Heterogeneous modification and reactivation of a cratonic margin beneath the
Korean Peninsula from teleseismic travel time tomography

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Key Points:

- Detailed upper mantle structure beneath the Korean Peninsula is imaged by teleseismic relative-time tomography
- A sharp transition in the degree of modification is observed between different Archean-Proterozoic continental lithospheres
- Strong low-velocity anomalies overlap with recent tectono-magmatic activities implying reactivation of the cratonic margin
Abstract
Phanerozoic plate tectonics have significantly modified the cratonic lithosphere in the eastern Eurasian plate. To reveal the detailed evolution of the cratons at their margins, we describe upper mantle structures beneath the southern Korean Peninsula (SKP) based strictly on teleseismic relative-time data. We found a thick (~150 km) high-velocity anomaly localized in the southwestern part of the SKP with large velocity contrasts ($dlnVp \approx 4.0\%$, $dlnVs \approx 6.0\%$) at its boundaries, suggesting the presence of a fragment of long-lasting cratonic lithosphere at the continental margin. Low-velocity anomalies beneath the Proterozoic Gyeonggi Massif and along the eastern margin primarily show regions with a shallower modified lithosphere. The strong contrast in velocity also indicates partial melting clearly corresponding to areas of Cenozoic basalts, higher heat flow, and high topography. Our results suggest heterogeneous break-up processes of old continental lithosphere and reactivation at its margins due to recent upper mantle dynamics.

Plain Language Summary
Constraining detailed structures at the margin of old continental lithosphere, i.e., the boundary at which modification processes of cratonic lithosphere intensely occur, is important to understand the evolution of continental lithosphere. The eastern part of the Eurasian plate is a natural laboratory that allows us to study processes due to strong effects from multiple episodes of continental collision and subduction of different oceanic plates since they first formed. By obtaining high-resolution images of the upper mantle to better understand these processes, we performed seismic tomography strictly using teleseismic relative-time data from stations in the southern Korean Peninsula (SKP) for the first time. We found sharp boundaries with a large lateral velocity gradient around a thick high-velocity structure in the southwestern SKP, which suggests a highly variable lithospheric structure. Relatively low-velocity anomalies found beneath the north and along the eastern margin indicate a high-temperature upper mantle that potentially contains a partial melt. In particular, the low-velocity anomalies show clear correlations with regions of Cenozoic volcanism, higher surface heat flow, and relatively high elevation with records of recent uplift. We suggest that the modification of cratonic lithosphere is highly heterogeneous, such that localized upper mantle dynamics may control lithospheric instability.
1. Introduction

Modification of the cratonic lithosphere is a salient process by which to understand the evolution and destruction of old continents (Artemieva & Mooney, 2002; Foley, 2008). Constraining the properties of continental lithosphere has indicated that marginal processes play an important role in altering entire cratonic structures (Artemieva, 2006; Thomas, 2006). Previous studies have suggested that interactions with slab subduction, mantle dynamics, and effects due to deeper structures may explain tectonic processes at the margin (e.g., Hu et al., 2018; Levander et al., 2014; Moresi et al., 2014; Schmandt & Lin, 2014). Persistent multiple episodes of such tectonic events generally overlap each other and blur the resulting features. Therefore, unraveling cratonic evolution and the modification of their margins is difficult (Rawlinson et al., 2014; Savage et al., 2017).

The continental lithosphere in the Eurasian eastern margin comprises several cratonic blocks (e.g., the North China Craton (NCC), South China Craton (SCC) and the Siberian Craton), which have experienced multiple tectonic events since the Paleozoic, including continental collisions and accretions, the subduction of various oceanic plates (e.g., Paleo-Asian, Paleo-Tethys, Mongol-Okhotsk, Izanagi, and Pacific slab), and subsequent back-arc extension and rifting (e.g., East Sea or Sea of Japan, Yellow Sea, and East and South China Sea) (Kusky et al., 2014; Tang et al., 2018). Previous studies have proposed that extensive reworking and destruction of the cratonic lithosphere mainly occurred in eastern China (Gao et al., 2004; Menzies et al., 1993) as evidenced by the loss of a significant portion of its deep root (from > 180 km to less than 100 km) during the Mesozoic and Cenozoic (e.g., L. Chen, 2010; Z. Huang et al., 2009). Widespread Mesozoic to Cenozoic magmatism indicates entirely non-cratonic geochemical and geophysical signatures (S. H. Choi et al., 2006; Kimura et al., 2018) accompanied by ubiquitous extensional basins (Chough et al., 2018; S. Liu et al., 2017; Ren et al., 2002). Due to the complexity, numerous different modes of lithosphere modification have been suggested for the cratons in China (e.g., Griffin et al., 1998; Wu et al., 2019; Zhang, 2005). In particular, current cratonic lithosphere margins have not been examined due to less sampling in the region, where dynamics in the upper mantle and effects due to nearby subducting slabs are significant.
Figure 1. (a) Regional tectonic map of northeast (NE) Asia. Major tectonic provinces are denoted by black dashed lines. Convergent plate boundaries are shown with red saw-toothed lines (Bird, 2003). Depths of the subducting oceanic slabs are indicated by brown dashed contours at 100-km intervals based on the Slab2 model (Hayes et al., 2018). Volcanoes are indicated by red triangles. CNCC: central North China Craton (NCC); ENCC: eastern NCC; KP: Korean Peninsula. (b) Map showing the geology of the KP (Chwae et al., 1995). Black solid lines indicate the boundaries of the different geologic provinces. GB: Gyeongsang Basin; GM: Gyeonggi Massif; IB: Imjingang Belt; JI: Jeju volcanic Island; NM: Nangrim Massif; OB:
Okcheon Belt; UI: Ulleung volcanic Island; YM: Yeongnam Massif. (c) Seismic stations used in this study. Stations from different networks are shown by different colors and symbols. KMA: Korea Meteorological Administration; KIGAM: Korea Institute of Geoscience and Mineral Resources; KINS: Korea Institute of Nuclear Safety; JMA: Japan Meteorological Agency; NEID: National Research Institute for Earth Science and Disaster Resilience; GSN: Global Seismic Network.

Tectonic affinities have been suggested in the cratonic lithosphere modification process between China and the Korean Peninsula (KP) (Cho et al., 2017; D. K. Choi, 2019; Oh et al., 2019). The KP is in the easternmost part of the Eurasian continent (Figure 1a). The basement of the KP consists of three late Archean to early Proterozoic massifs: the Nangrim (NM), Gyeonggi (GM), and Yeongnam (YM), from north to south (Figure 1b) (Chough et al., 2000; Cho et al., 2008). Two fold-and-thrust belts, i.e., the Paleozoic Imjingang belt (IB) and the late Precambrian-Paleozoic Okcheon belt (OB) separate the GM from the NM and YM, indicating collisional and accretional processes during their formation. Extensive Mesozoic-to-early Cenozoic plutonic emplacements followed in a post-collisional and subduction tectonic setting (S.W. Kim et al., 2011). Subsequent rollback of subducting oceanic plates affected the Precambrian lithosphere from eastern China to the YM, forming the Gyeongsang arc-backarc basin (GB) at the southeastern margin of the YM and rifted structures in the East Sea to the east of the KP (Ren et al., 2002; Chough and Sohn, 2010). Spatially non-uniform and complex tectonic structures indicate that the cratonic lithosphere in the KP was exposed to complicated modification processes (Chough et al., 2000).

In this study, we present high-resolution images of the upper mantle structure beneath the southern KP (SKP). Detailed 3-D P and S wave velocity variations were estimated using data from dense seismic arrays throughout the SKP with a recording time of more than five years. Unlike previous tomography studies in this region (e.g., C. Chen et al., 2017; Ma et al., 2018), we report velocity models based entirely on teleseismic relative-time residual data to exploit seismic rays that share common trajectories outside the model volume beneath the seismic array (Aki et al., 1977; Thurber, 2003). Therefore, we can estimate velocity anomalies in greater detail by measuring the small travel time differences caused by an area of interest. Together with a set of resolution tests, we suggest that heterogeneous marginal processes played a significant role in breaking up the continental lithosphere.
2. Data and Methods

We collected seismic data recorded by permanent seismic stations deployed in and around the SKP (Figure 1c), consisting of a total of 254 broadband, short-period, and accelerometer sensors. We selected 1,388 events between 2013 and 2018 with magnitudes greater than mb 5.4 at epicentral distances from 30–95° from the International Seismological Catalogue, identifying P and S waves from the vertical and tangential component data, respectively. The waveform data were corrected for corresponding instrument response and filtered with frequency bands from 0.1–5.0 Hz for P and 0.1–1.0 Hz for S waves. Interstation coherency in the P and S wave waveforms was used to measure the high-quality relative arrival time residuals (Rawlinson & Kennett, 2004). All teleseismic waveforms were visually inspected and noisy or incoherent data were eliminated to obtain more reliable measurements (Figures S1 and S2). Our final dataset comprised 96,273 rays from 684 events for P waves and 35,418 rays from 274 events for S waves (Figure S3). The average cross-correlation coefficient of our data set was 0.92 for P waves and 0.83 for S waves. The average uncertainties in the residual travel time observations (Rawlinson, Reading et al., 2006; Song et al., 2018) were 78 ms for P waves and 149 ms for S waves (see Text S1 in the supporting information for details).

Fast marching ray-tracing (Rawlinson, de Kool et al., 2006) and the subspace inversion scheme (Kennett et al., 1988) were used iteratively to perform the tomography. We defined a model space spanning ~800 km deep with a uniform grid spacing of ~10 km in the crust and ~20 km in the mantle. The crustal and mantle velocities were inverted simultaneously from starting models in Crust1.0 (Laske et al., 2013) and ak135 (Kennett et al., 1995), respectively, with a fixed Moho depth. Data misfit for the inversion was determined by the difference between the observation and travel time predictions based on the initial or recovered model with respect to the ak135 reference model (Rawlinson et al., 2016). Regularization factors were systematically determined by evaluating the trade-offs among data misfit, model smoothness, and model variance (Figure S4).

3. Model Estimation and Resolution Tests

Although the effect of velocity heterogeneities outside of the model space is negligible due to similar ray trajectories (Aki et al., 1977), strong velocity perturbations at less-constrained bottom depths could potentially bias amplitudes and patterns of the estimated anomalies above
(e.g., D. Zhao et al., 2013). In our study area, regional tomography at depths of 400–800 km suggested velocity perturbations of up to 5% in the $V_p$ and 7% in the $V_s$, mainly due to the stagnant Pacific slab in the mantle transition zone (C. Li & van der Hilst, 2010; Tao et al., 2018). We tested and minimized the impacts that the deeper anomalies had on our results. In comparison, a long-period pattern of similarity was observed as a function of the back-azimuth between the observed and synthetic residuals from a recent 3-D upper mantle velocity model for deeper depths (Tao et al., 2018) (Figures S5 and S6). We confirmed that this long-period pattern originated from deeper upper mantle structures. This effect caused insignificant (minor) velocity reductions ($d\ln V_p < 0.3\%$, $d\ln V_s < 0.5\%$) in our images for shallower areas (< 300 km), based on an additional inversion with a corrected data set (Figure S6c) generated by subtracting the synthetic data from the observations (Figure S7). As a conservative choice, we determined the results by inversion with the corrected data set as our final model for interpretations (Figure 2).
Figure 2. Models of $P$ and $S$ wave tomography. Light blue/red dashed contours in the depth profiles represent a 1% (1.5%) increase/decrease in $V_p$ ($V_s$), respectively. Tectonic divisions are indicated by black solid lines at a depth of 120 km. Regions with heat flow rates $> 70$ mW/m$^2$ are shown by black dotted contours at a depth of 60 km for the $V_p$. The locations of vertical cross-sections (a–d) are shown by black lines at a depth of 240 km for the $V_p$. Cenozoic volcanism is indicated by red triangles. Black lines in the vertical cross-sections indicate Moho interfaces. See figure 1 for the abbreviations.
The resolution of the velocity model was meticulously assessed by performing multiple synthetic recovery tests. All synthetic data were generated using an identical source-receiver combination as the actual data with Gaussian random noise, whose standard deviation was equivalent to the estimated residual uncertainties. First, we conducted conventional checkerboard tests using various scales (with diameters of 45, 60, 90, and 120 km) of high- and low-velocity anomalies: ±4% for $V_p$ (Figures 3a and S9) and ±7% for $V_s$ (Figures S8a and S9). The output models show good resolution, particularly beneath the SKP, and were able to resolve structures at 60 and 120 km checkers down to depths of 200 and 360 km, respectively. Second, a spike test was performed using discrete short-wavelength anomalies to better check the effects due to smearing (Rawlinson, Reading et al., 2006). We used spikes with a diameter of 60 km and maximum amplitudes of ±5% for $V_p$ (Figure 3b) and ±8% for $V_s$ (Figure S8b) distributed in similar positions and polarities in the observed anomalies (Figure 2). Input anomalies were generally well-identified, preserving their original polarities and locations without significant merging. Last, we tested the resolution with realistic structural input models, which applied simplified block anomalies to evaluate the vertical resolutions of interfaces (e.g., Youssof et al., 2015). The input models replicated the observed velocity contrast and pattern consisting of a positive anomaly ($d\ln V_p$ of +1.5% and $d\ln V_s$ of +2.0%) in the southwest and a negative anomaly ($d\ln V_p$ of −2.5% and $d\ln V_s$ of −3.5%) in the north and along the eastern areas of the SKP (Figures 3c and S8c) for two different thicknesses of 90 and 150 km. The original shape of the input structures was well-recovered with a velocity contrast of up to 65%. We interpret the resulting velocity anomalies (Figure 2) in the next section for the areas showing robust and reliable recovery in the resolution tests.
Figure 3. Resolution tests of the $P$ wave tomography. (a) Checkerboard test with diameters of 60 and 120 km and amplitudes of ±4% (see Figure S9 for anomalies with diameters of 45 and 90 km). (b) Spike test using anomalies with a diameter of 60 km and amplitudes of ±5%. Input models and their recoveries are shown at the top and bottom, respectively. Black dashed lines represent ±0.5% velocity contours. (c) Structural tests using high- and low-velocity blocks with amplitudes of +1.5% and −2.0%, respectively, with thicknesses of 90 (left) and 150 km (right). Black dashed lines represent +0.5% velocity contours.
4. Results and Discussion

We observed an overall pattern of relatively low velocities in the north and east of the SKP and relatively high velocities in the west and southwest (Figure 2). This pattern agrees well with previous tomography study results (e.g., Ma et al., 2018; Tao et al., 2018) that have shown a high-velocity block located beneath the SKP within a relatively slower upper mantle. In particular, previous studies have suggested the existence of slower upper mantle velocities (< – 1%) at shallow depths (< 300 km) beneath northeast (NE) Asia (e.g., Chang et al., 2015; Debayle et al., 2016; Legendre et al., 2015), which indicates the presence of a more thermally enhanced upper mantle (C. Li & van der Hilst, 2010), and mean velocities in our models that could be slower than the normal mantle. Recent regional studies have shown that upper mantle structures are likely complex in nature, suggesting the possibility of sharp contrasts in lateral velocities at the margins around the KP in the shallow upper mantle (J. Huang & D. Zhao, 2006; S. Kim et al., 2016). Our results provide greater details of the distribution of relatively high- and low-velocity anomalies beneath the KP and are consistent with previous observations.

A distinct high-velocity structure ($d\ln Vp$ of ~1.5% and $d\ln Vs$ of ~3.0%) extends ~150 km laterally and ~220 km vertically mainly at the southwestern part of the KP (cross-sections B–D in Figure 2). Synthetic tests using structural models (Figures 3c and S8c) showed that the vertical extent and amplitude of the high-velocity anomaly is well constrained. Together with an estimated low heat flow of < 50 mW/m$^2$ (Y. Lee, 2010), the Archean-Proterozoic surface geology (Figure 1b), and the possible affinity of the southwestern SKP to cratonic lithosphere in China, we note that the imaged high-velocity structure reflects relatively thick continental lithosphere preserved in the cratonic margin beneath the KP. Recent tomographic studies found similar small-scale positive velocity anomalies in the upper mantle beneath the eastern margin of the NCC, which were interpreted as fragments of thick lithosphere (e.g., He & Y. -F. Zheng, 2018; Xu et al., 2018). Paleoproterozoic subcontinental lithosphere signatures from distributed mantle xenoliths within reactivated regions of eastern China and the SKP also reflect the persistent but spatially heterogeneous feature of the continental lithosphere (S. R. Lee & Walker, 2006; K. -L. Wang et al., 2003; J. Zheng et al., 2001). The observed complexity of the cratonic lithosphere beneath the SKP based on our results and NE Asia from previous studies suggests a non-uniform modification process, which may correspond to episodic magma intrusions (K.
Wang et al., 2018; S. Li et al., 2019) and extension (Ren et al., 2002) due to the subduction and rollback of multiple slabs since the Mesozoic.

Figure 4. A three-dimensional plot of $P$ wave tomography indicating the main features interpreted. Structures in blue and red show +0.5 and −0.5% velocity isosurfaces, respectively. Red truncated cones indicate Cenozoic volcanism. Gray lines on the surface delineate tectonic boundaries. See figure 1 for the abbreviations.

Our results yield a striking localization of lithosphere modification among the different Precambrian massifs in the SKP. We note that the GM is missing its deeper part resulting in a clear velocity contrast at the boundary with the thicker lithosphere beneath the YM (cross-sections A and C in Figure 2, and Figure 4). Differences in the physical and chemical properties, including thickness, thermal state, and composition, affect the relative instability of the continental lithosphere (Foley, 2008; J. P. Zheng et al., 2015). Therefore, the loci of modification may be localized at an initial zone of lithospheric weaknesses due to collision, rifting, mobile belts or lithospheric interlayers (L. Chen, 2014; L. Liu et al., 2018b; Lu et al., 2011). Recently, it has been reported that the GM experienced multiple episodes of accretionary, i.e., arc-related and extensional, rift-related tectonic events through a series of pre- and post-collisional stages (e.g.,
Hyndman, 2019) since the Paleoproterozoic (Cho et al., 2017; Oh et al., 2019), which may be linked to the subsequent late Permian–Triassic collision between the NCC and SCC (S.W Kim et al., 2011; Oh & Kusky, 2007). Magmatic compositions in the NM and GM also support lithospheric modification, which suggests a coeval process with Mesozoic orogenies (J. H. Yang et al., 2010; Cheong et al., 2018). Inherent weaknesses beneath the GM during this period can explain the removal of thick lithosphere focused at this location, whereas there was less of an effect on the YM due to a long-lasting lithospheric core. However, a very recent and transient event (e.g., Levander et al., 2011; Bao et al., 2014) is more favorable due to less distinction in geological expression across the tectonic regions (Figure 1b) compared with the large lateral variation in our image (Figure 2). Previous studies have suggested that the localization of tectonic activities, including magmatism, extension, and uplift, can be attributed to mantle dynamic processes at the undulating bottom of the lithosphere (e.g., Artemieva, 2018; Rawlinson et al., 2017; Steinberger et al., 2019). In this case, weak properties could have maintained deeper regions of the GM and been removed during a short time period due to effects from a more recent event, such as the opening of the East Sea in the Miocene (e.g., Evanzenia et al., 2014; Shen et al., 2018) or the initiation of Philippine Sea plate subduction.

A similar large variation defines the eastern margin of the thicker lithosphere exhibiting a different level of lithosphere modification between the YM and GB (Figure 2). The continental and back-arc systems of the GB developed during the Early Cretaceous to early Tertiary due to the northward oblique subduction of the paleo-Pacific plate (Chough & Sohn, 2010). Subsequent rollback of the subducting slab yielded extensional stress and an injection of hot asthenosphere toward the mantle wedge, which resulted in the lithospheric basement thinning via thermal erosion or hydrous weakening (S. W. Kim et al., 2016; Kusky et al., 2014). In our image (Figure 2), a significant coincidence in the tectonic boundary between the YM and GM and a rapid velocity variation clearly captured the resulting difference in the lithospheric thickness, proving that subduction processes were the cause of lithospheric-scale modification in the southeastern SKP (Chough & Sohn, 2010).

We observed distinct areas of slower anomalies with $d\ln Vp < -2\%$ and $d\ln Vs < -3\%$ at shallow depths (60 and 120 km in Figure 2) along the eastern margin of the SKP. These areas are recognized as sub-lithospheric upper mantle due to the significant velocity contrasts ($d\ln Vp$ of ~4.0% and $d\ln Vs$ of ~6.0%) with the continental lithosphere in the YM. The magnitude of the
decrease in velocity is a minimum estimate based on synthetic tests using structural models (Figure 3c), where the low-velocity anomalies resolved up to 35% of the input velocity perturbation compared with 65% for the high-velocity anomalies. Taking into account the generally slower ambient upper mantle in NE Asia compared with the global average (~1% in Vp and ~2% in Vs), the observed reductions in velocity correspond to temperature increments of 300–350 K (Goes et al., 2000; Karato, 1993) with anelastic attenuation in the corresponding depth range (Adenis et al., 2017). Variations in the composition and grain size (velocity perturbation ~1%) (Cammanaro et al., 2003; Faul & Jackson, 2005) did not result in such a temperature increase even when incorporating relatively weak upper mantle anisotropy (< 1%) in this region (Kang & Shin, 2009; Wei et al., 2016). The presence of a small fraction of partial melt (1–2%) can explain the large drop in velocity (Hammond & Humphreys, 2000; Mavko, 1980), which has also been suggested in previous tomography studies (Simutė et al., 2016; Adenis et al., 2017). Prolonged subduction history of the current Pacific and former oceanic plates could have accommodated a hot and wet sub-lithospheric upper mantle in back-arc and adjacent continental margins (S. Kim et al., 2016; Tauxin et al., 2017). Subsequent continental rifting accompanied by upper mantle dynamics since the late Oligocene may have yielded partial melting at shallow depths, resulting in weakening and thinning along the eastern margin of the continental lithosphere.

We observed a clear correlation between the upper mantle low-velocity anomalies and areas characterized by Cenozoic basaltic eruptions (S. H. Choi et al., 2006; H.-O Choi et al., 2013; Won et al., 1994) and relatively high heat flow rate (> 70 mW/m²) (Figures 2 and 4). The topography of the lithosphere-asthenosphere boundary generally accumulated the ascending mantle flow beneath the thin lithosphere (Duggen et al., 2009; Steinberger et al., 2019). A steep gradient in lithospheric thickness also induces and maintains localized mantle upwelling and decompressional melting along the margins of the relatively thick lithosphere (Conrad et al., 2010; King & Anderson, 1998). The buoyant upper mantle located beneath the thinner lithosphere can disturb and reactivate the overlying continent (e.g., Karlstrom et al., 2012; X. Yang & Gao, 2018). This mechanism corresponds to the positive residual topography associated with the negative mantle gravity anomaly (Kaban et al., 2016), high elevation with geomorphic disequilibrium in the late Pleistocene (< 125 ka) (D.-E. Kim et al., 2016) observed in the eastern mountain ranges. Small-scale sub-lithospheric upper mantle convection can continuously play a
role in the reworking and modification of thicker lithosphere in a similar manner as asthenospheric swell or convective mantle infiltration through marginal heterogeneities (e.g., Liu et al., 2018a, 2018b; Z. Wang & Kusky, 2019). The observed sharp contrast in mantle velocity and spatial coincidences in recent surface magmatic activities exhibit a retaining destabilization process of the cratonic lithosphere margin.

5. Conclusions

High-resolution upper mantle seismic tomography constrained by teleseismic relative-time dataset revealed detailed continental lithospheric structure that has experienced heterogeneous modification and reactivation beneath the KP. Distinctive lithospheric features of the adjacent different Precambrian massifs indicate different responses of each lithosphere on the recent marginal tectonic processes. An anomalously thick high-velocity structure beneath the YM suggests a fragment of cold, resistant cratonic lithosphere. The absence of deeper lithosphere and mostly occupied by high-temperature, buoyant upper mantle beneath the GM, continental arc and back-arc system (GB) in the southeast, and along the eastern margin show highly modified regions. A clear spatial coincidence of low velocities and recent tectono-magmatic activities suggest persistent reactivation of cratonic margin possibly controlled by intensive interaction of the prominent lithospheric structures and convective upper mantle.

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