Intra-plate basalts of ~35–0 Ma in East Eurasia formed in a broad backarc region above the stagnant Pacific Plate slab in the mantle transition zone. These basalts show regional-scale variations in Nd-Hf isotopes. The basalts with the most radiogenic Nd-Hf center on the Shandong Peninsula with intermediate Nd-Hf at Hainan and Datong. The least radiogenic basalts occur in the perimeters underlain by the thick continental lithosphere. Shandong basalts possess isotopic signatures of the young igneous oceanic crust of the subducted Pacific Plate. Hainan and Datong basalts have isotopic signatures of recycled subduction materials with billions of years of storage in the mantle. The perimeter basalts have isotopic signatures similar to pyroxenite xenoliths from the subcontinental lithospheric mantle beneath East Eurasia. Hainan basalts exhibit the highest mantle potential temperature ($T_p$), while the Shandong basalts have the lowest $T_p$. We infer that a deep high-$T_p$ plume interacted with the subducted Pacific Plate slab in the mantle transition zone to form a local low-$T_p$ plume by entraining colder igneous oceanic lithosphere. We infer that the subducted Izanagi Plate slab, once a part of the Pacific Plate mosaic, broke off from the Pacific Plate slab at ~35 Ma to sink into the lower mantle. The sinking Izanagi slab triggered the plume that interacted with the stagnant Pacific slab and caused subcontinental lithospheric melting. This coincided with formation of the western Pacific backarc marginal basins due to Pacific Plate slab rollback and stagnation.

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The age of the stagnant slab is, therefore, 60–50 Ma at its leading edge and gradually increases in age to 125–160 Ma eastward in the current western Pacific (Fig. 1c right panel). The slab west of the subducted ridge was the Izanagi Plate. It is inferred that the Izanagi Plate slab sank to the bottom of the lower mantle and most of it now lies at the core–mantle boundary (Seton et al., 2015) (Fig. 1b).

East Eurasia was affected by extension in the Cenozoic, it was stretched to form the Sea of Japan and the South China Sea within the continental crust. These basins were initiated at ~35 Ma as continental rift zones. Subsequent seafloor spreading began about 35–25 Ma and ceased by ~15 Ma (Hall, 2002; Kimura et al., 2005; Sdolias and Müller, 2006; Tamaki et al., 1992; Taylor and Hayes, 1980; Yoshida et al., 2013) (Fig. 1) (see also Supplementary Material Appendix A1). Studies in the 1990s suggested that active asthenospheric upwelling beneath northeast China contributed to magmatism and extension (Tatsumi et al., 1990). The collision of India with Eurasia at ~50 Ma may also have contributed to marginal basin extension by trench rollback (Jolivet et al., 1994) or lateral injection of the asthenosphere beneath the region (Flower et al., 1992).

The roles of the subducted Pacific plate have been investigated recently using mantle tomography. The results have been interpreted...
to indicate that Pacific Plate slab stagnation occurred no earlier than 10–20 Ma (X. Liu et al., 2017). This model explains the upper plate stretching as a result of slab rollback and subsequent stagnation of the Pacific plate. However, the proposed stagnation age is younger than expected from the ~35 Ma extension history in this region.

East Eurasia extension could have been triggered by subduction of the Izanagi–Pacific Ridge at 60–50 Ma. The subsequent breakoff of the Izanagi Plate from the mantle transition zone at ~35 Ma may have caused the rollback of the Pacific Plate at 35–15 Ma to form the stagnant slab in the mantle transition zone and the overlying marginal basins (Seton et al., 2015) (Fig. 1a). In this context, the stagnant slab beneath the South China Sea is also the Pacific Plate rather than the subducted Philippine Sea Plate, the latter is too short and insufficient to form the entire stagnant slab beneath the South China Sea (Hall, 2002) (Fig. 1a and b cross Section 3).

2.2. Thickness of continental lithosphere

The depth of the lithosphere-asthenosphere boundary is estimated by the S-wave velocity model of An and Shi (2006). The thick lithosphere (>100 km) beneath the Tibetan Plateau in the Tethyan tectonic domain, including the Yangtze Block and beneath North Korea, extends to Sikhote-Alin (Fig. 2a). The depth of the lithosphere asthenosphere boundary is intermediate in the Paleo Asian tectonic domain to the north (Fig. 2a). The thinnest lithosphere (<80 km) occurs beneath Eastern Eurasia margins (An and Shi, 2006) (Fig. 2a) (see Supplementary Material Appendix A2).

2.3. Structure and age of East Eurasian continental lithosphere

East Eurasia is underlain by lithosphere with various ages as old as the Archean. Fig. 3a shows a simplified map of tectonic terranes in China (Zheng et al., 2013). The crust of East Eurasia can be simplified as four E–W trending strips of different age and origin. This crust is composed of the Central Asian Orogenic Belt in the north, the North China Craton of Archean age between 30° N–45° N, the Dabie-Sulu-Qintang, and the Proterozoic South China Block in the south (Fig. 3a).

We compiled pertinent information about the mantle and crustal xenoliths for Nd isotopes from the Geochemistry of Rocks of the Oceans and Continents (GEOROC) database (GEOROC, 2015) (see calculation method for Nd model \( \tau_{DM} \) ages in Fig. 3 caption; comprehensive reference in Supplementary Material Appendix B). The Nd model \( \tau_{DM} \) ages for peridotite are consistent with their Os TRD (single stage depletion model) ages of 2.5–0 Ga (Liu et al., 2011) (Fig. 3b) and no obvious patterns in lithospheric heterogeneity can be identified. In contrast, most granulite and 8 out of 17 pyroxenite \( \tau_{DM} \) ages are older than 2 Ga and are from the North China Craton (Fig. 3c and d). This suggests that the Archean lower crust and some Archean lithospheric mantle are preserved beneath the oldest exposed Precambrian terrane (Fig. 3a and c).

![Fig. 2. Depths of the seismic lithosphere and petrological asthenosphere. (a) Thickness of the seismic lithosphere according to the S-wave velocity model showing surface geology tectonic terranes. Figure adopted and modified from An and Shi (2006). (b) Lithospheric thickness and surface topography showing that thin lithosphere and low topography both occur in East Eurasia. Small dots show distributions of Late Cenozoic intra-plate basalt centers. Thick dashed lines show ~100 km depth contour lines. (c) Schematic cross section (thick yellow line in (a)) comparing lithospheric thickness and the average depth of magma segregation (\( P_{loc} \); top depth of asthenosphere) beneath the intra-plate basalt centers estimated by OBS1. Good correlation is shown for each center. (d) Correlation between the top depth of the asthenosphere (\( P_{loc} \)) and the mantle temperature at depth (\( T_{loc} \)). A common trend is shown with the exception of transitional Changbaishan and Hainan-QT with high \( T_{p} \) and thus \( T_{loc} \).]
2.4. Deep plumes

Receiver function analysis of the mantle transition zone beneath Hainan shows up- and down-warping structures in the 410 km and 660 km discontinuities, respectively (Fig. 4a). This has been interpreted as a hot mantle plume anomaly $\Delta T = 380 \, ^\circ C$ higher at 660 km and $\Delta T = 200 \, ^\circ C$ at 410 km (Huang et al., 2015). The expansion of the Hainan plume head is supported by the temporal-spatial spread of basalt centers since ~30 to 0 Ma in and around the South China Sea, including Hainan, southeast China, and eastern Vietnam (Ho et al., 2003). The Hainan plume is suggested to have upwelled through the window between the Pacific and Izu-Bonin slabs as shown by low-Vp body in the lower mantle (Figs. 1a and 4b).

A similar deep plume was also proposed for beneath the Baikal Rift based on a $V_p$ structure (Zhao et al., 2006). Deep mantle tomography beneath eastern China revealed a continuous low-$V_p$ body from 100 km to 400 km depth beneath Datong (Huang and Zhao, 2006). The low-$V_p$ body continues to 700 km depth and converges with the Baikal plume, and then a low-$V_p$ region vertically continues to $\sim 1300 \, \text{km}$ depth (Huang and Zhao, 2006). The deeper mantle tomography image also shows a low-$V_p$ perturbation beneath Datong–Baikal (Fig. 4c).

3. Basalt magmas

3.1. Late Cenozoic basalt chemistry

Regional basalt activity since 35 Ma is the focus for this Late Cenozoic magmatic province. Most of these Late Cenozoic alkali basalts occur over the stretched area although some appear on the edges of thick lithosphere (Fig. 2). They collectively form a large Late Cenozoic igneous province (Barry and Kent, 2013; Johnson et al., 2003; Tatsumi et al., 1990). Among these, large alkali basalt provinces are in Hainan–eastern Vietnam–southeastern China erupted ~15 Ma, after the opening of the South China Sea (Ho et al., 2003; Hoang and Flower, 1998), in the Baikal Rift–Mongolia erupted during 35–6.6 Ma (Barry et al., 2003; Johnson et al., 2003) and in the Chifeng erupted during 25–6 Ma (Wang et al., 2015). (Fig. 1a).
Magmatism in this age range also formed MORB-like tholeiitic basalt ocean floors of the Sea of Japan at 20–15 Ma and transitional basalt in the South China Sea at ~30–16 Ma (Expedition 349 Scientists, 2014; Fedorov and Koloskov, 2005; Ho et al., 2003; Yoshida et al., 2013). Post-opening stage marginal basin basalts were mostly alkali ocean island basalt-like and erupted since ~15 Ma (Cousens and Allan, 1992; Expedition 349 Scientists, 2014; Hirahara et al., 2015; Huang et al., 2013; Kimura et al., 2005; Wang et al., 2013; Zhang et al., 2017) similar to the alkali basalts from inland East Eurasia (Choi et al., 2006; Kimura et al., 2005; Wang et al., 2013; Yan et al., 2014).

All the Late Cenozoic (35–0 Ma) basalts show lower Sr and higher Nd isotopic compositions (e.g., \( \varepsilon_{\text{Nd}} = 3–13 \)) differing from isotopically enriched Cretaceous to Paleogene basalts with arc geochemical signatures (Choi et al., 2008; Kuritani et al., 2011, 2013; Liu et al., 2008; Sakuyama et al., 2013, 2014; Sun et al., 2017; Wang et al., 2011, 2015; Xu et al., 2017). The mechanism of slab flux transport through the >300 km-thick upper mantle is still debated. The mechanism could be diapirs of hydrous buoyant mantle or small-scale hydrous plume from the stagnant slab and associated mantle transition zone (Kuritani et al., 2011, 2013; Niu et al., 2015; Richard and Iwamori, 2010; Sakuyama et al., 2013; Sun et al., 2017; Thomson et al., 2016). Formation of a big mantle wedge (BMW) by this hydrous mantle was also explored (Zhao et al., 2009).

3.2. Role of asthenosphere and lithosphere

Studies of Late Cenozoic alkali basalts from southeast China to Korea before 2011 focused on the roles of the asthenosphere and lithosphere, including the timing of the extension and how this removed enriched lithospheric mantle from the continental keel (Tang et al., 2006; Xu et al., 2005; Zeng et al., 2010; Zhang et al., 2003; Zou et al., 2000). The role of subcontinental lithospheric mantle continues to be discussed, along with melting of the subcontinental lithospheric mantle pyroxene (Huang et al., 2013; J.-Q. Liu et al., 2017). The role of the subcontinental lithospheric mantle and depleted asthenosphere mantle were discussed for the enriched-type (E) and depleted-type (D) MORB-like backarc basin basalts in the Sea of Japan (Okamura et al., 2005; Pouclet et al., 1995; Shuto et al., 2004). Conversely, sediment subducted with the Pacific Plate may also be an alternative source of the E-type basalt (Cousens et al., 1994; Hirahara et al., 2015; Kuritani et al., 2011).

3.3. Role of stagnant slab

Studies after 2011 increasingly focused on the role of the stagnant Pacific Plate slab in the mantle transition zone at 440–660 km depth, which could have supplied elements to the basal source mantle via water, silicate melt, carbonated melt, or supercritical fluid from subducted igneous oceanic crust or sediment (Chen et al., 2017; J.-Q. Liu et al., 2016; Kuritani et al., 2011, 2013; S.-C. Liu et al., 2016; Sakuyama et al., 2013, 2014; Sun et al., 2017; Wang et al., 2011, 2015; Xu et al., 2017). The mechanism of slab flux transport through the >300 km-thick upper mantle is still debated. The mechanism could be diapirs of hydrous buoyant mantle or small-scale hydrous plume from the stagnant slab and associated mantle transition zone (Kuritani et al., 2011, 2013; Niu et al., 2015; Richard and Iwamori, 2010; Sakuyama et al., 2013; Sun et al., 2017; Thomson et al., 2016). Formation of a big mantle wedge (BMW) by this hydrous mantle was also explored (Zhao et al., 2009).

3.4. Role of plume

Geochemical arguments for a deep plume origin are reported for Hainan basalts and basalts from the South China Sea, where the stagnant slab is absent (Tu et al., 1991; Wang et al., 2011, 2013; Zhang et al., 2017) (Fig. 1a). A plume origin for the Baikal Rift basalts, including the Vitim Volcanic Field and basalts in Mongolia, have also been discussed (Barry et al., 2003; Johnson et al., 2003). These arguments use geochemistry to identify materials recycled from the deep mantle (Wang et al., 2013) or argue for high mantle potential temperatures, both expected for deep mantle plumes (Johnson et al., 2003; Wang et al., 2011). These may contradict to buoyant wet upwelling of
hydrated asthenosphere above the stagnant slab (S.-C. Liu et al., 2016; Wang et al., 2015).

3.5. Water in the mantle

Water content in the mantle source is a crucial test of the hydrous plume model. The estimated water contents in East Eurasian primitive basalts are at least 0.2–2.1 wt%, based on water contents in clinopyroxenes (Chen et al., 2017). Higher water contents, as much as 2–4 wt% in primary basalts, were suggested for basanitic to alkalic basalts (S.-C. Liu et al., 2016). However, they are mostly from low degree partial melts of the source mantle at 1 to 3% partial melting, which require only 300–600 ppm water in the source. The observations preclude extremely high mantle water contents shown for some backarc basin basalts containing up to 5000 ppm water (Dixon et al., 2004). Rather, the water contents are comparable to that in the hotspots with 300–800 ppm water (Dixon et al., 2004; Kimura et al., 2017), more akin to deep plume source mantle although water from the stagnant Pacific Plate slab is not negated (Ichiki et al., 2006) (see Supplementary Material Appendix A4).

The subcontinental mantle lithosphere beneath East Eurasia appears to be dry. The absence of hydrous minerals and the lack of traces of melting (Chen et al., 2001; Xu et al., 1996; Yu et al., 2003) are reported from mantle xenoliths. The low water, <300 ppm, in the xenoliths was also reported from water in clinopyroxenes (Hao et al., 2012). This range is comparable with that in mantle xenoliths beneath Hawaii that range from 50 to 100 ppm for peridotite and 250–470 ppm water for pyroxenite (Bizimis and Peslier, 2015).

3.6. Chemical geodynamic studies

The wide distribution of East Eurasian basalts and abundant isotopic data enabled examination of the geochemical structure of the upper mantle. An early effort included the use of Sr-Nd-Pb isotope compositions of Late Cenozoic basalts in East Eurasia and the Western Pacific by mapping the enriched mantle 1 (EM1) component (Flower et al., 2001), which is peculiar to the Indian Ocean asthenosphere (Mahoney et al., 1998; Miyazaki et al., 2015). The spatial distribution of the EM1 component is regarded as a remnant “high-tide mark” of lateral asthenospheric mantle flow of the Indian Ocean DMM since Neotethys closed (Flower et al., 2001) (Fig. 1c).

The many possible origins of the EM1 component preclude a unique solution for the chemical geodynamics of East Eurasian basalts. Researchers disagree about the origin of the EM1 component, which could be from either the Indian Ocean mantle asthenosphere (Dupré and Allègre, 1983; Mahoney et al., 1998) or the subcontinental lithosphere beneath the Gondwana supercontinent (Cousens and Allan, 1992; Hoernle et al., 2011; Kimura et al., 2016; Okamura et al., 2005; Tatsumoto and Nakamura, 1991). Alternatively, EM1 could also have originated from metasomatized mantle in the transition zone that was affected by an ancient (>1 Ga) slab sediment component (Kuritani et al., 2011, 2013; Sakuyama et al., 2013, 2014; Wang et al., 2017). This concept differs from that considering deep-recycled slab sediment modified at the subduction zone and held for more than 2 Gy near the core-mantle boundary for the origin of EM1 for a deep-plume source (Collerson et al., 2010).

The enriched mantle 2 (EM2)-like component is also identified in some basalts, such as those from Changbaishan (Kuritani et al., 2011), Hainan Island (Tu et al., 1991), Jeju Island (Choi et al., 2006) and the Sea of Japan E-type tholeiitic basalt (Cousens et al., 1994; Hirahara et al., 2015). They are thought to be influenced by young (<0.1 Ga) recycled sediment from the deep to stagnant slab beneath the backarc (Cousens et al., 1994; Hirahara et al., 2015; Kuritani et al., 2011), contamination of subcontinental lithospheric mantle (Tu et al., 1991), or large-scale mantle domain containing EM2 component (Choi et al., 2006).

Another important isotopic group is the Focal Zone (FOZO) isotopic component for basalts from Hainan to the South China Sea (Wang et al., 2013), the Baikal Rift (Barry and Kent, 2013; Johnson et al., 2003), and Chifeng flood basalts (Guo et al., 2016; Yu et al., 2015). The source of FOZO is likely to be prolonged (1–2 Gyr) storage of recycled oceanic plate slab consisting of a constant mixture between sediment, igneous oceanic crust, and metasomatized mantle (Kimura et al., 2016; Stracke, 2012). This may come from mantle as deep as the core-mantle boundary.

4. Chemical geodynamics

The sources of Late Cenozoic East Eurasian basalts may include the following geochemical components: (1) enriched mantle from the shallow subcontinental lithosphere, (2) depleted MORB source mantle, (3) stagnant slab in the mantle transition zone, or (4) plume from below the mantle transition zone. We examine the sources of East Eurasian Late Cenozoic basalts to take a new look at the geodynamics of East Eurasia.

4.1. Our approach

Late Cenozoic basalts erupted in East Eurasia between 20° N and 50° N between the Sea of Japan and the South China Sea. This area is underlain entirely by the stagnant Pacific Plate slab in the mantle transition zone except for under the southern Hainan–South China Sea and western Datong–Baikal, where a slab tear and the leading edge of the slab are located, respectively (Fig. 1a). The locations of the 22 examined basalts including South China Sea basalts are given in Fig. 5. The Late Cenozoic basalts of the same age range from Baikal, Chifeng Shikote-Alin, the Sea of Japan are compared later.

We examine Nd, Hf, and Pb isotope data to identify the magma sources. The Nd-Hf isotope systematics are robust against crustal contamination (see below) relative to Sr-Pb because of their high abundances in the basalts compared to the crustal rocks (Guo et al., 2016; Kimura et al., 2016; Sakuyama et al., 2013). Nevertheless, Pb isotope composition is also examined in order to better characterize the mantle sources.

We use the trace element abundances in the basalts to estimate the source mantle conditions and compositions, including mantle potential temperature (Tm); minimum pressure of basal segregation (Pm); temperature of the melt at Pm (Tm); degree of melting (F); water content in the mantle (XH2O); and fractions of peridotite and pyroxenite (Fp) in the mantle source. We use an inversion mass balance calculation model known as Ocean Basalt Simulator version 1 (OBS1) (Kimura and Kawabata, 2015), a tool for chemometric analysis of the basalt source conditions and compositions. These provide insights into the chemical geodynamics when combined with the isotopic compositions.

4.2. Geochemical dataset

We compiled data from the GEOROC database (GEOROC, 2015), including late Eocene to Quaternary basalt chemistry with Nd-Hf-Pb isotope data (see full references in Supplementary Material Appendix B). Our own analysis of Hf isotopes on a few Changbaishan samples are also used (Supplementary Material Table S1). The available samples spatially and geochemically (by isotope) cover almost all of the basalt provinces in East Eurasia, even if not all of the eruption centers are examined (Fig. 5a).

4.3. Nd-Hf-Pb isotope geochemistry

4.3.1. Regional Nd-Hf isotopic variation

The isotopic variations of Late Cenozoic East Eurasia basalts are extremely wide as shown by εNd–T厂区 plots (Choi et al., 2008; Sakuyama et al., 2013; Xu et al., 2017). The isotopic variations cover the global
isotopic trend defined by the ocean island basalts (OIBs) (Blichert-Toft and Albarède, 1997) extending from DMM through FOZO to the EM1 mantle endmember components (Stracke, 2012). In the εNd–εHf isotopic space, all the basalt data align along a quasi-linear trend with slight kinks showing at least four components (Fig. 5b). The components are assigned to DMM (Jiangsu), FOZO (Hainan), and EM1 (Wudalianchi or Yunnan), with an additional local HB (Hebei) component representing a common convergence point for Changbaishan, Wudalianchi, and Yunnan isotopic trends (Fig. 5b). This DMM–FOZO–HB–EM1 isotopic composition spatially varies approximately from NE to SW, from Shandong–Jiangsu through Hainan–Anhui to Yunnan (Fig. 5a and b). Similarly, basalts gradually become less radiogenic to the NE with increasing distance from the Shandong Peninsula, from Jiangsu–Shandong through Jeju–Jengok to Changbaishan–Ulleung–Wudalianchi (Fig. 5a and b). These collectively form a large quasi-concentric structure centered on the Shandong Peninsula (Fig. 5a).

Note that Hf isotope data are unavailable for Hainan basalts. Therefore, data from nearby Anhui and a South China Sea seamount are used to represent the Hainan plume mantle source (Ho et al., 2003) (see Fig. 5a and b) because Sr–Nd–Pb isotope compositions of the Hainan basalts are similar to these basalts.

Fig. 5. Late Cenozoic intra-plate basalt centers and their geochemical data. (a) Locations of Late Cenozoic intra-plate basalt centers. White dashed lines show the concentric structure of basalt chemistry centered on the Jiangsu (blue symbols)–Shandong group through the common trend group (red and orange symbols) to the perimeter group (black and purple symbols). (b) εNd–εHf isotope compositions of the intra-plate basalts corresponding to the concentric structure as in (a). (c) εHf isotope composition plotted versus MgO. Thin regression lines show chemical variations in the basalt provinces representing fractional crystallization and assimilation or source mixing. See fractional crystallization trend for reference. (d) εHf isotope composition plotted versus SiO2. Considerable SiO2 covariation is seen between the isotopically different basalt suites showing an arrangement similar to the εNd–εHf plots in (b). Circles show typical SiO2 range of melts from carbonated peridotite + mafic lithology for the origin of Shandong primary magma (Sakuyama et al., 2013), peridotite melt (Kogiso et al., 1998), and pyroxenite melt (Pertermann and Hirschmann, 2003) composition. The dataset was from literature. See Supplementary Material Appendix B for the references. For samples lacking radiometric ages, ε-values were calculated assuming \( t = 0 \) Ma. Most samples are from eruption centers > 15 Ma; therefore, the age effect is negligible for our purposes. Blue and red thin dashed lines indicate MORB and OIB fields, respectively. EM1: Enriched Mantle 1; EM2: Enriched Mantle 2; and FOZO: Focal Zone isotopic mantle components (Stracke, 2012).
4.3.2. Crustal assimilation

Previous studies found negligible crustal assimilation on the basis of Nd and Hf isotope compositions for East Eurasian intra-plate alkali basalts (Guo et al., 2016). However, the dataset shows a wide compositional variation from alkali basalt to basaltic andesite (SiO₂ = 39–57 wt%), so the possible effects of crustal assimilation need to be explored further. Fig. 5c plots εHf-MgO using MgO to monitor fractional crystallization, and trends in each basalt province are shown by thin regression lines. The regression lines are calculated based on the dataset of each suite. Decrease in MgO indicates fractional crystallization whereas decrease in εHf suggests crustal assimilation.

The trends are mostly flat or show a slight increase in εHf with decreasing MgO, confirming nil to slight modification of isotopic compositions as a result of crustal interactions. The extent of crustal assimilation is especially small in basalts with >7 wt% MgO. The MgO-rich basalts from key isotopic components of Jiangsu (DMM), Hainan (FOZO), Hebei (HB), and Wudalianchi (EM1) are unaffected by crustal assimilation (Fig. 5c). Changbaishan basalts with <7 wt% MgO show a slight increase in εHf; however, they are minor and form a flat trend together with the higher-Mg Guandong and Ulelung basalts, and they also have clear EM1 tendencies (Fig. 5c). Only Yunnan basalts show steep negative trends in εHf-MgO that can be interpreted to be a result of crustal assimilation (Fig. 5c).

4.3.3. Major element–εHf variations due to different source lithology

The major element compositions covary somewhat with εHf. SiO₂ content varies between 39 wt% in basanite to 58 wt% in trachyte and trachyandesite, irrespective of MgO contents (Figs. 5d and 6a). The DMM-type Shandong basalt with the highest MgO (~10 wt%) has the least SiO₂ (38 wt%). In contrast, the EM1-type Changbaishan basalt with the highest MgO content (~9.5 wt%) has a high SiO₂ content (~50 wt%), suggesting different source lithology. (Fig. 5d). The FOZO-type Hainan and HB-type Hainan basalts and other basalts between them plot in the middle of the DMM- and EM1-type basalts, with SiO₂ = 45–50 wt% for the unfractionated basalts with MgO >9 wt% (Figs. 5c, d, and 6). Fractional crystallization paths of representative MgO-rich basalts from Shandong, Hainan, and Changbaishan were calculated using MELTS (Ghiorso et al., 2002) at 0.5 GPa, 0.5 wt% H₂O, and proximity of the DMM-type basalts to the Izanagi–Paciﬁc ridge (~60 Ma) (Miyazaki et al., 2015). The isotopic composition of the DMM-type basalts shares similar characteristics with Shandong basalts, indicating a similar source lithology (Figs. 5c, d, 6a, and b).

The trends in εHf-MgO that can be interpreted to be a result of crustal assimilation (Fig. 5c). The Shandong basalts are extremely Fe-rich and silica-deﬁcient (e.g., FeO ~12 wt%, MgO of 8–12 wt%, and SiO₂ = 38–44 wt%) (Fig. 6b). Based on experimental results, this feature has been ascribed to the role of CO₂ during the melting of the pyroxenite-bearing peridotite mantle source (Sakuyama et al., 2013). The DMM-type Jiangsu basalts also show similar characteristics with Shandong basalts, indicating a similar source lithology (Figs. 5c, d, 6a, and b).

In contrast, partial melting of pyroxenite without CO₂ can form an SiO₂-rich basaltic andesite melt (FeO = 8–11 wt%, MgO = 2–6 wt%, and SiO₂ = 50–56 wt%) (Kogiso et al., 1998; Pertermann and Hirschmann, 2003). Such compositions overlap low-MgO basalts from Changbaishan and Wudalianchi, suggesting involvement of pyroxenite in the source. High MgO basalts from these areas are poor in SiO₂, indicating mixing between pyroxenite and peridotite melts (Figs. 5c, d, 6a, and b). They all share the same EM1-type εHf suggesting that the pyroxenite melt plays the dominant role (Figs. 5 and 6).

The FOZO- and HB-type basalts, including Hainan, have intermediate SiO₂ and FeO at a given MgO. They are intermediate in all aspects in terms of isotopes and major elements (Figs. 5 and 6). Silica content is comparable to that of melts from peridotite with or without pyroxenite (Kogiso et al., 1998; Mallik and Dasgupta, 2012).

4.3.4. Mantle sources

East Eurasia basalts were derived from different sources with different isotopic compositions and lithologies (Figs. 5 and 6). This source model is important when combined with regional geochemical variations (Fig. 5a).

The DMM-type basalts appear to have been inﬂuenced by the geochemical ﬂux from the stagnant Paciﬁc Plate (Sakuyama et al., 2013; Zou et al., 2000). A good geochemical analogue of the stagnant slab material is MORB-like metabasalts in the Shimanto greenrocks of the SW Japan trench accretionary prism, scraped off from the subducting Izu–Paciﬁc Ridge at ~60 Ma (Miyazaki et al., 2015). The isotopic proximity of the DMM-type basalts to the Izu–Paciﬁc Ridge MORBs (Fig. 7a) strongly supports the previous proposal for the basalt sources in and around the Shandong Peninsula (Sakuyama et al., 2013; Zou et al., 2000). The source mantle lithology should contain pyroxenite from the Izu–Paciﬁc Ridge MORBs. The considerable amount of CO₂ needed to produce silica-deﬁcient, Fe-rich basalts could have originated from subducted carbonated MORBs (Staudigel et al., 1996) or calcareous sediments (Guo et al., 2016; Sakuyama et al., 2013).

Fig. 6. Major element compositions of East Eurasia intra-plate basalts. Basalt dataset and groups are as in Fig. 5. Compositional ﬁelds of experimental basalt data are from (Sakuyama et al., 2013) for silica-deﬁcient Fe-rich basalt, peridotite melt (Kogiso et al., 1998) for peridotite source with or without pyroxenite, and (Pertermann and Hirschmann, 2003) pyroxenite melts. Blue and red lines with crosses show fractional crystallization paths of representative MgO-rich basalts from Shandong (DMM-type), Hainan (FOZO-type), and Changbaishan (EM1-type) calculated using MELTS algorithm (Ghiorso et al., 2002).
The plume influenced basalts are all FOZO-type, such as the South China Sea basalts (Expedition 349 Scientists, 2014; Tu et al., 1991; Wang et al., 2011, 2013; Yan et al., 2014; Zhang et al., 2017), the Baikal Rift–Mongolia basalts (Barry et al., 2003; Johnson et al., 2003), and the Chifeng flood basalt (Guo et al., 2016; Yu et al., 2015). They are isotopically indistinguishable from the basalts of Hainan, Jeju, and some from Datong (e.g., Baiyinchagan) (Fig. 7a). The FOZO-type basalts are believed to be from the deep mantle because of (1) low-Vp bodies in the lower mantle beneath the Hainan–South China Sea and Baikal–Chifeng–Datong areas (Huang et al., 2015; Zhao et al., 2006)(Fig. 4b).
and (2) the absence of the stagnant slab shown by the updated P-wave tomography images from Fukao et al. (2009) (Fig. 1a). All but one study ascribed the FOZO source to a deep mantle plume (Barry et al., 2003; Johnson et al., 2003; Tu et al., 1991; Wang et al., 2011, 2013, 2015; Zhang et al., 2017) (Fig. 7). The FOZO (or C: common component) is often associated with the plume-derived ocean island basalts (Hanan and Graham, 1996; Jackson et al., 2007) and possibly a constant mixture of recycled slab materials and variously depleted mantle sources (Stracke et al., 2003). The major element composition of FOZO-type basalts suggests peridotite melts with or without pyroxenite (Figs. 5 and 6). This source lithology is consistent with a constant mixture origin. Among these FOZO-type basalts, the Chifeng flood basalts is thought to be influenced by a hydrous plume from the stagnant Pacific Plate slab (Wang et al., 2015). We discuss the limited role of water in these basalts in a later section.

EM1-type basalts may have originated from the subcontinental lithospheric mantle, as has been suggested in many previous works (Tang et al., 2006; Xu et al., 2005; Zhang et al., 2003; Zou et al., 2000). Our examinations limit the role of the lithospheric mantle in the area of thick lithosphere, e.g., Wudalianchi-Changbaishan, Ulleung, and Yunnan, although lithospheric mantle beneath Ulleung in the stretched Sea of Japan back arc basin may be thinner. The best candidate source material is the pyroxenite mantle, such as xenoliths with extremely unradiogenic εNd−εHf (Fig. 7a) and εNd = 2–4 Ga (Fig. 3c–d). Such mantle material is suitable for the formation of the EM1-type basalts by mixing with the HB-type source mantle or its melt (Fig. 7a and b). The pyroxenite is associated with the Precambrian terranes (Fig. 3d), and their Archaean lithosphere survives against thermal erosion or delamination (Fig. 3). The pyroxenite xenoliths are from Hanuoba in Datong on the thick lithosphere. The DMM- and FOZO-type basalts always erupt on the thin lithosphere (Fig. 2a) with minimal effect of EM1 (Fig. 7a). This suggests the absence of Archaean lithosphere beneath the thin lithosphere (Fig. 3c and d). The ultra-depleted peridotite (Figs. 3d) covers all the post-Paleoproterozoic lithospheric (Fig. 3). This cannot account for the basaltic source due to ultra-radiogenic εNd−εHf (pale yellow field in Fig. 7a).

The HB-type basalts form a subset of the FOZO-type (orange symbols in Fig. 7a). The intermediate nature between FOZO- and EM1-types suggests minor source mixing of an EM1 source with dominant FOZO source, perhaps forming ambient asthenospheric mantle beneath East Eurasia (Figs. 6 and 7b). Similarly, intermediate basalts between FOZO- and the highest εNd DMM-types could represent source mixing, as with the Shandong basalts that fall on the FOZO–DMM array (Fig. 7b). These basalts are spatially between the FOZO and DMM eruption centers, forming parts of the great concentric geochemical structure (Fig. 5a).

Additionally, the depleted Indian Ocean MORB source mantle (Indian Ocean DMM) dominates mantle beneath the backarc basins above the Pacific Plate either stagnant or subducting (Hickey-Vargas et al., 1995; Miyazaki et al., 2015; Pearce et al., 2007; Woodhead et al., 2012). The D–MORB and E–MORB basalts from the Indian Ocean DMMs are two discrete magma types in the Sea of Japan (Cousens et al., 1994; Cousens and Allain, 1992; Hirahara et al., 2015; Okamura et al., 2005; Poulet et al., 1995) (Fig. 7a). They are commonly more radiogenic in εHf than the East Eurasia intra-plate basalts, reflecting the signature of the Indian Ocean DMM (Pearce et al., 2007; Woodhead et al., 2012) or the effect of subducted sediment (Chauvel et al., 2009; Hirahara et al., 2015) (Fig. 7a). The Indian Ocean DMM is also ubiquitous beneath the Kuriles (Martynov et al., 2012), northeast Japan (Kimura and Nakajima, 2014), Izu (Kimura et al., 2010; Tollstrup et al., 2010), and Mariana (Straub et al., 2015; Woodhead et al., 2012) arcs since 25 Ma (Miyazaki et al., 2015).

4.3.5. Variations in Pb isotopes

Pb isotopes for East Eurasian basalts are generally consistent with εNd−εHf apart from large variations in the proportion of EM1. The isotopic array formed by the DMM-type of Jiangesu through Shandong to FOZO-type Hainan align parallel to the Northern Hemisphere Reference Line (NHRL) (Zindler and Hart, 1986) with increasingly radiogenic Pb (Fig. 8a). There is another parallel array from FOZO-type through HB- to EM1-types when Yunnan basalts are disregarded (Fig. 8a and b). Although the mixing array is strongly deflected at FOZO-type Hainan, their mixing relationships are maintained (Fig. 8b). The South China Sea basalts fully overlap and the most radiogenic Baikal–Chifeng basalts plot close to the Hainan FOZO-type basalts. Crustal assimilation is reported for the Baikal basalts, explaining the wide variations seen for this suite (Barry et al., 2003; Johnson et al., 2003) (Fig. 8a). The ~60 Ma Izanagi–Pacific Ridge MORBs from the Shimanto greenocks have isotopic composition similar to the DMM-type basalts (Fig. 8a).

For a further test, we examined 206Pb/204Pb–εHf plots (Supplementary Material Appendix C). The results show that FOZO-type mantle from Hainan, South China Sea, Baikal, and Jeju (slightly affected by EM2, see also Fig. 8a) plot within a narrow compositional range and confirm mixing arrays that extend from FOZO-type to DMM-type and FOZO-type to HB- and EM1-types. Interestingly, EM1-type basalts of Yunnan, Changbaishan, and Wudalianchi stems from different portions of the FOZO–HB array indicating different mixing trajectories of a discrete EM1-type component with different Pb isotope composition (Supplementary Material Appendix C and Fig. 8b).

4.4. Trace element model

We used two numerical simulation models, PRIMACALC2 (Kimura and Ariskin, 2014) and OBS1 (Kimura and Kawabata, 2015). PRIMACALC2 is used to estimate primitive magma composition. OBS1 uses the modeled primitive magma to infer sources and melting conditions.

4.4.1. PRIMACALC2 model

The details for calculating primary magma compositions using PRIMACALC2 are given elsewhere (Kimura and Ariskin, 2014). A brief summary is in Supplementary Material Appendix D. The effect of crustal assimilation is not considered due to negligible isotopic variations in the fractionated basalts shown above. Estimation of a primary basalt from a pyroxenite-bearing peridotite source can also be made by examining Mg# (Mg# = Mg/[Mg + Fe] given by molar ratio) and Ni in olivine in equilibrium.

4.4.2. OBS1 model

OBS1 calculates the trace element mass balance between the primary basalt and model mantle melt on the basis of the adiabatic melting of dry to wet peridotite mantle with or without recycled pyroxenite (Kimura and Kawabata, 2015). The source mantle peridotite is a hypothetical mixture of DMM (Workman and Hart, 2005) and PM (McDonough and Sun, 1995), given by fractions of fDMM and fPM. The pyroxenite fraction (fPY) is also included assuming a normal (N-)MORB composition (Jenner and O’Neill, 2012). We note that in terms of OBS1 calculations, the trace element compositions of PM, DMM, and N-MORB pyroxenites are significant but their isotopic compositions are not. OBS1 uses fDMM, fPM, and fPY to identify the source lithology. The calculations use 26 incompatible trace element compositions of a primary basalt. The source conditions of Tp, Pmelt, F, and Tmelt are determined by fitting the calculated melt compositions. The AHe is initially given to explore the effect of water. The model is detailed elsewhere (Kimura and Kawabata, 2015). The method to estimate fPY of N-MORB sourced pyroxenite is given briefly in Supplementary Material Appendix E.

4.4.3. Rationale and drawback in Eurasian basalt analysis

The OBS1 model is applicable to the East Eurasia basalts because their chemistries are similar to those of ocean island basalts, and because they originate by adiabatic melting of a wet or hot mantle
plume (Kuritani et al., 2013; Sakuyama et al., 2013, 2014; Wang et al., 2011). The OBS1 model is best suited for FOZO- and DMM-type basalts. The N-MORB sourced pyroxenite is the initial assumption in OBS1 and recycling oceanic crust, either from the stagnant slab (DMM-type) or from the lower mantle (FOZO-type) that are in the basalts (Supplementary Material Appendix E).

The subcontinental lithospheric pyroxenite for the EM1-type source is not involved in adiabatic melting unless the lower part of the subcontinental lithosphere is entrained in plume upwelling. Nevertheless, OBS1 is still applicable. Pyroxenite xenoliths from East Eurasia formed in a supra-subduction environment (Xu, 2002; Yu et al., 2010). They have flat middle rare earth element (MREE) to heavy rare earth element (HREE) patterns and elevated fluid mobile elements, such as Rb, Ba, Th, U, K, Pb, and light rare earth elements (LREEs) (Supplementary Material Appendix F). A small pyroxenite melt fraction results in increased fluid mobile element contents in basalts; however, it does not affect OBS1 thermodynamic calculations using elemental mass balance of REE and high field strength (HFS) elements. We, therefore, can explore the $f_{py}$ of N-MORB sourced pyroxenite in the mantle source even if the EM1-type basalts were contaminated by lithospheric pyroxenite melts. One drawback is that estimates of mantle conditions and components should be less precise from the effect of lithospheric melt contamination (see more details in Supplementary Material Appendix F).

4.5. Model results

All of the basalts studied here are mildly alkalic (transitional tholeiite–alkali basalt) to alkalic (basanitic) with some trachytic compositions. These samples range in Mg# from 0.43–0.69, averaging 0.59 with SiO$_2$ = 40.99–54.22 wt%, averaging 48 wt%. High-MgO samples

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Fig. 8. The Pb isotopic variations of Late Cenozoic East Eurasian intra-plate basalts. (a) Pb isotope compositions of East Eurasia basalts and source mantle materials in and adjacent to the area. (b) Model mantle compositions of East Eurasian subcontinental lithospheric mantle and deep recycled mantle sources same as in Fig. 7b (Kimura et al., 2016). Data sources are the same as in Fig. 7 with fields of the Pacific sediment from Miyazaki et al. (2015) and the Baikal from Rasskazov et al. (2002).
were chosen when possible in order to minimize the effects from crustal assimilation and from high SiO₂-low MgO melts of lithospheric pyroxenite in the EM1-type (Fig. 5c and d). Changbaishan, Shandong, Jiangsu, and Hainan basalts have more than one type of basalt and were analyzed separately. Calculated primitive basalt compositions are listed in Supplementary Material Table S2. The OBSI results are given in Supplementary Material Table S3. The representative fitting results are in Fig. 9, showing excellent fits of the modeled magma compositions.

4.5.1. Trace element characteristics

All basalts are enriched in LREEs with variously steep REE slopes shown in multielement plots for the primary basalt compositions (Fig. 9). The HREE abundances also vary from 1 × primitive mantle (PM) values from the literature (McDonough and Sun, 1995) to several-times PM. Extended trace element patterns also vary from quasi-straight to MREE-humped patterns (Fig. 9), and the details of these patterns differ among basalts.

The DMM-type Shandong basalts have negative anomalies in Rb, Ba, and Pb with slight positive anomalies in Zr–Hf (Sakuyama et al., 2013; Zeng et al., 2010). Jiangsu basalts of DMM-type are similar to those of Shandong but do not have a strong negative Pb anomaly. There are two common types of trace element patterns in the two basalt centers. One group has a strongly MREE-humped pattern with depletions in Rb, Ba, and K (Fig. 9a and c). Another group has a straight REE pattern with or without negative spikes in K and Rb (Fig. 9b and d). These discrepancies are likely to originate from various degrees of depletion of those fluid-mobile elements from the N-MORB pyroxenite source due to subduction modification (Kimura et al., 2016) before pyroxenite was mixed with the DMM-type mantle source (see below for further discussion).

There are three different types of FOZO-type Hainan basalts: alkali olivine basalt (AOB), quartz tholeite (QT), and olivine tholeite (OT) (Wang et al., 2011) (Fig. 9e, f, and g). Trace element abundances increase from OT to AOB. They share common trace element features with nil to negative K spikes; those of other elements are relatively smooth except for Pb, which shows positive spikes.

The EM1-type Changbaishan basalts have tholeiitic and alkalic types with low and high trace element abundances, respectively (Fig. 9h and i). They commonly possess strongly positive spikes in LILEs such as Rb, Ba, K, Pb, and smaller Sr spikes relative to adjacent elements (Kuritani et al., 2011). There is no systematic spike or trough in Nb-Ta. The same trace element feature is present for EM1-type Wudalianchi basalts (Kuritani et al., 2013) (Fig. 9j). The Yunnan basalts have straight REE patterns that continue up to Nb-Ta, while troughs are shown in U-Th with elevated Rb. This pattern is distinctive, perhaps due to an unusual source or the mantle solidus in the crust (Fig. 9k).

There are compositional variations in the intermediate types. The Liaoning basalts are isotopically similar to FOZO but compositionally similar to the tholeiitic basalts of Changbaishan–Wudalianchi, except for showing a positive Sr spike (Fig. 9i). The Wuhanada basalts are close to an HB-type and are similar to Changbaishan–Wudalianchi basalts, but exhibit less prominent spikes (not shown). The Anhui basalts are also of the HB-type and have similar trace element patterns to the Hainan AOB (Fig. 9m). The Dongba basalts are similar to the DMM-type of Shandong, highly alkalic, and show strong positive troughs in Nb-Ta (Fig. 9n), sharing the same features with the Chifeng flood basalts (Yu et al., 2015).

4.5.2. Source mantle condition: $XH_2O$

The calculated $XH_2O$, $P_{melt}$, $T_p$, and $f_{PY}$ are in Supplementary Material Table S3. The results are given as average, minimum, and maximum solutions and are plotted in Fig. 10. The source mantle beneath East Eurasia is not extremely wet for a backarc setting. Rather, the range is comparable with that in ocean island tholeiite and alkali basalts with $XH_2O = 300–800$ ppm and 0.3–1.0 wt% water in basalts (Dixon et al., 2004). The model calculations assumed a constant $XH_2O$ of 300 ppm in the mantle source. The estimated water contents in the calculated primitive basalts were 0.2–2.1 wt% (Fig. 10a). This range is consistent with the minimum water contents in the primary basalts calculated from water in clinopyroxenes (Chen et al., 2017). Higher water contents, as much as 2–4 wt%, are suggested for Chinese basanitic to alkalic basalts. However, they are from low degree melts ($F = 0.01$) that require only 300–600 ppm $XH_2O$ in the source (S.-C. Liu et al., 2016).

4.5.3. Source mantle condition: $T_p$

Mantle potential temperature, $T_p$, is calculated to be 1280–1440 °C. The estimated $T_p$ distribution is robust and is high in the plume-influenced FOZO-type basalts from Hainan (1420 °C) and Dongba in Datong (1370 °C), while it is low in the stagnant slab-influenced DMM-type Shandong–Jiangsu basalts (1310 °C) (Fig. 10b). The high $T_p$ calculated for EM1-type Changbaishan basalt is less reliable due to the contribution of subcontinental lithosphere pyroxenite. The same may be true for Wudalianchi. The estimated $T_p$ from Hainan differs between Hainan-QT, -OT, and -AOB (Fig. 10b). This represents different eruption stage in the middle Miocene to the Holocene for the basalts examined here (Wang et al., 2011) or local heterogeneity of thermal structure in the plume (Hirahara et al., 2015). Difference in the estimated $T_p$ of Hainan-QT between ours (hottest 1440 °C) and Wang et al. (2011) (1500–1600 °C) originates from use of dry peridotite solids in the later work. Pyroxenite source estimated in the Hainan-OT and -AOB sources result in even lower $T_p$ in OBSI.

The $T_p$ is a function of $XH_2O$ in the OBSI mass balance model calculations. We can test the effect of $XH_2O$ using OBSI1. The transitional Hainan-QT basalt was estimated to be at $T_p = 1420$ °C when 300 ppm $XH_2O$ was in its source (Supplementary Material S3). If the hydrated asthenosphere above the stagnant slab upwellled and the mantle was near ambient geotherm ($T_p = 1330$ °C), 2500 ppm $XH_2O$ is needed in the mantle source to generate the basalt. The calculated water in the primary Hainan-QT basalt was 2.6–2.7 wt% (see “Hi water” columns in Supplementary Material Table S3). Such high water is not observed in the transitional basalts (Chen et al., 2015) precluding high $XH_2O$. Further, the estimated $T_p$ is lowered only 20–30 °C by increasing $XH_2O$ to 900 ppm as in the 50–Ma Xiaotuqiyuan transitional basalt (see “1% H2O” columns of Hainan-QT in Supplementary Material Table S3). Our estimates for $T_p$ are robust.

4.5.4. Source mantle condition: $F$

In general, alkalic basalts show a lower $F$ (e.g., the Hainan-AOB, Hainan-OT, and alkali Jiangsu basalt), while transitional basalts have higher $F$ (e.g., the Hainan-QT and transitional Jiangsu and Shandong basalts). With a constant $XH_2O$ assumed in the model, the water does not control $F$ at the top of the adiabatic melting column (Fig. 10a). Fig. 10c indicates that $T_p$ is the major control for $F$, and $f_{PY}$ has a secondary impact by lowering the source mantle solidus temperature. The estimates of $T_p$, $f_{PY}$, and $F$ are independent and robust (Supplementary Material Appendix E) (Kimura and Kawabata, 2015).

4.5.5. Source mantle condition: $P_{melt}$ and $T_{melt}$

The estimated depths of the top of the melting column, $P_{melt}$, correlate well with the seismic lithosphere–asthenosphere boundary (Fig. 2c). The FOZO-type transitional basalts from Hainan and Datong basalts have deeper and hotter $P_{melt}$ and $T_{melt}$, presumably reflecting a deeper plume source. The DMM-type Jiangsu basalts show the lowest $P_{melt}$ and $T_{melt}$ (Fig. 2d). The $P_{melt}$ anticorrelates with basalt $T_{melt}$, and the Hainan basalts show a linear negative slope between the highest $P_{melt}$ and $T_{melt}$ Hainan-QT and the lowest $P_{melt}$ and $T_{melt}$ Hainan-AOB. The isotopically intermediate HB-type basalts form a similar trend between Shandong and Datong (Dongba). The EM1-type Yunnan and Wudalianchi basalts overlap on the intermediate trend and the Changbaishan transitional basalt plot on the Hainan-QT. These, however, may be somewhat erroneous.
Fig. 9. Multielement plots of trace element compositions from representative East Eurasian intra-plate basalts. Black line: primary basalt composition estimated by PRIMACALC2 (Supplementary Material Table S2); thick red line: average basalt composition calculated by OBS1; and thin red line: those for minimum and maximum values by OBS1 (Supplementary Material Table S3). Values normalized to PM (McDonough and Sun, 1995).
4.5. Source composition: fPY and fPM

The fPY fraction of N-MORB pyroxenite origin differs between isotopic groups. The FOZO-type Hainan-QT with the highest Tp has almost nil fPY. The DMM-type basalts are rich in fPY (≤ 0.46). Isotopically intermediate basalts including HB-type have intermediate fPY (Fig. 10c and d).

Notably, the Dongba basalt from Datong is rich in fPY although other parameters, such as Tp and Pmt, are somewhat similar to those from Hainan-QT. The alkali basalts from Jiangsu have far less fPY in contrast to those in transitional basalt sources. The alkali basalts from Hainan-AOB have the highest fPY among the basalts from the area (Fig. 10c and d). The fPY fraction differs even in the same basalt province with the same isotope groups. This is natural because the distribution of the recycled pyroxenite blob differs in the plume source as suggested for the ocean islands, such as Hawaii (Sobolev et al., 2007) or the Columbia River Basalt (Takahahshi et al., 1998).

The fPY fraction of the EM1-type basalts is intermediate value (e.g., Wudalianchi) or almost nil (Changbaishan) (Fig. 10c and d). The OBS1 model detects pyroxenite of an N-MORB source only. The low fPY in Changbaishan basalt suggests no such source; nevertheless, pyroxene melt from the subcontinental lithospheric mantle is included as suggested by the elevated fluid mobile elements. Relatively high fPY in the Wudalianchi and Yunnan would reflect the role of asthenosphere sourced pyroxenite.

The estimated fPM–fDMM fractions are given in Fig. 10d. Although the errors are large, the source peridotite should have been fertile for all the basalts (fPM > 0.5). This result indicates that the basalts are not from Indian Ocean DMM (e.g., fPM = 0), which is ubiquitous beneath the Sea of Japan (Hirahara et al., 2015), northeast Japan arc (Kimura and Nakajima, 2014), and Kurile arc (Martynov et al., 2012). This fertile source lithology is consistent with the ocean island basalt type enriched 6NCS–6H Isotope compositions (Fig. 7a).

5. Roles of the plume, stagnant slab, and continental lithosphere

Based on the isotope geochemistry and trace element modeling, we discuss the roles of the plume, stagnant slab, and continental lithosphere for the origin of the late Cenozoic East Eurasian basalts. We explore the thermal condition in the mantle and the source mantle components and propose the tectono-magmatic evolution of the East Eurasia margins.

5.1. Thermal condition

5.1.1. Deep plume

The asthenosphere beneath Hainan has an exceptionably high mantle potential temperature (Tp = 1420–1440 °C) (Fig. 10). This result is consistent with the observed up- and down-warping of the 660 km and 410 km mantle transition boundaries which show 380 K and 220 K higher mantle temperatures, respectively, compared to the ambient mantle (Huang et al., 2015). This corresponds to a Tp ≥ 1450 °C in the mantle transition zone (Ono, 2008). These observations collectively indicate deep upwelling of a mantle plume from the lower mantle (Fig. 4a). Similar mantle potential temperature of Tp = 1450 °C has been estimated for the Sea of Japan basalts during opening (Ono, 2008). Hawaiian and other ocean island basalts also show Tp = 1400–1500 °C (Herzberg and Asimow, 2008; Kimura and Kawabata, 2015). These ocean island basalts are thought to be generated by the plumes from the core-mantle boundary (French and Romanowicz, 2015).
5.1.2. Lithosphere–asthenosphere boundary

The $T_m$ and $P_m$ at the lithosphere asthenosphere boundary beneath the basalt provinces negatively correlate, suggesting a regular mantle geotherm of $\Delta T_m = -5 \degree C/km$ in the asthenosphere (Fig. 11 orange geotherm). A steeper mantle geotherm of $\Delta T_m = -10 \degree C/km$ is found in Hainan (Fig. 11 red geotherm). The subcontinental lithospheric mantle geotherm has been estimated by thermobarometry of East China mantle xenoliths (Chen et al., 2001; Xu et al., 1996; Yu et al., 2003) (Fig. 11 black geotherm). The lithosphere–asthenosphere boundary exhibits a $\Delta T$ of 120–150 ˚C thermal gap between the molten asthenosphere and sublithospheric lithosphere (Fig. 11).

The regular (orange) asthenospheric geotherm is parallel to the solidus curve of a fertile peridotite containing 100–300 ppm $H_2O$, consistent with the assumption that 300 ppm water in the mantle was responsible for the low degree of partial melting ($F = 0.01-0.03$). The lithosphere is sublithospheric due to low-$T$ at ~1200 °C adjacent to the asthenosphere boundary. The lithospheric mantle is dry as suggested by the absence of hydrous minerals, lack of traces of melting (Chen et al., 2001; Xu et al., 1996; Yu et al., 2003), and the low $XH_2O < 300$ ppm in the xenoliths (Hao et al., 2012).

5.1.3. Melting of subcontinental lithosphere

The mantle geotherms (Fig. 11) show the importance of melting pyroxenite in the subcontinental lithosphere. The solidus of dry pyroxenite (Yasuda et al., 1994; Yaxley, 2000) is shown together with that for peridotite (Fig. 11 green dashed line). The pyroxenite solidus at 2–2.5 GPa ranges from 1200 to 1240 °C, indicating that melting of the lithospheric pyroxenite by heat from underplated or intruded basalts could generate $SiO_2$-rich EM1-type basalt. Nevertheless, other basalt-types may be asthenospheric mantle melts with or without pyroxenite from the stagnant Pacific Plate slab and/or deep upwelling mantle in their sources (Fig. 5).

5.2. Mantle source

5.2.1. Relationship of FOZO–DMM–EM1 components

To test the links between the isotopic composition and source lithology, Fig. 12 plots $\epsilon_{Nd}$ in basalt versus $F_{py}$. The typical FOZO-type Hainan-QT with $\epsilon_{Nd} = -4$ shows almost no $F_{py}$. The DMM-type basalts of Pacific IOC show mixing between EM1-DMM components (Fig. 12).

![Fig. 11. Comparison of P–T profiles in the lithosphere–asthenosphere mantle beneath East Eurasia. Lithosphere P–T data (dark blue dots) are from mantle xenoliths and Qilin and SE Australia geotherms, and mineral stability fields are from the literature (Chen et al., 2001; Xu et al., 1996; Yu et al., 2003). The P–T conditions from DMM-, FOZO-, and HB-types basalts (orange circles) and the high-T Hainan-QT basalt (black diamond) are from OBS1 calculations. The thermal gradient at the lithosphere–asthenosphere boundary is $\\Delta T = 120-150$ °C higher for the common areas. The Hainan basalts show exceptionally steep geotherm (thick red line) compared to others (thick orange line). Periodotite solidus lines with various water contents are from Katz et al. (2003). Solidus line for dry pyroxenite is from the literature (Yasuda et al., 1994; Yaxley, 2000).](image)

![Fig. 12. Correlation between pyroxenite fraction of MORB source ($f_{py}$) and $\epsilon_{Nd}$ in the source mantle materials. The intermediate basalts (red) plot between the DMM-type Shandong–Jiangsu (blue) and FOZO-type Hainan (red) endmembers, showing a linear mixing hyperbola between the Pacific Plate oceanic crust ($f_{py} = 1, \epsilon_{Nd} = 8$) and a deep plume source ($f_{py} = 0, \epsilon_{Nd} = 4$). The EM1-type Changbaishan basalts plot between the Hainan FOZO-type composition and a pyroxenite from the subcontinental lithosphere (MORB source $f_{py} = 0, \epsilon_{Nd} = -10$). The EM1-type Wudalianchi and Yunnan basalts plot on the mixing hyperbola drawn between the DMM-type basalt and the subcontinental lithosphere pyroxenite melt.](image)
Jiangsu–Shandong show \( f_{PV} = 0.4 \)–0.5 and plot in the middle of the Hainan–QT and Pacific Plate slab composition at \( f_{PV} = 1 \) and \( \varepsilon_{Nd} = -8 \). The EM1-type Changbaishan basalt plot at \( f_{PV} = 0 \) with lower \( \varepsilon_{Nd} = -2 \). Its \( \varepsilon_{Nd} \) is intermediate between Hainan–QT and pyroxenite from the subcontinental lithosphere, \( \varepsilon_{Nd} = -12 \) (see Fig. 7a).

The extremely low-\( \varepsilon_{Nd} \) is from the selective contribution from the subcontinental lithospheric pyroxenite. The pyroxenite melts by the heat from the asthenosphere-derived basalts, while lithospheric peridotite does not (see solidi in Fig. 11). This accounts for the significant contribution of lithospheric pyroxenite in the EM1-type basalts with lack of contributions from the lithospheric peridotites, with extremely high-\( \varepsilon_{Nd} \) (Fig. 7a). As noted earlier, the subcontinental lithospheric pyroxenites are distinguished from other N-MORB sourced pyroxenites by OBS1 modeling showing \( f_{PV} = 0 \) as in Changbaishan. However, other EM1-type Wudalianchi and Yunnan basalts possess \( f_{PV} = 0.3 \) and plot on the mixing line between the subcontinental lithospheric pyroxenite and the DMM-type basalts suggestive of the \( f_{PV} \) component in the asthenosphere-derived basalts of DMM-type (Fig. 12). The HB-type Wulanhada basalts plot between the FOZO- and EM1-types, and the intermediate isotope type basalts of Anhui and Dongha plot between the FOZO- and DMM-types, as expected from their \( \varepsilon_{Nd} \) and major elements (Figs. 5 and 6).

A further assessment of East Eurasian basalt sources is available. Isotopic variations between the FOZO- and DMM-types are regarded to be mixtures between the plume-sourced deep mantle and stagnant Pacific Plate slab components (Fig. 12). The variations between the FOZO- and HB-types are regarded to reflect mixing between a deep plume source and subcontinental lithospheric pyroxenite components (Fig. 12). Moreover, the wide isotopic variations in Yunnan and Wudalianchi may be mixtures of stagnant slab-influenced DMM-type melts and melts from subcontinental lithospheric pyroxenite (Fig. 12). The different slopes of \( \varepsilon_{Nd}–f_{PY} \) Trends drawn between the EM1–HB–FOZO–DMM-type components (Fig. 5a) are, therefore, explained by different mixing processes.

### 5.2.2. Spatial distribution of mantle sources

Fig. 13 summarizes the spatial distribution of the FOZO- (red), DMM- (gray), EM1-(purple), and HB- and intermediate (orange) East Eurasia basalt types, including the Baikal Rift, Mongolia, Cifheng flood basalt, and the South China Sea areas. The depleted Indian Ocean DMM basalts (white) are also shown for Sikhto-Alin, the Sea of Japan, and the sub arc mantle of northeast Japan, Kurile, although some DMM-type component exist beneath Hainan and the South China Sea (Figs. 5–7, Supplementary Material Appendix C). For comparison, isotope data for \( \sim 35 \) Ma basalts from southwest Japan, Ryukyu, and the Philippine Sea Plate are also shown. See Supplementary Material Appendix B for full references of the data sources.

The prominent first-order structure is the contrasting fertile FOZO-related (red and orange) in the west versus depleted Indian Ocean DMM (white) in the east. FOZO influences all of western East Eurasia and its influence extends into the great marginal basins in the western half of the Sea of Japan and the Philippine Sea. The Indian Ocean MORB source underlies the eastern Sea of Japan and the eastern margins of the Philippine Sea Plate, including the Izu-Bonin-Mariana (IBM) arc system (Fig. 13). The border between the two great mantle source regions lies between the NE and SW Japan arcs (Kimura et al., 2014) and between Sikhto-Alin and the Korean Peninsula. The FOZO-Indian Ocean DMM boundary is obscured in the Philippine Sea by the occurrence of FOZO-related basalts associated with the plume-influenced Benham Rise, Central Basin Fault, and Kinan Seamount Chain (Hickey-Vargas et al., 2006; Ishizuka et al., 2007, 2009) (Fig. 13).

The second-order structure is the great concentric isotopic distribution examined in this study (gray–red–orange–purple in Fig. 13a). The FOZO-type mantle is revealed in basalts from the Baikal Rift, Mongolia, Chifeng, Datong, Hainan, and South China Sea areas. These basalts are all thought to be derived from a deep plume that originated below the stagnant Pacific Plate (see stars for plume locations in Fig. 13a–d with mantle tomography slab images in the depth range 700–620, 500–480 and 410–350 km revised from Fukao et al., 2009). The Baikal–Datong plume rises along the western edge of the Pacific Plate slab in the mantle transition zone. The Chifeng flood basalt is underlain by the slab at 550–480 km depth, but the plume appears to extend above the slab, and its margin may reach to Chifeng (Fig. 13d). The Hainan plume is discrete from the Baikal–Datong plume and escapes upward through the slab tear (Fig. 13c). At 410–350 km depth, the Hainan plume spreads widely above the slab. It underlies the South China Sea and perhaps extends to beneath Jeju Island, where FOZO-type magma erupts (Fig. 13d). The presence of a plume is unclear beneath the Philippine Sea Plate; however, the FOZO-type mantle occurs above slab windows beneath the Benham Rise and the Central Basin Fault (Fig. 13b–d). The Kinan Seamount Chain and Iwo-Jima are underlain by the thick Pacific Plate (Fig. 13c). They may be affected by lateral mantle flows from the Hainan plume (Fig. 13d).

The third-order structure observed is the eruption of DMM-type basalts in and around the Shandong Peninsula. This may reflect entrainment of the stagnant Pacific Plate slab by upwelling mantle. The area is clearly underlain by the Pacific Plate slab in the mantle transition zone. It is notable that the slab in this area is deeper, ranging from 480 to 700 km depth (Fig. 13b–d). This part of the slab is surrounded by the Hainan and the Datong–Baikal plumes with \( T = 1400–1450 \) °C. The interaction of upwelling plumes and the stagnant slab may entrain subducted oceanic crust. Similar DMM-type basalts are reported in the Cifheng flood basalt (Wang et al., 2015; Yu et al., 2015) and Miocene South China Sea basalt (Zhang et al., 2017) (Fig. 13); these can be ascribed to the entrainment of the edge of the stagnant slab by the Datong–Baikal and Hainan plumes. These DMM-affected areas are all related to the leading edges of the stagnant slab. Entrainment of the DMM-type component might be enhanced in contrast to the areas where rigid and cold stagnant slab is present and strong upwelling of a hot deep plume is absent.

Finally, the fourth-order structure is that EM1-type sources occur only where thick continental lithospheric mantle remains (Yunnan, Wudalianchi, Changbaishan, and Ulleung Island; see Fig. 2). They are located in the perimeters of the concentric structure and above the margins of the stagnant Pacific Plate slab (Fig. 13c), but with no other relationship to the stagnant slab and two plumes. The Wudalianchi and Yunnan basalts have contributions from the DMM-type mantle source (Fig. 12). The stagnant slab component may be entrained in their asthenospheric mantle source, similar to Chifeng. Asthenospheric melts are mixed with melts from subcontinental lithospheric pyroxenite (Fig. 12). The Changbaishan source does not exhibit any traces of the Pacific Plate slab and is readily explained by mixing between HB- or FOZO-type basalt and lithospheric pyroxenite melt (Fig. 12). The FOZO source mantle occurs below Chifeng close to Changbaishan and may be derived from the Datong–Baikal plume, which extends to beneath the Changbaishan area.

Fig. 13e summarizes a schematic mantle cross section along the gray line on Fig. 13b–d. The FOZO-source mantle upwelled from the lower mantle along the leading edge and through a tear in the stagnant Pacific Plate slab. The ambient upper mantle was originally Indian Ocean DMM. It still remains beneath the Sea of Japan, the Kurile Basin, northeast Japan arcs, IBM arcs, and partially beneath Hainan. The upwelled plumes interacted with the stagnant Pacific Plate slab to form DMM-type source. Thick subcontinental lithosphere remained beneath the Tibetan Plateau and adjacent areas, as well as beneath the Korean peninsula and Sikhto-Alin. This resulted in basaltic melts that range from FOZO-type or DMM-influenced FOZO-type with added EM1-type signatures.

Dehydration of the subducted/stagnated Pacific Plate slab in 35–15 Ma would have supplied some water to form 500–1000 ppm water in the deep upper mantle based on electrical conductivity studies (Ichiki et al., 2006, 2008) (Fig. 13e). Part of the water in the plumes may originate from this hydrated mantle, but not a significant amount in the basalts sources as estimated from the water in basalts.
Fig. 13. Maps showing distribution of $\varepsilon$Nd in the basalts and the stagnant Pacific Plate slab in the mantle transition zone with a schematic cross section showing the isotopic mantle sources. (a) Isopleth map showing spatial distribution of $\varepsilon$Nd in basalts, (b) P-wave tomography map at 700–620 km depth, (c) the same map at 500–480 km depth, and (d) at 410–350 km depth. The locations of FOZO-type Hainan, Datong, and Baikal plumes are shown by red stars; FOZO-type basalts are shown by red circles. The highest $\varepsilon$Nd is in the Shandong–Jiangsu (DMM-type) basalts shown by gray symbols. The HB-type and intermediate basalts are shown by orange circles. The EM1-type centers shown by purple circles are Ulleung, Changbaishan, Wudalianchi, and Yunnan in the perimeters showing the lowest $\varepsilon$Nd. The white circles correspond to basalts of the Indian Ocean DMM-type with the most radiogenic $\varepsilon$Nd. Note that $\varepsilon$Nd contour lines are smoothed in order to show regional variation. Also note that colour in the circles do not exactly correlate with the colour scale in $\varepsilon$Nd. Green and red contour lines indicate depths to the Wadati–Benioff zone of the Pacific and the Philippine Sea Plate slabs, respectively. Thick dark blue lines show trenches. Black dashed lines show isopleths of $\varepsilon$Nd values as in panel (a). Gray line indicates location of the cross section for panel (e). (e) Cross section showing genetic relationship between the mantle sources of East Eurasian basalts. Dark gray: igneous oceanic crust source from the Pacific Plate. Yellow: ambient Indian Ocean DMM. Pink: deep plume source mantle. Gray, orange, and blue on the arrows represent DMM, FOZO, and EM1-type magmas, respectively. Dashed circles indicate interactions between asthenospheric magma to the Pacific Plate slab and subcontinental lithosphere. Light blue shaded area is the hydrated lower half of the upper mantle due to dehydration of the subducted Pacific Plate slab.
5.3. Origin of FOZO–DMM–EM1 components

5.3.1. Eurasian FOZO from a Paleoproterozoic subducted slab

Mantle wedge peridotite above a subducted slab (MwP) is soaked by the slab flux and forms a slab composite together with the igneous oceanic crust (IOC) (Kimura, 2017). If the slab composite is recycled deep into the mantle at different ages (e.g., 0.1, 2, and 2.5 Ga), they will evolve isotopically. Thick dark blue lines in Figs. 7b and 8b indicate an isotope growth model of the IOC–MwP slab composite for 0–2.5 Ga (Kimura et al., 2016). The slab composite can stay in the mantle transition zone by density crossover or sink to the core-mantle boundary due to its high density (Hirose et al., 1999). The East Eurasian FOZO mantle is explained by a slab composite that formed in the Late Paleoproterozoic to Mesoproterozoic and stored somewhere in the deep mantle (1.0–1.3 Ga for $^{206}\text{Pb}/^{238}\text{U}$ and 1.5–2.0 Ga for Pb isotopes) (Figs. 7b and 8b).

FOZO as a global mantle source is thought to be a near-constant mixture of recycled slab materials (Stracke, 2012). Such deeply recycled material would have been stored in the lower mantle for about 2.0–1.0 Gyr beneath the northern hemisphere (Kimura et al., 2016).

We prefer this ancient subduction recycling model for FOZO source rather than an FOZO source of 4.4–4.5 Ga age derived from an early silicate Earth reservoir (Wang et al., 2013) for two reasons: (1) the absence of a HIMU (high-$\varepsilon_Hf$ $\sim 5.8^{206}\text{Pb}/^{238}\text{U}$) mantle component in northern hemisphere basalts including East Eurasia; HIMU formed from subduction-modified igneous oceanic crust with $>2.0$ Ga recycling age (Hanju et al., 2014; Kimura et al., 2016); and (2) the absence of a high $\varepsilon_{4He}$ mantle source for the East Eurasia basalts (Barry et al., 2007; Chen et al., 2007) precluding the contribution of the unfractonated early silicate Earth reservoir (Jackson et al., 2007; Mundel et al., 2017).

The FOZO source is always associated with plumes in East Eurasia and is most probably from the core-mantle boundary. Small-scale low $\delta_V^+/V_p$ perturbation areas are detected in the lowest mantle (2800 km deep) beneath the Hainan and Baikal plumes (French and Romanowicz, 2015). These low velocity anomalies could be the site of deep FOZO storage and the source of the plumes.

5.3.2. Eurasian DMM from Pacific plate stagnant slab

All our observations suggest that the East Eurasian DMM-type source mantle differs from Indian Ocean DMM and is subduction-modified Pacific Plate igneous oceanic crust origin. DMM-type sourced basalts from Shandong Peninsula have strong MREE-humped patterns with depletions in Rb, K, and Pb and the HIMU signatures of ocean island basalts (Willbold and Stracke, 2006) (Fig. 9). Nevertheless, their isotopic compositions are not like HIMU, but are similar to ~60 Ma MORBs from ~30°E (Sakuyama et al., 2013) (Fig. 3b) is also consistent with a model that calls for long-term storage in the subcontinental lithosphere.

The modeled trace element composition of the RMP is also consistent with the EM1 component in the mantle transition zone for a longer time (Kuritani et al., 2011) and Wudalianchi (Wang et al., 2017). Retaining the EM1 component in the mantle transition zone for a longer time may be difficult due to vigorous whole mantle convection (Nakagawa et al., 2010).

5.3.3. Eurasian EM1 from ancient supra-subduction zone lithosphere

The EM1 mantle source can be derived from the subcontinental lithosphere (Hoernle et al., 2011) or lower crust (Willbold and Stracke, 2010). A supra-subduction zone mantle peridotite modified at $\sim$2 Ga can reproduce $^{143}\text{Nd}/^{144}\text{Nd}$ and Pb isotopic compositions of EM1-type pyroxenite source in the subcontinental lithosphere.

Model calculations show that the EM1 source can also form from residual mantle peridotite (RMP) in the upper mantle wedge of ancient subduction zones along with a deeply recycled Archean MwP (2.5 Ga). The green open stars with RMP in Figs. 7b and 8b show present day isotopic composition of RMP formed in a hot ($T_\text{m} = 1600$ °C) subduction zone at 2.0 Ga (Kimura et al., 2016). The observed compositions of lithospheric pyroxenite (green dots in Fig. 7b) can thus form from the RMP. The modeled trace element composition of the RMP is also consistent with the subcontinental lithospheric pyroxenite (Supplementary Material Appendix F). The Archean Re-Os age and $\tau_{\text{mantle}}$ age of $>2$ Ga from Chinese pyroxenite xenoliths (Fig. 3b) is also consistent with a model that calls for long-term storage in the subcontinental lithosphere. Moreover, mixing of ancient RMP (2 Ga) and ancient depleted uppermost mantle D-DMM ReLish (2–3 Ga) (Salties et al., 2011) would explain compositions of the East Eurasian mantle peridotite xenoliths (yellow dots with black circles in Fig. 7a, see green mixing hyperbolae between RMP and D-DMM ReLish).

The above conclusion precludes the origin of the EM1 component from the mantle transition zone proposed previously for Changbaishan (Kuritani et al., 2011) and Wudalianchi (Wang et al., 2017). Retaining the EM1 component in the mantle transition zone for a longer time may be difficult due to vigorous whole mantle convection (Nakagawa et al., 2010).

5.3.4. Where did the EM2 component go?

An additional note is that E-type MORB in the Sea of Japan is explained by involving sediment from the subducted Pacific Plate itself (Hirahara et al., 2015). The absence of a young sediment component in DMM-type basalts can be explained by the thin sediment cover along the Izanagi–Pacific Ridge or the removal of the sediment layer by offscraping or within the subduction zone owing to its buoyancy (Behn et al., 2011; Hirahara et al., 2015). The latter mechanism would explain the absence of a young sediment component in East Eurasian basalts and the sparse occurrence of an EM2 mantle component in the deep plume sources (Stracke et al., 2003). A slight shift of FOZO-type Jeju mantle toward EM2 composition (Choi et al., 2006) (Fig. 8a and Supplementary Material Appendix C) and possible presence of EM2-like component in the South China Sea mantle (Wang et al., 2013) could be ascribed to ancient subducted sediment recycled together with IOC–MwP composite because subducted young Pacific Ocean sediments does not account for the composition (Fig. 8a).

5.4. East Eurasia backarc tectonics and magmatism

5.4.1. Tectono-magmatic history

Late Cenozoic regional tectonics and magmatism dates back to when Izanagi–Pacific Ridge was subducted at 60–50 Ma (Fig. 14a). The Cenozoic stretching of the continental margin formed the Sea of Japan and the South China Sea at 35–15 Ma (Fig. 14b). The Yellow Sea and the East China Sea basins also formed during this episode, but the basalt were buried by thick sediments. The Cenozoic retreat of the western Pacific trenches was largely responsible for this extension. Backarc extension may also have been exacerbated by the breakoff of the Izanagi Plate from the Pacific Plate along the Izanagi–Pacific Ridge in the mantle transition zone and the subsequent sinking of the Izanagi Plate slab into the lower mantle (Fig. 14b).

The Izanagi-Pacific spreading ridge prior to subduction was underlain by hot MOR mantle that could have been retained even after ridge subduction (Fig. 14a). The retreat of the Japan Trench could be explained by mantle counterflow induced by the Izanagi Plate breakoff (Fig. 14b). The deplited Indian Ocean DMM beneath the Sea of Japan, Sirkhote-Alin, Kurile, and the eastern Philippine Sea (Fig. 13) might have been created by such counterflow above the stagnant Pacific Plate slab from the Indian Ocean MORB mantle beneath the slabs (Fig. 14b). The estimated high $T_m = 1450$ °C for the Sea of Japan backarc basin basalts is due to introduction of the pre-existing hot, deep mantle upwelling through the Izanagi–Pacific Ridge window (Fig. 14b).
During the same time, a deep FOZO mantle source formed transitional seafloor basalt in the South China Sea, reflecting the influence of the Hainan plume following the Indian Ocean DMM intrusion (Fig. 14b). The main stage of marginal basin formation (35–15 Ma) was related to the stretching of the subcontinental lithosphere induced by asthenospheric upwelling (Fig. 14b). Replacement of the uppermost upper mantle occurred by the input of the deep FOZO mantle source through the Hainan and Datong–Baikal plumes. Later, the FOZO mantle extended laterally above the stagnant slab (Fig. 14c).
Because of the high $T_p = 1420 \degree$ C plumes beneath Hainan and Datong–Baikal, part of the stagnant Pacific Plate slab was entrained in the plumes. The slab beneath Shandong Peninsula was surrounded by the plumes, and the leading edge of the Pacific Plate along the plumes was entrained. This resulted in a DMM-type mantle source in these limited areas (Figs. 13 and 14). The overall current upper mantle structure is shown graphically in Fig. 14d.

5.4.2. Hot versus wet plume

Water in the global FOZO-type mantle source is often hydrous, containing 300–800 ppm water (Dixon et al., 2002, 2004; Kimura et al., 2017). Therefore, the deep FOZO beneath East Eurasia can account for most of the magmatic water in the basalts. A hydrous mantle is often discussed as the EM1 source (Kuritani et al., 2011; Wang et al., 2017), but is not necessary because the estimated water content compares with that of the lithospheric pyroxenite from China (Hao et al., 2012) and elsewhere (Bizimis and Peslier, 2015). The hydrous mantle or fluid discussed with the origin of the DMM source mantle (Guo et al., 2016; Wang et al., 2015) is also not essential. The estimated $T_p (= 1320 \degree$ C, $XH_2O = 300$ ppm) for the Jiangsu–Shandong basalts (Fig. 10b) includes the ambient mantle temperature. If more water is contained in the source, the estimated mantle potential temperature should be lower, suggesting no significant water.

The dehydrating Pacific Plate slab released fluids to hydrate the mantle above it. Such metasomatized mantle should be weaker and more buoyant than ambient mantle and could form upwelling diapirs (Kuritani et al., 2011; Richard and Iwamori, 2010; Sakuyama et al., 2013). However, such small-scale upwelling does not explain the hemispheric scale marginal basin formation at 35–15 Ma. Regional upwelling of voluminous hydrated mantle has been proposed as the Big Mantle Wedge (BMW) hypothesis (Zhao et al., 2009). However, the proposed mechanism does not explain where the huge amount of infiltrating mantle came from, as required to compensate the mass deficits caused by slab rollback (Fig. 14).

Intense hydration is important for forming the mantle transition zone, especially given the hydrophilic dominant mantle transition zone mineral, wadsleyite, which can contain up to 2.5 wt% H$_2$O (Pearson et al., 2014). The interpretation of wet low-density mantle may be consistent with the extended low-$V_p$ (Chen et al., 2015; S.-C. Liu et al., 2016) and low-resistivity regions above the stagnant slab (Ichiki et al., 2006). However, we prefer a model of dynamic hot plume upwelling and lateral flow to explain the low-$V_p$ region above the stagnant slab (Fig. 13d) as proposed in the electrical conductivity study (Ichiki et al., 2006) (their second model; see also Figs. 13e and 14c). The mantle transition zone is also not much hydrous globally from the electrical conductivity point of view (Huang et al., 2005; Yoshino et al., 2008). We, however, do not solely negate the role of water from the Pacific Plate slab. Some water can be from the slab by deep dehydration during slab rollback and stagnation (Figs. 14a–c). We leave this question open, and the deep slab dehydration will be discussed elsewhere.

6. Conclusions

A chemical geodynamic examination of Late Cenozoic basalts in East Eurasia revealed that three major mantle sources form the hemispheric scale isotopic structure. East Eurasian basalts are derived from mantle sources with FOZO, DMM, and EM1 isotope characteristics associated with deep rooted plumes, the oceanic crust of the stagnant Pacific Plate slab, and subcontinental lithosphere pyroxenite origins, respectively. The FOZO source basalts occur above and around the Hainan and Datong–Baikal plumes, which slip out from the slab tear of the Pacific Plate and upwell along the leading edge of the slab, respectively. The DMM source basalts are from the Shandong Peninsula where the Pacific Plate slab is surrounded by the two plumes; some DMM-influenced basalts erupted along the leading edge of the slab. Distribution of the EM1 source basalts is limited to the region where thick continental lithosphere underlies. These spatial correlations with regional structure are consistent with our interpretation for mantle sources. Trace element modeling estimated the amount of MORB-sourced pyroxenite from the stagnant slab in the mantle sources: high in DMM, variable in FOZO, and low in EM1. Our model also estimates high $T_p = 1420 \degree$ C in the plume-sourced basalts, while stagnant slab-sourced basalts show lower $T_p = 1320 \degree$ C, close to the ambient mantle temperature and reflecting stagnant slab material entrained in the plume cooling the mantle source. Late Cenozoic (<35 Ma) magmatism may have been induced by the breakoff of the Izanagi Plate slab from the Pacific Plate slab and its sinking from the mantle transition zone into the lower mantle. The deep plume formed by mantle counterflow induced by the sinking Izanagi Plate. Subsequent trench retreat and stagnant slab formation of the Pacific Plate allowed upwelling of the FOZO deep mantle and replaced the depleted Indian Ocean MORB source mantle above the stagnant Pacific Plate slab. Plume emplacement could produce the observed seismic and geochemical structures of the mantle beneath East Eurasia. The estimated water in these mantle sources is 300–600 ppm comparable with that in other deep plume sources. The role of water from the stagnant slab has limited effect on the mantle sources, and its importance has been overestimated in previous studies. Future research is needed to constrain the amount and identify the source of water in the mantle sources.

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