

Magmatism and Extension: The Thermal and Mechanical Effects of the Yellowstone Hotspot

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Currently, there is a parabolic-shaped pattern of seismicity and latest Quaternary faulting which centers on the axis of the eastern Snake River Plain (SRP) and has its apex at the Yellowstone Plateau. The pattern of activity surrounds a region of aseismicity that includes the eastern SRP. Geologic evidence suggests that this pattern has migrated in tandem with rhyolitic volcanism induced by the Yellowstone hotspot during its 16 m.y. trek across western North America. Similarly there is a migration in the onset of basaltic lava flows and the emplacement of up to 12 km of midcrustal mafic intrusions that follows the track of rhyolitic volcanism. These observations suggest that the intrusion of magmas has a significant influence on the pattern of seismicity and faulting. We present a one-dimensional, finite-difference, thermomechanical model that accounts for the observed pattern of increased faulting followed by fault quiescence within the circum-eastern SRP. In this model, mafic magmas are intruded into a lithosphere that is already extending. The intrusions heat the surrounding rocks resulting in locally increased strain rates. As the intruded magmas solidify, the length of time required to return strain rates to their pre-intrusion level is then determined. The model assumes constant horizontal tectonic forces and maps strain rate as a function of yield strength and time since intrusion. We vary model parameters such as crustal thickness, initial geothermal gradient, and amount of magma intruded, in order to assess how they affect turnaround time for strain rates. The amount and timing of intruded material allowed by the model are constrained by seismic refraction data as well as by regional geochronologic and petrologic studies. We assume intruded rock rheology to be a Maryland diabase; mantle rheology we base on dry olivine; we represent the crust by both dry granite and dry anorthosite rheologies. Our results suggest that 2 to 3 m.y. appears to be a reasonable length of time for strain rates to return to levels present before a midcrustal mafic intrusion equivalent in magnitude to the Yellowstone intrusion. This corresponds closely to the length of time between the onset of along-axis accelerated faulting and subsequent fault quiescence, assuming a hotspot velocity of 3.5 to 4.0 cm/yr relative to the North American Plate.

INTRODUCTION

The eastern Snake River Plain (SRP) and the Yellowstone volcanic field are products of the southwesterly movement of the North American Plate over a hotspot [Morgan, 1972; Smith and Sbar, 1974; Suppe *et al.*, 1975; Armstrong *et al.*, 1975]. The track of the hotspot is defined by a 450 km migration in explosive rhyolitic eruptive centers from southeastern Oregon/northwestern Nevada at 16 Ma to the most recent eruption at 0.6 Ma on the Yellowstone Plateau [Armstrong *et al.*, 1975]. Centered on the track created by the hotspot is a pattern of seismicity and latest Quaternary faulting that is bounded by two non-concentric parabolas whose apexes are at Yellowstone Plateau [Anders *et al.*, 1989]. Myers and Hamilton [1964] first described this pattern of seismicity as a broad arc of epicenters and accelerated faulting which they called the "products of general collapse" (Figure 1); other workers subsequently have described the interior boundary of the seismic domain as a "wedge" [Smith and Sbar, 1974], and a "V" [Scott *et al.*, 1985; Smith *et al.*, 1985]. Smith and Sbar [1974] described the eastern SRP as a wedge-shaped opening or rift that

propagates eastward. They called on aseismic creep to accommodate extension within the eastern SRP. Conversely, Anders and Geissman [1983], working south of the eastern SRP, showed that faults within the interior region, which formed what they called a "collapse shadow," were inactive, and therefore not creeping aseismically. They further suggested that increases in the rate of faulting, followed by fault inactivity, are tied directly to the movement of the Yellowstone hotspot. Subsequently, this pattern of fault activity has been observed at numerous locations along the track of the Yellowstone hotspot [Scott *et al.*, 1985; Smith *et al.*, 1985; Pierce and Scott, 1986; Anders and Geissman, 1987; Anders *et al.*, 1989; Pierce and Morgan, 1990].

Here we present a thermomechanical model that can account both for the observed increase and decrease in fault displacement rates along-axis of the Yellowstone hotspot track as well as those observed off-axis. We will explain the origin of the parabolic distribution of seismicity and latest Quaternary faulting within the circum-eastern SRP in two parts. We propose first that the pattern of increased fault activity along-axis results from an initial increase in tectonic strain rates caused by hotspot magma heating the lithosphere. The observed along-axis reduction in strain rates results from the subsequent crystallization of these magmas. Second, we propose that the off-axis increase in fault displacement rates is the result of the thermal effects of hotspot magmas intruded into the upper mantle and/or lower crustal flow driven by buoyancy forces created by the hotspot.

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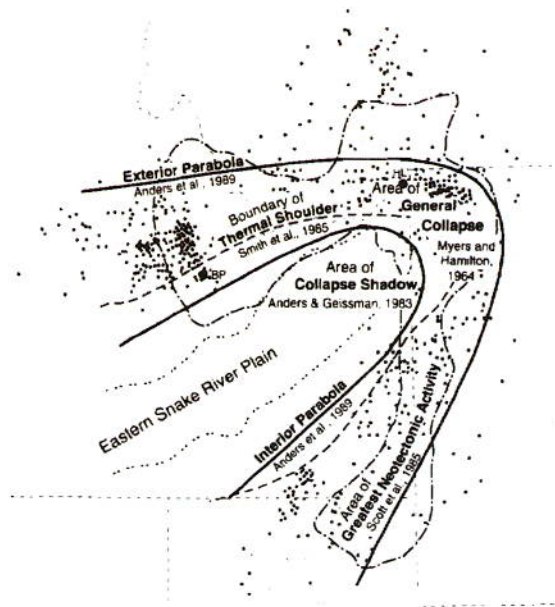


Fig. 1. Map showing the distribution of earthquakes greater than $M_s = 3$ (small dots) since 1961. Large solid circles represent the 1983 Borah Peak (BP) $M_s = 7.3$ earthquake and the 1959 Hebgen Lake (HL) $M_s = 7.5$ earthquake. Domains of seismic activity are defined by dashed and dotted lines and are discussed in the text.

MAGMATISM AND CRUSTAL STRUCTURE OF THE CIRCUM-EASTERN SRP

Our modeling of the geological processes which formed the Yellowstone hotspot track is constrained in depth by geophysical data and in time by absolute age determinations. Seismic and gravity data from the eastern SRP suggest that the Moho under the eastern SRP is about 45 km and 35-50 km under the peripheral regions [McCamy and Meyer, 1964; Sparlin et al., 1982; Braile et al., 1982]. These data also suggest a velocity anomaly under the eastern SRP, approximately 12 km thick and having a velocity of 6.5 km/s [Sparlin et al., 1982; Braile et al., 1982]. Sparlin et al. [1982] suggested that the anomaly corresponds to rock with a 2.88 g/cm³ density. The distribution of densities in the upper mantle and crust, as implied from the velocity structure, provides an important constraint on the magmatic evolution of the circum-eastern SRP.

Isotopic and geochemical data has been used by Hildreth et al. [1991] to suggest that SRP basalts are mantle derived. These magmas reside at the crust/mantle boundary until they have fractionated sufficiently to overcome the density contrast between the crust and the magma [Leeman, 1988], and/or until they heat up the lower crust enough to decrease its viscosity, thus allowing the buoyant magmas to ascend. The rhyolitic magmas extruded in the vicinity of Yellowstone show low Sr^{87}/Sr^{86} and Pb^{206}/Pb^{204} ratios, which suggest that the magmas are derived from the lower crust and have not been contaminated by upper crustal rocks or by more primitive mantle-derived rocks [Doe et al., 1982].

Armstrong et al. [1975] have shown that there is a progressive decrease in isotopic ages of silicic eruptions from southeastern Oregon/northwest Nevada at 16 Ma to the most recent eruption on the Yellowstone Plateau (Figure 2), about 600 ka [Christiansen and Blank, 1972]. Silicic eruptions in the eastern

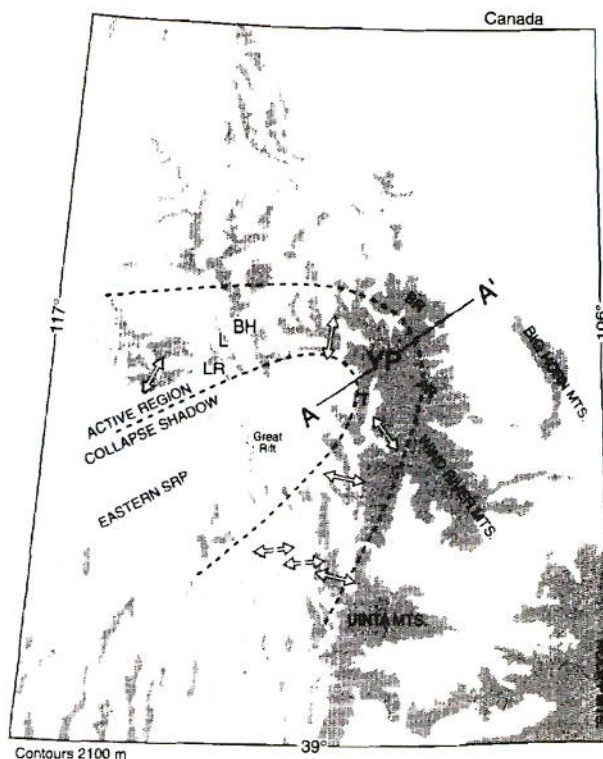


Fig. 2. Elevation map of an area slightly larger than the area depicted in Figure 1. Shaded areas represent where the elevation is higher than 2100 m. The eastern Snake River Plain forms a marked topographic depression, and there appears to be a rough elevation anomaly that extends outward from the outer parabola. This anomaly is difficult to objectively subtract from elevation rises that owe their origin to late Cretaceous and early Tertiary overthrusting, Eocene volcanism, and Cenozoic normal faulting footwall uplift. YP, BH, BR, AR, L, T and LR represent the Yellowstone Plateau, Beaverhead Mountains, and the Bear Tooth, Absaroka, Lemhi, Teton and Lost River Ranges respectively. The light hatched area is the Great Salt Lake. Double-headed arrows are extension directions determined from seismic moment tensors [Eddington et al., 1987]. Cross-section A-A' is shown in Figure 3.

SRP were not found to progress in an absolute linear manner from southwest to northeast. Rather, the trend was generally linear, punctuated by eruptions occurring ahead of or behind the assumed position of the sublithospheric mantle plume. This pattern of eruptions probably results from the complex interaction between the rising magmas and the lithosphere.

Armstrong et al. [1975] calculated the average northeastern migration rate to be about 3.5 cm/yr. Subsequently, Rodgers et al. [1990] recalculated the rate at 4.0 cm/yr. The rates are based on the deposition of volcanic rock in the extended region that surrounds the eastern SRP. These values obviously depend on the amount of extension in the region adjacent to that traversed by the hotspot, the accuracy of the dating techniques, and the position of the assumed center of eruption, which is difficult to locate because younger basalt flows often cover the older rhyolitic deposits. The spatial scatter of eruptive events is especially apparent during the last 6.5 m.y. [Morgan et al., 1984; Christiansen and Blank, 1972].

Initiation of extensive basaltic activity follows rhyolitic volcanism by about 2 to 3 m.y. This timing is based on the observation that most of the source of the 2.0 Ma Huckleberry Ridge Tuff is exposed, whereas all of the source of the 4.4 Ma

tuff of [Ember] has been resurfaced. The 8-10 m.y. mafic in Falls (F) source (Figure) northeast farther reported uncertainty critical Plateau. A thin magma midcrust: the obs Yellowstone con magm: contrib in the r similar of east chamb demon: added. Thes from v impede arrivec crustal and ge base c midcr rhyolit eastern

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tuff of Kilgore is buried under about a kilometer of basalt [Embree *et al.*, 1978]. Once initiated, basaltic volcanic activity has been continuous along most of the track of the hotspot and has resulted in the accumulation of 1 to 3 km of basalt at the surface. It is likely that midcrustal intrusion of mafic magma has been continuous, although seismic refraction studies show that the 8-12 km midcrustal velocity anomaly caused by midcrustal mafic intrusions is present under the sources of the 1.3 Ma Mesa Falls (Henry's Fork Caldera, Figure 3), and is present under the source of the 600 ka Lava Creek eruption (Yellowstone Lake in Figure 3) [Smith *et al.*, 1982]. Based on the same data, the northeastern limit of the midcrustal intrusion is interpreted to be farther to the southwest in Smith and Arabasz [1991] than reported in Smith *et al.* [1982]. This discrepancy suggests some uncertainty about the location of the midcrustal intrusion in the critical juncture between the eastern SRP and the Yellowstone Plateau.

A time lag is implied between the arrival of the rhyolite magmas at the surface and the emplacement of mafic magmas at midcrustal levels. This sequence of events is further supported by the observed lack of significant amounts of basalt in and around Yellowstone Plateau. Quenching of the mafic magmas caused by the contrast in solidus temperatures between the extant siliceous magma chambers and the rising mafic magma probably contributes significantly to the positioning of the mafic magmas in the midcrust. Lachenbruch *et al.* [1976] have suggested that a similar process must have occurred beneath the Long Valley caldera of east-central California. They argue that a silicic magma chamber would have solidified long before the 2 m.y. of demonstrable activity of the caldera unless heat was continually added.

These data taken together present a scenario in which magmas from within the mantle rise to the Moho and are subsequently impeded by density and/or rheological constraints. The newly arrived magmas heat the base of the crust, melting the lower crustal rocks. These magmas in turn rise to shallow crustal levels and generate explosive rhyolitic eruptions. After residing at the base of the crust for some time, the mafic magma rises to midcrustal levels and then to the surface, creating the post-rhyolitic basaltic eruptions that dominate the landscape of the eastern SRP.

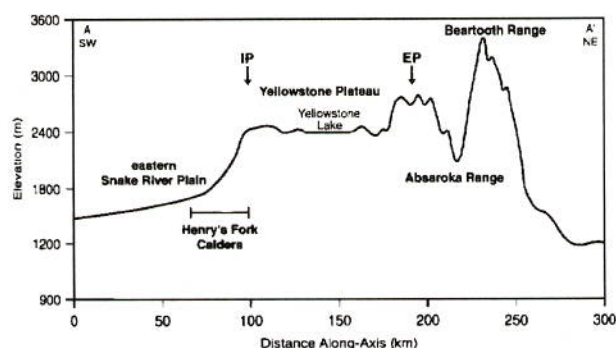


Fig. 3. Elevation plotted along-axis of the Yellowstone hotspot track. The trend of the cross-section is N55°E and is shown on Figure 2 as A-A'. The symbol IP marks the intersection of the Interior Parabola in Figures 1 and 2 and the trend of the cross-section; EP marks the intersection of the Exterior Parabola in Figure 1 and the cross-section. Both IP and EP are positioned near the apex of the respective parabolas. The Henry's Fork Caldera does not lie directly on this cross-section, but is a few kilometers to the northwest.

Leeman [1982] explained the timing of intrusions and the chemistry of volcanic rocks found in the circum-eastern SRP with a sequence of events like the one discussed above. In this paper, we extend the explanation to include the migrating parabolic patterns of fault displacement rates and seismicity occurring in the circum-eastern SRP during the last 16 Ma. The obvious surface features of the hotspot, volcanic eruptions and shallow intrusions, are not important to our explanation since rocks in this region have low brittle strength. Instead the rheology of the mid- to lower crust is more significantly affected by mafic intrusions replacing granitic rocks at midcrustal depths and subsequently crystallizing. The physics and timing of this process are the main topics of this paper.

OBSERVED PHYSIOGRAPHY

One of the most obvious features of the eastern SRP is the sharp contrast between the elevated mountains surrounding the eastern SRP and the elevation of the plain itself (Figure 2). Average elevation in the circum-eastern SRP exceeds 2000 m, yet the average elevation for the eastern SRP is less than 1500 m. The Yellowstone Plateau (Figure 3), a relatively low-relief feature, averages about 2500 m. Northeast of the Yellowstone Plateau elevation rises further with the numerous 3000+ m peaks of the Absaroka and Beartooth Ranges.

Smith and Shar [1974] first described the eastern SRP as a northeastward-migrating, wedge-shaped basin. Ruppel [1982] pointed out that there is a regional elevation arch that parallels the eastern SRP. Fritz and Sears [1989], Rodgers *et al.* [1990], and Pierce and Morgan [1990] all have suggested that a pattern of stream drainages that deflect toward the axis of the eastern SRP coincides with the arch observed by Ruppel.

Figure 3 shows the relief along the axis of the eastern SRP, including the topography just northeast of Yellowstone. This topographic profile includes the Absaroka and Beartooth Ranges, Precambrian to Eocene rocks thought to be uplifted during the late Cretaceous to early Tertiary Laramide Orogeny [see Birchfiel and Davis, 1975]. Recent apatite fission-track uplift studies indicate that the Beartooth Range (Figures 2 and 3) may have been continuously uplifting since between 15 and 5 Ma [Omar *et al.*, 1991].

Other mountain ranges bordering the eastern SRP have similar young uplift ages, all approximately timed to the arrival of the hotspot. For example, studies of tilted volcanic rocks show that the most recent uplift of the Teton Range (Figure 2) started between 2 and 4 Ma [Pierce and Morgan, 1990]. North of the eastern SRP the Beaverhead Mountains and the Lemhi Range (Figure 2) were uplifted some time between 10 and 6.5 Ma [Rodgers and Anders, 1990]. These mountains were all uplifted as an isostatic/elastic response to normal faulting in a manner similar to that suggested by Vening Meinesz [1950] and Zandt and Owens [1980]. Although a 350-km-diameter regional elevation rise associated with the position of the hotspot is observable [Suppe *et al.*, 1975], it is difficult to define that portion of total relief resulting from isostatic/elastic uplift induced by normal faulting and which portion of total relief is due to thermal expansion or previous tectonic events.

ISOSTASY AND THERMAL SUBSIDENCE IN THE EASTERN SRP

Brott *et al.* [1981] suggested that the depression associated with the eastern SRP is caused mostly by thermal contraction.

with isostatic loading by dense magmas intruded into the crust playing a less significant role. They incorporated small density changes at 175, 350, and 525 km from Yellowstone in their subsidence calculations to adjust the predicted elevation to actual elevation. We suspect that their coefficient of linear thermal expansion ($7.4 \times 10^{-5}/^{\circ}\text{C}$) is too great and that the isostatic effect of intruded magma has a much greater effect on subsidence.

The observed along-axis subsidence used by Brott *et al.* [1981] does not account for the 1 to 3 km of basalt added to the near-surface in the eastern SRP. Quaternary basaltic volcanism is evident along the entire length of the eastern SRP. The Henry's Fork Caldera on the southwest edge of the Yellowstone Plateau (Figure 3) marks the transition between the predominantly rhyolitic rocks of the Yellowstone Plateau and the predominantly basaltic rocks of the eastern SRP [Christiansen and Embree, 1987]. Southwest of the Henry's Fork Caldera basalts of the eastern SRP exhibit a southwestward thickening, yet only a few small Quaternary rhyolitic eruptives are present at the surface [Armstrong *et al.*, 1975]. Northeast of the Henry's Fork Caldera, within the Yellowstone Plateau, volcanism is predominantly rhyolitic, with only minor basaltic volcanism [Christiansen and Blank, 1972]. We suggest that the elevation drop from the Yellowstone Plateau to the eastern SRP (Figure 3) results largely from isostatic compensation associated with crustal loading by dense magma. As discussed above, the midcrustal velocity anomaly located directly under the region defined in Figure 1 as the collapse shadow, first appears in seismic refraction lines as an 8- to 10 km velocity anomaly under the Yellowstone Plateau [Smith *et al.*, 1982] which increases to 12 km thick under the eastern SRP [Sparlin *et al.*, 1982]. Later interpretations of these data suggest that the midcrustal anomaly thins to zero under the Yellowstone Plateau [Smith and Arabasz, 1991]. The along-axis thickening of the midcrustal velocity anomaly corresponds to the 1 km drop in elevation from the Yellowstone Plateau to the eastern SRP.

Studies of other hotspots like Hawaii can provide important constraints on the positioning of magmas. For example, the Hawaiian hotspot has a long-wavelength elevation anomaly perpendicular to its track that progressively increases in length down stream [see Sleep, 1987]. Superimposed on this long-wavelength feature is a shorter-wavelength downwarp caused by the surface loading of the volcanic edifice which has been used to assess the effective elastic thickness of the oceanic lithosphere [Crough, 1978; Detrick and Crough, 1978; Watts and ten Brink, 1989]. Within the circum-eastern SRP there is also a long-wavelength positive elevation anomaly [Suppe *et al.*, 1975; Pierce and Morgan, 1990; see Figure 2] which mimics the parabolas shown in Figure 1 and which is superimposed by a short-wavelength downwarp. We suggest that downwarping of the eastern SRP is analogous to the propagating line load of the Hawaiian Islands in that both downwarps are caused by the isostatic response to magmatic loading.

Despite the many similarities between the Hawaiian Islands and the eastern SRP, no similar earthquake pattern is associated with Hawaii. This may be because the strength of the oceanic lithosphere is significantly greater than the strength of the lithosphere in the circum-eastern SRP. Moreover, the middle of the Pacific Plate, on which the Hawaiian Island edifice sits, was not extending before the hotspot arrived, whereas the lithosphere in the circum-eastern SRP was extending before the arrival of the Yellowstone hotspot.

KINEMATIC MODEL OF FLOW FROM A PLUME

Sleep [1987] modeled the geometry of the flow of material away from a plume in the asthenosphere. Basically, the asthenosphere is dragged by the moving lithosphere and the underlying mantle, which is essentially fixed to the plume. The plume flow upstream from the hotspot is against the asthenosphere flow; as the upstream plume flow diminishes it eventually reaches a point where it stagnates. The crudely parabolic streamline through the stagnation point separates plume material on the inside from normal asthenosphere on the outside [Sleep, 1987]. Such parabolic stagnation lines also correspond to the observed parabola-like shaped positive elevation anomalies centered on the tracks of oceanic hotspots [Sleep, 1987, 1990].

We suggest that the outer parabola of the eastern SRP also marks a somewhat inboard position of a plume stagnation line. The elevation anomaly could mark the actual position of the stagnation line (Figure 4). As discussed above, the elevation anomaly is difficult to define in the circum-eastern SRP, yet elevation does drop off outboard of the exterior parabola, as seen in Figure 2.

THERMOMECHANICAL MODEL

We propose that the outwardly migrating exterior parabola is caused by an outward deflected hotspot mantle plume. The plume causes extension either by providing the gravitational energy

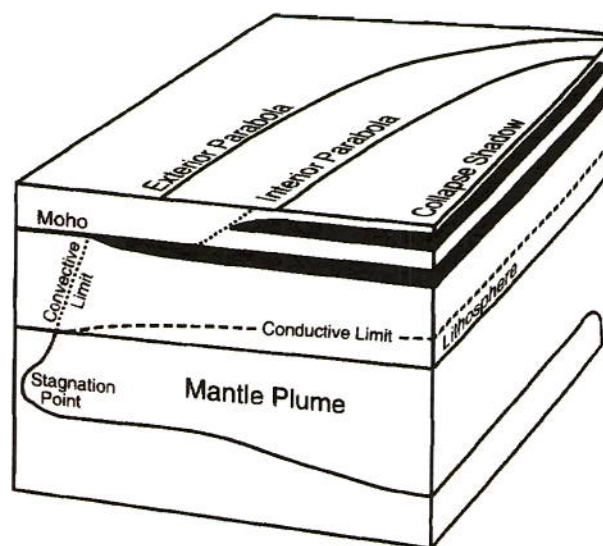


Fig. 4. Interpretive diagram of the lithosphere and upper asthenosphere under the circum-eastern Snake River Plain region. The depiction of the mantle plume, base of the lithosphere, and sub-Moho lithospheric intrusions is based on our interpretation of the emplacement of magma, assuming a hotspot model for the origin of volcanism in the circum-eastern Snake River Plain. The depth to the Moho and the shape and depth of the midcrustal intrusion are based on geophysical data [McCamy and Meyer, 1964; Braile *et al.*, 1982; Sparlin *et al.*, 1982]. The conductive limit describes the extent of thermal perturbation resulting from the presence of the asthenosphere plume; the convective limit roughly forms an envelope representing the delay in arrival times of rising magmas that results from a delay in the arrival of the outward deflected plume, which is the assumed source of lithospheric magmas [Sleep, 1987]. The stagnation point is the outer limit of lateral migration for the deflected plume. The stagnation line lies beyond the exterior parabola based on the elevation anomaly shown in Figure 2.

necessary to drive lower crustal channel flow or more directly through thermal weakening resulting from magma ascending to the upper mantle. On the other hand, the origin of the outwardly migrating region of fault quiescence or "collapse shadow" is more complicated. As we will suggest, it is the result of an interaction of the crust and upper mantle with intruded magmas. We will model the effects that such a crustal intrusion of mafic magma has on lithospheric strain rates and test whether or not the observed fault displacement rate patterns can be explained by magmatic intrusions.

Numerical Heat Flow Model

The finite-difference thermal model we use is constructed as a one-dimensional grid representing depth in the earth, on which temperature, volume heat content, and volume heat generation are assigned. Thermal conductivity and heat flow are defined for intermediate points. At various time intervals changes are made at those selected node points thought to best represent intruded mafic magmas. New values are subsequently calculated for all node points iteratively until the next change is made. From *Cawthorn* [1975], the heat content, Q , of a partial melt is the sum of the specific heat and the latent heat

$$Q = \rho c T + L f \quad (1)$$

where ρc is the volume specific heat, T is the temperature, L is the latent heat of melting per volume and f is the fraction of melt. Because the aspect ratios of the midcrustal eastern SRP intrusion are high and the time lengths involved are short (< 5 m.y.), the results will not significantly differ from those of a more complicated three-dimensional model (see *Gettings* [1988] for a three-dimensional model of a midcrustal intrusion).

Our numerical method treats conductive heat transfer and convective heat transfer in separate steps. The heat flow equation for a fixed medium is solved explicitly. Temperature is defined at N nodes, each separated by vertical distance Δx . The heat flow during a time step Δt is obtained from the temperatures of n and $n+1$ node points at the beginning of that time step

$$q_{n+1/2} = -k_{n+1/2} \frac{T_{n+1} - T_n}{\Delta x} \quad (2)$$

where the subscript $n+1/2$ indicates that conductivity k and heat flow are defined between the temperature nodes.

The change of enthalpy per volume Q at the n^{th} node point is given by

$$\Delta x \Delta Q_n = (q_{n-1/2} - q_{n+1/2} + H_n \Delta x) \Delta t \quad (3)$$

where H is heat generation per unit volume. This formulation conserves energy because the heat removed by conduction from the n^{th} node by $q_{n-1/2}$ is added to the $n-1^{\text{th}}$ node. Since this formulation becomes inaccurate and eventually unstable if the time step is too long, we used a time step of 2500 years and a nodal spacing of 500 m. The quantity $\kappa \Delta t / \Delta x^2$, where $\kappa \equiv k / \rho c$ is the thermal diffusivity, is 0.3, which is safely below the instability limit of 0.5 [e.g., *Richtmyer and Morton*, 1967, p. 189].

At the end of each time step, the enthalpy at each node is updated by adding ΔQ_n . New temperatures are obtained from this enthalpy. For temperatures below the solidus, the enthalpy is simply

$$Q_n(T_n) = \rho_n C_n T_n \quad (4)$$

where $\rho_n C_n$ is the volume specific heat of the n^{th} node. Simple relationships are also used for temperatures above the solidus. Our objective is to compute total yield strength of the lithosphere. It is not necessary to obtain precise temperatures near and above the solidus, where yield strength is negligible. It is important, however, to keep track of the total energy associated with latent heat of magma that eventually freezes. Between the liquidus and the solidus melt fraction is assumed to be linear with temperature. This allows freezing of basalt and melting of granite to proceed in a numerically convenient manner. The enthalpy is thus

$$Q_n(T_n) = \rho_n C_n T_n + L_n \left[\frac{T_n - T_s(n)}{T_l(n) - T_s(n)} \right] \quad (5)$$

where T_l and T_s are liquidus and solidus temperatures and L_n is the volume specific heat. The liquidus is only slightly exceeded in the calculations. For convenience the enthalpy for temperatures above the liquidus is assumed to be

$$Q_n(T_n) = \rho_n C_n T_n + L_n \quad (6)$$

A more sophisticated method would be necessary to investigate the petrology of intrusions and lavas. The convective motions associated with basalt intrusion are modeled by changing material between nodal points after a series of conductive time steps. The scheme conserves material and energy except for adiabatic effects, which are ignored. For example, the intrusion of 500 m of fresh basalt at 30-km depth is accomplished by changing the temperature and material properties of the nodal centered at 30 km to fresh basalt, which then displaces the assigned temperatures and material properties of each node below 30 km one node downward. Movement of 500 m of basalt between 30 km and 10 km is accomplished by raising the 30-km grid to 10 km and then lowering each node between 10 km and 29.5 down one node. The fresh basalt is assumed to come from below the domain of the calculation. That way the details of melting in the mantle do not have to be considered. Our method also requires that a minimum of one grid point be moved at one time. The details of the intrusion process are circumvented since hot rocks have little strength.

Magma Loading

We assume simple initial conditions and boundary conditions. An initial geothermal gradient is assumed with a steady-state conductive gradient overlying a 1370°C isothermal region at the intrusion temperature of basalt. This provides a realistic geothermal gradient in the crust and uppermost mantle, where significant strength exists. The unrealistic intersection of the isothermal and conductive gradients is unimportant, as little strength exists at those depths. The natural boundary condition of constant temperature 0°C is used for the free surface. An artificial bottom boundary of no heat flow is imposed at 60-km depth, far enough down that it does not affect the calculation in the shallower regions of interest.

Magma loading is performed to mimic the loading of the lithosphere based on geophysical and geochemical data from both the eastern SRP-Yellowstone and the Hawaiian Islands. In the salient loading history of Hawaii and the eastern SRP-Yellowstone, as discussed above, mafic magma intruded into the lithosphere rises until it encounters a buoyancy [*Leeman*, 1982, 1988] or rheological barrier at the base of the crust. After a time lag, magma rises to midcrustal depths. We choose to ignore the

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where H is heat generation per unit volume. This formulation conserves energy because the heat removed by conduction from the n^{th} node by $q_{n-1/2}$ is added to the $n-1^{\text{th}}$ node. Since this formulation becomes inaccurate and eventually unstable if the time step is too long, we used a time step of 2500 years and a nodal spacing of 500 m. The quantity $\kappa \Delta t / \Delta x^2$, where $\kappa \equiv k / \rho c$ is the thermal diffusivity, is 0.3, which is safely below the instability limit of 0.5 [e.g., Richtmyer and Morton, 1967, p. 189].

At the end of each time step, the enthalpy at each node is updated by adding ΔQ_n . New temperatures are obtained from this enthalpy. For temperatures below the solidus, the enthalpy is simply

$$Q_n(T_n) = \rho_n C_n T_n \quad (4)$$

where $\rho_n C_n$ is the volume specific heat of the n^{th} node. Simple relationships are also used for temperatures above the solidus. Our objective is to compute total yield strength of the lithosphere. It is not necessary to obtain precise temperatures near and above the solidus, where yield strength is negligible. It is important, however, to keep track of the total energy associated with latent heat of magma that eventually freezes. Between the liquidus and the solidus melt fraction is assumed to be linear with temperature. This allows freezing of basalt and melting of granite to proceed in a numerically convenient manner. The enthalpy is thus

$$Q_n(T_n) = \rho_n C_n T_n + L_n \left[\frac{T_n - T_s(n)}{T_l(n) - T_s(n)} \right] \quad (5)$$

where T_l and T_s are liquidus and solidus temperatures and L_n is the volume specific heat. The liquidus is only slightly exceeded in the calculations. For convenience the enthalpy for temperatures above the liquidus is assumed to be

$$Q_n(T_n) = \rho_n C_n T_n + L_n \quad (6)$$

A more sophisticated method would be necessary to investigate the petrology of intrusions and lavas. The convective motions associated with basalt intrusion are modeled by changing material between nodal points after a series of conductive time steps. The scheme conserves material and energy except for adiabatic effects, which are ignored. For example, the intrusion of 500 m of fresh basalt at 30-km depth is accomplished by changing the temperature and material properties of the nodal centered at 30 km to fresh basalt, which then displaces the assigned temperatures and material properties of each node below 30 km one node downward. Movement of 500 m of basalt between 30 km and 10 km is accomplished by raising the 30-km grid to 10 km and then lowering each node between 10 km and 29.5 down one node. The fresh basalt is assumed to come from below the domain of the calculation. That way the details of melting in the mantle do not have to be considered. Our method also requires that a minimum of one grid point be moved at one time. The details of the intrusion process are circumvented since hot rocks have little strength.

Magma Loading

We assume simple initial conditions and boundary conditions. An initial geothermal gradient is assumed with a steady-state conductive gradient overlying a 1370°C isothermal region at the intrusion temperature of basalt. This provides a realistic geothermal gradient in the crust and uppermost mantle, where significant strength exists. The unrealistic intersection of the isothermal and conductive gradients is unimportant, as little strength exists at those depths. The natural boundary condition of constant temperature 0°C is used for the free surface. An artificial bottom boundary of no heat flow is imposed at 60-km depth, far enough down that it does not affect the calculation in the shallower regions of interest.

Magma loading is performed to mimic the loading of the lithosphere based on geophysical and geochemical data from both the eastern SRP-Yellowstone and the Hawaiian Islands. In the salient loading history of Hawaii and the eastern SRP-Yellowstone, as discussed above, mafic magma intruded into the lithosphere rises until it encounters a buoyancy [Leeman, 1982, 1988] or rheological barrier at the base of the crust. After a time lag, magma rises to midcrustal depths. We choose to ignore the

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effects of rhyolitic plutons that rise from the base of the crust to the near-surface, because these plutons do not occupy a large area and have intruded into a region of the crust that has little relative strength.

In our model, magma rising into the midcrust is delayed by a phase shift of 500,000 yr. This phase shift is selected because the source of the 600 ka Lava Creek eruption is underlain by a velocity anomaly thought to be a mafic intrusion [Smith *et al.*, 1982] and the first magmas reaching the surface are thought to be of lower crustal origin uncontaminated by mafic magmas [Doe *et al.*, 1982]. Shifts in the lag time are significant only in that they increase or decrease the recovery time by approximately the amount of the shift. Midcrustal intruded magmas are taken from the center of the previous 125,000 yr subcrustal intrusion in a conservative effort to maximize the temperature and maintain temperature well above the solidus.

The amount of magma added to the midcrust is based on the size and estimated density corresponding to the high-velocity anomaly under the eastern SRP [Sparlin *et al.*, 1982; Braile *et al.*, 1982]. The total thickness of the modelled midcrustal intrusion is dependent on an assumed range of densities of $\rho = 3.2, 3.1$ and 2.9 g/cm^3 for intruded material. This range of densities results in a corresponding range of intrusion thicknesses of 4.5, 7.5 and 11 km, assuming the 6.5 km/s velocity anomaly corresponds to a 2.88 g/cm^3 density as suggested by Sparlin *et al.* [1982].

Intrusion into the midcrust is also staggered in time and thickness to approximate a Gaussian distribution. Emplacement depths were originally selected at random with the restriction that all final emplacement depths fall in the 8 to 20-km depth range observed for the midcrustal velocity anomaly. Several different emplacement schemes were tried; they yielded no significant difference in the calculated value of strain rate. The subcrustal intrusion thickness was first set assuming that there is a subcrustal intrusion under the eastern SRP similar to that observed under Hawaii, which is estimated to be about 8 km [Watts *et al.*, 1985; ten Brink and Brocher, 1987]. The resulting calculations, to be discussed in the Results section, suggest that, after some minimum threshold, the thickness of the underplated layer has little effect on strain rates in the short time span (<16 m.y.) associated with passage of the Yellowstone hotspot.

LITHOSPHERIC YIELD STRENGTH

Lithospheric strain rates for a fixed value of horizontal force are determined on the assumption that yield strength can be partitioned into friction-dependent rock strength and rock strength that is exponentially dependent on temperature [Brace and Kohlstedt, 1980; Sibson, 1982; Meissner and Strehlau, 1982; Smith and Bruhn, 1984]. This is likely an oversimplification [discussed in Hobbs *et al.*, 1986; Strehlau, 1986; Scholz, 1988] since there is a depth dependent semi-brittle region that is characterized by both brittle and nonlinear viscous flow. Because this type of behavior is currently beyond our ability to quantitatively model, we will instead use a discrete transition from brittle, frictional behavior to nonlinear viscous flow. In the brittle field we assume the Coulomb fracture criterion

$$|\tau| = S_0 + \mu \bar{\sigma}_n \quad (7)$$

where S_0 is the shear strength, μ the coefficient of internal friction, $\bar{\sigma}_n$ the normal stress and τ the shear stress. Substituting the general equations for biaxial stress

$$\sigma_n = 1/2 (\sigma_1 + \sigma_3) + 1/2 (\sigma_1 - \sigma_3) \cos \beta \quad (8)$$

$$\tau = 1/2 (\sigma_1 - \sigma_3) \sin \beta \quad (9)$$

where β is the angle between σ_1 and the normal to the plane of resolved stresses, yields equation (10) for rock strength as a function of depth assuming that μ , the coefficient of internal friction, is independent of rock type [Byerlee, 1968, 1978]

$$\sigma_1 = 2S_0 [(\mu^2 + 1)^{1/2} + 1] + \sigma_3 [(\mu^2 + 1)^{1/2} + 1]^2 \quad (10)$$

Because this paper deals only with extension, the value of σ_1 is assumed to be lithostatic. We also assume a similar value of deviatoric stresses at depth as Brace and Kohlstedt [1980] do, and we assume pore pressures to be hydrostatic. We have included the effects of pore pressure resulting in a gradient of 10 MPa/km in the brittle field. Although small amounts of water have significant effects on the rheology of rock [Tullis and Yund, 1980], we have used, in a manner similar to Buck [1991], dry quartz and anorthosite for crustal rheologies. At any given depth, strength is assigned to be the lesser value calculated from the power law creep equation (11) and by frictional dependent strength, equation (10), the lesser calculated value is used to assign a strength value for that depth. The power law creep equation is

$$(\sigma_1 - \sigma_3) = \left[\frac{\dot{\epsilon}}{A \exp\left(\frac{Q}{RT}\right)} \right]^{1/n} \quad (11)$$

where $\dot{\epsilon}$ is the strain rate, T is the absolute temperature, R is the gas constant, A , Q and n are material constants. The function corresponding to the lesser value of equation (10) and (11) is then integrated over the appropriate depth range

$$\int_0^h (\sigma_1 - \sigma_3) dx \quad (12)$$

where h is the depth to the region beyond which strength is negligible.

The parameters considered in this model are initial geothermal gradient, lithologies, thermodynamic and rheological properties of respective rock types, initial crustal thickness, total amount of magma intruded into the crust and upper lithosphere, timing of intrusions, temperature of intrusions, and the size and relative positioning of crustal intrusions. Of these parameters, we successively choose one that proves to be sensitive, hold the rest constant, and allow the sensitive variable to vary. The most sensitive variables are: the amount of magma intruded into the midcrust, the initial geothermal gradient, rheological properties of the crust, and initial crustal thickness.

Material constants used in strength determinations are selected on the basis of their ability to model the rheological properties of respective lithospheric rock types and on the availability of empirical data on these rock types. The most difficult rheological property to model is that of the lowermost crust, but it is likely that the base of the crust retains little strength. We therefore default to the quartzo-feldspathic rheological properties used for the upper crust. Midcrustal intrusion rheological properties were calculated using data from studies on diabase discussed below.

Strain rate. Strain rates are mapped by calculating integrated yield strength for a fixed value of $\dot{\epsilon}$ in equation (11). These values are plotted on integrated yield strength versus time plots, as shown in Figure 5. An amount of extension presumed to be reasonable in the circum-eastern SRP confines the range of strain rates to between $\dot{\epsilon} = 10^{-14}/\text{s}$ and $\dot{\epsilon} = 10^{-16}/\text{s}$. Strain rates of $\dot{\epsilon} = 10^{-14}/\text{s}$ to $\dot{\epsilon} = 10^{-16}/\text{s}$ range from 3% to 300% in 10 m.y. These strain

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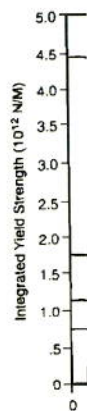


Fig. 5. different yielding over the amount a km crust crustal in of $\dot{\epsilon} = 10^{-14}$ 26 km. A effect on

rates correspond to strain rate determinations from seismic moment tensors for the northeastern Basin and Range province [Eddington *et al.*, 1987] and to those suggested for the Basin and Range province by Mercier [1980].

Figure 5 shows that there is little difference in relative yield strength as a function of time over this range of strain rates, yet a strain rate of $\dot{\epsilon} = 10^{-12}/s$ significantly changes the integrated yield strength profile. For a very rapid strain rate of $\dot{\epsilon} = 10^{-12}/s$ (Figure 5) there are values of total horizontal force for which the increase in strain rate is never recovered. For example, if there is a regional tectonic force of 3×10^{12} N/m acting on the northeastern SRP, then after 500,000 years the strain rate will always be greater than $\dot{\epsilon} = 10^{-12}/s$. For a more reasonable strain rate of $\dot{\epsilon} = 10^{-15}/s$, and an assumed tectonic force of 10^{12} N/m, the strain rate increases until about 2 m.y. and then decreases until it is fully recovered at about 5 m.y. Because $\dot{\epsilon} = 10^{-14}/s$ to $\dot{\epsilon} = 10^{-16}/s$ appears to be a reasonable range of strain rates, a median strain rate of $\dot{\epsilon} = 10^{-15}/s$ was chosen for our mapping of strain rates in Figures 6 and 7. Strain rates can be plotted for other values in Figures 6 and 7 as is done in Figures 5 and 8, but this is not done because the density of lines that would be present in these diagrams would be overwhelming.

The irregularities in the strain rate curves (the "wavy" pattern) are the result of uneven loading of magma. The instantaneous effect of loading magma is to increase the yield strength of the lithosphere by moving material downward. Subsequently the strength is reduced, as the intruded magmas heat the surrounding rock and the rock, driven to greater depths, is heated as the geothermal gradient attempts to re-equilibrate. Fluctuations in the strain rate curves occurring after 2 m.y. are the result of temperature differences between successively spaced time steps.

Initial thermal structure. It is not known what the geothermal gradient under the area that was to become the eastern SRP was at 16 Ma. Because the structure of the initial geothermal gradient plays a crucial role in the results of our thermal modeling, we have chosen a range of geothermal gradients which should

encompass the probable lithospheric geothermal gradient at the time of first intrusion.

A representative initial geothermal gradient for a stable continental interior is assumed to be $.023^\circ\text{K/m}$ (labeled "low initial geotherm" on accompanying diagrams) at $h = 0$. The initial geothermal gradient is then altered by increased heat production from radioactive elements in the upper crust. The distribution of radioactive heat production is

$$H(x) = -H \exp(-x/x_c) \quad (13)$$

where x_c is the characteristic depth (assumed to be 10 km) x is depth, and H (Table 1) is heat production. An initial value of $.033^\circ\text{K/m}$ is used to represent the high-end member (labeled "high initial geotherm" in accompanying diagrams) value, which is close to the "characteristic" geothermal gradient for the Basin and Range province assumed by Lachenbruch and Sass [1977]. If extension occurred in the circum-eastern SRP before 16 Ma, then it is likely that a higher geothermal gradient than $.023^\circ\text{K/m}$ is representative. At present, heat flow data from the circum-eastern SRP show values of 190 mW/m^2 for the center of the eastern SRP and 120 mW/m^2 on the margin of the SRP [Brott *et al.*, 1981]. These values are high, and if correct, reflect the mid-to-upper crustal residual heat from intrusions related to the hotspot.

The geographic region of greatest importance to our modeling of the thermal and mechanical effects of the hotspot lies within the "active region" (Figure 1). Not enough heat flow data exist from this region, excluding the Yellowstone Plateau, to be useful for our modeling. We presume that heat flow values from the active region, excluding the Yellowstone Plateau, are lower than the "characteristic" Basin and Range province values [see Brott *et al.*, 1981]. This may also be reflected by the greater crustal thickness within the active region [McCamy and Meyer, 1964; Braille *et al.*, 1982; Sparlin *et al.*, 1982]. Because we are not sure what the thermal structure of the lithosphere was before the arrival of the hotspot, an average value between the high and low values for the geothermal gradients is used to examine how an intermediate geothermal gradient affects the conclusions of our modeling. The resulting geothermal gradient is 0.028°K/m and is labeled "intermediate geothermal gradient" on accompanying diagrams.

Thermodynamic properties of intruded magma. Magma intrusion temperature is estimated to be 1370°C . This value is probably too high for magma intruded into the midcrust but reasonable for magma intruded into the upper mantle from a hot plume. Because the model is most sensitive to the temperature of intrusion in the midcrust, we have chosen a temperature that will retard the crystallization time for intruded magma, and will result in the greatest transfer of heat by conduction to the surrounding

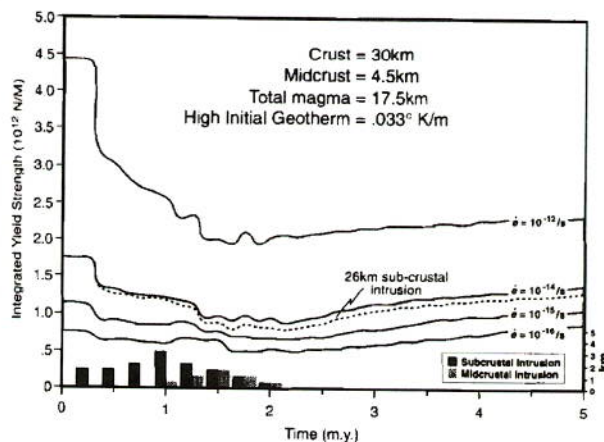


Fig. 5. Plot of integrated yield strength of the lithosphere for four different strain rates. The horizontal force required to cause extensional yielding of the lithosphere is determined by integrating the yield strength over the thickness of the lithosphere. The solid columns represent the amount and timing of mafic magmas intruded below the Moho for a 30-km crust. The shaded columns represents the amount and time of midcrustal intrusions. The dashed line is the strength profile for a strain rate of $\dot{\epsilon} = 10^{-14}/s$, where the total subcrustal intrusion is increased from 13 to 26 km. As indicated, doubling the amount of subcrustal intrusion has little effect on the total strength in 5 m.y.

TABLE 1. Constants

Symbol	Name	Value
ρ_c	volume specific heat	4.2 MW/m^3
L_i	latent heat	1260 MW/m^3
k	thermal conductivity	$4.0 \text{ W/m/s}^\circ\text{K}$
R	universal gas constant	$8.31/\text{mole}^\circ\text{K}$
H	radioactive heat source	$4.0 \times 10^{-6} \text{ W/m}^3$
T_s	granite solidus at $h=0$	850°C
	diabase solidus at $h=0$	1150°C
	solidus temperature gradient	2°C/km
T_l	granite liquidus at $h=0$	950°C
	diabase liquidus at $h=0$	1250°C
	liquidus temperature gradient	2°C/km

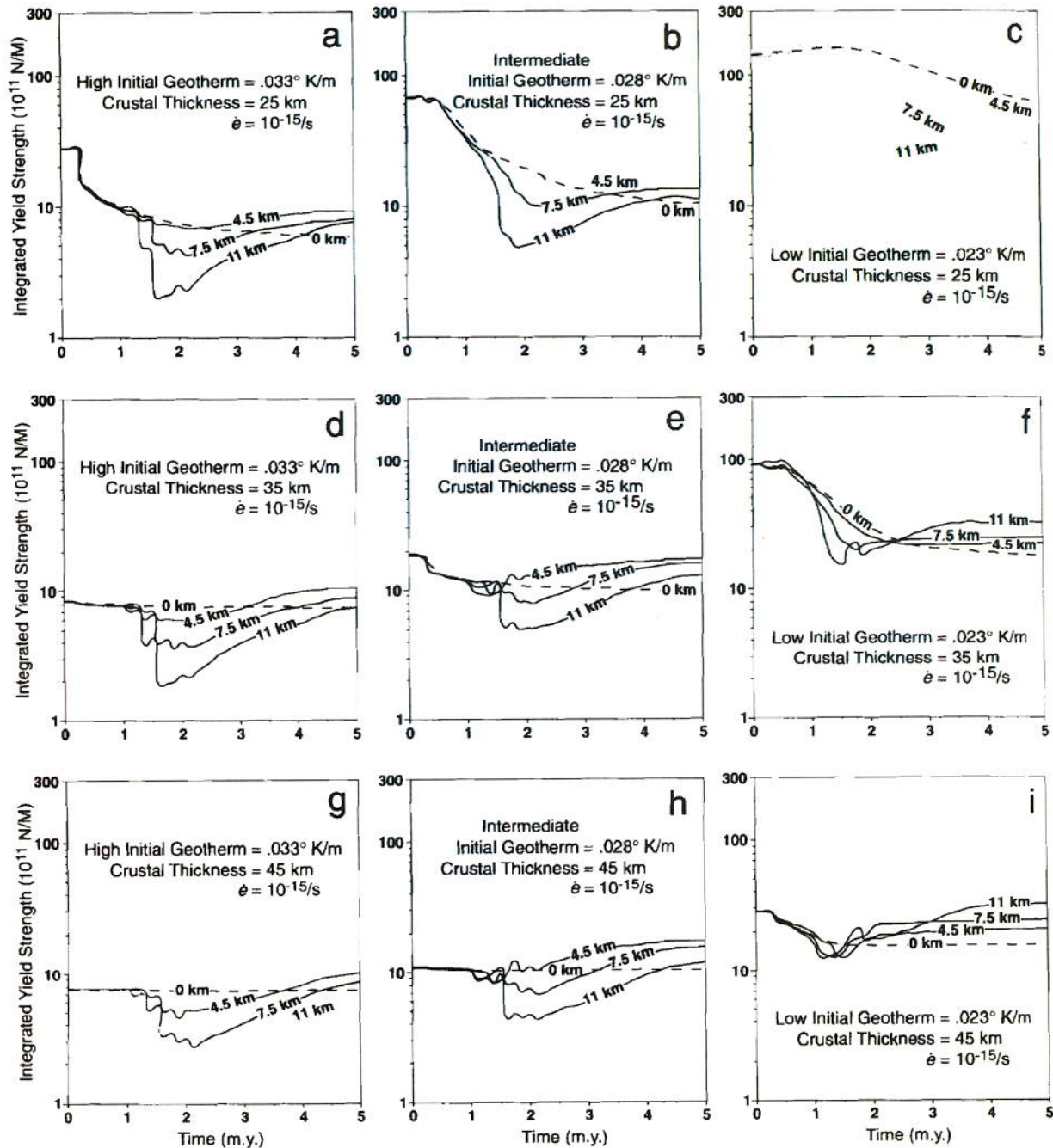


Fig. 6. Results from thermal modeling which assumes that the Westerly Granite [Carter *et al.*, 1981] represents the rheology of the crust. Each block comprises calculations for three amounts of midcrustal intrusions. All blocks within each column have the same initial geothermal gradient and all blocks within each row have the same initial crustal thickness. The distribution and timing of magma loading is shown in Figure 5. The amount of magma varies as indicated, but the peak loading is always 1 m.y. for the basal intrusion and 1.5 m.y. for the midcrustal intrusion. Subcrustal intrusion amounts are 13, 17.5, and 22 km, corresponding to midcrustal intrusions of 4.5, 7.5, and 11 km. Solid lines represent strain rate paths which allow for the observed patterns of deformation within the circum-eastern Snake River Plain; light dotted lines correspond to scenarios that fail to account for observed patterns. Horizontal forces required to initiate extensional yield are given in units of 10^{11} N/m.

rock. The adiabat was not considered because the errors in our initial assumptions exceed the approximately 20°C of cooling expected from pressure release.

The model is designed to add magma in varying amounts to the base of the crust at 125,000 yr intervals. Each successive 125,000 yr intrusion is placed in the center of the previous intrusion,

resulting in only a one-time interval juxtaposition of magma at the bottom of the crust and the top of the mantle. There is a phase shift of 1 m.y. between the first subcrustal intrusion and the first midcrustal intrusion. The heat and thermodynamic properties of rock representing subcrustal intrusion of magma is transferred after only 125,000 yrs residence time at the base of the crust, thus

Integrated Yield Strength (10^{11} N/m)

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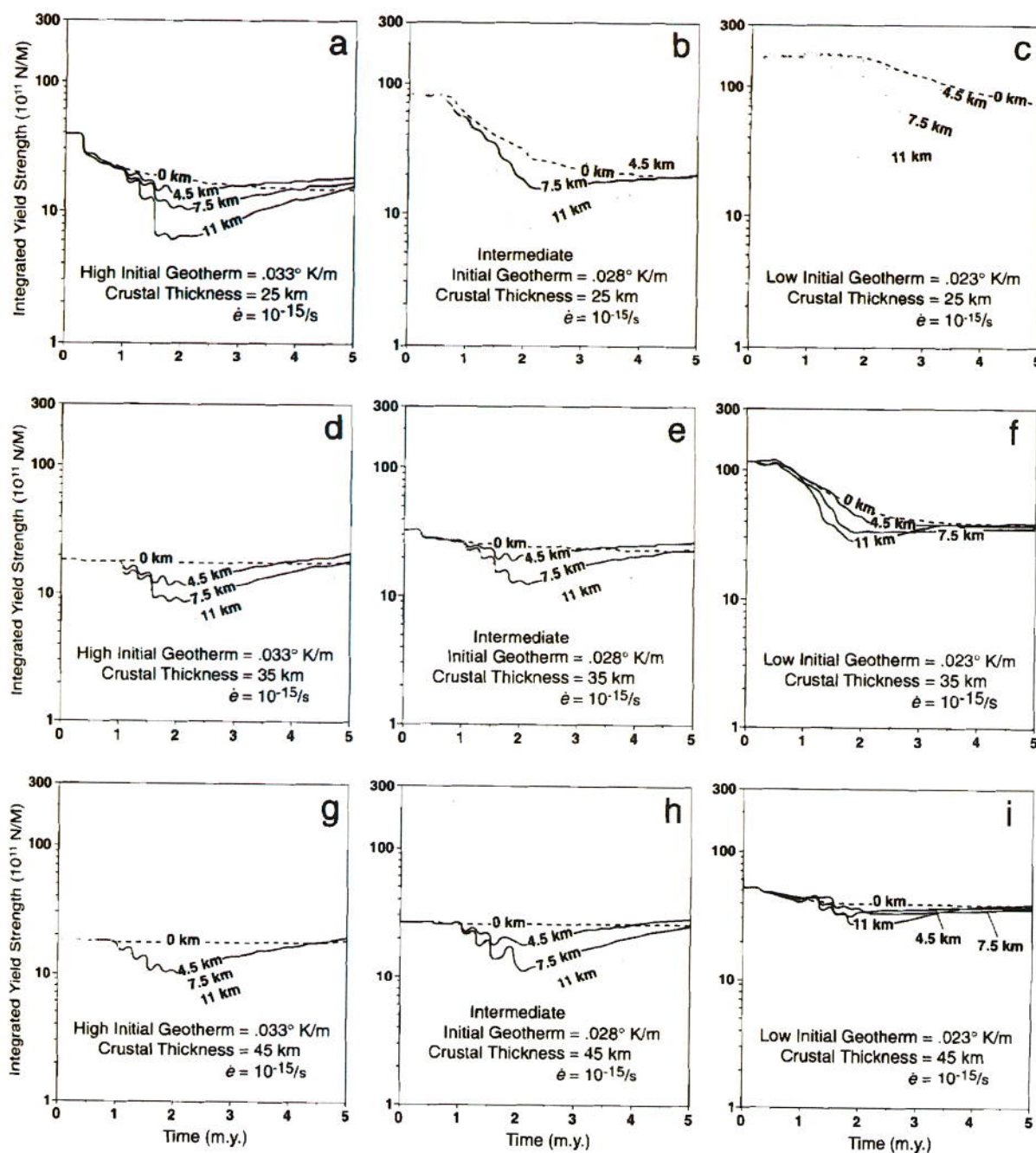


Fig. 7. Results from thermal modeling which assumes anorthosite [Shelton and Tullis, 1981] represents the rheology of the crust. Each block comprises calculations for three amounts of midcrustal intrusions. All blocks within each column have initial geothermal gradient and all blocks within each row have the same initial crustal thickness. The amount of magma varies as indicated, but the peak loading is always 1 m.y. for the basal intrusion and 1.5 m.y. for the midcrustal intrusion. Subcrustal intrusion amounts are 13, 17.5, and 22 km corresponding to midcrustal intrusions of 4.5, 7.5 and 11 km. Solid lines represent strain rate paths which allow for observed patterns of deformation within the circum-eastern Snake River Plain; light dotted lines correspond to scenarios that fail to account for observed patterns. Horizontal forces required to initiate extensional yield are given in units of 10^{11} N/m.

maximizing the temperature of intruded magma and hopefully simulating the process of magma intrusion associated with a continental hotspot. Transferred magma is always maintained above the solidus for the depth assigned. The volume specific heat value is $pc = 4.2$ MW/m³. Thermal conductivity is assumed

to be 4.0 W/m/s/ $^\circ$ K and the latent heat of crystallization is 1260 MW/m³ (Table 1).

Rheological parameters. Selection of rheological characteristics of materials representing the crust is important to the conclusions of our modeling of magmatic intrusions. At

little of the total lithospheric yield strength. This observation is commensurate with conclusions of previous workers, who suggest that thickened crust results in a weaker lithosphere [England and McKenzie, 1982; England, 1983; Sawyer, 1985; Houseman and England, 1986; Sonder *et al.*, 1987].

The difference between the results shown in Figures 6 and 7 is simply that the crust in Figure 6 is modeled as quartz-dominated rock and in Figure 7 it is feldspar-dominated. The family of curves representing the strain rate paths in Figure 6 shows greater sensitivity to the volume of magma intruded than does the family of curves for feldspar-dominated rheological properties in Figure 7, for similar boundary conditions. This is primarily due to the shallower depth of the transition from brittle, frictional behavior to nonlinear viscous flow as defined by a given thermal gradient for quartz-dominated rock as opposed to feldspar dominated rock. In all cases involving intrusion into the midcrust, the greater the thickness of intruded magma, the greater the increase in strain rate for a given horizontal force and the longer the length of time required to return to initial rates of strain.

In Figures 6 and 7, most 25-km and all 35- and 45-km crusts recover initial strain rates in less than 3 m.y. after the peak of the midcrustal intrusion. Many intrusion scenarios produce lower strain rates at some time after intrusion than existed before intrusion (e.g., Figures 6g-i). For 25-km crust with a low initial geothermal gradient (Figures 6c and 7c), the increase in strain rates as a result of intrusion is never reduced. In scenarios like these, involving initial low geothermal gradients and/or thin crust, the reduction of integrated yield strength in the upper mantle will swamp any contribution to total yield strength added by midcrustal intrusions. But, as will be discussed later, a return to initial or time zero strain rates is not necessary for our model to explain the migratory pattern of the parabolic distribution of seismicity and fault displacement rates.

DISCUSSION

Our thermomechanical model is constrained by the current understanding of the thermal and lithologic structure within the circum-eastern SRP, as well as by the assumed structure of this region prior to the arrival of the hotspot. Conditions necessary for our model to explain the migration of the parabolic distribution of seismicity and latest Quaternary faulting are as follows.

1. There must first be an increase in strain rates along-axis of the hotspot track that is of sufficient magnitude to match those observed within the seismically active areas of the circum-eastern SRP.

2. Strain rates must return to levels commensurate with those found outside the seismically active area. Furthermore, the turnaround time must be less than 2 to 3 m.y. That is, they must fall within the time necessary for the hotspot to move from a position under the apex of the inner parabola to the apex of the outer parabola (the strain rate at each apex should be approximately the same). The distance traversed should be between 70 and 120 km, depending on the assumed velocity of the North American Plate relative to the hotspot.

3. The outward migration of the outer parabola results from an increase in strain rates caused by the subcrustal intrusions or as discussed below, due to gravity driven lower crustal channel flow. If underplating the crust causes extension in the active region, then the pre-hotspot geothermal gradient must have been low enough for the circum-eastern SRP upper mantle to make a significant contribution to the total yield strength of the

lithosphere. The outward migration of the subcrustal intrusions, whether due to underplating or channel flow, are a prediction of dynamic models of plume/lithosphere interaction [Sleep, 1987].

4. The outward migration of the interior parabola corresponds roughly to the outward limit of midcrustal intrusions, as evidenced by seismic refraction/reflection studies [Braille *et al.*, 1982; Sparlin *et al.*, 1982]. Our modeling suggests that magma arrival at the distal portions of the midcrust is delayed. If it were not, the postintrusion decrease in strain rates would migrate inward, not outward. Based on the geometry of the current distribution of seismicity, over 90% of the outward cessation of activity occurs in less than 500,000 yrs.

Lower Limit Strain Rates

Although modeling is useful in assessing the relative strain rate profiles during magma intrusion, there are insufficient data to make a reasonable estimate of the state of horizontal force at the initiation of extension. The problem lies in the inability to assess strain rates for regions having little or no seismic activity and/or no evidence of deformed rocks upon which extensional strains can be determined. We suspect that strain rates of $\dot{\epsilon} = 10^{-17}/s$, determined from seismic moment data [Eddington *et al.*, 1987], may define a meaningful lower limit to strain rates [for further discussion of estimated strain rates for stable regions, see Lynch and Morgan, 1987]. However, based on the earthquake data from the region of active extension in the circum-eastern SRP, we have chosen a strain rate of $\dot{\epsilon} = 10^{-15}/s$ to track the history of extension shown in Figures 6 and 7. Although we have little control over the pre-hotspot strain rates, we can say that the initiation of extension is rapid and symmetrically distributed about the track of the hotspot [see Anders *et al.*, 1989].

Underplating the Crust

Thickness of the subcrustal intrusions has little additional effect on reducing strain rates for a given horizontal force once the intrusion exceeds 5-10 km. Increasing the thickness of the subcrustal intrusion from 13 to 26 km (dashed lines in Figure 5 corresponding to a 26-km subcrustal intrusion) has an almost imperceptible effect on the curve of constant strain rate for intervals less than 5 m.y. The timing of the peak in midcrustal intrusion affects only the relative temporal positioning of the highest strain rate value. Shifting the position of the peak intrusion does not significantly affect the strain rate recovery time although the recovery time may be affected by increasing the interval over which the midcrustal intrusions are emplaced.

In order for our model to explain the observed patterns of seismicity and fault displacement rates, there must be an increase in extension rates in the active region not caused by midcrustal intrusions. One possibility is that underplating of the crust with intruded magma affects extension rates substantially. To model this, calculations are performed in which only subcrustal intrusions are considered (dashed line labeled "0 km" in Figures 6 and 7). The results show that for high geothermal gradients and a thick crust extension rates increase minimally. This scenario is incompatible with the observation of crustal extension within the active region. If we assume that the circum-eastern SRP region had a uniform crustal thickness and geothermal gradient before the advance of the hotspot (which it most likely did not), then strain rate profiles that do not increase as a result of a basal intrusion cannot explain the patterns of enhanced seismicity.

Another category of strength calculation that can be discarded is that in which 3 to 5 m.y. after a midcrustal intrusion the strain

rate continues to increase or only moderately recovers. Examples of this are 25 km crust with a low initial geothermal gradient (Figures 6b (4.5 km) and 6c (4.5 km, 7.5 km); 7b (4.5 km), 7c (4.5 km, 7.5 km)). There are several examples of strain rate profiles that fail to return to pre-intrusion levels, yet exhibit a turn-around in total strength at some point (Figures 6a-6c and 6f; 7a-7c, 7f and 7i). A return to initial strain rates, as previously mentioned, is not a necessary condition because the initiation of extension most likely occurs below the state of horizontal force at time zero. The conditions necessary for our model to account for the migratory pattern of seismicity are the existence of a crossover point (labeled E in Figure 8) between lithospheric strain rate profiles for subcrustal and midcrustal intrusions only (dashed line in Figures 7a through 7i, and Figure 8), and a return of strain rates in less than 2 to 3 m.y. after midcrustal intrusion. The crossover points mark the time at which the calculated strain rates from midcrustal plus subcrustal intrusions exceed the calculated subcrustal-only intrusion strain rates. If there is no crossover point, a condition arises in which strain rates within the collapse shadow of the eastern SRP are always greater than those in the region of seismicity, a condition which is not consistent with the observed pattern of strain [Anders *et al.*, 1989, 1990]. The solid line labeled "10-15" in Figure 8 shows a typical scenario in which the necessary conditions exist, given an initial state of horizontal force between lines "A-B" and "C-D", for our model to explain the observed pattern of seismicity and migrating fault displacement rates in the circum-eastern SRP. Although underplating of the crust provides an explanation of the increased seismic activity off-axis of the hotspot track, another possible cause for increased seismicity in this region will be discussed below in the section on decoupling the crust.

In Figures 6 and 7 those initial conditions that can successfully account for the pattern of seismicity and faulting within the eastern SRP during the last 16 m.y. are represented by solid lines and those that we interpret as failures are represented by light dotted lines. Some scenarios appear better than others, because the relative increases and decreases in strain rates are large enough to allow more latitude on original assumptions. In general, those scenarios which are not dominated by an upper mantle contribution to yield strength are the ones most likely to fulfill the conditions necessary for the migrating parabolic distribution of seismicity and increased fault displacement rates. Those least likely to succeed are the unlikely combinations of thin crust and low initial geothermal gradient, or thick crust and high initial geothermal gradient. Eight of the 18 calculations in Figures 6 and 7 involving an 11-km midcrustal intrusion did not succeed in meeting our criterion, and those that did succeed required excessive time, often over 3 m.y., to return strain rates rapidly enough to successfully explain the observed deformation pattern within the circum-eastern SRP. This suggests that an 11-km intrusion is too great, which is possible because the 12-km density anomaly is likely to represent disseminated intrusions into granitic rock, not complete replacement of them.

Accumulated Strain Energy

Gettings [1988] has argued that midcrustal mafic intrusions may increase crustal strength to levels that permit localized strain concentrations, which in turn may account for loci of earthquakes in the southeastern United States. Gettings [1988], citing the work of Mitchell *et al.* [1984] on generation of differential stresses resulting from cooling intrusions, suggested that the mafic

intrusions can provide a mechanism for seismicity tens of millions of years after intrusion. The midcrustal intrusions associated with the eastern SRP are closely analogous to the one proposed by Gettings. The results of our study suggest that the duration of the increase in total yield strength is on the order of 5 m.y. or less, which is more rapid than Gettings calculated. This may be because we treated intrusions as small, 500-m-thick layers staggered in time and distributed within a 12-km zone in the midcrust, where Gettings treats intrusions as large pools of magma 20x20x5 km. The track of the hotspot has been intruded as long ago as 16 Ma, yet at present there is no seismicity [King *et al.*, 1987] except directly over the hotspot, suggesting no significant accumulation of strain energy from stresses generated by differential thermal contraction. Moreover, the magnitude of strain rates generated by thermal contraction [Mitchell *et al.*, 1984] is significantly lower than that generated by regional tectonics.

Extension and Decoupling the Crust

Although our model for the strain history of the circum-eastern SRP assumes that a lithospheric column extends uniformly, it is possible that a weak lower crust could flow independently of the middle to upper crust. Gans [1987], Buck [1988, 1991], Block and Royden [1990], and Kruse *et al.* [1991] have suggested that during extension the lower crust may strain at different rates than the middle to upper crust. Kruse *et al.* [1991] have suggested that gradients in gravitational potential energy or buoyancy forces drives lower crustal flow and that the movement of lower crustal material can in turn be treated as a channel flow problem. As first suggested by Suppe *et al.* [1975], there is a topographic anomaly associated with the Yellowstone hotspot (Figure 2). Thermal expansion of the lithosphere caused by a hotspot can produce such a parabola-shaped increase in topography [Sleep, 1987]. The excess gravitational energy provided by the topography surrounding the eastern SRP is sufficient [Kruse *et al.*, 1991] to drive the lower crust toward the collapse shadow. And, indeed, lower crustal flow in the circum-eastern SRP may account for the orientation of extension directions toward the eastern SRP (arrows in Figure 2). Lower crustal material flowing into the eastern SRP must be compensated for either by elevation changes or by migration of mass toward the highly extended Basin and Range province to the southwest. Lower crustal flow will create a subsidence gradient which is compensated for by faulting in the upper crust. Thermal contraction caused by cooling of the eastern SRP will contribute to lower crustal flow in two ways: first, along axis thermal subsidence will cause an elevation drop [Brott *et al.*, 1981] that will increase the buoyancy forces which drive the lower crustal flow; second, contraction will create space to accommodate flow toward the eastern SRP. Although a more quantitative assessment is necessary, lower crustal flow can potentially account for observed patterns in seismicity of the circum-eastern SRP. Again, our model is compatible with either lower crustal flow or upper mantle intrusion or a combination of these two mechanisms as an explanation of the off-axis extension.

Dueker and Humphreys [1990] use *P*-wave residual time data to suggest that the upper mantle in the off-axis active region is as cold (fast relative velocities) as the nonextending upper mantle ahead of the hotspot. If their interpretation is correct, then the off-axis active region is extending by some mechanism other than either magmatic intrusions into the upper mantle or thermal uplift generated by a mantle plume. Given a cold upper mantle in the

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off-axis active region, lower crustal flow could still account for the extension. Because, as discussed above, there are several variables that control topography in this region, the elevation pattern shown in Figure 2 may not serve as a definitive criterion for the dimension of a mantle plume. If elevated topography existed before the advent of the hotspot and if a significant amount of extension occurred along-axis of the hotspot track, then buoyancy forces could be generated without calling upon the thermal influences of the hotspot in the off-axis active region. With or without off-axis thermal influences, the region of the collapse shadow would not experience extension because of the large value of total yield strength created by middle and upper crustal intrusions. Although we prefer an off-axis thermal effect from the hotspot, the distinction between the two explanations for extension in the active region lies in better resolution of the topographic anomaly as well as better *P*-wave residual data.

Overpressuring by Magmatic Intrusion

Parsons and Thompson [1991] have argued that overpressuring from magmatic intrusions, such as those that produced the Great Rift, is responsible for the aseismicity of the collapse shadow. The Great Rift is the largest of several late Quaternary rift-type lava eruptions on the eastern SRP and is the only rift that extends across the eastern SRP [Kuntz *et al.*, 1982]. Also, the Great Rift has a trend consistent with extension within the northeastern Basin and Range province (Figure 2). Parsons and Thompson [1991] suggest that the feeder dikes to the rift eruptions regionally increase the least principal stress thus inhibiting failure in the upper crust. This mechanism would be likely to contribute to aseismicity of the collapse shadow, but in all likelihood it would not be solely responsible for it. In order to maintain the overpressure, new magmas must be constantly injected into the upper crust. The new magmas continue dilating the upper crust which in turn results in the aseismic extension of the eastern SRP.

It is important to note that there is a distinction between the region of the collapse shadow and the eastern SRP (see Figure 1). The area between the interior parabola and the eastern SRP has numerous faults that were once active and are now quiescent. With a few exceptions, these faults do not extend into the eastern SRP. Also, no dike structures are found in the region between the interior parabola and the eastern SRP, at least none that are exposed at the surface. Both models require midcrustal intrusions within the collapse shadow but the Parsons and Thompson model requires extension within the collapse shadow and ours does not. We argue that the test between these models as end-members in explaining the aseismicity lies in the predicted extension history of the faults which lay in the aseismic region between the eastern SRP and the interior parabola. Anders and Geissman [1983], Anders *et al.* [1989] and Anders [1990] have shown that a segment of a fault lying within the collapse shadow has a displacement rate two orders of magnitude below a segment of the same fault lying within the active region. Again, the two mechanisms may operate simultaneously in some regions of the collapse shadow, but only our mechanism explains the observed fault quiescence in that part of the collapse shadow that lies between the eastern SRP and the active region.

CONCLUSIONS

We have presented a model for magmatism and extension that we suggest explains reasonably the pattern of seismicity and latest Quaternary faulting within the circum-eastern SRP. Our

modeling depends on several assumptions, the most important of which are the timing and relative depth distribution of intruded magma. To model the magma intrusion process in the circum-eastern SRP we relied largely on the work of Leeman [1982, 1988] and on geophysical data from several studies done in the circum-eastern SRP [Braile *et al.*, 1982; Sparlin *et al.*, 1982; Smith *et al.*, 1982].

These studies suggest that magma rises through the upper mantle and ponds at or near the base of the crust because of the crustal density or rheology contrast between the magma and the lower crust. Subsequently, the lower crust melts and rises to the upper crust resulting in the violent rhyolitic eruptions that define the track of the hotspot. At some time after its ponding at the base of the crust some of the basaltic magma rises to the midcrust, where it is quenched by less refractory acidic magmas. Our modeling of this process suggests the strain rate, for a constant horizontal force condition, is most sensitive to changes in depth of the initial Moho, initial geothermal gradient, and amount of magma intruded into the midcrust.

The results of our modeling further suggest that the integrated yield strength of the upper mantle at the moment of first intrusion controls whether our model successfully explains the migratory pattern of seismicity and fault displacement rates in the circum-eastern SRP. Our model does not work when either the integrated yield strength of the lithosphere is not affected by subcrustal intrusions or when strain rates are not reduced after midcrustal intrusion. These situations arise when the crust is thick and the geothermal gradient is high, or when the crust is thin and the geothermal gradient is low. Neither of these situations is likely to have existed in the circum-eastern SRP during the last 16 Ma.

We suggest that magmatic intrusions associated with movement of the North American Plate over the Yellowstone hotspot are responsible for much of the historic seismicity found within the circum-eastern SRP. The presence of Neogene faulting throughout the northeastern Basin and Range province indicates that this area underwent extension prior to the introduction of hotspot magmatism. Thus the thermal activity associated with the hotspot is not the sole cause of circum-eastern SRP seismicity. Rather, the hotspot magmatism traversed an extending region that would have been seismically active even without the presence of the hotspot.

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