Relationship of geodetic velocities to velocities in the mantle

E. A. Hetland and B. H. Hager

Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts, USA

Received 7 June 2004; revised 23 July 2004; accepted 4 August 2004; published 8 September 2004.

[1] We examine the relationship of the velocities at the surface (i.e., geodetic velocities) to the velocities in the upper-most mantle throughout a seismic cycle. We model the rheology of the lower crust and mantle as linear viscoelastic, and vary the relative viscosities between the lower crust and mantle, as well as the distribution of viscosities and secular velocities in the mantle. We find that the geodetic velocities are related to the transient velocities at depth; the strength of the relationship decreases as the contrast between the lower crust and mantle viscosity increases. In these models we find that the geodetic velocities are not related to the secular velocities at depth. *INDEX TERMS:* 1236 Geodesy and Gravity: Rheology of the lithosphere and mantle (8160); 1208 Geodesy and Gravity: Crustal movements—intraplate (8110); 8120 Tectonophysics: Dynamics of lithosphere and mantle—general; 8159 Tectonophysics: Rheology—crust and lithosphere. **Citation:** Hetland, E. A., and B. H. Hager (2004), Relationship of geodetic velocities to velocities in the mantle, *Geophys. Res. Lett.*, *31*, L17604, doi:10.1029/2004GL020691.

1. Introduction

[2] There is a large body of research that uses surface velocities measured with geodetic techniques to infer the distribution of the rheological properties of the lower crust and mantle [e.g., *Kaufmann and Amelung*, 2000; *Kenner and Segall*, 2000; *Piersanti et al.*, 2001; *Pollitz*, 2001] and the distribution of secular velocities in the mantle [e.g., *Bourne et al.*, 1998; *Flesch et al.*, 2000, 2001]. Most common among the former studies are models of post-seismic relaxation, where transient velocities in non-elastic layers at depth are a function of the co-seismic stress changes, the earthquake rupture geometry, and the rheological structure. For a given earthquake, the transient velocities at depth depend on the distribution of rheology. Hence, when post-seismic geodetic observations are used to infer the distribution of rheology, the implicit assumption is that the surface velocities are related to the transient velocities at depth.

[3] In addition to studies of transient geodetic velocities, researchers have used the thin viscous sheet (TVS) approximation, where it is assumed that the velocities in the lithosphere do not vary with depth, to infer distributions of velocities in the mantle from geodetic observations [e.g., *Bourne et al.*, 1998]. *Bourne et al.* [1998] argued for the appropriateness of the TVS approximation by showing that fault slip rates calculated from geodetic observations using the TVS approximation matched the geologic slip rates in New Zealand and Southern California. *Savage et al.* [1999], using geodetic data from Northern California, demonstrated that the slip rates calculated using a model of viscoelastic relaxation from previous earthquakes matched the geologic rates as well as rates calculated using the TVS approximation, concluding that there is no compelling argument that the TVS approximation is appropriate for the lithosphere. The TVS approximation assumes a relationship between surface velocities and the secular velocities at depth.

[4] Several papers have argued that the distribution of surface velocities is independent of the distribution of secular velocities in the upper-most mantle. *Li and Rice* [1987] demonstrated that, for an elastic upper crust separated from the mantle by a lower crustal Maxwell viscoelastic layer, with a relaxation time smaller than that of the mantle, the surface velocities are not dependent on the particular distribution of velocities at the top of the mantle. *Savage* [2000], using the principle of correspondence, has shown that the geodetic velocities throughout a seismic cycle are identical for models of viscoelastic flow in lower layers and creep along the continuation of the fault at depth. *Zatman* [2000] used an analytic solution to investigate the relationship between an elastic upper crust and viscous lower stratum, and concluded that the velocities at the surface are not dependent on the steady motion of the viscous region, but noted that surface velocities could be used to investigate the transient velocities in the viscous strata due to post-seismic relaxation.

[5] In this paper, we investigate the relationship between geodetic velocities and those in a linear viscoelastic mantle, separated by a linear viscoelastic lower crust. When discussing this relationship, it is prudent to distinguish between the relationship of geodetic observations to distributions of *transient* velocities at depth, and the relationship to *secular* velocities at depth. If the surface velocities are distinct for different transient or secular velocities in the mantle, we say that the surface velocities are related to those in the mantle. If the surface velocities are identical for different mantle velocities, we say that there is no relationship.

2. Models

[6] We investigate two-dimensional finite element models composed of a strike-slip fault breaking an elastic upper crust overlaying a viscoelastic lower crust and viscoelastic mantle. We use the finite element program GeoFEST 2.3 [*Lyzenga et al.*, 2000]. The two-dimensionality of the model implies that the fault is infinite in extent. In the lower crust and mantle, we use linear Maxwell viscoelastic rheologies, where the elastic response is given by the shear modulus, and the time-dependent response is specified by the viscosity. While the lower crust and mantle viscosities are distinct, we do not vary the viscosities with depth in each of these layers. Additionally, we specify that the lower-crust

0094-8276/04/2004GL020691$05.00

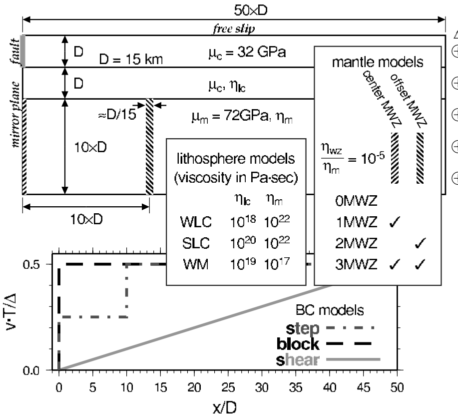

Figure 1. Model geometry used in this study, shear moduli of the crust (μ_c) and mantle (μ_m), and viscosities of the lower crust (η_{lc}), mantle (η_m), and MWZ (η_{wz}). The lower graph shows the basal boundary conditions (BC models), and the tables list the viscosities used in the lithosphere models, and the presence (check mark) of the MWZ in the mantle models. Due to anti-symmetry, only one half of the model is shown, Δ is the rupture displacement, T is the recurrence time, and v is velocity.

viscosity has no lateral variation. We control the lateral variation of viscosities in the mantle by including low-viscosity columns, which we refer to as mantle weak zones (MWZ). The viscosity of the columns is 10^{-5} of the viscosity of the background mantle, and the MWZ are located directly under the fault or offset on both sides of the fault (Figure 1). Due to anti-symmetry, we only model one half of the model, specifying a zero velocity boundary condition on the edge of the model containing the fault and a constant velocity boundary condition on the opposite edge, while the top of the model is free-slip (Figure 1).

[7] We group the models according to variations in: 1) the relative strengths between the lower crust and mantle, "lithosphere models", 2) the number of MWZ in the mantle, "mantle models", and 3) the basal boundary conditions applied to the model, "BC models". We consider three lithosphere models: a weak lower crust with a strong mantle (WLC), a strong lower crust with a stronger mantle (SLC), and a weak mantle with a stronger lower crust (WM). We consider four mantle models; the first does not contain any MWZ, yielding a homogeneous mantle (0MWZ), and the remaining three contain up to three MWZ, which we refer to as *n*MWZ where *n* is the number of MWZ in the mantle (Figure 1). To control the secular velocity distribution in the upper-most mantle, we specify three boundary conditions on the bottom of the model (BC models): a simple-shear velocity gradient (shear), a block-like velocity (block), and a stepped velocity (step, similar to bookshelf block-like model; Figure 1).

[8] We do not propose that the MWZ in these models are physically appropriate for models of the lithosphere; we only use them as means to control the secular velocities along the Moho. We use a sufficiently low viscosity contrast between the MWZ and the mantle such that the distribution of secular velocities along the Moho is distinct for each of the three BC models. We choose viscosities of the lower crust and mantle purely for demonstrative purposes. By considering three lithosphere, four mantle and three BC models, there are 36 possible models. However, in order to illustrate the relationship between surface and mantle velocities we present 18 models in this paper, as described below.

[9] We "spin-up" all of the fault models by cycling over successive fault ruptures until the displacements and stresses throughout a seismic cycle do not vary from one cycle to the next, reaching a cycle invariant state [e.g., *Lyzenga et al.*, 1991; *Savage*, 2000]. The number of seismic cycles that is required to attain cycle invariance is a function of the boundary conditions, the strength of the system, and the recurrence time of the ruptures, T (we used T = 100 years).

3. Results

[10] We present the results of the models in two sections. First we discuss the influence of the distribution of viscosities in the upper-most mantle on the velocities at the surface — the relationship of surface velocities to the *transient* velocities at depth. Second we discuss the influence of the distribution of mantle velocities on the velocities at the surface — the relationship of surface velocities to the *secular* velocities at depth.

3.1. Relationship to Transient Velocities

[11] To examine the relationship of surface velocities to the distribution of the transient velocities that develop for a particular distribution of viscosities in the mantle, we consider all four mantle models and the shear BC model. In the WLC, SLC and WM lithosphere models, the surface velocities are all distinct for the four mantle models (Figure 2). The velocities throughout the seismic cycle in

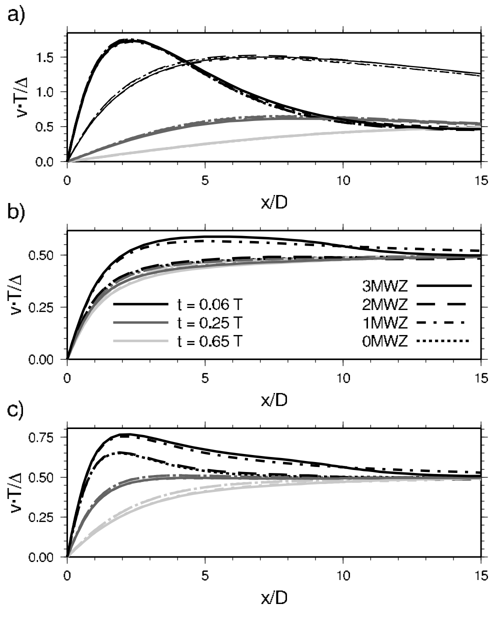

Figure 2. Velocities, v, at the surface throughout a seismic cycle using the shear BC model, varying mantle viscosity models and the WLC (**a**), SLC (**b**) and WM (**c**) lithosphere models. Thin lines in panel **a** are the velocities for a model with η_{lc} and η_m equal to 10^{21} and 10^{17} Pa·sec, respectively, at time 0.06T.

the WLC model are essentially indistinguishable for the four mantle models, since the co-seismic stresses relax in the lower crust much more rapidly than in the mantle (Figure 2a). The WLC models exhibit a weak relationship between the surface velocities and the transient velocities in the mantle.

[12] In the SLC lithosphere models, the surface velocities in the 3MWZ and 1MWZ models are quite distinct from the 0MWZ and 2MWZ models (Figure 2b). The surface velocities throughout most of the later seismic cycle in the 1MWZ and 3MWZ models are only slightly different than those of the 0MWZ model, while during the post-seismic period, the velocities are much larger, since the relaxation of coseismic stresses in the MWZ dominates the relaxation of stresses in the lower crust. For the SLC model, the 0MWZ and 2MWZ models show little variation of velocities throughout the seismic cycle. The 2MWZ model only contains the MWZ offset from the fault, and the co-seismic stresses do not cause significant shear across the offset MWZ, yielding surface velocities similar to the homogeneous mantle (0MWZ). In the 3MWZ model, the post-seismic velocities are constrained between the center and off-center MWZ, exhibiting an almost block-like post-seismic motion, whereas in the 1MWZ model the post-seismic velocities are more smoothly distributed, decreasing with distance away from the fault (Figure 2b). The SLC models exhibit a strong relationship between the surface velocities and the transient velocities in the mantle, but only in the immediate post-seismic period.

[13] In the WM model, the surface velocities are distinct for the four different mantle models (Figure 2c). As in the SLC models, in the WM models the post-seismic velocities in the 0MWZ and 2MWZ viscosity models are virtually identical. In general, post-seismic velocities are dominated by relaxation in the MWZ and the weak upper-most mantle. In the WM models, the transient velocities in the mantle during the post-seismic period are related to the surface velocities.

[14] As the contrast between the lower crust and mantle viscosities increases, the geodetic velocities are less closely related to the transient velocities at depth, regardless of whether the mantle is stronger or weaker than the lower crust. For example, for a model with a background mantle viscosity the same as the WM model, but with a lower crust viscosity of 10^{21} Pa · sec, the velocities early in a seismic cycle are virtually indistinguishable for each of the viscosity models (Figure 2a). As the relationship to transient velocities is most prominent in the post-seismic period, it is sufficient to demonstrate the lack of a relationship only in the post-seismic period.

3.2. Relationship to Secular Velocities

[15] To examine the dependence of the surface velocities on the distribution of secular velocities in the mantle, we use the mantle model 3MWZ and control the distribution of velocities in the upper-most mantle using varying BC models. The velocities in the upper-most mantle develop passively, and the amount of shear which develops across each MWZ depends not only on the basal boundary condition but also on the viscosities of the lower crust and mantle. We ensure that the distributions of velocities along the Moho are distinct for the three BC models. To illustrate the distinctness of the secular velocities in the upper-most mantle, we show the total displacements throughout a seismic cycle along the Moho (Figure 3). The displacements convey the degree that the basal boundary conditions propagate up to the Moho and lower crust. The propagation is not exact. For instance immediately below the Moho in the shear BC model, there is still some strain across the MWZ. However it is much less than the amount of strain across the MWZ in the block or step BC models (Figure 3).

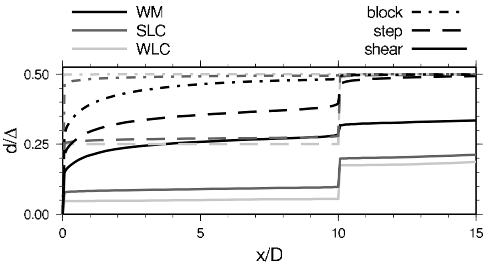

Figure 3. Cumulative displacements, d, throughout a seismic cycle along the Moho using mantle viscosity model 3MWZ, and indicated lithosphere and BC models.

[16] For all three lithospheric models — weak lower crust, strong lower crust and weak mantle — the velocities at the surface throughout the seismic cycle are identical and are the same as those shown for the 3MWZ models in Figure 2. Hence, the surface velocities are not dependent on the distribution of secular velocities in the upper-most mantle, and there is no relationship between the surface velocities and the distribution of secular velocities in the mantle.

4. Discussion

[17] The development of the transient velocities in the upper-most mantle depends on the particular distribution of viscosities in the lower crust and mantle, and the relationship of geodetic observations to transient velocities in the mantle is often exploited to infer rheologies of the lower crust and mantle using geodetic data of both post-seismic relaxation [e.g., *Kenner and Segall*, 2000] and viscoelastic rebound [e.g., *Kaufmann and Amelung*, 2000]. Although, in this study, the viscosity variations are limited to discrete weak zones in the mantle, the conclusions are likely to hold for other viscosity distributions. For instance, *Kenner and Segall* [2000] and *Pollitz* [2001] both evaluated velocities throughout the seismic cycle for models in which the lower crust contained low-viscosity zones of lateral dimensions comparable to the locking depth of the fault.

[18] Flesch and colleagues used geodetic data to infer the effective lithospheric viscosity throughout the western United States [*Flesch et al.*, 2000] and central Asia [*Flesch et al.*, 2001]. In this paper, we show that geodetic data may be in fact sensitive to the rheology of the sub-seismogenic layers. However, Flesch and coworkers did not use the geodetic data directly to infer effective viscosities; instead they used the geodetic data to infer secular strain rates using the TVS approximation. Then, calculating stresses from gravitational potential and tectonic boundary conditions,

they obtained effective viscosities [*Flesch et al.*, 2000, 2001]. Their studies implicitly assumed that the geodetic observations are related to the secular strain rates throughout the lithosphere. We demonstrate that, with linear viscosity, this assumption is not valid. *Flesch et al.* [2000] resolved a large gradient in effective viscosity roughly across the Central Nevada Seismic Belt in the western US, a region with significant transient velocities due to post-seismic relaxation [e.g., *Hetland and Hager*, 2003]. The failure to account for transient velocities leads to an incorrect inference of lithospheric strength.

5. Conclusions

[19] We evaluate the relationship of geodetic observations to the transient and secular velocities in the uppermost mantle. The velocities are related (not related) when surface velocities are distinct (identical) for distinct velocities in the mantle. We consider lithospheric models characterized by an upper mantle stronger than the lower crust, as well as a lower crust stronger than the mantle. We demonstrate that geodetic velocities are always related to the transient velocities in the mantle, although the relationship depends on the relative strengths of the lower crust and mantle. As the contrast between the lower crust and mantle viscosities increases, the relationship becomes weaker.

[20] We also demonstrate that the geodetic velocities are not related to the secular velocities at depth, even when the relaxation time of the lower layers is long compared to the seismic cycle. This contributes to the findings of several past studies [e.g., *Li and Rice*, 1987; *Savage*, 2000; *Zatman*, 2000], illustrating the lack of a relationship between surface velocities and secular velocities at depth using linear rheologies. Moreover, our models clarify the distinction between the relationship of geodetic data to secular and transient velocities at depth.

[21] **Acknowledgments.** The authors thank G. Lyzenga and J. Parker for the use of GeoFEST 2.3, available at www.openchannelsoftware.org. Insightful reviews by R. Bürgmann and one anonymous reviewer greatly improved this paper. All figures were produced using GMT 3.3 by *Wessel and Smith* [1998]. This work was supported by NSF grants 9814303 and 6895843.

References

Bourne, S. J., P. C. England, and B. Parsons (1998), The motion of crustal blocks driven by flow of the lower lithosphere and implications for slip rates of continental strike-slip faults, *Nature*, *391*, 655–659.

Flesch, L. M., W. E. Holt, A. J. Haines, and B. M. Shen-Tu (2000), Dynamics of the Pacific–North American plate boundary in the western United States, *Science*, *287*, 834–836.

Flesch, L. M., J. A. Haines, and W. E. Holt (2001), Dynamics of the India-Eurasia collision zone, *J. Geophys. Res.*, *106*, 16,435–16,460.

Hetland, E. A., and B. H. Hager (2003), Postseismic relaxation across the Central Nevada Seismic Belt, *J. Geophys. Res.*, *108*(B8), 2394, doi:10.1029/2002JB002257.

Kaufmann, G., and F. Amelung (2000), Reservoir-induced deformation and continental rheology in the vicinity of Lake Meade, Nevada, *J. Geophys. Res.*, *105*, 16,341–16,358.

Kenner, S. J., and P. Segall (2000), Postseismic deformation following the 1906 San Francisco earthquake, *J. Geophys. Res.*, *105*, 13,195–13,209.

Li, V. C., and J. R. Rice (1987), Crustal deformation in great California earthquake cycles, *J. Geophys. Res.*, *92*, 11,533–11,551.

Lyzenga, G. A., A. Raefsky, and S. G. Mulligan (1991), Models of recurrent strike-slip earthquake cycles and the state of crustal stress, *J. Geophys. Res.*, *96*, 21,623–21,640.

Lyzenga, G. A., W. R. Panero, and A. Donnellan (2000), The influence of anelastic surface layers on postseismic thrust fault deformation, *J. Geophys. Res.*, *105*, 3151–3157.

Piersanti, A., C. Nostro, and F. Riguzzi (2001), Active displacement field in the Suez-Sinai area: The role of postseismic deformation, *Earth Planet. Sci. Lett.*, *193*, 13–23.

Pollitz, F. F. (2001), Viscoelastic shear zone model of a strike-slip earthquake cycle, *J. Geophys. Res.*, *106*, 26,541–26,560.

Savage, J. C. (2000), Viscoelastic-coupling model for the earthquake cycle driven from below, *J. Geophys. Res.*, *105*, 25,525–25,532.

Savage, J. C., J. L. Svarc, and W. H. Prescott (1999), Geodetic estimates of fault slip rates in the San Francisco Bay Area, *Geophys. Res. Lett.*, *104*, 4995–5002.

Wessel, P., and W. H. F. Smith (1998), New, improved version of the Generic Mapping Tools released, *Eos Trans. AGU*, *79*(47), 579.

Zatman, S. (2000), On steady rate coupling between an elastic upper crust and a viscous interior, *Geophys. Res. Lett.*, *27*, 2421–2424.

B. H. Hager and E. A. Hetland, Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139, USA. (eah@mit.edu)