Mantle transition zone-derived EM1 component beneath NE China: Geochemical evidence from Cenozoic potassic basalts

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A B S T R A C T

The isotopic characteristics of the sub-oceanic mantle are well established, but in continental regions these properties are usually obscured, and therefore controversial, because of the potential effects of crustal contamination together with lithospheric mantle metasomatism and melting. The so-called EM1 (Enriched Mantle-1) signature, characterized by low 206Pb/204Pb and 143Nd/144Nd ratios, is particularly problematic in this respect because EM1-type OIB sources are commonly attributed to recycled crust and/or lithospheric mantle. In this paper we show that a suite of Cenozoic potassic basalts from NE China displays many previously unrecognized correlations between chemical and isotopic parameters, which tightly constrain the isotopic characteristics of an extreme EM1-type mantle source located in the asthenosphere. Its radiogenic isotopes are similar to, but even more extreme than, those of the oceanic endmember composition represented by the Pitcairn hotspot, namely 206Pb/204Pb ≤ 16.5, 143Nd/144Nd ≤ 0.5123 (or εNd ≤ −6.4), 176Hf/177Hf ≤ 0.2825 (or εHf ≤ −10.1). These characteristics require a source of recycled crustal material of Precambrian age (2−2.2 Ga). An important new constraint is the Mg isotopic composition of δ28Mg (≤−0.6‰), which is lower than normal mantle (δ28Mg = −0.25 ± 0.07‰) and lower crustal values (δ28Mg = −0.29 ± 0.15‰), but consistent with sedimentary carbonate (δ28Mg = −5.57‰ to −0.38‰). The endmember EM1 source produced high-SiO2 melts with low MgO, CaO/Al2O3 and δ28Mg values, exceptionally high K/U ≈ 50,000, Ba/Th ≈ 400, low U/Pb ≈ 0.06, and positive Zr and Hf anomalies. The chemical and isotopic parameters of this potassic basalt suite form binary mixing arrays, one end point of which is the inferred asthenospheric EM1 reservoir, whereas the other is a more ordinary, depleted mantle component, which is also sampled by local lithospheric mantle xenoliths. These binary arrays include well-developed correlations between Sr, Nd, Hf, Pb and Mg isotopes, negative correlations of 206Pb/204Pb with K2O, K/U, Hf/Hf*, positive correlations of δ28Mg with MgO, and 143Nd/144Nd with Fe2O3/T and U/Pb.

We propose that the EM1 reservoir contains recycled ancient carbonate-bearing sediments, subducted into the mantle transition zone, where K, Rb, Ba and Pb are sequestered by K-hollandite as suggested by Murphy et al. (2002) for the Gaussberg lamproites. Loss of small amounts of carbonate melt extracted Th, U and some of the LREE, while retaining K, Rb, Ba, Pb, Zr and Hf in the residue, thereby generating the observed trace element anomalies. In Cenozoic time, this deep EM1 reservoir ascended into the shallow asthenosphere and underwent low-degree partial melting, at pressures below the stability field of K-hollandite, thereby releasing K, Rb and Ba into the melt. The partial melts ascended through subcontinental lithosphere and were progressively modified by interaction with the lithospheric mantle, thus accounting for the linear chemical and isotopic trends noted above. This interaction imposed a progressively more depleted character on the erupted melt, both in terms of isotopic composition and trace element enrichment.

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1. Introduction

‘EM1 (Enriched Mantle-1)’ is Zindler and Hart’s (1986) nomenclature for oceanic basalts with particularly low 206Pb/204Pb and...
$^{143}\text{Nd}/^{144}\text{Nd}$ along with intermediate $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The Cenozoic potassic basalts in NE China (Fig. 1) are distinguished by some of the least radiogenic Pb and Nd isotopes seen anywhere in the world, and they can be regarded as continental equivalents of the EM1 type of oceanic basaltic volcanism seen, for example, on Pitcairn Island (Fig. 2). While the origin of the sub-oceanic EM1 signature is still debated, a gradual consensus is evolving that it is generated by recycling of continental material, either in the form of subducted sediments (e.g. Woodhead and McCulloch, 1989; Eisele et al., 2002; Delavault et al., 2016) or lower crustal material (Hanan et al., 2004; Willbold and Stracke, 2010), although delaminated and recycled subcontinental lithosphere has also been considered (McKenzie and O’Nions, 1983; Gibson et al., 2005; Class and le Roex, 2006; Geldmacher et al., 2008). By contrast, the origin of the EM1 signature under continental regions, as expressed by potassic basalts in NE China, has been the subject of intense debate.

Over the past decades, a bewildering number of models have been proposed to account for the EM1 characteristics of the potassic basalts in NE China. A clear majority of authors invoke some form of lithospheric metasomatism, and declare the metasomatized lithospheric mantle to be the immediate source of the erupted potassic melts (Basu et al., 1991; Zhang et al., 1991, 1995, 1998; Zou et al., 2003; Chen et al., 2007; Chu et al., 2013; Sun et al., 2014, 2015), even though none of the Chinese mantle xenoliths analyzed so far actually come close to the isotopic compositions required to confirm this proposition (e.g. Tatsumoto et al., 1992; Choi et al., 2005; Zhang et al., 2011). The suggested metasomatic agents include melts derived from ancient subducted sediments in the mantle transition zone (MTZ) (Sun et al., 2014, 2015), delaminated ancient lower crust (Chu et al., 2013), and an unspecific reservoir in the deep asthenosphere (Zhang et al., 1991, 1995; Zou et al., 2003). In any case, all these authors argue that none of the asthenospheric melts reach the surface directly. However, they do not give specific reasons why the ultimate melts should come from the lithospheric mantle, and we suspect that this interpretation has become an article of faith hardened by frequent repetition. In continental regions, the work of McKenzie (1989) provided a rationale for this by postulating the movement of extremely small melt fraction forming veins in the lithosphere. These veins would have low melting points, and melt could be generated by “activating” the lithosphere either through the impingement of a plume from below or simply through lithospheric extension and thinning (McKenzie and Bickle, 1988). Remelting of these veins plus some interaction with the surrounding peridotite might explain many of the exotic features displayed by highly enriched, potassic magmas (e.g. Foley, 1992).

In addition to the lithospheric metasomatism model outlined above, a few authors have postulated direct asthenospheric melt sources for the EM1-type potassic basalts. Choi et al. (2006) argued generally in favor of asthenospheric sources for both EM1 and EM2 type basalts occupying separate domains beneath East Asia. Zhao et al. (2014) suggested a K-rich, delaminated lithosphere, located in the asthenosphere as the source of the EM1-type melts. However, these authors provided little information about the actual origin of the EM1 source for the potassic basalts. Kuritani et al. (2013) pro-
posed an asthenospheric source based on the inferred high melting temperatures, and they suggested that the potassic- and EM1 characteristics originated from the MTZ which had been metamutli- atized by K-rich sediment fluids through a stagnation of an ancient slab. However, they did not attempt to explain how the strong fractionation of elements with similar incompatibility (e.g., K/Th) in the potassic basalts can be achieved during their multi-stage processes.

The contributions cited above have typically focused on specific geochemical characteristics deemed to be important for explaining the origin of the potassic EM1-type basalts. This means that there has been no comprehensive overview of the combined major plus trace element and isotopic characteristics of the entire suite of these basalts. Because of this, some in our opinion important features have escaped notice. In this study, we discuss new chemical and Sr–Nd–Pb–Hf–Mg isotopic data in the context of a comprehensive set of published data on all five suites of potassic rocks. On the basis of the combined chemical and isotopic variations seen in these potassic suites, we propose a new model in which the unique chemical characteristics of the EM1 source component are initially generated by partial melting processes in the presence of residual K-hollandite and majorite in the MTZ, but the binary correlations between chemical and isotopic compositions result from the interaction between asthenospheric potassic (EM1-type) melts and the depleted subcontinental lithospheric mantle.

2. Geological setting and sample description

2.1. Geological setting

The Greater Khingan Range and the Songliao Basin belong to the Xing'an–Mongolia Orogenic Belt (XMOB) (Fig. 1a). The XMOB is the eastern segment of the Central Asian Orogenic Belt, which formed by a series of subduction and collision events between the Siberia and Baltica cratons to the north and the Tarim and North China cratons to the south (e.g. Xiao et al., 2015). Continued subduction of the Paleo-Pacific plate during the early Mesozoic caused northwest movement of the north China–Mongolia block, and induced a compressional tectonic regime in NE China (e.g. Wang et al., 2006). During the late Mesozoic the tectonic regime changed from compression to extension, resulting in asthenospheric upwelling, magmatic underplating, and the formation of rift basins (Zorin, 1999; Wu et al., 2002; Meng, 2003). From then on, magmas erupted in close association with the development of graben basins and regional boundary faults in NE China (Liu et al., 1994).

Cenozoic intraplate basalts in NE China are mainly distributed along the north margin and flanks of the Songliao Basin (Fig. 1b). Among them, five potassium-rich intracontinental monogenetic volcanic fields (Wudalianchi, Erkeshan, Keluo, Nuominhe and Xiaogulihe) crop out at the boundary between the northwestern margin of the Songliao Basin and the Greater Khingan Range (Fig. 1b). The volcanic rocks in these fields cover an area of >3000 km² and were mainly erupted during the Quaternary (Zhang et al., 1998; Zhao et al., 2014). Mantle-derived spinel lherzolite and harzburgite xenoliths have been reported from all the potassium-rich volcanic fields except Xiaogulihe (Zhang M. et al., 1998, 2000; Zhang Y.L. et al., 2011; Sui et al., 2012), whereas garnet peridotite xenoliths, which are derived from ~70 km depth based on P−T estimates, have been found only in Nuominhe (Sui et al., 2012). According to the summary of Liu et al. (2016), the relative thicknesses of lithosphere beneath the potassic volcanic fields in NE China are as follows: Erkeshan < Wudalianchi (~120 km) < Keluo < Nuominhe (~150 km). Due to a lack of local geophysical data, the thickness of the lithosphere beneath Xiaogulihe is unclear.

2.2. Sample description

We collected seventeen fresh potassic basalts from the Wudalianchi volcanic field for whole-rock analysis including major and trace elements, and Sr, Nd, Pb, Hf isotopes. Eight of these samples, plus six samples from Nuominhe, were also selected for whole-rock Mg isotopic analysis. The samples analyzed in this study are very fresh with low loss on ignition values (~0.12 to 0.16 wt.%), and observations of thin sections revealed no evidence of postmagnetic alteration. The Mg isotope analytical procedures were established by An et al. (2014). All the analytical methods are given in the Supplementary Material.

The Wudalianchi samples have aphyric or porphyrytic texture and vesicular structure. Minor olivine and clinopyroxene phenocrysts (~15%) are dispersed in the groundmass consisting of volcanic glass, olivine, augite, feldspar microlites, and rare oxide minerals. Some olivine phenocrysts show euhedral and skeletal morphology. Clinopyroxene phenocrysts are euhedral or subhedral with diameters varying mainly from 0.3 to 1 mm. The petrography, elemental and Sr–Nd–Pb–Hf isotopic compositions of the six Nuominhe samples have been reported elsewhere (Liu et al., 2017), and a brief summary is given in the Supplementary Material.

3. Results

The results of the major and trace elements, and Sr–Nd–Pb–Hf isotopic analyses of Wudalianchi basalts are given in Table S1. For comparison, we compiled published elemental and Sr–Nd–Pb–Hf isotopic data for potassic (Wudalianchi, Erkeshan, Keluo and Nuominhe) and ultrapotassic rocks (Xiaogulihe) from NE China (Table S2). The magnesium isotopic compositions of the Wudalianchi and Nuominhe potassic basalts are listed in Table 1.

3.1. Whole rock major, trace elements and Sr–Nd–Pb–Hf isotopes

Wudalianchi rock samples have high contents of SiO₂ (48.51–53.83 wt.%) and high K₂O/Na₂O ratios (5.5–1.62), low to medium concentrations of MgO (5.20–8.17 wt.%), and low CaO contents (4.85–7.56 wt.%). These samples are classified as phonotephrite, tephriphonolite, basaltic trachyandesite and trachyandesite (Fig. S1). MgO correlates positively with Fe₂O₃ (total iron) and CaO/Al₂O₃, and negatively with Al₂O₃ (Fig. S2). SiO₂ and K₂O (not shown). TiO₂ remains constant with the variation of MgO (Fig. S1). Notably, Wudalianchi basalts have lower MgO, Fe₂O₃ (total iron) and CaO/Al₂O₃, and higher SiO₂, K₂O and Al₂O₃ than Nuominhe basalts (Figs. S1, S2).

Compatible trace elements (Ni, Cr, Sc and V) correlate positively with MgO (Fig. S2), and no correlation can be found between most incompatible elements and MgO. All samples exhibit strong LREE (light rare earth element) enrichments relative to the HREE (heavy rare earth element) ([La/Yb] = 55.3–97.8), with no obvious Eu anomaly. In a primitive-mantle-normalized element diagram, Wudalianchi samples show positive Ba, K, Pb and Sr anomalies and negative Th, U, Nb, Ta and Ti anomalies (Fig. 3). However, Zr and Hf vary widely from negative to positive anomalies. Compared to the Wudalianchi basalts, Nuominhe high-MgO potassic basalts show less enrichment in K, Ba and LREEs, and negative anomalies in Pb, Zr and Hf (Fig. S3).

Wudalianchi samples show enriched Sr–Nd–Hf isotopic compositions and extremely unradiogenic Pb isotopic compositions (87Sr/86Sr = 0.7051–0.7054, εNd = −3.7 to −5.7, εHf = −5.1 to −8.4; 206Pb/204Pb = 16.835–17.103, 207Pb/204Pb = 15.435–15.450, 208Pb/204Pb = 36.786–37.137), close to the EM1 endmember (Fig. 2). The new data in this study, combined with literature data for the entire suite of potassic basalts in NE China, show negative correlations on plots of 87Sr/86Sr versus 143Nd/144Nd
Table 1
Magnesium isotopic compositions of potassic basalts from Wudalianchi and Nuominhe, Northeast China.

<table>
<thead>
<tr>
<th>Sample name</th>
<th>$^{25}\text{Mg}$</th>
<th>2SD</th>
<th>$^{26}\text{Mg}$</th>
<th>2SD</th>
<th>N</th>
<th>$\Delta^{26}\text{Mg}^{c}$</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Wudalianchi basalts</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>WDLC01</td>
<td>−0.217</td>
<td>0.029</td>
<td>−0.451</td>
<td>0.049</td>
<td>5</td>
<td>0.018</td>
</tr>
<tr>
<td>WDLC02</td>
<td>−0.238</td>
<td>0.027</td>
<td>−0.456</td>
<td>0.031</td>
<td>5</td>
<td>0.000</td>
</tr>
<tr>
<td>WDLC03</td>
<td>−0.242</td>
<td>0.032</td>
<td>−0.474</td>
<td>0.041</td>
<td>8</td>
<td>0.004</td>
</tr>
<tr>
<td>WDLC05</td>
<td>−0.201</td>
<td>0.044</td>
<td>−0.573</td>
<td>0.046</td>
<td>8</td>
<td>0.007</td>
</tr>
<tr>
<td>WDLC05 replicate</td>
<td>−0.281</td>
<td>0.023</td>
<td>−0.569</td>
<td>0.052</td>
<td>8</td>
<td>0.015</td>
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<tr>
<td>WDLC06</td>
<td>−0.249</td>
<td>0.039</td>
<td>−0.490</td>
<td>0.050</td>
<td>5</td>
<td>0.007</td>
</tr>
<tr>
<td>WDLC07</td>
<td>−0.229</td>
<td>0.016</td>
<td>−0.466</td>
<td>0.035</td>
<td>5</td>
<td>0.014</td>
</tr>
<tr>
<td>WDLC07 replicate</td>
<td>−0.240</td>
<td>0.042</td>
<td>−0.466</td>
<td>0.043</td>
<td>4</td>
<td>0.003</td>
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<tr>
<td>WDLC08</td>
<td>−0.254</td>
<td>0.033</td>
<td>−0.496</td>
<td>0.033</td>
<td>7</td>
<td>0.004</td>
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<tr>
<td>WDLC13</td>
<td>−0.251</td>
<td>0.027</td>
<td>−0.503</td>
<td>0.039</td>
<td>5</td>
<td>0.011</td>
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<tr>
<td>WDLC13 replicate</td>
<td>−0.253</td>
<td>0.014</td>
<td>−0.493</td>
<td>0.021</td>
<td>4</td>
<td>0.003</td>
</tr>
<tr>
<td><strong>Nuominhe basalts</strong></td>
<td></td>
<td></td>
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<tr>
<td>13NM05</td>
<td>−0.195</td>
<td>0.024</td>
<td>−0.385</td>
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<td>5</td>
<td>0.006</td>
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<td>13NM13</td>
<td>−0.183</td>
<td>0.033</td>
<td>−0.379</td>
<td>0.032</td>
<td>4</td>
<td>0.014</td>
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<td>13NM18</td>
<td>−0.193</td>
<td>0.030</td>
<td>−0.396</td>
<td>0.030</td>
<td>4</td>
<td>0.014</td>
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<tr>
<td>13NM26</td>
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<td>0.029</td>
<td>−0.344</td>
<td>0.035</td>
<td>4</td>
<td>−0.001</td>
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<tr>
<td>13NM27</td>
<td>−0.232</td>
<td>0.037</td>
<td>−0.433</td>
<td>0.032</td>
<td>4</td>
<td>−0.004</td>
</tr>
<tr>
<td>13NM27 replicate</td>
<td>−0.217</td>
<td>0.041</td>
<td>−0.434</td>
<td>0.049</td>
<td>4</td>
<td>0.009</td>
</tr>
<tr>
<td>13NM34</td>
<td>−0.192</td>
<td>0.018</td>
<td>−0.381</td>
<td>0.036</td>
<td>4</td>
<td>0.007</td>
</tr>
</tbody>
</table>

*a* $\delta^{26}\text{Mg} = ((^{25}\text{Mg} - ^{24}\text{Mg})_{\text{sample}} / (^{25}\text{Mg} - ^{24}\text{Mg})_{\text{SM }	ext{C}} - 1) \times 1000$, where $X = 25$ or 26 and DSM3 is a Mg standard solution made from pure Mg metal (Galy et al., 2003).

*b* 2SD = twice the standard deviation of the population of N (N ≥ 4) repeat measurements of a sample solution.

$c$ $\Delta^{26}\text{Mg} = \delta^{26}\text{Mg} - 0.521 \times \delta^{25}\text{Mg}$, where $\delta^{25}\text{Mg} = 1000 \times \ln(^{25}\text{Mg} / ^{26}\text{Mg}) / 1000$, $X = 25$ or 26 (Young and Galy, 2004).

$d$ Replicate denotes repeating sample dissolution, column chemistry and instrumental analysis.

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Fig. 2. Variations in (a) $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$, (b) $^{176}\text{Hf}/^{177}\text{Hf}$ vs. $^{143}\text{Nd}/^{144}\text{Nd}$, (c) $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$, (d) $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ for potassic basalts from NE China. Data for Cenozoic sodic basalts (Fig. S1) in NE China are shown in (a), (c) and (d) for comparison, and the data for Halaha sodic basalts (Ho et al., 2013) are separated and highlighted. The six Nuominhe samples (Liu et al., 2017) which we analyzed Mg isotopes in this study are distinguished from others by using different legend (solid diamond). We delineate the ranges for Keeloo mantle xenoliths, and the data for these xenoliths are from Zhang et al. (2011). The detailed references for the published data of potassic rocks and sodic basalts can be found in the Supplementary Material. Data source for EM1-type (Pitcairn islands) and EM2-type oceanic island basalts (Samoa islands): http://georoc.mpch-mainz.gwdg.de/georoc, the detailed references are listed in the Supplementary Material. The average Sr, Nd and Hf isotopic compositions for DMM (Depleted MORB Mantle) are from Workman and Hart (2005) and references therein.
and $^{206}$Pb/$^{204}$Pb, and a positive correlation of $^{143}$Nd/$^{144}$Nd versus $^{206}$Pb/$^{204}$Pb (Fig. 2). Basalts from WEK (Wudalianchi–Erkeshan–Keluo) exhibit higher $^{87}$Sr/$^{86}$Sr, and lower $^{143}$Nd/$^{144}$Nd, $^{206}$Pb/$^{204}$Pb and $^{176}$Hf/$^{177}$Hf than those from Nuominhe. It is important to note that the Sr–Nd–Hf isotopic compositions of all the potassic basalts are enriched compared to those of lithospheric mantle xenoliths collected from Keluo, which show typically depleted Sr–Nd–Hf isotopic compositions (Fig. 2). Finally, the radiogenic isotope ratios of the potassic basalts are correlated with certain major elements ($K_2O$, $Fe_2O_3T$; Fig. 4) and incompatible trace element ratios ($K/UF$, $Hf/Hf^\ast$; $Hf/Hf^\ast = Hf/\left(SmN \times NdN\right)^{1/2}$, where the subscript N denotes primitive-mantle normalized; Fig. 5).

3.2. Magnesium isotopes

The Wudalianchi and Nuominhe basalts have $\delta^{26}$Mg values ranging from $-0.57\%$ to $-0.34\%$. In a plot of $\delta^{25}$Mg vs. $\delta^{26}$Mg (Fig. S3), all samples and the USGS standards fall along the mass-dependent fractionation line for Mg isotopes (Young and Galy, 2004). The $\Delta^{25}$Mg values of these samples range between $-0.004$ and $0.018$ (Table 1). The $\delta^{26}$Mg values of the potassic basalts in this study are significantly lower than those of the terrestrial mantle (average $\delta^{26}$Mg = $-0.25 \pm 0.07\%$, Teng et al., 2010; or $\delta^{26}$Mg = $-0.23 \pm 0.04\%$, Lai et al., 2015; Fig. 6). However, they are comparable to the Cenozoic sodic basalts from eastern China ($\delta^{26}$Mg = $-0.60\%$ to $-0.35\%$, Yang et al., 2012; Huang et al., 2015). The $\delta^{26}$Mg values ($\delta^{26}$Mg = $-0.57\%$ to $-0.45\%$) of the newly analyzed Wudalianchi samples are similar to the published data from Wudalianchi and Erkeshan ($\delta^{26}$Mg = $-0.57\%$ to $-0.46\%$, Tian et al., 2016) (Fig. 6), but clearly lower than the Nuominhe samples ($\delta^{26}$Mg = $-0.44\%$ to $-0.34\%$) (Fig. 6).

4. Discussion

In the following, we first assess possible effects of fractional crystallization and crustal contamination on the chemical compositions of the potassic basalts, and then discuss the nature of the observed chemical and isotopic correlations and their endmember compositions. In particular, we consider specifically how the
Fig. 5. Variations in K/U and Hf/Hf∗ versus 87Sr/86Sr and 206Pb/204Pb for potassic rocks from NE China. In (b) and (d), the pale yellow regions represent ranges of K/U and Hf/Hf∗ ratios for Keluo mantle xenoliths (data from Zhang et al., 2011). For DMM, the average K/U (19,000) ratio is from Arevalo et al. (2009), and the Hf/Hf∗ ratio (1.06; Hf/Hf∗ = HfN/(SmN × NdN)0.5, where the subscript N denotes primitive-mantle-normalized) is calculated based on the data of Workman and Hart (2005). Other data sources and symbols as in Fig. 2. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 6. Variations in δ26Mg versus (a) MgO, (b) SiO2, (c) 87Sr/86Sr, (d) 143Nd/144Nd for potassic basalts from NE China. The published Mg isotope data for Wudalianchi and Erkehebasan basalts are from Tian et al. (2016), and the major elements and Sr–Nd isotope data of these samples are from Chu et al. (2013). Low-δ26Mg sodic basalts from eastern China are shown for comparison. The major elements and Sr–Nd isotope data for low-δ26Mg Cretaceous and Cenozoic sodic basalts in NCC (North China Craton) are from Zhang et al. (2003) and Tang et al. (2006), respectively, and Mg isotope data of these samples are from Yang et al. (2012). The major elements and Mg isotope data for low-δ26Mg sodic basalts in SCB (South China Block) are from Huang J. et al. (2015), and their Sr isotope data are from Huang and Xiao (2016). The vertical gray bars in (c) and (d) represent data range of 87Sr/86Sr and 143Nd/144Nd for Keluo mantle xenoliths. Data for these xenoliths are from Zhang et al. (2011). The horizontal gray bar (53Mg = −0.23 ± 0.04%) and diagonally striped bar (53Mg = −0.25 ± 0.07%) stand for recommended 53Mg values of the terrestrial mantle, and the data are from Lai et al. (2015) and Teng et al. (2010), respectively. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
chemical peculiarities of the EM1 endmember can be generated by processes in the mantle transition zone using a two-stage model.

4.1. Evidence for an asthenospheric source

4.1.1. Assessment of fractional crystallization and crustal contamination

The positive correlation between MgO and Ni (Fig. S2) indicates the potassic basaltas have experienced olivine fractionation. At first glance, the kink (at MgO ca 9 wt.%) in the plots of FeO/Ti and CaO/Al2O3 versus MgO, and the decreased Fe2O3/Ti and CaO/Al2O3 could be due to extensive fractionation of Fe-oxides and clinopyroxene, respectively. However, the correlations presented by radiogenic isotope ratios with FeO/Ti and CaO/Al2O3 (Fig. 4) do not support this interpretation, because such correlations cannot be produced by simple fractionation processes. The TiO2 content remains unchanged with the variation of MgO for all the potassic basaltas (Tables S1, S2), precluding significant fractionation of Ti-bearing Fe-oxides. For Nuominhe basaltas, which are characterized by higher MgO content relative to other potassic basaltas (Fig. S2), the nearly constant FeO/Ti and CaO/Al2O3, regardless of changing MgO, indicate little fractionation of clinopyroxene. In addition, the absence of negative Eu and Sr anomalies (Fig. 3) indicates the lack of significant plagioclase removal. As a result, we conclude that the potassic basaltas have only experienced olivine and minor clinopyroxene fractionation. Since previous work has shown that fractional crystallization of olivine and pyroxene induces very limited (≤0.07% for δ26Mg) Mg isotope fractionation during basalt differentiation (Teng et al., 2007), the low δ26Mg values of the potassic basaltas and their correlations with major elements (e.g., MgO, SiO2, Fig. 6) are also unlikely to be the result of fractional crystallization. Some type of mixing process is required to explain the correlations between isotopes (Sr, Nd, Pb and Mg) and major elements (Figs. 4, 6).

A possible candidate for such a mixing process could be the contamination of primary mantle-derived magmas by crustal materials. However, abundant evidence shows that crustal contamination has negligible influence on the chemical compositions of the potassic basaltas in NE China: (1) The common occurrence of mantle xenoliths and xenocrysts hosted by these basaltas indicates that the magmas ascended rapidly (Zhang M. et al., 1998, 2000; Zhang Y.L. et al., 2011; Sui et al., 2012; Zhao et al., 2014), allowing for little interaction between the magma and crustal materials; (2) Significant excess 230Th/230Th = 1.24–1.33 also shows that the magmas ascended rapidly and survived such magma–crust interaction (Zou et al., 2003); (3) Detailed studies on crustal xenoliths and xenocrysts hosted by Wudalianchi lavas revealed that crustal contamination is extremely localized and limited (only micro-scale interactions occur at the contact zone between the xenolith and host lava) during lava petrogenesis (Mao et al., 2015; McGee et al., 2015); (4) Os isotope data rule out obvious addition of crustal material into the basaltic magma (Chu et al., 2013). Therefore, the mixing process must have occurred in the mantle rather than the crust, and crustal contamination could not have produced the observed light Mg isotope compositions.

4.1.2. Binary mixing in the mantle

The Sr–Nd–Pb isotopic compositions of the potassic rocks are strongly correlated with each other (Fig. 2), 87Sr/86Sr is positively correlated with K/U and Hf/Hf*, while 206Pb/204Pb is negatively correlated with them (Fig. 5). Since K and U exhibit similarly incompatible behavior during partial melting of mantle peridotite, the K/U ratio of basaltas provides a proxy to that of their mantle source (e.g. Arevalo et al., 2009). The same is true for Hf/Hf*, because Sm and Nd have peridotite/melt partition coefficients that bracket those of Hf (e.g. Sun and McDonough, 1989). Therefore, all these correlations require a mixing process of two endmembers: an EM1-like endmember with high K/U and Hf/Hf*, and a depleted endmember with low K/U and Hf/Hf*, and relatively depleted Sr–Nd–Hf isotopic compositions. Notably, the δ26Mg values of the potassic basaltas increase with increasing 143Nd/144Nd and decreasing 87Sr/86Sr (Fig. 6). Since Mg isotope fractionation is essentially negligible (e.g. −0.07% for δ26Mg) during mantle partial melting (Bourdon et al., 2010; Teng et al., 2010), the covariations of δ26Mg values with Sr–Nd isotopes also trace the mixing process in the mantle. The EM1-like endmember has the lower δ26Mg value, while the depleted endmember shows a normal mantle-like δ26Mg value (e.g. −0.25 ± 0.07%, Teng et al., 2010). In terms of radiogenic isotopes, both the asthenospheric mantle and the depleted lherzolitic mantle are possible candidates for the depleted endmember (Fig. 2). In NE China, all the Sr–Nd–Pb isotopic data for Cenozoic intraplate basaltas constitute a mixing array between EM1 and a depleted mantle component (Fig. 2). Among these basaltas, sodic basaltas from the Halahe volcanic field (Fig. 1b) have the most depleted Sr–Nd isotopic compositions (Fig. 2a). They have typical OIB-like geochemical characteristics, with positive Nb–Ta and negative Pb anomalies (Figs. 2, 3), and can be regarded to represent the depleted endmember in the asthenosphere in this region (Ho et al., 2013). The lherzolitic mantle beneath this region also shows depleted Sr–Nd–Hf isotopic compositions, as represented by mantle xenoliths found in the Keluo potassic basaltas (Figs. 2a, 2b; Zhang et al., 2011). On plots of Hf/Hf* and K/U versus 87Sr/86Sr and 206Pb/204Pb, the potassic basaltas lie on mixing lines between EM1 and the local lherzolitic mantle, and this differs significantly from the mixing line between EM1 and the depleted endmember in the asthenosphere (represented by the Halahe basalts) (Fig. 5). Based on this observation, we argue that the lherzolitic mantle, rather than the asthenospheric mantle, constitutes the depleted endmember of the potassic basalt mixing array. In addition, a comprehensive review of the xenolith Mg-isotope data (Fig. S4a) shows that the overwhelming majority of mantle xenoliths have terrestrial mantle-like δ26Mg values (e.g. −0.25 ± 0.07%, Teng et al., 2010). A few clinopyroxenite xenoliths show lower δ26Mg values (as low as −1.51 ± 0.07%), however, these garnet-free clinopyroxenite xenoliths are characterized by extremely high CaO contents and a negative correlation between CaO and δ26Mg (Fig. S4b), and they have therefore been intensely metasomatized by isotopically light carbonate melts/liquids (Hu et al., 2016; Wang et al., 2016). Such metasomatic signatures have never been observed in the mantle xenoliths entrained by the potassic basaltas (Zhang M. et al., 2000; Zhang Y.L. et al., 2011; Sui et al., 2012), the lherzolitic mantle beneath these potassic volcanic fields is expected to show terrestrial mantle-like δ26Mg value, and this is consistent with the correlations between δ26Mg and Sr–Nd isotopes shown by the potassic basaltas (Fig. 6). In summary, all the geochemical features are consistent with the inference that the depleted endmember of the various binary mixing arrays displayed by the potassic basaltas is normal lherzolitic mantle.

4.1.3. Melt–rock interactions in the lherzolitic mantle

Since all magmas must pass through the lherzolitic mantle before their eruption, melt–rock interaction becomes the most likely way to generate the observed mixing relationships in the potassic basaltas. Recently, Liu et al. (2016) found that the compositional variations (e.g. MgO, K2O/Na2O and 87Sr/86Sr) in the potassic basaltas from NE China are correlated with lherzolitic thickness. This can be explained by progressive interactions between silica- and potassium-rich melts and refractory lherzolitic mantle. This is confirmed by the new observation that the δ26Mg values increase with increasing lherzolitic thickness from Erkeshan and Wudalianchi to Nuominhe basaltas, as well as by the
accompanying increase of MgO and $^{143}$Nd/$^{144}$Nd, and decrease of SiO$_2$ and $^{87}$Sr/$^{86}$Sr (Fig. 6). All these correlations are easily explained by progressive interactions between low $\delta^{26}$Mg enriched potassic melts with normal, refractory lithospheric mantle. Because all of the chemical-isotopic trends supporting this proposition are linear within the scatter of the data, it is probably not possible to recover the details of this interaction, e.g. whether it is by assimilation or some form of reaction–precipitation. In any case, the most MgO-rich lavas, erupted on the thickest lithosphere are closest to normal peridotite melts in terms of MgO, CaO/Al$_2$O$_3$, radiogenic isotopes, $\delta^{26}$Mg, and K/U.

In conclusion, the lithospheric mantle beneath the potassic volcanic fields played a crucial role in modifying the geochemical compositions of the EM1 source-derived initial melts, but it did not act as their direct origin. Therefore, we argue that the potassic basalts in NE China require a sublithospheric mantle origin. In the absence of any evidence for a deep-mantle plume, we argue further that this source must be located either in the asthenosphere or in the MTZ near the base of the upper mantle.

4.2. EM1 mantle source: nature and origin

4.2.1. Nature of the EM1 mantle source

The Sr–Nd–Pb–Hf isotopic compositions of the isotopically most extreme and most highly potassic initial melts are very similar to, but even more extreme than, those of the oceanic EM1 endmember composition represented by the Pitcairn islands, namely $^{206}$Pb/$^{204}$Pb $\leq$ 16.5, $^{143}$Nd/$^{144}$Nd $\leq$ 0.5123 (or $\epsilon$Nd $\leq$ -6.4), $^{176}$Hf/$^{177}$Hf $\leq$ 0.2825 (or $\epsilon$Hf $\leq$ -10.1) (Fig. 2). This observation indicates that the suite of Cenozoic potassic basalts from NE China are derived from an extreme EM1-type mantle source. We can further constrain the chemical features of the initial potassic melts and their EM1-type mantle source, using the correlations between element compositions (e.g. MgO, SiO$_2$) or element ratios (e.g. K/U, Hf/Hf$^*$) and Sr–Nd–Pb–Hf–Mg isotopes (Figs. 4–6) shown by the potassic basalts. The distinctive chemical compositions of the initial potassic melts include low $\delta^{26}$Mg (ca. -0.6%), MgO (>5 wt.%), and Hf/Hf$^*$ (Fig. 6). In addition, the linear correlation between K/U and Ba/Th (Fig. 5) suggests that the initial potassic melts are also characterized by unusually high Ba/Th (~400) ratios. Since K/U, Ba/Th and Hf/Hf$^*$ are normally not sensitive to fractionation during partial melting, the high K/U (~50,000), Ba/Th (~400) and Hf/Hf$^*$ (~125) ratios should also be an intrinsic property of the EM1 source. Because U/Pb is likely to undergo some fractionation during partial melting, the source U/Pb is likely to be even lower than the ratios observed in the melts. It will be seen below that such a low U/Pb ratio is consistent with the highly unradiogenic Pb-isotopic composition of the basalt.

Tian et al. (2016) argued that the low $\delta^{26}$Mg feature and the EM1-type Sr–Nd–Hf isotopic compositions are decoupled and must be derived from different source regions in the mantle. They proposed that the potassic basalts were formed by interaction of carbonated, asthenosphere-derived, low SiO$_2$, low-$\delta^{26}$Mg original melts with enriched subcontinental lithospheric mantle (SCLM) having EM1-type Sr–Nd–Hf isotopic compositions. During that interaction, the initial MORB-like Sr–Nd–Hf isotopic compositions of the original melts must have been entirely transformed to EM1-like, while the low-$\delta^{26}$Mg feature of the melts remained intact. Meanwhile, the interaction elevated the SiO$_2$ and K$_2$O in the reacted melts (Tian et al., 2016). However, given our new data, the $\delta^{26}$Mg values of the potassic basalts are now seen to be well correlated with Sr–Nd isotopes (Fig. 6). It is therefore clear that the Mg isotopic variations are coupled with the EM1 characteristics defined by the Sr, Nd, Pb isotopes. Therefore, we argue that the low $\delta^{26}$Mg (e.g. $-0.6\%$) feature of the initial potassic melts is an intrinsic property of their EM1 mantle source. The low U/Pb ratios ($\sim$0.06) and the positive correlation between $\delta^{26}$Mg and U/Pb in the potassic basalts (not shown) are consistent with this conclusion.

4.2.2. Origin of the EM1 mantle source

We begin this discussion by an evaluation of two previously proposed models about the origin of the EM1 source of potassic basalts, one involving delaminated lower crust, the other calling on ancient sediments subducted and stored in the MTZ.

Chu et al. (2013) suggested that potassium-rich silicate melts originated from delaminated ancient, lower continental crust (LCC) and subsequently metasomatized the SCLM. This modified SCLM then remelted and ultimately produced the potassic EM1-type basalts. Lowermost continental crust possesses several of the characteristics associated with the potassic basalts. Among these are high K/U and Ba/Th ratios (Fig. 5). While the average lower crust given by Rudnick and Gao (2014) shows somewhat lower values for these parameters, namely K/U $\leq$ 25000 and Ba/Th $\leq$ 200, granulite-facies lower crustal xenoliths have been observed with much higher values for these ratios (Mansur et al., 2014). This is not the case, however, for Hf/Hf$^*$. Rudnick and Gao (2014) give a value of 0.86 for this parameter, and the xenoliths analyzed by Mansur et al. (2014) have similarly low Hf/Hf$^*$ values of 0.7 ± 0.4 (average value ±1SD, one standard deviation). Recent evidence from magnesium isotopes also argues against involvement of delaminated lower crust. The average LCC has mantle-like $\delta^{26}$Mg values ($\delta^{26}$Mg = -0.26 ± 0.06‰; Yang et al., 2016), and this value is consistent with the majority of LCC xenoliths (Teng et al., 2013; Wang et al., 2016; Yang et al., 2016). Since Mg isotope fractionation is negligible during metamorphism and crustal subduction (Li et al., 2014; Wang et al., 2014), delaminated LCC is expected to retain the original mantle-like $\delta^{26}$Mg values of LCC. Therefore, on balance, the geochemistry of (delaminated) lower crustal rocks now appears to be inconsistent with at least some of the critical evidence used to trace the origin of the EM1 component, both in China and in Gaussberg lamproites. We emphasize, however, that these constraints may not apply to all EM1-type mantle sources, although at least some Pitcairn basalts, for example, also show positive Hf/Hf$^*$ ratios, but lack the high K/U and Ba/Th ratios that characterize the EM1 sources discussed here, and insufficient Mg isotope data are currently available for EM1 oceanic basalts to evaluate this evidence.

As an alternative to delaminated lower crust, we investigate the possible role of subducted and recycled sediments. Such recycling of sediments to explain potassic magmas has been invoked by several previous authors (e.g. Nelson, 1992; Murphy et al., 2002; Kuritani et al., 2013; Sun et al., 2014, 2015). Recycled sediments have also been invoked to explain the sources of oceanic EM1 basalts (e.g. Woodhead and McCulloch, 1989; Woodhead and Devey, 1993; Eisele et al., 2002; Delavaux et al., 2016). New evidence supporting the presence of recycled sediments in the potassic basalts from NE China has been presented by Sun et al. (2015) in the form of high $\delta^{18}$O values in olivine phenocrysts from Xiaogulhe, and by Tian et al. (2016) in the form of low $\delta^{26}$Mg values in Wudalianchi and Erkeshan basalts. The $\delta^{26}$Mg data presented in the present paper and the correlations of $\delta^{26}$Mg with EM1 type compositional parameters in Fig. 6 provide further support for this interpretation. The EM1 source of the potassic basalts shows low $\delta^{26}$Mg ($\sim-0.6\%$) values that differ from normal, ‘unmodified LCC’ ($\delta^{26}$Mg = -0.29 ± 0.15‰; Wang et al., 2016; Fig. 5a) and mantle ($\delta^{26}$Mg = -0.25 ± 0.07‰; Teng et al., 2010). Instead, they fall in the range of sedimentary carbonates ($\delta^{26}$Mg = -5.57‰ to -0.38‰; e.g. Young and Galy, 2004; Wombacher et al., 2011; Huang K.J. et al., 2015 and references therein). It is reasonable that subducted sediments, if they en-
trained some carbonate, will show a low-δ²⁶Mg feature, as represented by the carbonate-bearing meta-sediments (δ²⁶Mg values reach as low as −0.75‰; Wang et al., 2015). Deeply subducted carbonate-bearing sediments thus provide a solution to the low-δ²⁶Mg (≤ −0.6‰) feature of the EM1 mantle source.

Although the weight of evidence summarized above points toward an EM1 source containing recycled sediments, the potassic basalts, as well as the Gaussian lamproites, show three striking geochemical anomalies that are inconsistent with subducted sediments or their direct partial melts. These are the very high K/U (≈50,000), Ba/Th (≈400) and Hf/Hf²⁺ ratios (≈125) (Figs. 5, SS). For example, the data compiled by Plank (2014) for sediments being subducted under 30 modern arcs yield a K/U = 11,000 ± 6000 and Hf/Hf²⁺ = 0.61 ± 0.27 (average value ±1SD), and simple partial melting in the upper mantle will not generate the specific anomalies seen in the potassic melts. Simply calling on the stability of a K–Ba rich mineral phase such as K-hollandite in the transition zone (Murphy et al., 2002; Rapp et al., 2008) will also not explain the extremely high Hf/Hf²⁺ or K/U ratios, because changing a phase assemblage does not by itself change the bulk composition of that assemblage, unless the phase in question, in this case K-hollandite, is preferentially removed or accumulated by some petrogenetic process. Finally, mobilization of sedimentary Th and U in preference to K, Ba, and Pb during subduction, as proposed by Murphy et al. (2002) is implausible in view of the more recent assessment of such mobilization budgets by Porter and White (2009). Their arc-by-arc evaluation of element fractions mobilized during subduction and sampled by arc volcanoes shows that in the great majority of arcs, K and Ba are mobilized in preference to both U and Th, leading to lower, not higher, K/U and Ba/Th in the residual slab entering the deeper mantle.

In summary, none of the previous contributions that have called on a deeply subducted sedimentary source component have satisfactorily explained the specific trace-chemical anomalies that distinguish the potassic lavas from NE China, as well as the lamproites from Gausberg. We will attempt to resolve these issues in the following section, introducing a two-stage model, which can reproduce the specific chemical anomalies seen in both the Gausberg lamproites and in potassic basalts from NE China.

4.3. Two-stage genetic model for the EM1 mantle source

4.3.1. Age of the EM1 reservoir

The extraordinarily radiogenic Pb isotopic compositions of the EM1 mantle source beneath NE China plot on the left-hand side of the 4.53 Gyr Geochron (Fig. 7), requiring long-term isolated evolution of a reservoir with low time-integrated μ (238U/204Pb) value. This is supported by the low U/Pb ratios (≈0.06, corresponding to a μ value of about 4.2) of the initial potassic melts (see section 4.2.1). In order to understand whether the present-day Pb isotopic compositions could be achieved by recycling of ancient sediment, we performed simulations using a three-stage Pb growth model with varying μ values (Fig. 7), similar to that of Eisele et al. (2002), but without carrying out a full Monte-Carlo type simulation, which is beyond the scope of this paper. Details about the
specific assumptions used in our calculations are given in the caption of Fig. 7 and in the Supplementary Material.

The first stage between $T_0$ at 4.53 Ga and $T_1$, set here at 3.7 Ga following Stacey and Kramers (1975), corresponds to the mantle evolution of the source material, with formation of a crustal reservoir at $T_1$. This stage is followed by a crustal history between $T_1$ and $T_2$. During the third stage, from $T_2$ (the time of sediment subduction) to $T_{\text{present}}$, the subducted sedimentary assemblages evolve with sediment characteristics until the present (Fig. 7a). The modeling result shows that, in order to obtain a present day $^{206}\text{Pb} / ^{204}\text{Pb}$ of 16.5 and $^{207}\text{Pb} / ^{206}\text{Pb}$ of 15.4 (the average value of Xiaogulie), the recycled sediments should have evolved with $\mu_3$ values of $\sim$4.1 starting at about 2.2 Ga ago (Fig. 7b). That is to say, the sediments were subducted into the mantle at $\sim$2.2 Ga ago, and isolated in some part of the mantle since then. Because of this drastic reduction of the U/Pb ratio (expressed in the calculations by the parameter $\mu = (238\text{U} / ^{206}\text{Pb})_{\text{present}}$), the growth of radiogenic Pb isotopes was severely retarded.

Plausible environments to isolate recycled materials for such a long time ($\sim$2.2 Gyr) may be the MTZ (e.g. Nishi et al., 2013), or the base of the lower mantle (e.g. Christensen and Hofmann, 1994), because both of them are likely to be slab ‘graveyards’ (e.g. Hirose et al., 1999; Chen and Brudzinski, 2001). However, there is no clear evidence for a deep mantle plume located in the lower mantle beneath NE China, given the absence of a low seismic velocity anomaly rooted in the lower mantle (e.g. Zhao, 2004; Huang and Zhao, 2006). There is, however, substantial evidence for the existence of a stagnant, subducted slab resting near the base of the upper mantle, that is, within the MTZ. Both global and regional tomography reveal a high-seismic anomaly, presumed to represent a stagnant subducted slab in the MTZ beneath NE China (e.g. Zhao, 2004; Huang and Zhao, 2006; Fukao et al., 2009). Seismic tomography does not reveal how long this stagnant slab might have rested in the MTZ, but given its potentially neutral buoyancy at the base of the upper mantle, we postulate that it (or part of this stagnant slab) may have been subducted in Proterozoic time, in order to satisfy the constraints provided by the lead isotopes (Fig. 7).

4.3.2. Origin of EM1 chemical anomalies

We pointed out above that several striking chemical anomalies, namely the high K/U, Ba/Th and Hf/Hf* ratios seen in the EM1 endmember melts cannot be produced by simple partial melting of sediment-bearing mantle sources. Evidence from transition-zone derived diamonds can help to understand how these trace element ratios might be fractionated in such mantle sources. Walter et al. (2008) and Thomson et al. (2016), using observations on diamond inclusions from Juina (Brazil), have shown that “slab geotherms will intersect in a deep depression along the melting curve of carbonated ocean crust” leading to low-degree partial melts of carbonatic composition, which will then react with reduced ambient mantle and produce diamond. Here we apply this model to carbonated subducted sediments and consider the effect of this deep decarbonation reaction on the trace element budget of the residual sediment (Fig. 8), using the experimental data by Grassi and Schmidt (2011). After the extraction of a small amount of carbonate melt in the MTZ (Figs. 8, 9a), the residual sedimentary assemblage mainly includes K-hollandite, stishovite, majorite garnet, and corundum (Grassi and Schmidt, 2011). The effect of K-hollandite is to strongly retain K, Rh, Ba, and Pb during carbonate melt extraction, while Th, U, Nb Ta and REE enter the carbonateite and are partly removed from the residue (Grassi et al., 2012). In addition, the residual majorite retains Zr and Hf in preference to the light and middle REE. The combined effect of these two residual high-pressure minerals is to create an assemblage with elevated (Ba, Rh, K, Pb)/(Th, U) and (Zr, Hf)/(Sm, Nd) (Fig. 8).

Fig. 8 shows the progressive evolution of these anomalies, as small amounts ($\leq$2.5%) of carbonateite melt are removed from the residual assemblage. Of particular importance in this context are the extremely low U/Pb ratios of the Xiaogulie basalts (0.06 ± 0.03, average value ±1SD), which are easily modeled by this pro-
cess (Fig. 8e), and which are lower than those of most of the sediments being subducted in present-day arcs (U/Pb = 0.11±0.09, average value ±15D; Plank, 2014). This process requires pressures in excess of about 10 GPa, because K-hollandite is not stable at lower pressures (e.g. Grassi and Schmidt, 2011), and it generates the anomalies that are characteristic for the potassic basalts from NE China and for Gausberg lamproites. We are not aware of another mechanism that can generate these anomalies in the upper mantle.

In order for the residual solid to generate EM1-type basaltic melts that can ascend into the lithosphere and/or directly to the surface, it must first rise into the shallow asthenosphere (Fig. 9b), where it may undergo low degrees of partial melting at pressures lower than the stability of K-hollandite and majorite garnet, because any partial silicate melts generated in the presence of these phases will not be potassic. Due to the strong disturbance by the subducting Pacific slab, the deep residual solid was probably incorporated into the mantle ascending flow and transported up into shallow asthenosphere. Evidence for such ascending flow beneath NE China has been provided by recent results (slow velocity anomalies above the stagnant slab) of seismic tomography (e.g. Zhao, 2004; Huang and Zhao, 2006; Dokht et al., 2016). In any case, partial melting of a sediment-bearing mantle source at depths of about 150 km is expected to generate melts that are relatively high in SiO$_2$ and low in MgO and CaO/Al$_2$O$_3$. These melts will essentially mimic the incompatible-trace element anomalies (K/U, Ba/Th, U/Pb, Hf/Hf$^+$) of their EM1-type source, and their HREE concentrations will be buffered at low values by residual pyrope garnet.

When these melts ascend through the subcontinental lithosphere, they interact with the depleted mantle peridotite to varying degrees that depend on the local lithospheric thickness (Fig. 9b; Liu et al., 2016, 2017). This interaction process generates the binary mixing arrays between the EM1-type asthenospheric source component and the relatively depleted lithospheric component seen in the combined major and trace elements and isotopic compositions.

### 4.4. Implications

The present work, together with previous studies (e.g. Murphy et al., 2002; Rapp et al., 2008; Kuritani et al., 2011, 2013; Sun et al., 2014), suggests that the MTZ can sequester subducted sediments without being admixed into the convecting mantle for long periods of time, and thus form an EM1 reservoir. We have built on this previously suggested idea to describe specifically how the geochemical peculiarities of a subcontinental EM1 reservoir in the MTZ can be generated by first extracting small amounts of carbonate melts from ancient subducted sediments and then storing the residual sedimentary assemblages for as long as about 2.2 billion years. The striking similarity of the geochemistry of the Chinese potassic basalts with that of the Gausberg lamproites suggests that the specific mechanisms operating in the MTZ beneath NE China also apply to at least some other potassic-ultrapotassic volcanic rocks in other parts of the world.

### 5. Conclusions

Isotopic evidence indicates that the potassic volcanic fields in NE China sample an extreme EM1 endmember composition. The Mg, Pb, Hf, Nd and Sr isotope and element data for the Cenozoic intraplate potassic basalts provide critical constraints on the genesis of these rocks, and more importantly, the origin of the EM1 component beneath NE China. The major geochemical variations in the potassic basalts are induced by interactions between the initial low $\delta^{26}$Mg potassic melts and the lithospheric mantle they pass through. The EM1 mantle source, which generates the initial potassic melts, is characterized by positive Zr, Hf anomalies (Hf/Hf$^+$ > 1), strong enrichment of K, Ba and Pb relative to Th and U (K/U $\geq$ 50,000, Ba/Th $\geq$ 400, U/Pb $\geq$ 0.06), and low $\delta^{26}$Mg (e.g. $< -0.6\%$) values. Such elemental and isotopic features are consistent with recycled ancient carbonate-bearing sediments. However, the sediments must have lost small amounts of carbonate melts when they arrived at MTZ depth, in order to generate the observed geochemical anomalies. To reproduce the extremely unradiogenic Pb isotopes, the residual sedimentary assemblage must have been isolated in the MTZ for about two billion years. This implies that recycled ancient crustal materials, if stored in the MTZ, can escape convective homogenization for a long period of time and thus form an isolated EM1 reservoir.

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