Ridge-push force and the state of stress in the Nubia-Somalia plate system

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ABSTRACT

We assessed the relative contribution of ridge-push forces to the stress state of the Nubia-Somalia plate system by comparing ridge-push forces with lithospheric strength in the oceanic part of the plate, based on estimates from plate cooling and rheological models. The ridge-push forces were derived from the thermal state of the oceanic lithosphere, seafloor depth, and crustal age data. The results of the comparison show that the magnitude of the ridge-push forces is less than the integrated strength of the oceanic part of the plate. This implies that the oceanic part of the plate is very little deformed; thus, the ridge-push forces may be compensated by significant strain rates outside the oceanic parts of the plate. We used an elastic finite element analysis of geoid gradients of the upper mantle to evaluate stresses associated with the gravitational potential energy of the surrounding ridges and show that these stresses may be transmitted through the oceanic part of the plate, with little modulation in magnitude, before reaching the continental regions. We therefore conclude that the present-day stress fields in continental Africa can be viewed as the product of the gravitational potential energy of the ridge ensemble surrounding the plate in conjunction with lateral variations in lithospheric structure within the continent regions.

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INTRODUCTION

The excess potential energy of a mid-ocean ridge generates a ridge-push force generally thought to be an order of magnitude less than the slab-pull force (Forsyth and Uyeda, 1975; Lithgow-Bertelloni and Richards, 1998; Conrad and Lithgow-Bertelloni, 2004; Schellart, 2004; Wessel and Müller, 2007; Faccena et al., 2012; van Summeren et al., 2012). While viscous and frictional resistances balance the slab-pull, basal shear traction resists the ridge-push force (Forsyth and Uyeda, 1975; Doglioni, 1990). Numerical and experimental studies have shown that basal shear tractions can be as dominant as slab-pull in driving plate motion (Rucker and Bird, 2007; Barba et al., 2008). In terms of the net slab-pull acting on the plates, almost 8%–12% of the slab-pull is used to pull the surface plate into the mantle, with the remainder split among driving mantle flow, bending the plate, and balancing the viscous resistance of the mantle (Schellart, 2004). However, the relative importance and contribution of the ridge-push force in assisting plate motion are unknown (Richardson, 1992). It is also not well understood whether this force can be transmitted in an oceanic plate and assist plate motions, or if it is dissipated in the form of intraplate deformation. The transmission of ridge-push force in an oceanic plate can be assessed, as a first-order analysis, by comparing the ridge-push force to the total strength of oceanic lithosphere, provided that the influences of other sources of tectonic stresses (basal shear traction, far-field forces, and stress associated with gravitational potential energy of elevated regions) are negligible.

The Nubia-Somalia plate system provides an ideal location in which to evaluate the ambient lithospheric state of stress in a continental plate given its unique boundary geometry of being nearly completely surrounded by mid-ocean ridges (Fig. 1). Evaluation of the state of stress in the Nubia-Somalia plate system is not complicated by boundary forces and is the product of upper mantle and lithospheric density forces arising from a combination of intraplate tectonic forces (e.g., ridge-push from the cooling oceanic lithosphere; see detailed discussions in Coblentz and Sandiford, 1994; Stamps et al., 2010) and basal tractions (e.g., upper-mantle convection; see discussion in Lithgow-Bertelloni and Silver, 1998; Moucha and Forte, 2011; Stamps et al., 2014). There is indication in mantle convection models that much of Africa's tectonic evolution (including the present-day high topography of the East African Rift and the South African plateau) is driven by buoyancy forces in the mantle (Lithgow-Bertelloni and Silver, 1998; Gurnis et al., 2000; Stamps et al., 2010, 2014; Moucha and Forte, 2011; Ghosh and Holt, 2012; Ghosh et al., 2013).

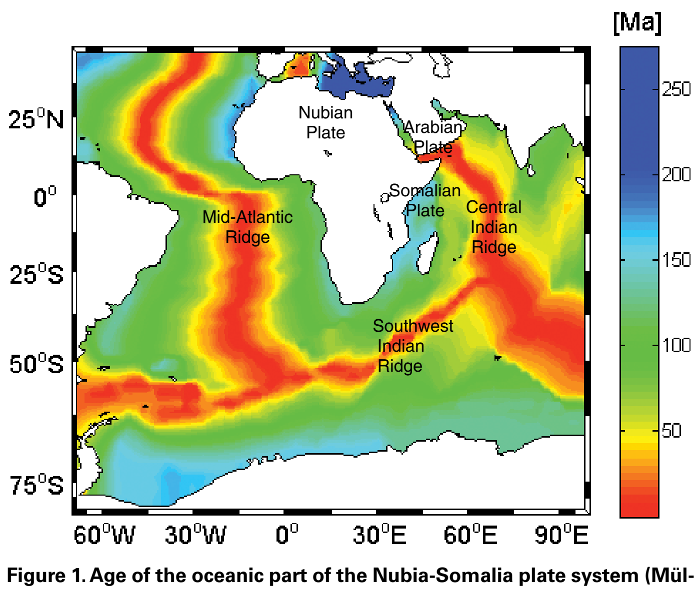

Figure 1. Age of the oceanic part of the Nubia-Somalia plate system (Müller et al., 2008).

In this study, we examine the relation between the present-day lithospheric strength and ridge-push transmission in the oceanic part of the Nubia-Somalia plate system. To assess the tectonic response of the oceanic part of the plate to the ridge-push–related stresses, the lithospheric strength and ridge-push force are compared. We estimated the strength of the lithosphere using a plate cooling model and relevant rheology, and we derived the ridge-push force from thermal state, crustal age, and seafloor depth data. The thermal structure of oceanic plates is constrained by bathymetry, surface heat flow, and geoid height data (Parsons and Sclater, 1977; Stein and Stein, 1992).

METHODOLOGY

Thermal State of Oceanic Lithosphere

Following the mathematical development of thermal models of oceanic lithosphere, a significant advancement was made to constrain the basal temperature and asymptotic thickness of oceanic plates using observed seafloor depth, surface heat flow, and geoid height data (cf. e.g., detailed discussions in Parsons and Sclater, 1977; Stein and Stein, 1992; McKenzie et al., 2005; Grose and Afonso, 2013). The thermal state of oceanic lithosphere mainly depends on age (Stein and Stein, 1992; McKenzie et al., 2005). We used physical parameters from the Global Depth and Heat-Flow Model (GDH1) and crustal age data to estimate the thermal structure of the oceanic lithosphere (Stein and Stein, 1992; Muller et al., 2008). The GDH1 model is well constrained by heat flow, bathymetry, and geoid height data (Stein and Stein, 1992). The temperature distributions T in the oceanic lithosphere were predicted using the plate cooling model and parameters in GDH1 (Eq. 1):

$$T = T_o + \left(T_b - T_o\right)\left[\frac{z}{H} + \frac{2}{\pi}\sum_{j=1}^{\infty}\frac{1}{j}\exp\left(-\frac{kj^2\pi^2 t}{H^2}\right)\sin\left(\frac{j\pi z}{H}\right)\right], \qquad (1)$$

where t is the age of the oceanic lithosphere, z is depth, H is the basal thickness of the lithosphere, k is thermal diffusivity, and T_o and T_b are the surface and basal temperatures, respectively (Carslaw and Jaeger, 1959, p. 100). Table 1 shows the parameters used to compute the thermal structure of the oceanic lithosphere. The crustal age data (Fig. 1) were taken from Muller et al. (2008) and have two arc-minutes resolution. The value of the basal temperature used for this study is based on the GDH1 model.

Lithospheric Geoid and Ridge-Push Force

The geoid anomaly of a thermally compensated oceanic lithosphere and the ridge-push force are linearly related to the dipole moment of density distributions (Haxby and Turcotte, 1978; Parsons and Richter, 1980). It is, therefore, theoretically possible to calculate the ridge-push force from observed geoid height data. This approach requires that the isostatic geoid anomaly of a cooling oceanic lithosphere be extracted from the total observed geoid before the ridge-push force can be determined. However, isolation of the isostatic geoid from the total observed geoid remains problematic. Several approaches have been explored to extract the isostatic geoid anomaly from the cooling oceanic lithosphere model. DeLaughter et al. (1999) used observed geoid signals between wavelengths of 2800 and 1000 km to predict the isostatic geoid anomaly of a cooling oceanic lithosphere. This corresponds to the spherical harmonic degrees of 14 and 39, respectively. The results of this analysis, however, depend on the input geoid model. Sandwell and Schubert (1980) used geoid slope (gradient of the geoid with increasing crustal age) to suppress long-wavelength anomalies and enhance the isostatic part of the geoid signal. In the present study, we used observed seafloor depth, thermal structure, and crustal age data to predict the geoid height associated with the cooling oceanic part of the Nubia-Somalia plate system. The depth data are from the global relief model of Earth's surface (ETOPO1) and have one arc-minute resolution (Amante and Eakins, 2009). The model integrates global shoreline, bathymetric, and topographic data sets.

The geoid height (ΔN), as derived from the plate cooling model, was used to estimate the ridge-push force (F_{RP}) in the oceanic part of the Nubia-Somalia plate system (Parsons and Richter, 1980; Turcotte and Schubert, 2002).

$$\Delta N = -\frac{2\pi G}{g}\left\{\frac{\left(\rho_m - \rho_w\right)w^2}{2} + \rho_m \alpha_v (T_b - T_o) H^2 \right.$$
$$\left.\left[\frac{1}{6} + \frac{2}{\pi^2}\sum_{m=1}^{\infty}\frac{(-1)^m}{m^2}\exp\left(-\frac{km^2\pi^2 t}{H^2}\right)\right]\right\}, \qquad (2)$$

$$F_{RP} = \frac{g^2}{2\pi G}\Delta N, \qquad (3)$$

where w is seafloor depth from the ETOPO1 global relief model (Amante and Eakins, 2009), α_v is the volume thermal expansion coefficient, g is gravitational acceleration, G is the universal gravitational constant, and ρ_w and ρ_m are densities of water and mantle, respectively.

Mechanical Behavior of Lithosphere

The computation of lithospheric strength is based on the thermal structure and relevant rheological parameters of the crust and lithospheric mantle. Olivine and plagioclase are the dominant minerals of the upper mantle and oceanic crust, respectively. Accordingly, the top 8-km-thick oceanic crust and the underlying lithospheric mantle were modeled using the rheology of diabase and peridotite, respectively. The cold part of the lithosphere deforms by shear failure, and, hence, the Coulomb-Byerlee's frictional failure criterion was used in estimating the strength of the brittle regime (Eq. 4; Sibson, 1974).

$$\sigma_1 - \sigma_3 = \beta\rho g z(1-\lambda), \qquad (4)$$

where, $\sigma_1 - \sigma_3$ represents the critical stress difference for shear failure, ρ is the average density of rock above depth z, g is gravitational acceleration, λ is the pore fluid factor (ratio of the pore fluid pressure to overburden pressure = 0.4), and β is a parameter for which the value depends on cohesion and friction coefficients. The value of the frictional coefficient on a fault, as inferred from earthquake mechanism solutions in some seismogenic regions, ranges from 0.2 to 0.3 (Reasenberg and Simpson, 1992; Iio, 1997). In the absence of information in the ocean, we used a frictional coefficient of 0.75 estimated by laboratory experiments (Byerlee, 1978). For negli-

TABLE 1. PARAMETERS USED IN THE COMPUTATION OF THE PREDICTED TEMPERATURE AND GEOID HEIGHT

Parameters	Values
Thermal expansion coefficient, α_v	3.1×10^{-5} K^{-1}
Plate thickness	95 km
Basal temperature	1723 K
Thermal conductivity, K	3.1 W m^{-1} K^{-1}
Specific heat, C_P	1.2 kJ kg^{-1} K^{-1}
Density of water, ρ_w	1000 kg m^{-3}
Density of mantle, ρ_m	3330 kg m^{-3}
Density of crust, ρ_c	2920 kg m^{-3}

gible cohesion and friction coefficient (0.75), the values of β are 0.75 and 3.0 for extensional and compressional tectonic regimes, respectively.

The strength in the ductile regime was computed using power-law creep (Eq. 5; Kirby, 1983; Ranalli, 1995) and the low-temperature plasticity flow law (Eq. 6; Goetze, 1978; Mei et al., 2010; Faul et al., 2011; Ranalli and Adams, 2013; Demouchy et al., 2013). The low-temperature plasticity flow law is based on experimental results on olivine in anhydrous conditions (Mei et al., 2010).

$$\dot{\varepsilon} = A\sigma^{n} \exp\left(-\frac{E}{RT}\right), \tag{5}$$

$$\dot{\varepsilon} = A_{P}\sigma^{2} \exp\left[-\frac{E_{k}(0)}{RT}\left(1-\sqrt{\frac{\sigma}{\sigma_{P}}}\right)\right], \tag{6}$$

where A, A_P, and n are rheological parameters; σ is deviatoric stress; E is activation energy; R is the gas constant; T is temperature; $E_k(0)$ is zero-stress activation energy; σ_P is Peierls stress, and $\dot{\varepsilon}$ is the average strain rate value of the mantle (10^{-15} s^{-1}). We used a range of strain rate values (10^{-14}, 10^{-15}, and 10^{-16} s^{-1}) in the computation of lithospheric strength. A change in strain rate by one order of magnitude increases or decreases the estimate of lithospheric strength by less than 0.3 TN m^{-1}. Detailed discussion on the upper and lower limits of lithospheric strength for a possible range of strain rate values (10^{-14} to 10^{-16} s^{-1}) can be found in Mahatsente et al. (2012). The effect of pressure on the creep strength is not significant for lithospheric depth scale modeling and, hence, is neglected. The strength at any depth is the lower stress difference associated with the predominant deformation mechanisms. The creep parameters are given in Table 2. To estimate the total lithospheric strength S, we integrated the stress $\sigma(z)$ over the thickness of the lithosphere H as

$$S = \int_{0}^{H} \sigma(z)\,dz. \tag{7}$$

RESULTS AND DISCUSSIONS

Thermal Structure

The plate cooling model has been extensively used to estimate the thermal structure of oceanic lithosphere (cf. e.g., Stein and Stein, 1992; McKenzie et al., 2005; Grose and Afonso, 2013). The model describes major features of observed seafloor depth and heat-flow variations, and it can be used to constrain strength variations in the oceanic and continental lithosphere. We used the plate cooling model to determine the temperature distribution in the oceanic part of the Nubia-Somalia plate system. The thermal model shows cooling and contraction of the plate (Fig. 2A and 2B). The lower isothermal boundary of the oceanic part of the plate is in the order ~1660 K (Fig. 2B). This isotherm defines the asymptotic plate thickness of the oceanic part of the Nubia-Somalia plate system at old ages (>75 Ma).

The models are first-order approximations and do not take into account the effects of hydrothermal circulation near the surrounding mid-ocean ridges. It is possible that the presence of fossil ridges may affect the thermal evolution of the oceanic lithosphere. In the Indo-Australian plate, for example, the fossil ridge segments of the Wharton and Central Indian Basins are documented to have ceased spreading between 65 and 42 Ma (Krishna et al., 2012), and the effects of the fossil ridge on the present thermal structure of the plate are not considered to be significant (unless existing fracture zones are reactivated by the occurrences of nearby earthquakes). In the Africa plate, however, there is a lack of fossil ridge segments, and therefore this effect is considered to be negligible.

TABLE 2. RHEOLOGICAL PARAMETERS USED IN THE COMPUTATION OF LITHOSPHERIC STRENGTH

Material	A (MPa^{-n} s^{-1})	n	E (kJ mol^{-1})	Reference
Oceanic crust				
Diabase (wet)	2.0×10^{-4}	3.4	260	Shelton and Tullis (1981)
Maryland-diabase (dry)	8	4.7	485	Mackwell et al. (1998)
Lithospheric mantle				
Peridotite (wet)	2.0×10^{3}	4.0	471	Chopra and Paterson (1981, 1984)
Peridotite (dry)	2.5×10^{4}	3.5	532	Chopra and Paterson (1981)
Material	**A_P (MPa^{-2} s^{-1})**	**E_K (kJ mol^{-1})**	**σ_P (MPa)**	**Reference**
Low-temperature plasticity				
Peridotite (dry)	1.4×10^{-7}	320	5.9×10^{3}	Mei et al. (2010)

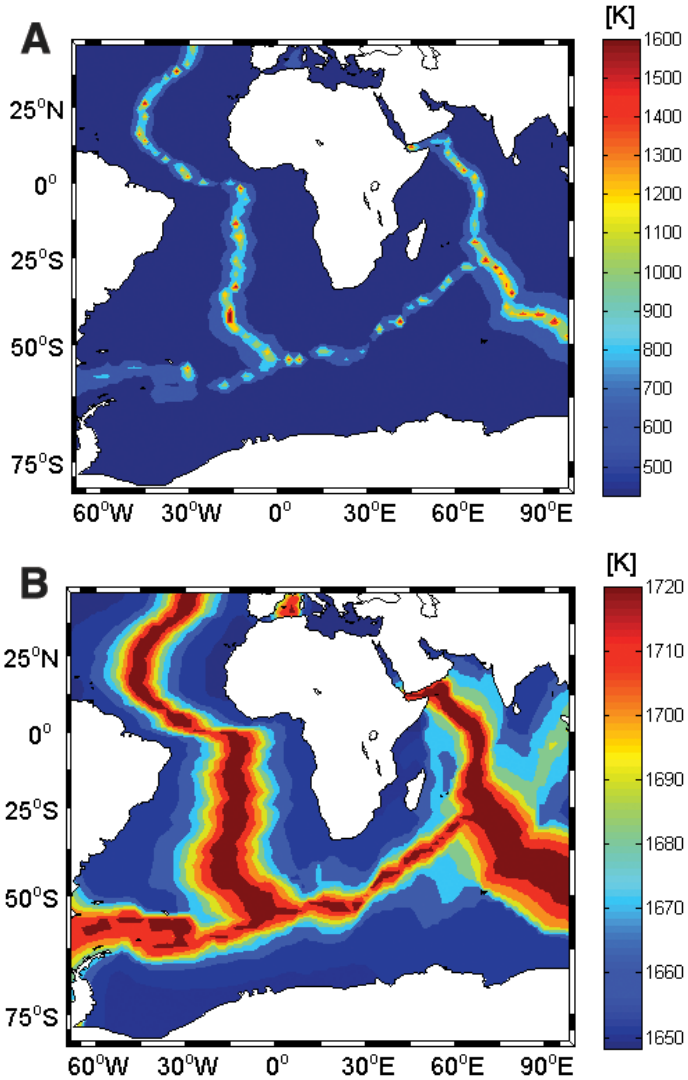

Figure 2. Models of thermal structure of the oceanic part of the Nubia-Somalia plate system at depths of (A) 8 km and (B) 95 km. We use the two thermal structures to infer the strength of the oceanic part of the plate.

We used the two thermal models and realistic rheological parameters to infer the strength of the oceanic part of the Nubia-Somalia plate system, and this is discussed in "Stress Fields and Lithospheric Strength."

Stress Fields and Lithospheric Strength

In the plate-tectonics paradigm, the lithospheric plates are the surface manifestation of large-scale flow in the mantle, with convection of the mantle primarily controlled by thin thermal boundary layers (Hess, 1962; Turcotte and Oxburgh, 1967; Moresi and Solomatov, 1998). The theory of plate tectonics has seen its greatest success in the explanation of the thermal evolution and dynamics of oceanic plates. To a very good approximation, the bathymetry of the ocean floors increases with square root of the age of the ocean floor (at least for ocean lithosphere younger than ca. 80 Ma), and the corresponding decrease in the heat flow with the square root of age gives rise to the well-established age–bathymetry–heat-flow relationship for cooling oceanic lithosphere. The density structure of the cooling oceanic plate can be computed using observed seafloor depth, surface heat flow, and geoid height variations (see detailed discussions in Parsons and Sclater, 1977; Stein and Stein, 1992; McKenzie et al., 2005; Grose and Afonso, 2013) and provides a way to compute the resulting "ridge-push" force with a high degree of fidelity.

The geoid drop from the ridge crest to the deep oceanic basin of the Nubia-Somalia plate system for the cooling oceanic plate model is ~10 m along all the ridge segments with gradients correlating with spreading (Fig. 3A). In contrast, the lithospheric (or upper-mantle) geoid (Fig. 3B), as defined in Chase et al. (2002) and Coblentz et al. (2011, 2015), exhibits a large number of long-wavelength features along the ridge segments that reflect upper-mantle density variations and are presumably associated with upper-mantle convection. The difference between the two geoid fields (Fig. 3C) can be as great as 5 m (along the Northwest Indian Ridge and along the southern Mid-Atlantic Ridge around 40°S), but overall the agreement is good.

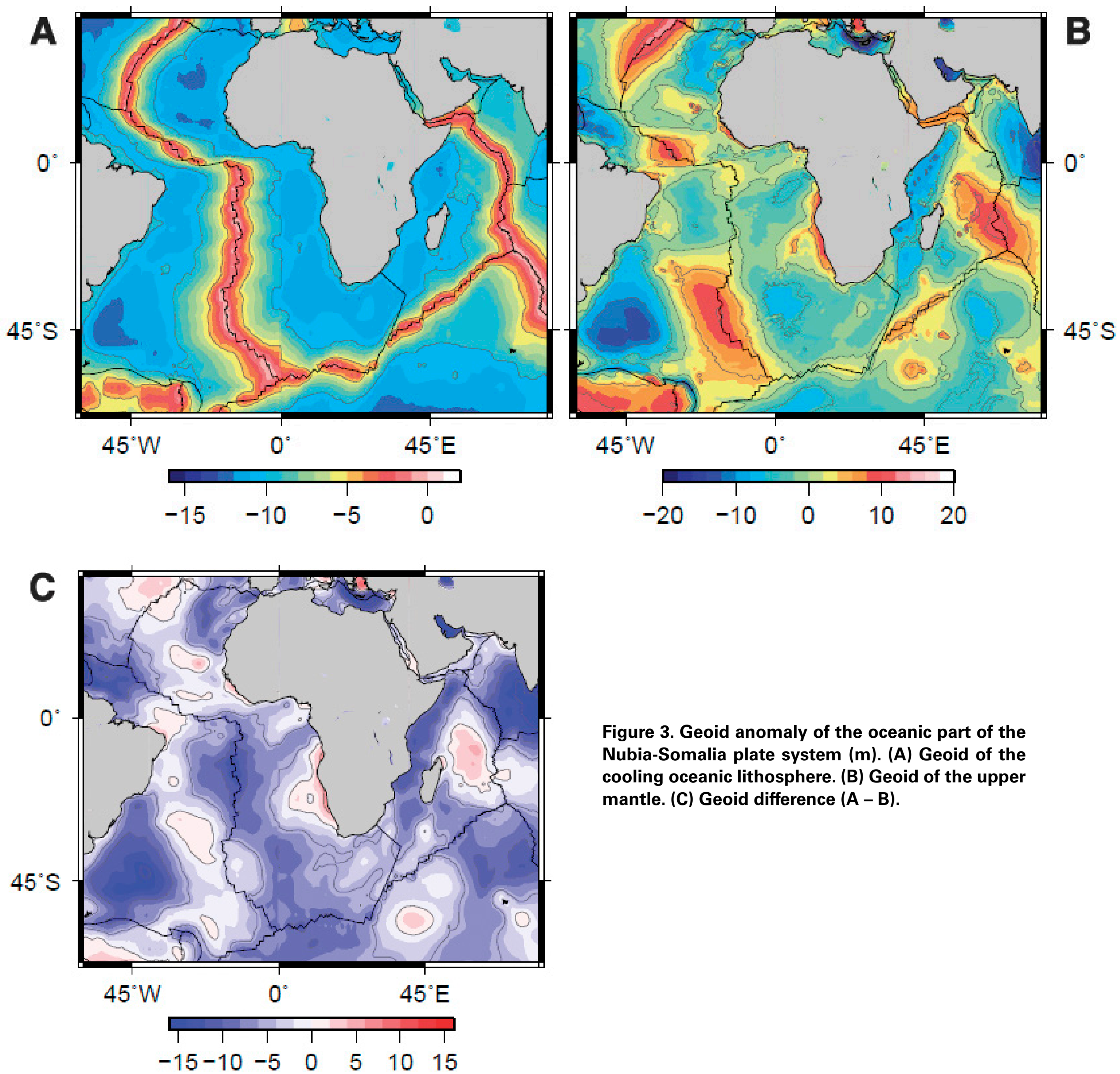

Figure 3. Geoid anomaly of the oceanic part of the Nubia-Somalia plate system (m). (A) Geoid of the cooling oceanic lithosphere. (B) Geoid of the upper mantle. (C) Geoid difference (A – B).

The difference in the predicted stresses between a formulation that represents the ridge-push force concentrated at the ridge crest (2.5×10^{12} N per meter of ridge segment) and an evolving ridge-push force that develops with the cooling oceanic lithosphere is known to be significant (see discussion in Richardson and Cox, 1984). Here, we used a finite-element analysis of the African intraplate stress field to determine constraints on the stress state resulting from variations in the gravitational potential energy of the lithosphere associated with lithospheric geoid gradients (for a discussion of the methodology, see Humphreys and Coblentz, 2007). We calculated the magnitude and orientation of the tectonic stresses, where tectonic stress refers to the horizontal stress component minus the reference state (Engelder, 1993). The lithosphere is represented by a spherical elastic shell composed of 1893 nodes in a network of 3576 constant-strain triangular elements. The spatial resolution of the finite-element grid is ~2° in both latitude and longitude. Thus, the sensitivity of the modeled stresses is limited to large-scale tectonic features with wavelengths of a few hundred kilometers. The stress magnitudes are presented as averages over a lithosphere of constant thickness, assumed to be 125 km. Substantial stress focusing may occur where there are variations in the thickness of the mechanical lithosphere, such as beneath the Afar region in the East African Rift, where the lithosphere is thought to be as thin as 30 km; thus the predicted stresses discussed here are lower limits on the state of stress.

The predicted stresses in the Nubia-Somalia plate system originating from the complete upper-mantle geoid are shown in Figure 4. The oceanic regions are in a state of compression with a magnitude in the range of 10–20 MPa. The potential energy due to the density variations associated with the continental margin and elevated continental lithosphere reduces the magnitude of the intraplates stresses in the continental regions to near-zero (in the range of 0–5 MPa), and the stress regime is an equal mix of strike-slip and compressional states of stress (Fig. 4B, inset). We note that the state of stress for continental regions with elevations greater than ~500 m is primarily strike slip to extensional.

The fact that the ambient state of stress in the African continental lithosphere is extensional has important implications for understanding the sources of intraplate tension. This prediction differs significantly from other investigator's assumptions that the state of stress within the continents is dominated by stresses transmitted from the mid-ocean ridges and is therefore compressional (e.g., Crough, 1983; Houseman and England, 1986; Zoback, 1992). These predictions are also consistent with the finding of Zoback and Mooney (2003) that compressive stresses in continental cratonic regions are associated with lithospheric roots that produce strong negative potential energy differences relative to surrounding regions (which introduce additional compressive stresses that are superimposed on the stresses derived from lithospheric gravitational potential energy forces; see discussion in Flesch and Kreemer, 2010). The combination of intracontinental gravitational potential energy forces and lithospheric root forces helps to explain the prevalence of a compressional state of stress in cratonic regions. In the case of Africa, our modeling shows that areas of high elevation and a thin mantle lid (East African Rift and South Africa) are predicted to be in extension, consistent with the observed stress regime in these areas, and indicative that the high elevations of these areas result in density moments that exceed the gravitational potential energy of the surrounding mid-ocean ridges (Bird et al., 2006). Furthermore, our prediction of deviatoric stresses associated with lithospheric geoid (and therefore gravitational potential energy gradients) in the range of 10–20 MPa is consistent with the recent work of Stamps et al. (2010, 2014), which found similar results using a thin viscous sheet approach for predicting the African intraplate stresses.

The deviatoric stress magnitudes (averaged over a 100-km-thick lithosphere) for the Nubia-Somalia plate system are shown in Figure 5. To facilitate a comparison, σ_1 (maximum compressive stress) is averaged over the plate for three cases of the ridge-push force formulation considered: (1) a ridge crest boundary force of 2.5×10^{12} N per meter of ridge segment; (2) forces from the gradient in the geoid derived with the cooling oceanic lithosphere model (based on seafloor depth and crustal age data); and (3) forces derived from the gradient of the upper mantle geoid. The net torque acting on the plate for the geoid gradient formulations (in 1.5 and 2.5×10^{25} Nm) are less than half of the net torque for the ridge crest case (~5×10^{25} Nm), suggesting that the formulation that captures the geometry of the young oceanic lithosphere is more closely in a state of mechanical equilibrium and therefore more closely approximates the ambient state of stress. The ridge crest formulation is an oversimplification of the ridge-push force (Richardson and Cox, 1984), and therefore the predicted magnitudes of σ_1 for this formulation (25 MPa in the oceanic basins and 15 MPa in the elevated continental regions) are overestimates of the compressive stress generated by the ridge-push force. Computation of the stresses

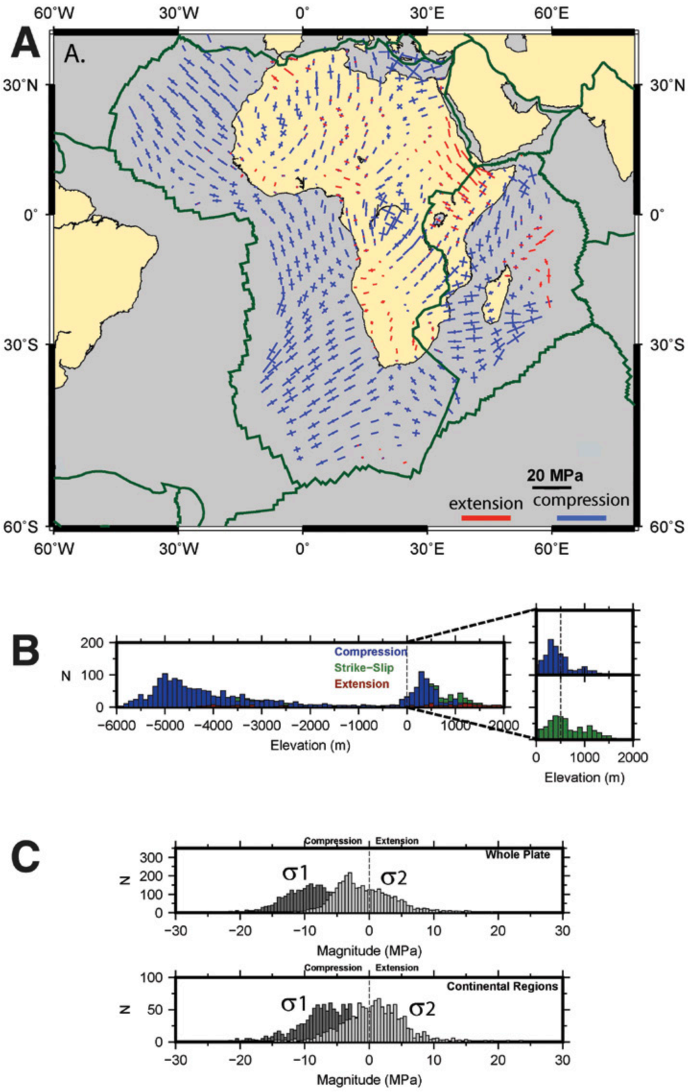

Figure 4. (A) Predicted stress fields in the Nubia-Somalia plate system derived from elastic finite element analysis of geoid gradients of the upper mantle, (B) distribution of stress magnitudes with elevation, and (C) distribution of stress magnitudes for the two principal stress directions σ_1 and σ_2.

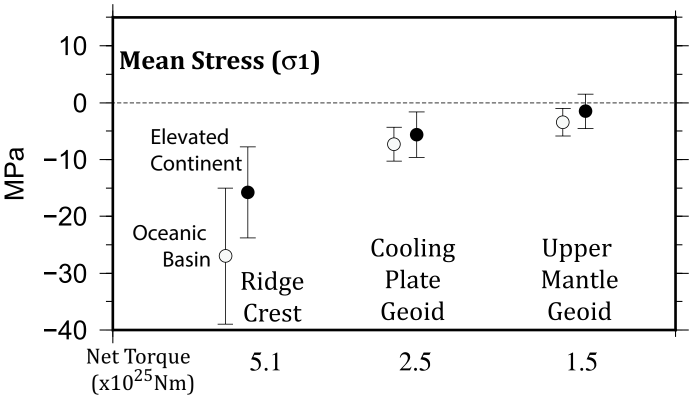

Figure 5. Mean stress magnitudes for σ_1 (most compressive stress magnitude) for three cases of the ridge-push force formulation acting on the Nubia-Somalia plate system: 1—ridge crest, force per unit length of ridge acting along the ridge crest; 2—cooling plate geoid, distributed force based on the cooling plate model; and 3—upper mantle geoid, distributed force based on the gradient of the upper-mantle geoid (Coblentz et al., 2015). White- and black-filled circles designate the mean stress value for the oceanic basin and elevated continental regions, respectively. Errors bars are the stresses for 1 standard deviation about the mean value. The net torque acting on the plate from the three formulations is also listed for comparison.

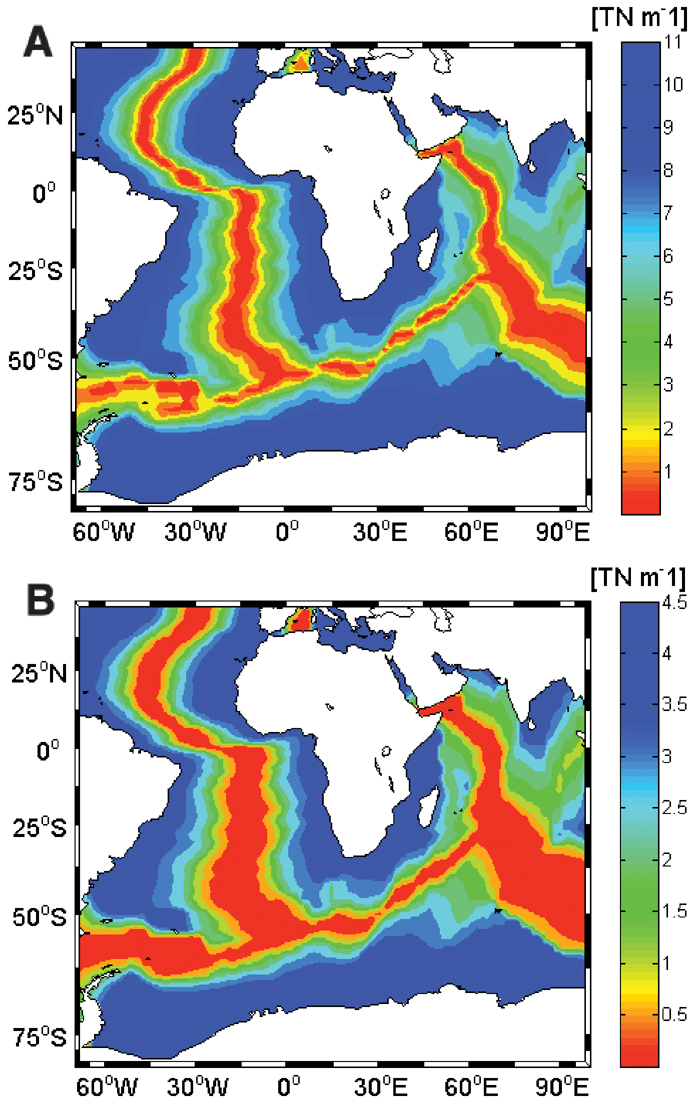

Figure 6. Models of the integrated lithospheric strength of the oceanic part of the Nubia-Somalia plate system for (A) dry and (B) wet rheology.

using the geoid gradient is more viable, and both the cooling oceanic plate model and the upper-mantle geoid predict average compressive stresses in the range of 5–10 MPa in the oceanic basins, which are reduced by about a third in the continental regions. Ridge-push based on the upper-mantle geoid (which captures other subtleties in the upper-mantle density structure that are difficult to explicitly include in a numerical model of the intraplate stress field) results in a predicted state of stress in the continental lithosphere that is neutral to slightly compressive and is more consistent with observation (e.g., Zoback and Mooney, 2003).

These stress fields can be used to constrain the strength of the oceanic part of the Nubia-Somalia plate system. Models of the integrated strength of the oceanic part of the Nubia-Somalia plate system for dry and wet rheology are shown in Figures 6A and 6B, respectively. The models are based on the assumptions that the ocean basin is in compression and the regions near the surrounding mid-ocean ridges and the rift zones are in extension. Rocks at a depth greater than 6 km are assumed to be impermeable and could block hydrothermal circulation (Morgan et al., 1987). For this reason, we consider both wet and dry olivine rheology as limiting cases. The strength of the oceanic part of the Nubia-Somalia plate system increases with age both for dry and wet rheology. The use of dry rheology increases the strength by a factor of 2. The older oceanic parts of the plate exhibit high strength. This is attributed to the cold geothermic condition in the old oceanic lithosphere. The regions near the axial zones of the mid-ocean ridges are characterized by low strength (~1 TN m^{-1}).

Role of Ridge-Push in the Stress State of the Nubia-Somalia Plate System

The Nubia-Somalia plate system moves at a slower rate and is completely surrounded by mid-ocean ridges. It is therefore a reasonable assumption that the effects of basal shear traction and far-field forces on the state of stress of the Nubia-Somalia plate system are negligible. Thus, the unique dynamics and tectonic setting of the Nubia-Somalia plate provide an opportunity to study the role of ridge-push force in the stress state of the Nubia-Somalia plate system. Figure 7 shows the ridge-push force anomaly (on a point-by-point basis) in the oceanic part of the Nubia-Somalia plate system, as derived from the isostatic geoid anomaly of the oceanic part of the Nubia-Somalia plate system (Fig. 3A) using a plate cooling model.

We assessed the role of ridge-push force in the stress state of the Nubia-Somalia plate system by comparing the net ridge-push force (Fig. 7) with the integrated strength of the oceanic part of the plate for dry and wet rheologies of the crust and upper-mantle structures (Figs. 6A and 6B). The strength of the oceanic part of the Nubia-Somalia plate system is higher than the ridge-push forces (Fig. 8). This may indicate that the oceanic part of the plate is very little deformed; thus, the ridge-push forces may be compensated by significant strain rates outside the oceanic parts of the plate. Lithospheric deformation occurs when the ambient force exceeds the integrated strength of the lithosphere (Sonder and England, 1986). The stresses associated with the ridge-push forces may be transmitted into the interior of the oceanic plate and influence the present-day stress state of the African continent. As discussed already, the stress fields in the Nubia-Somalia plate system based on the gravitational potential energy forces associated with the cooling oceanic lithosphere are dominantly compres-

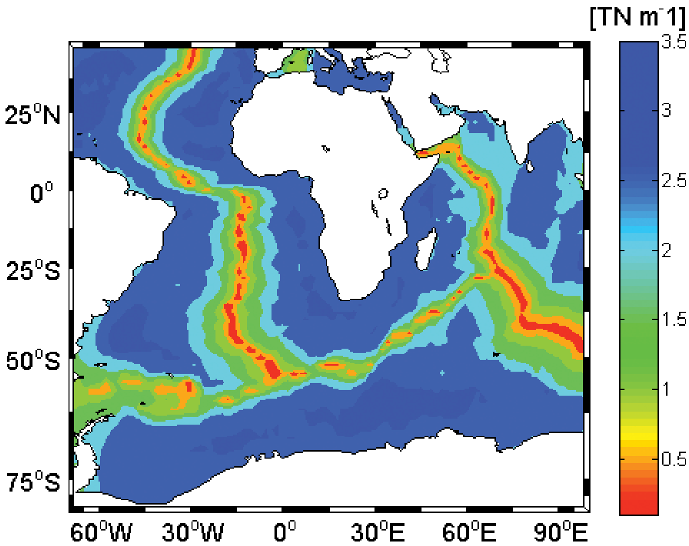

Figure 7. Ridge-push forces derived from isostatic geoid anomalies in the oceanic part of the Nubia-Somalia plate system. The ridge-push force for old plates (>75 Ma) is of the order of ~3.5 TN m^{-1}.

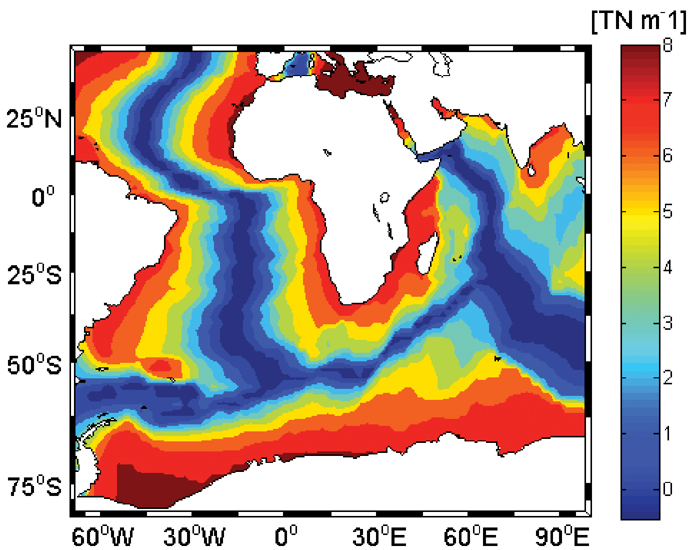

Figure 8. Difference between ridge-push (Fig. 7) and integrated strength of the oceanic part of the Nubia-Somalia plate system (Fig. 6A).

sional. It is, therefore, less likely that the ridge-push–related stresses are diffused due to intraplate deformation. Instead, the stresses are transmitted into the interior of the plate (and, in the case of the Nubia-Somalia plate system, into the continental regions). An important implication of the transmission of stresses into the continents is that potential energy variations (as defined by the geoid gradients) are communicated via the intraplate stress field throughout the plates (see discussion in Sandiford, 2010). This notion is supported by the observation that intraplate deformation (in the form of enhanced seismic moment release and neotectonic indicators) within the continental interiors is consistent with significant ongoing strain accumulation in the continental interior (e.g., Braun et al., 2009). This transmission (or communication) of stresses is dependent, however, on the magnitude of basal shear traction and other sources of tectonic forces. The relatively slow rate of movement of the Nubia-Somalia plate and the presence of the surrounding mid-ocean ridges suggest that the basal shear traction (at least due to the plate motion over the asthenosphere) is negligible.

We leave a more detailed evaluation of the effects of basal shear traction on the relation between lithospheric strength and ridge-push transmission in the oceanic part of the Nubia-Somalia plate system for future studies. For an asthenospheric strain rate of 10^{-14} s^{-1} and viscosity 10^{20} Pa S, basal shear traction can be in the order of a few TN m^{-1} (Mahatsente et al., 2012).

CONCLUSIONS

Ridge-push is one of the major forces acting on the plates in the plate-tectonics paradigm, and as such, it is one of the major contributors to the stress state of tectonic plates. Most of the continental plates (including North America, South America, and Indo-Australia) are relatively fast moving, and the stress generated by ridge-push forces may be counterbalanced by basal shear tractions and other sources of tectonic stresses. The Nubia-Somalia plate, however, is unique in its combination of slow absolute plate velocity and its plate-boundary geometry, which is nearly completely surrounded by mid-ocean ridges. It therefore offers an ideal location in which to evaluate the contribution of the ridge-push force to the stress state of continental plates. We assessed the role of ridge-push force in the stress state of the Nubia-Somalia plate system by comparing the magnitudes of ridge-push forces with the integrated strength of the oceanic part of the Nubia-Somalia plate. The integrated lithospheric strength and ridge-push forces are age dependent, and, hence, the comparison is made for all ages of the plate. The magnitude of the ridge-push force is significantly less than the integrated strength of the oceanic part of the Nubia-Somalia plate system. This suggests that the oceanic part of the plate is very little deformed. Thus, the stresses associated with the gravitational potential energy of the surrounding ridges may be transmitted into the interior of the oceanic and continental parts of the Nubia-Somalia plate system. The present-day stress fields in the Nubia-Somalia plate system may be caused by lateral variations of lithospheric structure and thicknesses in the African continent and gravitational potential energy forces associated with the surrounding mid-ocean ridges.

The effects of other sources of tectonic stresses on the stress state of the Nubia-Somalia plate system have not been quantitatively evaluated in this study. However, given the slow motion of the Nubia-Somalia plate system and its present tectonic setting, these effects may be negligible.

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