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

Abstract: Margins of old continental lithosphere are likely prone to ongoing modification processes. Therefore, constraining detailed structures beneath the margin can be essential in understanding the evolution of the continental lithosphere. The eastern margin of the Eurasian plate is a natural laboratory that allows us to study the strong effects from multiple episodes of continental collision and subduction of different oceanic plates since their formation. To reveal the detailed evolution of cratons at their margins, we describe, for the first time, the upper mantle structures beneath the southern Korean Peninsula (SKP) based strictly on teleseismic relative arrival time data from densely deployed local seismic arrays, which allows us to constrain the details of the lithospheric structures beneath the Archean-Proterozoic basement.

We imaged a thick (~150 km) high-velocity anomaly mainly beneath the Proterozoic Yeongnam massif with large velocity contrasts (dlnVp ≈ 4.0% and dlnVs ≈ 6.0%) at its boundaries, suggesting the presence of a long-lasting cratonic root in the southwestern SKP. On the other hand, low-velocity anomalies were found beneath the Proterozoic Gyeonggi Massif, Gyeongsang arc-back-arc basin, and along the eastern margin of the SKP, indicating significantly modified regions. The possible existence of a remnant cratonic root beneath the SKP and contrasting lithospheric structures across the different Precambrian massifs suggests the highly heterogeneous modification of cratonic lithosphere at the eastern Eurasian plate margin. Strong velocity reductions, which indicate a thermally elevated upper mantle with potential partial melts, correspond to areas of Cenozoic basalts, high surface heat flow, and high topography along the eastern KP margin. We interpret this coincidence as a result of recent reactivation of a craton margin, which is controlled by intense interaction between the convective upper mantle and heterogeneous continental lithosphere.

Response to Reviewers: Dear Editor:

We appreciate your consideration of our manuscript. We carefully reviewed all comments made by the editor and reviewers and present our responses. The corresponding modifications/additions are also marked in the revised manuscript (written in blue). We believe that our corrections have improved the original manuscript. Furthermore, following reviewer suggestions, we used a professional English editing service to thoroughly
check and revise the English of the revised manuscript. We hope that the manuscript conforms to the standards of Gondwana Research.

Before we present our responses to the comments from Reviewer #1 comments, we would first like to address the editor’s suggestion.

Editor's comment:
"I would also suggest that you add expanded discussion on the nature of craton destruction by consulting papers on the North China Craton where several geological and geophysical studies have shown the nature of extensive craton destruction, particularly in the eastern domain. You can also see a paper that integrates both geological and geophysical techniques to evaluate the decretionization in the North China Craton [Santosh, 2010, Precambrian Research, on Columbia]. There are also several recent papers that link the differential destruction of the craton with gold metallogey [Shengrong Li and others, papers in GR, OGR etc.]. Thus, you can expand your model into a more regional picture for the Sino-Korean craton as a whole, which will enhance the appeal of your excellent work."

Response to the Editor's comment:

We sincerely thank the editor for this constructive suggestion with respect to our original manuscript. As suggested, we expanded our discussion on the nature of craton destruction along the Korean Peninsula associated with the evolution of the North China Craton (NCC) and Sino-Korean craton. We have added related discussions with relevant references, including those that the editor has suggested.

We have added the following sentences to the revised manuscript (lines 336–353):

"Heterogeneous lithospheric structures have been observed in the NCC. An extensive but non-uniform modification process of craton lithosphere was suggested in the central and eastern NCC. Seismic tomography and anisotropy (e.g., Cheng et al., 2013; Jiang et al., 2013) and S wave receiver function analysis (e.g., Chen et al., 2009) revealed the heterogeneous structure of the craton lithosphere, indicating its complex nature due to deformation. An integrated thermomechanical model, jointly inverting heat flow data, surface wave dispersion curves, geoid height, and absolute elevation (Guo et al., 2016), demonstrated the highly heterogeneous physical state of the NCC developed by the uneven destruction of craton lithosphere due to sublithospheric mantle convection. Spatial distribution and temporal variation in geochemical characteristics of the Mesozoic ore deposits (e.g., gold metallogeny) (e.g., Li et al., 2014, 2017) were interpreted to have resulted from inhomogeneous decretionization. Therefore, the heterogeneous destruction and reactivation of the craton lithosphere occurred extensively along the Sino-Korean craton margin, which includes the KP and NCC. The structural heterogeneity of the craton lithosphere was likely much sharper in scale and thickness within a confined area, as observed at the current continental margin of the KP. A distinct spatial correlation between active surface tectono-magmatic processes and modified regions indicates that upper mantle dynamics played an important role in deforming the craton lithosphere, which was facilitated by multiple subduction-collision events of oceanic/continental plates at the continental margin (e.g., Santosh, 2010; Cai et al., 2018)."

References (newly added in the revised manuscript):

Reviewers' comments:

Reviewer #1: In this paper, the authors present and interpret new body wave teleseismic tomography results from the Korean Peninsula. Approximately 5 years of data from a large and dense array of seismometers is used to generate P-wave and S-wave images of the crust and upper mantle. One of the main findings is evidence of modification of the cratonic margin beneath the Korean Peninsula, with a particularly clear association of Cenozoic volcanism and low velocities in the upper mantle. Overall, I thought this was a very nice paper, with a particularly thorough and well done application of teleseismic tomography, which included a useful range of synthetic reconstruction tests. As such, I recommend publication in Gondwana Research subject to minor-moderate revisions. Note that I reviewed a previous version of this paper that was submitted to GRL; this review is an updated version of that review. It appears that the core of the paper is the same, but a few new figures and extra text has been added. It is notable that the additional text is not of the same standard as the original in terms of quality of prose/grammar.

(0) Highlight 4: What is meant by "Retained reactivation"?

(1) Lines 13-15: This sentence does not make sense, particularly the second half of it. Overall, the written English could do with some improvement; I will point out a few instances below, but not all of them.

(2) Line 19: Should be "seismic arrays", and later in the sentence "finding out fine details" is poor expression.

(3) Line 83: I wouldn't say that they "suffered different tectonic processes" - plate tectonics is not a disease. Perhaps "were subject to" instead of "suffered"
What are the responses of the short period instruments and accelerometers used? Do they seem noisier than the broadband stations at 10 s period? I am slightly surprised that the long-period cut-off for the P-wave and S-wave band pass filtering are the same - normally one would allow longer period S-waves than P-waves, but perhaps this is related to the use of heterogeneous instrumentation. It would be good to know the proportions of the three types of instruments used and their relative SNRs for the period ranges that are targeted.

This sentence needs to be fixed.

Are there local 3-D crustal models available e.g. from receiver functions or ambient noise tomography? I suspect that Crust 1.0 is unlikely to be very useful given the density of stations and hence the short-scale-length of structures that can be resolved.

I'm not sure that it can be claimed that heterogeneity outside the prescribed model region is negligible, but that is what is assumed in teleseismic tomography - rightly or wrongly.

I'm not sure that I feel entirely comfortable with what has been done here. If you subtract model predictions from the observations, and say that these are due to structure between 400-800 km depth that are poorly resolved, then what do you get in the 400-800 km depth range when you invert the adjusted dataset? Also, presumably the synthetic residuals will be influenced by structure in the 0-400 km range (or is this set to ak135 when you do the predictions?), so are you not just removing long-wavelength structural effects throughout the whole local model region?

The figure is greyscale, so it is not accurate to say that residuals are colour-coded.

This figure is ok - a bit busy perhaps and it is not entirely clear to me what the yellow/orange and light blue colours represent.

Not much is said about crustal structure, even though it is explicitly inverted for. I realise that the velocity perturbations likely define "bulk" properties in the sense that they probably represent depth-averaged perturbations through the full crustal thickness, but nonetheless they may be useful to understand.

What is implied by the statement "reactivate the overlying continent"?

Not sure what is meant by a "retaining destabilisation process".

We would like to thank Reviewer #1 for constructive comments and suggestions made for our original manuscript. We carefully went through all comments and present our responses below.

We would like to thank Reviewer #1 for constructive comments and suggestions made for our original manuscript. We carefully went through all comments and present our responses below.
We thank Reviewer #1 for pointing out the unclear meaning of this phrase. We intended to suggest that the upper mantle dynamics sustained a continuous reactivation of the craton margin. We removed the word "retained" in the revised manuscript to clarify the meaning. We also combined Highlights 3 and 4 in the original manuscript to conform to journal formatting.

Highlights 3 and 4 in the original manuscript:
Highlight 3. A sharp transition in the degree of modification among Archean-Proterozoic massifs
Highlight 4. Retained reactivation of the craton margin by upper mantle dynamics

Highlight 3 in the revised manuscript:
Highlight 3: Non-uniform modification and reactivation of the craton margin via mantle dynamics

Response to Reviewer Comment No. 1:
We appreciate the reviewer's comment. We obtained the services of a professional English language editing company to thoroughly check and revise the written English in the revised manuscript. We modified this sentence to clearly indicate the original meaning.

From the original text (lines 13-15):
"Constraining detailed structure at the margin of old continental lithosphere is important for understanding the evolution of continental lithosphere due to the most active modification processes of it."

To the revised text (lines 14-16):
"Margins of old continental lithosphere are likely prone to currently ongoing modification processes. Therefore, constraining detailed structures beneath the margin can be essential to understand the evolution of the continental lithosphere."

Response to Reviewer Comment No. 2:
We thank the reviewer for their comment. We have modified this sentence as suggested.

From the lines 18-21 in the original manuscript:
"... teleseismic relative arrival time data from densely deployed local seismic array for the first time, finding out fine details of lithospheric structure beneath the Archean-Proterozoic basement."

To the lines 20-22 in the revised manuscript:
"... teleseismic relative arrival time data from densely deployed local seismic arrays, which allows us to constrain the details of lithospheric structures beneath the Archean-Proterozoic basement."

Response to Reviewer Comment No. 3:
We appreciate this suggestion on the expression. We have corrected this by replacing "suffered" to "were subject to" as suggested in the revised manuscript (line 74).

Response to Reviewer Comment No. 4:
We used three types of instruments: broadband (Guralp CMG-3T or Kinemetrics STS-1 or 2), short period (Kinemetrics SS-1 or Guralp CMG-40T-1), and acceleration (Kinemetrics ES-T or ES-DH) sensors. We present examples of the instrumental responses for the corresponding sensors in Fig. R1 of the detailed response to reviewer file. To determine an appropriate frequency range, we tested different low cut-off frequencies (e.g., 0.03, 0.05, and 0.1 Hz). With a minimum frequency of 0.1 Hz, we were able to obtain more stable measurements that were less affected by noise. For all used data, we compared the relative mean SNRs between the different instruments for various low cut-off frequencies (i.e., 0.03, 0.05, and 0.1 Hz). The smallest variation in the SNR between the different types of sensors was estimated when using the 0.1 Hz corner frequency (Fig. R2 in the detailed response to reviewer file). We included a remark on SNR variations in the method section of the revised manuscript.

In the revised manuscript (lines 112-114):

"With these frequency ranges, the highest degree of waveform coherency was achieved while maintaining similar levels of signal-to-noise ratios among different types of instruments."

Response to Reviewer Comment No. 5:

We would like to thank the reviewer for their comment. We have modified the text as suggested.

From the original text in the original manuscript (lines 135-137):

"It is noted that the processed waveforms are highly coherent each other enough to measure relative delay times accurately."

To the modified text in the revised manuscript (lines 116-117):

"The waveforms of the processed data are similar in shape, which allowed us to measure the relative residuals with high precision."

Response to Reviewer Comment No. 6:

We would like to thank the reviewer for raising this question. We inverted our dataset with a local model of Moho depths (Chang and Baag, 2007; Kim et al., 2015) and a 3-D velocity structure (S. Kim et al., 2016). In the detailed response to reviewer file, Fig. R3 shows Moho depth maps with the sampled stations, and Figs. R4 and R5 show, respectively, the results for the P and S wave tomography inverted with the local crustal model compared with the inversion results based on the Crust1.0 model. We found that there were no significant differences in patterns and scales for the main features in the upper mantle.

We included additional content for the inversion results with a local crustal model in the third paragraph of section 2.Data and methods in the revised manuscript.

From the original text:

"Further tests were conducted for different configurations of the crustal structure using a 1-D initial model (S. Kim et al., 2011) (Figs. S1 and S2)."

To the modified text in the revised manuscript (lines 158-161):
"Further tests were conducted for different configurations of the crustal structure using a 1-D initial model (Kim et al., 2011) (Figs. S1 and S2) and a 3-D local crustal structure (S. Kim et al., 2016) with Moho interfaces (Chang and Baag, 2007; Kim et al., 2015)."

Response to Reviewer Comment No. 7:
We would like to thank the reviewer for pointing out this issue. We have modified the text to clarify this point.

From the original text in the original manuscript (lines 213–214):  
"Although the effect of velocity heterogeneities outside of the model space is negligible due to similar ray trajectories (Aki et al., 1977) ..."

To the modified text of the revised manuscript (lines 164–165):  
"Although the effect of velocity heterogeneities outside the model space is less significant due to similar ray trajectories (Aki et al., 1977) ...

Response to Reviewer Comment No. 8:
We would like to thank the reviewer for raising these questions. We calculated the synthetic residuals using the 3-D velocity structure only at depth ranges of 400–800 km from Tao et al., 2018, setting ak135 to between 0 and 400 km (Fig. S3). Thus, the calculated synthetic residuals only included the effects of velocity perturbations in the deeper depth ranges (400–800 km). We verified that we can reproduce the similar residual patterns when only using the velocity heterogeneities in the deeper depth range. We observed that the velocity heterogeneities at deeper depths (400–800 km) can be smeared upward, resulting in over or under estimating the velocity structures in the above model spaces (0–400 km) (Fig. S4). However, we further confirmed that these effects led to velocity changes of ~0.3% for dlnVp and ~0.5% for dlnVs, which is insignificant to interpretations of the upper mantle.

Response to Reviewer Comment No. 9:
We appreciate your comment. We have modified the text in the revised manuscript.

Figure 7 caption in the revised manuscript:
 "The residuals are shown in grayscale on the right based on their epicentral distances.”

Response to Reviewer Comment No. 10:
We would like to thank the reviewer for their comment. We have added a color-scale and modified the figure caption to clarify the image.

Figure 12 caption in the revised manuscript:
Fig. 12. A three-dimensional plot of the P wave tomography with interpretations of the main features. Structures in blue and red show +0.5 and -0.5% velocity isosurfaces, respectively, which are cut at the top by a horizontal velocity cross-section at a depth of 60 km with a color scale on the right. On the surface, red truncated cones indicate Cenozoic volcanism and gray and black lines delineate tectonic boundaries and coastlines, respectively. Vertical black dashed lines connect the
Cenozoic volcanism and their locations vertically projected onto the horizontal velocity cross-section. Surface relief is shown as a transparent layer.

Response to Reviewer Comment No. 11:

We would like to thank the reviewer for pointing out this issue. We have added the results of the crustal velocity structure (Figs. S7 and S8) to the revised manuscript.

In the revised manuscript (lines 221–226):
“Figs. S7 and S8 show the results of crustal velocity structure, where the velocity perturbations are superimposed on the initial crustal model. In the upper part of the crust (~10 km), relatively low velocities were mainly observed beneath the Cretaceous-Cenozoic volcano sedimentary deposits in the southern part of the SKP (e.g., GB and JI), which are consistent with a previous study (Kang and Shin, 2006). In the lower part of the crust (~20 km), we found relatively low velocities beneath the GM (Cho et al., 2006).”

Response to Reviewer Comment No. 12:

Thank you for raising this question. We removed the word “reactivate” and replaced the phrase “overlying continent” with “the surface of the continental lithosphere” to provide a clearer meaning.

In the revised manuscript (lines 324–325):
“The hotter and buoyant upper mantle, located beneath the thinner lithosphere, can disturb the surface of the continental lithosphere.”

Response to Reviewer Comment No. 13:

Thank you for this comment. As we modified “Highlight 4” in the first part of the reviewer comments, we removed the word “retaining” and modified the sentence to clarify the meaning.

In the revised manuscript (lines 333–334):
“The observed sharp contrast in mantle velocity and a spatial correspondence in recent surface magmatic activities potentially exhibit a destabilization process that characterizes the cratonic lithosphere margin due to intensive interactions between the convective upper mantle and lithospheric heterogeneity.”

References:


Research Data Related to this Submission

Title: Seismic data and velocity model of "Heterogeneous modification and reactivation of a craton margin beneath the Korean Peninsula from teleseismic travel time tomography"
Repository: figshare
https://doi.org/10.6084/m9.figshare.8980499.v2
Dear Editor:

We are submitting the revised manuscript titled “Heterogeneous modification and reactivation of a craton margin beneath the Korean Peninsula from teleseismic travel time tomography” by Jung-Hun Song, Seongryong Kim, and Junkee Rhie (Manuscript # GR-D-19-00421) for consideration for publication in *Gondwana Research*.

We carefully reviewed all comments made by the editor and reviewers, which significantly improved our original manuscript. Furthermore, following reviewer suggestions, we used a professional English editing service to thoroughly check and revise the English of the revised manuscript. We hope that the manuscript conforms to the standards of *Gondwana Research*.

Sincerely,

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Declaration of interests

☒ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

☐ The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:
Remnant cratonic root beneath the YM

Localized upwellings?

Thermally elevated upper mantle beneath the GM, GB, and along the eastern KP

GB: Gyeongsang Basin
GM: Gyeonggi Massif
KP: Korean Peninsula
YM: Yeongnam Massif
△: Cenozoic Volcanism
- Detailed lithospheric structure of the Korean Peninsula (KP) via seismic tomography
- Presence of a long-lasting cratonic root at the continental margin beneath the KP
- Non-uniform modification and reactivation of the craton margin via mantle dynamics
Heterogeneous modification and reactivation of a craton margin beneath the Korean Peninsula from teleseismic travel time tomography

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Abstract

Margins of old continental lithosphere are likely prone to ongoing modification processes. Therefore, constraining detailed structures beneath the margin can be essential in understanding the evolution of the continental lithosphere. The eastern margin of the Eurasian plate is a natural laboratory that allows us to study the strong effects from multiple episodes of continental collision and subduction of different oceanic plates since their formation. To reveal the detailed evolution of cratons at their margins, we describe, for the first time, the upper mantle structures beneath the southern Korean Peninsula (SKP) based strictly on teleseismic relative
arrival time data from densely deployed local seismic arrays, which allows us to constrain the
details of the lithospheric structures beneath the Archean-Proterozoic basement. We imaged a
thick (~150 km) high-velocity anomaly mainly beneath the Proterozoic Yeongnam massif with
large velocity contrasts ($dlnVp \approx 4.0\%$ and $dlnVs \approx 6.0\%$) at its boundaries, suggesting the
presence of a long-lasting cratonic root in the southwestern SKP. On the other hand, low-velocity
anomalies were found beneath the Proterozoic Gyeonggi Massif, Gyeongsang arc-back-arc
basin, and along the eastern margin of the SKP, indicating significantly modified regions. The
possible existence of a remnant cratonic root beneath the SKP and contrasting lithospheric
structures across the different Precambrian massifs suggests the highly heterogeneous
modification of cratonic lithosphere at the eastern Eurasian plate margin. Strong velocity
reductions, which indicate a thermally elevated upper mantle with potential partial melts,
correspond to areas of Cenozoic basalts, high surface heat flow, and high topography along the
eastern KP margin. We interpret this coincidence as a result of recent reactivation of a craton
margin, which is controlled by intense interaction between the convective upper mantle and
heterogeneous continental lithosphere.

**Keywords:** Korean Peninsula, Seismic Tomography, Cratonic Lithosphere, Continental Margin,
Heterogeneous Modification

### 1. Introduction

Modification of the cratonic lithosphere is a salient process by which we can
understand the evolution and destruction of old continents (Artemieva and Mooney, 2002; Foley,
Constraints on the properties of the continental lithosphere indicate that marginal tectonic processes play an important role in altering entire cratonic structures (Artemieva, 2006; Thomas, 2006; Boyce et al., 2019). Previous studies have suggested that interactions with slab subduction, collision, mantle dynamics, and effects due to deeper structures explain tectonic processes at margins (e.g., Levander et al., 2014; Moresi et al., 2014; Schmandt and Lin, 2014; Hu et al., 2018). Persistent multiple episodes of such tectonic events usually 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 Siberian Craton), which have experienced multiple tectonic events, 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 (Menzies et al., 1993; Gao et al., 2004) based on 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., Huang et al., 2009; Chen, 2010). Widespread Mesozoic to Cenozoic magmatism indicates entirely non-cratic geochemical and geophysical signatures (Choi et al., 2006; Kimura et al., 2018), accompanied by ubiquitous extensional basins (Fig. 1) (Ren et al., 2002; Liu et al., 2017; Chough et al., 2018). Due to the complexity, previous studies have suggested numerous different modes of lithosphere modification for the cratons in China (e.g., Griffin et al., 1998; Zhang, 2005; Guo et al., 2016; Wu et al., 2019). Specifically, the
Korean Peninsula (KP) is known to consist of Archean–Proterozoic continental lithosphere located in the forefront (Chough et al., 2000), where current dynamics in the upper mantle and effects due to nearby subducting slabs are the most significant. Despite this, current cratonic lithosphere margins have not been examined in detail due to reduced sampling in the region.

Previous studies have suggested the presence of tectonic affinities in the cratonic lithosphere modification process between China and the KP (Cho et al., 2017; Choi, 2019; Oh et al., 2019). The basement of the KP consists of late Archean to early Proterozoic massifs: from north to south, the Nangrim (NM), Gyeonggi (GM), and Yeongnam (YM) (Fig. 2) (Chough et al., 2000; Cho et al., 2008). As in China, each of the massifs in the KP were subject to different tectonic processes during amalgamations and breakups of supercontinents (Columbia, Rodinia, Gondwana, and Pangea) before the formation of the current setting since the late Paleozoic to early Triassic (Kim et al., 2014, 2018, 2019; Lee et al., 2019; Oh et al., 2019). Two fold-and-thrust belts, i.e., the Paleozoic Imjingang belt (IB) and 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, 2017). Subsequent oblique subduction and rollback of the oceanic plates affected the Precambrian lithosphere from eastern China to the YM, forming the Gyeongsang arc-back-arc basin (GB) at the southeastern margin of the YM and rifted structures in the East Sea to the east of the KP, respectively (Ren et al., 2002; Chough and Sohn, 2010). Spatially non-uniform and complex tectonic structures indicate that the cratonic lithosphere in the KP has been exposed to intensive, complicated modification processes at the continental margin during the Phanerozoic period (Chough et al., 2000).
To understand the detailed structure and tectonic evolution of a craton margin, we conducted high-resolution seismic tomography of the upper mantle beneath the southern KP (SKP). Detailed 3-D upper mantle $P$ and $S$ wave velocity variations were estimated for the first time using data from dense seismic arrays throughout the SKP with a recording time of more than five years. Unlike previous large-scale tomography studies in East Asia (e.g., Chen et al., 2017; Ma et al., 2018), we report velocity models based entirely on teleseismic relative arrival time data from the local seismic arrays to exploit seismic rays that share common trajectories outside the model volume (Aki et al., 1977; Thurber, 2003). Therefore, we can estimate local velocity heterogeneities in greater detail by measuring small travel time differences caused by an area of interest. Together with a set of resolution tests, we suggest that marginal tectonic 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 (Fig. 2b), 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 responses and filtered with frequency bands from 0.1–5.0 Hz for $P$ waves 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 and Kennett, 2004). To incorporate waveform data
from different instrumental settings and obtain more detailed results, we scrutinized the ranges of frequency bands for $P$ and $S$ waves by testing different pass-bands. With these frequency ranges, the highest degree of waveform coherency was achieved while maintaining similar levels of signal-to-noise ratios among different types of instruments. All teleseismic waveforms were visually inspected and noisy or incoherent data were eliminated to obtain more reliable measurements. Fig. 3 shows examples of $P$ and $S$ phases recorded by the local arrays. The waveforms of the processed data were similar in shape, which allowed us to measure the relative residuals with high precision. For the example data, the measured relative travel time residuals were generally positive at stations in the north and eastern margins of the SKP in contrast to negative values in the southwest. Although the time residuals depend on the direction of the incoming ray, the estimated amplitudes and polarities were consistent among the neighboring stations (Fig. 4). Our final dataset comprised 96,273 rays from 684 events for $P$ waves and 35,418 rays from 274 events for $S$ waves (Fig. 5). The average cross-correlation coefficient of our data set was 0.92 for $P$ waves and 0.83 for $S$ waves.

We determined the uncertainty in the travel time residuals based on waveform similarity (Rawlinson and Kennett, 2004). The estimated uncertainties were used as data weights in the tomographic inversion. The procedure for obtaining the absolute residual uncertainty for each relative travel time measurement consisted of two steps: (1) the estimation of an average root-mean-square (RMS) value of the uncertainty for each event dataset and (2) the calculation of absolute residual uncertainties for individual rays in the same event dataset based on relative uncertainty estimations with respect to the determined average RMS value. In the first step, we performed the recovery test suggested in the adaptive stacking method (Rawlinson and Kennett, 2004). We applied random time perturbations using a Gaussian distribution with a standard
deviation of 0.9 s for $P$ waves and 1.7 s for $S$ waves to the aligned traces, measuring RMS differences between the input time shifts and their recoveries. We repeated this test 50 times to obtain the average RMS uncertainty value for each event dataset. In the second step, the absolute uncertainties for each individual ray were obtained from the relative uncertainties, which were estimated by the adaptive stacking method based on the waveform coherence between the individual waveforms and stacked reference waveform (Rawlinson and Kennett, 2004). We adjusted the calculation limits of the absolute error level such that the RMS value of the relative residual uncertainties was equal to the predetermined average RMS uncertainty. We set the minimum individual residual uncertainty values to no less than 35 ms, which is equivalent to 70% of the sampling interval, to account for imperfect coherence in the records and data noise (Rawlinson et al., 2006b). The average uncertainties in the total residual travel time observations (Rawlinson et al., 2006b; Song et al., 2018) were 78 ms for $P$ waves and 149 ms for $S$ waves. We also tested different random time perturbations to measure the residual uncertainty but the comparison of the uncertainties among the different event datasets was consistent.

Fast marching ray-tracing (Rawlinson et al., 2006a) and the subspace inversion scheme (Kennett et al., 1988) were used iteratively to perform tomography. We defined a model space spanning a depth of ~800 km with a uniform grid spacing of ~10 km in the crust and ~20 km in the mantle. Regularization factors (damping ($\varepsilon$) and smoothing ($\eta$)) for the model were systematically determined by evaluating the trade-offs between the data misfit, model smoothness, and model variance (Rawlinson et al., 2006b) (Fig. 6). Data misfit for the inversion was determined based on the difference between the observed and predicted travel times starting from an initial model, which consisted of a 3-D crustal model over the mantle portion with the ak135 reference model (Kennett et al., 1995). To account for crustal effects, we simultaneously
inverted the model for crustal and upper mantle structures with fixed Moho depths. Instead of using local crustal models in the SKP, a global crustal model (Crust1.0; Laske et al., 2013) was adopted to cover regions beneath not only the SKP but also its margins and off-coasts. Further tests were conducted for different configurations of the crustal structure using a 1-D initial model (S. Kim et al., 2011) (Figs. S1 and S2) and a 3-D local crustal structure (S. Kim et al., 2016) with Moho interfaces (Chang and Baag, 2007; Kim et al., 2015). We found that patterns and amplitudes of upper mantle structures deeper than 60 km were consistent and robust, such that determining those depths was the primary focus of our interpretations.

Although the effect of velocity heterogeneities outside the model space is less significant due to similar ray trajectories (Aki et al., 1977), strong velocity perturbations at relatively less-constrained bottom depths can potentially bias amplitudes and patterns of the estimated anomalies above (e.g., 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 (Li and van der Hilst, 2010; Tao et al., 2018). We tested and minimized the impacts that the deeper anomalies had on our results. We first calculated synthetic residuals from a recent 3-D upper mantle velocity model for deeper depths (Tao et al., 2018) (Fig. S3) and compared them with our observations. As a result, a long-period pattern of similarity was observed as a function of the back-azimuth between the observed and synthetic residuals (Fig. 7), confirming that this pattern originates from deeper upper mantle structures. However, this effect caused only a small decrease in amplitude ($d\ln V_p < 0.3\%$ and $d\ln V_s < 0.5\%$) in our images for shallower areas ($< 300$ km) based on an additional inversion with a corrected dataset generated by subtracting the synthetic data from the
observations (Fig. S4). As a conservative choice, we determined the results by inversion with the corrected dataset as our final model.

### 3. Results

#### 3.1. Resolution tests

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 $V_p$ (Figs. 8a, S5a) and ±7% for $V_s$ (Figs. 8b, S5b). 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 verify the effects due to smearing (Rawlinson et al., 2006b). We used spikes with a diameter of 60 km and maximum amplitudes of ±5% for $V_p$ (Fig. 9a) and ±8% for $V_s$ (Fig. S6a) distributed in discrete positions with different polarities. 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 the interfaces (e.g., Youssof et al., 2015). The input models consisted of a positive anomaly ($dlnV_p$ of +1.5% and $dlnV_s$ of +2.0%) in the southwest and a negative anomaly ($dlnV_p$ of −2.5% and $dlnV_s$ of −3.5%) in the north and along the eastern areas of the SKP (Figs. 9b, S6b).
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 in the following section for the areas showing robust and reliable recovery in the resolution tests.

3.2. Tomographic model

Fig. 10 shows the results of the $P$ and $S$ wave tomographic models for the horizontal cross-sections while Fig. 11 shows the same results for the vertical cross-sections. Our models resulted in a reduction of data variance by 80.8% from 0.0841 to 0.0162 s$^2$ (from 290.0 to 127.1 ms in RMS) for $P$ wave tomography and 74.9% from 0.651 to 0.164 s$^2$ (from 806.9 to 404.4 ms in RMS) for $S$ wave tomography.

We observed an overall pattern of relatively high velocities in the west and southwest and relatively low velocities beneath the north and east in the upper mantle of the SKP (Fig. 10). A distinct high-velocity structure ($dlnVp$ of $\approx 1.5\%$ and $dlnVs$ of $\approx 3.0\%$) extends $\approx 150$ km laterally and $\approx 220$ km vertically mainly beneath the YM at the southwestern part of the KP (cross sections b–d in Fig. 11). Low-velocity anomalies are located beneath the GM, GB, and along the eastern margin (cross sections a–d in Fig. 11), showing a sharp contrast ($dlnVp$ $\approx 4.0\%$, $dlnVs$ $\approx 6.0\%$) with the observed high-velocity anomaly. Distinct areas of slower anomalies, with $dlnVp < -2\%$ and $dlnVs < -3\%$, appear at shallow depths (60 km in Fig. 10) within the generally low-velocity upper mantle in the north and eastern margin of the SKP.

Figs. S7 and S8 show the results of the crustal velocity structure, where the velocity perturbations were superimposed on the initial crustal model. In the upper part of the
crust (~10 km), relatively low velocities were found mainly beneath the Cretaceous-Cenozoic
volcano sedimentary deposits in the southern part of the SKP (e.g., GB and JI), which are
consistent with a previous study (Kang and Shin, 2006). In the lower part of the crust (~20 km),
we found relatively low velocities beneath the GM (Cho et al., 2006).

4. Discussion

The velocity patterns observed in our results agree with previous results from
tomography studies (e.g., Ma et al., 2018; Tao et al., 2018) that have shown a high-velocity
block located beneath the SKP within a relatively slow upper mantle. Specifically, previous
studies have suggested the existence of broad slower upper mantle velocities (< –1%) at shallow
depths (< 300 km) beneath northeast (NE) Asia (e.g., Chang et al., 2015; Legendre et al., 2015;
Debayle et al., 2016), which indicates the presence of a more thermally enhanced upper mantle
(Li and van der Hilst, 2010). This suggests that the mean velocities in our models are possibly
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 (Huang and Zhao, 2006; S. Kim et al.,
2016). Our results provide greater details with respect to the distribution of relatively high- and
low-velocity anomalies beneath the KP, consistent with previous observations.

Synthetic tests using structural models (Figs. 9b, S6b) showed that the vertical
extent (~220 km) and amplitude of the high-velocity anomaly in the southwest SKP is well
constrained. Together with an estimated low heat flow of < 50 mW/m² (Lee et al., 2010), the
Archean-Proterozoic surface geology (Fig. 2a), and the possible affinity of the southwestern SKP
to cratonic lithosphere in China, we note that the imaged high-velocity structure possibly reflects relatively thick continental lithosphere preserved in the cratonic margin beneath the KP. Recent tomographic studies found similar small-scale positive velocity anomalies beneath the central and eastern margin of the NCC, which were interpreted as fragments of thick lithosphere (e.g., Jiang et al., 2013; He and 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 (Zheng et al., 2001; Wang et al., 2003; Lee and Walker, 2006). 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 for the craton margin, which corresponds to episodic magma intrusions (Wang et al., 2018; Li et al., 2019) and extension (Ren et al., 2002) due to the subduction and rollback of multiple slabs since the Mesozoic.

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 (Figs. 11a, 11c, and 12). Differences in the physical and chemical properties, including thickness, thermal state, and composition, can affect the relative instability of the continental lithosphere (Foley, 2008; Zheng et al., 2015). Therefore, the loci of modification can be localized at an initial zone of lithospheric weakness due to collision, rifting, mobile belts, or lithospheric interlayers (Lu et al., 2011; Chen et al., 2014; Liu et al., 2018b). Recently, several studies have 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 (Kim et al., 2013; Cho et al., 2017; Oh et al., 2019), which are likely linked to the subsequent late Permian–Triassic collision between the NCC and SCC (Oh and Kusky, 2007; S.W. Kim et al., 2011). Re–Os isotopic data from the late Triassic kimberlite in the NM and O–Hf isotopic composition from the Triassic plutons in the GM indicate magmatic origins of metasomatized lithospheric mantle sources and the emplacement of asthenospheric mantle, supporting lithospheric modification as a coeval process with Mesozoic orogenies (Yang et al., 2010; Cheong et al., 2018). Inherent weaknesses beneath the GM during this period can be a possible mechanism to 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 recent and transient event (e.g., Levander et al., 2011; Bao et al., 2014) is more favorable due to less of a distinction in the geological expression across the tectonic regions (Fig. 2a) compared with the large lateral variation in our image (Fig. 10). 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., Rawlinson et al., 2017; Artemieva, 2018; Steinberger et al., 2019). In this case, it is possible that the weak properties have maintained deeper regions of the GM and been removed during a short period due to effects from a more recent tectonic event, such as the opening of the East Sea (e.g., Evanzia et al., 2014; Shen et al., 2018) or the initiation of Philippine Sea plate subduction during the Cenozoic.

A similar large variation defines the eastern margin of the thicker lithosphere characterized by a different level of lithospheric modification between the YM and GB (Fig. 2a). The continental arc and back-arc system at the GB developed during the Early Cretaceous to early Tertiary due to the northward oblique subduction of the paleo-Pacific plate (Chough and
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 thinning of the lithosphere via thermal erosion or hydrous weakening (Kusky et al., 2014; S.W. Kim et al., 2016). In our image, a significant coincidence in the tectonic boundary between the YM and GB (at a depth of 120 km in Fig. 10), as well as a rapid variation in the velocity, captured the resulting difference in lithospheric thickness, proving that subduction processes were the cause of lithospheric-scale modification in the southeastern SKP (Chough and Sohn, 2010).

Distinct areas of slower anomalies at shallow depths (60 km in Fig. 10) along the eastern margin of the SKP were recognized as sub-lithospheric upper mantle due to significant velocity contrasts ($dlnVp$ of ~4.0% and $dlnVs$ 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 (Fig. 9b), where the low-velocity anomalies resolved up to 35% of the input velocity perturbation compared with 65% for the high-velocity anomalies. Considering the generally slower 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 (Karato, 1993; Goes et al., 2000), with anelastic attenuation in the corresponding depth range (Adenis et al., 2017). Variations in the composition and grain size (velocity perturbation ~1%) (Cammarano et al., 2003; Faul and Jackson, 2005) did not result in such variations even when incorporating relatively weak upper mantle anisotropy (< 1%) in this region (Kang and Shin, 2009; Wei et al., 2016). The presence of a small fraction of partial melt (1–2%) can explain the large drop in velocity (Mavko, 1980; Hammond and Humphreys, 2000), which has also been suggested in previous tomography studies (Simutė et al., 2016; Adenis et al., 2017). The protracted subduction history of the current Pacific and former oceanic plates likely has
accommodated a hot and wet sub-lithospheric upper mantle in back-arc and adjacent continental 
 margins (S. Kim et al., 2016; Tauzin et al., 2017). Subsequent continental rifting, accompanied 
 by upper mantle dynamics, 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 (Won et al., 1994; Choi et al., 2006; Choi 
 et al., 2013) and relatively high heat flow rates (> 70 mW/m²) (Fig. 10). The topography of the 
 lithosphere-asthenosphere boundary generally accumulated the ascending mantle flow beneath 
 the thin lithosphere (e.g., Duggen et al., 2009; Steinberger et al., 2019). A steep gradient in 
 lithospheric thickness can induce and maintain localized mantle upwelling and decompressional 
 melting along the margins of the relatively thick lithosphere (King and Anderson, 1998; Conrad 
 et al., 2010). The hotter and buoyant upper mantle located beneath the thinner lithosphere can 
 disturb the surface of the continental lithosphere (e.g., Karlstrom et al., 2012; Yang and Gao, 
 2018). This mechanism corresponds to the positive residual topography associated with the 
 negative mantle gravity anomaly (Kaban et al., 2016) and high elevation with geomorphic 
 disequilibrium in the late Pleistocene (< 125 ka) (D.E. Kim et al., 2016) observed in the eastern 
 mountain ranges (Figs. 10, 12). 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; Wang and Kusky, 2019). The observed sharp 
 contrast in mantle velocity and a spatial correspondence in recent surface magmatic activities 
 potentially exhibit a destabilization process that characterizes the cratonic lithosphere margin due 
 to intensive interactions between the convective upper mantle and lithospheric heterogeneity.
Heterogeneous lithospheric structures have been observed in the NCC. An extensive but non-uniform modification process of craton lithosphere was suggested in the central and eastern NCC. Seismic tomography and anisotropy (e.g., Cheng et al., 2013; Jiang et al., 2013) and S wave receiver function analysis (e.g., Chen et al., 2009) revealed the heterogeneous structure of the craton lithosphere, indicating its complex nature due to deformation. An integrated thermomechanical model, jointly inverting heat flow data, surface wave dispersion curves, geoid height, and absolute elevation (Guo et al., 2016), demonstrated the highly heterogeneous physical state of the NCC developed by the uneven destruction of craton lithosphere due to sublithospheric mantle convection. Spatial distribution and temporal variation in geochemical characteristics of the Mesozoic ore deposits (e.g., gold metallogeny) (e.g., Li et al., 2014, 2017) were interpreted to have resulted from inhomogeneous decratonicization. Therefore, the heterogeneous destruction and reactivation of the craton lithosphere occurred extensively along the Sino-Korean craton margin, which includes the KP and NCC. The structural heterogeneity of the craton lithosphere was likely much sharper in scale and thickness within a confined area, as observed at the current continental margin of the KP. A distinct spatial correlation between active surface tectono-magmatic processes and modified regions indicates that upper mantle dynamics played an important role in deforming the craton lithosphere, which was facilitated by multiple subduction-collision events of oceanic/continental plates at the continental margin (e.g., Santosh, 2010; Cai et al., 2018).

5. Conclusions
High-resolution upper mantle seismic tomography constrained by teleseismic relative-time datasets revealed, for the first time in detail, continental lithospheric structures beneath the KP, which was inferred to have experienced heterogeneous modification and reactivation at the craton margin. Distinct lithospheric features of the different adjacent Precambrian massifs indicate different responses of each lithosphere to recent marginal tectonic processes. An anomalously thick high-velocity structure beneath the YM suggests the presence of a cold, resistant cratonic lithosphere fragment at the eastern margin of the Eurasian plate. In contrast, the absence of deeper lithosphere mostly occupied by high-temperature, buoyant upper mantle beneath the GM, continental arc and back-arc system of the GB in the southeast, and along the eastern margin indicate highly modified regions. A clear spatial coincidence between low velocities and recent tectono-magmatic activity suggests persistent reactivation of a cratonic margin by intensive interaction between the prominent lithospheric structures and convective upper mantle.

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**Fig. 1.** Regional tectonic map of northeast Asia. Major tectonic provinces are indicated 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). Cretaceous and Cenozoic extensional basins are shaded in gray (Ren et al., 2002). Volcanoes are indicated by red triangles. A pink dashed rectangular box shows the map boundary of Fig. 2. CNCC: central North China Craton; ENCC: eastern North China Craton; KP: Korean Peninsula.

**Fig. 2.** (a) Map showing the geology of the Korean Peninsula (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. (b) 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; NIED: National Research Institute for Earth Science and Disaster Resilience; GSN: Global Seismic Network.

**Fig. 3.** Examples of stacked P and S waveforms analyzed in this study. (a, b) Waveforms of the event that occurred *S* of Ndoi Island, Fiji (Mw = 6.9). (c, d) Waveforms of the event that occurred *NNE* of Kerman, Iran (Mw = 6.0). The time windows applied for waveform stacking are indicated with yellow shading. The locations of the events are indicated in Fig. 5.
Fig. 4. Relative arrival time residuals for $P$ and $S$ waves estimated using the waveforms shown in Fig. 3. The event back-azimuth (Baz.) and incidence angle of the incoming rays (Inc.) are shown by the white and black arrows on the bottom left of each panel, respectively. Stations without arrival time measurements are indicated with a cross.

Fig. 5. Distribution of teleseismic events used for $P$ (red dots on the left diagram) and $S$ (blue dots on the right diagram) wave tomography. Dashed black circles are plotted at 30° increments from the center of the SKP. Green stars and squares in each panel represent event locations of the example data from Figs. 3a–b and 3c–d, respectively.

Fig. 6. Determination of regularization factors (damping ($\varepsilon$) and smoothing ($\eta$)) for $P$ (top) and $S$ (bottom) wave tomography based on trade-off analyses (Rawlinson et al., 2006b). (Panels on the left) The optimum value of $\eta$ was determined based on the trade-off between model roughness and data variance measured by changing the $\eta$ while holding $\varepsilon$ at 1. (Panels in the middle) The optimum value of $\varepsilon$ was subsequently determined in a similar manner to the first step, changing $\varepsilon$ and holding $\eta$ at the value determined in the previous stage. (Panels on the right) The final values of $\eta$ were determined based on the relationship between model roughness and data variance, changing $\eta$ while holding $\varepsilon$ at the value determined in the previous step. Selected $\varepsilon$ and $\eta$ values in each step are indicated by red circles in each panel.

Fig. 7. Examples of the $P$ and $S$ wave residual plots for stations (a) SEO2 and (b) BUS2 as a function of the back-azimuth. The locations of each station are indicated in the inset maps of the panels, i.e., the observed $P$ wave residuals. The residuals are shown in grayscale on the right
based on their epicentral distances. Obs.: Observed residual; Syn.: Synthetic residual; Obs.-Syn.: Observed residuals subtracted by the synthetic residuals.

**Fig. 8.** Checkerboard resolution test results for the (a) $P$ and (b) $S$ wave tomography. Checkers with diameters of 60 and 120 km and amplitudes of ±4% for $Vp$ and ±7% for $Vs$ are shown (see Fig. S5 for anomalies with sizes of 45 and 90 km).

**Fig. 9.** Resolution tests of the $P$ wave tomography using spikes and structural anomalies. (a) Spike test using anomalies with a diameter of 60 km and amplitudes of ±5%. Input models and their recoveries are shown. The depths of each horizontal section are indicated in the upper right corner. Black dashed lines represent ±0.5% velocity contours. (b) Structural tests using high- and low-velocity blocks with amplitudes of +1.5% and −2.0%, respectively, with thicknesses of 90 and 150 km. Black dashed lines represent +0.5% velocity contours.

**Fig. 10.** Horizontal cross-sections through the $P$ ($Vp$; top) and $S$ ($Vs$; bottom) wave tomography at depths of 60, 120, and 180 km. The depths of each section are indicated in the upper right corner. The black solid lines at a depth of 120 km denote the tectonic divisions shown in Fig. 2. Regions with surface heat flow rates > 70 mW/m$^2$ are shown by gray dashed contours at a depth of 60 km of the $Vp$. Cenozoic volcanism is indicated by red triangles.

**Fig. 11.** Vertical cross-sections through the $P$ ($Vp$) and $S$ ($Vs$) wave tomography. The locations of each section (a–d) are indicated in Fig. 10. Cenozoic volcanism is indicated by red triangles.
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Black horizontal lines in the cross-sections indicate Moho interfaces. GB: Gyeongsang Basin; GM: Gyeonggi Massif; YM & OB: Yeongnam Massif and Okcheon Belt.

**Fig. 12.** A three-dimensional plot of the *P* wave tomography with interpretations of the main features. Structures in blue and red show +0.5 and −0.5% velocity isosurfaces, respectively, which are cut at the top by a horizontal velocity cross-section at a depth of 60 km with a color scale on the right. On the surface, red truncated cones indicate Cenozoic volcanism and gray and black lines delineate tectonic boundaries and coastlines, respectively. Vertical black dashed lines connect the Cenozoic volcanism and locations vertically projected onto the horizontal velocity cross-section. Surface relief is shown as a transparent layer.
Heterogeneous modification and reactivation of a craton margin beneath the Korean Peninsula from teleseismic travel time tomography

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Abstract

Margins of old continental lithosphere are likely prone to ongoing modification processes. Therefore, constraining detailed structures beneath the margin can be essential in understanding the evolution of the continental lithosphere. The eastern margin of the Eurasian plate is a natural laboratory that allows us to study the strong effects from multiple episodes of continental collision and subduction of different oceanic plates since their formation. To reveal the detailed evolution of cratons at their margins, we describe, for the first time, the upper mantle structures beneath the southern Korean Peninsula (SKP) based strictly on teleseismic relative
arrival time data from densely deployed local seismic arrays, which allows us to constrain the details of the lithospheric structures beneath the Archean-Proterozoic basement. We imaged a thick (~150 km) high-velocity anomaly mainly beneath the Proterozoic Yeongnam massif with large velocity contrasts \( d\ln V_p \approx 4.0\% \) and \( d\ln V_s \approx 6.0\% \) at its boundaries, suggesting the presence of a long-lasting cratonic root in the southwestern SKP. On the other hand, low-velocity anomalies were found beneath the Proterozoic Gyeonggi Massif, Gyeongsang arc-back-arc basin, and along the eastern margin of the SKP, indicating significantly modified regions. The possible existence of a remnant cratonic root beneath the SKP and contrasting lithospheric structures across the different Precambrian massifs suggests the highly heterogeneous modification of cratonic lithosphere at the eastern Eurasian plate margin. Strong velocity reductions, which indicate a thermally elevated upper mantle with potential partial melts, correspond to areas of Cenozoic basalts, high surface heat flow, and high topography along the eastern KP margin. We interpret this coincidence as a result of recent reactivation of a craton margin, which is controlled by intense interaction between the convective upper mantle and heterogeneous continental lithosphere.

**Keywords:** Korean Peninsula, Seismic Tomography, Cratonic Lithosphere, Continental Margin, Heterogeneous Modification

1. **Introduction**

Modification of the cratonic lithosphere is a salient process by which we can understand the evolution and destruction of old continents (Artemieva and Mooney, 2002; Foley,
Constraints on the properties of the continental lithosphere indicate that marginal tectonic processes play an important role in altering entire cratonic structures (Artemieva, 2006; Thomas, 2006; Boyce et al., 2019). Previous studies have suggested that interactions with slab subduction, collision, mantle dynamics, and effects due to deeper structures explain tectonic processes at margins (e.g., Levander et al., 2014; Moresi et al., 2014; Schmandt and Lin, 2014; Hu et al., 2018). Persistent multiple episodes of such tectonic events usually 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 Siberian Craton), which have experienced multiple tectonic events, 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 (Menzies et al., 1993; Gao et al., 2004) based on 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., Huang et al., 2009; Chen, 2010). Widespread Mesozoic to Cenozoic magmatism indicates entirely non-cratonic geochemical and geophysical signatures (Choi et al., 2006; Kimura et al., 2018), accompanied by ubiquitous extensional basins (Fig. 1) (Ren et al., 2002; Liu et al., 2017; Chough et al., 2018). Due to the complexity, previous studies have suggested numerous different modes of lithosphere modification for the cratons in China (e.g., Griffin et al., 1998; Zhang, 2005; Guo et al., 2016; Wu et al., 2019). Specifically, the
Korean Peninsula (KP) is known to consist of Archean–Proterozoic continental lithosphere located in the forefront (Chough et al., 2000), where current dynamics in the upper mantle and effects due to nearby subducting slabs are the most significant. Despite this, current cratonic lithosphere margins have not been examined in detail due to reduced sampling in the region.

Previous studies have suggested the presence of tectonic affinities in the cratonic lithosphere modification process between China and the KP (Cho et al., 2017; Choi, 2019; Oh et al., 2019). The basement of the KP consists of late Archean to early Proterozoic massifs: from north to south, the Nangrim (NM), Gyeonggi (GM), and Yeongnam (YM) (Fig. 2) (Chough et al., 2000; Cho et al., 2008). As in China, each of the massifs in the KP were subject to different tectonic processes during amalgamations and breakups of supercontinents (Columbia, Rodinia, Gondwana, and Pangea) before the formation of the current setting since the late Paleozoic to early Triassic (Kim et al., 2014, 2018, 2019; Lee et al., 2019; Oh et al., 2019). Two fold-and-thrust belts, i.e., the Paleozoic Imjingang belt (IB) and 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, 2017). Subsequent oblique subduction and rollback of the oceanic plates affected the Precambrian lithosphere from eastern China to the YM, forming the Gyeongsang arc-back-arc basin (GB) at the southeastern margin of the YM and rifted structures in the East Sea to the east of the KP, respectively (Ren et al., 2002; Chough and Sohn, 2010). Spatially non-uniform and complex tectonic structures indicate that the cratonic lithosphere in the KP has been exposed to intensive, complicated modification processes at the continental margin during the Phanerozoic period (Chough et al., 2000).
To understand the detailed structure and tectonic evolution of a craton margin, we conducted high-resolution seismic tomography of the upper mantle beneath the southern KP (SKP). Detailed 3-D upper mantle $P$ and $S$ wave velocity variations were estimated for the first time using data from dense seismic arrays throughout the SKP with a recording time of more than five years. Unlike previous large-scale tomography studies in East Asia (e.g., Chen et al., 2017; Ma et al., 2018), we report velocity models based entirely on teleseismic relative arrival time data from the local seismic arrays to exploit seismic rays that share common trajectories outside the model volume (Aki et al., 1977; Thurber, 2003). Therefore, we can estimate local velocity heterogeneities in greater detail by measuring small travel time differences caused by an area of interest. Together with a set of resolution tests, we suggest that marginal tectonic 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 (Fig. 2b), 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 responses and filtered with frequency bands from 0.1–5.0 Hz for $P$ waves 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 and Kennett, 2004). To incorporate waveform data
from different instrumental settings and obtain more detailed results, we scrutinized the ranges of
frequency bands for $P$ and $S$ waves by testing different pass-bands. With these frequency ranges,
the highest degree of waveform coherency was achieved while maintaining similar levels of
signal-to-noise ratios among different types of instruments. All teleseismic waveforms were
visually inspected and noisy or incoherent data were eliminated to obtain more reliable
measurements. Fig. 3 shows examples of $P$ and $S$ phases recorded by the local arrays. The
waveforms of the processed data were similar in shape, which allowed us to measure the relative
residuals with high precision. For the example data, the measured relative travel time residuals
were generally positive at stations in the north and eastern margins of the SKP in contrast to
negative values in the southwest. Although the time residuals depend on the direction of the
incoming ray, the estimated amplitudes and polarities were consistent among the neighboring
stations (Fig. 4). Our final dataset comprised 96,273 rays from 684 events for $P$ waves and
35,418 rays from 274 events for $S$ waves (Fig. 5). The average cross-correlation coefficient of
our data set was 0.92 for $P$ waves and 0.83 for $S$ waves.

We determined the uncertainty in the travel time residuals based on waveform
similarity (Rawlinson and Kennett, 2004). The estimated uncertainties were used as data weights
in the tomographic inversion. The procedure for obtaining the absolute residual uncertainty for
each relative travel time measurement consisted of two steps: (1) the estimation of an average
root-mean-square (RMS) value of the uncertainty for each event dataset and (2) the calculation
of absolute residual uncertainties for individual rays in the same event dataset based on relative
uncertainty estimations with respect to the determined average RMS value. In the first step, we
performed the recovery test suggested in the adaptive stacking method (Rawlinson and Kennett,
2004). We applied random time perturbations using a Gaussian distribution with a standard
deviation of 0.9 s for $P$ waves and 1.7 s for $S$ waves to the aligned traces, measuring RMS
differences between the input time shifts and their recoveries. We repeated this test 50 times to
obtain the average RMS uncertainty value for each event dataset. In the second step, the absolute
uncertainties for each individual ray were obtained from the relative uncertainties, which were
estimated by the adaptive stacking method based on the waveform coherence between the
individual waveforms and stacked reference waveform (Rawlinson and Kennett, 2004). We
adjusted the calculation limits of the absolute error level such that the RMS value of the relative
residual uncertainties was equal to the predetermined average RMS uncertainty. We set the
minimum individual residual uncertainty values to no less than 35 ms, which is equivalent to 70%
of the sampling interval, to account for imperfect coherence in the records and data noise
(Rawlinson et al., 2006b). The average uncertainties in the total residual travel time observations
(Rawlinson et al., 2006b; Song et al., 2018) were 78 ms for $P$ waves and 149 ms for $S$ waves. We
also tested different random time perturbations to measure the residual uncertainty but the
comparison of the uncertainties among the different event datasets was consistent.

Fast marching ray-tracing (Rawlinson et al., 2006a) and the subspace inversion
scheme (Kennett et al., 1988) were used iteratively to perform tomography. We defined a model
space spanning a depth of ~800 km with a uniform grid spacing of ~10 km in the crust and ~20
km in the mantle. Regularization factors (damping ($\epsilon$) and smoothing ($\eta$)) for the model were
systematically determined by evaluating the trade-offs between the data misfit, model
smoothness, and model variance (Rawlinson et al., 2006b) (Fig. 6). Data misfit for the inversion
was determined based on the difference between the observed and predicted travel times starting
from an initial model, which consisted of a 3-D crustal model over the mantle portion with the
ak135 reference model (Kennett et al., 1995). To account for crustal effects, we simultaneously
inverted the model for crustal and upper mantle structures with fixed Moho depths. Instead of using local crustal models in the SKP, a global crustal model (Crust1.0; Laske et al., 2013) was adopted to cover regions beneath not only the SKP but also its margins and off-coasts. Further tests were conducted for different configurations of the crustal structure using a 1-D initial model (S. Kim et al., 2011) (Figs. S1 and S2) and a 3-D local crustal structure (S. Kim et al., 2016) with Moho interfaces (Chang and Baag, 2007; Kim et al., 2015). We found that patterns and amplitudes of upper mantle structures deeper than 60 km were consistent and robust, such that determining those depths was the primary focus of our interpretations.

Although the effect of velocity heterogeneities outside the model space is less significant due to similar ray trajectories (Aki et al., 1977), strong velocity perturbations at relatively less-constrained bottom depths can potentially bias amplitudes and patterns of the estimated anomalies above (e.g., 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 (Li and van der Hilst, 2010; Tao et al., 2018). We tested and minimized the impacts that the deeper anomalies had on our results. We first calculated synthetic residuals from a recent 3-D upper mantle velocity model for deeper depths (Tao et al., 2018) (Fig. S3) and compared them with our observations. As a result, a long-period pattern of similarity was observed as a function of the back-azimuth between the observed and synthetic residuals (Fig. 7), confirming that this pattern originates from deeper upper mantle structures. However, this effect caused only a small decrease in amplitude ($dlnV_p < 0.3\%$ and $dlnV_s < 0.5\%$) in our images for shallower areas ($< 300$ km) based on an additional inversion with a corrected dataset generated by subtracting the synthetic data from the
observations (Fig. S4). As a conservative choice, we determined the results by inversion with the corrected dataset as our final model.

3. Results

3.1. Resolution tests

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 $V_p$ (Figs. 8a, S5a) and ±7% for $V_s$ (Figs. 8b, S5b). 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 verify the effects due to smearing (Rawlinson et al., 2006b). We used spikes with a diameter of 60 km and maximum amplitudes of ±5% for $V_p$ (Fig. 9a) and ±8% for $V_s$ (Fig. S6a) distributed in discrete positions with different polarities. 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 the interfaces (e.g., Youssof et al., 2015). The input models consisted of a positive anomaly ($dlnV_p$ of +1.5% and $dlnV_s$ of +2.0%) in the southwest and a negative anomaly ($dlnV_p$ of −2.5% and $dlnV_s$ of −3.5%) in the north and along the eastern areas of the SKP (Figs. 9b, S6b)
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 in the following section for the areas showing robust and reliable recovery in the
resolution tests.

3.2. Tomographic model

Fig. 10 shows the results of the $P$ and $S$ wave tomographic models for the
horizontal cross-sections while Fig. 11 shows the same results for the vertical cross-sections. Our
models resulted in a reduction of data variance by 80.8% from 0.0841 to 0.0162 s$^2$ (from 290.0
to 127.1 ms in RMS) for $P$ wave tomography and 74.9% from 0.651 to 0.164 s$^2$ (from 806.9 to
404.4 ms in RMS) for $S$ wave tomography.

We observed an overall pattern of relatively high velocities in the west and
southwest and relatively low velocities beneath the north and east in the upper mantle of the SKP
(Fig. 10). A distinct high-velocity structure ($d\ln V_p$ of $\approx 1.5\%$ and $d\ln V_s$ of $\approx 3.0\%$) extends $\approx 150$
km laterally and $\approx 220$ km vertically mainly beneath the YM at the southwestern part of the KP
(cross sections b–d in Fig. 11). Low-velocity anomalies are located beneath the GM, GB, and
along the eastern margin (cross sections a–d in Fig. 11), showing a sharp contrast ($d\ln V_p \approx 4.0\%,$
$d\ln V_s \approx 6.0\%$) with the observed high-velocity anomaly. Distinct areas of slower anomalies, with
$d\ln V_p < -2\%$ and $d\ln V_s < -3\%,$ appear at shallow depths (60 km in Fig. 10) within the generally
low-velocity upper mantle in the north and eastern margin of the SKP.

Figs. S7 and S8 show the results of the crustal velocity structure, where the
velocity perturbations were superimposed on the initial crustal model. In the upper part of the
crust (~10 km), relatively low velocities were found mainly beneath the Cretaceous-Cenozoic volcano sedimentary deposits in the southern part of the SKP (e.g., GB and JI), which are consistent with a previous study (Kang and Shin, 2006). In the lower part of the crust (~20 km), we found relatively low velocities beneath the GM (Cho et al., 2006).

4. Discussion

The velocity patterns observed in our results agree with previous results from tomography studies (e.g., Ma et al., 2018; Tao et al., 2018) that have shown a high-velocity block located beneath the SKP within a relatively slow upper mantle. Specifically, previous studies have suggested the existence of broad slower upper mantle velocities (< −1%) at shallow depths (< 300 km) beneath northeast (NE) Asia (e.g., Chang et al., 2015; Legendre et al., 2015; Debayle et al., 2016), which indicates the presence of a more thermally enhanced upper mantle (Li and van der Hilst, 2010). This suggests that the mean velocities in our models are possibly 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 (Huang and Zhao, 2006; S. Kim et al., 2016). Our results provide greater details with respect to the distribution of relatively high- and low-velocity anomalies beneath the KP, consistent with previous observations.

Synthetic tests using structural models (Figs. 9b, S6b) showed that the vertical extent (~220 km) and amplitude of the high-velocity anomaly in the southwest SKP is well constrained. Together with an estimated low heat flow of < 50 mW/m² (Lee et al., 2010), the Archean-Proterozoic surface geology (Fig. 2a), and the possible affinity of the southwestern SKP
to cratonic lithosphere in China, we note that the imaged high-velocity structure possibly reflects relatively thick continental lithosphere preserved in the cratonic margin beneath the KP. Recent tomographic studies found similar small-scale positive velocity anomalies beneath the central and eastern margin of the NCC, which were interpreted as fragments of thick lithosphere (e.g., Jiang et al., 2013; He and 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 (Zheng et al., 2001; Wang et al., 2003; Lee and Walker, 2006). 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 for the craton margin, which corresponds to episodic magma intrusions (Wang et al., 2018; Li et al., 2019) and extension (Ren et al., 2002) due to the subduction and rollback of multiple slabs since the Mesozoic.

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 (Figs. 11a, 11c, and 12). Differences in the physical and chemical properties, including thickness, thermal state, and composition, can affect the relative instability of the continental lithosphere (Foley, 2008; Zheng et al., 2015). Therefore, the loci of modification can be localized at an initial zone of lithospheric weakness due to collision, rifting, mobile belts, or lithospheric interlayers (Lu et al., 2011; Chen et al., 2014; Liu et al., 2018b). Recently, several studies have 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 (Kim et al., 2013; Cho et al., 2017; Oh et al., 2019), which are likely linked to the subsequent late Permian–Triassic collision between the NCC and SCC (Oh and Kusky, 2007; S.W. Kim et al., 2011). Re–Os isotopic data from the late Triassic kimberlite in the NM and O–Hf isotopic composition from the Triassic plutons in the GM indicate magmatic origins of metasomatized lithospheric mantle sources and the emplacement of asthenospheric mantle, supporting lithospheric modification as a coeval process with Mesozoic orogenies (Yang et al., 2010; Cheong et al., 2018). Inherent weaknesses beneath the GM during this period can be a possible mechanism to 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 recent and transient event (e.g., Levander et al., 2011; Bao et al., 2014) is more favorable due to less of a distinction in the geological expression across the tectonic regions (Fig. 2a) compared with the large lateral variation in our image (Fig. 10). 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., Rawlinson et al., 2017; Artemieva, 2018; Steinberger et al., 2019). In this case, it is possible that the weak properties have maintained deeper regions of the GM and been removed during a short period due to effects from a more recent tectonic event, such as the opening of the East Sea (e.g., Evanzia et al., 2014; Shen et al., 2018) or the initiation of Philippine Sea plate subduction during the Cenozoic.

A similar large variation defines the eastern margin of the thicker lithosphere characterized by a different level of lithospheric modification between the YM and GB (Fig. 2a). The continental arc and back-arc system at the GB developed during the Early Cretaceous to early Tertiary due to the northward oblique subduction of the paleo-Pacific plate (Chough and
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 thinning of the lithosphere via thermal erosion or hydrous weakening (Kusky et al., 2014; S.W. Kim et al., 2016). In our image, a significant coincidence in the tectonic boundary between the YM and GB (at a depth of 120 km in Fig. 10), as well as a rapid variation in the velocity, captured the resulting difference in lithospheric thickness, proving that subduction processes were the cause of lithospheric-scale modification in the southeastern SKP (Chough and Sohn, 2010).

Distinct areas of slower anomalies at shallow depths (60 km in Fig. 10) along the eastern margin of the SKP were recognized as sub-lithospheric upper mantle due to significant velocity contrasts ($d\ln V_p$ of ~4.0% and $d\ln V_s$ 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 (Fig. 9b), where the low-velocity anomalies resolved up to 35% of the input velocity perturbation compared with 65% for the high-velocity anomalies. Considering the generally slower upper mantle in NE Asia compared with the global average (~–1% in $V_p$ and ~–2% in $V_s$), the observed reductions in velocity correspond to temperature increments of 300–350 K (Karato, 1993; Goes et al., 2000), with anelastic attenuation in the corresponding depth range (Adenis et al., 2017). Variations in the composition and grain size (velocity perturbation ~–1%) (Cammarano et al., 2003; Faul and Jackson, 2005) did not result in such variations even when incorporating relatively weak upper mantle anisotropy (< 1%) in this region (Kang and Shin, 2009; Wei et al., 2016). The presence of a small fraction of partial melt (1–2%) can explain the large drop in velocity (Mavko, 1980; Hammond and Humphreys, 2000), which has also been suggested in previous tomography studies (Simutė et al., 2016; Adenis et al., 2017). The protracted subduction history of the current Pacific and former oceanic plates likely has
accommodated a hot and wet sub-lithospheric upper mantle in back-arc and adjacent continental margins (S. Kim et al., 2016; Tauzin et al., 2017). Subsequent continental rifting, accompanied by upper mantle dynamics, 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 (Won et al., 1994; Choi et al., 2006; Choi et al., 2013) and relatively high heat flow rates (> 70 mW/m²) (Fig. 10). The topography of the lithosphere-asthenosphere boundary generally accumulated the ascending mantle flow beneath the thin lithosphere (e.g., Duggen et al., 2009; Steinberger et al., 2019). A steep gradient in lithospheric thickness can induce and maintain localized mantle upwelling and decompressional melting along the margins of the relatively thick lithosphere (King and Anderson, 1998; Conrad et al., 2010). The hotter and buoyant upper mantle located beneath the thinner lithosphere can disturb the surface of the continental lithosphere (e.g., Karlstrom et al., 2012; Yang and Gao, 2018). This mechanism corresponds to the positive residual topography associated with the negative mantle gravity anomaly (Kaban et al., 2016) and high elevation with geomorphic disequilibrium in the late Pleistocene (< 125 ka) (D.E. Kim et al., 2016) observed in the eastern mountain ranges (Figs. 10, 12). 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; Wang and Kusky, 2019). The observed sharp contrast in mantle velocity and a spatial correspondence in recent surface magmatic activities potentially exhibit a destabilization process that characterizes the cratonic lithosphere margin due to intensive interactions between the convective upper mantle and lithospheric heterogeneity.
Heterogeneous lithospheric structures have been observed in the NCC. An extensive but non-uniform modification process of craton lithosphere was suggested in the central and eastern NCC. Seismic tomography and anisotropy (e.g., Cheng et al., 2013; Jiang et al., 2013) and S wave receiver function analysis (e.g., Chen et al., 2009) revealed the heterogeneous structure of the craton lithosphere, indicating its complex nature due to deformation. An integrated thermomechanical model, jointly inverting heat flow data, surface wave dispersion curves, geoid height, and absolute elevation (Guo et al., 2016), demonstrated the highly heterogeneous physical state of the NCC developed by the uneven destruction of craton lithosphere due to sublithospheric mantle convection. Spatial distribution and temporal variation in geochemical characteristics of the Mesozoic ore deposits (e.g., gold metallogeny) (e.g., Li et al., 2014, 2017) were interpreted to have resulted from inhomogeneous decratonization. Therefore, the heterogeneous destruction and reactivation of the craton lithosphere occurred extensively along the Sino-Korean craton margin, which includes the KP and NCC. The structural heterogeneity of the craton lithosphere was likely much sharper in scale and thickness within a confined area, as observed at the current continental margin of the KP. A distinct spatial correlation between active surface tectono-magmatic processes and modified regions indicates that upper mantle dynamics played an important role in deforming the craton lithosphere, which was facilitated by multiple subduction-collision events of oceanic/continental plates at the continental margin (e.g., Santosh, 2010; Cai et al., 2018).

5. Conclusions
High-resolution upper mantle seismic tomography constrained by teleseismic relative-time datasets revealed, for the first time in detail, continental lithospheric structures beneath the KP, which was inferred to have experienced heterogeneous modification and reactivation at the craton margin. Distinct lithospheric features of the different adjacent Precambrian massifs indicate different responses of each lithosphere to recent marginal tectonic processes. An anomalously thick high-velocity structure beneath the YM suggests the presence of a cold, resistant cratonic lithosphere fragment at the eastern margin of the Eurasian plate. In contrast, the absence of deeper lithosphere mostly occupied by high-temperature, buoyant upper mantle beneath the GM, continental arc and back-arc system of the GB in the southeast, and along the eastern margin indicate highly modified regions. A clear spatial coincidence between low velocities and recent tectono-magmatic activity suggests persistent reactivation of a cratonic margin by intensive interaction between the prominent lithospheric structures and convective upper mantle.

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**Fig. 1.** Regional tectonic map of northeast Asia. Major tectonic provinces are indicated 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). Cretaceous and Cenozoic extensional basins are shaded in gray (Ren et al., 2002). Volcanoes are indicated by red triangles. A pink dashed rectangular box shows the map boundary of Fig. 2. CNCC: central North China Craton; ENCC: eastern North China Craton; KP: Korean Peninsula.

**Fig. 2.** (a) Map showing the geology of the Korean Peninsula (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. (b) 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; NIED: National Research Institute for Earth Science and Disaster Resilience; GSN: Global Seismic Network.

**Fig. 3.** Examples of stacked P and S waveforms analyzed in this study. (a, b) Waveforms of the event that occurred S of Ndoi Island, Fiji (Mw = 6.9). (c, d) Waveforms of the event that occurred NNE of Kerman, Iran (Mw = 6.0). The time windows applied for waveform stacking are indicated with yellow shading. The locations of the events are indicated in Fig. 5.
Fig. 4. Relative arrival time residuals for $P$ and $S$ waves estimated using the waveforms shown in Fig. 3. The event back-azimuth (Baz.) and incidence angle of the incoming rays (Inc.) are shown by the white and black arrows on the bottom left of each panel, respectively. Stations without arrival time measurements are indicated with a cross.

Fig. 5. Distribution of teleseismic events used for $P$ (red dots on the left diagram) and $S$ (blue dots on the right diagram) wave tomography. Dashed black circles are plotted at 30° increments from the center of the SKP. Green stars and squares in each panel represent event locations of the example data from Figs. 3a–b and 3c–d, respectively.

Fig. 6. Determination of regularization factors (damping ($\varepsilon$) and smoothing ($\eta$)) for $P$ (top) and $S$ (bottom) wave tomography based on trade-off analyses (Rawlinson et al., 2006b). (Panels on the left) The optimum value of $\eta$ was determined based on the trade-off between model roughness and data variance measured by changing the $\eta$ while holding $\varepsilon$ at 1. (Panels in the middle) The optimum value of $\varepsilon$ was subsequently determined in a similar manner to the first step, changing $\varepsilon$ and holding $\eta$ at the value determined in the previous stage. (Panels on the right) The final values of $\eta$ were determined based on the relationship between model roughness and data variance, changing $\eta$ while holding $\varepsilon$ at the value determined in the previous step. Selected $\varepsilon$ and $\eta$ values in each step are indicated by red circles in each panel.

Fig. 7. Examples of the $P$ and $S$ wave residual plots for stations (a) SEO2 and (b) BUS2 as a function of the back-azimuth. The locations of each station are indicated in the inset maps of the panels, i.e., the observed $P$ wave residuals. The residuals are shown in grayscale on the right
based on their epicentral distances. Obs.: Observed residual; Syn.: Synthetic residual; Obs.-Syn.: Observed residuals subtracted by the synthetic residuals.

Fig. 8. Checkerboard resolution test results for the (a) $P$ and (b) $S$ wave tomography. Checkers with diameters of 60 and 120 km and amplitudes of ±4% for $V_p$ and ±7% for $V_s$ are shown (see Fig. S5 for anomalies with sizes of 45 and 90 km).

Fig. 9. Resolution tests of the $P$ wave tomography using spikes and structural anomalies. (a) Spike test using anomalies with a diameter of 60 km and amplitudes of ±5%. Input models and their recoveries are shown. The depths of each horizontal section are indicated in the upper right corner. Black dashed lines represent ±0.5% velocity contours. (b) Structural tests using high- and low-velocity blocks with amplitudes of +1.5% and –2.0%, respectively, with thicknesses of 90 and 150 km. Black dashed lines represent +0.5% velocity contours.

Fig. 10. Horizontal cross-sections through the $P$ ($V_p$; top) and $S$ ($V_s$; bottom) wave tomography at depths of 60, 120, and 180 km. The depths of each section are indicated in the upper right corner. The black solid lines at a depth of 120 km denote the tectonic divisions shown in Fig. 2. Regions with surface heat flow rates > 70 mW/m$^2$ are shown by gray dashed contours at a depth of 60 km of the $V_p$. Cenozoic volcanism is indicated by red triangles.

Fig. 11. Vertical cross-sections through the $P$ ($V_p$) and $S$ ($V_s$) wave tomography. The locations of each section (a–d) are indicated in Fig. 10. Cenozoic volcanism is indicated by red triangles.
Black horizontal lines in the cross-sections indicate Moho interfaces. GB: Gyeongsang Basin; GM: Gyeonggi Massif; YM & OB: Yeongnam Massif and Okcheon Belt.

**Fig. 12.** A three-dimensional plot of the $P$ wave tomography with interpretations of the main features. Structures in blue and red show $+0.5$ and $-0.5\%$ velocity isosurfaces, respectively, which are cut at the top by a horizontal velocity cross-section at a depth of 60 km with a color scale on the right. On the surface, red truncated cones indicate Cenozoic volcanism and gray and black lines delineate tectonic boundaries and coastlines, respectively. Vertical black dashed lines connect the Cenozoic volcanism and locations vertically projected onto the horizontal velocity cross-section. Surface relief is shown as a transparent layer.
Click here to download e-component: Supplementary.docx
Dear Editor:

We appreciate your consideration of our manuscript. We carefully reviewed all comments made by the editor and reviewers (written in black) and present our responses (written in blue) in this letter. The corresponding modifications/additions are also marked in the revised manuscript (written in blue). We believe that our corrections have improved the original manuscript. Furthermore, following reviewer suggestions, we used a professional English editing service to thoroughly check and revise the English of the revised manuscript. We hope that the manuscript conforms to the standards of Gondwana Research.

Before we present our responses to the comments from Reviewer #1 comments, we would first like to address the editor’s suggestion.

**Editor’s suggestion:** I would also suggest that you add expanded discussion on the nature of craton destruction by consulting papers on the North China Craton where several geological and geophysical studies have shown the nature of extensive craton destruction, particularly in the eastern domain. You can also see a paper that integrates both geological and geophysical techniques to evaluate the decratonization in the North China Craton [Santosh, 2010, Precambrian Research, on Columbia]. There are also several recent papers that link the differential destruction of the craton with gold metallogey [Shengrong Li and others, papers in GR, OGR etc.]. Thus, you can expand your model into a more regional picture for the Sino-Korean craton as a whole, which will enhance the appeal of your excellent work.

We sincerely thank the editor for this constructive suggestion with respect to our original manuscript. As suggested, we expanded our discussion on the nature of craton destruction along the Korean Peninsula associated with the evolution of the North China Craton (NCC) and Sino-Korean craton. We have added related discussions with relevant references, including those that the editor has suggested.
We have added the following sentences to the revised manuscript (lines 336-353):

“Heterogeneous lithospheric structures have been observed in the NCC. An extensive but non-uniform modification process of craton lithosphere was suggested in the central and eastern NCC. Seismic tomography and anisotropy (e.g., Cheng et al., 2013; Jiang et al., 2013) and S wave receiver function analysis (e.g., Chen et al., 2009) revealed the heterogeneous structure of the craton lithosphere, indicating its complex nature due to deformation. An integrated thermomechanical model, jointly inverting heat flow data, surface wave dispersion curves, geoid height, and absolute elevation (Guo et al., 2016), demonstrated the highly heterogeneous physical state of the NCC developed by the uneven destruction of craton lithosphere due to sublithospheric mantle convection. Spatial distribution and temporal variation in geochemical characteristics of the Mesozoic ore deposits (e.g., gold metallogeny) (e.g., Li et al., 2014, 2017) were interpreted to have resulted from inhomogeneous dekratonization. Therefore, the heterogeneous destruction and reactivation of the craton lithosphere occurred extensively along the Sino-Korean craton margin, which includes the KP and NCC. The structural heterogeneity of the craton lithosphere was likely much sharper in scale and thickness within a confined area, as observed at the current continental margin of the KP. A distinct spatial correlation between active surface tectono-magmatic processes and modified regions indicates that upper mantle dynamics played an important role in deforming the craton lithosphere, which was facilitated by multiple subduction-collision events of oceanic/continental plates at the continental margin (e.g., Santosh, 2010; Cai et al., 2018).”

References (newly added in the revised manuscript):


Comments by Reviewer #1:

In this paper, the authors present and interpret new body wave teleseismic tomography results from the Korean Peninsula. Approximately 5 years of data from a large and dense array of seismometers is used to generate P-wave and S-wave images of the crust and upper mantle. One of the main findings is evidence of modification of the cratonic margin beneath the Korean Peninsula, with a particularly clear association of Cenozoic volcanism and low velocities in the upper mantle. Overall, I thought this was a very nice paper, with a particularly thorough and well done application of teleseismic tomography, which included a useful range of synthetic reconstruction tests. As such, I recommend publication in Gondwana Research subject to minor-moderate revisions. Note that I reviewed a previous version of this paper that was submitted to GRL; this review is an updated version of that review. It appears that the core of the paper is the same, but a few new figures and extratext has been added. It is notable that the additional text is not of the same standard as the original in terms of quality of prose/grammar.

We would like to thank Reviewer #1 for constructive comments and suggestions made for our original manuscript. We carefully went through all comments and present our responses below.

(0) Highlight 4: What is meant by "Retained reactivation"?

We thank Reviewer #1 for pointing out the unclear meaning of this phrase. We intended to suggest that the upper mantle dynamics sustained a continuous reactivation of the craton margin. We removed the word "retained" in the revised manuscript to clarify the meaning. We also combined Highlights 3 and 4 in the original manuscript to conform to journal formatting.

Highlights 3 and 4 in the original manuscript:

Highlight 3. A sharp transition in the degree of modification among Archean-Proterozoic massifs

Highlight 4. Retained reactivation of the craton margin by upper mantle dynamics
Highlight 3 in the revised manuscript:

Highlight 3: Non-uniform modification and reactivation of the craton margin via mantle dynamics

(1) Lines 13-15: This sentence does not make sense, particularly the second half of it. Overall, the written English could do with some improvement; I will point out a few instances below, but not all of them.

We appreciate the reviewer's comment. We obtained the services of a professional English language editing company to thoroughly check and revise the written English in the revised manuscript. We modified this sentence to clearly indicate the original meaning.

From the original text (lines 13–15):

“Constraining detailed structure at the margin of old continental lithosphere is important for understanding the evolution of continental lithosphere due to the most active modification processes of it.”

To the revised text (lines 14–16):

“Margins of old continental lithosphere are likely prone to currently ongoing modification processes. Therefore, constraining detailed structures beneath the margin can be essential to understand the evolution of the continental lithosphere.”

(2) Line 19: Should be "seismic arrays", and later in the sentence "finding out fine details" is poor expression.

We thank the reviewer for their comment. We have modified this sentence as suggested.

From the lines 18–21 in the original manuscript:

“... teleseismic relative arrival time data from densely deployed local seismic array for the first time, finding out fine details of lithospheric structure beneath the Archean-Proterozoic basement.”
To the lines 20-22 in the revised manuscript:

“... teleseismic relative arrival time data from densely deployed local seismic arrays, which allows us to constrain the details of lithospheric structures beneath the Archean-Proterozoic basement.”

(3) Line 83: I wouldn't say that they "suffered different tectonic processes" - plate tectonics is not a disease. Perhaps "were subject to" instead of "suffered"

We appreciate this suggestion on the expression. We have corrected this by replacing "suffered" to "were subject to" as suggested in the revised manuscript (line 74).

(4) Lines 123-124: What are the responses of the short period instruments and accelerometers used? Do they seem noisier than the broadband stations at 10 s period? I am slightly surprised that the long-period cut-off for the P-wave and S-wave band pass filtering are the same - normally one would allow longer period S-waves than P-waves, but perhaps this is related to the use of heterogeneous instrumentation. It would be good to know the proportions of the three types of instruments used and their relative SNRs for the period ranges that are targeted.

We used three types of instruments: broadband (Guralp CMG-3T or Kinematics STS-1 or 2), short period (Kinematics SS-1 or Guralp CMG-40T-1), and acceleration (Kinematics ES-T or ES-DH) sensors. We present examples of the instrumental responses for the corresponding sensors in Fig. R1.
To determine an appropriate frequency range, we tested different low cut-off frequencies (e.g., 0.03, 0.05, and 0.1 Hz). With a minimum frequency of 0.1 Hz, we were able to obtain more stable measurements that were less affected by noise. For all used data, we compared the relative mean SNRs between the different instruments for various low cut-off frequencies (i.e., 0.03, 0.05, and 0.1 Hz). The smallest variation in the SNR between the different types of sensors was estimated when using the 0.1 Hz corner frequency (Fig. R2).
**Fig. R2.** The relative signal-to-noise ratio between different instruments with different long-period cut-off frequency limits that were applied to the teleseismic S wave data (0.03, 0.05, and 0.1 Hz).

We included a remark on SNR variations in the method section of the revised manuscript.

In the revised manuscript (lines 112–114):

"With these frequency ranges, the highest degree of waveform coherency was achieved while maintaining similar levels of signal-to-noise ratios among different types of instruments."

(5) Line 136: This sentence needs to be fixed.

We would like to thank the reviewer for their comment. We have modified the text as suggested.

From the original text in the original manuscript (lines 135–137):

"It is noted that the processed waveforms are highly coherent each other enough to measure relative delay times accurately."

To the modified text in the revised manuscript (lines 116–117):

"The waveforms of the processed data are similar in shape, which allowed us to measure the relative residuals with high precision."

(6) Line 195: Are there local 3-D crustal models available e.g. from receiver functions or ambient noise tomography? I suspect that Crust 1.0 is unlikely to be very useful given the density of stations and hence the short-scale-length of structures that can be resolved.

We would like to thank the reviewer for raising this question. We inverted our dataset with a local model of Moho depths (Chang and Baag, 2007; Kim et al., 2015) and a 3-D velocity structure (S. Kim et al., 2016). Fig. R3 shows Moho depth maps with the sampled stations. Figs. R4 and R5 show,
respectively, the results for the $P$ and $S$ wave tomography inverted with the local crustal model compared with the inversion results based on the Crust1.0 model. We found that there were no significant differences in patterns and scales for the main features in the upper mantle.

Fig. R3. Moho depths of the southern Korean Peninsula constrained by the receiver function analysis. Black solid curves indicate the Moho depth contours in km below sea level at 1-km intervals. White triangles and squares are the sampled sites of Chang and Baag (2007) and Kim et al. (2015), respectively.
Fig. R4. A comparison of the inversion results with different initial crustal models (a local crustal model and Crust1.0) for $P$ wave tomography.
Fig. R5. A comparison of the inversion results with different initial crustal models (a local crustal model and Crust1.0) for S wave tomography.

We included additional content for the inversion results with a local crustal model in the third paragraph of section 2 (Data and methods) in the revised manuscript.

From the original text:
"Further tests were conducted for different configurations of the crustal structure using a 1-D initial model (S. Kim et al., 2011) (Figs. S1 and S2)."

To the modified text in the revised manuscript (lines 158–161):
"Further tests were conducted for different configurations of the crustal structure using a 1-D initial model (Kim et al., 2011) (Figs. S1 and S2) and a 3-D local crustal structure (S. Kim et al., 2016) with Moho interfaces (Chang and Baag, 2007; Kim et al., 2015)."

(7) Line 214: I'm not sure that it can be claimed that heterogeneity outside the prescribed model region is negligible, but that is what is assumed in teleseismic tomography - rightly or wrongly.

We would like to thank the reviewer for pointing out this issue. We have modified the text to clarify this point.

From the original text in the original manuscript (lines 213–214):
“Although the effect of velocity heterogeneities outside of the model space is negligible due to similar ray trajectories (Aki et al., 1977) ...”

To the modified text of the revised manuscript (lines 164–165):
“Although the effect of velocity heterogeneities outside the model space is less significant due to similar ray trajectories (Aki et al., 1977) ...”

(8) Lines 219-228: I'm not sure that I feel entirely comfortable with what has been done here. If you subtract model predictions from the observations, and say that these are due to structure between 400-800 km depth that are poorly resolved, then what do you get in the 400-800 km depth range when you invert the adjusted dataset? Also, presumably the synthetic residuals will be influenced by structure in the 0-400 km range (or is this set to ak135 when you do the predictions?), so are you not just removing long-wavelength structural effects throughout the whole local model region?
We would like to thank the reviewer for raising these questions. We calculated the synthetic residuals using the 3-D velocity structure only at depth ranges of 400–800 km from Tao et al., 2018, setting ak135 to between 0 and 400 km (Fig. S3). Thus, the calculated synthetic residuals only included the effects of velocity perturbations in the deeper depth ranges (400–800 km). We verified that we can reproduce the similar patterns when only using the velocity heterogeneities in the deeper depth range. We observed that the velocity heterogeneities at deeper depths (400–800 km) can be smeared upward, resulting in over or under estimating the velocity structures in the above model spaces (0–400 km) (Fig. S4). However, we further confirmed that these effects led to velocity changes of ~0.3% for $dlnVp$ and ~0.5% for $dlnVs$, which is insignificant to interpretations of the upper mantle.

(9) Figure 7 caption: The figure is greyscale, so it is not accurate to say that residuals are colour-coded.

We appreciate your comment. We have modified the text in the revised manuscript.

Figure 7 caption in the revised manuscript:

“The residuals are shown in greyscale on the right based on their epicentral distances.”

(10) Figure 12: This figure is ok - a bit busy perhaps and it is not entirely clear to me what the yellow/orange and light blue colours represent.

We would like to thank the reviewer for their comment. We have added a color-scale and modified the figure caption to clarify the image.
Fig. 12. A three-dimensional plot of the $P$ wave tomography with interpretations of the main features. Structures in blue and red show +0.5 and −0.5% velocity isosurfaces, respectively, which are cut at the top by a horizontal velocity cross-section at a depth of 60 km with a color scale on the right. On the surface, red truncated cones indicate Cenozoic volcanism and gray and black lines delineate tectonic boundaries and coastlines, respectively. Vertical black dashed lines connect the Cenozoic volcanism and their locations vertically projected onto the horizontal velocity cross-section. Surface relief is shown as a transparent layer.

(11) Section 4: Not much is said about crustal structure, even though it is explicitly inverted for. I realise that the velocity perturbations likely define "bulk" properties in the sense that they probably represent depth-averaged perturbations through the full crustal thickness, but nonetheless they may be useful to understand.

We would like to thank the reviewer for pointing out this issue. We have added the results of the crustal velocity structure (Figs. S7 and S8) to the revised manuscript.

In the revised manuscript (lines 221-226):
“Figs. S7 and S8 show the results of crustal velocity structure, where the velocity perturbations are superimposed on the initial crustal model. In the upper part of the crust (~10 km), relatively low velocities were mainly observed beneath the Cretaceous-Cenozoic volcano sedimentary deposits in the southern part of the SKP
Fig. S7. Crustal $P$ wave velocity structure of (left column) the initial model (Crust1.0) and inverted model at depths of 10 (top) and 20 km (bottom). Major tectonic divisions are indicated by black solid lines.
Fig. S8. Crustal $S$ wave velocity structure of (left column) the initial model (Crust1.0) and inverted model at depths of 10 (top) and 20 km (bottom). Major tectonic divisions are indicated by black solid lines.

(12) Line 404: What is implied by the statement "reactivate the overlying continent"?

Thank you for raising this question. We removed the word “reactivate” and replaced the phrase “overlying continent” with “the surface of the continental lithosphere” to provide a clearer meaning.

In the revised manuscript (lines 324–325):
“The hotter and buoyant upper mantle, located beneath the thinner lithosphere, can disturb the surface of the continental lithosphere.”

(13) Line 413: Not sure what is meant by a "retaining destabilisation process".

Thank you for this comment. As we modified “Highlight 4” in the first part of the reviewer comments, we removed the word “retaining” and modified the sentence to clarify the meaning.

In the revised manuscript (lines 333–334):

“The observed sharp contrast in mantle velocity and a spatial correspondence in recent surface magmatic activities potentially exhibit a destabilization process that characterizes the cratonic lithosphere margin due to intensive interactions between the convective upper mantle and lithospheric heterogeneity.”

References:

Figure 7

(a) SEO2

P wave residuals

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S wave residuals

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Back Azimuth (Deg)

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Figure 9

Click here to download high resolution image
Remnant cratonic root beneath the YM

Localized upwellings?

Thermally elevated upper mantle beneath the GM, GB, and along the eastern KP