Thermal regime of the Southeast Indian Ridge between 88°E and 140°E: Remarks on the subsidence of the ridge flanks


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1. Introduction

[2] The flanks of the Southeast Indian Ridge are characterized by anomalously low subsidence rates for the 0–25 Ma period: less than 300 m Ma\(^{-1}\) between 101\(^\circ\)E and 120\(^\circ\)E and less than 260 m Ma\(^{-1}\) within the Australian-Antarctic Discordance (AAD), between 120\(^\circ\)E and 128\(^\circ\)E. The expected along-axis variation in mantle temperature (~50°C) is too small to explain this observation, even when the temperature dependence of the mantle physical properties is accounted for. We successively analyze the effect on subsidence of different factors, such as variations in crustal thickness; the dynamic contribution of an old, detached slab supposedly present within the mantle below the AAD; and depletion in \(\varphi_m\), a parameter here defined as the “ubiquitously distributed melt fraction” within the asthenosphere. These effects may all contribute to the observed, anomalously low subsidence rate of the ridge flanks, with the most significant contribution being probably related to the depletion in \(\varphi_m\). However, these effects have a deep-seated origin that cannot explain the abruptness of the transition across the fracture zones that delineate the boundaries of the AAD, near 120\(^\circ\)E and near 128\(^\circ\)E, respectively.


2. Geological and Geophysical Setting

[4] Seafloor spreading at the Southeast Indian Ridge (SEIR) started during late Cretaceous [e.g., Cande and Mutter [1982]; Tikku and Cande [1999]]. Until mid-Eocene, spreading was very slow (~10 km Ma\(^{-1}\)) and oblique. After about Chron 20 (~43 Ma), spreading was established at an intermediate rate, with no significant change in characteristics (direction and rate) during the last 36 Ma [Royer and Sandwell, 1989], except within the AAD, where kinematics events have been documented during Miocene time [Marks et al., 1999].

[5] Near 78\(^\circ\)E, the SEIR is influenced by the presence of the Amsterdam and Saint Paul (ASP) Islands which sit on the Antarctic plate, less than 40 km away from the spread-
ing center [Scheirer et al., 2000]. As the ASP hot spot was captured by the SEIR, about 3.5 Ma ago, excess volcanism formed a plateau of \( \sim 150 \times 200 \) km\(^2\), which stands 1 to 3 km above the surrounding seafloor. The influence of the ASP is thus relatively localized compared to that of the more distant but larger Kerguelen hot spot. Although it is located more than about 1100 km from the nearest SEIR segment, the Kerguelen anomaly influences the ridge axial depth and morphology over distances of several thousands of kilometers [Small, 1995; Ma and Cochran, 1997].

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Between \( \sim 118^\circ \)E and \( 128^\circ \)E (Figure 2), the Australian-Antarctic Discordance (AAD) was first recognized as a region of rough topography and deep regional bathymetry [Hayes and Conolly, 1972] centered on the Southeast Indian Ridge. It was proposed that it represents the surface expression of regionally cooler mantle temperatures [Weissel and Hayes, 1974]. The AAD is located in the center of a depth anomaly that extends from Antarctica to Australia [e.g., Marks et al., 1990] that persisted at least over the last 100 Ma [e.g., Veevers, 1982; Gurnis et al., 1998]. The AAD thus appears to be related to a long-term mantle anomaly,
the surface expression of which evolves through time, as evidenced by the fact that rough seafloor topography and fracture zone offsets in the AAD have increased, about 30–25 Ma ago [e.g., Weissel and Hayes, 1974; Vogt et al., 1983; Christie et al., 1998; Okino et al., 2004]. Early seismological work [e.g., Montagner, 1986; Forsyth et al., 1987; Kuo, 1993; Roult et al., 1994] indicated the presence of high seismic mantle velocities below the AAD. Gurnis and Müller [2003] have reviewed results from three different global seismic inversion models at three different depths (100, 400, and 900 km): model SB4L18 [Masters et al., 2000]; SAW24B16 [Megnin and Romanowicz, 2000]; and S20RTS [Ritsema and van Heijst, 2000]. Although the horizontal resolution is >1000 km, all three models show a north-south trending high shear wave velocity anomaly in the lower mantle and near the transition zone (670 km) beneath the AAD. Using broadband surface wave group and phase velocity measurements, Ritzwoller et al. [2003] have recently resolved a NW-SE trending anomaly (Figure 3), termed as the Australian-Antarctic Mantle Anomaly (AAMA). On the basis of surface wave tomography, the depth extent of the AAMA appears to be confined to the upper ~120 km of mantle [Ritzwoller et al., 2003].

3. Estimates of Mantle Temperature Variations Beneath the Southeast Indian Ridge

[8] From melting models [e.g., McKenzie, 1984] and from models on the formation of axial topography [e.g., Chen and Morgan, 1990], it can be inferred that the observed contrasts in axial depth and morphology along the SEIR require thicker crust and warmer upper mantle temperatures both to the east and west of the AAD. This prediction is confirmed by direct measurements based on OBS data from different segments of the SEIR. Within the AAD, thin crustal thicknesses of 3.6 km and of 4.2 km have been measured, respectively, near 125°E [Kojima et al., 2003] and near ~127 and ~128°E [Tolstoy et al., 2002]. In contrast, east of the AAD, the crustal thickness is greater than ~7–7.5 km [Tolstoy et al., 2002]. West of the AAD, the crustal structure changes significantly and the total
thickness increases from 4.2 km near 118°E to 7.2 km near 101°E [Holmes et al., 2005].

Shah and Sempe [1998] have reviewed five models relating crustal thickness and mantle temperature: two melting models [McKenzie, 1984; Klein and Langmuir, 1987] and three models based on fluid flow dynamics [Su et al., 1994; West et al., 1994; Chen, 1996]. In the present paper, we only examine the results from the melting models, which use the temperature of the melting column at the ridge axis and the temperature at the base of the lithosphere, consistent with what is used in the simple model of plate creation [McKenzie, 1967], which successfully accounts for the variation of depth and oceanic heat flow with age [Parsons and Sclater, 1977].

McKenzie [1984] proposed a detailed model to compute the crustal thickness (e.g., the total amount of melt) generated by the isentropic upwelling of a mantle column. The model calculations depend on the entropy difference \( \Delta S \) between solid and liquid per unit mass and on the variation of melt fraction by weight, \( X \), with Pressure \( P \) and temperature \( T \). For instance, for \( \Delta S = 362 \text{ J kg}^{-1} \text{ K}^{-1} \), the temperature increase (at the depth where melting begins) required for 1 km change in crustal thickness is 20°C when the \( T, P \) (in GPa) and \( X \) are linked by the following relationship: \( T = 1100°C + 100P + 600X \). When the heat transport by melt and the gravitational energy are taken into account, the temperature increase is equal to 14°C per additional km of crust. Recently, [McKenzie et al., 2005] ascribe an increase of potential temperature of 12.5°C to each km increase in crustal thickness. The potential temperature is defined to be the temperature that an element of fluid would have if it were moved adiabatically from a reference depth \( z_0 \) to a given depth \( z \). If the mantle is in adiabatic equilibrium everywhere and if the reference depth is the Earth’s surface (\( z_0 = 0 \)), then the relation between the actual temperature \( T \) at depth \( z \) and the potential temperature \( T_p \) is

\[
T_p = T \exp\left(-\frac{g \alpha z}{C_p}\right)
\]

where \( \alpha \) is the thermal expansion coefficient (for solid and magma, notations \( \alpha_s \) and \( \alpha_f \) are, respectively, used) and \( C_p \) is the specific heat at constant pressure. With \( \alpha = 3 \times 10^{-5} \text{ K}^{-1} \) and \( C_p = 10^3 \text{ J kg}^{-1} \text{ K}^{-1} \), the difference between potential and real temperature amounts to about 4%. The
order of magnitude of the expected mantle temperature variation beneath the Southeast Indian Ridge corresponding to \(3.5\) km change in crustal thickness between \(88^\circ\) and \(128^\circ\) could be \(50^\circ\)C.

This estimate is consistent with basalt geochemistry based on the analysis of major elements. [Klein and Langmuir, 1987] recognized a negative correlation between segment-scale averaged axial depth and Na\(_8\) values, and between Na\(_8\) and Fe\(_8\) values for mid-ocean ridge basalt (MORB) glasses worldwide (Na\(_8\) and Fe\(_8\) are the Na\(_2\)O and FeO, respectively, content normalized at a weight percentage of MgO equal to 8%), reflecting variations in the temperature of a relatively uniform mantle. The Na\(_8\) content of basaltic glasses from the Southeast Indian Ridge [e.g., Christie et al., 1995] increases as the axial seafloor deepens between \(88^\circ\) and \(118^\circ\) (Figure 4), in response to a decrease in the temperature of initial melting that could be of the order of \(50^\circ\)C (from \(1325 \pm 50^\circ\)C to \(1375 \pm 50^\circ\)C), based on the work by Klein and Langmuir [1987].

### 4. Analysis of Subsidence Variations

#### 4.1. Reassessment of Subsidence Rates

The relationship between basement depth and crustal age provides a theoretically straightforward way to assess variations in mantle temperature below mid-ocean ridge crests. Assuming that the physical properties of the upper mantle are constant and that the mid-ocean ridge is isostatically compensated, seafloor depth (\(z\)) theoretically increases linearly with the square root of seafloor age (\(t\)). For young lithosphere, the relationship can be written as [e.g., Davis and Lister, 1974]

\[
z = z_0 + 2\rho_w\alpha(T_m - T_0)\sqrt{\frac{k}{\pi}} \frac{\sqrt{t}}{(\rho_m - \rho_w)} \tag{2}
\]

where \(z_0\) is axial depth, \(k\) and \(\alpha\) are the mantle thermal diffusivity and expansion coefficient, \(\rho_m\) and \(\rho_w\) are the densities of mantle and water, respectively, \(T_0\) is the surface temperature, and \(T_m\) is the temperature of the mantle column below the rise crest. This expression, analytically derived from the half-space model, also is a good approximation for the plate model [e.g., McKenzie, 1967; Parsons and Sclater, 1977] for ages younger than \(40–50\) Ma, when the temperature at the base of the plate is ascribed to be equal to \(T_m\).

Previous studies have shown that the flanks of the SEIR are characterized by lower than normal subsidence rates [e.g., Cochran, 1986; Kane and Hayes, 1994; Sykes, 1995]. Here, we reassess subsidence rates during the \(0–25\) Ma period (since the onset of the AAD-related fracture zones), using data sets that were not available prior to 1994. For bathymetry, we use the global model (topo_8.2.jpg) of
Smith and Sandwell [1997], based on satellite altimetry derived gravity and ship soundings. Bathymetry is inferred from gravity using a transfer function that accounts for the effect of seafloor and Moho topography. The latter effect is computed using an average crustal thickness of $C24$ 7 km and multiplied by a band pass filter that is the same everywhere. However, because the bathymetric prediction relies on gravity only at wavelengths shorter than 160 km, the fitted subsidence rate should have no influence of altimetry (W. H. F. Smith, personal communication, 2007). A deterministic approach was then followed, by carefully selecting, between $176^E$ and $176^E$, a set of flow lines apparently unaffected by structural features, such as fracture zones, off-axis seamounts, intraplate volcanism, propagator trails, etc (Figures 1 and 2).

For ages, the plate kinematics parameters of Royer and Sandwell [1989] were used to compute flow lines, except between $176^E$ and $28^E$, because it does not reflect the complexity of the AAD. Within the AAD (Figure 4) we have used the age determination of Marks et al. [1999], who carefully identified a series of magnetic anomaly picks from closely spaced aeromagnetic profiles [e.g., Morgan et al., 1979; Vogt et al., 1983]. (Figure 5)

In order to correct ridge flank depths for sediment thickness, we have systematically reanalyzed all available seismic reflection lines (Figure 6). These data were primarily collected during the mid 1960s and early 1970s, mainly with R/V Eltanin, R/V Vema, and R/V Conrad. We digitized sediment thickness along ship tracks, converted the measurements from time to depth and applied loading corrections to account for the isostatic effects of the sediments [Crough, 1983]. A new, regional map of sediment thickness was produced (Figure 7), which differs only slightly from the previous map of Hayes [1991]. Drill hole data from ODP Leg 187 [Christie et al., 2004] indicate the presence of sediments 150 to 200 m thick, which suggests that the sediment cover within the AAD is unresolved by the seismic data, due to the diffraction of seismic waves on the rough seafloor topography (Table 1). The underestimation of the sediment thickness may result in errors in estimating the basement depth and subsequent underestimation of subsidence rates. Although we cannot totally preclude this hypothesis, we think that errors in sediment thickness do not change the major characteristics of the sediment pattern.
Figure 6. Available seismic lines used in the present study. Colors over track lines represent sediment thickness.
[16] Plots of corrected basement depth versus age \(1/2\) along these flow lines are shown in Figure 8 (anomalous flow line 1 within the AAD, Figure 8b, has not been used to compute the averages below). Subsidence rates (Figure 9) are somewhat different from one flow line to the next and are generally lower on crust younger than \(\sim 3-10\) Ma than on older crust.

[17] On the northern flank, subsidence rates for the \(0-25\) Ma period are equal to \(373 \pm 28\) (RMS) m Ma\(^{-1/2}\) between longitudes 90 and 101\(^{1/2}\)E; \(295 \pm 25\) (RMS) m Ma\(^{-1/2}\) between 101 and 120\(^{1/2}\)E; \(257 \pm 13\) (RMS) m Ma\(^{-1/2}\) between 120 and 128\(^{1/2}\)E; and \(347 \pm 15\) (RMS) m Ma\(^{-1/2}\) between 128\(^{1/2}\)E and 140\(^{1/2}\)E.

[18] On the southern flank, we obtain for the same age range \([0-25\) Ma]: \(358 \pm 37\) (RMS) m Ma\(^{-1/2}\) between longitudes 90 and 101\(^{1/2}\)E; \(262 \pm 28\) (RMS) m Ma\(^{-1/2}\) between 101 and 120\(^{1/2}\)E; \(254 \pm 10\) (RMS) m Ma\(^{-1/2}\) between 120 and 128\(^{1/2}\)E; and \(361 \pm 33\) (RMS) m Ma\(^{-1/2}\) between 128\(^{1/2}\)E and 140\(^{1/2}\)E.

[19] West of 101\(^{1/2}\)E, subsidence rates are close to the average documented for the world’s ocean basins [e.g., Parsons and Slater, 1977]. We will thus hereafter use these values as reference for “normal” subsidence: 373 and 358 m Ma\(^{-1/2}\) for the northern and southern flank, respectively.

[20] Subsidence values found by Kane and Hayes [1994] are possibly affected by uncontrolled errors due to structural effects such as ridge jumps, propagating rifts, etc. However, besides some differences, the general characteristics of the subsidence pattern are the same as those found by Kane and Hayes [1994] and Hayes and Kane [1994], except within the AAD. The gross trend is that the average subsidence rate is lower between 101\(^{1/2}\)E and 120\(^{1/2}\)E by about \(\sim 70-80\) m Ma\(^{-1/2}\). However, within the AAD, our results differ significantly from those of Kane and Hayes [1994]: We find that the subsidence rate is consistently lower between 120\(^{1/2}\)E and 128\(^{1/2}\)E than between 101\(^{1/2}\)E and 128\(^{1/2}\)E, whereas there is no such trend in the work by Kane and Hayes [1994], who used crustal ages based on the magnetic lineations of Cande et al. [1989], which do not reflect the complexity of the tectonic history within the AAD during the Miocene [Marks et al., 1999].

[21] To summarize, our results indicate one, large-scale regional domain that extends from 101\(^{1/2}\)E and 128\(^{1/2}\)E, with subsidence rate less than 300 m Ma\(^{-1/2}\), and one subdomain, with subsidence rate less than 260 m Ma\(^{-1/2}\), delineated by

**Figure 7.** Revisited map of sediment thickness (in seconds two-way traveltime) with hand-contoured isopachs based on seismic lines displayed in Figure 6. Within the AAD, the sediment cover is unresolved by the seismic data. Between 90\(^{1/2}\)E and 105\(^{1/2}\)E, the sediment thickness resolved by the seismics is \(\sim 200-300\) m on crust younger than 15 Ma but tends to increase from east to west, except in some specific, unsedimented areas.

<table>
<thead>
<tr>
<th>Hole</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth, m</th>
<th>Age</th>
<th>Sediment Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>1152A</td>
<td>41°53.9’S</td>
<td>127°0.4’E</td>
<td>5066</td>
<td>25.5</td>
<td>22</td>
</tr>
<tr>
<td>1153A</td>
<td>41°16.5’S</td>
<td>129°48.9’E</td>
<td>5592</td>
<td>28</td>
<td>267</td>
</tr>
<tr>
<td>1154A</td>
<td>41°28.7’S</td>
<td>131°19.0’E</td>
<td>5747</td>
<td>28</td>
<td>233</td>
</tr>
<tr>
<td>1155A</td>
<td>41°57.5’S</td>
<td>127°59.7’E</td>
<td>4986</td>
<td>25</td>
<td>177</td>
</tr>
<tr>
<td>1156A</td>
<td>42°44.0’S</td>
<td>127°53.3’E</td>
<td>4878</td>
<td>23</td>
<td>118</td>
</tr>
<tr>
<td>1157A</td>
<td>43°15.7’S</td>
<td>128°53.2’E</td>
<td>5080</td>
<td>22</td>
<td>200</td>
</tr>
<tr>
<td>1158A</td>
<td>43°56.9’S</td>
<td>128°49.7’E</td>
<td>5167</td>
<td>21</td>
<td>198</td>
</tr>
<tr>
<td>1159A</td>
<td>45°57.4’S</td>
<td>129°60.0’E</td>
<td>4515</td>
<td>14</td>
<td>145</td>
</tr>
<tr>
<td>1160A</td>
<td>44°0.6’S</td>
<td>134°59.9’E</td>
<td>4636</td>
<td>22</td>
<td>166</td>
</tr>
<tr>
<td>1161A</td>
<td>44°17.2’S</td>
<td>129°31.1’E</td>
<td>5031</td>
<td>19</td>
<td>116</td>
</tr>
<tr>
<td>1162A</td>
<td>44°38.0’S</td>
<td>129°11.3’E</td>
<td>5475</td>
<td>18</td>
<td>333</td>
</tr>
<tr>
<td>1163A</td>
<td>44°25.5’S</td>
<td>126°54.5’E</td>
<td>4365</td>
<td>17</td>
<td>160</td>
</tr>
<tr>
<td>1164A</td>
<td>43°44.9’S</td>
<td>127°44.9’E</td>
<td>4809</td>
<td>19</td>
<td>138</td>
</tr>
</tbody>
</table>

Figure 8a. Basement depth plotted versus the square root of crustal age along selected flow lines crossing the SEIR between 90°E and 133.6°E. Flow lines are based on the plate kinematics parameters of Royer and Sandwell [1989]. Zero-age coordinates are indicated for every flow line. Note the very low subsidence rates between 0 and ~5–9 Ma as well as the slightly arcuate shape of the curves, which indicates that the subsidence rate progressively increases with age in the 0–25 Ma period.
two prominent fracture zones: near 120°E and 128°E, respectively. The change in subsidence rate is very abrupt across both fracture zones.

4.2. Temperature Dependence of the Mantle Physical Properties

Using commonly accepted, constant thermal parameters \( (\alpha = 3 \times 10^{-3} \text{ K}^{-1}, \kappa = 10^{-6} \text{ m}^2 \text{ s}^{-1}, \rho_m = 3300 \text{ kg m}^{-3} ) \) [Lister, 1977] and equation (1), temperature variations of about 350°C would be required to explain variations in thermal subsidence of 100 m Ma\(^{-1/2}\), which is unrealistic. Low values of \( \alpha \) and \( \kappa \) between 101°E and 128°E may partially explain part of the observed low subsidence rates. This ad hoc explanation is possible, but not satisfying because it cannot be supported by direct estimates of the mantle thermal parameters, while indirect estimates are affected by very large uncertainties [Patriat and Doucoure, 1992].

The temperature dependence of the mantle physical properties \( (k, \alpha, \rho, C_p) \) and the variable initial temperature of the melting column beneath the ridge axis do affect the subsidence rate [McKenzie et al., 2005]. We follow the approach of McKenzie et al. [2005] to analytically compute an overestimate of the subsidence rate variation that would result in response to a temperature change at the base of the plate. If active heat generation within the lithosphere and horizontal heat conduction are ignored, the temperature \( T(z, t) \) within a cooling plate satisfies

\[
\frac{\partial T}{\partial t} = \frac{1}{\rho C_p} \frac{\partial^2 G}{\partial z^2} - \frac{T}{\rho C_p} \frac{\partial (\rho C_p)}{\partial t} \tag{3}
\]

The equation being nonlinear, McKenzie et al. [2005] introduce the integral

\[
G = \int k(T) dT
\]

and write equation (3) as

\[
\frac{\partial T}{\partial t} = \frac{1}{\rho C_p} \frac{\partial^2 G}{\partial z^2} - \frac{T}{\rho C_p} \frac{\partial (\rho C_p)}{\partial t} \tag{5}
\]

The second term on the right-hand side being considerably smaller than the first, it is hereafter ignored. McKenzie et al. [2005] derive the temperature field by solving the resulting equation using standard methods. Then, they approximate the subsidence \( s(t) \) below the depth of the ridge by assuming isostatic compensation:

\[
s(t) = \frac{1}{(\rho_m - \rho_a)} \left\{ \int_0^l \rho[T(0, z)] dz - \int_0^l \rho[T(t, z)] dz \right\} \tag{6}
\]

The expression proposed is accurate to O(\( \alpha T \)). The subsidence rate per square root of age is

\[
a(t) = \frac{\partial s(t)}{\partial \sqrt{t}} = 2 \sqrt{t} \frac{\partial s(t)}{\partial t} = -2 \sqrt{t} \left( \frac{1}{(\rho_m - \rho_a)} \int_0^l \frac{\partial T(t, z)}{\partial t} \frac{\partial \rho(T)}{\partial T} dz \right) \tag{7}
\]
Then, ignoring the second term on the right-hand side of equation (5), we obtain

$$a(t) = -2\sqrt{t} \frac{1}{(\rho_m - \rho_w)} \int_0^L \frac{1}{C_p(T)} \frac{\partial^2 G}{\partial z^2} \frac{\partial \rho(T)}{\partial T} \, dz$$

(8)

Using the same relation between $\alpha(T)$ and $\beta(T)$ as McKenzie et al. [2005], we derive

$$a(t) = 2\sqrt{t} \frac{1}{(\rho_m - \rho_w)} \int_0^L \frac{\alpha(T)}{C_p(T)} \frac{\partial^2 G}{\partial z^2} \, dz$$

(9)

[28] From the definition of $G$ (equation (4)), we get

$$a(t) = 2\sqrt{t} \frac{1}{(\rho_m - \rho_w)} \int_0^{T_m} \frac{\alpha(T)}{C_p(T)} \frac{\partial}{\partial T} \left[ k(T) \frac{\partial T}{\partial T} \right] \, dT$$

(10)

where $T_m$ is the mantle temperature at the base of the plate.

Let us now evaluate the difference in subsidence rate for two different thermal configurations within the mantle, A and B, characterized by different plate thicknesses and basal temperatures: $L_A$, $T_{m,A}$ and $L_B$, $T_{m,B}$ respectively, with $T_{m,B} > T_{m,A}$ (as pointed out by McKenzie et al. [2005], the crustal thickness fixes the potential temperature of the mantle; hence only the thickness of the lithosphere remains as an adjustable parameter to fit the depth and heat flow.

Figure 9. (bottom) Bathymetric profile (plotted versus longitude) along the axis of the South Indian Ridge. Indications on axial morphology are after Cochran et al. [1997]. Average characteristic width ($w$) and length ($l$) of abyssal hills are in km, based on work by Goff et al. [1997]. (top) Along-axis subsidence rates variations plotted versus longitude. (middle) The [0–42 Ma] age range [after Hayes and Kane, 1994]. (top) This study subsidence rates, reassessed for the 0–25 Ma age range. Triangles and inverted triangles are for the northern and southern flanks, respectively.
Table 2. Temperature Dependence of $\alpha(T)$, $k(T)$, $\rho(T)$ and $C_p(T)$ Used by McKenzie et al. [2005]*

<table>
<thead>
<tr>
<th>Physical Properties</th>
<th>Dependence on Temperature</th>
<th>Coefficients</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thermal expansion coefficient</td>
<td>$\alpha(T) = \alpha_0 + \alpha_1 T$</td>
<td>$\alpha_0 = 2.832 \times 10^{-5}$ K$^{-1}$</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>$k(T) = \frac{\rho m}{C_p(T)} T + d_0 + d_1 T^2 + d_3 T^3$</td>
<td>$\rho = 3.79 \times 10^{-5}$ K$^{-1}$</td>
</tr>
<tr>
<td>Density</td>
<td>$\rho(T) = \rho_0 + \exp(-[\alpha_0(T - T_0) + \frac{1}{2}(T^2 - T_0^2)])$</td>
<td>$c = 0.0015$ K$^{-1}$</td>
</tr>
<tr>
<td>Specific heat at constant pressure</td>
<td>$C_p(T) = k_0 + \frac{k_1}{T^2}$</td>
<td>$d_0 = 1.753 \times 10^{-7}$ W m$^{-1}$ K$^{-1}$</td>
</tr>
</tbody>
</table>

* $T$ is expressed in K. For specific heat, experimental values for forsterite are used. To convert J mol$^{-1}$ into J kg$^{-1}$, the molecular weight of forsterite was taken equal to 140.69.

Observations wherever the crustal thickness is known; where it is not known, both $L$ and $T_m$ are required. Let $T_A(z, t)$ and $T_B(z, t)$ be the respective temperature fields within the plate, solutions of equation (5); $a_A(t)$ and $a_B(t)$ are the corresponding subsidence rate. We write

$$a_B(t) - a_A(t) = 2\sqrt{\frac{T_m}{\rho_m - \rho_o}} \left\{ \int^{T_2}_0 \frac{\alpha(T)}{C_p(T)} \left( k(T) \frac{\partial(T_B - T_A)}{\partial z} \right) dT + \int^{T_A}_{T_2} \frac{\alpha(T)}{C_p(T)} \left( k(T) \frac{\partial(T_B - T_A)}{\partial z} \right) dT \right\}$$

(11)

[30] The second term of the sum of the right-hand side in equation (11) is much smaller than the first and consequently, it is ignored. The derivative with respect to $T$ in the integral of the first term of the sum has a constant, negative sign between 0 and $T_m$. Hence the difference in subsidence rate between configurations A and B can be approximately overestimated as follows:

$$|a_B(t) - a_A(t)| \leq 2\sqrt{\frac{T_m}{\rho_m - \rho_o}} \max_{[0,T_2]} \left\{ \frac{\alpha(T)}{C_p(T)} \left( k(T) \frac{\partial(T_B - T_A)}{\partial z} \right) \right\}$$

(12)

[31] Because the term in the square bracket is a difference, it can be approximated (to the first order) using the analytical solution, $T_B^{PM}(z, t)$, proposed by Parsons and Sclater [1977] for a plate of thickness $L_A$ at temperature $T_m$ (for configuration B, exchange subscripts, from A to B). We thus obtain the following approximation:

$$|a_B(t) - a_A(t)| \leq 2\sqrt{\frac{T_m}{\rho_m - \rho_o}} \max_{[0,T_2]} \left\{ \frac{\alpha(T)}{C_p(T)} \left( k(T) \frac{\partial(T_B^{PM} - T_A^{PM})}{\partial z} \right) \right\}$$

(13)

[32] At great depths, the thermal gradient decreases to zero, so that the term in brackets can be approximated. We get

$$|a_B(t) - a_A(t)| \leq 2\sqrt{\frac{T_m}{\rho_m - \rho_o}} \max_{[0,T_2]} \left\{ \frac{\alpha(T)}{C_p(T)} \right\} |q_B^{PM} - q_A^{PM}|$$

(14a)

where $q_B^{PM}$ stands for the theoretical surface heat flow based on the Parsons and Sclater [1977] model. For ages younger than about ~40–50 Ma, the analytical expression for temperature based on the half-space model approximation [e.g., Davis and Lister, 1974] can be used to obtain

$$|a_B(t) - a_A(t)| \leq 2\sqrt{\frac{T_m}{\rho_m - \rho_o}} \max_{[0,T_2]} \left\{ \frac{\alpha(T)}{C_p(T)} \right\} \cdot \sqrt{\frac{k(T)}{\pi T_m}} |T_B^{PM} - T_A^{PM}|$$

(14b)

[33] Using $T_m = 1300^\circ$C and the numerical values (Table 2) based on the work by McKenzie et al. [2005], we obtain that a change in mantle temperature of ~50°C at the base of the plate cannot produce a change in subsidence rate any greater than about 35 m Ma$^{-1/2}$. Hence the observed variations in subsidence rate along the SEIR for the 0–25 Ma period cannot be explained solely by changes in mantle temperature (Table 2).

4.3. Crustal Thickness Variations

[34] Crustal thickness variations can dramatically affect residual depth anomalies [Marks et al., 1990], but not necessarily subsidence rates. Abrupt crustal thinning at a given age theoretically affects seafloor deepening by producing a characteristic saw tooth in the depth–age$^{1/2}$ profile, and by slightly affecting the slope of the curve in response to the mantle temperature variation that produced the crustal thinning. Basalts geochemistry consistently indicates that Indian mantle lavas have been derived by smaller degrees of melting than Pacific mantle lavas throughout the last ~28 Ma [Christie et al., 2004]. Hence the spatially abrupt change of subsidence rate across the eastern bound-
Figure 10. Seafloor depth versus square root of age (Ma), computed with a plate model, assuming that the mantle temperature progressively decreased between 14 and 7 Ma, from 1300°C to ~1230°C. The resulting theoretical subsidence curve does not reflect the shape of the observed curves (Figure 8).

4.4 Dynamic Effects

[36] Since the early 1970s, the AAD has been proposed to be the locus of deep seated dynamic effects [e.g., Weisell and Hayes, 1974]. If such effects are present, the SEIR topography is not compensated beneath the AAD and the half-space or the plate cooling models do not apply (equation (2) does not hold). The major problem is that the quantitative assessment of dynamic topography greatly depends on the physical model that is used to explain its existence. Because the Australian-Antarctic Depth Anomaly across the ocean basin reflects a cold anomaly within the mantle that persisted since at least ~100 Ma, our preferred hypothesis is that of Gurnis et al. [2002], according which the AAD could be located above a remnant, fossil slab that subducted to the southwest below the Australian Continent (Figure 11a), which was then part of Gondwanaland, during upper Cretaceous times [Gurnis et al., 1998, 2000]. Recently, Gurnis and Müller [2003] proposed model including a mantle wedge above the subducting slab that they state is consistent with (1) thinner crust and chaotic seafloor topography suggesting that a cold spot is presently being sampled within the AAD; (2) the continuous sampling by the SEIR of a cold anomaly since at least 45 Ma followed by a pulse resulting in the abrupt onset of fracture zone at ~20 to 25 Ma; and (3) the existence of a high-velocity (presumably cold) trapped transition zone anomaly.

[37] On the basis of the work by Gurnis and Müller [2003], the computed contribution to the observed seafloor topography (Figure 11), however, appears to be significant only within the AAD; west of 115°E, it dramatically decreases to zero. Therefore, while AAD-related dynamic effects may partially contribute to anomalous seafloor deepening between 120°E and 128°E, they cannot explain the subsidence pattern to the west of the AAD, e.g., between 101°E and 115°E, nor the abruptness of the subsidence rate variation across the eastern boundary of the AAD.

4.5 Effect of “Ubiquitously Distributed” Melt Within the Asthenosphere

[38] Partial melt is generally accepted to be a primary factor that influences mantle shear wave velocity anomalies (see for instance the recent, detailed discussion of Priestley and McKenzie [2006]. Forsyth [1992, p. 29] noted that “a velocity anomaly of 0.35 km/s (derived from observed travel time differences) in the 20-40 km depth range below the AAD would require an unrealistic mantle temperature difference...without the involvement of melt. In contrast, it could be explained by a change in melt fraction...requiring a temperature change of less than 100°C.” Following Turcotte and Schubert [1982, p. 183], let us hence assume that (1) the lithosphere thickness y_L(t) at a given age t corresponds to the depth at which the mantle is entirely solid and (2) the asthenospheric mantle, below y_L(t), contains an homogeneous melt fraction, hereafter noted φ_m, a regionally averaged melt fraction in volume (Figure 12). Then the density of lithosphere (\rho_L) and wet asthenosphere (\rho_m^w) are

\[ \rho_m^w \approx \rho_m [1 - \varphi_m] + \rho_m^f \varphi_m \]  

\[ \rho_L(T) \approx \rho_m [1 - \alpha(T - T_m)] \]

respectively, where \rho_m and \rho_m^f stand for the solid and liquid mantle densities (the expansion coefficient \alpha is for solid mantle). For young ages (<40 to 50 Ma), the temperature field within the boundary layer is again expressed using the half-space model approximation. Assuming isostatic compensation at the base of the plate, seafloor depth
can be written as follows [Turcotte and Schubert, 1982, Pioneer 183]:

\[
\begin{align}
\frac{r}{r_m/C_0} & = z_0 + \frac{\rho_m \alpha (T_m - T_0)}{r_m \left( 1 - \frac{\rho_m}{\rho_i} \right) \left( 1 - \frac{\rho_m - \rho_i}{\rho_m - \rho_w} \right)} \\
& \times 2 \sqrt{\frac{\kappa t}{\pi}} + \frac{\varphi_m (\rho_m - \rho_i)}{r_m \left( 1 - \frac{\rho_m - \rho_i}{\rho_m - \rho_w} \right)} y(t)
\end{align}
\]

where \( \lambda_1 \) is solution of

\[
\varphi_m \frac{H_1 \sqrt{\pi}}{C_p T_m} = e^{-\lambda_1} \frac{\lambda_1 \text{erf}(\lambda_1)}{\lambda_1}
\]

[39] Using the physics of sill solidification, Turcotte and Schubert [1982, pp. 172–174] derive the depth of the solidification interface, \( y_L(t) \), by balancing the heat conducted away from this interface and the heat released by solidification:

\[
y_L(t) = 2\lambda_1 \sqrt{\kappa t}
\]
and $H_L$ and $C_p$ are the mantle latent and specific heat, respectively. Note that in equation (17), $y(t)$ is proportional to the square root of age to the first order. The linear relation between seafloor depth and $\sqrt{t}$ is thus preserved in equation (16), which can be rewritten as follows:

$$z = z_0 + \left[ -\frac{\rho_m \alpha (T_m - T_0)}{(\rho_m - \rho_i) \left( 1 - \frac{\rho_m - \rho_i}{\rho_m - \rho_w} \right)} \frac{2 \sqrt{\kappa}}{\pi} \right] \left[ -\frac{\phi_m (\rho_m - \rho_i)}{(\rho_m - \rho_w) \left( 1 - \frac{\rho_m - \rho_i}{\rho_m - \rho_w} \right)} \frac{2 \lambda \sqrt{t}}{\kappa} \right]$$  \hspace{1cm} (19)

[40] The first term ($S_1$) of the right-hand side of equation (19) corresponds to the subsidence in absence of partial melt, while the second term ($S_2$) is the contribution of partial melt. Parameter $\phi_m$, as defined by Turcotte and Schubert [1982], is a “homogeneous melt fraction,” which would be equally distributed within the asthenosphere. This parameter depends on mantle temperature, $T_m$. Of regional significance, it is used to characterize the average density of the asthenosphere and has some bearing to seismic wave velocities. Hence it can be used to correlate seismic velocity and seafloor subsidence anomalies below ocean basins. This parameter $\phi_m$ must not be mistaken with the partial melt fraction, a function of $P$ and $T$, that characterizes the zone below the ridge from which melt is extracted to produce the axial crust. Mantle density variations due to partially molten rocks within this zone have hardly any effect on seafloor depth after a few millions years.

[41] Including the effect of ubiquitously distributed partial melt, $\phi_m$, adds term ($S_2$) to ($S_1$), the subsidence rate obtained in absence of melt. Fitting equation (19) to the data thus requires the adjustment of 4 parameters, instead of 3: $\alpha$, $\kappa$, $T_m$, but also $\phi_m$. The absolute value of $\phi_m$ needs to be known, while $\alpha$ and $\kappa$ need to be readjusted ($\alpha$ and $\kappa$ must be lower than expected when no melt is present). Such an enterprise is beyond the scope of the present paper. Absolute estimates of $\phi_m$ are not easy; if not impossible to obtain, but variations in melt fraction ($\Delta \phi_m$) can be approximated using general relationships between seismic velocity and melt fraction within the mantle. On the basis of the work by Ritzwoller et al. [2003], shear wave velocity negative anomalies at 100 km depth below the AAD do not exceed 8%, which, according to model calculations, can be ascribed to variations in melt fraction of 1% or less [e.g., Hammond and Humphreys, 2000]. For $\kappa = 10^{-6}$ m$^2$, $H_L = 400$ kJ kg$^{-1}$, $T_m = 1300$ °C and $\Delta \phi_m < -1\%$, we obtain $S_2 < -25$ m Ma$^{-1/2}$.

[42] Depletion of ubiquitously distributed melt within the asthenosphere is consistent with the existence of the Australian-Antarctic mantle anomaly resolved by Ritzwoller et al. [2003] extending to the west of the AAD. It may significantly contribute to the low subsidence rates observed within the AAD for the [0–25 Ma] period. However, it does not explain the sharpness of the subsidence rate variation across the eastern boundary of the AAD.

5. Conclusions

[43] Anomalously low subsidence rates characterize the flanks of the Southeast Indian Ridge for the 0–25 Ma period (subsidence of seafloor is less than 300 m Ma$^{-1/2}$ between 101°E and 118°E and less than 260 m Ma$^{-1/2}$

Figure 12. Sketch explaining the significion of parameter $\phi_m$: an “homogeneous melt fraction,” which would be equally distributed within the asthenosphere (crossed area). This parameter is supposed to depend on the mantle temperature, $T_m$, at the base of the plate. This parameter $\phi_m$ must not be mistaken with the partial melt fraction (a function of $P$ and $T$) that characterizes the zone below the ridge from which melt is extracted to produce the axial crust (grey area).
between 120°E and 128°E) while geophysical and geochemical evidence suggest that the expected along-axis variation in mantle temperature below the ridge crest probably does not exceed ~50°C. The variation of α and k with T affect the deepening of the seafloor with age, but the expected mantle temperature variations are too small to explain the full range of subsidence rate variations between 101°E and 128°E. These cannot be explained by one single effect but by a combination of factors in addition to mantle temperature.

[44] We successively considered four different factors: (1) the temperature dependence of the mantle physical properties; (2) variations in crustal thickness in response to an abrupt or progressive (between 7 and 14 Ma) mantle temperature decrease inferred from AAD basalt geochronology; (3) dynamic effects possibly created by an old, detached slab which subducted below Eastern Gondwana, creating a cold zone within the mantle below the AAD [Gurnis et al., 2000]; and (4) depletion in the “ubiquitously distributed melt fraction” (ϕm) that characterizes the asthenosphere.

[45] These effects may all contribute to the observed, anomalously low subsidence rate of the ridge flanks, with the most significant contribution being probably related to the depletion in ϕm. However, these effects have a deep-seated origin within the upper mantle, resulting in long-wavelength geophysical variations. They altogether probably explain that the subsidence is anomalously low between 101°E and 128°E. None of them can explain the abruptness of the transition across the fracture zones that delineate the boundaries of the AAD, near 120°E and near 128°E, respectively.

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References


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