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Pacific plate deformation from horizontal thermal contraction

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ABSTRACT

The central approximation of plate tectonics is that the plates are rigid, which gives the theory its rigor and predictive power. Space geodetic measurements are consistent with the rigidity of stable plate interiors, but some failures of plate-circuit closure, in particular of oceanic plates, indicate that plates may be measurably non-rigid. We explore the hypothesis that horizontal thermal contraction causes deformation of oceanic plates. Here we show significant expected displacement fields due to thermal contraction for the Pacific plate based on a previously proposed relationship between sea-floor age and strain rate and on two end-member assumptions on how strain compatibility is enforced. The predicted maximum 2.2 mm/yr south-eastward motion of the north-eastern part of the plate relative to the Pacific-Antarctic Rise may contribute to a large part of the non-closure of the Pacific-North America plate motion circuit. Our predicted displacement rates cannot (yet) be confirmed by current space-geodetic data and will require seafloor geodesy with 1 mm/yr accuracy. The spatial distribution of predicted moment rate agrees reasonably well with that of intraplate earthquakes epicenters, similar to what is observed for plate boundary zones. Our results suggest that plate-scale horizontal thermal contraction is significant and that it may be partly released seismically.

INTRODUCTION

Plate tectonics has transformed our view of how our planet works. The central assumption or approximation of plate tectonics is that the plates are rigid. The assumption of plate rigidity is what gives plate tectonics its rigor and predictive power. There are two main quantitative tests of plate rigidity: plate-circuit closure and space geodetic measurements. The hypothesis of plate rigidity has survived many critical tests (e.g., DeMets et al., 1990; Argus and Gordon, 1996; Dixon et al., 1996; Beavan et al., 2002).

Nonetheless, there is mounting evidence that tectonic plates, in particular oceanic plates, are not rigid. For example, DeMets et al. (2010) find the largest misfit in the 3-plate circuit (i.e., inverting data only along the 3 mutual boundaries of those 3 plates) occurs for the Pacific-Nazca-Cocos circuit, namely 14 ± 5 mm/yr (95% confidence limit). Another important circuit that fails closure is the 4-plate Pacific-Antarctica-Nubia-North America circuit (i.e., using data only along the Pacific-Antarctica boundary, the Antarctica-Nubia boundary, and the Nubia-North America boundary) used in plate reconstructions to estimate motion between the Pacific and North American plates. The sum of relative plate motions can be compared with geodetic estimates of Pacific-North America motion. This circuit fails closure with a misfit of 5 ± 3 mm/yr (DeMets et al., 2010).

There are several possible explanations for the non-closure of plate circuits; e.g., motion of tectonic plates over the non-spherical Earth causes deformation of the plates (McKenzie, 1972; Turcotte, 1974). The explanation we explore here, however, is that some or all of the circuit misfits are caused by horizontal thermal contraction of the lithosphere. That such contraction occurs is assumed in the work of many who have studied the thermal evolution of the lithosphere and thermo-elastic stresses (e.g., Parmentier and Haxby, 1986; Sandwell and Fialko, 2004). The focus in most of these prior studies has been the difference in horizontal contraction between the upper and lower portions of the competent oceanic lithosphere (i.e., the lithosphere above the brittle-plastic transition (BPT)) and the bending moments applied to the lithosphere and the resulting flexure. Here we investigate the effect of vertically averaged horizontal thermal contraction in young oceanic lithosphere on plate-scale deformation by modeling the associated strain rate field. We focus on the Pacific plate.

HORIZONTAL THERMAL CONTRACTION

Kumar and Gordon (2009) presented the assumptions and methods used to determine the strain rate in cooling lithosphere due to thermal contraction. In one dimension, infinitesimal strain rate $(\varepsilon\,())$ due to a cooling rate $(\partial T/\partial t)$ is $=\alpha_l(\partial T/\partial t)\varepsilon$ where *T* is temperature, *t* is time, and α_l is the linear coefficient of thermal expansion, which has a value of ≈10^{-5} K^{-1}. We assume that the 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. We assume that stresses are relaxed by solid-state flow below the BPT. Kumar and Gordon (2009) showed that the depth-averaged cooling rate for a column of competent lithosphere to equal $T_m[\exp(-A^2)-1]/2At\sqrt{\pi}$ where $A=erf^{-1}(T_l/T_m)$, T_l is the temperature of the isotherm bounding the base of the competent lithosphere, and T_m is the initial temperature in a half-space cooling model (Turcotte and Oxburgh, 1967). Estimates of T_m range from 1300 °C to 1350 °C (Hillier and Watts, 2005; McKenzie et al., 2005). Following Kumar and Gordon (2009), we choose the lower value for T_m as is it leads to lower contraction rates than the alternative and thus we consider it to the be the more conservative option. McKenzie et al. (2005) and Wiens and Stein (1983) showed that the limiting depth of earthquakes approximately follows 600 °C. Martinod and Molnar (1995) assumed the BPT to occur at a depth ≈20% greater than that of the limiting depth of earthquakes. We therefore set T_l to 700 °C . This temperature also gives the best fit to the amplitude of geoid anomalies due to thermal stresses on profiles across fracture zones (Parmentier and Haxby, 1986). With the chosen values for T_l , T_m , and α_l , we find $\dot{\varepsilon} \approx -167.24 \cdot 10^{-5}/t$. We do not consider alternative cooling models as those differ only from the half-space cooling model for ages >70 Ma where the expected strain rates will be an order-of-magnitude lower than in the youngest lithosphere. We take $t = t' + t_0$ with t' being the age of the lithosphere and t_0 being a parameter that accounts for the nonzero thickness of lithosphere along the axis of mid-ocean ridge segments. We assume t_0 equals 0.1 Ma, which corresponds to a zero-age lithospheric thickness of ≈2 km (Kumar and Gordon, 2009). Kumar and Gordon (2009) discussed the strain rates and displacement rates expected from horizontal thermal contraction in a general way. Here we make explicit predictions for the Pacific plate.

RESULTS

We use the sea-floor age model (v. 3.6) of Müller et al. (1997) to estimate strain rates in grid cells 0.2° by 0.25° in longitude and latitude, respectively. Because of the $1/t$ dependency of strain rate, we estimate the average age in each grid cell as the inverse average of the reciprocal age values in that cell. The total model of 197,343 cells spans the entire Pacific plate, with its boundaries defined by ridge-transform segments in the east and south (C. DeMets, pers. comm., 2010) and predominantly subduction zone trenches in the west and north (Bird, 2003). We set the age of areas with unknown age to 200 Ma.

We apply the method of Haines and Holt (1993) to determine a continuous model of strain rates and associated velocities that best fits the expected age-based strain rate. For each grid cell the strain rates, and corresponding *a priori* covariance matrix, are set according to biaxial, age-derived, shortening both parallel and perpendicular to the local age gradient. If we simply allowed the cells to shrink without enforcing strain compatibility, gaps would open up everywhere in our model with the largest gaps opening up in the youngest lithosphere. If we forced all the gaps to close in young lithosphere, the result would be overlap in old lithosphere. Thus a sensible compromise is required for balancing extensional strains and contractional strains while enforcing strain compatibility along the boundaries of all the cells.

We consider two end-member cases in which the continuous model should best fit the expected strain rates. In Model 1, the *a priori* covariance matrix is set such that the modeled continuous area strain rate amplitudes best match the expected values for each grid cell, but the shortening rates parallel and perpendicular to the age gradient are not required to be equal. In Model 2, the *a priori* covariance matrix is set such that the model continuous tensor field best satisfies the expectation of isotropic shortening, but the area strain rates do not fit as well as in Model 1. In effect, the difference between the models is on how we set the a priori correlation between the two strain axes (0 for Model 1 and 1 for Model 2). The strain rates for Models 1 and 2 are broadly similar but differ in detail (Fig. 1A and B, respectively). The deviations from the level of expected area strain rates and the isotropy of shortening are shown in the Electronic Supplement. For both models, the straining of the Pacific plate is dominated by horizontal thermal contraction but also include regions of extensional strain evidently required by strain compatibility and the other assumptions of our models (Figs. 1A and 1B). The regions of extension tend to be in older lithosphere and might be much smaller or disappear altogether if we assumed that lithospheric strength increased with age.

The associated velocity field results from the spatial integration of strain rates across the Pacific lithosphere. In young lithosphere the ridge-parallel component of the velocity field is potentially larger than the ridge-normal component, because there is more lithosphere with relatively large strain rate across which to integrate. The ridge-parallel component is, however, controlled by offsets of ridge segments by transform faults, limiting the velocity. For Model 2, Figure 1C shows the velocity field in a reference frame that minimizes the predicted velocities at a set of GPS stations in old Pacific lithosphere and Figure 1D shows the velocity field in a reference frame that minimizes the velocities of points on the 0.78 Ma isochron (Müller et al., 1997) along the Pacific-Antarctic Ridge. The velocity fields for Model 1 are shown in the GSA Data Repository and are fairly similar to those for Model 2. To test whether the thermal contraction can be observed geodetically, we also analyze data from available continuous GPS stations (see GSA Data Repository). Residual horizontal GPS velocities (after subtraction of a rigid-body rotation) are shown in Figure 1C.

DISCUSSION

Most GPS residual velocities differ insignificantly from zero at the 95% confidence level after we solve for and remove a rigid-body rotation (Fig. 1C). The reduced χ^2 is 2.4. We did not use the velocity for the stations at Guadalupe and Chatham Islands to calculate the rigid-body rotation. The station on Guadalupe Island is one of the few sites that has a considerable, but insignificant (95% C.L.), motion of 1.2 ± 1.3 mm/yr toward S25°E. Previous estimates for this station were larger and significant (Plattner et al., 2007; Argus et al., 2010). Either way, that station's motion cannot be explained by thermal contraction, which is 0.2–0.3 mm/yr toward SW at this location. For the area containing the Chatham Rise and Campbell Plateau, SE of New Zealand, our model predicts the largest velocities relative to the GPS locations on the old lithosphere. The 0.6 mm/yr eastward and 0.6 mm/yr southward motion for stations CHAT and CHTI on Chatham Island are roughly consistent with our predictions, but with only the former observation being significant. At Campbell Island (i.e., southwestern most Pacific island) we predict 1.3–1.7 mm/yr to the SE, inconsistent with a campaign velocity there of 3.7 mm/yr toward the NW (Beavan et al., 2002). If we transform that study's velocity field onto ours, however, the observed motion of Campbell Island becomes insignificant. A real test of whether our prediction is correct requires the installation of a CGPS station on Campbell Island (and other nearby islands) or seafloor geodesy with 1 mm/yr accuracy.

We also present the predicted velocity field relative to the 0.78 Ma isochron along the Pacific-Antarctic Ridge (Fig. 1D) to see how thermal contraction of the Pacific plate modifies the results found assuming plate rigidity in the Pacific-Antarctic-Nubia-North America plate motion circuit. In this approximate reference frame, places offshore Baja California move 1.7 (Model 1) to 2.2 (Model 2) mm/yr toward the SE. The GPS station at Guadalupe Island could be used to test for that motion, but only if we had several GPS stations on the southernmost Pacific plate to provide a geodetic reference. Unfortunately, there are nearly no islands there. Although the up-to-2.2 mm/yr motion off western North America may help to explain the failure to close the Pacific-Antarctic-Nubia-North America circuit by 5 ± 3 mm/yr (DeMets et al., 2010), the thermal contraction in the Antarctic, Nubia, and North American plates needs to be modeled to address the problem fully. In any case, it is unlikely that thermal contraction can entirely explain a non-closure as large as the 14 ± 5 mm/yr observed for the Pacific-Nazca-Cocos circuit (DeMets et al., 2010), suggesting that additional deformation processes are involved in that three-plate circuit.

Is there other evidence that thermal contraction exists and controls the Pacific plate's intraplate strain rate field? To address this, we convert predicted strain rates to "tectonic" moment rates (which are proportional to strain rate times volume) and compare them with earthquake distributions. The number of events should be proportional to the tectonic moment rate if seismicity follows a Gutenberg-Richter distribution and the b-value and maximum expected moment are the same everywhere (e.g., Molnar, 1979). We use a catalog of 39 intraplate events and explain the catalog in the GSA Data Repository.

We order our model grid cells from oldest to youngest age and count the cumulative number of events, until 5 Ma (we exclude younger lithosphere to avoid possibly counting ridge-transform events as being intraplate). The normalized graph is shown in Figure 2A. The same figure also shows for Model 1 and 2 the normalized cumulative moment rate (M_0) for the same cell ordering. For the strain-to-moment conversion we assume that the seismogenic thickness follows the 700ºC isotherm. We also calculate the moment rates for alternative conversions in which the maximum seismogenic thickness is limited to 40 km. Our results show that all curves have the same concave-up shape. A general decrease in seismicity with age was already observed by Wiens and Stein (1983) and the similarity in the curves suggest that horizontal thermal contraction provides a first-order control on this decrease. Because of the limited number of available earthquakes, it is not possible to prefer one particular moment rate model over another. From Figure 1A it is clear that there is a change in seismicity rate around 30 Ma. If a correlation between earthquake numbers and M_0 exists, we would expect it to be most obvious in the younger lithosphere. We therefore also show a similar comparison but only for ages between 5 and 30 Ma (Fig. 2B), which contains 21 events. For this case both cumulative earthquakes and predicted M_0 form a similar linear relationship with age.

The correlation we observe between tectonic moment rate and number of events has also been found for most subduction zones and continental deformation zones (Kreemer et al., 2002; Kagan, 1999). These correlations do not, however, indicate the possibility that some, but not all, of the tectonic moment could be released aseismically. If all tectonic moment is released seismically, an unrealistically high maximum magnitude of >9 is needed to explain the earthquake rate and observed b-value (Okal and Sweet, 2007); thus most of the thermal strain rate in the Pacific is probably released aseismically. Moreover, it is expected that the moment release mainly reflects the differential rates of cooling with depth (Wessel, 1992) and not the vertically averaged cooling we consider here.

CONCLUSIONS

Strain rates from horizontal thermal contraction are much smaller than those occurring in narrow plate boundaries and generally below the level of direct detection with GPS measurements. When integrated across the plate away from the Pacific-Antarctic Ridge, however, velocities can become 2.2 mm/yr off the coast of western North America and can contribute significantly toward observed non-closure of the Pacific-North America plate circuit. Moreover, the correlation between the distribution of earthquakes and predicted strain or moment rate supports that thermal strain rates are real and released seismically, at least partly. We conclude that horizontal thermal contraction should be considered in future development of the plate tectonic theory.

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FIGURE CAPTIONS

Figure 1. A) Contours of log second invariant of model strain rate tensor for Model 1 for which areal strain rates best match the expected values. Linear strain rates saturate at $10^{-11.5}$ yr^{-1}. White circles are epicenters of selected Pacific intraplate earthquakes from 1960 to 2008 (Engdahl et al., 1998) (see Electronic Supplement). Orthogonal bars indicate the orientations of principal strain rate directions with the lengths of the bars indicating the relative magnitudes of the two horizontal principal strain rates, black if contractional and white if extensional. B) same as A but for Model 2 for which the strain rates best satisfy isotropy. C) Residual GPS velocities (white vectors), after removal of rigid-body rotation, with 95% confidence error ellipses. Black vectors and contours are predicted velocities from thermal contraction (Model 2) are shown in the same frame as GPS, D) Predicted velocities (vectors and contours) from thermal contraction (Model 2) relative to the 0.78 Ma isochron along the Pacific-Antarctic Ridge.

Figure 2. A) Normalized cumulative number of earthquakes from oldest to youngest (i.e., 5 Ma old) lithosphere. Also shown are normalized cumulative moment rate for both end-member models. Moment rate is defined as: $2\mu V\left[\left|(\dot{\varepsilon}_{xx}+\dot{\varepsilon}_{yy})/2\right|+\sqrt{\left[(\dot{\varepsilon}_{xx}-\dot{\varepsilon}_{yy})/2\right]^2+\dot{\varepsilon}_{xy}^{2}}\right]$, with the shear modulus, μ, set to 30 GPa, and volume, *V*, is the area times a thickness set to the depth of the 700ºC isotherm (or is alternatively limited to be no more 40 km.). B) same as A, but only for lithosphere between 5 and 30 Ma. old.

[1] GSA Data Repository item 2014xxx, [Details on GPS data analysis and earthquake catalog], is available online at www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.