Invited review

On the chemical markers of pyroxenite contributions in continental basalts in Eastern China: Implications for source lithology and the origin of basalts

Zong-Feng Yang *, Jie Li, Wen-Fei Liang, Zhao-Hua Luo

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ABSTRACT

Although it is widely recognized that mantle peridotite is compositionally heterogeneous, primitive mantle-like peridotite is commonly considered the potential peridotite source when identifying the source lithology of oceanic and continental basalts. Here, a simple equation that compatible with thermodynamic result (Ol (wt%) = 144.27ln((Fe + Mg + Ca + 2Na) / Si), mol ratio) is introduced to distinguish peridotitic and pyroxenitic rocks. Meanwhile, the basic assumptions, alternative interpretations and uncertainties of many of the literature markers for pyroxenite contributions in basalts are briefly summarized, and some detailed comments on these markers are presented using literature data on olivines and basalts from Eastern China. Both high-CaO (>0.1 wt%) and low-CaO (<0.1 wt%) olivines in these continental basalts are likely generated by a xenocryst diffusive re-equilibration process, suggesting that the olivine chemistry method may not be suitable for the source lithology identification of the basalts. For the Cenozoic and Mesozoic continental basalts in Eastern China, nearly all of the markers for pyroxenite sources based on whole-rock major and/or trace elements might be compromised by referitized/metasomatized peridotite sources. These uncertainties also exist for many other oceanic and continental basalts. However, the FC3MS (FeO*/CaO-3*FeO/SiO2, all in wt%) value, combined with those of Na2O+K2O, Mg#, MgO and La/Yb, suggests that Cenozoic OIB-type basalts in Eastern China are mostly primary pyroxenite-derived products, and caution should be paid when interpreting the petrological and geological implications of these basalts, particularly that of the low-MgO pyroxenite-derived basalts. At the normal mantle potential temperature (~1350–1400 °C), the pyroxenite source can produce melts that explain all of the compositional characteristics of the Cenozoic basalts, suggesting that the origin temperature and pressure of the Cenozoic basalts determined by the traditional peridotite model are overestimated. For the Mesozoic basalts, compositionally diverse peridotite sources can account for their petrogenesis, but the pyroxenite source cannot be excluded. We acknowledge that the pyroxenite source may represent crustal materials (continental or oceanic crust) recycling and/or mantle metasomatism, which have important geodynamic implications, but further work is needed to determine the origin of the pyroxenite.

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Olivine
Peridotite
Pyroxenite
Xenocrysts
Primary magma
1. Introduction

Identification of the source lithology of basalts plays a fundamental role in our understanding of the magmatic processes and origin of mantle-derived magmas, enabling us to gain insights into crustal material recycling and/or mantle metasomatism that contribute to mantle heterogeneity. Recently, a variety of markers have been experimentally and theoretically proposed to infer pyroxenite-derived melt contributions in both continental and oceanic basalts (Hauri, 1996; Herzberg, 2006, 2011; Lambart et al., 2013; Le Roux et al., 2010; Liu et al., 2008; Sobolev et al., 2007; Sobolev et al., 2005; Yang and Zhou, 2013), but the source lithology identification of basalts is still a debated issue because all of the markers are largely influenced by the compositional diversity of the source rocks and melting and crystallizing conditions of the basaltic magmas (Herzberg et al., 2014; Lambart et al., 2013; Matzen et al., 2013; Niu et al., 2011; Putirka et al., 2011; Yang and Zhou, 2013). Although mantle peridotite and pyroxenite from various locations around the world show large variations in bulk composition (Compositional data from Eastern China, special attention should be paid to test whether refertilized/metasomatized peridotite is well constrained by refertilized/metasomatized peridotite (Davis et al., 2013; Niu et al., 2011; Putirka et al., 2011). In addition, olivines crystallized from refertilized peridotite melts exhibit many chemical similarities compared with olivine-poor or olivine-free pyroxenite melts (Herzberg et al., 2014). Consequently, the major issue of the identification of the source lithology of basalts is to distinguish refertilized/metasomatized peridotite-derived basalts from pyroxenite-derived basalts.

Intraplate continental basalts are widely distributed in Eastern China, and numerous geochemical and petrological studies have been carried out over the years (Basu et al., 1991; Chen et al., 2007; Zhi et al., 1990; Zou et al., 2000), with particularly high-quality data accumulated in recent years (Liu et al., 2015; Liu et al., 2008; Qian et al., 2015; Wang et al., 2012; Xu, 2014; Zeng et al., 2011). However, there is some debate about the source lithology of these basalts, which in turn introduces significantly different interpretations of their petrogenetic and geological implications (Huang et al., 2013; Liu et al., 2008; Niu et al., 2011; Wang et al., 2012; Xu et al., 2014; Yang and Zhou, 2013; Zhang et al., 2009). The major evidence supporting the pyroxenite source of these basalts is largely taken from those that are used to infer the source lithology of oceanic basalts (Gao et al., 2008; Hong et al., 2013; Liu et al., 2015; Liu et al., 2008; Qian et al., 2015; Wang et al., 2012; Wang et al., 2011; Xu et al., 2014; Zeng et al., 2011; Zhang et al., 2009), which may also suggest a refertilized/metasomatized peridotite source (Niu et al., 2011; Yang and Zhou, 2013; Guo et al., 2014; Guo et al., 2016). Thus, to determine the source lithology of the basalts from Eastern China, special attention should be paid to test whether refertilized/metasomatized peridotite source can explain their geochemical and mineralogical characteristics. This work aims to briefly summarize the chemical markers of the pyroxenite contribution in basalts and discuss their prerequisites, alternative interpretations and

Table 1

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* PM, McDonough and Sun (1995).
* DMM, Workman and Hart (2005).
* HZ, Fallon and Danyushevsky (2000).
* HPY, Fallon et al. (1988).
* PHN1611, Kushiro (1996).
* KG1 and KG1, Kogiso et al. (1998).
* MIX1G, Hirschmann et al. (2003).
* MS-40, Lambart et al. (2009).
* Px-1, Sobolev et al. (2007)
uncertainties, as well as provide some detailed comments on the chemical markers for the pyroxenite contribution in basalts from Eastern China. Finally, we present a pyroxenite model for basalts from Eastern China.

2. Relationship between the bulk composition and lithology of mafic and ultramafic rocks

Refertilized/metamobilized peridotite may occasionally become olivine-poor or olivine-free pyroxenite. Before identifying the source lithology of basalts, it is useful to find a relationship between the bulk composition and lithology of mafic and ultramafic rocks. From a petrological perspective, peridotite and pyroxenite are lithologically controlled by the relative phase proportions of pyroxene versus olivine, and they are thermodynamically controlled by the bulk composition, temperature and pressure. Pyroxene and olivine are the two major rock-forming minerals that commonly used to distinguish peridotite from pyroxenite. Garnet could control the major-element composition of both peridotite- and pyroxenite-derived melts. However, garnet-pyroxyenite will become spinel-pyroxenite under low pressure, thus the major difference between peridotite and pyroxenite should be pyroxene versus olivine. Peridotite and pyroxenite can have the same mineral assemblage, but peridotite have ~40% olivine in volume and pyroxenite have ~40% olivine in volume. Although it is better to distinguish peridotite from pyroxenite based on the modal mineralogy when they equilibrate at a specific pressure and temperature, it is generally believed that peridotite has >30–35 wt% MgO and that pyroxenite has <30–35 wt% MgO (Downes, 2007; Kogiso et al., 2004; Lambart et al., 2013; Yang and Zhou, 2013) at the typical upper mantle temperatures and pressures under which basaltic magmas are commonly produced. However, the MgO content is not a good indicator of the lithology because pyroxenite may have a high MgO content, such as orthopyroxenite, which will have 40.1 wt% MgO if it consists of enstatite, whereas peridotite may have a low MgO content, such as dunite, which will have 23.4 wt% MgO if it consists of Fo30 olivine (FeMgSiO4). The classical division between a pyroxenite and peridotite is based on the relative proportions of olivine versus pyroxene; however, temperature and/or pressure, as well as other phases, such as spinel, garnet and Fe-Ti oxide, could also affect the relative proportions of olivine versus pyroxene. For typical pyroxenite, olivine-bearing spinel pyroxenite may become olivine-free garnet pyroxenite at elevated pressures (Kogiso et al., 2003). Moreover, garnet peridotite will have an elevated olivine/pyroxene ratio with increasing pressure because of the expanded stability field of garnet at the expense of pyroxene (Walter, 1998). This suggests that the lithology can only be safely determined for a given bulk composition at a specific temperature and pressure. However, the temperature and pressure are generally not easily obtained, and basaltic magmas are typically produced at a variety of temperatures and pressures. Thus, given that the bulk composition is typically used to represent the source rocks of basalts, it is better to find a simple compositional parameter that has only slight temperature and/or pressure effects to distinguish olivine-poor/olivine-free pyroxenite from olivine-rich peridotite.

The major end-member components of pyroxene, i.e., Wo (Ca2Si2O6), En (Mg2Si2O6), Fs (Fe2SiO4), Di (CaMgSi2O6), Jd (NaAlSi2O6), CATS (CaAl2SiO6), and garnet (Ca, Mg, Fe)3Al2Si3O12, all have (Fe + Mg + Ca + 2Na) / Si = 1 (mol ratio). Olivine Fo (Mg2SiO4) and Fa (Fe2SiO4) have (Fe + Mg + Ca + 2Na) / Si = 2 (mol ratio). Therefore, the relative proportions of olivine versus pyroxene and garnet for a given bulk composition can largely be determined by the (Fe + Mg + Ca + 2Na) / Si ratio. Here, we propose a simple empirical equation that Ol (wt%) = 144.27Ln[(Fe + Mg + Ca + 2Na) / Si], which can be used to discriminate peridotite from pyroxenite for a given bulk composition (Fig. 1a). The olivine contents calculated by this simple equation are in good agreement with those estimates of abyssal peridotite based on whole-rock CIPW norm calculation (Niu, 1997), mass balance calculation (Chu et al., 2009) and of experimental peridotite and pyroxenite based on pMELTS calculations (Ghiorso et al., 2002) (Fig. 1a and Supplementary Fig. 1), although our model olivine phase involves spinel and Fe-Ti oxides. Our model olivine content for a given bulk composition is generally slightly higher than its real content (Supplementary Fig. 1). Silica-excess and -deficient pyroxenite have (Fe + Mg + Ca + 2Na) / Si < 1 and > 1, respectively, and they can all become olivine-bearing pyroxenite at low pressure and olivine-free pyroxenite at high pressure.

The relationship between the olivine modal abundance calculated by Ol (wt%) = 144.27Ln[(Fe + Mg + Ca + 2Na) / Si] and the bulk MgO content suggests that both peridotite and pyroxenite show large variations in the MgO content with a constant olivine content (Fig. 1b). Our model suggests that if >30 wt% MORB is mixed with DMM peridotite, the mixed product will have ~40 olivine and can be called peridotite (Fig. 1b). TiO2 enrichment in ocean island basalts requires only ~1–10% MORB mixed with DMM peridotite (Prytulak and Elliott, 2007; Putirka et al., 2011), suggesting that TiO2 enrichment in basalts does not exclusively suggest a pyroxenite source from the lithological perspective. When the average amount of MORB added to the

![Fig. 1. Olivine modal abundance (wt%) as a function of (Fe + Mg + Ca + 2Na) / Si (mol) in (a) and relationship between calculated olivine modal abundance and MgO (wt%) in (b). Abyssal peridotite from Niu (1997). Xenoliths (peridotite and pyroxenite) in basalts from Chu et al. (2009). Experimental peridotite and pyroxenite from the data compiled by Yang and Zhou (2013). Detailed phase proportions of the representative experimental peridotite and pyroxenite shown in (a), see Table S7 of Yang and Zhou (2013). The DMM + MORB compositions are the same as those from Kogiso et al. (1998). The KR4003 + daitc melt compositions are the same as those from Herzberg et al. (2014). Negative olivine contents in (b) indicate (Fe + Mg + Ca + 2Na) / Si < 1.](image-url)
peridotite reaches approximately 70%, the mixture becomes nearly olivine-free pyroxenite, similar to the experimental pyroxenites MIX1G and M5–40 (Fig. 1b). Pyroxenite may also be produced if acid melts are mixed with peridotite, and as shown in Fig. 1b, ~10% dacite added to peridotite to reach 40%, the mixture becomes olivine-free pyroxenite, consistent with the conclusion of Herzberg et al. (2014). It worth noting that the mafic pyroxenite is often confused with the ultramafic pyroxenite in discussion on the origin of OIB-like basalts. The former is common in excess of silica and thus a kind of mafic eclogites, whereas the latter is deficient in silica and often represents ultramafic metasomatites or cumulates. The silica-excess mafic pyroxenite such as MORB typically depleted in melt-mobile incompatible elements and produce acid melts that cannot be source materials of OIB-like basalts. In contrast, ultramafic metasomatites in the forms of either refertilized peridotite or pyroxenite would have trace elements enriched source that can be a better candidate of the source of OIB-like basalts (Niu and O’Hara, 2003). Our current simple method is still arbitrary from a strict petrological perspective, but to our knowledge, it provides the simplest direct estimate of the olivine modal abundance of any given mafic and ultramafic bulk compositions and can thus help us to understand the lithological variation of the source materials of basalts.

3. Uncertainties and alternative interpretations of literature markers for the contribution of pyroxenite in basalts

Nearly all of the commonly used markers for pyroxenite contribution in basalts and their basic assumptions, alternative interpretations and uncertainties are briefly summarized in Table 2, and some detailed rigorous comments on most of these markers are presented below based on literature data for olivines and basalts from Eastern China. Some new alternative interpretations of pyroxenite source signatures of these continental basalts are explored, which may have important implications for source lithology and the origin of basalts.

3.1. Whole-rock CaO-MgO relations

It has been argued that at a given MgO content, pyroxenite can produce a melt with a lower CaO content than a peridotite melt (Herzberg, 2006; Herzberg and Asimow, 2008), which has been widely used to identify the source lithology of basalts (Fig. 2) (Gao et al., 2008; Hong et al., 2013; Liu et al., 2015; Liu et al., 2008; Nikulin et al., 2012; Timm et al., 2009; Zeng et al., 2011). However, this argument holds only for peridotite KR4003; KL-B1 if the solidus is at or above 3 GPa, which can be easily compromised by peridotite compositional diversity and magma mixing (Putirka et al., 2011; Yang and Zhou, 2013). The CaO-MgO relation was originally used by Herzberg (2006) to argue for a pyroxenite source for Hawaiian tholeiite. For alkalic melts, peridotite can produce low-CaO melts, as emphasized by Herzberg (2006) in his supplementary materials. However, this alternative interpretation was not considered in many subsequent studies, and numerous alkali basalts are inappropriately plotted in the CaO-MgO diagram to argue for a pyroxenite source (Gao et al., 2008; Hong et al., 2013; Liu et al., 2015; Liu et al., 2008; Nikulin et al., 2012; Timm et al., 2009; Zeng et al., 2011). More importantly, because CaO content in peridotite melts has a significantly negative correlation with Na2O + K2O content (Supplementary Fig. 2), this simple correlation results in a significantly negative correlation between CaO-13.81 + 0.274*MgO and Na2O + K2O (Fig. 2). The experimental melts of peridotite KR4003 ( Walter, 1998) and near-solidus melt of peridotite KL-B1 (Davis et al., 2011) with ~3.5% Na2O + K2O and their CaO-13.81 + 0.274*MgO = 0 are consistent with the argument that peridotite produces high-CaO melts. However, low-Na2O + K2O harzburgite melts and high-Na2O + K2O lherzolite

Table 2
Major markers of pyroxenite contribution in basalts and their alternative interpretations.

| Marker | Assumed peridotite source | Alternative interpretations and/or potential uncertainties | Potential peridotite source |a |
|--------|---------------------------|----------------------------------------------------------|-----------------------------|
| Olivine with low Ca, Mn and high Ni, Fe/Mn at a given Fo | PM and DM | 1. Temperature effect. 2. Magma mixing. 3. Xenocryst diffusive re-equilibration. 4. Olivine crystallized from low CaO peridotite melts or from metasomatized/refertilized peridotite melts. 5. Clinopyroxene fractionation. | PHN1611, HPY, MORB + DMM, MORB melts + PM, PS6.1, KPS2.1, KPS3.2, KPS4, MBK, MBK+ |
| Melts with high Fe/Mn | PM-like peridotite (KR4003) | 1. Temperature and pressure effect. 2. High Fe/Mn (iron-rich) peridotite source. 3. Clinopyroxene fractionation. | PHN1611, HK-66, KL-B1, MORB melts + PM |
| Melts with high Zn/Fe | PM-like peridotite (KR4003) | 1. Pressure effect. 2. High Zn/Fe peridotite source. 3. Clinopyroxene fractionation. 4. TiO2-rich peridotite. | PHN1611, MORB melts + PM |
| Melts with high TiO2 | PM and DM | 1. Carbonated peridotite. 2. Iron-rich peridotite source | PHN1611, HPY, MORB + DMM, MORB melts + PM |
| Melts with high Si2O3 and high FeO | Volatile-free DMM1 | Carbonated peridotite | KL-B1 + 2.5 wt% CO2, PERC, PERC3, |
| Melts with high CaO/Al2O3 | Volatile-free PM-like peridotite | Harzburgite melts | HZ |
| Melts with high SiO2 | Harzburgite melts | KLB-1 + 2.5 wt% CO2, PERC, PERC3, |

a Herzberg (2006).  
b Sobolev et al. (2005, 2007) and Herzberg (2011).  
c Herzberg (2011).  
d Le Roux et al. (2010).  
e Pyrtulak and Elliott (2007).  
† Lambert et al. (2013).  
‡ Lambert et al. (2013).  
§ Hauri (1996).  
¶ For detailed discussions of the alternative interpretations and/or potential uncertainties, see the text and Yang and Zhou (2013).  
® Carbonated peridotite PERC and PERC3 from Dasgupta et al. (2007), KLB-1 + 2.5 wt% CO2 from Hirose (1997). K2O-rich peridotite PS6.1, KPS2.1, KPS3.2 and KPS4 from Davis and Hirschmann (2013), and MBK and MBK+ from Laporte et al. (2014). Other peridotite sources are the same as in Table 1.
melts, including tholeiitic and alkaline basalts, can have low CaO contents and CaO-13.81 + 0.274 MgiO (Herzberg and Asimov, 2008) versus Na2O + K2O of experimental peridotite melts and natural basalts. Peridotite melts with MgO < 16 wt% and C-OIB (Ocean island basalt-like basalts in Eastern China and Mongolia) with MgO > 7.5% from data compiled by Yang and Zhou (2013). Peridotite KLB-1 3 GPa near-solidus melts from Davis et al. (2011), Peridotite KR4003 3–7 GPa melts from Walter (1998), Harzburgite 1.5–2.5 GPa melts from Falloon and Danyushevsky (2000); Peridotite + 1% and +2% CO2 melts from Dasgupta et al. (2007) and Dasgupta et al. (2013). Cenozoic and Mesozoic basalts from Eastern China that have been considered as pyroxenite-derived low-CaO melts from Gao et al. (2008), Hong et al. (2013), Liu et al. (2015), and Zeng et al. (2011); C-OIB (MgO > 7.5 wt%), Continental OIB-type basalts from Eastern China and Mongolia compiled by Yang and Zhou (2013). As CaO in olivine is highly incompatible, Ca-O-MgO relation of natural basalts would not significantly change when olivine-addition method is applied for primary magma correction. Thus, for basalts with MgO > 7.5–10%, olivine fractionation and accumulation would not change the results in identification of source lithology of natural basalts using the Ca-O-MgO relation.

Fig. 2. CaO-MgO relation and CaO-13.81 + 0.274 MgiO (Herzberg and Asimov, 2008) versus Na2O + K2O of experimental peridotite melts and natural basalts. Peridotite melts with MgO < 16 wt% and C-OIB (Ocean island basalt-like basalts in Eastern China and Mongolia) with MgO > 7.5% from data compiled by Yang and Zhou (2013). Peridotite KLB-1 3 GPa near-solidus melts from Davis et al. (2011), Peridotite KR4003 3–7 GPa melts from Walter (1998), Harzburgite 1.5–2.5 GPa melts from Falloon and Danyushevsky (2000); Peridotite + 1% and +2% CO2 melts from Dasgupta et al. (2007) and Dasgupta et al. (2013). Cenozoic and Mesozoic basalts from Eastern China that have been considered as pyroxenite-derived low-CaO melts from Gao et al. (2008), Hong et al. (2013), Liu et al. (2015), and Zeng et al. (2011); C-OIB (MgO > 7.5 wt%), Continental OIB-type basalts from Eastern China and Mongolia compiled by Yang and Zhou (2013). As CaO in olivine is highly incompatible, Ca-O-MgO relation of natural basalts would not significantly change when olivine-addition method is applied for primary magma correction. Thus, for basalts with MgO > 7.5–10%, olivine fractionation and accumulation would not change the results in identification of source lithology of natural basalts using the Ca-O-MgO relation.

Olívine Ni partitioning, which would successfully explain the Hawaiian high-Ni olivine phenocrysts (Matzen et al., 2013). This suggests that the temperature effect is still an alternative interpretation of high-Ni olivine. However, more high-quality experimental work is needed to quantify the effect of temperature on olivine Ni partitioning for liquids with various MgO contents.

The chemical features of olivine crystals occurring in basalts from Eastern China have also been used to infer a pyroxenite source (Gao et al., 2008; Hong et al., 2013; Liu et al., 2015; Qian et al., 2015; Wang et al., 2012; Wang et al., 2011). However, whether the chemical signals result from the melt composition or are simply characteristics of mantle olivine was not adequately resolved, as discussed below. Many Cenozoic and Cenozoic basalts from Eastern China contain varying amounts of olivine xenocryst that disaggregated from mantle peridotite (Chu et al., 2013; Fan and Hooper, 1991; Gao et al., 2008; Qian et al., 2015; Sakuyama et al., 2013; Wang et al., 2012), which were identified mainly based on their CaO content of <0.1 wt%. However, olivine from magmatic cumulates and low-CaO melts may also contain <0.1 wt% CaO (Kamenetsky et al., 2006; Li et al., 2012). Mantle olivines can also be distinguished from magmatic olivines by other minor elements such as Ti and Cr (Foley et al., 2013), but Ti and Cr contents are too low to precisely determine, which may be used to identify origin of olivine high quality data obtained for basalts in Eastern China. Moreover, textural characteristics, such as resorption and kink-bands, that have been taken to imply a xenocrystic origin may be produced from magmatic processes (Erdmann et al., 2014; Welsch et al., 2013). Thus, it is generally difficult to determine the origins of the olivine crystals that commonly occur in basalts. Here, we do not intend to determine the origins of the olivine that occur in basalts from Eastern China; instead, we present an alternative interpretation of their chemical characteristics.

Chemical zoning, typically representing elemental diffusion and reequilibration processes, is a common feature of olivine. The diffusion patterns of olivine from natural samples and experimental work suggest that olivine cores generally have high Fo and Ni contents and low Ca and Mn contents. The Mg-Fe diffusion zone is typically comparable with that of Ni but is generally wider than those of Ca and Mn because of their different diffusion coefficients (Costa et al., 2008; Costa and Dungan, 2005; Qian et al., 2010). For mantle olivine xenocrysts present in typical basaltic rocks, the diffusion characteristics of Fo, Ni and Ca can be simply illustrated as in Fig. 3a. Accordingly, several CaO- Fo and NiO-Fo co-variation trends can be easily obtained from a qualitative perspective (Fig. 3b, c).
The model CaO-Fo and NiO-Fo co-variation trends of typical mantle olivine xenocrysts that have been re-equilibrated with their host low-MgO melts can successfully explain the chemical characteristics of most low-CaO olivines in basalts from Eastern China (Fig. 3b). The olivine data provided by Wang et al. (2011) and Qian et al. (2015) show consistent results, and they are in good agreement with the chemical diffusion trends between typical mantle olivine xenocrysts and the low-MgO basaltic melts. Because the CaO diffusion rate is commonly slower than those of Ni and Fe-Mg (Costa et al., 2008; Costa and Dungan, 2005; Qian et al., 2010), mantle olivine will typically have <0.1 wt% CaO even when Fo decreases to 85. This diffusive re-equilibration model can explain why some slightly high-CaO olivines (>0.1 wt%), which have been considered as magmatic olivines from a pyroxenite source, have a low Fo and high Ni content. However, this alternative explanation is underestimated, and only olivines with <0.1 wt% CaO are considered mantle xenocrysts (Gao et al., 2008; Liu et al., 2015; Qian et al., 2015; Wang et al., 2012). In contrast, olivines from Hainan, Sihetun and Feixian basalts exhibit a large variations in CaO content, ranging from ~0% to ~0.2%, even when Fo > 90 (Fig. 3a). Although olivine with <0.1 wt% CaO has been excluded in these case studies when using olivine chemical data to infer source lithology (Gao et al., 2008; Wang et al., 2012), high-CaO olivine may also derive from mantle peridotite, as discussed below. The mantle olivine CaO content is mainly controlled by the equilibration temperature and pressure of mantle peridotite, and high-CaO olivine (>0.1 wt%) commonly occurs in high-temperature peridotite at a given pressure (De Hoog et al., 2010; Köhler and Brey, 1990; Witt-Eickschen and O’Neill, 2005) (Supplementary Fig. 4). pMELTS calculations (Ghiorso et al., 2002) for normal peridotite KLB-1 suggest that mantle olivine will have a CaO content of >0.1 wt% when the peridotite mineral assemblage re-equilibrates at 2 GPa with temperature > ~1200 °C, and they will have >0.2% CaO when the temperature is > ~1400 °C (Fig. 3b and Supplementary Fig. 4). Typical mantle olivine and olivine xenocrysts carried by basalts, mainly with <0.1 wt%, may originate from a random sampling process, and spinel peridotite mainly equilibrates at temperatures between 800 and 1200 °C (Rudnick et al., 2004; Wang et al., 2015; Witt-Eickschen and O’Neill, 2005). However, basaltic magmas may accidentally pick up some mantle olivine that has been re-equilibrated at higher temperatures in the vicinity of the source melting region, such as the high-temperature spinel peridotite reported by Witt-Eickschen and O’Neill (2005). Therefore, if both low- and high-temperature mantle olivine were picked up by basaltic magmas when in transit in the upper mantle, typical low-CaO and slightly higher-CaO mantle xenocrysts would be observed in basaltic rocks. In contrast, the NiO content of mantle olivine remains nearly constant during the peridotite re-equilibration processes at the near-solidus temperature, which is consistent with the NiO-Fo pattern of olivine in Cenozoic basalts (Fig. 3c). It worth noting that CaO and Ni diffusion patterns shown in typical olivine xenocyst may not occur in every xenocyst because diffusive re-equilibration, depending on the olivine size and residence time, can decrease or even erase elemental distribution patterns of olivine xenocrysts. Moreover, cut section effect of granular olivine may also affect their geochemical patterns observed in thin section.

For Sihetun and Feixian olivines in Mesozoic picrites, the large variation in NiO content appears to be inconsistent with the mantle olivine diffusive re-equilibration model. However, a closer examination of the olivine data reported by Gao et al. (2008) suggests that the olivine in peridotite also exhibits a large variation in NiO content, and some have a very high NiO contents at a given Fo (Supplementary Fig. 5), displaying chemical features similar to those high-CaO olivines that have been considered magmatic olivine. In addition, a mantle olivine xenocryst included in a hybridized hydrous dioritic magma suggests that the NiO...
content can vary from ~0.4% to ~0.65% with Fo ~ 90, even when CaO < 0.1 wt% (Qian et al., 2010). This suggests that the hosting melt composition may significantly influence the chemical diffusion patterns of the olivine xenocryst. Chromite separates of Gao et al. (2008) have significantly lower 187Os/188Os than wholerock and one chromite separate has sub-chondritic Os isotope composition (187Os/188Os = 0.112). This also imply that most olivines of Gao et al. (2008) with a xenocrystic origin. The Cenozoic basalts and Mesozoic picrites have significantly different bulk compositions, which might explain the different diffusive re-equilibration patterns of CaO-Fo and NiO-Fo in olivine. Therefore, CaO and NiO in olivine are likely inappropriately used in the source lithology identification of basalts in Eastern China, although this method has been successfully used in oceanic basalts.

3.3. Fe/Mn ratio and Mn content

High Fe/Mn ratios have frequently been used as an indicator of pyroxenite-derived melt contributions for basalts in Eastern China (Gao et al., 2008; Liu et al., 2008; Wang et al., 2012; Wang et al., 2011; Zhang et al., 2009), mainly based on the modeling and interpretations of normal peridotite melts (Herzberg, 2011; Sobolev et al., 2007). However, it is unrealistic to assume that a homogeneous peridotite is the potential source rock of any basalt because many observations suggest that the Fe/Mn ratio in peridotite varies considerably (Hirose and Kushiro, 1993; Kushiro, 1996; Le Roux et al., 2010; Lee, 2004; McDonough and Sun, 1995). In contrast, the Fe/Mn ratio in basaltic rocks has been considered a useful indicator of melting pressure (Niu et al., 2011). However, using the same partition coefficient of Fe and Mn and the peridotite melting equations as Niu et al. (2011), our calculation suggests that the pressure effect is not sufficient to account for the large variations in the Fe/Mn ratio of C-OIB (Fig. 4). Moreover, recent experimental determinations of the partition coefficients for Mn and Fe between basaltic melt and olivine, garnet, pyroxenes, and spinel at 3 GPa also suggest that the pressure effect cannot explain the high Fe/Mn ratios in OIB (Davis et al., 2013). Thus, the high Fe/Mn ratio in basalts should be mainly used as a compositional indicator of their source materials (Fig. 4a). The higher Fe/Mn ratios of basalts in Eastern China compared to typical MORB are mainly due to iron enrichment (Liu et al., 2008). This iron enrichment and high Fe/Mn can be achieved by iron-rich and high Fe/Mn peridotite (Davis et al., 2013; Humayun et al., 2004; Putirka et al., 2011) (Fig. 4b), such as peridotite KLB-1 and PHN1611, which does not necessarily require a mantle plume model or pyroxenite source.

Similar to the effect of temperature on Ni, the Mn content in olivine may be greatly influenced by the temperature effect on Mn partitioning between the olivine and melt. The temperature-dependent equation 2d of Putirka et al. (2011) has a lower prediction error than the composition-dependent equation of Herzberg and O’Hara (2002) in describing Mn partitioning between the olivine and melt even though the former captured 93 data points and the latter captured 143 data points. Moreover, the Mn partition coefficient of olivine calculated by the temperature-dependent equation 2d proposed by Putirka et al. (2011) is in good agreement with the recent high-precision experimental results (Davis et al., 2013; Le Roux et al., 2011). For olivines in basalts from Eastern China, Mn depletion and the high Fe/Mn ratio have not been widely used to infer the source lithology, partially because of the lack of high-precision data (Xu et al., 2012). Nevertheless, before using this method, the mantle olivine xenocryst diffusion model, as discussed in the previous section, should be carefully considered because mantle olivine generally has a low Mn content, and the chemical re-equilibration process may produce a low Mn content similar to that of olivine from pyroxenite-derived melts. Therefore, the high Fe/Mn in basalts and olivines cannot be safely used to identify the source lithology of basalts from Eastern China.

3.4. Zn/Fe ratio

The high Zn/Fe ratio in basaltic rocks from Eastern China has also been used to support pyroxenite-derived melts (Liu et al., 2015; Qian et al., 2015; Wang et al., 2012; Xu et al., 2012). However, as PM-like (Zn/Fe = 8.8 × 10\(^{-4}\)) peridotite produces low-Zn/Fe (~10 × 10\(^{-4}\)–1.2 × 10\(^{-4}\)) melts, high-Zn/Fe melts can be produced from a peridotite source if the peridotite has been metasomatized and enriched in Zn (Le Roux et al., 2011) (Fig. 5). Many trace element studies suggest that OIB-type basalts from Eastern China are mainly derived from enriched mantle, not a PM-like mantle source (Basu et al., 1991; Chen et al., 2007; Zou et al., 2000). Thus, it is reasonable to assume that the potential source peridotite has a high Zn content and high Zn/Fe ratio. Importantly, some high-precision data provided by Le Roux and co-workers suggests that mantle peridotite has Zn/Fe values from ~6 × 10\(^{-4}\) to ~11 × 10\(^{-4}\) (Le Roux et al., 2010). This wide range may explain why OIB has a large variation in the Zn/Fe ratio for the peridotite model (Davis et al., 2013). Interestingly, high-Fe/Mn peridotite PHN1611 that has been metasomatized by undersaturated silicate melts has a high Zn/Fe ratio (10.4 × 10\(^{-4}\)) (Smith et al., 1993), which can produce both high-Fe/Mn and high-Zn/Fe melts (Fig. 5). However, high-Zn/Fe peridotite can have either high Fe/Mn or low Fe/Mn ratios, and vice versa (Le Roux et al., 2010). This suggests that peridotite melts can have large variations in both Fe/Mn and Zn/Fe, as shown in Fig. 5b. Therefore, as suggested by Davis and co-workers, the Zn/Fe ratio in
basalts is not a sensitive indicator of non-peridotite source lithology (Davis et al., 2013).

4. Pyroxenite model of basalts from Eastern China based on the FC3MS value

As discussed above, when identifying the source lithology of basalt from Eastern China, the olivine chemistry method may be compromised by xenocryst diffusive re-equilibration processes. High Fe/Mn, Zn/Fe and low CaO may be characteristics of refertilized or metasomatized peridotite melts (Table 2). Moreover, in the commonly used Harker diagrams and phase projections, partial melts from a number of peridotitic rocks collectively show indistinguishable major element signatures compared with pyroxenite melts when olivine-dominated fractional crystallization occurs (Lambart et al., 2013; Putirka et al., 2011; Yang and Zhou, 2013). As shown in Fig. 6 for diverse peridotite melts and two representative pyroxenite melts, the most noticeable signature of pyroxenite melts is that they generally have lower MgO contents than peridotite melts, but olivine fractional crystallization can easily decrease the MgO content of peridotite melts, producing major element features that are collectively indistinguishable from those of pyroxenite melts. Moreover, if magma mixing occurs for both peridotite- and pyroxenite-derived melts, these melts will become more complex in their major element patterns. Consequently, it seems impossible to identify the source lithology of the Cenozoic and Mesozoic basalts from Eastern China based on their major elements (Fig. 6). Nevertheless, using a parameterization of multi-decadal melting experiments on a variety of peridotite and pyroxenite, Yang and Zhou (2013) proposed a parameter called the FC3MS value (FeO/Y/CaO×3+MgO/SiO₂, all in wt%), which can identify most pyroxenite-derived basalts. This marker was obtained based on the assumption that any peridotitic and pyroxenitic rocks that commonly occur in the upper mantle in both volatile-free and volatile-present conditions may produce melts contributing to basaltic magmas. The FC3MS value, which is mainly controlled by the source bulk composition and melting degree, has nearly no temperature or pressure effects, suggesting that the OB-type magmas from Eastern China and Mongolia are mainly derived from pyroxenite sources (Yang and Zhou, 2013).

Here, we firstly pay more attention to the source lithology and petrological implications of the Mesozoic and Cenozoic continental basalts with SiO₂ = 45–53 wt% from Eastern China because this type of basalt is very similar as both peridotite melts and pyroxenite melts (Fig. 6), and is widely distributed in not only Eastern China but also other countries and has been used to deduce many petrological and geological implications based on different assumptions of source lithology (Chen et al., 2007; Fan and Hooper, 1991; Liu et al., 2015; Timm et al., 2010; Zhang et al., 2009; Zhi et al., 1990). As discussed above, all of the geochemical characteristics of these basalts can be accurately explained by the peridotite model under the assumption that mantle peridotite is heterogeneous in bulk composition (Fig. 6). Although both the Cenozoic and Mesozoic basalts are geometrically similar as peridotite melts in every major element (Fig. 6), the high FC3MS value of the Cenozoic basalts at given MgO, Na₂O + K₂O, Mg# (MgO / (MgO + FeO), molar fraction) and Fe/Mn contents cannot be explained by any peridotite melts if olivine-dominated fractionation occurs (Fig. 7). Hydrous and carbonated peridotite-derived basaltic melts, not shown in Fig. 7, have identical and lower FC3MS values compared with volatile-free peridotite-derived melts, respectively (Yang and Zhou, 2013). Recent experimental petrology suggests that garnet peridotite and metasomatized peridotite-derived melts (Davis and Hirschmann, 2013; Grove et al., 2013; Laporte et al., 2014; Pilet et al., 2008) do not have a sufficiently high FC3MS value to explain the Cenozoic basalts from Eastern China. We noticed that some basalts from Hainan and Wudalianchi have sub-chondritic Os isotopes (187Os/188Os = 0.11869–0.12954) (Chu et al., 2013; Wang et al., 2013), but their high FC3MS (0.64–1.11) values suggest that their source lithology is pyroxenite. This seems inconsistent with their peridotite-like sub-chondritic Os isotopes, however, refertilized peridotite may become pyroxenite by infiltration of large amount of asthenosphere-derived melts, and preserve the sub-chondritic Os isotopes.

One may argue that the high FC3MS value of Cenozoic basalts is due to large amount of clinopyroxene fractionation, which appears to be consistent with the co-variation of FC3MS versus Na₂O + K₂O, Mg# and MgO (Fig. 7a–c). However, this interpretation is inconsistent with the typical evolution processes of peridotite-derived melts because the ascent and decompression of peridotite melts would result in olivine oversaturation and pyroxene undersaturation, and clinopyroxene crystallization commonly occurs when MgO > 6–10 wt%. In addition, a number of previous studies suggest that the geochemical features of Cenozoic basalts from Eastern China are mainly controlled by source materials, partial melting and olivine-dominated fractionation when the MgO content > 6–10 wt% (Fan and Hooper, 1991; Huang et al., 2013; Liu et al., 2015; Liu et al., 2008; Wang et al., 2012; Zeng et al., 2011; Zhang et al., 2009). Moreover, most of these basalts carry a variety of mantle-derived xenoliths, indicating that magma evolution is limited during the ascent in crustal conduits. Alternatively, pyroxenite melts can reasonably explain all of the co-variations of FC3MS versus Na₂O + K₂O, Mg# MgO and Fe/Mn (Fig. 7a–c). Large analytical errors in the MnO content of both peridotite and pyroxenite melts results in large variations in Fe/Mn, and peridotite melts can have high Fe/Mn ratios but low FC3MS values (Fig. 7d). Importantly, the pyroxenite M5–40 model melts that equilibrium with garnet and clinopyroxene can more
Fig. 6. Major element characteristics of experimental melts of peridotite and pyroxenite and representative basalts from Eastern China. Data for Cenozoic basalts from Huang et al. (2013), Liu et al. (2015), Qian et al. (2015), Wang et al. (2012), Wang et al. (2011), Xu et al. (2012), and Zhang et al. (2009) (Supplementary Table 1). Data for Mesozoic basalts from Gao et al. (2008) and Liu et al. (2008). Experimental pyroxenite melts including pyroxenites KG1 (1.5–3 GPa) (Kogiso et al., 1998) and MS–40 (1–2.5 GPa) (Lambart et al., 2012; Lambart et al., 2009; Lambart et al., 2013) melts (Supplementary Table 2). Experimental peridotite melts including PM-like peridotite KLB-1 (1–3 GPa) (Davis et al., 2011; Hirose and Kushiro, 1993) and KR4003 (3–7 GPa) (Walter, 1998), iron-rich peridotite HK-66 (1–3 GPa) (Hirose and Kushiro, 1993), iron-rich and K2O-rich PHN1611 (0.5–3 GPa) (Hirose and Kawamura, 1994; Hirose and Kushiro, 1998; Kushiro, 1996), Hawaiian peridotite HPY (1–3 GPa) (Falloon et al., 1988), depleted harzburgite HZ (1.5–2.5 GPa) (Falloon and Danyushevsky, 2000) and K2O-rich peridotite (1 and 3 GPa) (Davis and Hirschmann, 2013; Laporte et al., 2014) melts (Na2O + K2O > 12% melts are not shown) (Supplementary Table 3). Peridotite HK-66 has <40 wt% olivine in both our model (Fig. 1) and the natural sample (Hirose and Kushiro, 1993) and should be considered as a transitional lithology between pyroxenite and peridotite. L-OI, typical olivine fractionation trend. Only basalts with MgO > 7.5 wt% from Eastern China are shown in plots of FeO versus SiO2 and CaO–13.81 + 0.274 * MgO versus Na2O + K2O.
Fig. 7. FeO$^+$/CaO-3*MgO/SiO$_2$ versus MgO, Na$_2$O+K$_2$O, Mg#, Fe/Mn and La/Yb of experimental peridotite and pyroxenite melts and representative basalts from Eastern China. Experimental peridotite and pyroxenite melts and Mesozoic and Cenozoic basalts in (a–d) are the same as those in Fig. 6. (e) is modified from Yang and Zhou (2013). Hannuoba basalts (Liu et al., 2015; Zhi et al., 1990) that have the most evident compositional variations are indicated in (e). Hornblende peridotite melts from Pilet et al. (2008). Olivine fractionation and accumulation trends, and fractional (FC) and equilibrium (EC) clinopyroxene fractionation trends are shown (Yang and Zhou, 2013). Two melting curves of average pyroxenite with low-La/Yb (1.42) and high-La/Yb (2.78) sources are also shown. Metasomatized and PM-like peridotite have different major element characteristics but the same assumed source La/Yb (1.58). The La/Yb values of peridotite and pyroxenite melts are calculated using a non-modal batch melting model. FeO$^+$/CaO-3*MgO/SiO$_2$ of peridotite and pyroxenite melts are calculated using their quantitative relationships with melting degree (F%). 1σ of FeO$^+$/CaO-3*MgO/SiO$_2$ in (b) and (e) are mainly due to experimental melt errors and the small pressure effect. The broken curves in (a–e) represent the highest FC/MS value of peridotite-derived melts if olivine-dominated fractionation occurs. For other details, see Yang and Zhou (2013). The Mn content of experimental peridotite melts has large analytical error compared with Fe and other major elements. Thus there are no significant pressure relations with Fe/Mn ratio of melts for a specific source rock. Importantly, recent experimental work (Davis et al., 2013) and our calculation shown in Fig. 4 suggest that Fe/Mn are largely controlled by source composition.
reasonably account for the overall positive correlation of FC3MS and La/Yb than simple clinopyroxene fractional crystallization trends (Fig. 7e). In some cases, the pyroxenite M5–40 model melting trends are in good agreement with the correlation of FC3MS and La/Yb of continental OIB-type basalts when MgO = 5–12% (see Fig. 3c in (Yang and Zhou, 2013)) and the Hannuoba basalts (Liu et al., 2015; Zhi et al., 1990) when MgO = 5–11% (see Fig. 7e). Note that low-MgO (<6–10 wt%) basalts are commonly considered as highly evolved products and excluded in discussing their petrogenesis, but the low-MgO basalts shown here and our previous work (Yang and Zhou, 2013) with higher FC3MS value and higher La/Yb than cogenetic high-MgO basalts, which consistent with low degree partial melts of pyroxenite source. Alternatively, peridotite-derived melts can also produce comparable geochemical trends between FC3MS and La/Yb if garnet and clinopyroxene-dominated high pressure fractional crystallization occur, but this mechanism needs high temperature MgO-rich melts produced first. As we discussed below, at normal mantle potential temperature garnet pyroxenite would easily partially melted even when surrounding peridotite still subsolidus. Therefore, the low-MgO and high-MgO Cenozoic OIB-type basalts from Eastern China are almost primary pyroxenite melts, indicating that caution should be used when inferring the source lithology and petrogenesis of both low- and high-MgO basaltic rocks. This also suggests that the commonly used criteria for primary mantle-derived magmas, such as high MgO and Mg#, (Frey et al., 1978), are not suitable for these primary pyroxenite-derived mantle magmas. A similar conclusion has been reached by Lambart et al. (2009). For Mesozoic basalts in Eastern China, all of their compositional characteristics can be explained by partial melts from heterogeneous peridotite, but pyroxenite sources cannot be fully excluded because high-MgO pyroxenite melts can also account for the compositional characteristics of the basalts (Figs. 6 and 7).

There are also some Cenozoic alkaline basalts with very low SiO2 content (40–45 wt%) in Eastern China, and they are commonly have higher FeO# contents than those high SiO2 (>45 wt%) basalts (e.g. Sakuyama et al., 2013; Xu, 2014; Xu et al., 2014; Zeng et al., 2010). These low silica and high iron basalts have been experimentally produced by melting of pyroxenitic rocks of hornblende (Pilet et al., 2008) and carbonated eclogite (Dasgupta et al., 2006), and of peridotitic rocks of carbonated peridotite (Dasgupta et al., 2007; Dasgupta et al., 2013). Although carbonated peridotite has been suggested as the source lithology of basalts in Eastern China (Sakuyama et al., 2013; Zeng et al., 2010), their partial melts are too low in FC3MS value (<0.2) (Dasgupta et al., 2007; Dasgupta et al., 2013) to explain the high FC3MS (0.2–0.8, when MgO > 7.5 wt%) basalts in Eastern China (e.g. Sakuyama et al., 2013; Xu, 2014; Xu et al., 2014; Zeng et al., 2010) (See Fig. 3 in Yang and Zhou, 2013). Therefore, either pyroxenite source (hornblende, carbonated eclogite or low silica pyroxenite) or large amount of pyroxenite-derived melts are mixed with carbonated peridotite-derived melts can explain the major elements characteristics of these low silica basalts in Eastern China. Because typical low FC3MS (<0.2) carbonated peridotite-derived basalts have not been reported along with these high FC3MS basalts, the pyroxenitic rocks are the most likely source materials of basalts in Eastern China.

Using the olivine-addition method of Hainan basalts, Wang and co-workers propose that the mantle potential temperature varies from 1500 to 1580 °C in the Hainan area (Wang et al., 2012). They argue that this high mantle potential temperature and the high-foil olivine phenocrysts and recycled oceanic crust in their source region provide independent petrological evidence for the Hainan plume model. However, their model primary Hainan basalts, with the FC3MS value varying from 0.11 to 0.74 (Mg# = 72–73) (Wang et al., 2012), are highly inconsistent with peridotite melts (Fig. 7c). In addition, high-foil (89–91) olivines in Hainan basalts show a large variation in CaO content (Fig. 3), regardless of their textural characteristics (Wang et al., 2012). Although olivines with CaO < 0.1% have been considered mantle xenocrysts, high-CaO olivines with CaO = 0.1–0.25% may also disaggregated from high-temperature re-equilibrated peridotite (Fig. 3). Note that high-Fo olivine-bearing primitive samples (MgO > 9 wt% and CaO > 8 wt%), which have been used to calculate the primary melt compositions of Hainan basalts, have FC3MS values varying from 0.56 to 0.89 (Na2O + K2O = 3.34–4.96 wt%) (Wang et al., 2012), which are the typical pyroxenite melt characteristics and cannot be accounted for by peridotite melts (Fig. 7a, b). Thus, the model partial melting temperature and pressure of primary Hainan basalts based on a peridotite source exhibit high uncertainties. Trace element ratios, such as La/Yb and Zr/HF, have been frequently used to trace the garnet signature in basalt sources. Using this method, Huang et al. (2013) argued that Cenozoic OIB-like basalts from Guangdong in Southeast China may represent the partial melting of the asthenospheric mantle beneath the thickened lithosphere. This argument holds only for peridotite sources. However, these OIB-like basalts (MgO = 8–12%, Na2O + K2O = 4.6–6.6, Mg# = 55–63) exhibit FC3MS values (0.53–0.92) (Huang et al., 2013) that are not sufficiently low to be explained by peridotite-derived melts (Fig. 7a–c). If the basalts are pyroxenite-derived products, the typical garnet signature implied by the LREE/HREE ratios will not suggest a high-pressure garnet peridotite source because garnet can be stable at low pressures for pyroxenite sources. Therefore, the primary magma composition and its origin temperature and pressure cannot be accurately constrained if pyroxenite-derived basalts are erroneously considered peridotite sources.

For our pyroxenite model, at normal mantle potential temperatures of 1350–1400 °C, adiabatic upwelling of the mantle will result in 0–40% partial melting of the pyroxenite M5–40-like source rock when melting occurs between 2 and 3 GPa, but the peridotite is still near-solidus at the same temperature and pressure (Fig. 8). The high degree partial melting of pyroxenite seems cannot explain the enriched trace elements of the Cenozoic basalts from Eastern China, however, low degree partial melts derived from MORB and/or asthenosphere may react with peridotite to generate trace element-enriched pyroxenite, which can produce basaltic melts with enriched trace elements under very high degree partial melting (Fig. 7e). Different degrees of partial melts may erupt at the surface depending on the final extracting pressure, and all of these melts will present typical garnet signatures even when partial melting occurs at ~2 GPa. In the source region, the peridotite wall rock may re-equilibrate at the subsolidus temperature because of the adiabatic upwelling of the mantle. These near-solidus peridotite and other low-temperature mantle xenoliths may be accidentally picked up by
pyroxene-derived basaltic magmas during their ascent through the upper mantle. Below the solidus of peridotite, partial melting of low-MgO pyroxenite would yield primary melts not in equilibrium with the hosting mantle peridotite, this would lead pyroxene-derived melt interact with surrounding mantle, the major elements characteristic of pyroxenite primary magma would not be preserved any more if they reach equilibrium with mantle peridotite. However, the pyroxenite source signature can be preserved if the magmas migrate to surface rapidly when the surrounding mantle is partial melted (Lambart et al., 2012). Therefore, for Hainan basalts and the other Cenozoic basalts from Eastern China shown in Fig. 7, pyroxene-derived melts contaminated by small amount of disaggregated xenocrysts from peridotite can explain why many low-MgO basalts carry high Fo olivines that show different minor element characteristics compared with olivines crystallized from typical peridotite melts.

5. Concluding remarks

The commonly used markers for the pyroxenite contribution in basalts are briefly summarized, and some new alternative interpretations using a peridotite model are explored to explain the pyroxenite signatures displayed by literature markers of the Cenozoic and Mesozoic basalts from Eastern China. The uncertainties of the commonly used markers for pyroxenite contribution in basalts from Eastern China may also occur for many other oceanic and continental basalts. Although the Cenozoic basalts with SiO2 = 45–53 wt% collectively show compositional characteristics that are indistinguishable from those of peridotite-derived melts, their whole-rock FC3MS values, combined with Na2O + K2O, Mg#, MgO and La/Yb, suggests that they are mainly compositional characteristics that are indistinguishable from those of the Cenozoic and Mesozoic basalts from Eastern China shown in Fig. 7, pyroxenite-derived melts contaminated by small amount of disaggregated xenocrysts from peridotite can explain why many low-MgO basalts carry high-Fo olivines that show different minor element characteristics compared with olivines crystallized from typical peridotite melts.

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Appendix A. Supplementary data

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References


