

Horizontal thermal contraction of oceanic lithosphere: The ultimate limit to the rigid plate approximation

Ravi R. Kumar¹ and Richard G. Gordon¹

Received 30 October 2007; revised 8 September 2008; accepted 14 October 2008; published 16 January 2009.

[1] There is a contradiction between two widely accepted pillars of global tectonics, (1) the central plate tectonic assumption of plate rigidity and (2) the explanation of the relief of the seafloor as being due to lithospheric subsidence from thermal contraction. Here we quantify the rate of predictable horizontal thermal contraction of the lithosphere using depth averages of widely accepted thermal models. Depth-averaged cooling rate, and thus depth-averaged contraction rate, is proportional to t^{-1} , where t is the age of the lithosphere. Depth-averaged thermal contraction rate of old (i.e., 167 Ma old) lithosphere is 10^{-5} Ma^{-1} ($3 \times 10^{-19} \text{ s}^{-1}$), which is an order of magnitude greater than the average strain rate inferred from seismic moment release. Newly created (i.e., 0.1 Ma old) lithosphere is displaced by thermal contraction toward lithosphere in old ocean basins at a rate that is 1.35% of the half rate of seafloor spreading, giving displacement rates of 0.1 to 1.1 km Ma^{-1} (or, equivalently, 0.1 to 1.1 mm a^{-1}). Predicted displacement rates parallel to midocean ridges depend strongly on lithospheric age and are proportional to the distance along which contractional strain rates are integrated. Displacement rates can be significant ($\geq 1 \text{ km Ma}^{-1}$) for young lithosphere. In particular the displacement of oceanic lithosphere adjacent to southern Baja California relative to lithosphere near the Pacific-Antarctic Rise has an indicated north-south displacement rate between ≈ 3 and $\approx 10 \text{ km Ma}^{-1}$. This suggests that global plate motion circuits based on the assumption of plate rigidity may be biased by a substantial velocity.

Citation: Kumar, R. R., and R. G. Gordon (2009), Horizontal thermal contraction of oceanic lithosphere: The ultimate limit to the rigid plate approximation, *J. Geophys. Res.*, 114, B01403, doi:10.1029/2007JB005473.

1. Introduction

[2] The central assumption of plate tectonics is that the plates are rigid [Wilson, 1965; McKenzie and Parker, 1967; Morgan, 1968; Gordon, 1998]. In contrast to this assumption, a great success of the mobile view of the solid Earth has been the explanation of the broad features of submarine topography in terms of the cooling and subsidence of the lithosphere with age [McKenzie, 1967; Turcotte and Oxburgh, 1967; Parsons and Sclater, 1977; Stein and Stein, 1992; McKenzie et al., 2005; Hillier and Watts, 2005]. Quantitative models of lithospheric subsidence explicitly assume that the lithosphere contracts with age. Coefficients of thermal expansion for olivine differ little in crystallographic orientation [Bouhifd et al., 1996], which indicates that contraction should be dominantly isotropic. Thus these two widely accepted pillars of global tectonics are contradictory. It follows that oceanic plates cannot be rigid.

[3] Surprisingly, we are aware of no prior efforts to quantify the predictable horizontal thermal contraction of the lithosphere on a plate-wide scale. Here we use widely accepted models for subsidence of the oceanic lithosphere

to do so. We find that contraction rates are large enough to require small but significant modifications to the rigid-plate hypothesis.

2. Methods and Assumptions

[4] We apply half-space cooling models [Turcotte and Oxburgh, 1967], which indicate that the temperature T in the cooling half-space is given by

$$T(z, t) = T_m \text{erf}\left[\frac{z}{(2\sqrt{\kappa t})}\right] \quad (1)$$

where z is depth, t is the age of the lithosphere, κ is thermal diffusivity, and T_m is initial temperature. Given our focus on the cooling of the crust and uppermost mantle, a half-space cooling model should give essentially the same result as a plate model.

[5] Haxby and Parmentier [1988] examine thermal contractional stresses due to temperature changes in a thin, semi-infinite, perfectly elastic, rectangular plate bounded by the spreading center axis. Their results suggest that the thermoelastic stresses are small enough to neglect except very near ridges and transform faults. To simplify the analysis, we neglect the effects of thermoelastic stresses.

[6] Relevant thermal parameters, including the coefficients of thermal contraction and thermal diffusivity, are known to vary with temperature and pressure. To obtain a

¹Department of Earth Science, Rice University, Houston, Texas, USA.

Table 1. Critical Ages for Reference Depth-Averaged Thermal Contraction Rates^a

Log of Strain Rate (Ma ⁻¹)	T_l (°C)	$T_m = 1300$ (Ma)	$T_m = 1350$ (Ma)
-3	600	1.45	1.46
	700	1.67	1.68
	800	1.88	1.89
-3.5	600	4.60	4.61
	700	5.29	5.31
	800	5.94	5.97
-4	600	14.5	14.6
	700	16.7	16.8
	800	18.8	18.9
-5	600	145	146
	700	167	168
	800	188	189

^a T_l is the assumed temperature at the base of the competent lithosphere. T_m is assumed initial temperature in a half-space cooling model.

simple analytic solution, we use values that are the approximate average over the range of temperatures and pressures that we investigate.

[7] A coefficient of linear thermal expansion $\alpha_l = 10^{-5} (\text{°C})^{-1}$ has been used in all calculations. Initial temperatures, T_m , of 1300°C and 1350°C, are assumed (Table 1), consistent with recent estimates [McKenzie *et al.*, 2005; Hillier and Watts, 2005; Lee *et al.*, 2005]. Our plots use the lower of these two values of T_m , which gives slightly lower contraction rates than would be found from the higher value (Table 1).

[8] Rates of cooling, and hence contraction, vary with depth in the lithosphere. We assume that the potentially observable horizontal contraction of oceanic lithosphere will be a response to thermal stress averaged from the surface of the lithosphere down to the base of the competent lithosphere. Hereinafter we refer to this vertically averaged thermal stress as the contractional stress (Figure 1) [Parmentier and Haxby, 1986]. We assume that stresses are relaxed by creeping flow below the competent lithosphere.

[9] The surface of the lithosphere is instantaneously cooled to 0°C. Thus the surface has already fully contracted thermally at the ridge axis. Deeper in the competent lithosphere, the plate continues to cool, with the most rapid cooling occurring at its base. To relieve the contractional stress, the upper lithosphere must deform with the greatest contractional deformation, a strain of $\approx 0.4\%$, occurring at the surface. The near-surface lithosphere is relatively weak, as indicated by Byerlee's law, and porous, and thus would readily accommodate the assumed deformation. If, as we assume, the contractional stress is relieved by contractional strain, the upper competent lithosphere will be in horizontal deviatoric compression and the lower competent lithosphere in horizontal deviatoric tension (Figure 1), which is confirmed by earthquake mechanisms [Wiens and Stein, 1984; Bergman and Solomon, 1984; Wessel, 1992]. That the upper lithosphere contracts is further supported by the agreement of geoid anomalies across the Clarion and Udintsev fracture zones with predictions of models for flexure near fracture zones because such models assume that contractional straining fully relieves the contractional stress [Parmentier and Haxby, 1986]. Models for bending and cracking of the lithosphere in response to cooling [Gans *et al.*, 2003; Sandwell and Fialko, 2004] also assume that bending is superposed on the free horizontal contraction of the lithosphere. As the competent lithosphere contracts and bends, the uppermost lithosphere fails in compression resulting in a permanent contraction [Parmentier and Haxby, 1986; Wessel, 1992; Gans *et al.*, 2003; Sandwell and Fialko, 2004].

[10] We assume that the base of the competent lithosphere follows an isotherm at temperature T_l [Turcotte and Oxburgh, 1967] for which we consider three alternative temperatures, 600°C, 700°C, and 800°C. The choice of the 600°C isotherm as the lowest temperature follows from the work of McKenzie *et al.* [2005], which showed that the base of the seismogenic lithosphere approximately follows this temperature. Thus it gives a minimum thickness for a

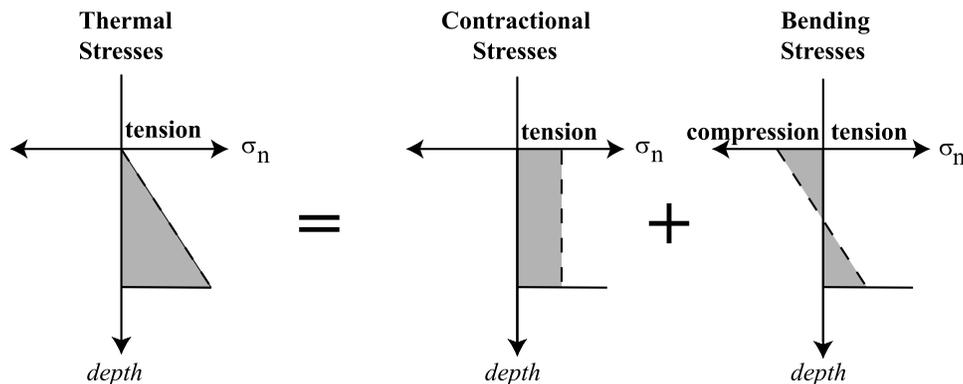


Figure 1. Simplified schematic of thermal stresses in a plate growing by accretion at its base and cooling from above. The material added to the bottom cools most rapidly, setting up tensional stresses. The thermal stresses can be decomposed into two parts, the uniform (vertically averaged) stress (contractional stresses) and the nonuniform (bending) stresses. Herein, following many others [Parmentier and Haxby, 1986; Wessel, 1992; Gans *et al.*, 2003; Sandwell and Fialko, 2004], we assume (except in very young lithosphere) that the contractional stresses are mainly relieved by free horizontal contraction of the lithosphere and that the remaining bending stresses flex the lithosphere. As the lithosphere contracts and flexes, the weak uppermost lithosphere fails in compression and is permanently shortened. After Wessel [1992].

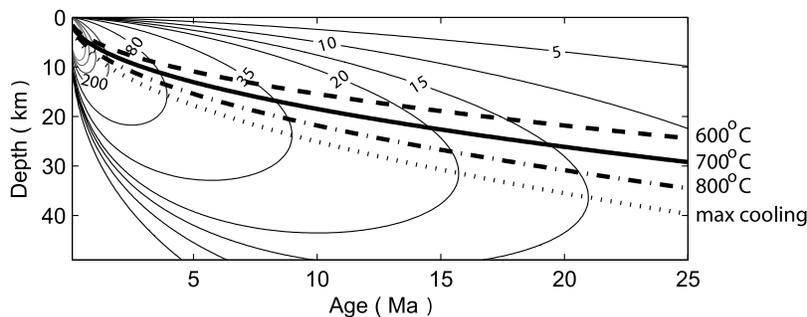


Figure 2. Thin solid curves depict contours of cooling rate ($^{\circ}\text{C Ma}^{-1}$) as a function of depth and age. Contours converge at zero depth in the youngest part of the lithosphere. Dashed curve, bold solid curve, and dash-dotted curve, respectively, show the 600°C , 700°C , and 800°C isotherm. Dotted curve represents the locus of fastest cooling, which corresponds to an isotherm of 888°C . After *Bratt et al.* [1985] and *Wessel* [1992].

temperature-controlled competent layer. The limiting depth of earthquakes is thought to be the boundary between unstable sliding and stable sliding. How far beneath this transition lies the brittle-plastic transition (i.e., the transition from semibrittle deformation to creeping flow [*Kohlstedt et al.*, 1995]) is poorly known, but in some prior work it has been assumed to occur at a depth about 20% greater than that of the base of the seismogenic zone [e.g., *Martinod and Molnar*, 1995]. We consider this possibility through models with limiting depths for the competent layer at temperatures of 700°C and 800°C , which respectively lie at depths 20% and 41% greater than the depth of the 600°C isotherm. Our preferred value is 700°C , which *Parmentier and Haxby* [1986] found to give the best fit to the amplitude of geoid anomalies due to thermal stresses on profiles across fracture zones. Reorganizing equation (1), l , the thickness of the competent lithosphere, is given by

$$l(t) = 2\sqrt{\kappa t} \operatorname{erf}^{-1}(T_i/T_m) \quad (2)$$

[*Collette*, 1974; *Sandwell*, 1986; *Wessel*, 1992]. It follows that the depth-averaged rate of cooling for a lithosphere bounded by a constant isotherm is given by

$$(1/l) \int_0^l (\partial T/\partial t) dz = T_m [\exp(-A^2) - 1] / (2At\sqrt{\pi}) \quad (3)$$

where $A = \operatorname{erf}^{-1}(T_i/T_m)$ [*Collette*, 1974; *Sandwell*, 1986; *Wessel*, 1992]. Thus the depth-averaged cooling rate is proportional to t^{-1} and independent of κ .

[11] Below we ignore all the cooling of oceanic lithosphere with ages between 0 Ma and 0.1 Ma to avoid the breakdown of the validity of the conductive cooling model as the axis of the midocean ridge is approached. The choice of 0.1 Ma, which corresponds to a ≈ 2 km thickness of the competent lithosphere, approximately including layers 2A and 2B, is reasonable but arbitrary. It does, however, correspond to an age and distance ten times as great as the outer boundaries of the crustal accretion zone defined by *Macdonald* [1982] and thus is an order of magnitude wider than the active accreting plate boundary. A more elegant approach, which is equally arbitrary and effectively similar to our approach, is to replace age, t , with $t + t_0$ in equations (1) to (3). *Parmentier and Haxby* [1986] chose t_0 to give a zero-age lithospheric thickness of 3 km and *Wessel* [1992]

chose t_0 to give a zero-age thickness of 1 km, bracketing our choice of ≈ 2 km for initial lithospheric thickness.

3. Results

3.1. Rate of Cooling as a Function of Depth and Age

[12] For any non-zero lithospheric age, the rate of cooling increases with depth until it reaches a maximum, which occurs at the 888°C isotherm (Figure 2). This temperature exceeds the values of 600°C to 800°C that we assume for T_i . Thus in the competent lithosphere, the rate of cooling is greatest at the base and least at the surface [*Wessel*, 1992].

3.2. Depth-Averaged Rate of Cooling

[13] Depth-averaged rate of cooling for the competent lithosphere decrease with age as t^{-1} and is 10^2 $^{\circ}\text{C Ma}^{-1}$ for 1.67-Ma-old lithosphere, $10^{\circ}\text{C Ma}^{-1}$ for 16.7-Ma-old lithosphere, and $1^{\circ}\text{C Ma}^{-1}$ for 167-Ma-old lithosphere (Figure 3). Unsurprisingly, the higher the temperature assumed for the base of the competent lithosphere, the greater the depth-averaged rate of cooling (Figure 3).

3.3. Thermal Contractual Strain Rates

[14] Using a value for α_1 of 10^{-5} ($^{\circ}\text{C}^{-1}$), the curves of rates of cooling can be simply transformed to one-dimensional rates of contractional strain. For our preferred model, contractional strain rate is 10^{-3} Ma^{-1} for 1.67-Ma-old lithosphere, is 10^{-4} Ma^{-1} for 16.7-Ma-old lithosphere, and 10^{-5} Ma^{-1} for 167-Ma-old lithosphere (Figure 3 and Table 1).

[15] The depth-averaged cooling rates and contraction rates as a function of age can be used to transform standard global catalogs of the age of the seafloor [*Müller et al.*, 1997] into a map of depth-averaged cooling rate or contraction rate (Figure 4). The maps indicate that most of the ocean basins contract at strain rates between 10^{-5} Ma^{-1} (≈ 167 -Ma-old lithosphere) and 10^{-4} Ma^{-1} (16.7-Ma-old lithosphere), with a substantial fraction also between 10^{-4} Ma^{-1} and 10^{-3} Ma^{-1} (1.67-Ma-old lithosphere), the latter roughly corresponding to the youngest age for which thermoelastic stresses can be neglected.

3.4. Thermal Contractual Displacement Rates

[16] Displacement rate across a contracting region can be estimated by integrating predicted strain rate over a given

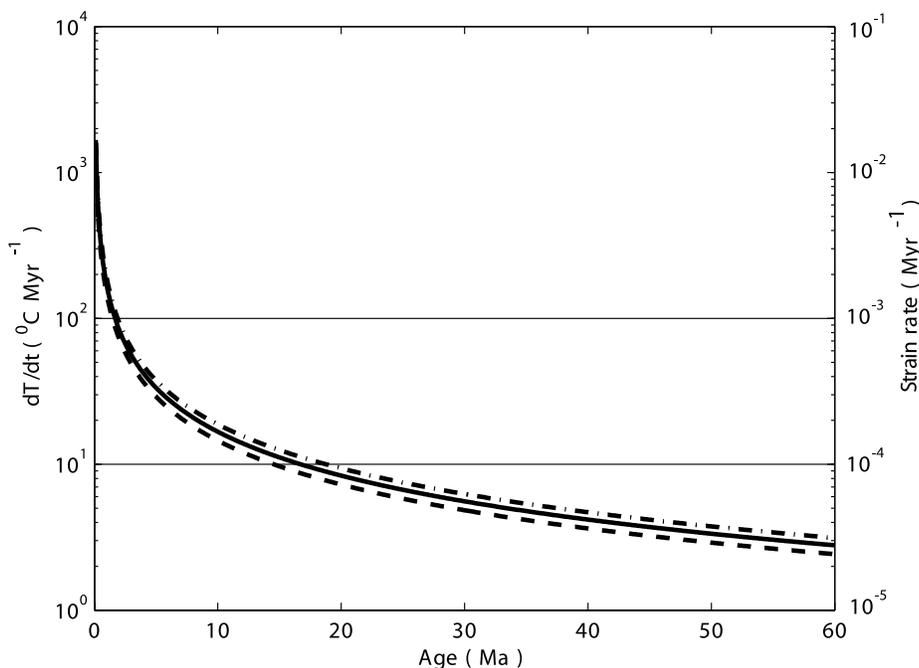


Figure 3. Depth-averaged rate of cooling versus age of the oceanic lithosphere. Dashed, solid, and dash-dotted curves correspond to models for which the base of the competent lithosphere is defined by the 600°C, 700°C, and 800°C isotherm, respectively.

distance. Most oceanic lithosphere is contracting at rates between 10^{-5} Ma^{-1} and 10^{-4} Ma^{-1} . If these typical thermal contraction rates are used along a hypothetical circumferential great circle path (i.e., $\approx 40,000 \text{ km}$ long), then a contractional displacement rate of 0.4 to 4 km Ma^{-1} (equivalent to 0.4 to 4 mm a^{-1}) would result from these typical values of thermal contractional strain rate. A misfit at the high end of this range would presumably cause observable nonclosure of global plate motion circuits. Local or regional paths of integration typically give lower displacement rates due to thermal contraction, but some can be higher and may lead to larger nonclosures of plate motion circuits.

[17] For each plate pair separating along a midocean ridge, the displacement rates due to horizontal thermal contraction can be decomposed into ridge-parallel and ridge-perpendicular components. The assumption that spreading rate is constant over time for any plate pair

permits a simple, universal, quantitative assessment of the ridge-perpendicular component of displacement rate due to horizontal thermal contraction. The ridge-perpendicular component of displacement rate of 0.1-Ma-old lithosphere relative to lithosphere of any older age can be expressed as a percentage of the spreading half rate. This percentage increases monotonically with age and is 1.35% for 80-Ma-old lithosphere (Figure 5). For the ultrafast segments of the East Pacific Rise (spreading half rate of 80 km Ma^{-1}), this gives a displacement rate of only 1.1 km Ma^{-1} (2.2 km Ma^{-1} full rate), which is small, but potentially detectable with space geodetic data (assuming the existence of enough sites on islands in optimum locations). At the other extreme, for ultraslow spreading (half rate of 5 km Ma^{-1}), the displacement rate is minuscule, 0.07 km Ma^{-1} (0.14 km Ma^{-1} full rate), and presumably is too small to be observed.

[18] Because of ridge-perpendicular horizontal thermal contraction, plate velocities estimated from marine magnetic

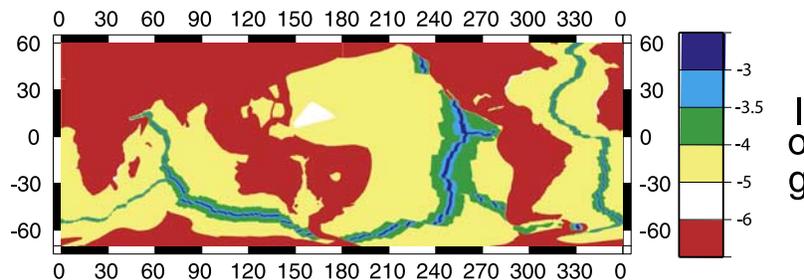


Figure 4. Depth-averaged thermal contractional strain rates (10^x Ma^{-1}) predicted for oceanic lithosphere. Red indicates continents or regions of oceanic lithosphere lacking age estimates.

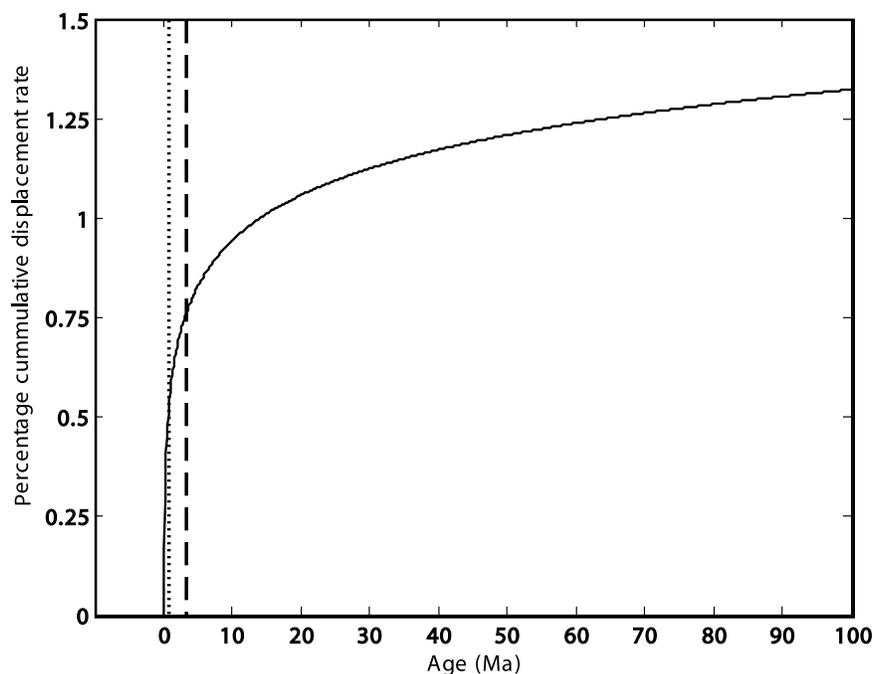


Figure 5. Solid curve: cumulative displacement rate as percentage of plate velocity as a function of lithospheric age. Dotted line: the age (0.78 Ma before present) for the old edge of the central anomaly. Dashed line: the age (3.2 Ma before present) for the middle of marine magnetic anomaly 2A, which was used in estimating the NUVEL-1A global set of relative plate angular velocities. Because of thermal contraction, plate interiors move $\approx 1\%$ slower (i.e., 0.1 to 0.8 mm a^{-1}) than estimated from spreading half rates determined from marine magnetic anomalies.

anomalies across midocean ridges are biased estimators of the relative velocities of plate interiors. The bias is not the full value of 1.35%, however, because significant thermal contraction occurs in the across-ridge interval spanned by the magnetic anomalies used to estimate the spreading rate. If this interval spans 3.2 Ma, as is the case for the NUVEL-1A set of relative angular velocities [DeMets *et al.*, 1994], then the bias is the displacement rate of 3.2-Ma-old lithosphere relative to 80-Ma-old lithosphere, which is 0.60% (1.35% for 80-Ma-old lithosphere minus 0.75% for 3.2-Ma-old lithosphere; Figure 5). Thus we predict that the velocities of stable plate interiors are on average about 0.6% slower than those predicted by NUVEL-1A or any other estimate constructed from spreading rates averaged over the past 3.2 Ma. For fast-moving plate pairs, this is potentially detectable from space geodetic data. For example, the effect on Pacific-Nazca spreading rate is 1.0 km Ma^{-1} ($=0.6\% \times 160 \text{ km Ma}^{-1}$).

[19] An alternative magnetic anomaly used in some analyses of current spreading rate is the outside edge of the Central Anomaly, corresponding to the Brunhes-Matuyama reversal boundary (0.78 Ma). In this case, the bias is the displacement rate of 0.78-Ma-old lithosphere relative to 80-Ma-old lithosphere, which is 0.85% (1.35% for 80-Ma-old lithosphere minus 0.50% for 0.78-Ma-old lithosphere; Figure 5). Thus we predict that the velocities of stable plate interiors are on average about 0.85% slower than predicted by any estimates from spreading rates averaged over the past 0.78 Ma.

[20] Because of this simple relation to spreading rate, if contraction occurred only in the plate motion direction (i.e., approximately in the ridge-perpendicular direction), thermal

contraction would not cause nonclosure of plate circuits but would resemble a worldwide apparent acceleration in plate motions.

4. Discussion

4.1. Isochron-Parallel Contraction and the Pacific Plate

[21] Displacement rates parallel to midocean ridges depend strongly on the age of the lithosphere and are proportional to the distance along which the strain rate is integrated. Displacement rates are greatest where the lithosphere is the youngest. Because of the fast spreading along the East Pacific Rise and the Pacific-Antarctic Rise, the Pacific basin has Earth's largest concentration of young, rapidly contracting oceanic lithosphere (Figure 4). Consider the isochron (5.3 Ma) in the Pacific plate corresponding to a thermal contractional strain rate of $3 \times 10^{-4} \text{ Ma}^{-1}$ (" $10^{-3.5}$ ", the border between light blue and green in Figure 4). This isochron spans about 10,000 km in latitude. The implied north-south contractional displacement rate thus is 3 km Ma^{-1} although the rate may not be fully realized because of the offset of this isochron along fracture zones and the absorption of some of the contraction by widening of transform faults [Collette, 1974; Sandwell, 1986]. Pacific plate seafloor to the west of this isochron contracts at rates between about $1 \times 10^{-5} \text{ Ma}^{-1}$ and $3 \times 10^{-4} \text{ Ma}^{-1}$. Over north-south distances of 10,000 km (say from 60°S to 30°N), these contractional strain rates correspond to displacement rates of 0.1 to 3 km Ma^{-1} . In contrast, Pacific plate lithosphere east of this isochron but west of the

1.67 Ma isochron contracts at rates between $3 \times 10^{-4} \text{ Ma}^{-1}$ and 10^{-3} Ma^{-1} . Over north–south distances of 10,000 km, these correspond to displacement rates of $\approx 3 \text{ km Ma}^{-1}$ to $\approx 10 \text{ km Ma}^{-1}$.

[22] The greatest hypothetical displacement rates would occur along narrow strips of very young seafloor. A great circle path from Baja California roughly parallel to isochrons suggests a total displacement rate of at least several millimeters per year, which may be detectable in estimates of geologically instantaneous plate motions [e.g., *DeMets et al.*, 1990, 1994] or relative to plate motions estimated from space geodetic data.

[23] Unfortunately, there are no geodetic sites on the Pacific plate near the Pacific–Antarctic Rise for making a direct test. It is nonetheless interesting that recent geodetic work shows that the island of Guadelupe, which lies west of Baja California on the youngest seafloor of any Pacific plate site, moves at $2.0 \pm 0.6 (\pm 1\sigma) \text{ mm a}^{-1}$ toward 149° relative to the older part of the Pacific plate [see Figure 3 of *Plattner et al.*, 2007]. This velocity may be consistent with the pattern of strain due to thermal contraction that we postulate. Baja California also moves relative to the Pacific plate at $\approx 5 \text{ mm a}^{-1}$ to the south–southeast. *Plattner et al.* [2007] convincingly argue for a separate Baja microplate, but part (i.e., $\approx 2 \text{ mm a}^{-1}$) of the difference between the motion of Baja California and the older part of the Pacific plate may be due to thermal contraction of young Pacific oceanic lithosphere as well. It is also of interest that the northern and southern GPS station networks on Baja California are converging [*Plattner et al.*, 2007], which might be related to the predicted thermal contraction of young oceanic lithosphere adjacent to Baja. The observed nominal strain rate ($\approx 2 \times 10^{-3} \text{ Ma}^{-1}$) is larger than is predicted for thermal contraction, but may be consistent with predictions within uncertainties.

[24] If our arguments are correct, plate reconstructions of Pacific–North America motion millions to tens of millions of years before present [e.g., *Atwater and Stock*, 1998; *Wilson et al.*, 2005; *Royer et al.*, 2006] may be biased by $\approx 40 \text{ km}$ ($\approx 0.4\%$ of 10^4 km) by the horizontal thermal contraction of young Pacific plate lithosphere.

4.2. Detection of Plate Nonrigidity Due to Horizontal Thermal Contraction

[25] We argue above that high rates of horizontal thermal contractional strain occur in young oceanic lithosphere and, therefore, that the plates are not strictly rigid. Insofar as the nonrigidity is caused by uniform thermal contraction, no earthquakes or other indicators of deformation are expected in response to the uniform part of the contraction, although earthquakes and deformation are observed in response to the nonuniform part of the contraction (Figure 1) [*Wiens and Stein*, 1984; *Bergman and Solomon*, 1984; *Wessel*, 1992; *Gans et al.*, 2003; *Sandwell and Fialko*, 2004]. Is there evidence for plate nonrigidity that may be caused by thermal contraction?

[26] Plate-circuit closures involving three or more plates test the assumption of plate rigidity and thus may detect nonrigidity due to thermal contraction. The most reliable tests of circuits consisting of three plates arise when all three plate pairs are separated by a boundary consisting of midocean ridge segments and transform faults. Four such

plate circuits were investigated by *DeMets et al.* [1990]: (1) Africa–South America–Antarctica (Bouvet triple junction), (2) Africa–Australia–Antarctica (Rodrigues triple junction), (3) Cocos–Nazca–Pacific (Galapagos triple junction), and (4) Pacific–Antarctic–Nazca (Juan Fernandez triple junction). *DeMets et al.* [1990] found that two of these four, Bouvet and Galapagos, failed closure significantly. We hypothesize that the nonclosure of these two plate circuits is caused by horizontal thermal contraction of oceanic lithosphere. Of particular interest is the failure of the Galapagos (Pacific–Nazca–Cocos) circuit, because it includes the Cocos plate, which is the plate with the youngest lithosphere on average incorporated into any circuits in NUVEL-1. The oldest part of the Cocos plate is predicted to contract at a rate of $\approx 10^{-4} \text{ Ma}^{-1}$ and much of it at rates substantially higher.

4.3. Comparison of Horizontal Thermal Contractional Strain Rates With Other Strain Rates

[27] The magnitude of nonrigidity of plate interiors is known only within broad limits. A lower bound on strain rate of about 10^{-6} Ma^{-1} can be established from the rate of moment release of intraplate earthquakes [*Gordon*, 1998]. This rate may be unrepresentative of typical lithosphere, however, if the moment release is concentrated in a few narrow belts. On the other hand, the upper bound on strain rate of stable plate interiors is more than two orders of magnitude greater than this. *Argus and Gordon* [1996] use geodetic results from very long baseline interferometry to place an upper bound on the laterally averaged strain rate of a large portion of the North American plate of $\approx 10^{-17} \text{ s}^{-1}$ ($\approx 3 \times 10^{-4} \text{ Ma}^{-1}$). The actual spatially averaged deformation rate could be much less than this upper bound.

[28] Thus the thermal contractional strain rate that we calculate may be better known than the background intraplate strain rate associated with earthquakes and faulting. Relevant strain rates are compared in Figure 6. What is new in Figure 6 (compared with prior versions) is the inclusion of the range of horizontal contractional strain rates due to cooling of oceanic lithosphere. In our estimates, the smallest depth-averaged rate of horizontal thermal contractional strain in oceanic lithosphere is $\approx 10^{-5} \text{ Ma}^{-1}$, which corresponds to 167-Ma-old lithosphere. Thus the smallest depth-averaged thermal contractional strain rate anywhere in oceanic lithosphere is an order of magnitude greater than the average strain rate inferred from seismic moment release. The depth-averaged thermal contractional strain rate for 16.7-Ma-old oceanic lithosphere is two orders of magnitude greater, and that of 1.67-Ma-old lithosphere is three orders of magnitude greater, than the strain rate released through intraplate oceanic earthquakes. These calculations give new lower bounds, substantially greater than found before, on the average background strain rate (i.e., nonrigidity) of stable oceanic plate interiors [*Gordon*, 1998].

5. Conclusion

[29] Because of the predictable effects of thermal contraction, we conclude that oceanic lithosphere, and especially young oceanic lithosphere, is not rigid. The ocean-wide-averaged, depth-averaged strain rate due to thermal contraction of oceanic lithosphere is about one to three orders of

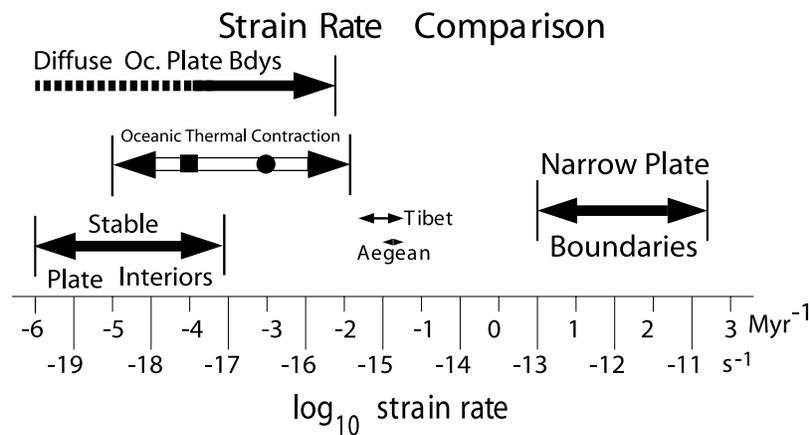


Figure 6. Comparison of the range of thermal contractional strain rate to the ranges of regionally averaged strain rates associated with large-scale tectonic processes (modified from Gordon [2000]). Square: Depth-averaged thermal contractional strain rate for 16.7-Ma-old oceanic lithosphere. Circle: Depth-averaged thermal contractional strain rate for 1.67-Ma-old oceanic lithosphere. The low end of the range of thermal contraction strain rates is the depth-averaged rate for the oldest oceanic lithosphere on Earth and defines the lower bound on nonrigidity of the stable interior of oceanic plates.

magnitude greater than the average strain rate inferred from earthquake moment release in intraplate oceanic lithosphere. Thermal contractional strain rates are greatest for young lithosphere. Thermal contraction likely leads to displacement rates large enough to be measured in the analysis of some plate motion circuits as may be indicated by the nonclosure of the Cocos-Nazca-Pacific plate circuit (Galapagos triple junction). Moreover, the small misfit to Pacific-North America plate velocities documented by C. DeMets, R. G. Gordon and D. F. Argus (manuscript in preparation, 2008) is also likely due, at least in part, to ongoing thermal contraction of Neogene Pacific plate lithosphere created along the East Pacific Rise and Pacific-Antarctic Rise. Horizontal thermal contraction appears to be the ultimate limit to the rigid plate approximation.

[30] **Acknowledgments.** We thank an anonymous reviewer, Norm Sleep, and associate editor Don Forsyth for helpful comments on the manuscript. This work was supported by National Science Foundation grant OCE-0453219.

References

- Argus, D. F., and R. G. Gordon (1996), Tests of the rigid-plate hypothesis and bounds on intraplate deformation using geodetic data from very long baseline interferometry, *J. Geophys. Res.*, *101*, 13,555–13,572.
- Atwater, T., and J. M. Stock (1998), Pacific-North America plate tectonics of the Neogene southwestern United States—An update, *Int. Geol. Rev.*, *40*, 375–402.
- Bergman, E. A., and S. C. Solomon (1984), Source mechanisms of earthquakes near mid-ocean ridges from body waveform inversion: Implications for the early evolution of oceanic lithosphere, *J. Geophys. Res.*, *89*, 11,415–11,441.
- Bouhifd, M. A., D. Andraut, G. Fiquet, and P. Richet (1996), Thermal expansion of forsterite up to the melting point, *Geophys. Res. Lett.*, *23*, 1143–1146.
- Bratt, S. R., E. A. Bergman, and S. C. Solomon (1985), Thermoelastic stress: How important as a cause of earthquakes in young oceanic lithosphere?, *J. Geophys. Res.*, *90*, 10,249–10,260.
- Collette, B. J. (1974), Thermal contraction joints in a spreading seafloor as origin of fracture zones, *Nature*, *251*, 299–300.
- DeMets, C., R. G. Gordon, D. F. Argus, and S. Stein (1990), Current plate motions, *Geophys. J. Int.*, *101*, 425–478.
- DeMets, C., R. G. Gordon, D. F. Argus, and S. Stein (1994), Effect of recent revisions to the geomagnetic reversal timescale on estimates of current plate motions, *Geophys. Res. Lett.*, *21*, 2191–2194.
- Gans, K. D., D. S. Wilson, and K. C. Macdonald (2003), Pacific plate gravity lineaments: Diffuse extension or thermal contraction?, *Geochem. Geophys. Geosyst.*, *4*(9), 1074, doi:10.1029/2002GC000465.
- Gordon, R. G. (1998), The plate tectonic approximation: Plate nonrigidity, diffuse plate boundaries, and global plate motions, *Annu. Rev. Earth Planet. Sci.*, *26*, 615–642.
- Gordon, R. G. (2000), Diffuse oceanic plate boundaries, in *The History and Dynamics of Global Plate Motion*, edited by M. Richards, R. G. Gordon, and R. Van der Hilst, *Geophys. Monogr. Ser.*, vol. 121, pp. 143–159, AGU, Washington, D. C.
- Haxby, W. F., and E. M. Parmentier (1988), Thermal contraction and the state of stress in the oceanic lithosphere, *J. Geophys. Res.*, *93*, 6419–6429.
- Hillier, J. K., and A. B. Watts (2005), Relationship between depth and age in the North Pacific Ocean, *J. Geophys. Res.*, *110*, B02405, doi:10.1029/2004JB003406.
- Kohlstedt, D. L., B. Evans, and S. J. Mackwell (1995), Strength of the lithosphere: Constraints imposed by laboratory experiments, *J. Geophys. Res.*, *100*, 17,587–17,602.
- Lee, C.-T., A. Lenardic, C. Cooper, F. Niu, and A. Levander (2005), The role of chemical boundary layers in regulating the thickness of continental and oceanic thermal boundary layers, *Earth Planet. Sci. Lett.*, *230*, 379–395.
- Macdonald, K. C. (1982), Mid-ocean ridges: Fine scale tectonic, volcanic and hydrothermal processes within the plate boundary zone, *Ann. Rev. Earth Planet. Sci.*, *10*, 155–190.
- Martinod, J., and P. Molnar (1995), Lithospheric folding in the Indian Ocean and the rheology of the oceanic plate, *Bull. Soc. Geol. Fr.*, *166*, 813–821.
- McKenzie, D. P. (1967), Some remarks on heat flow and gravity anomalies, *J. Geophys. Res.*, *72*, 6261–6273.
- McKenzie, D. P., and R. L. Parker (1967), The North Pacific: an example of tectonic on a sphere, *Nature*, *216*, 1276–1280, doi:10.1038/2161276a0.
- McKenzie, D. P., J. Jackson, and K. Priestly (2005), Thermal structure of oceanic and continental lithosphere, *Earth Planet. Sci. Lett.*, *233*, 337–349.
- Morgan, W. J. (1968), Rises, trenches, great faults, and crustal blocks, *J. Geophys. Res.*, *73*, 1959–1982.
- Müller, R. D., W. R. Roest, J.-Y. Royer, L. M. Gahagan, and J. G. Sclater (1997), Digital isochrons of the world's ocean floor, *J. Geophys. Res.*, *102*, 3211–3214.
- Parmentier, E. M., and W. F. Haxby (1986), Thermal stresses in the oceanic lithosphere: Evidence from geoid anomalies at fracture zones, *J. Geophys. Res.*, *91*, 7193–7204.
- Parsons, B., and J. G. Sclater (1977), An analysis of the variation of ocean floor bathymetry and heat flow with age, *J. Geophys. Res.*, *82*, 803–827.
- Plattner, C., R. Malservisi, T. H. Dixon, P. LaFemina, G. F. Sella, J. Fletcher, and F. Suarez-Vidal (2007), New constraints on relative motion between the Pacific plate and Baja California microplate (Mexico) from GPS measurements, *Geophys. J. Int.*, doi:10.1111/j.1365-246X.2007.03494.x.
- Royer, J.-Y., R. G. Gordon, and B. C. Homer-Johnson (2006), Motion of Nubia relative to Antarctica since 11 Ma: Implications for Nubia-Somalia,

- Pacific-North America, and India-Eurasia motion, *Geology*, *34*, 501–504, doi:10.1130/G22463.1.
- Sandwell, D. T. (1986), Thermal stress and the spacings of transform faults, *J. Geophys. Res.*, *91*, 6405–6418.
- Sandwell, D., and Y. Fialko (2004), Warping and cracking of the Pacific plate by thermal contraction, *J. Geophys. Res.*, *109*, B10411, doi:10.1029/2004JB003091.
- Stein, C. A., and S. Stein (1992), A model for the global variation in oceanic depth and heat flow with lithospheric ages, *Nature*, *359*, 123–129.
- Turcotte, D. L., and E. R. Oxburgh (1967), Finite amplitude convective cells and continental drift, *J. Fluid Mech.*, *28*, 29–42.
- Wiens, D. A., and S. Stein (1984), Intraplate seismicity and stresses in young oceanic lithosphere, *J. Geophys. Res.*, *89*, 11442–11464.
- Wessel, P. (1992), Thermal stresses and the bimodal distribution of elastic thickness estimates of the oceanic lithosphere, *J. Geophys. Res.*, *97*, 14,177–14,193.
- Wilson, J. T. (1965), A new class of faults and their bearing on continental drift, *Nature*, *207*, 343–347.
- Wilson, D. S., P. A. McCrory, and R. G. Stanley (2005), Implications of volcanism in coastal California for the Neogene deformation history of western North America, *Tectonics*, *24*, TC3008, doi:10.1029/2003TC001621.

R. G. Gordon and R. R. Kumar, Department of Earth Science, Rice University, Mail Stop 126, P.O. Box 1892, Houston, TX 77251-1892, USA. (rgg@rice.edu)