RICE UNIVERSITY

Diffuse Oceanic Plate Boundaries: Kinematic Observations and Rheological Constraints

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A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

Master of Science

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MARCH 2006
ABSTRACT

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We use a thin viscous sheet model to interpret the observed shortening along a seismic reflection profile collected on the Phèdre cruise (on the deforming oceanic lithosphere in the central Indian Ocean). We determine the 95% confidence region of the 1/e fall-off distance of strain rate to be between 7.1º and 20.4º, the vertically averaged power-law exponent to be between 1.6 and 18.5, and the ratio of the strength of the upper lithosphere to the lower lithosphere to be between 1 and 5.5. We calculate a simplified strength envelope of a two layered oceanic lithosphere based on the lower lithosphere deforming in accordance with the Dorn equation. This model estimates a range of between 0.2 and 20 TN m⁻¹ for the strength of the oceanic lithosphere. The ratio of the strength of the upper lithosphere with respect to the lower lithosphere ranges between 8.6 and 13.98.
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Chapter 1
Rheological constraints on deforming oceanic lithosphere in the central Indian ocean
1 Summary

We treat the deforming lithosphere in the equatorial Indian Ocean as a thin viscous sheet of fluid with a vertically averaged power-law rheology. We assume that the Indian component plate acts as an indenter on the region of deformation causing high strain rates adjacent to the Indian plate, which decay with distance southwards. We estimate the 1/e fall-off distance and total convergence by fitting the expected decay of a thin viscous sheet to the observed shortening as inferred from a seismic reflection profile collected on the Phèdre cruise, a 2100-km-long north-south profile along 81.5°E. This gives large 95% confidence intervals of 3.9° to 25.8° (432.9 km to 2863.8 km) for the 1/e fall-off distance and 15 km to 82 km for the total convergence across the zone of deformation. These limits on total convergence are much larger than those (22 km to 37 km) estimated by the original workers [Chamot-Rooke et al. 1993] and reconciles the apparent discrepancy between the original estimate of convergence from the seismic profiles and the amount of convergence indicated by plate reconstructions [DeMets et al. 2005]. The estimates of the total convergence and the 1/e fall-off distance strongly covary and we obtain narrower 95% limits of 7.1° to 20.4° (788 km to 2269 km) for the 1/e fall-off distance by limiting the convergence to the interval 41.3 km ± 2.3 km, as found from plate reconstructions [DeMets et al. 2005]. Assuming that the indenting lithosphere continues along strike of the deforming zone from the India-Capricorn pole of rotation at 74°E to the Sumatra trench at ≈100°E, results in a 95% confidence interval of 1.6 to 18.5 for the power-law exponent for the vertically averaged rheology of deforming oceanic lithosphere. At a reference strain rate of $10^{-16} \text{s}^{-1}$ these values indicate the ratio of the
strength of the upper lithosphere to that of the lower lithosphere is between 1 and 5.5, which indicates that the lower lithosphere supports at least $\approx 18\%$ of the force per unit length that the Indian and Capricorn plates exert on one another.

2 Introduction

A large part of the converging portion of the India-Capricorn diffuse oceanic plate boundary is blanketed with sediments of the Bengal Fan, which provide a medium in which deformation by thrust faulting can be imaged. From a seismic profile that traverses the length of this deforming oceanic lithosphere, Chamot-Rooke et al. (1993) estimated the amount of shortening along the profile by estimating the throws and dips of thrust faults. Among their results, they showed that deformation starts abruptly near the north end of the profile and that the north-south shortening was greatest in the north and decays with distance to the south. They estimated total north-south convergence of 22 to 37 km. In contrast, DeMets et al. [2005] used plate reconstructions to estimate north-south convergence since $\approx 8$ Ma along the same longitude of 41.3 km $\pm 2.3$ km (95 % confidence limits). The cause of this difference has been unclear.

Here we interpret this shortening profile in terms of deformation of a thin viscous sheet of fluid, which is a model that has previously been applied to deforming continental lithosphere [e.g., England and McKenzie, 1982]. We consider the following questions: (1) Can deforming oceanic lithosphere be usefully modeled as a thin viscous sheet of fluid? (2) If so, what is the best estimate (and uncertainty) of the total convergence along the north-south line sampled by the Phèdre profile?
Moreover, what causes the apparently significant difference between the convergence estimated from seismic profiles (22–37 km [Chamot-Rooke et al. 1993]) and from plate reconstructions (41.3 km ± 2.3 km [DeMets et al. 2005])? (3) What is the best estimate (and uncertainty) of the 1/e fall-off distance of strain as a function of distance in the thin viscous sheet? (4) What is the best estimate of strain rate and its spatial variation in the India-Capricorn diffuse plate boundary? (5) What is the best estimate (and uncertainty) for the power-law exponent of the vertically averaged rheology of deforming oceanic lithosphere? (6) What bounds does the value of the power-law exponent place on the relative strengths of the upper lithosphere, which deforms brittlely or semi-brittlely, and of the lower lithosphere, which deforms by creeping flow?

3 Background

3.1 Thin Viscous Sheet Model

Wide regions of large-scale deformation occurring inside traditional rigid plates have come to be called diffuse plate boundaries [Gordon and Stein, 1992] and the adjacent plates (inside a traditional plate) are called component plates [Royer and Gordon 1997; Gordon 1998]. Past researchers have considered deforming continental lithosphere as a thin sheet of viscous fluid with an indenter on one side [England et al 1985]. This approximation involves balancing the buoyancy forces against the vertically averaged deviatoric stresses required to deform the lithosphere. In the case of oceanic lithosphere the stresses are averaged across two layers (the upper
lithosphere, which deforms brittlely or semi-brittly [Kohlstedt et al. 1995], overlying a layer that deforms by creeping flow, which can be modeled as a power-law fluid with an exponent of \( \approx 3 \) to calculate the vertically averaged deviatoric stress. Gordon [2000] showed that the deforming oceanic lithosphere can be considered as a power law fluid for strain rates from \( 10^{-17} \text{s}^{-1} \) to \( 10^{-15} \text{s}^{-1} \), which span the strain rates of interest in these regions.

England et al [1985] considered a deformation model of a thin viscous sheet of fluid. They use a velocity boundary condition of \( v = V_0 \cos(2\pi x/\lambda) \), where \( x \) is distance along the strike of the edge of the indenter, \( \lambda \) is the characteristic along-strike wavelength, and \( V_0 \) is the maximum velocity. This boundary condition describes the action of an indenter causing deformation along \( y=0 \) (Fig.1). From this model they calculated the \( y \) component of velocity at a distance \( y \) from the edge of the indenter

\[
v_y = V_0 e^{-\sqrt{n} \pi y / \lambda}
\]  

where \( n \) is the power-law exponent of the fluid. The derivative of velocity with respect to \( y \) gives one component of the strain rate tensor, namely

\[
\dot{e}_{yy} = \frac{\partial v_y}{\partial y}
\]  

(2)

Now if we consider instantaneous time and integrate strain rate with respect to time we get strain

\[
\epsilon_{yy} = \int \dot{e}_{yy} \, dt
\]  

(3)

Integrating this strain this strain along \( y \) gives the cumulative displacement.
\[ C = C_0 \left( 1 - e^{-\sqrt{n\pi} y / \lambda} \right) \]  

(4)

where \( C_0 \) is the total convergence across the deformation zone and we have required that \( C (y=0) = 0 \) and that \( C (y\to\infty) = C_0 \). The across-strike width within which all but 1/e of the total convergence occurs is \( y_0 \) where \( y_0 \) is given by (England et al 1985)

\[ y_0 = \frac{\lambda}{\pi \sqrt{n}} \]

(5)

Substituting this into equation 4 gives

\[ C = C_0 (1 - e^{-y/y_0}) \]  

(6)

3.2 India-Capricorn Diffuse Plate Boundary

The India-Capricorn diffuse plate boundary is a large region of oceanic deformation occurring in the equatorial Indian Ocean. It is bounded by the Indian component plate on the northern side and the Capricorn plate on the southern side (Fig. 1). This pole of rotation of the Indian component plate relative to the Capricorn component plate lies near 5°S, 74°E; thus we expect north-south contractional deformation to occur east of \( \approx 74^\circ \)E and north-south extensional deformation to occur west of \( \approx 74^\circ \)E, as is observed.

Some researchers believe that the deformation in the central Indian ocean was driven, and continues to be driven by force exerted by the Tibetan plateau onto the Indian plate (after the Indo-Tibet collision), thus causing high stress in the Indo-Australian plate, which deforms the Indo-Australian plate creating diffuse boundaries between the Indian, Capricorn and Australian plate.
3.3 The Phèdre Line

Two seismic lines apparently run across the length of the Central Indian Ocean deforming region: the Phèdre line at 81.5°E (Chamot-Rooke et al., [1993]) and the Conrad line at 78.8°E (Van Orman et al., [1995]). Here we consider only the Phèdre line because the uncertainty involved in fitting the Conrad profile to the model is very high, which makes it less useful (Appendix 1). The Phèdre cruise on the French research vessel Marion Dufresne conducted seismic reflection profiling across the region of contractional deformation region in the equatorial Indian Ocean (India-Capricorn diffuse oceanic plate boundary). The longest of their lines was a 2100-km-long profile that ran along 81.5°E from 14°S to the coast of Sri Lanka. One hundred and thirty four faults or folds were identified in the sediments along this profile. Vertical throw was measured using the maximum vertical uplift of the sedimentary reflectors at the base of the sedimentary cover, which occurs at some distance from the fault [Chamot-Rooke et al. 1993]. To estimate horizontal throw from vertical throw, an estimate of fault dip is needed. Forty two of the 134 faults or folds were associated with a dipping crustal reflector interpreted as a thrust fault. By using bounds of 5 km s\(^{-1}\) to 7 km s\(^{-1}\) for the crustal wavespeed, Chamot-Rooke et al [1993] estimated the dips of these faults assuming a plane-fault geometry. For the other 92 presumed faults, Chamot-Rooke et al [1993] used the mean value of the dip of the 42 imaged faults, giving rise to extreme values of 36° for a wavespeed of 5 km s\(^{-1}\) and 45° for a wavespeed of 7 km s\(^{-1}\). Here, following Van Orman et al. [1995], we assumed a plane-fault geometry and a dip of \(\approx 40^\circ\) for all faults in constructing our best estimate of the horizontal throws and cumulative convergence (Figure 5). We
then use the limiting values of Chamot-Rooke et al [1993] in estimating uncertainties on the horizontal throws and cumulative convergence.

4 Methods

Here we consider the deforming lithosphere in the central Indian Ocean as a thin sheet of viscous fluid. We assume that India acts as an indenter onto this fluid causing deformation. Using the faulting and folding observed along the Phèdre line (in this region of deformation) as a measure of strains we incorporate this data into the thin viscous sheet model. To do this we need to fit the observed cumulative convergence versus distance (Figure 5: representing the observations along the Phèdre line) using the model described in equation 6 (representing the thin viscous sheet model),

\[ C = C_0 \left(1 - e^{-y/y_0}\right) \]

( where \( y_0 = 4D/\pi \sqrt{n} \) )

The point of contact between the Indian indenter and the deforming fluid is unknown, we modify the previous equation by adding another parameter \( y_D \), such that the uncertainty in this region is defined in a band of 0.5° N to 0.5° S.

\[ C = C_0 \left(1 - e^{(y_0-y)/y_0}\right) \quad (7) \]

This equation has three adjustable parameters, \( C_0 \) (the integral of north-south strain along the Phèdre line), the total convergence, \( y_0 \), the 1/e fall-off distance of strain rate, and \( y_D \), the location of the edge of the indenter along the y axis. By adjusting these three parameters we obtain the best possible fit to the Phèdre data, thus
essentially minimizing the errors in fitting the data to the equation. This gives us the 95% confidence limits for these three different parameters.

4.1 Curve Fitting and Uncertainties

4.1.1 Uncertainties

In this section we fit the Phèdre convergence curve data using the equation above. The uncertainties involved in fitting the data to the model are listed below:

1. The Phèdre data that we use consists of 134 estimated vertical throws that were recorded by seismic observations.

2. Each of these 134 vertical throws are converted into horizontal throws using fault dips (based on planar geometry) which are unknown in all but 42 of these cases and even these 42 cases have uncertainties associated with them.

3. The cumulative convergence at each fault is estimated by adding the horizontal throws of all the faults between \( y \) and \( y_D \).

4.1.1.1 Covariance Matrix

We fit the cumulative curve (Fig. 5), where each estimate of convergence at a fault is found by summing many individual estimates of horizontal throw upto, and including, that fault. Thus the errors at every fault add on to the next fault which causes the errors from fault to fault to be highly correlated. A diagonal covariance matrix would give meaningless estimates of the uncertainties in the estimated parameters because it is appropriate only if there is no covariance between different estimates. To solve this problem, we not only assign uncertainty values to every data
point but we also incorporate uncertainty in every point due to all other correlated points. Thus, to do this we use the covariance matrix, cov, where

\[
\begin{bmatrix}
1 & 1 & 1 & 1 & \ldots & 1 \\
1 & 2 & 2 & 2 & \ldots & 2 \\
1 & 2 & 3 & 3 & \ldots & 3 \\
1 & 2 & 3 & 4 & \ldots & 4 \\
1 & 2 & 3 & 4 & 5 & \ldots & 5 \\
\vdots & \vdots & \vdots & \vdots & \ddots & \vdots \\
1 & 2 & 3 & 4 & 5 & \ldots & n \\
\end{bmatrix}
\]

where \( \sigma \) is the uncertainty in the estimate of any horizontal throw. We assume that all throws have the same value of \( \sigma \) since we attribute them to observational errors. To fit the data to the model that we have described in the previous sections we use a three dimensional grid representing the three parameters and use the values at every node in this grid as a possible model to fit the data. At each of these points we calculate the resultant error in the fit between the model (represented by the grid node) and the data. The least square error or misfit vector, \( E \), between the data and a set of displacements is calculated for equation 7 which has an additional adjustable parameter \( y_D \), the offset in latitude. The sum-squared error, SSE, for a particular value of total convergence \( C_0 \), 1/e fall-off distance, \( y_0 \), and the offset in latitude, \( y_D \) (a point or node in our grid) is given by the expression \( E \text{cov}^{-1}E^T \), which is the total misfit to our model (with respect to the data) taking data correlation into account. We estimate the value of \( \sigma \) by requiring that SSE equals 131 (i.e., the number of data, 134, minus the number of adjustable parameters, 3) when \( C_0, y_0 \) and \( y_D \) have the values that minimize SSE.
4.1.1.2 Uncertainty in Dip

We consider the value of the mean dip to vary between 36° and 45°. Changes in the value of dip contribute to changing the value of the horizontal displacement calculated at each fault, which thus causes a change in the value of cumulative displacement. The gradient of convergence with respect to the dip angle is always negative, in other words an increase in the dip angle would always cause the cumulative displacement to drop while a decrease in dip angle would always cause it to increase. Thus, by using the extreme values of the dip angle we are essentially plotting the extreme possible uncertainty ellipses for the two parameters. Thus, a change in dip would change the total convergence but will not affect the 1/e fall-off distance.

4.1.2 Curve Fitting:

After arriving at our best fit value (our best fit model has a standard deviation of 180.6 m), which is at y_D = 0.05°S, C = 27 km and y_o = 7° (these values provide the minimum SSE of 131 as mentioned in the previous section) we calculated a three dimensional uncertainty ellipsoid for the 95% confidence region (the uncertainty in dip will be considered later) of the three constraining parameters by allowing the SSE to vary between 131 and 138.8 (which is the limit for calculating the 3D 95% confidence region). Thus, all values of y_D, C and y_o which give values of SSE in this range all considered to be part of the 3D uncertainty ellipsoid. Figure 6 shows the 95% confidence limits of the two parameters y_o and C_o found by projecting the 3D uncertainty region onto the y_D = 0.05°S plane. Plotting the best statistical fit data to
the observed cumulative convergence indicates a misfit (Fig. 7), particularly for higher values of \( y \), which is potentially due to the correlated nature of the data points. In other words, the uncertainty in the initial data points is relatively low, but since each of these points get cumulatively added to all the previous points the uncertainty keeps adding on as well. Thus, the initial points are less uncertain than the latter points because of which the model would give a higher preference to fit the initial points, thus causing higher misfits to the latter points. Points 1 through 4 (Figure 6) are extreme points on the 95% confidence region; the curves calculated from these four pairs of parameter values are illustrated in Figures 8 a-d. Because of the highly correlated uncertainties in points along the cumulative convergence curve, the best least-squares fit may be very different from the best visual fit. So far we have discussed the fits ignoring the uncertainties in dips. Fig 9 shows the increase in the 95% confidence limit if the uncertainty in dip is taken into account. This change in total convergence though occurs at the same values of the 1/e fall-off width, which indicates that the 1/e fall-off width is independent of the dips.

5 Results and Discussions

5.1 Strain rate

Fig.11a shows calculated strain rates for our best-fit model ignoring the constraints from plate reconstructions. These rates range from \( 1.4 \times 10^{16}\text{s}^{-1} \) (equivalent to a strain of 3.18% over 8 Myr) at the northernmost limit of deformation where the Indian indenter acts on the diffuse plate boundary (thin viscous sheet) to \( 4.2 \times 10^{17}\text{s}^{-1} \)
(equivalent to a strain of 1.06% over 8 Myr) at 8°S (the southernmost limit of faulting observed in seismic profiles). Fig. 11b shows calculated strain rates for our best-fit model with the constraints from plate reconstructions. A convergence estimate of 41 km and a 1/e fall-off width of 13° leads to strain rates ranging from $1.12 \times 10^{-16}$ s$^{-1}$ (equivalent to 2.83% over 8 Myr) at the northernmost limit of deformation to $6.1 \times 10^{-17}$ s$^{-1}$ (equivalent to 1.54% over 8 Myr) at 8°S. These strain rates are on average about an order of magnitude higher than the upper bound on strain rates in stable plate interiors (Gordon [2000]). The strain rates determined from the plate tectonic constraints have a narrower range since the total convergence falls off to its 1/e value over a comparatively larger distance.

5.2 Convergence estimates

The 95% confidence limits that we find for total convergence are 15 km to 82 km, much larger than the 22–37 km limits estimated by the original investigators [Chamot-Rooke et al. 1993] or the 22–32 km limits estimated by Van Orman et al. [1995]. Our limits are larger mainly for two reasons. First, from the dispersion of the data about the best-fitting model, we estimate substantial uncertainty ($\sigma = 180.6$ m) in individual estimates of horizontal throw. This causes our lower limits to be smaller, and our upper limits to be greater, than prior limits. Second, the acceptable values for the 1/e fall-off distance imply that substantial deformation occurs south of the southern limit (≈8°S) of deformation imaged on the Phèdre profile. Chamot-Rooke et al. [1993] note that the sediment thickness is almost negligible south of 8°S and that it becomes difficult to locate the faults. Our unconstrained best-fit curve indicates that
about 37% (1/e) of the convergence occurs south of 8°S but at strains substantially lower than the average north of 8°S. Our constrained best-fitting curve indicates that even more of the convergence is accommodated south of 8°S. Unlike the prior analyses, our estimates of total convergence (15–82 km) are consistent with those from plate reconstructions since 8 Ma (41.3 ± 2.3 km) [DeMets et al 2005].

As mentioned in the previous section, using our best-fitting model (convergence = 27 km, 1/e fall-off width = 7°) we have calculated the strain rate profile from north to south (Fig. 11) which gives us a value of $4.2 \times 10^{-17} s^{-1}$ of strain rate at 8°S. Assuming that all deformation occurred in the last 8 million years, the equivalent strain would be 1.06%. If the faults along the Phèdre line are evenly spaced every 7 km, the horizontal component of displacement would be expected to be $\approx 77$ m at 8°S and the corresponding vertical throw would be 56 m to 77 m for the range of dips (36° to 45°) that we have considered. Throws greater than $\approx 50$ m were resolved during the Phèdre cruise and an absence of observable faulting to the immediate south of 8°S was noted by Chamot-Rooke et al [1993]. A possible reason for this absence is a lack of sediments to the south of 8°S [Chamot-Rooke et al 1993]. According to our calculations based on our best-fit model the minimum average vertical throws at 8°S are $\approx 56$ m and any shortening further south should on an average be less than this. Thus, if most of the throws to the south are less than what was seismically resolvable, this would be another possible cause for the absence of any observable faulting (using seismic reflection profiles) to the south of 8°S.
5.3 1/e falloff width

The 95% confidence region calculated for the distance at which strain rate falls to 1/e of its original value, \( y_o \), is 3.9° to 25.8° (taking the uncertainty in dips into account). To constrain our model further we have used the plate tectonic constraints on convergence (41.3 ± 2.3 km). This results in a narrower band of values (95% confidence region) for the 1/e falloff width (Fig 9) between 7.1° and 20.4° (788 km to 2269 km).

5.4 Power law Exponent

The power law exponent of the fluid (which here is an average over the entire thickness of the oceanic lithosphere) can be estimated from the map view aspect ratio of the region of deformation as seen by equation 5.

\[
y_o = \frac{\lambda}{\pi \sqrt{n}} \quad \Rightarrow \quad n = \frac{\lambda^2}{\pi^2 y_o^2}
\]

The value of \( y_o \) is between 7.1° and 20.4° as seen in the previous section. Also as shown in Fig.2, \( \lambda \) is defined as twice the entire length of the indenter which includes both the compressional and the tensional parts. Thus, \( \lambda \) is four times the length of the compressional part of the indenter which we shall call \( D \) which in this case is the length of the Indian plate acting on the diffuse oceanic plate boundary and causing compressional faulting. We take the along-strike length, \( D(=\lambda/4) \), of this indenter to be 24°, from the pole of India-Capricorn rotation at 74°E to the Sumatra trench at ≈98°E. This length includes N-S shortening to the west of ≈ 86°E and NW-SE shortening to its east. Thus, taking only the N-S component would mean a 12°
value for D, while taking both the N-S and the NW-SE component would mean a D of 24°. Thus the value of D is likely to be at least 12° but cannot be more than 24°. The value of y₀ has already been calculated to be between 7.1° and 20.4°. Thus the power-law exponent lies between 1.6 and 18.5 in the case of D being 24° while it lies between 0.55 and 4.63 if D is considered to be 12°. The lithosphere has a vertically averaged rheology of a upper brittle layer with a power-law exponent of ∞ and of a lower layer with a power law exponent of 3. Thus, the minimum combined power-law exponent of the fluid should be 3. Since the lower limit is well constrained, we are more interested in constraining the upper limit of this value and thus choose to consider the higher case of D being 24° which gives limits of 1.6 and 18.5 for the power-law exponent.

5.5 Strength of the upper versus the lower lithosphere

A power-law exponent of 3 would suggest a relatively weak upper lithosphere while a higher value indicates a stronger upper lithosphere. Gordon [2000] defines R to be the ratio of the strength of the upper lithosphere relative to the strength of the lower lithosphere (at a reference strain rate of 10⁻¹⁶ s⁻¹). R for our calculated range of the power-law exponent lies between 0 and 5.5 (Figure 13). A low value of n and a corresponding low value of R imply a strong lower lithosphere as compared with the upper lithosphere, in particular a value of R less than 1 implies that a greater stress would be required to deform the lower lithosphere at a constant strain rate (in this case 10⁻¹⁶ s⁻¹) than would be required to deform the upper lithosphere. We have already placed a lower bound on the power law exponent of 3 in the previous section;
this is equivalent to placing a lower bound of 0 on the value of \( R \). Thus, only the upper bound of 5.5 provides useful information.

6 Conclusion

Prior work indicates that the convergence observed on the Phèdre profile (Chamot-Rooke et al. [1993]) of 22 to 37 km is lower than that expected from plate tectonic reconstructions (41.3 ± 2.3 km). Here we find that this discrepancy is caused either by the absence of seismically resolvable (on the Phèdre line) faults to the south of 8°S or due to the lack of sediments beyond 8°S making reflection seismic imaging impossible. The calculated strain rates based on the best-fitting model are an order of magnitude higher than the upper bounds on those of stable plate interiors, which is consistent with our understanding of diffuse oceanic plate boundaries. When constrained to a total convergence of 41.3 km ± 2.3 km (as found from plate tectonic reconstructions) the 95 % confidence limits on the width at which strain rate falls off to 1/e of its maximum value in the India-Capricorn diffuse plate boundary is between 4.2° to 15.8°, which constrains the power-law exponent of the vertically averaged rheology of the oceanic lithosphere to be between 1.6 and 18.5 (95 % confidence limits). The strength of the upper lithosphere is calculated to be up to 5.5 times the strength of the lower lithosphere at a reference strain rate of \( 10^{-16} \text{s}^{-1} \), which implies that the lower lithospheric strength should be added while calculating the total strength of the lithosphere.
Fig 1 The Indo-Australian composite plate with the diffuse plate boundaries in gray. The region of deformation in the Central Indian Ocean is shown in light gray. I-C = Indian Capricorn boundary, C-A= Capricorn Australia boundary modified from S. Zatman and R G. Gordon (unpublished manuscript)
Fig 2  Cartoon showing the velocity field caused by an Indenter acting on a thin viscous sheet of material. The applied velocity field follows a sine curve in shape.
Fig 3 Reflection seismic data collected from the Phedre cruise along 81.5°E included vertical throws of seismically observable faults. This data is represented by the red dots. Vertical throws of 134 faults were identified during the survey (modified from Chamot-Rooke et al (1993)). The x axis goes from north to south. Negative throws imply south dipping faults while positive ones indicate north dipping faults. This data has been obtained by digitizing Fig. 3 of Chamot-Rooke et al (1993) and replotted here.
Fig 4 Reflection seismic data collected from the Phedre cruise along 81.5°E included vertical throws of 134 seismically observable faults. By assuming a planar geometry and a constant dip angle of 40°, we have calculated the horizontal slip at each fault based on the vertical throw data. This horizontal slip is shown as red dots.
Fig 5 Vertical throw data for 134 faults was collected along 81.5°E during the Phedre cruise going north-south. This vertical throw data is converted into horizontal slip by assuming a constant dip (previous figure). By adding the horizontal slip of every point with its previous point as we go from north to south (right to left) we calculate the total shortening or convergence until 8°S latitude. In the above figure the total convergence is shown by the red dots.
Fig 6  Projection of the 95% uncertainty ellipsoid onto a 2D plane. The 3 parameters: a.) $y_0$, the 1/e fall off distance of strain b.) $C_0$, the total convergence, c.) $y_D$, the offset in the x-axis. The 3D projection is onto the $y_D=0.05^\circ$ plane which is the best fit value for $y_D$ which has a range from -1 to 1. The fitting equation is $C = -C_0e^{-y/y_0} + C_0$ where $C$, the convergence at a point is the summation of horizontal displacements of all the points before it. Since the errors covary, we use a covariance matrix to estimate the least squares best fit for the model. The blue star is the best fit and represents a 27 km convergence at a 1/e fall off width of 7°.
Fig 7 a Based on the best fit values of convergence (27 km) and 1/e falloff width (7°) represented by the blue star in the previous figure, a graphical view of this fit is shown. The red dots represent convergence across the diffuse plate boundary and the blue line represents the model with the least statistical error to fit the red dots. The model does not represent the best visual fit as the errors in each data point are interdependent and as we go south (right hand side in the above figure), the errors get magnified as they are the sum of errors of all the points to its left or in other words the weighing of the points is much higher to the left and thus our curve fitting model tries to fit them as best as possible and as a consequence does not visually fit the latter points very well.
Fig 7 b The red points represent horizontal slip calculated from the 134 fault observations along the Phedre line at 81.5°E. The green points are calculated from the best fit model (represented by the blue star in Fig.6) and are calculated by subtracting the value of convergence of every point from its previous point and thus getting a horizontal slip for the best fit model. The lines joining these points show the mismatch between observed values of horizontal slip and the values derived from the best fit model. A black line indicates a higher value of the observed horizontal slip as compared to the value calculated from the best fit model and a blue line is the opposite.
Fig 8 a.) Fit of data to the star 1 in Figure 6. A convergence value of 18 km and a 1/e falloff of 3.5° is used to fit the model to the data. The data are in red and the model based on these values is in blue.
Fig 8 b) Fit of data to the star 2 in Figure 6. A convergence value of 48 km and a 1/e falloff of 13° is used to fit the model to the data. The data are in red and the model based on these values is in blue.
Fig 8 c) Fit of data to the star 3 in Figure 6. A convergence value of 69 km and a 1/e falloff of 25° is used to fit the model to the data. The data are in red and the model based on these values is in blue.
Fig 8 d) Fit of data to the star 4 in Figure 6. A convergence value of 39 km and a 1/e falloff of 16° is used to fit the model to the data. The data are in red and the model based on these values is in blue.
Fig 9 Projection of 3D uncertainty ellipsoid onto a $y_D = 0.05^\circ$. To calculate the uncertainty in three parameters (convergence, 1/e falloff width and offset) used to fit the observed data along the phedre line we had used a fault planar dip of 40°. But, in fact the mean of the dips varies from 36° to 45°. Thus the ellipses show the 2D projection of the 95% confidence regions. While fitting the data when we consider the dip angle to be 36° (blue ellipse), 40° (black ellipse), 45° (red ellipse). Independent estimates of total convergence from plate tectonic data (DeMets et al [2005]) are used to limit the convergence (the gray rectangle).
Fig 10  a. Shows the lower end fit (width = 7.1°, convergence = 39 km) between the data and the model in fig 9. The red dots represent the data and the blue line represents the model calculated using the two parameters (width and convergence).
Fig 10  b. Shows the higher end fit (width = 20.44°, convergence = 43.6km) between the data and the model in fig 9. The red dots represent the data and the blue line represents the model calculated using the two parameters (width and convergence).
Fig 11a  Plot of north-south contractional strain rate versus distance along the phedre profile. Using the best fit model (convergence = 27 km, 1/e falloff width = 7° and offset = 0.05°S) we can plot a convergence vs distance graph. By differentiating this curve with respect to distance we get a graph of strain vs distance and by assuming the deformation to have occurred in the last 8 Myr we have calculated the strain rate vs distance graph as shown.
Fig 11b Plot of north-south contractional strain rate versus distance along the phedre profile. Constraining the best fit model to plate tectonic constraints of convergence as seen in Fig. 9 (convergence = 41 km, 1/e falloff width = 13° and offset = 0.05°S) we can plot a convergence vs distance graph. By differentiating this curve with respect to distance we get a graph of strain vs distance and by assuming the deformation to have occurred in the last 8 Myr we have calculated the strain rate vs distance graph as shown.
Fig 12. Strength envelope for oceanic lithosphere extrapolated from lab experiments on dry rocks (Kohlstedt et al. 1995).
Fig 13 Strain rate versus force per unit length curve for the theoretical model. The strain rates of interest (for deforming oceanic lithosphere) are shown. The corresponding values of the power law exponent and 'R' are shown for each curve. 'R' is the ratio of the yield strength of the plastic layer (upper lithosphere) to the force required to deform the creeping-flow layer (lower lithosphere) at a reference strain rate of $10^{-16}\text{s}^{-1}$. The fluid layer is assumed to obey a power-law rheology with a power-law exponent of 3. After Gordon (2000).
Chapter 2

Strength profile of oceanic lithosphere
1 Summary

We calculate strength envelope curves for the oceanic lithosphere by using a two layered model [Gordon 2000]: 1.) a brittle upper layer, 2.) a lower layer deforming by dislocation creep based on the Dorn equation (values from Karato et al [2003] are used for the different parameters). The strength envelopes are found to be very sensitive to the value of the activation energy used. These strength envelopes provide a range of 0.2 and 20 TN m$^{-1}$ for the strength of the oceanic lithosphere. The value of R (ratio of the strength of the upper lithosphere with respect to the lower lithosphere) ranges between 8.6 and 13.98 which are much higher than the range of 0 to 5.5 which we have estimated in the previous chapter. The 1/e fall-off distance is estimated to be between 4.75 and 5.8 which are lower than our estimates of 7.1° to 20.4° in the previous Chapter. These differences could be explained by changes in the values of the parameters used, in particular the activation energy.

2 Introduction

Olivine is the dominant constituent of the lithospheric part of the mantle. Many experiments have been conducted in the past to understand the creeping flow of olivine (Karato et al 2003). These experiments have been conducted at much higher strain rates (by about 10 orders of magnitude) than geological strain rates, because geological strain rates are difficult to replicate in a laboratory. These studies have been used to calculate the strength profile of the lower lithosphere (Kohlstedt et al 1995) providing us with information about lithospheric strengths. These calculations depend on theoretically
determined creep flow equations and laboratory determined parameters like activation energies.

Here we calculate the strength profile for the oceanic lithosphere based on a two layered model [Gordon 2000] by using lab determined parameters (based on olivine) and the Dorn equation (empirically determined equation for dislocation creep). We consider the following: (1) What are the limits to the strength of oceanic lithosphere? (2) What are the relative strengths of the upper versus the lower lithosphere? (3) How sensitive are these models to changes in different parameters like temperature and activation energy? (4) How do the values of R (ratio of the strength of the upper lithosphere to the lower lithosphere) compare with the range of values calculated in chapter 1? (5) What are the possible causes of any divergence between the two results?

3 Background

3.1 Strength Envelopes

In this section we will quantify the strength of the oceanic lithosphere based on which we will build our strength envelopes. We consider the oceanic lithosphere to consist of two layers [Gordon 2000]:

1. The upper part where temperature and pressure are comparatively low, the mode of deformation is based on faulting, which implies brittle or semi-brittle failure. Thus, we assume a constant strength of the upper lithosphere all the way down to 50 km (which is approximately the depth beyond which earthquakes in the oceanic lithosphere (excluding subduction zones) are not observed, in other words
this would define a boundary between the region of brittle deformation and the region of deformation by creeping flow), with the assumption that stresses equivalent to the strength of the lithosphere would cause failure of the entire upper layer.

2. The lower lithosphere where the temperature and pressure are high enough for the predominant mode of deformation to be creeping flow. There are two types of creep flow: diffusion and dislocation creep. Diffusion creep is more prominent where the temperature and stresses are relatively lower while dislocation creep is more prominent when they are higher. Figure 12b shows the dependence of creep flow as a function of temperature and pressure (Turcotte and Schubert 1982), which also shows that for the strain rates that we are considering, the dominant creep flow would be dislocation creep, which is what we will consider henceforth.

3.2 Creep Flow

Many laboratory experiments have been conducted to understand the high temperature and pressure deformation of olivine, which being the predominant constituent in the mantle, determines creeping flow in the mantle. A single creep mechanism is believed to fail to dominate across the ranges of temperature and pressure in the lithosphere, but, for the purposes of this study we will follow the Dorn equation which is an empirical flow law that is commonly used to describe dislocation creep. This law determines the relationship between differential stress and strain rate as a function of temperature and pressure:

$$ \dot{\varepsilon} = A_D \sigma^n e^{\left( \frac{E + pV}{RT} \right)} $$
where \( \dot{\varepsilon} \) = strain rate

\( A_D \) = pre-exponential factor

\( \sigma \) = differential stress

\( n \) = power law exponent

\( E \) = activation energy

\( p \) = pressure

\( V \) = activation volume

\( R \) = gas constant

\( T \) = temperature

4 Model

In this section we will build strength profiles (with depth) of oceanic lithosphere based on our model and use data from laboratory experiments for our parameter values. To solve the Dorn equation for stress we have used the following values and models for the following parameters and parameter fields:

Temperature:

We have considered two separate thermal models to calculate the temperature profile used in the Dorn equation:

a. Temperature profile for a 70 Myr old oceanic lithosphere calculated by using Fig. 7 of McKenzie et al [2005] which is based on analytical solutions by Parsons et al [1977]. A key factor considered here is the temperature dependence of
thermal conductivity. This profile predicts a temperature of 753.17°C at a depth of 50km (Fig 14).

b. ) Temperature profile for a 70 Myr old oceanic lithosphere using a half-space cooling model [Turcotte and Schubert 1982]. We have constrained the temperature at a depth of 50 km to be 753.17°C (the same as the previous model), which gives us a temperature of 1397.5°C at the base (mantle).

\[ A_D, E, V: \]

Experimentally determined values (Karato et al [2003]) are used for activation volume, the pre-exponential factor \( A_D \) and activation energy. The range of these values from Karato et al [2003] is shown in the following table.

<table>
<thead>
<tr>
<th></th>
<th>( A_D ) (s(^{-1})(MPa)(^{n}))</th>
<th>E (kJ/mol)</th>
<th>V (cm(^3)/mol)</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Limit</td>
<td>( 10^{6.3} = 2 \times 10^6 )</td>
<td>540</td>
<td>16</td>
<td>3.1</td>
</tr>
<tr>
<td>Lower limit</td>
<td>( 10^{5.9} = 7.9 \times 10^5 )</td>
<td>480</td>
<td>12</td>
<td>2.9</td>
</tr>
</tbody>
</table>

These upper limit set of values are similar to those used by Kohlstedt et al. [1995] (\( E = 535 \) kJ/mol, \( V = 17 \) cm\(^3\)/mol) to construct their strength envelopes. We assume the power-law exponent for the upper mantle to be 3. A constant strength model of the upper lithosphere is used to a depth of 50 km. The stress variation of depth for the lower lithosphere (which deforms by dislocation creep) is only constrained by the empirical flow law and thus provides a value of differential stress at a depth of 50 km that we have used as a constraint for the strength of the upper lithosphere (which is the reason behind equating the temperatures obtained from both the thermal models at 50 km). The strain
rates of interest in deforming oceanic lithosphere are between $10^{17}$ s\(^{-1}\) and $10^{15}$ s\(^{-1}\). Thus we have calculated a strength envelope for olivine under dry conditions for a strain rate of $10^{16}$ s\(^{-1}\) using mean lab values and for both thermal models (Figs. 15 and 16).

5 Results

The strength profile calculated by using the half space cooling model gives a lower value of the strength of the upper lithosphere of 1.6 TN m\(^{-1}\) and R of 9.4 as compared with the McKenzie et al [2005] model which gives a value of 2 TN m\(^{-1}\) and 13 respectively. By varying the base temperature used in the half-space cooling model by $\pm 10^\circ$C (Fig 17 and 18) we find a slight decrease in both the strength of the upper lithosphere (13% decrease) and R (1% decrease) by increasing the temperature and a similar percentage increase in these values by decreasing the temperature by $10^\circ$C.

This strength envelope is very sensitive to values of activation energies used (we have used 480-540 kJ/mol) as can be seen from Figs. 19 and 20 (for the half space cooling model where the strength of the upper lithosphere as well as R drop from 5.9 TN m\(^{-1}\) and 10.15 to 0.5 TN m\(^{-1}\) and 8.66 respectively) and Figs. 25 and 26 (for the McKenzie et al [2005] model where the strength of the upper lithosphere as well as R drop from 6.6 TN m\(^{-1}\) and 13.95 to 0.6TN m\(^{-1}\) and 12.12 respectively) which show more than an order of magnitude difference in the values of the differential stress which in turn imply more than an order of magnitude difference in terms of calculating the strength of the lithosphere. This envelope is also susceptible to changes in the pre-exponential parameter $A_D$, but these changes are limited to about $\pm 30\%$ in terms of the strength of the lithosphere for both the temperature models but this parameter does not affect the value
of R as seen in Figures 21 and 22 and Figures 27 and 28. The activation volume similarly affects the strength of the lithosphere by about ±16% for both the models while its effect on R is minimal (±1%). From these strength envelopes, we have calculated the limiting values of R, the ratio of the strength of the upper lithosphere relative to the lower lithosphere deforming at a constant strain rate of $10^{-16}$ s$^{-1}$ and found them to be between 8.6 and 13.98. In the range of strain rates that correspond to deforming oceanic lithosphere this would corresponds to a vertically averaged power-law exponent ranging from 27 to 41.5. If we consider the aspect ratios of the Indian Capricorn diffuse oceanic plate boundary, these values of the power law exponent would imply a 1/e fall off of strain rate at a distance of 4.75°-5.8° from the northern extremity of the diffuse oceanic plate boundary which are significantly lower than those that we have estimated in Chapter 1 (7.1° – 20.4°).

By varying all the parameters (including strain rates) within their uncertainty limits we have calculated the maximum and minimum possible strengths for the lithosphere based on these strength envelopes and found it to be between 0.2 TN m$^{-1}$ for a strain rate of $10^{-17}$ s$^{-1}$ TN m$^{-1}$ and 20 TN m$^{-1}$ for a strain rate of $10^{-15}$ s$^{-1}$. Figs. 31, 32 and 33 show that this large range is essentially due to a change in activation energy which seems to play a large part in determining the strength of the lithosphere. Gordon [2000] gives a lower bound of 8 TN m$^{-1}$ for the strength of the oceanic lithosphere in the central Indian ocean. We will thus constrain our results to values greater than 5TN m$^{-1}$ (to account for any potential uncertainty). This essentially leaves us with values derived at a strain rate of $10^{-16}$ s$^{-1}$ and $10^{-15}$ s$^{-1}$ with activation energies of 540 kJ/mol.
Figs. 31, 32 and 33 show that a change in strain rate does not affect R but significantly changes the strength of the lithosphere as expected, since a higher strain rate (at a given stress) implies a lower strength. The impact of a change in activation energy is also noted irrespective of the thermal model in use. A higher activation energy significantly increases the values of both the strength of the lithosphere and the value of R. Thus the activation energy is a key parameter for these kinds of equation-based model building.

6 Discussion

The values that we had obtained in Chapter 1 are significantly different from the values that we calculated from laboratory determined results. A key difference is in the relative strengths of the upper versus lower lithosphere which in Chapter 1 we had estimated to be less than 5.5 while here we have a range between 8.6 and 14 which is significantly higher implying a stronger upper lithosphere. One possible cause of these differences is the fact that lab experiments are conducted at high strain rates relative to the real earth strain rates of deforming oceanic lithosphere and maybe the extrapolation to higher strain rates might require different values of the existing parameters (or maybe even more parameters), in particular probably a lower value of the activation energy. If we reduce the activation energy to 353 kJ/mol (which is significantly lower than the mean experimental value of 510 kJ/mol) while keeping all the other parameters constant, we find that the value of R comes out to 5.5, which is about what our upper limit is. Thus a lower value of activation energy seems to be a possible solution to the discrepancy between our results as compared to the lab results (neglecting any covariances between
the different parameters). But, a lower value of activation energy would imply an implausibly weak lithosphere (~ 0.0045 TN m⁻¹) if all other values are kept the same. These values are implausibly low (Gordon [2000] gives a lower bound of 8 TN m⁻¹). It follows that a lower activation energy alone cannot resolve the discrepancy. But, if we significantly change the other parameters in particular the viscosity dependent $A_D$ (decrease by two orders of magnitude) then the lithospheric strength is about 1 TN m⁻¹ (which is still implausibly weak). Thus, it is difficult to predict as to which of these parameters might be causing the differences as they may be correlated. There have been other field based evidences (based on the flexure of lithosphere due to seamount loading) for potentially lower activation energies than those suggested by lab based experiments (Watts et al 2000), these studies assume deformation by diffusion creep and calculate an activation energy estimate of 120 kJ/mol which is much lower than the 240-300 kJ/mol predicted by lab based studies. These different estimates seem to suggest that the lab based constraints on activation energies for the deformation of oceanic lithosphere may be overestimated for both kinds of deformation mechanisms and that the parameters like activation energy, temperature, strain rates etc co-vary, thus leading to significantly different results.
Table 1 (HC = temperature profile derived from half space cooling model with a base temperature of 1397.5°C ; McK = temperature profile from McKenzie et al 2005)

<table>
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<tr>
<th>Strain rate s⁻¹</th>
<th>A₀ (s⁻¹(MPa)⁻¹)</th>
<th>E (kJ/mol)</th>
<th>V (cm³/mol)</th>
<th>Strength of Upper Lithosphere (TN/m)</th>
<th>R</th>
<th>T</th>
<th>Strength of lower Lithosphere (TN/m)</th>
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<td>A(_D) (s(^{-1})MPa(^{-n}))</td>
<td>E (kJ/mol)</td>
<td>V (cm(^3)/mol)</td>
<td>Strength of Upper Lithosphere (TN/m)</td>
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Fig 12 b Deformation map for a dry upper mantle. The deviatoric stress $\sigma$ is given as a function of temperature $T$ for several strain rates. The dashed line separates the dislocation creep regime from the diffusion creep regime. (From Turcotte & Schubert). For the strain rates observed in deforming Oceanic lithosphere the deformation mechanism would be dislocation creep.
Fig 13 Strain rate vs force per unit length curve for the theoretical model. The strain rates of interest (for deforming oceanic lithosphere) are shown. The corresponding values of the power law exponent and ‘R’ are shown for each curve. ‘R’ is the ratio of the yield strength of the plastic layer (upper lithosphere) to the force required to deform the creeping-flow layer (lower lithosphere) at a reference strain rate of $10^{-16}\text{s}^{-1}$. The fluid layer is assumed to obey a power-law rheology with a power-law exponent of 3. After Gordon (2000).
Fig 14 Temperature profile with depth using two different models:
a.) in blue is a model from McKenzie et al 2005 which is based on a
temperature dependant thermal conductivity, b.) in red is a
temperature profile based on the half space cooling model with the
temperature at a depth of 50 km constrained to 753.17°C (1026.2 K)
in order to match the blue curve at that same depth. The
temperature at the base for the half space cooling model thus comes
out to 1397.5°C (1670.5K)
Fig 15 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}$ s$^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^{6.1}$ s$^{-1}$ (MPa)$^n$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere = 1.5979e+012 and the corresponding $R = 9.4051$. Input values mentioned are from (Karato et al. [2003]). The temperature profile used is based on the half space cooling model with the temperature at the base being 1397.5$^\circ$C.
Fig 16 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}\,\text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or R. The temperature profile used is from McKenzie et al [2005] based on a temperature sensitive thermal conductivity. An activation energy of 510 kJ/mol is used for our calculations. Other values used are : pre-exponential factor $A_D=10^{6.1}\,\text{s}^{-1}(\text{MPa})^n$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere= 2.0698e+012 and the corresponding R = 13.0387. Input values mentioned are from (Karato et al [2001]).
Fig 17 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}$ s$^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or R. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^{6.1}$ s$^{-1}$ (MPa)$^n$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere = 1.4354e+012 and the corresponding R = 9.3621. Input values mentioned are from (Karato et al. [2001]). The temperature profile used is based on the half space cooling model with the temperature at the base being 1407.5°C.
Fig 18 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16} \text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^{6.1} \text{s}^{-1} (\text{MPa})^n$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere= $1.7807 \times 10^{12}$ and the corresponding $R = 9.4482$. Input values mentioned are from (Karato et al [2001]). The temperature profile used is based on the half space cooling model with the temperature at the base being 1387.5°C.
Fig 19 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}\text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. An activation energy of 540 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_{D}=10^{6.1}\text{s}^{-1}(\text{MPa})^{n}$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere= 5.941e+012 and the corresponding $R = 10.1455$. Input values mentioned are from (Karato et al [2001]). The temperature profile used is based on the half space cooling model with the temperature at the base being 1397.5$^\circ$C.
Fig 20 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{16} \text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. An activation energy of 480 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^6 \text{s}^{-1}(\text{MPa})^n$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere = 5.0121e+011 and the corresponding $R = 8.6604$. Input values mentioned are from (Karato et al [2001]). The temperature profile used is based on the half space cooling model with the temperature at the base being 1397.5°C.
Fig 21 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}$ s$^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or R. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^5$ s$^{-1}$ (MPa)$^{-n}$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere = 1.863e+012 and the corresponding R = 9.4051. Input values mentioned are from (Karato et al [2001]). The temperature profile used is based on the half space cooling model with the temperature at the base being 1397.5°C.
Fig 22 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16} \text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^6.3 \text{s}^{-1} (\text{MPa})^n$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere = 1.3705e+012 and the corresponding $R = 9.4051$. Input values mentioned are from (Karato et al [2001]). The temperature profile used is based on the half space cooling model with the temperature at the base being 1397.5°C.
Fig 23 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}\text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_0=10^6\text{s}^{-1}(\text{MPa})^{-n}$, activation volume $= 16 \text{cm}^3/\text{mol}$. The calculated strength of the upper lithosphere $= 1.7702e+012$ and the corresponding $R = 9.3463$. Input values mentioned are from (Karato et al [2001]). The temperature profile used is based on the half space cooling model with the temperature at the base being $1397.5^\circ\text{C}$.
Fig 24 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16} \text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^{6.1} \text{s}^{-1} (\text{MPa})^n$, activation volume = 12 cm$^3$/mol. The calculated strength of the upper lithosphere = $1.4423e+012$ and the corresponding $R = 9.4633$. Input values mentioned are from (Karato et al [2001]). The temperature profile used is based on the half space cooling model with the temperature at the base being 1397.5°C.
Fig 25 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}\text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. The temperature profile used is from McKenzie et al [2005] based on a temperature sensitive thermal conductivity. An activation energy of 540 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^{6.1}\text{s}^{-1}(\text{MPa})^{-n}$, activation volume $= 14 \text{ cm}^3/\text{mol}$. The calculated strength of the upper lithosphere $= 6.6964e+012$ and the corresponding $R = 13.9536$. Input values mentioned are from (Karato et al [2001]).
Fig 26 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{16}\text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. The temperature profile used is from McKenzie et al [2005] based on a temperature sensitive thermal conductivity. An activation energy of 480 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D = 10^{6.1}\text{s}^{-1}\text{(MPa)}^n$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere = 0.6398e+012 and the corresponding $R = 12.1217$. Input values mentioned are from (Karato et al [2001])+012.
Fig 27 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}$s$^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. The temperature profile used is from McKenzie et al [2005] based on a temperature sensitive thermal conductivity. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^{5.9}$s$^{-1}$(MPa)$^{-n}$, activation volume $= 14$ cm$^3$/mol. The calculated strength of the upper lithosphere $= 2.4133e+012$ and the corresponding $R = 13.0387$. Input values mentioned are from (Karato et al [2001]).
Fig 28 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{16}\text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. The temperature profile used is from McKenzie et al. [2005] based on a temperature sensitive thermal conductivity. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_0=10^{6.3}\text{s}^{-1}(\text{MPa})^n$, activation volume = 14 cm$^3$/mol. The calculated strength of the upper lithosphere = 1.7753e+012 and the corresponding $R = 13.0387$. Input values mentioned are from (Karato et al [2001]).
Fig 29 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{-16}\text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or $R$. The temperature profile used is from McKenzie et al [2005] based on a temperature sensitive thermal conductivity. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^{6.1}\text{s}^{-1}(\text{MPa})^n$, activation volume = 16 cm$^3$/mol. The calculated strength of the upper lithosphere is $2.2961e+012$ and the corresponding $R = 13.01$. Input values mentioned are from (Karato et al [2001]).
Fig 30 An approximate strength envelope for oceanic lithosphere. The upper lithosphere has a brittle rheology while the lower lithosphere deforms by creeping flow at a constant strain rate of $10^{16}\,\text{s}^{-1}$. The ratio of the areas contained within the two parts of the curve determine the relative strengths of the upper and lower oceanic lithosphere or R. The temperature profile used is from McKenzie et al [2005] based on a temperature sensitive thermal conductivity. An activation energy of 510 kJ/mol is used for our calculations. Other values used are: pre-exponential factor $A_D=10^{6.1}\,\text{s}^{-1}(\text{MPa})^{-n}$, activation volume $= 12\,\text{cm}^3/\text{mol}$. The calculated strength of the upper lithosphere $= 1.8659e+012$ and the corresponding $R = 13.0673$. Input values mentioned are from (Karato et al [2001]).
Fig 31 Strength of the lithosphere as a function of $R$ at a reference strain rate of $10^{-15}\,\text{s}^{-1}$. These values are calculated based on the Dorn equation using lab based parameters. The blue line represents calculations based on the half space cooling model as the base thermal model, while the red considers the Mckenzie et al [2005] thermal model. The stars represent the values determined by an input activation energy of 540 kJ/mol while the circles represent an activation energy of 480 kJ/mol. The red stars and circles are based on the McKenzie et al [2005] thermal model while the blue stars and circles are based on the half space cooling model.
Fig 32 Strength of the lithosphere as a function of $R$ at a reference strain rate of $10^{-16}$ s$^{-1}$. These values are calculated based on the Dorn equation using lab based parameters. The blue line represents calculations based on the half space cooling model as the base thermal model, while the red considers the Mckenzie et al [2005] thermal model. The stars represent the values determined by an input activation energy of 540 kJ/mol while the circles represent an activation energy of 480 kJ/mol. The red stars and circles are based on the McKenzie et al [2005] thermal model while the blue stars and circles are based on the half space cooling model.
Fig 33 Strength of the lithosphere as a function of $R$ at a reference strain rate of $10^{-17}$ s$^{-1}$. These values are calculated based on the Dorn equation using lab based parameters. The blue line represents calculations based on the half space cooling model as the base thermal model, while the red considers the Mckenzie et al [2005] thermal model. The stars represent the values determined by an input activation energy of 540 kJ/mol while the circles represent an activation energy of 480 kJ/mol. The red stars and circles are based on the McKenzie et al [2005] thermal model while the blue stars and circles are based on the half space cooling model.
REFERENCES


Gordon, R.G., C.DeMets, and D.F.Argus, Kinematic constraints on distributed lithospheric deformation in the equatorial Indian Ocean from present motion between the Australian and Indian plates, Tectonics, 9, 409-422, 1990.


APPENDIX 1

The Conrad Profile at 78.8°E:

The Conrad profile (Van Orman et al. [1995]) is another north-south profile that goes across the deforming region in the central Indian Ocean. This profile ranges from 0.8°N to 6.6°S. A statistical analysis similar to that carried out on the Phedre profile is attempted on the Conrad profile. One of the biggest problems in doing this for the Conrad profile is the high uncertainty in the fit. Also we believe that the Indian plate indents onto the diffuse plate boundary causing high strains to its immediate south. But, the Conrad profile shows very low strains at the northernmost extreme, we believe that this deformation may not be associated with our model and are unsure of as to where to start. Fig 1 shows a statistical fit taking all the observed strains into consideration which gives very high uncertainties. Removing a few points at the start as in fig 2 gives a much better fit (smaller uncertainties) but the uncertainty regions do not overlap the plate tectonic predictions for the convergence whose 95% confidence region is shown in red. Taking out a few more points as in figure 3 does not help. Thus, more constraints regarding the starting points are required.
Fig 1 a The Conrad profile distance vs convergence. b. The 95% confidence region for the fit.
Fig 2a The Conrad profile distance vs convergence, points in blue were omitted. b. The 95% confidence region for the fit.
Fig 3  a. The Conrad profile distance vs convergence, only the points in green were considered  b. The 95% confidence region for the fit.
APPENDIX  2

By using an empirical flow law (the dorn equation) for a strain rate of $10^{-16}$s$^{-1}$ which is the value that we have calculated based on our model, we have determined a stress vs. depth relationship for the lower lithosphere. By assuming an upper lithosphere of constant stress (Fig 14) we can calculate the ratio of the upper and lower lithospheric strengths or R. Now the empirical flow law depends on the values of the activation energy, activation volume, the pre-exponential coefficient and the strain rate. If we fix all the other values except activation energy (the equation is very sensitive to the activation energy) to lab derived values, then we can have a one on one relation between activation energy and R. More information would be needed to justify this assumption as we cannot just vary one lab determined value and fix all others. But, we do this for now due to lack of further constraints. Using the Phedre profile we have already limited the upper bound of R to be about 5.5 based on this value and the graph below we also fix an upper bound for activation energy to be about 355 kJ/mol. This value is much lower than that predicted by the lab results.
Fig 4 The variation of activation energy of olivine in the empirical dorn equation with the ratio of the strengths of the upper vs. lower oceanic lithosphere. These are calculated based on parameters from Karato et al 2001
APPENDIX 3

A part of a seismic line along 81.5°E from Chamot-Rooke et al [1993] is shown in Fig. 5. This line indicates regions of thrust faulting which have been measured for vertical throw by Chamot-Rooke et al [1993]. We measure the vertical offset (in time) in the sedimentary layers in five different places as shown in Fig. 5 where we notice thrust faulting. Assuming an average sediment velocity of 4 km/s in these sediments (which are close to the basement) we have calculated the vertical offset in depth. We have neglected any affects due to differential water bottom. Based on these calculation as shown in Table 1, we note that our results are very consistent with Chamot-Rooke et al [1993]’s results.
Fig. 5 Part of a seismic line along 81.5°E from Fig. 2 of Chamot-Rooke et al [1993]. Vertical throw is calculated on the highlighted faults marked from 1 to 5.
Table 1

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Effects due to differential water bottom have been neglected.