Strength of fracture zones from their bathymetric and gravitational evolution

Chad E. Hall and Michael Gurnis
Seismological Laboratory, California Institute of Technology, Pasadena, California, USA

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[1] Fracture zone evolution is investigated using dynamic models that allow the fault zones to freely slip. This is an improvement over past formulations where bathymetric offsets were imposed kinematically. The models use a viscoelastoplastic rheology that incorporates the influence of fault friction on fracture zone slip history. Using viscoelastic plates, we assess the role of small-scale convection on removal of the lowermost thermal lithosphere beneath fracture zones. Through a comparison of synthetic gravity to free-air gravity across fracture zones we find that the amplitude of the gravity jump across fracture zones is best fit by models with weak faults that have depth-averaged yield strengths <10 MPa. Fracture zones with such low strengths can convert to subduction zones with ∼100 km of convergence. Many fracture zones do not fit plate subsidence models with locked or slipping faults but are better fit by systems that are tectonically deformed by modest amounts of extension.

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1. Introduction

[2] Fracture zones, common features of the oceanic lithosphere that bound seafloor of discontinuous age, are the intraplate extension of transform faults where active strike-slip motion occurs. After transform faults evolve into fracture zones, the contact separating discontinuous lithosphere may "heal" or strengthen [*Wessel and Haxby*, 1990], so that fracture zones may not be weaker than normal oceanic lithosphere [*Haxby and Parmentier*, 1988]. In contrast, tectonic uplift [*Bonatti*, 1978] and volcanic activity [*Lowrie et al.*, 1986] localized at fracture zones suggest that they are zones of weakness that may facilitate changes in plate motions during stress reorientation. Moreover, the Izu-Bonin-Mariana subduction zone [*Uyeda and Ben-Avraham*, 1972], the Puysegar-Fiordland subduction zone [*Collot et al.*, 1995], and the Hjort Trench [*Meckel et al.*, 2003] may have each initiated along an old transform fault or fracture zone. Determining the strength of fracture zones is critical for developing quantitative tests of such hypotheses.

[3] A leading argument for strong fracture zones comes from the preservation of scarp offsets formed when a transform fault first transitions to a fracture zone. At such "zero-age" fracture zones, isostatic equilibrium produces a large bathymetric step between newly formed lithosphere and the older side that has already subsided. As the lithospheric segments separated by the fracture zone cool, there are two end-member scenarios for local bathymetric evolution. If a fracture zone is sufficiently weak, slip readily occurs, the lithospheric segments act as uncoupled blocks that subside isostatically, and the bathymetric step decays. Alternatively, if a fracture zone is sufficiently strong, there is no slip at depth below the fracture, and the bathymetric step remains. However, away from the fracture zone, plate subsidence continues unimpeded, with the older, colder plate subsiding more slowly than the younger plate. Because a strong, locked fracture zone resists the differential subsidence which would act to erase the bathymetric step, the oceanic lithosphere elastically flexes [*Sandwell and Schubert*, 1982; *Sandwell*, 1984], so that the younger side is deflected upward, and the older side is locally depressed. The elastic thickness of oceanic lithosphere is temperature- and thus age-dependent, with this model predicting that the older-side troughs should be broader in comparison to the sharper, short-wavelength peaks on the younger side.

[4] All previous models of fracture zone evolution assume a kinematic boundary condition at the lithospheric age offset that implicitly fixes the topographic step at its initial value [*Sandwell and Schubert*, 1982; *Sandwell*, 1984]. In contrast, we study dynamic models in which the fracture zone at depth is free to slip, and we study a suite of cases in which the material properties of the fracture zone and oceanic lithosphere are varied. We show that oceanic lithosphere can be locally weakened at a fracture zone and can still preserve long-lived bathymetric steps that are in agreement with observed profiles of bathymetry and gravity across fracture zones. Furthermore, we expand on previous work by using models which transition from elastoplastic to viscous material behavior, allowing us to explore whether edge-driven convection may influence fracture zone evolution. Finally, we show that many of the previously unexplained

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features of fracture zones are reproduced if small amounts of extension or compression are applied.

2. Modeling Fracture Zone Evolution

[5] Our models are two-dimensional cross sections of fracture zones (Figure 1), beginning with an initial state immediately after a transform fault has transitioned to a fracture zone that is no longer offset by strike-slip motion. Initial temperature conditions are given by solutions for conductively cooled half-spaces. Three-dimensional (3-D) temperature perturbations caused by heat transport along the transform offset are not included in the initial condition, but 3-D effects do not generate substantial deviations in surface topography or create a net downward warping near a fracture zone [*Phipps Morgan and Forsyth*, 1988]. We ignore the influence of shear heating due to transform fault slip on the initial thermal field, except for one sensitivity test described in section 3. The mesh is perturbed near the free surface, so that initial topography is determined by isostatic equilibrium. The influence of an overlying water layer on isostasy is included after the calculation by scaling the model topography by $\rho_m/(\rho_m - \rho_w)$. Initially, a weakened fault zone 7 km wide by 16 km deep is placed between the two plates (Figure 1). The side and bottom boundaries have zero normal velocities and are stress-free while the top boundary is a free surface.

[6] The initial bathymetric condition used here and previously by *Sandwell and Schubert* [1982] and *Sandwell* [1984] has several implicit assumptions. First, the active transform fault is assumed to be weak such that there is no transmission of stress across it; otherwise, flexure from differential subsidence would have deflected the older plate while it bordered the active transform fault. Second, as we do not model the older plate's history before it reached the ridge transform intersection, we cannot account for flexure due to thermal bending moments during this earlier evolution [*Parmentier and Haxby*, 1986; *Haxby and Parmentier*, 1988]. Assuming that no stress transmission occurs across