SUPPLEMENTAL INFORMATION

Initial Burst of Oceanic Crust Accretion in the Red Sea Due to Edge-Driven Mantle Convection

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SUPPLEMENTARY METHODS

1. Gravity data reduction

Shipboard gravity measurements from the NGDC database (Fig. S1) have been complemented after discretization by those in analogue form of Izzeldin (1982, 1987). A total of 10872 gravity measurements were used. The overall statistics at 333 cross-over points give a mean and standard deviation of 2.5 and 13.1 mGal, respectively. These data were collected over a 15-year period, during which recording instruments and navigation systems greatly improved. In order to integrate these different data sets before merging, the ship gravity for each line was adjusted in a least-square sense by applying a series of corrections aimed at reducing the errors at cross-over points and at matching the satellite-derived marine gravity field (Figs S2 and S3). The compensation of systematic errors outlined above, performed with techniques described in Huang et al. (1999), has a twofold effect: it reduces the cross-over differences and it adjusts ship gravity to the satellite-derived gravity field. After adjustment, the mean and rms differences between ship and satellite gravity are 0.4 and 5.6 mGal, and the overall cross-track discrepancies are reduced to zero mean and standard deviation of 3.1 mGal, equivalent to an accuracy of 2.2 mGal. The adjusted ship gravity is then included with satellite-derived gravity to derive a second gravity field (Fig. S2). Different techniques have been developed during the last few years to combine ship-borne and satellite-derived gravity data, such as least squares collocation (Hwang and Parsons, 1995), least-squares adjustment in the frequency domain (Barzaghi et al., 1993) and input-output system theory (Sideris, 1996; Li and Sideris, 1997). Studies on simulated (Li and Sideris, 1997), and real data (Tziavos et al, 1998) have shown that the best results are obtained by the input-output system theory (IOST) method, mainly when the power spectral densities are estimated directly from the data. The use of IOST requires the field of observables be given on the same grid with known error power spectral densities. Ship gravity anomalies were computed on the same grid of satellite-derived
global gravity (release 18.1) of Sandwell and Smith (1997) for the area of interest. We assume gravity anomalies affected by a gaussian random noise with zero mean and variance of $9.61 \text{ mGal}^2$ and $25.0 \text{ mGal}^2$ for ship-board and satellite-derived gravity data, respectively. The IOST estimated values for the free air anomalies are given by:

$$F\{\Delta g\} = F\{\Delta g_s\} + F\{\Delta g_a\}$$

where $F$ is the 2-D Fourier transform operator, $P_{g_s}$, $P_{g_a}$, $E_{g_s}$, and $E_{g_a}$ are the power spectral densities of ship-board and satellite-derived gravity data and noise. Figures S2 and S3 show the results of the procedure we have applied.

1.1 Sediment thickness

The northern Central Red Sea region is underlain mainly by Miocene and younger sediments (Sestini, 1965; Whiteman, 1968). In fact, the connection with the Mediterranean Sea was activated during Miocene and lagoonal conditions developed throughout, with thick evaporites (up to 6 km) deposited from shore to shore (Sestini, 1965; Said, 1969). Extensive salt diapirism is clearly shown on seismic reflection profiles across the Red Sea (Ross & Schlee, 1973; Izzeldin, 1987). By the end of Miocene, the Red Sea was separated from Mediterranean and erosional conditions prevailed. Neritic sedimentation started in the Pliocene when the Red Sea was connected to the Gulf of Aden. The unconformity at the base of post-evaporitic sediments, verified during DSDP Leg 23 (Whitmarsh et al., 1974), is very well documented in seismic profiles and has been designated as reflector “S”.

Thickness of hemipelagic cover and evaporitic layers has been estimated from refraction data of Tramontini and Davis, (1969), our own multi-channel (RS05) and single-channel seismic (MR79, MR83) reflection data and those of Izzeldin (1982, 1987, 1989) adopting an average interval seismic velocity of 2 km/s and 3.75 km/s, respectively.
(Tramontini & Davis, 1969). The post-evaporitic sequence has a more or less uniform thickness of about 0.2 km, except in the centre of the Nereus-Thetis intertrough where the thickness can increase up to 0.6 km and near the axial trough where it can thin down to 0.1 km. On both sides of the axial trough it is thinner than average and remarkably disturbed (Fig. S4b). The Miocene evaporites lie below reflector “S” and show two different seismic facies: regions rich of reflectors corresponding to layered evaporites (from borehole data they are made up of anhydrite, salt, mudstone and sandstone) with an average seismic velocity falling between 3.0 and 3.5 km/s, and regions almost devoid of reflectors corresponding to rock salt with an average velocity of 4.2-4.6 km/s. The layered evaporites are always surrounded and underlain by rock salt with salt increasing toward the axial trough (Izzeldin, 1982; Izzeldin, 1987). The thickness of the Miocene section is maximum under the flanks of the main trough and in the centre of the Thetis-Nereus inter-trough, where it can reach more than 5 km. Thinning occurs near the axial trough (Fig. S2c).

1.2 Mantle Bouguer anomaly

The components of the gravity field due to variations in crustal thickness or crustal and upper mantle density anomalies have been calculated removing the predictable signals produced by density contrast between water, sediments and rocks at the seafloor, and by density variations associated with the temperature field. The contribution of topography, sediments and crust to the local gravity field has been computed from the grids of bathymetry (Figs 1 and 2a), post-evaporitic unconsolidated sediments (Fig. S2b) and Miocenic evaporites (Fig. S2c) by a FFT technique based on the method of Parker (1973), that uses a series expansion of the Fourier transformed powers of the base of each layer to represent the Fourier transform of the gravity anomaly. The Bouguer correction was obtained by replacing the water layer (density of 1040 kg/m$^3$), unconsolidated sediments (density of 1900 kg/m$^3$), evaporites (density 2200 kg/m$^3$) and a 5 km constant thickness crust (density of 2670 kg/m$^3$) limited by and parallel to the base of the evaporites, with a
layer of mantle material (density of 3330 kg/m$^3$). The first nine terms of the series expansion were retained in our calculation to account for the non-linear gravitational attraction of the large topographic relief in this region. Grids of bathymetry and sediment thickness were produced at 0.15 km of spatial resolution in order to perform up-ward continuation also in the shallowest portion of the Red Sea (depth range > 100 m). Each grid was mirrored to avoid edge effects introduced by the implicit periodic assumption of the FFT routine. The crustal interface attraction predicted values, computed at bathymetry grid points, were interpolated onto combined ship-satellite gravity grid points and then subtracted from the corresponding free air anomalies. The new values were then re-gridded at 0.3 km to obtain the complete mantle Bouguer anomaly (MBA). The zero level is arbitrary and corresponds to the center of the range in anomaly amplitudes (Fig. 2b).

1.3 Mantle thermal gravity anomaly and residual mantle Bouguer anomaly

Assuming that compositional variations are negligible, the contribution of upper mantle density variations to the local gravity can be inferred from the mantle temperature field. The temperature field has been calculated by the steady-state advection-diffusion equation:

$$\kappa \nabla^2 T = \mathbf{v}_s \cdot \nabla T + \mathbf{v}_s \cdot \mathbf{z} \alpha$$

(2)

where $\kappa$=mantle thermal diffusivity, 8.04 $10^{-7}$ m$^2$/s; $\mathbf{v}_s$=solid mantle velocity vector; $\alpha$=adiabatic temperature gradient, 0.0003°C/m and $\mathbf{z}$=unit vector along z-axis. Mantle temperatures, through the 3D-domain of mantle flow calculations (see next Chapter), have been computed by the over-relaxation upwind finite difference method described in Phipps Morgan and Forsyth (1988), using a variable grid spacing (512x256x101) with the highest grid resolution (1 km) in the proximity of the plate boundaries. Temperature solutions were found assuming constant temperatures of 0 °C at the surface and 1350 °C at 150 km depth.
The predicted thermal mantle contribution (MTGA) to the gravity field was obtained stacking the gravity signals from all the horizontal layers defined by the finite difference grid. The gravity signal $\Delta g_i$ from each layer was computed converting the temperature field into density variations by a thermal expansion coefficient, then converted into the gravity signal observed at the sea level employing the FFT technique:

$$F\{\Delta g_i(x, y, z_0)\} = -2\pi Ge^{-kz_0}(e^{kz_i} - e^{kz_{i-1}})F\{\rho(x, y, z_i)\}/k$$

where $z$ is positive upward, $z_0 = 150$ km is the surface of observations, $z_i$ and $z_{i-1}$ are top and bottom depths of the $i$th layer, $k = \sqrt{k^2_x + k^2_y}$ with $k_x$ and $k_y$ wavenumbers, $\rho(x, y, z_i) = \{1 - \alpha[T(x, y, z_{i-1}) + T(x, y, z_i)]/2\}$ is the horizontal density distribution and $\alpha = 3.25 \times 10^{-5}/^\circ C$ is the thermal expansion coefficient.

The predicted thermal gravity field is subtracted from the MBA to create the residual anomalies (Fig. S5). The computed residual low pass filtered mantle Bouguer anomalies, (wavelengths shorter than 12 km were cut off), were downward continued to a depth of 8 km to infer crustal thickness variations (Fig. 3a).

### 1.4 Mantle Flow Velocity Field

In order to estimate the mantle thermal structure below the central northern Red Sea we adopted different mantle flow models: thin and thickening plate passive flow, and a three-dimensional modified form of the upwelling divergent flow model of Kuznir and Karner (2007). We considered a steady-state mantle flow induced by motion of the overlying rigid plates in an incompressible, homogeneous, isoviscous mantle beneath an accretionary plate boundary geometry and half spreading rate (6.1 mm/yr) that duplicates the northern central Red Sea rift system. We modelled the flow induced by seafloor spreading in a computational frame 1024x512 km wide and 150 km deep (1x1 km spaced grid points for each 1 km depth increment).
The steady-state three-dimensional passive mantle flow (thin plates and plates that thicken away from the ridge (Morgan and Forsyth, 1998; Blackman and Forsyth, 1992; Shen and Forsyth, 1992) has been solved via the Fourier pseudo-spectral technique outlined in Ligi et al. (2008). Given the ultraslow spreading rate of the Red Sea, passive flow models produce an inadequate mantle temperature field beneath the ridge segments, with a small amount of melt generated and a small thermal mantle contribution to the gravity field. Figure S5a shows the predicted thermal mantle gravity field for the plate-thickening passive flow model.

The upwelling divergent flow model outlined in Kuznir and Karner (2007) is based on the two-dimensional corner-flow solution of Reid and Jackson (1981) and simulates both passive and active mantle flows depending on the boundary conditions. Boundary conditions include the velocity field at the surface, defined by the half spreading rate $V_x^0(x,z=0)$ and the axial upwelling rate $V_z^0(x=0,z)$, defined by the $V_z^0/V_x^0$ ratio. For $V_z^0/V_x^0$ ratios less than 1.5, the model produces passive mantle flow solutions ($V_z^0/V_x^0 = 2/\pi$ is the thin plate solution); for higher ratio values the model simulates active flows (Fletcher et al., 2009). In order to calibrate the model against degree of melting inferred from the Na$_8$ of Nereus and Thetis basalts, we modelled temperature field and melt generation assuming an half spreading rate of 6.1 mm/yr (Chu and Gordon, 1998) and different values for the $V_z^0/V_x^0$ ratio. Ratios within range 3-4 fit well the observed Na$_8$ basalt content, suggesting an active component in mantle upwelling velocities.

Melting of the mantle that upwells beneath spreading centres may induce significant viscosity changes. Melting leads to rapid dehydration of the mantle and can increase the viscosity of the residue by two orders of magnitude (Hirth and Kohlstedt, 1996; Braun et al., 2000). On the other hand, the presence of small amounts of interstitial melt can favour grain boundary sliding reducing the effective viscosity. Short term temporal variations of mantle upwelling below a segment of the Mid Atlantic Ridge have been ascribed by
Bonatti et al. (2003) to active components due to non-uniform mantle rheology, with a sub-ridge low-viscosity zone between layers with higher viscosity. The rheological stratification may be due to the loss of H$_2$O from the upwelling mantle. The increase of viscosity in the upper part of the melting region (where dry melting occurs) induced by dehydration, limits buoyant upwelling; as a consequence, solid flow within the upper layer is mostly driven by plate separation. Thus, we modified the passive flow half-space solutions of Ligi et al. (2008) in a flow model that is passive within an upper layer and active below (given by the upwelling divergent flow beneath a mean spreading axis). We assumed as boundary condition for the lower limit of the upper layer mantle velocities from the upwelling divergent flow. Figure S5b shows the predicted thermal mantle gravity field we used to calculate the RMBA, assuming plate separation velocities at the base of the lithosphere as shown in the figure. In addition, at a depth of $z = 60$ km (base of the dry melting region), we assumed that passive flow velocities are those of an upwelling divergent flow model computed at the same depth, with $V_z^0/V_x^0 = 4$.

References


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Figure S1. Location of geophysical profiles used to produce basin-wide gridded magnetic, gravimetric, bathymetric and sediment thickness data. Red, green, blue, and black solid lines indicate profiles of RS05, MR79 and MR83 cruises and NGDC database, respectively.
Figure S2. Free air gravity grids. Grid step, 0.3 km. a, free air satellite-derived gravity anomalies (from Sandwell and Smith, 1997). b, combined shipborne and satellite-derived gravity field. c, residuals between b and a. Dotted red lines indicate locations of across-axis gravity profiles shown in Figure S3.
Figure S3. Across-axis gravity profiles showing differences between satellite-derived and ship-borne free air anomalies. Line locations are indicated in Figure S2c. Orange solid line, satellite-derived; red dashed line, ship-borne; and blue line, combined satellite and ship-borne gravity data resulted the IOST method of Li and Sideris (1997).
Figure S4. Maps showing magnetic anomalies, Plio-Quaternary and Miocene sediment isopachs, superimposed on bathymetry. a, observed magnetic anomalies. RS05 magnetic data were corrected for diurnal variations and for IGRF-2005 and were then gridded at 1 km step together with data from cruises MR79 and MR83, and NGDC database, previously filtered and validated. b, Plio-Quaternary hemipelagic cover thickness. We assumed reflector “S” as the base of post-evaporite sediments. c, Thickness of Miocene sediments including salt and layered evaporites. Thickness of the sedimentary sequences was estimated from our own multichannel (RS05) and singlechannel seismic (MR79 and MR83) reflection data, and those of ref. 1 and 2.
Figure S5. Mantle thermal gravity and residual mantle Bouguer anomalies. Arrow indicates half spreading rate. **a**, predicted thermal mantle contribution to the gravity field beneath a plate boundary geometry simulating that of the northern central Red Sea obtained assuming plate-thickening passive mantle flow (Blackman and Forsyth, 1992; Shen and Forsyth, 1992). **b**, predicted mantle thermal gravity field obtained assuming a reduction in plate separation velocity at the base of lithosphere beneath the Nereus-Thetis intertrough due to lithospheric stretching, and adopting the mixed thickening-plate driven (Ligi et al., 2008) and upwelling-divergent (Kuznir and Karner, 2007; Fletcher et al., 2009) mantle flow model. Gray shaded area represents the sector where spreading velocity is tapered down to a minor fraction of plate separation velocity. **c**, residual mantle anomalies obtained by subtracting from the