DATA REPOSITORY ITEM 2008111

NUMERICAL MODELING METHODS

Deformation and thermal evolution were modeled in a 2-D visco-elastic-plastic layer using the fast Lagrangian analysis of continua (FLAC) technique (Cundall, 1989). This explicit hybrid finite element–finite difference approach has been used to simulate localized deformation (i.e., faulting) in a variety of extensional environments (Hassani and Chéry, 1996; Poliakov and Buck, 1998; Lavier et al., 2000; Behn et al., 2006) and is described in detail by Poliakov (1993) and Lavier and Buck (2002). Material behavior is a function of temperature, strain-rate and accumulated plastic strain throughout the model space. In regions where deformation is visco-elastic, the material behaves as a Maxwell solid. Viscous deformation is incompressible and follows a non-Newtonian temperature- and strain-rate dependent power law assuming a dry diabase rheology (Mackwell et al., 1998). Plastic yielding is controlled by Mohr-Coulomb theory, where cohesion is a function of the total accumulated plastic strain (Poliakov and Buck, 1998; Lavier et al., 2000). In our models, cohesion decreases linearly from 44 MPa to 4 MPa when a critical fault offset of 500 m is reached. This critical fault offset has been shown to produce large-offset faults in thin lithosphere under amagmatic conditions (Lavier et al., 2000), although that modeling did not consider advective heat transport as described below. Following Poliakov and Buck (1998), we also include an annealing time in our calculations in which plastic strain decays over 10^{12} s. This annealing time reduces the broadening of fault zones due to numerical diffusion caused by regridding.

Magma injection is imposed by kinematically widening a vertical column of elements in an injection zone at the center of the model space (Buck et al., 2005; Behn et al., 2006). The rate of injection is described by the parameter M, which is defined as the ratio of the rate of injection-zone opening to the rate of far-field extension. The injection zone extends from the surface of the model space to a depth of 6 km. Here we envision that magma is supplied to the injection zone by along-axis transport from the center of the ridge axis. Thus, we assume that the injection zone remains fixed in the across-axis direction even though the thermal structure may migrate relative to its position.

Previous studies have used models with fixed thermal structure to investigate extensional faulting, with or without the mechanical effects of magma injection (Poliakov and Buck, 1998; Lavier et al., 2000; Buck et al., 2005; Behn et al., 2006). In contrast, we explicitly model temperature evolution associated with mantle flow and magma injection. This approach allows the axial lithospheric structure to adjust to the imposed spreading rate, rate of magma injection, and the resulting deformation field. This is particularly important for simulating the growth of large-offset faults, in which the axial thermal structure becomes highly asymmetric due to the advection of warm material into the footwall beneath the active fault.

Temperature evolution is modeled using a Lagrangian formulation, in which the advective component of heat transport follows the deforming grid. At each time step we then use explicit finite differences to solve the heat equation (Lavier and Buck, 2002). Heat is added to the ridge axis during magma emplacement due to the injection temperature of the magma and the latent heat of crystallization (Sleep, 1975; Phipps Morgan and Chen, 1993). The effects of hydrothermal circulation on temperature are simulated by increasing the thermal conductivity by a factor (Nusselt number, *Nu*) above a threshold depth of 7 km where T is also < 600ºC (e.g., Phipps Morgan et al., 1987). In the models presented here, *Nu*=8, which results in a brittle layer thickness of ~5 km, consistent with average maximum depth of seismicity at the Mid-Atlantic Ridge (Kong et al., 1992; Wolfe et al., 1995; Barclay et al., 2001).

The numerical domain is 60 km wide by 20 km deep, with a maximum grid resolution of 0.25 km x 0.25 km at the ridge axis, which gradually coarsens to 2 km with distance from the ridge. The decrease in grid resolution off-axis results in smoothing of the topography due to numerical diffusion during regridding (see below). Deformation is driven by applying a uniform rate of far-field extension along the sides of the model space corresponding to a full spreading rate of 50 mm/yr (i.e., at a rate near the transition from slow to intermediate spreading rates). A hydrostatic boundary condition is assumed for the base of the model space and the top boundary is stress-free. The top of the model space is set to 0ºC, while the bottom is defined to follow an error function with an initial temperature of 1300ºC at time t = 0. This bottom boundary condition allows us to use a relatively thin numerical domain while still modeling temperature accurately throughout the model space. All models except M=0 begin with laterally uniform temperature structure. For M=0 (Video DR6), the model starts with slightly elevated temperatures at the center of the model so as to focus strain in the middle of the model domain; otherwise strain would concentrate at the model boundaries.

The FLAC method employs an explicit time-marching scheme, in which the time step is set to be the minimum of the Maxwell relaxation time ($2\eta/E$) and the time required for an elastic P-wave to propagate across a distance equal to the local grid spacing. Because of the high grid resolution used in our calculations, the elastic propagation time would result in extremely short time steps and large computational times. To circumvent this problem we employ an adaptive density scaling method (Cundall, 1982). This approach assumes that in situations where the inertial forces are small, the inertial density and hence the time step can be increased. We also chose a ratio of the imposed boundary velocity to the P-wave velocity of 5 x 10^{-5} (Lavier et al., 2000), resulting in a time step of 1–5 years.

Regridding is necessary to overcome problems due to the degradation of numerical accuracy when elements become highly distorted due to either faulting or intrusion. The initial mesh consists of quadrilateral elements that are subdivided into triangles. Regridding occurs when the minimum angle in any of the triangular elements drops below 5º. During remeshing, strains are transferred from the old (deformed) to the new (undeformed) grid using linear interpolation (Lavier and Buck, 2002). This interpolation results in out-of-balance forces at the nodes, producing artificial accelerations that decay over several hundred time steps; this is observed in intermittent 'flash frames' in videos of model extension (e.g., near the 1 m.y time-step for M=0.5, see Video DR2). To dampen these artificial accelerations more rapidly, we decrease the time step by an order of magnitude immediately following a remeshing event and then increase it linearly to its original value over 1000 time steps.

Videos DR1 to DR6 show QuickTime movies for a total of 1.5 m.y. of extension for models with M values ranging from 0.7 down to 0.

ANALYSIS OF MEGAMULLIONS

We analyzed multibeam bathymetric data from the RIDGE Multibeam database (http://www.marine-geo.org/rmbs/) and from the published literature. Megamullions formed by long-lived detachment faults were identified on the basis of their characteristic morphology (domed shape and large-scale corrugations parallel to spreading direction). Where identification was uncertain, data are represented as open circles connected by dashed vertical lines in the plots of Fig. 3. Megamullion frequency was determined for spreading segments that are defined by first- and second-order discontinuities (transform faults and non-transform discontinuities with offsets greater than a few kilometers, respectively). Where available, data from both ridge flanks were used for each segment. In a few cases where off-axis traces of non-transform discontinuities were ambiguous, or where only minor discontinuities were present at fast-spreading ridges, determinations were made on combined segments.

For each spreading segment, we determined average axial depth along-axis between the centers of bounding transform or non-transform discontinuities. Ideally, instead of using axial depth, we would have used crustal thickness modeled from RMBA gravity to estimate the state of magma supply at the axis of each ridge segment. Unfortunately, only a few of the segments examined have gravity data available, and where such data have been published, variations in the assumed modeling parameters preclude consistent comparisons among segments.

Potential sources of error in the megamullion frequency plots include the following. First, the full lengths of spreading segments were not always covered by multibeam bathymetry. In these cases the available data were used if they covered a substantial portion of the segment and appeared to be representative of the segment as a whole; we consider this not to be a significant source of error. Second, some multibeam survey areas cover only a few thousand km^2, so frequency values could change significantly with the addition or deletion of one or two megamullions; also, most of the highest frequency values are associated with survey areas less than ~4500 km^2. Thus, details of frequency values may not be significant, although we judge the overall pattern to be robust. Finally, because we have no way to establish former axial depth for any megamullion that is now off-axis, we assume that conditions of tectonic vs. magmatic extension have not changed within the spreading segment and that the average depth at the present spreading axis is representative of the time when the megamullion formed; in all of these instances, values plot well within the axial-depth range of near- and on-axis megamullions, so this assumption appears not to introduce any significant error.

It is notable that all data plotting at average axial depths greater than 5000 m in Fig. 3A are for the abandoned spreading ridge in the Parece Vela backarc basin. These data were corrected for thermal subsidence, based on an age of 13 Ma at the abandoned rift axis (Ohara et al., 2001). Because the data available for conditions of very low magma supply (axial depths >5000 m) are so limited, there may be important aspects of megamullion formation in this range that presently are undetected.

VIDEO CAPTIONS

Video DR1 - QuickTime movie showing 1.5 m.y. of extension for M = 0.7 and a full spreading rate of 50 mm/yr. Panels show plastic strain (top), $\log_{10}$ strain-rate (middle), and temperature (bottom). High-strain zones simulate faults. Arrows in the middle panel show velocity at each time step (note that grid spacing is significantly finer than spacing of velocity vectors). The black line in the bottom panel marks the 600° isotherm (approximate brittle/plastic transition). The model is presented with no vertical exaggeration.

Video DR2 - QuickTime movie for M = 0.5. See Video DR1 caption for full description.

Video DR3 - QuickTime movie for M = 0.4. See Video DR1 caption for full description.

Video DR4 - QuickTime movie for M = 0.3. See Video DR1 caption for full description.

Video DR5 - QuickTime movie for M = 0.2. See Video DR1 caption for full description.

Video DR6 - QuickTime movie for M = 0. See Video DR1 caption for full description.

REFERENCES CITED

Barclay, A.H., Toomey, D.R., and Solomon, S.C., 2001, Microearthquake characteristics and crustal Vp/Vs structure at the Mid-Atlantic Ridge, 35° N: Journal of Geophysical Research, v. 106, p. 2017-2034.

Behn, M.D., Buck, W.R., and Sacks, I.S., 2006, Topographic controls on dike injection in volcanic rift zones: Earth and Planetary Science Letters, v. 246, p. 188-196, doi:10.1016/j.epsl.2006.04.005.

Buck, W.R., Lavier, L.L., and Poliakov, A.N.B., 2005, Modes of faulting at mid-ocean ridges: Nature, v. 434, p. 719-723, doi:10.1038/nature03358.

Cundall, P.A., 1982, Adaptive density-scaling for time-explicit calculations, 4th International Conference on Numerical Methods in Geomechanics, Edmonton, Alberta, Canada, v. 1, p. 23-26.

Cundall, P.A., 1989, Numerical experiments on localization in frictional materials: Archive of Applied Mechanics, v. 58, p. 148-159.

Hassani, R., and Chéry, J., 1996, Anelasticity explains topography associated with Basin and Range normal faulting: Geology, v. 24, p. 1095-1098.

Kong, L.S., Solomon, S.C., and Purdy, G.M., 1992, Microearthquake characteristics of a mid-ocean ridge along-axis high: Journal of Geophysical Research, v. 97, p. 1659-1685.

Lavier, L.L., and Buck, W.R., 2002, Half graben versus large-offset low-angle normal fault: Importance of keeping cool during normal faulting: Journal of Geophysical Research, v. 107, doi: 10.1029/2001JB000513.

Lavier, L.L., Buck, W.R., and Poliakov, A.N., 2000, Factors controlling normal fault offset in an ideal brittle layer: Journal of Geophysical Research, v. 105, p. 23,431-23,442, doi: 10.1029/2000JB900108.

Mackwell, S.J., Zimmerman, M.E., and Kohlstedt, D.L., 1998, High-temperature deformation of dry diabase with application to tectonics on Venus: Journal of Geophysical Research, v. 103, p. 975-984.

Ohara, Y., Yoshida, T., Kato, Y., and Kasuga, S., 2001, Giant megamullion in the Parece Vela backarc basin: Marine Geophysical Researches, v. 22, p. 47-61, doi:10.1023/A:1004818225642.

Phipps Morgan, J., and Chen, Y.J., 1993, The genesis of oceanic crust: Magma injection, hydrothermal circulation, and crustal flow: Journal of Geophysical Research, v. 98, p. 6283-6297.

Phipps Morgan, J., Parmentier, E.M., and Lin, J., 1987, Mechanisms for the origin of mid-ocean ridge axial topography: Implications for the thermal and mechanical structure of accreting plate boundaries: Journal of Geophysical Research, v. 92, p. 12,823-12,836.

Poliakov, A.N.B., 1993, An explicit inertial method for the simulation of viscoelastic flow: An evaluation of elastic effects on diapiric flow in two- and three-layers models, *in* Stone, D.B., and Runcorn, S.K., eds., Flow and Creep in the Solar System: Observations, Modeling and Theory, Kluwer Academic Publishers, p. 175-195.

Poliakov, A.N.B., and Buck, W.R., 1998, Mechanics of stretching elastic-plastic-viscous layers: Applications to slow-spreading mid-ocean ridges, *in* Buck, W.R., Delaney, P.T., Karson, J.A., and Lagabrielle, Y., eds., Faulting and Magmatism at Mid-Ocean Ridges, Geophysical Monograph 106, American Geophysical Union, Washington, D. C., p. 305-323.

Sleep, N.H., 1975, Formation of the oceanic crust: Some thermal constraints: Journal of Geophysical Research, v. 80, p. 4037-4042.

Wolfe, C.J., Purdy, G.M., Toomey, D.R., and Solomon, S.C., 1995, Microearthquake characteristics and crustal velocity structure at 29°N on the Mid-Atlantic Ridge: The architecture of a slow spreading segment: Journal of Geophysical Research, v. 100, p. 24,449-24,472.