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The African Plate: A history of oceanic crust accretion and subduction since the Jurassic

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ABSTRACT

We present a model for the Jurassic to Present evolution of plate boundaries and oceanic crust of the African plate based on updated interpretation of magnetic, gravity and other geological and geophysical data sets. Location of continent ocean boundaries and age and geometry of old oceanic crust (Jurassic and Cretaceous) are updated in the light of new data and models of passive margin formation. A new set of oceanic palaeo-age grid models constitutes the basis for estimating the dynamics of oceanic crust through time and can be used as input for quantifying the plate boundary forces that contributed to the African plate palaeo-stresses and may have influenced the evolution of intracontinental sedimentary basins. As a case study, we compute a simple model of palaeo-stress for the Late Cretaceous time in order to assess how ridge push, slab pull and horizontal mantle drag might have influenced the continental African plate. We note that the changes in length of various plate boundaries (especially trenches) do not correlate well with absolute plate motion, but variations in the mean oceanic crust age seem to be reflected in acceleration or deceleration of the mean absolute plate velocity.

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1. Introduction

The African plate is at the present the third largest tectonic plate (~60 million km^2) with approximately half of it covered by land. This plate comprises several old cratonic units and accreted younger crust, representing a period of more than 2.5 billion years of continental and oceanic crust growth (e.g. Burke, 1996). Initially part of Gondwana since 550 million years (Ma) and Pangea since 320 Ma, and now surrounded almost entirely by spreading centers, the African plate moved relatively slowly for the last 150 Ma (Lithgow-Bertelloni and Richards, 1998; Torsvik et al., 2010). However, its continental interior experienced many changes throughout this time including rifting and variations in sedimentary basin subsidence, most of them in regions situated thousands of kilometers away from plate boundaries. The African plate was also partly underlain by mantle with above the average temperature—either induced by a series of hotspots or a superswell, or both—that contributed to episodic volcanism (including several Large Igneous Provinces—LIPS), basin-swell topography, and consequent sediment deposition, erosion, and structural deformation (e.g., Bumby and Guiraud, 2005).

Long-term intra-continental crustal deformation may be the related to local and regional tectonic events (e.g., Cloetingh and Burov, 2011; Zoback et al., 1993) and mantle–lithosphere interaction (e.g. Heine et al., 2008). Far-field stresses related to changes in plate boundaries are able to propagate within the lithospheric plates over thousands of kilometers and therefore may trigger rifting, folding and changes in sedimentary basin subsidence rate in remote regions (e.g. Xie and Heller, 2009). The relationship between plate boundary forces and observed deformation within plate interiors has been studied for several large tectonic plates and numerous authors attempted modelling and quantification of ensuing present day and palaeo-stresses resulted from this connection: North America (e.g., Faure et al., 1996), South America (e.g., Meijer and Wortel, 1992), Africa (Meijer and Wortel, 1999), Eurasia (e.g., Nielsen et al., 2007; Warners-Ruckstuhl et al., 2012), Australia (e.g., Coblentz et al., 1995; Dyksterhuis and Muller, 2004), and Pacific (e.g., Faccenna et al., 2012; Wortel et al., 1991).

A systematic study of plate boundaries for the African plate since the opening of surrounding oceanic basins is lacking. This is mainly because geophysical data were insufficient and various models proposed for different oceanic basin formation were not properly linked. The publication of regional and global geophysical datasets in the last couple of years, including magnetic and gravity data acquired by satellites (e.g., Maus et al., 2007; Sandwell and Smith, 2009), prompted us to systematically

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reconstruct the ages and extent of oceanic crust around Africa for the past 200 Ma. Location of continent ocean boundary (COB) and old oceanic crust (Jurassic and Cretaceous) are updated in the light of new data and models of passive margin formation. Data, methods and a short review of available recent studies used to construct a model for the oceanic crust of the African plate are presented in Section 2. The model is presented in detail for selected times in Section 3.

This study aims to take a step further in understanding the causal link between the evolution of one of the largest tectonic plates, the African plate (Fig. 1), and its continental interior at times when important tectonic changes have been recorded. In Section 4, a new model for the evolution of the African plate boundaries from the Jurassic to Present is used to assess whether changes in plate geometry and plate boundaries correlate with deformation in the continental interior. Finally, as a case study, we show how ridge push, slab pull and horizontal mantle drag might have influenced the interior of the African plate in the Late Cretaceous by computing a simple model of palaeo-stresses.

As more detailed observations and measurements point to far-field stresses playing an important role in sedimentary basin evolution (e.g. Abadi et al., 2008; Bosworth et al., 2008) this study can contribute not only to a better understanding of the long-term tectonic history of Africa, but also can be used as an exploration tool for the hydrocarbon industry.

2. Plate boundaries within and around the African continent

2.1. Continent–ocean boundaries

Mapping the transition between the continental and oceanic crust (Continent–Ocean Boundary (COB) or Continent–Ocean Transition Zone (COTZ/OCTZ) proves to be a difficult task taking into account the structural complexities of some of the continental margins as highlighted in numerous recent studies (see Table 1 for an overview of the African continental margin studies). Particularly for the so-called non-volcanic margins, where the presence of exhumed (and serpentinized) continental mantle and isolated blocks of extended continental crust exist for at least tens of kilometers (e.g., Manatschal, 2004), a clear boundary between extended continental crust and oceanic crust is almost impossible to identify. In those cases, the COB is rather defined as the (oceanward) boundary that marks the onset of true oceanic crust. An integrated set of geophysical data is usually necessary for defining these boundaries, because unequivocal evidences resulted from drilling are rarely available. Seismic reflection and refraction data, heatflow and potential field data (for a review see e.g., Geoffroy, 2005; Minshull, 2009), as well as plate reconstructions are usually employed to define and test COBs and COTZs.

The passive margin community is currently in the process of reviewing the strict terminology of *volcanic* and *non-volcanic* margin types, as the criteria so far applied are insufficient and possibly obsolete. For example, a considerable amount of volcanic material is now observed in detailed seismic studies of presumed *non-volcanic* margins, (e.g. Peron-Pinvidic et al., 2010), whereas hyperextension and subcontinental mantle exhumation can also occur in volcanic margins (e.g. Lundin and Doré, 2011). In the case of the so-called *volcanic margins*, the presence of seaward dipping reflectors (SDRs)—volcanic material deposited on rifted and newly formed margins—hinders imaging techniques to fully resolve the exact position of the transition between the broken extended continental crust and newly formed oceanic crust. Similar problems are encountered on highly extended passive margins where subsequent salt deposition may also obstruct imaging deeper structures. While being aware of these complexities, we will use the terminology that defines the two extreme end-members of passive margins when discussing various COB segments in this paper: the magma-dominated (or magma-rich) and magma-poor margins by

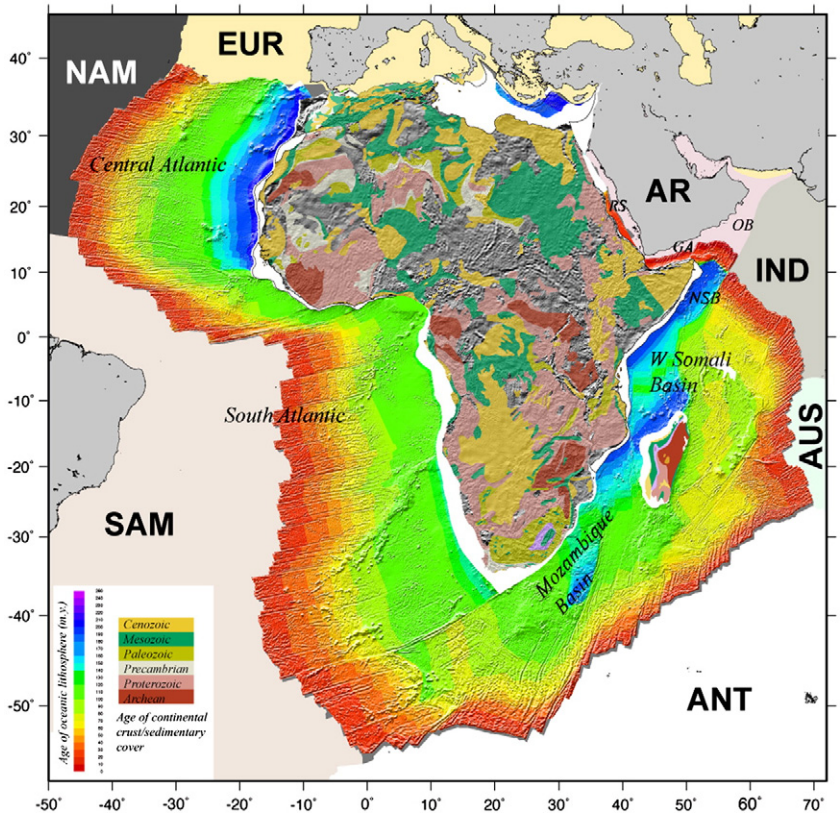


Fig. 1. Present day extent of the African plate. Age of oceanic crust (this study) and ages of continental basement and sedimentary cover (Persits et al., 2002) are draped on shaded relief of free air gravity anomaly (Sandwell and Smith, 2009). The white region surrounding present day coastlines is interpreted as extended continental crust and its oceanward limit is considered the continent–ocean boundary (COB). Neighboring tectonic plates are: Eurasia (EUR), Arabia (AR), India (IND), Australia (AUS), Antarctica (ANT), South America (SAM) and North America (NAM). Other abbreviations: GA—Gulf of Aden, NSB—North Somali Basin, RS—Red Sea, OB—Oman Basin.

Table 1
A summary of studies on the passive margins around the African continent with an emphasis on mapping and modeling the transition between the continental and oceanic area based on geophysical data (SDR – Seaward dipping reflectors; LCB – Lower crustal body; COT – Continent – Ocean Transition Zone).

Margin name/country	SDR	LCB	SALT	COT (km)	Seismic refraction	Seismic reflection	Magnetic	Gravity	References
Central Atlantic/Morocco	**X**		**X**	120		x	x	x	Jaffal et al. (2009); Labails et al. (2009, 2010); Maillard et al. (2006); Sahabi et al. (2004))
Central Atlantic/Western Sahara	**X**		**X**		x	x	x	x	Labails et al. (2009); Klingelhoefer et al. (2009)
Central Atlantic/Mauritania	**X**					x			Vear (2005)
Central Atlantic/Senegal/Gambia			**X**						Roussel and Liger (1983)
Gulf Guinea/Guinea-Bissau									
Siera Leone			**X**		x	x			Bennett and Rusk (2002); Sheridan et al. (1969)
Liberia						x	x	x	Behrendt et al. (1974)
Cote D'Ivoire									Mascle and Blarez (1987) Mascle et al. (1995); Sage et al. (1997)
Ghana/Togo/Benin									Mascle et al. (1995); Sage et al. (1997)
South Atlantic/Nigeria						x		x	Briggs et al. (2009)
South Atlantic/Cameroon						x			Rosendahl et al. (1991)
South Atlantic/Gabon	**X**	**X**				x	x	x	Dupre et al. (2011); Mascle et al. (1995); Nemcok et al. (2012); Rosendahl et al. (1991); Watts and Stewart (1998)
South Atlantic/Congo									Contrucci et al. (2004); Unternehr et al. (2010)
South Atlantic/Angola		**X**	**X**	~150	x	x		x	Blaich et al. (2011); Contrucci et al. (2004); Moulin et al. (2005); Torsvik et al. (2009); Unternehr et al. (2010)
South Atlantic/Namibia	**X**	**X**		~100	x	x	x	x	Bauer et al. (2000); Fernandez et al. (2010); Gladczenko et al. (1998); Heine et al. (2013); Stewart et al. (2000); Torsvik et al. (2009)
South Atlantic/South Africa (SE)	**X**				x	x		x	Hirsch et al. (2009)
Agulhas Bank/South Africa (SW)					x				Parsiegla et al. (2009)
Mozambique Basin/Mozambique	**X**				x	x	x	x	Leinweber and Jokat (2011)
Somali Basin/Tanzania			**X**			x	x	x	Labails and Gaina (2011)
Somali Basin/Kenya			**X**			x	x	x	Labails and Gaina (2011)
N Somali Basin/Somalia (E)							x	x	Labails and Gaina (2011)
Gulf Aden/Somalia (N)				~10–40	x	x	x	x	d'Acremont et al. (2005); Fournier et al. (2010); Leroy et al. (2004, 2010)
Red Sea/Eritrea	**X**						x	x	Bosworth et al. (2005); Cochran and Karner (2007)
Red Sea/Sudan	**X**						x	x	Bosworth et al. (2005); Cochran and Karner (2007)

considering that the *volcanic* margins do exhibit far more volcanic material than the *non-volcanic* margins (Figs. 2 and 3).

For our regional study, we preferred to consistently re-interpret the transition between the continental/extended continental/transitional crust and "true" oceanic crust as a first-order boundary between the continental and oceanic crust that can be used for kinematic reconstructions. Different physical properties reflected in the potential field data (Fig. 2), such as the density and magnetic values of oceanic and continental crust, were the first criteria used to construct a new, systematically defined COB for the African continental margins (Fig. 3).

Potential field data have been proven useful to complement the interpretation of 2D seismic data and assist in areas where seismic information is lacking. Numerous studies offer different algorithms to be used in efficiently capturing structural trends in various tectonic environments by using gravity and magnetic data (e.g., Kimbell et al., 2004; Lyngsie et al., 2006). We have used standard techniques that aimed to isolate gravity anomaly signals that may correspond to changes in crustal density and thickness at passive margins, where extended continental crust (less than 30 km thick) is juxtaposed with oceanic crust (~7 km). Bouguer and isostatic corrections using a crustal density of 2670 kg/m^3, a mantle density of 3330 kg/m^3, sediment infill density of 2300 kg/m^3 and water density of 1030 kg/m^3 (Fig. 2c) were applied to the free air gravity (Fig. 2a) to remove the gravitational effects of the masses that compensate topography mass effect (ETOPO1, Fig. 2b). Using terrain-corrected gravity we have computed a series of isostatic residuals assuming different flexural rigidities for the entire region (and equivalent lithospheric elastic thicknesses, Te). It has been shown that the gravity signal observed along several continental margins does not reflect a perfect Airy isostatic regime (that assumes zero thinning during rifting, Te = 0), and the flexural rigidity of the lithosphere has to be taken into account when modelling the gravity effect of extended margins and COBs (Kimbell et al., 2004; Watts and Stewart, 1998). The second derivative of the upward continued (20 km) isostatic residual computed with an elastic thickness of 20 km seems to coincide with the COB as defined by other geophysical methods (see Table 1). A horizontal gradient of this grid has been computed, and we have visually identified the COB location in the proximity of the slope of the oceanward maxima (Fig. 3b).

The interpretation of COB segments has been checked against published results of (Sahabi et al., 2004) in the Central Atlantic, a modified COB of (Torsvik et al., 2009) in the South Atlantic, and a new COB interpretation for the SE, E and NE of Africa guided by recent studies (Table 1) and derivatives of gravity anomalies (Fig. 3). For comparison we show the global COB location from Müller et al. (2008a) and the landward limit of the oceanic crust (LaLOC) interpreted by Heine et al. (2013) along the South Atlantic margin (Fig. 3b).

2.2. Plate boundaries within continental areas

Before Pangaea breakup and seafloor spreading between NW Africa and North America (around 195 Ma), and besides the northern complex plate boundary (consisting of mid ocean ridges, transform faults and trenches e.g., Stampfli and Borel (2002)), the nascent African plate was bordered to the east, south and west by rifted regions, of which some evolved later into mid-ocean ridges (Burke, 1996; Guiraud et al., 2005). Even after seafloor spreading was established in the South Atlantic and Indian oceans, major rifted areas also developed within the African continent without evolving into mid-ocean ridges; these regions acted as diffuse plate boundaries at various time intervals dividing the main African plate into several individual tectonic blocks. In the last 200 million years the following regional rifts developed (or continued to develop) within the African continent:

1) the Permian–Early Jurassic Karoo rifts that affected the southern, central and eastern part of the African continent (e.g. Catuneanu et al., 2005),

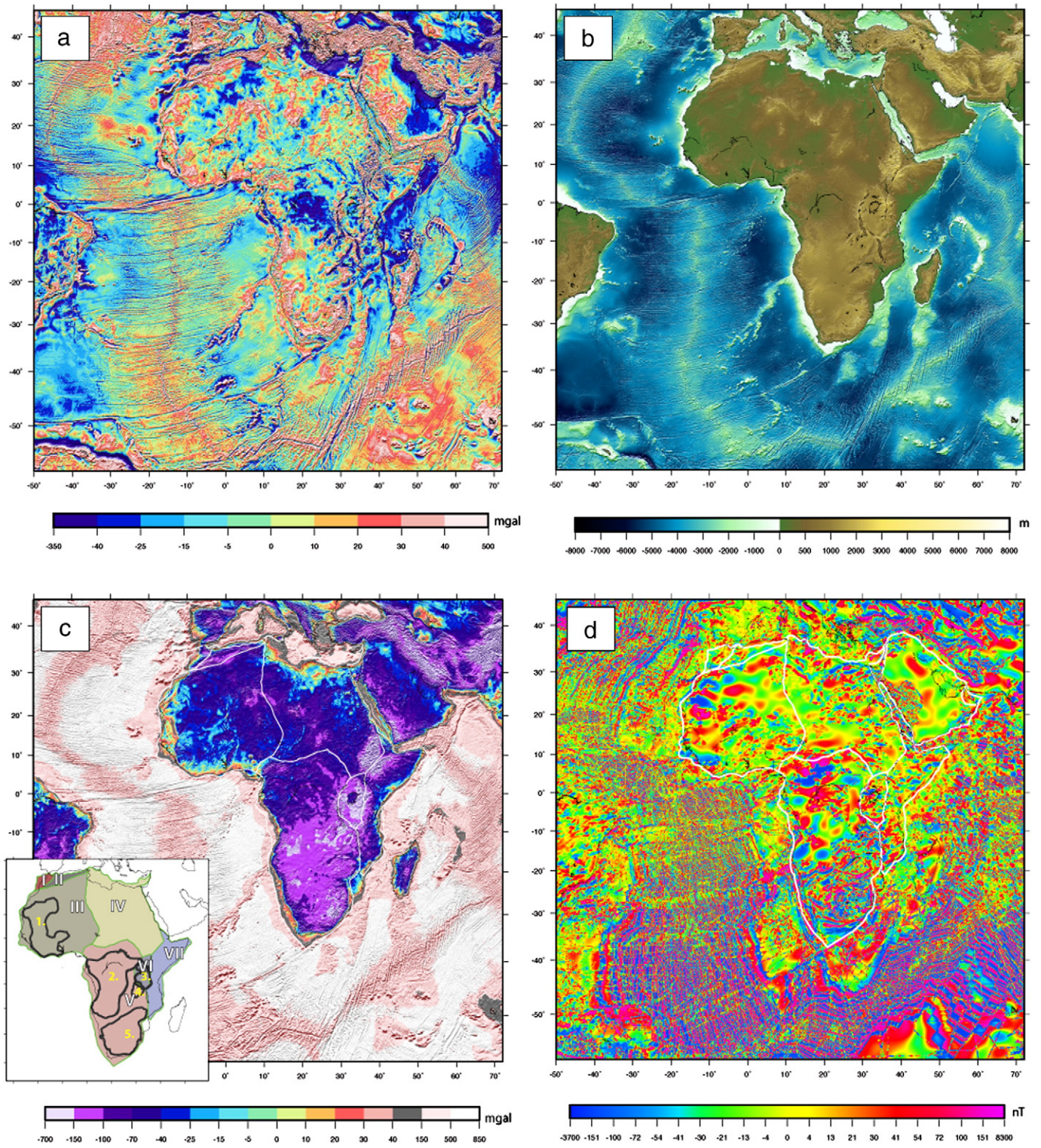


Fig. 2. a) Free air gravity (Sandwell and Smith, 2009) complemented onshore with the EGM2008 model (Pavlis et al., 2012); b) topography and bathymetry (ETOPO1, Amante and Eakins (2009); c) Gravity anomaly Bouguer corrected using ETOPO1 present day topography and bathymetry and a Bouguer correction of 2670 kg/m^3 and d) World Digital Magnetic Anomaly (Maus et al., 2007). White outlines show the present day COB around the African continent and intra continental plate boundaries that were active in Mesozoic and/or Cenozoic. Inset 3c: Main intra-continental plate boundaries and tectonic blocks that moved independent from each other at certain stages from the Jurassic to present day: I—Moroccan Meseta, II—Oran Meseta, III—NW Africa, IV—NE Africa, V—S Africa/Nubia, VI—Victoria, VII—Somalia. Thick black lines show outlines of cratonic areas: 1—West African, 2—Congo, 3–4. Tanzanian, 5—Kalahari (modified after Gubanov and Mooney, 2009).

2) the Late Jurassic–Early Cretaceous Central African Rift System (CARS) (Guiraud et al., 1992) and part of the East African Rift System (EARS) (Guiraud et al., 2005) followed by a pervasive early Senonian–late Santonian compressional event that inverted rift structures in northern and central Africa (Guiraud and Bosworth, 1997), and
3) the Tertiary East African Rift system (Bosworth, 1992; Chorowicz, 2005) that also extended offshore (Mougenot et al., 1986) and

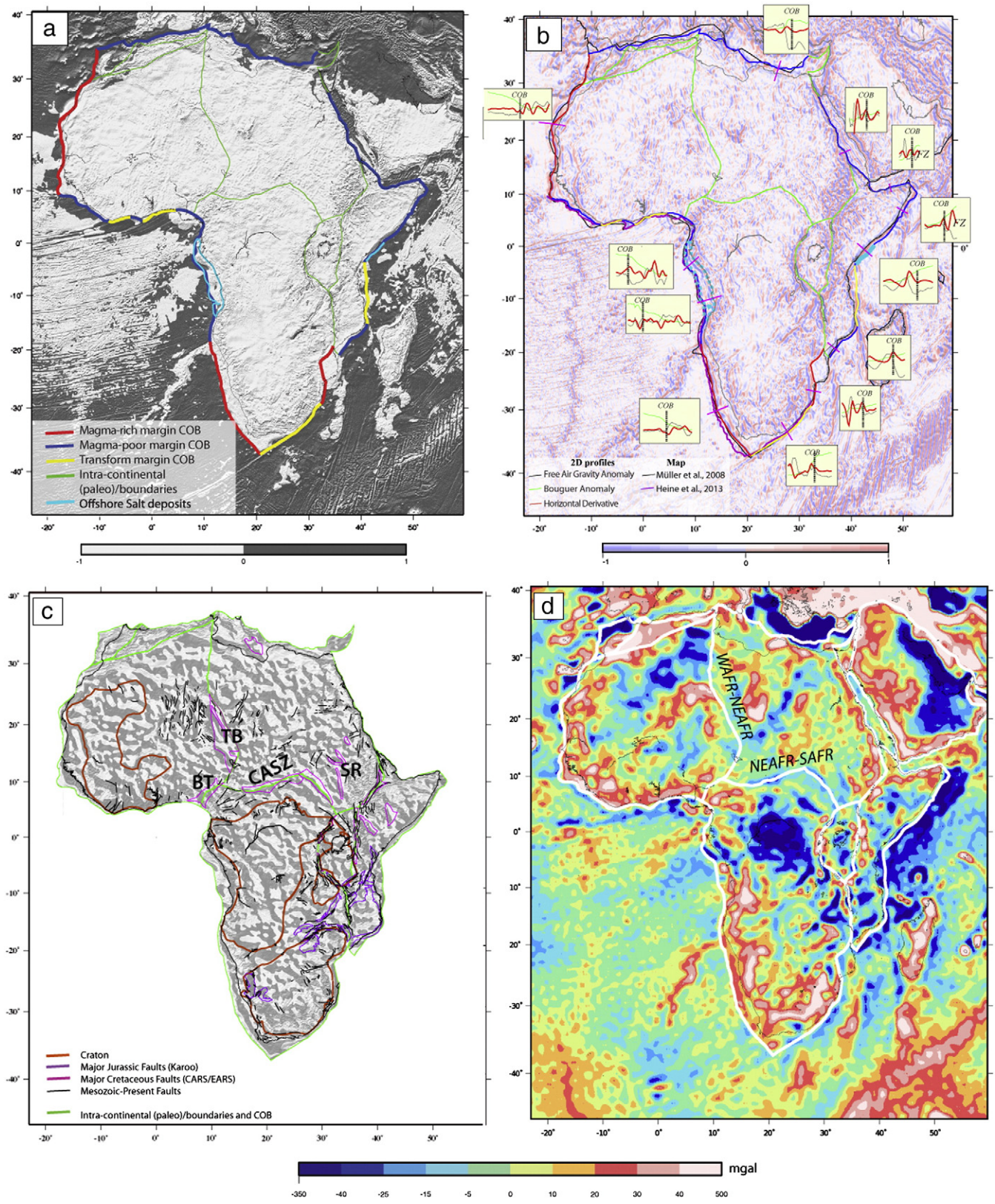


Fig. 3. Derivatives of isostatic gravity residuals (a–c) and satellite gravity anomaly (GOCE) used to map the COB and internal plate boundaries and major faults within the continental area. a) Residual gravity anomaly computed for an Airy Isostasy case (Te = 0); b) Horizontal gradient of second vertical derivative of upward continued (20 km) residual anomaly (Te = 20 km), present study identified COBs and published COBs by Müller et al. (2008a) and Heine et al. (2013); c) Horizontal gradient of second vertical derivative of upward continued (30 km) residual anomaly (Te = 50 km); and d) Satellite gravity anomaly (GOCE) downward continued at 4000 m. Abbreviations are BT—Benue Trough, CASZ—Central African Shear Zone, MR—Muglad Rift, TB—Tenere Basin, NEAFR—Northeast Africa, WAFR–West Africa, SAFR—South Africa.

subsequently led to the formation of the independent Somali plate (Horner-Johnson et al., 2007) and several microplates: Victoria, Rovuma and Lwandle, (Stamps et al., 2008) (Fig. 2).

Before presenting the evolution of the oceanic area around the African continent and defining the kinematics of oceanic crust accretion and destruction, we will describe the zones of lithospheric deformation that evolved into regional plate boundaries within the continental area. These plate boundaries were used in defining the outlines of main tectonic blocks that moved during the opening of various oceanic basins around the continental area. In these regions compression or extension accommodated the offshore tectonic development.

Locations of intra-continental diffuse plate boundaries were published in several studies (e.g., Gregory et al., 2009; Guiraud et al., 1992; Moulin et al., 2010; Torsvik et al., 2010)—some of them interpreted as linear trends in the potential field data, other utilized evidences from geological mapping, or both. The more recent studies (e.g., Moulin et al., 2010; Torsvik et al., 2009) focused on the South Atlantic opening, and for our study we therefore revisit the interpretation of these intra-continental boundaries with an "African plate" emphasis. We start with Torsvik et al.'s (2010) global model that shows the present day simplified plate boundaries of tectonic blocks that moved relative to each other for specific time intervals. The EGM2008 global gravity model is used for onshore areas (Fig. 2a). This is a spherical harmonic model of Earth's gravitational potential based on satellite-derived data (GRACE) merged with terrestrial and airborne gravity data (Pavlis et al., 2012). For checking/defining the location of the main intra-continental boundaries, we have used a residual gravity anomaly grid computed for an elastic thickness of 50 km, which was upward continued 30 km. The horizontal gradient of its derivative was used to establish the final outlines of the tectonic blocks (Fig. 3c). The Te value of 50 km has been selected based on the regional study of Perez-Gussinye et al. (2009) as the median Te value of continental Africa that characterizes regions bordering the cratons and/or crustal heterogeneities inherited from Palaeozoic and older tectonic events subsequently modified by Mesozoic and Cenozoic rifting.

The EGM08 model can resolve features as small as 10 km, as long as terrestrial data are available (Pavlis et al., 2012). As significant parts of the African continent are poorly covered by public-domain airborne or terrestrial data, satellite derived gravity anomaly data were also used to check our interpretation of regional tectonic feature outlines. We show a model based on the gravity anomaly recorded at 250 km by the latest satellite mission, GOCE (Gravity and Steady-State Ocean Circulation Explorer). The model presented here was computed using the time-wise (TIM) modelling approach for the altitude of 4 km (where most of the African topography is removed) (Braitenberg et al., 2011) (Fig. 3d).

Two intra-continental plate boundaries of Torsvik et al. (2009) were modified according to trends in the potential field data as following: one between the NW and NE Africa (in short *WAFR-NEAFR*), running from North Africa through the Tenere Basin (TB, Fig. 3c) towards the Benue Trough (BT, Fig. 3c), and a second one between NE Africa and South Africa (in short *NEAFR-SAFR*) that runs along the Central African Shear Zone (CASZ) continuing along the southern part of the Sudan Rifted area (SR, Fig. 3c) (Guiraud et al., 2005). Potential field data show numerous almost N–S trending lineations in the TB area, which may suggest that the Mesozoic *WAFR-NEAFR* boundary was either a diffuse area or a "jumping" boundary that relocated in time within short distances and created new rifts. We have selected the easternmost part of this rifted region as a plate boundary. In the case of *NEAFR-SAFR*, the gravity anomaly and derivatives show a very clear, almost E–W lineation along the CASZ, with two distinct NW–SE trends, one along the Muglad Rift (MR, Fig. 3c), the southern branch of the Sudan Rift system, and another one south of it, along the NE part of the exposed Archean crust of the Congo craton, trends that were described by earlier studies of Bermingham et al. (1983) and Browne and Fairhead (1983). As most of the Mesozoic plate boundaries seem to follow some inherited discontinuities in the crustal structure (Guiraud et al., 1987), we have chosen the northern trend observed in the gravity anomaly that runs south of MR. The selected boundary location coincides with an inferred Archean–Paleoproterozoic shield boundary and has been also chosen as the active Mesozoic plate boundary between NE Africa and S Africa in the Tectonic Map of Africa (Milesi et al., 2010).

2.3. Plate boundaries in the oceanic domain

Mid-ocean ridges acted as plate boundaries between two tectonic plates while upwelling magmatic material created divergent forces and led to plate growth. Mapping past plate boundaries within an oceanic domain is usually based on trends and ages of (positively and negatively) magnetized basaltic bodies that form the oceanic crust, in combination with oceanic fracture zones that offset these bodies parallel to the direction of movement between the tectonic plates. We interpret the magnetic anomalies derived from data acquired by ships or airplanes in the oceanic domain to infer the ages of oceanic crust and, together with fracture zone trend interpretation (using free air gravity anomaly or its vertical derivative Sandwell and Smith (2009)), construct an isochron model. Magnetic anomalies derived from satellite measurements usually cannot resolve the oceanic spreading anomalies, although some of the latest models seem to be more sensitive to larger bodies of magnetized basaltic layers in the oceanic area (e.g., Maus et al., 2008). A comprehensive model for the oceanic crust ages around the African continent is part of the global age grid published by Müller et al. (2008a). This model is mostly based on the previous global compilation of oceanic crust ages from marine geophysical data (Müller et al., 1997) and has not been updated since. Both in the Atlantic as well as in the Indian oceans, a wealth of studies has been published in the last decade; therefore we review these studies together with our own interpretation, and construct a new circum-African age grid (Fig. 1) and palaeo-age grids. In particular we have included an additional set of nine isochrons for the Eocene interval (C25y, C24y, C23y, C22o, C21y, C21o, C21y, C19o, C18y, C18o, C16y, C15o, C13y) based on a new Paleocene–Eocene global anomaly picks dataset (Gaina et al., 2011) that include Cande et al.'s (2010) identifications in the Indian Ocean, Müller et al.'s (1999) magnetic anomaly picks in the Central and South Atlantic, and new fracture zone segment interpretation based on Sandwell and Smith (2009) gravity anomaly grids. The timescale used in this study is Gee and Kent (2007) for times younger than M26, and Sager et al. (1998) for older times. Our model is illustrated by the new set of isochrons (shown in present day coordinates in Fig. 4) and the rotation model from Table 2. In the following we will briefly describe the regional kinematic models that were incorporated in this model.

2.3.1. Central Atlantic

The classical interpretation of Klitgord and Schouten (1986) for the opening of the Central Atlantic around 175 Ma has been challenged by recent studies. Instead of an earlier ridge jump towards the North American margin, these studies postulate that an earlier break-up was followed by slow seafloor spreading between northwest Africa and northeast America (Labails et al., 2010; Sahabi et al., 2004). Here we have constructed the Central Atlantic isochrons based on this new interpretation for times older than M0 (for which we adopt a 120.6 Ma age according to the Gee and Kent (2007) timescale), whereas for younger times we take Müller et al.'s (2008a) isochrons and interpret additional isochrons as described above.

2.3.2. South Atlantic

The older isochrons in the South Atlantic have been revised based on recent data and new interpretations summarized by Moulin et al. (2010). This study used an improved magnetic anomaly identifications from the new marine magnetic data from the NGDC (www.ngdc.noaa.

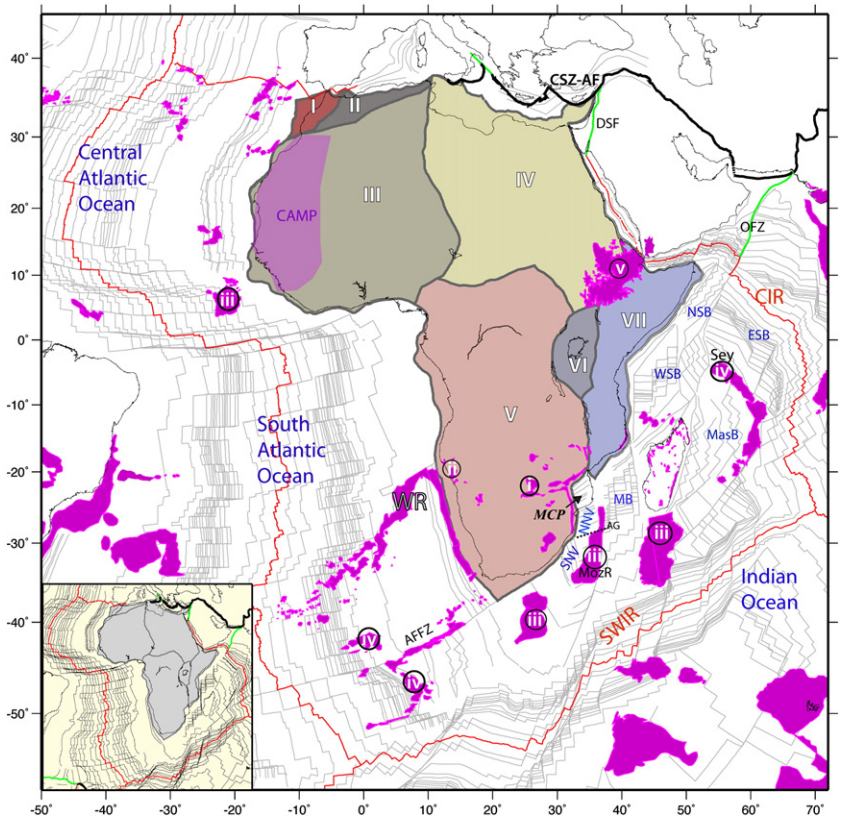


Fig. 4. Circum-Africa isochrons used in this study (grey lines) and volcanic provinces in the continental and oceanic areas (magenta) (Burke and Torsvik, 2004). The active mid-ocean ridges (MOR) are shown in red, trenches and compressional boundaries in black and transform fault boundaries in green. The inset figure shows Müller et al.'s (2008a) isochron model for the same area. Timing of major episodes of intra-plate volcanism is numbered as following: i) Karoo LIP, ii) Parana-Etendeka LIP SE Atlantic, iii) Indian Ocean (Madagascar, Agulhas LIP), iv) SW Indian Ocean and SE Atlantic Ocean (Conrad Rise and Del Cano Rise, Meteor Rise and Discovery seamounts), and v) Afar related volcanism. Roman numbers I–VII are like in Fig. 3c inset map. Abbreviations: AFFZ—Agulhas—Falkland Fracture Zone, AG—ESB, CIR—Central Indian Ridge, CSZ—AF—Cyprus Subduction Zone—Anatolian Fault, DSF—Dead Sea Fault, ESB, NSB and WSB—East, North and West Somali basins; MB—Mozambique Basin; MasB—Mascarene basin, MCP—Mozambique Central Plains; MozR—Mozambique Ridge, OFZ—Owen Fracture Zone, Sey—Seychelles.

gov/mgg/mggd.html) and BGR (B. Schreckenberger, pers. comm.) databases complemented with confidential data. Compared to the model of Müller et al. (2008a) that has only two Early Cretaceous isochrons (M0ry and M4y, with "y" and "o" referring to youngest and oldest side of normal magnetized chron), Moulin et al. (2010) identified a series of denser spaced isochrons south of the Walvis Ridge that span from M7o (127.5 Ma) to M4y (125.7 Ma), M2y (123.5 Ma) and M0o (121 Ma) (ages changed following the timescale of Gee and Kent, 2007). The oldest isochron is situated oceanward of a wide, double-sided anomaly called "Large Marginal Anomaly" (LMA) that is supposed to represent the transitional crust and the oldest oceanic crust covered with SDRs. The oceanward side of this magnetic anomaly is dated to 129 Ma (using the converted LMA age of Moulin et al., 2010) and we have used it as the oldest isochron in the southernmost part of the South Atlantic Ocean. North of the Walvis Ridge there are no identified M anomalies as the oldest generation of oceanic crust started probably at the beginning of the Cretaceous Normal Superchron (CNS, 83.5–120.6 Ma). Isochrons for Mid-Cretaceous to Present were constructed in the same way as for the Central Atlantic.

2.3.3. West Indian Ocean

2.3.3.1. Mozambique and West Somali Basins, South and North Natal Valley. The oldest part of the West Indian Ocean, the Mozambique and the West and North Somali basins (MB, WSB and NSB, Fig. 4) recorded the early history of Gondwana break-up and are floored by Jurassic and Cretaceous crust (e.g., Norton and Sclater, 1979). The West Somali Basin is an old Jurassic–Cretaceous inactive basin that formed as a result of Somali/Madagascar plate separation (Figs. 1 and 4). Detailed studies have been published in the 1980s when a complete set of geophysical data and drilling cores were collected in this area (e.g., Coffin and Rabinowitz, 1987; Coffin et al., 1986; Segoufin and Patriat, 1980). Based on these data, most authors concurred that breakup occurred in the Mid-Late Jurassic followed by seafloor spreading until the Aptian (~M0) (Segoufin and Patriat, 1980) or Barremian (~M10) (Coffin and Rabinowitz, 1987). A study that aimed to link the evolution of the Mozambique Basin with the opening in the WSB (Eagles and Konig, 2008), reinterpreted fracture zone and vintage magnetic anomaly data between Africa and Madagascar and suggested that the WSB seafloor spreading became extinct at ~M10, a solution similar to the Coffin and Rabinowitz (1987) interpretation, but the breakup in both basins occurred probably earlier, at ~170 Ma or slightly older.

New gravity data, vintage and new industry seismic data and vintage magnetic data together with published studies on the rifted African and Madagascan margins have been used to re-evaluate the location and timing of the COB and early seafloor spreading in the WSB (Gaina et al., 2010a; Labails and Gaina, 2011). These studies suggest a late Toarcian–Aalenian age for the break-up and seafloor spreading commencing around 170 Ma, with M41 (167.5 Ma—Sager et al., 1998 timescale) as the oldest Jurassic magnetic anomaly and M2 the youngest, followed by the mid ocean ridge extinction around 123 Ma. Our isochrons are constructed based on the above mentioned interpretation (Fig. 4).

Several recent scientific cruises collected geophysical data in the Mozambique basin that was used to construct a more detailed model

Table 2
Circum-African isochron model—finite rotations.

Age (Ma)	Chron	Rotation		
		Latitude	Longitude	Angle
I. Central Atlantic: North America relative to NW Africa				
10.949	C5n.2o	81.0	22.9	2.85
20.131	C6no	80.6	24.4	5.54
33.058	C13ny	76.0	6.0	9.76
35.343	C16ny	75.3	3.1	10.68
40.130	C18n.2no	74.5	−1.2	12.61
43.789	C20no	74.3	−3.6	14.09
47.906	C21no	74.9	−4.6	15.61
49.714	C22no	75.9	−3.5	16.15
52.364	C24n.1ny	77.3	−1.6	16.95
55.904	C25ny	80.6	6.6	17.90
67.735	C31ny	82.3	−1.8	21.53
83.50	C34ny	76.8	−20.6	29.51
120.60	C34no	65.9	−20.5	54.56
125.67	M4ny	65.9	−19.8	56.26
130.49	M10Nn.3no	66.0	−19.1	57.76
137.85	M16no	66.6	−18.1	59.34
146.56	M21no	66.0	−18.7	62.29
154.0	M25no	67.1	−15.9	64.23
Breakup (195)		64.3	−15.2	77.09
II. South Atlantic: South America relative to South Africa				
10.949	C5n.2o	61.2	−39.7	3.70
20.131	C6no	58.5	−37.1	7.53
33.058	C13ny	56.2	−33.6	13.39
35.343	C16ny	56.6	−33.3	14.46
38.426	C18n.1ny	57.1	−33.0	15.91
40.130	C18n.2no	57.1	−32.5	16.61
41.521	C19no	57.2	−32.2	17.16
43.789	C20no	57.3	−31.8	18.07
46.264	C21ny	57.4	−31.4	19.05
47.906	C21no	57.5	−31.2	19.70
49.714	C22no	58.1	−31.2	20.38
50.946	C23n.1no	58.5	−31.2	20.84
52.364	C24n.1ny	58.9	−31.2	21.37
55.904	C25ny	61.4	−32.2	22.27
67.735	C31ny	63.7	−33.5	25.41
83.5	C34ny	61.9	−34.3	33.51
120.60	C34no	51.4	−35.0	52.54
121.00	M1ny	51.4	−35.0	52.60
125.70	M4ny	50.6	−33.8	54.20
127.79	M8ny	50.1	−33.6	54.98
128.93	M10ny	49.0	−33.3	55.66
Breakup (130)		48.4	−33.3	56.18
III. Indian Ocean: Antarctica relative to South Africa				
10.949	C5n.2o	6.5	−47.8	1.86
20.131	C6no	9.2	−47.1	2.97
33.058	C13ny	11.8	−44.4	5.77
40.130	C18n.2no	13.1	−43.6	7.25
43.789	C20no	11.3	−42.1	8.07
47.906	C21no	9.3	−40.6	9.04
49.714	C22no	8.7	−40.6	9.42
51.743	C23n.2no	8.9	−41.5	9.82
53.347	C23n.3no	9.7	−43.2	10.17
55.904	C25ny	9.5	−45.1	10.70
57.554	C26ny	10.2	−47.3	10.98
67.735	C31ny	1.7	−43.9	12.21
83.50	C34ny	−1.4	−34.7	18.00
120.60	C34no	−10.3	−26.4	41.85
123.55	M2ny	−7.8	−29.5	42.15
125.67	M4ny	−8.9	−28.8	43.54
127.79	M8ny	−8.6	−29.0	44.47
128.93	M10ny	−8.5	−29.2	45.07
130.84	M11ny	−8.3	−29.4	45.90
133.73	M12Any	−8.0	−29.4	47.04
140.51	M18ny	−7.6	−30.6	49.12
143.36	M20n.1ny	−7.3	−31.1	49.80
147.06	M22n.1ny	−6.9	−31.5	50.72
148.67	M22n.2no	−6.7	−31.8	51.22
155.32	M26no	−8.2	−30.9	53.72
158.12	M29no	−11.1	−29.0	58.87
Breakup (170)				
IV. West Somali Basin: Madagascar relative to Nubia				
124.05	M2no	2.08	−42.04	0.38
125.67	M4ny	2.08	−42.04	0.96
130.49	M10Nn.3no	1.36	−44.40	2.32
132.76	M12Any	1.64	−43.48	3.34
140.51	M18ny	1.98	−42.89	5.67
155.32	M26no	−6.35	−33.38	9.98
Breakup (170)	M40ny	−14.44	−21.93	15.3
V. North Somali Basin: India relative to Nubia				
135.00		−28.8	−140.9	56.62
137.85	M16no	−25.1	−141.7	58.46
140.51	M18ny	−21.4	−144.2	60.19
155.32	M26no	−21.1	−144.7	60.31
Breakup (165)		−20.6	−145.5	60.53
VI. Mascarene Basin: Seychelles/India relative to Nubia/Madagascar				
67.735	C31ny	0.3	−144.4	14.42
73.619	C33ny	0.2	−144.4	24.11
83.5	C34ny	−9.7	−148.3	26.20
Breakup (90)		−4.0	−141.4	35.44
VII. Central Indian Ocean: India relative to Seychelles/Madagascar/Nubia				
10.949	C5n.2o	−23.6	−150.8	4.30
20.131	C6no	−26.0	−149.3	8.43
33.058	C13ny	−20.6	−137.6	16.05
40.130	C18n.2no	−20.4	−135.5	19.61
42.536	C20ny	−20.8	−136.0	20.54
43.789	C20no	−21.3	−136.4	20.94
46.264	C21ny	−20.4	−137.0	22.13
47.906	C21no	−20.3	−137.7	23.20
49.714	C22no	−20.4	−140.3	24.42
51.743	C23n.2no	−18.5	−141.9	27.11
53.347	C23n.3no	−18.3	−144.5	29.04
55.904	C25ny	−18.2	−147.7	31.82
57.554	C26ny	−17.3	−148.3	34.15
60.920	C27ny	−15.5	−148.9	38.42
62.499	C28ny	−14.9	−149.2	40.49
Breakup (64)		−15.5	−151.5	41.94
VIII. Greater Adria relative to Iberia				
160		41.5	154.0	0.88
170		73.5	−5.5	5.38
IX. Vardar Ocean: Pontides relative to Taurides/North Africa				
180		50.6	1.4	46.97
200		49.4	−3.2	26.29
210		49.9	−4.4	12.96
Breakup (225)		−41.6	−148.3	0.57
X. Neotethys: Iran/Lut/Afghan relative to Arabia				
140		45.9	−4.2	50.53
160		47.0	−2.2	47.04
180		47.4	−2.9	44.07
200		45.8	−2.5	36.59
220		44.2	3.9	22.57
240		44.4	4.1	12.98
Breakup (260)		45.6	5.1	3.87

for the early separation of Africa and Antarctica (König and Jokat, 2010; Leinweber and Jokat, 2011, 2012). Leinweber and Jokat (2011) postulated an older timing for the Gondwana break-up (~185 Ma), about 20 million years older than in previous studies (Jokat et al., 2003). They also interpreted as oceanic crust the Mozambique Coastal Plains (MCP)—an elevated area north of the Northern Natal Valley (Fig. 4) that was previously considered as continental crust (but see also Klausen, 2009). Almost all recent studies, which are based on new geophysical datasets and improved kinematic reconstructions (Eagles and Konig, 2008; König and Jokat, 2010; Leinweber and Jokat, 2012), suggest an Early Jurassic Gondwana breakup, a time closer to the age of

the Karoo LIP formed by a magmatic event that affected the South and East African continent (Jourdan et al., 2007). The Mozambique Basin and the West Somali Basin opened almost simultaneously (Eagles and Konig, 2008; Segoufin and Patriat, 1980), while the African/Somali plates moved northward, leaving the Antarctic and Madagascar blocks behind. As the model of Leinweber and Jokat (2011) of early Gondwana breakup between Africa and Antarctica assigned an Early Jurassic break-up time for the opening of Mozambique basin, and we inferred a COB age of ca. 170–175 Ma for the West Somali Basin, we here assign a conservative breakup age for these basins of 170 Ma. The Mozambique Basin isochrons are constructed following the magnetic anomaly interpretation of König and Jokat (2010) up to M25 (154 Ma). For the time interval from M25 to M41 (167.5 Ma) we constructed isochrons taking into account the basin geometry and assuming symmetric seafloor spreading.

In our model we have adopted an oceanic origin for the MCP and Mozambique Ridge (MozR) (Fig. 4) and regarded this corridor as part of the Mozambique Basin opening until M0 when the ridge between the MCP/NNV and MR became extinct. We have considered the Ariel Graben (AG, Fig. 4), which has a pronounced negative gravity anomaly, as the only extinct ridge in this area, although Leinweber and Jokat (2011) postulated additional southward ridge jumps. Early to Mid Cretaceous isochrons (132–121 Ma) have been constructed for the Southern Natal Valley (SNV), as postulated by previous authors (Goodlad et al., 1982).

2.3.3.2. North Somali Basin (NSB). The NSB is a deep oceanic basin situated north of the WSB and south of the Cenozoic Gulf of Aden oceanic basin (Figs. 1 and 4). The age of oceanic crust in this basin has been only inferred based on depth of basement, sediment ages and the regional tectonic framework (e.g., Mountain and Prell, 1990). In the same fashion, we have inferred a Jurassic age for the NSB, probably similar to that in the WSB, taking into account that India, which was juxtaposed to the Somali plate margin, was fixed to Madagascar while both were separated from the East African margin during the WSB opening (e.g., Coffin and Rabinowitz, 1987). Following this inference, oceanic crust north of the NSB (offshore SE Arabia, in the position of today's Oman Basin, Fig. 1), would have been the continuation of the NSB (and consequently of the same age). However, the age of the modern Oman Basin oceanic crust might be younger than NSB as suggested by basement depth, sediment ages and geophysical data (Mountain and Prell, 1990). In addition, the eastern Oman margin contains ophiolites (e.g. the Masirah ophiolite) that obducted onto East Arabian margin in the latest Cretaceous to Eocene time during transpressional motion between India and Arabia (~65–55 Ma according to Gnos et al., 1997; Immenhauser et al., 2000). The Masirah ophiolite section has a Late Jurassic (~150 Ma) MORB-type oceanic crust and contains a classical ophiolite sequence, albeit with a very thin gabbroic crustal section (~200–500 m) suggesting that it formed in a magma-poor environment (Peters, 2000). Hence, although the ophiolite record demonstrates that Jurassic seafloor spreading occurred also along the Oman margin, the inception of that spreading, as well as the pre-Cretaceous architecture of this basin, have been obliterated by subsequent compression and transpression. Our isochrons in the NSB are constructed using the kinematic parameters for the Nubia–India relative motion (Table 2) and the interpreted trends of NNW–SSE fracture zones from gravity anomaly models.

2.3.3.3. Late-Cretaceous–Cenozic Indian Ocean. Isochrons in the Mascarene Basin are based on a simplified model of Bernard and Munschy (2000), and isochrons for Cenozoic oceanic crust are following Cande et al. (2010) kinematic model.

2.3.3.4. Gulf of Aden and Red Sea. For the youngest oceanic basins that are part of the African plate: the Gulf of Aden and the Red Sea (Fig. 1), our isochron model followed Cochran (2005) and Fournier et al. (2010) with an extrapolated age of the COB of 15 and 21 Ma in the Gulf of Aden and ca. 11 Ma in the Red Sea respectively.

2.3.4. Mediterranean/Neotethys region

Mapping past plate boundaries east, south and west of the African plate is facilitated by the presence of oceanic crust whose age and structure can be identified relatively straightforward. In contrast, the northern part of the African plate underwent a very complex history involving numerous tectonic blocks and jumping plate boundaries. Apart from limited amount of oceanic crust left *in situ*, the evidence that documents the past motion of these blocks resides in ages and structure of obducted ophiolites and upper crustal nappes that were offscraped from the subducted African lithosphere and accreted to overriding oceanic (ophiolitic) and continental margins (e.g., van Hinsbergen et al., 2005) together with inferred ages of tectonic events including collision and subduction. Here we present a simplified interpretation of the northern African plate boundary through time that complies with regional kinematic framework and first order interpretation of geological data.

2.3.4.1. Preserved oceanic crust of the northern African plate. The Ionian abyssal plain (actively subducting at the Calabrian trench to the west and at the Hellenic trench to the east, Fig. 5) and part of the East Mediterranean basin (subducting northward below Cyprus, Fig. 1) are now the only "in situ" remnants of the southern branch of the Neotethys Ocean that developed as part of the African plate. Accretionary prisms (the submerged Calabrian and Mediterranean bathymetric highs, normally called "ridges") confine these oceanic domains from the NW, NE and N respectively (Finetti, 1976; Huguen et al., 2001; Mascle et al., 1986; ten Veen et al., 2004) (Fig. 5). The oceanic nature of the Ionian crust was inferred from seismic studies (e.g. Di Luccio and Pasyanos, 2007) and gravity modelling constrained by seismic data (Makris and Yegorova, 2006). A recent study documented in detail the structure of the old oceanic crust in the Ionian Abyssal plain (Gallais et al., 2011). The oceanic nature of some of the crust in the East Mediterranean basin has been proposed based on geophysical data, mainly for the western part of the basin (Gesret et al., 2010; Koulakov and Sobolev, 2006). Recent studies show that the easternmost part of the East Mediterranean basin, the Levantine basin, offshore Israel, has the structure of extended continental crust (Gardosh and Druckman, 2006).

2.3.4.2. Subducted oceanic crust that belonged to the African plate. A conceptual simplification of the Mediterranean region subdivides the Jurassic paleogeography between Africa and Eurasia into a microcontinental domain underlain by Precambrian ("Pan-African") and late Paleozoic ("Hercynian") basement—here termed "Greater Adria". Remnants of Greater Adria include all continental platforms and basins whose rocks are now found in the Apennines, the Southern and Austro-Alpine Alps, Dinarides, Hellenides and Anatolide–Taurides (Fig. 5b). Greater Adria was separated from Africa by the aforementioned still existing oceanic crust of the southeast Mediterranean basin. To the west and north, Greater Adria was separated from Eurasia by the Jurassic Alpine Tethys (Frisch, 1979) and to the northeast by a northern branch of the Neotethys represented by the Sava–Izmir–Ankara suture zone (Schmid et al., 2008; Sengor and Yilmaz, 1981, also see below).

The Jurassic opening of an Alpine Tethys (or Piemonte–Ligurian Ocean) domain between Africa, Greater Adria and Eurasia follows directly from differences in opening rates in the Central and North Atlantic ocean (Dercourt et al., 1986; Stampfli and Borel, 2002), and from Mid-Jurassic fragments of oceanic crust and mantle lithosphere with mostly Mid Ocean Ridge-MOR type geochemistry and 169–148 Ma ages (e.g., Smith, 2006) that are widespread in accretionary prisms preserved in the Apennines, Alps and Carpathians. Fig. 5 shows Jurassic to Early Cretaceous motion paths and estimated extension between Africa–Iberia (AFR-IB) and Africa–Eurasia (AFR-EUR) according

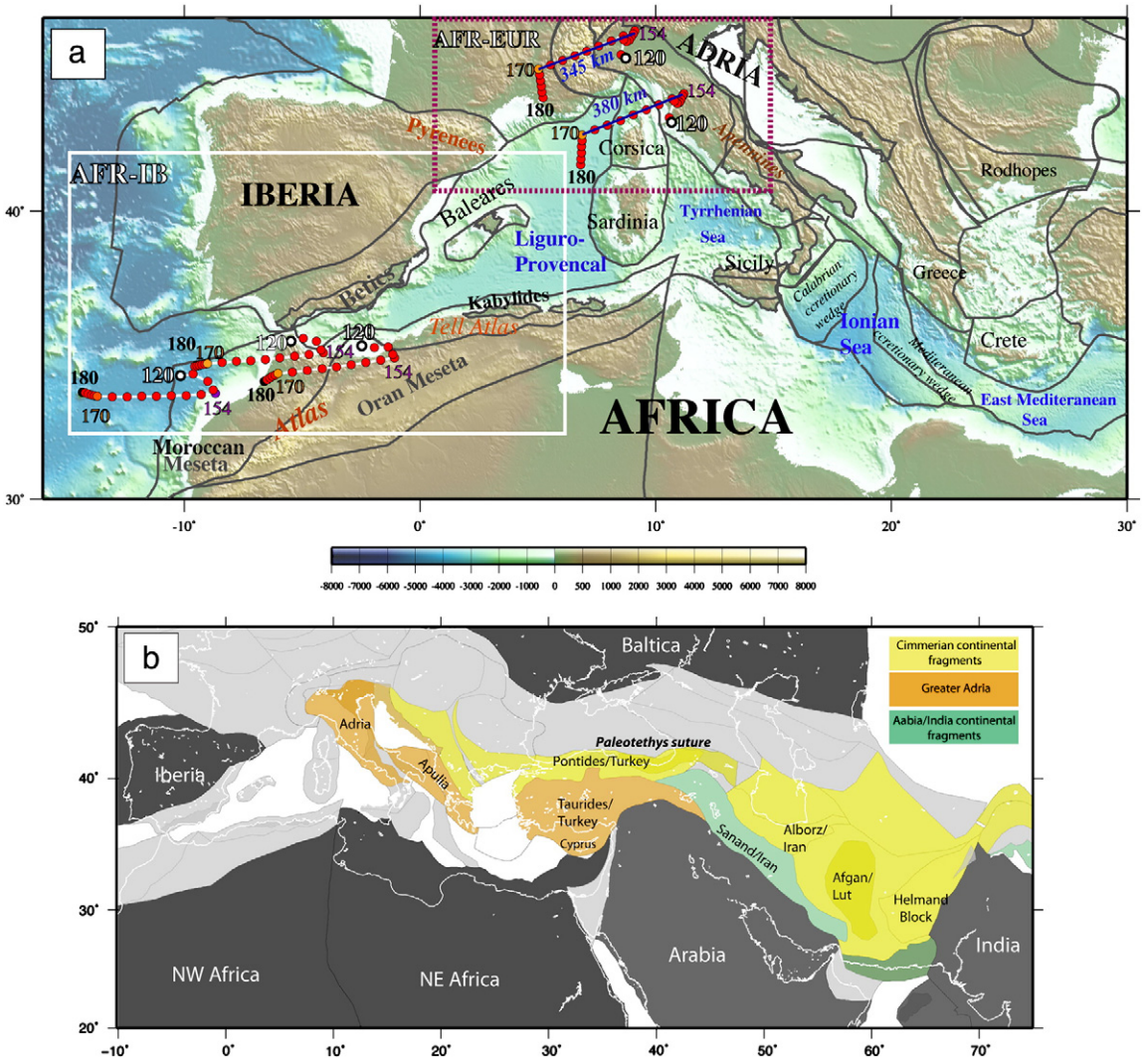


Fig. 5. a. Estimated extension between the Iberian plate and neighboring margins in the North Africa (motion path in the white box) and Iberia and (Greater) Adria (motion paths in the red box) that led to formation of oceanic crust between North Africa, Iberia and fragments of Adria in the Mesozoic. Transtension followed by strike-slip motion between Iberia and NW Africa from 180 to 154 Ma and extension between Iberia and (Greater) Adria from 170 to 154 Ma (~350 km) suggest that a transform plate boundary linked Central Atlantic to the Alpine Tethys (Ligurian-Piemont ocean)—a region that consisted of oceanic crust only between Iberia and (Greater) Adria. b. Present day location of main tectonic blocks that were displaced and deformed during the opening and closure of the Neotethys ocean.

to our composite rotational model. In this model, using the recent Iberia–Eurasia kinematic restorations of Vissers and Meijer (2012a,b) less than 400 km of extension between major tectonic plates is expected (Fig. 5), that is substantially less than in other models. For simplicity, we have treated Greater Adria as an African promontory (i.e. we keep it fixed to northern Africa), except for the last 4 million years when it experienced a small counter-clockwise rotation relative to Africa (van Hinsbergen and Schmid, 2012).

Transtension followed by strike-slip motion between Iberia and NW Africa from 180 to 154 Ma and extension between Iberia and (Greater) Adria from 170 to 154 Ma (~350 km) suggest that a transform plate boundary linked the Central Atlantic to the Alpine Tethys—a region that may have consisted of oceanic crust only between Iberia and (Greater) Adria, with extended continental margins dominating farther to the northeast in the Alpine region (Mohn et al., 2012).

A review of the controversy regarding the timing and geometry of the Eastern Mediterranean Basin is presented by de Lamotte et al. (2011) who, after analyzing the contrasting arguments brought forward for a Permian (Stampfli and Borel, 2002) and Late Triassic–Early Jurassic break-up age (Garfunkel, 1998; Robertson, 2006, 2012) concluded that the East Mediterranean basin should be rather a Late Triassic–Early Jurassic branch of the Neotethys opened in a north–south direction as the Taurides block (i.e., the eastern part of Greater Adria, Fig. 5) rifted away from the northern African margin. This conclusion was recently also put forward by Speranza et al. (2012) based on marine magnetic data. More or less simultaneously, the Pontides rifted away from Greater Adria opening the northern branch of the Neotethys as suggested by the oldest ages of oceanic crust found in the Izmir-Ankara suture zone in Turkey (e.g., Tekin et al., 2002). For the south-eastern Mediterranean basin, we adopt the interpretation of de Lamotte et al. (2011) and assign an age of ~220 to ~170 Ma to its oceanic crust whose remnants are now preserved in the Ionian Abyssal plain and in some regions of the East Mediterranean basin. The assumed transition to the continental crust follows the interpretation from the studies by Longacre et al. (2007) and Makris and Yegorova (2006).

Rifting along northern and north-eastern Arabia (which remained part of Africa until the onset of rifting in the Red Sea, ~35 Ma, (Bosworth et al., 2005)) occurred well before the separation of Greater Adria from Africa, in the Late Permian time. This age is inferred from the stratigraphic analysis of the northeastern Arabian passive margin rocks and paleomagnetic data (e.g., Muttoni et al., 2009; Robertson, 2007; Ziegler et al., 2001). Rifting led to the separation of a continental sliver

(or slivers) now found in Iran and Afghanistan (Sanand, Alborz, Lut and Helmand blocks, see Fig. 5). Seafloor spreading probably continued until the Late Triassic collision of these blocks with Eurasia (e.g., Muttoni et al., 2009). It has been suggested that pulses of rifting affected the northern part of the Arabian block, along the Oman margin, since the Carboniferous, but we model the inception of seafloor spreading starting in the Late Permian (~260 Ma) following Robertson (2007) and Ziegler et al. (2001). The Jurassic to Late Cretaceous intra-oceanic tectonic history of the Neotethys between Arabia and Eurasia is poorly constrained. More or less continuous arc volcanism along the southern Iranian margin (Sanandaj–Sirjan arc) occurred since ~150 Ma, attesting to northward subduction below the Eurasian margin since this time (Agard et al., 2011). Around ~95 Ma, a second, intra-oceanic subduction zone formed northeast of Arabia and ended with the obduction of ophiolites over the Arabian margin in the Late Cretaceous (~70 Ma) from Oman through Iran and Iraq to Syria (Agard et al., 2011; Al-Riyami et al., 2002; Dilek and Furnes, 2011; Searle and Cox, 2009). These ophiolites typically have supra-subduction zone geochemical signature and attest a phase of ocean spreading during the Late Cretaceous subduction, either in a back-arc fashion (e.g. Dilek and Furnes, 2011; Stampfli and Borel, 2002), or close to the trench in the forearc (Dewey and Casey, 2011; Stern et al., 2012). Given their similar age and setting (e.g., Celik et al., 2006), it seems most likely that the Cretaceous ophiolite belts of the Anatolide–Tauride–Cyprus region (Fig. 5) and north-eastern Arabia were derived from a laterally semi-continuous intra-oceanic subduction zone (e.g. Pedersen et al., 2010). Following ophiolite obduction, Arabia converged with Eurasia, and started to collide in Oligocene time (~27–28 Ma or thereafter (McQuarrie and van Hinsbergen, 2013)).

3. Absolute and relative motion of the African Plate

The African plate grew from the Jurassic time onwards by oceanic crust accretion eastward, southward and westward and partially northward. Following the Jurassic inception of subduction in the eastern Mediterranean region, subduction started to consume oceanic (and continental) crust of the northern African plate margin leaving only the small remnants in the East Mediterranean basin today. Our new circum-African isochron and kinematic model is used here to construct oceanic crust palaeo-age grids and furthermore analyze the age and geometrical parameters of the plate at key geological times in the African plate history. In addition, we re-evaluate the direction of motion of this plate relative to the mantle and aim to discuss the results within a regional geological perspective.

3.1. Relative plate motion: Circum-African palaeo-age grids

The isochron and rotation model based on existing and updated kinematic models for the evolution of oceanic crust around the African continent described above forms the basis for constructing palaeo-age grids. This technique has been used to make the present day global oceanic age grid (Müller et al., 2008a) and regional (e.g., Gaina and Müller, 2007) and global oceanic palaeo-age grids (Müller et al., 2008b). We select a series of six reconstructions and associated palaeo-age grids to discuss the position and areal extent of the African plate and its boundaries at times that coincided or preceded major plate boundary reorganizations.

3.2. The African plate absolute motion

Global absolute plate motion models aim to express the connection between the lithospheric plates and a reference frame usually connected to the Deep Earth. The most common references used so far are the Earth's spin axis and hotspots. We present and discuss the position of the African and surrounding plates and microplates relative to the underlying mantle (the "absolute" motion) based on a new global reference frame model (Doubrovine et al., 2012). This model uses relative plate motions based on paleomagnetic data and marine geophysical data plus estimation of both Pacific and Indo-Atlantic hotspot motion relative to the mantle to construct a self-consistent global model for relative and absolute motions for the past 124 Ma. This model has been merged with a true-polar wander (TPW) corrected paleomagnetism-based global reference frame for times older than 124 Ma, and we will call this model GMHRF^{h}2012.

Besides the recent global reference frame model, the absolute plate motion for selected geological times is also presented in some of the previously published (and still widely used) global models: 1) Müller et al. (1993)—FHRF1993 and 2) Torsvik et al. (2008)—HYBRF2008. The FHRF1993 model is based on hotspot tracks in the Indian and Atlantic oceans and considers that these hotspots were fixed relative to the mantle for the last 130 million years. The "moving hotspots" models superseded this model, as many evidences have shown that hotspots are deflected from vertical ascent through the mantle by the mantle circulation (or "mantle wind", see Steinberger and O'Connell, 1998). Despite these findings, the FHRF1993 is still used in many recent studies and therefore we will use it in our analysis for demonstrating differences between the models. Model HYBRF2008, like our preferred model, the GMHRF^{h}2012, is based on the "moving hotspot" hypothesis, and it takes the global hotspot tracks and a global mantle convection and hotspot motion model to compute the African plate motion relative to the moving hotspots for the last 100 million years. For older time a longitude adjusted (5°) paleomagnetism-based reference frame was considered.

Finally, we also compute the African plate absolute motion in the slab-fitted reference frame—SFRF2010 (van der Meer et al., 2010) that used the true-polar wander corrected paleomagnetism-based reference frame of Torsvik et al. (2008) and Steinberger and Torsvik (2008)—which in absence of palaeo-longitudinal control assumes a stable longitude for Africa—and adjusted the plate circuit in longitude to match subduction zones in the plate circuit to corresponding lower mantle slabs assuming vertical slab sinking. SFRF2010 applies a westward correction of up to 18° in the Late Jurassic compared to Africa's longitude today, decreasing in magnitude with older and younger times. This is the first absolute reference frame obtained with this method, and we will use it for illustrating similarities and differences with other models.

The mean absolute motion of the African plate in 10 million years interval as described by these four reference frame models is shown in Fig. 6. A peak in the absolute plate velocity from 110 to 100 Ma shown by HYBRF2008 and SFRF2010 is partly an artefact due to merging hotspot and paleomagnetic based reference frames at a time of significant true polar wander (i.e. when the whole solid Earth and mantle were tilted relative to the spin axis).

3.3. Snapshots of the African Plate from Jurassic to the Present

In the following we will present the evolution of the African plate as oceanic crust was accreted and subducted at its margins and discuss its interaction with the underlying mantle through time as illustrated by the absolute plate motion. We have chosen the time of reconstructions that were meaningful for important first-order changes within the continental area in an attempt to identify any connections between the evolution of distant plate boundaries and interior deformations.

3.3.1. Mid Jurassic (Bajocian, 170 Ma)

Around 170 Ma, the spreading rate doubled between NW Africa and North America, and the development of normal oceanic crust was established in the Central Atlantic (Labails et al., 2010). From the Central Atlantic, the plate boundary continued in the Alpine Tethys/Mediterranean domain along a transtensional/transcurrent fault south of Iberia along the northwestern tip of the African plate. The intra-Tethyan trench that was in the northeast and the transform

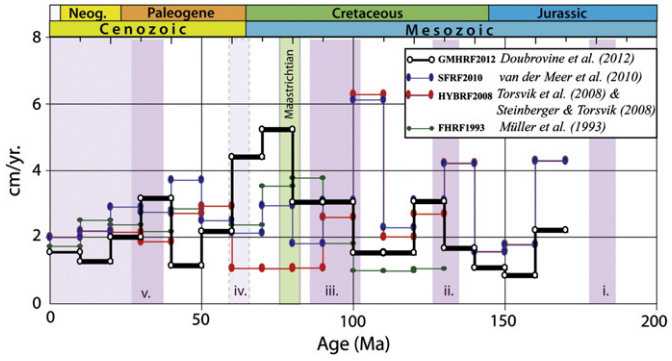


Fig. 6. Mean absolute plate motion of the African plate based on four global reference frames. Timing of major episodes of intra-plate volcanism is indicated by magenta boxes as following: i) Karoo LIP, ii) Parana-Etendeka LIP SE Atlantic, iii) Indian Ocean (Madagascar, Agulhas LIP), iv) SW Indian Ocean and SE Atlantic Ocean (Conrad Rise and Del Cano Rise, Meteor Rise and Discovery seamounts), and v) Afar related volcanism.

fault between the African plate and Iberia in the west are conceptually connected by another transtensional transform system along northern Greater Adria—a simplified plate boundary geometry following a more complex scenario postulated by Schmid et al. (2008).

In the present model, the mid ocean ridge between Greater Adria and the Pontides ceased shortly before 170 Ma. We have modelled the opening of the Neotethys northern branch as a symmetric seafloor spreading system and this assumption predicted the location of the extinct mid ocean ridge at 170 Ma approximately 600 km north of the Greater Adria margin. According to geological evidences (~130 Ma ophiolites which were obducted along the Greater Adria northern margin, e.g., Schmid et al., 2008), a new subduction zone was formed shortly afterward to the south of this extinct ridge, and we have placed the newly formed trench in the oceanic crust northeast of Greater Adria (Fig. 7, upper panels).

The plate boundary NE of the African plate (north of Arabia) is enigmatic. Collision of the Cimmerian blocks of Iran with Eurasia occurred in Late Triassic time, and if ocean spreading continued to the northeast of Arabia, as conceptually shown in Fig. 7, this requires subduction below the Iranian margin in the Jurassic time. We connect the intra-Tethyan ridge plate boundary north of Arabia through a transform to the trench north of Greater Adria, and another (relatively large) transform to the intracontinental plate boundary between the African plate and India/Madagascar/Antarctica/South America. Incipient seafloor spreading probably started between Madagascar/Antarctica and East Africa (see Section 2.3.3) at this time; also note that some of the intra-continental Karoo rifts (orange basin outlines in Fig. 7) situated in East Africa and SW Madagascar, entered the last phase of their development in the Toarcian–Aalenian (Geiger and Schweigert, 2005).

Along the SE and SW margins of South Africa, sedimentary basins were already formed in the Late Jurassic and Upper Jurassic north–south oriented grabens were described in offshore sedimentary basins located on the south-western African margin (e.g. Gerrard and Smith, 1982). South of this area, the Falkland island region behave as an independent microplate and may have rotated ca. 180° between Early and Mid Jurassic (Mitchell et al., 1986). Intra-continental rifting and margin extension that preceded the opening of the South Atlantic Ocean also started around end Jurassic time (for a review see Heine et al., 2013). Although a long-lived, well-established subduction zone along the western margin of South America may have acted as the plate boundary before breakup split South America and Africa along the South Atlantic ridge, we suggest that a diffuse plate boundary may have already been active in the same time between South America and Africa; that boundary would have linked the incipient seafloor spreading in the Mozambique–Riiser-Larsen Sea, transform faults along the Falkland plateau and the mid ocean ridge in the central Atlantic (Fig. 7 upper panels).

3.3.2. Mid Early Cretaceous (Hauterivian–Barremian, 130 Ma)

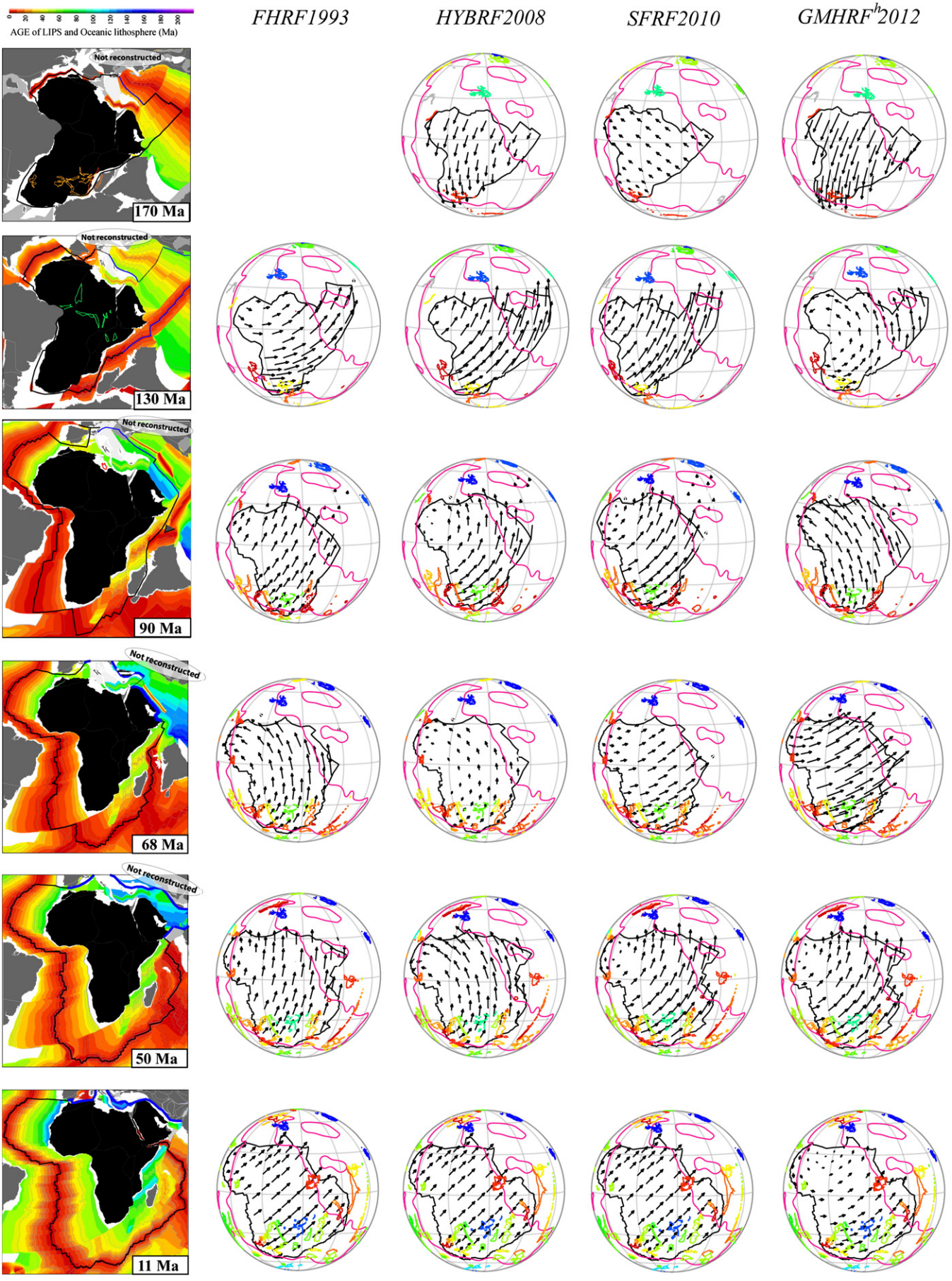
At this time a vigorous system of mid ocean ridges developed E and NE of Africa in the Mozambique/Riiser-Larsen Sea between the African plate and Antarctica, in the Somali Basin, between the African plate and Madagascar, and probably as a new intra-oceanic basin between two older regions of Tethys, as a result of Indian plate being rifted from the Antarctic plate and starting to move northward alongside the eastern Madagascar margin (Fig. 7). This continuous mid-ocean ridge was connected northward with the inferred spreading centre north of Arabia until shortly before 130 Ma. At Hauterivian–Barremian time ophiolites were emplaced along NE Greater Adria (Hellenides, Dinarides, e.g., Schmid et al., 2008), and possibly the northern distal margin of the Taurides (Çelik et al., 2011). Between Arabia and the Iran/Afghan blocks, subduction started in the Late Jurassic (Agard et al., 2011), and it remains unknown whether spreading continued within the Neotethys after this time. We conceptually show the plate boundary in the Early Cretaceous time as a northward subduction zone below the Eurasian margin, linked to the terminating obduction plate boundary in the Mediterranean region, by a north–south transform fault.

Northwest of the African plate, new oceanic crust was created between Iberia and Greater Adria mainly between 170 and ca. 150 Ma (see Fig. 5 and Section 2.3.4.2) and incipient seafloor spreading started in the southern South Atlantic (Moulin et al., 2010; Torsvik et al., 2009). The African plate interior started to be subjected to internal deformation, with rifting between the NW, N, NE and central part of the African continent (see Section 2.2, also for a review see Heine et al., 2013).

3.3.3. Late Cretaceous (Turonian–Coniacian, 90 Ma)

After Barremian time, seafloor spreading in the South Atlantic propagated northward and, by 100 Ma, connected with the Central Atlantic. Several million years later (probably around 96 Ma) it also connected to the mid ocean ridge south of the long Agulhas–Falkland fracture zone (AFFZ, Fig. 4). Seafloor spreading between East Africa and Madagascar ceased around 123 Ma (see Section 2.3.3) and the plate boundary relocated to the east, accommodating strike-slip motion between Madagascar and India. At the same time, spreading in the Mascarene Basin was probably initiated (Bernard and Munschy, 2000; Royer et al., 1992). This plate boundary continued northward as a mid ocean ridge or a transform fault accommodating transtensional motion between NW India/Kabul block and NE Africa/Arabia. In the Tethys domain, intra-oceanic north-eastward dipping subduction was probably initiated shortly before, at ca. 95 Ma, along the entire Tethyan domain and ophiolites created at this time were underthrusted by (i.e. obducted onto) continental crust of Arabia and eastern Greater Adria in Late Cretaceous time along the entire north-eastern margin of the African plate.

The Iberian plate acted independently from ~124 to 83 Ma and its plate boundaries accommodated Albian–Aptian counterclockwise rotation (Gong et al., 2008) creating a mixed regime of strike-slip, transpression and transtension north of Africa (Vissers and Meijer, 2012b). To the north, Greater Adria was separated from Eurasia by a southward dipping subduction zone in the Alpine region (Handy et al., 2010; Schmid, 2004). The Alpine segment connected with a transform fault to a northward dipping subduction zone in the Dinarides–Hellenic segment (Schmid et al., 2008; van Hinsbergen et al., 2005), and continued eastwards with the intra-oceanic subduction zone below the Cretaceous ophiolites mentioned above. This configuration of the African plate was setting the scene for the so-called the "Santonian event" (e.g. Guiraud and Bosworth, 1997) probably generated by far-field stresses induced by changes in plate boundaries (Bosworth et al., 1999); this event caused


AGE of LIPS and Oceanic lithosphere (Ma)
0 20 40 60 80 100 120 140 160 180 200
FHRF1993
HYBRF2008
SFRF2010
GMHRF^h2012
Not reconstructed
170 Ma
130 Ma
90 Ma
68 Ma
50 Ma
11 Ma

internal deformation of the northern African and Arabian continental area.

3.3.4. Late Cretaceous (Maastrichtian, 68 Ma)

In the Santonian (around 84 Ma) and Maastrichtian (ca. 68 Ma) the northern margin of Africa and Arabia were affected by two strong tectonic events. In the Maastrichtian, compressional phases were recorded from Morocco to Oman (Guiraud and Bosworth, 1999) and in SW Africa (Andreoli et al., 2009) and extensional regimes existed in central, southern and northern Africa (Viola et al., 2012) although no major intracontinental plate boundaries developed. The main changes in the African plate boundaries occurred in the northern part along the Arabian margin where ophiolite obduction marked the cessation of the intra-Neotethyan subduction zone that started ~95–90 Ma ago. This plate boundary was linked, probably through a transform system, to the neighboring trench along northern Greater Adria (Fig. 7), where (probably continental) subduction continued forming high-pressure belts in central Anatolia (e.g., Pourteau et al., 2010). The Hellenic–Dinaric region underwent continuous northward subduction of Greater Adriatic continental lithosphere (van Hinsbergen et al., 2005). Another set of transform faults south and east of Iberia linked the Central Atlantic mid-ocean ridge to the trench situated north of Greater Adria in the Alpine region. Africa–Europe convergence in the western Mediterranean was still mostly accommodated in the Pyrenees along the northern Iberian margin, with the Africa–Iberia boundary being mostly transcurrent (Vissers and Meijer, 2012b).

3.3.5. Eocene (Ypresian, 50 Ma)

After the Late Cretaceous ophiolite emplacement along the northern Arabian margin, the subduction zone that had continuously been active along the Iranian margin since the Jurassic, became the plate boundary of the African plate (Fig. 7). The northern branch of the Neotethys had completely subducted below the Pontides, except perhaps in eastern Anatolia (Sengor et al., 2008). Continental lithosphere of Greater Adria had been subducting from the Late Cretaceous time in the Aegean and Anatolian regions until the Late Eocene (Anatolia) and Miocene (Aegean) time, respectively (van Hinsbergen and Schmid, 2012; van Hinsbergen et al., 2005, 2010). Along the Dinarides–southern Alps region, continental underthrusting continued until the Present (e.g. Ustaszewski et al., 2008). The Early Eocene plate boundary in this region is thus an intra-continental trench located in the Tauride fold-thrust belt, the Pindos basin and the Dinarides fold-thrust belt. Through the southern Alps, the plate boundary links to a NW-ward dipping subduction zone along the south-western Iberian margin, with a change in subduction polarity in the Alps–Corsica/Sardegna–Pyrenean junction (van Hinsbergen et al., in review). The pole of Africa–Iberia rotation is located just southwest of Gibraltar, and minor (several tens of km) of convergence occurred between the African plate and Gibraltar in the Eocene (Vissers and Meijer, 2012a). To the west, the plate boundary connected to the Gibraltar Fracture zone.

In the Indian Ocean, the mid-ocean ridge between Madagascar and India became gradually extinct in the Mascarene Basin (Fig. 4) at about 59 Ma, when it moved northward, between the Seychelles and India (Ganerød et al., 2011). This ridge was probably connected northward to the subduction zone south of Eurasia by a transform fault—a precursor of the Owen fracture zone (OFZ, Fig. 4) with a transpressional component reflected by the final thrusting of ophiolites over the western Indian (Allemann, 1979; Gnos et al., 1997; Khan et al., 2007; Mahmood et al., 1995). In the South Atlantic, a ridge jump toward the South American plate around 60 Ma (Hartnady and le Roex, 1985) transferred South American oceanic crust to the African plate. This ridge jump ended the activity of the long Agulhas–Falkland Fracture Zone (AFFZ), a feature that accommodated strike-slip motion between Africa and South America for about 60 million years (Barker, 1979).

3.3.6. Miocene (Tortonian, 11 Ma)

With the westward propagation of the Central Indian Ridge (CIR, Fig. 4) into the Gulf of Aden (GA, Fig. 1), a considerable amount of African continental lithosphere started to be rifted away to become a separate plate—the Arabian plate, whereby the northern segment of the Dead Sea Fault (DSF, Fig. 4) is only Pliocene in age and may be considered intraplate deformation rather than a fully developed plate boundary (e.g., Searle et al., 2010). The Dead Sea fault connects to the Cyprus subduction zone and the East Anatolian Fault (CSZ-AF, Fig. 4), along which the African plate boundary continues westward (Fig. 7).

Most of Greater Adria in the eastern Mediterranean have been subducted leaving only its upper crust as part of the Dinaride–Hellenide–Tauride belt (e.g., Handy et al., 2010; Jolivet and Brun, 2010; van Hinsbergen et al., 2005, 2010); subduction of old oceanic crust from the Eastern Mediterranean Sea commenced in the Oligocene–Miocene. The last remaining parts of old oceanic crust of the Alpine Tethys started to be subducted below the Corsica/Sardegna block, followed by retreat of the Calabrian subduction zone towards Greater Adria opening the Tyrrhenian Sea (Cifelli et al., 2007). European crust had collided with North Africa in middle Miocene time in the Kabylides (Michard et al., 2006). This was followed by slab break-off and development of subduction transform faults (STEP) faults along the northern African margin, accommodating westward retreat of the Gibraltar subduction zone and eastward retreat of the Calabrian subduction zone (e.g., Govers and Wortel, 2005; Jolivet et al., 2009; Spakman and Wortel, 2004).

4. The African plate architecture, plate boundary forces and plate–mantle interactions

4.1. African plate boundaries since the Jurassic

Quantification of regional and global plate-driving forces has been used as a method to predict or unravel mechanisms that lead to plate motions or changes in plate motions (e.g. Forsyth and Uyeda, 1975; Jurdy and Stefanick, 1991). Ridge push and slab pull are the main forces that act upon tectonic plates, while oceanic crust is accreting at mid-ocean ridges and is subducting along trenches. In addition, due to the motion of the plate relative to the underlying mantle, mantle drag is also contributing to the plate driving forces. It has been postulated that the direction of the mantle drag vectors illustrates how the tectonic plate moves relative to the mantle, which results from the hypothesis that the plate is the upper, rigid layer of the convective cell. However, it has been suggested that this is not the case for all tectonic plates, since the mantle drag direction and magnitude may depend on the size and nature (i.e. continental vs. oceanic) of the tectonic plate (Artemieva and Mooney, 2002). Motion along transform faults is also regarded as a resistive force, but its magnitude is rather low compared to other driving forces (Forsyth and Uyeda, 1975). Assessment of the nature and lengths of each type of plate boundary may provide a first order estimate of forces exerted along plate margins (Faccenna et al., 2012; Forsyth and Uyeda, 1975; Meijer and Wortel, 1999).

Fig. 7. Circum-African oceanic paleo-age grids (left column), plate boundaries, and absolute plate motions in four different global reference frames in columns 2, 3 and 4 from the left (FHRF1993—Fixed Hotspot Reference Frame, HYBRF2008—Hybrid, SFRF2010—Subduction calibrated, GMHRF^{h}2012—Global Moving Hotspot, see text for references). Large igneous provinces are colored according to approximate age of eruption using the age grid color palette, plate boundaries show mid-ocean ridges and fracture zones as black lines and trenches in blue. Black arrows show direction of absolute plate motion for the African plate at 2° × 2° grid. The zero contour of the Large Low Shear Velocity that presumably maps the extent of a thermochemical pile situated at the core–mantle boundary (Torsvik et al., 2006) helps to illustrate how the position of the African plate relative to a fixed reference frame (the core mantle boundary) changed through time.

We have used the new isochron and rotation model and resulting palaeo-age grids to estimate the length of different plate boundaries around the African plate, the amount of oceanic area, and the mean oceanic crust age. Fig. 8 and Table 3 present the African plate geometry for the six time intervals presented in Section 3.3. showing the computed length of three types of plate boundaries (mid ocean ridges—MOR, transform faults and fracture zones—TF and trenches-TR), plate area, amount of oceanic crust, mean age of oceanic crust and mean absolute velocity in the GMHRF^{h}12 reference frame. The lengths of MOR and TF have been calculated based on this study's isochrons using the method of Seton et al. (2009) and are mostly based on preserved oceanic crust and therefore plate boundaries. The length of trenches is an estimate based on plate boundary geometry from this study. Consequently, the accuracy of calculation is poorer than in the case of MOR and TF plate boundaries. We have not computed these values for our oldest time frame (170 Ma) because besides the older oceanic crust in the Neotethys, only a small amount of oceanic crust was formed within the African plate (in the Central Atlantic) and the errors related to these (mostly inferred) plate boundaries are too large. Instead, we chose the 140 Ma reconstruction to illustrate the geometry of the African plate at the time when several Jurassic oceanic basins were already well established, but still before 130 Ma, when major changes in plate boundaries commenced. Analyzing plate boundaries of present day tectonic plate configuration Forsyth and Uyeda (1975) found no clear correlations between global tectonic plate average velocities and total plate area, continental area, or length of mid-ocean ridges or transform faults. They suggested that the only factor that seemed to have contributed to the average absolute velocity was the length of trenches. In our study we observe that a considerable change in the type of plate boundaries around the African plate from extensional (mid ocean ridges) to compressional (possible intra-oceanic subduction in the northeastern part) did not resulted in a change in absolute plate motion. On the other hand, the highest mean absolute velocity computed here (at around 70 Ma) coincides with a sharp decrease in the mean oceanic crust age which is probably due to the removal of large part of older oceanic crust in the Tethys domain between 90 and 68 Ma by subduction within the Neotethys margins. An increase in the mean oceanic age also corresponds to a rapid decrease in the mean absolute velocity at around 50 Ma (Table 3). We therefore suggest that, in the case of the African plate, the oceanic crust mean age variations play a role in the plate–mantle interaction.

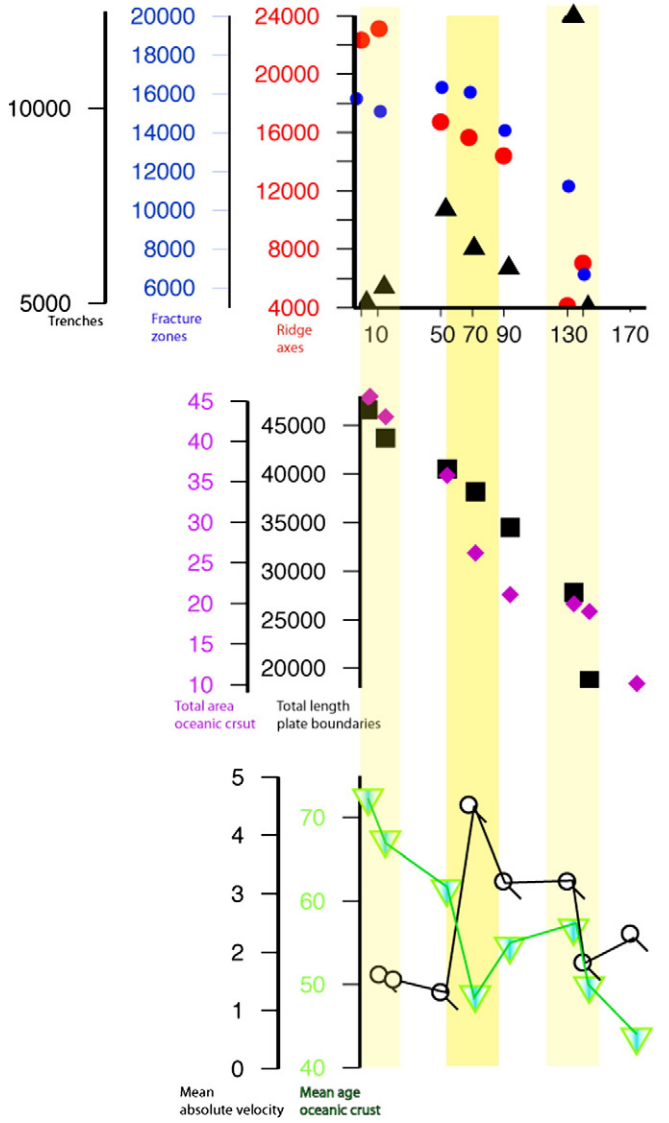


Fig. 8. The African plate geometry at selected times: total length of active ridge segments, fracture zones, trenches, total length of plate boundaries, total area of oceanic crust, mean age of oceanic crust and man absolute plate velocity.

4.2. The African plate absolute motion and intra-plate volcanism

The absolute motion of the African plate was computed for our selected time frames using the four global reference frames described in Section 3.2. (Figs. 6 and 7). Here we will briefly present the main magmatic events that were observed on the African plate for times younger than 200 million years and attempt to establish whether there are any correlations in time with changes in the absolute plate motion of this plate.

After the Central Atlantic Magmatic province (CAMP)—a ca. 200 Ma LIP that was formed shortly before breakup and incipient seafloor spreading in the Central Atlantic (Marzoli et al., 1999)—the African plate was disturbed and modified by extensive volcanic activity in the following time-intervals: i) in the Early Jurassic mostly in the southern Africa by the so-called Karoo event dated from ca.177 to 183 Ma (Jourdan et al., 2005); ii) in the Early Cretaceous, before breakup and early seafloor spreading in the South Atlantic—when Parana-Etendeka Large Igneous Province was formed in Namibia and northern South Africa, iii) in the Late Cretaceous, when the Agulhas LIP formed in the SW Indian ocean and in the SE Atlantic ocean (Gohl et al., 2011), but also manifested in the Central Atlantic by the Sierra Leone Rise volcanism (Kumar, 1979); iv) in the Early Cenozoic in the SE Atlantic ocean (south of Agulhas fracture zone, and v) from the Eocene/Miocene to recent as the Afar volcanism in the NE African continent (Figs. 4 and 6). Most of these large magmatic events are linked to mantle plumes (Coffin and Eldholm, 1994), most likely sourced from the core–mantle boundary (Burke and Torsvik, 2004). As the ascent of mantle plumes through the mantle and their impingement at the base of the lithosphere may induce a change in the motion between the tectonic plate and underlying mantle, large scale volcanism and absolute plate motion may be linked, at least at the inception of the plume–lithosphere interaction or emplacement of large igneous provinces.

According to the GMHRF^{h}2012 model, large volcanic events coincide with an increase in the absolute plate velocity at ca. 130 Ma and 100 Ma. The less-well constrained magmatic events in the Early Cenozoic may have occurred at the time of the African plate slowing down, and the same seems to be the case for the Eocene/Miocene Afar plume activity. The absolute plate motion may be also influenced by changes in the subduction regime, particularly in the event of a large slab break-off. Interestingly, the most noticeable deviation from a slow motion (1–4 cm/yr) relative to the mantle is shown by the GMHRF^{h}2012 model for the Late Cretaceous time (80 to 60 Ma) (Fig. 6). For that time severe continental internal deformation were described by (Guiraud and Bosworth (1997) and Guiraud et al. (2005) as the "Santonian event" (also see Section 2.2).

Table 3
The African Plate geometry at selected Lines. RI TOT-total ridge lenght, FZ TOT-total fracture zones and transform faults, TR TOT-total lenghts of trenchy, PBTOT-total Length of plate boundaries.

Age (Ma)	RI TOT (km)	FZ TOT (km)	TR TOT (km)	PB TOT (km)	Area ocean (km^2)	Mean age (Ma)	Mean vel (cm/yr)	Mean vel $\pm 1\ \sigma$
0	22,322 (48%)	19,127 (41%)	5117 (11%)	46,566	44,865,798	72.1		
11	23,098 (53%)	15,057 (34.5%)	5500 (12.5%)	43,655	42,440,661	67.27	1.5	0.3
50	16,729 (41.3%)	16,284 (40.2%)	7500 (18.5%)	40,513	35,198,749	61.42	1.2	0.1
68	15,674 (41%)	16,029 (42%)	6500 (17%)	38,203	25,608,553	48.68	4.4	0.8
90	14,420 (42%)	14,054 (40.5%)	6000 (17.5%)	34,474	20,509,182	54.6	3.1	0.5
130	4131 (15%)	11,200 (40%)	12,500 (45%)	27,831	19,328,471	56.8	3.1	1.3
140	7049 (37.7%)	6674 (36.3%)	5000 (26%)	18,723	18,362,285	49.8	1.7	0.67
170					9,528,510	43.6	2.2	0.72

Apart from the occurrences of massive volcanism probably associated with plume activity, smaller-scale volcanic events may have also been related to changes in absolute plate motion. For example, in the Benue Trough (Fig. 3) region situated in West Africa, age and orientation of regional dykes point to three periods of magmatic activities: (1) Late Jurassic to Albian (147–106 Ma); (2) Cenomanian to Santonian (97–81 Ma), and (3) late Maastrichtian to Eocene (68–49 Ma) (Maluski et al., 1995). The volcanic activity has been linked to various stages of South Atlantic opening (Maluski et al., 1995), but we note here that the intervals of CCW rotation of the African plate relative to the mantle also coincide with magmatic outbursts in the Benue Trough. At 130 Ma, all global reference models show a counter clockwise (CCW) rotational like in absolute motion, which in the GMHR^{h}2012 reference frame has a pole situated in the Benue Trough region. The same CCW rotation or SW–NE motion is observed for the Mid-Late Cretaceous time (90 Ma), but with a rotation pole NW (HYB2008) or SW (GMHR^{h}2012) of Benue Trough, and also for the Late Cretaceous (68 Ma) for certain reference frames (FHRF1993 and HYB2010), but changes to a more WSW–ENE in SFRF2010 and GMHR^{h}2012. We suggest that lithospheric stresses triggered by changes in absolute plate motion might have also contributed to the weakening of the crust in the central African continent at least at the time when the Central African Rift started to develop in the Cretaceous, as evidenced by dyke emplacement.

4.3. The African plate palaeo-stresses—an example of plate–mantle interaction in the Maastrichtian

One of the goals of our African plate study was to quantify how the evolution of this plate (by oceanic crust accretion and subduction and interaction with the underlying mantle) affected its continental interior through time. Large-scale rifting within the African continent produced diffuse plate boundaries from the Triassic to Early Jurassic (the Karoo system), in the Mid and Late Cretaceous and (Central African Rifts and older East African Rifts), and from the Oligocene to present day (East African Rift system, see also Section 2). No major tectonic movements between the main African blocks were recorded in the Campanian–Maastrichtian, a relatively quiet interval that followed the pervasive compressive event registered in the Santonian time, especially in northern Africa (Guiraud and Bosworth, 1997). However, changes in the stress regime at that time have been documented in the African interior in areas situated far from active plate boundaries (Bosworth et al., 1999; Viola et al., 2012).

We have chosen a time slice in the Campanian–Maastrichtian time interval (at 68 Ma that is close to the identified marine magnetic anomaly C31ny) for modelling the stress regime of the African plate as a case study that employs the new relative and absolute African plate model and its geometry at the time when 1) no internal plate boundaries were active, 2) the topography of the continental Africa was relatively flat as it precedes the arrival of the Afar plume at around 30 Ma (Burke and Gunnell, 2008), and 3) a change in stress regime was registered in regions situated at large distances from active plate boundaries. In addition, the availability of new data and studies on paleo-stresses in the SW Africa for that time interval (Viola et al., 2012) can be used to ground-truth the results of the paleo-stress model presented here.

4.3.1. Simplified model of the African lithosphere at 68 Ma

Following Steinberger et al. (2001), we constructed a simple model of the African plate lithosphere at 68 Ma to evaluate a stress pattern. The isostatically balanced lithosphere model is subjected to three main mechanisms: (1) lateral variations of the gravitational potential energy (GPE) within the plate (due to variations of lithospheric density structure) and along the boundaries (ridge push due to hot asthenosphere rising), (2) basal drag from the convecting mantle, and (3) pull from the subducting Tethys oceanic lithosphere.

This simple lithosphere model includes a crust of uniform density and a mantle lithosphere described by a half-space cooling model. Our new oceanic paleoage grid (Fig. 7) has been therefore converted into a simplified palaeobathymetric grid by using the Parsons and Sclater (1977) age-dependent thermal subsidence relationship and assuming that the oceanic crust is 7 km thick. In our simple model the continents have a uniform thickness of 37 km (isostatically balanced to have a uniform topography 300 m above sea-level for the continental area) and the continental mantle lithosphere is considered equivalent to 100 Ma old oceanic lithosphere. Thus, even though GPE is uniform within the continental part of the plate, variations of GPE across continent–ocean boundary and within the oceanic part cause non-trivial distribution of stresses within the entire plate.

The basal drag from the convecting mantle was derived from the convection velocity field at 100 km depth reconstructed back to 68 Ma (Fig. 9) using the technique described in Steinberger and Calderwood (2006). Subduction was modelled by setting outward velocity boundary condition along the northern and north-eastern boundary of our plate model (all other segments of the model domain are pinned).

This model significantly differs from a model used in the modern stresses analysis. Apart from the three mechanisms mentioned above, lithospheric stresses strongly depend on topographical variations, crust heterogeneities, variations of sub-crustal mantle density, and rheological variations of the lithosphere (e.g. Bird et al., 2006; Gaina et al., 2010b; Stamps et al., 2010). Whereas those parameters can be estimated for the present-day African plate with some level of confidence, reconstruction of these variables for geological time is more difficult. A complex model that analyses the African plate at present day whose outcome can reproduce 60% of the observed present day stress orientations and 80% of the observed stress regimes was briefly presented by Gaina et al. (2010a). Some of the present-day constraints and parameter optimization will be used to improve the palaeo-stress modelling presented in this study. This approach cannot guarantee that the predicted results will match the observed palaeo-stress pattern, but give the confidence that the correct type of mechanisms (and associated parameters) were taken into account in modelling.

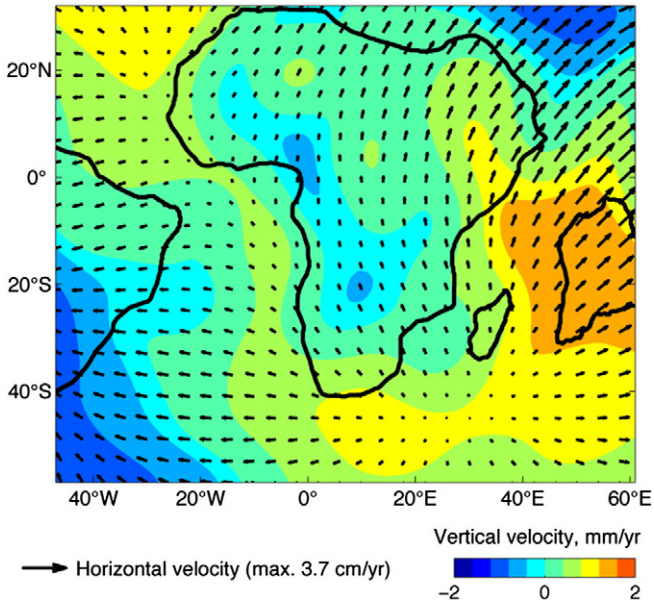


Fig. 9. Velocities of mantle flow at 100 km depth for the African plate at 68 Ma (model by B. Steinberger using Steinberger and Calderwood's (2006) technique). These velocities are used to calculate the basal drag in the palaeo-stress modelling for Maastrichtian time (see Section 4).

4.3.2. Modelling technique: the ProShell approach

The calculations were performed using the in-house numerical suite ProShell. Analysis of stresses within such a large plate requires taking into account the lithosphere non-flatness and the potential interaction between in-plane and bending deformation. ProShell is designed to analyze stresses and deformation in the lithosphere which can be described as a thin shell of arbitrary shape and treats a shell as a set of finite elements where each element is being considered as a flat (shear-deformable) plate with specific orientation in space (Kwon and Bang, 2000). The different orientation of individual elements may cause transformation of bending in one element into in-plane deformation of the neighboring element. Thus, the shell element accounts for balance of both, moments (bending) and in-plane force. The calculation of stresses in the lithospheric plate follows the Earth's global curvature and the local curvature associated with large perturbations of topography/bathymetry. The numerical approach was also optimized using the techniques described in Dabrowski et al. (2008).

For a given structure of the continental and oceanic crust, this program computes the isostatic position of the lithosphere and calculates the gravitational potential energy (GPE). The traction at the base of the lithosphere is modelled using the approach of Ellis et al. (1995). The maximum density variations within the lithospheric mantle (96 kg/m^3), the thermal correction of the mantle lithosphere due to crustal thickness variations (9 °C/km), and parameters of intensity of the basal drag such as viscosity of asthenosphere (2×10^{21} Pa · s) were taken from our modern stress analysis study (Gaina et al., 2010b) which was tested against the observed patterns of lithospheric stress (Bird et al., 2006; Bosworth et al., 2008; Delvaux and Barth, 2010; Heidbach et al., 2008).

4.3.3. Modelling results

To understand the influence of each mechanism contributing to palaeo-African stresses, we apply them individually and analyze the resulting stress patterns. The calculations were performed on a regular finite-element grid of approximately 450 × 500 linear elements, resulting in the system with approximately 1 million degrees of freedom.

Firstly, we analyzed the stresses resulting from the lateral variations of GPE. The assumption of low topographical variations at 68 Ma for continental Africa precludes analysis of the intra-continental variations of GPE, and thus this set of calculations presents the influence of push from mid-oceanic ridges only (Fig. 10). The stress regime map shows that the entire African continent is in a compressional regime due to the surrounding ridge push forces.

In Model 2 (Fig. 10) we simulated the influence of mantle flow drag by adding a basal traction term. Following Ellis et al. (1995) we assume that the lithosphere is affected by traction from Quette-like flow. This flow is controlled by the difference between the lithosphere and convecting mantle velocities (calculated at 100 km depth using the approach of Steinberger and Calderwood, 2006) and by the viscosity of the asthenosphere (taken as ~2×10^{21} Pa · s). The effect of this mechanism seems to be a severe change in the orientation of most compressive stresses in the northern and south-eastern parts of the African continent and the creation of an extensional stress regime in north-east Africa.

The two previous models used no-slip boundary conditions for velocities. Model 3, in contrast, sets outward directed boundary velocities in the middle part of the northern domain boundary. That simulates slab pull, associated with subduction of the old Tethys oceanic lithosphere (Fig. 10). We applied a simplified kinematic boundary condition and vary the magnitude of the subduction velocity from low (velocity of 1–3 cm/yr, Model 3a) to moderate values (4–7 cm/yr, Model 3b). Those variations, however, do not produce significant changes in the stress pattern (Fig. 10), illustrating the robustness of the model to rate of subduction uncertainties. Adding this mechanism does not lead to large changes in the stress field orientation, but significantly alters the stress regimes of Northern Africa (Fig. 10). The extensional regime seems to affect the entire NE Africa and its offshore part, being bordered by regions of strike-slip.

Our first order results correspond relatively well to the palaeo-stress regimes inferred from geological studies. A recent dataset of palaeo-stress indicators including measured striated fault planes was collected in the SW Africa (Namaqualand) and modelled into palaeo-stress tensors that suggest a compressional regime in the Maastrichtian (Viola et al., 2012). This timing is constrained by the dated markers situated close to the measured faulting directions and matches our results. In the northern and north-eastern part of Africa, after the compressional Santonian/Senonian event, NW–SE rifting recommenced in parts of the northern African–Arabian margin in the early Campanian and continued to the late Maastrichtian (Guiraud and Bosworth, 1997). Large NW–SE trending rift basins (Fig. 3c) were rejuvenated during Late Senonian–Paleocene times and magmatic activity occurred along the E–W trending Nubian fault swarm of southern Egypt–northern Sudan (Wilson and Guiraud, 1992). The palaeo-stress regime predicted by our model is consistent with these geological evidences that suggested an extensional/strike-slip regime in the NE and N of the African plate. Further modelling of the African plate palaeo-stresses for relevant geological time frames will be presented elsewhere (Medvedev et al., in prep).

5. Summary and conclusions

In this contribution we present a systematic study of the African plate boundaries since the opening of surrounding oceanic basins, starting at 195 Ma in the Central Atlantic, that aids further understanding of the geodynamics of the African plate. The accretion and subduction of oceanic crust around the African continent is described in the light of available geophysical and geological data and published models.

The geometry and location of the African plate in time and space are analyzed for several time frames that correspond to episodes of intra-continental deformation. The evolution of plate boundaries and lithosphere–mantle connections revealed by the absolute motion of the African plate are discussed in the attempt to explore a correlation among these and far-field stresses that may have influenced the African continental interior.

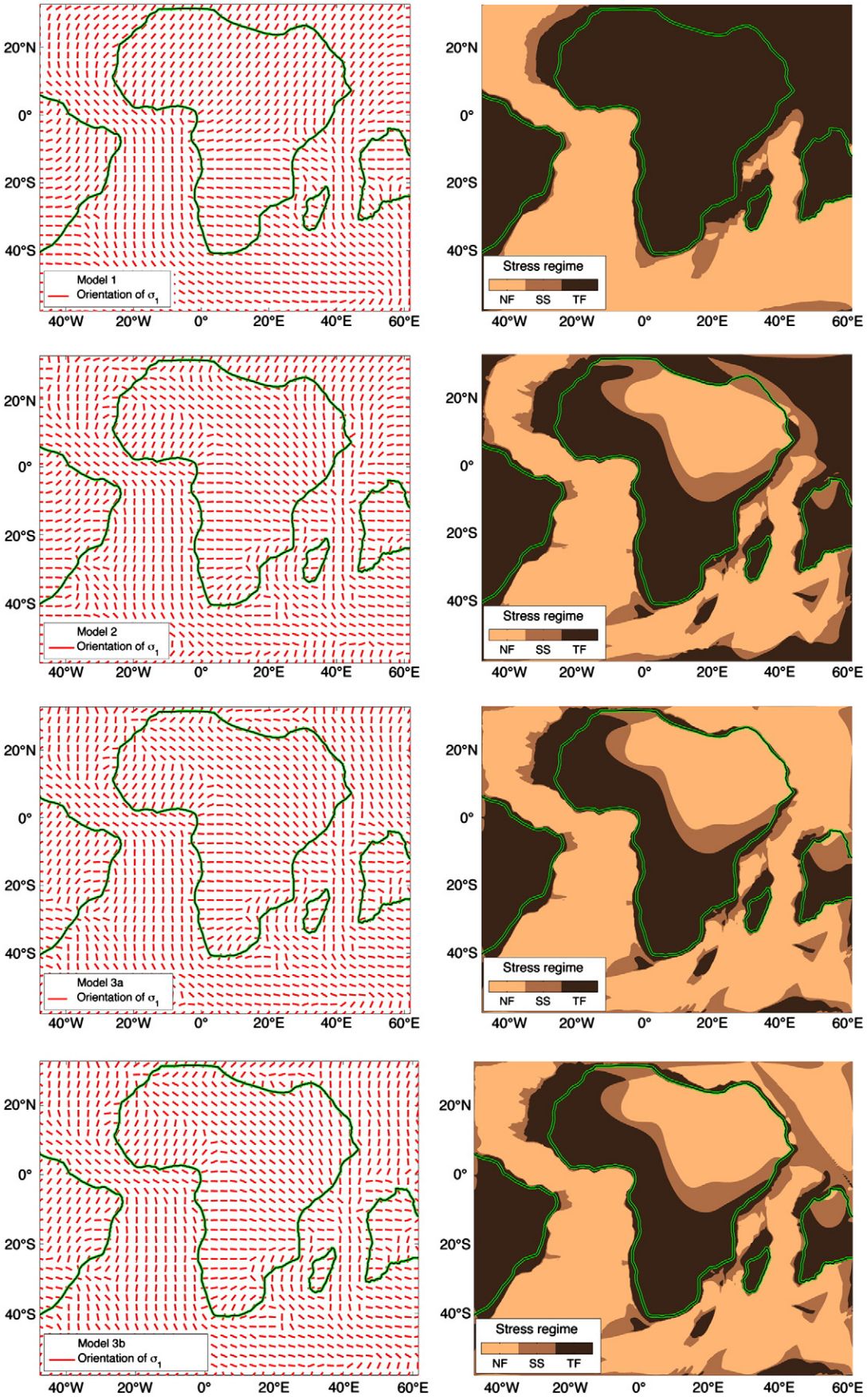


Fig. 10. Palaeo-stress modelling results: orientation of maximum horizontal compressive stress (left panels) and stress regime (right panels) for three models, (1) base model (top panel), (2) base and basal drag model (middle panel), and (3a, b) base, basal drag and subduction along the northern and northeastern borders using a subduction velocity of 7 cm/yr and 2 cm/yr (lower two panels). The regimes were calculated using techniques of Delvaux et al. (1997); NF is normal faulting, SS is strike-slip, and TF is the thrust faulting regime.

We observe that changes in the lengths and type of plate boundaries do not correlate with variations in the absolute plate velocities as suggested in studies based on today's plate configuration (Forsyth and Uyeda, 1975), but variations in the mean age of oceanic crust seem to be linked to changes in the mean absolute velocities. A counter-clockwise rotation of the African plate relative to the mantle around a pole situated in the Benue Trough or its vicinity seems to coincide to episodes of volcanic activities described in that region and rifting in the Central and Northern Africa.

A simple model of the African plate lithosphere at 68 Ma was used to calculate palaeo-stress regimes taking into account the influence of ridge push, slab pull and mantle drag forces. This model predicts a compressional regime in South Africa and an extension and strike-slip in NE and N Africa. These predictions are in accordance with the computed palaeo-stress tensors based on geological evidences in south-western Africa (Viola et al., 2012) and agree with published models on the geological evolution of the Northern part of the African continent (e.g. Guiraud and Bosworth, 1999).

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