume rigid behaviour of northwest Africa during the breakup of Pangaea. In fact, there is strong geological and geophysical evidence that support the existence of at least one separate Moroccan Plate, first during the late Triassic and the early Jurassic and later during the Oligocene–early Miocene. However, in this instance, the lack of continuity and uncertainties in the geometry of the S1 anomaly prevent the direct application of geometrical best-fit algorithms for the determination of the oceanic closure reconstruction pole. Therefore, we will use an indirect approach, based on the analysis of magnetic anomaly data from the central Atlantic, to determine the reconstruction pole of Morocco relative to North America at the time of formation of the first oceanic crust. Then, the pre-rift configuration will be estimated as before through crustal balancing of the stretched continental margins of Morocco and Nova Scotia.

Apart from the different pre-breakup geometry of Pangaea, which simply represents an improvement of previous Pangaea type-A models, the reconstructions discussed in this paper propose a significantly different timing for the rifting and spreading events in the whole central Pangaea domain, including the central Atlantic and Atlas regions.

Klitgord & Schouten (1986) proposed the following estimation for the time of formation of the first oceanic crust in the central Atlantic. They observed that the velocity of full spreading between chron M21 (147.7 Ma) and the oldest crust drilled at DSDP Leg 76 Site 534 (Fig. 1; middle Callovian, according to Gradstein & Sheridan 1983) was constant and equal to 38 mm yr^{-1}. Then, by extrapolation of this rate to the location of the BSMA, they obtained an age of 170 Ma. This time was not considered as the age of the BSMA, but the time of the eastward ridge jump that left a remnant

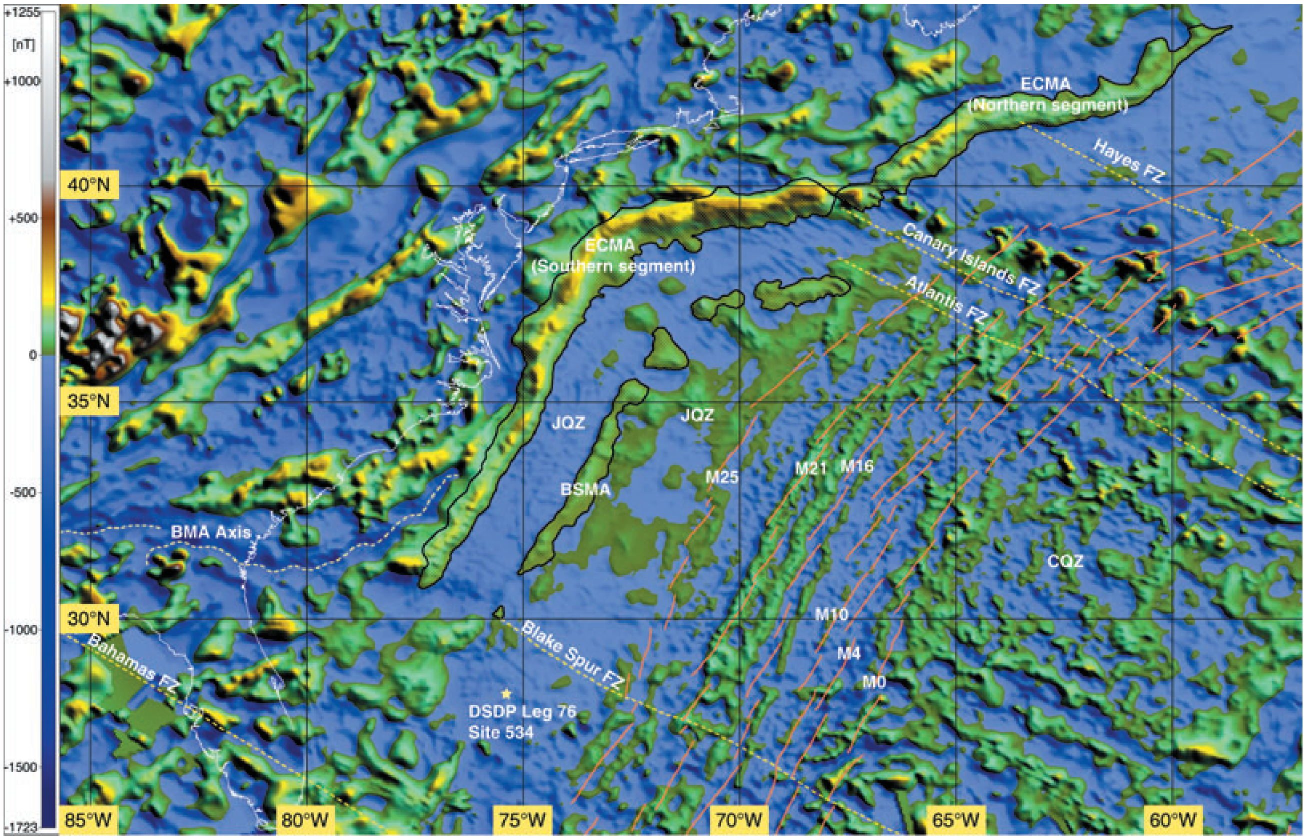

Figure 1. Magnetic anomalies of the eastern North American margin and the western central Atlantic, extracted from the global data set of Korhonen *et al.* (2007). The 0 nT contour lines bounding the segments of the ECMA and BSMA are shown as black lines. JQZ and CQZ are, respectively, the Jurassic and the Cretaceous quiet zones. BMA is the Brunswick magnetic anomaly. M-series lineations are from NGDC (1993).

oceanic seaway between the BSMA and the ECMA. Finally, further extrapolation of the 38 mm yr^{-1} spreading rate allowed to estimate an age of 175 Ma for the first oceanic crust in the proto-Atlantic.

Today accurate crustal, structural and stratigraphic studies of the eastern North American and northwest African margins, published since the 1990s, not only allow a better determination of the amount of crustal extension during the rifting stages but also provide a precise timing of the tectonic events that occurred during the breakup of Pangaea. For instance, it has been shown that both rifting and onset of SFS at the eastern North American margin were spatially diachronous events (Withjack *et al.* 1998; Schlische *et al.* 2002; Schlische 2003). In particular, we now know that (1) the crust below the ECMA in the zone between the Blake Spur fracture zone and the Georges Bank Basin (Fig. 1), which was emplaced during the first pulse of SFS as an anomalous oceanic crust (Sheridan *et al.* 1993; Holbrook *et al.* 1994a,b; Talwani *et al.* 1995; Oh *et al.* 1995; Talwani & Abreu 2000), is correlated with a tectonic inversion of the rift structures of the eastern margin of North America during the latest Rhaetian (∼200 Ma; Schlische *et al.* 2002); (2) this event is also coeval with the magmatic activity in the Central Atlantic Magmatic Province (CAMP), because the age of the oldest lava is close to the Triassic–Jurassic boundary (Fowell *et al.* 1994); (3) more to the north, in the Fundy rift zone, basin inversion occurred at a later time (∼185 Ma; Withjack *et al.* 1998), and (4) according to Steiner *et al.* (1998), the age of the N-MORB flows and breccias at the base of the Fuerteventura sequence in the Canary Islands, which is located to the west of the S1 magnetic anomaly, is Toarcian.

These observations suggest that the first oceanic crust in the central Atlantic formed at ∼200 Ma in the area bounded by the Blake Spur and Atlantis fracture zones, whereas to the south and to the north of this region, rifting was still in progress. Furthermore, the stratigraphic record of the Fundy Basin requires a drift/rift transition at ∼185 Ma for the North American and Moroccan conjugate margins and a corresponding age for the northern segment of the ECMA and the S1 anomaly. We assume that this event was coeval with the ridge jump that isolated the proto-Atlantic area as an oceanic remnant between the ECMA and BSMA anomalies. Finally, the middle Callovian age of the oldest crust drilled at DSDP Leg 76 Site 534 (Gradstein & Sheridan 1983) suggests that the onset of SFS in the area of the Blake Plateau occurred even later at ∼170 Ma, in agreement with the estimation of Klitgord & Schouten (1986).

In the next sections, we will first determine the pre-rift configuration of Pangaea in the central Atlantic, through the establishment of a reliable fit of North America, Morocco and northwest Africa during the late Triassic. Then, four tectonic phases and associated plate reconstructions will be discussed to illustrate the changing evolution of the central Atlantic and Atlas regions in the time interval from the late Ladinian to the Tithonian and during the Oligocene–early Miocene building of the Atlas orogen. The geomagnetic time scales of Cande & Kent (1995) and Grandstein *et al*. (1994) were used respectively for anomalies younger than Chron 34 (83.5 Ma) and for older times.

2 THE FIT OF THE CONTINENTS AROUND THE CENTRAL ATLANTIC

The southern segments of the ECMA and the BSMA (Fig. 1) play a key role in the reconstruction of the early stages of opening of the

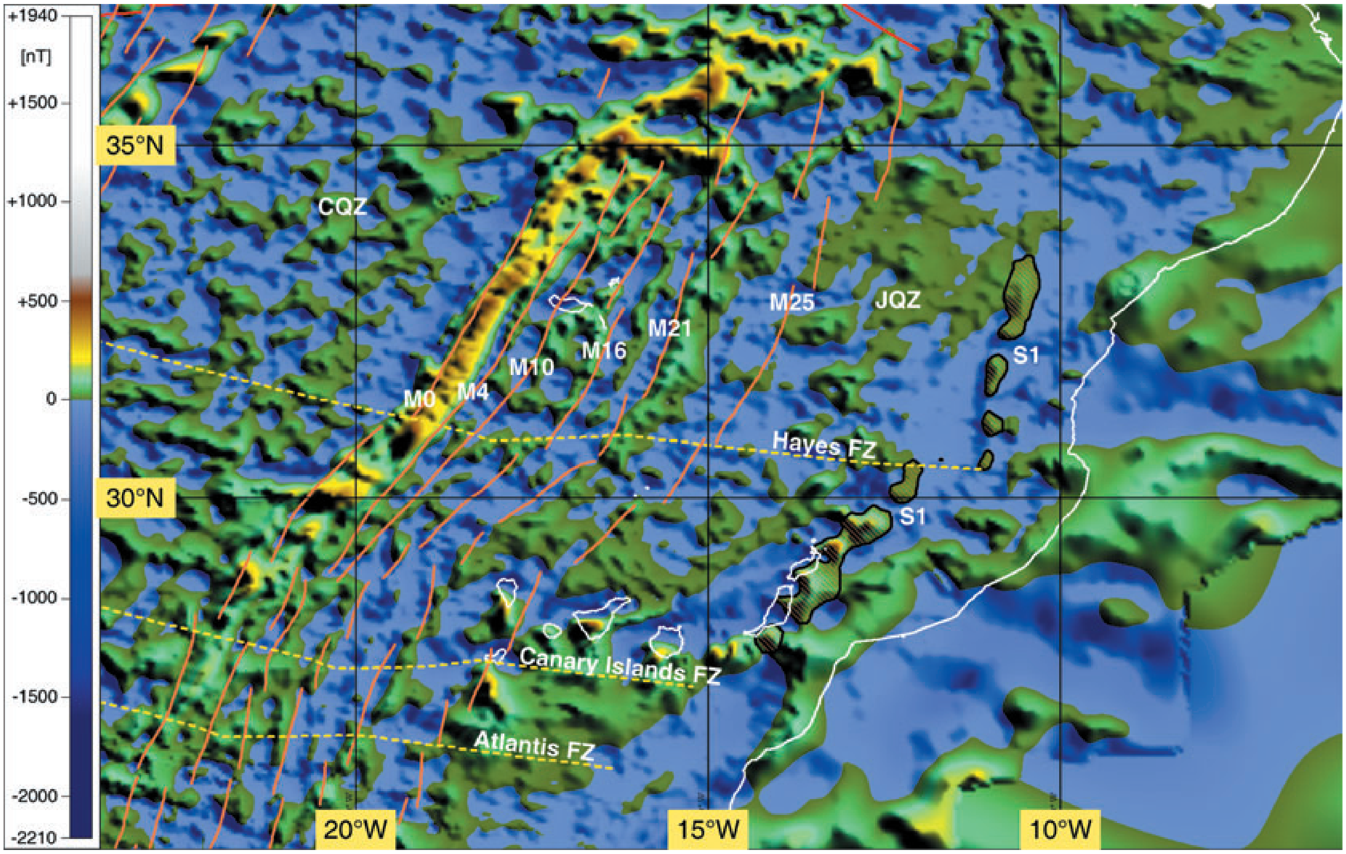

Figure 2. Magnetic anomalies offshore Morocco, extracted from the global data set of Korhonen *et al.* (2007). The 0 nT contour lines bounding the segments of the S1 anomaly are shown as black lines. M-series lineations are from NGDC (1993).

Atlantic Ocean. The origin of these anomalies has been revealed through the interpretation of several seismic profiles along the eastern North American continental margin. Holbrook *et al.* (1994a) analysed a seismic profile (Line 6) offshore South Carolina, showing that the ECMA was associated with a transitional zone of thinned crust having high seismic velocities (6.3–6.9 Km s^{-1}) in the basement, high densities (2870–3090 Kg m^{-3}) and high magnetizations (2.6–5.0 A m^{-1}). Furthermore, the upper part of these bodies formed a wedge of seaward dipping reflectors (SDR), which could be interpreted as basaltic flows. Some time later Talwani *et al.* (1995) and Talwani & Abreu (2000) reviewed seismic reflection and refraction data acquired by the EDGE project (Line 801, Sheridan *et al.* 1993) off New Jersey to show that the crust lying above the Moho, and underlying the carbonate bank was an anomalous oceanic crust composed by extrusive and intrusive rocks emplaced during the initial phase of drifting. Also in this interpretation, the upper part of the anomalous body was represented by SDR responsible for the formation of the ECMA. More to the north, at the latitude of the Nova Scotia margin, the continent–ocean transition zone of eastern North America assumes a non-volcanic character. For this area, Wu *et al.* (2006) have recently shown that a low-amplitude ECMA (northern segment) could be associated with the transition from extremely thinned continental crust to exhumed serpentinized mantle.

In summary, the available data strongly support the idea that the ECMA and, by consequence, the BSMA are true oceanic anomalies as suggested by Roest *et al.* (1992). In fact, from the point of view of plate kinematics the magnetic anomaly associated with the emplacement of the anomalous oceanic crust observed by Talwani *et al.* (1995) is perfectly equivalent to the anomaly generated by an exhumed serpentinized mantle at the northern non-volcanic margin. Furthermore, in the interpretation of Vogt (1973), the ECMA and BSMA are conjugate anomalies that can be used to constrain the relative positions of North America and Africa at the time of formation of the first oceanic crust. Accordingly, we used the recent world digital magnetic anomaly map (WDMAM) of Korhonen *et al.* (2007; Fig. 1) to determine, accurately, the geometry of these anomalies. The application of an improved version of the Bullard's *et al.*'s (1965) geometrical best-fit algorithm (see Appendix) to the 0 nT contours of the ECMA and the BSMA furnished an Euler pole $\boldsymbol{e} = (+24.51°\text{N}, 47.00°\text{E})$, a rotation angle $\Omega = +1.49°$ and a misfit $\chi^2 = 0.1745$. The choice of the 0 nT contour lines for the fitting procedure is justified by the fact that only the zero contour line of the BSMA has sufficient continuity to produce a reliable fit. These data determine location and angle of the stage pole of northwest Africa relative to North America between the onset of SFS (200 Ma) and the time of ridge jump (185 Ma).

Fig. 3 illustrates the fit of BSMA against the ECMA, obtained using the method described in the Appendix. This Figure also shows the position of BSMA with respect to ECMA in the reconstruction of Klitgord & Schouten (1986). We note that the stage pole of Klitgord & Schouten (1986) is in agreement with the prolongation of the fracture zone trend within the Jurassic Quiet Zone (JQZ). However, this reconstruction furnishes a very poor fit of the ECMA–BSMA pair and predicts flow lines that are discordant with respect to the pattern of gravity anomalies. This implies that the trend of fracture zones within the proto-Atlantic area cannot be obtained by prolongation of the trend observed outside this region.

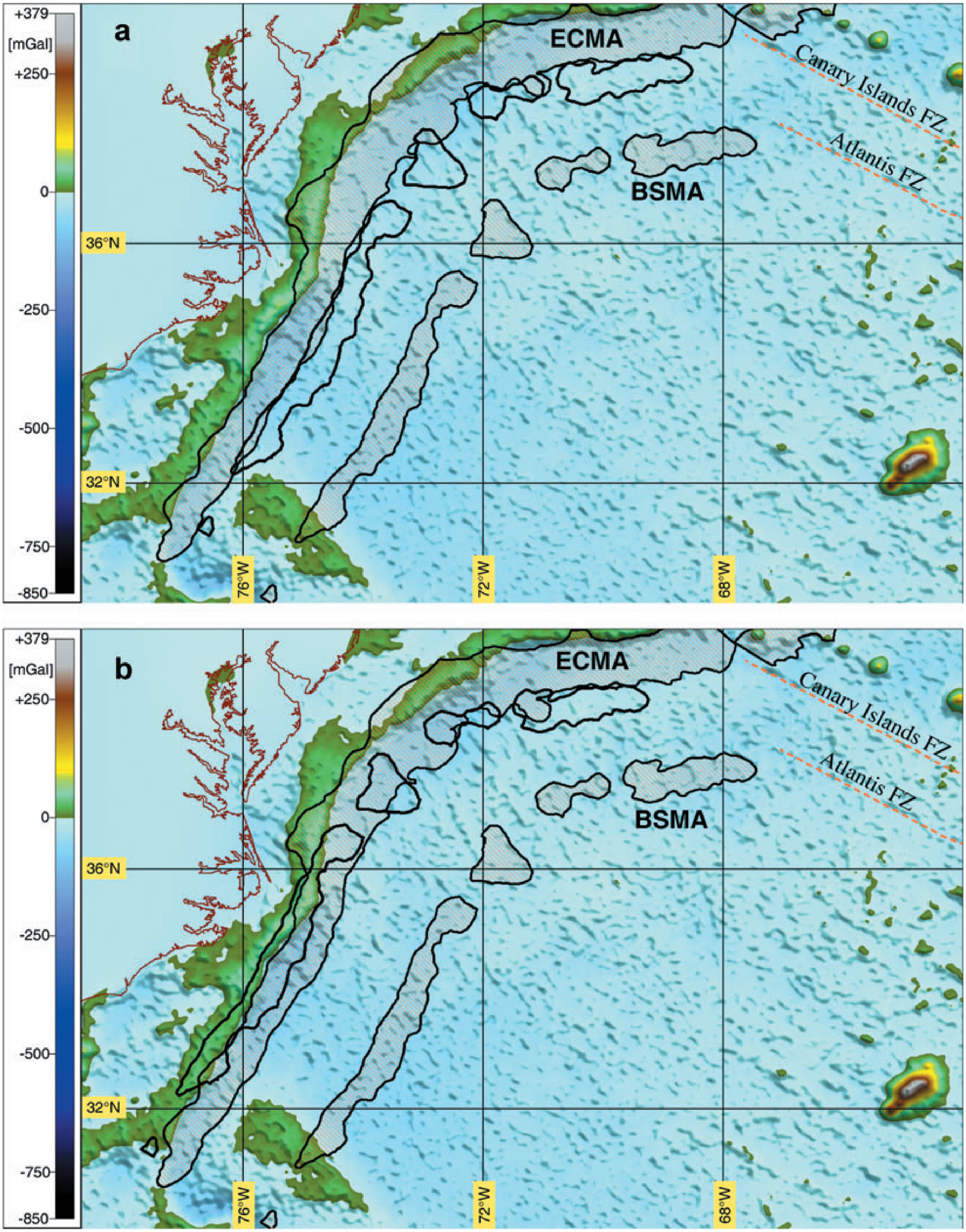

Figure 3. Geometrical fit of BSMA against ECMA proposed in this paper (a) and fit of the two anomalies based on the stage pole of northwest Africa with respect to North America in the model of Klitgord & Shouten (1986) (b). The background pattern of free-air gravity anomalies (Sandwell & Smith 1997) shows that the trend of fracture zones cannot be extrapolated to the proto-Atlantic region; in this area, the pattern is in substantial agreement with the stage pole proposed in this paper.

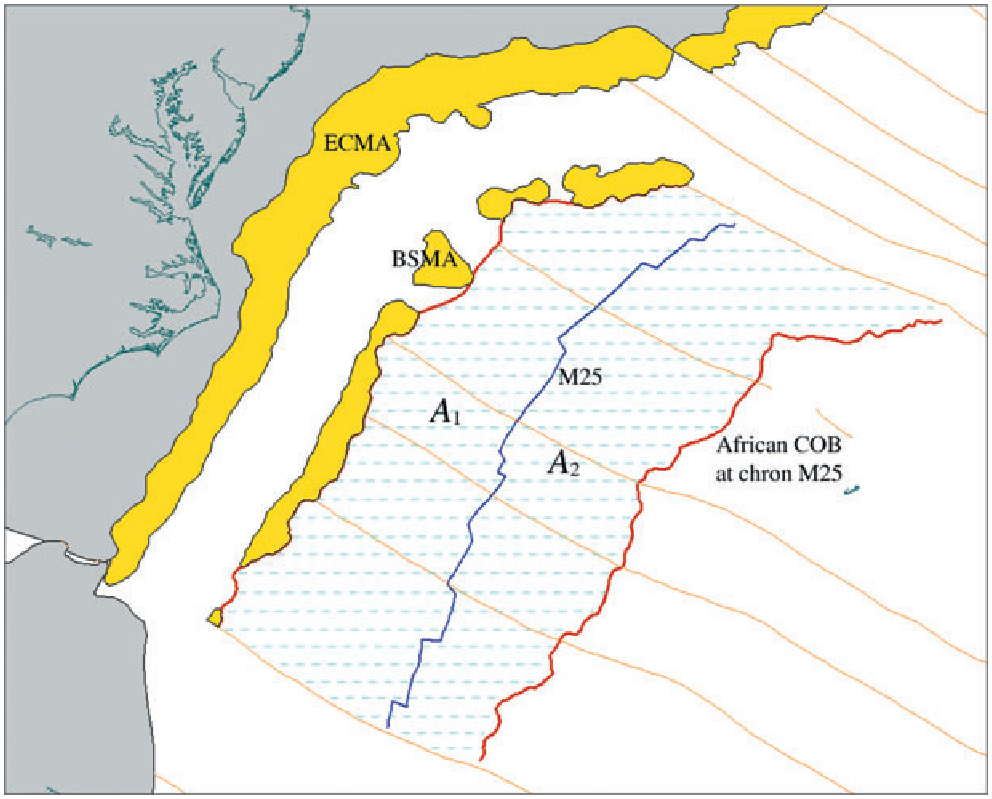

Figure 4. African COB at chron M25 in the North American reference frame. This boundary has been built by rotation of the eastern margin of the BSMA about the stage pole of northwest Africa relative to North America between 185 Ma and chron M21, in such a way that the areas A_1 and A_2 of the oceanic lithosphere created to the east and to the west of the spreading centre are equal (dashed areas).

To complete the reconstruction of the position of North America with respect to northwest Africa at the time of formation of the first oceanic crust, we must now fit the BSMA against the northwest African continental margin. This operation cannot be performed accurately by direct match with the present day COB because there is large uncertainty about the location of this boundary offshore northwest Africa. The best geological constrain regarding the location of the northwest African COB is represented by salt deposits in the deepest half-grabens of the stretched margin, as they formed during the last stage of rifting, immediately prior to the onset of SFS (Tari & Pamić 1998; Davison 2005). However, these data only determine the landward limit of a region where the true COB is located and 'not' the COB itself. Therefore, we use kinematic constraints to determine the closure Euler pole of the central Atlantic between North America and northwest Africa.

Consider the plate tectonic configuration of the central Atlantic region at chron M25 (154.3 Ma). The analysis of fracture zone trends shows that we do not introduce severe errors in the analysis considering the time interval between the BSMA ridge jump (185 Ma) and chron M21 (147.7 Ma) as a single stage, characterized by a unique stage pole. In this instance and in absence of other constraints, we can assume that the location of the African margin at chron M25 was symmetric, in North American coordinates, to the BSMA about the location of the North American M25 isochron (Fig. 4). From Fig. 4, it is evident that we cannot determine a fit between the BSMA and anomaly M25. This observation suggests that an important change occurred at 185 Ma in the spreading direction, which caused the re-orientation of ridge segments through propagation and asymmetric spreading, and confirms that the fracture zone trends cannot be extrapolated to the proto-Atlantic zone. Therefore, a curve that is symmetric with respect to the BSMA, representative of the African COB at chron M25, can only be found by rotating the BSMA about the BSMA-M21 stage pole by an angle Ω, so that the amount of oceanic lithosphere on both sides of the M25 spreading centre is the same. This is equivalent to select a rotation angle Ω that is twice the average angle carrying individual points from the eastern flank of the BSMA onto the M25 anomaly. We computed the stage pole of northwest Africa with respect to North America using the finite rotations of Klitgord & Schouten (1986). Then, the rotation angle resulted to be $\Omega = 4.75^\circ$. Therefore, the complete stage pole for the time interval from 185 to chron M25 is (60.7°N, 30.6°E, 4.8°).

We can estimate the geometry and location of the present northwest African COB by rotating the curve specular to BSMA, which was obtained using the procedure described above, through the finite reconstruction pole of North America with respect to northwest Africa at chron M25. Fig. 5 illustrates the resulting shape of the margin as well as the location of two features that some authors have considered as markers of the African COB. Regarding the first of them, the seaward limit of the salt basins, we already stated that this boundary does not constrain the precise location of the COB but only the landward margin of the region, where the COB is located. The second feature is represented by the West African Coast magnetic anomaly (WACMA). Roussel & Liger (1983) erroneously interpreted this anomaly as the landward boundary of the JQZ, conjugate to the ECMA, hence the magnetic marker of the African COB. However, geophysical evidence suggests that the anomaly conjugate to the ECMA is the BSMA. In our view, no magnetic anomaly is expected at the northwest African margin as an expression of the magmatic bodies that were emplaced at the breakup time. Therefore, we will follow the interpretation of Klitgord & Schouten (1986), who considered the WACMA as a

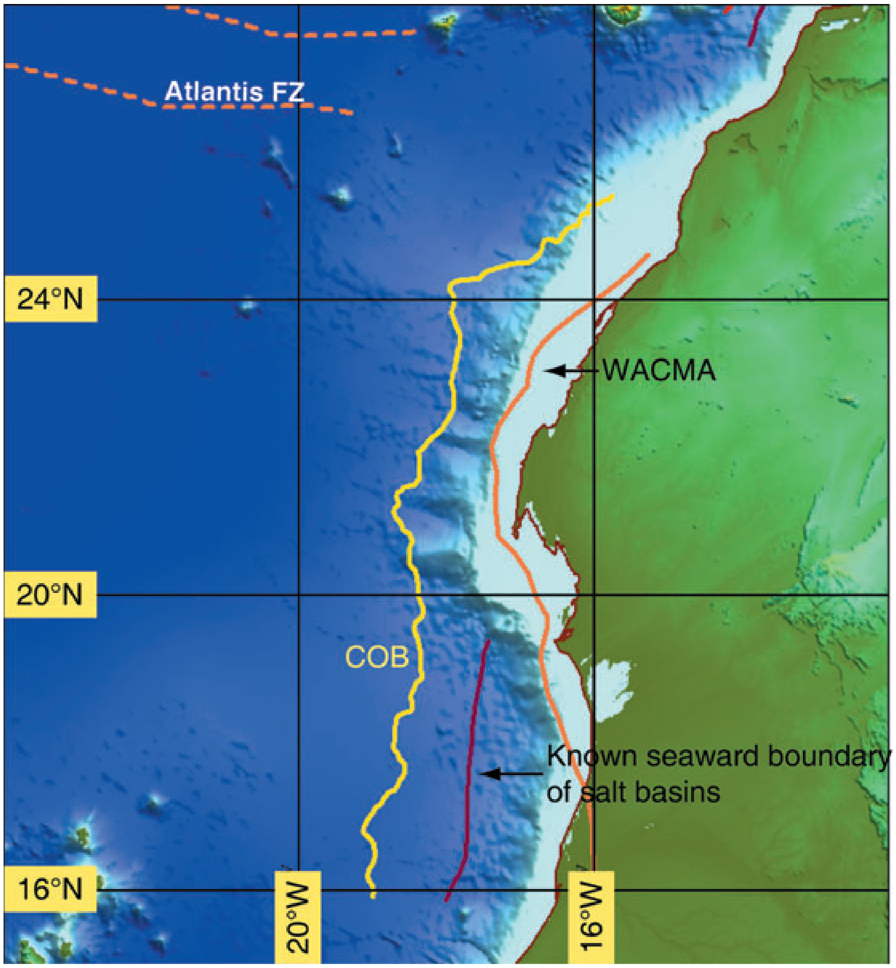

Figure 5. Location of the northwest African COB, inferred from the geometry of BSMA and kinematic analysis. The map also illustrates the bathymetry along the margin, the location of the West African Coast magnetic anomaly (WACMA) and the seaward boundary of the salt basin offshore Mauritania, according to Tari & Pamić (1998).

marker of the basement hinge zone. Fig. 5 shows that the northern part of the COB proposed here practically coincides with the base of the continental scarp, whereas the southern segment is located 90–120 km seaward of the salt basin. This discrepancy could imply that (1) the M25–M21 stage pole cannot be extrapolated to the BSMA, (2) the M25 finite reconstruction pole is wrong or (3) there was no symmetry about the ridge at chron M25. All these hypothesis should be investigated in future work, and new geophysical surveys offshore Mauritania are necessary to assess the validity of the COB location shown in Fig. 5, which is based on theoretical considerations alone.

The finite reconstruction pole of BSMA relative to northwest Africa can be easily calculated combining the stage pole of closure for the oceanic area between M25 and BSMA and the finite reconstruction pole of N. America relative to northwest Africa at M25. The result is listed in Table 1. To complete our fit of central Pangaea, we must now consider the northern segment of ECMA and its Moroccan conjugate, the S1 anomaly.

3 FITTING MOROCCO TO NORTH AMERICA

We mentioned the fact that a geometrical fit of the ECMA and S1 anomalies produces an unreliable result because of the discontinuity of S1 and some uncertainty about its geometry. Therefore, we adopted a fit based on SFS data and kinematic considerations. Our starting point is represented by two observations: (1) a continental block separated from northwest Africa during the late Triassic and early Jurassic, leading to the formation of the Atlas Rift, a large trough extending from Morocco to Tunisia (Laville & Piqué 1991; Laville *et al.* 1995, 2004; Piqué & Laville 1995; Piqué *et al.* 1998a; Le Roy & Piqué 2001), with a syn-rift sequence up to 12 km thick and an estimated stretching factor $\beta = 2.15$ (Beauchamp *et al.* 1996); and (2) the rift structures were inverted during the Cenozoic, forming the present Atlas mountain belt (Beauchamp *et al.* 1996; Frizon de Lamotte *et al.* 2000; Piqué *et al.* 2002), whose elevations reach 4000 m in the western region. Obviously, the independent motion of the Moroccan Plate (including Algeria and Tunisia) relative to northwest Africa during the late Triassic–early Jurassic time interval and, later, during the Cenozoic influenced the pattern of SFS anomalies in the central Atlantic. In particular, considering the region extending from the northern ECMA to the S1 anomaly, we expect a larger area of Cenozoic oceanic crust with respect to the region between the BSMA and the northwest African COB. This is because when the Moroccan Plate was offset to the east during the Atlas orogeny, a greater spreading rate with respect to North America must have accommodated this eastward motion.

The boundaries of the oceanic part of the Moroccan Plate can be established as follows. The northern boundary is clearly

Table 1. Early Mesozoic and Tertiary rotation model for the central Atlantic and Atlas regions.

Age (Myr)	Latitude (°)	Longitude (°)	Angle (°)	References
		BSMA–northwest Africa		
20.1	79.57	37.84	5.29	Klitgord & Schouten (1986)
33.1	76.41	7.12	9.81	Klitgord & Schouten (1986)
147.7	66.50	341.90	61.92	Klitgord & Schouten (1986)
154.3	67.15	344.00	64.70	Klitgord & Schouten (1986)
185.0	68.01	347.20	69.22	This paper
		North America–BSMA		
185.0	0.00	0.00	0.00	This paper
200.0	24.51	47.00	1.49	This paper
230.0	24.51	47.00	3.74	This paper
		Southern Florida–northern Florida		
154.3	0.00	0.00	0.00	This paper
185.0	−54.87	207.94	0.57	This paper
230.0	−55.17	152.55	0.95	This paper
		Blake Plateau–northern Florida		
175.0	0.00	0.00	0.00	This paper
185.0	17.42	272.77	1.90	This paper
		Morocco–northwest Africa		
19.0	0.00	0.00	0.00	This paper
33.1	−65.93	162.92	2.87	This paper
185.0	−65.93	162.92	2.87	This paper
		Morocco–North America		
185.0	−67.98	166.91	72.09	This paper
230.0	−68.35	170.43	73.19	This paper

represented by the Azores–Gibraltar Fracture Zone, which starts at the Azores triple junction and presently separates Africa from Eurasia. Conversely, at first glance, the southern boundary is not so well constrained—we only know that the eastward motion of Morocco relative to northwest Africa must have been accommodated along one or more fracture zones north of the BSMA. This implies that the eastern tract of one or more fracture zones was converted into a right-lateral strike-slip fault during the Tertiary, and that the eastern termination of this fault was aligned with compressive structures in western Morocco. The sketch in Fig. 6 illustrates the expected pattern of deformation of the oceanic isochrons in the eastern central Atlantic region, in the hypothesis that Morocco behaved as a single rigid plate.

Three fracture zones are candidate to represent the southern boundary of the Moroccan Plate. From the south to the north, they are the Atlantis FZ, the Canary Islands FZ and the Hayes FZ (Fig. 7). The westernmost tract of the Atlantis FZ bounds to the north the proto-Atlantic ocean. Therefore, it separates at the North American margin the southern segment of the ECMA, which is conjugate to the BSMA, from a northern segment conjugate to the S1. The easternmost tract of this fracture zone is located in the Boujdour Basin of Morocco. An important characteristic of both the Atlantis and Hayes FZ is represented by the presence of eastward offsets of the Mid-Atlantic Ridge (MAR) at the crossing point. This characteristic is interesting because we expect that the Cenozoic eastward escape of Morocco caused dextral offset of the spreading centre between Morocco and North America, as illustrated in Fig. 6. This feature is not present in the Canary Islands FZ. A second key characteristic, which, this time, is shared by the Canary Islands and Hayes FZ, is associated with the evident alignment of their eastern terminations with offshore and on land transpressive structures.

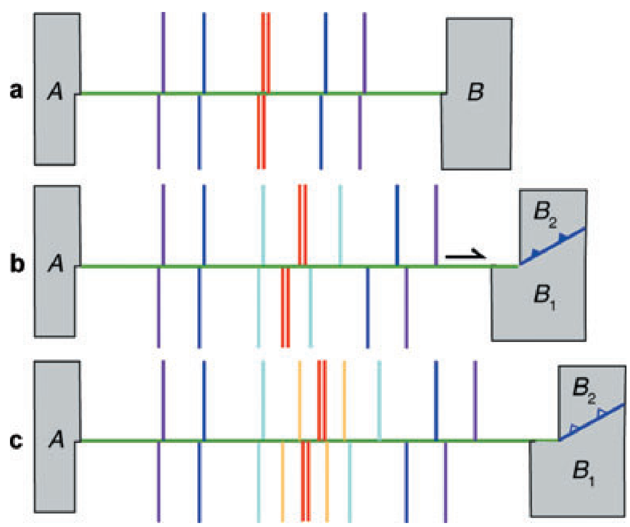

Figure 6. Sketch map illustrating the process of deformation of oceanic isochrons associated with the formation of the Atlas orogen. In (a), *B* is a single plate and magnetic lineations have a constant original offset at the fracture zone. At time (b), *B* starts deforming through the eastward escape of its northern part B_2; eastern isochrons are deformed, and the offset between spreading segments is increased. From time (c) onward, newly formed isochrons are undeformed and have the same offset of the ridge segments.

We now consider the geological evidence of compressive and transpressive structures onshore and offshore western Morocco, which could be linked to right-lateral strike-slip motion in the central Atlantic. The geology of the Canary Islands (Fig. 7),

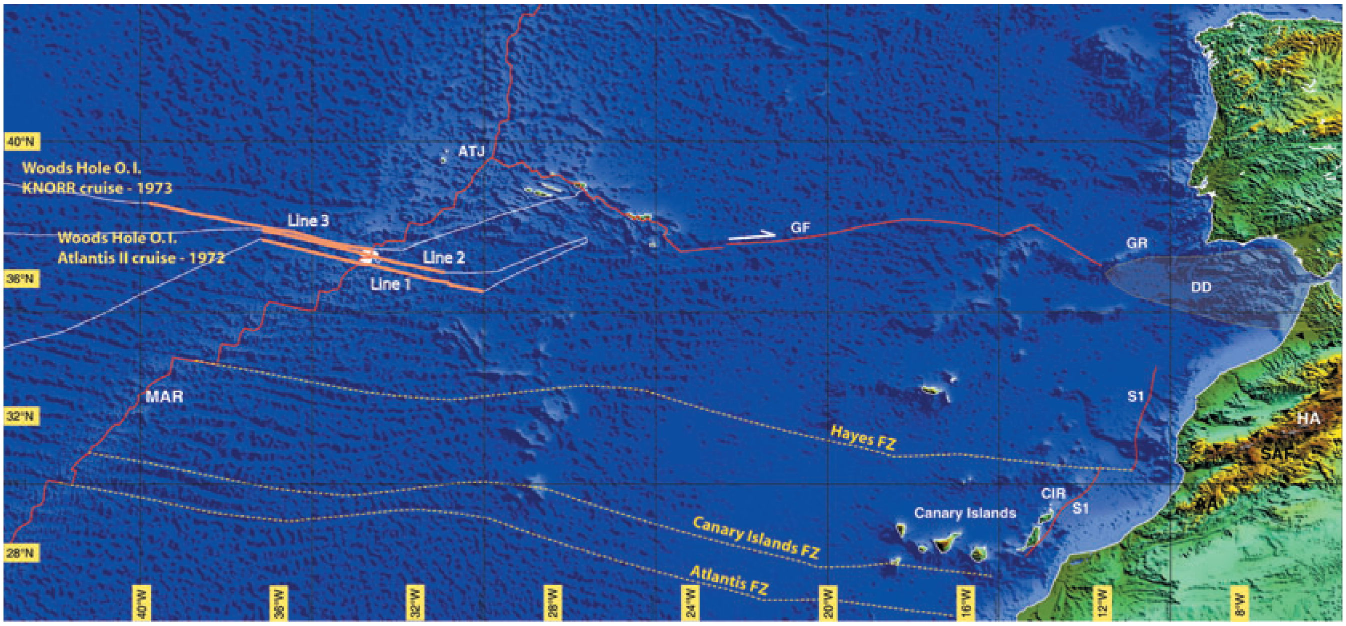

Figure 7. Tracks of available magnetic cruises north of the Hayes FZ (white lines). The base map shows bathymetry and relief in the area of interest. Traces of used magnetic anomaly data are shown in orange. MAR, Mid-Atlantic Ridge; ATJ, Azores Triple Junction; GF, Gloria Fault; GR, Gorringe Ridge; DD, area of diffuse deformation; CIR, Canary Islands Ridge; AA, Anti-Atlas; HA, High Atlas; SAF, South Atlas Fault.

located at the eastern termination of the corresponding fracture zone, has been investigated since the 1960s. Their crustal structure was first studied by Bosshard & MacFarlane (1970) using seismic refraction and gravity data, whereas the sedimentary sequence and Mesozoic palaeogeographic setting have been described by Steiner *et al.* (1998), on the basis of the Fuerteventura stratigraphic record. Anguita & Hernan (1975) were the first authors to propose a tectonic origin for the islands chain. They assumed that the islands were aligned along the seaward prolongation of the dextral transpressive South Atlas Fault (SAF) during the Atlas orogeny. In their hypothesis, Cenozoic volcanism would be associated with tensional intervals between compressive pulses. The alignment of the Canary Islands with the seaward prolongation of the SAF was also used by Le Pichon *et al.* (1977) in their fit of the continents around the central Atlantic. It is interesting to note that these authors proposed a Kimmeridgian–Oxfordian reconstruction, which was based on the fit of the seaward edges of the JQZ south of the SAF and its North American counterpart (the so-called 40°N fault). An interpretation of the Canary Islands as uplifted crustal blocks in a compressional or transpressive tectonic context was also stressed by Robertson & Stillman (1979), Price (1980), Fernandez *et al.* (1997), Anguita & Hernan (2000) and Gutiérrez *et al.* (2006). In summary, the available geological evidence supports the hypothesis that the NE–SW oriented Canary Islands Ridge is a compressive structure linked to the Atlas orogeny. In this instance, the E–W directed western Canary Islands would be aligned along a dextral transpressive fault.

The existence of Cenozoic tectonic activity at the termination of the Atlantis and Canary Islands FZ is also evidenced by major erosional events at the Moroccan continental margin (Arthur *et al.* 1979; Hinz *et al.* 1982) and by the structural pattern of the Anti-Atlas region of Morocco. It is generally accepted that the 2500 m uplift of the impressive Anti-Atlas fold belt, of Palaeozoic age, occurred during the Neogene (Helg *et al.* 2004; Burkhard *et al.* 2006), whereas Malusà *et al.* (2007) have recently shown that this region was affected by NW–SE transtension during the early Jurassic and NW–SE shortening in a right-lateral transpressional context during the Neogene. Therefore, the Anti-Atlas could have accommodated part of the deformation associated with the eastward escape of Morocco.

More to the north, the most important thrust structures of western Morocco are clearly represented by the High Atlas belt, whose western end is located at the termination of the Hayes FZ (Fig. 7). Figs 1 and 7 also show that the S1 anomaly appears to be composed of two lineations, which are separated by the Hayes FZ; the northern segment of this anomaly is clearly displaced to the east. The southernmost structure of the Atlas belt is the SAF, a major transpressive fault running from Morocco to Tunisia, which also represented the southern boundary of the Atlas rift during the late Triassic–early Jurassic phase of extension. The structural features of the SAF have been described in detail by Frizon de Lamotte *et al.* (2000). We assume that this fault accommodated most of the shortening associated with the eastward escape of the Moroccan Plate, whereas a minor amount of compression was partitioned between the Canary Islands Ridge and the Anti-Atlas region. Therefore, the geological evidence shows that the Moroccan Plate behaved as a single rigid plate only at first approximation, whereas in the central Atlantic right-lateral strike-slip occurred mainly along the Hayes FZ and, at a lesser extent, along the Atlantis FZ. A realistic scenario could be the following. An initial phase of dextral strike-slip motion could have occurred along the Atlantis FZ. In this instance, the deformation would have been transferred from the Atlantis FZ to the Canary Islands FZ close to the eastern termination, at about 20°W, giving rise to the Canary Islands and Anti-Atlas uplifts. Then, some time later, when right-lateral motion along the Atlantis FZ ceased, the eastward escape of Morocco would have been accommodated by right-lateral strike-slip along the Hayes FZ, triggering the main Atlas orogenic phase.

To investigate the consequences of the Atlas orogeny on the pattern of SFS in the central Atlantic, we first applied the Shaw's algorithm (1987) to generate synthetic fracture zones starting from the kinematic model. In fact, as already mentioned, the hypothesis of an eastward escape of Morocco relative to northwest Africa requires the existence of a larger area of oceanic lithosphere between Morocco and North America with respect to the amount predicted by the North America–northwest Africa reconstruction poles. In particular, the length of synthetic fracture zones, generated using these Euler poles, should be less than the observed distance between the conjugate COBs. The application of the Shaw's algorithm to three fracture zones on the Moroccan Plate and the conjugate oceanic region in North America showed that model fracture zones, built using rotation poles of the northwest Africa–North America pair, did not reach the conjugate COBs (northern ECMA and S1) and revealed a deficit in length of 150–180 km. Apart from the geological evidence of an independent Moroccan Plate, represented by the formation of the Atlas Rift in the late Triassic-early Jurassic and the Atlas mountain belt in the Cenozoic, this represents a first geophysical evidence that northwest Africa must be separated in at least two blocks. However, a definitive word about this topic can only be given through the analysis of marine magnetic anomalies in the central Atlantic.

To do this, we analysed three magnetic profiles in the area north of the Hayes FZ (Fig. 7), having an orientation compatible with observed fracture zone trends (Lines 1–3, Fig. 7). The assumed declination and inclination of the geomagnetic field at the survey times (1972 and 1973) were, respectively, $D = 17.47°$W $I = 55.31°$N, $D = 18.10°$W $I = 56.43°$N and $D = 19.30°$W $I = 58.25°$N. The thickness of the magnetized layer was set to 0.5 km. The mean palaeolatitude and direction of the blocks at the magnetization time was set on the basis of the global model of Schettino & Scotese (2005). Finally, the bathymetry was considered to be coincident with the external surface of the magnetized sources. Fig. 8 shows the model and measured magnetic anomalies along the eastern side of Line 1. This profile reaches anomaly 8 on Moroccan Plate. We excluded the western trackline (on North American Plate) because the magnetic data set was incomplete. Fig. 9 shows the model and measured magnetic anomalies along the Line 2, which reaches anomalies 13 on North America and 7 on the Moroccan side. Finally, Fig. 10 shows the interpretation of Line 3, which is entirely placed on North American Plate and reaches anomaly 18. The excellent matches of the model and measured curves were obtained through the introduction of many stages, characterized by different spreading rates, for the time interval from chron C18n to the present. Conversely, standard models of the central Atlantic comprise only four stages from chron C21 to the present (e.g. Klitgord & Schouten 1986; Royer *et al.* 1992

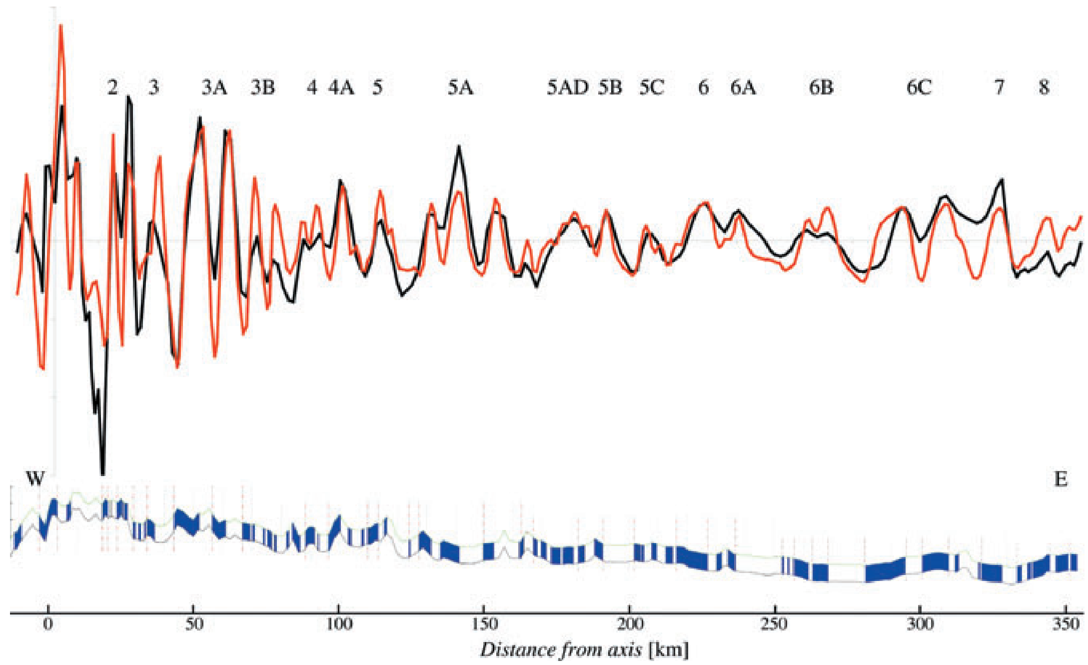

Figure 8. Magnetic anomaly profile along Line 1 of Fig. 7. Measured anomalies are shown in black; model anomalies are represented in red.

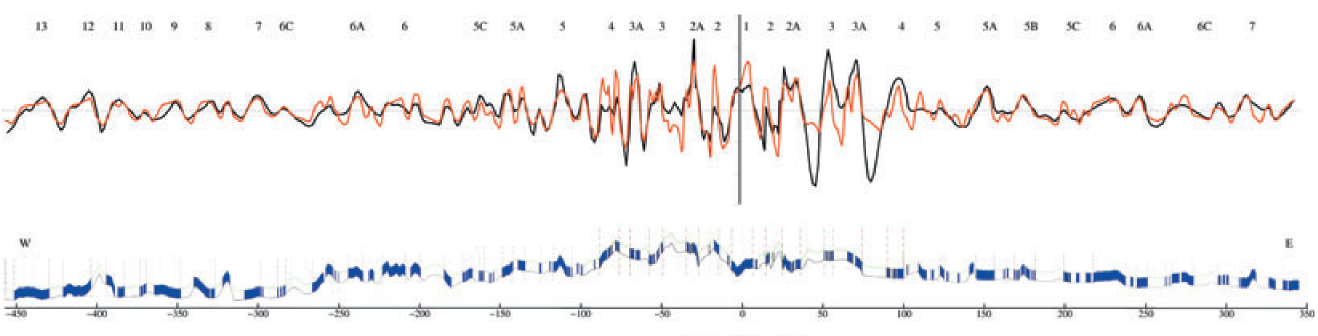

Figure 9. Magnetic anomaly profile along Line 2 of Fig. 7. Measured anomalies are shown in black; model anomalies are represented in red.

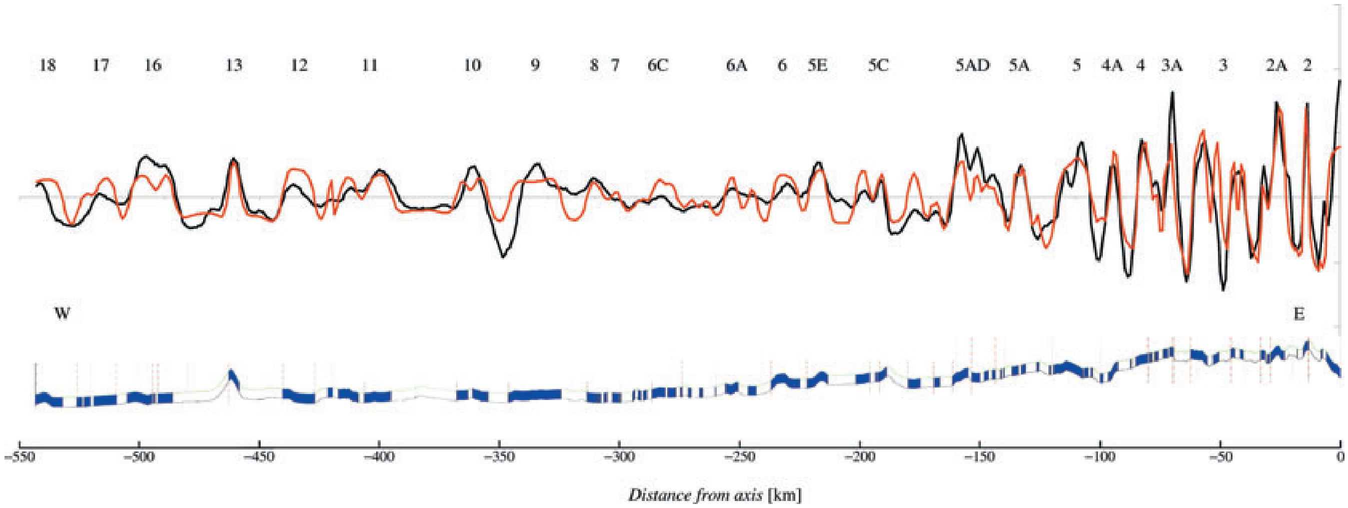

Figure 10. Magnetic anomaly profile along Line 3 of Fig. 7. Measured anomalies are shown in black; model anomalies are represented in red.

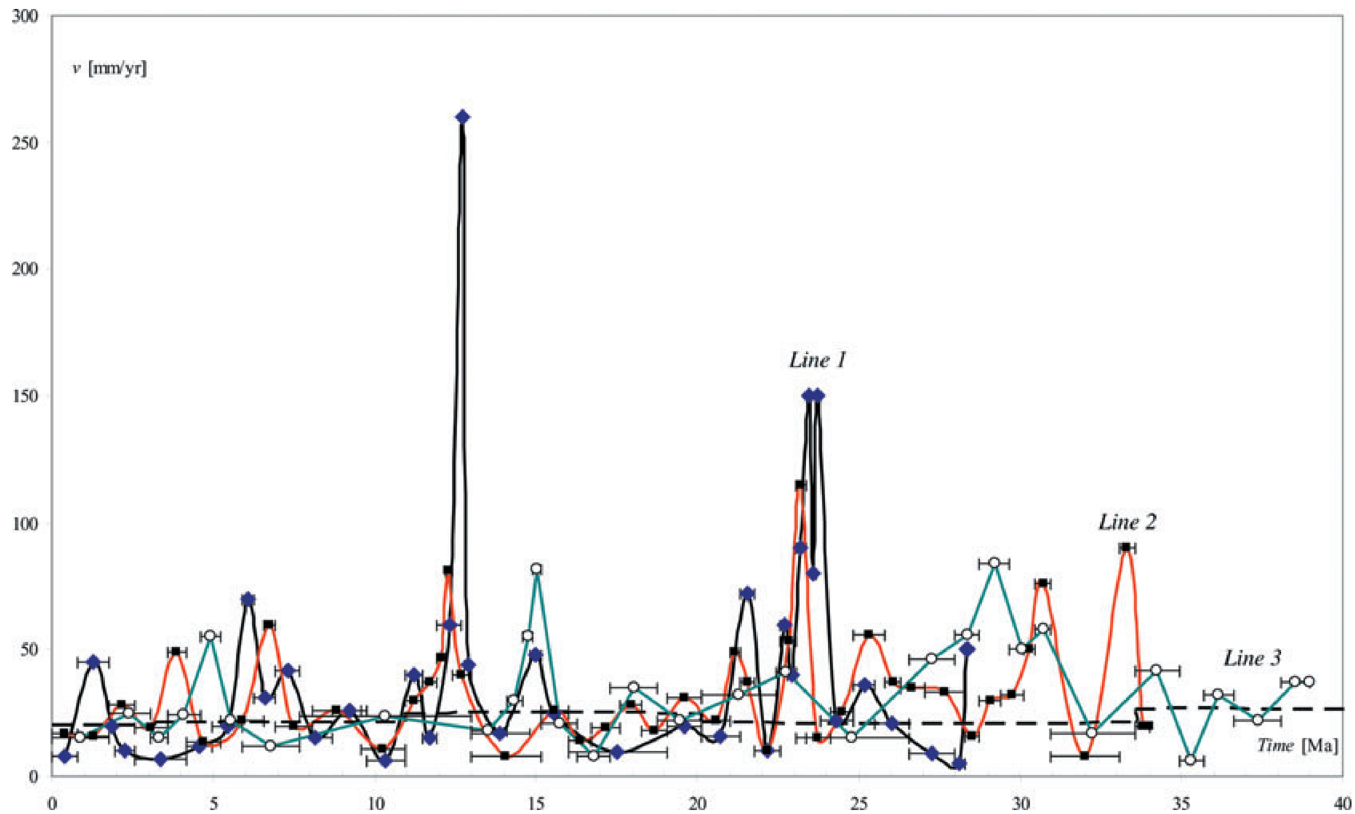

Figure 11. Full spreading velocities along Lines 1–3 of Fig. 7 (green, red and black solid lines) compared with the predicted velocity of northwest Africa relative to North America along the same profiles (black dashed line). Each point is plotted at the mean age of the corresponding stage. Error bars are stage lengths.

Müller *et al.* 1997). Fig. 11 shows a plot of the predicted full spreading rates, according to the interpretation of the three magnetic anomaly profiles. For times preceding chron 13 (33.1 Ma), the curve relative to Line 3 simply oscillates about the velocity predicted for northwest Africa. We interpret this result as evidence that Morocco and northwest Africa were a single plate from the Pliensbachian (185 Ma) to the early Oligocene (33.1 Ma). Then, the three plots show some prominent peaks and velocities that are, on average, greater than the predicted spreading rate of northwest Africa. This phase stops during chron C6n and is followed by a time interval with velocities that are close to the predicted spreading rate of northwest Africa, although three very short spikes of anomalous high spreading rate can be observed at ∼15 , 12 and ∼6 Ma. Hence, the main phase of the Atlas orogeny ceased at chron C6n (19 Ma), and we do not introduce severe errors in the model assuming that from this time the Moroccan Plate remained fixed to northwest Africa. We exclude that the higher spreading rates shown in Fig. 11 resulted from asymmetric spreading that favoured

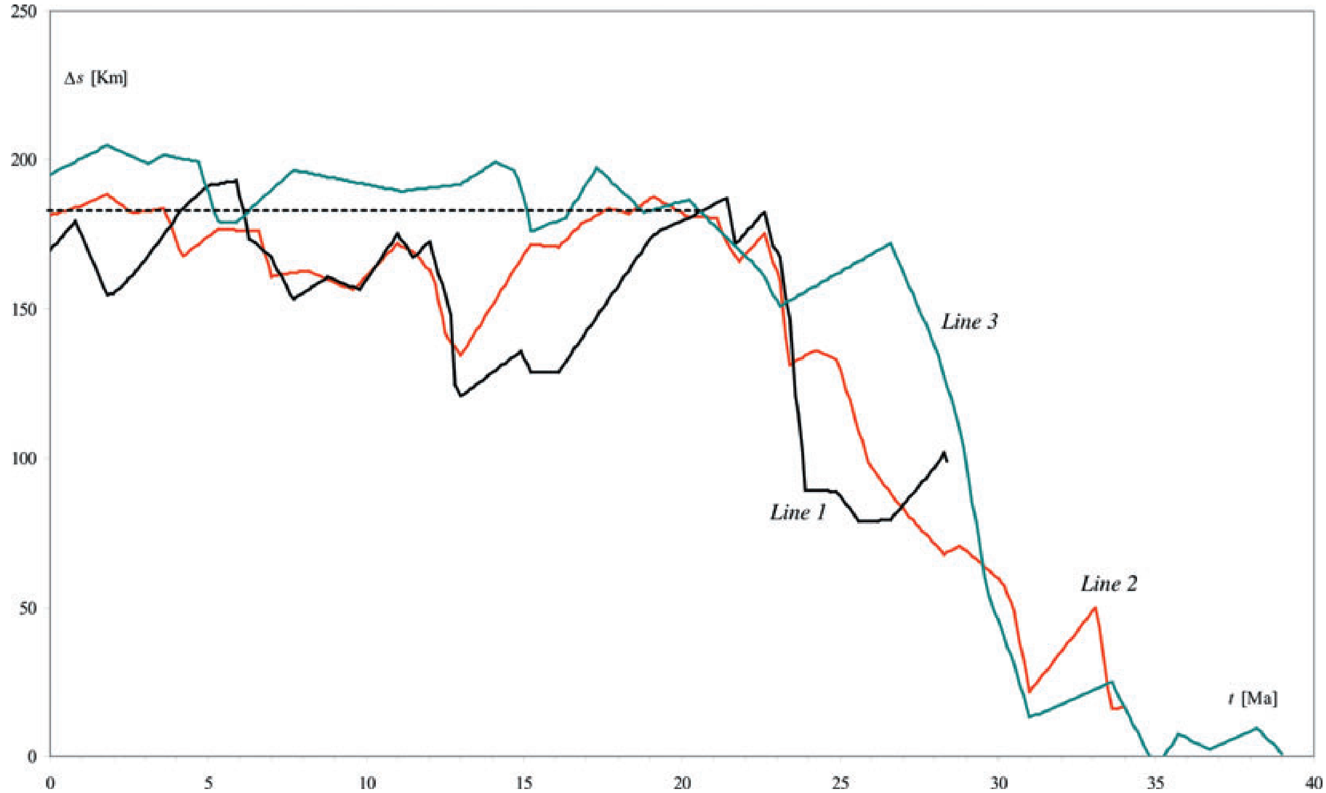

Figure 12. Amounts of right-lateral escape of Morocco relative to northwest Africa from chron C18n to the present, predicted from the analysis of Lines 1–3. To compare the three results, the plot $\Delta s_2(t)$ associated with line 2, which starts at $t = 34$ Ma, has been offset by the quantity $\Delta s_3(34) = 16.6$ km. Similarly, plot $\Delta s_1(t)$ associated with Line 1, which starts at $t = 28.4$ Ma, has been offset by the quantity $[\Delta s_3(28.4) + \Delta s_2(28.4) + \Delta s_3(34)]/2 = 96.2$ km. On an average, ∼180 km displacement of Morocco is required by the analysis of the three magnetic anomaly profiles.

accretion to the North American or the Moroccan plates before chron C6. In fact, we have already shown through the analysis of fracture zones that an anomalous higher amount of oceanic lithosphere is present between Morocco and North America. Therefore, the analysis of three magnetic profiles independently points to the same result as obtained before, in addition to furnishing a constraint on the timing of the eastward escape. Regarding the amount of eastward escape, we can calculate this quantity by the following formula:

$$\Delta s\,(t) = \int_{t_0}^{t} \left[v_{\mathrm{MOR,NAM}}\left(t'\right) - v_{\mathrm{NWA,NAM}}\left(t'\right)\right] \mathrm{d}t'. \quad (1)$$

Here $\Delta s(t)$ is the dextral offset of Morocco with respect to northwest Africa at time t, t_0 is the profile starting time (43.4 Ma) and $v_{\mathrm{NWA,NAM}}$ and $v_{\mathrm{MOR,NAM}}$ are, respectively, the velocities of northwest Africa and Morocco relative to North America. The curves $\Delta s = \Delta s(t)$ for the time interval from C18n (39 Ma, Bartonian) to the present are shown in Fig. 12. This figure illustrates even more clearly the timing of the Atlas orogeny. We estimate a total offset of 170–180 km during the time interval from the early Oligocene (C13n, 33.1 Ma) to the Burdigalian (C6n, 19 Ma), which is consistent with the previous analysis of fracture zones. Therefore, apart from the demonstration that Morocco behaved as an independent plate during the geological past, we obtained a precise dating and amount of convergence for the Atlas orogeny.

We now discuss another important evidence for the existence of an independent Moroccan Plate during the Cenozoic. As stated above (see Fig. 6), we expect a deformed pattern of SFS magnetic anomalies in the area between Morocco and the MAR. Let δr_{C13} be the original total dextral offset at the MAR, during chron C13, of the segment north of the Hayes FZ with respect to the spreading segment south of the Atlantis FZ. Then, the total dextral offset at chron C6 will be given by

$$\delta r_{C6} = \frac{1}{2}\int_{t_0}^{t_1} \left[v_{\mathrm{MOR,NAM}}\left(t'\right) - v_{\mathrm{NWA,NAM}}\left(t'\right)\right] \mathrm{d}t' + \delta r_{C13} = \frac{1}{2}\Delta s\,(t_1) + \delta r_{C13}, \quad (2)$$

where the integration is performed between $t_0 = 19.0$ Ma and $t_1 = 33.1$ Ma. This equation allows to estimate the quantity δr_{C13} if we assume that δr_{C6} coincides with the present ridge offset (that is, if we assume that the eastward escape of the Moroccan Plate ended by chron C6, as suggested by Fig. 10). As shown in Fig. 12, the present total offset at the MAR is $\delta r_{C6} = 188$ km. Therefore, using the 170 km estimate of eastward escape we obtain $\delta r_{C13} = 188 - 85 = 103$ km. A similar calculation allows to estimate the expected total offset ζ of isochrons older than chron C13 in the eastern central Atlantic. It results in

$$\zeta = \int_{t_0}^{t_1} \left[v_{\mathrm{MOR}}\left(t'\right) - v_{\mathrm{NWA}}\left(t'\right)\right] \mathrm{d}t' + \delta r_{C13} = \frac{1}{2}\Delta s\,(t_1) + \delta r_{C6}. \quad (3)$$

Therefore, we obtain $\zeta = 85 + \delta r_{C6} = 273$ km. This value is in very good agreement with the 271 km offset observed along the C31 isochron (Fig. 13). In summary, not only the total offset of the C31 isochron is much greater than the present MAR offset, but the relationship between ζ and δr_{C6} through eq. 3 is almost exactly that predicted by the magnetic profile modelling discussed above. We conclude that geophysical and geological evidence requires the

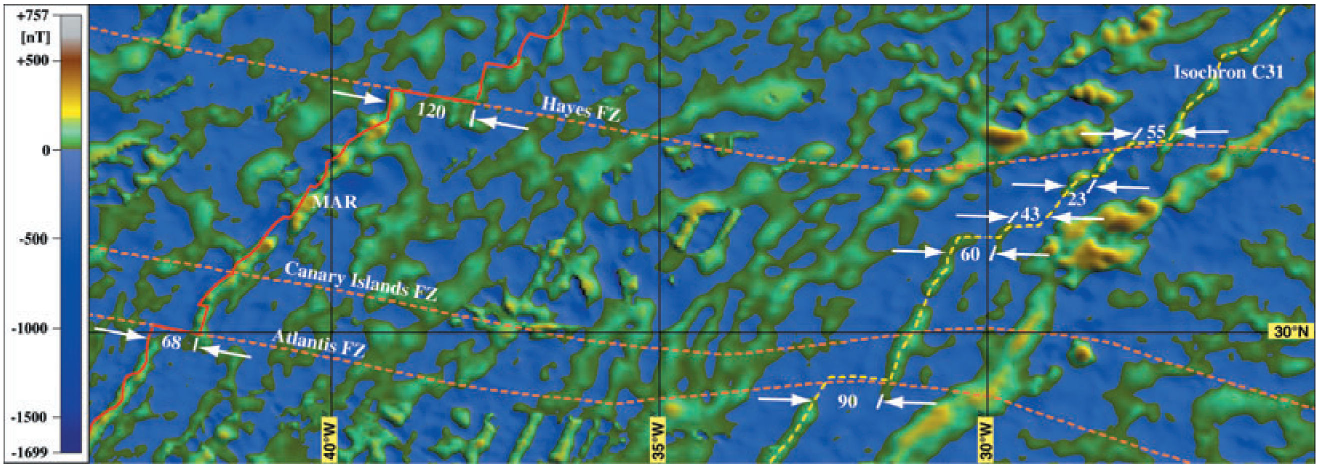

Figure 13. Offsets (in km) along the MAR and the chron 31 isochron from the Atlantis FZ to the Hayes FZ. The base map shows magnetic anomalies from the WDMAM 2007 (Korhonen *et al.* 2007).

existence of an independent Moroccan Plate, first during the late Triassic and early Jurassic, and later during the Oligocene–early Miocene time interval.

To determine finite rotations for the Moroccan Plate, we must now consider this as an approximately rigid block bounded to the south by the Atlantis FZ and the SAF. The reconstruction pole with respect to northwest Africa for times preceding chron 13 can be found if we know the stage pole of motion during the Atlas orogeny. The fracture zones trend indicate that Morocco and northwest Africa moved about the same stage pole, though with different spreading rates, during this time interval. Therefore, we can apply the following approximation:

$$\boldsymbol{s}_{\mathrm{MOR,NAM}}(33.1, 19.0) \approx \boldsymbol{s}_{\mathrm{NWA,NAM}}(33.1, 20.1) = (+68.2, 341.3), \quad (4)$$

where $\boldsymbol{s}$ indicates a stage pole location. Regarding the rotation angle, we observe that at the latitude of line 3, 468 km of new oceanic crust were created during this stage. This corresponds to a rotation $\Omega = 7.92^\circ$ about the stage pole. Therefore, the full stage pole of the Moroccan Plate with respect to North America is $\boldsymbol{S}_{\mathrm{MOR,NAM}}(19.0,33.1) = (+68.2^\circ\mathrm{N}, 341.3^\circ\mathrm{E}, -7.9^\circ)$. This stage pole allowed us to calculate the finite reconstruction poles at chron C13n relative to both North America and northwest Africa. Finally, the last reconstruction pole allowed us to determine the relative position of Morocco with respect to northwest Africa at the time of maximum extension in the Atlas Rift (185 Ma) because no motion occurred in this region between the Pliensbachian and the early Oligocene.

We have seen that the Pliensbachian is also the time of formation of the first oceanic crust between Morocco and North America. Therefore, the finite reconstruction poles obtained so far allow a complete description of the configuration of Pangaea at the time of extinction of the proto-Atlantic ridge and onset of drifting between Morocco and Nova Scotia. To complete our fit of Pangaea, we only need to estimate the pre-rift positions of Morocco and northwest Africa with respect to North America at the time of initiation of the rifting process (∼230 Ma, late Ladinian). This estimation was accomplished by crustal balancing of three available seismic profiles: profile SIS04 offshore Morocco (Contrucci *et al.* 2004), profile SMART 2 at the Nova Scotia margin (Wu *et al.* 2006) and profile EDGE Line 801 in the Baltimore Canyon Trough (Talwani & Abreu 2000).

To perform a correct balancing of these crustal profiles, we projected the data onto small circle arcs of the estimated stage poles of motion during the rift stage. These Euler poles were assumed to coincide with the location of the first stage poles during the drifting stage because the corresponding flow lines are compatible with the azimuths of rift structures in eastern North America and Morocco. In particular, the location of the stage pole of North America with respect to Morocco, between the late Ladinian (230 Ma) and the Pliensbachian (185 Ma), was approximated by the corresponding stage pole of North America relative to northwest Africa. From the analysis of profile SIS04, we obtained a stretching factor $\beta = 1.75$ and a corresponding angle of rotation $\Omega_1 = 0.75^\circ$. Similarly, balancing of SMART 2 furnished $\beta = 1.89$ for the North American margin and a rotation angle $\Omega_2 = 1.20^\circ$. Therefore, the estimated total angle of rotation from the pre-drift configuration to the pre-rift fit of Morocco against North America was $\Omega = \Omega_1 + \Omega_2 = 0.8^\circ + 1.2^\circ = 2.0^\circ$. This implies a mean stretching rate of 4.8 mm yr^{-1} between the late Ladinian (230 Ma) and the Pliensbachian (185 Ma), considerably lower than the initial mean spreading rate in this area (∼14.1 mm yr^{-1}).

We consider, now, the conjugate margins of North America and northwest Africa. In this instance, we only had reliable crustal profiles from the North American margin. Therefore, the total angle of rotation was estimated by doubling the angle recovered by these data and adding the angle necessary to remove the ECMA and BSMA anomalies. From the analysis of profile EDGE Line 801, we obtained $\beta = 1.18$ and a corresponding angle $\Omega_1 = 0.05^\circ$. Therefore, the total angle was estimated to be $\Omega = 2\Omega_1 = 0.10^\circ$. Finally, this quantity was incremented by 2.2° to take into account the space occupied by the ECMA and BSMA, giving a total angle $\Omega = 2.3^\circ$. Therefore, the mean stretching rate was 8.2 mm yr^{-1} between the late Ladinian (230 Ma) and the latest Rhaetian (200 Ma), which is not much different from the estimated slow spreading rate in the proto-Atlantic (∼10.9 mm yr^{-1}). The finite reconstruction poles resulting from the previous analyses are listed in Table 1. The next section illustrates the geological consequences of the kinematic model discussed above.

4 RIFTING EVENTS DURING THE LATE TRIASSIC

A fit of central Pangaea at 230 Ma (late Ladinian), which is based on the kinematic model discussed in the previous sections, is

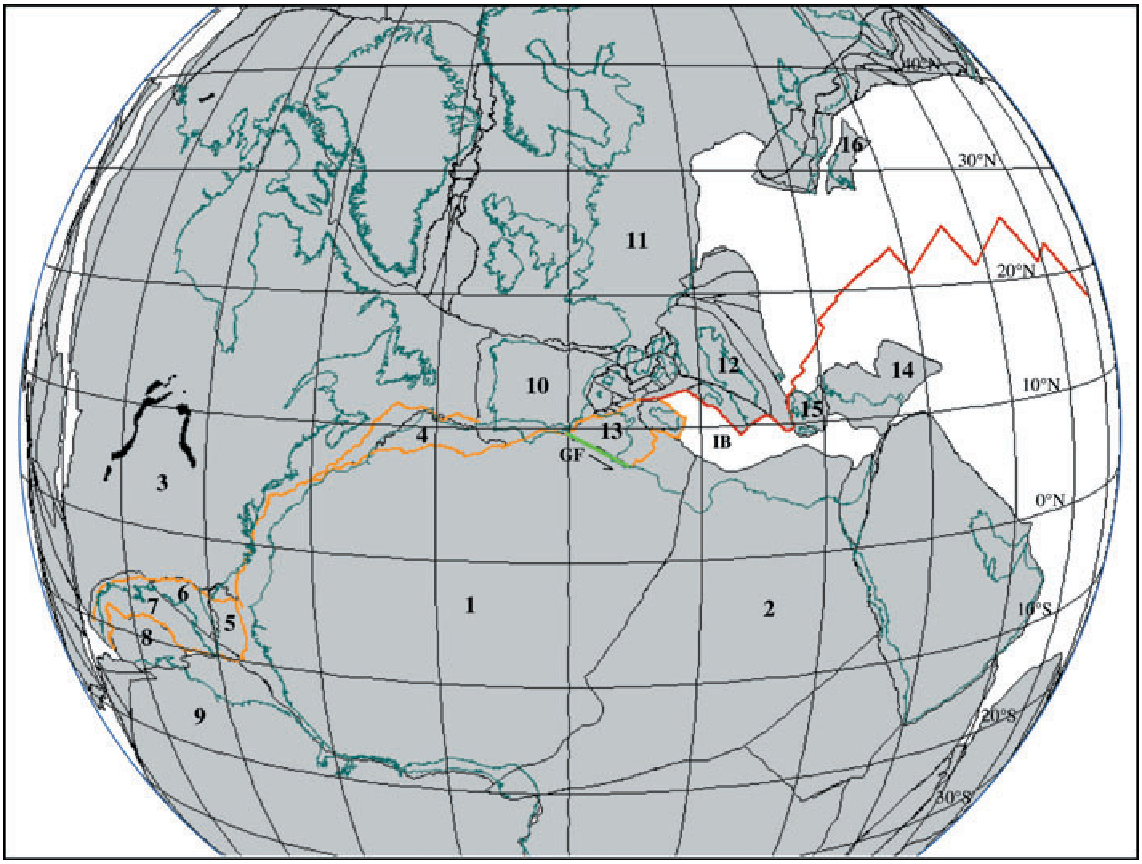

Figure 14. Fit of central Pangaea at 230 Ma (late Ladinian). Large igneous provinces (LIPs) are shown in black, whereas the areas affected by the first rifting events are bounded by the orange lines. Transform faults are shown in green. Dark red line is an incipient spreading centre. GF, Gafsa Fault; IB, Ionian Basin; 1, northwest Africa; 2, northeast Africa; 3, North America; 4, Morocco; 5, Blake Plateau; 6, N. Florida; 7, S. Florida; 8, Yucatan; 9, Brazilian craton; 10, Iberia; 11, Eurasia; 12, Adria; 13, Tunisia; 14, Menderes–Taurides Platform; 15, southern Greece; 16, Sakarya.

illustrated in Fig. 14. This reconstruction shows that a WSW–ENE directed system of rifts established during the late Ladinian, which cut central Pangaea from the Caribbean to the Tethys, where it was possibly converted into an oceanic spreading centre. The subsequent tectonic evolution of the central Atlantic and Atlas regions during the first phases of breakup and onset of SFS is constrained by the M25 and M21 anomalies. We divide the ∼82 Myr from the late Ladinian to chron M21 (147.7 Tithonian) in three time intervals, or phases, which group one or more consecutive tectonic stages that are characterized by the same configuration of the plate boundaries. Therefore, whereas the limits of usual stages are associated with significant changes in the location of the poles of relative motion, phase transitions also imply the formation or the extinction of plate boundaries.

Starting from the fit illustrated in Fig. 14, the first phase includes the entire late Triassic and terminates at the time of formation of the first oceanic crust in the proto-Atlantic, close to the Triassic–Jurassic boundary. Fig. 15 shows a plate tectonic reconstruction at end of this time interval (200 Ma). The phase started with the formation of the first rift structures in eastern North America (Klitgord *et al.* 1988; Benson 1992; Schlische 1993, 2003; Withjack *et al.* 1998; Schlische *et al.* 2002), northwest Africa (Roussel & Liger 1983; Davison 2005), western Morocco (Piqué & Laville 1995; Le Roy *et al.* 1998; Piqué *et al.* 1998b; Le Roy & Piqué 2001), and the Atlas (Laville & Piqué 1991; Laville *et al.* 1995, 2004; Piqué *et al.* 1998a, 2002; Ait Brahim *et al.* 2002; Ellouz *et al.* 2003).

The western part of the rift system shown in Fig. 15 is represented by the extensional structures of southeastern North America (South Georgia and Florida) and the Caribbean zone. The basic mechanism of deformation of these regions was represented by a combination of (1) left-lateral strike-slip across the Bahamas FZ (Klitgord *et al.* 1984), which cuts peninsular Florida in NW–SE direction (Fig. 1), (2) extension and emplacement of mafic magmas along the axis of the Brunswick magnetic anomaly (Fig. 1) (McBride & Nelson 1988; McBride *et al.* 1989; Heatherington & Mueller 1991; Lizarralde *et al.* 1994) and (3) extension between the Blake Plateau and northwest Africa. This partitioning of deformation caused a delayed start of spreading in the southern segment of the central Atlantic (∼170 Ma). Phase 1 terminated at the Triassic–Jurassic boundary, with the onset of SFS in the proto-Atlantic and cessation of spreading in the Ionian Basin and the western Tethys. From this time onward, Adria was part of the African Plate.

5 RIFTING AND SPREADING DURING THE JURASSIC

The second phase started at 200 Ma and terminated 15 Myr later (Pliensbachian, 185 Ma), when extension ceased in the Atlas region and the proto-Atlantic. Fig. 16 illustrates the Plate configuration at the end of the phase. The formation of the first oceanic crust between Morocco and Nova Scotia, and consequently of the northern ECMA

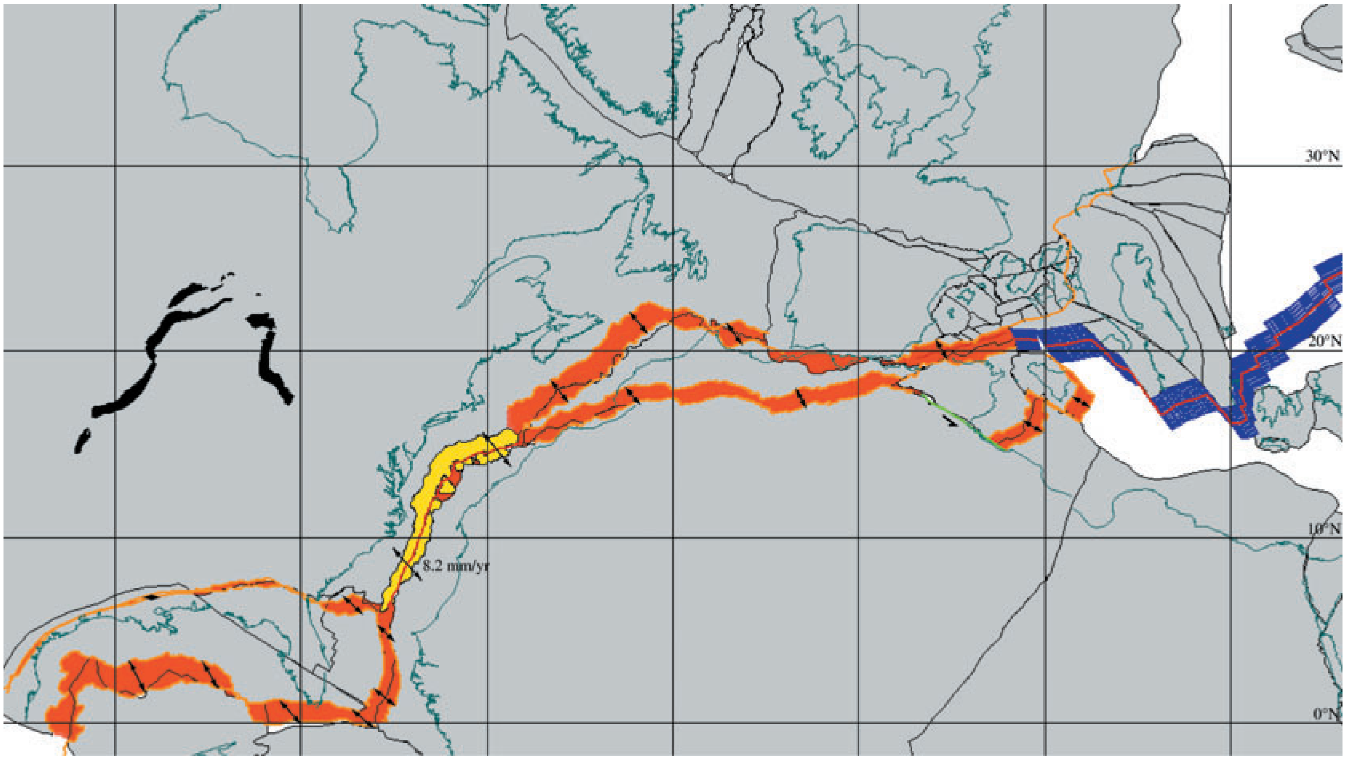

Figure 15. Plate reconstruction at 200 Ma (Triassic–Jurassic boundary). Initial oceanic crust associated to the ECMA and BSMA is shown in yellow. Present-day coastlines are shown for reference. Orange lines bound rift zones. Red lines are spreading centres. Blue lines are synthetic isochrons representing the oceanic crust formed during the phase 1. Transform faults are shown in green. Arrows represent direction and amount of relative motion. Areas affected by active rifting and thinning are shown in light brown.

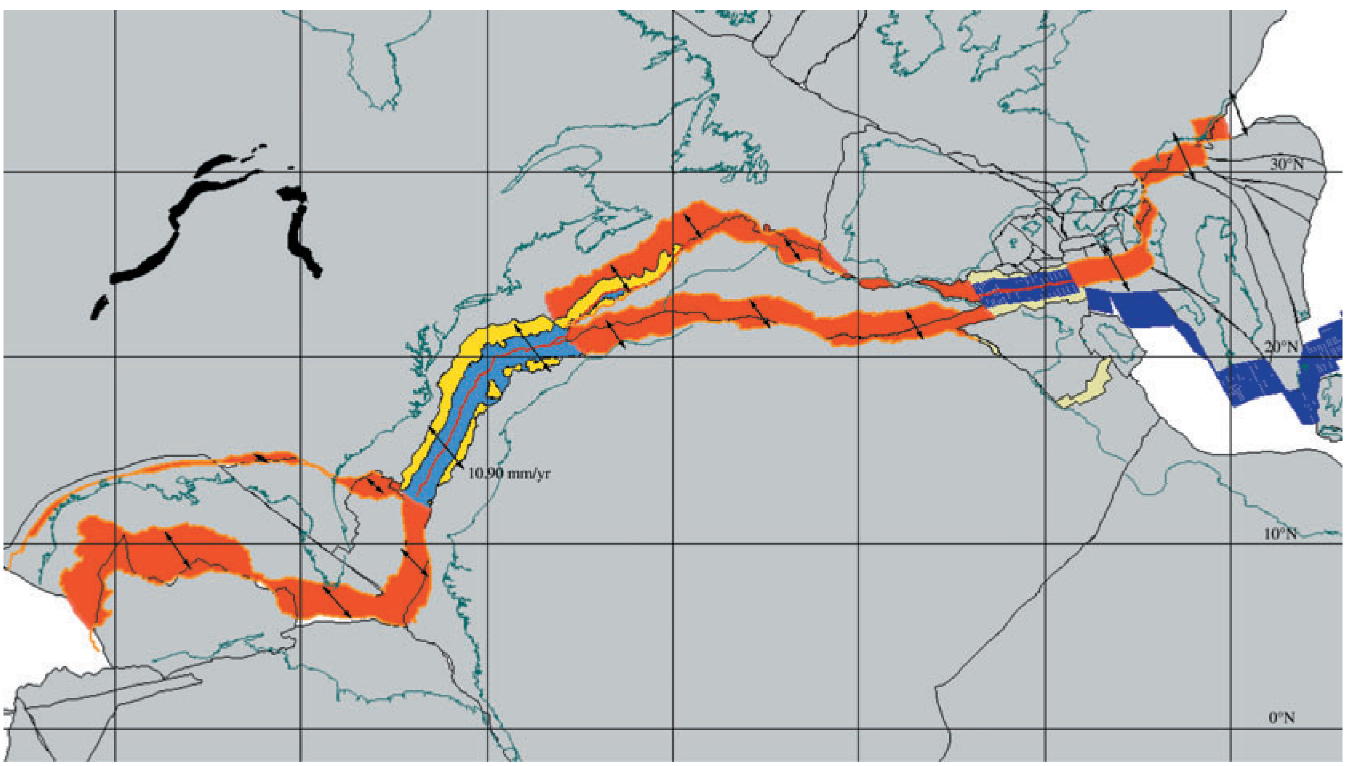

Figure 16. Plate reconstruction at 185 Ma (Pliensbachian). Initial oceanic crust associated with the S1, BSMA and the two segments of the ECMA is shown in yellow. Main areas with thinned crust are shown in dark khaki.

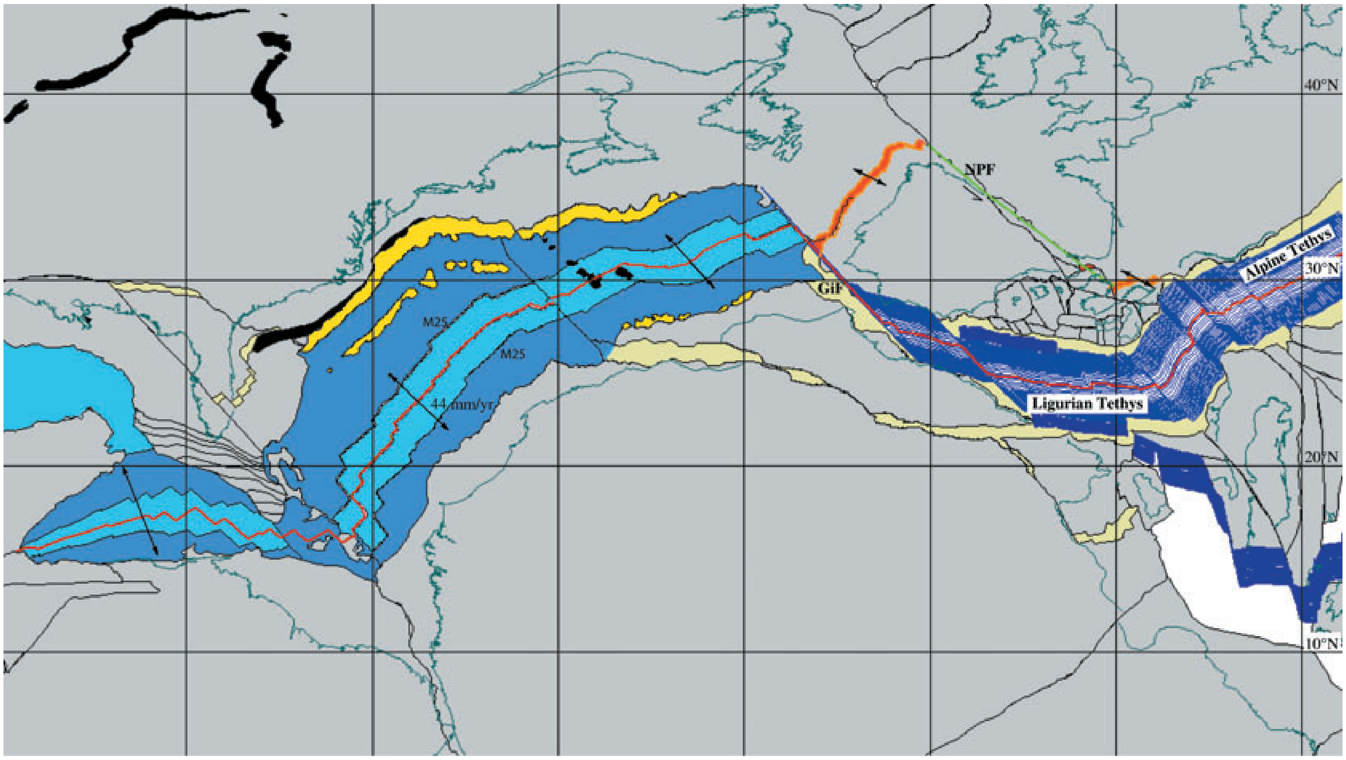

Figure 17. Plate reconstruction at 147.7 Ma (chron M21, Tithonian). GiF, Gibraltar Fault; NPF, North Pyrenean Fault; Blue lines are synthetic isochrons relative to the opening of the Ligurian and Alpine Tethys oceans.

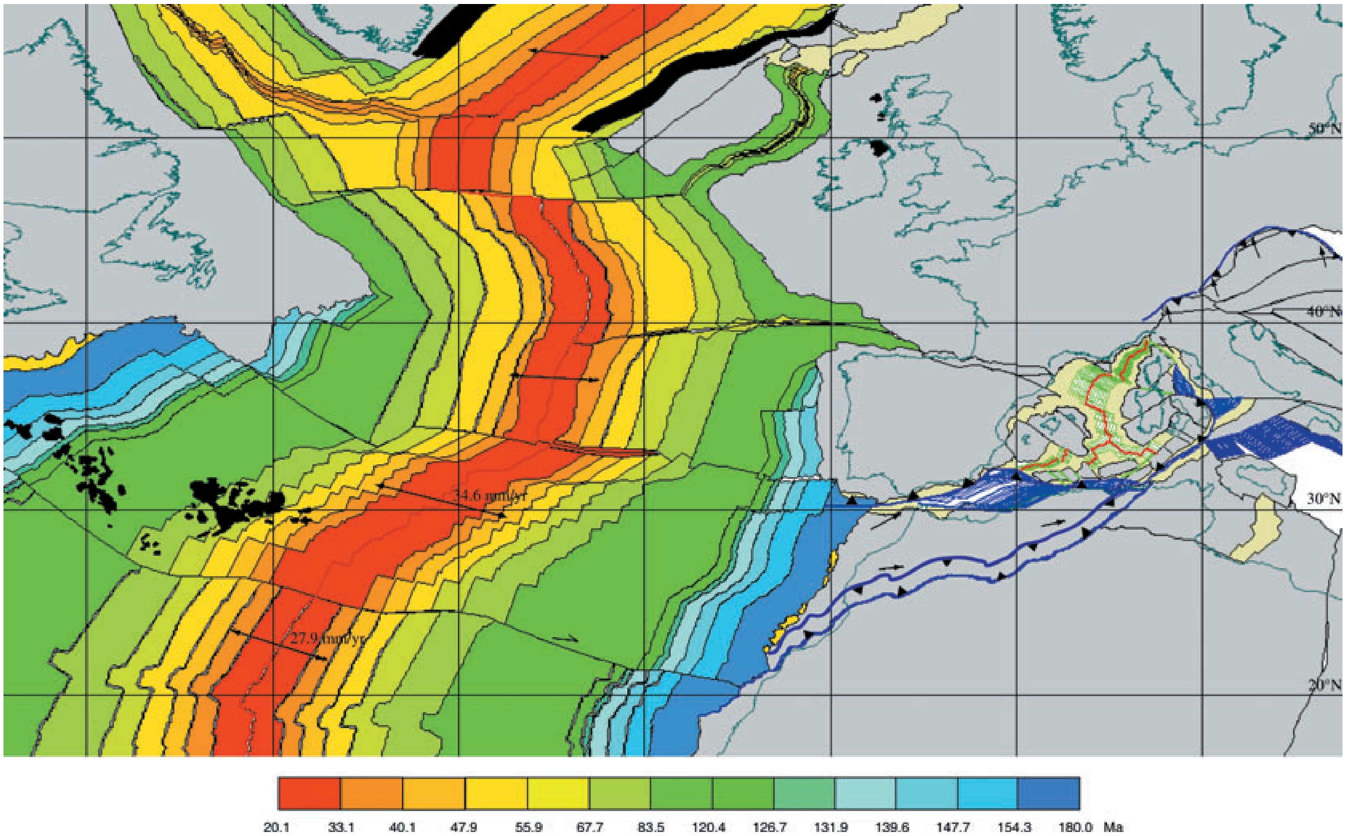

Figure 18. Plate reconstruction at 20.1 Ma (Chron C6n, early Burdigalian). The middle Atlas deformation zone is not shown. Green lines are synthetic isochrons bounding the distribution of oceanic crust in the western Mediterranean (after Schettino & Turco 2006).

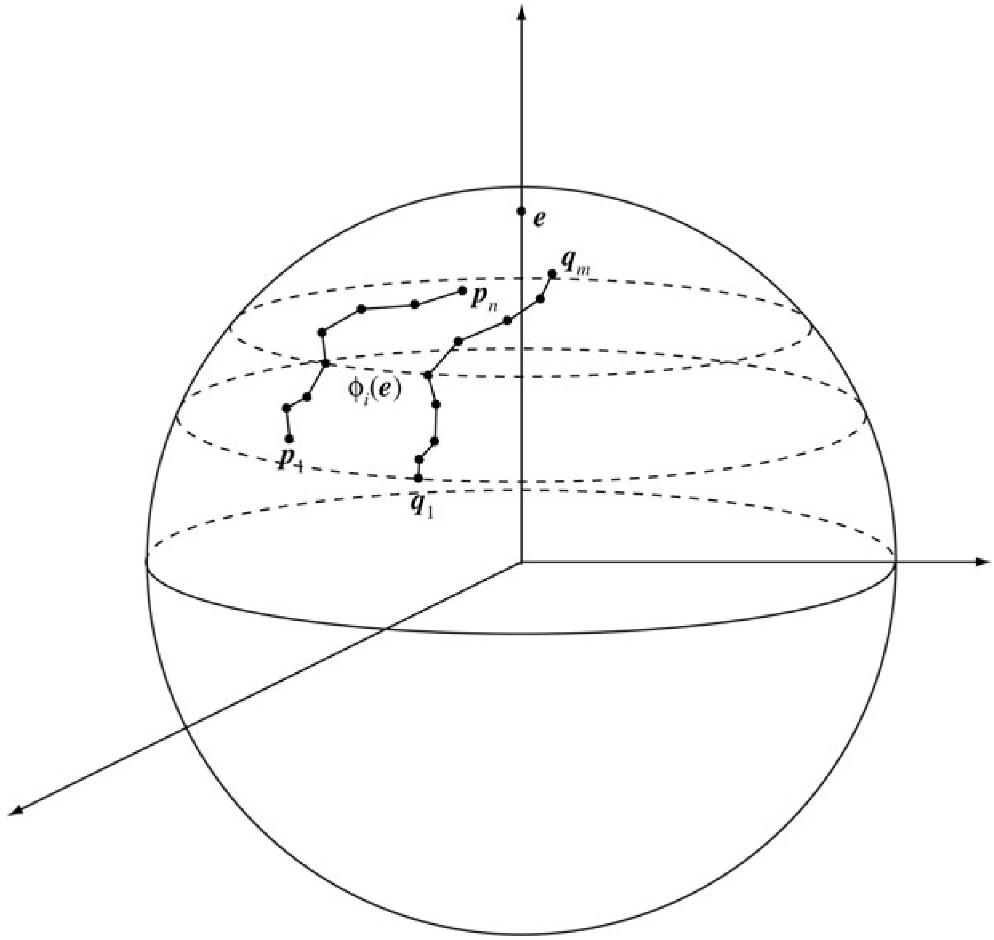

Figure 19. Geometrical fitting of conjugate lines. In a reference frame where the test Euler pole ***e*** has been carried to the North Pole, the fitting procedure is reduced to finding the best rotation about the North Pole, by angle ϕ_0, that minimizes the squared sum of longitude misfits.

and S1 magnetic anomalies, is considered coeval with the ridge jump that caused the end of spreading in the proto-Atlantic.

During this time interval, rifting in the Atlas region continued, but extensional structures became progressively more important at the northern boundary of Morocco, separating this plate from Iberia and the Newfoundland region of North America. It is also likely that the southern Ligurian Basin, between Iberia and Tunisia, reached the spreading stage during this phase.

Starting from ∼185 Ma and until the Oligocene–early Miocene event, when the Atlas orogeny closed the Triassic and early Jurassic rift valleys, the Moroccan Plate remained fixed to northwest Africa. Therefore, a phase transition occurred during the Pliensbachian (185 Ma), and the beginning of the third phase is characterized by the transfer of the Atlantic kinematics to the Tethyan realm through a new plate boundary located between Iberia and Morocco—the so called 'Gibraltar Fault'. The upper limit of this time interval is placed at 147.7 Ma (chron M21, Tithonian), which is the time of a further northward jump of the transfer zone, from the Gibraltar area to the North Pyrenean Fault Zone (Schettino & Scotese 2002). A plate reconstruction at the upper boundary of phase 3 is shown in Fig. 17.

6 ATLAS OROGENY

In a previous section, we have shown that the building of the Atlas mountain belt encompasses the Oligocene–early Miocene time interval. As already discussed, the Moroccan Plate initiated to escape eastwards with respect to northwest Africa during chron C13n (33.1 Ma). In the Atlantic Ocean, at least one Ridge-Ridge-Fault triple junction formed at the MAR to allow higher spreading rates of the ridge segments facing the Moroccan Plate (i.e. North of the Atlantis FZ). This motion caused the formation of compressional structures at the eastern termination of the Atlantis, Canary Islands and Hayes FZ, along the western continental margin of Morocco and in the Atlas and Anti-Atlas regions. In particular, the Triassic and Jurassic rift structures of the Atlas region were reactivated as inverse faults, determining the Atlas uplift (Beauchamp *et al.* 1996; Frizon de Lamotte *et al.* 2000; Piqué *et al.* 2002). Fig. 18 shows a plate reconstruction at 20.1 Ma (anomaly 6, early Burdigalian). For simplicity, Morocco has been considered in this paper as a single rigid plate, although a more detailed reconstruction would have considered motions along the middle Atlas during both the Triassic–Jurassic rifting phase and the Tertiary orogeny, as well as slip partitioning during the eastward escape. The phase of Atlas orogeny terminated at 19 Ma (chron C6n, early Burdigalian), when both the Atlas uplift and extension in the western Mediterranean ceased and new plate boundaries formed in North Africa and in the central Mediterranean.

7 CONCLUSION

A new plate motions model has been discussed for the central Atlantic and Atlas regions from the late Ladinian to the Tithonian.

A consistent part of this paper has been devoted to describe a new accurate pre-breakup fit of central Pangaea. Therefore, in contrast to published fits based on extrapolated SFS data alone (e.g. Klitgord & Schouten 1986), our kinematic model between the initial rifting and the first identified magnetic anomaly (M25) is not a simple backward prosecution of the M21–M25 stage. We have seen that this kinematics predicts a specific style of deformation in North Africa during the late Triassic–early Jurassic time interval. However, it also predicts a specific deformation of the SFS anomalies in the central Atlantic during the Atlas orogeny. This pattern has been fully confirmed by the analysis of magnetic anomaly profiles and fracture zone geometry in the zone north of the Atlantis FZ. The analysis has allowed both a precise dating of the Atlas uplift (33–19 Ma) and a confirmation of the goodness of fit for the central Pangaea region. In particular, we have shown that the amount of oceanic crust between Morocco and North America is greater than that predicted by the Klitgord & Schouten's model because synthetic fracture zones generated using their rotation parameters do not reach the conjugate COBs. Furthermore, the analysis of three magnetic anomaly profiles on the North American and Moroccan plates demonstrated the existence of a higher spreading rate with respect to the one predicted by the Klitgord & Schouten's model. In this instance, a high spreading rate associated with an asymmetric spreading which favoured accretion to the North American or the Moroccan plates before chron C6 is excluded by presence of an excess oceanic crust between the conjugate COBs. In fact, asymmetric spreading simply distributes the new crust differently between the conjugate plates, without affecting the total amount of accreted lithosphere. Furthermore, the excess oceanic crust predicted by the magnetic anomaly modelling is compatible with the observed amount of lithosphere between Morocco and North America, which definitely excludes asymmetric spreading. The proposed model has important implications regarding the plate kinematics of the Tethyan region during the late Triassic–early Jurassic time interval, which will be published elsewhere. A computer animation is available as Supporting Information in the online version of this paper, which illustrates the tectonic evolution of the central Atlantic and Atlas regions since the late Triassic.

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REFERENCES

Ait Brahim, L. *et al*., 2002. Paleostress evolution in the Moroccan African margin from Triassic to Present, *Tectonophysics,* **357,** 187–205.

Anguita, F. & Hernan, F., 1975. A propagating fracture model versus a hot spot origin for the Canary Islands, *Earth planet Sci. Lett.,* **27,** 11–19.

Anguita, F. & Hernan, F., 2000. The Canary Islands origin: a unifying model, *J. Volcanol. Geotherm. Res.,* **103,** 1–26.

Arthur, M.A., von Rad, U., Cornford, C., McCoy, F. & Sarnthein, M., 1979. Evolution and sedimentary history of the Cape Bojador continental margin, northwestern Africa, in *Initial Reports of the DSDP,*Vol. 47, pp. 773–816, eds von Rad, U. *et al*., Texas A & M University (Ocean Drilling Program),College Station, TX, USA.

Beauchamp, W., Barazangi, M., Demnati, A. & El Alji, M., 1996. Intracontinental rifting and inversion: Missour Basin and Atlas Mountains, Morocco, *Am. Assoc. Petrol. Geol. Bull.,* **80**(9), 1459–1482.

Benson, R.N., 1992. Map of exposed and buried early Mesozoic rift basins/synrift rocks of the U.S. middle Atlantic continental margins, *Miscellaneous Map Series*Vol. 5, Delaware Geological Survey.

Bosshard, E. & MacFarlane, D.J., 1970. Crustal structure of the western Canary Islands from seismic refraction and gravity data, *J. geophys. Res.,* **75**(26), 4901–4918.

Bullard, E.C., Everett, J.E. & Smith, A.G., 1965. The fit of the continents around the Atlantic: a symposium on continental drift. *Phil. Trans. R. Soc. Lond. A,* **258**(1088), 41–51.

Burkhard, M., Caritg, S., Helg, U., Robert-Charrue, C. & Soulaimani, A., 2006. Tectonics of the Anti-Atlas of Morocco, *Comptes Rendus Geosciences,* **338**(1–2), 11–24.

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

Contrucci, I., Klingelhöfer, F., Perrot, J., Bartolome, R., Gutscher, M.A., Sahabi, M., Malod, J. & Rehault, J.-P., 2004. The crustal structure of the NW-Moroccan continental margin from wide-angle and reflection seismic data, *Geophys. J. Int.,* **159,** 117–128.

Davison, I., 2005. Central Atlantic margin basins of North West Africa: geology and hydrocarbon potential (Morocco to Guinea), *J. Afr. Earth Sci.,* **43,** 254–274.

Ellouz, N., Patriat, M., Gaulier, J.-M., Bouatmani, R. & Sabounji, S., 2003. From rifting to Alpine inversion: Mesozoic and Cenozoic subsidence history of some Moroccan basins, *Sediment. Geol.,* **156,** 185–212.

Fernandez, C., Casillas, R., Ahijado, A., Perello, V. & Hernandez-Pacheco, A., 1997. Shear zones as a result of intraplate tectonics in oceanic crust: the example of the Basal Complex of Fuerteventura (Canary Islands), *J. Struct. Geol.,* **19**(1), 41–57.

Fowell, S.J., Cornet, B. & Olsen, P.E., 1994. Geologically rapid Late Triassic extinctions: Palynological evidence from the Newark Supergroup, *Geol. Soc. Am. Spec. Paper,* **288,** 197–206.

Franchetau, J., 1973. Plate tectonics model of the opening of the Atlantic Ocean south of the Azores. In *Implications of Continental Drift to the Earth Sciences,* pp. 197–202, eds Tarling, D.H. & Runcorn, S.K., Academic Press, New York.

Frizon de Lamotte, D., Saint Bezar, B., Bracène, R. & Mercier, E., 2000. The two main steps of the Atlas building and geodynamics of the western Mediterranean, *Tectonics,* **19**(4), 740–761.

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

Gutiérrez, M., Casillas, R., Fernandez, C., Balogh, K., Ahijado, A., Castillo, C., Colmenero, J.R. & Garcia-Navarro, E., 2006. The submarine volcanic succession of the basal complex of Fuerteventura, Canary Islands: a model of submarine growth and emergence of tectonic volcanic islands, *Geol. Soc. Am. Bull.,* **118**(7–8), 785–804.

Heatherington, A.L. & Mueller, P.A., 1991. Geochemical evidence for Triassic rifting in southwestern Florida, *Tectonophysics,* **188,** 291–302.

Heirtzler, J.R., Dickson, G.O., Herron, E.M., Pitman, W.C., III & Le Pichon, X., 1968. Marine magnetic anomalies, geomagnetic field reversals, and motions of the ocean floor and continents, *J. geophys. Res.,* **73**(6), 2119–2136.

Helg, U., Burkhard, M., Caritg, S. & Robert-Charrue, C., 2004. Folding and inversion tectonics in the Anti-Atlas of Morocco, *Tectonics,* **23,** TC4006, doi:10.1029/2003TC001576.

Hinz, K., Dostmann, H. & Fritsch, J., 1982. The continental margin of Morocco; seismic sequences, structural elements and geological development, in *Geology of the Northwest African Continental Margin,* pp. 34–60, eds von Rad, U., Hinz, K., Sarnthein, M. & Seibold, E., Springer-Verlag, Berlin, Germany.

Holbrook, W.S., Reiter, E.C., Purdy, G.M., Sawyer, D., Stoffa, P.L., Austin, J.A., Jr., Oh, J. & Makris, J., 1994a. Deep structure of the U.S. Atlantic continental margin, offshore South Carolina, from coincident ocean bottom and multichannel seismic data, *J. geophys. Res.,* **99**(B5), 9155–9178.

Holbrook, W.S., Purdy, G.M., Sheridan, R.E., Glover, L. III, Talwani, M., Ewing, J. & Hutchinson, D., 1994b. Seismic structure of the U.S. Mid-Atlantic continental margin, *J. geophys. Res.,* **99**(B9), 17 871–17 891.

Keen, C.E. & Potter, D.P., 1995a. The transition from a volcanic to a nonvolcanic rifted margin off eastern Canada, *Tectonics,* **14**(2), 359–371.

Keen, C.E. & Potter, D.P., 1995b. Formation and evolution of the Nova Scotian rifted margin: evidence from deep seismic reflection data, *Tectonics,* **14**(4), 918–932.

Klitgord, K.D. & Schouten, H., 1986. Plate kinematics of the central Atlantic, in *The Geology of North America: The Western North Atlantic Region,* Vol. M, pp. 351–378, eds Vogt, P.R. & Tucholke, B.E., Geological Society of America, Boulder, CO.

Klitgord, K.D., Popenoe, P. & Schouten, H., 1984. Florida: a Jurassic transform plate boundary, *J. geophys. Res.,* **89**(B9), 7753–7772.

Klitgord, K.D., Hutchinson, D.R. & Schouten, H., 1988. U.S. Atlantic continental margin; structural and tectonic framework, in *The Atlantic Continental Margin, The Geology of North America,* Vol. I-2, pp. 19–55, eds Sheridan, R.E. & Grow, J.A., Geological Society of America.

Korhonen, J.V. *et al*., 2007. *Magnetic Anomaly Map of the World (and associated DVD), Scale: 1:50,000,000,* 1st edn, Commission for the Geological Map of the World, Paris, France.

Laville, E. & Piqué, A., 1991. La Distension crustale atlantique et atlasique au Maroc au debut du Mesozoique; le rejeu des structures hercyniennes, *Bull. Soc. Géol. de France,* **162**(6), 1161–1171.

Laville, E., Charroud, A., Fedan, B., Charroud, M. & Piqué, A., 1995, Inversion négative et rifting atlasique: l'exemple du bassin triasique de Kerrouchène (Maroc), *Bull. Soc. Géol. de France,* **166**(4), 365–374.

Laville, E., Piqué, A., Amrhar, M. & Charroud, M., 2004. A restatement of the Mesozoic Atlasic Rifting (Morocco), *J. Afr. Earth Sci.,* **38,** 145–153.

Le Pichon, X. & Fox, J.P., 1971. Marginal offsets, fracture zones, and the early opening of the North Atlantic, *J. geophys. Res.,* **76,** 6294–6308.

Le Pichon, X., Sibuet, J.-C. & Francheteau, J., 1977. The fit of the continents around the North Atlantic ocean, *Tectonophysics,* **38,** 169–209.

Le Roy, P. & Piqué, A., 2001. Triassic-Liassic Western Moroccan synrift basins in relation to the Central Atlantic opening, *Mar. Geol.,* **172,** 359–381.

Le Roy, P., Guillocheau, F., Piqué, A. & Morabet, A.M., 1998. Subsidence of the Atlantic Moroccan margin during the Mesozoic, *Can. J. Earth Sci.,* **35,** 476–493.

Lizarralde, D., Holbrook, W.S. & Oh, J., 1994. Crustal structure across the Brunswick magnetic anomaly, offshore Georgia, from coincident ocean bottom and multi-channel seismic data, *J. geophys. Res.,* **99**(B11), 21 741–21 757.

Malusà, M.G., Polino, R., Cerrina Feroni, A., Ellero, A., Ottria, G., Baidder, L. & Musumeci, G., 2007. Post-Variscan tectonics in eastern Anti-Atlas (Morocco), *Terra Nova,* **19,** 481–489.

McBride, J.H. & Nelson, K.D., 1988. Integration of COCORP deep reflection and magnetic anomaly analysis in the southeastern United States: implication for origin of the Brunswick and East Coast magnetic anomalies, *Geol. Soc. Am. Bull.,* **100,** 436–445.

McBride, J.H., Nelson, K.D. & Brown, L.D., 1989. Evidence and implications of an extensive early Mesozoic rift basin and basalt/diabase sequence beneath the southeast Coastal Plain, *Geol. Soc. Am. Bull.,* **101,** 512–520.

Morel, P. & Irving, E., 1981. Paleomagnetism and the evolution of Pangea, *J. geophys. Res.,* **86**(B3), 1858–1872.

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

National Geophysical Data Center (NGDC), 1993. Global Relief Data on CD-ROM, NGDC Data Announcement 93-MGG-01, Boulder, CO.

Oh, J., Austin, J.A., Phillips, J.D., Coffin, M.F. & Stoffa, P.L., 1995. Seaward-dipping reflectors offshore the Southeastern United States; seismic evidence for extensive volcanism accompanying sequential formation of the Carolina Trough and Blake Plateau basin, *Geology,* **23**(1), 9–12.

Olivet, J.L., Bonnin, J., Beuzart, P. & Auzende, J.M., 1984. *Cinématique de l'Atlantique nord et central* (Rapp. Sci. Tech.), Vol. 54, pp. 108–112, Centre National pour l'Exploration des Oceans.

Pitman, W.C, III & Talwani, M., 1972. Sea floor spreading in the North Atlantic, *Geol. Soc. Am. Bull.,* **83**(3), 619–646.

Piqué, A. & Laville, E., 1995. L'Ouverture initiale de l'Atlantique central, *Bull. Soc. Géol. de France,* **166**(6), 725–738.

Piqué, A. *et al*., 1998a. Évolution structurale des domaines atlasiques du Maghreb au Méso-Cénozoique; le rôle des structures héritees dans la déformation du domaine atlasique de l'Afrique du Nord, *Bull. Soc. Géol. de France,* **169**(6), 797–810.

Piqué, A., Le Roy, P. & Amrhar, M., 1998b. Transtensive synsedimentary tectonics associated with ocean opening; the Essaouira-Agadir segment of the Moroccan Atlantic margin, *J. Geol. Soc.,* **155**(6), 913–928.

Piqué, A., Tricart, P., Guiraud, R., Laville, E., Bouaziz, S., Amrhar, M. & Ouali, R.A., 2002. The Mesozoic–Cenozoic Atlas belt (North Africa): an overview, *Geodinamica Acta,* **15,** 185–208.

Price, I., 1980, Gravity tectonics on a passive margin: Deep Sea Drilling Project Site 415 in relation to regional seismic data, in *Initial Reports of the Deep Sea Drilling Project,* pp. 759–771, eds Stout, L.N. & Worstell, P., Texas A & M University (Ocean Drilling Program), College Station, TX, USA.

Robertson, A.H.F. & Stillman, C.J., 1979. Late Mesozoic sedimentary rocks of Fuerteventura, Canary Islands: implications for West African continental margin evolution, *J. Geol. Soc. Lond.,* **136,** 47–60.

Roeser, H.A., von Rad, U., Hinz, K., Sarnthein, M. & Seibold, E., 1982. Magnetic anomalies in the magnetic quiet zone off Morocco. In *Geology of the Northwest African Continental Margin,* pp. 61– 68, eds von Rad, U. *et al*., Springer-Verlag, New York.

Roeser, H.A., Steiner, C., Schreckenberger, B. & Block, M., 2002. Structural development of the Jurassic magnetic quiet zone off Morocco and identification of Middle Jurassic magnetic lineations, *J. geophys. Res.,* **107**(B10), 2207, doi:10.1029/2000JB000094.

Roest, W.R., Dañobeitia, J.J., Verhoef, J. & Collette, B.J., 1992. Magnetic anomalies in the Canary basin and the Mesozoic evolution of the central North Atlantic, *Mar. geophys. Res.,* **14**(1), 1–24.

Roussel, J. & Liger, J.L., 1983. A review of deep structure and ocean-continent transition in the Senegal Basin (West Africa), *Tectonophysics,* **91,** 183–211.

Royer, J.-Y., Müller, R.D., Gahagan, L.M., Lawver, L.A., Mayes, C.L., Nürnberg, D. & Sclater, J.G., 1992. A global isochron chart, Technical Report No. 117, 38 pp., Institute for Geophysics, University of Texas, Austin, TX.

Sahabi, M., Aslanian, D. & Olivet, J.-L., 2004. Un nouveau point de départ pour l'histoire de l'Atlantique central, *Comptes Rendus Geosciences,* **336**(12), 1041–1052.

Sandwell, D.T. & Smith, W.H.F., 1997. Marine gravity anomaly from Geosat and ERS 1 satellite altimetry, *J. geophys. Res.,* **102,** 10 039–10 054.

Schettino, A. & Scotese, C.R., 2002. Global kinematic constraints to the tectonic history of the Mediterranean region and surrounding areas during the Jurassic and Cretaceous, in: *Reconstruction of the evolution of the Alpine-Himalayan orogen,* eds Rosenbaum, G. & Lister, G.S.. *J. Virtual Expl.,* **7,** 147–166.

Schettino, A. & Scotese, C.R., 2005. Apparent polar wander paths for the major continents (200 Ma–present day): a palaeomagnetic reference frame for global plate tectonic reconstructions, *Geophys. J. Int.,* **163**(2), 727–759.

Schettino, A. & Turco, E., 2006. Plate kinematics of the Western Mediterranean region during the Oligocene and early Miocene, *Geophys. J. Int.,* **166**(3), 1398–1423.

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

Schlische, R.W., 2003. Progress in understanding the structural geology, basin evolution, and tectonic history of the eastern North American rift system, in *The Great Rift Valleys of Pangea in Eastern North America,* Vol. 1: Tectonics, Structure, and Volcanism, pp. 21–64, eds LeTourneau, P.M. & Olsen, P.E., Columbia University Press, New York, USA.

Schlische, R.W., Withjack, M.O. & Olsen, P.E., 2002. Relative timing of CAMP, rifting, continental breakup, and inversion: tectonic significance, in *The Central Atlantic Magmatic Province: Insights from Fragments of Pangea*, *Geophysical Monograph,* Vol. 136, pp. 33–59, eds Hames, W.E., McHone, G.C., Renne, P.R. & Ruppel, C.R., American Geophysical Union, Washington, DC, USA.

Sclater, J.G., Hellinger, S. & Tapscott, C., 1977. Paleobathymetry of the Atlantic Ocean from the Jurassic to the present, *J. Geol.,* **85**(5), 509–552.

Shaw, P.R., 1987. Investigations of relative plate motions in the South Atlantic using SEASAT altimeter data, *J. geophys. Res.,* **92**(B9), 9363–9375.

Sheridan, R.E. *et al*., 1993. Deep seismic reflection data of EDGE U.S. Mid-Atlantic continental margin experiment: implications for Appalachian sutures and Mesozoic rifting and magmatic underplating, *Geology,* **21,** 563–567.

Steiner, C., Hobson, A., Favre, P., Stampfli, G.M. & Hernandez, J., 1998. Mesozoic sequence of Fuerteventura (Canary Islands): witness of Early Jurassic sea-floor spreading in the central Atlantic, *Geol. Soc. Amer. Bull.,* **110**(10), 1304–1317.

Sundvik, M.T. & Larson, R.L., 1988. Seafloor spreading history of the western North Atlantic Basin derived from the Keathley sequence and computer graphics, *Tectonophysics,* **155**(1–4), 49–71.

Talwani, M. & Abreu, V., 2000. Inferences regarding initiation of oceanic crust formation from the U.S. East Coast Margin and Conjugate South Atlantic Margins, in *Atlantic Rifts and Continental Margins, Geophysical Monograph,* Vol. 115, pp. 211–233, eds Mohriak, W. & Talwani, M., American Geophysical Union, Washington, DC, USA.

Talwani, M., Ewing, J., Sheridan, R.E., Holbrook, W.S. & Glover, L. III, 1995. The EDGE experiment and the US Coast Magnetic anomaly, in *Rifted Ocean-Continent Boundaries, NATO/ARW Series Book,* Vol. 3, pp. 155–181, eds Banda, E., Talwani, M. & Torne, M., Kluwer, Amsterdam.

Tari, V. & Pamić, J., 1998. Geodynamic evolution of the northern Dinarides and the southern part of the Pannonian Basin, *Tectonophysics,* **297,** 269–281.

Van Der Voo, R., 1983. Paleomagnetic constraints on the assembly of the Old Red Continent, *Tectonophysics,* **91**(3), 271–283.

Van Der Voo, R., 1990. Phanerozoic paleomagnetic poles from Europe and North America and comparisons with continental reconstructions, *Rev. Geophys.,* **28**(2), 167–206.

Verhoef, J., Collette, B.J., Dañobeitia, J.J., Roeser, H.A. & Roest, W.R., 1991. Magnetic anomalies off West-Africa (20–38°N), *Mar. Geophys. Res.,* **13,** 81–103.

Vogt, P.R., 1973. Early events in the opening of the North Atlantic, in *Implications of Continental Rift to the Earth Sciences,* pp. 693–712, eds Taling, D.H. & Runcorn, S.K., Academic Press, New York.

Withjack, M.O., Schlische, R.W. & Olsen, P.E., 1998. Diachronous rifting, drifting, and inversion on the passive margin of central eastern North America: an analog for other passive margins, *Am. Assoc. Petrol. Geol. Bulletin,* **82**(5A), 817–835.

Wu, Y., Louden, K.E., Funck, T., Jackson, H.R. & Dehler, S.A., 2006. Crustal structure of the central Nova Scotia margin off Eastern Canada, *Geophys. J. Int.,* **166**(2), 878–906.

APPENDIX

The difference between our algorithm and the method of fitting used by Bullard *et al.* (1965) can be summarized as follows. Assume that the lines to be fitted are represented by two series of unit vectors, respectively $(\mathbf{p}_1, \mathbf{p}_2, \ldots, \mathbf{p}_N)$ and $(\mathbf{q}_1, \mathbf{q}_2, \ldots, \mathbf{q}_M)$, in a reference frame where a test Euler pole $\boldsymbol{e} = (\lambda_e, \varphi_e)$, originally located at latitude λ_e and longitude φ_e, is placed at the North Pole (Fig. 19). Transformation of standard geographic coordinates to this new reference frame is easily obtained by rotating each position vector about an Equatorial pole at $(0°, \varphi_e + 90°)$ by angle $-\theta_e = -(\pi/2 - \lambda_e)$. For each point $\mathbf{p}_i$ on the first line, which can be carried onto the second line at position $\mathbf{p}'_i$ by rotation about the North Pole, let $\phi_i(\boldsymbol{e})$ be the longitude difference between the two locations. Similarly, for each point $\mathbf{q}_j$ on the second line, which can be carried back onto the first one at position $\mathbf{q}'_j$ by rotation about the North Pole, let $\phi'_j(e)$ be the longitude difference. In general, only $n \le N$ points of the first line can be projected onto the second line, and only $m \le M$ points of the second line can be projected back onto the first line. If we rotate the western line by angle ϕ_0 about the North Pole, the individual point misfits are given by $\phi_i(\boldsymbol{e}) - \phi_0$. Similarly, if we rotate the eastern line by angle $-\phi_0$ about the North Pole, we obtain individual misfits $\phi'_j(\boldsymbol{e}) - \phi_0$. The total mean square misfit is given by

$$\chi^2(\boldsymbol{e}) = \frac{N}{n^2}\sum_{i=1}^{n}\left[\phi_i(\boldsymbol{e}) - \phi_0\right]^2 + \frac{M}{m^2}\sum_{j=1}^{m}\left[\phi'_j(\boldsymbol{e}) - \phi_0\right]^2. \qquad \text{(A1)}$$

This formula has some differences with respect to the one proposed by Bullard *et al.* (1965). In fact, the formula of these authors assumed that the same number of points was projected between the two lines. This assumption is adequate when the two lines are considered as perfectly conjugate, that is, when each line must be fitted against the whole conjugate line and not against a subset of the input data. In their work, Bullard *et al.* (1965) had complete information on the geometry of the two conjugate margins. Therefore, their assumption of a common number of projections appears to be adequate. Conversely, in the fit of the ECMA and BSMA anomalies, we could have missing information from one of the two lines (most likely the BSMA). In this instance, we must search for a best fit of one line against a subset of the second line, not necessarily a whole geometrical fit. Eq. (1) takes into account the possibility of missing information from one of the two lines. However, the search must also try to maximize the percentage of matched segments from each line, that is, the number of projected points, because we could find wrong Euler poles that furnish very good fits of small segments of the two lines. This problem is solved in eq. (A1) by multiplying the squared misfit of each line, respectively, by N/n and M/m.

Expression (A1) reaches a minimum when rotation angle $\phi_0 = \phi_0(\boldsymbol{e})$ is given by

$$\phi_0(\boldsymbol{e}) = \frac{\frac{N}{n^2}\sum_{i=1}^{n}\phi_i(\boldsymbol{e}) + \frac{M}{m^2}\sum_{j=1}^{m}\phi'_j(\boldsymbol{e})}{\frac{N}{n} + \frac{M}{m}}. \qquad \text{(A2)}$$

Then, the fitting procedure will search for the Euler pole $\boldsymbol{e}$, which minimizes the misfit χ^2 in Expression (A1).

SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Movie S1. An animation illustrating the tectonic evolution of the Central Atlantic and Atlas regions since the late Triassic.

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