Invited review

Lithosphere thickness controls the extent of mantle melting, depth of melt extraction and basalt compositions in all tectonic settings on Earth – A review and new perspectives

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ARTICLE INFO

Keywords: Unifying governing variable on global basalt magmatism
Lid effect
Lithospheric thickness control
Basalt compositions
Mid-ocean ridges basalts
Intra-plate ocean island basalts
Volcanic arc basalts
Continental interior basalts
Large igneous provinces
Paradigm change

ABSTRACT

Basalts and basaltic rocks are the most abundant igneous rocks on the earth and their petrologic and geochemical studies have formed our knowledge base on the thermal structure and composition of the mantle with which we have developed workable models on the chemical differentiation of the earth. All this would not have been possible without innovative and painstaking experimental petrology on mantle peridotite melting, basaltic magma generation and evolution largely done in the period of 1960s -1980s. However, the ~30 year lively debate on the nature of “primary magma” among experimental petrologists and the petrology community during this time had inadvertently shelved the development of consensus models on mantle melting in the context of plate tectonics. Continued experimental petrology in parallel with worldwide sampling and study of mid-ocean ridge basalts (MORB) brought about new insights, culminating with a model in 1980s that mantle potential temperature (T_MP) variation controls the extent and pressure of mantle melting and basalt compositions. The tenet of this model is that hotter rising mantle begins to melt deeper and thus has greater decompression depth interval to melt more with the melt having the petrological signature of higher extent and pressure of melting than cooler mantle. This model has gained wide acceptance in MORB studies and has also been invoked in the study of intra-plate basalts in ocean basins and in continental settings. Basalt generation above subduction zones, on the other hand, has been generally accepted as resulting from slab-dehydration induced mantle wedge melting since early 1980s, but recent studies also advocate mantle temperature variation as the primary control on the extent of mantle melting. All these views with laudable merits have formed a paradigm on mantle melting and basaltic magmatism. In this paper, I review the historical developments towards this paradigm and demonstrate in simple clarity that it is the lithosphere thickness, not T_MP, that controls the extent of mantle melting, depth of melt extraction and basalt compositions, i.e., the lid effect. The lithospheric lid caps the rising melting mantle, thus limiting the extent of decompression melting and equilibrium pressure/depth of melt extraction, which is well registered in the compositions of MORB, intra-plate ocean island basalts (OIB), volcanic arc basalts above subduction zones (VAB) and basalts in continental interiors (CIB). Hence, lithosphere thickness is the governing variable that controls mantle melt compositions in all tectonic settings on earth. Major element compositions (e.g., Si-Mg-Fe) of erupted basalts have no memory of initial depth of melting because of effective and efficient melt-solid (e.g., olivine [Mg,Fe]_2SiO_4) equilibration in the rising melting mantle. Therefore, basalt-olivine based thermobarometry, albeit useful, supplies no information on T_MP. It is also the lithosphere thickness that controls whether “mantle plumes” can surface or not and the large igneous provinces (LIPs) serve as effective manifestations for thin or thinned lithosphere at the time of emplacement. This new understanding based on global observations, well-understood experimental petrology and rigorous analysis is fundamental and requires a major change to the current paradigm.

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https://doi.org/10.1016/j.earscirev.2021.103614

Received 24 December 2020; Received in revised form 19 March 2021; Accepted 27 March 2021
Available online 31 March 2021

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1. Introduction

Basalts and basaltic rocks are the most abundant igneous rocks on the earth and their petrology and geochemistry have been used to infer the thermal structure and composition of the mantle and to research the chemical differentiation of the earth (e.g., Zindler and Hart, 1986; Hofmann, 1988, 1997; Wylie, 1988a, 1988b; Sun and McDonough, 1989; McDonough and Sun, 1995; Herzberg and O’Hara, 2002; Herzberg et al., 2007; Putirka, 2005, 2008; Lee et al., 2009). However, it was unclear until 1960s that the mantle consists of peridotites whose partial melting produces basaltic magmas thanks to experimental petrology (e.g., Ringwood, 1962; Green and Ringwood, 1963, 1964, 1967, 1970; O’Hara and Yoder Jr., 1963, 1967; O’Hara, 1963, 1965, 1967, 1968a, 1968b, 1970; Green, 1968; Kushiro, 1968, 1973; Presnall et al., 1979).

Yet debate continued on what may actually cause the partial melting until 1970s when decompression melting became gradually accepted as the important mechanism (Carmichael et al., 1974; Yoder, 1976) although this concept was conceived in 1950s (Verhoogen, 1954) and developed further by comparing natural basalts with experimental melts in 1960s (Green and Ringwood, 1967). These early experimental studies have formed a solid foundation for many aspects of our present-day knowledge on basalt genesis in terms of peridotite compositions, pressure and temperature conditions, effect of water and phase equilibria, but had not developed into a consensus paradigm in the context of plate tectonics because of the ~30 year lively debate on the nature of “primary magma” among experimental petrologists and in the basalt petrology community, i.e., (1) picritic liquids formed by 20–30% melting at pressures of ~20–30 kbar or (2) tholeiitic liquids formed by 10–12% melting at shallow depths (~10 kbar) (see review by Niu, 1997, pages 1061–1062).

Continued experimental petrology (e.g., Green, 1971, 1973; Jaques and Green, 1980; Falloon and Green, 1987, 1988; Falloon et al., 1988) in parallel with worldwide sampling and study of mid-ocean ridge basalts (MORB; see sampling expeditions and data given in Klein and Langmuir, 1987, 1989; Brodholt and Batiza, 1989; Niu and Batiza, 1993) brought about new insights, culminating with theoretical models in 1980s that mantle potential temperature ($T_{\text{MP}}$) variation controls the extent and pressure of mantle melting (Dick et al., 1984; McKenzie, 1984; Klein and Langmuir, 1987, 1989; McKenzie and Bickle, 1988; Niu and Batiza, 1991a; Kinzler and Grove, 1992; Langmuir et al., 1992). The most influential model by Langmuir and co-authors (Klein and Langmuir, 1987, 1989; Langmuir et al., 1992; Gale et al., 2014; Dalton et al., 2014) details that hotter rising mantle begins to melt deeper, has taller melting column to melt more, produces thicker crust and shallower ridge depth with the melt having the signature of higher extent and pressure of melting. All this is opposite for ridges above colder mantle. The same ideas have been invoked in the study of intraplate magmatism in ocean basins (e.g., Yang et al., 2003; Putirka, 2008; Armitage et al., 2008; Sager et al., 2016; Jennings et al., 2019) and in continental interiors (e.g., Li et al., 2008; Wang et al., 2008; Plank and Forsyth, 2016).

This model with laudable merits has thus become a paradigm on mantle melting and basaltic magmatism, i.e., $T_{\text{MP}}$ variation controls the extent of mantle melting and basalt compositions. However, this model has basic problems (Niu, 1997, 2004): (1) the petrological parameter used to infer the initial depth of melting (i.e., $F_{\text{Fe}} = \text{FeO wt\% at } \text{MgO } = 8.0 \text{ wt}\%$) is invalid (see Niu and O’Hara, 2008; Niu et al., 2016a); (2) the assumption that decompression melting continues all the way to the Moho igneous crust is not supported by the effect of conductive thermal boundary layer (CTBL) atop the mantle (Niu and Hekinian, 1997a, 1997b; Niu, 2016a); and (3) erupted basalts have no memory of initial depth of melting, but preserve the signature of final depth of melt equilibration at the base of the CTBL (Niu, 2016a). Given the fundamental importance of the basalt problem in addressing issues of mantle dynamics in a global context, it is time to review the progress in the study of basalt petrogenesis, which I have been involved in and my research has dedicated to over the past 30 years. Basalt compositions are determined by [1] fertile mantle source compositions, [2] conditions of mantle melting, and [3] complex magma differentiation processes largely in crustal magma chambers before eruption. In this review, I focus on [2] although [1] can have deterministic effect on [2] (Niu et al., 2001; Niu and O’Hara, 2008). I demonstrate in simple clarity that the first order global MORB systematics is a consequence of the lithospheric thickness variation, termed lid effect. The lid effect is well registered in the compositions of basalts in all tectonic settings, not only in MORB, but also in intra-plate ocean island basalts (OIB), volcanic arc basalts above subduction zones (VAB) and basalts in continental interiors (CIB). It is also the lithospheric lid that controls whether “mantle plumes” can surface or not and large igneous provinces (LIPs) serve as effective manifestations for thin or thinned lithosphere at the time of their emplacement. Because erupted basalts, which only record the final depth of melting and melt equilibration at the base of the CTBL, have no memory of initial depth of melting in terms of olivine-making elements SiO$_2$, FeO and MgO, basalt-olivine-based thermobarometry, albeit useful, provides no information on the initial depth of mantle melting and $T_{\text{MP}}$.

The unifying lid effect as the governing variable that controls the extent of mantle melting and basalt composition in all tectonic settings on earth demands a fundamental change on the current paradigm in order to use basalt petrogenesis as a tool to advance our understanding of global mantle dynamics. In the following, I start by reviewing basic concepts that may not be well connected for many when considering basaltic magmatism in terms of experimental petrology and in the context of global tectonics. I will then show the data and guide the reader to appreciate the lithospheric thickness variation, the governing variable, that controls the compositions of basalts in all tectonic settings on earth. To ease the readability, all the technical details are given in relevant figure captions.

2. Basic geological concepts relevant to mantle melting and magma generation

2.1. Why basalt magmas erupt where they do?

Fig. 1 shows the familiar Earth cross section with the internal layered structure and seismic P-wave and S-wave velocity variation as a function of depth. The equations in [a] state that the material properties such as density ($\rho$), bulk modulus ($K$) and shear modulus ($\mu$) determine P-wave ($V_p$) and S-wave ($V_s$) seismic velocities. The observation that $V_s = 0$ because of $\mu = 0$ tells us that the outer core is entirely liquid. By inference, the fact that at no depth in the mantle $V_s = 0$ but $V_s > 0$ states clearly that the Earth’s mantle is entirely solid with no volumetrically significant melt anywhere and at any depth. Hence, in a global context, mantle melting with magma formation is a highly localized shallow phenomenon. These highly “localized” localities of mantle melting are well understood in the framework of plate tectonics to be associated with plate boundaries (Fig. 2). For example, at divergent boundaries such as mid-ocean ridges, mantle melting produces mid-ocean ridge basalts (MORB) that create the ocean crust, which covers about two-thirds of the earth’s surface (e.g., Macdonald, 1982). At convergent boundaries such as the Izu-Bonin-Mariana island arcs, mantle wedge melting produces volcanic arc basalts (VAB) (e.g., Perfit et al., 1988; Gill, 1981; Arculus, 1981; Tatsumi and Eggins, 1995), which is thought to be responsible for continental crust accretion (e.g., Taylor, 1967; Arculus, 1981) despite the debate (Niu et al., 2013). However, there is also widespread basaltic magmatism in plate interiors away from plate boundaries such as Hawaiian volcanoes. These within-plate basalts have thus been considered as mantle melting “anomalies” associated with anomalously hot mantle or “hotspots” (Wilson, 1965), which were later interpreted as the surface expressions of deep-rooted mantle plumes derived from the lower mantle or core-mantle boundary (Morgan, 1971). Whether mantle plumes exist or not in the earth has been the subject of hot debate (Anderson, 2004; Foulger and Natland, 2003; Davies, 2005; Foulger, 2005, 2010; Campbell, 2005;
Our current understanding of mantle melting and basalt magma formation owes a great deal to the experimental petrology as discussed above. The work by Jacques and Green (1980) is particularly instrumental in demonstrating for the first time how the compositions of mantle melts vary systematically as a function of fertile mantle compositions and the extent and pressure of melting, which laid the foundation for a number of petrological models for MORB petrogenesis (e.g., Klein and Langmuir, 1987; McKenzie and Bickle, 1988; Niu and Batiza, 1991a, 1991b; Kinzler and Grove, 1992; Langmuir et al., 1992). Fig. 4 shows major element compositions of primary MORB melts produced by decompression polybaric melting using one of these models (Niu, 1997) improved upon by incorporating diamond-aggregate constrained higher-pressure experimental melts (Hirose and Kushiro, 1993; Baker and Stolper, 1994). These models well capture the experimental melt compositions in terms of fertile mantle peridotite composition and the extent and pressure of melting, but their direct application to actual polybaric decompression mantle melting is not straightforward because

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2.2. Mechanisms of mantle melting

Fig. 3 illustrates, in pressure-temperature (P-T) space, the ways in which mantle melting may occur. The concepts of solidus, conductive

geothermal gradient, convective geothermal gradient (i.e., the adiabatic or adiabatic geotherm) and mantle potential temperature (T_M) are all self-explained. The effect of minute volatiles on the solidus is insignificant for volumetrically important magma generation and can thus be neglected here for clarity (Niu and Green, 2018). The concepts that need clarifying are as follows:

[1] The mantle peridotite below the solidus is always solid, and it can partially melt only when placed onto or above the solidus.

[2] The slope of the adiabat is less steep than that of the solidus (i.e., \( \text{d}T/\text{d}P \text{(ADIABAT)} < \text{d}T/\text{d}P \text{(SOLIDUS)} \)), making decompression melting possible because the adiabatically rising mantle can cross over the solidus at \( P_0 \) (see below).

[3] Adiabatically rising mantle with higher \( T_{MP} \) (e.g., \( T_{MP[hot]} \) beneath intra-place settings; Fig. 3d) intersects the solidus and begins to melt deeper (\( P_D \)) than the rising mantle with lower \( T_{MP} \) (e.g., \( T_{MP[normal]} \) beneath ocean ridges (Fig. 3c).

[4] The lithosphere is the conductive thermal boundary layer (CTBL), whose slope \( dT/dP \text{(CONDUCTIVE)} \) decreases with increasing lithosphere thickness.

[5] Decompression melting begins when the adiabatically rising mantle intersects the solidus at \( P_0 \) (see [3] above) but ends when the melting mantle encounters the base of the lithosphere at \( P_L \), which is the base of the CTBL and is the same as the lithosphere-asthenosphere boundary (LAB; Niu and Green, 2018). The extent of melting (F) is thus proportional to the decompression depth interval, i.e., \( F \propto P_L-P_0 \), which is the case for MORB, OIB, VAB and CIB (see below).

[6] Mantle upwelling beneath ocean ridges is a passive response to plate separation, i.e., plate separation creates a gravitational void to allow the hot and buoyant asthenosphere to rise (McKenzie and Bickle, 1988).

[7] Mantle upwelling beneath intra-place settings is considered active or dynamic because of the hot and buoyant “mantle plumes” (e.g., Sleep, 1990; Davies, 1999).

[8] Because the solidus is a material property, its position and topology in P-T space vary, depending on the compositions, especially the effect of \( \text{H}_2\text{O} \)-dominated volatiles (Fig. 3e; also see Niu and Green, 2018 and Fig. 14c below). The parcel of mantle indicated with a blue open star at that depth in Fig. 3c will remain solid without melting, but introduction of water into the mantle will change the solidus into the wet solidus as shown in Fig. 3e, where the blue open star will become located above the wet solidus, causing melting to occur. Hence, VAB above subduction zones are understood as slab-dehydration induced mantle wedge melting or fluid flux melting, which is consistent with the VAB geochemistry in having elevated abundances of H, O-soluble chemical elements although decompression facilitated melting must also be at work (see Section 5 below).

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the model melts represent the integrated melt compositions without considering continued melt-solid equilibration during the \( P_\text{T}_\text{M} \)-to-\( T_\text{M} \) decompression melting process whereas the erupted basalts do record such equilibration. This is particularly important for \( \text{SiO}_2 \), \( \text{FeO} \) and \( \text{MgO} \) because they are sensitive to the pressure of melt-solid equilibration and because the equilibration is effective and efficient, controlled by olivine (\( ([\text{Mg,Fe}]_\text{SiO}_4) \)), which is the most abundant mineral (> 55 wt%) in the melting mantle. On the other hand, elements such as \( \text{Na}_2\text{O} \) (also incompatible elements such as \( \text{TiO}_2 \), \( \text{P}_2\text{O}_5 \), \( \text{K}_2\text{O} \), \( \text{K}_2\text{O}/\text{TiO}_2 \) and \( \text{CaO}/\text{Al}_2\text{O}_3 \)) are incompatible in olivine and are thus insensitive to the melt-solid equilibration. Consequently, the erupted basalts are expected to have the signatures of the extent of melting (see \( \text{Na}_2\text{O} \) and \( \text{CaO}/\text{Al}_2\text{O}_3 \) in Fig. 4) and final depth/pressure (\( P_\text{F} \)) of melt-solid equilibration (i.e., melt extraction depth) if preserved (see below).

### 2.4. Basalts have no memory of initial depth of melting (i.e., \( P_\text{O} \)) in terms of \( \text{SiO}_2 \), \( \text{MgO} \) and \( \text{FeO} \)

Although it is conceptually apparent that hotter rising mantle intersects the mantle solidus deeper (Fig. 3) and the melt would have the compositional signature of higher pressure of melting (i.e., higher \( \text{MgO} \), \( \text{FeO} \) and lower \( \text{SiO}_2 \)) than cooler mantle. This is unlikely to be true in practice because of the continued and inevitable melt-solid equilibration during the \( P_\text{T}_\text{M} \)-to-\( T_\text{M} \) decompression melting. The pressure signature in basalts, if recorded and discernable, would be the final depth of melting (\( P_\text{F} \)), which is the depth of melting cessation and melt extraction (Niu, 1997, 2016a). This can be illustrated through Fig. 5 using the well-understood ocean ridge mantle melting processes (Niu, 1997, 2004, 2016a).

**Fig. 5** shows plate separation induced passive mantle upwelling and decompression melting beneath ocean ridges. The decompression melting begins when the upwelling mantle intersects the solidus at \( P_\text{O} \) (corresponding to \( T_\text{O} \approx T_\text{MP} = T_\text{O} - \text{P}_\text{O} \times 1.8 \text{ °C} \); Niu and O’Hara, 2008), continues in the melting region \( \bigcirc \) and stops at \( P_\text{F} \) (corresponding \( T_\text{F} \)) at which the upwelling mantle encounters the CTBL (region \( \bigtriangledown \)). Because of the buoyancy contrast, the melt (red-arrowed dash lines) is extracted from the residue (blue-arrowed think lines) to form the ocean crust. The residue contributes to the new accretion of the mantle lithosphere, which, when tectonically exposed on the seafloor, is sampled as abyssal peridotites (Dick et al., 1984; Dick, 1989; Niu, 1997, 2004). Mantle melting is considered to be close to fractional polybaric melting and the decompression melting mantle would have very small melt porosity (or melt retention), widely believed to be < 1% (e.g., McKenzie, 1985; Johnson et al., 1990; Langmuir et al., 1992; Spiegelman and Elliott, 1993; Lundstrom et al., 1995; Sims et al., 2002). To be conservative, we can assume that in the \( P_\text{T}_\text{M}-P_\text{F} \) decompression melting region \( \bigcirc \), there exists 1–2% melt retention (Niu, 1997) in close physical contact with 98–99% solid matrix dominated by olivine (\( ([\text{Mg,Fe}]_\text{SiO}_4) \)). Such small melt/solid ratio ensures effective and efficient melt-solid re-equilibration, at the very least for elements \( \text{Si} \), \( \text{Fe} \) and \( \text{Mg} \) controlled by olivine. The 1000’s of years of melting time is really long enough to ensure complete melt-solid re-equilibration that is readily achieved in 10’s of hours in peridotite melting experiments. We thus cannot avoid the conclusion that erupted basalts have no memory of initial depth of melting (\( P_\text{O} \)) in terms of olivine-making elements \( \text{Si} \), \( \text{Fe} \) and \( \text{Mg} \), but can record the final depth of melting or melt-solid equilibration (\( P_\text{F} \)) at which the upwelling mantle encounters the CTBL, stops to melt and ends the above-mentioned melt-solid re-equilibration. It is therefore simply invalid to use \( \text{FeO} \) as proxy for \( T_\text{MP} \) in discussing mantle melting (see Fig. 4 and below). This analysis based on ocean ridge mantle melting applies to mantle melting in all other tectonic settings because there always exist CTBLs that cap the decompression melting (see Sections 3–6 below). It is necessary to emphasize in this context that fractional melting is equilibrium (not disequilibrium) melting, and the smaller the melt retention (i.e., melt...
chemical elements incompatible in olivine. Olivine-making elements Si, Mg and Fe although this may not apply to escape from porosity; e.g., where generally understood mechanisms of mantle melting is illustrated in pressure-temperature (P-T) space (Fig. 3).

Y. Niu residue immediately as soon as it is produced no matter how small. By definition, in the process of fractional melting, the melt will leave the residue immediately as soon as it is produced no matter how small. Yet, the basalt-olivine-based thermobarometry, albeit useful, has no significance when discussing TMP, the basalt-olivine-based thermobarometry can thus provide the information on P８ and T８ that are not applicable to chemical elements incompatible in olivine.

Hence, erupted basalts have no memory of the initial depth (P₀) of melting in terms of Si, Mg and Fe, and thus provide no information on T₀ and T₀. Therefore, the basalt-olivine-based thermobarometry, albeit useful, has no significance when discussing TMP. Because basalts can record the initial depth of melting (P₀), the basalt-olivine-based thermobarometry can thus provide the information on P₂ and T₂, thickness of the mantle lithosphere and its thickness increases with decreasing spreading rate (Niu, 1997, 2004, 2016a; Niu and Hekinian, 1997a), MORB may not even record the information about P₂. This statement is significant because abyssal peridotites, which are considered as subridge mantle melting residues, are not simple melting residues, but contain excess olivines and elevated abundances of incompatible elements, ascertaining the fact of MORB melt cooling and olivine crystallization as well as melt referentilization during ascent through the advanced residues in the CTBL (Niu, 1997, 2004; Niu and Hekinian, 1997b; Niu et al., 1997). Hence, MORB melts have varying re-equilibration histories in the CTBL at varying depths shallower than P₂. Therefore, using MORB SiO₂, FeO and MgO (in whatever corrected forms such as SiO₂-FeO, SiO₂-FeO-MgO) to discuss P₂, T₂ and mantle melting processes is invalid and has no significance (Niu, 2016a).

3. The lid-effect on mantle melting at mid-ocean ridges and MORB petrogenesis

The most referenced hypothesis on MORB magmatism at present is the work by Langmuir and co-workers (Klein and Langmuir, 1987, 1989; Langmuir et al., 1992). After Dick et al. (1984), Klein and Langmuir (1987) (abbreviated as KLB87) showed correlated variations of MORB FeO porosity; e.g., << 1–2%) is, the more efficient in reaching equilibrium. By definition, in the process of fractional melting, the melt will leave the residue immediately as produced so it must not produce any chemical elements incompatible in olivine.
Earth-Science Reviews 217 (2021) 103614

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Fig. 4. Model compositions (selected major elements) of mantle melts produced by decompression melting (Niu, 1997; also see Niu and O’Hara, 2008) to illustrate that the compositions of primary mantle melts vary as a function of the initial depth of melting, i.e., the depth at which the rising mantle intersects the solidus (e.g., \( P_0 = 25, 20 \) and 15 kbar), and the increasing extent of melting as the mantle continues to rise. The decompression melting is arbitrarily stopped at 8 kbar for all three melting paths. The point is that SiO\(_2\) (inverse; also weakly Al\(_2\)O\(_3\) not shown), FeO (positive) and MgO (positive) of the primary melts are sensitive to pressure of melting, while CaO/Al\(_2\)O\(_3\) (positive) and Na\(_2\)O (inverse; also TiO\(_2\), K\(_2\)O, P\(_2\)O\(_5\) not shown) are sensitive to the extent of melting.

That is, the compositions of the erupted basalts contain the information on the extent of melting and final depth of melt equilibration, which can be extracted if the effects of shallow level melt evolution can be corrected for and if the effect of mantle source compositional variation can be properly evaluated. More complex and thermodynamics-based models such as pMELTS (Asimow and Ghiorso, 1998; Ghiorso et al., 2002; Smith and Asimow, 2005) are available, but the model data presented here are effective to capture the compositional systematics of primary mantle melts by decompression melting. See text for caveats in applying these models. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

3.1. The merit and errors of KL87 on global MORB petrogenesis

Langmuir et al. (1992) further elaborated quantitatively that the global MORB FeO\(_8\) variation can be used to calculate the mantle solidus temperature (\( T_0 \propto T_{MP} \)). Fig. 6 summarizes the ideas and errors of KL87. By interpreting the inverse correlation of ridge segment averaged MORB FeO\(_8\) with ridge depth and the positive correlation of MORB Na\(_2\)O with ridge depth (Fig. 6c), KL87 states that MORB erupted at shallow ridges represent hot mantle that begins to melt deeper with higher extent of melting whereas MORB erupted at deep ridges reflect cooler mantle that begins to melt shallower with lower extent of melting, resulting in thicker crust beneath shallow ridges than beneath deep ridges (Fig. 6a, b). While this interpretation is simple and attractive, the fundamental assumptions are in error (Niu, 1997, 2004, 2016a; Niu and Hekinian, 1997a; Niu and O’Hara, 2008) despite its popular reference and influence.

3.1. The merit and errors of KL87 on global MORB petrogenesis

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1. The assumption that decompression melting continues all the way to the Moho ignores the near-seafloor cooling and thus ignores the presence and effect of the CTBL (e.g., Niu, 1997; Niu et al., 1997).

2. The assumption that MORB mantle has uniform major element compositions neglects the effect of mantle source heterogeneity on the observed MORB chemistry.

3. The assumption that FeO (or Fe\(_8\)) in MORB melts directly indicates the initial depth of mantle melting ignores the inevitable melt-solid equilibration in the melting mantle (see Fig. 5 and above). Hence, the calculated \( P_0, T_0 \) and \( T_{MP} \) using the single variable Fe\(_8\) have no significance (Fig. 6).
This concept applies to all settings: erupted basalts have no memory of $T_0$ (thus no $T_0$ and $T_{MP}$ information), but can preserve $P_F$.

Note: MORB record $P_F$ or variably less because of melt cooling during transport in the CTBL recorded in abyssal peridotites.

Fig. 5. Schematic (left) and qualitative (right) illustrations of sub-ridge thermal structure with all the elements self-explained. The plate separation induced decompression melting begins when the upwelling mantle intersects the solidus at $P_0$ (corresponding to $T_0$) and continues until the upwelling melting mantle reaches $P_F$ (corresponding to $T_F$), the final depth of melting or melt-solid equilibration, which is the very base of the conductive ("cold") thermal boundary layer (CTBL; region Ω) beneath the ridge. Because of the buoyancy contrast, the melt (red-arrowed thin dash lines) ascends faster than the solid residue (blue-arrowed think lines). The melt is extracted to form the ocean crust and the residue contributes to the accretion of the lithospheric mantle. The latter, when tectonically exposed on the seafloor, is sampled as abyssal peridotites. The globally large mantle temperature variation of $ΔT_{MP}$ from $P_0$ to $P_F$, the 1–2% (or much less) melt in physical contact with 98–99% (or much more) solid matrix ensures effective and efficient melt-solid re-equilibration, especially for elements Si, Mg and Fe controlled by olivine, the most abundant mantle mineral. The 1000’s of years of melting time ensure complete melt-solid re-equilibration that is readily achieved in 10’s of hours in peridotite melting experiments. Therefore, if MORB melts indeed record pressure signature of melting, the signature must be $P_F$ not $P_0$ (Niu and O'Hara, 2008; Niu et al., 2011; Niu, 1997, 2004, 2016a). This unavoidable conclusion based on objective, logical and rigorous analysis has been overlooked because readers conveniently stick to the widely referenced earlier model (Klein and Langmuir, 1987) without having thought about the experimental petrology. The plain language is as follows: [1] erupted MORB melts have no memory of $P_0$ in terms of SiO$_2$, FeO and MgO, but can preserve the signature of $P_F$; [2] parameters such as Fe$_{90}$ have no significance in discussing $T_{MP}$ and the claimed global $ΔT_{MP}$ ($ΔP_{MP}$) $≈$ $250$ K results from a petrologic misunderstanding (see Fig. 6); [3] the concept and conclusion rigorously illustrated here apply to mantle melting and basaltic magmatism in all tectonic settings on the earth; [4] MORB melts may not even record $P_F$ because MORB melt, during ascent, crystallizes and adds olivine in the advanced residues in region Ω as revealed from abyssal peridotites (Niu, 1997, 2004; Niu and Hekinian, 1997b; Niu et al., 1997). The concept of $P_F$ at depth deeper than the Moho was introduced by Niu and Batiza (1991a) (for interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

4] The use of MORB Fe$_{90}$, corresponding to variably evolved MORB melts with Mg# $= 0.56–0.68$ to discuss mantle melting processes, violates basic petrological principles; MORB melts with Mg# $≥ 0.72$ in equilibrium with mantle olivine of Fo$_{90}$ can be used to discuss mantle processes (Niu and O’Hara, 2008). This simple and objective analysis further substantiates that Fe$_{90}$-based calculations for Fo$_{90}$, $T_0$, $T_{MP}$ and $T_{SP}$ have no significance. Note that Gale et al. (2014) re-corrected the data to show Fe$_{90}$ = Fe$_{90}$ (MORB melt in equilibrium with mantle olivine Fo$_{90}$), but this Fe$_{90}$ = Fe$_{90}$ is entirely incorrect because it is petrologically not possible as illustrated in simple clarity in Fig. 3 of Niu (2016a).

5] Therefore, the conclusion on the basis of MORB Fe$_{90}$ (or any other form such as Fe$_{90}$) that global ocean ridge $T_{MP}$ variation (up to $ΔT_{MP} = 250$ K) controls the initial depth/pressure and extent of melting is entirely misleading.

One may wish to relate MORB Na$_{90}$ to modal clinopyroxene (Cpx) in spatially related abyssal peridotites (AP; Fig. 7a) as evidence in support of $T_{MP}$ control on the extent of mantle melting because both are expected to decrease with increasing extent of melting towards beneath shallow ridges (see Niu et al., 1997). However, this correlation can be readily understood as inherited from the sub-ridge fertile mantle compositional variation reflecting previous melting and melt extraction histories with compositionally depleted (low Na$_{2}$O, low Cpx mode) and physically buoyant mantle beneath shallow ridges (Fig. 7b; Niu, 2004, 2016a). The AP samples are from fracture zone walls 0.5 to 14 Ma old, yet the spatially associated basalts are from the present-day ridge axis (Dick et al., 2007). Despite the compositional complementarity between MORB (melt) and AP (residue), it is unlikely for the ~14 Ma old AP to be melting residues of the present-day MORB. Mantle temperature anomalies could be long-lived but will decay away with time towards increased ridge depths. Compositional anomalies (buoyant mantle beneath shallow ridges), however, will be long-lasting and will not decay away with time (Niu, 2016a).

3.2. The lid-effect on ridge mantle melting: evidence from the correlation of MORB chemistry with spreading rate

With the understanding that ocean ridges are passive features and that sub-ridge mantle melting results from plate-separation induced asthenospheric upwelling and decompression (McKenzie and Bickle, 1988), it is thus logical to predict that plate separation rate must affect the extent of mantle melting and MORB composition. This is because the rate of mantle upwelling is proportional to the rate of plate separation (Reid and Jackson, 1981; Phipps Morgan, 1987; Niu, 1992, 1997). To test this hypothesis, Niu and Hekinian (1997a) analyzed the global MORB data and limited AP data available then and demonstrated that the extent of ocean ridge mantle melting increases with increasing plate
spreading rate. Langmuir and co-workers denied this finding and stated that “There is no correlation between the chemical parameters and spreading rate” (Gale et al., 2014). In response to this denial, I used the updated global MORB data compiled by these authors and substantiated the average compositions of global MORB (in terms of K$\text{O}$ as of MgO $\approx$ 8 wt%) with ridge depth and the positive correlation of MORB NaO ($\text{Na}_2\text{O}$ at MgO $\approx$ 8 wt%) with ridge depth (c), KL87 states that MORB erupted at shallow ridges represent hot mantle that begins to melt deeper (high FeO) with higher extent of melting (low NaO) whereas MORB erupted at deep ridges reflect cooler mantle that begins to melt shallower (low FeO) with lower extent of melting (high NaO), resulting in thicker crust beneath shallow ridges than beneath deep ridges (a,b). While this interpretation has been popular, [1] the assumption that decompression melting continues all the way to the Moho by ignoring the near-surface conductive cooling and [2] the misuse of MORB FeO as proxy for mantle solidus depth and temperature have misguided the community on ocean ridge mantle melting (see Niu and O’Hara, 2008; Niu, 2016a). See text for details.

3.3. The lid-effect on ridge mantle melting: evidence from the correlation of MORB chemistry with ridge depth

While the ~60,000 km long globe-encircling ocean ridge system is considered to be the largest mountain ranges in ocean basins, the ridge depth varies significantly from near sea level around Iceland to up to 5500 m below sea level in the Cayman Trough. On average, the shallower ridge depth along the fast-spreading East Pacific Rise and the deeper ridge depth at many slow-spreading ridge segments suggest that ridge depth may increase with decreasing spreading rate, but this expected correlation is essentially absent with large amplitude of ridge depth variation towards slow-spreading ridges (Niu, 2016a). Hence, any correlation between spreading rate and ridge depth, if present, is insignificant. The large ridge depth variation must be isostatically compensated by the density variation of the underlying mantle material. Therefore, alternative variables must be explored. Mantle temperature variation and compositional variation are possible causes because earth’s surface elevation, including seafloor depth variation, must correlate with density structure of the subsurface (below seafloor) rocks as a function of temperature or composition. Except for ridges around Iceland where mantle temperature may be high, there is no evidence for large sub-ridge mantle temperature variation on a global scale, perhaps, no more than 50 K if any (Niu and O’Hara, 2008). The interpreted 250 K mantle temperature variation advocated by Langmuir’s research team (see Fig. 6) does not exist because it is an artifact as the result of using MORB FeO as proxy for mantle solidus depth and temperature, defined as proxy for mantle solidus depth and temperature, that is invalid and has no petrological significance (Niu and Hékinian, 1997a; McKenzie and Bickle, 1988). By interpreting the inverse correlation of ridge segment-averaged MORB FeO ($\text{Fe}_2\text{O}_3$ at MgO $\approx$ 8 wt%) with ridge depth and the positive correlation of MORB NaO ($\text{Na}_2\text{O}$ at MgO $\approx$ 8 wt%) with ridge depth (c), KL87 states that MORB erupted at shallow ridges represent hot mantle that begins to melt deeper (high FeO) with higher extent of melting (low NaO) whereas MORB erupted at deep ridges reflect cooler mantle that begins to melt shallower (low FeO) with lower extent of melting (high NaO), resulting in thicker crust beneath shallow ridges than beneath deep ridges (a,b). While this interpretation has been popular, [1] the assumption that decompression melting continues all the way to the Moho by ignoring the near-surface conductive cooling and [2] the misuse of MORB FeO as proxy for mantle solidus depth and temperature have misguided the community on ocean ridge mantle melting (see Niu and O’Hara, 2008; Niu, 2016a). See text for details.

Fig. 6. Cartoons and interpretations of the widely referenced model of ocean ridge mantle melting that stresses the primary control of mantle potential temperature variation (Klein and Langmuir, 1987 [KL87]; McKenzie and Bickle, 1988). By interpreting the inverse correlation of ridge segment-averaged MORB FeO ($\text{Fe}_2\text{O}_3$ at MgO $\approx$ 8 wt%) with ridge depth and the positive correlation of MORB NaO ($\text{Na}_2\text{O}$ at MgO $\approx$ 8 wt%) with ridge depth (c), KL87 states that MORB erupted at shallow ridges represent hot mantle that begins to melt deeper (high FeO) with higher extent of melting (low NaO) whereas MORB erupted at deep ridges reflect cooler mantle that begins to melt shallower (low FeO) with lower extent of melting (high NaO), resulting in thicker crust beneath shallow ridges than beneath deep ridges (a,b). While this interpretation has been popular, [1] the assumption that decompression melting continues all the way to the Moho by ignoring the near-surface conductive cooling and [2] the misuse of MORB FeO as proxy for mantle solidus depth and temperature have misguided the community on ocean ridge mantle melting (see Niu and O’Hara, 2008; Niu, 2016a). See text for details.
beneath deep ridges upwells reluctantly in response to plate separation, modal garnet, jadeite/diopside ratio in clinopyroxene, and pyroxenes/shallow (see Niu and Oshikawa, 2004, 2016a). This significant correlation between the two is apparently consistent with their interpretation that high extent of melting (low Na$_{8}$) and melt extraction (low Cpx in AP) is associated with shallow ridges above hot mantle whereas low extent of melting (high Na$_{8}$) and melt extraction (high Cpx in AP) is associated with deep ridges above cool mantle. The diagram is based on the Cpx-Na$_{8}$ data by Dick et al. (1984). The 18 localities and the data used are given in Niu et al. (1997). [b] offers an alternative interpretation that the correlation in [a] does not tell anything about the mantle temperature variation and varying extent of melting but is inherited from the sub-ridge fertile mantle compositional variation reflecting previous melting and melt extraction histories (Niu, 2004, 2016a).


Fig. 9c illustrates consistently the above predicted correlations. The mantle source is progressively more enriched (or less depleted) from beneath shallow ridges to beneath deep ridges, pointing to increasing modal garnet, jadeite/diopside ratio in clinopyroxene, and pyroxenes/olivine ratio, thus a progressively denser mineral assemblage towards beneath deeper ridges. This explains straightforwardly why deep ridges underlain by fertile and dense asthenospheric mantle are deep and why shallow ridges underlain by depleted and less dense asthenosphere are shallow (see Niu and O’Hara, 2008). In addition, dense fertile mantle beneath deep ridges upwells reluctantly in response to plate separation, which leads to limited extent/amplitude of upwelling, allowing conductive cooling to penetrate to a great depth, making a thickened conductive thermal boundary layer (CTBL, region $\odot$), forcing melting to stop at a deep level (P$_{d}$), thus having a short melting interval (P$_{o}$-P$_{d}$), melting less, and producing probably a thin magmatic crust relative to the more refractory mantle beneath shallow ridges. The effect of mantle solidus depth variation due to mantle compositional fertility variation is insignificant (see the analysis and Fig. 15 of Niu and O’Hara, 2008). Therefore, the correlations in Fig. 9a,b also result from the lid-effect. Hence, the global MORB compositional systematics (Fig. 9a,b) are the net effect of [1] fertile mantle source inheritance and [2] varying extent of melting controlled by the varying amplitude of upwelling and decompression melting as the result of mantle density variations ultimately still controlled by fertile mantle compositional variation. The recent study by Dick and co-workers supports this understanding (Zhou and Dick, 2013; Dick and Zhou, 2015). Fertile mantle compositional control on mantle melting processes and MORB compositions have been reported in many regional and local studies (e.g., Schilling et al., 1983; Langmuir et al., 1986; Natland, 1989; Sinton et al., 1991; Perfit et al., 1994; Niu and Batiza, 1994; Castillo et al., 1998; Niu et al., 1999, 2001; Michael et al., 2003; Gill et al., 2016). It is particularly important to note that for a given ridge segment or fertile mantle domain, MORB derived from an isotopically and incompatible element enriched source have low FeO and CaO/Al$_{2}$O$_{3}$ (e.g., Batiza and Niu, 1992; Niu et al., 2002), which echoes the observations in Fig. 9a,b. Importantly, a recent near-ridge seamount MORB study shows significantly correlated variations of Fe isotope ratios with radiogenic isotopes and the abundances and ratios of incompatible elements (Sun et al., 2020a).

It is important to note that for a thermal expansion coefficient of $\alpha = 3 \times 10^{-5}$ K$^{-1}$ for mantle peridotites, the effect of temperature variation on mantle density is rather small compared to the effect of compositional variation. For example, mantle density decrease of ~1% for compositionally depleted mantle as a result of prior melt extraction would be equivalent to raising mantle temperature by 330 K (Niu and Batiza, 1991b; Niu et al., 2003; Niu and O’Hara, 2008). It is also worth to emphasize that it is straightforward that the enriched (or less depleted) mantle with higher Fe/Mg (e.g., low Mg$^{+} = Mg/[Mg + Fe^{2+}] < 0.89$) is readily understood to be denser than the depleted (or less enriched) mantle with slightly higher Mg/Fe (e.g., high Mg$^{+} = Mg/[Mg + Fe^{2+}] > 0.90$), but the effect of higher Al$_{2}$O$_{3}$ in the enriched (or less depleted) mantle is far more important because of the formation and stability of garnet, which has the greatest density of all the mantle minerals; 1.0 wt % Al$_{2}$O$_{3}$ can make up to 5 wt% garnet (see Niu et al., 2003; Niu and O’Hara, 2008) as illustrated in Fig. 9c. To help readers better appreciate the geodynamic significance of compositional (vs. "thermal") buoyancy contrast, we can compare the largest topographic contrast on the Earth: the compositionally buoyant continental lithosphere (the crust and mantle lithosphere) and the compositionally dense oceanic lithosphere despite the fact that the oceanic lithosphere is "hotter" than continental lithosphere with mean per-square-meter heat flow ratio of 105.4/70.9 $\approx$ 1.49, which is also understood by the steeper geotherm beneath oceans (i.e., $dT/dP_{Oceanic} > dT/dP_{(Continental)}$) (Davies and Davies, 2010; Niu et al., 2003; Niu, 2016a). Although the latter is related to the thin lithosphere beneath oceanic realms than beneath continents, the fact that the thinned lithosphere (\(< 80$ km\) beneath eastern continental China since the Cenozoic (e.g., Guo et al., 2020; Sun et al., 2020b) is thinner than the narrow seafloor lithosphere (\(< 80$ km\)) of eastern China (\(< 200$ m above sea level) and the high plateaus in western China (\(< 1500$ m above sea level) further illustrates in simple clarity the subsurface compositional buoyancy contrasts that determines the surface elevation (see Section 6 below).
4. The lid-effect on mantle melting beneath intra-plate ocean islands for OIB

The recognition of the lid effect on ocean ridge mantle melting (Niu and Hekinian, 1997a; Niu and O’Hara, 2008), along with the work in the literature (Ellam, 1992; Haase, 1996), offers an impetus for further testing the hypothesis. We thus carried out a global analysis of OIB (Humphreys and Niu, 2009; Niu et al., 2011) because the thickness ($T$) of oceanic lithospheric lid varies significantly as a function of seafloor age ($t$), at least for seafloor younger than ~70 Ma, i.e., $T \propto t^{1/2}$ as the result of conductive heat loss to the seafloor and thermal contraction with time and because the thickness ($T$) or age ($t$) of the lithosphere at the time of OIB volcanism is known or can be well constrained (e.g., Parsons and McKenzie, 1978; Sclater et al., 1980; Stein and Stein, 1992; Niu and Green, 2018). Fig. 10a shows the correlated variations of global average OIB compositions with the lithosphere thickness at the time of OIB volcanism (Humphreys and Niu, 2009; Niu et al., 2011), which is wholly consistent with the lid effect, i.e., OIB erupted on thin oceanic lithosphere ($T_{OL}$) have a petrological signature of high extent of melting ($F \propto P_{O} - P_{F}$) and low pressure of melt equilibration (i.e., shallow $P_{F}$ melt extraction), whereas OIB erupted on thick $T_{OL}$ have the signature of low extent of melting ($F \propto P_{O} - P_{F}$) and high pressure of melt equilibration (i.e., deep $P_{F}$ melt extraction) as illustrated in Fig. 10b. The base of the lithospheric lid is conceptually the lithosphere-asthenosphere boundary (LAB; Fig. 3) in terms of mantle melting as it is the pargasite (amphibole) dehydration solidus (e.g., Green et al., 2010; Niu and Green, 2018).

We must point out that intraplate OIB such as basalts erupted on Hawaiian Islands are widely interpreted as resulting from decompression melting of thermally buoyant upwelling mantle plumes. If this interpretation is correct, then the hot upwelling mantle would intersect the solidus at great depths (i.e., shallow $P_{F}$ melt extraction), whereas OIB erupted on thick $T_{OL}$ have the signature of low extent of melting (low $F \propto P_{O}$) and high pressure of melt equilibration (i.e., deep $P_{F}$ melt extraction) as illustrated in Fig. 10b. The base of the lithospheric lid is conceptually the lithosphere-asthenosphere boundary (LAB; Fig. 3) in terms of mantle melting as it is the pargasite (amphibole) dehydration solidus (e.g., Green et al., 2011; Niu and Green, 2018).

However, the OIB data (Fig. 10a) do not show evidence for such varying $P_{O}$ and thus $T_{MP}$, but
Increasing extent of melting

Ridge Axial Depth (m)

Shallowest Ridge

Deepest Ridge

Pyroxene/olivine (decreasing Fe, Mg)
Jadeite/diopside (increasing Na, Al, decreasing Ca)
Garnet (increasing Al)
Garnet rich lithologies - most likely garnet peridotite

Bulk density
Upwelling amplitude
CTB
Melting interval (P_o-P_f)
Extent of melting, F
Ridge deepens

Decreasing ridge axial depth
Increasing extent of melting

(caption on next page)
5. The *lid effect* on mantle wedge melting above subduction zones for VAB: A new and consistent perspective

I use the phrase “a new and consistent perspective” to inform the community that the apparently complex subduction-zone magmatism may in fact be simple thanks to the data compilation by Turner and Langmuir (2015a, 2015b) although my understanding of the data differs from theirs.

Volcanic arc basalts (VAB) are widely accepted as resulting from slab dehydration induced mantle wedge melting. This is also called flux melting because the melting is triggered by influx of water-dominated fluids from the subducting slab, which lowers the mantle wedge solidus and makes melting possible (Fig. 3e). This interpretation is reasonable because VAB have distinct compositions enriched in water and water-soluble elements (e.g., U, K, Sr, Pb) but depleted in water-insoluble elements (e.g., Ti, Nb, Ta) (e.g., Tatsunami and Eggins, 1995). However, whether the extent of mantle wedge melting may vary between VAB systems is unclear, and if it does, what may actually control the extent of melting is also unclear. By emphasizing water-facilitated melting, it is expected that the extent of mantle wedge melting must be higher than sub-ridge mantle melting for MORB because of significantly lowered wet solidus (e.g., Stolper and Newman, 1994), but the extent of melting is also expected to be constrained by the lithospheric lid as discussed above (Figs. 3, 8-10). Indeed, Plank and Langmuir (1988) analyzed the then available global VAB data and showed the presence of correlated variations of arc-averaged VAB major element compositions with arc crustal thickness, i.e., Na2O, Na2O corrected to MgO = 6.0 wt% increases whereas CaO (CaO corrected to MgO = 6.0 wt%) decreases with increasing crustal thickness. The authors interpreted these correlations as resulting from varying extent of melting: thick crust limits melting to a deep depth, giving a VAB petrological signature of low extent of melting (high Na2O and low CaO), whereas thin crust allows melting to stop at a shallow depth, giving a VAB signature of high extent of melting (low Na2O and high CaO). Using the updated global data, Turner and Langmuir (2015a, 2015b) confirmed the same observations using more chemical parameters consistent with varying extent of melting (see Fig. 11a), which is apparently consistent with the *lid effect* although a paradoxical issue must be addressed (see below).

These authors choose to interpret the VAB systematics (Fig. 11a) as controlled by mantle wedge temperature variation as advocated in recent studies (e.g., England and Wilkins, 2004; England et al., 2004; England and Katz, 2010). Following their study of the Chilean Southern volcanic zone VAB (Turner et al., 2016), Turner and Langmuir (2015a, 2015b) interpret the global VAB systematics (Fig. 11a) as resulting from mantle wedge temperature structure variation as illustrated in Fig. 11b-d, i.e., the mantle wedge is up to 250 K hotter beneath thin lithosphere than beneath thick lithosphere with respect to the dry solids. Hence, the hot mantle wedge beneath thin lithosphere melts more than the cold mantle wedge beneath thick lithosphere. This apparently reasonable interpretation is in fact conceptually misleading because the interpreted mantle wedge temperature structure has nothing to do with mantle wedge asthenosphere where melting takes place and whose temperature is controlled by Tmp of the convective mantle wedge asthenosphere and the subducting slab which drives mantle wedge convection while also serving as a heat sink. Objectively and rigorously, the VAB systematics (Fig. 11a) is not caused by mantle wedge temperature variation but is a straightforward consequence of the *lid effect*, which requires that decomposition melting be the dominant mode of mantle wedge melting in addition to flux-melting for the onset of mantle wedge melting. This can be elucidated below with conceptual clarifications and caveats.

5.1. The nature of the lithospheric lid overlying the mantle wedge

From the observation (Fig. 11a) to the interpretation (Fig. 11b-c) is the hidden assumption by Turner and Langmuir (2015a, 2015b) that the arc crust and lithosphere are “equivalent”, which cannot be true unless the crust is in direct contact with the asthenosphere without a lithospheric mantle root. The latter cannot be true without fast lithospheric extension.
Approximate Seafloor Age (Ma)

- Si$_{72}$ (R = 0.825, > 99.0%)
- P$_{72}$ (R = 0.745, > 99.0%)
- Ti$_{72}$ (R = 0.901, > 99.5%)
- [Sm/Yb]$_N$ (R = 0.819, > 99.0%)

Lithosphere-asthenosphere boundary (LAB) depth (km)

Increasing extent of melting

Seafloor spreading and ageing

OIB melt signature of:
- Low F (small $P_f - P_m$)
- High P (shallow $P_f$)

OIB melt signature of:
- High F (large $P_f - P_m$)
- Low P (shallow $P_f$)

Approximate dry mantle solida $P_f$, adiabatically upwelling asthenosphere begins to melt.

Sub-solidus asthenosphere

(caption on next page)
and rifting, which is not the case beneath modern volcanic arcs above subduction zones. The fact that a number of VAB suites contain mantle peridotite xenoliths (e.g., Parkinson and Arculus, 1999; Ionov et al., 2012; Tollan et al., 2015) manifests the presence of mantle lithosphere beneath arc crust. The process of mantle xenolith inclusion can be explained using understood mechanism of mantle xenolith incorporation in some OIB and continental alkali basalts with abundant volatiles. Primitive VAB melts formed in the asthenospheric mantle wedge contain abundant dissolved water, whose exsolution during ascent due to decompression (reduced solubility in the melt; Sparks et al., 2000; Cashman and Blundy, 2000; Blundy and Cashman, 2005) results in volume expansion and elevated viscosity to develop destructive power to incorporate the conduit walls of the lithospheric mantle as xenoliths carried to the surface during VAB eruption (see Sun et al., 2017, 2020b for mechanism). Hence, it is the thickness of the entire lithosphere, i.e., the LAB depth, that caps the mantle wedge melting not the crust (Fig. 11b,c). In this case, the correlation of VAB geochemistry with crustal thickness (Moho depth) may mean the correlation with the LAB depth, i.e., the arc crustal thickness is proportional to the lithosphere thickness (i.e., Moho depth × LAB depth). This is possible because continental arc with thicker continental crust (purple solid circles in Fig. 11a) is expected to have a thicker lithospheric mantle root whereas thinner oceanic arc crust (blue solid circles in Fig. 11a) is expected to have a thinner lithospheric mantle root. This is likely to be true in general inferred from global surface elevation and isostasy, but it is imperative to acquire high resolution seismic data on both LAB and Moho depths to verify this hypothesis in the future.

5.2. The nature of volcanic arc crust

We should note the contrast between the ocean crust and arc crust. The ocean crust, with a few localized exceptions at some slow-spreading ridges (Niu, 1997), is solidified magmas formed at ocean ridges. The observation (Fig. 11a) says that the thicker arc crust is, the lesser of which is of arc magmatic origin as a result of mantle wedge melting. This inverse correlation of the extent of mantle melting with arc crustal thickness contradicts the belief that arc crust is an arc magmatic construct. This contradiction reflects widespread misconceptions because much of the arc crust must be pre-arc crustal basement of long and complex histories already in place at the time of subduction initiation (Niu et al., 2003; Niu, 2016b). One may argue that thicker arc crust may mean longer arc volcanism in one place for tens of millions of years or that the initial arc crust was exceptionally voluminous. However, there is no evidence to support such arguments. For example, there is no evidence that the thicker arc crust (~35 km) above the continental basement of the Japanese arcs records longer volcanic history than the thinner arc crust (~17–20 km) above the “oceanic” basement of Tonga and Mariana arcs with similar timing of subduction initiation at ~52 Ma (Stern, 2002). The young Ryuku arc (<12 Ma) above the continental basement (Chinese continental shelf) has a thick crust (~25 km) with no evidence of exceptionally voluminous magmatism. Hence, this observation and conceptual understanding here are of far-reaching significance for models of subduction initiation (Niu et al., 2003; Niu, 2014) and for reevaluating the widely accepted “island-arc model” for continental crust accretion (see Niu et al., 2019).

5.3. What controls mantle wedge thermal structure?

Turner and Langmuir (2015a, 2015b) implicitly assume that the lithosphere thickness (inferred from crustal thickness, i.e., LAB depth × Moho depth) controls mantle wedge temperature structure (Fig. 11b,d) and ascribe the extent of mantle wedge melting inferred from the VAB compositions (Fig. 11a) as resulting from mantle wedge temperature structure control with temperature variation of up to 250 K (Fig. 11d). This nested interpretation is confusing and is erroneous in both concept and practice. The LAB is likely the amphibole (parasite) dehydration solidus, which is an isotherm of ~1100 °C at all pressures <3 GPa (Niu and Green, 2018). So, the conductive geotherm of the thin lithosphere (dT/dP|THICK ALI) is steeper than that of the thick lithosphere (dT/dP|THICK ALI) (Fig. 11e,f), making the convective asthenosphere pushed down at greater depth beneath thick lithosphere than beneath thin lithosphere. The thickness of the conductive lithosphere is irrelevant to, and thus cannot be used to measure, the temperature of the underlain convective asthenosphere. If we followed the logic in Fig. 11b,d, we would say that the mantle beneath ocean ridges with thin lithosphere must be hotter than the mantle beneath Hawaiian Islands with thick lithosphere, which is likely wrong. The concept of TDP was introduced (McKenzie and Bickle, 1988) to discuss lateral mantle temperature variation of convective asthenosphere at similar depth with dT/dP|adia at, which is irrelevant to dT/dP|Conductive of lithospheric lid.

The logical discussion on mantle wedge temperature must be based on initial depth of melting, which remains debated, but must take place right below active arc volcanoes some distance above the slab surface controlled by the interplay of [1] slab-derived fluids, [2] TMP and [3] effect of the subducting slab (Gill, 1981; Arculus, 1994; Tatsumi and Eggins, 1995; Stern, 2002; Elliott, 2003; England et al., 2004; Syracuse and Abers, 2006; Syracuse et al., 2010; England and Katz, 2010; Grove et al., 2012).

[1] The effect of slab-derived fluids is understood to develop water-saturated wet solidus that triggers mantle wedge melting, which is deep in the asthenosphere and does not know the thickness of the overlying lithospheric lid.

[2] The role of TMP maintained by the convective asthenosphere with dT/dP|adia is necessarily important and is unaffected by the lithospheric lid above with dT/dP|Conductive and the cold subducting slab below.
Fig. 11. [a] Arc-averaged compositions of modern global volcanic arc basalts (VAB) above subduction zones show significant correlations with crustal thickness (adapted from Turner and Langmuir, 2015a, 2015b) as recognized previously using limited dataset (Plank and Langmuir, 1988). The subscript 6.0 refers to values corrected to MgO = 6.0 wt%. In reproducing these plots, I colour-code the data separating arcs built on oceanic (blue) and continent (purple) basement (Clift and Vannucchi, 2004; Syracuse et al., 2010). The authors argue that these correlations are not caused by crustal level processes but reflect mantle melting processes and the extent of mantle wedge melting increases with decreasing crustal thickness. [b,c] Among alternative interpretations, the authors (Turner and Langmuir, 2015a) seem to show their preference for the possible effect of the lithosphere thickness on the extent of mantle wedge melting and VAB chemistry by implicitly assuming the arc crustal thickness in [a] as a proxy for lithosphere thickness without elaboration, yet emphasize in [d] that mantle wedge temperature structure exerts the primary control on the extent of mantle wedge melting, where hot mantle wedge (scenario A) is argued to be associated with thin lithosphere whereas cold mantle wedge (Scenario B) is associated with thick lithosphere with globally between-wedge temperature variation of up to 250 K with respect to the dry solidus. I acknowledge the importance of the observations (a) but stress that the interpretation (d) is problematic (see text). [e,f] illustrates my understanding in P-T space, the lid effect, that is consistent with the observation (a), petrological concepts and geodynamic likelihood. The initial depth of mantle wedge melting takes place right below active arc volcanoes some distance above the slab surface controlled by the interplay of slab-derived fluids, mantle potential temperature and thermal effect of the subducting slab cooling. The mantle wedge melting is best understood as flux-melting, diapir formation, diapiric upwelling and decompression melting. [1] The slab-derived fluids define the wet mantle wedge solidus (also see Fig. 14); [2] initial wet melting facilitates the development of diapirs; [3] buoyancy-driven rise and growth of diapirs; [4] diapiric upwelling and decomposition melting; [5] decomposition melting stops when capped at the LAB, the base of the conductive lithospheric lid. Hence, lithospheric thickness variation, the lid effect, exerts the primary control on the extent of mantle wedge melting and VAB composition (Fig. 11a). We should note that at ocean ridges, plate separation induces passive mantle upwelling and decompression melting, but in the mantle wedge, slab-dehydration induces flex-melting develops into buoyant and growing diapirs that rise and melt by decompression, which is well captured by the lithospheric lid effect. Note that the diapiric decompression melting mantle is likely to have steeper dT/dP [ADIABAT] than shown because the melt has higher heat capacity than the solid phases. Note that for conceptual clarity I choose the same $P_0$ for [e] and [f] but $P_0$ is expected to increase with increasing slab dip $\theta$ and $\sin(\theta)$,
in favor of enhanced decompression melting. The illustration in e and f can be further improved when more data become available. $T_{\text{AL}}$ stands for arc lithosphere thickness. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

The effect of the subducting slab is often described by a slab thermal parameter ($\Phi$), which is the product of plate age [$A$], convergence velocity [$V$] and the sine of the slab dip angle [$\theta$], i.e., $\Phi = A^* V^* \sin(\theta)$ (Syracuse et al., 2010). The greater the $\Phi$ value is, the more cold-slab mass is subducted per unit time and thus serves as a heat sink to cool the mantle wedge, but the physical significance of this parameter is, if any, not obvious (see below).

However, the subducting slab that drives mantle wedge convection that maintains the hot mantle wedge with $dT/dP_{\text{(Adiabat)}}$, otherwise the mantle wedge would freeze. The fast descent of the slab will induce fast lateral hot/warm asthenospheric material supply towards the slab (i.e., the corner flow; England et al., 2004), so [3] cools the mantle wedge while also inducing heat supply to maintain [2]. Thus, the thermal effects of [2] and [3] tend to be cancelling, depending on the depth and extent of slab-wedge decoupling (e.g., Wada and Wang, 2009; Syracuse et al., 2010). Nevertheless, the combined net thermal effect of [2] and [3] could still play some role (England and Katz, 2010), but this is unlikely to be important in controlling the extent of mantle wedge melting because the observations (Fig. 11a) do not record such combined thermal effect.

Turner et al. (2016) show that the estimated extent of melting varies along the Chilean South Volcanic Zone ($\sim 33^\circ$-$34^\circ$S) as a function of the Moho depth with a constant thermal parameter (i.e., $\Phi \approx 0$), which denies the effect of slab cooling, but supports the lid effect on the VAB magmatism. Turner and Langmuir (2015b) argued, however, with confusing inference that the difference between observed $N_{\text{Na,0}}$ and $N_{\text{Si,0}}$ predicted from regression against crustal thickness (Fig. 11a) correlates with a composite parameter $V^* \sin(\theta)$ (where $V$ and $\theta$ are the same as defined above) and interpret this correlation as slab cooling effect on mantle wedge temperature. The slab with smaller dip angle is interpreted to cool mantle wedge more with lower extent of melting. This interpretation contradicts that of Turner et al. (2016) and is physically incorrect because mathematically $V^* \sin(\theta)$ is the same as $\Phi/A$ as defined above, and the key variable here is $\sin(\theta)$, which has nothing to do with cooling (see above), but is proportional to the depth of the slab surface beneath arc volcanoes, and is potentially proportional to depth interval of decomposition melting $F \propto P_{\text{O}}-P_{\text{P}}$, depending on actual $P_{\text{P}}$. With increasing $\theta$ and $\sin(\theta)$, $P_{\text{O}}$ tends to increase and thus $F \propto P_{\text{O}}-P_{\text{P}}$ increases. This is conceptually and physically the same as the lid effect, not thermal effect (a detailed analysis will be presented elsewhere; also see below).

5.4. Diapiric upwelling, decompression melting and lid effect on the extent of melting and VAB compositions

The observations (Fig. 11a) demonstrate that it is the lithosphere thickness variation that controls the extent of mantle wedge melting and the global VAB compositions as illustrated schematically for scenarios of thin ($e$) and thick ($f$) lithosphere with $dT/dP_{\text{THIN AL}} > dT/dP_{\text{THICK AL}}$ and $L_{\text{AT}}(\text{THIN AL}) < L_{\text{AT}}(\text{THICK AL})$. Mantle wedge melting begins at $P_{\text{O}}$ and the melting mantle will rise because of buoyancy, and the rising melting mantle continues to melt progressively more by decompression until capped by the lithospheric lid at the LAB. As expected, VAB erupted on thick lithosphere have the signature of low extent of melting (Fig. 11a) because of shorter decompression melting interval [3] whereas VAB erupted on thin lithosphere have the signature of high extent of melting (Fig. 11a) because of taller decompression melting interval [3]. We should note that at ocean ridges, decompression melting results from plate separation induced passive mantle upwelling and decompression melting, but there is no obvious mechanism for mantle wedge upwelling. Fig. 11e,f shows schematically that the upwelling results from initial flux-melting that develops into growing and buoyant diapirs (e.g., Green and Ringwood, 1967; Wyllie, 1971) driven by the Stakes-law for continued decompression melting.

5.5. VAB preserves the garnet signature

We note from Fig. 11a that the VAB preserve the garnet signature whose intensity (high Dy/Yb and low Sc) decreases with increasing extent of melting from beneath thick lithosphere (less dilution) to beneath thin lithosphere (more dilution). One may interpret the varying Sc and Dy/Yb as the signature clinopyroxene, but this is a clear garnet signature because $K_{\text{Dcpx}}(\text{Di}) < K_{\text{Dcpx}}(\text{Gn}) \approx 3.5$ and $K_{\text{Dcpx}}(\text{Di}) \approx 3 < K_{\text{Dcpx}}(\text{Gn}) \approx 4$ (see Niu et al., 1996), indicating that clinopyroxene will not cause Dy/Yb variation, but garnet does. This recognition offers important insights: [1] mantle wedge melting does start in the garnet peridotite stability field with garnet as a residual phase, and the garnet signature is progressively more diluted with increasing extent of diapiric decompression melting in the spinel peridotite stability field beneath thin lithosphere as seen in OIB (Fig. 10); [2] it is also probable that eclogite in the subducting ocean crust may participate in and contribute to mantle wedge melting most likely at the onset of melting at $P_{\text{O}}$, which requires further investigation.

6. The lid-effect on mantle melting beneath continents and CIB

Cenozoic basalt volcanism in continental interiors occurs where lithosphere is thin or recently thinned caused by water introduction through seafloor subduction currently or in no distant past such as eastern China, eastern Australia, western Mediterranean and western USA (Niu, 2014). Eastern China is a type region for studying intracontinental mantle melting because it has a history of widespread lithosphere thinning in the Mesozoic by means of basal hydration weakening associated with Pacific-Pacific plate subduction (see Niu, 2005b, 2014; Niu et al., 2015), which developed oceanic type seismic low velocity zone (LVZ) beneath the region (Niu, 2014) that is well maintained in the Cenozoic by dehydration of the present-day Pacific plate stagnant in the mantle transition zone (Karason and van der Hilst, 2000; Niu, 2014). The widespread Cenozoic basalt volcanism in eastern China is the consequence of this LVZ (Figs. 12-13).

Fig. 12 summarizes the topography of continental China and the adjacent regions and the interpreted basaltic magmatism in eastern China in the context of plate tectonics. The great gradient line (GGI; white dashed line) marks the contrast in elevation, gravity anomaly, crustal thickness and mantle seismic velocity from high plateaus in the west to hilly low plains of eastern China, which manifests varying
(c) Origin of Cenozoic basalts in eastern continental China (see [a])

GGL: Steep gradient in elevation, morphology, crustal thickness, gravity anomaly and heat flow - all resulting from sudden lithosphere thickness change - the ISOSTATIC EFFECT

**West China Plateau**

**East China lowland/plain**

OIB-like Cenozoic basalts highly enriched in incompatible elements, but isotopically depleted with $\varepsilon_{Nd} > 0$

Two causes related to western Pacific subduction:

1. Trench retreat induced eastward continental drift and localized extension;
2. mantle wedge suction and asthenospheric supply

Region of asthenospheric flow with decompression component, and thus decompression melting for Cenozoic volcanism

Water and hydrous melt resulting from dehydration of the stagnant Pacific slab lying horizontally in the 410D-660D mantle transition zone

(capTION on next page)
lithosphere thickness from $\geq 150$ km beneath the plateaus in the west to $\leq 80$ km beneath eastern China (Niu, 2005b; Niu, 2014), which is also responsible for the widespread alkali basalt magmatism in eastern China (Niu, 2005a, 2014; Niu et al., 2015). (c) Cartoon interpreting that the lithosphere is thick beneath the high plateaus in western China, but thin in eastern China across the sharp GGL. Western Pacific subduction induced corner-flow requires asthenospheric material replenishment from the west (i.e., remote mantle wedge suction). In response, the eastward asthenospheric flow experiences decompression melting, which explains the widespread Cenozoic alkali basalt volcanism in eastern China (e.g., Sun et al., 2020b). The western Pacific trench retreat causes eastward continental drift of Eurasia and continental extension in eastern China, which can also facilitate localized decompression melting for the Cenozoic alkali basalt volcanism in eastern China (Niu, 2014). We note that the transition-zone slab released water in the form of incipient hydrous melt percolates upwards and metamorphizes the upper mantle, contributing to the source of the alkali basalts widespread in eastern China (Niu, 2005b; Guo et al., 2014; Sun et al., 2017). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

7. The lid-effect on mantle melting for large igneous provinces (LIPs) on land and in ocean basins

Many intra-plate basaltic volcanoes away from plate boundaries have been interpreted as resulting from decompression melting of dynamically upwelling hot mantle plumes originated from the hot thermal boundary layer in the D* region at the core-mantle boundary (e.g., Morgan, 1971; Campbell and Griffiths, 1990; Griffiths and Campbell, 1990; Duncan and Richards, 1991; Sleep, 1990, 1992; Coffin and Eldholm, 1994; Davies, 1999). But such lower-to-upper mantle material transfer cannot happen without a plume head with sufficient buoyancy (Hill et al., 1992; Campbell and Griffiths, 1990; Duncan and Richards, 1991; Davies and Richards, 1992; Davies, 2005; Niu et al., 2017). Decompression melting of plume heads produces large igneous provinces (LIPs) with large basalt volumes emplaced in short time periods, forming ocean plateaus in ocean basins (e.g., Ontong Java Plateau in the Pacific and Kerguelen Plateau in the Indian Ocean) and continental flood basalts provide (e.g., Siberian Trap, Deccan Trap, Columbia River Basalt) (Fig. 14a,b). Following the emplacement of LIPs are long-lived volcanic activities that may produce volcanic chains (or hotspot tracks) such as age-progression seamount chains in ocean basins if the plates move fast relative to the volcanic centers. The most significant in this context concerns the question whether mantle plume heads can cause continental breakup, especially breakup and dispersal of supercontinents because there has been abundant literature in recent decades that correlates continental breakup in time and space with mantle plumes (Richards et al., 1989; Hill, 1991; Hill et al., 1992; Storey, 1995; Li et al., 1999, 2008; Condie, 2004; Zhong et al., 2007; Buiter and Torsvik, 2014; Zhang et al., 2018). I have recently demonstrated that continental breakup is a straightforward consequence of plate tectonics without requiring mantle plumes, and mantle plumes, if needed, may be of help at early rifting stage, but cannot lead to complete breakup, let alone to drive long distance dispersal of broken continents (Niu, 2020a). From the perspective of scientific developments, I predict that many advocates will continue to disagree on my objective, logical and rigorous analysis. So, it is necessary and particularly pertinent here to discuss the issue in the context of the lid effect on mantle melting and global basin magmatism.

Fig. 14c shows in P-T space the mantle solidus (McKenzie and Bickle, 1988) and adiabat for two scenarios with $T_{\text{MP}} = 1600 °C$ for mantle plumes of dynamic upwelling and $T_{\text{MP}} = 1350 °C$ for passive upwelling beneath ocean ridges for comparison. To illustrate the concept, we can choose $T_{\text{MP}} = 1600 °C$ for the plume scenario, which is likely hotter than widely assumed. $P_{20}$ is the depth of rising mantle plume that intersects the solidus and begins to melt. For comparison, we also show the depth of $P_{20}$ for cooler mantle with $T_{\text{MP}} = 1550 °C$ and $1500 °C$. Fig. 14d shows three scenarios of varying thickness of continental lithosphere when impacted by a rising mantle plume head (after Niu, 2020a). The key concepts and logical reasoning are given below:

1. The adiabatically rising mantle plume head begins to melt when intersecting the solidus at $P_{20} \approx 140$ km.

2. The rising mantle plume head continues to melt until capped by the lithosphere at $P_{20} \approx 120$ km, which is the LAB equivalent to average continental lithosphere thickness.

3. Melting cannot happen in the lithosphere as it is under solidus conditions (except for volumetrically small metasomatic veins or veins of lower solidus temperature).

4. A spherical plume head 1000 km across ($R = 500$ km) that flattens to a disk 2000 km across ($r = 1000$ km) when reaching the lithosphere (see Fig. 14a,b; Campbell, 2007) will have thickness of $\approx 167$ km.

5. To be conservative, we assume the flattened lithosphere to be about half of the thickness $\sim 84$ km.

6. Whether melting actually occurs or not and if so, to what extent, strictly depends on the lithospheric thickness (see above; Niu et al., 2011) as quantified (Watson and McKenzie, 1991; White and McKenzie,
[a] Compositional systematics of Cenozoic basalts across the GGL along the traverse A in Fig. 12 [a], showing the lid effect

Increasing topographic elevation
Increasing lithosphere thickness (from ~ 80 km to ~ 120 km)
Increasing depth (pressure) of melt extraction
Decreasing extent of melting

[b] Compositional systematics of Cenozoic basalts as a function of final equilibration pressures recorded in the contained clinopyroxene megacrysts, showing the lid effect (locations indicated as B in Fig. 12 [a])

Increasing final depth of melt equilibration at the LAB
i.e., increasing depth (pressure) of melt extraction
Increasing garnet signature
Decreasing extent of melting

(caption on next page)
stretch of ~250 km from southeast to northwest across the GGL, the elevation increases from ~600 m to ~1400 m above sea level, corresponding to lithosphere thickness (the LAB depth) increase from ~80 km to ~120 km (Gao et al., 2020). The systematic variation of the petrological parameters as a function of relative distance from southeast to northwest is a simple manifestation of the lid effect, i.e., increasing pressure/depth of melt extraction (decreasing SiO₂ and increasing FeO, MgO, Sm/Yb, and decreasing extent of melting (increasing Ti₂O, P₂O₅, La/Sm, and Sm/Yb)) from beneath thin lithosphere to beneath thick lithosphere. The compositional systematics of the 10 suites of Cenozoic alkali basalts containing clinopyroxene megacrysts from eastern China with a north-south spatial coverage in excess of 2500 km (localities labeled as B in Fig. 12a), plotted as a function of crystallization pressure of clinopyroxene megacrysts, which is best understood as crystalized from a “stable magma reservoir” close beneath the LAB (Sun et al., 2020b). These systematics are manifestations of the lid effect, i.e., with increasing pressure of clinopyroxene megacryst crystallization (i.e., the LAB depth), the compositions of the melt indicate increasing pressure of melt extraction (decreasing SiO₂ and Al₂O₃ and increasing FeO, MgO, and Sm/Yb) and decreasing extent of melting (increasing Ti₂O, P₂O₅, Sm/Yb). The strength of the garnet signature, as predicted, increases with increasing pressure (increasing Sm/Yb and decreasing Sc) because Sc and Yb are highly compatible in garnet. The definition of the parameters is the same as in Figs. 8-10).

8. Summary

The idea that mantle (potential) temperature (Tₚ) variation controls the extent of mantle melting and basaltic composition (i.e., “temperature control”) developed in 1980s has since become the paradigm on basalt petrogenesis. My research over the past 30 years on the subject demonstrates repeatedly that this paradigm is unsupported by observations and needs change. With the principle that large scale Earth processes are likely very simple, but the key skill to discover the simplicity is to correctly identify the primary variables that control the processes (Niu, 2020b), I devoted effort on global observations to seek key variables controlling mantle melting, which leads to the conclusion that lithosphere thickness variation (i.e., the lid effect) exerts the primary control on the extent of mantle melting, depth/pressure of melt extraction and basalt compositions in all tectonic settings on Earth: mid-ocean ridge basalts (MORB), intra-plate ocean island basalts (OIB), volcanic arc basalts above subduction zones (VAB) and basalts in continental interiors (CIB). Below are condensed summary of rigorous analyses, key observations, and future work.

1. The common perception that solid rock melting needs heating gives the impression of temperature control and thus the popular acceptance of the paradigm. But the understood mantle melting mechanisms (Fig. 3) do not require excess heat or temperature but require conditions that place the asthenospheric mantle onto or above the solidus (dry or wet solidus) in P-T space although the initial depth of melting (P₀) may vary.

2. The extent of mantle melting (F) is not controlled by P₀ but controlled by decomposition interval P₀-Pₚ. That is, the final depth of decomposition melting Pₚ at the LAB is important as it is the base of the conductive thermal boundary layer (CTBL) that varies between tectonic settings and can also vary vastly on all scales for a given tectonic setting, depending on the geological history.

3. The effect of P₀ on the extent of melting F ∝ P₀-Pₚ can be readily understood by comparing the scenarios of Iceland and Hawaii that are considered as two type hot mantle plumes with similarly hot and deep P₀, but hugely different Pₚ (Iceland) ≈ 90 km > Pₚ (Hawaii) ≤ 20 km. If P₀ ≈ 100 km, then the decomposition melting interval would be P₀-Pₚ (Iceland) ≈ 80 km > P₀-Pₚ (Hawaii) ≥ 10 km. Hence, the extent of melting beneath the thin lithosphere in Iceland can be up to 8 times greater than that beneath the thick lithosphere in Hawaii (linear scaling is assumed here for conceptual clarity), which is consistent with their
petrological and geochemical contrast (see discussion given in Niu et al., 2011). Therefore, lithosphere thickness variation, i.e., the lithospheric lid effect, is predicted to be the primary variable that controls the extent of mantle melting, pressure/depth of melt extraction at the LAB, and basalt compositions.

(4) Basalt compositions are known to vary as a function of fertile mantle composition, extent and pressure of melting and crustal level magma evolution. Once the effects of crustal level processes can be corrected for (largely removed), the data can be used to discuss mantle sources and processes. The significantly correlated

Fig. 14. (a) The idea of mantle plume initiation at the core-mantle boundary (CMB), its rise and growth into a spherical plume head, the flattening and impact of the plume head upon reaching the lithosphere and beginning to melt by decomposition (Campbell and Griffiths, 1990; Campbell, 2005; figure after Saunders et al., 1992). (b) An advanced scenario presented and described by Campbell (2007) as follows. A plume head of ~1000 km diameter rises beneath continental crust, flattens and melts by decomposition to form a flood basalt province. Arrival of the plume head also leads to uplift, which places the lithosphere under tension, as shown by the arrows. The final diameter of the flattened plume head is claimed to reach 2000–2500 km. Tension introduced by the plume head can lead to run-away extension and the formation of a new ocean basin, drawing the hot plume head into the spreading center leading to the formation of thickened oceanic crust represented by the seaward dipping (seismic) reflectors (SWDRs) as seen on both sides of the North Atlantic genetically associated with the Iceland plume (White and McKenzie, 1989; Saunders et al., 1998; Larsen et al., 1999) and many other passive continental margins (Storey et al., 1992; Coffin and Eldholm, 1994; Ernst, 2014). (c) Modified after Niu (2020a) and Niu et al. (2015) (with the wet solidus and dehydration solidus adapted from Green et al. (2010) and Green (2015), showing in P-T space the mantle solidus (McKenzie and Bickle, 1988) and adiabat for two scenarios with $T_{MP} = 1600°C$ for mantle plumes of dynamic upwelling and $T_{MP} = 1350°C$ appropriate for passive upwelling beneath ocean ridges for comparison. To illustrate the concept, I choose $T_p = 1600°C$ for the plume scenario, which is probably hotter than widely assumed. In the lower T and P portion (upper left) are wet solidus and dehydration solidus relevant to mantle wedge melting (Fig. 11) and lithosphere-asthenosphere boundary (LAB) phase equilibria beneath ocean basins and thinned continental lithosphere with a seismic low velocity (LVZ) (Figs. 10, 12,13). (d) Showing three scenarios of varying thickness of continental lithosphere when impacted by a rising mantle plume head (after Niu, 2020a). The inevitable conclusions are: [1] mantle plumes cannot melt by decomposition to produce LIPs beneath thickened cratonic lithospheric lid; [2] LIPs in the geological record indicate thin lithosphere at the time of volcanism; [3] it is the size, thickness and strength of the continental lithosphere that determines whether a mantle plume can surface and whether a mantle plume can break up the continents, not the other way around; [4] if there are/were many more mantle plumes and plume heads beneath continents at present and probably also in Earth’s history, only those arriving beneath thin or thinned lithosphere could surface through basaltic magmatism. Therefore, arrival of mantle plume heads beneath stable continents will not thin, weaken and break lithosphere, but is predicted to thicken the lithosphere and may thus be an important mechanism to cause craton stabilization (Niu, 2020a). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
It is important to note that different from OIB and CIB, the lid effect with OIB erupted on thick lithosphere having the petrological signature of low extent (low F, P = deep LAB) of melting and high pressure (high P = deep LAB) of melt extraction, whereas the opposite is true for OIB erupted on thin lithosphere (Fig. 10b). The same is true for CIB as demonstrated in Figs. 12-13. We should note that the compositional scatter about the systematic trends must be combined effect of mantle source compositional variation, initial depth (P0) of melting due to TMP variation or source fertility variation, and errors associated with fractionation correction. We do not at all ignore all these factors, but they are secondary and insignificant because they are overshadowed by the lid effect.

The observations and analysis in (4) inform us explicitly that OIB MORB compositional systematics as a function of ridge spreading (Fig. 8a) is a straightforward consequence of the lid effect. While slab-dehydration induced mantle wedge melting or flux melting is the understood major mechanism of magma generation above subduction zones, the global VAB compositional systematics reported by Turner and Langmuir (2015a, 2015b) is a straightforward consequence of the lid effect. Although these authors presented the correlations with arc crustal thickness (i.e., Moho depth vs. LAB), it is predicted that Moho depth ∝ LAB depth and more effort is needed to obtain high quality seismic data to resolve LAB depths beneath global volcanic arcs to further quantify the Moho ∝ LAB depth relationship. Note that Turner and Langmuir (2015a, 2015b) incorrectly interpret the lid effect as mantle wedge temperature structure control by advocating hot mantle beneath thin arc lithosphere and cool mantle beneath thick arc lithosphere, which is the same as arguing for hot mantle beneath ocean ridges with thin lithosphere and cool mantle beneath ocean islands like Hawaii with thick lithosphere.

The observations that erupted basalts record P0 (i.e., the LAB depth), not P0, in all settings (MORB, OIB, VAB and CIB) reiterate the fundamental understanding of effective melt-solid equilibration in the melting mantle (Fig. 5; Niu, 1997, 2016a). This smoking-gun evidence that erupted basalts have no memory of initial depth of melting at least in terms of olivine-making elements Si, Mg and Fe. It follows in simple clarity that basalt-olivine-based thermobarometers provide no information on P0 and TMP. In other words, the calculated P0 and TMP using such thermobarometers have no significance unless the lid effect is properly corrected for if possible (Niu et al., 2011). This new understanding requires thorough reevaluation of many discussions on P0 and TMP in the literature.

Mantle plumes or plume heads, no matter how big and how hot, will not melt beneath thickened cratonic lithosphere lid, which is below the solids, will not surface and will not cause continental breakup. If anything, the arrival of mantle plumes and plume heads will contribute to new accretion of the lithosphere thickening, and importantly may facilitate stabilization of cratons, rather than causing continental breakup as popularly believed (Fig. 14; see Niu, 2020a, 2020b). Therefore, LIPs as a result of mantle plume head decompression melting must indicate the thin or thinned lithosphere at the time of LIP volcanism.

Following all the above, we can add here that komatiite as a result of very high extent of mantle melting (high F = P0 - P0) requires not only deep initial melting (high P0) but also shallow melting cessation (low P0) under thin or very thin lithospheric lid although Archean komatiites are often preserved in association with cratonic shields of thick lithosphere. This inference offers an additional perspective on understanding the petrogenesis of yet mysterious komatiites (see McKenzie, 2020).

It may be too big wording to say paradigm shift, but the “temperature control” paradigm on basalt petrogenesis that is inconsistent with all the observations needs change in order to promote scientific progress forward. As the change from “temperature control” to “lid effect” is fundamental and may be unaccustomed to many, it will take time to convince the community through further debate. I endeavor to actively participate in this debate through this publication and future communications. To facilitate such debate, I offer the following three statements.

1) Global MORB (Figs. 8, 9), OIB (Fig. 10), VAB (Figs. 11) and CIB (Fig. 13) compositions all show the lid effect (i.e., the Pf control), but do not show the effect of “temperature control” (i.e., P0 or TMP). The latter may not be important at all in reality or, if any, must have been obliterated because of effective and efficient melt-solid equilibration in the melting mantle. [2] Objectiveness and open-mindedness (vs. “Confirmation bias”) are requisite twins for insights and discoveries.
Declaration of Competing Interest

The author declares no conflict of interest nor competing financial interests.

Acknowledgments

I thank Professor Gillian Foulger for invitation, Professor Yan Yang for editorial handling, and two journal reviewers for their detailed constructive comments that have helped improve the clarity of the paper. One reviewer was politely critical of my self-citations, but the latter reflects my 30-year dedication on the “basalt problem”, for which I am obligated to write this review. This work is supported by grants from National Natural Science Foundation of China (NSFC 91958215, 41630968), NSFC-Shandong Joint Fund for Marine Science Research Centers (U1606401) and 111 Project (B18046).

References


Arculus, R.J., 1981. Island arc magmatism in relation to the evolution of the crust and mantle. Tectonophysics 73, 123–133.


J. Geol. 99, 767–775.

J. Geophys. Res. 97, 6970.


J. Geophys. Res. 98, 7887.

J. Geol. 43, 783–786.

