Annual Report, 1998: Continuum Mechanics Models of Blind Thrusts in the LA Basin; B. H. Hager

A primary goal of the Geodesy Group is to develop models of interseismic velocities that are plausible from both a geological and a continuum mechanics perspective. New graduate student Brendan Meade has learned how to use and to modify the software that Bonnie Souter developed for estimating block motions from geologic and GPS data, and for predicting kinematically consistent GPS velocities from these models. Based on new geodetic results in the China Lake area, in the Imperial Valley (see the progress report of Herring et al.), and because of our interaction with Meghan Miller (on sabbatical at MIT this academic year) on the Mojave region, we have decided to substantially revise the block boundaries in the Souter/Hager model. We are now making these changes, before doing a "final" inversion for publication. (Meade received a merit-based department fellowship, so his education in the fundamentals has not been at SCEC expense.)

With Greg Lyzenga, Andrea Donnellan, and Danan Dong of JPL, Hager completed the manuscript "Reconciling Rapid Strain Accumulation with Deep Seismogenic Fault Planes in the Ventura Basin, California." GPS measurements across the east-central Ventura basin before the 1994 Northridge earthquake show exceptionally high strain rates, with velocity gradients of 5 mm/yr over distances of 12 km. Interpreting such rapid strain accumulation using the usual model of deep slip on a dislocation in a uniform elastic halfspace is problematic because it requires slip to extend to within 5 km of the surface. Such shallow slip is difficult to reconcile with the substantial coseismic displacement to depths in excess of 15 km during the Northridge earthquake. We model the displacement and velocity fields throughout the earthquake cycle using a 2-D finite element model with a viscoelastic rheology. Displacements are driven by far-field and basal velocity boundary conditions and by imposed periodic earthquakes on the thrust faults bounding the basin. The thrust faults rupture through an elastic upper crust to a depth of 15 km, comparable to the depth of substantial coseismic displacement for the 1994 Northridge earthquake. After a transient stage, during which stresses build up to quasi-equilibrium values, the behavior of the model becomes periodic.

For the horizontal component, the sum of the coseismic displacement divided by the repeat interval, plus the average interseismic velocity, is equal to the geologic velocity. The temporal variation in surface velocity depends mainly on the Elsasser relaxation time, which is sensitive to the ratio of Maxwell relaxation time of the ductile lower crust to its thickness. There are large temporal variations in interseismic velocity if the relaxation time is much less than the earthquake repeat time, and negligible variations if the relaxation time is comparable to the repeat time. Late in the cycle, models with longer relaxation times have larger velocity gradients across the basin, but not large enough to satisfy the geodetic observations. We are able to match the observed horizontal strain rate if we include the variations in elastic modulus associated with the deep basin sediments. The model reconciles geologic, geodetic, and seismological observations of deformation. There are trade-offs among the far-field convergence rate, the Elsasser time, the earthquake repeat time, and the time into the earthquake cycle. Acceptable convergence rates range from 8 mm/yr, for a relaxation time of the lower crust of 300 yr, to 12 mm/yr, for a 30 yr relaxation time. Observations of velocity variations early in the seismic cycle, such as are becoming available for Landers and Northridge, are crucial for constraining the rheology.

The almost tautological result of these viscoelastic models, that (horizontal) deformation on geologic time scales equals the sum of interseismic deformation plus coseismic deformation, means that total horizontal interseismic displacements do not depend on the rheological description of the model, but only on the difference between geologic and coseismic displacements. Thus, although we used viscoelastic models, we would have obtained the same horizontal interseismic displacements had we used the more standard model of creep on semi-infinite dislocations in an elastic halfspace. This result has important implications for modeling Landers postseismic displacements (see below).

Predicting vertical motions is more problematic because vertical motions are driven both by (visco)elastic stress and gravity. Furthermore, the surface topography that loads the surface depends not just on accumulated coseismic displacements, but also on erosion and sedimentation, processes that are particularly vigorous in the Ventura basin region, where topography is degraded at almost as high a rate as it is created. Under SCEC funding, MIT Research Scientist Ming Fang and Hager have begun to investigate implementing self-gravitation and load redistribution in viscoelastic models. Our preliminary results have made us question the validity of previous approaches (including our own) to computing interseismic velocities when convergence normal to block boundaries occurs.

It has long been recognized that gravitational coupling plays an important role in slow deformation and post-seismic relaxation near fault zones, especially in the vertical direction. The *m*-th component of deformation u_m induced by a cluster of dislocations with fault plane distribution $\boldsymbol{\Sigma}(\mathbf{r}')$ can be expressed by a general Somigliana's relation

$$u_m(\mathbf{r}) = \oint_{\Sigma} T^m_{l,k}(\mathbf{r},\mathbf{r}') U_k(\mathbf{r}') n_l(\mathbf{r}')\, d\Sigma(\mathbf{r}') \quad (1)$$

where U_k denotes the *k*-th component of the displacement discontinuity across the normal direction n_l of a fault. The key issue here is to estimate the Green tensor $T^m_{l,k}$. Generally speaking, $T^m_{l,k}$ can be determined by $T^m_{l,k} = G^m_{l,k} + S^m_{l,k}$, where $G^m_{l,k}$ is a homogeneous general solution and $S^m_{l,k}$ a special singular solution, usually in the entire space, representing the point source of an elementary dislocation. The complete solution for $G^m_{l,k}$ in a non-gravitational half space and $S^m_{l,k}$ in a non-gravitational whole space have been given by Ben-Menahem & Singh (1968) and Singh (1970). The general solution $G^m_{l,k}$ in a self-gravitational half space was first derived by Rundle (1980). For a long time, authors estimated $T^m_{l,k}$ by using Rundle's (1980) gravitational $G^m_{l,k}$ plus Ben-Menahem & Singh's (1968) non-gravitational $S^m_{l,k}$. The inconsistency is obvious. Efforts have been made recently in solving $S^m_{l,k}$ for a self-gravitating spherical Earth (Pollitz, 1992, 1997; Piersanti et al, 1995). This global approach does not have adequate resolution for a complex of local and regional distribution of fault zones, and thus is not suitable for our objectives.

Aimed at analyzing detailed space-geodetic observations in regional fault zones, we made progress in solving the $S^m_{l,k}$ problem in a self-gravitational whole space. The complete solution is massive and still an on going effort. As a demonstration, we discuss the most important vertical component $S_{3,3}^3$. In a cylindrical coordinate (r,φ,z), we can prove that for a point source at (0, 0, h)

$$S_{3,3}^3 = \int_0^\infty W_0(z,k)\, J_0(kr)\, k\, dk \qquad (2)$$

On a non-gravitational entire space, Ben-Menahem & Singh's (1968) solution gives

$$W_0 = K\frac{e^{-k|z-h|}}{4\pi}, \qquad K = \varepsilon + k\gamma(z-h) \qquad (3)$$

where γ is a function of elastic constants and ε is the unit discontinuity function. In a gravitational whole space with the gravity along the $\hat{\mathbf{z}}$ direction, we have

$$W_0 = \frac{1}{4\pi}\left(K(\varepsilon,k)e^{-k|z-h|} + A(\varepsilon,k)e^{-a|z-h|} + B(\varepsilon,k)e^{-b|z-h|}\right) \qquad (4)$$

Here *a*, and *b* in (4) are identical to Rundle's (1980) characteristic values a_1 and a_2. We save the detailed expressions for the coefficients *K*, *A*, and *B* for later publications. However, it is worth noting that these coefficients are independent of depth z, while the coefficient *K* in (3) explicitly depends upon z. This is a direct consequence of degeneracy breaking down. The gravity field along the $\hat{\mathbf{z}}$ direction beaks down the isotropy (spherical symmetry) of the homogeneous space, and hence breaks the degenerate solution (3) into a full solution (4). We believe that this analysis warrants further investigation.

Prepared by our viscoelastic modeling of Northridge interseismic velocities, we have been keenly interested in the observations and models of postseismic displacements following the Northridge and, more abundantly, the Landers earthquakes. Particularly startling is the statement by Deng et al (1998), "afterslip can only explain one horizontal component of the postseismic deformation, whereas viscoelastic flow can explain [both] the horizontal [components]." This statement seems to be in direct contradiction to the "tautology" discussed above and by others (e.g., Savage) that, for horizontal displacements, it should be mathematically possible to use either a viscoelastic or elastic dislocation model. The issue at question is whether an elastic dislocation model with dip-slip displacement can cause surface motions perpendicular to the trace of the dislocation. In their Figure 2b, Deng et al show that a uniform dislocation from, say, 10 – 15 km depth cannot. Yet if the dislocation is allowed to extend much deeper than 15 km, fault normal displacement in the sense observed is predicted. Thus, while their particular model of relatively shallow afterslip can be rejected, a more general model with deeper afterslip may actually provide an acceptable fit to the data. Models that incorporate rate-state friction laws (e.g., Tse and Rice) predict downward propagating afterslip on the fault plane. A more useful approach than an "either-or" approach to the mechanism is to determine what parameter ranges can be excluded by the available data.

Publications and Products

During the past year, one paper and one thesis were submitted, and one AGU talk was supported under SCEC funding to Hager. Also, Hager is on the SCIGN Advisory Board.

Reconciling rapid strain accumulation with deep seismogenic fault planes in the Ventura basin, California, B. H. Hager, G. A. Lyzenga, A. Donnellan, and D. Dong, *J. Geophys. Res.*, under revision, 1998, SCEC contribution #459.

Comparisons of geologic models to GPS observations in southern California, B. J. Souter, Ph. D. thesis, Massachusetts Institute of Technology, February, 1998.

Hager, B. H., Research opportunities and challenges in earth sciences for the next decade, *AGU 1998 Fall Meeting*, *79*, 45, F14, 1998.

Note: The following report by Elizabeth Hearn (EH), U. Oregon, is included because she will be a postdoc at MIT and partial support is requested for her in the MIT Science proposal (B. H. Hager, PI). Also relevant to that proposal is work reported in the progress report of the investigation of T. Herring, R. King, S. McClusky, and R. Reilinger.

Following the 1992 M 7.3 Landers earthquake, a postseismic strain transient occurred with about 15% of the coseismic moment release and a characteristic decay time of 5 - 80 days (Wyatt et al., 1994; Savage & Svarc, 1997 [SS97]; Shen et al., 1994). This rapid postseismic response was superimposed on a slower strain transient whose characteristic decay time has not been resolved (SS97). Because of its size and the quality of GPS and strain data available from before and immediately after the event, Landers is an ideal case for evaluating distributed viscoelastic relaxation (Deng et al, 1998), fault afterslip (SS97), and poroelastic (Peltzer et al., 1998) models. Although the viscoelastic model (D98) reproduces the pattern and magnitude of the measured postseismic surface displacement field, it fails to replicate the rapid strain transient. Existing kinematic afterslip models yield vertical surface displacements opposite in polarity to observations (P98). Poroelasticity can overcome the vertical response issue, but requires an (unreasonable?) poroelastic response for the entire upper crust (P98), and has not been shown to reproduce the temporal behavior. No postseismic deformation model has yet explained the two relaxation times.

EH used the 3-D viscoelastic finite-element program GAEA (Saucier, 1991), to dynamically model postseismic deformation following the Landers earthquake. GAEA can model deformation associated with prescribed boundary and fault slip displacements, and gravitational forces arising from density and elevation variations. Slip on faults is either specified ("split nodes", Melosh and Raefsky, 1981) or calculated ("slippery" nodes, Melosh and Williams, 1989). Fault slip is parameterized with piecewise quadratic sections, which precludes discontinuities between adjoining elements. EH modified GAEA to include afterslip and viscoelasticity with biviscous rheology. Afterslip is modeled as a pseudoviscous response to shear stress on the deep extension of the rupture plane (Sibson, 1982). As slip progresses, the shear stress decays, which slows and eventually stops the afterslip. Biviscous rheology (Ivins, 1996; Ivins and Sammis, 1996) was incorporated in GAEA because of the short characteristic times for postseismic strains following the Landers earthquake. EH tested many afterslip and viscoelastic models to find the ones that best match temporal and spatial postseismic displacement patterns, assuming viscous afterslip and lower crustal VE relaxation. (No poroelastic effects and no gravitation is included in the models discussed here.)

The (irregular) model mesh has split nodes representing the coseismic rupture along the traces of the Landers, Big Bear and Joshua Tree ruptures. Faults are modeled as curvilinear, vertical surfaces. The 500 by 700 by 55 km mesh is 5 elements deep and the model layers are 5, 5, 5, 10, and 20 km thick. The top three layers are modeled as elastic and represent the seismogenic upper crust; the bottom two layers are modeled either as elastic or viscoelastic, and represent the lower crust. Coseismic ruptures occur within the top three layers, and afterslip is restricted to the third layer (at 15 km depth) and below.

Effective fault viscosity trades off with assumed fault zone thickness. Though different viscosity values could be used at each node, we model a viscosity function varying with depth only. (Modeling frictional faulting, sensitive to Coulomb stresses, is not yet included.) Afterslip is calculated dynamically, so there is no guarantee that the afterslip distribution results in the same surface displacements as a kinematic lower crustal relaxation model. Thus our modeling can determine the conditions under which kinematically required slip functions make dynamic sense.

Postseismic displacements modeled assuming viscoelastic relaxation have directions and magnitudes of surface displacements similar to the observations of SS97, but seem to be 10-20 degrees counterclockwise of those of Shen et al. (1994). West of the rupture and at the nearest stations on either side, this model replicates the SS97 velocity orientations better than does their dislocation model. For sites at the east end of the GPS alignment, however, the viscoelastic model under predicts displacement magnitudes, and the orientations are rotated up to 20 degrees counterclockwise of observed displacements (relative to SANH). Fault-normal displacements occur in our viscoelastic model, driven by purely strike-slip coseismic displacements, so we do not think dip slip (D98) is needed. Surface displacements west of the rupture are larger than to the east relative to the far field because of the curvature of the rupture; coseismic stresses throughout the crust are greater to the west than to the east, resulting in larger postseismic displacements.

The afterslip model also appears to model surface displacement orientations reasonably well, although displacement orientations near the rupture are off of the SS97 displacement orientations by up to 45 degrees. Overall, the afterslip model matches the Shen data better because the displacements appear to be rotated 10-20 degrees clockwise from the vectors given by the VE lower crust model. The displacement magnitudes decline with distance from the rupture at both ends of the array, inconsistent with GPS observations, unless the fault viscosity decreases with depth. Constant fault zone viscosity fails to give the fast initial relaxation, similar to viscoelastic lower crust. We get more fault-normal motion than SS97, probably because our model results in more afterslip on the N-S oriented part of the rupture, south of the GPS array.

We have not looked carefully at gravitational forces yet, but it seems that the VE lower crust model is more consistent with vertical displacements measured by inSAR than the afterslip model. Postseismic vertical displacements are opposite in sign to the coseismic displacements. Some of this may be poroelastic but some may be the response to coseismic gravitational PE changes.

Finally, the actual orientations of postseismic deformation vectors relative to a stable far-field reference are important, particularly for a curved rupture with a variable slip. Also, the characteristic time of rapid decay vs. distance from the rupture looks about the same everywhere, which argues for a thick biviscous lower crust and against a thin weak layer or afterslip. Although differences to the characteristic decay times of fast transients may be too small to see with existing GPS data, they may be related to strain reversal at PFO.