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LITHOSPHERIC GEODYNAMICS OF THE ARABIAN MARGIN

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by
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ABSTRACT

The Arabian Margin experienced intense volcanism over the last 10 Ma, including volcanic eruptions as recent as 600 years ago. What is more, two earthquakes with magnitude > 5 have been recently reported with normal faulting along the Arabian Margin, suggesting that the Arabian Margin is undergoing active deformation. Due to the limited number of GPS stations within the Arabian plate, investigating the intraplate deformation was challenging. A new set of GPS data with 87 stations is used in this thesis to investigate the Arabian margin rigidity and intraplate deformation. This new GPS velocities show higher residuals along the Arabian margin that produces dilatational and shear strain rate patterns within the Arabian margin, in the vicinity of the Makkah-Madinah Transtensional zone. Both anomalies can be correlated with the recent earthquake and volcanic activities that occurred along preexisting structures attributed to the northern Red Sea diffuse extension. The causes of these GPS residuals along the Arabian Margin are unknown. In this thesis, we use the finite element modeling approach (GTECTON platform) to highlight the mechanical deformation processes along the Arabian margin and test their driving forces. These candidate forces are related either to the edge forces, the Arabian Margin interior forces as introduced by calculating the Gravitational Potential Energy, or the basal tractions as driven by sub-lithospheric topography and mantle flow. Our results indicate that the GPS residuals are not likely linked with the Gravitational Potential Energy forces. Instead, the basal tractions along an asthenospheric channel is the potential driving force for the observed deformation along the Arabian margin.
Although the asthenospheric channels and northern Red Sea diffuse extension appear to play an important role in accommodating the basal tractions and recent deformation, their origin is unknown. To address this problem, we adopt a comprehensive approach that combines the structural changes along the Red Sea with the regional tectonic forces evolution of the Afro-Arabian plate boundaries. To demonstrate, we test the effect of the Mediterranean tectonic evolution (the northern boundary of the Afro-Arabian plate) and the nature of rifting along the southern Red Sea and Sirhan rifts on the northern Red Sea rifting and the Arabian margin evolution. Our numerical results indicate that the African slab roll-back in the Mediterranean, as developed the Aegean extension, caused Africa to slow down with respect to Eurasia, allows the extensional processes along the southern Red Sea to transfer to the Sirhan rift through the Makkah Madinah transtensional zone. This motion transfer appears to play a significant role in initiating a relict boundary between Arabia and Africa at ~ 27 Ma that inhabited afterwards by what is known as the Makkah Madinah volcanic line. Accelerating the eastern Mediterranean while keeping the Aegean block fixed could provide the required boundary conditions to initiate northern Red Sea diffuse extension at 25 Ma. The region of diffuse extension is bounded by the volcanic provinces in NE Egypt to the west and Sirhan rift to the east, Makkah Madinah transtensional zone to the southeast and the Mediterranean to the north. Last, our results indicate that the northward propagation of the dike intrusions along the Red Sea can localize the rifting along the northern Red Sea at ~ 21 Ma.
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Chapter 1

Introduction

Since the 1980s, many researchers have investigated western Arabian geology, particularly focusing on the geodynamic evolution of the Arabian margin in a plate tectonics framework (Bosworth et al., 2005, references therein). The tools used in these analyses have primarily been either surface geologic mapping or studying deep lithospheric imaging (e.g. Hansen et al., 2006; Park et al., 2007; Tang et al., 2018). Although many outstanding questions remain, these previous studies provide a framework to place Arabia’s evolution within the context of global tectonics.

The Arabian margin (Figure 1.1) has had a history of volcanism, earthquakes and tectonic uplift (Bosworth and Stockli, 2016; El-Hadidy, 2015; Omar and Steckler, 1995). Particularly, the Lunayyir volcanic eruption is one of at least twenty one different eruptions estimated (Camp et al., 1987) to have occurred within the Arabian margin over the past 1500 yr (Duncan and Alamri, 2013). This volcanic field experienced more recent deformation in the form of earthquake swarms that occurred in 2009 with a surface fault rupture that extended for 8 km (Pallister et al., 2010). The Rahat volcanic eruption occurred in the vicinity of Madinah city on 1256 (Camp et al., 1987) preceded by a historic earthquake (Figure 1.1) event in the Khyber Harrat in 1068, as inferred from the
Figure 1.1: Conceptual map of the Arabian margin. This map shows the extensional style changes along the Red Sea strike. The areas of asymmetrical diffuse extension, central Red Sea transform, and Makkah-Madinah transtensional zone (MMTZ) is emphasized. Yellow and red stars denote the Tabuk 2004 earthquake and the Lunayyir 2009 earthquake, respectively. Green stars represent Rahat 1068 earthquake.
simultaneous collapse of three dams that had been built between 660 – 680 (Saudi Geological Survey report). and Tabuk earthquake on 2004 (Aldamegh et al., 2009). Although these events are geologically recent, they appear to localize on preexisting structures (such as Tabuk 2004). These structures highlight an earlier phase of deformation along the Arabian margin (Roobol and Stewart, 2019) that occurred contemporaneous with the Red Sea rifting evolution (Figure 1.2). The goal of this work is to quantify the present-day deformation and evaluate the associated tectonic driving forces, and incorporate these results into a proposed tectonic evolution of the region. This tectonic model highlights the relationship between the present-day deformation and the earlier deformation phase associated with the Red Sea rifting evolution.

In general, the present-day tectonic events have been assessed as discrete events, disconnected in time. In 2002, a new collaboration was established between King Abdulaziz City for Science and Technology (KACST) and the Massachusetts Institute of Technology (MIT). Under this collaboration, GPS stations were deployed around Arabia (Figure 1.3) to provide a continuous time series for the motion of Arabia, particularly along the Red Sea (ArRajehi et al., 2010). This new deployment was designed to span the width of Arabia – about 2000 km. As a result GPS stations were distributed approximately 800 km apart, leaving significant gaps in coverage. These stations were used to define the rigid-plate reference frame of the Arabian plate.
Figure 1.2: Geologic map of the northwestern Arabian margin as modified after Roobol and Stewart (2019). This map highlights the diffuse extensional structures in red and dike intrusions in green. Both features are occurred in response to the northern Red Sea rifting evolution. On the upper left corner is the general map of Arabia, and in the lower left corner is the map key.
In 2014, a new project was established by the Saudi General Commission for Survey (GCS) to deploy more than 100 additional GPS stations around Arabia. Although the first phase of this project was aimed at covering the Saudi major cities, including Riyadh, Jeddah, Makkah and the eastern province, it also provides sufficient coverage to study surface deformation within the western Arabian margin. In 2016, we were granted access to this data set, via an official channel through KACST. We processed and analysed the new GPS data using the GNSS-Inferred Positioning System and Orbit Analysis Simulation Software package (GIPSY-OASIS 6.2), developed by the Jet Propulsion Laboratory (JPL). This processing was conducted at the University of South Florida under the supervision of Dr. Rocco Malservisi.

In Chapter 2, we provide results from the processing of data from this new set of GPS stations using the GIPSY-OASIS 6.2 package. We are able to obtain a significantly improved velocity field for the Arabian plate. In this analysis, we assess the GPS time series for the stations across the GPS network and determine its uncertainties. With this velocity field, we analyze the dilatational and shear strain rates, following the approach proposed by Hackl and others (2009) to test the concept of a rigid Arabia plate. We find that the region of western Arabia deviates from the rigid plate assumption, and we correlate the observed strain rate anomalies with reported earthquakes focal mechanisms and other geologic deformation indicators.
Figure 1. 3: Map of GPS velocities with stable Africa reference frame (ArRajehi, et al., 2010). This map shows 10 GPS stations within Saudi Arabia, mainly distributed along the Red Sea as their study aimed to investigate the Red Sea extension.
In chapters 3 and 4, I investigate the mechanisms and physical processes that can drive this observed deformation along the Arabian margin. We use the finite element method (modeling code GTECTON (version 2017.3; Govers and Wortel, 1993; Govers & Meijer, 2001; Özbakir et al., 2017)) on a spherical shell with discrete faults and triangular element mesh, as a tool to evaluate the deformation processes acting in the region of the Arabian Margin.

In Chapter 3, we explore the potential driving forces for the observed strain rate anomalies that were reported in Chapter 2. We calculate the displacements on a viscoelastic spherical shell. The potential driving forces were used as boundary conditions on the original model domain. We compare the deformational consequences of these forces to our observed GPS data. We use this misfit analysis between the observed GPS velocities and model predicted velocities, to assess the reliability of the force considered.

In Chapter 4, we test models of the evolution of Arabian plate boundary forces and their possible role in producing the observed geologic and tectonic evolution of the western Arabian margin. In particular, we explore how the Arabian margin and Sirhan rift are affected by interactions among ongoing tectonic processes in the Afar mantle plume, southern Red Sea, Gulf of Aden, and within the Mediterranean, which have led to the formation of the Aegean rift.
The importance of this work comes from the need to re-assessing the natural hazards within the Arabian margin. This can be done by observing the deformation field and evaluating the driving forces. Furthermore, investigating the tectonic evolution of the Red Sea rifting and the Arabian margin is the base for future studies that either explore natural resources or investigate natural hazards within the Arabian margin.
References


Chapter 2

How rigid is the Arabian Margin? Tectonic strain patterns from a new GPS network

Abstract

The deformation field of the Arabian Margin was previously investigated using a limited network of Global Navigation Satellite System (GNSS) stations. That analysis supported a rigid plate assumption for Arabian plate (ArRajehi et al., 2010). In this study, we evaluate the Arabian plate deformation field using an expanded network totaling 87 GNSS stations distributed within Saudi Arabia. We process the data using GIPSY-OASIS 6.2 and analyze the resulting time series for long term, annual, and semiannual signals. We calculate uncertainties for our GPS velocities using the Hector package. The new set of GPS data shows shear strain as expected in the vicinity of the Dead Sea Transform, which also extends southward along the northern Red Sea margin. In contrast to previous analyses, dilatational strain is observed within the western Arabian Margin, which is correlated with recent earthquake activity and recently active volcanism. Central and eastern Arabia exhibit rigid body motion as our GPS velocities are in agreement with the velocities calculated using the Euler pole set described by the International Terrestrial Reference Frame (ITRF)-2008 (Altamimi et al., 2012). Our results suggest that there is a transition from essentially rigid behavior within central Arabia to non-rigid behavior in
western and northwestern Arabia, consistent with geological observations in the form of distributed grabens and dike intrusions.

2.1 Introduction

Quantifying plate rigidity is important in understanding the links between tectonic deformation and the plate tectonic system (DeMets et al. 1990; Sella et al. 2002, Malservisi et al., 2013; Njoroge et al., 2015). Although the concept of rigid plate behavior is typically assumed, it may be less applicable at or near the margins of slowly spreading rift systems (Dixon et al. 1996; Gordon 1998). Improvements in geodetic observing systems now allow us to measure displacements with an uncertainty approaching 1-2 mm/yr (Drewes 1982; Dixon et al. 1996; Sella et al. 2002). This allows us to assess plate rigidity at a level not possible previously.

The Red Sea Rift formed with the separation of Arabia from Nubia approximately 23-21 Ma (Bosworth et al., 2005), although there is evidence of northern Red Sea extension as early as early Oligocene (Omar and Steckler, 1995; Tubbs et al., 2014). Western Arabia lies within the Arabian shield, with outcrops indicating that the upper crust is composed of crystalline Precambrian basement, Cenozoic alkalic volcanic fields (Harrat) and small sedimentary basins and grabens (Stern and Johnson, 2010). Although continental rifting typically occurs as either narrow localized or wide diffuse rift zones (Christensen and Mooney, 1995), the Red Sea rift architecture exhibits both rifting styles along strike (Figure 2.1), from localized rifting in the south to a more diffuse rift zone in the north.
Figure 2.1: Conceptual map of the Arabian margin. This map shows the extensional style changes along the Red Sea strike. The areas of asymmetrical diffuse extension, central Red Sea transform, and Makkah-Madinah transtensional zone (MMTZ) is emphasized.
(Johnson, 1998), which occurs in the form of diffuse grabens and dike intrusions. The variation in rifting style have been assumed to be as a result of differences in lithospheric strength, localized rifting associated with strong, old, stable lithosphere, and diffused rifting more typically associated with weaker and more deformable lithospheric conditions (Kogan et al., 2012).

Previous studies using a relatively small number of GPS stations concluded that Arabia could be considered as rigid based on 12 GPS stations distributed across the Saudi Arabian portion of the Arabian plate (ArRajehi et al., 2010; McClusky et al., 2010). Most present-day earthquakes and recent volcanism, evidence of active tectonics (e.g., Uwayrid, Ishara, and Lunayyir Harrats) are distributed within the northwestern Arabian Margin (Figure 2.1) and along a diffuse set of Red Sea syn-rift structures. Some of these structures appear to be active (i.e., they have hosted recent moderate earthquakes: the Tabuk 2004 earthquake (Aldamegh et al., 2009) and the Lunayyir 2009 earthquake (Pallister et al., 2010)). This region was poorly sampled in the previous GPS studies.

Here we investigate the present extent of Arabian plate rigidity using an updated geodetic network. This GPS network data is provided by the Saudi General Survey Commission (GSC), who deploy more than 101 GPS stations distributed across Saudi Arabia (Figure 2.2). We update the previous analyses for displacements and develop strain rate maps to investigate the rigidity of the Arabian margin.
Figure 2: Map of GPS velocities with stable Nubia reference frame. This map shows 87 GPS out of 101 provided by the Saudi General Survey Commission. We plot the GPS velocities with respect to stable Nubia reference frame as calculated using the UNAVCO plate motion calculator with Euler pole ITRF2008. The black vectors denote our best data quality for the time series that span for more than 18 months. The blue vectors denote our moderate quality data for the time series that span for less than 18 months. Interestingly, both categories exhibit very similar azimuths with different magnitudes. The data is processed using Gipsy-OASIS 6.4, which is a code developed by the Jet Propulsion Laboratory (JPL).
2.2 Data analysis

Our GPS data set is provided by the Saudi General Survey Commission, using data from 101 GPS stations. Of these GPS stations, 14 GPS stations were removed due to their high uncertainty, leaving 87 GPS stations distributed across Saudi Arabia in our analysis (Figure 2.2). The Saudi General Survey Commission chose these GPS locations to be primarily in the vicinity of major Saudi Arabian cities for surveying purposes. In spite of this, an analysis of the time series show that the data quality is similar to geodetic level time series (Hackl et al., 2011).

Our raw GPS RINEX files were processed using GIPSY-OASIS 6.4, which is a code developed by the Jet Propulsion Laboratory (JPL). We produce a daily solution for each single station estimating fiducial free position by employing a precise point positioning (PPP) strategy (Zumberge et al., 1997). We perform a phase ambiguity resolution using the single-receiver algorithm described by Bertiger et al. (2010). FES2004 ocean-loading corrections are applied, compatible with the JPL orbits calculations. We account for the tropospheric delay through VMF1 mapping functions. We correct for the second order ionosphere effects using the IONEX model (ftp://cddis.gsfc.nasa.gov/pub/gps/products/ionex/; Bassiri and Hajj, 1993; Kedar et al., 2003). We transform the daily solutions into the ITRF2008 No Net Rotation reference frame (Altamimi et al., 2012) using the UNAVCO plate motion calculator (https://www.unavco.org/software/geodetic-utils/plate-motion-calculator/plate-motion-calculator.html).
Following the approach of Njoroge and others (2015) and Malservisi and others (2015), the time series are analyzed for long-term trends to compute velocity components of each station. We visually inspect each time series to identify other unknown data jumps. We then fit each time series component using the equation prescribed by Njoroge and others (2015). Daily positions that differ by more than five times the uncertainty from the time series are removed to re-fit the data in an iterative way, until no outliers remain.

2.3 Data Quality

Most of our GPS data span different periods of time from mid-2016 to May 2018, with a few of the stations deployed in 2015. Although our time series are relatively short, they are sufficiently long enough to have significant results (Bennett et al., 2007). Moreover, the density of our GPS network enables us to infer a spatial pattern, even if our signal is still slightly noisy due to the length of the time series.

We calculate the GPS velocities with stable Nubia (Figure 2.2; Table 1) using the Euler pole set described by ITRF-2008 (Altamimi et al., 2012). We place our GPS vectors into two categories based on their time series duration and quality (Figure 2.2). Continuous GPS stations with time series span of more than 18 months are classified within the high quality category A (Figure 2.2, black vectors), while stations with time series duration of less than 18 months (Figure 2.2, blue vectors) are classified within the moderate quality category B (see table 1 for details).
In Figure 2.3, we show the time series for four stations representative of our GPS network. We plot up, north, and east component of each station with respect to ITRF 2008 No Net Rotation reference frame (Altamimi et al., 2012). These stations are MK87 (western region), RY96 (central region), HL04 (100 km away from one of the volcanic fields), and HL02 (northern region). All of these stations are classified as quality A except HL02, which is classified as quality B station. The velocity uncertainties of these time series were evaluated following the Hackl et al. (2011, 2013) approach, where the noise of our time series is characterized using the Hector package (Bos et al., 2012). Using this package, we find that our uncertainties using white plus flicker noise are of the order of mm/yr or lower for the horizontal. The analysis of the color of the noise using a power spectrum approach (Hackl et al., 2013) shows that the noise is very close to flicker or with even lower time correlation. The corresponding power spectrum density plots for the time series presented in figure 2.3 are shown in figure 2.4. This is similar to what has often been observed (Hackl et al., 2011 and Njoroge et al., 2015) in atmospherically stable areas (e.g. Kalahari desert in South Africa). The uncertainties of these stations are probably lower than normally expected for their short time series because of the dry environment (Malservisi et al., 2013).
Figure 2.3: Time series plots of four GPS stations. The time series in the ITRF 2008 for the sites "HL04, RY96, HL02, MK87". The data variation over the period of recording shows insignificant error in comparison with the observed deformation signal. This collection of GPS stations is chosen to represent the verity of our velocities, from rigid body-like motion as reported by RY96 station to Red Sea parallel residuals as indicated by MK87 station, and to further internal deviation as produced by of HL04 and HL02 stations.
Figure 2.4: The corresponding power spectrum density plots for the GPS stations presented in figure 2.3. The purple crosses denote the residual time series. The x-axis denote cycles per year while y-axis the power spectrum magnitudes. The green line is the best model as characterized with white plus flicker noise.
2.4 Results

2.4.1 GPS residuals with Stable Arabia Reference Frame

We calculate our GPS residuals (Table 2) with respect to a stable Arabia reference frame using the Euler pole set described by ITRF-2008 (Altamimi et al., 2012). These residuals are used to investigate the internal deformation within the Arabian Margin (Figure 2.5). As shown in Table 2, where we list our GPS stations with their corresponding residuals and azimuth, our GPS data indicate significant residuals along the Arabian margin striking in a direction NNW approximately parallel to the Red Sea. In contrast, central and eastern Arabia exhibit rigid body motion as our GPS velocities are in agreement with the velocities calculated using the rigid plate Euler pole set as described earlier. Overall, our new GPS network lacks ideal coverage in the area of interest, but in spite of this limitation, we are able to observe ongoing deformation in the form of GPS residuals within the Arabian margin.

2.4.2 Strain rate analysis

2.4.2.1 Interpolation

We use these new geodetic observations to investigate the Arabian margin deformation through the use of regional strain rate maps. There are several previously developed
Figure 2.5: Map of GPS velocities with stable Nubia reference frame. This map shows the GPS residuals with respect to a rigid Arabia reference frame calculated using the UNAVCO plate motion calculator with Euler pole ITRF2008. The black vectors denote our best data quality for the time series that span for more than 18 months. The blue vectors denote our moderate quality data for the time series that span for less than 18 months. Quality A data show two distinct patterns along the Arabian margin, and within the Arabian plate. The data is provided by the Saudi General Survey Commission and processed using Gipsy package.
inversion approaches used to produce strain rate maps. Spakman and Nyst (2002) determined strain rate in a discretized region by using different paths of relative displacement between pairs of observation points. Kreemer and others (2000) included geological and geophysical information (e.g. fault locations, etc.) in their evaluation of strain in a geodetic network by minimizing the local strain rate and velocity field residuals through an Eulerian pole inversion.

Here, we modified the Hackl and others (2009) approach to use the gpsgriddertool (Sandwell and Wessel, 2016), which incorporates a 2-D elastic interpolation. This approach uses spline interpolation (Wessel and Bercovici, 1998) to obtain a continuous strain rate map without requiring any prior information about regional faults (see, e.g., Wdowinski et al., 2001; Allmendinger et al., 2007; Kahle et al., 2000; Hackl et al., 2009). This method produces a continuous velocity field by using analytic Green’s functions for the in-plane response of a 2-D elastic body to in-plane forces (Sandwell and Wessel, 2016). Coupling between the two horizontal velocity components of the GPS models is incorporated in this method and controlled by our choice of Poisson ratio; however, sensitivity tests of our results indicate that they are insensitive to such coupling, perhaps due to the large spatial scale of our GPS network.

The results of this interpolation can be used to calculate the strain rate tensor ($\dot{\varepsilon}_{ij}$) and antisymmetric rotation rate tensor ($\dot{\omega}_{ij}$) using the following (Eq. 2.1):

$$\dot{\varepsilon}_{ij} = \frac{1}{2} \left( \frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right)$$

$$\dot{\omega}_{ij} = \frac{1}{2} \left( \frac{\partial v_i}{\partial x_j} - \frac{\partial v_j}{\partial x_i} \right)$$

Eq. 2.1
Where \( i \) and \( j \) represent the northerly and easterly velocity field directions, respectively. The trace of the strain rate tensor (Eq. 2.2) represents the dilatational strain.

\[
\delta = \dot{\varepsilon}_{ii} + \dot{\varepsilon}_{jj} \quad \text{Eq. 2.2}
\]

We obtain a fuller description of the deformation along the Arabian Margin by using an eigenspace analysis. This also provides the shearing component of the strain rate. Particularly, defining the maximum and minimum shear strain rate magnitude and direction using eigenvalues \((\lambda_1, \lambda_2)\) and eigenvector analysis, respectively. The maximum shear strain rate is computed by a linear combination of the maximum and minimum eigenvalues, (Eq. 2.3):

\[
\dot{\varepsilon}_{\text{max shear}} = \frac{\lambda_1 - \lambda_2}{2} \quad \text{Eq. 2.3}
\]

The direction of the maximum shear makes an angle of 45 with the eigenvector associated with the largest eigenvalue. Hence, there are two maximum shear strain vectors perpendicular to each other, which can't be distinguished from each other without additional constraints. Equation 2.4 gives the shear strain direction:

\[
\theta_{1,2} = \frac{1}{2} \arctan \left( \frac{2\dot{\varepsilon}_{ij}}{\dot{\varepsilon}_{ii} - \dot{\varepsilon}_{jj}} \right) \pm 45 \quad \text{Eq. 2.4}
\]

### 2.4.2.2 Strain rate maps

We find significant dilatational strain within western Arabia. Another smaller anomaly occurs within the southern Arabian margin (Figure 2.6). The GPS network coverage in
Figure 2.6: Map of dilatational strain rate pattern within the Arabian margin. We correlate the dilatational strain pattern with earthquake (Mw > 4.5) distribution (using the IRIS database (www.iris.edu)) and focal mechanisms (using the Global CMT Catalog (Dziewonski et al., 1981; Ekström et al., 2012)).
southern Arabia is relatively sparse reducing the importance we can place on the southern Arabian margin anomaly. Our GPS velocities within the stable Arabia reference frame show large magnitudes along the Red Sea margin. Several GPS stations (i.e. HL04, HL02, and HL01) show a northward motions deviating from the regional NNW pattern. This produces the observed dilatational strain rate anomaly within the Arabian margin.

The anomaly is correlated with the displacement of HL04 station, which is classified as a category A station. The northward motion of HL04 deviates from the general NNW strike pattern along the Red Sea producing extensional strain within the Arabian margin. The strain vectors as denoted in figure 2.6 indicate an extensional axis parallel to the Red Sea. Several GPS stations support HL04 displacement, such as HL01 station, which is classified as category a B station, that is moving to the north. Moreover, although the HL02 station, which is classified as category A station, is about 300 km away from HL04, it exhibits similar azimuth and a slightly lower magnitude. We calculate dilatational strain rate maps excluding the HL04 station, and we got the same dilatational strain rate anomaly with lower magnitude. More, we calculate the dilatational strain rate map without all HL stations, and we got no dilatational strain rate anomaly within the Arabian margin. This analysis highlights the importance of all HL stations, not only HL04, in the observed dilatational strain rate anomaly within the Arabian margin.

Based on our strain rate analysis using GPS stations within Arabia, we find that the shear strain localizes within Arabia (Figures 2.7); to two main locations, south and north of the Makkah Madinah Transtensional Zone (MMTZ). The resulting shear strain rate vectors
for both observed anomalies are semi-orthogonal to the Red Sea. The shear strain rate values are smaller in comparison to the observed dilatational strain rate. Similar to the dilatational strain rate, the observed shear strain rate anomaly (Figure 2.7) is correlated with the HL04 station. Similarly to the dilatational strain rate calculations, all HL stations, including HL 04, are important as well in producing the observed shear strain rate anomaly.

To highlight the effect of the Arabian plate boundary along the Dead Sea Transform, we include here the UNAVCO stations (e.g. RAMO and RASH) in our analysis. We find the shear strain rate, as expected, is high in the vicinity of the Dead Sea Transform (Figures 2.8). This shear strain anomaly is slightly larger in magnitude than the shear strain rate anomaly calculated with our stations. These stations highlight the effect of the plate boundary on the northwestern Arabian margin. The resulting shear strain vectors for the observed anomaly are parallel to the Dead Sea Transform.

As we using our GPS network for both shear strain rate maps (Figures 2.7 and 2.8), the magnitude of the shear strain rate anomaly within the Arabian margin reduced significantly by introducing the Dead Sea Transform. Moreover, the observed shear strain rate along the Dead Sea Transform localized along the transform, while the observed shearing along the central Arabian margin is not well linked to any known structures due to the sparse GPS coverage. This will potentially decrease the possibility of having
Figure 2. 7: Map of shear strain rate anomalies within the Arabian margin. This map shows two shear strain rate pattern within the Arabian margin. We correlate the dilatational strain pattern with earthquake (Mw > 4.5) distribution (using the IRIS database (www.iris.edu)) and focal mechanisms (using the Global CMT Catalog (Dziewonski et al., 1981; Ekström et al., 2012)).
Figure 2.8: Map of shear strain rate anomaly along the Dead Sea Transform. This map shows high shear strain in the vicinity of the Dead Sea transform fault. We correlate the dilatational strain pattern with earthquake (Mw > 4.5) distribution (using the IRIS database (www.iris.edu)) and focal mechanisms (using the Global CMT Catalog (Dziewonski et al., 1981; Ekström et al., 2012)).
2.5 Discussion

2.5.1 GPS Velocities

The observed GPS residuals with respect to rigid Arabia along the Arabian margin indicate a non-rigid behavior. This is seen as a systematic pattern of displacement striking in a direction NNW approximately parallel to the Red Sea. These observed displacements are robust as they are constrained by a full network solution that takes into account the motion of multiple stations. In terms of magnitude, the length of the time series may not be quite long enough to fully estimate the magnitude of the displacements. The uncertainties, however, are small enough not to significantly affect the observed overall pattern.

The velocity magnitudes of category B stations (short time series) are generally larger than nearby category A stations; however, they show a similar regional displacement azimuth within the Arabian margin. This confirms our earlier analysis that the uncertainties are small enough that the pattern of deformation stands out above the noise. However, our GPS network exhibits higher velocities than the limited UNAVCO GPS network both along the Red Sea margin and within Arabia (Figure 2.2). For instance, in central and eastern rigid Arabia, the velocity of category A stations exhibit higher magnitudes as compared with the SOLA station, which is the only UNAVCO station in
central Arabia, with respect to stable Arabia reference frame. This difference is seen at several other stations with time series extending more than 24 months.

Since our GPS uncertainties are low, the differences between our GPS velocities and UNAVCO velocities may be caused by the ITRF 2008 reference frame. Altamimi and others (2012) built their reference frame calculations on 206 GPS stations, four of them within the Arabian Plate (HALY, SOLA, BHR2, and YIBL). They emphasize the effect of large rotation rate misfit (0.083 mas/a ≈ 2.5 mm/a) between the geological models and ITRF2008, which suggests that the accuracy of the No Net Rotation ITRF2008 realization is not better than 2 mm/a. We argue that reducing our GPS velocities systematically with the amount suggested by Altamimi and others (2012) will not affect the observed pattern, rather than changing the magnitude. Since we are more interested in the observed dilatational strain, the calculated rate could be reduced.

2.5.2 Correlations with Regional Tectonics

Although the chosen grid size used in our analysis can affect the amplitude or geometry for the observed dilatational strain rate anomaly, the direction of the dilatational strain rate (Figure 2.6) is robust in terms of the chosen grid size. The dilatational strain rate shows a dependence on the HL stations and less dependence on the rest of our GPS network, including the Dead Sea Transform stations (i.e. RAMO station). In contrast, the existence of the observed shear strain rate pattern is more dependent on the chosen grid
size. Both strain rate patterns, dilatational and shear, are not influenced by our choice of the value of Poisson’s ratio value in the gpsgriddler code (Sandwell and Wessel, 2016).

These strain rate patterns (with respect to a rigid Arabia) differ from previous ideas about the Arabian margin’s present-day rigidity, and suggest there is ongoing deformation. This observed region of dilatational and shear strain rate (Figures 2.6, 2.7) can be spatially correlated with the locations of recent earthquakes and volcanic activity within the Arabian Margin, such as the Lunayyir earthquake in 2009 (Pallister et al., 2010) and the volcanism along the MMTZ, particularly along the northern segment of Rahat Harrat (Figure 2.9). The observed direction of the dilatational strain rate shows is perpendicular to the volcanic cones strike long the northern segment of Rahat volcanic field. This suggests a strong correlation between the observed dilatational strain rate pattern and long term deformation within the Arabian margin.

The northern shear strain rate anomaly (Figure 2.8) can be correlated with earthquakes in the northwestern Arabian margin, such as the Tabuk earthquake in 2004 (Aldamegh et al., 2009). Integrating both observed dilatational and shear strain rate anomalies within the Arabian margin (Figures 2.6, 2.7) shows that the region of dilatational strain is bounded by two shear strain rate regions, which suggests a transtensional system along the MMTZ. Overall, we argue for a combined effect of shear and dilatational strain within the western Arabian margin.
Figure 2.9: Main volcanic sequence within the Arabian Margin modified after Downs and others (2018). In this maps, we project the observed dilatational strain rate directions on a high resolution topography for the volcanic fields. The red dots within the volcanic field denote the volcanic cones.
The shear strain rate results as shown in figure 2.8 indicate shearing along the Dead Sea Transform and within the northwestern Arabian margin. This shearing strain may utilize pre-existing structures, such as Great Ja`dah Dike intrusions (Roobol and Stewart, 2019), and may produce earthquakes with relatively large magnitudes, such as the 5.1 Mw Tabuk earthquake on 2004 (Aldamegh et al., 2009). However, due to the large distances between the GPS stations in the northwestern Arabian Margin, it is difficult to precisely determine the details of the active shear zones.

Our results suggest ongoing deformation within the Arabian margin, particularly along the northern edge of the MMTZ. Since our GPS stations are sparse, with typically more than 200 km spacing, in the regions where these anomalous strain rates are observed, strain rate trajectories are, by necessity, extrapolated. Nevertheless, our comparison of the observed strain rate locations with the reported volcanic and earthquake events indicates a potential agreement between both sets of data. However, a denser GPS network is needed within the northwestern Arabian Margin to be able to relate the observed strain rate anomalies to specific active structures.

2.5.3 Conclusions

We have presented a new set of GPS data for the Arabian plate. This set consist of 87 GPS stations distributed across Saudi Arabia. Our calculated velocities, with respect to stable
Arabia, show a non-rigid behavior along the Red Sea margin and rigid plate motion in central and eastern Arabia. Our strain rate analysis identified a dilatational strain rate region bounded with two shear strain regions within the Arabian margin and a shear strain rate region (as expected) in the vicinity of the Dead Sea Transform. The driving tectonic forces for the observed deformation within the Arabian margin are not known and need to be evaluated using our new data set.
Table 2. 1: Table of GPS velocities with respect to stable Nubia reference frame. This table shows the GPS velocities in north (Nvel) and east (Evel) directions and their associated uncertainties (Esig and Nsig) with stable Nubia reference frame. The approximated duration for every site is provided in separate column.

<table>
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<th>Site</th>
<th>lon</th>
<th>lat</th>
<th>Evel (mm/yr)</th>
<th>Nvel (mm/yr)</th>
<th>Esig</th>
<th>Nsig</th>
<th>Period (mo)</th>
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Table 2.2: Table of GPS residuals with respect to rigid Arabia. This table lists the magnitude and azimuth of GPS residuals.

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| ES08 | 47.681829 | 27.511797 | 2.126074317 | -55.11395215 |
| ES96 | 49.565777 | 25.355789 | 3.903549667 | -21.17450637 |
| ES97 | 48.833933 | 25.848942 | 3.880070618 | -63.15758553 |
| ES98 | 48.725401 | 26.876 | 3.065047471 | 85.13390038 |
| ES99 | 49.530774 | 26.265264 | 3.499201337 | -14.54854392 |
| HL02 | 42.902927 | 28.240933 | 4.48952258 | 21.32980117 |
| HL04 | 40.818018 | 25.996758 | 9.547137791 | 9.793827358 |
| MK85 | 40.265325 | 20.132454 | 12.91491916 | -36.35615535 |
| MK86 | 40.138996 | 20.838344 | 10.98973507 | -33.78405658 |
| MK89 | 40.834274 | 21.353297 | 12.05997119 | -33.33297827 |
| MK90 | 40.397699 | 21.282321 | 11.23062674 | -19.43041304 |
| MK92 | 40.021786 | 22.084102 | 9.054114921 | -33.22495761 |
| MK93 | 39.371511 | 21.971325 | 8.181245932 | -42.73630778 |
| MK94 | 39.629403 | 22.579031 | 10.58422661 | -23.27322489 |
| MK95 | 38.842014 | 23.117595 | 9.098072378 | -37.31325124 |</p>
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**Category B**

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2.6 References


Chapter 3

On the driving forces of the Arabian Margin deformation field: A mechanical model study

Abstract

GPS stations show in Saudi Arabia high residuals within the Arabian Margin with respect to a stable Arabia reference frame, with dilatational strain along the Makkah Madinah Transtensional zone (chapter 2). The causes of these GPS residuals and the strain they represent along the Arabian Margin have not been previously understood. In this work, we use the finite element method (GTECTON code) to assess mechanical deformation processes affecting the Arabian Margin, and determine whether they could be caused by Red Sea rifting-related forces, Arabian Margin interior forces (such as the differences in the gravitational potential energy), or mantle-related basal tractions. Our modeling misfit results range from 10.2 mm/yr for the lithospheric forces to 3.45 mm/yr for basal tractions including an asthenospheric channel underneath the Arabian margin. These finite element models suggest that the GPS residuals (deformation) are not closely linked with Red Sea rift processes nor with the gravitational potential energy forces. Instead, basal tractions that relate to an asthenospheric channel along the Makkah-Madinah Transtensional Zone seem to be a potential driving force to explain the observed GPS residuals.
3.1 Introduction

The present deformation field of the Arabian plate was developed (chapter 2) by including a new set of Global Navigation Satellite System (GNSS) stations. We found that the GPS displacements differed from those expected with respect to stable Arabia along the Red Sea margin (Figure 3.1). As we concluded in chapter 2, these residuals (Table 3.1) indicate a non-rigid behavior in that region of the Arabian plate. This is seen as a systematic pattern of displacement striking in a direction NNW approximately parallel to the Red Sea. Strain rate maps, computed (chapter 2) using this new GPS derived displacement field, show a dilatational strain rate anomaly along the western and northwestern Arabian Margin (Figure 3.2) and a shear strain rate anomaly along the northwestern edge of Arabian margin. The shear strain rate is expected because it represents interplate deformation in the vicinity of the Dead Sea Transform, which is the plate boundary between the Arabia and the Sinai and eastern Mediterranean plates. Understanding the possible causes of the observed intraplate deformation is a focus of this work.

The general concept of a rigid Arabian plate was (ArRajehi et al., 2010) based on GPS observations from the UNAVCO network along the Red Sea and central Arabia. That
Figure 3.1: Map of GPS residuals. These GPS velocities are with respect to stable Arabia reference frame. The black arrows represent the observed GPS velocities. These velocities are listed in Table 3.1. The red lines represent approximate Arabian Plate boundaries, which serve as our numerical model domain boundaries. The right hand map focuses on the western Arabian Margin, where the significant GPS residuals were observed.
Figure 3. 2: Dilatational strain rate determined using GPS results in Arabian plate. This dilatational strain rate distributed along the Arabian Margin is plotted with strain rate crosses. Focal mechanisms for earthquakes (Mw > 4.5) as identified using the IRIS database (www.iris.edu) and the Global CMT Catalog (Dziewonski et al., 1981; Ekström et al., 2012) are plotted. A potential correlation may exist between the dilatational strain rate and earthquakes in central Arabian margin.
network had minimum coverage in the area of northwestern Arabia. Historical and recent earthquake activity (Aldamegh et al., 2009; Pallister et al., 2010; El-Hadidy, 2015) suggest that the western and northwestern Arabian margin and the Makkah-Madinah transtensional zone (MMTZ) are regions of active deformation (Figure 3.3). Harrat Lunayyir, one of youngest volcanic fields in western Arabia, experienced an earthquake swarm in 2009 with the largest event of magnitude 5.8 and surface fault rupture (Pallister et al., 2010). The observed fault rupture of that event is 8 km long with ~ 1 m of offset. It has a strike of 330-340°, sub-parallel to the Red Sea axis. Modeling of the InSAR observations (Xu et al., 2016) suggests that this swarm initiated with a dike injection that subsequently involved normal faulting.

During the last 13 Ma, this region of the western Arabian margin has been the site of numerous volcanic eruptions (Figure 3.3) along the MMTZ, as well as along the northwestern Arabian Margin and in the vicinity of the Sirhan rift (Bosworth and Stockli, 2016). This regional pattern of volcanic activity suggests that an underlying upper mantle perturbation may exist in this region (White and McKenzie, 1989). A possible configuration, which we will test here, is that this perturbation may be along an asthenospheric channel, that is a localized region of thinned lithosphere that coincides with the MMTZ (reported as the MMN volcanic line by Camp and Roobol, 1992; Park et al., 2007; Faccenna et al., 2013) and along the Red Sea (Park et al., 2007). Previous seismic analyses have been interpreted to indicate mantle flow beneath the Arabian Margin (Chang et al., 2011); similar flow patterns have been inferred from dynamic
Figure 3.3: Conceptual model for the Arabian margin. The MMTZ, Lunayyir Harrat, and Sirhan rift are emphasized in this map. Notice that the distribution of volcanic eruptions within the Arabian margin could be correlated to the Red Sea structural architecture changes, particularly the transition between continuous to segmented sea floor spreading.
topography models (Daradich et al., 2003). Global mantle circulation computations (Faccenna et al., 2013), which use boundary conditions inferred from seismic tomography and slab models, indicate mantle flow underneath the Arabian Margin that is slightly oblique (clockwise) to the Red Sea rift axis.

3.1.1 Local Driving Forces

The motion and internal deformation of a tectonic plate is largely controlled by the combination of forces acting on it. The overall plate motion of the Arabian plate is largely governed by the negative buoyancy of the Neo-Tethys slab (slab pull) which has subducted beneath the Eurasian plate (Bosworth et al., 2005). In this analysis, we remove the rigid plate motion of Arabia from the GPS data (ITRF 2008 reference frame, Altamimi et al., 2012). The resulting displacement/velocity field is therefore with respect to rigid Arabia. As a result, we can assume that this essentially removes the effect of plate-scale driving forces such as slab pull. This allows us to focus on the local tectonic driving forces that potentially are generating the observed deformation within the plate. We evaluate the effects of three candidate forces: (1) Edge forces (i.e. additional effects of the Red Sea rift), (2) Lithospheric body forces (i.e. driven by the horizontal gradient of the gravitational potential energy (GPE) associated with the elevated topography), and (3) Mantle-related tractions, imposed along the base of the lithosphere. The consideration of mantle traction force is motivated by patterns observed in the seismic shear wave velocity and anisotropy results, and the occurrence of Arabian Margin volcanism.
3.1.2 Edge forces ($F_E$)

The Red Sea rift is a heterogeneous extensional system, where the continuous oceanic generation of crust is observed in the southern portion, discontinuous patches of oceanic crust separated by strike slip faults are reported along the central Red Sea, and crust of an unknown nature is found along the northern part of the Red Sea (Cochran and Martinez, 1988). Szymenski and others (2016) found, based on fission track analysis, that there is no evidence of an increase in geothermal gradient prior to the Red Sea Rift initiation, and concluded that Red Sea rifting was driven by far field extensional forces. While most rift systems are characterized by elevated axial heights (e.g. mid-Atlantic ridge) that exert horizontal forces (“ridge push”) perpendicular to the rift axis, the Red Sea appears to be an evolving rift system, where the mid-oceanic ridge is still forming (Ligi et al., 2015). The spreading center is characterized, instead, by an axial trough extending along the southern and central portions of the rift. Thus, the “push” force exerted by the Red Sea “ridge” is insignificant. Nevertheless, there could be an edge force, defined by basal tractions produced by mantle flow underneath the rift axis, driving additional deformation.

3.1.2.1 Lithospheric Body Forces ($F_B$)

In addition to the edge forces associated with the spreading center, there are two other forces, the horizontal tractions associated with (a) the lithospheric gravitational potential
energy (GPE) differences, and (b) horizontal basal tractions driven by mantle flow (Ghosh et al., 2009). Such lithospheric body forces are considered the principal driving force in the East African rift system (Stamps et al., 2015), which appears to be dynamically driven and supported by a mantle superplume (Nyblade and Robinson, 1994; Lithgow-Bertelloni and Silver, 1998). This model, as developed for the East African rift extension, is built on the fact that the Somalian plate is not attached to a slab that could exert additional pull forces (McClusky et al., 2010), and thus there is a need for an additional force to drive extension. In contrast, since the Arabian plate is attached to the Neo-Tethys slab that has subducted beneath the Eurasian plate, there are possible sources for far field forces to drive the motion of the Arabian plate (Bosworth et al., 2005, references therein). In this case, we investigate what role, if any, these GPE forces could have in producing the observed intra-plate deformation.

3.1.2.2 Basal Tractions ($F_M$)

Distributed basal shear tractions can also act as driving forces for large plates (Bird et al., 2008). These basal tractions are inferred to play a major role in various tectonic systems around the world (Ghosh and Holt, 2012) and in most cases, these global basal tractions are inferred to be driving forces rather than resistive (Ghosh et al., 2013). For Arabia, such mantle tractions acting at the base of the lithosphere (assuming lithospheric thickness to be approximately 100 km) would exert a horizontal traction of about 10 MPa in the vicinity of the Afar mantle plume (Wang et al., 2015). We infer the geographical
Figure 3.4: Asthenospheric channel domains as reported by seismic tomography models. Yellow outline denote a regional anomaly with more than 100 km. Green outline represent a more localized channel with shallower depth. It is important to note that seismic stations coverage was improved significantly for the period between 2008 to 2019.
extent of basal tractions in the vicinity of the Arabian plate from seismic tomography maps (Figure 3.4), which map regions of thinned lithosphere or warm mantle (Park et al., 2007; Tang et al., 2019) that indicate possible flow direction. Figure 3.4 indicates two low shear-wave-velocity anomalies, one is deep regional anomaly that aligns with the MMTZ, with an inferred thickness of the asthenospheric channel ranging between 70-190 km (Park et al., 2007; Park et al., 2008; Tang et al., 2019). The other shear wave velocity anomaly is shallow (25 km) and localizes in the northern segment of the MMTZ (Tang et al., 2019).

The orientations of the basal tractions are defined by the published shear wave splitting azimuths. The shear wave splitting database for Arabia (Qaysi et al., 2018) shows a uniform pattern (Figure 3.5) within the central and southern Arabian margin. These azimuths are not confined along the reported channel, instead, they are distributed within the Arabian plate. Although the observed shear wave splitting azimuths can indicate either an anisotropic medium or mantle flow, Faccenna and others (2013) reproduced these azimuths by mantle convection flow models with an asthenospheric channel beneath the Arabian margin. Furthermore, the conversion of the observed azimuths into specific basal tractions requires a sublithospheric topographic feature (Höink et al., 2012). We argue that a regional mantle flow, even if independent of the Afar plume (Konrad et al., 2016), can produce basal tractions along a preexisting asthenospheric channel (Figure 3.4) beneath the Arabian margin.
Figure 3.5: Shear wave splitting database for the Arabian plate (Qaysi et al., 2018). These velocities are reported in part as well by Hansen and others (2006), and Chang and others (2011). The shear wave splitting velocities show a regional northward pattern.
In summary, we explore the potential for various tectonic driving forces, including the gravitational potential energy of the Arabian Margin, Red Sea rifting forces, and Mantle-related tractions to produce the observed patterns of deformation in western Arabia. We can test the importance of each potential mechanism by computing the misfit between the observed GPS displacement residuals (Figure 3.1) and our predicted velocities. We can also compare our model generated stress field with the geologic structures, distribution of volcanism, and earthquake focal mechanisms within the Arabian Margin.

3.2 Modeling

Understanding the processes driving intraplate deformation is difficult; particularly in regions where one needs to disentangle different tectonic components. Several approaches have been developed to tackle this problem (Nijholt et al., 2018; Warners-Ruckstuhl et al., 2013; Plattner et al., 2009). Their general approaches use mechanical models to quantify the deformation resulting from the tectonic forces. The key element of their models depends on the goal of their study (e.g. resistance to slip on major regional fault zones (Nijholt et al., 2018)). Here, we follow these approaches in the sense of testing each tectonic force one at a time to evaluate their effect on the observed deformation.
3.2.1 Force Models

In the discussion above of potential driving forces for the Arabian margin (section 3.1.1), we excluded edge forces from our potential candidates and indicated we would focus our analysis on lithospheric body forces and basal tractions. Here, we develop models of the effects of these two force components, independent of each other, to test their potential role in producing the observed deformation.

3.2.1.1 Lithospheric Body Forces Model

Gravitational potential energy (GPE) describes the potential energy of a unit mass in a field of gravitational attraction. In a tectonic setting, the lithospheric gravitational potential energy is the product of lithospheric thickness, density and gravitational acceleration. Higher topographic elevation produces larger potential energy values. The spatial gradient of GPE values within a plate can produce a traction field that acts as a lithospheric body force, acting in the direction of maximum gradient.

The gravitational potential energy (GPE) for the Arabian Margin (Figure 3.6) is calculated following the approach proposed by Warners-Ruckstuhl and others (2012). This approach assumes isostatic equilibrium with a uniform plate thickness throughout the Arabian Plate. Although the isostatic assumption is not strictly met and the thickness varies, this approach still provides a useful estimate for the GPE of the Arabian Margin (Figure 3.6). The GPE-related tractions that would drive deformation are derived from
Figure 3.6: Map of GPE field and derived tractions. The calculated GPE is projected on our model domain mesh. On the top of the GPE field we plot the GPE derived forces in black vectors as computed by taking the horizontal derivative of the GPE in the direction perpendicular to the Red Sea. These GPE forces show a gradual increase southward along the Red Sea, as correlated with the topographic elevation in which it reaches the highest peak in the southern Arabian margin with ~3000 m.
the horizontal gradients of the GPE in reference frame orthogonal to the Red Sea and shown in black vectors on the GPE map (Figure 3.6).

### 3.2.1.2 Basal Traction Model

Basal tractions were approximated using dynamic uplift calculations, assuming that the uplifted margins are driven by the mantle basal tractions (Ghosh et al., 2013). Although these basal tractions can play an important role at global tectonic scales (Bird, 1988), their effect in smaller scales requires further investigation with additional constraints, such as the sub-lithospheric topography, asthenospheric flow, and strong lithosphere that allows these basal tractions to propagate to the upper plate. The interaction among these conditions controls the overall effectiveness of the basal tractions in producing upper plate deformation.

The potential for lithospheric thinning under the MMTZ and channeling of sub-lithospheric mantle underneath western Arabia is supported by previous research (Park et al., 2007; Faccenna et al., 2013; Tang et al., 2019; Yao et al., 2017; Koulakov et al., 2016). The thickness of the asthenospheric channel ranges between 70-190 km depth (Tang et al., 2019), extending northwardly across western Arabia with a width of 200 km as inferred from shear wave velocity models (Figure 3.5; Chang et al., 2011) or corresponding volcanism (Camp and Roobol, 1992). Mantle flow is assumed to
Figure 3.7: Basal tractions distribution that provides the best fit with our GPS velocities. This map in agreement with the Asthenospheric channeling proposed by Park and others (2007) and Faccenna and others (2013). The basal tractions were derived based on the GPS residuals (with respect to rigid Arabia) best fit, which describe the basal tractions residuals along the Red Sea. This suggests western Arabia slowing down as compared to the rest of Arabia as the whole plate is dragged to the east by the Neo-Tethys slab pull.
propagate northward through the asthenospheric channel as indicated by the observed shear wave splitting azimuth (Hansen et al., 2006; Qyasi et al., 2019).

We implement this asthenospheric channel in our model (Figure 3.7). This channel, as inferred from seismic tomography and aligned with the volcanic provinces, accommodates basal tractions that are parallel to the observed shear wave splitting. To test the effect of the basal tractions along this channel and the Red Sea margin on the observed GPS residuals, we adjust the magnitude and direction of the associated tractions to provide a best fit. We can then compare the model conditions that produce this “best fit” model with tectonic constraints.

3.1.1 Model Setup

To address the role these forces could play in the observed deformation, we develop a suite of numerical deformation models for the Arabian Plate, including our area of interest (Figure 3.1, inset). We use the GTECTON platform, which is a finite element modeling code (version 2017.3; Govers and Wortel, 1993; Govers & Meijer, 2001; Özbakir et al., 2017), to investigate the mechanical deformation processes along the Arabian margin and test their driving forces. We use a viscoelastic spherical shell with discrete faults and a triangular element mesh. We represent the model domain with a homogeneous Maxwell viscoelastic material (Young’s modulus = 7 GPa; Poisson’s ratio = 0.30; lithospheric viscosity $\eta = 1e19$ Pa s). Since we run our model for one year to
produce a set of velocities that can be compared with the observed GPS residuals, the and Maxwell relaxation time for the chosen lithospheric viscosity is 10 years, our results are not dependent on lithospheric viscosity values higher than 1e19 Pa s. This is, although we use a linear viscoelastic rheology to provide a time domain for the applied tectonic forces, the short time duration of our models results in minimal viscous relaxation and our results are essentially fully elastic.

Our model domain (Figure 3.1) is defined by the Arabian plate boundaries, as bounded by the Red Sea rift system (to the west), the Dead Sea Transform (to the northwest), Zagros collision zone (to the east), Anatolia collision zone (to the north), and the Gulf of Aden (to the south). We are interested in western Arabia, which is highlighted in the inset (Figure 3.1) where the GPS residuals are observed, and so we constrain our misfit calculations to the western Arabia region. Our boundary conditions are fixed (i.e. no displacement), as we are focused on the causes of the internal deformation in an otherwise stable Arabia reference frame. The proposed tectonic forces (section 3.2.1) are implemented in our models as initial conditions.

The model is a two-dimensional plane stress spherical shell, which is assumed to be thin in the vertical direction. We compute velocity, stress (σ) and strain (ε) rates using the mechanical equilibrium equations (Eqs. 3.1, 3.2, 3.3, and 3.4) for plane stress:
\[ \dot{\varepsilon}_{xx} = \frac{1}{E} (\dot{\sigma}_{xx} + \nu \dot{\sigma}_{yy}) + \frac{1}{6\eta} (2\sigma_{xx} - \sigma_{yy}) \quad \text{Eq. 3. 1} \]

\[ \dot{\varepsilon}_{yy} = \frac{1}{E} (\dot{\sigma}_{yy} + \nu \dot{\sigma}_{xx}) + \frac{1}{6\eta} (2\sigma_{yy} - \sigma_{xx}) \quad \text{Eq. 3. 2} \]

\[ \dot{\varepsilon}_{xy} = \frac{1+\nu}{E} (\dot{\sigma}_{xy}) + \frac{1}{2\eta} (\sigma_{xy}) \quad \text{Eq. 3. 3} \]

\[ \dot{\varepsilon}_{zz} = \frac{-\nu}{1-\nu} (\dot{\varepsilon}_{xx} + \dot{\varepsilon}_{yy}) \quad \text{Eq. 3. 4} \]

These mechanical equilibrium equations for plane stress are solved using the finite element modeling platform GTECTON. This platform assumes that each individual element is a thin plane, and all of them are connected together to describe the outline of a spherical shell (Govers & Meijer, 2001). In the above equations, E is Young’s modulus, ν denotes Poisson’s ratio, η indicates lithospheric viscosity, and the dot indicates rates.

The tectonic forces are converted from three dimensions into two dimensions by traction boundary conditions. This assumption is adopted by numerous researchers as adequate to capture the lithospheric behavior (Özbakir et al., 2017 and references therein). In pursuit of simplicity, we do not account for the variations in geology and their associated mechanical heterogeneity, nor the variations in the crustal thickness, in spite of the fact that lithospheric thinning was reported along the MMTZ and the Red Sea Margin (Tang et al., 2018).
3.1.2 Model misfit

We consider the surface GPS velocities as representative for the entire lithospheric thickness. We determine our data (model) misfit to the observed GPS velocities using the weighted root mean square (WRMS) misfit (Eq. 3.5) as used by Stamps and others (2015):

\[
WRMS = \sqrt{\frac{\sum_{i=1}^{N} \left( \frac{(v_o - v_m)^2}{\sigma^2} \right)}{N-1}}
\]

Eq. 3.5

Where \( \sigma \) is the uncertainty on the GPS velocities, \( v_m \) is the predicted velocity, \( v_o \) the observed velocity, and \( N \) the number of observations. The WRMS quantifies the average over \( N \) stations. A perfect fit at all stations would give \( WRMS = 0 \) mm/yr.

To test our modeling tool, we calculate the misfit between rigid Arabia motion with respect to stable Nubia, as calculated by ITRF 2008 (Altamimi et al., 2012), and our observed GPS velocities (chapter 2) with respect to stable Nubia reference frame. The resulting misfit is 9.5 mm/yr, which will be used as reference value for the following misfit calculation. This misfit is notably high in comparison with other tectonic systems (Stamps et al., 2015). This test elaborates our previous conclusions (chapter 2) that the Arabian Margin exhibits higher GPS residuals along the Red Sea Margin, while the rest of the Arabian plate seems to be rigid. As we test each potential driver, we will compute the WRMS for each force component, to evaluate their effect on the observed deformation.
3.2 Results

3.2.1 Lithospheric Body Forces ($F_B$)

To test the effect of GPE variations throughout the Arabian Plate, we impose the GPE derived forces on the model as boundary conditions. Although GPE tractions are not technically shear tractions, in the plane stress approximation, this way of dealing with a lithospheric body forces is valid (Nijholt et al., 2018). Here, we attribute the uplifted margin to the GPE buoyancy rather than mantle driving forces.

In figure 3.8, predicted velocities derived from the GPE forces are orthogonal to the Red Sea. The magnitude of these velocities decreases northwardly along the Red Sea strike and easterly toward the plate interior. These GPE velocities generate a WRMS misfit of 10.2 mm/yr, which is even larger than our original misfit reference value (9.5 mm/yr). This is expected, since the forces associated with the GPE are perpendicular to the Red Sea rift axis (Figure 3.6), in the direction of maximum gradient, while the observed GPS residuals are semi-parallel to the Red Sea axis. These analyses suggest that these GPE forces are not the primary contributor to the observed GPS residuals along the Arabian Margin.
Figure 3.8: This map shows the misfit between the observed GPS residuals in black and our predicted GPE velocities in blue. As our model aims to test the sensitivity of our velocities to the GPE forces that exerted due to the topographic changes along the Red Sea. The total misfit is 10.2 mm/yr as computed by the weighted root mean square method, which suggests that the GPE forces are not significant.
3.2.2 Mantle Basal forces ($F_M$)

The magnitude of mantle basal tractions and direction along the proposed asthenospheric channels (Figure 3.7) that produce the best fit with our GPS residuals were found through iterative forward modeling (Figure 3.9). The calculated WRMS misfit for this preferred model is 3.45 mm/yr, which is lower than the reference misfit value (9.5 mm/yr). This best fit suggests that the largest basal tractions exist along the Red Sea. A second set of tractions exist within Arabia, along the previously inferred asthenospheric channel that is localized beneath the MMTZ. This set of tractions are acting northwardly along the channel.

The existing of an asthenospheric channel beneath the Arabian margin incorporating the basal tractions described above generates an extensional strain anomaly within western Arabia. The extensional strain rate value is dependent on the basal traction contrast between the basal tractions along the Red Sea and those along the asthenospheric channel. The principal axes of the predicted extensional strain are dependent on the azimuth contrast between basal tractions along the Red Sea and within the Arabia margin.
Figure 3.9: Predicted basal tractions velocities map compared with the observed GPS residuals. The GPS residuals are plotted in black and predicted velocities in blue. The predicted velocities are calculated on the basis of basal tractions in Figure 3.7. The WRMS misfit is 3.45 mm/yr, which is reduced significantly in compare to the original misfit.
Our predicted velocities show a northward gradual decay in the magnitude of our basal tractions along the Red Sea margin. This implies that the Dead Sea Transform is tectonically independent of these proposed basal tractions. Moreover, these basal tractions decay gradually as well into the plate interior. This limits the mantle flow effect to western Arabia, where most of the present day deformation is taking place. Although the basal tractions domains are separated with a sharp contrast, a series of iterative forward models indicate (via the calculated misfit) this is our preferred geometry.

We validate our basal tractions model by comparing the predicted minimum horizontal stress (SH-min) field with other observed deformation indicators. Unfortunately, the Arabian margin lacks stress indicator data. Nonetheless, stress trajectories within the Arabian Margin have been interpreted by Bosworth and Durocher (2017), with the SH-max directions defined from the alignment of volcanic provinces. However, this interpretation does not quantify the stress field. It is classified as a low precision data category (Heidbach et al., 2008).

Our model results, with the thin-sheet approximation, show that the integrated deviatoric stresses approximate the stress values and directions on the upper plate. The predicted SH-min (Figure 3.10) is sampled every 1.5 degrees to smooth out irregularities and capture the regional trend. The stress field shows extensional stresses along the Red Sea, particularly in the northern Arabian margin, while the MMTZ shows transtensional stresses. The magnitudes of our predicted stresses, similarly to our basal tractions, decrease as we approach the plate interior or Dead Sea Transform.
Figure 3.10: \( \text{SH-min} \) stress field as calculated on the basis of basal tractions model (Figure 3.7) and the resulting velocity field (Figure 3.9). These stresses are denoted with crosses on the top of the Arabian margin map. Cenozoic volcanic fields designated with purple color. The extensional stresses distributed in the western and northwestern Arabian margin, while compressional stresses seem to localize along the southern Arabian margin. The comparison between predicted \( \text{SH-min} \) stress patterns with the Arabia Margin volcanism and recent earthquake focal mechanism highlights the role of the asthenospheric channel underneath western Arabia on the upper plate deformation. In particular, the transtensional stresses along the MMTZ line, extensional stresses in the northwestern corner of Arabia, and the compressional stresses in the vicinity of highest topographic elevation in the southwestern Arabia.
3.3 Discussion

In this work we have evaluated a set of possible forces that potentially could produce the observed deformation in western Arabia (chapter 2). We have compared the observed GPS velocities with our model predicted velocities, and determined the misfit associated each tectonic force candidate. Here, we will assess the relative importance of each of to the potential driving forces.

3.3.1 The Potential Tectonic Driving Force

In our models (3.2), we tested two possible forces, the lithospheric body forces associated with GPE and basal tractions. Our misfit calculations show a residual WRMS misfit of 10.2 mm/yr for the lithospheric body forces and 3.45 mm/yr for the basal tractions (Figure 3.11). Basal (Figure 3.7) show an alignment with the reported shear wave splitting azimuth (Figure 3.5) within Arabia (Qyasi et al., 2019, references therein), and they exhibit westward deviation with respect to the inferred azimuths along the Red Sea. This implies that the cause of the Arabian Margin deformation can be that western Arabia motion is potentially slower than the rest of Arabia, allowing an extensional zone along the MMTZ.
Figure 3.11: GPE and basal traction results compared with observed GPS residuals. This map made to correlate both velocities to provide the best fit with the observed GPS residuals. The GPE predicted velocities denoted in red, the basal traction in blue, and the observed GPS residuals in black. The scale is 1 cm applied for the all plotted velocities.
tractions along the Red Sea and within a channel beneath the western Arabian plate appear to be the primary driving force for the observed deformation. These basal tractions

Our GPE calculations show that the GPE forces are not acting in the direction of the observed GPS residuals (Figure 3.1). This precludes the likelihood that this lithospheric body force is the main driving force for the observed deformation. Nevertheless, the GPE tractions may act as a supplementary role in addition to the proposed basal tractions (Figure 3.11). In figure 3.11, the addition of GPE and basal tractions velocities could potentially contribute to a better fit with the observed GPS residuals in the asthenospheric channel region.

In figure 3.7, the basal tractions along the Red Sea margin are not parallel to the inferred shear-wave-splitting azimuths (Figure 3.5), while the basal tractions along the proposed asthenospheric channel show strong alignment. The large misfit along the Red Sea could be attributed to the reference frame of basal tractions. To clarify, as we calculate the basal tractions based on the GPS residuals with respect to stable Arabia, the misfit with shear wave splitting azimuth along the Red Sea margin could be attributed to a mantle flow residuals rather a regional mantle flow. In other words, a regional mantle flow along the Red Sea could be correlated with the observed GPS velocities with stable Nubia reference frame, in which they are parallel to the observed shear wave splitting azimuths.
3.3.2 Sensitivity of Basal Traction

Having an asthenospheric channel in our model domain was important in producing the extensional strain rate anomaly within the Arabian margin. In order to evaluate the sensitivity of the assumed geometry and structure of the asthenospheric channel, we assess a set of conditions. Our predicted velocities are sensitive to the asthenospheric channel geometry, and the magnitude and direction of basal tractions. Since the asthenospheric channel geometry in our model was defined by previous seismic studies (Park et al., 2007; Faccenna et al., 2013), and the flow direction was defined by the observed shear wave splitting direction (Qyasi et al., 2018; Hansen et al., 2006), the primary adjustable parameter is the magnitude of the proposed basal tractions. We chose these values to provide the best fit for our GPS velocities by iterative forward modeling. Hence, our predicted dilatational strain rate in western Arabia is directly related to our choice of the magnitude of the basal tractions. As illustrated in chapter 2, although the duration of our GPS stations is short, their time series exhibit low uncertainties. Nevertheless, if their magnitudes are subjected to higher uncertainties, their observed displacement pattern along the Red Sea strengthen our data set and potentially suggests lower strain rates instead of questioning the existence of the strain anomalies. From our modeling we can infer that the existence of the dilatational strain anomaly in western Arabia is dependent on the asthenospheric channel geometry and the azimuth of basal tractions rather than their magnitudes.
We have tested the effects of the geometry (azimuth and width) of the asthenospheric channel width and azimuth to evaluate their effect on the observed GPS residuals. Although, the GPS coverage along the asthenospheric channel region is limited, we find that increasing the width of the asthenospheric channel effects the velocities azimuth of the surrounding stations (i.e. QS02, MD15, and MD03), in which they deviate toward the channel. While changing the asthenospheric channel azimuth has less effect because of the sparse coverage of our GPS network in the vicinity of the proposed asthenospheric channel.

### 3.3.3 Lunayyir Harrat

In 2009, an earthquake swarm struck Lunayyir Harrat with a surface fault rupture with a strike of 330-340°, that is sub-parallel to the Red Sea axis, Red Sea Tertiary dikes, and the volcanic cones alignment within Lunayyir Harrat (Pallister et al., 2010). More recently, a mantle seismicity swarm was recorded in 2014, with fault strike of N103E (Blanchette et al., 2018). These events occurred below the Moho, and strike differently from the 2009 earthquake swarm.

Our predicted stress field (Figure 3.10) suggests two stress domains associated with main driving forces, the Red Sea and the MMTZ asthenospheric channel. The upper plate deformation, as seen in the surface rupture of 2009 swarm suggests a Red Sea rifting dominance, as the normal faulting strike is parallel to the rift axis. The deeper plate
deformation, as represented in the 2014 mantle seismicity, suggests an asthenospheric channel dominance based on their strike. Two dynamic scenarios are possible, (1) upper plate deformation may have localized along the Tertiary dikes and preexisting structures (Roobol and Stewart, 2019; references therein), resulting in a stress field that is not changing significantly over depth. (2) Alternatively, the stress field is changing over depth, and shallow deformation is driven by Red Sea processes and deeper deformation is driven by the basal tractions along the asthenospheric channel. Nevertheless, neither hypothesis can be tested here, more constraints are needed for future investigations with three dimensional models.

3.4 Implications

Our predicted SH-min field (Figure 3.10) shows good agreement with the long-term deformation, which is represented in the spatial distribution of the volcanism within the Arabian Margin. In detail, the predicted SH-min stress field exhibits a transtensional pattern along the Makkah-Madinah transtensional zone (MMTZ), which is in agreement with what Camp and Roobol (1992) have already reported in light of the volcanic eruptions along the MMTZ. Moreover, the predicted extensional stress pattern along the Red Sea is in agreement with the recent earthquake focal mechanisms in western and northwestern Arabian margin (Aldamegh et al., 2009).
3.4.1 Nature of Asthenospheric Channel Coupling

Understanding the nature of coupling between asthenospheric flow and upper plate deformation is an important component of this analysis. Höink and others (2012) concluded, based on their numerical models, that asthenospheric channeling can increase in-plate stresses as it constrains the orientation of adjacent sub-lithospheric flow. Moreover, besides sub-lithospheric topography, lithospheric mantle viscosity can play a significant role in the degree of coupling between the lithosphere and asthenosphere (O’Neill et al., 2010).

Recent work relates basal shear tractions with low viscosity channel thickness (Höink et al, 2012) where thinner viscous layer thickness will likely promote higher basal shear stress. This may imply decoupling the lithosphere and asthenosphere along the low viscosity channel. Hence, a lower viscosity asthenospheric channel as reported by Park and others (2007), beneath the MMTZ, would lead to a relative lack of coupling. Nonetheless, testing this nature of coupling requires dense network of stations along the MMTZ, however, we have few GPS stations coverage along the volcanic provinces, and so the nature of coupling remains as an open question.

3.4.2 Tectonic Consequences of the Asthenospheric Channel Existence

Our basal tractions geometry is designed based on reported seismic tomography results for the Arabian asthenospheric channels (i.e., Camp and Roobol, 1992; Park et al., 2007;
Faccenna et al., 2013). Although our basal tractions domain is bounded with a sharp contrast, the calculated misfit seems to prefer this geometry. This suggests that the asthenospheric channel formed as recent as the Cenozoic. This conclusion is supported by recent seismic reflection profiles in northwestern Arabian basin that question the presence of a long lived Arch (sub-lithospheric topographic feature) in the western Arabia (Khalil, 2016). These observations and analyses open the possibility for a new geodynamics hypothesis that explains the formation of the MMTZ in light of the Red Sea rifting evolution. What makes this a difficult task is the absence of surface structural expressions for this channel, not to mention the lack of seismic cross-sections which can be used to validate our numerical models. Deciphering the asthenospheric channels formation needs a comprehensive approach that integrates the lithospheric thermal evolution timing constraints with the spatial and temporal evolution of the tectonic forces around the Arabian Plate. These analyses are beyond the scope of this chapter, a new hypothesis with illustrated figures and numerical models will be proposed in Chapter 4.

3.5 Conclusions

In conclusion, our calculations of model misfit suggest the observed GPS residuals are driven by the combination of the GPE and basal tractions. These basal tractions, which considered the primary driving force, localize along an existing asthenospheric channel that localizes the proposed basal tractions. The asthenospheric channel geometry defined based on reported seismic tomography models. The direction of these tractions within
Arabia are chosen based on the observed shear wave splitting azimuth. However, their
direction along the Red Sea margin deviates from the observed shear wave splitting
towards the plate boundary. These results describe the first order causes of the observed
deformation within the Arabian margin. A denser GPS network is needed to expand our
investigation into smaller scales within the deformation zones.
Table 3.1: List of the GPS residuals used in chapter 3 with respect to rigid Arabia. We limited our misfit calculation to the GPS stations listed here and highlighted in our domain as shown in the inset of Figure 3.1.

<p>| Site | lon   | lat   | |V| (mm/yr) | Azimuth |
|------|-------|-------|----------|----------|
| HL02 | 42.902927 | 28.240933 | 4.48952258 | 21.32980117 |
| HL04 | 40.818018  | 25.996758  | 9.547137791 | 9.793827358  |
| MK85 | 40.265325  | 20.132454  | 12.91491916 | -36.35615535 |
| MK86 | 40.138996  | 20.838344  | 10.98973507 | -33.78405658 |
| MK89 | 40.834274  | 21.353297  | 12.05997119 | -33.33297827 |
| MK90 | 40.397699  | 21.282321  | 11.23062674 | -19.43041304 |
| MK92 | 40.021786  | 22.084102  | 9.054114921 | -33.22495761 |
| MK93 | 39.371511  | 21.971325  | 8.181245932 | -42.73630778 |
| MK94 | 39.629403  | 22.579031  | 10.58422661 | -23.27322489 |
| MK95 | 38.842014  | 23.117595  | 9.098072378 | -37.31325124 |
| MK96 | 39.113472  | 22.280506  | 10.72985   | -43.48571048 |
| MK97 | 39.242909  | 21.556079  | 12.17326949 | -32.44592324 |
| MK98 | 39.540059  | 21.45125   | 12.40761137 | -36.94747659 |
| RY07 | 43.919548  | 23.403861  | 9.403474092 | -25.1070766  |
| RY09 | 44.889996  | 20.423727  | 10.12579681 | -39.79542832 |
| RY85 | 44.759185  | 23.760308  | 6.394578485 | -31.62452504 |
| TB05 | 36.368634  | 26.366944  | 9.026373857 | -38.18396533 |</p>
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Chapter 4

Early Red Sea Multiphase Rifting: Spatial and temporal extensional changes along the Red Sea rift

Abstract

The occurrence of earthquakes and recent volcanic activity along and within the Arabian margin can be associated with basal tractions along asthenospheric channels beneath the Red Sea and Makkah-Madinah Transtensional Zone (MMTZ) (chapter 3). How such asthenospheric channels formed beneath the Arabian margin remains unclear. Here we explore mechanisms that could lead to formation and evolution of the Red Sea rift and MMTZ in light of regional tectonic forces and local structural architecture. We use GTECTON, a finite element modeling package, to explore the effects of boundary conditions, as representative of far field forces, on deformation along the Red Sea and within the Arabian margin. These finite element models suggest that the development of the northern Red Sea rift and the MMTZ can be linked to a competition between two regional effects: (a) extension associated with the Aegean rift driven by African slab rollback, and kinematics of the eastern Mediterranean block during the last 40 Ma, and (b) evolution of the southern Red Sea rift from a passive to an active rift. By focusing on the period between 30 Ma to 20 Ma, we find three rifting stages:
1. ~27 Ma: MMTZ formed as a result of slowing motion of Africa due to a reduction in the African slab pull. In this stage, Sirhan and southern Red Sea rifts are characterized by passive rifting;

2. ~25 Ma: diffuse extension along the northern Red Sea is driven by differential motion between the Aegean block and eastern Mediterranean, while the southern Red Sea and Sirhan rifts remain passive, and

3. ~20 Ma: northern Red Sea localization as a result of the southern Red Sea evolving from a more mature rifting system and the forming of the northern Red Sea by northward propagation of rifting.

These stages indicate that early Oligocene to early Miocene reorganization of regional forces and kinematics produce the changes in deformation styles within the Arabian margin seen along the central and northern Red Sea.

4.1 Introduction

The Arabian Plate became independent of the Nubian plate with the formation of the Red Sea Rift about 21-23 Ma ago (Bosworth et al., 2005) after a period of more than 10 Ma of regional extensional deformation along the northern Red Sea. The Arabian plate consists of two main crustal blocks, which meet along a boundary that is oriented ~ north-south and is located east of the Central Arabian Magnetic Anomaly (Thomas et al, 2015). The western block basement exposures are considered to be the Arabian Shield while the eastern block is covered by a Phanerozoic sedimentary sequence and is exposed only in Oman (Johnson,
2011). The Arabian Shield outcrops reflect an upper crust that is composed of crystalline Precambrian basement, and Cenozoic alkalic volcanic fields with small sedimentary basins and grabens (Stern and Johnson, 2010).

Rigid motion of the Arabian plate (Figure 4.1), as determined by the MORVEL plate motion model (DeMets et al., 1990), shows extension along the Red Sea at a rate that ranges between 7 mm/yr to 14 mm/yr from north to south (ArRajehi et al., 2010), sinistral motion along the Dead Sea Transform at an average rate of 7 mm/yr, spreading along the Gulf of Aden at an average rate of 20 mm/yr, convergence between Arabia and Eurasia at an average rate of 20 mm/yr, subduction beneath the Makraan margin at an average rate of 20 mm/yr, and collision/extrusion along Anatolia at an average rate of 10 mm/yr. These motions were calculated based on the assumption of a rigid Arabia, which was initially proposed (ArRajehi et al., 2010) on the basis of observations from the UNAVCO GPS network along the Red Sea and central Arabia. However, in this network, which spanned the whole width of Arabia – about 2000 km – GPS stations were approximately 800 km apart, leaving the region of western Arabia and the active rifting poorly covered. Since 2015, the Saudi General Survey Commission started deploying GPS stations within Saudi Arabia. At the time we were granted access to the new data set it which had more than 100 stations. By analyzing data from these GPS stations (chapter 2) we found a discernable deviation with respect to stable Arabia along the Red Sea margin. These residuals indicate non-rigid behavior in that region of the Arabian plate seen as a systematic displacement striking in a direction NNW, approximately parallel to the Red
Figure 4.1: Modified regional tectonic map after Kumar (2008). This map highlights the boundaries of the Arabian plate and their tectonic nature. We plot the rigid plate motion along these boundaries with respect to stable Africa (DeMets et al., 1990). The Arabian plate as illustrated on the map is bounded by the Red Sea rift to the west, Gulf Of Aden and Owen F.Z to the south, the Makraan subduction zone and Zagros Collision zone to the east, Dead Sea Transform to the northwest, and Anatolia extrusion to the north.
Sea. Dilatational and shear strain were observed using this GPS data in this region of western Arabia (chapter 2). Present-day earthquakes and recent volcanic activity within the Arabian margin (e.g. Rahat Harrat) show a strong correlation with the observed dilatational strain rate anomaly (chapter 2). This relationship motivated us to investigate the driving forces for the observed deformation in chapter 3. We found that when an asthenospheric channel, which is correlated geographically with the Makkah-Madinah Transtensive Zone (MMTZ), is introduced (Figure 4.2) into our model, we localize the basal tractions and reproduce the observed dilatational anomaly. Basal tractions within this asthenospheric channel provide a primary driving force for the observed GPS residuals.

The recent tectonic activity can be linked with the tectonic history of the region. In particular an inherited region of diffuse extension formed during the early northern Red Sea extensional processes coincides with the current region of non-rigid behavior (Figure 4.3). This region of diffuse extension, which is observed in the form of dikes, grabens, and volcanic eruptions, falls within an area that is bounded to the east by the Sirhan Rift, to the west by the Nile river, to the south by the MMTZ and to the north by the Mediterranean oceanic lithosphere. Although the diffuse extensional processes ceased 15 Ma ago (Szymanski et al., 2016), the inherited weakness appears to still play a significant role in the present-day lithospheric strength and, therefore, in the patterns of deformation. Most present-day earthquakes and the areas of recent volcanic activity (e.g., Uwayrid, Ishara, and Lunayyir Harrats) are distributed within the northwestern Arabian Margin and along a
Figure 4.2: Proposed basal tractions geometry map. This map shows the needed basal traction distribution that provides the best fit with our GPS velocities (chapter 3). This pattern is in agreement with the asthenospheric channeling proposed by Park and others (2007) and Faccenna and others (2013). The basal tractions were determined based on the best fit to the GPS residuals (with respect to rigid Arabia). This suggests western Arabia is slowing down as compared to the rest of Arabia as the plate dragged to the east by the Neo-Tethys slab pull.
diffuse set of Red Sea syn-rift structures (e.g., the Tabuk earthquake (Aldamegh et al., 2009) and the Lunayyir earthquake (Pallister et al., 2010)). Recently, Szymenski and others (2016) attributed the diffuse extension to the decoupling between an asthenosphere channel beneath the MMTZ, which is believed to be a long-lived structure formed during the early Paleozoic (Coleman, 1974), and the Arabian Plate lithosphere. However, results of some recent seismic reflection profiles question the presence of the Hail Arch in the northwestern Arabian basin (Mesbah, 2016), which suggests that this asthenospheric channel formed in the Cenozoic and should be explained in light of the evolution of tectonic forces of the Arabian plate boundaries. This leads to our new hypothesis that incorporates the history of diffuse extensional processes with an asthenospheric channel formed contemporaneously with the early Red Sea rifting, and in particular links between the diffuse extension evolution to MMTZ formation.

With this perspective, our key question becomes, what tectonic process could lead to the formation of this asthenospheric channel, which serves to localize the basal tractions that drive the observed present-day deformation within the Arabian margin, and is this tectonic process linked to the fundamental change in extensional style along the Red Sea? In this chapter, we investigate the implications of asthenospheric channel formation for the tectonic evolution of the Red Sea rift over the last 30 Ma, particularly the northern Red Sea region of diffuse extension. We focus on the geodynamic evolution of the Arabian margin within the context of the regional tectonic forces, Arabian margin volcanism, and Red Sea rifting.
4.1.1 Regional Tectonic Forces

The Arabia-Eurasia collision along the Zagros mountains (to the east) formed with the closure of the Neo-Tethys ocean at 27 Ma (Pirouz et al., 2017). The motion of Arabia with stable Eurasia has been relatively constant (McQuarrie et al., 2003), at 20 to 30 km/Ma, since 56 Ma. The Dead Sea Transform (to the northwest) formed at ~20 Ma (Nuriel et al., 2017) soon after the dike intrusion event along the Red Sea (Sebai et al., 1991). The onset of spreading along the Gulf of Aden (to the south) dates to ~17 Ma (Bosworth, 2005). The northern Africa plate edge can be divided into two subduction domains, (a) the eastern Mediterranean as dominated by the Hellenic subduction system, which was active since the Late Cretaceous, and the eastern Mediterranean slab roll-back region, which led to the formation of the Aegean basin, estimated to begin by 40 Ma (Agostini et al., 2010); and (b) the western Mediterranean domain constraining the Apennines–Maghrebides and Carpathians subduction systems since 50 Ma (Carminati et al., 2012). McQuarrie and others (2003) suggested that an event occurred at 25 Ma that reduces Africa-Eurasia motion to less than 10 mm/yr, changing the direction of Africa-Eurasia motion from northeast, between 56 to 25 Ma, to north, since 25 Ma.

4.1.2 Arabian Margin

Rifted margins have been classified (Manatschal 2004; Lavier and Manatschal, 2006) into three types, magma rich, magma poor, and shear margins. The Red Sea rift architecture exhibits dramatic changes along strike in extensional style and volcanism (Figure 4.3) as the tectonic setting changes across the Barakah suture zones (Johnson, 1998). These
Figure 4. 3: Present-day conceptual map of the Arabian margin. This map shows the extensional style changes along the Red Sea strike. The areas of asymmetrical diffuse extension, central Red Sea transform, and Makkah-Madinah transtensional zone (MMTZ) is emphasized.
structural architectural differences are observed in the form of a coastal escarpment, the existence of transfer zones, the absence of an axial trough, and the shift from a southern localized to a northern diffused style of deformation (Johnson, 1998). This extensional heterogeneity may reflect the effect of Arabian lithospheric strength variations as reported by Steckler (1986).

The southern Red Sea rift is an example of a magma rich margin and narrow continental rifting system (Figure 4.3), characterized by strong, old and stable lithosphere (Johnson et al, 2011), cold lithosphere (Buck, 2006) and thick elastic thickness of the southern Arabian margin (Chen et al., 2015). In contrast, the northern Red Sea rift is described as hyperextended crust and a magma poor margin (Stockli and Bosworth, 2018) with asymmetrically diffuse extension (Johnson, 1998) and dike intrusions (Roobol and Stewart, 2009). This is interpreted to indicate young and more deformable lithospheric conditions (Christensen and Mooney, 1995). Previous magma poor margin models attribute the pattern of asymmetrical dike intrusions to the lithospheric necking location (Govers and Wortel, 1993; Lavier and Manatschal 2006).

Arabian margin outcrops indicate that the upper crust is composed of crystalline Precambrian basement, Cenozoic alkalic volcanic fields (Harrat), sedimentary basins, and grabens (Stern and Johnson, 2010). The Cenozoic volcanic fields in the western Arabian margin are classified into three phases: pre-, syn- and post-Red Sea rifting volcanism (Chazot et al., 1998). Pre-rift volcanism began with eruptions ~ 30 Ma in Afar and the volcanic fields within the Sudanese margin, followed by the Harrat Ash Shaam volcanism
which began 28 Ma and continued through 22 Ma. Mafic volcanism at Harrat Ash Shaam returned following a hiatus from 22-13 Ma (Ilani, et al., 2001). Harrat Hadhan first erupted ~ 28 Ma and continued to be active until 15 Ma (Stockli and Bosworth, 2018). In the northwestern Arabian plate margin, the Ash Shaam Harrat formed as a result of Sirhan rift activities (Segev et al., 2014), which accommodated volcanic eruptions that started 28 Ma and ceased for 9 Ma between 22-13 Ma. This gap in volcanic activity is attributed to a regional tectonic rearrangement within the Red Sea region (Ilani et al., 2001).

4.1.3 Northern Red Sea Rifting Onset – Pre Localization Stage

The onset of the Red Sea rifting is recorded during the early Oligocene by a crustal cooling event (i.e. unroofing) in the northern segment (Omar and Steckler, 1995), which contrasts with a more regional Miocene cooling event in the southern segment of the Red Sea rift (Bohannon et al., 1989). A seismic stratigraphy section for Midyan, located within the northern Red Sea, supports this interpretation of the onset of rifting in the early Oligocene (Tubbs et al., 2014). Results of apatite fission track analyses from the Neoproterozoic basement in the eastern desert in Egypt indicate that a cooling phase started in the Oligocene (Bojar et al., 2002). Apatite fission track analyses from the Arabian shield in Jordan also indicate an early Oligocene cooling event, despite the samples’ location being 200 km away from the Red Sea rift (Feinstein et al., 2013). The wide cooling event within the northwestern Arabian margin corresponded with a steady geotherm along the central Arabian margin, as inferred by fission track analysis data (Szymenski et al., 2016). Furthermore, Hughes and others (1995) dated the onset of
ripping in the central Red Sea during the early Oligocene using radiometric dating (K-Ar).

Zabargad, which is a peridotite island within the Red Sea, is dated by Bosch and Bruguier (1998) to be ~ 22 Ma based on a high temperature metamorphic event which seems to be coeval with dike intrusions along the Red Sea, suggesting potential coupling between mantle and upper crust.

Lithospheric strength variations are normally assumed to control the tectonic processes at plate boundaries (Steckler, 1986) and in the plate interior. Hansen and others (2007) reported a thinner lithospheric thickness for the northwestern Arabian margin as compared with the southwestern margin. This is consistent with elevated heat flow values (60 mW/m^2) being observed in the northwestern Arabian margin (Forster et al., 2007; Gosnold, 2011) in comparison to lower heat flow values (35-45 mW/m^2) observed within the southern Arabian margin (Gettings and Showail, 1982). The elevated heat flow within the northwestern Arabian margin appears to predate the Red Sea dike intrusions (Forster et al., 2010). Foster and others (2010) suggested, based on their thermal history models, high pre-Miocene surface heat flow (~ 55 and 60 mW/m^2) in the northern Arabian-Nubian shield. In contrast, the Egyptian margin exhibits relatively lower present-day heat flow values inland (35 and 55 mW/m^2) and high heat flow (75 – 100 mW/m^2) in the vicinity of the northern Red Sea (Morgan et al., 1985).

The diffuse extensional pattern in the northern Red Sea is observed in the form of parallel grabens and normal faults spread out within the northwestern Arabian margin (Figure 4.3), where the most dominant grabens are the Tabuk, Fayha and Tayma grabens
(Johnson, 1998; Roobol and Stewart, 2008). The history of diffuse extension is also preserved in the form of dike intrusions (Figure 4.3) along the northern Red Sea (Roobol and Stewart, 2009) and the southern Red Sea margins (Sebai et al., 1991). Roobol and Kadi (2008) placed the faulting and dike intrusions in the central Arabian margin into two ages, pre-rift and syn-rift. The pre-rift faults and grabens, such as Tabuk, Fayha and Tayma grabens, are parallel to the Red Sea rift, with extension accommodated on high-angle normal faults, while the syn-rift extension is accommodated on normal rotational faults known as a master listric fault, which separates the northern Red Sea rift basin from the Arabian Shield crystalline rocks. The pre-rift grabens are dated by Roobol and Kadi (2008) at 20-25 Ma. They are succeeded by dike intrusions, such as the Great Ja’adah Dike that underlies the Tabuk graben for 400 km, as seen on aeromagnetic maps. Such gabbroic dike intrusions were also seen in the Fayha and Tayma grabens.

Although the diffuse extensional structures parallel the northern Red Sea (Roobol and Kadi, 2008; Stockli and Bosworth, 2018; Johnson, 1998; Szymenski et al., 2016) an active seismic reflection experiment across the northern Red Sea margin identified two main rift necking zones. They extend onshore, in a direction not parallel to the coastline (Sinadinovski et al., 2017). This change in lithospheric necking direction could reflect two distinct processes, diffuse rifting followed by northern Red Sea formation.
4.1.4 Red Sea Rifting Localization

The relatively low values of the Arabian margin surface heat flow, which is currently ~ 45 mW/m², led Buck (2006) to conclude that the Arabian-Nubian shield is too cold and strong to break up by an extensional force alone, and therefore localized dike intrusions are required to weaken the lithosphere and break up the continent along the Red Sea rift. Bosworth and others (2005) reported that these dikes mark the Red Sea rift initiation at 23-21 Ma. Coleman and others (1983) noted two compositional groups of the proto-Red Sea rifting dikes that reflect their distance from the Red Sea. The dikes near the Red Sea are characterized as tholeiitic composition, and those within the Arabian plate as alkalic composition. Sebai and others (1991) reported that the tholeiitic dike intrusions occurred simultaneously along the whole length of the Red Sea between 25-21 Ma (Figure 4.3). However, of their 23 samples, only one sample was located along the northern Red Sea; it has an age of 22 Ma. This allows the possibility that the inferred northward of the propagation dike intrusions is not well constrained by their data set. Synchronous with the dike intrusions along the Red Sea, mildly alkalic volcanic eruptions occurred in northeastern Africa at 24 Ma (Endress et al., 2011). Zumbo and others (1995) report tholeiitic dike swarms in southern Yemen with an age of ~ 25 Ma. In the southwestern Arabian margin, an age of 25–23 Ma is reported as well by Bohannon and others (1989) for the dike intrusions and a cooling event, with an inferred increase in geothermal gradient prior to the Red Sea rift initiation.
4.1.5 Our Approach

Understanding the underlying cause of this diffuse extension is key to resolving many of the regional geodynamic problems including the formation of the MMTZ and Dead Sea transform. This lithospheric-scale diffuse rifting across the northern Red Sea appears to have started at 23 Ma and lasted for four million years. It extends over a width of 1200 km, from the Nile river to the west to the Sirhan rift to the east (Stockli and Bosworth, 2018).

The large-scale regional tectonic forces acting on the Red Sea area are Neo-Tethys slab pull, African slab roll-back, and mantle related processes (i.e. Afar plume or other mantle flow). We investigate the interaction among these forces over time and their effects on preexisting structures and weaknesses within the Arabian margin. The investigation of these regional tectonic forces and their effects on the spatial and temporal distribution of diffuse extension can provide clues to improve our understanding of the geodynamic evolution of the northern Red Sea rifting and MMTZ volcanism, and the links between those processes to the observed deformation within the Arabian margin.

We used a finite element modeling approach to test the effects of the evolving plate boundary conditions on the predicted deformation along the Red Sea and within the Arabian margin, particularly, the MMTZ formation and the asymmetrical wide rifting of the northern Red Sea. Our approach is to take geodynamic snapshots at chosen times to
investigate the evolution. Each snapshot is designated to incorporate the corresponding boundary conditions.

4.2 Modeling

Here we present our proposed tectonic evolution of the Arabian margin and the Red Sea rift and test it in light of the tectonic forces acting along the Afro-Arabian plate boundaries for the last 40 Ma in three distinct stages, 27 Ma, 25 Ma, and 20 Ma. These stages reflect conditions at the time of MMTZ formation, northern Red Sea diffuse extension, and northern Red Sea localization. By doing so, we focus on the Red Sea rifting initiation and evolution processes. We define our model domain, equations and constraints to test these processes in light of the tectonic forces described above in section 4.1.1.

4.2.1 Proposed Tectonic Evolution

We follow a chronological order for the tectonic activities in the Mediterranean. For our modeling, the beginning, as reported in the previous section (section 4.1.1) occurs with the initiation of roll-back of the African slab, which led to the formation of the Aegean basin between 40-30 Ma. This event associated with a slow down of the African plate to with respect to the Eurasian plate (Figure 4.4). This event, as illustrated in our schematic map, was contemporaneous with rifting along both the Sirhan and the southern Red Sea rifts. As illustrated in our schematic map, the northward motion of the eastern
Figure 4. Tectonic map of the plate boundary between Arabia and Nubia at ~27 Ma. This plate boundary denoted mainly by pre-rift volcanic eruptions distribution. The motion of Arabia with respect to Nubia transferred from the southern Red Sea to Sirhan rift through a transtensional zone that occupied later on by intensive volcanic eruptions.
Mediterranean block ceased, and a transfer in plate motion from the southern Red Sea rift to the Sirhan rift developed within the Arabian margin, where both rifts were active and characterized by magma rich margins. This change in motion began the formation of the plate boundary between Arabia and Nubia around 27 Ma. We hypothesize that this differential motion was accommodated by a sinistral strike slip zone along what is known today as the MMTZ, where syn-rift volcanism appears to localize along a preexisting weakness zone (Figure 4.4). For the period around 25 Ma, there is a differential motion change within the Mediterranean, in particular between the Aegean block and the rest of the eastern Mediterranean (Figure 4.5). This differential motion is modeled (Özbakir et al., 2017) by defining the convergence of Africa, Arabia, stable Eurasia, and roll-back of the Hellenic trench. Özbakir and others argue that the Cyprus subduction is 42% locked, which promotes northward displacement that could help drive the diffuse extension during the northern Red Sea rifting evolution. Hence, we assume that this change in displacement may potentially exist since 25 Ma.

The boundary between these two domains could be the Nile river intersection with the Mediterranean. At this time the southern Red Sea rift was driven by far field forces. We suggest that this differential change in motion is the driving mechanism for the diffuse extension process. The diffuse extension domain as illustrated in our schematic map (Figure 4.5) is bounded by Nile river (to the west), Sirhan rift (to the east), MMTZ (to the southwest), and eastern Mediterranean (to the north).
Figure 4. Tectonic map of the Arabia and Nubia plates at 25 Ma. This figure emphasizes the Red Sea extensional style changes as function of the Arabian margin structural architecture and far field forces. We are hypothesizing that dike intrusions were propagating northward and the diffuse extension was driven by slowness of Nubia due to the African slab role back and the preexisted structural architecture within the Arabian margin.
Figure 4. Tectonic map of the Arabia and Nubia plates at 20 Ma. This figure emphasizes the dike intrusion event as propagated for the whole length of the Red Sea. The dike intrusions were following in their distribution syn-rifting structural pattern. We are hypothesizing that most of the volcanic eruptions along the Makkah-Madinah transtensional zone occupied a relict plate boundary.
The dike intrusions along the Red Sea may have occurred in response to a regional change in the stress field, as their occurrence comes after the Arabia-Eurasia collision event, which has been dated by Koshnaw and others (2018) at 26 Ma. Coeval with the Arabia-Eurasia collision, the Afar plume has been active since 30 Ma (Bosworth, 2005, references therein). This synchronicity may indicate a potential interaction between these two far field forces on the tectonic style and timing of the southern Red Sea rift system. This leads us to further infer that the northward dike propagation is a potential process that affects the formation of the northern Red Sea (Figure 4.6). Conceptually, these far field forces, the Arabia-Eurasia collision and a regional mantle flow could be driven by the Afar plume, can affect the Red Sea rift initiation, can utilize the weaknesses associated with the preexisting structures along the Red Sea.

4.2.2 Model Setup

Our goal is to implement the regional tectonic forces along our model domain boundaries in the form of velocity boundary conditions, and compare the resulting strain rate patterns in the region of interest with the observed deformation. To achieve our goal, we develop a mechanical model for the Afro-Arabia Plate system, including our area of interest (Figure 4.7, inset A). Our model’s kinematic boundary conditions are defined and change for different models based on the tectonic spatial and temporal changes along Afro-Arabia plate edges during the last 40 Ma.
The model is a two-dimensional plane stress spherical shell, which is assumed to be thin in the vertical direction. We compute velocity, stress ($\sigma$) and strain ($\varepsilon$) rates using the mechanical equilibrium equations (Eqs. 4.1, 4.2, 4.3, and 4.4) for plane stress:

$$\varepsilon_{xx} = \frac{1}{E}(\sigma_{xx} - \nu \sigma_{yy}) \quad \text{Eq. 4.1}$$

$$\varepsilon_{yy} = \frac{1}{E}(\sigma_{yy} - \nu \sigma_{xx}) \quad \text{Eq. 4.2}$$

$$\varepsilon_{xy} = \frac{1+\nu}{E} (\sigma_{xy}) \quad \text{Eq. 4.3}$$

$$\varepsilon_{zz} = \frac{-\nu}{1-\nu} (\varepsilon_{xx} + \varepsilon_{yy}) \quad \text{Eq. 4.4}$$

Where $E$ is Young’s modulus, and $\nu$ denotes Poisson’s ratio. The GTECTON platform assumes that each individual element is a thin plane, and all of them are connected together to describe the outline of a spherical shell. This allows the use of a planar cartesian description of the mechanical equilibrium equations. We represent the model domain with a homogeneous elastic material (Young’s modulus = 7.5 GPa; Poisson’s ratio = 0.30). The mechanical equilibrium equations for plane stress are solved using the FEM platform GTECTON (version 2017.3; Govers and Wortel, 1993; Govers & Meijer, 2001; Özbakir et al., 2017) on an elastic spherical shell with discrete faults and triangular element mesh.

In our reference model, we fix the western edge and relate the boundary and initial conditions to major geologic or tectonic features: the Neo-Tethys slab pull and Indian mid-oceanic ridge to the east, Mediterranean subduction zones to the north and Gulf of
Figure 4. Deformational model domain. This map shows the model domain for the Afro-Arabia plate for the period between 30 – 20 Ma. Because of this, the northern boundary of our domain does not coincide with the present geographic outlines for Africa. The white arrows represent our boundary conditions with stable Eurasia reference frame. The green triangles denote our fixed boundaries. The insect A labels the domain where we focus for the coming figures (18, 20, 21).
Aden, and Red Sea and Sirhan rifts within the model domain. The constraints on the plate boundary kinematics have been approximated mainly from relative Afro-Arabia plate motion with respect to the Eurasian plate. Our model boundary conditions along the eastern Mediterranean block and southern Red Sea change temporally over the life of the Red Sea rift. We are interested in testing the effect of these temporal changes on the predicted strain rate patterns along the Red Sea and within the Arabian margin. Our models do not depend on time; rather, each model represents a set of boundary conditions in a particular time. The following sections detail the Red Sea rift evolution via this set of conceptual and numerical models.

4.3 Results

We present three sets of kinematic boundary conditions corresponding to the three stages we are exploring of the initiation and evolution of the Red Sea. These boundary conditions are chosen to test our proposed tectonic evolution for the Red Sea and Arabian margin. As we are interested in the strain patterns within the northern Red Sea region, effects of step change in boundary conditions along northern boundary region of our domain, which is ~500 km away from the area of interest, here a minimal effect on our results.
4.3.1 Makkah-Madinah Transtensional Zone formation (~ 27 Ma)

We test the tectonic consequences of the kinematic boundary conditions along the Arabian plate edges at 27 Ma (Figure 4.8). In this model, we fix the Aegean block, which is part of the eastern Mediterranean, and assign velocity boundary conditions along the northern and eastern edges, including the rest of eastern Mediterranean, Anatolia, Zagros, Makran, and along the boundary between African and Indian plate. The velocity values vary between 2 km/Ma along the eastern Mediterranean and 8 km/Ma for the rest of plate boundaries. We introduce slippery nodes structures within the Afro-Arabia plate in the southern Red Sea, Gulf of Aden, and Sirhan rift. The resulting strain localizes along a zone that links both rifts, southern Red Sea to Sirhan. This zone exhibits a combination of dilatational strain rate with sinistral strike slip motion, i.e. a transtensional strain rate zone. The dilatational strain rate magnitude (Figure 4.8) range between 0.1-0.3 strain per million years. The highest strain rates concentrate around the edges of the slippery nodes, which may considered an edge effect. Nevertheless, the dilatational strain rate shows a pathway to transfer the deformation from the southern Red Sea rift to the Sirhan rift. This dilatational strain rate pattern appears to be controlled by the slowing of the African plate associated with the onset of Aegean extension due to the African slab roll-back, and passive rifting along the southern Red Sea and Sirhan.
Figure 4.8: Strain rate map of the Arabian margin around 27 Ma. The dilatational strain rate localizes along a transtensional zone with sinistral strike slip motion as proposed in Figure 4.4. We hypothesize that this strain rate pattern driven by the slowness along the northern boundary, which caused by the onset of Aegean rift due to the African slab role back, and passive rifting along the Red Sea.
4.3.2 Northern Red Sea Diffuse Rifting (25 Ma)

Here, we accelerate the eastern Mediterranean velocity while keeping the Aegean block fixed due to the ongoing slab roll-back processes. We continue to include slippery nodes in the plate interior structures, indicating the passive nature of the southern Red Sea and Sirhan rifts. The resulting dilatational strain rate (Figure 4.9) suggests a wide extensional zone within the northern Red Sea region. The zone, as can be seen in our model, is bounded by the Nile river to west, the Sirhan rift to the east, the MMTZ to the south, and the eastern Mediterranean to the north.

The magnitude of the predicted strain rate values (Figure 4.9) is dependent on the applied boundary conditions. Here, the strain rate values range between 0.1-0.3 strain per million years. If we assume 100 km as the original width, the strain rate values may imply an extension of 10 km/Ma within the Arabian margin. This includes several implied assumptions, such as the diffuse extension is uniformly distributed, there are no preexisting structures in the region that can localize the deformation, and there is no influence from sub-lithospheric processes. Overall, to validate our predicted total extension we need to compare it with the regional reconstruction models.

4.3.3 Northern Red Sea Rifting Localization (20 Ma)

Our third stage focuses on the localization of rifting in the northern Red Sea. As we proposed in the tectonic evolution section (section 4.2.1), we suggest that northward dike
Figure 4. 9: Strain rate map of the time window around 25 Ma. This figure emphasizes the boundary conditions needed for producing the diffuse extension within the Arabian margin. The deformation pattern proposed for period between 27-25 Ma. The dilatational strain rate diffuses from the Nile river to the Sirhan rift as proposed by Figure 4.5. In seek of simplicity we are using homogenous elastic domain in which the dilatational strain seems to be uniformly distributed. In reality, the diffuse extension partitioned within the Arabian margin in the form of normal faulting and grabens. We hypothesize that this strain rate pattern driven by increasing the velocity of eastern Mediterranean block and remaining the Aegean block fixed due to the ongoing rifting, where southern Red Sea remains under passive rifting.
propagation can serve as a driver for the localization of rifting in the northern Red Sea. To test this hypothesis, within the regional forces context, we maintain the slippery node boundary condition along the Sirhan rift, and consider the southern Red Sea and Gulf of Aden as active faults with 8 km/Ma slip rate. The Aegean block remains fixed, and the rest of eastern Mediterranean block continues with a high velocity (8 km/Ma).

The resulting dilatational strain rate (Figure 4.10) localizes along the northern Red Sea, leading to the initiation of a narrow extensional region within a wide magma-poor rifting environment. This suggests that the pattern of strain rate prior to actual rift formation is one of the major differences between the northern and southern Red Sea rift. This then imprints the plate boundary with a different structural architecture, leading to differences in thermal evolution for the two rift segments.

4.4 Discussion

Our MMTZ formation hypothesis is in general agreement with the earlier suggestion of Camp and Roobol (1992), who concluded there was a sinistral strike slip motion along the MMTZ, or what they call MMN volcanic line. Their interpretation is based on two main observations, the en echelon vent segments along the MMTZ, and the counterclockwise rotation, due to seafloor spreading in the Gulf of Aden, based on paleomagnetic data as reported by Tarling (1970).
Figure 4. 10: Strain rate map of the time window around 20 Ma. This figure emphasizes the northern Red Sea formation due the northward rifting propagation from southern Red Sea towards eastern Mediterranean block. The process occurred around 20 Ma where intensive dike intrusions were taking place along the Red Sea rift. The dilatational strain rate seems to localize along the northern Red Sea as proposed by Figure 4.6. We hypothesize that these dikes were facilitating the northward rifting propagation, which is prescribed in this figure in the form of dilatational strain rate localization. The boundary conditions needed for the northern Red Sea formation dependent on the southern Red Sea evolution from a passive to an active rifting, which could be attributed to the Afar plume.
The time duration between the slowing down of both the eastern Mediterranean and Aegean blocks and the accelerating eastern Mediterranean block defines the amount of shear and extension that can occur along the MMTZ. Nevertheless, the strain rate magnitudes indicate (Figure 4.8) approximately 10 km of extension per million years across the MMTZ. The MMTZ does not show large pull apart basins, suggesting a relatively short transition period between the slowing of the Aegean and the accelerating eastern Mediterranean block.

The total extension over the diffuse extension region is 35 km (Steckler et al., 1988; Bosworth, 1995) in the southern Suez rift with stretching factor ($\beta$) of 2 along the rift axis and a regional stretching factor of 1.6 (Gaulier et al., 1988). With these values, and limiting the diffuse extension duration to 4 million years, as proposed by Stockli and Bosworth (2018), requires us to assume boundary displacement rates along the northern and eastern edges of Arabia on the order of 8 km/Ma. However, McQuarrie and others (2003) reported that the Arabia-Eurasia convergence rate has been constant, at 20 to 30 km/Ma, since 56 Ma. Based on this, either the regional stretching factor needs further investigation, or these boundary displacements suggest a three to four times shorter time window for the diffuse extension processes.

The gap in volcanic activities along Shaam Harrat, which lasted for 9 Ma between 22 -13 Ma (Ilani, et al, 2001), can be attributed to the rifting competition between northern Red Sea and Sirhan rifts. Prior to rifting localization along the Red Sea rift, the Sirhan rift was
competing with the northern Red Sea diffuse extension. The northward propagation of
dike intrusions along the Red Sea (~ 25-22 Ma) may favor the deformation along the
northern Red Sea and leading to the cessation of extension along Sirhan rift.

4.5 Implications

Our proposed tectonic evolution hypothesis implies potential temporal and spatial effect
on the regional tectonic systems and on the nature of coupling between lithosphere and
asthenosphere. Here we discuss these implications for the Dead Sea Transform and on the
formation of central Red Sea transforms.

4.5.1 Dead Sea Transform Evolution

The Dead Sea Transform is considered to have formed due to the relative strength of the
Mediterranean oceanic lithosphere in contrast to the northwestern Arabian margin
(Steckler, 1986). Garfunkel (2014) reported that the Dead Sea Transform initiated at 20-
17 Ma, which is contemporary with seafloor spreading onset in the Gulf of Aden and the
reduction in the rate of Gulf of Suez faulting. Nuriel and others (2017) placed the Dead
Sea Transform onset at 20.8 Ma, based on strain analysis for calcite deposits along the
shear zone.

The formation and propagation of the Dead Sea Transform has been argued to play an
important role in the cessation of the Sirhan rift and the change in the extensional style
along the Red Sea from normal to oblique (Segev et al., 2017). The total accumulated displacement along the Dead Sea Transform is estimated to be ~100 km (Bosworth, et al., 2005). The current GPS velocities along the Dead Sea Transform jump significantly from 5.9 to 6.8 cm/yr across the intersection of the Sirhan rift with the Dead Sea Transform (Gomez et al., 2006).

Although the third phase of volcanic eruptions in the northwestern Arabian margin shows an alignment with the Dead Sea Transform, the second phase, which was simultaneous with the Red Sea formation, of volcanic eruptions in Ash Shaam and Homs Harrats in the northwestern Arabian plate edge shows an alignment with the strike of the Red Sea Rift (Weinstein and Garfunkel, 2014). Furthermore, changes in Dead Sea Transform style from transtensional to transpression formed during 20 My (Sobolev et al., 2005). These changes could be attributed to the effect of the northern Red Sea diffuse extension on the subsequent transform processes. Overall, these observations suggest that the early Dead Sea Transform processes were affected by northern Red Sea diffuse rifting, with the coupling disappearing once the Dead Sea Transform became more developed.

Since the Dead Sea Transform formation preceded the northern Red Sea diffuse rifting (Figure 4.5), the inherited weakness in the northwestern Arabian margin may have played a significant role in controlling the deformational behavior along the Dead Sea Transform. Recent seismic tomography studies for the northwestern Arabian margin conclude that deformation associated with the Dead Sea Transform is restricted to the upper 40 km of the lithosphere (El Khrepy et al, 2016). Red Sea rift-associated dike
intrusions were subsequently folded by Dead Sea Transform displacement (Figure 4.3), indicating that the Red Sea dikes predate the Dead Sea Transform (Roobol and Stewart, 2009).

4.5.2 Central Red Sea Axial Curvature

The central Red Sea rift is characterized (Figure 4.3) by multiple transforms (e.g. Garson and Krs (1976), Ghebreab (1998), El-Isa (2015), and Almalki et al. (2015)). These transforms are interpreted from the regional aeromagnetic data between Sudan and Saudi Arabia that were mapped by the Saudi-Sudanese Red Sea Commission (RSC) (Izzeldin 1987). However, a recent geomorphic study (Augustin et al., 2016) argues instead, based on new bathymetric data, for a continuous trough since most of the transform offsets appear to be absent. Nevertheless, both data sets confirm an axial curvature in the central Red Sea.

Our proposed tectonic evolution implies that the central Red Sea rift can act as a transition between the two end members of extensional styles along the southern and northern Red Sea. The nature of this transition can be inferred based on our proposed evolution from the diffused extension stage at 25 Ma (Figures 4.5, 4.8) into the localized deformation stage along the northern Red Sea at 20 Ma (Figures 4.6, 4.9). Based on these models, we view the central Red Sea axial curvature as a natural response to the geodynamic evolution of the rift system, in particular the spatial transition from southern localized into northern diffused, and the temporal transition from diffused deformation
along the northern Red Sea into localized due to the dike intrusions event. Overall, our proposed tectonic evolution can provide a new frame work to investigate the observed transition along the central Red Sea axial curvature.
4.6 References


Feinstein, S., M. Eyal, B. P. Kohn, M. S. Steckler, K. M. Ibrahim, B. K. Moh'd, and Y. Tian. "Uplift and denudation history of the eastern Dead Sea rift flank, SW Jordan:


Chapter 5

Conclusions

The Red Sea rift initiation has been reported by many researchers to be between 30 - 20 Ma, based on a broad spectrum of field data; meanwhile, its evolution passed through several rifting stages, observed within the Arabian margin in the form of dike intrusions, volcanic eruptions, grabens, and uplift events. Since the 1980s, many researchers have proposed links between these tectonic activities and regional tectonic forces; however, very few numerical models have been developed. Hence, in this work, we use the finite element modeling approach (GTECTON platform) to highlight mechanical deformation processes that describe mechanism for the initiation and evolution of Red Sea rifting as a function of regional tectonic forces.

Our finite element models implement the kinematic evolution of the Afro-Arabia plate edges in the form of boundary and slippery node conditions, to test the sensitivity of these conditions on the predicted strain rate patterns along the northern Red Sea and within the Arabian margin, during various geological time frames. Our results show that the differential motion among both Aegean and eastern Mediterranean blocks, with respect to the Eurasian plate, played an important role in the formation of the MMTZ and the later diffuse extension along the northern Red Sea. The evolution of the southern Red Sea
from a passive to an active rift system was the driving mechanism for the northern Red Sea formation.

In detail, the MMTZ formed to allow the transfer of motion from the southern Red Sea to the Sirhan rift, which occurred due to slowing of Africa with respect to Eurasia as a result of the African slab roll-back in the Mediterranean. This slowing was followed by acceleration of the eastern Mediterranean block, while the Aegean basin was undergoing extension. This led to the initiation of a wide zone of rifting along the northern Red Sea, bounded by the Nile river and Cairo volcanic provinces to the west, Sirhan rift zone to the east and MMTZ to the south. We consider these regional tectonic forces to be the driving forces for both MMTZ formation and northern Red Sea diffuse extension. By introducing faulted node conditions along the southern Red Sea and Gulf of Aden, we can amplify the resulting stresses along the northern Red Sea and thus promote northward propagation of the rift system. As a consequence of this early diffuse extension phase, the southern Dead Sea Transform exhibits transtensional structures, while the northern Dead Sea Transform, particularly beyond the eastern edge of the diffuse extension region, displays transpressional structural styles.

The implications of this tectonic evolution, we hypothesize, is that it relies on the successive phases of deformation within the Arabian margin. The regional stress field changes due to either the rifting breakup or transform initiation may utilize lithospheric weak zones that formed during the tectonic evolution of the Red Sea and Arabian margin.
Based on this study, the recent deformational phase, as represented in the Cenozoic volcanic eruption episodes or earthquake activities within the Arabian margin, is attributed to the interaction between the inherited lithospheric weaknesses with the sub-lithospheric tractions as potentially driven by mantle processes. This conclusion was built on the basis of analyzing a new set of GPS velocities. The observed GPS velocities produce dilatational and shear strain rate anomalies within the Arabian margin.
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- (2017) AGU Fall Meeting, New Orleans: “Arabian Plate Deformation: The role of inherited structures in the localization of strain in the Red Sea extensional system”
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