Elemental and Sr–Nd–Pb isotopic compositions of late Cenozoic Abaga basalts, Inner Mongolia: Implications for petrogenesis and mantle process

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The Abaga volcanic field is one of the largest and least known areas of late Cenozoic intraplate igneous activities in eastern China. Twenty-seven Abaga basaltic rocks were analyzed for major and trace element contents, Sr–Nd–Pb isotopic compositions, and K–Ar dating. These basalts predominantly consist of alkali basalt with subordinate transitional olivine tholeiite and rare quartz tholeiite. Combining our data of ten K–Ar dates with previously published data showed that the time of the principal volcanic eruption was from the Miocene to the Quaternary (14.57–0.19 Ma). A large amount of volcanic activity is believed to have begun in the northwestern portion of the AVF and gradually migrated southeastwards with geological time.

The Abaga basalts exhibit chondrite-normalized REE patterns, incompatible elements and isotopic ratios (⁸⁷Sr/⁸⁶Sr of 0.703654–0.704286 and ¹⁴³Nd/¹⁴⁴Nd of 0.512845–0.512891) affiliated with oceanic island basalts (OIBs). In general, they have relatively homogenous Pb isotopic compositions (²⁰⁶Pb/²⁰⁴Pb of 18.409–18.521, ²⁰⁷Pb/²⁰⁴Pb of 15.514–15.546, and ²⁰⁸Pb/²⁰⁴Pb of 38.259–38.447), indicating that these lava suites have a similar source. But the alkali basalts are relatively enriched in incompatible elements when compared to the tholeiites. This compositional difference may be attributable to the different degrees of partial melting in the mantle source. The Sr–Nd–Pb isotopic data indicated that the mantle sources of the late Cenozoic Abaga basalts display a DMM–EM2 array similar to those of Southeast Asia. The ²⁰⁸Pb/²⁰⁴Pb vs. ²⁰⁶Pb/²⁰⁴Pb plot was also significantly displaced above the Northern Hemisphere reference line in a pattern clustering near the less-depleted end of the field for the Indian Ocean mid-oceanic ridge basalt (MORB) that shows Dupal signatures. Therefore, we suggest that the Abaga basaltic magmas were dominantly derived from an Indian Ocean MORB-like depleted asthenospheric source with fewer contributions of an enriched mantle component (EM2) from the subcontinental lithospheric mantle.

Keywords: geochronology, geochemistry, Abaga basalts, late Cenozoic, Inner Mongolia

INTRODUCTION

Cenozoic volcanism in East Asia is generally attributed to mantle upwelling induced by passive extension or asthenosphere extrusion following the collision of India into Eurasia or the subduction of the Pacific Plate into the Eurasian Plate (Chung *et al.*, 1994). Previous research on Cenozoic intraplate volcanic rocks of East Asia showed that the basaltic magma may have resulted from the mixing of different proportions of depleted MORB mantle (DMM) or mid-oceanic ridge basalt (MORB) components with an enriched mantle component (EM1 or EM2). Most of these basalts have the Dupal-like Pb isotopic anomaly,

whereas the mantle sources for the basalts have a DMM–EM1 array for North/NE China and a DMM–EM2 array for southeastern Asia (e.g., Xie *et al.*, 1989; Zhou and Zhu, 1992).

In general, most studies have supported the notion of the DMM-end member source being in the asthenospheric mantle, but the origin and occurrence of enriched EM1 and EM2 mantle components are still equivocal. Although many workers (e.g., Song *et al.*, 1990; Basu *et al.*, 1991; Tatsumoto *et al.*, 1992; Zhang *et al.*, 1996) proposed the existence of EM1 and/or EM2 components in the subcontinental lithosphere mantle, recently Choi *et al.* (2006) argued that both EM-type end-members mixing with the DMM may be derived from the shallow asthenospheric mantle. In addition, from available lead isotopic data, the ²⁰⁶Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb ratios of Cenozoic intraplate basaltic rocks decrease from SE China through North China to NE China, and an EM2-type end-member ap-

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Fig. 1. Map showing the distribution of late Cenozoic basaltic rocks and sample locations in the Abaga area, Inner Mongolia (modified from Deng and Macdougall, 1992). The inset shows major tectonic divisions of China (after Zhao et al., 2001), where QDSL and KL denote the Qinling-Dabie Sulu Orogen, and Kunlun Orogen, respectively.

pears to be restricted to the southeastern Asia region (Xie et al., 1989).

The Abaga volcanic field (AVF) and its extensive lava region, the Dariganga Plateau in SE Mongolia, straddling the border between China and Mongolia, are the largest areas of late Cenozoic intraplate lava fields in eastern Asia (Fig. 1). Tectonically, the AVF lies on the Xing'an-Mongolia Orogenic Belt, which is a Paleozoic/early-Mesozoic fold belt between two Archaean continental nuclei: the North China Craton to the south and the Siberian Craton to the north. These extensive basaltic rocks provide a good opportunity to further examine these unsettled issues. For AVF lavas, some K-Ar dating data and

major elemental and Sr-Nd isotopic analyses have been published, but almost no data are yet available on their trace element and Pb isotope compositions in the international literature. Therefore, the purposes of this study were to (1) present the temporal-spatial distribution of the volcanic rocks based on K-Ar ages; (2) identify the geochemical character of the magma source regions; (3) discuss the origin of the Abaga basalts on the basis of their major and trace elements and Sr-Nd-Pb isotope geochemistry; and (4) assess the origin and occurrence of enriched mantle components of late Cenozoic basalts across eastern China.

ANALYTICAL METHODS

Ten samples used for K–Ar dating were free from xenoliths and xenocrysts, and were prepared as follows. Each fresh sample was crushed into fragments of 20–80 mesh in size and 203.5–245.5 mg in weight for the extraction and measurement of argon by isotope dilution, and fragments of an aliquot were ground into powder in an Al₂O₃ Spex mill for the potassium analysis. Isotopic analysis of the extracted argon was performed in substantially modified MM1200 mass spectrometers at State Key Laboratory of Earthquake Dynamics, Institute of Geology, China Earthquake Administration, Beijing, China. Potassium was determined using an atomic absorption spectrophotometer. The decay constants and conversion factors used were those recommended by Steiger and Jäger (1977).

Twenty-seven samples were analyzed for major and trace element abundances. All elements except Si were determined using solutions prepared by dissolving 0.5 g of rock powder in a mixture of ultrapure HF and HNO₃ in Teflon beakers under clean room conditions. The Si solution was prepared by NaOH fusion of 0.05 g of rock powder in nickel crucibles followed by water leaching and HCl acidification. Solutions were analyzed using a Perkin-Elmer Optima 2000DV inductively coupled plasma-atomic emission spectrometer (ICP-AES; for Si, Al, Fe, Mg, Ca, Na, K, Ti, P, and Mn) and a Perkin-Elmer Elan 6100 inductively coupled plasma mass spectrometer (ICP-MS; for Ba, Co, Cr, Cu, Cs, Hf, Li, Nb, Ni, Pb, Rb, Sc, Sr, Th, U, V, Y, Zn, Zr, La, Ce, Pr, Nd, Sm, Eu, Tb, Dy, Ho, Er, Tm, Yb, and Lu) at the National Museum of Natural Science, Taichung and Research Center for Environmental Changes, Academia Sinica, Taipei, Taiwan. Calibration curves were constructed using US Geological Survey rock standards AGV-1, BCR-1, W-1, and G-2, and Geological Survey of Japan rock standard JB-1. Values for these rock standards were adapted from compilations by Govindaraju (1994). The precision was estimated to be around $\pm 2\%$ for the ICP-AES method and better than $\pm 5\%$ for all ICP-MS analyses. Details of the analytical techniques were discussed by Ho (1998) and Ho *et al.* (2003).

The Sr and Nd fractions were separated by ion exchange chromatography, and isotopic compositions were measured using a MAT 262 mass spectrometer at National Cheng Kung University, Tainan, Taiwan. Nd isotopic ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ and are reported relative to 0.511855 for the La Jolla standard. Sr isotopic ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and were relative to 0.710240 for the SRM987 standard. The precisions of the Nd and Sr isotopic compositions were both better than ± 0.000010 (2σ). Details of the analytical procedures are given in Smith and Huang (1997).

In addition, about 100 mg of whole rock powder of each sample was completely decomposed in a mixture of HF–HNO₃ for the Pb isotopic analysis. Pb was separated in Teflon® columns containing $\sim 80\ \mu\text{l}$ AG1-X8, 100–200 mesh and employing HBr–HCl washing and elution procedures. The procedural blank was $<100\ \text{pg}$ for Pb. For measuring the isotopic ratios, Pb was loaded with a mixture of Si-gel and H₃PO₄ onto a single-Re filament and measured at around 1300°C. The NBS 981 Pb standard was used to determine the thermal fractionation, and the measured isotopic ratios of samples were corrected with a value of 0.1% per atomic mass unit. Isotopic ratios were measured on a Finnigan MAT-262 thermal ionization mass spectrometer (TIMS) at the Laboratory for Radiogenic Isotope Geochemistry, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China. Technique details on chemical separation and measurement are described in Chen *et al.* (2000, 2002).

DISTRIBUTION AND K–Ar AGES OF ABAGA BASALTS

The AVF consists of late Cenozoic basaltic rocks and lava platforms covering an area of about 9300 km² (43°20′–45°15′ N, 114°–117°E, Fig. 1). Development of these rocks may have essentially been controlled by a set of NEE- and NW-trending deep faults. Based on the geographic distribution of basaltic lavas, the AVF can be subdivided into two large lava platforms: the Abaga in the northwest and the Hueiterngshilii in the southeast.

On the basis of stratigraphic relationships, four eruptive episodes in the eastern part of Abaga Qi are recognized, which are separated from one another by sedimentary formations. The products of the first, second, and fourth episodes were mainly alkali basalts. In the third episode, the eruptive lavas were dominantly tholeiites (Bureau of Geology and Mineral Resources of Inner Mongolia Autonomous Region, BGMR/IMAR, 1991). These volcanic activities were previously considered to be early Pleistocene. However, K–Ar analyses of basaltic rocks can provide some evidence of eruptive ages, and a limited radiometric dating data set is available for the timing of activity in the AVF. For example, Luo and Chen (1990) reported 21 K–Ar age data from surface outcrop and drilling borehole samples. In order to determine the time of volcanic eruption in different areas, we reviewed the existing chronological data and also selected ten representative surface samples for K–Ar dating. The detailed chronological data of Luo and Chen (1990) and Liu *et al.* (1992) and the analytical results of this study are summarized in Table 1.

Abaga lava platform

Based on the K–Ar ages, it appears that the volcanic activity of the Abaga lava platform commenced around

Table 1. K–Ar ages of whole rock of basalts from the Abaga area, Inner Mongolia

Sample No.	K (%)	$^{40}\text{Ar}_{\text{radiogenic}}$ (mole/g)	$^{40}\text{Ar}_{\text{radiogenic}}$ (%)	Age ($\pm 1\sigma$, Ma)	References
Abaga lava platform					
B47	1.41	3.578E-11	39.02	14.57 ± 0.36	Luo and Chen (1990)
B40	1.16	2.664E-11	7.02	13.20 ± 0.50	Luo and Chen (1990)
B42	1.40	3.124E-11	30.10	12.82 ± 0.43	Luo and Chen (1990)
B45	1.26	2.225E-11	34.25	10.16 ± 0.28	Luo and Chen (1990)
B48	0.84	1.038E-11	11.72	7.11 ± 0.48	Luo and Chen (1990)
B36	0.81	9.370E-12	27.78	6.63 ± 0.25	Luo and Chen (1990)
B37	0.76	8.467E-12	17.59	6.41 ± 0.64	Luo and Chen (1990)
B50	1.29	1.421E-11	19.69	6.34 ± 0.38	Luo and Chen (1990)
IM54A	1.33	1.452E-11	79.92	6.29 ± 0.14	this study
IM55A	0.66	7.031E-12	63.34	6.13 ± 0.21	this study
B39	1.25	1.295E-11	9.72	5.96 ± 0.31	Luo and Chen (1990)
B49	1.01	1.133E-11	27.19	5.93 ± 0.18	Luo and Chen (1990)
X-5	1.63	1.541E-11	48.76	5.44 ± 0.15	Liu <i>et al.</i> (1992)
IM56A	1.30	1.174E-11	72.93	5.20 ± 0.13	this study
IM48A	0.83	6.546E-12	72.19	4.54 ± 0.14	this study
IM57A	0.94	7.289E-12	60.07	4.47 ± 0.15	this study
B36	1.09	7.295E-12	20.09	3.86 ± 0.25	Luo and Chen (1990)
IM49A	0.88	5.143E-12	49.83	3.37 ± 0.07	this study
IM51A	1.73	7.668E-12	67.47	2.55 ± 0.07	this study
Hueiterngshilii lava platform					
B22	1.26	3.384E-11	20.08	15.12 ± 0.92	Luo and Chen (1990)
B30	1.17	6.642E-12	19.39	3.27 ± 0.25	Luo and Chen (1990)
B25	1.11	6.234E-12	11.86	3.24 ± 0.30	Luo and Chen (1990)
B27	0.80	4.461E-12	22.67	3.21 ± 0.23	Luo and Chen (1990)
B26	0.68	3.280E-12	14.12	2.78 ± 0.23	Luo and Chen (1990)
B29	1.21	4.377E-12	13.82	2.08 ± 0.08	Luo and Chen (1990)
BSD-2	1.47	4.708E-12	20.25	1.85 ± 0.08	Liu <i>et al.</i> (1992)
BSD-1	0.97	2.933E-12	18.55	1.74 ± 0.08	Liu <i>et al.</i> (1992)
IM58A	1.16	2.831E-12	53.08	1.41 ± 0.06	this study
B24	1.06	2.170E-12	8.41	1.18 ± 0.20	Luo and Chen (1990)
B21	1.90	4.751E-12	10.29	1.15 ± 0.11	Luo and Chen (1990)
IM45A	1.56	2.717E-12	44.08	1.00 ± 0.02	this study
BSD-3	1.18	1.815E-12	10.39	0.89 ± 0.07	Liu <i>et al.</i> (1992)
B31	1.00	8.163E-13	3.07	0.47 ± 0.23	Luo and Chen (1990)
B28	1.92	1.105E-12	2.05	0.33 ± 0.25	Luo and Chen (1990)
IM44A	2.13	7.125E-13	19.13	0.19 ± 0.01	this study
X-2	1.68	4.704E-12	2.69	0.16	Liu <i>et al.</i> (1992)

Parameters for ^{40}K : $\lambda = 5.543 \times 10^{-10} \text{ year}^{-1}$; $\lambda_e = 0.581 \times 10^{-10} \text{ year}^{-1}$; $\lambda_\beta = 4.962 \times 10^{-10} \text{ year}^{-1}$; $^{40}\text{K}/\text{K} = 1.167 \times 10^{-4} \text{ atom \%}$ (Steiger and Jäger, 1977).

Standard LP-6 analyzed value is $126.5 \pm 1.2 \text{ Ma}$ (recommended value: $128.9 \pm 1.4 \text{ Ma}$).

14.57 Ma and ceased as late as 2.55 Ma, however, the most extensive volcanism took place in the late Miocene (close to and younger than 7.1 Ma). Published data suggest that late Cenozoic volcanism ended in the middle Pliocene ($3.86 \pm 0.25 \text{ Ma}$, Luo and Chen, 1990); however, one K–Ar datum ($2.55 \pm 0.07 \text{ Ma}$) obtained from Abaga Qi of this study indicated that basaltic volcanism continued into the late Pliocene. It is worth noting that no Quaternary basalts are exposed in this region. In fact, the volcanic activity of the Abaga lava platform shows a similar eruptive age and active intensity as those of the Dariganga Plateau (Kononova *et al.*, 2002).

Hueiterngshilii lava platform

Except for one sample from a 170-m deep-drill core (No. B22), with an age of $15.12 \pm 0.92 \text{ Ma}$ (Luo and Chen, 1990), all basaltic rocks in this area erupted in the middle Pliocene ($<3.3 \text{ Ma}$). In Table 1, a basanite yields $0.19 \pm 0.01 \text{ Ma}$ with a similar K–Ar age (0.16 Ma) from Liu *et al.* (1992), which is younger than the previous date of $0.33 \pm 0.25 \text{ Ma}$ by Luo and Chen (1990) deduced from the same dating method. K–Ar data displaying the timing of volcanic activity in the Hueiterngshilii lava platform that may be divided into two episodes: the first occurred in the middle Miocene and the second in the mid-

dle Pliocene to the Quaternary. Middle Miocene basalts have been found in only one drill hole, and these are covered by late Miocene–Quaternary sediments and basaltic rocks. In contrast, a considerably large amount of the second-episodic basaltic rocks are exposed on the surface. Pliocene volcanic activity was quite intense, and it continued into the Quaternary.

In summary, late Cenozoic volcanic activity of the AVF began roughly 15 Ma and continued to as recently as 0.16–0.19 Ma. Basalts from the middle to late Miocene mainly erupted in the Abaga lava platform, whereas most Pliocene volcanism took place in the Hueiterngshilii lava platform, and Quaternary basalts are only concentrated in the southeastern part of the AVF. The younger K–Ar ages of basaltic lava in each platform may constrain the ending time of the magmatism; therefore, it is obvious that the volcanic eruptions migrated from the northwest southeastward with geological time.

BULK CHEMISTRY

Major and minor elements

Most Abaga basaltic rocks contain olivine and clinopyroxene phenocrysts, but some samples are aphyric. The abundances of major elements of the representative basaltic rocks are listed in Table 2. Based on CIPW normative calculations and the Ne–Ol–Di–Hy–Qz tetrahedron of Yoder and Tilley (1962) as well as the classification scheme of alkali basalts proposed by Chih (1988), these samples could be classified into four types (Figs. 2A and B): quartz tholeiite, olivine tholeiite, alkali olivine basalt, and basanite. The most abundant rock type was basanite which constituted about 41% of the rock samples analyzed. The alkali olivine basalt and olivine tholeiite comprised about 26% and 22%, respectively, of the total samples. Only few quartz tholeiites (11%) were found in the AVF, and they fell in the ternary Qz–Di–Hy diagram (Fig. 2A).

Considered as a group, the Abaga basalts show wide variations in major element compositions. For example, analyzed samples had SiO_2 values ranging from 42.37% to 52.83%. Although the alkali basalts and tholeiite had similar Na_2O and K_2O contents, the alkali basalts generally had systematically higher MgO , ΣFeO , CaO , and TiO_2 and lower SiO_2 and Al_2O_3 contents than did the tholeiites. The total alkali content ($\text{Na}_2\text{O} + \text{K}_2\text{O}$) of the Abaga basalts exceeded 3.4%, and the $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios varied from 1.4 to 3.1, which are typical of sodium-rich basaltic rocks.

Several important criteria have been outlined as justification for assuming a possible primary origin. These include the high Mg values (68–75) and Ni (>300 ppm) and Cr (>500 ppm) contents (Frey *et al.*, 1978; Wilkinson and Le Maitre, 1987). However, the Mg values

Fig. 2. (A) Normative compositions of basaltic rocks from the Abaga area. The nomenclature of volcanic rocks followed Yoder and Tilley (1962). (B) Plots of normative olivine against normative nepheline compositions of Abaga alkali basalts. Field boundaries are based on those of Chih (1988).

(57.9–66.3) of the Abaga basalts indicate that they might not represent primitive basalts, but probably have undergone a certain degree of fractional crystallization. Variation diagrams were plotted for the major elements against SiO_2 (Fig. 3), a useful index of differentiation for basaltic magma. It is apparent that ΣFeO , MgO , CaO , and TiO_2 generally decrease and Al_2O_3 increase with increasing SiO_2 contents in the Abaga basalts. These regular variations indicate that the fractional crystallization of Mg-olivine, clinopyroxene, and Ti–Fe oxide minerals may have occurred after the formation of the initial liquid. In these diagrams, no correlation of SiO_2 with Na_2O or K_2O was observed, which suggests that fractional crystallization of feldspar was not significant in these samples.

Table 2. Major (wt%) and trace element (ppm) compositions of Abaga basalts compared with volcanic rocks from the Hannuoba and Wudalianchi-Erkeshan-Keluo areas, North/NE China

Sample No Rock type	IM 44A B	IM 44B B	IM44C B	IM 45A B	IM 45B B	IM 48A AOB	IM 48B AOB	IM 48C B	IM 49A B	IM 49B B	IM 50 OT
SiO ₂	42.62	42.37	42.69	44.22	44.22	43.73	43.89	43.80	44.29	43.57	48.95
Al ₂ O ₃	12.35	12.61	12.42	12.21	12.66	12.11	12.46	11.71	12.07	12.22	11.46
ΣFeO	13.38	13.49	13.64	13.48	13.35	12.82	12.91	12.67	11.65	11.79	11.09
MgO	10.23	10.37	10.10	10.15	10.30	9.07	9.07	10.36	9.56	10.16	9.38
CaO	9.86	9.84	10.29	9.91	9.79	10.17	10.05	10.47	10.83	10.93	9.05
Na ₂ O	3.99	4.08	3.95	3.54	3.67	2.41	2.98	2.65	3.45	3.28	2.77
K ₂ O	2.61	2.53	2.45	1.93	1.87	0.99	0.96	1.96	1.14	1.25	1.51
TiO ₂	3.44	3.44	3.50	3.26	3.27	3.67	3.69	3.54	2.83	2.78	2.60
P ₂ O ₅	0.99	0.97	1.11	0.79	0.79	0.81	0.83	0.81	0.98	0.97	0.45
MnO	0.184	0.182	0.194	0.183	0.182	0.177	0.241	0.192	0.244	0.277	0.149
L.O.I.	0.10	0.20	0.11	0.10	0.30	3.22	2.50	1.50	2.30	2.51	1.51
Total	99.754	100.082	100.454	99.773	100.402	99.177	99.581	99.662	99.344	99.737	98.919
MG	61.66	61.78	60.90	61.29	61.87	59.81	59.64	63.23	63.31	64.44	64.01
Li	8.6	8.7	8.7	7.1	7.8	34.8	24.3	8.3	12.8	13.5	8.5
Sc	13.4	13.7	19.9	15.2	14.6	15.6	15.4	14.9	17.9	18.0	20.3
V	209	213	235	216	205	217	216	202	205	199	222
Cr	164	162	162	219	196	209	200	201	300	299	322
Co	53	52	64	52	50	53	54	52	56	56	55
Ni	202	213	308	222	221	227	228	241	260	302	263
Cu	65	57	58	69	58	73	63	58	65	63	48
Zn	134	140	125	134	132	133	137	132	141	145	110
Rb	53.6	52.6	49.2	39.4	37.1	16.5	17.0	42.9	14.4	14.4	31.5
Sr	948	956	543	836	844	888	870	781	1001	993	819
Y	25.2	25.4	26.9	24.3	23.6	24.9	24.6	23.7	27.7	26.9	22.3
Zr	242	244	272	223	218	208	204	197	224	220	226
Nb	71	72	69	58	57	55	54	52	65	64	41
Cs	0.57	0.61	0.52	0.44	0.46	0.33	0.52	0.39	0.38	0.45	0.46
Ba	663	673	655	531	540	863	771	677	664	672	575
La	65.7	66.8	50.3	52.8	52.2	53.6	52.3	49.5	49.8	49.6	31.6
Ce	120.5	123.4	90.1	100.0	98.6	102.1	100.4	95.5	97.2	98.6	62.8
Pr	14.3	14.6	12.4	12.2	11.9	12.1	11.8	11.3	11.6	11.4	7.6
Nd	51.9	52.4	43.9	44.6	43.3	44.2	43.2	41.6	42.2	41.9	29.7
Sm	13.00	13.26	10.00	11.54	11.18	11.83	11.52	10.97	10.39	10.39	7.91
Eu	4.00	4.08	3.25	3.54	3.44	3.76	3.63	3.42	3.17	3.17	2.68
Tb	1.71	1.73	1.31	1.56	1.50	1.55	1.52	1.46	1.32	1.31	1.06
Dy	8.03	8.27	6.44	7.63	7.39	7.48	7.29	7.01	6.21	6.27	5.30
Ho	1.34	1.38	1.03	1.31	1.20	1.28	1.24	1.19	1.03	1.03	0.90
Er	3.21	3.30	2.49	3.21	3.15	3.08	3.03	2.88	2.51	2.52	2.23
Tm	0.41	0.42	0.31	0.43	0.42	0.41	0.40	0.38	0.32	0.33	0.30
Yb	2.11	2.17	1.68	2.27	2.23	2.17	2.09	2.01	1.73	1.74	1.66
Lu	0.28	0.29	0.22	0.31	0.30	0.30	0.29	0.27	0.24	0.24	0.23
Hf	8.53	8.76	5.80	7.94	7.71	6.97	6.78	6.48	5.39	5.52	6.09
Pb	4.8	4.7	4.1	4.6	5.0	3.6	3.1	3.5	3.0	3.1	4.4
Th	7.95	8.22	7.36	6.15	6.21	5.76	5.47	5.20	5.84	5.82	4.42
U	2.25	2.26	1.94	1.66	1.67	1.35	1.32	1.33	1.62	1.80	1.13
(La/Yb) _N	22.34	22.08	21.48	16.68	16.79	17.72	17.95	17.67	20.65	20.45	13.66

The major element compositions of the alkali basalts and tholeiites from the AVF together with volcanic rocks from the Hannuoba (Zhi *et al.*, 1990) and Wudalianchi-Erkeshan-Keluo (WEK, Zhang *et al.*, 1995) are plotted in Fig. 3 for comparison. Intraplate magmatism of the Hannuoba in North China and WEK in NE China has been well established, and these are two type examples of late

Cenozoic sodium and potassic volcanic fields, respectively. In general, no significant major element compositional difference appears to exist between basalts from the Abaga and Hannuoba areas, with the exception of the alumina and titania contents. In Fig. 3, most samples collected from the AVF had higher TiO₂ and lower Al₂O₃ values at a given SiO₂ concentration compared to

Table 2. (continued)

Sample No Rock type	IM 51A B	IM 51B B	IM51C B	IM 54A AOB	IM 54B AOB	IM54C AOB	IM 55A OT	IM 55B OT	IM55C OT	IM 56A AOB	IM 56B OT
SiO ₂	44.54	44.34	45.64	46.94	47.42	46.54	50.63	48.93	48.34	46.84	50.68
Al ₂ O ₃	12.12	11.54	11.60	11.46	11.21	11.54	11.84	11.90	12.41	11.74	13.21
ΣFeO	11.60	11.91	11.72	12.60	12.79	12.78	12.82	11.58	11.80	12.23	11.52
MgO	10.51	10.99	10.94	10.02	9.66	9.63	8.72	7.97	8.33	9.00	7.94
CaO	10.66	10.34	10.16	9.51	9.70	9.83	9.16	10.13	10.54	9.50	8.65
Na ₂ O	3.36	3.05	3.29	3.02	2.93	2.96	2.95	2.99	2.96	2.68	3.02
K ₂ O	2.18	2.05	2.04	1.61	1.52	1.51	1.08	1.09	1.02	1.65	1.12
TiO ₂	2.85	2.92	2.80	3.30	3.33	3.39	2.73	2.61	2.59	3.13	2.73
P ₂ O ₅	1.09	1.13	1.07	0.75	0.76	0.87	0.43	0.43	0.51	0.63	0.42
MnO	0.17	0.17	0.17	0.175	0.191	0.198	0.174	0.177	0.214	0.155	0.158
L.O.I.	0.96	1.28	0.88	0.32	0.30	0.10	0.18	1.30	0.39	2.48	0.40
Total	100.04	99.72	100.31	99.705	99.811	99.348	100.714	99.107	99.124	100.035	99.848
MG	65.58	65.99	66.25	62.58	61.37	61.31	58.86	59.14	59.75	60.75	59.18
Li	9.8	11.6	10.2	7.2	12.1	10.4	7.8	7.3	7.9	7.8	7.7
Sc	19.5	18.3	19.6	21.7	22.3	24.9	20.6	22.8	26.0	18.2	20.3
V	235	223	223	242	260	244	203	205	203	210	196
Cr	492	457	461	340	353	301	229	283	260	270	254
Co	54	54	53	58	60	61	58	52	57	53	50
Ni	262	282	272	338	378	353	309	282	289	337	313
Cu	52	51	34	69	73	52	63	71	65	63	62
Zn	125	125	124	167	178	160	188	166	168	183	202
Rb	40.6	35.5	36.0	29.0	26.4	23.9	20.7	21.1	14.2	31.7	22.0
Sr	1016	955	956	815	816	513	487	485	446	688	497
Y	28.3	27.0	26.9	31.9	31.1	30.8	23.9	25.2	24.5	26.8	26.6
Zr	247	237	239	259	245	255	172	174	172	247	186
Nb	76	74	73	48	46	43	30	30	27	49	28
Cs	0.56	0.48	0.51	0.23	0.20	0.21	0.08	0.08	0.07	0.19	0.60
Ba	597	602	598	417	434	412	283	289	313	421	289
La	52.2	52.5	53.6	43.6	42.7	37.7	23.2	24.4	21.2	40.0	25.5
Ce	100.5	100.4	101.5	88.6	86.3	73.1	48.8	51.5	42.5	80.6	53.9
Pr	12.9	12.7	12.9	11.2	10.9	10.5	5.9	6.2	5.8	9.9	6.6
Nd	52.7	52.5	53.3	41.6	40.7	37.6	24.1	25.2	21.9	37.0	26.3
Sm	10.27	10.13	10.15	11.61	11.44	9.75	7.19	7.48	6.23	9.98	7.78
Eu	3.17	3.11	3.16	3.56	3.57	3.24	2.37	2.45	2.21	3.10	2.51
Tb	1.24	1.22	1.21	1.66	1.64	1.39	1.15	1.20	1.00	1.40	1.24
Dy	5.94	5.79	5.80	8.19	8.26	7.21	6.12	6.37	5.45	7.09	6.59
Ho	0.99	0.96	0.95	1.40	1.39	1.20	1.06	1.12	0.93	1.20	1.17
Er	2.38	2.27	2.29	3.46	3.50	2.98	2.68	2.81	2.35	2.99	2.95
Tm	0.29	0.28	0.28	0.47	0.46	0.39	0.38	0.40	0.32	0.41	0.42
Yb	1.65	1.57	1.59	2.52	2.51	2.15	2.10	2.18	1.80	2.17	2.32
Lu	0.20	0.19	0.19	0.35	0.34	0.29	0.29	0.30	0.24	0.29	0.33
Hf	5.12	4.84	4.86	7.92	7.78	5.62	5.40	5.56	3.79	7.83	6.03
Pb	5.0	3.8	3.5	3.1	3.2	3.1	2.3	2.1	2.1	6.8	4.3
Th	6.58	6.08	6.07	4.55	4.31	4.05	3.08	3.21	2.70	5.80	4.22
U	1.73	1.53	1.55	1.24	1.19	1.08	0.83	0.89	0.74	1.46	0.95
(La/Yb) _N	22.69	23.99	24.18	12.41	12.20	12.58	7.92	8.03	8.45	13.22	7.88

$\Sigma\text{FeO} = \Sigma\text{Fe}_2\text{O}_3/1.1$, and assuming that $\text{Fe}_2\text{O}_3 = 0.2\text{FeO}$ (Middlemost, 1989).

$\text{MG}(\text{Mg-value}) = 100 * \text{Mg}/(\text{Mg} + \text{Fe}^{+2})$.

Rock types: QT, Quartz tholeiite; OT, Olivine tholeiite; AOB, Alkali olivine basalt; B, Basanite.

Hannuoba basalts. In contrast, WEK volcanic rocks are highly potassic and considerably differ in chemical characteristics from late Cenozoic basalts of eastern China (Zhang *et al.*, 1995). At a given SiO₂ content, therefore, the potassic volcanic rocks from the WEK area display the highest K₂O and lowest ΣFeO concentrations.

Rare earth elements (REEs) and other trace elements

Table 2 also lists REE and other trace element abundances. Chondrite-normalized REE patterns of the Abaga basalts are shown in Fig. 4. All of the Abaga basalts had similar REE patterns (Fig. 4), exhibiting subparallel smooth curves from La to Lu and displaying moderate to

Table 2. (continued)

Sample No Rock type	IM56C AOB	IM 57A QT	IM57B QT	IM 58A QT	IM58B OT	IM 46A*	IM 46B*	IM46C*	H**	WEK***
SiO ₂	46.40	52.83	50.90	52.26	49.38	53.86	53.60	52.70	44.14~51.81	43.47~55.08
Al ₂ O ₃	11.80	12.21	12.96	12.87	13.39	14.40	14.44	14.97	13.54~16.11	10.55~14.52
ΣFeO	13.18	10.55	11.20	10.86	11.74	11.24	11.01	11.88	9.88~13.54	7.19~9.72
MgO	9.00	6.89	7.57	7.84	8.69	3.80	3.71	4.06	5.69~10.62	3.86~14.07
CaO	9.51	8.02	8.80	8.91	9.47	5.90	5.72	5.61	7.49~10.4	5.05~10.33
Na ₂ O	2.64	2.79	2.82	3.02	3.03	4.70	4.81	4.62	2.04~5.75	2.95~4.42
K ₂ O	1.52	0.99	0.96	1.28	1.26	0.88	0.79	0.92	0.59~3.43	3.58~6.96
TiO ₂	3.09	2.57	2.63	2.52	2.58	2.78	2.72	2.67	1.76~2.89	2.06~3.03
P ₂ O ₅	0.74	0.37	0.45	0.37	0.46	0.41	0.44	0.57	0.28~1.27	0.66~1.3
MnO	0.190	0.128	0.151	0.147	0.172	0.210	0.196	0.229	0.11~0.21	0.09~0.17
L.O.I.	2.39	1.80	1.89	0.11	0.10	2.73	2.22	2.61		
Total	100.460	99.148	100.331	100.187	99.612	100.910	99.656	100.829		
MG	58.95	57.87	58.70	60.29	60.89	41.56	41.48	41.82		
Li	7.4	5.6	5.4	8.0	8.6	15.1	15.0	14.1		
Sc	19.8	20.4	23.0	20.7	22.8	21.4	21.7	33.2	9.4~22	8.9~22
V	202	196	190	203	199	241	241	286	103~199	102~222
Cr	222	228	200	232	203	2	20	4	77~240	34~487
Co	57	47	51	53	57	20	21	28	40~61	22~51
Ni	329	275	240	262	257	10	18	11	69~272	43~360
Cu	51	67	43	70	62	31	28	29		23~55
Zn	185	185	182	186	184	187	122	129	99~198	84~124
Rb	26.6	18.3	12.9	23.8	17.2	24.0	17.6	27.9	3.2~62.9	70.7~132
Sr	507	458	431	523	451	305	168	366	400~1595	1104~2030
Y	25.6	27.6	25.8	21.2	20.5	40.4	40.0	45.6	16~24	18.9~29.8
Zr	244	186	180	159	156	164	158	174	115~389	269~695
Nb	43	26	22	28	24	8	8	7	15~100	29~101
Cs	0.17	0.20	0.16	0.13	0.13	3.83	3.44	3.99		
Ba	415	242	233	284	295	227	180	226	196~1136	1363~2152
La	34.4	23.2	19.8	20.6	17.8	19.9	21.1	17.5	11.7~67	60~125
Ce	66.5	49.9	40.9	43.7	36.5	53.0	56.3	41.8	26~144	122~231
Pr	9.3	6.1	5.6	5.3	5.0	6.8	7.3	6.3		14~25
Nd	33.2	24.8	21.1	22.0	18.9	26.8	28.1	23.1	14.9~58	54~91
Sm	8.40	7.70	6.22	6.44	5.26	8.96	9.24	7.31	4.2~11.9	9.3~15
Eu	2.80	2.46	2.19	2.11	1.87	2.76	2.85	2.44	1.52~3.73	2.5~4.1
Tb	1.18	1.25	1.01	0.99	0.82	1.73	1.80	1.44	0.61~1.27	0.7~1.6
Dy	6.11	6.79	5.63	5.24	4.39	10.42	10.59	8.99		3.8~6.2
Ho	1.00	1.20	1.00	0.92	0.75	2.12	2.16	1.77		0.7~1.2
Er	2.49	3.06	2.50	2.29	1.89	5.82	5.84	4.92		1.6~2.6
Tm	0.32	0.44	0.35	0.34	0.26	0.90	0.91	0.75		
Yb	1.79	2.40	1.96	1.84	1.49	5.18	5.17	4.46	0.83~1.75	0.89~1.7
Lu	0.24	0.34	0.27	0.26	0.21	0.76	0.76	0.65	0.11~0.25	0.12~0.25
Hf	5.39	6.06	4.06	4.95	3.35	5.82	5.70	4.03	2.8~8.4	4.5~16
Pb	3.9	3.4	3.3	3.8	3.3	10.6	8.4	11.7		7~23
Th	5.15	3.88	3.27	3.36	2.73	4.21	4.27	3.93	0.9~8.6	4~12
U	1.25	0.92	0.77	0.68	0.53	1.21	1.23	1.10		0.8~1.9
(La/Yb) _N	13.78	6.93	7.25	8.03	8.57	2.76	2.93	2.81	5.71~52.98	43.04~85.36

*Mesozoic basaltic andesites (unpublished data); **Hannuoba basalts (Zhi *et al.*, 1990); ***WEK potassic volcanic rocks (Zhang *et al.*, 1995).

steep sloping with light REE (LREE) enrichment. In general, the alkali basalts were relatively enriched in LREEs compared to the tholeiitic basalts. The (La/Yb)_N ratios increase with the undersaturated character from 6.9~8.0 in the quartz tholeiites, to 7.9~13.7 in the olivine tholeiites, 12.2~18.0 in the alkali olivine basalts, and 16.7~22.3 in the basanites. As a whole, the LREE abun-

dance in the Abaga basalts was higher than those of E-type and N-type MORBs, but was comparable with oceanic island basalts (OIBs; Sun and McDonough, 1989). In Fig. 4, the absence of a negative Ce anomaly can be observed. This possibly suggests that these rocks were generally not affected by low-temperature alterations (Zou *et al.*, 2000). In addition, no negative Eu anomaly was

Fig. 3. Major elements vs. SiO_2 plots for the Abaga basalts. The chemical data for the volcanic rocks of Hannuoba (Zhi et al., 1990) and Wudalianchi-Erkeshan-Keluo (WEK) areas (Zhang et al., 1995) are shown for comparison.

found, indicating that plagioclase fractionation was insignificant during the generation and evolution of these magmas.

Other incompatible elemental contents are also good indicators of different basaltic suites. Most alkali-rich

basalts show higher Ba, Hf, Nb, Rb, Th, U, and Zr variation trends than do tholeiitic suites. In general, the significance of incompatible element abundances in basaltic rocks is its relationship to the degree of partial melting.

Fig. 4. Chondrite-normalized rare earth element distribution of representative basaltic rocks from the Abaga area, compared to OIB, E-MORB and N-MORB (data and the normalizing values are from Sun and McDonough, 1989). Note that the samples show strong LREE-enriched patterns with $(La/Yb)_N$ ratios of 6.93–24.18.

The trace element contents of the Abaga basaltic rocks show similar abundances and ranges of variability compared to those of the Hannuoba area, except for significantly higher Cr, Ni, and V contents for the some Abaga basaltic rocks (Table 2). They have an average La content of 38 ppm which approximates those of the average basalt from oceanic islands (37 ppm, Sun and McDonough, 1989). Compared to WEK potassic volcanic rocks, the most distinctive feature of the Abaga basalts is that they are depleted in incompatible elements (e.g., Ba, Rb, Sr, Zr, and LREEs) and highly enriched in compatible elements (e.g., Co, Cr, Ni, Sc, and V). In addition, the slight to moderate depletion of Nb on the spider diagram patterns displayed by WEK potassic rocks distinguishes them from Abaga basaltic rocks, which show Nb enrichment similar to OIBs.

Fig. 5. Plots of $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ for the Abaga basalts. Data for the late Cenozoic basalts of Hannuoba (Song *et al.*, 1990) and Fujian-Taiwan regions (Chung *et al.*, 1995; Zou *et al.*, 2000, 2004) are shown for comparison. The oceanic island basalt (OIB) field is from Zindler and Hart (1986).

Sr–Nd–Pb isotopic composition

Nine samples from each of the lava platforms encompassing the full range of chemical diversity were selected for Sr–Nd isotope analysis (Table 3). As a group, basaltic rocks of the AVF clustered within narrow ranges of Sr–Nd isotopic composition with $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.703654–0.704286 and $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.512845–0.512891. Our results closely agree with Sr–Nd isotope analyses of ten Abaga Qi-Xilinhot basalts reported by Deng and Macdougall (1992). On the $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ diagram (Fig. 5), the basalts show depleted characteristics and values of ϵ_{Sr} and ϵ_{Nd} are negative and positive, respectively. In Fig. 5, various basaltic rocks are shown together with OIBs for comparison. The Sr and Nd isotopic composition data from the Abaga basalts are similar to those of the Hannuoba alkali basalts and Fujian–Taiwan basalts (Song *et al.*, 1990; Chung *et al.*, 1995; Zou *et al.*, 2000, 2004), but they have lower $^{87}\text{Sr}/^{86}\text{Sr}$ and higher $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than most Hannuoba tholeiites. In addition, $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the Abaga basalts are within the range of OIBs (Zindler and Hart, 1986) and are closely related to the mantle array.

Lead isotopes were determined in five basaltic rocks with the following compositional range: $^{206}\text{Pb}/^{204}\text{Pb}$ of 18.409–18.521, $^{207}\text{Pb}/^{204}\text{Pb}$ of 15.514–15.546, and $^{208}\text{Pb}/^{204}\text{Pb}$ of 38.259–38.447 (Table 3). In the $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 6A), the Abaga basalts lie to the right of the 4.55-Ga geochron. In the $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 6B), the Abaga data points lie well above and generally subparallel to the Northern Hemisphere reference line (NHRL) of Hart (1984), possessing a feature known as the DUPAL Pb isotopic anomaly.

Table 3. Sr–Nd–Pb isotopic compositions of representative basaltic rocks from the Abaga area, Inner Mongolia

Sample No.	Rock type	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	ϵ_{Nd}^* (0 Ma)	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$	$\Delta 7/4^{**}$	$\Delta 8/4^{***}$
IM44A	B	0.703654 ± 12	0.512887 ± 12	+4.9	18.448	15.515	38.273	+2.5	+34.2
IM45A	B	0.703989 ± 12	0.512869 ± 11	+4.5					
IM48C	B	0.704286 ± 12	0.512880 ± 11	+4.7	18.455	15.514	38.259	+2.3	+31.9
IM50	OT	0.704191 ± 10	0.512858 ± 11	+4.3					
IM51A	B	0.704261 ± 14	0.512845 ± 11	+4.0	18.409	15.525	38.280	+4.0	+39.7
IM54A	AOB	0.704020 ± 12	0.512889 ± 11	+4.9					
IM55A	OT	0.704062 ± 11	0.512891 ± 11	+4.9	18.521	15.546	38.447	+4.8	+42.8
IM56A	AOB	0.704179 ± 12	0.512876 ± 10	+4.6					
IM58A	QT	0.704271 ± 12	0.512854 ± 10	+4.2	18.514	15.544	38.389	+4.7	+37.8

* $\epsilon_{\text{Nd}} = ((^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}} / (^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} - 1) \times 10^4$; $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}^{0 \text{ Ma}} = 0.512638$.

** $\Delta 7/4 = ((^{207}\text{Pb}/^{204}\text{Pb})_{\text{DS}} - (^{207}\text{Pb}/^{204}\text{Pb})_{\text{NHRL}}) \times 100$; $(^{207}\text{Pb}/^{204}\text{Pb})_{\text{NHRL}} = 0.1084(^{206}\text{Pb}/^{204}\text{Pb}) + 13.491$ (Hart, 1984).

*** $\Delta 8/4 = ((^{208}\text{Pb}/^{204}\text{Pb})_{\text{DS}} - (^{208}\text{Pb}/^{204}\text{Pb})_{\text{NHRL}}) \times 100$; $(^{208}\text{Pb}/^{204}\text{Pb})_{\text{NHRL}} = 1.209(^{206}\text{Pb}/^{204}\text{Pb}) + 15.627$ (Hart, 1984).

DISCUSSION

Geochemical characteristics of the Abaga basalts and implications for their mantle sources

(A) *Crustal contamination?* Since the crustal components probably played a significant role in the genesis of volcanism in some regions, e.g., some Vietnamese basalts (Hoang *et al.*, 1996) and Changbaishan trachytic rocks (Xie and Wan, 1992), we used Sr-isotope and trace element ratios such as Nb/U, Rb/Ba, and Rb/Sr to assess the effects of crustal contamination during magmatic ascent.

In general, Sr-isotope ratios are sensitive to crustal contamination, especially from radiogenic continental crust composition. Like the Penghu Island basalts, $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the present study do not show a good correlation with SiO_2 contents (Fig. 7), and together with the low Sr isotopic ratios in the Abaga basaltic rocks, significant contamination by upper continental materials can be ruled out. Remarkable differences in trace element ratios also exist between Abaga basalts and the continental crust. For example, the average Nb/U ratio of Abaga basaltic rocks (41) is close to average values of OIBs and MORBs (of 47 ± 10 , Hofmann *et al.*, 1986). This ratio is significantly higher than the value for the average continental crust (12, Taylor and McLennan, 1985) or upper continental crust (4, Condie, 1993). In addition, Rb/Ba (0.018–0.084; average, 0.061) and Rb/Sr (0.014–0.091; average, 0.041) ratios of the Abaga basalts are also lower than those of the continental crust (Rb/Ba of 0.128 and Rb/Sr of 0.123) and upper continental crust (Rb/Ba of 0.204 and Rb/Sr of 0.32, Taylor and McLennan, 1985).

The Sr-isotope and trace element ratios precluded the possibility that the magmas had been contaminated by crustal materials before they erupted onto the surface. In fact, upper mantle xenoliths such as spinel lherzolite were

found within alkali olivine basalt and basanite of the AVF (Deng and Macdougall, 1992), indicating that the magma may have rapidly ascended from deep levels and did not reside at crustal levels for a long time. Therefore, we suggest that the elemental and isotopic characteristics of the Abaga basalts are basically features of their mantle source region.

(B) *High-Ti basalts* The Abaga basalts have relatively high TiO_2 contents (2.52–3.69 wt%) and high Ti/Y (558–899), Nb/Y (0.85–2.83), and Nb/La (1.03–1.46) ratios which are similar to chemical features of high-Ti basalts in continental flood basalt (CFB) provinces of the world (Fig. 8). Similar geochemical features can also be observed in late Cenozoic basalts from Chifeng and Fanshi (Han *et al.*, 1999; Tang *et al.*, 2006), where high-Ti basaltic magmas appear in the Archean North China Craton. In general, high-Ti basalts are abundant and widespread in the main CFB provinces, e.g., Parana, Siberia, Deccan, Karoo, and Ethiopian. Geochemical studies of these CFB provinces have shown that the high field strength elements (HFSE) contents of the lavas are related to the nature of the mantle sources and/or crustal contaminants (Lightfoot *et al.*, 1990, 1993; Peate and Hawkesworth, 1996; Stewart and Rogers, 1996; Pik *et al.*, 1998, 1999). A high-Ti content has commonly been interpreted as evidence for the involvement of asthenospheric mantle source components in the petrogenesis of the magmas (Barry *et al.*, 2003). Accordingly, the high-Ti characteristic of the Abaga basaltic magma indicates an origin in the deep mantle and of a product of partial melting of the asthenosphere.

(C) *OIB-like geochemical characteristics* Trace element ratios of volcanic rocks have been found to be useful for separating different sources when discussing the petrogenesis of the magmas (Wilson, 1989). Figure 9 shows variations of Ce/Nb vs. Ce. The Abaga basalts are

Fig. 6. $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams for the Abaga basalts. Fields for the Wudalianchi (Zhang *et al.*, 1998; Zou *et al.*, 2003), Hannuoba (Song *et al.*, 1990; Basu *et al.*, 1991), Fujian-Taiwan (Chung *et al.*, 1995; Zou *et al.*, 2000, 2004), South China Sea seamount (Tu *et al.*, 1992), Indian Ocean MORB, and Pacific and North Atlantic MORBs (Barry and Kent, 1998; Zou *et al.*, 2000) are shown for comparison. Fields of EM1 and EM2 are adopted from Zindler and Hart (1986). Symbols are same as those in Fig. 5.

essentially characterized by Ce-enrichment, but low Ce/Nb ratios relative to those of island-arc and mid-ocean ridge basalts, and fall within the OIB field. Similarly, other trace element ratios such as Zr/Nb (3.20–8.18), Zr/Y (6.74–10.13), Th/Nb (0.08–0.15), and La/Nb (0.69–0.97) calculated for the Abaga basalts from the chemical data listed in Table 2 are also close to average ratios of OIBs.

Primitive mantle-normalized incompatible element diagrams of the Abaga basalts are quite similar to those of typical OIBs (Sun and McDonough, 1989), although a few alkali basalts show “troughs” for Rb and K relative to neighboring elements (Fig. 10). Small depletions of Rb found in certain samples, e.g., IM-48A, 48B, and 49A, usually have slightly high loss of ignition values, indi-

Fig. 7. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vs. SiO_2 contents for the Abaga basalts compared to other late Cenozoic volcanic rocks. Fields for the Penghu Islands are from Lee (1994). The data for Vietnam are from Hoang *et al.* (1996).

cating that they may have been affected by alterations in the post-eruption stage. A few basanites and alkali olivine basalts have slight negative K anomalies, but their Ba/Rb ratios are similar to those of intraplate basalts (~12, Sun and McDonough, 1989), which may be interpreted as being due to K-bearing minerals in the residual phases (Ho *et al.*, 2003). In general, if the degree of partial melting of the mantle source increases, then the percentage of partial melting of K-bearing minerals in the mantle source will also increase. Therefore, Ho *et al.* (2003) suggested that tholeiitic basalts have lower incompatible element contents than alkali basalts and show no distinct negative K anomalies which may have been derived from a larger degree of partial melting.

LREEs are highly incompatible in the mantle-melt system, and LREE ratios should be close to those of the mantle source under a batch melting condition. In chondrite-normalized REE patterns (Fig. 4), the Abaga basalts exhibit strong LREE enrichment with $(\text{La}/\text{Yb})_{\text{N}}$ varying from 6.9 to 22.3. In addition, the La/Ce (0.47–0.56) and Ce/Sm (6.5–10.0) ratios of these samples also appear to be higher than those of primitive mantle (0.39 and 4.0, respectively, Sun and McDonough, 1989), strongly reflecting their derivation from a fertile mantle source.

Fig. 8. Nb/La and Nb/Y ratios vs. Ti/Y for Abaga basalts compared to continental flood basalt (CFB) provinces. (1) Low-Ti basalts from Parana, Siberia, and the Poladpur unit of Deccan; (2) high-Ti basalts from Siberia; (3) high-Ti basalts from southern Ethiopian and Yemen traps; (4) high-Ti basalts from north-western Ethiopia. Data are from Lightfoot *et al.* (1990, 1993), Peate and Hawkesworth (1996), Stewart and Rogers (1996), and Pik *et al.* (1998, 1999). MORB and OIB compositions are from Sun and McDonough (1989).

Fig. 9. Ce/Nb vs. Ce diagram illustrating the OIB-like character of late Cenozoic Abaga basalts. The fields of OIB, MORB, and island-arc basalt (IAB) are adopted from Ryerson and Watson (1987).

Fig. 10. Spider diagrams showing primitive mantle-normalized incompatible element abundances for selected basaltic rocks from the Abaga area. Note that the samples are enriched in highly incompatible trace elements similar to those of OIB. OIB, E-MORB, and N-MORB data and trace element abundances of the primitive mantle are from Sun and McDonough (1989).

Lead isotopic characterization of the mantle source

Xie *et al.* (1989) argued that eastern China basalts display geographically controlled variations in the Pb isotopic compositions whereby the $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios decrease from SE China through North China to NE China. In the $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams (Figs. 7A and B), lead isotopic compositional variations of Wudalianchi (Zhang *et al.*, 1998; Zou *et al.*, 2003), Hannuoba (Song *et al.*, 1990; Basu *et al.*, 1991), Fujian-Taiwan (Chung *et al.*, 1995; Zou *et al.*, 2000, 2004), South China Sea seamount (Tu *et al.*, 1992), Indian Ocean MORB, and Pacific and North Atlantic MORB (Barry and Kent, 1998; Zou *et al.*, 2000) are also plotted for comparison. It should be noted that the radiogenic Pb isotopic ratios of the Abaga basalts are distinctly higher than the data for other volcanic rocks from North and NE China (e.g., Wudalianchi, Hannuoba),

but they cluster near the less-depleted end of the field for Indian Ocean MORBs, and overlap with previously published data for late Cenozoic basalts from SE China (e.g., Fujian-Taiwan and the South China Sea seamount) extending toward the EM2 end-member. The difference in Pb isotopic compositions between Abaga basalts and other North/NE China volcanic rocks is remarkable since the EM1-signature has consistently been observed in these areas.

Source of the EM2 component

In Inner Mongolia, many episodes of continental magmatism have occurred since the late Paleozoic, following closure of the Paleo-Asian Ocean (BGMR/IMAR, 1991). Based on geological investigations, considerable amounts of volcanic rocks were formed during late Paleozoic, Mesozoic, and Neogene–Quaternary times. However, there are significant variations in volcanic rock compositions due to changes in geodynamic regimes from the late Paleozoic to the late Cenozoic. In the late Carboniferous–early Permian, the AVF was near the continental margin, with a developing island arc and back-arc basin (Ruzhenstev and Pospelov, 1992; Wiechert *et al.*, 1997). Therefore, in the Paleozoic and Mesozoic eras, there were many volcanic rocks of calc-alkaline affinity in the AVF; this observation is supported by major and trace element compositions of basaltic andesite samples (~180 Ma, unpublished data) of IM46A, IM46B, and IM46C (Table 2). These samples, for example, have a significantly negative Nb anomaly on the spider diagram with an average Nb/U ratio of 6.5, which is similar to that of the upper continental crust. We have found no available isotopic data on the continental crust of the AVF. But Mesozoic basaltic andesites in general exhibit higher Pb ratios ($^{206}\text{Pb}/^{204}\text{Pb}$ of 18.822, $^{207}\text{Pb}/^{204}\text{Pb}$ of 15.589, and $^{208}\text{Pb}/^{204}\text{Pb}$ of 38.730, unpublished data) than those of the Abaga basalts (Fig. 8).

Late Cenozoic intraplate alkali basaltic rocks and basaltic pyroclastics captured a few spinel peridotite xenoliths in the AVF. In general, the peridotite xenoliths may provide evidence of the nature, compositions, and geological processes of the continental lithospheric mantle. Wiechert *et al.* (1997) studied LREE-enriched clinopyroxenes and melt pockets of the Dariganga lava plateau, a region adjacent to the AVF, and obtained Sr–Nd isotope signatures consistent with derivation from a mixture of DMM and EM2 sources. In addition, low Cu and S contents and high $\delta^{34}\text{S}$ values in whole-rock peridotites were also explained as interactions with oxidized fluids that may have been derived from subducted oceanic crust (Wiechert *et al.*, 1997). The lack of a negative Nb anomaly on the spider diagrams of Abaga basaltic rocks (Fig. 10), cannot be explained by sediment contamination of the asthenospheric mantle. We, therefore,

Fig. 11. $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams for the Abaga basalts. The fields for the Wudalianchi, Changbaishan, Hannuoba, Fujian-Taiwan, South China Sea seamount, Indian Ocean MORB and Pacific and North Atlantic MORBs are shown for comparison. Wudalianchi and Changbaishan are from Basu *et al.* (1991), ETO from Jahn (1986) and Chung and Sun (1992); other data sources are the same as those in Fig. 6. The fields of DMM, EM1 and EM2 are adopted from Zindler and Hart (1986).

suggest that the enriched mantle component (of the EM2 type) in Abaga basalts was caused by expelled fluid from sediments of the paleo-subduction plate, and the source of EM2 component may have been derived from the sub-continental lithosphere.

Petrogenesis of late Cenozoic Abaga basalts

A mantle plume with depleted isotopic signatures has been proposed as the cause of intraplate magmatism in NE China (Zhou *et al.*, 1988; Song *et al.*, 1990). Deng *et al.*

al. (2004) further adopted the “continental roots-plume tectonics” model to describe the configuration and dynamic condition of the subcontinental lithosphere and upper mantle beneath China, and it was proposed that the Abaga Qi-Zhangjiakou volcanic field was one “subplume” (defined as a second-order feature that developed from the primary plume). However He isotopic data showed no evidence of a high- $^3\text{He}/^4\text{He}$ mantle plume involved in NE China volcanism (Chen *et al.*, 2007). Furthermore, because of the lack of unequivocal tomographic images delineating distinct, deep-seated thermal anomalies that could be interpreted as plumes, Choi *et al.* (2006) suggested that the late Cenozoic alkalic magmatism in East Asia originated in the shallow asthenosphere. But a few Dariganga xenoliths have high $^3\text{He}/^4\text{He}$ ratios which are similar to those found in Hawaiian hot-spot basalts; thus, Ionov and Wood (1992) suggested that the Dariganga Plateau may be an intracontinental hot-spot.

The Indian Ocean MORB or the East Taiwan Ophiolite was suggested as being the depleted mantle component of East Asian basalts (Jahn, 1986; Chung and Sun, 1992; Zhou and Mukasa, 1997). Plots of $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ for the Abaga basalts (Fig. 11) are used to depict the binary mixing of the Indian Ocean MORB (DMM) and EM2-type component. It is clear that Sr–Pb and Nd–Pb isotopic correlations of the Abaga basalts fall outside but close to the range of Indian Ocean MORBs. Therefore, we suggest that the upwelling asthenospheric mantle was the main source for the Abaga basalts; in contrast, the enriched (EM2-type) mantle component may have only partially mixed with the magma.

The bulk chemistry of selected samples from the AVF can be subdivided into alkali basalt and tholeiite suites. The general geochemical features of the tholeiites are similar to alkali basalts, but a remarkable feature is that the tholeiites are more depleted in LREEs and other incompatible elements. This difference is probably related to their mantle source and related geological processes. However, limited Sr–Nd–Pb isotopic data have not shown significant isotopic variations between various rock types in the AVF. These isotopic characteristics are comparable to those found in basalts from the Penghu Islands, Taiwan Strait (Fig. 7). Therefore, the isotopic data suggest a similar mantle reservoir for both the alkali basalt and tholeiite magma, although they might not be directly related.

There is a general continuum of major and trace element compositions between tholeiites and alkali basalts. The mechanisms giving rise to successive compositional changes in volcanic rocks are believed to depend on the degree of partial melting of the mantle source and on crystal-liquid fractionation processes at shallow depths. Tholeiitic magmas are thought to originate through relatively large amounts of partial melting at relatively shallow depths in the mantle, whereas alkalic magmas are

generally agreed to originate from lesser amounts of partial melting at relatively deeper levels in the mantle. In addition, based on major- and trace-element studies, these samples can be produced from parental magma by fractional crystallization of olivine, clinopyroxene, and Fe–Ti oxide minerals under high-pressure conditions. Therefore, a variety of magmas in the AVF could have been produced by various amounts of partial melting and fractional crystallization during ascent.

Spatial lead isotopic provinciality of late Cenozoic volcanic rocks across eastern China

During the Cenozoic, eastern China was in a rift setting, where divergent plate motion and extensional structures dominated (Ma and Wu, 1987). Therefore, intraplate basalts strongly erupted during the Neogene and are widely distributed in eastern China. A heterogeneous mantle source has been proposed for the Cenozoic basalts from eastern China on the basis of Sr–Nd–Pb isotope data. Based on the Pb isotopic ratios of volcanic rocks, late Cenozoic samples from eastern China have been divided into two major groups: one representing a mixture of the DMM and EM1 components, and the other a mixture of DMM and EM2 (e.g., Basu *et al.*, 1991; Zhang *et al.*, 1991; Tu *et al.*, 1992; Chung *et al.*, 1994; Tang *et al.*, 2006). However, this general feature is questioned recently by Chen *et al.* (2007) who argued that not all basalts from NE China can be explained by mixing of DMM with EM1. Sr–Nd–Pb isotopic data of potassic basalts from NE China were interpreted to result from mixing between a FOZO end-member and a LoMu end-member.

The identification of isotopically different mantle components and the discussion of their distributions and origins are important to our understanding of mantle evolution and magma genesis (Zhang *et al.*, 1991). In eastern Asia, the enriched component EM1 and/or EM2 types or LoMu end-member may have been derived from the subcontinental lithosphere. Late Cenozoic basalts in SE China, including those of Zhejiang-Fujian, the Leiqiong area, and the South China Sea Basin, have high $^{208}\text{Pb}/^{204}\text{Pb}$ (37.610~39.260) and $^{207}\text{Pb}/^{204}\text{Pb}$ (15.437~15.649) ratios which plot above the NHRL, indicating a Dupal anomaly in the EM2 direction (Peng *et al.*, 1986; Zhu and Wang, 1989; Basu *et al.*, 1991; Tu *et al.*, 1991, 1992; Lan *et al.*, 1994; Zou *et al.*, 2000). Chung *et al.* (1994) suggested that the EM2 component derived from the continental lithospheric mantle was related to the subduction of the Pacific plate beneath the Eurasian plate in the Mesozoic. Ho *et al.* (2003) further pointed out that the EM2 component of the basalts may have been derived from the continental lithospheric mantle such as Tungchihsu group II pyroxenitic rocks.

In contrast, the EM1 signatures require a mantle source with a time-integrated history of depletion in U relative to Pb and enrichment in REE and Rb relative to HREEs

Fig. 12. Outline of the mantle source components of late Cenozoic basaltic rocks from East Asia (modified after Choi *et al.*, 2006). Distributions of the basalts are after Hoang *et al.* (1996), Barry *et al.* (2003), Ho *et al.* (2003), and Choi *et al.* (2006). 1, Wudalianchi-Erkeshan-Keluo; 2, Dariganga; 3, Abaga; 4, Chifeng; 5, Hannuoba; 6, Fanshi; 7, Datong; 8, Changbaishan; 9, Baengnyeong Is.; 10, Ganseong; 11, Ulleung Is.; 12, Dok Is.; 13, Jogokni; 14, Jeju Is.; 15, NW Taiwan; 16, Penghu Islands.

and Sr, respectively. Basu *et al.* (1991), Zhang *et al.* (1991), and Xu *et al.* (2005) suggested that late Cenozoic basalts in North China can be attributed to mixing between DMM and EM1 reservoirs. The EM1 component has commonly been thought to be related to an ancient subcontinental lithosphere (Song *et al.*, 1990; Basu *et al.*, 1991) or to the metasomatized lower portion of an old subcontinental lithosphere (Tatsumoto and Nakamura, 1991). On the other hand, based on trace element and isotopic compositions of basaltic rocks from UDGJ areas (Ulleung Island–Dok Island–Ganseong area–Jogokni) on the middle and northeastern portions of the Korean Peninsula and within the Sea of Japan, which have OIB-like characteristics, Choi *et al.* (2006) ruled out the East Asia Cenozoic volcanic rocks having an origin in the mantle lithosphere. Because these basaltic rocks have a vertical array on the Pb–Sr and Pb–Nd isotopic correlation diagrams and neither lithospheric province is composed of the mantle end members, EM1 or EM2 (Choi *et al.*, 2006), they advocated that the spatial distributions of the two large-scale distinct mantle domains, the DMM-EM1 and DMM-EM2 arrays, represent shallow asthenospheric mantle sources.

The distributions of isotopic data and mantle components indicate that the mantle sources of Cenozoic basalts in East China and neighboring regions are rather complicated (Figs. 11 and 12). New Pb isotopic data obtained in this study show that AVF lava in Inner Mongolia has SE China basalt-like DMM-EM2 array characteristics. Additionally, it is noteworthy that Miocene basalts in northwestern Taiwan, located at the SE China continental margin, exhibit an EM-type lead isotope signature (Chung *et al.*, 1995). Early Miocene basalts (23–20 Ma) have uniform Sr–Nd–Pb isotopic compositions comparable with those of the other SE China basalts pointing toward the EM2 component, while late Miocene basalts (13–9 Ma) show distinct isotopic characteristics indicating additional involvement of an EM1-type mantle source. In comparison with basaltic rocks of UDGJ areas, those of NW Taiwan show significant isotopic variations, from DMM-EM2 to DMM-EM1, at about 7 Ma. It is difficult to use a model of “two large-scale distinctive asthenosphere mantle domains” to explain their origin. In fact, postulating a heterogeneous mantle or layered mantle from investigations of ultramafic xenoliths in alkali basalts from different regions in the world has been suggested by many authors (e.g., Ottonello *et al.*, 1980; Nohda *et al.*, 1991; Tatsumoto *et al.*, 1992; Griffin *et al.*, 2004). Thus Chung *et al.* (1995) proposed that the EM1 and EM2 components of Miocene basalts in NW Taiwan may have resided at different levels of the continental lithospheric mantle.

CONCLUSIONS

The late Cenozoic volcanic activity of the AVF was entirely basaltic and occurred as relatively large eruptions widely dispersed in space and time. Based on K–Ar dating, incipient volcanism may have taken place in middle Miocene times (at around 15 Ma) and gradually increased in tempo toward the late Miocene Epoch. Volcanic activities were most extensive during the Pliocene to early Pleistocene, and declined and ended in the late Pleistocene (the youngest is 0.16 Ma). It was found that a systematic relationship appears to exist between the ages of the basalt eruptions and geographic locations. Large amounts of volcanic activity are believed to have begun in the northwest portion of the AVF and gradually migrated southeastwards with geological time. In the Quaternary, volcanism was limited to the southeast portion of this region.

Abaga lavas are comprised of basanites and alkali olivine basalts, with subordinate olivine tholeiites as well as rare quartz tholeiites. In comparison with the tholeiites, alkali basalts are notably enriched in incompatible elements. These differences are common in intraplate basaltic rocks and most likely resulted from different degrees of partial melting from a mantle source. The TiO_2 con-

tents of these rocks are consistently higher than 2.5%, and the Na₂O/K₂O ratios of all samples studied exceeded unity, indicating their High-Ti and sodium-enriched character. On the whole, the geochemical features of these rocks show a strong affinity to OIBs, which is especially exemplified by their trace element abundances, REE patterns, and Sr–Nd–Pb isotopic ratios.

The Abaga basalts have uniformly depleted Sr and Nd isotopic ratios, and there is no obvious geochemical evidence for significant crustal contamination from the continental crust during magma generation. Therefore, the chemical and isotopic compositions of these basaltic rocks may reflect those of the mantle source regions. From available lead isotopic data of intraplate Cenozoic volcanic rocks from East China, it was reported that EM2 appears to be restricted to the SE China region. However, the Abaga basaltic lavas indicate that the distribution of EM2 extends into the Inner Mongolia area. Although the ⁸⁷Sr/⁸⁶Sr vs. ¹⁴³Nd/¹⁴⁴Nd data (Fig. 5) show depleted characteristics, the Sr–Pb and Nd–Pb plots (Fig. 11) clearly demonstrate that additional mixing of a limited EM2-type component is required during magma generation. In the AVF area, prior to the Cenozoic continental magmatism, it is believed that subduction of the paleo-plate underneath the Xing'an–Mongolian Orogenic Belt may have been a major tectonic process over a long period of time. The lithospheric mantle source may have been modified by sediments associated with this paleo-subduction zone. We thus suggest that the EM2-type isotope signature of Abaga basalts was derived from the continental lithosphere mantle, and these rocks can be attributed to mixing processes between the two sources: asthenosphere-derived plume magmas with little contamination by the subcontinental lithospheric materials en route to the surface.

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