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Key Points:

- Full-wave seismic tomography reveals low-velocity anomalies below the Wisconsin Dome and Illinois and Michigan Basins below 50–60 km depth
- These scattered low velocities beneath the thinned lithosphere reflect emplaced hot asthenospheric materials after lithospheric delamination
- The topographic uplift due to lithospheric delamination, as shown in geodynamic models, is sufficient to end subsidence in cratonic basins

Supporting Information:

Supporting Information may be found in the online version of this article.

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Lithospheric Delamination Below the North American Midcontinent Ceased Subsidence in Cratonic Basins

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Abstract Cratonic lithospheres carry a long history of tectonic modifications that result in heterogeneous structures, as revealed by an increasing number of geophysical observations. The existence of cratonic basins indicates protracted periods of tectonic modification, causing subsidence within global continental interiors. An enigmatic aspect of this process is the cessation of subsidence in cratonic basins with unclear mechanisms. Here, using full-wave ambient noise tomography, we reveal distinct seismic low-velocity anomalies below 60 km beneath the Illinois and Michigan Basins, where subsidence terminated in the late Paleozoic to the early Mesozoic. These low-velocity volumes, surrounded by distinctly higher velocities, are attributed to asthenospheric materials upwelling to shallow mantle depths during lithospheric foundering or delamination. This lithospheric modification may be associated with a major regional tectonic exhumation in the early Mesozoic that could have terminated basin subsidence and unroofed upper portions of basin stratigraphy. This timing coincides with the passage of this region over mantle plumes, which likely triggered lithospheric delamination and asthenospheric upwelling. Geodynamic modeling shows that the emplacement of these buoyant asthenospheric materials would lead to an uplift of about 3.5 km, sufficient to terminate the subsidence in the cratonic basins within this region. These findings document evidence of lithospheric delamination in the North American midcontinent and present important links between geodynamic drivers and geological records of the evolution of the cratonic lithosphere in North America and beyond. They also offer broader implications for understanding how deep Earth processes shape surface environments, influencing resource distribution and long-term landscape evolution.

Plain Language Summary Cratons are ancient and stable continental regions. Cratonic basins are depressed areas within these stable continental cores and would stop sinking over time, though the reason for the abrupt cessation of subsidence is not well understood. We used an advanced seismic imaging method to investigate the subsurface structure below the North American midcontinent, covering the Illinois and Michigan Basins, where sinking stopped between 350 and 300 million years ago. Below 60 km depth, we found volumes through which seismic waves propagate more slowly, surrounded by regions where seismic waves travel faster, particularly below the Illinois and Michigan Basins. These lower velocity regions suggest that material from the Earth's asthenosphere has replaced older continental lithosphere, potentially due to a process called lithospheric delamination, where parts of the Earth's lithosphere peel away. Our numerical model shows that the ascending mantle material could have caused the surface to rise by about 3.5 km, enough to stop the basins from sinking further and remove the uppermost layers of sedimentary rocks. This study highlights the role of lithospheric delamination in shaping cratonic basins and provides new insights into how Earth's deep processes influence surface geology over time.

1. Introduction

Cratonic basins are localized subsidence features within cratons with prolonged tectonic subsidence histories (hundreds of millions of years). These basins are widely distributed within global cratons in various sizes, shapes, and alignments (Figure 1a; Allen & Armitage, 2012; Daly et al., 2018). Various subsidence mechanisms have been proposed to explain the formation of cratonic basins (Allen & Armitage, 2012). Stratigraphic records at basins serve as a primary archive of Earth's sedimentation history, preserving signals of tectonic, climatic, and eustatic changes over geologic time. Variations in sediment thickness, facies, and unconformities allow

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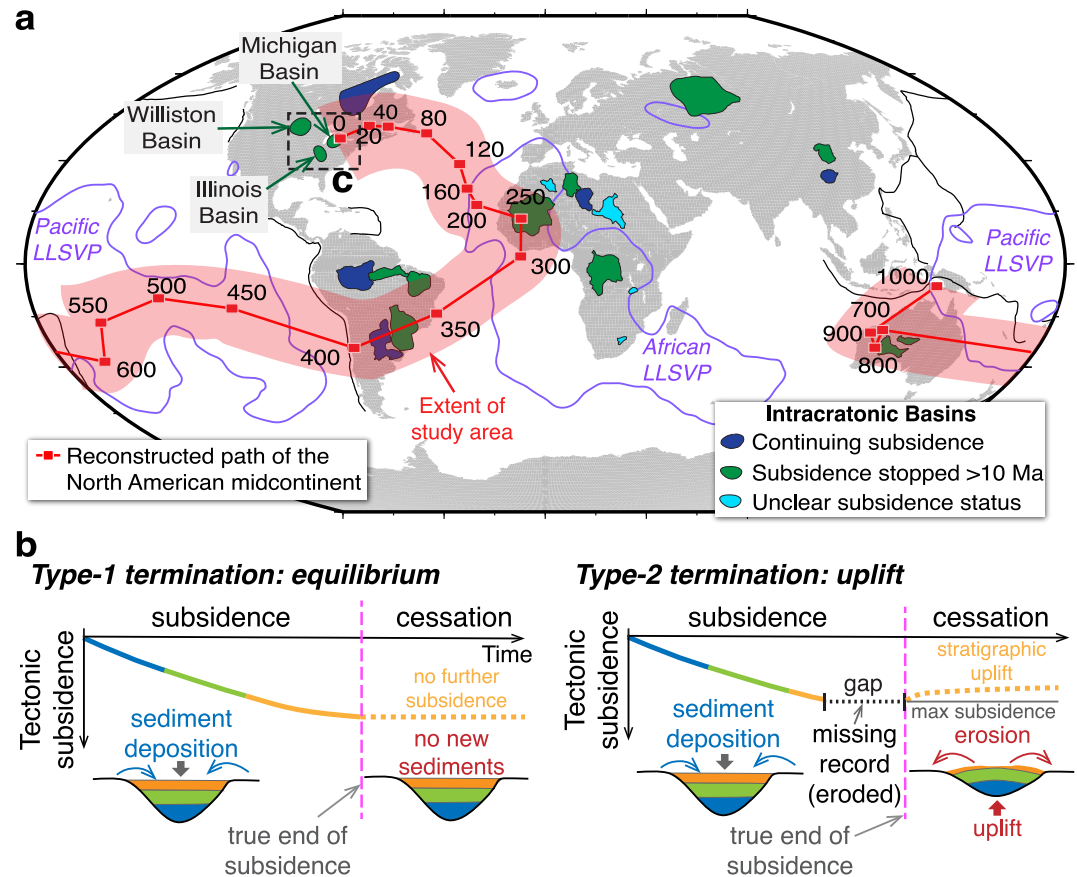


Figure 1. Cratonic basins and models of subsidence cessation. (a) Global distribution of cratonic basins extended from Daly et al. (2018). The subsidence status is based on published subsidence curves (Albert-Villanueva et al., 2016; Allen & Armitage, 2012; Galeazzi et al., 2010; Galushkin & Eloghbi, 2014; Haddad et al., 2001; Korsch et al., 1991; Linol et al., 2015; Underdown et al., 2007). The red squares are the reconstructed locations of the North American midcontinent at the labeled time in Ma (Merdith et al., 2021; Y. Wang et al., 2023). The purple outlines are the isochrons of -1% shear-wave velocity anomalies at the core-mantle boundary from model GLAD-M25 (Lei et al., 2020), outlining the Pacific and African Large Low-Shear-Velocity Provinces (LLSVP). (b) Two scenarios of subsidence termination at cratonic basins with equilibrium (Type-1) or as the consequence of uplift (Type-2).

researchers to reconstruct subsidence patterns, infer past surface processes, and link basin evolution to broader geodynamic and environmental changes. The timing and duration of subsidence in cratonic basins inferred from sedimentation patterns preserve information about the tectonic and geodynamic modification of continental interiors. The termination of tectonic subsidence and even partial inversion of cratonic basins are important records of the dynamic evolution of cratonic lithospheres (Albert-Villanueva et al., 2016; Allen & Armitage, 2012; Galeazzi et al., 2010; Galushkin & Eloghbi, 2014; Haddad et al., 2001; Korsch et al., 1991; Linol et al., 2015; Underdown et al., 2007). However, these records are often buried in the subsurface, making them hard to observe. The mechanisms that lead to subsidence termination in cratonic basins are understudied globally.

Geological evidence for the termination and inversion of cratonic basins provides a temporal context to evaluate lithospheric structures. We summarize the termination of subsidence in cratonic basins in two ways (Figure 1b): (a) the lithosphere reaches isostatic equilibrium without creating new accommodation space (DeRito et al., 1983) (Type-1) and (b) the lithosphere undergoes uplift, resulting in basin inversion and surface erosion (Albert-Villanueva et al., 2016; Galeazzi et al., 2010; Haddad et al., 2001) (Type-2). The *inversion* in this paper refers to the flip in the sense of vertical motion in the basin area and does not necessarily come with the inversion of specific fault styles. These two scenarios predict distinctly different endings of the tectonic subsidence histories in the cratonic basins (Figure 1b). Basins that terminated after reaching isostatic equilibrium (Type-1) would preserve the stratigraphic records that represent a complete history of subsidence, with the subsidence curve gradually

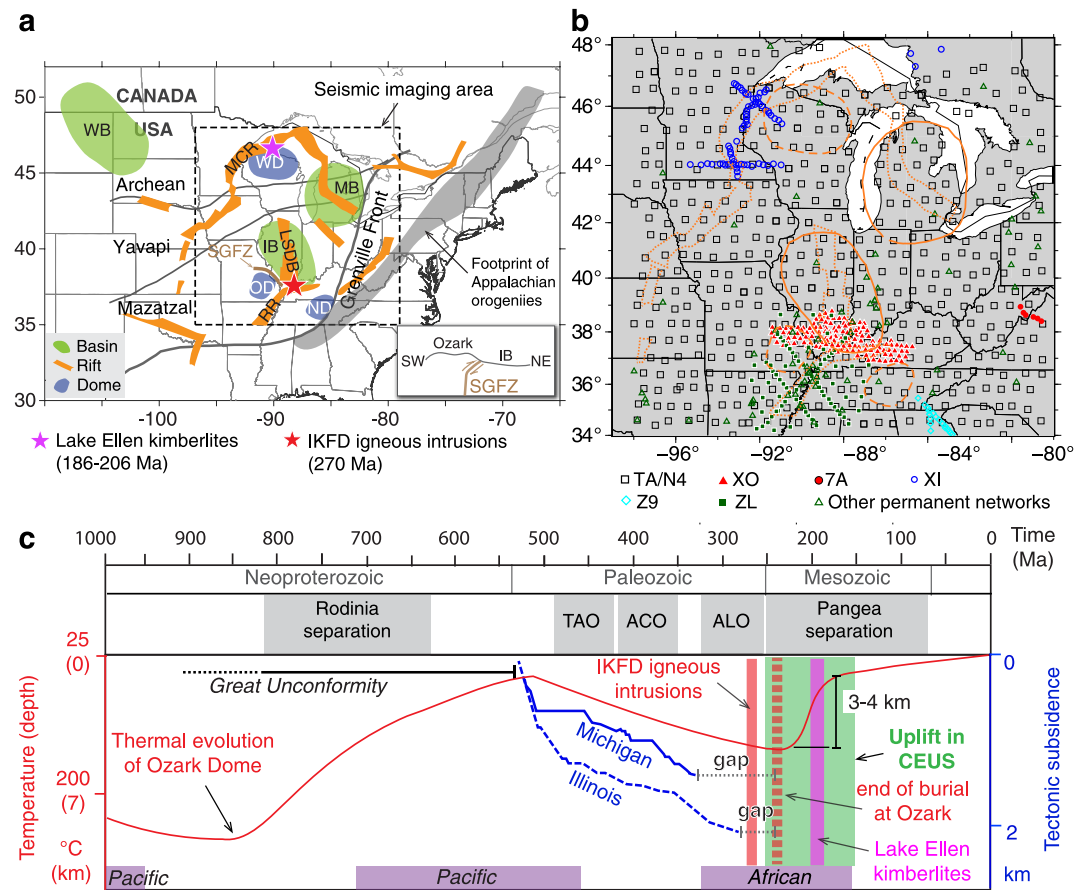


Figure 2. Tectonic settings and seismic stations. (a) Major tectonic features within the North American midcontinent (Bickford et al., 2015; Marshak et al., 2017; Stein et al., 2018). The red and purple stars mark the locations of the Illinois-Kentucky Fluorspar District (IKFD), with igneous intrusions at around 270 Ma (Moorehead, 2013), and the Lake Ellen (Michigan) kimberlites with an age of 186–206 Ma (Heaman et al., 2004; McGee et al., 1983). The inset diagram shows the geometry and Carboniferous faulting along the Ste. Genevieve Fault Zone (SGFZ; Nelson & Lumm, 1985). IB—Illinois Basin, MB—Michigan Basin, WB—Williston Basin, MCR—Midcontinent Rift, RR—Reelfoot Rift, LSDB—La Salle Deformation Belt (initiated as a rift in Neoproterozoic; Marshak & Paulsen, 1996). WD—Wisconsin Dome, OD—Ozark Dome, ND—Nashville Dome. (b) Seismic stations from multiple permanent and temporary networks. See Table S1 in Supporting Information S1 for the list of seismic networks and detailed descriptions. (c) Tectonic subsidence of the Illinois and Michigan Basins (Allen & Armitage, 2012) (blue lines) and simplified thermal evolution of the Ozark Dome (red line) from low-temperature thermochronology (DeLucia et al., 2018). The magenta boxes mark the time ranges the study area was atop the Pacific and African LLSVPs based on (a). The green box brackets the regional uplift in the central and eastern U.S. (CEUS) at 250–150 Ma (Y. Wang et al., 2023). Appalachian orogenic events (Hatcher, 2010): TAO—Taconic Orogeny, ACO—Acadian Orogeny, ALO—Alleghenian Orogeny.

flattening toward the end. In contrast, for the Type-2 scenario, the inversion of the basin would erode the youngest stratigraphy that records the final period of subsidence. This leads to inconsistency in the end times of the retrieved tectonic subsidence curves: the actual termination of the subsidence episode versus a point before the end of subsidence and basin inversion, resulting in a gap between the last sedimentary record and the actual time of subsidence termination (the gap in Figure 1b). There is no notable flattening of the subsidence curve based on stratigraphical records. Identifying the temporal gap between these events is critical in examining whether (a) the basin has maintained equilibrium with no uplift or (b) uplifts caused basin inversion and the termination of subsidence.

Our seismic imaging area is located in the North American midcontinent, covering several tectonic features, including the Illinois and Michigan Basins, the Ozark, Nashville, and Wisconsin Domes, the Midcontinent and Reelfoot Rifts, and the La Salle Deformation Belt (Figures 2a and 2b). The Midcontinent and Reelfoot Rifts are two failed continental rifts formed in the Neoproterozoic and early Paleozoic, respectively (Nelson &

Marshak, 1996; Stein et al., 2018). Although debated, the La Salle Deformation Belt in the Illinois Basin may have been initiated as a rift in the Neoproterozoic and later reactivated by reverse faulting (Marshak & Paulsen, 1996). The Ozark Dome has documented two episodes of exhumation in the latest Neoproterozoic to earliest Paleozoic and in the early Mesozoic (DeLucia et al., 2018) (Figure 2c). The Nashville Dome was exhumed in the early Paleozoic but was later eroded and covered with sediments (Stearns & Reesm, 1986). The Wisconsin Dome formed in the middle Proterozoic with the Precambrian basement exposed at the surface (Peterman et al., 1985), although the tectonic evolution since then is unclear.

The cratonic Illinois and Michigan Basins, the focus of this study, have different sizes, shapes, and long-axis orientations (Figure 2a). Both basins began to subside in the early Cambrian (Allen & Armitage, 2012; Figure 2c). The tectonic subsidence curves constructed using preserved stratigraphy imply that the Illinois and Michigan Basins stopped subsiding in the Carboniferous at about 280 Ma and 330 Ma, respectively (Figure 2c; Allen & Armitage, 2012; Bond & Kominz, 1984; Buschbach & Kolata, 1990; Catacosinos et al., 1990). The history of reconstructed tectonic subsidence of the two basins (Allen & Armitage, 2012) temporally coincides with a burial episode observed using low-temperature thermochronology in the Ozark Dome southwest of the Illinois Basin (Figure 2c; DeLucia et al., 2018). This burial phase at the Ozark Dome ended in the early Mesozoic at about 240–220 Ma. It was followed by an episode of exhumation of about 3–4 km (DeLucia et al., 2018), superimposed on a regional uplift in the central and eastern US in the early Mesozoic at about 250–150 Ma (Y. Wang et al., 2023; Figure 2c).

The thermal history at the Ozark Dome shows continued heating after the end of stratigraphically preserved subsidence in the Illinois and Michigan Basins, suggesting continued sedimentation in the basins until the early Mesozoic. Furthermore, widespread unconformities above the Carboniferous units in the Illinois and Michigan Basins (Buschbach & Kolata, 1990; Catacosinos et al., 1990; Swezey, 2008, 2009) have been documented, suggesting basin-scale erosion from Permian to early Cretaceous and localized erosion thereafter. However, the exact magnitude of the eroded stratigraphy is difficult to estimate. The subsidence curves for both basins do not show flattening toward the ends. These collective lines of evidence suggest that Type-2 subsidence termination is possible in these basins (Figure 1b). Sediment accumulation likely continued in both the Illinois and Michigan Basins through the early Mesozoic but was later removed during the Triassic-Jurassic exhumation (Figure 2c). There is a lack of evidence for prominent post-Carboniferous fault movements along the Ste. Genevieve Fault Zone between the Ozark Dome and the Illinois Basin to accommodate a vertical offset of 3–4 km (Nelson & Lumm, 1985; Figure 2a). This implies that the 3–4 km exhumation suggested by the thermal history at the Ozark Dome should be part of a broader, regional uplift in the central and eastern U.S. (e.g., Boettcher & Milliken, 1994; Poag & Sevon, 1989; Y. Wang et al., 2023). Lateral heterogeneities in the uplift amplitude are expected to exist across the region.

Buckling under compression due to the Appalachian orogenic events can not explain the episode of uplift in the early Mesozoic in the central and eastern U.S. First, the timing of the Taconic and Acadian events all happened before the uplift and termination of the basins, when the basins were undergoing subsidence (Figure 2c). Second, although the Alleghenian orogeny occurred after the end of the subsidence curve for the Michigan Basin, the Illinois Basin was still undergoing subsidence (Figure 2c). The influence of the Alleghenian Orogenic event can not explain the >50 Myr difference in the ending times of the two basins, given that they are located at a similar distance from the orogeny (Figure 2a). Third, dynamic modeling work has shown the potential link between the uplift of arches between the Illinois, Michigan, and Appalachian Foreland Basins (Quinlan & Beaumont, 1984), instead of the termination of subsidence in the basins.

To understand the potential structural signature of geologically documented uplift and the associated mechanical processes leading to the termination of cratonic basin subsidence, we constructed a 3-D lithospheric seismic velocity model below the North American midcontinent, using full-wave ambient noise tomography. The velocity model covers the Illinois and Michigan Basins and their vicinities for approximately the top 200 km (Figure 2a). This study area is targeted for its excellent coverage of seismic data (Figure 2b) and well-studied surface geology. We explore simplified geodynamic models constrained by the imaged modern lithospheric structure to quantify the surface topographic evolution associated with the inferred lithospheric modification. We then compare the modeled topographic changes with geological and geochronological observations of tectonic subsidence and uplift in the two basins and the surrounding region (Allen & Armitage, 2012; DeLucia et al., 2018).

2. Data and Methods

2.1. Extraction of Rayleigh Wave Empirical Green's Functions

We downloaded vertical-component seismic waveforms recorded between 2011 and 2013 as 12-hr segments from 872 seismic stations within the study area from the EarthScope Data Management Center (Hutko et al., 2017; Figure 2). The seismic networks include the EarthScope Transportable Array (TA), N4, XO, NM, Z9, ZL, 7A, LD, XI, CN, IM, IU, OK, PE, PN, SS, and US. See Table S1 in Supporting Information S1 for the list of seismic networks, detailed descriptions, and associated references. We removed the instrument responses and downsampled the data to 5 Hz. We sliced the 12-hr waveform segment into 6-hr short-term segments with a two-hour overlap between segments. To avoid spectral leakage, the 6-hr time windows were tapered at the two ends using a Tukey window with a cosine fraction of 0.05. We then demeaned and detrended the short-term segments. We removed transient signals in the ambient noise with anomalous amplitudes. Specifically, we calculated the maximum absolute raw noise amplitude of each segment and the standard deviation of the raw noise amplitudes of all segments. We discarded waveform segments with peak amplitudes above 10 times the standard deviation.

Before cross-correlating the waveforms between station pairs, we normalized the short-term waveform segments (6-hr segments) with a moving-window average method. We computed the waveform cross-correlations in the frequency domain with the *SeisGo* Python toolbox (Yang, Bryan, et al., 2022; Yang, Zuffoletti, et al., 2022). After computing the short-term cross-correlations, we extracted the Rayleigh wave empirical Green functions (EGFs) as the negative time derivatives of the waveform cross-correlations (Gao & Shen, 2014; Yang & Gao, 2020). We then used the *Robust stacking* method (Yang, Bryan, et al., 2022) to stack the EGFs across 30-day segments. We convolved the EGFs with a Gaussian wavelet with a standard deviation of 2 s, the same as the Gaussian source time function for the waveform simulations in the full-wave tomography workflow described below. The scripts to extract EGFs and the resultant shear-wave velocity model are archived on Zenodo (Yang et al., 2025).

2.2. Full-Wave Ambient Noise Tomography

The full-wave ambient noise tomography method includes three main steps as documented in Gao and Shen (2014) and Yang and Gao (2020). Here, we briefly describe the key steps and parameters.

Step-1: finite-difference wave simulation. We simulated wave propagation in a 3-D spherical Earth using the nonstaggered-grid, finite-difference method (W. Zhang et al., 2012). The simulation model space is from -98° to -78.5° in longitude, from 34° to 48° in latitude and from the Earth's surface down to 1,006 km depth. The simulation model domain has a grid size of 0.05° by 0.05° in the longitudinal and latitudinal directions. The vertical grid size increases with depth from 1.8 km for the top 25 km to about 6 km at the depth of 100 km. The initial reference velocity model used in wave simulation consists of two parts: a global shear-velocity model of the crust and upper mantle for the upper 400 km (Shapiro & Ritzwoller, 2002) and the 1-D IASP91 model (Kennett & Engdahl, 1991) at greater depths down to the bottom of the model.

Step-2: measuring Rayleigh wave phase delays. We measured the difference in Rayleigh wave phase arrival times between the observed EGFs and the synthetics by searching for the time delay that maximizes the correlation coefficient between the two waveforms. We measured the phase delays in eight overlapping period bands, including 75–125 s, 60–100 s, 40–75 s, 30–60 s, 20–40 s, 15–30 s, 10–20 s, and 7.5–15 s. To include the uncertainty, we considered the phase delays measured from the substacks of the EGFs over 30 days and then averaged them for each station pair at each period band for inversion. We required the signal-to-noise ratio of the EGF to be ≥ 7 and the correlation coefficients between synthetics and observed EGFs to be ≥ 0.75 .

Step-3: inversion for velocity perturbations. We calculated the three-dimensional, finite-frequency sensitivity kernels of Rayleigh waves at the above eight period bands. We considered the influence of both compressional and shear wave velocities on the propagation of Rayleigh waves (Z. Zhang & Shen, 2008), which provides additional constraints for the shallow crust, reducing the influence of P-wave velocity uncertainties on deeper structures. Using 3-D finite-frequency sensitivity kernels allows us to resolve structures smaller than the dominant wavelength, as demonstrated by previous studies (Gao &

Shen, 2014; Tape et al., 2009; W. Wang et al., 2020). We solved for velocity perturbations using a damped least-squares scheme (Gao & Shen, 2014; Paige & Saunders, 1982; Yang & Gao, 2020) to match the phase delay measurements, using the conjugate gradient method to search for the optimized solution. We chose 8 as the damping and smoothing coefficient for all iterations based on tradeoff curves considering multiple coefficients, including 2, 4, 8, 12, and 16. The choice of damping and smoothing coefficients, as a common practice in geophysical inversions, balances the L2 norm of the velocity perturbations and the misfit between the observed and predicted phase delays. We used variance reduction and chi-square to quantify the model misfit. Based on our tests and trade-off analyses, the variation of damping and smoothing coefficients would only change the amplitudes but not the existence and overall geometry of major seismic velocity anomalies in our model.

To account for the uncertainties in the tomographic processing, we considered the error of phase delay measurements for each frequency range in the least-squares inversion. The error for frequency range k , e_{ph}^k , was computed as:

$$e_{ph}^k = \sqrt{e_{sub}^2 + e_{syn}^2}, \quad (1)$$

where e_{sub} is the standard error of the mean phase delays measured using the 30-day substacks and e_{syn} is the error of the synthetic Green's functions. e_{sub} was computed as:

$$e_{sub} = \frac{\sigma_{sub}}{\sqrt{N_{sub}}} + \hat{\delta}, \quad (2)$$

where σ_{sub} is the standard deviation of the phase delays measured using the 30-day substacks, N_{sub} is the number of substacks, and $\hat{\delta}$ is the resampling time interval in measuring the phase delays (0.2 s in our study). e_{syn} can be approximated as the time traveling across one half of the nominal horizontal grid size in the latitudinal direction ($d = 0.5 \times 0.05^\circ \times 110 \text{ km/degree} = 2.75 \text{ km}$), computed using an average velocity of 3.75 km/s (the midpoint between 2 km/s and 5.5 km/s). With these values, $e_{syn} = d/\hat{v} \approx 0.73 \text{ s}$.

Full-wave ambient noise tomography is an iterative imaging method, with each iteration containing the above steps. The velocity model was progressively updated for seven iterations when the changes between iterations became negligible. For iterations 1 to 5, we used 752 stations without the XI network covering the Mid-Continent Rift. These stations were added for iterations 6 and 7. The final velocity model after seven iterations shows significant improvements in Rayleigh wave phase delays, which are concentrated much closer to zero compared to the phase delays computed using the initial velocity model (Figure 3 and Figure S1 in Supporting Information S1). Since we are using Rayleigh waves, the resultant shear-wave velocity model reflects the velocity structure of vertically polarized shear waves.

2.3. Determination of Lithosphere-Asthenosphere Boundary Depth

Previous studies have explored approaches to extract the lithosphere-asthenosphere boundary (LAB) from seismic velocity models (e.g., Lloyd et al., 2020; Pasyanos et al., 2014; Priestley & McKenzie, 2013). For a highly heterogeneous velocity structure like ours, extracting the LAB depth by simply searching for the maximum negative velocity gradient (or the fastest velocity decrease) may fail. To consider all possible cases of vertical velocity patterns (see Figure S2 in Supporting Information S1 for four representative cases), we designed our algorithm by combining multiple criteria, including the maximum negative gradient, the amplitude of the velocity drop, the overall range of absolute velocity, and the presence of a shallower secondary velocity drop. We searched for LAB signals within the 40–250 km depth range after interpolating the velocity profiles with a depth interval of 0.5 km.

Our approach to identifying the base of high-velocity mantle lithosphere includes the following four steps. (a) Detect the first maximum in velocity at mantle depths searching downward, called the “peak.” (b) Detect the first turning point (the depth where the velocity starts to increase) below the first peak, called the “trough.” If there is no increase in velocity below the first peak, assign the trough as the bottom of the velocity profile. (c) If the velocity drop from the peak to the trough is too small, with a threshold of 0.3 km/s in our detection, then look for a second velocity decrease at greater depths below the first peak detected in step 1. (d) Once the final “peak” and

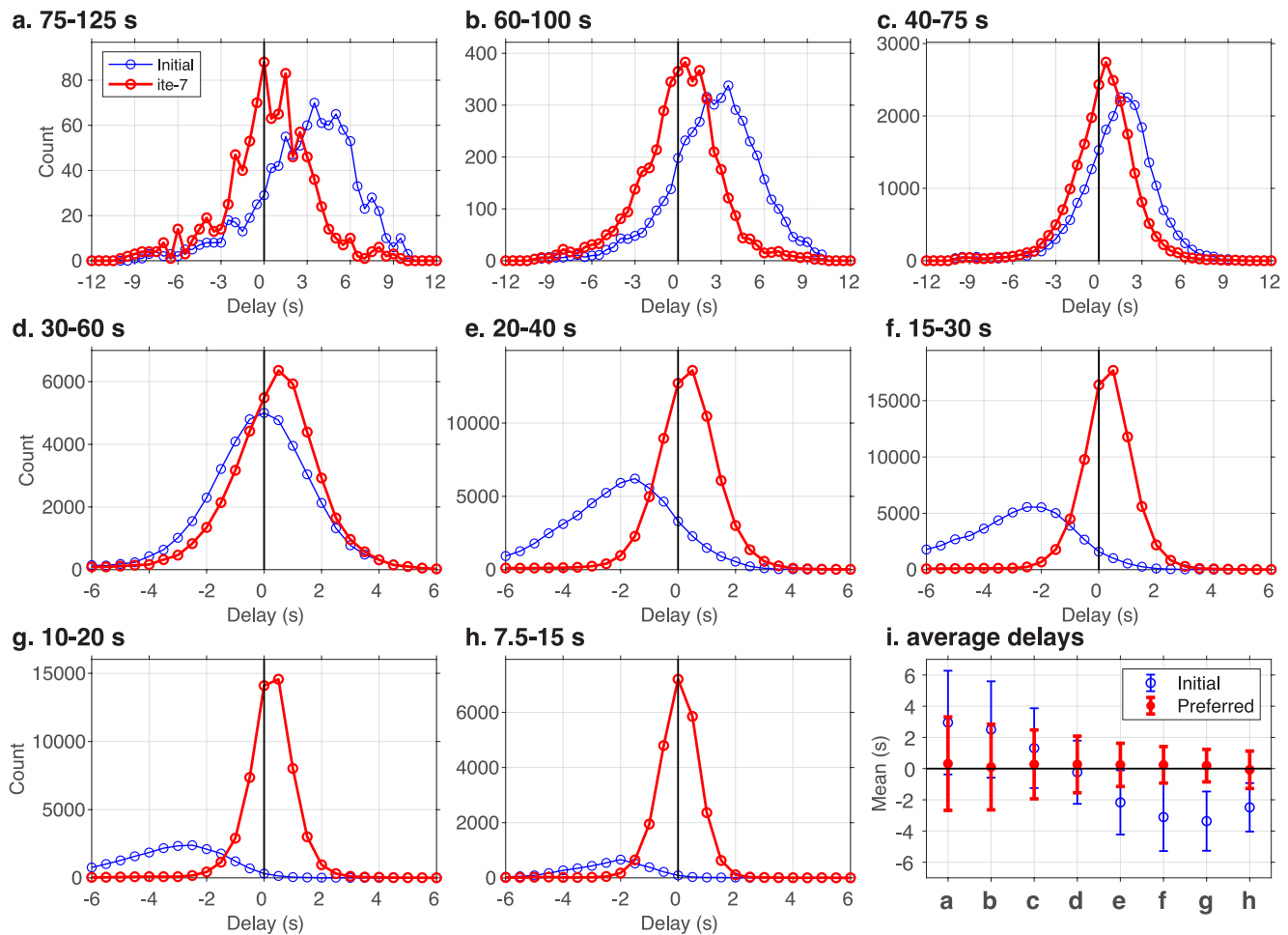


Figure 3. Distributions of phase delays at eight period ranges computed using the reference (blue) and final (red) models. (a–h) Phase delay histograms computed using the initial model (thinner blue) and the final preferred model after seven iterations (thicker red). (i) Comparison of average phase delays for each period computed using the reference model and the final preferred model (after seven iterations). The error bars show the standard deviations of the phase delays.

“trough” points are determined, extract the LAB as the depth with a maximum negative velocity gradient between the peak and trough. For regions with overall low velocities (<4.6 km/s) below the first peak, the LAB signal with a velocity drop lower than the above threshold (0.3 km/s) is also acceptable. See Text S1 in Supporting Information S1 for the implementation of these procedures in MATLAB.

3. Results

3.1. Key Seismic Velocity Features

The shear-wave velocity model reveals distinct high-velocity volumes within the mantle lithosphere, predominantly outside of the rifts and basins in the study area (Figures 4–6). At 60–120 km, the high-velocity anomalies are located primarily at the margins or in-between the main tectonic features, such as along the edges of the Illinois and Michigan Basins (H1–H7 in Figures 4 and 5). The H7 anomaly is dipping to the west and northwest and is connected with a larger high-velocity volume to the west (Figures 4c, 5, and 6a). The H8 high-velocity anomaly is beneath the central Illinois Basin, with lower velocities above and below it (L2; Figures 4c and 5). This forms a vertically slow-fast-slow structure (Figure 4c). The lateral extent of the H8 anomaly coincides with the span of the deeper negative velocity gradient identified by Chen et al. (2018) (Figure S5d in Supporting Information S1). Below 100 km, most high-velocity volumes are imaged to the west and northwest of the Illinois and Michigan Basins (Figure 6a). The average velocity in the high-velocity regions decreases from 120 to 160 km depth (Figure 6b).

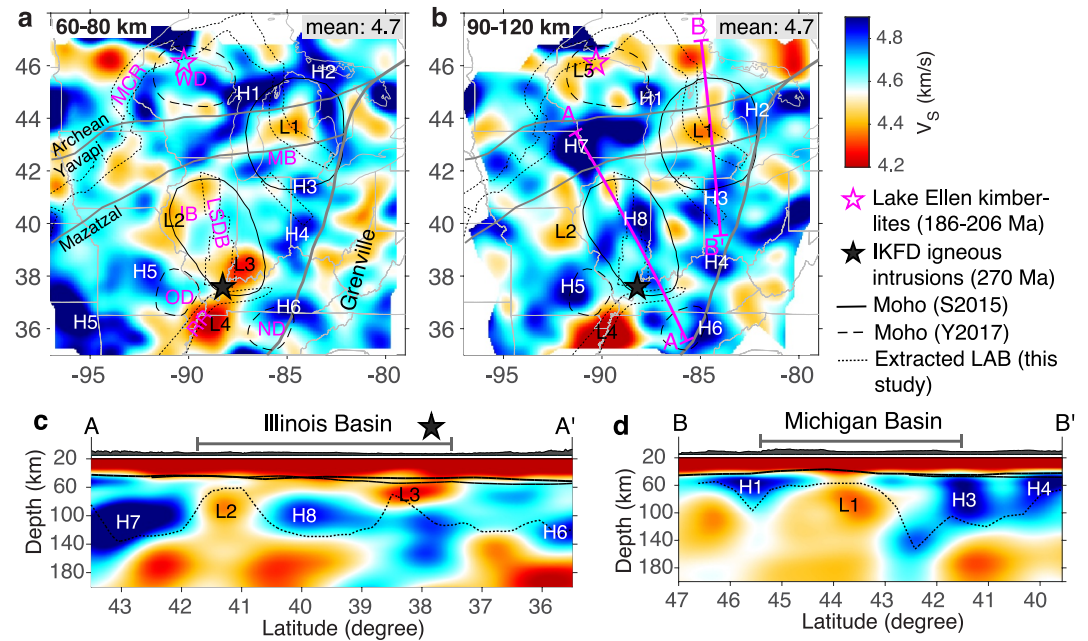


Figure 4. Shear-wave velocity model. (a, b) Average shear-wave velocities within the depth ranges of (a) 60–80 and (b) 90–120 km. Major high- and low-velocity features are labeled as H1–H9 and L1–L5, respectively. (c, d) Vertical cross-sections across the Illinois and Michigan Basins. Moho from Schmandt et al. (2015) (S2015) and Yang et al. (2017) (Y2017) are projected for reference. The dotted line is the extracted LAB from this study (see Section 2.3 and Figure 7 for details). The Lake Ellen kimberlites and IKFD igneous intrusions are marked as an open magenta star and a filled black star in (a–c), respectively. See (b) for profile locations. Figure S3 in Supporting Information S1 shows all panels in this figure with a non-divergent colormap (Cramer et al., 2020).

Outside of regions with high-velocity volumes, we observe multiple low-velocity anomalies above 120 km (Figures 4–6a). These low-velocity volumes, with velocities of approximately 4.2–4.4 km/s, are scattered in patches below multiple places. Most of them are concentrated below the central Michigan Basin (L1), the northwestern (L2) and southern (L3) Illinois Basin areas, the Reelfoot Rift (L4), the north-northwestern Wisconsin Dome area (L5), and to the north and northeast of the Michigan Basin close to the edge of our model (unlabeled). Among these low-velocity anomalies, L3 and L4 have the lowest velocities of 4.2–4.3 km/s, while L1 and L2 are about 4.4 km/s (Figure 4). L1 and L2 are connected with the broader low-velocity anomalies below about 140 km (Figures 4c–5). L1 partially overlies the H3 high-velocity feature (Figures 4d and 5). The lateral

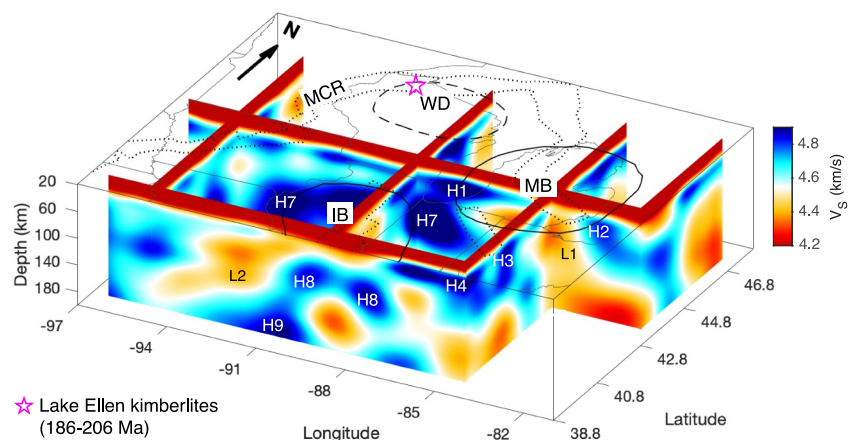


Figure 5. Perspective view of selected vertical cross-sections. The labels of major velocity features are the same as in Figure 4.

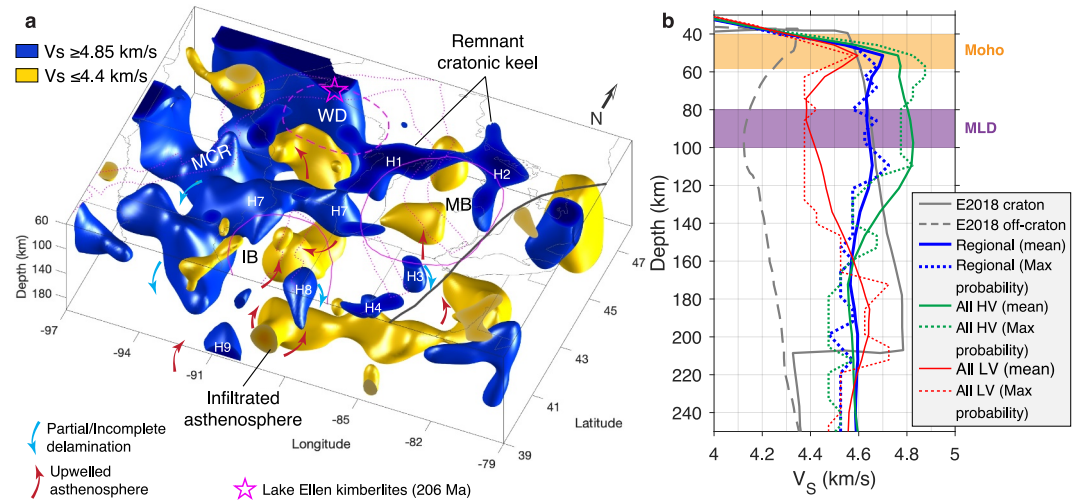


Figure 6. Perspective view of major velocity anomalies. (a) Major high (≥ 4.85 km/s; blue patches) and low (≤ 4.4 km/s; yellow patches) velocities. (b) Average 1-D velocity profiles within the footprints of the high-velocity volumes (HV; green), low-velocity volumes (LV; red), and across the entire imaging area (blue). We used the merged footprints of the HV (≥ 4.75 km/s) and LV (≤ 4.45 km/s) anomalies at 60–80 and 90–120 km. The velocity profiles below the western Canada cordillera (off-craton; dashed gray line) and the craton to the east (solid gray line) from Eaton et al. (2018) (E2018) are plotted for comparison. The regional Moho (Schmandt et al., 2015) and Mid-Lithospheric Discontinuity (MLD) (L. Liu & Gao, 2018) are labeled for reference. See Figure S4 in Supporting Information S1 for the probability distribution of the 1-D velocity profiles.

extent of the L2 anomaly partially overlaps with the extent of the negative velocity gradient at 60–90 km identified by Chen et al. (2018) (Figure S5c in Supporting Information S1), cross-validating the presence of a velocity drop in the uppermost mantle (bottom of H8) with limited lateral extent. The L3 anomaly in the southern Illinois basin is near the Illinois-Kentucky Fluorspar District (Figures 4a and 4c).

3.2. Variations of Lithospheric Thickness

The estimated LAB depth shows strong lateral heterogeneity (Figure 7a), ranging from 50 to 250 km, with most values concentrated at 50–150 km (Figure 7b). Because of the consideration of multiple characteristics of the 1-D

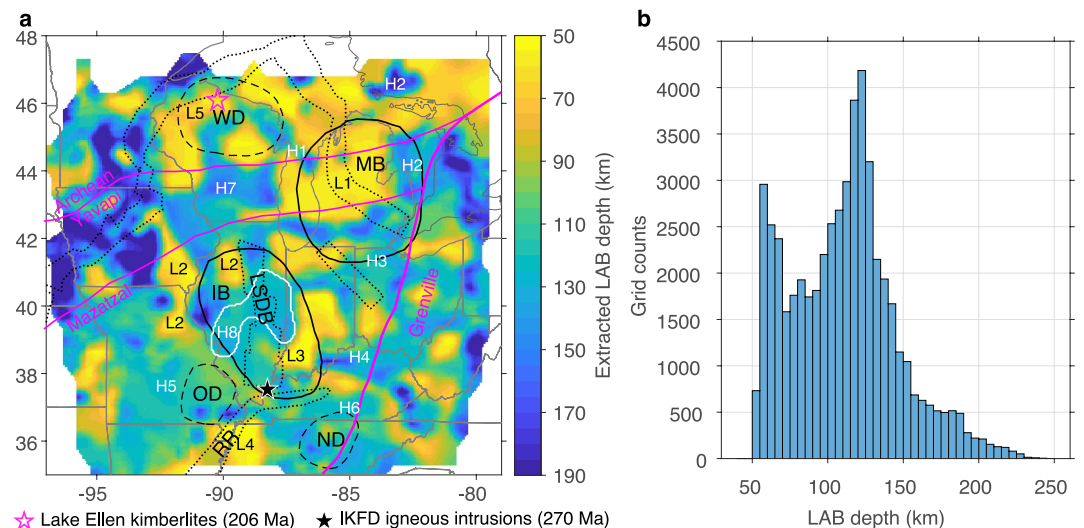


Figure 7. LAB depth determined from our velocity model. (a) Map view of the LAB depth distribution. We applied a 7-point average smoother (with a width of about 38.5 km) to the LAB result. The white line around H8 is the velocity isochron at 4.75 km/s. (b) Histogram of the LAB depths at all grids, computed using the smoothed model.

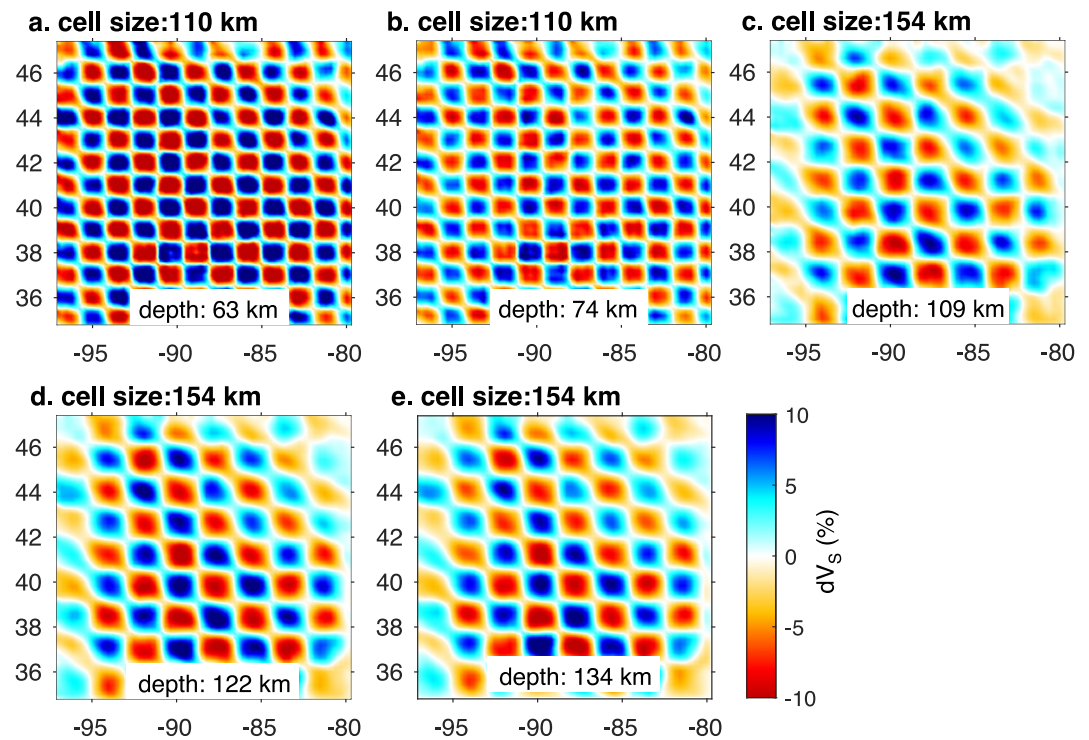


Figure 8. Checkerboard recovery results with different horizontal block sizes. These are a subset of all checkerboard tests to highlight the minimum resolvable horizontal scales of the velocity anomalies. See Figures S6–S8 in Supporting Information S1 for more resolution tests.

velocity profiles when searching for the LAB (see examples in Figure S2 in Supporting Information S1), there are very sharp lateral contrasts in some places, such as the shallow LAB (about 50–60 km) at the L2 low-velocity anomaly and the much deeper LAB (>190 km) immediately to the west (Figure 7a). This sharp edge in LAB depth is an artifact, though the relative deepening of LAB westward from the L2 anomaly is a robust feature. In our analyses below, we focus on the relative LAB depth patterns across the study area, instead of the absolute sharpness of lateral contrasts.

Regions with low-velocity anomalies also correspond to anomalously thin lithospheres within the study area (Figure 7a). The LAB depths show a bi-modal distribution, with two peaks at around 50–70 km and 110–130 km. The shallower peak correlates with regions above the imaged low-velocity anomalies (L1–L5). In particular, the LAB is only about 50–60 km deep below most of the central and northern Michigan Basin, extending to the north and northeast beyond the outline of the basin (Figure 7a). The deeper peak represents the regional average lithospheric thickness, consistent with the drop in velocity from 110 to 160 km along the regional velocity profile (Figure 6b). This depth range also marks the bottom of the imaged high-velocity anomalies (H1–H7; Figure 7a). Within the footprint of the H8 anomaly, although the velocity overlying H8 (about 4.4 km/s) is much lower than the ambient mantle (about 4.6–4.7 km/s), there is no noticeable velocity drop above H8 (Figure 4c). Therefore, the extracted LAB is at the bottom of the H8 anomaly. The deepest LAB is observed below and adjacent to the Midcontinent Rift, where large volumes of relatively high velocities are imaged (Figure 6a). However, these velocities are slightly lower than those of the high-velocity anomalies H1–H8 (Figures 4a, 4b, 5, and 6a).

3.3. Resolution Analysis

We conducted separate resolution tests to assess the horizontal and vertical model resolutions (Figure 8 and Figures S6–S8 in Supporting Information S1). These tests used synthetic data without noise, the addition of which may affect the percentage of amplitude recovery and the detailed geometry of the retrieved anomalies, but should not change our conclusion about the overall model resolution. To evaluate the horizontal resolution, we used standard checkerboard input models with alternating positive and negative velocity anomalies in horizontal grids in various scales (Figure 8). The vertical checkerboard pattern alternates every 20–50 km, increasing with depth.

We computed the synthetic phase delays using the sensitivity kernels for the tomographic inversion of the real data. We then inverted the noise-free synthetic phase delays to recover the velocity perturbations, following the same inversion procedure as for the real data. The checkerboard tests show that the shear-wave velocity perturbations can be well resolved with horizontal scales of >110 km at the depth range of approximately 60–100 km and >154 km at greater depths (Figure 8). These scales are smaller than the major seismic features discussed in this paper.

We used homogeneous layered models to examine the vertical resolution of our tomographic model. The layered models have vertically alternating fast and slow velocity patterns with different layer thicknesses (Figures S6 and S7 in Supporting Information S1). We followed the same procedure as in the checkerboard tests to recover the input velocity anomalies. We can recover layers with a thickness of >20 km at a depth of <80 km with an amplitude recovery of $>50\%$. At greater depths, the minimum resolvable thickness is 40 km with laterally variable amplitude recovery. Furthermore, Figure S8 in Supporting Information S1 shows checkerboard tests with vertical blocks at scales of 30–50 km below 50 km depth, increasing with depth. Based on these tests, at an amplitude recovery of $>50\%$, the vertical resolution is about 30 km shallower than 100 and 50 km at greater depths. All features discussed in the main text can be well-resolved with $\geq 50\%$ amplitude recovery in the horizontal and vertical directions.

4. Discussion

4.1. Comparison With Previous Velocity Models

Seismic imaging efforts over the past two decades have produced multiple seismic velocity models covering the North American midcontinent at regional, continental, and global scales. In Figures S9 and S10 in Supporting Information S1, we compare our velocity model with eight previously published and digitally available shear-wave velocity models (Cui et al., 2024; Golos et al., 2018; Hua et al., 2025; Schmandt et al., 2015; Shen & Ritzwoller, 2016; Simmons et al., 2010; Xie et al., 2018; Zhou et al., 2022) at depths around 70 and 100 km. The exact depth is dependent on the depth of the nearest model grid. Except for the GyPSuM model (Simmons et al., 2010), all other velocity models in this comparison used the EarthScope Transportable Array stations within the study area.

Variations in seismic velocities with different anomaly amplitudes are revealed by all of the plotted (Figures S9 and S10 in Supporting Information S1; Cui et al., 2024; Golos et al., 2018; Hua et al., 2025; Schmandt et al., 2015; Shen & Ritzwoller, 2016; Simmons et al., 2010; Xie et al., 2018; Zhou et al., 2022) and earlier (e.g., Bedle & van der Lee, 2006; Bedle & Van Der Lee, 2009; Chen et al., 2016) models. Most of the continental-scale models show a dominant velocity anomaly of $\pm 2\%$ to $\pm 3\%$. The two examined global models by Simmons et al. (2010) and Cui et al. (2024) have lower overall velocity perturbations than most of the other models, with a velocity perturbation of $\pm 1\%$ (Figures S9a, S9g, S10a, and S10g in Supporting Information S1). The velocity anomalies from this study (Figures S9i and S10i in Supporting Information S1) are predominantly around $\pm 6\%$, with some places reaching $\pm 9\%$.

In these models, the sporadic zones of low- and high-velocity anomalies are imaged with different scales, distributions, geometries, and amplitudes. In particular, low-velocity anomalies below the Reelfoot Rift region are imaged by all examined models (Cui et al., 2024; Golos et al., 2018; Hua et al., 2025; Schmandt et al., 2015; Shen & Ritzwoller, 2016; Simmons et al., 2010; Xie et al., 2018; Zhou et al., 2022). The L3 anomaly in the southeastern Illinois Basin region at 60–80 km in our velocity model is imaged by all but Zhou et al. (2022). Localized low-velocity anomalies below the Michigan Basin (L1 anomaly in our model) are also observed in the models by Schmandt et al. (2015), Shen and Ritzwoller (2016), Xie et al. (2018), Zhou et al. (2022), and Hua et al. (2025). Extensive low velocities below the Illinois Basin are imaged at various depths in the models by Simmons et al. (2010), Schmandt et al. (2015), Shen and Ritzwoller (2016), Xie et al. (2018), Golos et al. (2018), Cui et al. (2024), and Hua et al. (2025). The high velocity features H2–H6 in our model are also imaged by Hua et al. (2025), though generally at larger scales and lower perturbations.

While sharing some major velocity features, the velocity model by this study (Figures S9i and S10i in Supporting Information S1) reveals stronger velocity variations than other models. The examined continental-scale velocity models all reveal the contrast between the generally lower velocities to the west of the Rocky Mountains and the higher velocities to the east within the North American midcontinent (Cui et al., 2024; Golos et al., 2018; Hua

et al., 2025; Schmandt et al., 2015; Shen & Ritzwoller, 2016; Simmons et al., 2010; Xie et al., 2018; Zhou et al., 2022). However, they significantly underestimate the degree of internal heterogeneities within the mid-continent, as demonstrated by the comparison in Figures S9 and S10 in Supporting Information S1. A more accurate estimate of velocity perturbations is important in interpreting the mechanism of velocity variations and the associated mechanical processes (see detailed analyses in Section 4.2). In addition, our velocity model reveals more details in the scale, location, and geometry of the seismic features (e.g., H3, H4, H5, H6, and L1-L5) compared to the much smoother models, such as the models by Simmons et al. (2010), Schmandt et al. (2015), Shen and Ritzwoller (2016), Xie et al. (2018), Golos et al. (2018), Zhou et al. (2022), and Cui et al. (2024). Furthermore, our model also reveals new seismic features, such as the high velocities outside the Michigan Basin (H1 and H2), between the Midcontinent Rift, the Michigan Basin, and the Illinois Basin (H7), and below the central Illinois Basin (H8). These improvements by our velocity model enable the comparison between the outlines and extents of surface tectonic features and the seismic features at depth.

4.2. Mechanisms of Velocity Reduction in Mantle Lithosphere

The average velocity within the low-velocity domains at 70–120 km is about 4.4 km/s (Figure 6b), comparable to the global average asthenospheric velocity below Phanerozoic regions at similar depths (Fischer et al., 2020; French & Romanowicz, 2014). This velocity is approximately 0.4 km/s lower than the 4.8 km/s high-velocity volumes and 0.25 km/s lower than the regional average of 4.65 km/s (Figure 6b). Therefore, the average velocity anomaly within the low-velocity regions is approximately -8.3% relative to the high-velocity mantle or -5.4% relative to the regional average. In the following paragraphs, we discuss the possible causes of these observed velocity reductions relative to the surrounding higher velocities.

Multiple factors can lead to a reduction in shear-wave velocities in the upper mantle. Among these factors, the increase in temperature could significantly contribute to the observed velocity reduction. Assuming the velocity reduction is purely due to temperature increase, we can estimate the required thermal anomaly using scaling factors based on experimental studies (e.g., Goes et al., 2000; Jackson & Faul, 2010; Steinberger & Calderwood, 2006). Specifically, the observed low-velocity anomaly, with a velocity reduction of 5.4% relative to the regional average, corresponds to a temperature anomaly of approximately 208–359°C (see Table S2 in Supporting Information S1 for individual estimates using different scaling factors). It is about 320–552°C hotter than the surrounding high-velocity volumes. Lateral temperature differences of this magnitude can be created by variations in lithospheric thickness between 60 and 180 km across the study area (Figure 7), even for a constant mantle potential temperature. This is consistent with the consequence of compositional variations between lithospheric and asthenospheric materials, as detailed below.

The contributions from other factors would reduce the magnitude of the required thermal anomaly. Specifically, smaller grain sizes decrease shear-wave velocity through anelastic effects (e.g., Havlin et al., 2021; Jackson & Faul, 2010) and could help explain their velocity contrast with the high-velocity regions. However, the impact of grain size variations decreases as the temperature declines. Compositional variation within the mantle, such as enriched mantle peridotite that is typical of the asthenosphere (imaged low velocities) versus depleted mantle typical of cratonic lithosphere (imaged high velocities), could also contribute to seismic velocity contrasts in the relevant depth range (70–120 km). This would further reduce the required thermal anomaly. This scenario is particularly true for lithospheric erosion due to plume interactions (Eaton & Frederiksen, 2007). The potentially enriched mantle associated with the imaged low velocities would reflect the emplacement of asthenospheric materials, as a consequence of structural replacement. Partial melts (e.g., Chantel et al., 2016; Hammond & Humphreys, 2000; Havlin et al., 2021) and the presence of volatiles, such as water, carbon dioxide, and hydrous minerals (e.g., Karato & Jung, 1998; Saha & Dasgupta, 2019), could also significantly reduce shear-wave velocities. Volatiles may be from the subducted Farallon slab that reached the study area at 70–85 Ma and is currently stagnant below the study area within the mantle transition zone (Hua et al., 2025; Nyamwandha et al., 2016). The presence of volatiles could also help stabilize partial melts by lowering the required temperature (Saha & Dasgupta, 2019). However, the effects of partial melts and volatiles, if they exist, would most likely be restricted to the lowest-velocity regions within the low-velocity zones, such as the L3 and L4 anomalies below the southeastern Illinois Basin and the Reelfoot Rift areas (Figures 4a and 4b). Even when some water content (e.g., up to 300 ppm by weight) is allowed in the mantle to depress the solidus, significant effects from partial melt are not expected where shear-wave velocity is above ~ 4.35 km/s (Hua et al., 2023).

Furthermore, a relatively strong radial anisotropy, up to 9%, is imaged below the Illinois and Michigan Basins (Figure S11 in Supporting Information S1; Zhou et al., 2022), suggesting horizontal fabrics (e.g., sheets of partial melt Kawakatsu et al., 2009 or remnants of lateral mantle flow) with faster horizontally polarized shear waves than vertically polarized waves. Since the model from this study reflects the velocity of vertically polarized shear waves, radial anisotropy would also contribute to the observed velocity reduction. However, the observed radial anisotropy would, at most, account for 0.08 km/s (i.e., ~13%) of the total observed velocity reduction (Text S2 and Figure S11 in Supporting Information S1).

Overall, although the exact contribution of each factor is hard to estimate, the combined effects of factors other than temperature should be relatively small, unless there is infiltration of asthenospheric materials or partial melts due to structural alterations. These mechanisms are not completely independent. For example, the infiltration of asthenospheric materials could increase the temperature and may bring partial melts. High temperature anomalies may also lead to partial melt. Therefore, the reduction in velocity in the low-velocity zones likely reflects a significant structural modification with strong thermal anomalies and the intrusion of asthenospheric materials.

4.3. Widespread Localized Lithospheric Modification by Delamination

The widespread low velocities below about 60 km and associated lithospheric thinning found across the study area (Figures 4–6a) are unlikely to be related to processes during the accretion of this cratonic lithosphere at >1.6 Ga (Whitmeyer & Karlstrom, 2007), considering that thermal diffusion since then (with a temperature decrease of up to 240°; Bedini et al., 2004; Michaut & Jaupart, 2004) would tend to smooth sharp variations in lithospheric thickness. Furthermore, there is no notable correlation between the distributions of major seismic velocity features and lithospheric thicknesses and the extents of major Proterozoic terranes (Figures 4a, 4b, and 7a). Therefore, the localized thinning of the modern lithosphere is more likely a consequence of post-accretion modification, most intensive below the Wisconsin Dome and the Illinois and Michigan Basins (Figures 4–7).

We interpret the imaged intermittent high-velocity volumes along the edges of sedimentary basins and other major tectonic features (Figures 4–6a) as the remnants of cratonic mantle keels that have survived thermal and mechanical modifications by post-accretion tectonic events. The existence of low-velocity anomalies below regions of thinner lithosphere (Figures 4c and 4d) indicates the absence of mantle keels in these regions. Lateral variations in seismic velocity and thickness of the lithosphere can be attributed to multiple mechanical processes (L. Liu et al., 2025), including stretching and thinning under extension, buckling and thickening under compression, and erosion or delamination due to plume interaction or other disturbances. However, both stretching and buckling would mostly produce linear tectonic features, such as lithospheric thinning below rifts and lithospheric thickening below mountain ranges. Yet, the seismic velocity features and lithospheric thickness variations all show scattered, localized patterns instead of linear patterns (Figures 4–6a). Except for the Reelfoot Rift region, there is no clear correlation between lithospheric thinning and rifts (Figure 7a). The lithospheric thickness along the Midcontinent Rift ranges from <50 km to >190 km. The thinning below the Michigan Basin extends to a much broader region than the width of the underlying rift. Additionally, there are multiple locations with thin lithospheres without rifting structures, such as the footprints of the L2, L3, and L5 low-velocity anomalies (Figure 7a). Therefore, rifting is likely not the dominant mechanism that formed the widespread, localized lithospheric thinning. However, these faulting and rifting events may have created weak zones within the mantle lithosphere, facilitating later structural modifications. Buckling, on the other hand, would lead to lithospheric thickening and can not explain the observed thin lithosphere at some locations. In summary, lithospheric erosion or delamination is the most viable mechanism responsible for localized lithospheric thinning as observed in our velocity model. The low velocities below thinner lithospheres likely reflect the emplacement of asthenospheric materials, consistent with the discussion in Section 4.2.

The delamination of cratonic mantle lithosphere has been proposed as an important modification mechanism globally, based on seismic imaging (Bao et al., 2014; Celli et al., 2020; Eaton et al., 2018; Shen et al., 2018; H. Zhang et al., 2023) and geodynamic modeling (L. Liu et al., 2025; Peng et al., 2022; Shi & Morgan, 2024; Shi et al., 2021; Y. Wang et al., 2023). Cratonic mantle keels are traditionally considered thick mantle roots that are melt-depleted, seismically fast, and mechanically resistant to deformation (Eaton et al., 2018). Although a long-held view is that the cratonic mantle is neutrally buoyant (Jordan, 1978), recent studies suggest that these keels are denser than the ambient mantle, as required by topographic evolution and geoid data (L. Liu et al., 2019; Y. Wang et al., 2022b). These densified cratonic keels may result from their temporal change in temperature and

composition (Eaton & Perry, 2013; L. Liu et al., 2019). Negatively buoyant mantle keels may detach from the overlying lithosphere upon sufficient mechanical and thermal perturbations (L. Liu et al., 2025; H. Wang & Currie, 2015; Y. Wang et al., 2023), a process called delamination. Depending on the size, density, and viscosity of the delaminated mantle lithosphere relative to the ambient mantle, it can take 20–50 Myr for the delaminated lithospheric fragments to sink to the mantle transition zone (Peng et al., 2022). These detached mantle keels may eventually relaminate back to the base of the continental lithosphere, possibly with lateral offset, when being reheated in the deeper mantle and becoming buoyant, a process that may take over 100 Myr (e.g., Hu et al., 2018; Peng et al., 2022; Shi & Morgan, 2024; Z. Wang et al., 2022; Y. Wang et al., 2023). This cyclic delamination and relamination of the cratonic mantle lithosphere could play a critical role in the evolution of the cratonic lithosphere (L. Liu et al., 2025; Peng et al., 2022; H. Wang & Currie, 2015; Y. Wang et al., 2022b, 2023; Z. Wang et al., 2022).

The velocity reduction with depth associated with the low-velocity volumes (Figure 6b) resembles the velocity profiles of other regions around the globe with proposed lithospheric delamination, such as the Western Canadian Cordillera (Bao et al., 2014; Eaton et al., 2018) and West Antarctica (Shen et al., 2018). With a reduction of 0.6 km/s compared to the intact mantle keels, the imaged scattered low-velocity anomalies reflect strong localized structural modifications. They are overlain by a 50–60 km thick lithospheric lid with higher velocities (Figure 6b), reflecting a significantly thinned lithosphere (Figure 7a). Lithospheric thinning is most prominent below the central and northern Michigan Basin, the Wisconsin Dome and its surroundings, and the margins of the Illinois Basin (Figure 7a). Below the central Illinois Basin, although the detected LAB is comparable to the regional average, low-velocity materials (below 4.4 km/s) have infiltrated above the H8 anomaly and reached the crust (Figures 4c and 5).

We suggest that the distributed high-velocity volumes are the remains of cratonic mantle keels at various stages after heterogeneous lithospheric erosion or delamination. Specifically, these high-velocity features could reflect: (a) intact mantle keels (H1–H6), such as the high-velocity mantle lithosphere of the African craton (Celli et al., 2020), (b) partially detached lithosphere (the dipping H7), such as those observed below the western US (Schmandt & Humphreys, 2010), and (c) partially or completely detached lithosphere (H8 and H9). The H8 and H9 volumes may be portions of a larger mantle keel that is still in the process of detaching, having failed to detach completely during the most intensive delamination stage. They could be connected to the intact keel of anomaly H4 (e.g., Figure 5). Alternatively, H8 and H9 may be segments of keel materials that are now relaminating after reheating (e.g., Peng et al., 2022; Z. Wang et al., 2022; H. Zhang et al., 2023). The occurrence and rate of relamination are strongly controlled by the density anomaly and mantle potential temperature (Peng et al., 2022). If relaminated, a lateral offset is expected between their current and original positions (Peng et al., 2022). The lower velocities above H7 and H8 (Figures 4c and 5) could be asthenospheric materials infiltrating the shallow mantle after partial lithospheric detachment.

4.4. Mechanism and Timing of Inferred Lithospheric Delamination

Several mechanisms have been proposed to drive the delamination of the cratonic lithosphere, including plume interaction (Shi & Morgan, 2024; Shi et al., 2021; Y. Wang et al., 2023), boundary compression (Göğüş et al., 2011), lithospheric extension (Y. Wang et al., 2023), subduction-driven mantle convection (Göğüş et al., 2011), edge-driven mantle convection (Bao et al., 2014; King & Ritsema, 2000; van Wijk et al., 2010), and overall gravitational instability due to negative buoyancy (Elkins-Tanton, 2007; H. Wang & Currie, 2015; Y. Wang et al., 2022b). However, for localized delamination, as in our case, craton-plume interactions (Y. Wang et al., 2023) and gravitational instability (H. Wang & Currie, 2015) are more plausible. Negative buoyancy may come from the relatively dense mantle lithosphere (Y. Wang et al., 2022b) as a result of metasomatism (Griffin et al., 2009) and the possible presence of eclogitization (Jull & Kelemen, 2001).

Although we cannot entirely rule out lithospheric foundering driven by more recent processes, here we explore the idea that the gravitational instability was initiated by interactions with mantle plumes in the late Paleozoic to early Mesozoic and enabled by the presence of dense mantle lithosphere ((Y. Wang et al., 2023)). This gravitational instability, in the form of lithospheric delamination, was responsible for the heterogeneous remains of the cratonic lithosphere in our study region. It marked a substantial structural modification of the mantle lithosphere. An attractive feature of this hypothesis is that it explains the cessation of subsidence in the cratonic basins in this region in the late Paleozoic to early Mesozoic (Figure 2c), as explained in detail in Section 4.5.

The plume-craton interaction below the study area is supported by the overlap of the reconstructed plate history and the locations of the Pacific and African Large Low-Shear-Velocity Provinces (LLSVP) at the core-mantle boundary, which are commonly believed to be the source of mantle plumes (Heyn & Conrad, 2022; Heyn et al., 2020; Torsvik et al., 2014). The North American midcontinent passed over the Pacific and African LLSVPs at approximately 700–425 and 325–160 Ma (Merdith et al., 2021; Y. Wang et al., 2023) (Figures 1a and 2c), respectively. The mantle plumes of these LLSVPs could have heated the lithosphere and thermally weakened it, especially along metasomatized weak zones. Weakening could further destabilize the lithosphere and trigger delamination (Shi et al., 2021; Y. Wang et al., 2023). The delamination would then lead to surface uplift with the emplacement of hotter and lighter materials below the thinned lithosphere (Roy et al., 2009; Y. Wang et al., 2023).

The proposed lithospheric delamination below the North American midcontinent could have been initiated when this lithosphere passed over the African LLSVP from the late Paleozoic to the early Mesozoic. Although the lithosphere might have been preconditioned for subsidence by plume interaction in the late Proterozoic and early Paleozoic (Sleep, 2009), the structural signature of this earlier event in the mantle lithosphere would likely be more obscured by cooling and subsidence throughout most of the Paleozoic. Therefore, we explore the idea that the characteristics of the modern lithosphere revealed in our velocity model largely reflect its modification due to its later interaction with mantle plumes from the African LLSVP. This later interaction also coincides with the most recent evidence of major tectonic activity in this region, which includes a regional uplift at the Ozark Dome (DeLucia et al., 2018) and the broader central and eastern US area (Y. Wang et al., 2023) (Figure 2c). This time range also coincides with the sparse magmatism documented by the Lake Ellen kimberlites (186–206 Ma) northeast of the Wisconsin Dome (Heaman et al., 2004) and igneous intrusions (270 Ma) in the southern Illinois Basin within the Illinois-Kentucky Fluorspar District (e.g., Moorehead, 2013) (Figures 2a and 2c). The pressure conditions of the 186–206 Ma Lake Ellen kimberlite, which is located above a lithosphere of 50–60 km thick at present (Figure 7), suggest a lithospheric thickness of 140 km or more when it erupted (Heaman et al., 2004; McGee et al., 1983). This implies that the lithosphere in this region was undergoing thinning around this time, consistent with the process of lithospheric delamination. Intracratonic magmatism is one of the most important predictions of lithospheric delamination (e.g., Celli et al., 2020; Wu et al., 2019). The presence of this evidence of magmatism in the North American midcontinent in the late Paleozoic to early Mesozoic, although sparse, reflects a severe perturbation of the cratonic lithosphere, with lithospheric delamination being the most likely mechanism (Wu et al., 2019). The sparse evidence of magmatism suggests that either (a) most of the magma due to delamination may have been trapped below the relatively cold and rigid crust (e.g., Pieterek et al., 2024; Thybo & Artemieva, 2013) resulting in crustal underplating (e.g., Xiao et al., 2022, 2025; Yang et al., 2017) or (b), alternatively, the igneous extrusions were covered by Quaternary sediments across the majority of the study area (Soller & Garrity, 2018). The latter may explain the relatively fast seismic velocity in the middle crust below the Illinois and Michigan Basins (Shen & Ritzwoller, 2016). Therefore, the scarcity of surface magmatism does not preclude the occurrence of asthenospheric emplacement due to lithospheric delamination.

4.5. Surface Uplift Due To Lithospheric Delamination

To quantify the surface topographic changes caused by the inferred lithospheric delamination, we built a series of 2-D geodynamic models with a scale comparable to the seismic observations of the main velocity anomalies using the CitcomS finite-element code (Zhong et al., 2008). See Text S3 and Table S3 in Supporting Information S1 for a detailed description of the model configuration and parameters; T. Gerya, 2019; T. V. Gerya & Meilick, 2011; Hirth & Kohlstedt, 1996; L. Liu & Hasterok, 2016). The numerical models use simplified geometry for the delaminated mantle keel to highlight the overall geodynamic effect on topographic evolution, without considering the detailed geometry of the imaged velocity anomalies as in Figures 4–6. In reality, as indicated by the velocity model, the delamination may have occurred in small fragments scattered throughout the study area between the imaged high-velocity mantle keels. More sophisticated modeling in the future is desired to consider the detailed geometry of delamination and the variability of all modeling parameters.

To trigger delamination, the Sub-Cratonic Lithospheric Mantle (SCLM) has to be denser than the ambient mantle. The topography and gravity data (Y. Wang et al., 2022a, 2022b) suggest an excess density of 0.5%–1.2% for the SCLM, likely due to mafic compositions within the SCLM, such as garnet-lherzolite or eclogite (Y. Wang et al., 2023). Taking into account the relatively high velocity of the intact mantle lithosphere compared to the underlying asthenosphere (Figure 6b), we choose a density anomaly of 1.0% for the SCLM in our final model. Furthermore, the MLD within the cratonic lithosphere, likely enriched in hydrated minerals (Fu et al., 2022;

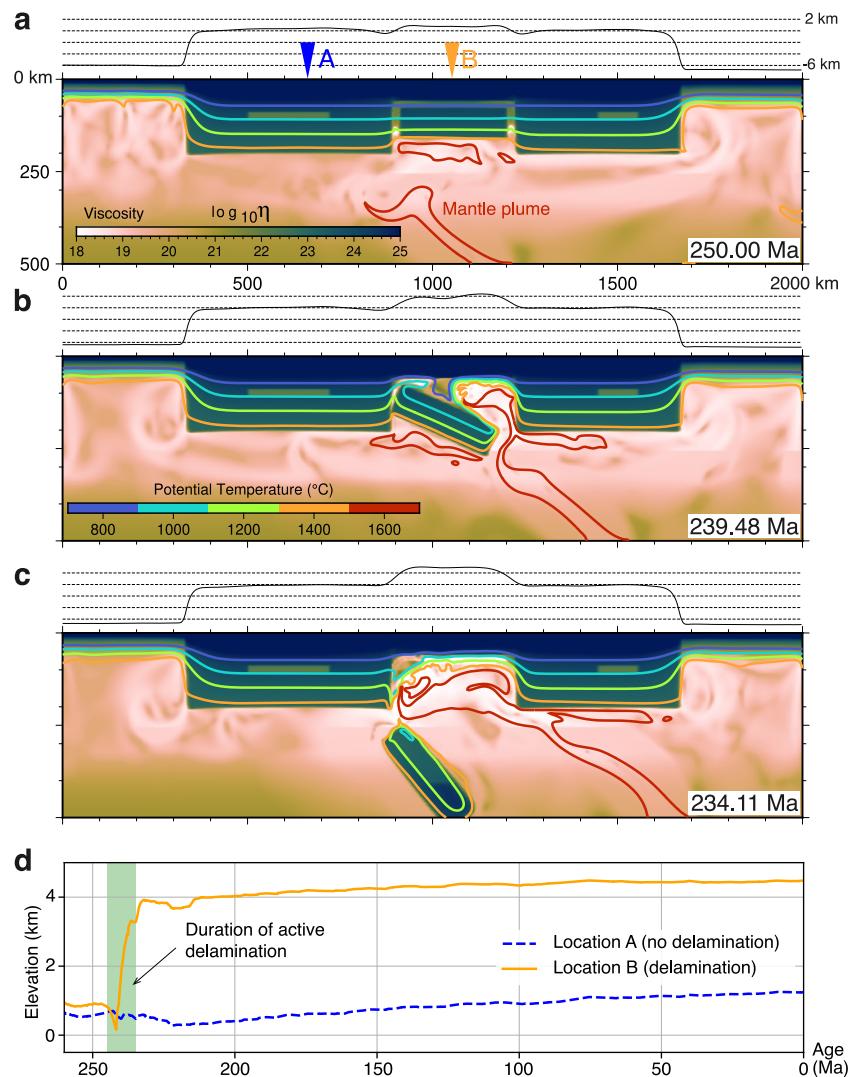


Figure 9. Geodynamic modeling of surface topography due to delamination. (a–c) Snapshots of the model evolution with delamination of the Sub-Cratonic Lithospheric Mantle. The colored contours represent the temperature field. The background image shows the viscosity field. The curve atop each snapshot shows the surface topography. The horizontal dashed lines above each snapshot have an interval of 2 km in elevation. See Figure S13 in Supporting Information S1 or snapshots from 185 Ma to 0 Ma. (d) Evolution of surface elevation over time at locations without delamination (location A) and with delamination (location B). See the top snapshot in (a–c) for the two locations. The green shaded area between 245 Ma and 235 Ma in (d) represents the duration of active delamination.

Z. Wang & Kusky, 2019), is a weak zone with a relatively low melting temperature. It plays a key role in the delamination of the lower SCLM (e.g., Z. Wang & Kusky, 2019; Y. Wang et al., 2023). We set the MLD as a hydrous peridotite, which is much weaker than the neighboring SCLM. It should be noted that MLDs are spatially intermittent in our model, which is also consistent with seismic observations (e.g., Hopper & Fischer, 2015; Krueger et al., 2021). Within the craton, the initial surface temperature is 0°C , linearly increasing to 580°C at the Moho (40 km depth) and 1350°C at the bottom of the cratonic lithosphere (200 km depth). This thermal profile is consistent with previous work on the continental thermal structure (McKenzie et al., 2005). Finally, we introduce a plume that is about 250°C hotter than the ambient mantle at the bottom of the lithosphere (Herzberg et al., 2007), equivalent to a temperature of 1600°C , and moves at a rate of approximately 3 cm/year (estimated based on Figure 1a). The moving plume is to simulate the passage of the North American plate over a fixed plume from the African LLSVP between 250 and 200 Ma (Figure 2c). See Figure S12 in Supporting Information S1 for the initial distributions of temperature, viscosity, and composition.

The geodynamic model from 280 Ma to the present day (Figure 9, Figures S13 and S15 in Supporting Information S1, and Movie S1) suggests that the topographic uplift induced by delamination provides a viable mechanism for crustal exhumation, sufficient to end subsidence in the Illinois and Michigan Basins and partially exhume the uppermost stratigraphy. When the hot plume arrives below the continent, it induces melt (red stippling in Figures 9a–9c) from the MLD material. The melt quickly migrates upward and propagates along the MLD, causing delamination of a portion of the SCLM around 239–234 Ma (Figures 9b and 9c). During delamination, the surface above the delaminated area is elevated by more than 3 km (Figures 9b–9d), due to the removal of the dense SCLM segment below the MLD and the infiltration of the hot and buoyant asthenospheric mantle (Peng et al., 2022; Y. Wang et al., 2023). The topographic evolution over time (Figure 9d) shows a prominent contrast of about 3.5 km between areas without (location A) and with delamination (location B). This surface uplift in the region with delamination relative to the surrounding region is sufficient to slow down and even invert the subsidence in the basin. It is also comparable to the exhumation of 3–4 km in the early Mesozoic at the Ozark Dome inferred from low-temperature thermochronology (Figure 2c). Although the Ozark Dome is located to the southwest of the Illinois Basin, it provides an upper limit on the estimated uplift for the Illinois Basin as part of the regional uplift in the early Mesozoic (Figure 2c). As the lithosphere moves away from the mantle plume, the rate of surface uplift gradually decreases, with relatively minor changes in surface elevation since about 235 Ma (Figure 9d). The long-term variation of surface elevation is consistent with the thermal history at the Ozark Dome (Figure 2c). A similar scenario of surface topographic evolution with delamination and post-delamination stabilization also applies to the Michigan Basin.

The modeling result shows the post-delamination cooling of the emplaced asthenospheric materials (Figures 9a–9c and Figure S13 in Supporting Information S1). After the passage of the mantle plume, the temperature of the emplaced asthenospheric materials decreases from 1600°C at around 240 Ma (Figures 9b and 9c) to 1200–1400°C at the present day (Figure S13c in Supporting Information S1) with localized higher temperatures (>1400°C; Figure S13c in Supporting Information S1). The cooling rate is about 83–167°C per 100 Myr, which is lower than the range of 267–350°C per 100 Myr modeled by Eaton and Frederiksen (2007) but overlaps with the rate of 100°C per 100 Myr by J. Liu et al. (2021). Compared to the intact mantle SCLM, the temperature anomaly below the thinned lithosphere after delamination is about 200–300°C, increasing with depth (Figure S13c in Supporting Information S1). This range of thermal anomaly is lower than the estimated temperature anomaly of 320–552°C associated with the imaged low velocities below 50–60 km. This is consistent with the discussion in Section 4.2 that other factors are required to fully account for the velocity reduction, such as the emplacement of enriched mantle from the asthenosphere and trapped volatiles. This aligns with the consequence of lithospheric delamination.

It should be noted that our model does not consider surface erosion or sedimentation. If these surface processes are considered, there are two competing effects: (a) the erosion of the uplifted area and the sedimentation around the surroundings will make the elevation contrast smaller; and (b) due to the isostatic adjustment, the erosion of the elevated area will induce more uplift, and sedimentation in the surroundings will cause more subsidence in the surroundings, increasing the resultant elevation contrast. These two competing effects may cancel each other out. Accounting for them should not alter the key results of the modeling that a prominent surface uplift should coincide with the onset of lithospheric delamination. A detailed quantitative evaluation of these two competing processes is beyond the scope of this paper. In addition, both locations A and B experience a slight elevation drop (0.3–0.4 km) at around 224 Ma when the plume moves away from the continent and is under the oceanic region to the right. The plume could lift the ocean floor slightly by thermal heating and dynamic upwelling. Since our code calculates the surface elevation by stress balance, the uplift of the right-side ocean will cause a slight relative subsidence (0.3–0.4 km) of the continent at around 224 Ma, similar to a seesaw. This minor artifact does not significantly change the modeled surface uplift due to delamination.

4.6. Implications for Lithospheric Evolution at Cratonic Basins

The new shear-wave velocity model (Figures 4–6) provides structural evidence of localized lithospheric delamination below the North American midcontinent. We argue that this imaged lithospheric modification is a consequence of the Mesozoic passage over the African LLSVP. The Michigan Basin is underlain by the eastern arm of the Midcontinent Rift, which is a failed rift initiated in the Neoproterozoic about 1.1 Ga (Stein et al., 2018). The Illinois Basin is underlain by the La Salle Deformation Belt, which was proposed to be initiated as a rift (Marshak & Paulsen, 1996) but later reactivated into an anticlinal feature (Hamburger & Rupp, 1988). These

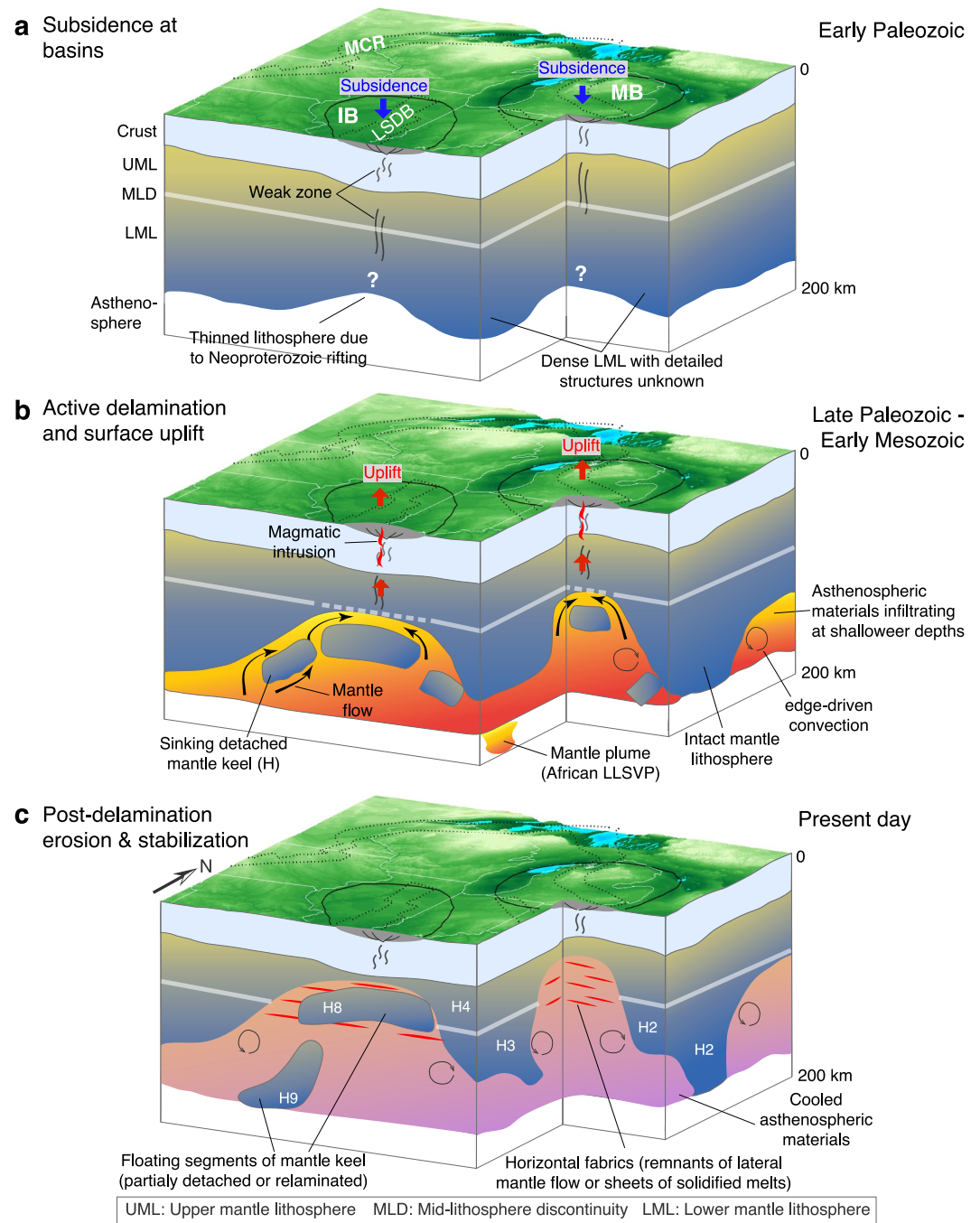


Figure 10. Lithospheric delamination causing the termination of basin subsidence. (a) Lithospheric structure during the subsidence at the Illinois (IB) and Michigan (MB) Basins in the early Paleozoic. (b) Delamination of the mantle lithosphere in the late Paleozoic to early Mesozoic. The resultant uplift eventually terminated the basin subsidence. (c) Post-delamination erosion and stabilization with cooling of the emplaced asthenospheric materials below the thinned lithosphere. The remnant cratonic keels H1-H9 correspond to the imaged high-velocity features in Figure 5.

preexisting lithospheric weak zones, together with the complexity of the plume head, may have facilitated the lithospheric delamination to concentrate below the Illinois and Michigan Basins. Although the detailed lithospheric structure before the onset of delamination is unknown, we argue that it presumably had a dense root, considering the prolonged subsidence, and was thinner below the Illinois and Michigan Basins due to the

Neoproterozoic rifting events associated with the initiation of the Midcontinent Rift and, possibly, the La Salle Deformation Belt (Figure 10a).

The distributed lithospheric delamination in the late Paleozoic to early Mesozoic would have induced asthenospheric upwelling and small-scale mantle convection, leading to surface uplift (Figure 10b). More buoyant asthenospheric materials infiltrated the space below the thinned mantle lithosphere, explaining the surface volcanism in the late Paleozoic in the southeastern Illinois Basin (Moorehead, 2013) and in the early Mesozoic in Lake Ellen northeast of the Wisconsin Dome (Figures 2a and 2c). The buoyant materials below the thinned remnant lithosphere would have pushed it up, resulting in surface uplift. The distributed asthenospheric upwelling caused a regional surface uplift with laterally variable amplitudes. With continued surface uplift, subsidence in the cratonic basins eventually stopped with possible basin inversion and erosion (Figure 10b). Small-scale convection below the newly thinned lithosphere could have been sustained after delamination, as shown in our dynamic model (Movie S1) and previous work involving a sharp contrast in lithospheric thicknesses (e.g., King, 2007; Peng et al., 2022; Roy et al., 2009; van Wijk et al., 2010). The delamination-induced uplift below the Illinois and Michigan Basins is consistent with the Type-2 termination of basin subsidence where the end of basin subsidence in the Triassic is followed by partial inversion in both basins (Figure 1b). The inversion of the basin would have removed the late Paleozoic to the early Triassic stratigraphic record of subsidence. This explains the gap between the end of subsidence retrieved from stratigraphic records in the Illinois and Michigan Basins and the exhumation observed in thermochronology data of the Ozark Dome and the broader central and eastern US (Figure 2c).

In the present day, active lithospheric delamination and asthenospheric upwelling could have ceased or at least be largely complete. Partially delaminated mantle lithosphere fragments are surrounded by the newly emplaced asthenosphere (Figure 10c). The relatively strong radial anisotropy below the Illinois and Michigan Basins (Zhou et al., 2022) (Text S2 and Figure S11 in Supporting Information S1), where low-velocity asthenospheric materials are imaged (Figure 7), suggests strong horizontal fabrics. They may reflect lateral mantle flow at the plume head when interacting with the overlying lithosphere (Ito et al., 2014; Movie S1) or due to the presence of horizontal sheets of solidified partial melt trapped below the thinned lithosphere after delamination (Kawakatsu et al., 2009) (Figure 10c). These horizontal fabrics could be remnants of past active asthenospheric upwelling during the lithospheric delamination (Figure 9). Without actively upwelling, the asthenospheric materials that infiltrated the space after delamination could be mostly stationary now relative to the surrounding lithosphere (Figures 6a and 10c). However, we expect that small-scale edge-driven convection still exists at places with sharp changes in lithospheric thickness (Figure 10c), which is shown in the numerical modeling results (Figure S13 in Supporting Information S1 and Movie S1). The sustained small-scale edge-driven convection may have helped the thermal anomaly to persist.

The lack of prominent post-delamination subsidence may have resulted from the competing processes of subsidence due to cooling and exhumation after erosion. Previous studies have proposed a cooling rate of approximately 40–105°C/Gyr (Bedini et al., 2004) or 50–150°C/Gyr (Michaut & Jaupart, 2004) for the continental lithosphere. This means that, without reheating, the temperature of the remnant lithosphere after delamination would have decreased for approximately 10–35°C since 235 Ma. The thermal subsidence of the thinned lithosphere due to this temperature decrease should be negligible, consistent with the geodynamic modeling result (Figure 9d). In addition, the thermal anomaly and buoyancy of the emplaced asthenosphere should have persisted to the present day (Figure S13 in Supporting Information S1). The exhumation at the Ozark Dome since the early Mesozoic (DeLucia et al., 2018; Figure 2c) indicates that erosion and isostatic exhumation may have eclipsed predicted thermal subsidence. Geological studies using detrital zircon geochronology (Gehrels et al., 2011; Finzel, 2014; Chapman & Laskowski, 2019; W. Wang & Bidgoli, 2019) and provenance analysis of sedimentary rocks (Stevens Goddard et al., 2023; Leary et al., 2020) indicate a westward continental-scale sediment transport system across North America from the Permian in the late Paleozoic to the Cretaceous. This system likely transported a large (but difficult to quantify) amount of sediments, including those in the Illinois and Michigan Basins, to the Ancestral Rocky Mountain basins (Gehrels et al., 2011; Leary et al., 2020). The relatively low (<1 km) modern surface elevation at the basins compared to the predicted uplift (3–4 km) suggests that erosion has dominated the post-delamination surface processes.

A thermal anomaly due to the emplaced asthenospheric materials is not reflected in surface heat flow measurements. The average surface heat flow within the North American midcontinent is approximately 40–50 mW/

m^2 , which is much lower than that of the Phanerozoic orogenic provinces outside the craton (Figure S14 in Supporting Information S1; Blackwell et al., 2011). However, when the data are examined without interpolation, some heat flow values within the Illinois and Michigan Basins reach 70–80 mW/m^2 (Figure S14 in Supporting Information S1). These values are comparable to some heat flow measurements within the central African craton (Lucazeau, 2019), where active lithospheric erosion due to plume interaction has been proposed (Celli et al., 2020). Nonetheless, the predominantly low surface heat flow measurements within the study area do not, in general, signify strong temperature anomalies within the mantle lithosphere. Therefore, despite some relatively high measurements, the overall surface heat flow pattern within the study area may be controlled predominantly by heat productivity within the crust, rather than by the temperature of the mantle lithosphere. The 1-D velocity profiles (Figure 6b) show much higher average velocities (4.6–4.8 km/s) above 50–60 km than those at greater depths. This suggests that the interpreted delamination and the associated thermal perturbations are largely limited to depths below 50–60 km. Furthermore, sedimentary cover, particularly within the basins, can also significantly decrease surface heat flow measurements due to its blanketing effect (Kim et al., 2020). On the other hand, the low velocities below the thinned lithosphere are higher than the velocities below the Basin and Range province in the western U.S. (Golos et al., 2018; Schmandt et al., 2015; Shen & Ritzwoller, 2016) and the Canadian Cordillera (Bao et al., 2014; Eaton et al., 2018), where much higher (e.g., >70 mW/m^2) surface heat flow is observed (Figure S14 in Supporting Information S1). In these regions, the shear-wave velocities are generally below 4.1 km/s. Alternatively, if contributions from other factors to velocity reduction, such as the emplacement of enriched mantle and the infiltration of volatiles (see Section 4.2), are more dominant, the required thermal anomaly would be decreased.

We note that dynamic topography due to mantle convection may also impact the vertical motion history of the study area, but may not be a dominant factor. For example, the relatively low present-day elevation in the broader region containing the Illinois and Michigan Basins is largely related to the Cenozoic dynamic subsidence at much larger scales driven by the sinking Farallon slab below the central and eastern US (L. Liu et al., 2008). However, because of its transient nature, longer wavelengths, and smaller amplitudes, dynamic topography may not reverse the long-term effect of lithospheric processes associated with the cratonic basins. Consequently, both the stratigraphic record and the thermal history of the basin sediments should reflect the localized deformation of the cratonic lithosphere below the basins.

5. Conclusions

We investigated the mechanism for the cessation of subsidence in cratonic basins. Using full-wave ambient noise tomography, we constructed a 3-D shear-wave velocity model of the North American midcontinent down to ~200 km. The new velocity model reveals strong heterogeneity within the mantle lithosphere, including low-velocity anomalies predominantly beneath the Illinois and Michigan Basins, where subsidence terminated in the late Paleozoic to the early Mesozoic. We proposed a model in which these low-velocity volumes, which are surrounded by distinctly higher velocities, reflect the infiltrated asthenospheric materials that were emplaced following lithospheric delamination. This lithospheric modification may be associated with the early Mesozoic tectonic exhumation observed in adjacent structures, such as the Ozark Dome to the southwest of the Illinois Basin, when this region passed over mantle plumes. The plume interaction likely triggered the lithospheric delamination. We exploited simplified geodynamic modeling to quantify the surface uplift due to the proposed lithospheric delamination. The numerical modeling shows that the emplacement of buoyant asthenospheric material would lead to a surface uplift of about 3.5 km, sufficient to terminate the subsidence in the cratonic basins. The findings of this study document delamination in the North American midcontinent and demonstrate the important role of lithospheric delamination in the evolution of cratonic basins. The distributed delamination provides a mechanical process that can drive basin-scale lithospheric modification, leading to changes in surface topography. Our findings also contribute to understanding the link between deep Earth processes and surface environments and how they influence resource distribution and long-term evolution of the topography.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

The continuous seismic waveforms are available through the EarthScope Data Management Center at <https://ds.iris.edu/ds/nodes/dmc/> and (Hutko et al., 2017). The velocity model is archived on Zenodo (Yang et al., 2025) and available through EarthScope Earth Model Collaboration (EarthScope DMC, 2011; Hutko et al., 2017) at <https://doi.org/10.17611/dp/emc.2025.nacratonfwant.1>. The open-source Python toolbox *SeisGo* for extracting empirical Green's functions from ambient noise is available on Zenodo (Yang, Zuffoletti, et al., 2022). The scripts for processing and plotting are archived on Zenodo (Yang et al., 2025). The full-wave ambient noise tomography codes are available on GitHub (<https://github.com/xyangp/FWANT>). The mantle convection code CitcomS is available at <https://geodynamics.org/resources/citcoms>.

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