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Deep mantle forces and the uplift of the Colorado Plateau

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[1] We introduce a quantitative model of global mantle convection that reconstructs the detailed motion of a warm mantle upwelling over the last 30 Ma towards the interior of the southwestern USA from observed present-day mantle heterogeneity. The onset and evolution of uplift in the central Basin and Range province and Colorado Plateau during this time is determined by tracking the topographic swell due to this mantle upwelling. We show that: (1) the extension and basaltic volcanism (post 25 Ma) in the central Basin and Range coincides with the arrival and eastward progression of this upwelling, and (2) dynamic uplift of the southern Colorado Plateau, totaling about 1 km, transpired in the last 20 Ma. Since 10 Ma, the center of uplift continued northeastward from the southwestern rim of the plateau consistent with a young Grand Canyon model and eastward sweep of magmatism in the western Colorado Plateau. **Citation:** Moucha, R., A. M. Forte, D. B. Rowley, J. X. Mitrovica, N. A. Simmons, and S. P. Grand (2009), Deep mantle forces and the uplift of the Colorado Plateau, *Geophys. Res. Lett.*, *36*, L19310, doi:10.1029/2009GL039778.

1. Introduction

[2] Since the advent of plate tectonics, it has been speculated that the northern extension of the East Pacific Rise (EPR), specifically its mantle source, has been over-ridden by the North American Plate in the last 30 Myr [e.g., *Menard*, 1960; *Wilson*, 1973; *Jacobs et al.*, 1974; *Dixon and Farrar*, 1980; *Fletcher et al.*, 2007]. Consequently, it has also been postulated that the opening of the Gulf of California, the extension in the Basin and Range (BR) province, and the uplift of the Colorado Plateau (CP) are the resulting continental expressions of the over-ridden mantle source of the EPR. In contrast, entirely plate-based models infer no deeper mantle dynamics contribution to mid-ocean ridges and hence, view western US deformation to be plate-driven, for example by collapse of high topography and/or stress related to relative plate motions [e.g., *Humphreys and Coblentz*, 2007]. However, only qualitative models based solely on surface observations and heuristic, simplified conceptions of mantle convection have been used in support of (or against) either hypothesis. In this study we utilize a numerical model of time-dependent backward mantle convection that reconstructs the position of the EPR mantle source from present-day mantle heterogeneity and focus on its implications for the late Cenozoic tectonic evolution of the southwestern US and the uplift of the CP and adjoining regions.

[3] The CP is part of the Rocky Mountain orogenic plateau bounded by the BR province to the west and south, the Rio Grande rift and the Great Plains to the east and the southern and central Rocky Mountain orogen to the north-east and north (see Figure 1a) [e.g., *McMillan et al.*, 2000]. At the end of the Cretaceous the CP, as well as the Great Plains of the USA, was covered by the Western Interior Seaway.

[4] Today, the Late Cretaceous marine strata of the Western Interior Seaway rest at a mean elevation of about 2 km above sea level atop the CP – thus constraining the amount of post-Cretaceous uplift [*Spencer*, 1996]. As much as 600 m of the 2 km elevation gain is due to isostatic support of Cretaceous sediments associated with subduction-controlled continental tilting [*Mitrovica et al.*, 1989]. The remaining tectonic uplift (1.4 km) of this region has been the focus of considerable debate. The amount of upper crustal shortening in the CP (<few %) cannot explain this anomalous elevation via crustal thickening and Airy isostasy. Instead, models for multiple causes and times of uplift seem to be needed to explain the regional orogenic plateau's elevation. Arguments for uplift components include: (a) regional Laramide (80–40 Ma) uplift due to a mechanism that modifies, or delaminates, the underlying lithosphere by low angle or "flat slab" subduction of the Farallon plate beneath the North American plate [e.g., *Bird*, 1988; *Zandt et al.*, 1995], (b) more focused mid-Tertiary (40–25 Ma) components related to the foundering of the Farallon slab and the ignimbrite flare up [e.g., *Spencer*, 1996; *Humphreys et al.*, 2003; *Roy et al.*, 2009], and (c) regional middle-to-late Cenozoic (25 Ma – present) dynamic uplift driven by mantle convection (or a plume) [e.g., *Wilson*, 1973; *Jacobs et al.*, 1974; *Dixon and Farrar*, 1980; *Fitton et al.*, 1991; *Parsons et al.*, 1994; *Karlstrom et al.*, 2008; *Moucha et al.*, 2008a].

[5] In this regard, we note that recent tomographic imaging of mantle structure below La RISTRA seismic array reveals the presence of a warm mantle anomaly underneath the CP [*Sine et al.*, 2008]. Sine et al. interpret this anomaly in terms of upward, passive return flow generated by the foundering of the Farallon slab at about 40–20 Ma [*Humphreys et al.*, 2003] and *Karlstrom et al.* [2008] attribute this anomaly to small scale convection. In contrast, a recent global geodynamic model of present-day flow demonstrated that the CP currently overlies a strong

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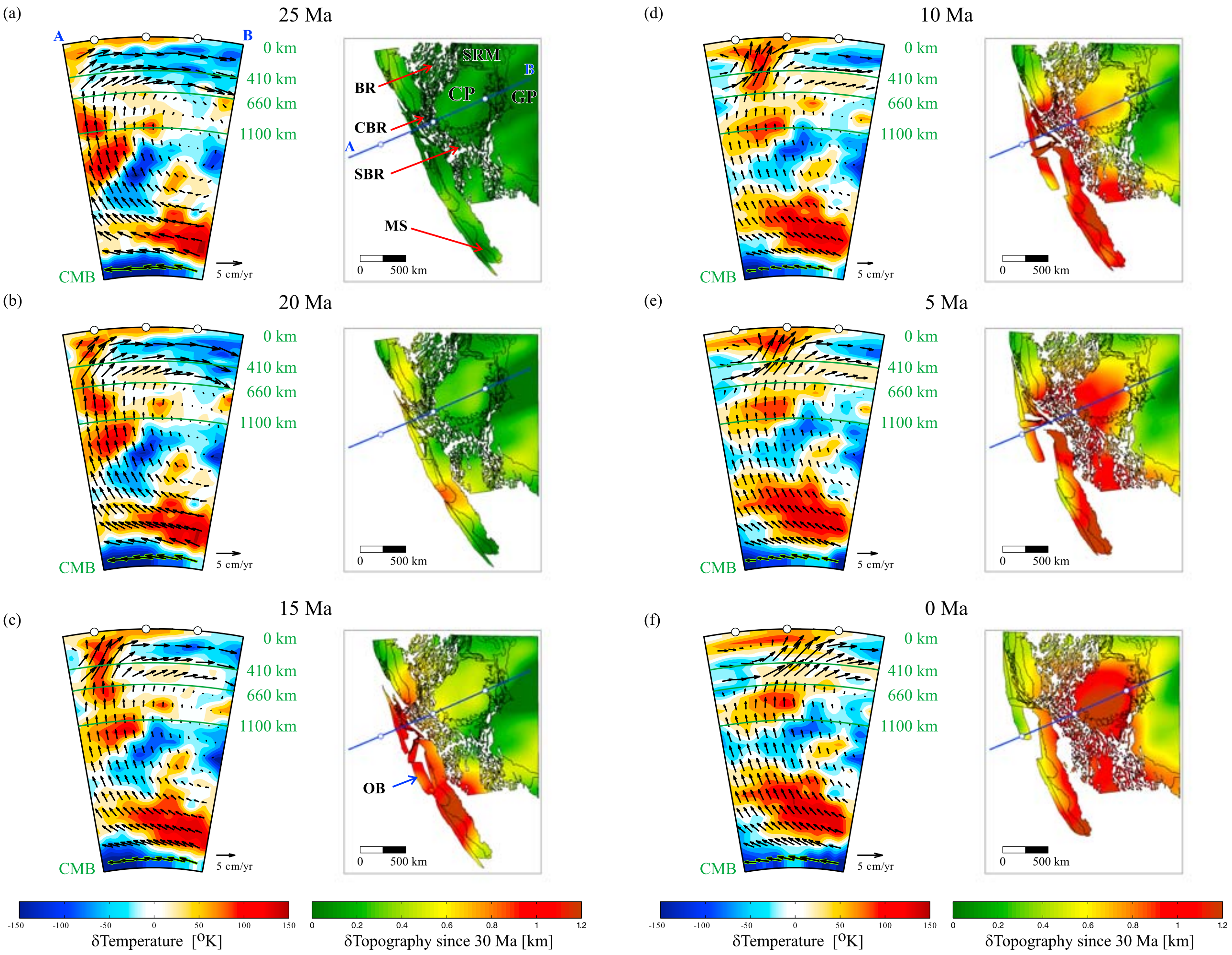

Figure 1. Evolution of dynamic topography in the southwestern USA in a fixed North-American reference frame relative to 30 Ma for (a) 25 Ma, (b) 20 Ma, (c) 15 Ma, (d) 10 Ma, (e) 5 Ma, and (f) 0 Ma. Rotations of individual blocks within the fixed North-American reference frame are obtained from tectonic reconstructions [*McQuarrie and Wernicke*, 2005]. Results are shown for viscosity V1 and the TX2007 density model and are relative to dynamic topography at 30 Ma (see auxiliary Figure S1). Radial cross-sections of the reconstructed mantle temperature variations from surface to core-mantle-boundary (CMB) along the line A–B, affixed to the rigid North American plate, are shown on the left for each time frame. Superimposed on these cross-sections are the corresponding mantle flow velocity vectors. The vectors are extracted from global flow field in the mantle's no-net rotation frame of reference along a cross section fixed relative to North America. BR, Basin and Range; CBR, Central Basin and Range; SBR, Southern Basin and Range; SRM, Southern Rocky Mountain orogen; CP, Colorado Plateau; GP, Great Plains; OB, Outer Borderland which contains the Patton Ridge; MS, Magdalena Shelf. The Rio Grand rift lies between the CP and the GP.

mantle upwelling that originates at the core-mantle boundary [*Moucha et al.*, 2008a]. Consequently the CP overlies a localized topography high within the western US.

[6] Recently, *Roy et al.* [2009] used a regional thermoelastic model of the lithosphere to argue that a transition from cooler mantle-wedge conditions beneath the plateau to hotter mantle temperatures, tied to the ignimbrite flare up, could be responsible for as much as 1.6 km of rock uplift since 40 Ma.

[7] There is consensus that the uplift of the CP followed a southwest-to-northeast progression. However, while various paleoaltimetry methods attempted to constrain the timing of uplift, little consensus has emerged on this issue [e.g., *Wolfe et al.*, 1998; *Pederson et al.*, 2002; *Sahagian et al.*, 2002; *McMillan et al.*, 2006; *Flowers et al.*, 2008; *Karlstrom et al.*, 2008].

2. Tomography Based Mantle Flow Model

[8] In this study we directly address this ongoing controversy by inferring the temporal evolution of global mantle flow from backward mantle convection, with a focus on the southwestern US. We quantify its effect on the CP and adjacent regions topography over the last 30 Ma. We present results from a global time-dependent numerical model of topography that is supported by convectively maintained vertical stresses generated by viscous-flow in the mantle (henceforth termed dynamic topography). These

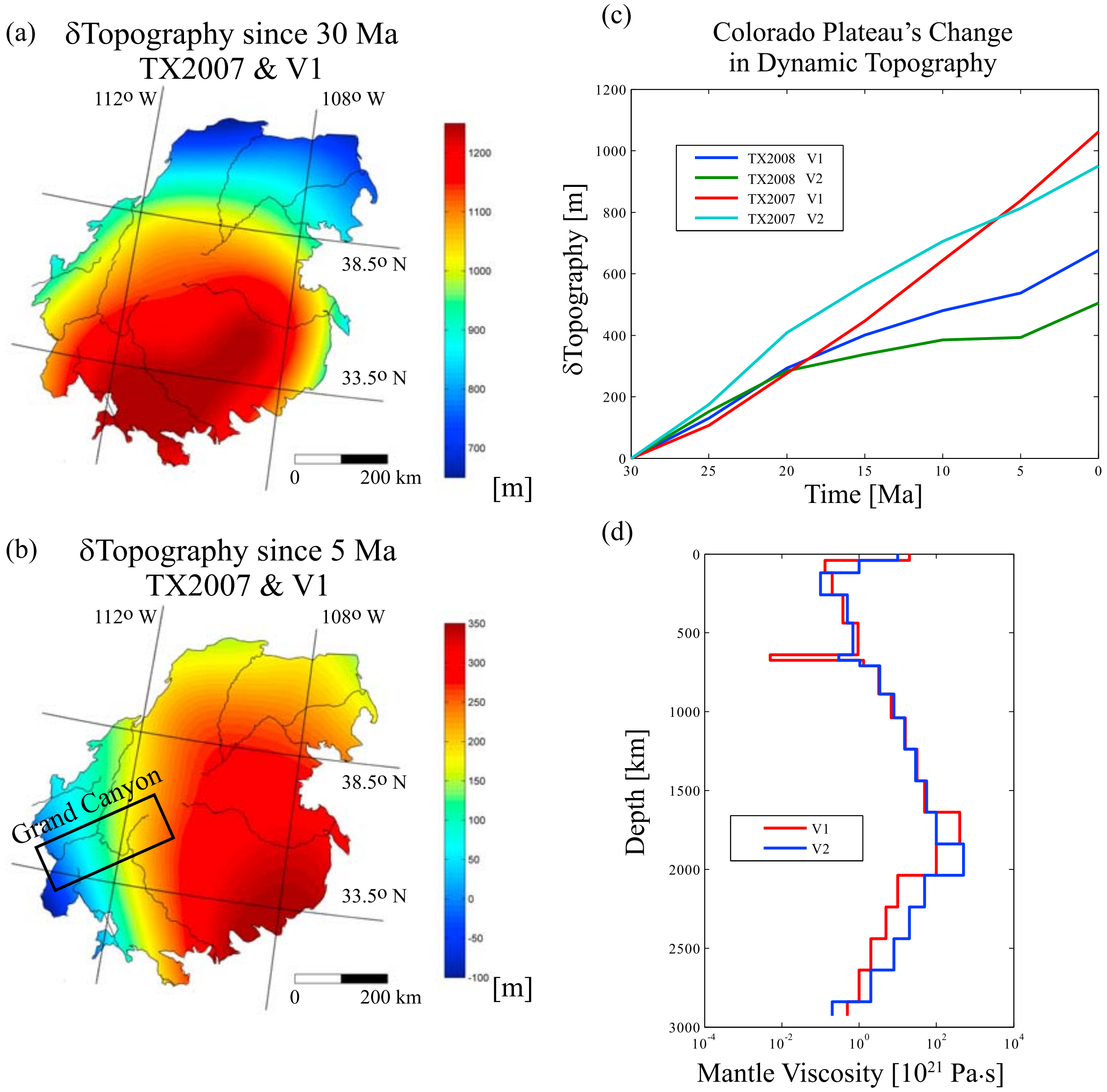

Figure 2. Change in Colorado Plateau dynamic topography with respect to (a) 30 Ma and (b) 5 Ma in a fixed North American reference frame accounting for the slight rotation of the Colorado Plateau [*McQuarrie and Wernicke*, 2005]. (c) Evolution of the average (area integral) dynamic topography change in the Colorado Plateau with respect to 30 Ma for each instant in time. (d) Plotted are the results for two versions of a global density model TX2007 and TX2008 and two viscosity models V1 and V2 (see text and auxiliary material).

vertical stresses originate from buoyancy forces residing in both the lithosphere and the sublithospheric mantle. The globally distributed present-day mantle and lithospheric density variations are inferred from joint seismic-geodynamic inversions that include mineral physical constraints on the scaling of seismic shear-wave speed to density [*Simmons et al.*, 2009]. A significant result of the joint seismic-geodynamic inversions is that the inferred scalings vary in all three dimensions and thus incorporate intrinsic changes in mantle chemistry and possible partial melting effects on density. These effects yield cratonic roots with near-neutral buoyancy and reductions of buoyancy in anomalously hot regions in the shallow mantle. The final density model provides an excellent fit to present-day geodynamic observables while preserving excellent fit to global seismic data. Herein we use two iterations of such density models – one termed 'TX2007', our preferred model, by *Simmons et al.* [2007] and the other termed 'TX2008' by *Simmons et al.* [2009] to quantify the inherent uncertainty of the starting model (see auxiliary material).[1]

[9] The convection simulation incorporates Newtonian rheology with a viscosity profile that is constrained by global joint inversions of convection-related surface observables and data associated with the response of the Earth to ice-age surface mass loading [*Mitrovica and Forte*, 2004]. This profile is labeled 'V1' (Figure 2d). Since the evolution of dynamic topography is sensitive to the adopted viscosity profile, we consider a second viscosity profile, labeled 'V2', that also fits same geodynamic and ice age observations as V1. Compared to V1, V2 is distinguished by a stiffer

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[1] Auxiliary materials are available in the HTML. doi:10.1029/2009GL039778.

lithospheric mantle, a lower asthenospheric viscosity, an absence of a low viscosity notch at the base of the upper mantle, and a stiffer lower mantle.

3. Evolution of Southwestern US Dynamic Topography and Mantle Flow

[10] The time-dependent reconstruction of dynamic topography is obtained via "backward" global mantle convection simulations. We adopt the approach used in *Moucha et al.* [2008b] where the direction of buoyancy-induced flow in the mantle is numerically reversed by using negative time. To this end, the initial (present-day) temperature distribution is advected backwards using boundary conditions that are consistent with a new Indo-Atlantic plate reconstruction model in the no-net-rotation reference frame. Details of the numerical method are given by *Moucha et al.* [2008b]. The inherently high surface resolution (~100 km) of our global geodynamic model enables us to quantify the role of mantle convection in the evolution of regional tectonic settings such as the southwestern US.

[11] Figure 1 depicts the modeled evolution of the southwestern US dynamic topography in a fixed North American reference frame over the last 30 Ma at 5 Ma intervals. Specifically, only the changes in dynamic topography with respect to 30 Ma are shown, because the surface topography at 30 Ma is unknown. The modeled reference dynamic topography at 30 Ma and at present-day are shown in the auxiliary material (Figure S1). We emphasize that a change in dynamic topography corresponds to tectonic uplift (or subsidence) driven by mantle convection. Although additional thermal-elastic effects are not included [e.g., *Roy et al.*, 2009], the near-surface thermal isostatic contributions to topography are incorporated in our convection calculation. Other effects on rock-uplift driven by surface tectonics such as extension, or erosion/deposition are also not included. Therefore, these effects and earlier extension and magmatic components (pre-30 Ma) that resulted in complex surface deformation would need to be added to the modeled dynamic topography in Figure 1 to obtain the total surface topography. Moreover, our plotting of these results only accounts for the extension and tectonic re-organization of the southwestern US by incorporating the motion of individual blocks according to a recent tectonic reconstruction by *McQuarrie and Wernicke* [2005]. That is, our global backward mantle convection simulations do not retrodict these reconstructions.

[12] In addition to dynamic topography, Figure 1 also shows the reconstructed evolution of mantle heterogeneity (temperature variations) and the associated flow field along a radial cross-section that is fixed to the rigid North American plate across the CP and neighboring regions. From the temporal evolution along this cross-section, it is evident that a deep-seated warm mantle upwelling has been overridden by the westward motion of the North American plate in the Indo-Atlantic frame of reference. This mantle upwelling coincides with the location of the northern extension of the reconstructed position of the EPR as first proposed by *Menard* [1960] and others [*Wilson*, 1973; *Jacobs et al.*, 1974; *Dixon and Farrar*, 1980; *Eaton*, 1987; *Wilson*, 1988; *Fletcher et al.*, 2007].

[13] In the fixed North American frame of reference, the northeastward-migrating mantle upwelling was beneath the southwestern coast of the US at about 20 Ma, near the present-day location of Los Angeles (Figure 1b). At this time, the position of the upwelling also coincided with the past intersection of the Mendocino transform fault with the EPR (Mendocino and Rivera triple junction) [e.g., *Atwater*, 1970; *Dixon and Farrar*, 1980]. According to these calculations, the southwestern coast of California and the Baja Peninsula were uplifted by this mantle upwelling between 20 and 15 Ma, (Figure 1c). This prediction agrees well with sediments cored from the Patton Ridge, located on the Outer Borderland block of California, that suggest that the ridge was uplifted and possibly exposed sometime between 20–16 Ma [*Marsaglia et al.*, 2006]. Similarly, *Fletcher et al.* [2007] proposed that the Magdalena fan (and filling of the trench) likely resulted from regional uplift of the Magdalena shelf that began roughly at 16 Ma and further subaerial exposure of the shelf and its erosion occurred between 15–12 Ma.

[14] At about 15 Ma, the central BR began a period of extension and mafic magmatism indicating a possible influx of warm mantle beneath the region [e.g., *Fitton et al.*, 1991; *Zandt et al.*, 1995]. The cross-section (Figure 1c) reveals that the bulk of the warm upwelling mantle was indeed located beneath the central BR at this time. As this mantle upwelling propagated eastward, the associated center of the topographic swell began to uplift the southwestern edge of the CP at about 10 Ma with a maximum uplift at about 5 Ma (Figures 1d and 1e). This relatively recent uplift of the southwestern portion of the CP agrees well with a ca. 6 Ma model of the Grand Canyon proposed by *Karlstrom et al.* [2008] as well as the apparent west-to-east tilting of the CP [e.g., *Sahagian et al.*, 2002; *McMillan et al.*, 2006; *Flowers et al.*, 2008].

4. Dynamic Topography and the Colorado Plateau

[15] A detailed look at the change in the CP's dynamic topography is shown in Figure 2. Since 30 Ma, vertical mantle flow has increased the dynamic topography of the CP by over a 1000 m in the south to no less than 600 m in the north (Figure 2a). In the last 5 Myr, a west-to-east gradient across the Grand Canyon has emerged due to the eastward progression of the mantle upwelling – the eastern block of the Grand Canyon was uplifted by about 200 m relative to the western block (Figure 2b). This is about half the amount of the uplift estimated from the differences in eastern-versus-western Grand Canyon incision rates [*Karlstrom et al.*, 2008]. The additional amount of uplift required to match this observation may be due to effects on topography that we are not modeling, such as erosional isostasy, local small scale edge-driven convection across the step of the plateau's root [e.g., *Karlstrom et al.*, 2008], or lateral viscosity variations (i.e., locally reduced viscosity that may enhance the rate of uplift). Moreover, the eastward progression of the mantle upwelling towards the eastern edge of the CP and the Rio Grande rift region over the last 5 Ma fits well with recent magmatic activity in this region whose trace element geochemistry suggest a mantle source that is similar to oceanic hotspots [*McMillan et al.*, 2000].

[16] Uncertainties in our geodynamic modeling originate from uncertainties in both the adopted density and mantle viscosity models (see auxiliary material). In Figure 2c, we plot the evolution of the integrated average (over the CP) of dynamic topography for viscosity models V1 and V2 (see Figure 2d) and two density models TX2007 and TX2008. Because the TX2008 is essentially a damped version of the TX2007, the amplitude of the modeled dynamic topography is consequently decreased.

[17] Comparison of the predicted dynamic topography for the two viscosity models suggests that our predictions are only slightly sensitive to variations in viscosity. The stiffer lithosphere and lowermost mantle of V2 cause both the rate of change as well as the amplitude of dynamic topography to decrease. In effect, these different viscosity models, which are both constrained by the same geodynamic observations, provide an estimate of the potential influence of lateral viscosity variations on our results and they appear to be less than the order of the uncertainties in the density model. It is important to note that this globally constrained convection model is not tuned to regional observations via shear-wave to density scaling or viscosity, but its predictions fall well within the range of current estimates of recent topographic evolution of the southwestern US. The essential ingredient in this successful reconciliation of mantle dynamics and surface geology is a robust mapping of the buoyancy forces and their temporal evolution, hence the importance of the new joint seismic-geodynamic inferences of the 3-D mantle density structure [*Simmons et al.*, 2009].

5. Conclusions

[18] We estimate from the uncertainties of our model (Figure 2c) that the average change in the dynamic topography of the Colorado Plateau is in the range of 400 to 1100 m over the last 30 Ma. The magnitude and timing of this uplift is consistent with data from a thermochonometry study of *Flowers et al.* [2008] that can account for ~1 km of surface uplift in Early Tertiary time (pre-30 Ma). Thus, it appears that the uplift history of the Colorado Plateau is a complex superposition of multiple events caused by different mechanisms that occurred at different times since the Late Cretaceous.

[19] In the last 5 Ma the average change in dynamic topography was 100–300 m. Though modest, the impact of this recent change in dynamic topography may have played an important role in the formation of the Grand Canyon by establishing a 200 m gradient in the flow direction of the Colorado River along the Grand Canyon (Figure 2b). Observations support this predicted gradient [*Karlstrom et al.*, 2008]; however, relating this gradient to surface faulting and differential incision requires a more detailed investigation.

[20] Our numerically reconstructed path of a warm mantle upwelling throughout the Neogene period also provides compelling support for the idea that the tectonic and magmatic evolution of the southwestern US has been driven by this mantle upwelling enhanced by the positive mantle buoyancy under this region [*Menard*, 1960; *Wilson*, 1973; *Jacobs et al.*, 1974; *Dixon and Farrar*, 1980; *Eaton*, 1987; *Wilson*, 1988; *Fitton et al.*, 1991; *Parsons et al.*, 1994; *Fletcher et al.*, 2007]. Indeed, our reconstructed path traces the plate kinematic inference by *Dixon and Farrar* [1980] of where the intersection of the Mendocino transform fault with the East Pacific rise would have been located if it were not over-ridden by the North American plate.

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