CHAPTER VI

MODEL STUDIES: THERMAL STRUCTURE OF THE OCEANIC LITHOSPHERE AND FREE-AIR GRAVITY MODEL STUDIES AT 79°E FRACTURE ZONE
CHAPTER VI

MODEL STUDIES: THERMAL STRUCTURE OF THE OCEANIC LITHOSPHERE AND FREE-AIR GRAVITY MODEL STUDIES AT 79°E FRACTURE ZONE

6.1 THERMAL MODELS OF OCEANIC LITHOSPHERE

The broad structure of the oceanic lithosphere was initially derived from the regional ocean floor topography and heat flow data. Based on these data, models have been proposed which fall into two categories. One class of models (plate model) are represented by a slab of finite and constant thickness which cools with distance from the axis of accretion with the base of the slab held at a constant temperature (McKenzie, 1967; Sclater and Francheteau, 1970). The second category (boundary layer model) assumes that the lithosphere is a semi-infinite half space, which increases in thickness with age (Parkér and Oldenburg, 1973; Forsyth, 1977). These models have explained the general decrease in heat flow with age and the increase in depth away from the ridge axis. The raw heat flow data are highly scattered, hence more emphasis has been placed on the ocean floor topography in developing the quantitative models of the oceanic lithosphere. Sclater et al. (1971) have shown that most mid ocean ridges are at a depth of 2500 m and the depth increases with age indicating a general subsidence. It was shown that for the crust younger than 80 Ma the depth increases linearly with the
square root of age (Davis and Lister, 1974). This relation is not valid for older ocean floor and it was found that beyond 80 Ma the depth increases exponentially to a constant value of 6400 m (Parsons and Sclater, 1977). This two stage relationship between the depth and age was explained in terms of formation of a thermal boundary layer. The magma cools as it moves away from the spreading center and the thickness of the rigid layer thus created increases (Figure 6.1). Parker and Oldenburg (1973), predicted that the thickness of the rigid layer increases as the square root of age. Furthermore Parsons and Sclater (1977) have demonstrated that for the crust younger than 80 Ma, both plate and boundary layer models approximate each other in defining the isothermal surfaces and they differ principally in their lower boundary conditions (Figure 6.1). The boundary layer cooling model predicts an error function geotherm which is a good approximation of the actual geotherm for ages less than 70 Ma (Parsons and Sclater, 1977).

As per the thermal models the lower boundary of the lithosphere can be represented by an isotherm $T_M$, at temperature greater than $T_M$ mantle rocks behave like fluid and at temperature below $T_M$, the rocks of the mantle and crust are essentially rigid at low stress levels (Turcotte and Oxburgh, 1967, 1972). The value of $T_M$ was suggested to be about 1000° ± 200°C and the average thickness of the lithosphere about 100 km (Turcotte, 1974).
Figure 6.1 Schematic thermal models of the oceanic lithosphere. A) Boundary layer model. B) Constant plate model. (after Sclater et al., 1980). $T_M$ indicates the deep mantle temperature. The solid lines represent isothermal surfaces within the cooling lithosphere.
Studies of lithospheric bending at Hawaii (Walcott, 1970) and at trenches (Hanks, 1971) have shown that the oceanic lithosphere exhibits elastic behavior on time scales of millions of years. Based on flexural rigidity values it was suggested that the upper 25 km of the lithosphere behaves like an elastic brittle material on geological time scales. In the depth range 25-100 km the oceanic lithosphere exhibits plastic behavior, at low stress levels this region acts as a stress guide and plastic deformation takes place when subjected to large stress (above 10 k bars) (Walcott, 1970). The boundary of the top layer of the lithosphere which behaves like an elastic brittle material can be represented by an isotherm $T_E$, the value of $T_E$ was suggested to be about 300°C (Turcotte, 1974). Based on these observations, Turcotte (1974) proposed a broad division of the oceanic lithosphere (away from the ridge axis) depending on the deformation mechanism. The thickness of the elastic brittle layer is termed as the effective elastic thickness ($H_E$) of the lithosphere.

<table>
<thead>
<tr>
<th>DEPTH RANGE IN km</th>
<th>DEFORMATION MECHANISM</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 - 25</td>
<td>Elastic Brittle</td>
</tr>
<tr>
<td>25 - 100</td>
<td>Plastic</td>
</tr>
<tr>
<td>&gt; 100</td>
<td>Fluid</td>
</tr>
</tbody>
</table>
Thus the oceanic lithosphere can be viewed as a combination of mechanical and thermal boundary layers (Sclater et al., 1980; Parsons and McKenzie, 1978).

Mechanical models have been further refined from the studies of flexure of the oceanic lithosphere at Hawaiian-Emperor chain (Watts, 1978; Caldwell and Turcotte, 1979; Watts et al., 1980; Bodine et al., 1981). These studies have shown that the effective elastic thickness of the lithosphere is approximately proportional to the square root of the age.

6.2 THERMAL STRUCTURE OF THE LITHOSPHERE AT FRACTURE ZONES

The various thermal and mechanical models of the lithosphere discussed above demonstrate a prominent age-dependent thermal structure of the lithosphere. Fracture zones in the ocean basins which represent inactive traces of the transform faults depict unique asymmetrical contact between separate sections of the same lithospheric plate having substantially different thermal structures. The difference in the thermal structures arises from the age offset of the lithospheric sections which are in contact at the fracture zones. Topography along the fracture zones consist of long ridges, troughs and scarps which separate regions of different depth (Menard and Atwater, 1969). This topographic expression of the fracture zone is preserved and can be traced for thousands of kilometers. The persistence of the
topographic expression suggests that the generation and evolution of the oceanic lithosphere proximal to the transform boundaries is different from the normal oceanic lithosphere (Gallo et al., 1986). Even though there is persistence of the topography due to the relative vertical subsidence of the crust with age, if there is sufficient sedimentation the fracture zone may get buried. The diminished bathymetric signature of the Clipperton fracture zone in the Pacific ocean is due to the sediment in-fill and this in turn resulted in small amplitude geoid anomalies (Wessel and Haxby, 1990). The persistence of ridge and trough topography along the fracture zones has been attributed to the thermomechanical interactions in the lithosphere driven by the age contrast across the fracture zone (Sandwell and Schubert, 1984). Figure 6.2 shows the thermomechanical model of the lithosphere at a fracture zone.

6.2.1 FLEXURAL RESPONSE OF THE LITHOSPHERE

As a consequence of the differential subsidence due to the age contrast across the fracture zone, the lithosphere adjacent to the fracture zone undergoes flexure and maintains the topographic expression (Sandwell and Schubert, 1982). These authors have demonstrated that the lithospheric flexure occurs across the Pioneer and Mendocino fracture zones in the north west Pacific. Figure 6.3 illustrates the evolution of the fracture zone and the flexural response of the lithosphere. At the ridge axis (A-A', Figure 6.3) the newly created lithosphere on the younger side separates the older
Figure 6.2 Thermomechanical model of the oceanic lithosphere at a fracture zone (after Sandwell, 1984).
lithosphere. As the crust generated by the two segments evolve to B-B'(Figure 6.3), the depth far from fracture zone on the younger lithospheric segment increases at a higher rate than the depth far from fracture zone on the older segment (Menard and Atwater, 1969; Delong et al., 1977; Sibuet and Mascle, 1978). This mechanism is depicted by the subsidence curves shown at the right side of the Figure 6.3. By the time the two segments evolved to B-B' the overall change in depth across the fracture zone, $h_B$, is less than the initial bathymetric step $h_A$. However, the height of the scarp at the fracture zone remains equal to $h_A$. If there is no vertical slip at the fracture zone, the lithosphere must flex to maintain the expression of the scarp as shown on the lower portion of the Figure 6.3. The amplitude of this flexure $\delta_B$ is the change in the differential subsidence from A to B, i.e., $h_A - h_B$.

The shape of the flexural topography depends on the effective elastic thickness $H_E$ of the lithosphere (Sandwell and Schubert, 1982). Since $H_E$ is proportional to the square root of the age, the flexure is asymmetric about the fracture zone with the younger lithosphere flexing at a shorter wave length than the older lithosphere. The effective elastic thickness again depends on the thermal structure controlled by the depth to the isotherm $T_E$. These observations suggests that the thermal structure across the fracture zone strongly influences the overall fracture zone topography.

If the lithosphere is in local isostatic equilibrium, the effects of lateral heat conduction and local isostacy results in rapid smoothening of the topography with age. However, it is found that the topographic expression
Figure 6.3 Schematic illustration of the flexural response of the lithosphere at a fracture zone. Subsidence curves of the ridge segments are shown at the right.
persists along the fracture zones. Component of lithospheric flexure is responsible for the persistence of ridge and trough topography (Sandwell, 1984). Lithospheric flexuring implies that the isostatic compensation is regional rather than local.

The density structure within the lithosphere is sensitive to the temperature changes. In the present study the thermal structure of the lithosphere is computed to predict the probable density variations within the lithospheric section considering a thermomechanical model (Figure 6.2) of the fracture zone.

6.3 COMPUTATION OF THERMAL STRUCTURE AT A FRACTURE ZONE

In this section the thermal structure of the oceanic lithosphere at the fracture zone is computed based on the thermomechanical model considering the lateral heat conduction across the fracture zone, following the methods proposed by Louden and Forsyth (1976) and Sandwell (1984).

In the region outside the active transform fault, the two dimensional thermal structure is found by solving the time dependent heat conduction equation:

\[ \frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2} = \frac{1}{k} \frac{\partial T}{\partial t} \quad --- (1) \]
Where $k$ is the thermal diffusivity, which is assigned a value of $8 \times 10^{-7}$ sq. m per sec. \cite{Parsons and Sclater, 1977}.

Initially at $t=0$ age, at the ridge axis near the transform fault, the geotherm on the right side of the fracture zone (Figure 6.2) corresponds to an age of $t_0$ and the left side has a constant temperature $T_M$. Thus, initially, the thickness of the elastic layer on the left (younger) side is zero and a continuous elastic layer develops across the fracture zone away from the transform fault as the fracture zone evolves.

The initial temperature distribution is:

$$T(x, z, t_0) = \begin{cases} 
T_S + (T_M - T_S) \text{erf} \left[ \frac{z}{2 \sqrt{kt_0}} \right] & \text{if } x > 0 \\
T_M & \text{if } x < 0 
\end{cases} \quad (2)$$

The surface boundary condition is $T(x, 0, t) = T_S$

Where $T_S$ is the sea water temperature ($\approx 0$ °C), $T_M$ is the temperature of the deep mantle, 1365°C \cite{Parsons and Sclater, 1977} and $t$ is the time.
Away from the initial conditions the thermal structure across the fracture zone is a function of the age contrast. At a given depth and distance from the fracture zone, the isotherms can be computed.

For the initial temperature distribution given above, the convolution integral can be evaluated analytically (Sandwell and Schubert, 1982), and is:

\[
T(x, z, t) = T_S + \frac{(T_M - T_S)}{2} \left[ \operatorname{erfc} \left( \frac{x}{2\sqrt{k(t-t_0)}} \right) \operatorname{erf} \left( \frac{z}{2\sqrt{kt}} \right) + \operatorname{erfc} \left( \frac{x}{2\sqrt{k(t-t_0)}} \right) \operatorname{erf} \left( \frac{z}{2\sqrt{k(t-t_0)}} \right) \right]
\]  

--- (3)

The equations (2) and (3) are modified from that of Sandwell (1984) by changing the sign of \(x\), in order to simulate the effects of a right laterally offset fracture zone, wherein, the right side of the fracture zone represents older crust (Figure 6.4).

The function \(\operatorname{erfc}(x)\) ranges from 2 to 0 as \(x\) varies from \(-\infty\) to \(\infty\). Thus, far from the fracture zone the geotherm follows the error function of the boundary layer convection model (Turcotte and Oxburgh, 1967). As a result of lateral heat conduction, temperatures vary continuously across the fracture zone (Louden and Forsyth, 1976).
The base of the lithosphere is an isotherm equal to a fraction of the mantle temperature, given as 0.9$T_M$ (Sandwell, 1984). The base of the elastic layer as defined by the stress relaxation temperature $T_E$, is suggested to be in the range of 300°C to 600°C (Watts et al., 1980). A value of 450°C for $T_E$ (Sandwell and Schubert, 1982) has been considered in the present study.

In Figure 6.4, the computed isotherms 0.9 $T_M$ and $T_E$, obtained from the equation (3), for an age offset of 2.08 My across the 79°E fracture zone are shown. The physical constants and the model parameters are summarized in Table 6.1

6.3.1 FLEXURAL RIGIDITY

The elastic properties of the lithosphere are usually expressed in terms of flexural rigidity (Cochran and Talwani, 1979). The flexural rigidity can be computed using the relation:

$$D = \frac{E H^3}{12 (1 - \nu^2)}$$  \hspace{1cm} (4)
Where,

- \( D \) - Flexural Rigidity
- \( E \) - Young's Modulus
- \( H_E \) - Effective elastic layer thickness
- \( \nu \) - Poisson's ratio

The variations of the flexural rigidity across the 79°E fracture zone are computed using the above relation. The effective elastic thickness of the lithosphere was obtained from the isotherm defining \( T_E \) and the constants Young's modulus and Poisson's ratio were assumed (Table 6.1). The variation of the flexural rigidity of the lithosphere across 79°E fracture zone is shown at the bottom part of the Figure 6.4.

### 6.3.2 Density Structure

The variations in the density are related to the temperature changes, due to the thermal expansion. The following equation of state is used to compute the density variations corresponding to the temperature changes.

\[
\sigma_n(x, z, t) = \sigma_M \left[ 1 - \alpha_n (T(x, z, t) - T_M) \right] \quad (5)
\]
### TABLE T6.1

Physical constants and model parameters used

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Definition</th>
<th>Value/ Units</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha$</td>
<td>Thermal expansion coefficient</td>
<td>$3.1 \times 10^{-5} , \text{k}^{-1}$</td>
<td>Parsons and Sclater (1977)</td>
</tr>
<tr>
<td>$E$</td>
<td>Young's modulus</td>
<td>$6.5 \times 10^4 , \text{MPa}$</td>
<td>Sandwell and Schubert (1982)</td>
</tr>
<tr>
<td>$T_E$</td>
<td>Stress relaxation temperature</td>
<td>$450^\circ\text{C}$</td>
<td>Watts et al. (1980)</td>
</tr>
<tr>
<td>$T_M$</td>
<td>Mantle temperature</td>
<td>$1365^\circ\text{C}$</td>
<td>Parsons and Sclater (1977)</td>
</tr>
<tr>
<td>$T_S$</td>
<td>Surface temperature</td>
<td>$0^\circ\text{C}$</td>
<td></td>
</tr>
<tr>
<td>$\nu$</td>
<td>Poisson's ratio</td>
<td>$0.25$</td>
<td></td>
</tr>
<tr>
<td>$D$</td>
<td>Flexural rigidity</td>
<td>newton-meter</td>
<td></td>
</tr>
<tr>
<td>$\delta$</td>
<td>Flexural amplitude</td>
<td>meters</td>
<td></td>
</tr>
<tr>
<td>$H_E$</td>
<td>Effective elastic thickness</td>
<td>meters</td>
<td></td>
</tr>
<tr>
<td>$E$</td>
<td>Young's modulus</td>
<td>megapascals</td>
<td></td>
</tr>
</tbody>
</table>
Figure 6.4 Computed isotherms at 79°E fracture zone. Stress relaxation temperature (Te) 450°C and deep mantle temperature (Tm) 1365°C are considered. The bottom part of the figure shows the variation in flexural rigidity across the 79°E fracture zone.
Where $\sigma_M$ is the density of the mantle at temperature $T_M$ and is assumed to be 3400 kg/m$^3$, and $\alpha$ is the thermal expansion coefficient and is considered as $3.1 \times 10^{-1}$K$^{-1}$ (Parsons and Sclater, 1977).

Densities obtained from the above relation are utilized to model the free-air gravity anomaly to deduce probable density structure and to estimate lithospheric thickness across 79°E fracture zone.

6.4 FREE-AIR GRAVITY MODEL STUDIES

The response of the differential subsidence and the lithospheric flexure across the fracture zone result in step like free-air gravity anomalies (Sibuet et al., 1974) and the geoid anomalies (Parmentier and Haxby, 1986; Sandwell, 1984; Wessel and Haxby, 1990). Several authors have suggested the use of gravity anomalies as a means of predicting the probable density structure and thickness of the lithosphere, associated with the cooling plate model (Le Pichon et al., 1973; Sibuet et al., 1974; Dorman, 1975; Sibuet and Mascle, 1978).

The free-air gravity anomalies observed across the 79°E fracture zone are modeled following the procedure of Talwani et al., (1959), assuming two dimensional bodies. The gravitational edge effect anomalies that are observed across the fracture zones are best reflected over profiles perpendicular to the strike of the fracture zone. Therefore, data along an E-W
line (AA-AA', Figure 2.8) was considered by retrieving the observed gravity and water depth values from the closely spaced (5 miles) N-S lines for the purpose of model studies. The free-air gravity along the profile AA-AA' (Figure 6.5) thus represents data points spaced at 9.2 km and reflects long wavelength character. The average depth on the western and eastern sides of the fracture zone are 5250 and 5400 m respectively. The age of the oceanic crust as derived from the magnetic studies along this profile are 59.47 Ma on the western (younger side) and 61.55 Ma on the eastern (older) side of the fracture zone. The age offset across the fracture zone along the profile AA-AA' is 2.08 Ma.

The gravity effect of the lithospheric section is computed by assuming three layers, the sediment, the oceanic crust and the upper mantle. In the absence of any refraction data a sediment layer of 500 m with a density of 2260 kg/m$^3$ was assumed. These values are assumed considering the sediment thickness reported at the DSDP sites 214 and 238. The sediment layer is followed by the oceanic crust for which a crustal thickness of 6.0 km and a density of 2800 kg/m$^3$ was assumed. Global average thickness of the oceanic crust was suggested to be about 6.0 km by Spudich and Orcutt (1980). The remaining part of the lithospheric section is divided into layers of constant density differing by 10 kg/m$^3$ starting from 3400 Kg/m$^3$ (Figure 6.5), following a similar approach as that of Sibuet et al. (1974).
Figure 6.5  Free-air gravity model showing the lithospheric section at 79°E fracture zone. (a) The isotherms of $T_E$ and $T_M$ are shown. (b) Computed and observed free-air gravity along a profile AA-AA' (see figure 2.8). The bottom part of the figure shows the difference between the computed and observed free-air gravity for different assumed thicknesses of the lithosphere.
The average layer thickness within the upper mantle was constrained by the thermal structure. The depth at which a given density occurs has been determined by successive iterations using the thermal model equations (2) and (3) and the corresponding density by eq. (5). These density and layer thicknesses were incorporated in the model and a best fit was obtained (Figure 6.5). In order to satisfy the gravity low at the fracture zone more thickness of the sediment and crustal layers were invoked.

After having achieved the probable density structure and the lithospheric thickness estimate, the response of the computed gravity for various assumed thicknesses of the lithosphere was tested. These tests were made by varying the thickness of the lithosphere, while all other parameters were kept constant. The bottom part of the Figure 6.5 shows the difference between the observed and the computed gravity values, for various assumed thicknesses of the lithosphere. The least difference is observed for a lithosphere with a thickness of 100 km (Figure 6.5).

*Parker and Oldenburg, (1973)* proposed an empirical relationship to estimate the lithospheric thickness as $Z = 9.4 \times \sqrt{\text{age}}$. This relationship approximates the lithospheric thickness within ± 25 km. In the present study area the thickness of the lithosphere works out to 72.48 km for a crustal age of 59.47 Ma, using this empirical relation. The estimated lithospheric thickness obtained from the thermal and free-air gravity model is 100 km.
Sibuet et al. (1974) demonstrated that the estimates of lithosphere thickness can be obtained from the observed gravitational edge effect across the fracture zone. Estimations of lithospheric thickness were obtained across the Mendocino fracture in the Pacific (Sibuet et al., 1974) and Romanche and Chain fracture zones in Atlantic (Sibuet and Masque, 1978), using the thermal conduction model, with the assumption of no horizontal heat conduction across the fracture zones. In the present study, the estimates of lithospheric thickness were computed from the gravitational edge effect observed across the 79°E fracture zone, considering the thermal structure including the effects of horizontal heat conduction. These estimates give a thickness of about 100 km, which within the acceptable limits, matches the value obtained by the empirical formula suggested by Parker and Oldenberg (1973). More realistic estimates can be achieved by including the crustal information, from seismic refraction studies.