Evolution of late Cenozoic magmatism and the crust–mantle structure in the NE Japan Arc

TAKEYOSHI YOSHIDA1*, JUN-ICHI KIMURA2, RYOICHI YAMADA1, VALERIO ACOCELLA3, HIROSHI SATO4, DAPENG ZHAO5, JUNICHI NAKAJIMA5, AKIRA HASEGAWA5, TOMOMI OKADA5, SATORU HONDA4, MASAHIRO ISHIKAWA6, OKY DICKY ARDIANSYAH PRIMA7, TAKESHI KUDO8, BUNICHIRO SHIBAZAKI9, AKIKO TANAKA10 & TOSHIFUMI IMAIZUMI1

1Institute of Earth Sciences, Graduate School of Science, Tohoku University, Sendai 980-8578, Japan
2Institute for Research on Earth Evolution, Japan Agency for Marine-Earth Science and Technology, Yokosuka 237-0061, Japan
3Dipartimento Scienze Geologiche, Universita Roma TRE, Rome 00146, Italy
4Earthquake Research Institute, University of Tokyo, Tokyo 113-0032, Japan
5Research Center for Prediction of Earthquakes and Volcanic Eruptions, Tohoku University, Sendai 980-8578, Japan
6Graduate School of Environment and Information Sciences, Yokohama National University, Yokohama 240-8501, Japan
7Faculty of Software and Information Science, Iwate Prefectural University, Iwate 020-0193, Japan
8College of Engineering, Chubu University, Kasugai 487-8501, Japan
9International Institute for Seismology and Earthquake Engineering, Building Research Institute, Tsukuba 305-0802, Japan
10Geological Survey of Japan, AIST, Tsukuba 305-8567, Japan

*Corresponding author (e-mail: tyoshida@m.tohoku.ac.jp)

Abstract: We review the evolution of late Cenozoic magmatism in the NE Japan arc, and examine the relationship between the magmatism and the crust–mantle structure. Recent studies reveal secular changes in the mode of magmatic activity, the magma plumbing system, erupted volumes and magmatic composition associated with the evolution of crust–mantle structures related to the tectonic evolution of the arc. The evolution of Cenozoic magmatism in the arc can be divided into three periods: the continental margin (66–21 Ma), the back-arc basin (21–13.5 Ma) and the island-arc period (13.5–0 Ma). Magmatic evolution in the back-arc basin and the island-arc periods appears to be related to the 2D to 3D change in the convection pattern of the mantle wedge related to the asthenosphere upwelling and subsequent cooling of the mantle. Geodynamic changes in the mantle caused back-arc basin basalt eruptions during the back-arc basin opening (basalt phase) followed by crustal heating and re-melting, which generated many felsic plutons and calderas (rhyolite/granite phase) in the early stage of the island-arc period. This was followed by crustal cooling and strong compression, which ensured vent connections and mixing between deeper mafic and shallower felsic magmas, erupting large volumes of Quaternary andesites (andesite phase).

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The NE Japan arc is a typical island arc associated with a cold subduction zone. It is one of the most well-known island arcs on Earth and provides a useful model for our understanding of how subduction zones produce magmas, including andesites. The NE Japan arc is related to the westward...
subduction of the Pacific plate beneath the Eurasian plate at a rate of 8–9 cm a\(^{-1}\) with a relatively shallow dip angle of about 30\(^\circ\) (DeMets 1992). Integrated study of geology, petrology and geophysics is valuable for revealing processes such as the formation of crust–mantle structures, the origin of the magmas and the geodynamic evolution of the subduction zone system (Iwamori & Zhao 2000; Stern 2002; van Keken 2003; Sato et al. 2004; Hasegawa et al. 2005; Yoshida et al. 2005; Kimura & Yoshida 2006; Yoshida 2009).

Secular changes in the late Cenozoic NE Japan magmatic arc are key factors involved in the arc’s evolution, and changes in (1) the mode of magmatic activity, (2) the magma plumbing system, (3) the erupted volume and (4) the magmatic composition have all been investigated in relation to the tectonic history of the arc (Sato 1994; Yoshida et al. 1995; Umeda et al. 1999; Yoshida 2001, 2009; Yamada & Yoshida 2011; Yamada et al. 2012). The evolution of late Cenozoic magmatism in the NE Japan arc can be divided into three major periods – (1) the continental margin (66–21 Ma), (2) the back-arc basin period (21–13.5 Ma) and (3) the island-arc (13.5–0 Ma) periods (Ohguchi & Yoshida 2011; Hasegawa et al. 2005; Yoshida et al. 2005; Umeda et al. 1999; Yoshida 2001, 2009; Yamada & Yoshida 2011; Yamada et al. 2012). The continental margin period is characterized by mafic to felsic subaerial volcanism, with thick ignimbrites in the continental rift systems. The back-arc basin period is characterized by submarine bimodal volcanism under extensional tectonics, and the island-arc period is characterized by felsic caldera-forming events followed by andesitic stratovolcano-forming volcanism under neutral to compressional stress fields (Ohguchi et al. 1989; Sato & Yoshida 1993; Sato 1994; Yoshida 2001). The characteristic magma of subduction zones is andesite, although this is usually associated with basaltic to rhyolitic magmas (Gill 1981). Calc-alkaline andesites are typical, particularly in mature arcs constructed on a thick crust (Kuno 1968, and references therein). The mafic magmas derived from the mantle wedge may be underplated and stagnating around the Moho beneath the thick crust, especially under horizontal compression. Differentiated magmas can stagnate at levels of neutral buoyancy in the crust (Ryan 1987; Takada 1989, and references therein). There are many models for andesite genesis in the subduction zone of Japan, including differentiation by fractional crystallization of mafic magmas (low-K tholeiite from Adatara volcano: Fujinawa 1988), magma mixing/co-mingling between mafic and felsic end-members (low-K andesite: Toya et al. 2005; medium-K andesite: Ban & Yamamoto 2002; Hirotani & Ban 2006; high-K andesite: Inoue & Ban 1996), assimilation and fractional crystallization (DePaolo 1985) or crustal contamination (Kobayashi & Nakamura 2001; Hirotani et al. 2009), re-melting of lower-crustal hydrous mafic rocks (Takahashi 1986; Kimura et al. 2001, 2002) and melting of the mantle wedge or slab (high-magnesian andesite and adakite: Hanyu et al. 2006; Yamamoto & Hoang 2009; Hoang et al. 2009). Possible processes which could produce the felsic end-member of mixed magmas are: the fractional crystallization from mafic end-member magmas (Kanisawa & Yoshida 1989; Miyagi et al. 2012); the melt extraction or re-melting of partially or fully crystallized precursory mafic magmas in the upper to middle crust (Hildreth & Wilson 2007; Hirotani et al. 2009; Miyagi et al. 2012); the partial re-melting of mafic end-member material solidified to become hornblende gabbro or amphibolite (Feeley et al. 1998; Hansen et al. 2002; Toya et al. 2005); and the partial re-melting of crustal basement rocks (Shuto et al. 2006, 2008). All of these crustal processes forming calc-alkaline andesites work effectively in mature arcs on thick crust.

To understand the origin of island-arc andesites, it is very important to know in detail the structure of the magma plumbing system from the upper mantle through the entire crust. Moreover, the observation that andesite-dominated volcanism may occur only at specific stages of the evolution of a subduction zone justifies a comprehensive study of the evolutionary history of the NE Japan arc and the crust–mantle structures throughout its evolution. The purpose of this paper is to review the evolutionary history of late Cenozoic magmatism in the arc, including its thermal and tectonic evolution, and to examine the relationship between magmatism and the crust–mantle structures at each evolutionary stage.

**Tectonic and magmatic evolution in the NE Japan arc**

**Tectonic outline**

The NE Japan arc consists of multiple tectonic segments separated by topographic depressions (Fig. 1a). Horst and graben-like structures developed in the Early to Middle Miocene, simultaneously with the opening of the back-arc basin of the Japan Sea (Okamura et al. 1995; Yoshida et al. 2005; Yamada & Yoshida 2011). Detailed seismological profiles of the arc enable the examination of the deep structures in the island-arc crust. From these profiles, two rift systems, which correspond to the surface topography and the geology, can be recognized (Fig. 1b: Sato et al. 2004); the
Fig. 1. (a) Digital elevation model showing tectonics and the distribution of Quaternary volcanoes in the NE Japan arc (modified from Sato 1994 and Yoshida et al. 2005). (b) Seismic $V_p$ structure across the NE Japan arc (Iwasaki et al. 2001; Nishisaka et al. 2001; Takahashi et al. 2004; Sato et al. 2004). Numbers in (b) correspond to km s$^{-1}$. The two dashed lines in (a) show the western and eastern edges of Quaternary volcanism; the eastern edge is the present volcanic front, and the western edge is the (Quaternary) volcanic rear edge (Yoshida et al. 2005). Tanakura tectonic line (TTL) is the main boundary fault between the basement of NE and SW Honshu. The Hatagawa tectonic line (HTL) divides the basement of NE Honshu into the Abukuma Belt and the Kitakami Belt. Other symbols are explained in the upper-right box. K, Kitakami Mountains (Massif); A, Abukuma Mountains; L, Kitakami River Valley; B, Ou Backbone Range; I, Intermountain Basins; D, Dewa hills; T, Tobishima Basin (Akita–Niigata Basin); O, Oga–Avashima fault zone (Tobishima–Funakawa uplifted zone); M, Mogami Trough; S, Sado Ridge; TGTL, Tonegawa tectonic line; ISTL, Itoigawa Shizuoka tectonic line.
Fig. 2. (a) Horizontal gradient distribution of short-wavelength components of Bouguer gravity anomalies (Kudo et al. 2010; Yamamoto et al. 2011) and the distribution of late Cenozoic calderas (white circles) and Quaternary volcanoes (white triangles) in the NE Honshu arc. Topographic density is assumed to be 2.67 g cm$^{-3}$. The distribution of late Cenozoic large calderas corresponds strongly with the distribution of gravity depressions, and it is likely that the density structure of these calderas is a major cause of the gravity undulations of this region (Kudo et al. 2010). Section X–Y and the rectangle, show the study section from Figure 14a, b, and the study area in Figure 14c, respectively. (b) Distribution of the late Cenozoic large calderas and Quaternary volcanoes superimposed on a map of positive openness which is one of the thematic maps derived from the 50 m grid-size digital elevation models (Yokoyama et al. 1999; Prima et al. 2006). Adapted from Yoshida et al. (2005). QVF, Quaternary volcanic front; VRE, Quaternary volcanic rear edge; TTL, Tanakura tectonic line; HTL, Hatagawa tectonic line. Short lines represent active faults (Active Fault Research Group 1991).
Yamato Basin rift and the Northern Honshu rift. The Yamato Basin rift forms half-graben, bounded by westward-dipping Miocene normal faults. The thickness of the arc crust decreases westward owing to early Miocene crustal stretching, and is related to the opening of the Japan Sea. The Northern Honshu rift is younger (Middle Miocene), and formed a deep sedimentary basin by simple shear. Basalts were erupted along the rift axis associated with a thinning of the upper crust (Yagi et al. 2001; Sato et al. 2004). Very active felsic volcanism, which created more than 80 calderas concentrated along the topographic high of the Ou Backbone range (Ito et al. 1989; Sato & Yoshida 1993; Sato 1994; Yoshida et al. 1999a), occurred at the main stage of the island-arc period in NE Honshu (Fig. 2). During the late Miocene to late Pliocene, the Ou Backbone range was characterized by widespread north–south-trending dextral faults and NE–SW-trending normal faults (Acocella et al. 2008). During the Quaternary, the tectonic setting was characterized by a strong east–west compression (i.e. Sato 1994), and more andesites erupted from along-arc and east–west-aligned stratovolcanoes.

Temporal evolution of the NE Japan magmatic arc

Tectono-magmatic periods and spatio-temporal variation of magmatism. The three major evolutionary periods of the NE Japan arc can be further subdivided into 13 stages (Table 1 and Fig. 3) based on modes of volcanism, tectonic events and the inferred regional stress field (Ohguchi et al. 1989; Sato 1994; Yoshida et al. 1995; Nakajima et al. 2006c; Yoshida 2001, 2009). Figure 3 shows the migration of the position of the volcanic front in the NE Japan arc over the past 60 myr, where the age of each volcanic event is plotted against the distance from the Quaternary volcanic front (QVF). The volcanic front was located deeper in the back-arc in the time period immediately before the back-arc basin spreading, and it had quickly migrated to the trench side at a position approximately 40 km trench-ward of the QVF until the end of Stage 7. Through Stages 8–13 the volcanic front migrated back to the rear-arc and had settled in its present QVF position by Stage 13. It is clear that the rapid trench-ward migration of volcanism was coeval to the deepening of the back-arc basins under strong extension (Fig. 3), and it is possible that the formation of rift graben may have activated the back-arc basin volcanism.

The late Cenozoic NE Japan arc is subdivided into eight tectonic segments (Fig. 4a), based on gravity anomalies, volcanic stratigraphy and the age of volcanic rocks (Yamada & Yoshida 2011). Figure 4b shows the East–West cross-section through the forearc to the back-arc basin of the NE Japan arc, and Figure 5 shows the volcanic stratigraphy over the past 30 myr (Yamada & Yoshida 2011). Continental margin volcanism first occurred at c. 35 Ma in the western continental rift zone (Kano et al. 2007), and volcanism during this stage was dominated by subaerial andesite activity. Back-arc basin volcanism was active in the Yamato Basin between 21 and 17 Ma (Kaneoka et al. 1992), followed by rift volcanism in the Aosawa Rift at 17–15 Ma (Yagi et al. 2001), and in the Kuroko Rift at 15–13.5 Ma (Yamada & Yoshida 2011). The first half of the back-arc basin period was characterized by the effusion of large amounts of seafloor basalt during the opening of the Yamato Basin, whereas the second half was characterized by spatially varying amounts of graben-fill basalt.

Table 1. The 13 stages of the NE Japan arc evolution

| Period | Stage | Characteristics |
|--------|-------|----------------|
| (1) Continental margin period (66–21 Ma) | Stage 1 (66–49 Ma) | Acidic to intermediate igneous activity |
| | Stage 2 (49–35 Ma) | Hiatus with no volcanic activity. |
| | Stage 3 (35–27 Ma) | Alkaline volcanism of within-plate type in the back-arc side |
| | Stage 4 (27–21 Ma) | Calc-alkaline volcanism with alkaline rhyolites in the back-arc side |
| (2) Back-arc basin period (21–13.5 Ma) | Stage 5 (21–17 Ma) | Yamato basin spreading |
| | Stage 6 (17–15 Ma) | Early Northern Honshu rift (Aosawa rift) stage |
| | Stage 7 (15–13.5 Ma) | Later Northern Honshu rift (Kuroko rift) stage |
| (3) Island-arc period (13.5–0 Ma) | Stage 8 (13.5–10 Ma) | Oceanic island arc under neutral stress field |
| | Stage 9 (10–8 Ma) | Oceanic island arc under weak compressional stress field |
| | Stage 10 (8–5.3 Ma) | Early caldera-dominated arc (NE–SW compression) |
| | Stage 11 (5.3–3.5 Ma) | Late caldera-dominated arc (NE–SW compression) |
| | Stage 12 (3.5–1.0 Ma) | Transitional stage to strong compressional stress field |
| | Stage 13 (1.0–0 Ma) | Andesitic stratovolcanoes (east–west strong compression) |
and rhyolite effusions, as part of the bimodal volcanism in the Northern Honshu rift (Yagi et al. 2001; Yamada & Yoshida 2011). Rhyolitic magmas are mainly distributed in the forearc side, which is underlain by the inland continental margin crust (Fig. 4b). This activity in the back-arc basins was closely followed by a volcanic quiescence between 14 and 12 Ma. After the Japan Sea stopped opening at about 13.5 Ma, island-arc volcanism occurred in the eastern rift zones. Volcanism in the island-arc period is mainly characterized by caldera-forming felsic volcanism between 8 and 1.7 Ma (Yoshida et al. 1999a; Yamada & Yoshida 2002, 2011), which was accompanied by subordinate basaltic to andesitic volcanism (Acocella et al. 2008). After 1.7–1.0 Ma, many andesitic stratovolcanoes formed under strong compression (Umeda et al. 1999).

Erupted volume and regional stress field. The erupted magma volume reflects both the magma production rate in the mantle–crust section and the regional stress field (Umeda et al. 1999; Acocella et al. 2008; Yamada & Yoshida 2011). Compiled results (Fig. 6) show that erupted volumes of major volcanism changed from 4600 km$^3$ of basalts (per 1 myr in an arc length of 200 km) in the early back-arc basin period (21–19 Ma) to a volume of 2000–1000 km$^3$ from bimodal submarine volcanism in the late back-arc basin period (19–13.5 Ma), then to 200 km$^3$ from felsic pyroclastic eruptions in the early island-arc period (13.5–3.5 Ma), and finally to 500 km$^3$ of andesites in the latest island-arc period.

Changes in the erupted volume of magma along the volcanic front in the NE Japan arc since 2.0 Ma were estimated by Umeda et al. (1999) (Fig. 6b). Three stages are evident in the distinct types of volcanism, volume of eruptive magma and the locations of eruptive centres (Fig. 6c: Umeda et al. 1999): (1) felsic volcanism associated with large calderas (2.0–1.0 Ma); (2) felsic volcanism terminating at around 1.0 Ma followed by andesitic stratovolcanoes that became the dominant form of volcanism (1.0–0.6 Ma); (3) a continuation of andesite-dominated volcanism with a return of subordinate felsic volcanism. The eruptive rate increased drastically after 0.6 Ma (Fig. 6a, b). The change of eruption style from felsic pyroclastic flows associated with caldera formation, to andesitic stratovolcanism along the volcanic front after 2 Ma, was synchronous with the tectonic change from a
Fig. 4. (a) Tectonic framework (after Yamada & Yoshida 2011), and (b) east-to-west traverse profile (MITI 1985, 1998; Sato & Amano 1991; Yamada & Yoshida 2011) from the back-arc to the volcanic front of the NE Japan arc.
NE–SW-oriented compression to an ENE–WSW-(east–west-) oriented compression (Umeda et al. 1999; Acocella et al. 2008).

Sato (1994) synthesized the regional tectonic stress field, basin development and crustal deformation of the NE Japan arc between the Oligocene and the Quaternary, based on dyke, vein and fault orientation data, and compiled regional geology (Takeuchi 1980; Sato et al. 1982; Tsunakawa 1986; Otsuki 1990; Yamamoto 1991). Dyke orientations are parallel to the horizontal compressive principal stress axis $\sigma_{\text{Hmax}}$ (Anderson 1951; Nakamura 1977; Tsunakawa 1983). Results from veins and outcrop scale faults also highlight the regional tectonic stress field summarized in Figure 7 (Otsuki et al. 1977; Mimura 1979; Sato 1986, 1994; Oishi & Takahashi 1990; Otsuki 1990).

### Temporal evolution of tectonic background

It is probable that the change from Stage 1 to Stage 2 (Fig. 3) was related to the change in the motion of the Pacific plate from NW to WNW at 50 Ma (Eocene plate reorganization: Patriat & Archache 1984; Gordon & Jurdy 1986; Maruyama & Seno 1986). Subsequent continental margin magmatism can be regarded as being due to the westward subduction of the Pacific Plate (e.g. Muller et al. 2008). The continental rifting which occurred deep in the back-arc region (Stages 3–4 in Fig. 3), before the opening of the Yamato Basin, is considered to have been a precursor of the back-arc basin opening (Kano et al. 2007, Nohda 2009).

The late Eocene Monzen formation (Stage 3) contains alkaline to subalkaline within-plate type mafic to felsic volcanic rocks that erupted on land or in shallow water (Kano et al. 2011; Yamada et al. 2012). Sato (1994) showed that an extensional stress field became prevalent during the beginning of Stage 3, and major extension-related normal faulting (NNE to NE directed $\sigma_{\text{Hmax}}$) started at 25–20 Ma (Stage 4). This was followed by the back-arc opening of the Yamato and Aosawa Basins at 21–15 Ma (Stage 5–6). Normal faulting propagated from the present Japan Sea coast at 17–15 Ma (Stage 6) towards the forearc at 15–13.5 Ma (Stage 7), which followed the trench-ward migration of the volcanic field (Fig. 3). From 21 to 15 Ma (Stages 5–6), normal faults under NW- to NWW-directed $\sigma_{\text{Hmax}}$ developed owing to the opening of the Japan Sea and the counter-clockwise rotation of NE Japan (Otofuji & Matsuda 1983, Fig. 5. Late Cenozoic volcanic stratigraphy along a transect from the Yamato Basin to the forearc range (Fig. 4a). Synthesis from radiometric ages of volcanic rocks, foraminiferal and micro-fossil ages in the intercalated sediments, and others (Yamada & Yoshi 2011 for detailed data source). A–H, Middle Miocene tectonic segments shown in Figure 4a.)
The formation of the Japan Sea would have begun in the central region of the rift zone (the Japan Basin and Yamato Basin) and rifting propagated subsequently to the east (Fig. 5). The clockwise rotation of SW Japan beginning in 16 Ma (Stage 6) produced a NW–SE-directed transtensional stress regime in the NE Japan arc at 15–13.5 Ma (Stage 7), as shown in Figure 7. Owing to the crustal stretching associated with pull-apart formation, the NE Japan back-arc subsided rapidly to middle bathyal depths (Sato 1994).

After the Japan Sea finished opening in about 13.5 Ma, a neutral stress regime prevailed between 13.5 and 10 Ma (Stage 8). The neutral to weak compressional stress regime with NE- to ENE-directed $\sigma_{Hmax}$, including both weak tension and compression, can be divided into two phases: 10–8 Ma (Stage 9) and 8–3.5 Ma (Stages 10–11). Lithospheric cooling led to thermal subsidence of the back-arc region at 13.5–10 Ma (Stage 8), and it is likely that the uplift of the axial zone of the volcanic arc was caused by igneous underplating during the back-arc basin period and the related crustal felsic magmatism at 13.5–8 Ma (Stage 9). The increase in the westward motion velocity of the Pacific plate at around 4 Ma (Pollitz 1986) produced stronger compression across the arc at 3.5–1 Ma (Stage 12), reactivating most of the Miocene normal faults, and uplifting the volcanic arc. The strong compressional stress regime since 3.5 Ma (Stage 12–13) is characterized by a $\sigma_3$-directed normal to the trench with a vertical $\sigma_3$ (Sato 1994). The greatest crustal shortening occurred within the volcanic arc at 1–0 Ma (Stage 13) in areas that were stretched the most during the Miocene, which implies tectonic inversion (Sato 1994).

Figure 8 shows the distribution and types of volcanoes in the NE Japan arc from the late Miocene to the Quaternary (Yoshida et al. 1999a; Acocella et al. 2008). The late Miocene–Pliocene and the Quaternary tectonic setting, estimated from fault analysis, are also shown in Figure 8 (adapted from Acocella et al. 2008). The Miocene–Pliocene deformation pattern is consistent with a NE–SW maximum compression, resulting from the NE–SW collision with the Kuril sliver, and the Quaternary deformation pattern is consistent with a WNW–ESE (east–west) maximum compression controlled by the motion of the Pacific plate (Acocella et al. 2008).

**Crust–mantle structures and associated control of arc magmatism**

Magmatism in the NE Japan arc is also controlled by the crust–mantle structures beneath the arc. Recent developments in seismological studies of the NE Japan arc have revealed heterogeneities, both in the mantle and in the arc crust. In this section, we summarize the heterogeneous structure of the arc crust and the underlying mantle, and then correlate this with the late Cenozoic magmatism.

**Basement structures**

**Basement terranes.** The distribution of volcanoes and the spatial variation in the composition of magma are closely related to the basement structures which exert a strong influence on the magma's composition (Kersting et al. 1996; Kimura & Yoshida 2006). Pre-Palaeogene basement rocks beneath the NE Japan arc (Geological Survey of Japan 1977) consist of Carboniferous to Jurassic metamorphic rocks of the Mino Belt, Cretaceous sedimentary rocks and granitoids of the Abukuma Belt, and an Ordovician to Cretaceous complex of the Kitakami Belt (Fig. 9a, b). The Mino and Abukuma Belts are in contact along the Tanakura Tectonic Line (TTL in Fig. 9a, b), which is the main boundary fault between the basement of NE and SW Honshu (Isozaki 1996). Another major fault, the Hatagawa Tectonic Line (HTL in Fig. 9a, b), separates the Abukuma and Kitakami Belts subparallel to the TTL.

The late Cenozoic basalts are isotopically enriched to the west of the TTL relative to those erupted to the east (Sato et al. 2007). The late Cenozoic rhyolites to the west of the TTL are generally richer in alkalies than the associated mafic rocks, and those to the east are usually compositionally similar to the associated mafic rocks (Shuto et al. 2008; Yamada et al. 2012).

**Basement granitoids.** Granitoids of various ages (Cretaceous to Palaeogene) intrude into the basement terranes. These granitoids have been classified based on Sr–Nd isotope geochemistry into the Kitakami, North, Transitional, South and Sado zone granitoids (Fig. 9a, c; Kagami et al. 2000; Kagami 2005). The Kitakami zone granitoids are the most depleted isotopically, and the South zone granitoids are the most enriched. The distribution of the geochemical zoning is subparallel to the major faults in the basements, even though the zone boundaries do not follow the tectonic lines exactly (Fig. 9a). Provided that these granitoids are of lower crustal melt in origin, the spatial correlations between granitoid chemistry and the basement terranes suggest a coupled upper and lower crust (Kagami et al. 2000; Kagami 2005). The enriched North and South zones are extensions of the isotopic zones identified in the SW Honshu arc (Kagami 2005), and the Kitakami and Sado zones, together with the Quaternary rear arc (QRA), are plotted on the mantle array (Fig. 9c).
Fig. 6. (a) Temporal changes in the volume (km³; per 1 Ma, for 200 km of arc length along axis) and composition of erupted magma from the back-arc to the island-arc periods (modified from Yamada & Yoshida 2011). (b) Temporal changes in eruptive magma volume (km³) for stratovolcanoes (upper) and large calderas (lower) during the last 2.0 Ma along the volcanic front in the NE Japan arc. (c) Distribution of eruptive centres for last 2 Ma. (b, c) Modified from Umeda et al. (1999).
Fig. 6. Continued.
Fig. 7. Regional stress field in NE Japan since Stage 3 (modified from Sato 1994). QVF, VRE, TTL and HTL as in Figure 2.
Fig. 8. (a) Distribution and types of volcanoes in the NE Japan arc from late Miocene to Quaternary (adapted from Yoshida et al. 1999a). (1) Hatched area represents emergent areas of the NE Japan arc in the specified time period. (2) Orientation of inferred regional maximum compression, based on direction of dyke swarms and other data (Kano et al. 1991; Sato 1994). (3) Calderas refer to the relative time periods. (4) Quaternary stratovolcanoes. (b) The late Miocene–Pliocene structural data: (1) relative abundance of the four types of fractures; (2) schematic diagram showing the mean trend of the four types of fracture, and the mutual relationships. The Mio–Pliocene deformation pattern is consistent with a NE–SW maximum compression (after Acocella et al. 2008). (c) The Quaternary structural data: (1) relative abundance of the four types of fractures; (2) schematic diagram showing the mean trend of the four types of fracture, and the mutual relationships. The Quaternary deformation pattern is consistent with a WNW–ESE (east–west) maximum compression (after Acocella et al. 2008).
Fig. 9. (a) Regional isotope zoning of basement granitoids (Kitakami, North, Transitional, South, and Sado zones) by Kagami et al. (2000) and Kagami (2005), and tectonic boundaries of Kitakami (KT), Abukuma (AB), Mino (MN) and Shimanto (SMT) belts. These basement structures are oblique to the present NE Japan trench-arc system. (b) Distribution of Quaternary volcanoes (QVF-I, II, III, and QRA) in the NE Japan arc and basement tectonic boundaries (KT, AB, MN and SMT) (modified from Kimura & Yoshida 2006). (c) Nd–Sr isotopic compositions of basement granitoids (Kagami et al. 2000; Kagami 2005) and Quaternary lavas from the volcanic front and rear arc. Patterned fields show the isotopic compositions of the basement granitoids. Spatial correlations in Nd–Sr isotope composition are evident between the basement granitoids and the lavas for the frontal-arc volcanoes (Kimura & Yoshida 2006), although the rear arc lavas and the Kitakami and the Sado zone granitoids show a more depleted trend on mantle array than the frontal arc volcanics. BE, Bulk Earth.
Crustal structure

Lower crustal rocks beneath the NE Japan arc. Nishimoto et al. (2005) compared the \( V_p \) measurements for each xenolith from the Ichino-megata volcano with the \( V_p \) profile of the NE Japan arc given by Iwasaki et al. (2001). The rocks used in their study cover nearly the entire range of lithological variations of the Ichino-megata xenoliths, and are considered as representative rock samples of the lower crust and upper mantle of the back arc side of the NE Japan arc. The relatively low \( V_p \) (6.6–7.0 km s\(^{-1}\)) in the 15 km-thick lower crustal layer of the NE Japan arc is due to the high abundance of hornblende and magnetite in gabbros and amphibolites with a basic to ultrabasic composition (Fig. 10). Takahashi (1986) suggested that the thick lower crust of the NE Japan arc is composed of hornblende (+ pyroxene) gabbro and/or amphibolite. In addition, Sakuyama (1983), Fukuyama (1985) and Yoshida et al. (1997) suggested that the Ichino-megata basic to ultrabasic lower crustal xenoliths are highly fractionated cumulates derived from calc-alkaline magmas. Kanisawa & Yoshida (1989) showed that cumulates of low-K dacites from a Quaternary frontal volcano have the composition of hornblende gabbros (S in Fig. 10). The composition of hornblende gabbros and amphibolites of the Ichino-megata xenolith is plotted on the trend lines of the Quaternary volcanic rocks in Figure 10 Yoshida et al. (1997) suggested that these basic to ultrabasic cumulates comprise the lower crust of the NE Japan arc. Some amphibolite xenoliths have very similar \(^{87}\text{Sr}/^{86}\text{Sr} \) ratios (0.7030–0.7032) to the host magmas (0.7030–0.7033; Fig. 11; Zashu et al. 1980) and a similar bulk-rock chemistry to the hornblende gabbro xenoliths. It is therefore likely that these amphibolites are originally derived from the late Cenozoic calc-alkaline magmas. Gabbroic xenoliths from Ichino-megata volcano have higher \(^{87}\text{Sr}/^{86}\text{Sr} \) ratios (0.7048–0.7035) than the host magmas (Zashu et al. 1980).

![Fig. 10. Total alkali–SiO\(_2\) diagram for the Quaternary volcanic rocks of the NE Japan arc and Ichino-megata lower crustal to mantle xenolith. Diagram includes data from Kuno (1967), Aoki (1971), Aoki & Shiba (1973), Fukuyama (1985), Aoki & Yoshida (1986), Nishimoto et al. (2005, 2008) for Ichino-megata xenolith, Yamamoto (1984) for Oshima–Oshima volcano of the Chokai volcanic zone, Hayashi et al. (1991) for Kamu volcano of the Chokai volcanic zone, Nakagawa (1983) for Moriyoshi volcano of the Moriyoshi volcanic zone, Yoshida et al. (1983) for Hachimantai volcano of the Sekiro volcanic zone, Togashi (1977) for Osore volcano of the Aoso–Osore volcanic zone and Kanisawa & Yoshida (1989) for Adachi volcano of the Aoso–Osore volcanic zone. S, B and L denote calculated cumulate (S), bulk rock (B), and liquid composition (L) of the low-K calc-alkaline dacite of Adachi volcano (Kanisawa & Yoshida 1989). Abbreviations 2–4 show the most primitive basalt compositions of Iwate volcano (2: Kuritani et al. 2013), Funagata volcano (3: Kimura & Yoshida 2006) and Sanno-megata volcano (4: Yoshinaga & Nakagawa 1999). Estimated primary magma composition from Iwate primitive basalt (2) is shown in the figure as 1 (Kuritani et al. 2013). The compositional relation between lower crustal xenoliths and the Quaternary volcanics shows that hornblende gabbro to hornblendeite and amphibolite from the lower crust of the NE Japan arc are basic to ultrabasic cumulates or restites of hydrous calc-alkaline magma. There are few mafic xenoliths directly solidified from basaltic magma.](image-url)
et al. 1980). Mantle peridotites from Ichino-megata volcano also have a wide range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7030–0.7053: Zashu et al. 1980). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of gabbro and peridotite cover the range of isotopic composition from the late Cenozoic continental margin to the island-arc basalts from the back-arc to the frontal region in the NE Japan arc (Fig. 11).

Lower crustal heterogeneity. Nishimoto et al. (2008) examined the $dV_p$ and $dV_s$ values from the lower crust of NE Honshu on the $V_p – V_s – V_p/V_s$ deviation diagram, in the temperature range of 500–700°C at 0.8 GPa, and compared the data with those of typical minerals and rocks (Fig. 12c). It was concluded that: (1) the seismically high-$V_p$ and -$V_s$ regions beneath the Tobishima Basin (T in Fig. 1) consist of hornblende pyroxene gabbro; (2) hornblende gabbro is the predominant rock type beneath the Dewa Hills (D in Fig. 1) and the Ou Backbone range (B in Fig. 1); (3) the low-$V_p$ and high-$V_s$ regions beneath the Kitakami Mountains (K in Fig. 1) consist of quartz-plagioclase-bearing rocks with intermediate compositions (granodiorite to diorite).

Tomographic images of $V_p$ and $V_s$ at 25 km depth beneath the eastern margin of the Japan Sea to NE Japan (Zhao et al. 2011) are shown in Figure 12a & b, where the nine velocity anomalies are recognized and labelled A–I. Figure 12c plots the observed average $dV_p$ and $dV_s$ data at a depth of 25 km (Zhao et al. 2011). Estimated lithologies are: (A) (high $V_p$, high $V_s$, low $V_p/V_s$) as peridotites;
Fig. 12. (a, b) P-wave (a) and S-wave (b) velocity images at depths of 25 km in the NE Japan back-arc region (after Zhao et al., 2011). Red and blue colours denote low and high velocities, respectively. The velocity perturbation scales are shown at the bottom. The nine velocity anomalies discussed in the text are labelled (A–I) in (a) and (b). QVF, VRE, TTL and HTL are the same as in Figure 2. (c) The nine labels of velocity anomalies are plotted in the ‘\(V_p - V_s/V_p\) deviation diagram’, together with \(dV_p\) and \(dV_s\) values of the various minerals and rocks at 0.8 GPa (Nishimoto et al., 2008). Anomaly A, peridotite; B, pyroxene gabbro; C, pyroxene hornblende gabbro – amphibolite; D, gabbro-norite to amphibolite; E, hornblende pyroxene gabbro to amphibolite; F, hornblende gabbro to amphibolite; G, partially molten lower crust; H, granodiorite; and I, granite.
Fig. 13. (a) Vertical cross-section of $S$-wave velocity perturbation along the line shown in the insert map (after Nakajima et al. 2001). White circles and black lines show microearthquakes and seismic velocity discontinuities, respectively. The black bar and reverse triangle at the top denote the land area and the location of the volcanic front (VF), respectively. Red triangles in the insert map show Quaternary volcanoes. The rectangle indicates the area shown in (b) and (c). (b, c) Magma plumbing systems in the forearc region of the NE Japan arc superimposed on the vertical cross section of $S$-wave (b), and $P$-wave (c) velocity perturbation (after Nakajima et al. 2006a), respectively (modified from Yoshida 2009). The locality of these sections is shown in Figure 14c along the section east–west. NRF denotes the Nagamachi–Rifu fault. Red and blue colours denote low- and high-velocity perturbation. The velocity perturbation (%) scale is shown at the bottom of (b). Crosses indicate shallow microearthquakes determined by Yoshimoto et al. (2000).
(B) (high \(V_p\), high \(V_s\), medium \(V_p/V_s\)) as hornblende pyroxene gabbro; (C) (high \(V_p\), medium \(V_s\), high \(V_p/V_s\)) as hornblende gabbro to amphibolite; (D) (medium \(V_p\), high \(V_s\), low \(V_p/V_s\)) as gabbro–norite to amphibolite; (E) (medium \(V_p\), medium \(V_s\), medium \(V_p/V_s\)) as pyroxene hornblende gabbro to amphibolite; (F) (medium \(V_p\), low \(V_s\), high \(V_p/V_s\)) as hornblende gabbro to hornblende–gabbro–norite; (G) (medium \(V_p\), low \(V_s\), high \(V_p/V_s\)) as partially molten lower crust; (H) (low \(V_p\), medium \(V_s\), low \(V_p/V_s\)) as granodiorite; and (I) (low \(V_p\), low \(V_s\), low \(V_p/V_s\)) as granite.

It is likely that hornblende gabbro and amphibolite (anomalies C, E and F in Fig. 12) are the dominant lower crustal rocks, and that they are distributed along the central axis of the NE Japan arc. In the rifted back-arc basin, hornblende pyroxene gabbro to gabbro–norite and/or pyroxene amphibolite (anomalies B and D) are dominant (Nishimoto et al. 2008). The \(dV_p\) and \(dV_s\) values of the rifted continental crust (Tamaki et al. 1992: anomaly I located at the eastern margin of the Japan Sea to the west of TTL) suggest that these crustal blocks consist mainly of granitic rocks. These are different to those of the granodioritic lower crust (e.g. anomaly H) and to the tonalitic lower crust of the Kitakami Mountainlands (Nishimoto et al. 2008). Present results show that there are regional variations of pre-Palaeanogeic lower crustal plutonic rocks in the NE Japan arc, from the northeastern forearc tonalite to the southwestern back-arc granitoid with transitional granodiorite. Lower crustal pyroxene–hornblende to hornblende gabbro in the central axis of the NE Japan arc, and hornblende pyroxene gabbro in the rifted back-arc basin, might be emplaced in the Cenozoic to the pre-Palaeanogeic rifted continental margin basement which consists of tonalite–granodiorite–granite.

Crustal structure and magma plumbing system. Figure 13b, c shows high-resolution tomographic images of the forearc region near Sendai in NE Japan (rectangle in Fig. 13a; Nakajima et al. 2006a). Nakajima et al. (2006a) explained the complex velocity structure using \(V_p\), \(V_s\) and Poisson’s ratio. Figure 13d, e plots the \(V_p\) and \(V_s\) data for the numbered sections 1–10 in Fig. 13c. Estimated lithologies based on \(V_p\) and \(V_s\) data, using the method of Nishimoto et al. (2008) and geological data (Sato et al. 2002), are: (1) pre-Palaeanogeic granitic rocks; (2) Cenozoic sediments; (3) late Miocene caldera-fill pyroclastic rocks to shallow fluid-rich felsic pluton (Sato et al. 2002); (4) the felsic part of the Pliocene magma reservoir; (5) pre-Palaeanogeic basement rocks; (6) middle-crustal mafic plutonic rocks; (7) diorite to hornblende pyroxene gabbro/ pyroxene amphibolite; (8) hornblende gabbro to partially molten lower crust; (9) Miocene hornblende gabbro; and (10) the mafic part of the Pliocene magma reservoir/hornblende gabbro.

Figure 14a, b shows the modelled density structure along the across-arc profile (X–Y line in Fig. 2a). Relatively denser materials are obtained in the area beneath the eastern rift border zone of the Northern Honshu rift (shown as the ‘rift border zone’ in Fig. 14c), which is located at the easternmost part of the Kitakami river valley (Fig. 14a). This high-density body corresponds to the Miocene hornblende gabbro (Section 9 in Fig. 13c), and it is possible that this is a mafic dyke complex which intruded into the eastern rift border zone.

The island-arc crust of the NE Japan arc is deformed under compressional stress in the direction of plate motion. The Ou Backbone range and the eastern rift border zone of the Northern Honshu rift (Fig. 14c) are two parallel seismically active zones which are NS elongated, and form the focused deformation zones (Miura et al. 2004). Okada et al. (2010) examined the deep structure of the Ou Backbone range and, by performing regional-scale seismic tomography, found that the distinct seismic low-velocity regions are continuously distributed from the mantle wedge to the middle crust just below the Ou Backbone range.
(Fig. 14c). On this basis, the low-velocity zone beneath the Ou Backbone range could be interpreted as a region of partial melting, which would suggest that crustal fluids separating and upwelling from deeper depths are closely related to the active seismic zones. Around the Onikobe caldera in the middle of the Ou Backbone range, the distribution of seismic low-velocity areas, areas of high seismicity and earthquake focal mechanisms are closely related to the distribution of the late Cenozoic calderas (Onodera et al. 1998; Umino et al. 1998; Nakajima & Hasegawa 2003b; Yoshida et al. 2005). There are three levels of low-velocity bodies in the Ou Backbone range: the lower crust, middle crust and upper crust (Fig. 14d), which presumably reflect the partially molten lower crust, the middle crustal magma reservoirs and the caldera-fill pyroclastic rocks with occasionally central plutonic bodies, respectively.

Figure 15a shows the geotherms (Kushiro 1987; Tatsumi et al. 1994) and rock types in the crust of the NE Japan arc adapted from Yoshida (2001). The solids of water-saturated granitic magma crosses the geotherm at a depth of 15–20 km. It is possible that the water-saturated felsic magma under the volcanic area of the Ou Backbone range could be stable in the middle to lower crustal conditions of amphibolite to hydrous granulite facies in the NE Japan arc. The brittle–ductile transition (Fig. 15b), which strongly affects the emplacement of igneous intrusions (Aizawa et al. 2006), is of a shallow depth near the Conrad discontinuity beneath the Ou Backbone range and the Dewa Hill, but is deeper than the Moho beneath the eastern margin of the Japan Sea.

Figure 14c shows the two low-velocity belts at depths of 24 km. As described above, the eastern low-velocity belt could be interpreted as a remnant of the middle Miocene magma plumbing system, which formed at the eastern rift border zone of the Northern Honshu rift and marked the volcanic front (RBL, rift border line in Fig. 14c) at that time. The western low-velocity belt, beneath the Ou Backbone range, reflects the present magma plumbing system from which many felsic caldera-forming magmas and stratovolcano-forming andesitic magmas have been derived. The high-resolution tomographic image around Sendai (Fig. 13b, c) confirms the existence of low P- and low S-wave velocity bodies at depth of 3–7 and 10–15 km beneath the late Miocene and Pliocene caldera complexes (Nakajima et al. 2006a), which are interpreted being still-hot, solidified pluton with a fluid-saturated top (Yoshida 2001; Sato et al. 2002; Nakajima et al. 2006a). The chemical composition of pumice from the late Miocene and Pliocene calderas indicates differing equilibration depths, as estimated by the An–Qz–Ab + Or–H2O system (Luth et al. 1964). The estimated depths correlate well with the depth range of the low-Vp and low-Vs bodies observed in the tomographic images (Fig. 13b, c). Some Pliocene calderas have deeper magma reservoirs (10–15 km) than those of the Late Miocene (3–7 km).

**Mantle structure**

Seismic structure of the mantle wedge. Since the early 1990s, investigations using seismic tomography images of the sub-arc mantle wedge have been
Fig. 15. (a) Geotherms (Kushiro 1987; Tatsumi et al. 1994) and rock types in the crust of the NE Japan arc, solidus of dry and wet (water-saturated) peridotite and basalt, and wet granite solidus with Qz (quartz)-in line (Wyllie 1971; Robertson & Willie 1971); adapted from Yoshida (2001). Plus signs connected by tie-lines show each temperature range estimated from: the petrologic studies of Megata xenoliths (Arai 1980); the thermodynamical studies of the Onikobe hydrothermal systems of solidified plutons (Yamada 1988); and the velocity perturbations of P-waves beneath volcanic and non-volcanic areas of the Ou Backbone Range (Hasegawa et al. 2000) in the NE Japan arc, respectively. Dotted fields show pressure and temperature conditions of the representative metamorphic facies. The estimated geotherm from Ichino-megata xenoliths is consistent with the geotherm at the present Ou Backbone Range and early Miocene back-arc rift axis, and is higher than the present back-arc geotherm (Tatsumi et al. 1994). (b) Strength profiles for the crust and upper mantle beneath: the Ou Backbone Range; the Dewa Hill; the forearc region; and the eastern margin of the Japan Sea (after Shibazaki et al. 2008). Shibazaki et al. (2008) assigned the two layers of crust (upper quartz diorite and lower wet diabase) and the upper mantle of wet olivine for modelling. Hot regions are set beneath the Ou Backbone Range and Dewa Hill based on the geothermal gradient beneath NE Japan (Yano et al. 1999; Tanaka et al. 2004).
performed (e.g. Hasegawa et al. 1991, 2000, 2005; Hasegawa & Nakajima 2004; Nakajima et al. 2001, 2005; Zhao et al. 1990, 1992, 2009, 2012, and references therein). The dense seismic network in NE Japan enables the world’s finest images of these structures. The essential structure of the mantle wedge beneath the NE Japan arc consists of a low-V\textsubscript{p}/V\textsubscript{s} structure lying in the middle of the mantle wedge, subparallel to the down-going Pacific Plate slab (Fig. 11a: Sato & Hasegawa 1996; Nakajima et al. 2001, 2005; Zhao et al. 2012).

The temperature (Nakajima & Hasegawa 2003a) and porosity (Nakajima et al. 2005) of the low-velocity mantle wedge have also been investigated, and a high-temperature wedge core with the presence of melts (Sato & Hasegawa 1996; Nakajima et al. 2005) has been assumed (Fig. 11a). The top depth of the low-velocity mantle reaches immediately below the arc Moho, and low-frequency earthquakes occur in the surrounding areas (Fig. 11b: Xia et al. 2007). This further suggests melt coalescence and migration from the partially molten mantle wedge (Hasegawa & Nakajima 2004). The top becomes deeper towards the rear-arc side and the low-velocity region continues beneath the Japan Sea (Zhao et al. 2012), although the large anomaly appears to diminish off the Japan Sea coast, across the Quaternary volcanic arc (Fig. 11). The low-velocity bodies show heterogeneities in the mantle wedge, which is thicker with a strong low-velocity amplitude beneath the Quaternary volcanic centres, and thinner and faster beneath non-volcanic areas (Hasegawa & Nakajima 2004), although a continuous low velocity region is imaged just below the Moho depth along the entire volcanic front.

Seismic structure of the slab. Using the tomographic image based on the precise analyses of the seismic locations, the dehydration and re-hydration profiles of the down-going slab and overriding forearc mantle wedge have been estimated (Kita et al. 2006; Tsuji et al. 2008). The down-going Pacific Plate slab has a relatively high V\textsubscript{p}/V\textsubscript{s} at a depth of about 80 km. In contrast, the overriding forearc peridotite has a lower V\textsubscript{p}/V\textsubscript{s} than that expected from serpentinite. The results led Tsuji et al. (2008) to conclude that the slab does not dehydrate significantly until a depth of c. 80 km depth, and that slab dehydration is largely completed between depths of 70 and 90 km. Kita et al. (2006) and Hasegawa et al. (2007) also examined slab seismicity with precise re-location of the foci. Their results support the conclusion of Tsuji et al. (2008), and indicate that the deepening of the top downdip seismicity surface from the slab–mantle interface, correlates with the dehydration front of the metamorphic slab that is controlled by the inverted thermal gradient of the slab (Fig. 11c).

Kawakatsu & Watada (2007) examined the receiver function structure of the slab and the mantle beneath NE Japan. They also proposed that the slab dehydrated at around 80 km deep and that the released hydrous fluids migrated upwards into the mantle wedge to form serpentinite. The hydrated serpentinite layer reaches a depth between 120 and 150 km. Together with the geodynamic model (e.g. van Keken 2003), the seismic observations beneath the NE Japan arc provide constraints to the slab–mantle thermal structure and origin of the fluids released from the slab. The finest seismic images of the mantle wedge and the slab provide strong insights into the origin of subduction volcanisms.

Along-arc mantle heterogeneity

Attenuation tomography reveals the low-attenuation (high-Q) subducting Pacific slab and the high-attenuation (low-Q) anomalies in the crust and mantle wedge (Umino & Hasegawa 1984; Tsumura et al. 2000). The location and geometry of low-Q zones in the mantle wedge agree with those of low-V anomalies imaged by tomography at a depth of the uppermost mantle, although they cross with the NE Japan arc at a depth of 60–80 km (Fig. 16d: Yoshida et al. 1999b; Tsumura et al. 2000). Tsumura et al. (2000) have shown that the distribution of the mantle wedge low-Q zones is different in the southwestern part compared with the northeastern part of NE Honshu. In the southwestern part (west of the TTL), low-Q zones are located beneath the active volcanoes or the volcanic front, and are continuously distributed from the crust to the mantle wedge, deepening westward. In contrast, several low-Q zones exist in the northeastern part (east of the TTL), and appear isolated from each other. They are located just beneath the active volcanoes at shallower depths, and deepen to the west of the volcanic front (Tsumura et al. 2000). Q-Values in the mantle wedge vary not only across the arc, but also along the arc (Tsumura et al. 2000). The spatial variation of Q-values is consistent with the basement structure bounded by the TTL, and suggests that the TTL has a deep mantle root (Yoshida et al. 1999b, 2005). This mantle heterogeneity along the TTL, oblique to the present volcanic front, extends to the Japan Sea side, as shown by dV\textsubscript{p} distributions in Figure 16b, c (Zhao et al. 2012). This implies that the TTL extends to the Japan Sea side with a deep mantle root.

The temperature structure in the mantle wedge at a depth of 40 km inferred from seismic wave attenuation (Fig. 16e; Nakajima & Hasegawa 2003a) corresponds to the distribution of Quaternary
volcanoes. However, the temperature structure at a depth of 60–80 km (Fig. 16f) does not correspond with the distribution of Quaternary volcanoes at the surface (Yoshida et al. 1999b; Tsumura et al. 2000; Nakajima & Hasegawa 2003a; Yoshida et al. 2003). The estimated mantle temperature is higher on the SW Honshu side than on the NE Honshu side, divided by the TTL. If the temperature distribution in the mantle wedge is constrained only by the geodynamic models (van Keken 2003; Peacock 2003), the observed spatial variation of $Q$-values crossing the volcanic front requires explanation using other factors, including that of the heterogeneous distribution of $H_2O$ (Karato 2003) that borders the TTL. Yoshida et al. (1999b) suggested that the counter-clockwise rotation of the continental hydrous lithosphere beneath SW Honshu, at the back-arc opening of the Japan Sea, formed this mantle heterogeneity. A high-temperature or a hydrous mantle wedge could also result in low viscosities (Honda & Yoshida 2005a).

**Fig. 16.** (a–c) Map views of $P$-wave tomography at depths of (a) 40, (b) 60 and (c) 80 km. Black triangles represent the distribution of Quaternary volcanoes (after Zhao et al. 2012). (d) Seismic $Q$ structure at a depth of 60 km (Tsumura et al. 2000; Yoshida et al. 1999b, 2005). (e, f) Relations between distribution of volcanoes and mantle temperature at depths of (e) 40 and (f) 80 km, in the NE Japan arc (Nakajima & Hasegawa 2003a; Yoshida et al. 2005). The estimated mantle temperature is higher on the SW Honshu side than on the NE Honshu side, divided by the TTL. These mantle structures are oblique to the present volcanic front and extend to the Japan Sea side, as shown by $dV_p$ distribution in (b) and (c). TTL and HTL are the same as in Figure 1.
similar to the low-\(Q\) zone in the southwestern part of the NE Japan arc (west of the TTL). The high-temperature anomaly obtained was almost slab-parallel, larger in the shallow part (volcanic front side) and weaker in the lower part (back-arc side). This result is consistent with the distribution of the Quaternary volcanoes in the southwestern part of the NE Japan arc (west of the TTL).

**Heterogeneity of seismic velocity anisotropy.** Shear-wave splitting analyses reveal that the low-\(V\) and low-\(Q\) zones in the mantle wedge show strong seismic anisotropies (Okada et al. 1995; Nakajima & Hasegawa 2004, Nakajima et al. 2006b). Anisotropic bodies are revealed in the crust, mantle wedge and the subducting Pacific slab (Hall et al. 2000; Nakajima & Hasegawa 2004; Nakajima et al. 2006b; Wang & Zhao 2008; Huang et al. 2010). Nakajima & Hasegawa (2004) have clearly shown that the azimuth in the forearc region is trench-parallel, whereas in the back-arc region the azimuth is trench-normal (Fig. 17b). The trench-normal orientation in the back-arc side might originate from the orientation of olivine by mantle wedge corner flow, associated with slab subduction (Hall et al. 2000). The trench parallel orientation in the forearc mantle can be explained by the preferred orientation of olivine under wet conditions (Katayama et al. 2004), although Wang & Zhao (2008) suggested that the trench-parallel anisotropy beneath the volcanic-front area in the mantle wedge results from the complex 3D flow in the

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**Fig. 17.** (a) \(P\)-wave anisotropic tomographic images at a depth of 40 km under the eastern margin of the Japan Sea (modified from Huang et al. 2010). Red and blue colours denote low and high velocities, respectively, and velocity perturbation scales are shown at the bottom. The azimuth and length of bars represent the fast-velocity direction and the anisotropic amplitude, respectively. Black triangles are the Quaternary volcanoes. The fast direction of the seismic velocity anisotropy is trench-normal (\(\gamma\)-direction) on the NE Honshu side, but is trench-oblique (\(z\)-direction) in the back-arc side of the SW Honshu side of the TTL. QVF, VRE, TTL and HTL are as in Figure 2. (b) Distribution of the fast \(S\)-wave polarization directions in the mantle wedge (modified from Nakajima & Hasegawa 2004; Nakajima et al. 2006b). Azimuth and length of each bar correspond to the average fast polarization direction and delay time at each station, respectively. The delay-time scale is shown in the upper-left corner. The pattern of their seismic anisotropy in the back-arc side is consistent with the results of Huang et al. (2010), and the azimuth in the forearc region is trench-parallel (\(x\)-direction).
Fig. 18. (a) Regional distribution of $K_{60}$ ($K_2O\%$ at 60% $SiO_2$) of the Quaternary volcanic rocks (modified from Yoshida et al. 1999b, 2005). The size of the circle is correlated to the range of the value in $K_{60}$ and plotted on the location of each Quaternary volcano. The $K_{60}$ values correlate with the boundaries of the four Quaternary volcanic zones of...
mantle wedge, owing to oblique convergence or slab rollback (Hall et al. 2000; Kincaid & Griffiths 2003).

Huang et al. (2011) determined high-resolution P-wave isotropic and anisotropic velocity structures under the Japan Sea off NE Japan, using 175 425 high-precision P-wave arrival times from 2833 local earthquakes relocated with SP depth phase. Their results show that strong velocity variations exist in the crust and the uppermost mantle under the eastern margin of the Japan Sea. P-Wave azimuthal anisotropy is complex under the Japan Sea, which could also indicate the existence of a complex lithospheric structure. On the NE Honshu side, the fast direction of the seismic velocity anisotropy in the mantle wedge is trench-normal, whereas it is trench-oblique in the back-arc side of the SW Honshu side of the TTL (Fig. 17a). The east–west-trending trench-normal anisotropy shown in the NE Honshu back-arc is related to the recent slab-driven corner flow in the NE Japan mantle wedge (Hall et al. 2000). The NW–SE trend is parallel to the 20–15 Ma dyke direction (Fig. 7), which reveals the direction of σ1max at that time. The NW–SE trend is also normal to the Yamato Basin axis, so this direction is consistent with the mantle wedge corner flow during the Yamato Basin opening. It is possible that the NW–SE-trending seismic velocity anisotropy beneath SW Honshu formed as trench-normal anisotropy during the Yamato Basin opening at 21–15 Ma, and this could still be maintained because of the weak mutual rotation and weak magmatism in the back-arc region to the west of the TTL after the back-arc opening.

**Genesis of Quaternary magmatism in the NE Japan arc**

Figure 18a shows the regional distribution of the normalized K$_2$O% at 60% SiO$_2$ ($K_{60}$) of Quaternary volcanic rocks (Yoshida et al. 1999b). The composition of Quaternary volcanic rocks changes across and along the arc. Distribution of $K_{60}$ generally increases from the volcanic front to the back-arc side, and the value of $K_{60}$ is higher in SW Honshu than in NE Honshu at the volcanic front. Typical island-arc volcanism, with clear across-arc chemical variations in the Quaternary volcanic rocks, characterizes NE Honshu to the east of the TTL. In contrast, the pattern of across-arc chemical variations is not so simple in SW Honshu to the west of the TTL, and the distribution of the Quaternary volcanoes is narrow and limited in the region near the volcanic front (Fig. 18a).

The distribution of Quaternary volcanoes is closely related to the distribution of low-V$_s$ regions of the mantle wedge (Fig. 18b), which are interpreted as partial melting zones (Hasegawa & Nakajima 2004). This close relation suggests that the Quaternary magmatic plumbing system is directly connected from the upper mantle to the crust (Hasegawa & Nakajima 2004; Hasegawa et al. 2005; Acocella et al. 2008), and it is clear that the distribution pattern of the inclined low-V$_s$ zones in the mantle wedge forms branches and a trunk (Fig. 18b: Kondo et al. 1998, 2004; Tamura et al. 2002; Prima et al. 2006).

The Quaternary volcanoes on the NE Honshu side of the TTL are divided into four across-arc volcanic zones according to their K-level (Fig. 18): A (Aoso-Osore volcanic zone); S (Sekiryo volcanic zone); M (Moriyoshi volcanic zone); and C (Chokai volcanic zone) (Nakagawa et al. 1986). The K-level in volcanic rocks is lowest at the volcanic front side and highest at the back-arc side (Fig. 18a). The Sekiryo volcanic zone is an axial zone of arc volcanism, in which the temperature of andesitic magmas is the highest and the eruptive materials are the most voluminous (Fig. 18d, e: Yoshida et al. 2005).

Sakuyama (1977) classified intermediate to felsic volcanic rocks into three types on the basis of...
Fig. 19. (a, b) Injection of depleted asthenospheric mantle into enriched continental lithosphere during: (a) Stage 5 (21–17 Ma) to (b) Stage 6 (17–15 Ma), forming the Yamato Basin rift and the Northern Honshu rift, respectively (after Sato et al. 2004). (c) Schematic cross-sections depicting the magmatic evolution and tectonic setting of the middle to late Miocene NE Japan arc (Sato & Yoshida 1993; Yoshida et al. 1995). DMS, depth of magma segregation; QVF, Quaternary volcanic front. The arrow with QVF in (c) denotes the land area. (d) Schematic diagram showing the location of slab dehydration, fluid pathways and partial melting for frontal- (QVF) and rear-arc (QRA) basalts in the NE Japan arc, after Kimura & Yoshida (2006). A P-wave perturbation seismic tomography image (Nakajima et al. 2005) is superimposed, with lighter colours showing greater perturbation. SED, sediments; DM, depleted mantle; and F, fluid flux and degree of melting. Black bar under QVF and QRA denotes the land area. Dashed arrows show fluid pathways from different slab source regions to the magma source regions.
of hydrous phase assemblage, and showed that the volatile contents decrease from the back-arc to the volcanic front in the Quaternary volcanoes of the NE Japan arc. Figure 18c shows a schematic isobaric phase diagram at a pressure of about 5 kbar (Sakuyama 1979; Yoshida et al. 2005). Volcanoes that are composed of rocks without hornblende and biotite phenocrysts are dominantly distributed in the frontal volcanic zone. Those with hornblende and no biotite phenocrysts occur mainly in the back-arc side of the volcanic front, and those with biotite and hornblende phenocrysts mostly appear in the area furthest from the trench. The most primitive Sekiryo low-K tholeiitic basalts, however, have a higher water content (Kuritani et al. 2013) than the Sekiryo calc-alkaline andesites, and this means that the calc-alkaline mafic end-member needs to degas before forming andesites by any process (Miyagi et al. 2012).

**Across-arc geochemical variation**

After establishment of the island-arc magmatism at around 13.5 Ma, a prominent across-arc variation dominated. The across-arc geochemical variation of Quaternary lavas (1.7–0 Ma) in the NE Japan arc has been examined rigorously since the 1950s. Total alkalis (Kuno 1966), K2O (Fig. 18a) and incompatible element contents increase gradually from the volcanic front to the rear-arc lavas (Kawano et al. 1961; Aoki et al. 1981; Yoshida & Aoki 1984; Sakuyama & Nesbitt 1986; Nakagawa et al. 1988; Yoshida et al. 1997; Kimura & Yoshida 2006, and references therein). Sr and Pb are more radiogenic in the volcanic front, and Nd is less radiogenic (Fig. 9c: Shibata & Nakamura 1991; Kimura & Yoshida 2006; Takahashi et al. 2012). These are typical of across-arc variations commonly found in arcs (e.g. Dickinson & Hatherton 1967; Kimura & Stern 2008). Because of the relatively low slab-dip angles, the volcanic arc is wider in NE Japan than elsewhere, and the across-arc variations are more prominent compared with those of other subduction zones (Kimura & Stern 2008).

The origins of across-arc variations have been examined, and different depths of melt segregation (Kuno 1959), different degrees of partial melting in the source mantle (Miyashiro 1974), different slab flux (Shibata & Nakamura 1991) and different sources of the mantle (Nohda et al. 1988, 1992; Togashi et al. 1992; Ujike & Tsuchiya 1993) have been proposed. Kimura & Yoshida (2006) and Kimura et al. (2009) examined the various parameters involved in the melting of the slab liquid-fluxed mantle across the NE Japan arc, and proposed a common depleted mantle source which is fluxed by different slab fluids from the shallow slab beneath the volcanic front and from the deep slab

beneath the rear-arc volcanoes. They also estimated the higher slab flux fraction beneath the volcanic front and the lower slab flux fraction beneath the rear-arc, which resulted in a larger extent of partial melting of the mantle beneath the volcanic front than the rear-arc. Magma segregation depths were also estimated to be shallower at c. 1 GPa beneath the volcanic front, but deeper at c. 2 GPa beneath the rear-arc (Fig. 19d). Kimura and co-authors concluded that seismic observations are consistent with the results of their model based on the intensive–extensive variables (e.g. slab dehydration depth, slab temperature, flux fraction to the mantle wedge, degree of partial melting, and melting pressure and temperature), derived from the geochemical mass-balance calculation package called the Arc Basalt Simulator (Kimura & Yoshida 2006; Kimura et al. 2009, 2010).

**Along-arc geochemical variation**

Along-arc isotopic variation is unusually large at the volcanic front in the Quaternary NE Japan lavas. A southward increase in Sr and Nd isotope ratios has been identified (Fig. 9b: Notsu 1983; Kersting et al. 1996; Kimura & Yoshida 2006; Takahashi et al. 2012). The spatial variation correlates closely with the basement terranes (Kersting et al. 1996) and the isotopic composition of basement granitoids (Kimura & Yoshida 2006), which suggests a correlation to the compositional variation of the mantle lithosphere (Kersting et al. 1996), the basement granitoids (Kimura & Yoshida 2006) and the lower crustal amphibolite, through the contamination of the partial melts (Takahashi et al. 2012). Tatsumi et al. (2008) and Takahashi et al. (2012) also found that the strong spatial variation is more prominent in the tholeiitic (TH by Miyashiro 1974) suite of basaltic andesite to andesite magma, rather than in the calc-alkaline (CA by Miyashiro 1974) magma suite, and suggested lower crustal amphibolite origin of the TH lavas. Kimura & Yoshida (2006), however, suggested that the most primitive TH magmas along the volcanic front are upper mantle melts in origin, and that they were later affected by contamination with the lower crustal amphibolite melts.

**Role of the heterogeneous lower crust**

The distribution of both zones of basement terranes and basement granitoids are oblique to the alignment of the Quaternary volcanoes (Fig. 9a). For example, the Mino (south), Abukuma (middle) and Kitakami (north) belts are divided by the TTL and the HTL. An arbitrarily parallel distribution of the South, Transitional and North geochemical zones of the basement granitoids (Kagami et al. 2000;
Fig. 20. (a) Across-arc variation of the SiO₂-normalized alkali (Na₂O + K₂O% at 57.5% SiO₂) level, projected to the 40° N latitude line of NE Japan, and its secular change (modified from Yoshida et al. 1995). Tatsumi et al. (1983) showed that the depth of the magma segregation beneath the NE Japan arc is shallow (1.1 GPa) beneath the volcanic front, but...
Kagami (2005) is also oblique to the volcanic distribution (Fig. 9a). These reflect the changes in allocation of the arc-trench system before and after the opening of the Japan Sea; the latter was NE–SW (oblique to the present NE Japan arc) before the opening, and NNE–SSW (parallel to the present NE Japan arc) after the opening until the present (Otufuji et al. 1985; Ohki et al. 1993; Yoshida et al. 1999b). The present-day mantle structure, including the low velocity and mantle anisotropy structures (Nakajima & Hasegawa 2004; Nakajima et al. 2006b), is principally parallel to the present arc-trench system, although there are some mantle heterogeneities in the areas of SW Honshu and NE Honshu bordered by the TTL. Therefore, it is likely that the along-arc spatial variation of the Quaternary lava chemistry is mainly affected by the overlying lithosphere, and the most plausible candidate for affecting this appears to be the lower crustal melts, which can contribute to both the origin of the basement granitoids and to the Quaternary lavas.

Role of the mantle wedge

In contrast to the volcanic front lavas, the isotopic compositions of the Quaternary rear-arc lavas commonly show a contribution from a depleted source similar to those for mid-ocean ridge basalts (MORBs) or the back-arc basin basalts, irrespective of the overlying basement terranes (Kimura & Yoshida 2006). The resistance of the rear-arc lavas to crustal contamination has been ascribed to the high element concentrations in the primary mantle-derived basalts, which include Sr, Nd and Pb (Kimura & Yoshida 2006), and to a stable magma plumbing system in the thick, low-temperature, brittle crust of the back-arc region (Fig. 15b: Shibazaki et al. 2008). The high element abundances of incompatible elements (excluding heavy rare earth elements) could reflect a low degree of melting of the rear-arc mantle in the garnet stability field (Kimura & Yoshida 2006; Kimura et al. 2010), and this would be consistent with the deep top-surface of the low velocity mantle beneath the rear-arc (Fig. 11).

The common features of the rear-arc lavas, and the overall across-arc geochemical variations, correlate closely with the mantle wedge and slab structures running parallel to the arc-trench system. This indicates that the establishment of the volcanic arc was fundamentally controlled by the slab–mantle infrastructures. The proposed basic structure is a focused slab dehydration at around 80–90 km deep, and a high-temperature and melt-bearing mantle wedge developed parallel to the slab between a depth of 30 and 150 km (Fig. 19d: Kimura & Yoshida 2006; Kimura et al. 2009). This suggests that the contribution of slab-derived materials to the generation of frontal-arc magmas is significantly higher than the contributions to the rear-arc magmas. Moreover recent results of analysis of water contents in glass inclusions in phenocrysts suggest that frontal-arc basalts have high water contents, and that the magmatic water content does not always show an across-arc variation (Walker et al. 2003; Portnyagin et al. 2007; Kimura et al. 2010; Kuritani et al. 2013).

Development of the NE Japan arc in the late Cenozoic

The detailed images of the slab–mantle–crust seismic structures are intimately related to the spatial variation in the Quaternary magmas (e.g. Kimura & Yoshida 2006; Kimura et al. 2009). These enabled the most detailed discussions pertaining to the magma genesis model for magmatic arcs. The sub-arc structures and the present magmatism (the latest island-arc period), however, are evidently the products of the preceding continental margin and back-arc basin periods, as envisaged by the tectonic history and mantle–crust structure of the arc. Therefore, the developmental history of the subduction system and magmatism in these preceding periods has also been rigorously studied. This study now summarizes the results of these studies in a temporal order. The following sections deal with: (1) the drastic change in the thermal structure and source materials of the mantle in the back-arc basin period; (2) the transitional nature of the back-arc basin to island-arc periods in the early stages of the island-arc period; and (3) the youngest island-arc period.

**Fig. 20.** (Continued) deeper (2.3 GPa) beneath the back-arc, and corresponds with a change in magmatic composition from olivine tholeiite (THB) to high-alumina basalt (HAB) and then to alkali olivine basalt (AOB) at a similar temperature of about 1300 °C. This figure shows the secular changes within an estimated depth of magma segregation; the thermal structure in the mantle wedge through the Cenozoic Era. The abbreviations, K–S are the same as in Figure 1. The black bar above the cross-section denotes the land area. (b) Secular changes of across-arc variation of 

SiO2-normalized alkali level (modified from Yoshida et al. 1995). The parallelograms show, for reference, the field of Quaternary (1.7–0 Ma) volcanics. SiO2-normalized alkali contents were plotted as a function of the distance of the volcanic centre from the Quaternary volcanic front (QVF).
Secular changes in the mantle

Estimation of mantle melting conditions and thermal structure. Changes in the magmatic composition could reflect changes in: (1) the source mantle composition; (2) the slab-flux; (3) the thermal structure of the mantle; and (4) the heterogeneous overlying crust. Experimental studies have investigated the melting conditions with the assumed water content in the source. Tatsumi et al. (1983) conducted multiphase saturation experiments for the Quaternary basalt compositions and showed that the depths of magma segregation beneath the NE Japan arc are shallow (1.1 GPa) beneath the volcanic front, but deeper (2.3 GPa) beneath the back-arc. This corresponds with a change in magmatic composition from olivine tholeite (THB) to high-alumina basalt (HAB) and then to alkali olivine basalt (AOB) at a similar temperature of about 1300°C. Following the experimental results, Yoshida et al. (1995) formulated the relation between segregation depth and SiO$_2$-normalized total alkali content.

Tatsumi et al. (1994) also conducted experiments on the early to middle Miocene MgO-rich primitive basalt compositions in the NE Japan arc, from both the frontal-arc and the back-arc sides, and showed that the depths and temperatures of the mantle sources were constant at 1.3 ± 0.1 GPa and 1320 ± 10°C, respectively, beneath the entire NE Japan arc. Yamashita & Fujii (1992) performed experiments on the early Miocene MgO-rich basalts erupted during the back-arc basin period, which had been sampled by drilling in the Japan Sea oceanic floor, and estimated pressure and temperature of 1.4 GPa and 1340°C. With these data, Tatsumi et al. (1994) concluded that the lithosphere thickness beneath the forearc had not changed at a depth of about 30–40 km (c. 1.1–1.3 GPa) during the past 20 Ma, but that the back-arc lithosphere has thickened through cooling at a depth of between about 40 km (c. 1.3 GPa) to around 70 km (c. 2.3 GPa; Fig. 15). The conclusion of the study suggests that asthenospheric upwelling during the back-arc basin opening period, and the subsequent deepening of the back-arc basin, may have caused cessation of volcanism in the back-arc basin, together with the thermal subsidence of the Japan Sea (Yamaji & Sato 1989). The asthenospheric deepening in the back-arc occurred through the establishment of the across-arc thermal structure of the mantle wedge in the island-arc period (Fig. 19c).

Secular changes of thermal structure in the mantle. Figure 20a summarizes the secular changes in the depth of magma segregation through the continental margin, the back-arc basin and the island-arc periods, as estimated by SiO$_2$-normalized total alkali content using the formula proposed by Yoshida et al. (1995). In the continental margin period, it is estimated that the asthenosphere was at its shallowest beneath the volcanic front, and that it deepened gently towards the back-arc. The depth of asthenosphere beneath the volcanic front, however, was deeper than today. In the back-arc basin period, the axis of shallower asthenosphere was not located beneath the volcanic front at the time, but was located along the present-day coastline of NE Japan owing to the asthenospheric upwelling during the back-arc basin opening. In the island-arc period, the asthenospheric depth steeply inclined towards the back-arc, and the depth beneath the volcanic front was shallower than in the previous period (Fig. 20b: Yoshida et al. 1995).

Isotopic variation of the source mantle and temporal development. Back-arc basin volcanism in the Yamato Basin started with the eruption of isotopically enriched tholeiites during 21–17 Ma (LD-group in Fig. 21a: Nohda 2009), and was followed by isotopically depleted tholeiites during 17–15 Ma (D-group in Fig. 21a: Nohda 2009). The same drastic change in Sr and Nd isotope ratios from enriched to depleted compositions, by the injection of a depleted asthenospheric mantle into an enriched continental lithosphere, was observed in: SikhoteAlin at around 35 Ma (Nohda 2009); the Oga Peninsula at 22 Ma (Kurasawa & Konda 1986); the eastern margin of the Japan Sea (Aosawa rift) at 16.5 Ma (Ujike & Tsuchiya 1993; Yagi et al. 2001); and the inland NE Japan (Kuroko rift) at 15 Ma (Yagi et al. 2001; Nohda 2009; Yamada et al. 2012). Nohda (2009) explained this relationship by an eastward asthenospheric flow during the continental arc to the back-arc basin periods (Fig. 19a). The back-arc basin opening ceased at around 15 Ma, and subsidence of the back-arc region began at 14.5 Ma, perhaps owing to cooling of the injected hot mantle (Yamaji 1990).

In contrast, there was little to no change in the Sr and Nd isotope ratios at the volcanic front, and enriched magmas erupted throughout the back-arc basin period, although the Hf isotope ratios appear to have become more radiogenic with time (Nohda et al. 1988; Tatsumi et al. 1988; Shuto et al. 1993; Hanyu et al. 2006). The 23 Ma high-Mg andesite at Choshi has the lowest $\varepsilon_{Hf}$ values, while the 5 Ma Mitaki and Quaternary Iwate have the highest $\varepsilon_{Hf}$ values, and the 16–12 Ma Ryozen, Takadate and Tomari volcanics show intermediate $\varepsilon_{Hf}$ values (Hanyu et al. 2006). Hanyu et al. (2006) concluded that the high $\varepsilon_{Hf}$ value of late-stage volcanism at the volcanic front can be explained by the addition of fluids from subducted sediments and the oceanic crust to the mantle wedge.
Thermal subsidence with lithospheric thickening.

The change from extensional to neutral tectonics was associated with a rapid decrease in erupted volumes (Fig. 6a). This stage (Stage 8: oceanic island-arc stage) has been regarded as being a thermal subsidence stage under a neutral stress field (Yamaji & Sato 1989; Sato 1992, 1994). The eruption of alkali basalts in the back-arc side during this stage is generally explained by the increasing depths of magma segregation, and the decreasing degree of partial melting of parental magmas (Tatsumi et al. 1983, 1994; Sato & Yoshida 1993; Yoshida et al. 1995). The mantle isotherm steepened in the back-arc side during Stage 8. As a result, just after the end of the back-arc basin period, the thermal structure in the uppermost mantle wedge became very similar to the present one. The transition from a back-arc basin to an island-arc setting is explained by the lithospheric thickening beneath the back-arc side of the NE Honshu arc (Fig. 19c: Sato & Yoshida 1993; Tatsumi et al. 1994, Yoshida et al. 1995), owing to a temperature drop associated with a decrease in erupted volumes when the extension stopped (Yamada & Yoshida 2011). It is possible that the cooling of the back-arc upper mantle was mainly the result of a large volume of magmatic effusion in the back-arc basins. As a result of lithospheric thickening in the back-arc side, the magmatic composition changed from enriched low-K tholeiites to depleted...
1: N type-MORB and volcanic arc basalts
2: Within plate tholeiites and volcanic arc basalts
3: E type-MORB
4: Within plate alkali basalts and within plate tholeiites
5: Within plate alkali basalts

Melting model
- DM fluxed melting
- PM
- PM 0.5% extract
- DM
- DM 0.5% extract

- Yamato Basin
- Aosawa Rift
- Babame Rift
- Kuroko Rift
- Basalt and andesites in the Oga Peninsula
- Island-arc basalts in the fore-arc range (Mitaki, Jyogi and Araya basalt)

Temporal trend

Frontal Volcanics
- 21~34 Ma
- 14~16 Ma
- 8~12 Ma
- 0~3 Ma

Lesser Antilles
Sunda
Kermadec
New Britain
Sottocchi
Izu-Mariana
alkali basalts between 13.5 and 10 Ma (Fig. 21a). In the volcanic front, the magma composition evolved from high high field strength element (HFSE) magmas in the back-arc basin period to low HFSE magmas in the island-arc period, together with an increase in $e_{\text{Hf}}$ value (Shuto et al. 1992; Hanyu et al. 2006). Felsic magma chambers became shallower with a change from high-temperature aphyric–or plagioclase–phyric rhyolites to low-temperature plagioclase–quartz phryic rhyolites during the tectonic change from a back-arc basin to an island-arc setting (Yamada & Yoshida 2004, 2011). It is possible that this petrographical change within the felsic magma was related to a shallowing of the emplacement level of felsic magma reservoirs from the deep crust to the shallower crust.

Temporal evolution of basaltic magma in the NE Japan arc. The chemical composition of late Cenozoic volcanic rocks from the NE Japan arc suggests the presence of three types of mantle sources: the subcontinental enriched mantle; the sub-arc lithospheric-enriched mantle (Ujike & Tsuchiya 1993); and the asthenospheric-depleted mantle (Nohda et al. 1988; Tatsumi et al. 1989; Ohki et al. 1994; Shuto et al. 1995, 2006; Yoshida et al. 1995). Ujike & Tsuchiya (1993) showed that the geochemical signatures of the magmas, including the Sr and Nd isotope ratios and Zr/Y ratios, changed drastically from a continental margin type to an oceanic arc type (Pearce 1983) in the Aosawa rift at around 16 Ma. Shuto et al. (1995) also showed that the late Cenozoic basaltic rocks in the NE Japan arc exhibit a systematic decrease in Zr/Y ratios with age (from within-plate basalts to volcanic arc basalts), from 35 Ma to the present. Yamada et al. (2012) showed the temporal evolution of basaltic magmas on the Nb–Zr–Y discrimination plot of basalts (Fig. 22b, c: Meschede 1986), and revealed that basalts from the Oga Peninsula of the continental margin period plot in the field of within-plate alkali basalts. After the opening of the Sea of Japan, the basaltic magmas evolved from low-Nb N-type MORB basalts (Yamato Basin) to within-plate tholeiite basalts (Aosawa Rift), and ultimately to volcanic arc basalts (Babame Rift and younger volcanic rocks).

To examine the source material and partial melting of the source material controlling the HFSE enrichments, Yamada et al. (2012) calculated the nature of the changes in the Zr–Nb–Y compositions of basalt melts during the adiabatic ascent of a primitive mantle (PM: Sun & McDonough 1989) and a depleted mantle source (DM: Workman & Hart 2005). Model calculations of adiabatic decompression melting (Phipps-Morgan 1999) were performed at mantle potential temperature $T_p = 1450 ^\circ C$, with the mineralogical mode during ascent and melting based on Ghiorso et al. (2002). Slightly depleted PM and DM source compositions were also calculated using the ABS3 model (Kimura et al. 2010), assuming the extraction of a 0.5% basalt melt from PM and DM sources. The results are shown in Figure 22b.

The basalts from the PM source mantle change from within-plate alkali basalt to within-plate tholeiites and then from E-type MORB to N-type MORB, with decreasing depth of magma segregation. Similarly, the basalts from the DM source mantle follow the field of within-plate tholeiite to N-type MORB (Fig. 22b). The Oligocene Oga alkalic within-plate basalts of the continental-margin period, which have an isotopically enriched mantle source, are interpreted to have originated from the continental lithospheric mantle, and deep-level melting of a thick, enriched lithospheric
mantle (see the 3.0–2.8 GPa symbols of PM and the 0.5%-melt-extracted PM plots in Fig. 22b) can form such alkaline basalts. In contrast, the N-type MORB of the Yamato Basin and the Aosawa within-plate tholeiite may have originated from a common DM or a 0.5%-melt-extracted DM source mantle, with shallower depths for the Yamato Basin basalts (Fig. 19a) and greater depths for the Aosawa Rift basalts (Fig. 19b).

Volcanic arc basalts from the Babame and Kuroko rifts, which lie between the DM and PM models, could originate from a shallow depth. The elevated Nb values may indicate a more fertile source mantle than that of the Yamato Basin and Aosawa Rift basalts. The most radiogenic Nd isotopic composition from the Yamato Basin basalts yields ${^{143}\text{Nd}}/{^{144}\text{Nd}} = 0.51316$ (Cousens & Allan 1992), which is comparable to values obtained for Quaternary basalts of the back-arc side, but higher than those obtained for basalts of the Babame and Kuroko rifts, suggesting a relatively fertile mantle source. It is also possible that the addition of slab melt components to the shallow (2–1 GPa) DM source mantle could result in an increased Nb value (shaded area in Fig. 22b). This fluxed melting calculation was performed using the ABS2 model (Kimura et al. 2010). Yamada et al. (2012) explained the fertile shallow mantle source of basalts for the Babame and Kuroko rifts, by a shallow DM melting induced by slab flux from dehydration of the Pacific Plate slab. This is because the tectonic setting of the Babame and Kuroko rifts was in a time of transition from the late stage of back-arc basin opening to the island-arc stage, and the compositions of basalts of the early island-arc stage are almost identical to those of the Babame and Kuroko rifts (see Fig. 22c).

**Palaeomagnetic studies and rotational history.** Takahashi & Saito (1997) reviewed palaeomagnetic studies and discussed Miocene intra-arc bending at the arc–arc collision zone in central Japan.

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**Fig. 23.** (a) Past distribution of volcanoes in the NE Japan arc (data from Kondo et al. 1998). Yellow dashed lines are the surface projections of ‘hot fingers’ by Tamura et al. (2002). The data are divided into two periods: 11–5 and 5–0 Ma (modified from Honda & Yoshida 2005a). QVF, VRE, TTL and HTL are as in Figure 2. (b) Map of volcanic ages v. distance from the Quaternary volcanic front, adapted from Honda & Yoshida (2005a). The age data for area P (Yamada & Yoshida 2002) and area Q (Kondo et al. 2004) are plotted against the distance from the volcanic front to the sampling sites. Black circles are the radiometric ages, and the pale-blue-dotted hatched zone is the age ranges estimated from stratigraphic data. Each arrow shows the possible migration of volcanism from back-arc to the volcanic front side.
Palaeomagnetic results reveal the Miocene rotational history in central, SW and NE Japan (Fig. 21b: Takahashi & Saito 1997). As shown in Figure 21b, differential rotations in the Japanese islands took place between 20 and 10 Ma. The clockwise rotation of SW Japan occurred at c. 15 Ma, while the Kanto Mountains (K in Fig. 9b) rotated during the Middle Miocene (15–10 Ma). The small declinations in the Nohi area (N in Fig. 9b) indicate that differential rotation between SW Japan and the Nohi area occurred at c. 15 Ma. However, the palaeomagnetic deflections of NE Japan show a counter-clockwise direction during the Early Miocene. The major rotations of the SW and NE Japan arcs ceased at 14 Ma, while the Kanto Mountains rotated until the Late Miocene (Fig. 21b). The major change occurred in the stress axis in the NE Japan arc from a NNW–SSE to a NE–SW direction (Fig. 7) around 15 Ma in association with the rapid rotation of SW Japan. It is likely that this rapid rotation locked the NE Japan block to the Kuril forearc sliver, as a collision of the Kuril forearc sliver and the NE Japan arc had begun at this time (Kimura 1986; Acocella et al. 2008). The events at 15 Ma are consistent with the change in shear movement along the TTL, from sinistral (the back-arc opening stage) to dextral (the forearc sliver collision stage; Awaji et al. 2006; Yoshida 2009).

Mantle dynamics and arc magmatism

The origin of geochemical variations in volcanic rocks in late Cenozoic NE Japan arc has been discussed in the preceding sections of this paper. It is evident that the most significant event was the opening of the back-arc basin and the subsequent subduction volcanism in the island-arc period. These changes are closely related to the mantle structure, including that of: (1) the upwelling-subsidence of the asthenosphere during the back-arc basin period; and (2) the mantle wedge corner flow in the island-arc period. Such a temporal development of the mantle has been investigated by
several geodynamic models (Honda & Yoshida 2005a, b; Honda et al. 2007; Zhu et al. 2009).

Spatio-temporal variations of volcanisms and mantle convection. Volcanoes were distributed uniformly immediately after the cessation of the opening of the Japan Sea, but, in the island-arc period, they had a tendency to be concentrated and finally to form branches and a trunk structure (Figs 18b & 23a; Yoshida et al. 1995; Kondo et al. 1998, 2004; Yamada & Yoshida 2002; Prima et al. 2006). The trunk structure consists of volcanoes located along the volcanic front, and the branches coincide almost with the ‘hot fingers’ (Tamura et al. 2006). The trunk structure consists of volcanoes whose ages are examined the spatial and temporal evolution of volcanism. When the ages of the volcanoes are plotted against the distance from the volcanic front (Yamada & Yoshida 2002; Kondo et al. 2004), the spatio-temporal plots show the migration of volcanism from back-arc to volcanic front (Fig. 23b: Honda & Yoshida 2005a). The most recent migration (5–0 Ma) suggests a velocity of migration of $c. 2 \text{ cm} \cdot \text{a}^{-1}$ (Honda & Yoshida 2005a), which may be correlated with dynamic movement in the mantle wedge.

Honda & Yoshida (2005a) performed a geodynamic simulation of the mantle wedge and examined the spatial and temporal evolution of volcanism. When the ages of the volcanoes are plotted against the distance from the volcanic front (Yamada & Yoshida 2002; Kondo et al. 2004), the spatio-temporal plots show the migration of volcanism from back-arc to volcanic front (Fig. 23b: Honda & Yoshida 2005a). The most recent migration (5–0 Ma) suggests a velocity of migration of $c. 2 \text{ cm} \cdot \text{a}^{-1}$ (Honda & Yoshida 2005a), which may be correlated with dynamic movement in the mantle wedge.

Tamura et al. (2002) explained the distribution pattern of Quaternary volcanoes in the NE Japan arc with a ‘hot fingers’ model, by assuming the low-velocity mantle wedge beneath the volcanoes to have high-temperature anomalies. Honda & Yoshida (2005a), however, suggested that the distribution pattern of the volcanoes was not preserved over a long period, and that the finger-like structure flip-flopped at around 5 Ma. These phenomena have been explained by the existence of small-scale convolutions beneath the arc. Alternatively, Zhu et al. (2009) simulated the 3D dynamics of hydrous thermo-chemical plumes in the oceanic subduction zones, and computed the spatial and temporal patterns of melt generation intensity above the slab, then compared results with data from the NE Japan arc. This suggested that the clustering of volcanic activity could potentially be related to the activity of thermal–chemical plumes in the mantle wedge. Even though the models vary, the important insights gained by the geodynamic model perspectives are that: (1) the mantle convection patterns can fundamentally control the magma genesis; and thus (2) control the distribution of the volcanic activity and magma production rate.

Temporal evolution of mantle convection. Temporal variations in the dynamics of the continuously cooling low-viscosity mantle wedge have been examined (Fig. 24a–c: Honda & Yoshida 2005a). It is

Fig. 25. (a) Distribution of late Cenozoic calderas and Quaternary volcanoes (Yoshida et al. 1999a) superimposed on coloured map of water temperatures of hot spring (modified from Yano et al. 1999). Red and blue colours denote the high and low temperatures, respectively. The temperature scale is shown at the bottom. QVF, VRE, TTL and HTL are as in Figure 2. Legends for the other symbols are shown in the upper-right box of Figure 1. The section X–Y and rectangle in (a) show the study section in (c) and (d), and the study area in (b), respectively. (b) Location of reflection points (blue circles) of each S-wave reflector in the depth ranges of: (1) 0–5 km; (2) 5–10 km; (3) 10–15 km; (4) 15–20 km; and (5) 20–25 km (Hori et al. 2004), at around the Onikobe caldera swarm superimposed on the map of thermal water temperatures (Yano et al. 1999) and modified from Yoshida et al. (2005). S-wave reflectors distributed beneath and around the caldera complex are assumed to be fracture systems filled with high temperature hydrothermal fluids. The distribution pattern of S-wave reflectors changes with depth, suggesting that it is controlled by the distribution of unexposed plutons under the caldera structures and the related fracture system. It may also be related to the crustal thermal structure and non-symmetrical updouming of the Ou Backbone Range. (c) Relationship between geomorphology, major fault systems and recent crustal movements (modified from Imaizumi 1999). Recent tectonic movements have been strongly controlled by a reverse reactivation of the Miocene basin-bounding normal faults under WNW–ESE (to east–west) compression, owing to the westward subduction of the Pacific plate since the late Pliocene (Nakamura 1992; Sato 1994; Sato et al. 2002; Kato et al. 2004). Black arrows indicate the uplift and subsidence of the each tectonic unit, estimated from geological and geomorphological data (Japan Association for Quaternary Research 1987, Imaizumi 1999; Tajikara & Ikeda 2004). (d) Accumulated stress (Pa) distribution after 3000 years using a finite element analysis for an X–Y section of the NE Japan arc (Shibazaki et al. 2008). Red and blue colours denote high and low stress concentrations. The scale (Pa) is shown at the bottom. W-shaped stress concentration (shear faulting) zones in the upper crust are created along the locations of major fault zones by the non-uniform thermal structure of the crust and uppermost mantle beneath the NE Japan arc. The W-shaped stress concentration zones arise from a shortening deformation associated with nonlinear viscous flow in the two low-viscosity regions (hot regions) beneath the Ou Backbone Range and the Dewa Hills, surrounded by the high-viscosity region in the lower crust and the uppermost mantle (see Fig. 15b). The solid grey lines indicate the inland area. Upper and lower white lines indicate the boundaries between the upper crust and the lower crust, and the Moho discontinuity, respectively. Red arrows indicate the locations of the three major active fault zones at the surface. Hypocentres of earthquakes (1 January 2003 to 31 December 2005) are also shown in the section by black circles (Shibazaki et al. 2008).
evident that three distinctive phases exist during cooling, including: (1) phase 1 – a steady state convection; (2) phase 2 – a constant cooling with 2D convection; and (3) phase 3 – an unstable 3D convection (Honda & Yoshida 2005a). The first two phases form a homogeneous thermal structure along the arc, but the final phase forms a trunk-and-branches structure (Fig. 24). The volcano distribution in the back-arc basin period of the NE Japan arc was in alignment with arc-parallel rift systems, as noted above, and thus would have related to the 2D convection. In contrast, the heterogeneous distribution of volcanoes in the late island-arc period may have related to the 3D convection (Fig. 24d).

The long-term developmental history of the mantle wedge could have been constrained by the surface manifestation of the volcanic activities, although the role of the deep-seated slab and the island-arc crust should also be considered as the driving force of the movement and production of disturbing effects to the ascending magmas, respectively. Detailed studies of erupted magmas would contribute essential knowledge towards this type of study, and would provide knowledge of additional strong constraints involved in the deep processes.

Island-arc evolution and arc magmatism

An important tectonic change associated with the change in magmatic activity occurred in the back-arc basin period and was then followed by the island-arc period (Sato 1994; Jolivet et al. 1994; Nakajima et al. 2006c; Yoshida 2009). Both the surface tectonic and deep mantle source conditions in the two periods were different, as noted above. The bimodal magmatism continued to the early island-arc period, and was followed by two stages of caldera-dominated magmatism, and lastly by andesite-dominated magmatism. This implies that andesite-dominated arc volcanism has been limited to the last 3.5 myr.

Tectono-magmatic evolution in the island-arc period

The island-arc period after 13.5 Ma can be divided into four phases of igneous activity: (1) the formation of oceanic island chains with submarine volcanism (13.5–8 Ma); (2) a late Miocene caldera-forming phase with weak updoming of the Ou Backbone range (8–5.3 Ma); (3) a Pliocene caldera-forming phase under weak compression (5.3–1.7 Ma); and (4) the creation of a late Quaternary highly compressional volcanic arc with andesitic stratovolcanoes (1.7–0 Ma). Nakajima et al. (2006c) argued that there were three phases of uplift in the surrounding the Ou Backbone range after the rapid subsidence (Aosawa and Kuroko 1989). These magmas would then have fractionated, re-mobilized or re-melted from solidified

Caldera-dominated magmatisms

The uplift of the present Ou Backbone Range began in the island-arc period at 10 Ma, and was associated with an increase of horizontal compression (Nakajima et al. 2006c). Between 8 and 1.7 Ma, active felsic volcanism created more than 80 calderas associated with subordinate andesites to basalts in the NE Japan arc (Figs 2 & 25a; Ito et al. 1989; Sato & Yoshida 1993; Sato 1994; Yoshida 1999a; Prima et al. 2012). There are two peaks in caldera formation interrupted by a short period of dormancy at 5–4 Ma, which is related to a short transgression, and the number and size of collapse calderas decreased from the late Miocene to the Pliocene (Fig. 8a).

The calderas have an average diameter of about 10 km and an average aspect ratio of 1.24 in diameter. They are divided into three groups related to their diameter size (about 5, 10 and over 14 km), and are mainly classified into piston-cylinder type with subordinate funnel type. The spatial and size distributions of calderas are comparable with those of the Cretaceous granitic plutons from the Kitakami Mountainland in NE Honshu (Yoshida et al. 1999a). The collapse of such calderas would have formed in a neutral to weakly compressive stress field (Yoshida et al. 1993; Sato & Yoshida 1993; Sato 1994), and this would have resulted in the rise of felsic magmas into the crust where large intra-crustal magma chambers were formed (Aizawa & Yoshida 2000; Aizawa et al. 2006).

NE–SW compression to ENE–WSW (east–west) compression. It has been argued that the regional stress field controlled the volcanic activity in the NE Japan arc (Sato & Yoshida 1993; Yoshida et al. 1993, 1997, 1999a; Accocella et al. 2008), and that basaltic magmas derived from the mantle wedge, underplated and stagnated near the Moho, which acts as a density barrier (e.g. Ryan 1987; Takada 1989). These magmas would then have fractionated, re-mobilized or re-melted from solidified
precursors or the pre-existing arc crust, to form the felsic magmas in the inland area of the NE Japan arc, with a thick crust (Fig. 19c; Sato & Yoshida 1993). Such an event is confirmed by the existence of large felsic effusives in the eastern margin of the back-arc basin rift system (Figs 4 & 5; Yamada et al. 2012). During the neutral stress condition between 13.5 and 10 Ma, the felsic magmas would have risen diapirically through the ductile lower crust owing to their buoyancy (Aizawa & Yoshida 2000; Aizawa et al. 2006), and the mode of ascent would have changed in the brittle upper crust to dyke or sheet (Fig. 15b). An increase in the compressional stress field occurred between 10 and 8 Ma, and it is likely that this increase led to the formation of sills and laccolithic shallow reservoirs in the upper crust (Fig. 19c: Sato & Yoshida 1993; Aizawa & Yoshida 2000). The regional change in the stress field was, therefore, the major control of caldera-dominated volcanism with laccolithic shallow reservoirs that occurred in the earlier half of the island-arc period. Felsic magma at this level could then have intruded along subsurface low-angle thrust sheets (Fig. 11b), and it is possible that magma migration along the thrust sheets caused the uplift of the Ou Backbone Range (Yoshida et al. 1993; Sato & Yoshida 1993; Sato 1994).

The clockwise rotation of SW Honshu (Otofuji & Matsuda 1983) and the collision with the Kuril forearc sliver (Kimura 1986) caused an oblique (NE–SW-trending) compression of the NE Japan arc during the Mio-Pliocene, and triggered felsic magmatism along the areas of localized extension (Acocella et al. 2008). After about 5 Ma, the Pacific plate accelerated (Pollitz 1986). Pollitz (1986) suggested that the change in Pacific plate motion introduced a large component of compression normal to the Japan trench. This ENE–WSW (east–west) strong compression closed the caldera-feeding systems and favoured the development of stratovolcanoes with deeper magma plumbing systems directly connected to the basaltic mantle source region. This strong coupling between the Pacific plate slab and the overriding NE Japan arc crust has continued since 3.5 Ma, perhaps with an occasional east–west stress release as envisaged by the Mw 9.0 Tohoku–Oki earthquake, which occurred on 11 March 2011 (Fio & Matsuzawa 2012), and has returned to the original NE–SW compression (Acocella et al. 2008; Kosuga et al. 2012). During the Quaternary, more andesites erupted, mainly from along-arc and east–west-aligned stratovolcanoes (Fig. 6c). The north–south thrust faults and the subordinate east–west-aligned volcanic structures produced by the east–west compression may reflect the convergence vector of the Pacific plate (Acocella et al. 2008).

Subcaldera plutons and S-wave reflectors. Many S-wave reflectors have been found along the caldera rims or around caldera complexes (Figs 13b, c & 25b) in the wide area of the NE Japan arc (Umino et al. 2002; Hori et al. 2004; Yoshida et al. 2005), although some of these may be related to active faults (Hori et al. 2004). These S-wave reflectors are presumed to reflect fracture systems filled with hydrothermal fluids, and seem to be generated from cooled magma reservoirs near wet solidus beneath the calderas (Fig. 25b). The shallow reflectors at depths of 0–5 km around the calderas are nearly vertical and parallel to the caldera walls, and are mainly distributed inside the caldera complexes. A number of reflectors at a depth of 5–10 km, with a gentle or flat dip, have also been found along the caldera rims or around caldera complexes. At a depth of 10–20 km, the reflectors fade away from the area of the large caldera complexes and are found on the northern to eastern sides of the caldera complexes. Those deeper than 20 km are located in an area beneath the caldera complexes, and the number of reflectors has decreased. The depth distribution of S-wave reflectors has two peak depths at 6–11 and 16 km (Hori et al. 2004). Figure 13b, c clearly shows that many reflectors are derived from the basic part of the middle crustal magma reservoirs at around the depth condition of the wet solidus of basaltic magma (Fig. 15a) to the eastern side of reservoir. The changes in the distribution of the reflectors with depth in the wide area of the NE Japan arc could be explained by the non-symmetrical cone and ring fractures formed around the cooling magmatic reservoirs (Anderson 1936). It is possible that low angle non-symmetrical cone fractures might have formed around subcaldera middle crustal plutons in a horizontally compressional stress field. At shallow depths, pre-existing caldera structures with ring fractures would then be used as conduits of hydrothermal fluids.

Change of intra-crustal thermal structure. The intra-crustal thermal structure was affected by the intensive intrusion of magma into the crust, which was in turn controlled by the mantle thermal structure (e.g. Honda & Yoshida 2005a), as well as by the magma storage systems related with large calderas formation (Fig. 19c). The crustal thermal anomalies are seismologically detectable (Figs 11, 13 & 14: Sato et al. 2002; Nakajima et al. 2006a; Xia et al. 2007). The distribution of thermal waters in NE Japan (Yano et al. 1999) coincides well with the dense distribution of late Cenozoic calderas (Fig. 25a). Such evidence suggests that still-hot plutons, solidifying from felsic caldera-forming magma reservoirs, may exist beneath many of these calderas (Yamada 1988; Sato et al. 2002; Yoshida
et al. 2005; Nakajima et al. 2006a; Takada & Fukushima 2013).

The thermal history of the NE Honshu arc, as inferred from petrological studies of the volcanic rocks (Ban et al. 1992; Yoshida et al. 1995) and studies of the thermal structure of forearc sedimentary basins (Yamaji 1994), shows that the geothermal gradient decreased from the Miocene (50–60 °C km⁻¹) to the Quaternary (20–30 °C km⁻¹). Miocene to Quaternary magma plumbing systems migrated to the back-arc side as the volcanic zone shrunk, and the depths of the magma reservoirs increased from the late Miocene to the Quaternary, inducing the cooling of the shallower reservoirs. The regional westward retreat of the volcanic front (Fig. 3; Ohguchi et al. 1989; Yoshida et al. 1995), together with the geodynamic model by Honda & Yoshida (2005a), is consistent with the regional cooling of the NE Honshu arc crust in the island-arc period of the NE Japan arc.

Andesite-dominated magmatism

Transition in eruptive mode from felsic calderas to andesitic stratovolcanoes. In the last stage of the island-arc period, the horizontal stress increased considerably, and felsic caldera magmatism with subordinate andesitic to basaltic eruptions changed to andesite-dominated magmatism. In this stage, the mantle temperature had decreased by small-scale convection associated with mantle corner flow, induced by long-lived subduction (Honda & Yoshida 2005a). The crust also became more rigid as it cooled, although magmatism had converged to the volcanic frontal area (Fig. 23b).

The east–west-aligned Quaternary volcanoes are the shallowest expression of the east–west-trending hot mantle branches (Fig. 24d), suggesting that mantle–crust coupling is important for magma ascent (Acocella et al. 2008). Such coupling favours the extrusion of, and mixing between, deeper mafic and shallower felsic magmas (Ban & Yamamoto 2002; Ban et al. 2007; Hirotani et al. 2009). The strong compressional crustal stress then increased pressure in the deep magma reservoirs and accelerated the upwelling of the deeper mafic end-member of calc-alkaline andesites. As a result, the total erupted volume of mixed calc-alkaline andesitic magma was twice that of the felsic caldera forming stage with more ductile upper crust owing to the presence of many large shallow magma reservoirs.

Recent tectonic movements and major fault systems across the NE Japan arc. Recent tectonic movements have been strongly controlled by the reverse reactivation of the Miocene basin-bounding normal faults under a WNW–ESE (to east–west) compression, owing to the westward subduction of the Pacific plate since the late Pliocene (Nakamura 1992; Sato 1994; Sato et al. 2002; Kato et al. 2004). Figure 25c (modified from Imaizumi 1999) shows the relationship between geomorphology, major fault systems and recent crustal movements. Black arrows indicate the uplift and subsidence of the each tectonic unit estimated from geological and geomorphological data (Japan Association for Quaternary Research 1987; Imaizumi 1999; Tajikara & Ikeda 2004). The development of S-wave reflectors in the eastern side of the caldera complexes at a depth of 10–20 km (Figs 13b, c & 25b) suggests that a strong compression around caldera-forming middle crustal magma reservoirs caused underground thrust sheet fractures towards the trench side. The segregated magmatic fluids then moved along the sheet fractures, and this may have formed the large number of S-wave reflectors. Shibazaki et al. (2008) calculated the accumulated stress distribution using finite element analysis (Fig. 25d). W-shaped stress concentration (shear faulting) zones in the upper crust are created along the locations of the major fault zones by the non-uniform thermal structure of the crust and the uppermost mantle beneath the NE Japan arc. The W-shaped stress concentration zones arise from the shortening deformation associated with nonlinear viscous flow in the two low-viscosity regions (hot regions) beneath the Ou Backbone range and the Dewa Hills, which are surrounded by the high-viscosity region in the lower crust and uppermost mantle (see Fig. 15b). The absolute steady stress is concentrated at the bottom of the upper crust beneath the Ou Backbone range and the Dewa Hill, as this area is marked by a high temperature gradient. The formation of many S-wave reflectors around middle crustal magma reservoirs is consistent with this shortening deformation associated with a nonlinear, viscous flow.

Shibazaki et al. (2008) have shown that a large-scale upward movement occurs beneath the Ou Backbone range and the Dewa Hill, caused by a shortening deformation in the upper mantle. Age data of volcanism have shown the migration of volcanism from the back-arc area to the volcanic front (Yamada & Yoshida 2002; Kondo et al. 2004; Honda & Yoshida 2005a), and the most recent migration started at around 5 Ma in the back-arc side (Fig. 23b). If the thermal condition in the uppermost mantle controls the crustal deformation as well as controlling volcanic activity, the uplift and shortening movement from the eastern margin of the Japan Sea to the Ou Backbone range during Pliocene to Quaternary (e.g. Ikeda et al. 2012) could be explained by a shortening in the upper mantle associated with the migration of volcanism correlated with the dynamic movement in the mantle wedge (Honda & Yoshida 2005a).
Magma mixing and H$_2$O degassing. Miyagi et al. (2012) proposed that the Quaternary frontal medium-\(K\) series magma from Hokkaido was made by mixing of low-\(K\) basalts and medium-\(K\) rhyolites, whereby the rhyolites were partial melts of previously intruded, semi-solidified low-\(K\) basalts. The observed chemical variation of melts was explained by fractional crystallization and subsequent re-melting or re-mobilization of low-\(K\) basalt. In addition, the depleted Sr, Nd and Pb isotopic compositions of the basalt and rhyolite support the interpretation that the two rock types derived from a common source, with little or no evidence of contamination by crustal rocks. It is possible that the source of the felsic end-member of magma mixing is derived via 90% fractional crystallization, or from re-melting, or re-mobilization of a solidified or semi-solidified precursor basaltic magma. In this case, the basaltic magma chamber would have been located closely beneath the rhyolitic magma chamber, at a depth of 10–15 km or more. They also estimated that the total volume of basalt supplied beneath the volcanoes was approximately 10 times the volume of erupted rhyolite magma. The solidified basalt, with about 5 wt% H$_2$O, supplied a much larger amount of magmatic volatiles (‘excess magmatic volatiles’) to the overlying felsic magma than could be dissolved. They concluded that most of the excess magmatic volatiles were lost by degassing before the eruption.

Tectonically controlled lateral variation of phenocryst assemblage. Quaternary calc-alkaline andesites from the NE Japan arc have shown a clear lateral variation in phenocryst assemblages (Fig. 18c). Although the volatile content of the magma and its crystallization temperature are related to the magmatic alkalinity, an estimated lower H$_2$O content (Sakuyama 1977) in the frontal calc-alkaline andesites (Fig. 18c) suggests that the mafic end-member of the mixed calc-alkaline magma from frontal volcanoes degassed prior to producing andesite by any process. As discussed above, a large number of \(S\)-wave reflectors are found along caldera rims, or to the eastern side of the caldera complexes. These \(S\)-wave reflectors could have formed from the excess volatile degassing of the deep-seated mafic intrusives. Under strong compressional stress conditions, the excess volatiles which were degassed from the deep-seated mafic magma reservoirs could have risen directly upwards, or could have escaped to the trench side along thrust sheets, forming seismically active focused deformation zones (Okada et al. 2010). These processes, or other degassing processes, could have been more effective at the volcanic front to the forearc region than in the back-arc side, owing to the strong shortening deformation associated with nonlinear viscous flow (Fig. 25d).

Concluding remarks

The NE Japan arc is a typical island arc associated with a cold subduction system. Cenozoic magmatism in the NE Japan arc can be divided into three major periods of volcanic activity: continental margin, back-arc basin and island arc. Recent studies reveal secular changes in the magmatic activity, magma plumbing system, erupted volumes and magmatic composition associated with the evolution of crust–mantle structures related to the tectonic evolution of the NE Japan arc.

Each volcanic period has its unique pattern of across-arc alkali variations, which are closely related to the thermal structure and the asthenosphere–lithosphere boundary in the uppermost mantle wedge. In the back-arc basin period, the mantle isotherm was very flat, and the back-arc opening by strong extension associated with basin subsidence repeatedly expanded the area of mafic volcanism.

A drastic change from enriched to depleted magma is interpreted as being the result of an asthenospheric injection into the subcontinental lithosphere that occurred in: the continental Sikhote-Alin at 35 Ma; the Oga Peninsula at 22 Ma; the Yamato Basin at 17 Ma; the Aosawa rift at 16.5 Ma; and the inland Kuroko rift area at 15 Ma. These relations are explained by the eastward asthenospheric flow (Nohda 2009). The magmatic evolution in the NE Japan arc in the back-arc basin and island-arc periods would have been related to the change in convection pattern of the mantle wedge from 2D to 3D related to the asthenosphere upwelling and subsequent cooling of the mantle. The geodynamic changes in the mantle would have caused back-arc basin basalt eruptions during the back-arc basin opening (basalt phase) followed by crustal heating and re-melting, which generated many felsic plutons and calderas (rhyolite/granite phase) in the early phase of the island-arc period. This was further followed by crustal cooling and strong trench-normal compression, which ensured vent connections and mixing between deeper mafic and shallower felsic magmas, erupting larger volumes of Quaternary calc-alkaline andesites (andesite phase).

Because of the voluminous magmatic effusions during back-arc basin opening, the temperature of the uppermost mantle beneath the back-arc decreased. Consequently, the lithosphere thickened beneath the back-arc, and the erupted magmatic composition changed from enriched low-\(K\) tholeiites to depleted alkali basalts at 13.5–10 Ma. In
the volcanic front, the composition of magmas evolved from high HFSE magmas to low HFSE magmas, together with a change in theHF isotope ratios. This is explained by fluxed melting, with the addition of fluids from subduction sediments and oceanic crust to the mantle wedge (Hanyu et al. 2006; Yamada et al. 2012).

During the neutral stress conditions at 13.5–10 Ma, felsic magmas ascended diapirically owing to their buoyancy within the ductile lower crust, and in the brittle upper crust the mode of magma ascent changed to that of dykes or sheets. The increase in horizontal tectonic stress at around 10–8 Ma might have formed sills or laccolithic shallow reservoirs in the upper crust followed by two peaks of late Miocene to Pliocene caldera-dominated magmatism, associated with subordinate low-K tholeiites. The depth of the caldera-forming magma reservoirs increased from the late Miocene to the Pliocene, and there was a regional cooling of the upper crust associated with a stress axis change from a NE–SW to an east–west compression.

It is possible that calc-alkaline andesites of the last stage of the island-arc period were formed by magma mixing or co-mingling between deep-seated mafic and shallow felsic end-members. The processes, which could possibly be responsible for producing the felsic end-member of mixed magma, are: fractional crystallization from the mafic end-member magma; melt extraction or re-melting of partially or fully crystallized precursor mafic magma in the middle crust; partial re-melting of mafic end-member material which solidified to hornblende gabbro or amphibolite; and partial re-melting of crustal basement rocks.

In the last stage of the island-arc period, strong compressional stress increased the pressure in the deep magma reservoirs and accelerated the upwelling of deeper mafic end-member of mixed calc-alkaline andesites. As a result, there was double the total erupted volume of mixed andesite magma than that of the felsic caldera-dominated magmatism. Although phenocryst assemblages in calc-alkaline andesites suggest that the H2O contents in magmas tended to increase from frontal-arc to the rear-arc, recent studies show a high water content in frontal primitive basalts. Under strong compressional stress conditions, the excess volatiles which had degassed from the deep-seated mafic magma reservoirs could rise directly upwards or escape trench-side along thrust sheets, forming S-wave reflectors and seismically active focused deformation zones. The strong degassing process of the mafic end-member would have worked more effectively at the volcanic front to the forearc region than in the back-arc side, through the shortening associated with nonlinear viscous flow in the uppermost mantle to the lower crust.

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EVOLUTION OF LATE CENOZOIC MAGMATISM, NE JAPAN ARC

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