**UNCERTAINTY IN DENSITY MODELS**

A more complete discussion of uncertainties in density models derived similarly to ours, and the sources thereof, is given by Levandowski et al. (2017). Here, we present a brief discussion of the variability of densities in our accepted 3-D models and then turn our attention to systematic biases that may have been introduced by our assumptions and modeling procedure.

Uncertainty is readily quantified from our modeling: we have 1000 estimated densities at each point on a 20x20 km grid and in 15 different layers. Considerable care should be taken in deciding what a meaningful measure of uncertainty is, however. The density in any one 5x20x20 km cell in the lower crust of our model is not important. The meaningful quantity is the density across multiple layers in our model, say the 20–40 km range. Also, because our primary goal is to investigate the long-wavelength topographic gradient that has developed on the Plains in the Cenozoic, we are chiefly interested in average densities along a line of longitude. As such, we should consider that an individual 100-km wide swath of lower crust subsumes a grid ~55 nodes from north to south, 5 nodes from east to west, and 4 layers thick. If the densities of these ~1100 cells are independent of one another (not a terrible approximation, as determined from statistical tests that are not shown here), then the uncertainty of that volume is roughly 1/33 the uncertainty of any given cell. Considering the differences in modeled density across the Plains (>100 kg/m3), one would require that the uncertainty in a typical cell be 3300 kg/m3, which is non-physically large: It would require the assertion that our modeling cannot tell whether there is air or asthenosphere at any given point.

We now discuss the uncertainty in the density over a given depth range at any of our 20x20 km columns. Beyond the upper few km, uncertainty is highest near the Moho, so we will discuss results from this depth range. The density in, say, the 40-50 km depth range varies in any given column across the 1000 simulations, mainly controlled by how uncertain the Moho depth is. For a typical point, 950 of the 1000 density models are within 30-40 kg/m3 of the mean value from across all of the 1000 models. If we consider the depth range from 30-50 km, uncertainty is typically ±25 kg/m3. The 20-50 km depth range primarily discussed in the text typically has uncertainties of 15-20 kg/m3. Again, these values are the 95% confidence range. With regard to the variations shown in figure 3C of the main text, the difference between any two individual points (not to mention the longitudinal averages that are actually of interest in this study) can be considered statistically robust if it is greater than ~25 kg/m3.

**POTENTIAL IMPACTS OF OUR ASSUMPTIONS: SYSTEMATIC BIASES**

**Material below 150 km**

Estimates of modern lithospheric thickness on the Plains place the lithosphere-asthenosphere boundary at or even below the maximum depth that we consider, 150 km. We use this depth because the seismic velocity models of Shen et al. (2013) lose resolution rapidly below 150 km. It is important, however, to consider what the effect of additional, unaccounted mantle lithosphere beneath the Plains would be.

Away from the Oklahoma Aulacogen, the topography and gravity of the Plains is rather closely reproduced by our initial model. Therefore, the presence of additional, thermally dense lithosphere would cause the integrated density to be too high, with a more severe impact on topography residuals than on gravity residuals. Synthetic tests confirm the logical suspicion that ensuing adjustments to the density model are characterized by negative changes to mantle density (implicitly reflecting compositional buoyancy). For example, Schutt et al. (2011) estimate average Moho temperatures—at ~45 km depth—of 800°C on the Plains. Assuming a linear geotherm from the Moho to the base of the lithosphere at 1300°C and examining a lithospheric thickness of 200 km (i.e., our models truncate the lower 50 km of lithosphere), the temperature would be 1138°C at 150 km and increase to 1300°C at 200 km. The average temperature over the lower 50 km of the Plains’ lithosphere is then 1219°C. With a coefficient of thermal expansion of 3x10-5/°C, the omitted 50 km of lithosphere would average ~8 kg/m3 denser than asthenosphere beneath the CP at the same depth range. By equation 1, this difference increases the thermal buoyancy of the CP mantle relative to the Plains’ by 0.125 km (for a total of just over 500 meters). Synthetic tests that extend the model space to 200 km, using the initial model in the upper 150 km, adding such a body beneath the Plains, and adding uniform 3200 kg/m3 asthenosphere elsewhere are characterized by a negative density adjustments to the Plains’ mantle that average (excluding southern Oklahoma) -3.1 kg/m3 over the depth range from 60 to 200 km, with nearly uniform adjustments, on average, over this depth range.

In our framework, negative adjustments to mantle density would be interpreted as compositional buoyancy, perhaps reflecting melt depletion of the Plains’ lithospheric mantle. By equation 1, the synthetic results show that a +135 meter buoyant height due to mantle composition would be modeled beneath the Plains if the mantle lithosphere indeed extends to 200 km and has a thermal structure similar to that described above. By contrast, careful examination of the CP in Figure ADJS reveals very small, positive adjustments (as great as +15 kg/m3, average of ~+2 kg/m3) to the seismically derived initial density estimate. These signals are compatible with a very minor compositional antibuoyancy (e.g., from melt infiltration; Roy et al., 2016) in the CP. Thus, if the lithospheric mantle beneath the Plains does extend to 200 km depth, we would interpret a modestly greater thermal buoyancy of the CP mantle relative to the Plains and a somewhat negative compositional buoyancy of the CP relative to the Plains.

**Grain Size**

Our estimates of thermal variations in mantle density (equation 3) are derived for 1 mm grains, but that choice is arbitrary. With increasing grain size, the effect of temperature on velocity decreases (Jackson and Faul, 2010). Therefore, the velocity variations reported by Shen et al. (2013) would correspond to proportionally greater density variations. Using 1 cm grains, we derive a relationship—compare to equation 3—between velocity and density of approximately:

$Δρ=Δv\_{s} ×\left(8.5-\frac{z}{90 km}+\frac{Δv\_{s}}{5}\right); 0\leq Δv\_{s}\leq 6\%$(8a)

$Δρ=Δv\_{s} ×\left(9.7-\frac{z}{90 km}-\frac{6(Δv\_{s}-6)}{35}\right); Δv\_{s}\geq 6\%$ (8b)

In practice, the difference between 1 mm and 1 cm grains is minor. For example, a +5% $Δv\_{s}$ at 100 km would be modeled as a +7.55 kg/m3 perturbation in 1 mm grains but a +8.39 kg/m3 perturbation in 1 cm grains. Relative to the eastern Plains, thermal mantle buoyancy would support an additional ~40 meters in the CP and ~10 meters on the western Plains.

**Solidus Velocity**

We follow Levandowski et al. (2014) in estimating the velocity at the solidus as 4.5 km/s. That estimate examines the highest velocities reported by Shen et al. (2013) at 120 km depth, ~4.75 km/s, which are found beneath the Wyoming Craton. Using estimates of thermal boundary layer thickness (Schutt et al., 2011) and assuming a linear geotherm in the mantle, the temperature of this material is ~820°C. Next, assuming solidus temperature of 1300°C and scaling that 480°C difference between inferred temperature beneath the Wyoming Craton and asthenosphere to a velocity perturbation yields 0.25 km/s, implying an asthenospheric temperature of 4.5 km/s.

Following this rather long list of assumptions, it becomes clear that the solidus velocity is poorly constrained. The impact of a lower velocity is minor in relation to objectives of the present work, however. The primary impact of choosing a lower solidus velocity is additional thermal antibuoyancy in the Plains’ mantle. As was the case for thick lithosphere, a component of mantle chemical buoyancy becomes necessary. Therefore, the contribution of mantle temperature to CP uplift increases but this is countered by relative compositional antibuoyancy. Higher solidus velocity decreases the role of mantle temperature.