Small-scale convection at the edge of the Colorado Plateau: Implications for topography, magmatism, and evolution of Proterozoic lithosphere

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ABSTRACT

The Colorado Plateau of the southwestern United States is characterized by a bowl-shaped high elevation, late Neogene–Quaternary magmatism at its edge, large gradients in seismic wave velocity across its margins, and relatively low lithospheric seismic wave velocities. We explain these observations by edge-driven convection following rehydration of Colorado Plateau lithosphere. A rapidly emplaced Cenozoic step in lithosphere thickness between the Colorado Plateau and adjacent extended Rio Grande rift and Basin and Range province causes small-scale convection in the asthenosphere. A lithospheric drip below the plateau is removing lithosphere material from the edge that is heated and metasomatized, resulting in magmatism. Edge-driven convection also drives margin uplift, giving the plateau its characteristic bowl shape. The edge-driven convection model shows good consistency with features resolved by seismic tomography.

INTRODUCTION

The Colorado Plateau (CP) is a tectonically stable physiographic province of the southwestern United States (Fig. 1A), consisting of largely extant Proterozoic lithosphere (ca. 1.7 Ga; Wendlandt et al., 1996; Gilbert et al., 2007) that was uplifted to its current elevation (~1.8 km) during the Cenozoic. Dynamic topography (Moucha et al., 2009; Liu and Gurnis,

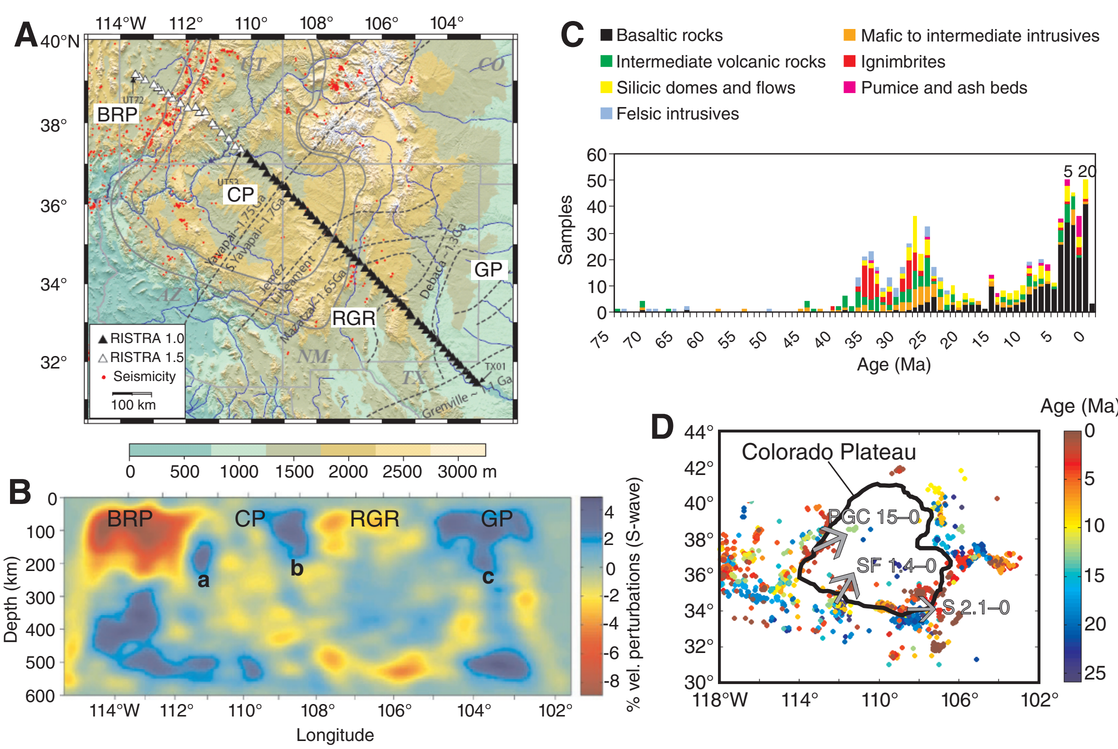

Figure 1. A: Map of the Colorado Plateau and surrounding regions. Black and white triangles show the station location of LA RISTRA seismographs, and red dots indicate characteristic seismicity (1973–2000). Tectonic provinces: CP—Colorado Plateau; BRP—Basin and Range province; GP—Great Plains; RGR—Rio Grande rift. States: UT—Utah; CO—Colorado; TX—Texas; NM—New Mexico; AZ—Arizona. B: LA RISTRA relative S-wave velocity body wave tomography (Sine et al., 2008). High-velocity tomographic features a, b, and c are interpreted as downwelling structures (this study; van Wijk et al., 2008). C: K-Ar and $^{40}Ar/^{39}Ar$ igneous rock ages in New Mexico, redrawn from Chapin et al. (2004). Vertical axis shows number of events or stratigraphic units. Numbers above columns show number of additional basalts that were not plotted. D: Distribution and timing of magmatism since 26 Ma in the CP region (major element data from NAVDAT database, http://www.navdat.org). Arrows indicate younging direction for individual volcanic fields (age range in Ma indicated): PGC—Pahranagat–San Raphael and Grand Canyon volcanic belts (Nelson and Tingey, 1997); SF—San Francisco volcanic field (Conway et al., 1997); S—Springerville volcanic field (Connor et al., 1992).

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2010) and static thermal uplift, in response to removal of the Farallon slab (Bird, 1984; Beghoul and Barazangi, 1989; Roy et al., 2009), have been suggested to contribute to the plateau's current elevation. The plateau features a superimposed bowl-shaped topography, where edges are elevated with respect to the interior by ~400 m (Fig. 1A). A large seismic S-wave velocity contrast of ~12% between the CP and the Basin and Range province (BRP) has been imaged across a narrow distance (~100 km), indicating that the CP western edge temperature gradient is steep and large (Sine et al., 2008). Since the late Neogene, magmatism has been widespread at the CP edges (Fig. 1D). This magmatism differs from middle Cenozoic magmatism in composition and distribution (Fig. 1C): late Neogene–Quaternary magmatism is fundamentally basaltic (Chapin et al. (2004), whereas middle Cenozoic magmatism is intermediate to silicic and is virtually absent on the CP. Late Neogene–Quaternary magmatism and the middle Cenozoic pulse are also separated by a magmatic quietus and are interpreted as separate geodynamic events (Humphreys et al., 2003; Christiansen and Yeats, 1992; Lipman, 1992).

We show that convective destabilization of the CP linked to extension of its surrounding regions (BRP and Rio Grande rift [RGR]) is a plausible mechanism for generating the Neogene magmatic pulse, the large variation in seismic anomalies across the CP edge, and the CP bowl-shaped topography. Specifically, an edge-driven convection cell is predicted to develop at the CP edge when variations in lithosphere thickness are as large as those between the CP and its surrounding extended provinces. We use numerical models of upper mantle flow to show that mantle lithosphere is removed from the CP edge, which produces both the differential elevation of the CP and magmatism along the CP margin. The models predict a large temperature contrast between the CP and the BRP, which, when mapped to synthetic seismic wave velocities, corresponds well to the tomographically imaged seismic structure (Sine et al., 2008).

CONVECTION MODEL FOR THE COLORADO PLATEAU MARGINS

Proterozoic CP lithosphere extends to a depth of 120–140 km (West et al., 2004). In contrast, the lithosphere of the RGR and BRP surrounding the CP has been dramatically thinned by several tens of km due to Cenozoic extension to a current thickness of ~60–80 km (Wernicke, 1992; Zandt et al., 1995; DePaolo and Daley, 2000; West et al., 2004). The thickness of BRP and RGR lithosphere adjacent to the CP prior to Cenozoic extension is unknown, but assuming that the better constrained Cenozoic crustal extension and thinning is approximately equal to that of the mantle lithosphere produces a pre-extension lithosphere thickness of ~130 km in the BRP adjacent to the CP. Once the BRP began to extend, the step between the CP and BRP lithosphere developed; we include a Neogene step in lithosphere thickness between the CP and its surrounding provinces.

Modeling was performed using the CitCom finite-element code (Moresi and Gurnis, 1996; GSA Data Repository[1]). Mass, momentum, and thermal energy conservation equations are solved in two dimensions, for an incompressible infinite Prandtl number fluid with Boussinesq approximations (van Hunen et al., 2005; van Wijk et al., 2008; Data Repository). The LA RISTRA seismic array (Gao et al., 2004; Sine et al., 2008) crossed the CP western margin (Fig. 1A), and to compare our two-dimensional model calculations with these tomographic results, one of our models was chosen with identical geometry (model III). We additionally modeled the south-southeastern edge of the CP that is located adjacent to highly extended lithosphere (models I and II, Fig. 2). We performed a series of tests (Data Repository) where initial lithosphere thickness, rheology, and BRP extension velocity were varied. Convection was induced in all cases (Fig. 2), and as is characteristic of edge-driven convection, the convection cells show upward flow of asthenosphere beneath the thinned areas of the BRP. Below the CP interior, a cold downwelling develops. Mantle lithosphere is removed at the edge as a drip and sinks into the asthenosphere. A lower activation energy (180 versus 360 kJ/mol) promotes the formation of instabilities at the base of the lithosphere, and the cold drip becomes both colder and larger in size. Despite lateral heat conduction, the combined effect of continuous BRP thinning and advective heat transport via convection maintains an abrupt (~500 °C over 60 km) lateral temperature contrast between the CP and surrounding extended

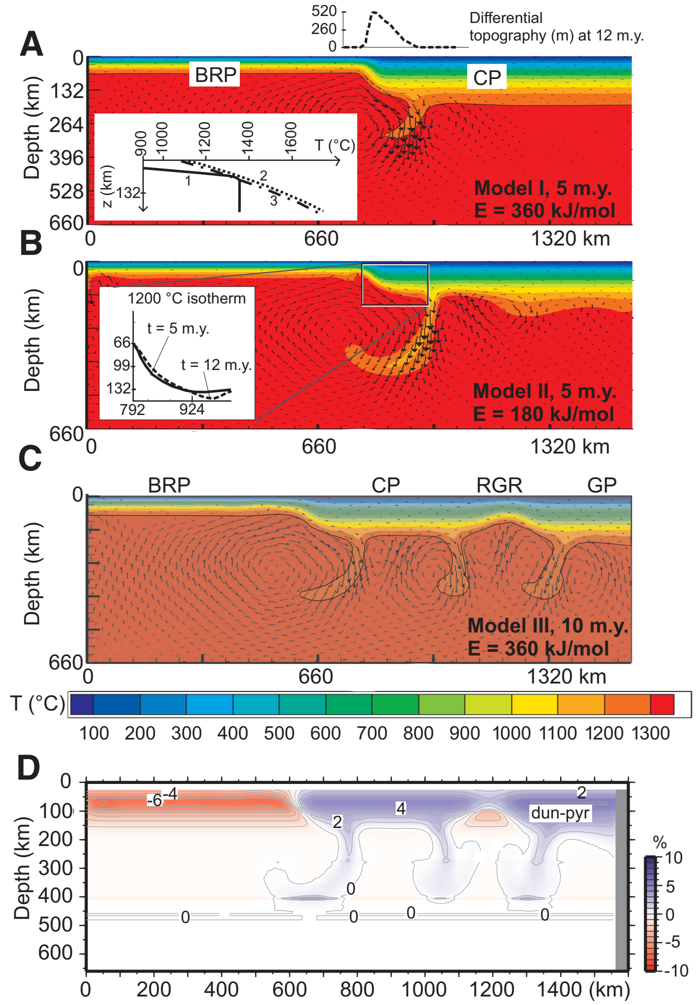

Figure 2. Temperature and flow field modeling (Sine et al., 2008) for initial Basin and Range province (BRP) thickness ~65 km, with the 1300 °C isotherm indicated by black contour. Models I and II are set up to cross the Colorado Plateau (CP) western edge, model III is set up along the LA RISTRA line. A: Model I with E = 360 kJ/mol, at 5 m.y. Inset: 1—geotherm at x = 810 km; 2—dry solidus (Hirschmann, 2000); 3—hydrous solidus (Katz et al., 2003). Above A is shown the topography resulting from lithosphere thinning after 12 m.y. at the CP edge. This gives the CP its characteristic bowl shape. B: Model II with E = 180 kJ/mol, at 5 m.y. Inset: 1200 °C isotherm after 5 m.y. (solid line) and 12 m.y. (dashed line). C: Model III with E = 360 kJ/mol, set up along the LA RISTRA line, crossing CP, at 10 m.y. D: S-wave velocity anomalies predicted from the thermal structure in Figure 2C by modeling the lithosphere with temperatures below 1200 °C as depleted (dunitic), and the rest of the domain as pyrolitic in composition (details in the GSA Data Repository [see footnote 1]).

[1]GSA Data Repository item 2010167, model description and seismic conversion method, is available online at www.geosociety.org/pubs/ft2010.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

terrains, which is an important condition for consistency with the observed steep seismic velocity gradient (Fig. 1B).

We also performed tests with a domain crossing the entire CP along the LA RISTRA array (Fig. 1A). Lithospheric drips form below the CP on both western and eastern sides, as well as below the Great Plains (GP) edge to the east of the RGR (van Wijk et al., 2008). The drip below the GP edge (Fig. 2C) is closer to the RGR than suggested by LA RISTRA tomography (Fig. 1B) and earlier modeling (Gao et al., 2004; Wilson et al., 2005; van Wijk et al., 2008); this may be because this drip is a longer-lived feature than simulated here, forming ca. 25 Ma with the onset of RGR rifting, migrating eastward during its development (van Wijk et al., 2008). Beneath the CP center the models predict upwelling (Fig. 2C), which could heat and ablate the metasomatized lithosphere there and increase elevation.

TECTONIC AND MAGMATIC ACTIVITY

The above convection modeling predicts a magmatic pulse at the CP edge as the CP lithosphere edge is heated by lateral heat conduction and mantle lithosphere is thinned by convective removal, resulting in hot asthenosphere replacing lithosphere (Fig. 2). The solidus (Fig. 2A) is crossed below the BRP and below the CP edge; the resultant amount of modeled melt depends on the initial thickness of the BRP lithosphere, the extension rate, and the amount of convectively removed mantle lithosphere. CP edge magmatism produced by this process can explain the late Neogene to present magmatism of the plateau's southern and western edges, such as the San Francisco volcanic field (Fig. 1D), and predicts that late Neogene–Quaternary mafic rocks are derived from melting of mantle lithosphere and/or asthenospheric sources. During the late Neogene–Quaternary phase, magmatism was dominantly bimodal, with the basaltic end member comprising both subalkaline (tholeiitic) and alkali olivine compositions (Fig. 1C). Mafic rocks were derived from melting of relatively anhydrous lherzolite of OIB affinity from mantle lithosphere and/or asthenospheric sources.

This late Neogene–Quaternary magmatic phase was preceded by a period of relative quiescence that followed the middle Cenozoic pulse. The middle Cenozoic phase was characterized by widespread and voluminous intermediate to silicic magmatism, ranging from basaltic andesite to rhyolite, with calc-alkaline affinity (Lipman, 1992). Fluid-enriched melting of the lithosphere played a substantial role in this phase of magmatism (Livaccari and Perry, 1993), while having lesser importance in the late Neogene–present phase. Only minor alkalic and silica-undersaturated magmatism occurred on the CP between 28 and 19 Ma. Nonetheless, the lower lithosphere may have been affected by related metasomatic activity at this time, even within the interior of the CP, as seen by the relatively low velocities of the CP interior (Fig. 1B).

Numerous studies (e.g., Christensen and Yeats, 1992; Dickinson, 1997) have documented broad spatial-temporal patterns for (middle) Cenozoic magmatism through the BRP. To examine late Cenozoic spatial-temporal patterns from 25 to 0 Ma in the CP and surrounding regions, we used the NAVDAT database (http://www.navdat.org) (Fig. 1D). Detailed field studies (e.g., Connor et al., 1992; Nelson and Tingey, 1997; Conway et al., 1997) document local progressions in ages of late Neogene–present basaltic lavas on the western, southern, and southeastern margins of the plateau (Fig. 1D). These age progressions are generally eastward or northeastward, as expected from inward migration of edge convection.

SEISMIC STRUCTURE

The large change in S-wave velocity (~12% between the CP and the BRP) imaged using the LA RISTRA array occurs over a short lateral distance (~100 km; Sine et al., 2008) across the CP edge (Fig. 1B). We used model temperatures to predict seismic velocity structures using the extensively tested methods of Cobden et al. (2008) (Data Repository). We find that the predicted maximum thermally induced, V_S contrast between CP and BRP lithosphere is up to 14% in V_S and up to 9% in V_P and may be enhanced by another 1.5%–2% by lithospheric depletion (Fig. 2D; Fig. DR1). This range is similar in magnitude to that imaged (Sine et al., 2008). However, tomographically imaged amplitudes may be an underestimate of the true ones due to regularization, hence an additional contribution due to melt is likely required (see above). The percentage of melt necessary to fit the seismic observations is difficult to quantify given the smoothing effects of regularization and limited tomographic resolution, uncertainties in the seismic signature of melt without knowledge of melt geometry, and uncertainties in the exact contribution of anelasticity. The dipping fast structure "a" (Fig. 1B), which is suggestive of downwelling, strongly contributes to the CP-BRP seismic velocity contrast with an ~4%–5% amplitude (Sine et al., 2008). Our modeled velocity structure of the downwelling is less strong (<~3%). The model for the entire CP (Fig. 2C) predicts fast velocity structures below the CP western and eastern sides, as well as below the GP, corresponding to the locations found in tomography.

The CP has a depleted lithosphere, but its base has been affected by rehydration and metasomatism by Farallon subduction (Humphreys et al., 2003; Li et al., 2008), compatible with the evidence for fluids significantly affecting the first phase of Cenozoic magmatism (see above). This would lower CP lithosphere viscosity and make it more conducive to destabilization and recycling (Li et al., 2008). Metasomatism and rehydration may also explain relatively low seismic wave velocities of the CP interior lithosphere compared to those below the GP (Fig. 1B), which was not affected by Farallon subduction. We suggest that such refertilization and subsequent destabilization may be a key driver for late Neogene volcanic encroachment onto the southern parts of the plateau. In our models, not surprisingly, only the CP thermal boundary layer, which would be conducive to destabilization, is eroded. We did not perform tests of destabilization of rehydrated CP lithosphere, but suggest that more weakened CP lithosphere may be subject to removal by this process (Li et al., 2008).

UPLIFT

The removal of basal mantle lithosphere from the CP edge by edge-driven convection, and thinning of CP lithosphere by dissociation (Li et al., 2008), result in uplift as metasomatized mantle lithosphere is replaced by less dense asthenosphere (Fig. 2A). We calculate a first-order estimate of CP uplift associated with this process using mantle lithosphere density of 3300 kg/m^3 and an asthenosphere density of 3200 kg/m^3. A maximum of 20 km of Cenozoic lithosphere thinning is inferred from xenolith data (Riter and Smith, 1996) and tomography (West et al., 2004), which would result in ~600 m of CP uplift centered above the central CP upwelling (Fig. 2C). The amount of lithosphere removal by edge convection varies in different models between ~9 km and ~17 km after 12 m.y. of model evolution, resulting in ~270–520 m of additional uplift at the CP edges (Fig. 2A), as is seen in the observed CP bowl-shaped topography. This suggests that thick lithosphere likely no longer exists beneath the outermost edge of the plateau. Colorado River incision data suggest a maximum of ~700 m of Neogene uplift of the CP edge with respect to the BRP (Karlstrom et al., 2008), with the predicted amplitude of the uplift largest close to the CP edge, decreasing toward the interior (Fig. 2A). The Neogene timing of CP edge uplift is also consistent with incision observations (Karlstrom et al., 2008). A relative crustal depression of ~150 m maximum is further predicted directly above the drips in the interior CP.

CONCLUSIONS

We propose that CP differential topography and late Neogene–Quaternary magmatism result from edge-driven convection. Our model predicts sharp temperature contrasts in the shallow mantle across the CP edge, in good agreement with results from seismic tomography. Proterozoic mantle lithosphere is presently being removed from the CP edge, resulting in margin uplift that gives the plateau its characteristic bowl shape. Partial melting of lithosphere and asthenosphere driven by this process causes the present phase of edge magmatism. Uplift associated with CP dissociation following rehydration may account for ~600 m of the overall high CP elevation.

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