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Macroscopic strength of oceanic lithosphere revealed by ubiquitous fracture-zone instabilities

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ABSTRACT

The origin of plate tectonics is one of the most fundamental issues in earth and planetary sciences. Laboratory experiments indicate that the viscosity of silicate rocks is so strongly temperature-dependent that the entire surface of the Earth should be one immobile rigid plate. The rheology of oceanic lithosphere is, however, still poorly understood, and there exist few constraints on the temperature dependency of viscosity on the field scale. Here we report a new kind of observational constraint based on the geoid along oceanic fracture zones. We identify a large number of conspicuous small-scale geoid anomalies, which cannot be explained by the standard evolution model of oceanic lithosphere, and estimate their source density perturbations using a new Bayesian inversion method. Our results suggest that they are caused most likely by small-scale convection involving temperature perturbations of $\sim$300 K $\pm$ 100 K. Such thermal contrast requires the activation energy of mantle viscosity to be as low as 100 ± 50 kJ mol^{-1} in case of diffusion creep, and 225 ± 112 kJ mol^{-1} in case of dislocation creep, substantially reducing the thickness of the stiffest part of oceanic lithosphere. Oceanic lithosphere may thus be broken and bent much more easily than previously thought, facilitating the operation of plate tectonics.

1. Introduction

The rheology of oceanic lithosphere plays a major role in the operation of plate tectonics, a phenomenon only observed on our planet and still considered enigmatic (Bercovici et al., 2000; Tackley, 2000; Schubert et al., 2001). Oceanic lithosphere, which is the top boundary layer of mantle convection, is generally thought to be very stiff, and thus difficult to break or bend. The Earth's mantle is composed of silicate rocks, whose viscosity is strongly temperature-dependent (Karato and Wu, 1993; Hirth and Kohlstedt, 2003). This dependency is quantified by the activation energy, which is usually estimated to range from 240–375 kJ mol^{-1} (diffusion creep) to 470–510 kJ mol^{-1} (dislocation creep) for the upper mantle in wet and dry conditions respectively, based on the deformation experiments of olivine aggregates (Mei and Kohlstedt, 2000a, 2000b; Karato and Jung, 2003; Hirth and Kohlstedt, 2003). With this value, the bulk of the oceanic lithosphere is too stiff to become convectively unstable, prohibiting the initiation of subduction (Solomatov, 1995). In order to generate plate tectonics, therefore, some additional mechanism is required to compensate temperature-dependent viscosity, but what this mechanism could be is still unresolved (Bercovici, 2003; Korenaga, 2013). So far proposed mechanisms include the feedback between shear localization and grain size evolution (Kameyama et al., 1997; Braun et al., 1999; Landuyt et al., 2008), the pre-existing zone of weakness such an oceanic fracture zone (Toth and Gurnis, 1998; Hall et al., 2003; Gurnis et al., 2004), the higher water content of oceanic lithosphere (e.g. Regenauer-Lieb et al., 2001), hydration by thermal cracking (Korenaga, 2007), and rheological weakening by a secondary orthopyroxene phase (Farla et al., 2013).

Observational constraints on the temperature dependency of viscosity have been difficult to establish. The geodynamic study of seamount loading history suggests the activation energy of 120 kJ mol^{-1} for diffusion creep (Watts and Zhong, 2000), but this is usually thought to represent the temperature dependency of the Peierls mechanism (Goetze and Evans, 1979), not that of high-temperature creep. The Peierls mechanism can operate only under low temperatures and very high stresses (>100 MPa), so it is not relevant to the destabilization of oceanic lithosphere by low convective stresses, which is often believed to be critical for

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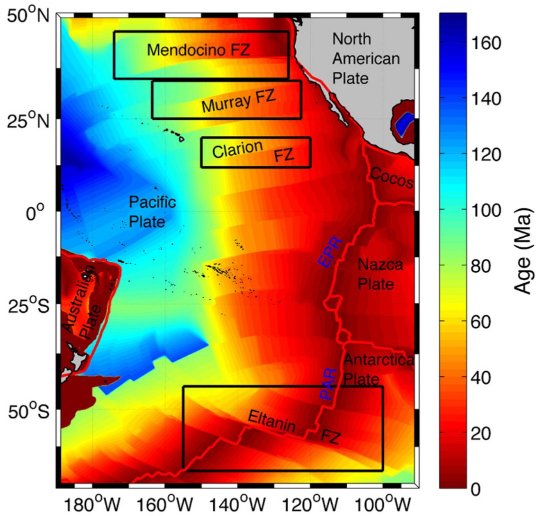

Fig. 1. Age map of the Pacific ocean floor (Müller et al., 2008). Black boxes represent the study areas and red lines denote plate boundaries. EPR: East Pacific Rise, PAR: Pacific-Antarctic Rise and FZ: Fracture Zone. (For interpretation of the references to color in this figure, the reader is referred to the web version of this article.)

the initiation of subduction (Solomatov, 2004). The seismically derived thermal structure of the Pacific lithosphere, which deviates from the standard cooling model, may also constrain the activation energy (Ritzwoller et al., 2004). The lithospheric seismic structure can be explained with thermo-mechanical erosion by convective instabilities with either diffusion creep with the activation energy of $\sim$120 kJ mol^{-1} or dislocation creep with the activation energy of 360–540 kJ mol^{-1} (van Hunen et al., 2005). As pointed out by Korenaga and Korenaga (2008), however, the old Pacific seafloor under which lithospheric thinning is inferred, is heavily populated with hotspot islands and oceanic plateaus, and it is not clear whether the thinning is due to intrinsic convective instabilities. Here we show that a new kind of observational constraint on the activation energy of high-temperature creep can be extracted from the subtle signatures of the geoid along oceanic fracture zones. A recent study of regional geoid anomalies along the Mendocino fracture zone indicates the occurrence of small-scale convective instabilities (Cadio and Korenaga, 2014). Because the amplitude of involved temperature variations can be related to the activation energy, we analyze four major fracture zones in the Pacific (Fig. 1) and provide a new observational constraint on activation energy, by establishing the characteristic amplitude of temperature variations associated with small-scale convection.

2. Data analysis and inversion

Oceanic lithosphere gradually cools and thickens as seafloor ages (Turcotte and Schubert, 2002). Consequently, a fracture zone, which juxtaposes two lithospheric segments of different ages (and thus of different thicknesses), produces lateral density variations detectable in geoid signal (Crough, 1979). Such variations in the thermal structure of oceanic lithosphere could also give rise to convective instabilities at its base (Huang et al., 2003; Dumoulin et al., 2008). This geodynamic process yields an additional geoid component, based on which we can constrain the nature of instabilities. To isolate this additional convective component, we first estimate the geoid contribution of lithospheric cooling from the half-space cooling model (HSC), which linearly relates the thickness of oceanic lithosphere with the square root of seafloor age (Turcotte and Oxburgh, 1967). To that aim, we use the 3D numerical model of oceanic lithosphere developed by Cadio and Korenaga (2012) in order to take into account both vertical and lateral density variations across the fracture zone. An analytical solution for a theoretical geoid exists for the half-space cooling (HSC) model (Haxby and Turcotte, 1978), but it ignores any lateral density perturbations, which is inappropriate especially when considering geoid signals around a fracture zone. Our model of lithosphere, assumed in local isostatic equilibrium, is composed of an array of constant density prisms, for which the density and location are coupled to temperature variations predicted by the HSC model. The values of thermal parameters used in the theoretical calculation are given by Cadio and Korenaga (2012). A geoid signal in every surface point of the model space is calculated by adding contributions from all individual prisms. Among standard evolution models, the HSC model predicts the greatest geoid contribution for old seafloor (Haxby and Turcotte, 1978) and thus gives us the minimal amplitude of residual signals after correction for cooling. Consequently, density contrasts and thermal variations derived from our study can be seen as lower bounds.

The identification of the cooling component in the total geoid is considerably improved here by using the continuous wavelet transform (Cadio and Korenaga, 2014). Such analysis provides a detailed description of signals both in spatial and spectral domains, and thus highlights the signal components at each scale and position. We use spherical Poisson multipole wavelets that are particularly well suited to analyze potential fields (Holschneider et al., 2003). The simultaneous analysis of the EGM2008 geoid (Pavlis et al., 2008) and theoretical geoid in the wavelet domain, at scales varying from 100 to 500 km, shows that the characteristic scale of cooling process in the vicinity of fracture zones is $\sim$100 km. This estimate is relevant because the expected depths of density sources range from the surface to about 100–150 km, which corresponds to the depth extent of oceanic lithosphere.

We apply this approach to four major fracture zones in the Pacific: Mendocino, Clarion, Murray, and Eltanin (Fig. 1). Because of their large age offsets, they display significant geoid signals and are optimal place to initiate convective instabilities. The location and the geometry of these fracture zones also allow us to identify a large number of residual geoid anomalies, covering seafloor as old as 100 Ma (Figs. 1–2 and Figs. S1–S3). We focus only on anomalies localized at 100 km scale, and those satisfying the following two criteria (Cadio and Korenaga, 2014): (1) they are not correlated with any topographic structure (Figs. 2 and S1–S3) and (2) their spectral content indicates an extremum at 100 km scale, ensuring that this spatial scale is characteristic of a signal under consideration (Fig. S4). With this screening, localized residual geoid anomalies may safely be regarded to originate in perturbations to the density structure of normal oceanic lithosphere, not in dynamic or flexural topography. If the lithosphere is treated as purely elastic and the fracture zone behaves as a locked fault, for example, differential subsidence across the fracture zone could indeed deflect the lithosphere, producing a ridge on the younger side and a trough on the older side (Sandwell and Schubert, 1982; Sandwell, 1984). Flexural deformation could contribute to geoid anomalies, but such anomalies would be correlated with flexural topography. Furthermore, the residual geoid anomalies are corrected for locally compensated residual topography assuming Airy compensation (Haxby and Turcotte, 1978), so the crustal contribution to the residual anomalies is minimized. After subtracting this isostatic geoid, we invert the residual anomalies for the statistical distribution of their source density perturbations.

Our inversion method is based on Bayesian statistics and is implemented by combining forward modeling with Markov chain Monte Carlo sampling (Cadio and Korenaga, 2014). We approximate each density perturbation as a right rectangular prism and calculate its geoid signature in a static Earth's mantle by solving the Poisson's equation (e.g. Nagy et al., 2000). We do not account the geoid contribution of the surface topographic deformation potentially induced by the presence of such density per-