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Generation and Evolution of Basaltic Magmas: Some Basic Concepts and a New View on the Origin of Mesozoic-Cenozoic Basaltic Volcanism in Eastern China

(玄武岩浆起源和演化的一些基本概念以及对中国
东部中-新生代基性火山岩成因的新思路)

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Abstract. Some basic concepts of basaltic magma generation and evolution are discussed in the context of global tectonics. These concepts need better understanding before invoking elusive possibilities in igneous petrogenesis on all scales and in all tectonic environments. A hypothesis for the Mesozoic lithosphere thinning and Mesozoic/Cenozoic basaltic volcanism in eastern China is presented. This hypothesis is consistent with observations and complies with basic physics. While the eastern China volcanism can be defined as “intra-plate” volcanism, it is in fact a special consequence of plate tectonics. The Mesozoic lithosphere thinning in eastern China is best explained by a process that “transformed” the deep portion of the lithosphere into convective asthenosphere by hydration. The water that did so may come from dehydration of subducted Pacific (or predecessor) oceanic lithosphere that is presently lying horizontally in the transition zone beneath eastern Chinese continent as detected by seismic tomographic models. The Mesozoic volcanism may be genetically associated with the lithospheric thinning because the basaltic source is ancient isotopically enriched ($\epsilon_{Nd} < 0$) lithosphere - being converted to the asthenosphere. The NNE-SSW Great Gradient Line (GGL) marked by the sharp altitude, gravity anomaly, crustal thickness, and mantle seismic velocity changes from high plateaus in the west to hilly plains of eastern China is an expression of variation in lithospheric thickness from probably $> 150\text{-}200$ km thick beneath the plateaus in the west to probably < 80 km thick beneath eastern China. The “remote” western Pacific subduction systems induce asthenospheric flow from beneath eastern China towards the subduction zones (i.e., “wedge suction”), which, in turn, requires asthenospheric material replenishment from beneath the plateaus to eastern China. As a result, such eastward asthenospheric flow experiences upwelling and decompression (from beneath thickened to thinned lithosphere), which causes the flowing asthenosphere (e.g., $\epsilon_{Nd} > 0$) to partially melt and produce the Cenozoic eastern China basaltic volcanism. Such volcanism may have actually begun at the end of the Mesozoic lithosphere thinning in the late Cretaceous. This simple concept is currently being substantiated with detailed petrologic and geochemical data.

Keywords: Mantle peridotite, partial melting, magma generation and evolution, eastern China, lithosphere thinning, Mesozoic and Cenozoic volcanism, a special case of plate tectonics

摘要: 本文以全球大地构造为背景讨论了玄武岩浆起源和演化的一些基本概念。这些概念的正确理解有助于合理解释各种环境中火成岩的形成机制, 同时也有助于依靠野外岩石组合来判别古构造环境。在此基础上并结合已有资料及观察对中国东部中生代岩石圈减薄, 中-新生代基性火山岩成因提出了一些新解释。这些解释与地质观察相吻合, 且符合基本的物理学原理。虽然中国东部基性火山活动可称为“板内”火山活动, 但它实际上是板块构造的特殊产物。中国东部中生代岩石圈减薄是其下部被改造为软流层的缘故。这种改造是加水“软化”所致。该水来源于中国东部地幔过渡带(410-660km) 内古太平洋(或其前身) 俯冲板块脱水作用。其过程将岩石圈底部改造为软流层实际上就是岩石圈减薄的过程。这是因为软流层是地幔对流的重要部分, 而大陆岩石圈则不直接参与地幔对流。中生代玄武岩具有 $\epsilon_{Nd} < 0$ 的特征, 说明其源于新近改造而成的软流层, 亦即原古老岩石圈之底部。中国大陆北北东-南南西向的海拔梯度突变界线与东-西部重力异常, 陆壳厚度变化, 以及地幔地震波速变化梯度吻合。为方便起见, 该北北东-南南西向梯度线可被称为“东-西梯度界”。该界东-西海拔高度差异(西部高原与东部丘陵平原), 陆壳厚度差异(西部厚而东部薄), 和 100-150km 深度范围地幔地震波速差异(西部快而东部慢) 均受控与上地幔重力均衡原理。这表明西部高原岩石圈厚度 $>150\text{-}200\text{km}$ 而东部丘陵平原岩石圈厚度 $<80\text{km}$ 。“遥远”的西太平洋俯冲带具有自然的地幔楔吸引作用。该吸引作用会引起中国东部“新生”软流层东流。该软流层东流必然要引起西部高原底部软流层的东向补给(流动)。这一过程必然会导致东移软流层的减压, 亦即从西部的深源(深于 $>150\text{-}200\text{km}$ 岩石圈) 到东部的浅源(深于 $\sim 80\text{km}$ 岩石圈)。该东移软流层的减压分熔可合理解释具有软流圈地球化学特征($\epsilon_{Nd} > 0$) 的新生代中国东部基性火山活动及玄武岩。这些对中国东部中-新生代地质过程的解释也许甚为简单, 但这些解释与观察吻合, 并符合基本的物理学原理。更为仔细的, 以岩石学和地球化学为主的讨论正在准备之中。

关键词: 地幔杆栏岩, 地幔分熔, 玄武岩浆的成因与演化, 中国东部, 岩石圈减薄, 中-新生代基性火山作用, 板块构造的特例。

1. Introduction

The advent of plate tectonics theory ~ 35 years ago has revolutionized Earth Science thinking, and provided a solid framework for understanding how the Earth works. For example, plate tectonics theory elegantly explains the occurrences and distributions earthquakes and voluminous volcanic eruptions along plate boundaries: (1) at seafloor spreading centers where two plates pull apart, the ocean crust is being continuously created by volcanism; (2) at convergent boundaries where the oceanic plate returns into the Earth's deep interior through subduction zones, volcanic arcs such as the "Pacific ring fires" are being built. However, plate tectonics theory, by its original definition, cannot explain earthquakes and volcanic activities occurring within plate interiors. Hotspots or deep-rooted mantle plumes have been widely invoked to be responsible for "intra-plate" volcanism. Intraplate volcanism is indeed widespread and is thus an important mode of mantle melting. Many of the intraplate volcanic activities are apparently associated with widely perceived mantle plumes/hotspots such as the Hawaii, Samoa, Tahiti volcanic islands, but such association is not clear in many other cases. For example, Cenozoic volcanic activities widespread in eastern Australia [see Johnson *ed.*, 1989], eastern China [see Deng et al., 1998; Zhang et al., 1998; Liu, 1999], western and central Europe [see Wilson & Patterson, 2001], the well-known Cameroon volcanic line straddling the Atlantic-African passive continental margin [see Fitton & Dunlop, 1985; Halliday et al., 1988] and numerous seamounts scattered throughout much of the Earth's ocean floor [Batiza, 1982] away from plate boundaries cannot be readily explained by either plate tectonics theory or mantle plume hypothesis. Mantle source "wet spots" [Green & Falloon, 1998] may explain some of the "intraplate" melting "anomalies", but such mechanism alone cannot account for the aforementioned widespread and large scale volcanic activities.

Therefore, much effort remains necessary to study the generation and evolution of basaltic magmas and origin of basaltic rocks. The significance of such study is several-fold: (1) Magma generation is a consequence of Earth's thermal evolution, and is arguably the most effective process leading to chemical differentiation of the Earth over Earth's history from a compositionally more or less uniform proto-earth (chondritic?) to the present-day layered Earth (i.e., the metallic core, and the silicate mantle and crust) with each layer being compositionally heterogeneous on all scales. (2) Magma evolution in the course of cooling and also through interactions with wall-rocks at shallow levels is the very process leading to formation of igneous rocks of varying composition, mineralogy, textures, and thus different rock series and rock types, recognizing compositionally different primitive magmas, as a result of different fertile source rock compositions and different melting conditions, also playing a governing role in igneous rock diversity. (3) Igneous petrogenesis-associated mineralization is the by-product of magma evolution/assimilation. Hence, understanding the basic concepts of magma generation and evolution is essential for a geologist. It is worth to emphasize that modern analytical techniques have been instrumental in our understanding many aspects of igneous rock petrogenesis that were not possible previously, yet the new technologies cannot replace basic mineralogy, petrography, petrology and in particular field observations. Specifically, the validity of geochemical data (for example, trace elements and isotopes) interpretations become questionable without having a clear understanding of the fundamentals of magma genesis and first-hand field and petrographic observations.

In this invited contribution, I review some **very well** understood, yet **not so well** perceived, concepts for basaltic magma generation and evolution. I emphasize why asthenospheric mantle flow is important in the generation of basaltic magmas. While I focus on basaltic magmas, the origin of some granitic magmas will be mentioned in relevant contexts. Finally, I present some new views that may explain the Mesozoic lithospheric thinning and related Mesozoic/Cenozoic "intraplate" volcanism in eastern China. By no means do I claim these views to be correct, but they are testable hypotheses, by which I mean to **cast a brick to attract jade**. I also hope that this paper may provide some food for thought for our younger generations of research students in developing testable hypotheses.

2. The concepts of magma generation

2.1. Partial vs. total melting

Basaltic magmas are derived from more magnesian picritic melts produced by partial melting of mantle peridotites in the upper mantle. For convenience in the following discussion, basaltic magmas/melts also include magnesian picritic melts. The present-day thermal status of the Earth's mantle does not allow total melting. For reasons of physical melt extraction and heat budget, total melting is indeed neither likely in the mantle nor in the crust. As a result, magma generation is the result of partial, not total, melting regardless of the nature of source rocks or magma types. Partial melting has two products, the melt and the residue. In

simple words, the melt represents the easily melted component of the source rock, whereas the residue is the more refractory component of the source rock. The bulk composition of the resultant melt can only be more felsic (i.e., higher SiO₂) or less mafic (i.e., lower FeO and MgO, and lower MgO/FeO ratio) than the source rock. In other words, basaltic melts result from partial melting of mantle peridotites; partial melting of basaltic rocks (basalts, diabases, gabbros etc.) cannot produce basaltic melts, but more silicic melts (trochjemitic, tonalitic, andesitic and granitic melts, for example). By contrast, the melting residues become less felsic and more mafic (higher MgO/FeO) and refractory than the source rock. In summary, partial melting is the process that “transforms” a single and perhaps uniform rock (source) into two compositionally very different rocks: (1) the melt (or magma), which, upon cooling/assimilation, evolves to igneous rocks of various mineral assemblages, textures and bulk compositions, and which is enriched in the so-called incompatible elements while depleted in the so-called compatible elements. (2) the refractory melting residue, which is depleted in incompatible elements and enriched in compatible elements. Because magma generation is a consequence of Earth’s thermal evolution, and takes place today and throughout Earth’s history on a global scale, we can state that magma generation is the most effective process leading to chemical differentiation of the Earth.

2.2. Mantle source materials for basaltic magmas

Source rocks for basaltic magmas are mantle peridotites dominated by olivine, plus orthopyroxene (opx), clinopyroxene (cpx) and one Al₂O₃-rich mineral. The latter mineral, depending on pressure (depth), can be plagioclase (plag, <~ 30 km), spinel (~30-80 km) or garnet (> ~ 80 km). Mantle peridotites are also subdivided into lherzolite (> 10 vol. % cpx) or harzburgites (< 10 vol. % cpx). In practice, lherzolite is the primary source for most basaltic magmas whereas, under most circumstances, harzburgites are melting residues. Pyroxenite, as dikes or veins dispersed in mantle peridotites, may contribute to basaltic magmas, but is volumetrically insignificant. Minerals such as amphiboles and phlogopite (also minor sulfides, oxides etc.) may be present in mantle peridotites, contributing to geochemically enriched basaltic rocks. Note that these rocks are common in both asthenosphere and lithosphere. By definition, the lithosphere approximates the plate, the cold thermal boundary layer. Hence, the lithospheric mantle is *not* the source of most basaltic magmas – too cold to melt unless the lithospheric mantle may be disturbed under some **special** circumstances (see below). Asthenosphere is the very source of most mantle derived basaltic magmas. The mantle melts so derived vary in compositions depending on detailed source mineralogy and depth of melting. If the mantle is sufficiently hot, like in the Archean, melting begins at great depths (e.g., > 120 km in the garnet lherzolite stability field), high MgO komatiite melt may be produced (Fig. 1). If melting occurs in a depth range of ~ 80 to 110 km, high MgO picrite melts can be produced (Fig. 1). If the mantle is cooler and begins to melt at shallower depths, alkali olivine basaltic melts or olivine tholeiite melts may be produced (Fig. 1). Note that one may find abundant literature stating basaltic magmas formed in the plagioclase lherzolite stability field. This is possible experimentally (Fig. 1), but is conceptually misleading in practice because this depth range is within the so-called “cold” thermal boundary layer beneath ocean ridges and elsewhere within the “cold” lithosphere. Therefore, rarely basaltic magmas may be generated in the plagioclase stability field [Niu, 2004].

Having understood the above, we emphasize that lithospheric mantle does contribute to magma generation in certain situations even though it is not the major source of basaltic magmas. Growth or thickening of the lithospheric mantle is achieved through basal accretion (see below) due to conductive heat loss, i.e., by transforming topmost asthenospheric mantle material into lithospheric material [see Niu & O’Hara, 2003]. This is straightforward for oceanic lithosphere accretion, but the same is true to a first order for subcontinental lithosphere even though the latter is in general older and has much more complex histories. The process of converting the topmost asthenospheric mantle to the base of the growing lithospheric mantle is always accompanied by “mantle metasomatism” [Niu & O’Hara, 2003]. In other words, the deep portions of the lithosphere, in particular the more ancient subcontinental lithosphere, contains metasomatized dikes or veins enriched in volatiles (e.g., CO₂, H₂O etc.), metasomatic phases (e.g., amphiboles, phlogopite etc.) and incompatible elements. These enriched metasomatic materials have low melting temperatures and are ready to melt if physical conditions permit. Reactivation of deep-rooted lithospheric faults or fault zones associated with accreted orogenic collision zones will produce volumetrically small, yet highly enriched magmas. Many of those highly enriched magmas in Tibet and the peripheral regions of the Tibetan plateau [see Yu & Zhang, 1998; Yu et al., 2004] may be genetically associated with such metasomatized lithospheric materials and processes. Highly enriched magmas during

early stages of continental rifting are also sourced from metasomatized lithospheric mantle [Barberi et al., 1982]. Highly enriched basaltic magmas found on many ocean islands likely also result from assimilation with metasomatized oceanic lithosphere [Niu & O'Hara, 2003]. In the context of mantle plume-lithosphere interactions, the lithospheric mantle may also participate in the melting processes as a result of "thermal erosion" or assimilation if the heat from the plume is adequate [O'Hara, 1998]. The important message here is that lithosphere is unimportant for the generation of most basaltic magmas, but lithospheric materials do participate in magma genesis, in particular, portions of ancient subcontinental lithosphere with abundant metasomatic assemblages. The latter will result in alkalic mafic melts highly enriched in volatiles and all sorts of incompatible elements as well as isotopes (e.g., radiogenic Sr and unradiogenic Nd).

2.3. Causes of partial melting – magma generation

Theoretically, there are four possible mechanisms through which a solid rock can partially melt to form magmas: (1) heating; (2) decompression; (3) volatile addition; and (4) compression. Heating is the most readily perceived process to cause melting of a rock, yet it is practically unimportant for basaltic magma generation, but important for granitic magma formation when hot, mantle-derived basaltic melts rise and intrude continental crust. Raising pressure (i.e., compression) is a thermodynamically sound possibility, but its actual role in magma generation is neither well documented nor better realized. In terms of partial melting of mantle peridotites to produce basaltic magmas, decompression [Carmichael et al., 1974; Yoder, 1975] and volatile addition are the two most important processes. We focus here on the concept, and will provide examples in a later section.

The first three mechanisms can be readily understood in P-T space as illustrated in Figure 2a. A piece of rock at position *A* beneath the solidus (subsolidus condition) remains solid regardless of its exact position. If this piece of rock is placed above the liquidus (super-liquidus condition) at position *C*, this rock will melt completely or undergo total melting. If this rock is placed between the solidus and liquidus such as at position *B*, only "part" of the rock will melt, hence the concept of **partial melting**. Theoretically, the amount of melt produced or the extent of melting (mass fraction of the melt over the mass of the source rock prior to melting) depends on the position of *B*, varying from 0% to 100% as illustrated in Figure 2a. As total melting is unlikely (see above), the liquidus does not apply in practice in the context of magma generation. Hence, we focus closely on the solidus. The solidus is a material property and cannot in practice be a straight line in P-T space for multi-phase Earth rocks (see Fig. 1), but a straight line is simple enough to correctly illustrate the concept.

The Earth's asthenospheric mantle, which is the peridotite source of most basaltic magmas, is largely solid as indicated by position *A* (Fig. 2). In order to produce melts from the solid rock, we need to bring the rock from *A* onto or above the solidus. There are three ways to do so (Fig. 2b): (1) heating ($+\Delta T$) the source rock, i.e., to move the rock from *A* horizontally to the right onto or above the solidus; (2) lowering the pressure ($-\Delta P$) of the source rock by "lifting" it from *A* vertically up onto or above the solidus (decompression); (3) as the solidus is a material property, its position in the P-T space depends on bulk composition. For a source rock of the same composition at position *A*, addition of water ($+\Delta P_{H_2O}$, also alkalis and other volatiles) will change the physical property of the rock, and change the solidus to a lower temperature, for example, the curve in Figure 2b. In other words, the "dry" rock at position *A* would now be located above the new and wet solidus, leading to partial melting (Fig. 2b).

The fourth mechanism that can cause a solid rock to melt is to increase pressure as illustrated in Figure 3. Figure 3a compares mantle peridotite dry solidus with the solidus of peridotite that is saturated with CO_2 and H_2O . Obviously, a subsolidus rock located at the open star would begin to melt if it can be moved across the "wet" solidus against pressure. This is theoretically possible, but it is unknown yet under what mantle conditions a solid material could move against pressure (downwards). While subduction of cold oceanic lithosphere into the hot asthenosphere through subduction zone meets the physical requirement of "increasing pressure", the requirement of high temperature (e.g., $< 1000^\circ C$) cannot be readily met by the cold subducting slab. Perhaps, the basal portions of the subducting lithosphere may melt as it descends. The latter is possible because the temperature is getting higher and because the basal portion is likely to be enriched in volatiles [Niu & O'Hara, 2003], thus may reach the wet CO_2+H_2O -saturated solidus during oceanic lithosphere subduction. Should such melting take place, which is inaccessible and unobservable, it would have far-reaching geodynamic significance. The melt so produced may serve as lubricant to ease slab subduction or perhaps even weaken the slab by reducing its thickness and negative buoyancy. The trade-off

is unknown, but these various scenarios and possible effects need to be considered in models of Earth dynamics.

While I focus on basaltic magma genesis, Figure 3b provides insights into the granitic (or granitoids) magma genesis as a result of “increasing pressure”. Granitic magmas have long been considered to result from partial melting of crustal materials as a result of excess heat applied to the source rocks, whether the heat comes from basaltic magmas or is accumulated from radioactive decays of heat-producing elements in the source rocks in various continental settings. Granitic magma generation due to pressure increase has never been considered as a mechanism although phase relationships in P-T space (Fig. 3b) have been known for ~ 3 decades. Figure 3b states that continental crustal lithologies such as basaltic rocks, granitic rocks, intermediate rocks like tonalite, and sediments derived from these rocks, if water-saturated, have the solidus topology with locally negative P-T slope at relatively low pressures (< 10 kbars). This is important because a crustal rock at the position of the open star may experience melting when it moves across the solidus against pressure. The question is under what Earth conditions this process may take place. One possibility is that rapid burial process may lead to deeply buried rocks to melt under elevated pressures. This may indeed explain the origin of some granitic magmas that may have been hardly considered as a mechanism except in the context of the amphibolite facies regional metamorphism and migmatitization or granitization. Importantly, Figure 3b can very well explain the young granitic magmas in regions associated with continental collision. For example, underthrusting of the Indian plate crustal materials (e.g., terrigenous sediments) under Tibet certainly experiences a process of increasing pressure. This brings the sediments from subsolidus condition to super-solidus condition – leading to partial melting and granitic magma generation. I consider that the Miocene leucogranites in the higher Himalaya [Windley, 1995] and southern Tibet [Harrison et al., 1998; Murphy & Harrison, 1999] may have formed this way. This idea, if correct, is geodynamically important and needs independent tests.

2.4. Decompression (- ΔP) melting and adiabat

Except for basalts genetically associated with oceanic lithosphere subduction (i.e., island arc basalts or IAB), the majority of the terrestrial basalts, such as those formed beneath global ocean ridges (i.e., mid-ocean ridge basalts or MORB), those associated active mantle “hotspots” (i.e., ocean island basalts or OIB), and continental flood basalts (CFB), are all derived from basaltic magmas generated by decompression partial melting of the asthenospheric mantle. That is, it is the **ascent** of the asthenospheric mantle that results in such volumetrically important basaltic magma production. In this context, we should realize that it is the asthenospheric mantle flow, and in particular upward flow (decompression), that triggers partial melting and magma generation (Fig. 2). The asthenospheric flow requires a driving force. The very driving force is the presence of some form of pressure gradient – materials flow from regions of high pressures to regions of low pressures, either laterally or vertically. Buoyancy contrast facilitates dense materials to sink, but less dense materials to rise if such buoyancy forces are large enough to overcome viscosity resistance. We will discuss details of these later, but herein focus on upward asthenospheric mantle flow, decompression melting and two basic concepts: (1) adiabat and (2) mantle potential temperature.

Figure 4 illustrates the concepts. The **adiabat** is a thermal gradient or path in P-T space along which the rising asthenospheric mantle material takes without losing heat. The small temperature drop with decreasing pressure (~ 1.8°C/kbar) largely results from volume expansion (Fig. 4). Melting begins when the ascending asthenosphere intersects the solidus. Upon melting, as the melt has greater heat capacity than the solid, the adiabat of the ascending/melting mantle is steeper (~ 6°C/kbar; see Fig. 4). Obviously, if part of the mantle is hot, such as materials within widely assumed mantle plumes, this rising mantle intersects the solidus deeper and melts deeper. If the mantle is relatively cooler such as beneath ocean ridges, this rising mantle intersects the solidus at a shallower level and melts at a shallower depth. As mantle temperature increases with depth, it is unclear to say hot vs. cold mantle. A conceptual reference is needed, which led to the concept of **mantle potential temperature (T_p)**. It is defined as the temperature that a solid parcel of mantle would have if brought to the surface along the adiabat [McKenzie & Bickle, 1988]. For example if we linearly extend the subsolidus adiabat for “Plumes/Hotspots” (Fig. 4) to the surface, it would have a temperature of 1500°C, i.e., $T_p = 1500^\circ\text{C}$. If we do the same for “Ocean Ridges”, it would have a temperature of 1350°C, i.e., $T_p = 1350^\circ\text{C}$. While these concepts are explicitly defined and well accepted, neither the values of the adiabat nor mantle potential temperatures are well constrained. For example, McKenzie & Bickle [1988] suggested $T_p = 1280^\circ\text{C}$ for MORB mantle, whereas Fang & Niu [2003] consider

$T_p = 1350^\circ\text{C}$ for MORB mantle. Green & Falloon [1998], however, believe that T_p is essentially the same beneath ocean ridges and within mantle plumes, approximately, $T_p = 1400^\circ\text{C}$.

2.5. Isobaric and polybaric melting reactions

Fertile mantle peridotites in the depth range relevant to basaltic magma generation are dominantly lherzolites (Fig. 1). Basaltic magmas are generated at depth ranges equivalent to spinel and garnet peridotite stability fields (Fig. 1). Partial melting of natural multiphase systems are incongruent melting. That is, to generate melt, one or two minerals will be created at the expense of other minerals. For example, isobaric experimental work shows that partial melting of garnet lherzolites [see Herzberg, 1992] is characterized by the reaction of the form:



where a , b , c and d are mass fractions of the respective mineral phases, all < 1.0 . This reaction says that in order to produce one mass unit of melt, a mass unit of cpx, b mass unit of ol (olivine) and c mass unit of gnt (garnet) must melt whereas d mass unit of opx must be produced. For partial melting in the spinel lherzolite stability field, isobaric melting experiments [see Kinzler & Grove, 1992; Baker & Stolper, 1994; Walter et al., 1995] shows the following reaction:



where the mass fraction coefficients a , b , c and d have the same significance as above with $a > b$. However, for a natural system undergoing *decompression* melting in the spinel lherzolite stability field, the form of the reaction is the same, but $b > a$ [Niu, 1997, 1999, 2004]. That is, continuous melting as a result of decompression, opx melts faster than cpx (see Figs. 6 and 7 of Niu [1997]). In practice, we know much more about melting in the spinel lherzolite stability field than in the garnet stability field, partly because more and better constrained experiments have been done under these conditions, and partly because of the notion that MORB are mostly generated in the spinel lherzolite stability field [McKenzie & Bickle, 1988; Niu & Batiza, 1991]. Figure 5 shows how primary melts as well as the melting residue change as a function of initial melting depths (e.g., $P_0 = 25, 20$ and 15 kbars), which are depths the adiabatic upwelling mantle intersects the solidus (Figs. 1-4), and the extent of melting as the melting mantle rises (decompression) calculated using the model of Niu [1997] for a preferred fertile lherzolite source (Fig. 5). In all three scenarios, the decompression melting stops at a depth of ~ 8 kbars. The initial depth of melting or the depth at which the adiabatically rising mantle intersects the solidus depends on T_p as well as detailed source mineralogy (Fig. 6). The greater the T_p is, the deeper the ascending mantle intersects the solidus. For example, mantle plumes are often considered to be hotter, thus intersects the solidus deeper than upwelling mantle beneath ocean ridges (Fig. 4). As a result, if decompression melting stops at the same level, the hotter mantle that begins melting deeper will melt more, and produce more melt than the cooler mantle. Note also the initial depth of melting also depends on the composition of the fertile mantle lherzolite [Niu et al., 2001].

3. Basaltic magma genesis and tectonic settings

The plate tectonics theory and mantle plume hypothesis well explain the occurrence of basaltic volcanisms and distribution of most basaltic rocks on the Earth's surface. The plate tectonics theory describes relative motions and tectonic activities along the boundary between two adjacent plates. The classic theory of plate tectonics assumes that plate interiors are rigid, which has led to the definition of three types of plate boundaries: (1) the divergent boundaries (e.g., ocean ridges, backarc spreading centers, continental rift systems), (2) the convergent boundaries (e.g., subduction zones of oceanic-oceanic plates, oceanic-continental plates, and zones of continental-continental collision), and (3) the transform fault systems. Recognizing that many plates are not rigid, but exhibit internal deformation, has led to a fourth type of plate boundaries, termed diffuse plate boundaries [Gordon, 1998]. Many of the orogenic belts in western China, for example, are diffuse boundaries because they are readily reactivated as indicated by frequent earthquakes and some young volcanisms although they are all within the "single" Eurasian plate. Voluminous volcanisms occur along the first two types of plate boundaries, yet rarely at transforms, and even less so with the so-called diffuse boundaries. This observation in fact tells us the way in which magmas form in the context of plate tectonics.

3.1. Magma generation at divergent plate boundaries.

3.1.1. Mid-ocean ridge basalts (MORB) result from decompression melting. In the context of plate tectonics, ocean ridges are mostly passive features because mantle upwelling is largely caused by plate separation [see McKenzie & Bickle, 1998]. The plate separation creates a gravitational “void”, which draw asthenospheric mantle material to ascend passively to fill the “void”. This passive ascent brings the subsolidus material (position *A* in Fig. 2b) along the adiabat (Fig. 4) almost vertically onto the solidus, i.e., decompression. Hence, MORB are produced by decompression melting of mantle material that rises in response to plate separation (Fig. 6a,b). Continued plate separation leads to continued mantle upwelling, decompression melting, and ocean crust formation.

Basalts formed along back arc basin spreading centers (backarc basin basalts or BABB) share a common origin with MORB although the initiation of backarc basins remains mysterious and BABB composition differs from MORB, mostly due to contributions of a fluid phase in the melting region most likely associated with subduction-zone processes.

3.1.2. Magmas associated with continental rifting or lithospheric extension result from decompression melting. While rifts (e.g., the Rio Grande rift in New Mexico, the East African rift system) or regions of obvious lithospheric extension (e.g., the Basin and Range Province in western US) in continents are often considered as intra-plate features, some of these rifts have the potential to develop into new ocean basins such as the Red Sea. In this case, it is convenient to discuss rift/extension associated magmatism under the heading of divergent plate boundaries. Although northern end of the East African Rift may be affected by the so-called Afar hotspot, volumetrically small volcanisms associated with most rift systems or regions of continental extensions largely result from decompression melting of passive rising asthenosphere as is the case for MORB. However, rift-associated magmas are mostly enriched in alkalis, volatiles and other incompatible elements [Barberi et al., 1982]. These characteristics are consequences of two additional factors. (1) Compared to ocean ridges where lithosphere is thin, continental lithosphere in regions of extension or early stage rift is rather thick. This, plus the very slow rate of extension/rifting (< a few mm/yr), leads to limited decompression melting. As a result, the deep and very low degree melting favors generation of melts enriched in volatiles, alkalis and other incompatible elements. (2) Deep portions of subcontinental lithosphere have often been refertilized over its history with metasomatic melts in the form of fertile and easily-melted lithologies (e.g., small dikes or veins of phlogopite, amphibole bearing garnet pyroxenites or even entrapped melts along grain boundaries), and these easily-melted component will contribute to the magmas formed in continental rifting systems (see above). This inference is also supported by the fact that as soon as the rift begins to develop into an ocean basin, the depleted MORB mantle prevails. Red Sea is indeed a good example.

3.2. Magma generation at convergent plate boundaries.

3.2.1. Island arc basalts (IAB) result from subducting-slab dehydration induced mantle wedge melting. As island arc volcanic rocks are dominated by basalts, their source rocks are mantle peridotites. As subduction zones are coldest regions in the upper mantle, heating or thermal anomalies can be safely ruled out as being important. The flow field in the mantle wedge is characterized by the trench-ward and then downward flow dictated by the subducting slab (see Fig. 7) [see Davies & Stevenson, 1992]. Significant decompression is thus unlikely to be important as would be the case beneath ocean ridges. Hence, neither decompression nor heating contribute significantly to IAB magma genesis. Because IAB contain abundant water, water must play a key role in IAB magma generation. These observations and inferences have led to the notion that IAB result from subducting-slab dehydration induced partial melting of mantle wedge peridotites (Figs. 2b & 7) [see Gill, 1981]. The subducting oceanic lithosphere has acquired its water from hydrothermal alteration during its accretion at ocean ridges and subsequent weathering on the seafloor. This interpreted melting mechanism for IAB is *apparently* consistent with the petrology and geochemistry of IAB [see McCulloch & Gamble, 1991; Pearce & Peate, 1995; Davidson, 1996; Niu & O’Hara, 2003].

Figure 7 is schematic, but several elements of conceptual significance should be noted, in particular the significance of the big arrow to the left. This arrow means that there is a lateral asthenospheric flow towards mantle wedge corner, which is required by the corner flow to replenish the “vacancy” caused by slab-dragged downward mantle flow near the subducting slab. This suggests that mantle wedge “corner flow” represents a low-pressure region that draws asthenospheric materials to move towards the corner (or “corner suction”). This lateral flow is also required by IAB compositions – while slab dehydration and “slab

component” is geochemically important, without continuous lateral fertile asthenospheric material supply, mantle wedge would become too depleted and too refractory to melt, let alone to produce apparently long-lived arc volcanism and observed arc magma compositions.

We here also need to conceptually distinguish the convective asthenospheric mantle wedge and the overlying arc lithosphere that is isolated from the convection. However, slab-released water may hydrate the overlying lithosphere, at least the forearc lithosphere beneath which subducting slab just begins to dehydrate. This should and will weaken this part of the lithosphere by reducing its viscosity. This may become a process that transform part of the lithosphere into convective asthenosphere. This converted materials, when dragged down to depths, may contribute to the arc volcanism. Therefore, arc magmas may have source components that include the familiar slab component (including sediment input), asthenospheric mantle wedge materials, and also previous sub-arc lithospheric materials. The latter is particularly important in understanding IAB petrogenesis.

3.2.2. Volcanic rocks observed at active continental margins such as the Andes along the Pacific coast in the South America are mostly andesites and other silicic varieties. It is important to realize that these are not primary magmas, but differentiated from “primary melts” by cooling and assimilation with the thickened crust on their way up. The “primary” melts are basaltic melts resulting from slab-dehydration induced partial melting of mantle wedge peridotites as is the case for IAB.

3.2.3. Adakites are **geochemically distinctive** calc-alkaline andesites/dacites interpreted to result from partial melting of subducting slab. Subduction zones are coldest regions of the upper mantle because of cold oceanic lithosphere subduction. This lithosphere is often too cold to melt, but plays a key role in supplying fluids and water-soluble elements in IAB magma genesis (see above). However, when an ocean ridge is spatially close to a trench, and if the subducting lithosphere is young (e.g., < 25 Ma), this lithosphere may still be **warm**, and **could** well be heated up quickly during its early/shallow subduction [Defant & Drummond, 1990]. This heated slab, plus the dehydration effect, may reach the solidus of less refractory portion of the subducting slab, i.e., the crustal lithologies atop the slab such as basalts and gabbros that have metamorphosed to blueschists and eclogites. This physical scenario has been **interpreted** to be capable of explaining the geochemical characteristics of **adakites**, coined by Defant & Drummond [1990] after the name of Adak island of Aleutian arc. The geochemical characteristics include: andesite/dacite major element composition, high concentrations of Sr, light rare-earth elements (REEs), low in Y and heavy REEs, low in alkali elements (Rb, K etc.), and very steep REE patterns (i.e., high La/Yb) [Defant & Drummond, 1990]. The more silicic bulk composition is consistent with partial melting of mafic rocks (basaltic composition). The high Sr is consistent with partial melting of the high Sr protoliths of oceanic basalts/gabbros. The low K is consistent with K-depleted nature of oceanic crustal rocks. The steep REE patterns or the high La/Yb ratio is also consistent with the source rock being eclogitic under the conditions of partial melting, where garnet as a residual phase holds back much of the heavy REEs. All these **consistencies** only suggest, but do not prove, that **adakites** result from partial melting of subducting ocean crust [Garrison & Davidson, 2003].

Note also that partial melting of eclogites with basalt/gabbro protoliths will produce melts with compositions similar to adakites in many ways without providing any tectonic information, whether they are formed in a subduction setting or in fact formed deep in the continental crust or crust/mantle transition zones where previously underplated basaltic rocks may have metamorphosed to eclogites. Partial melting of these eclogites must be associated with more recent mantle thermal events. In this case, such adakite-like rocks may be enriched in alkalis depending on the bulk composition of the protoliths vs. alkali-depleted normal oceanic crust. TTG (tonalite-trondhjemite-granodiorite) suites are likely the products and derivatives of partial melting of basaltic eclogites or garnet-bearing amphibolites (deep portions of the proto crust?) in the Archean and perhaps over much of the Earth’s history.

In this context and on a broader scale, we should ask ourselves “Is it really appropriate to name rocks on the basis of **process**, or **origin** or **trace elements**?”¹ As “the history of petrogenesis has been the history of re-interpretations of origins”², which means the rock origins are often poorly known, it is logical to name rocks descriptively based on petrographic observations. Also, I. Bindeman and P.J. Wyllie³ noted that in the past few years the number of papers using name **adakites** has increased enormously, **mostly** in Chinese publications. Are there more **adakites** in China? Have we discovered more in China? Or do we intend to

^{1,2,3} Personal Communication with P.J. Wyllie (California Institute of Technology)

surpass the international-level of **adakite** understanding? The observation by Bindeman and Wyllie is worth thinking.

3.2.4. Boninites are geochemically highly depleted MgO-rich andesites. The origin of boninites is hotly debated. Partial melting of subducting/subducted slab, multistage melting of previously melted MORB residues, mantle plume/hotspot-influenced melting etc. are a few examples [see Crawford, 1989; Falloon & Danyushevsky, 2001]. There are also attempts to divide boninites into high-Ca (high temperature) and low-Ca (low temperature) boninites, and some more subdivisions using subtle major element compositional variations [Crawford et al., 1989]. Apparently, the large compositional variability of boninites in major elements, trace elements and radiogenic isotopes is likely the reason for diverse petrogenesis models. However, a first order petrogenesis model is possible if we consider a number of unique aspects of boninites: (1) exclusively associated spatially with subduction zones (arcs, forearcs, and backarcs); (2) rich in water from ~ 1 to up to 8 wt.%; (3) high Mg/[Mg+Fe²⁺] (Mg[#] > 0.75 mostly); (4) high Mg olivine (Fo > 0.90) and high Cr/[Cr+Al] (Cr[#] often > 0.7) spinel phenocrysts; (5) opx as a characteristic liquidus phase; (6) depletion in Nb, Ta, Zr, Hf, Ti and most REEs with concave upwards REE patterns; (7) relatively enriched in water-soluble incompatible elements (Ba, Rb, Sr etc.); and (8) variable but radiogenic Sr, Pb and unradiogenic Nd isotopic ratios relative to MORB. All these characteristics are consistent with subduction-zone slab dehydration induced partial melting of highly depleted harzburgitic mantle peridotites. The radiogenic Sr and Pb and unradiogenic Nd isotopes depend on the nature and histories of the harzburgites, slab-derived fluids, and perhaps also the contribution of subducted terrigenous sediments. The key here is the nature of the harzburgitic source rocks, which may be part of the depleted subarc asthenosphere or even perhaps the subarc lithospheric root. The latter may either be remnants of ancient subcontinental lithosphere or compositionally depleted and physically buoyant oceanic plateau roots [Niu et al., 2003].

3.2.5. Can collision produce magmas? Continent-continent collision is one type of convergent plate boundaries. The present-day type example is the collision of the Indian plate with the Eurasian plate along the Himalayas. The collision began with the closing of the Tethys ocean following the subduction of the oceanic lithosphere portion of the Gondwana beneath the Eurasia. The present-day “subduction” is largely resisted by the continental collision because neither side in bulk is necessarily denser than the other. However, seismic data suggest deep portion of the Indian plate continues to **underthrust** beneath Tibet. The lack of subduction-zone like volcanisms such as IAB at island arcs (e.g., in the western Pacific) or continental arcs (e.g., the Andes) reflect the fact that the present-day **underthrusting** lithosphere of the Indian plate is continental, and is too dry (vs. oceanic lithosphere) to have any significant fluid to release. The observation/inference here also counters the old theory of friction-heat induced melting in subduction zones. The question is whether collision by itself can cause mantle melting. The answer is simply NO. This is because collision (1) cannot generate sufficient heat to cause mantle melting, (2) cannot in any sensible way induce decompression of mantle materials, and (3) cannot provide any excess water needed to trigger mantle melting (see Fig. 2). Rapid uplift and erosion (or “unroofing”) may remove the overlying load and have the effect of “decompression” for deep seated rocks, but whether such “decompression” can cause mantle melting remains to be explored (e.g., detailed evaluation of rate of “unroofing” vs. rate of conductive heat loss).

The next question is if collision can cause crustal melting and granitic magma formation. The terminologies such as “syn-collisional granites”, “post-collisional granites” and “post-orogenic granites” etc. [Pearce et al., 1984; Harris et al., 1986] have led to the perception that there must be some sort of genetic link between continental collision and granitic magmatism. There could be some links, but confusions arise because we do not know the exact meanings of these various types of granites in terms of physical mechanisms through which they may form in the context of continental collision. Two apparent difficulties are: (1) our dating techniques may not have sufficient resolving power to distinguish between “syn-” and “post-” collisional processes; and (2) there is no theoretical foundation yet known why “syn-collisional” and “post-collisional” granites should be distinct in petrology and geochemistry. Therefore, any attempt to distinguish these granites, if they did exist as defined, is at best an arbitrary exercise with no scientific significance. The most fundamental problem is the fact that we actually do not understand why there should be magmas formed at all as a result of continental collision. Continental collision as such cannot in any straightforward way provide needed conditions for magma generation not only in the mantle (see above), but also in the crust in terms of widely perceived mechanisms (Fig. 2). Personally, I hypothesize that subducting or underthrusting crustal materials (e.g., from the Indian plate) saturated with water (crustal rocks including terrigenous sediments are not dry, but full of hydrous minerals such as amphiboles, micas, chlorite and all

sorts of clays) may lead to “compression” induced partial melting of these materials (Fig. 3b), giving rise to the Miocene leucogranites in higher Himalaya [see Windley, 1995] and southern Tibet [Harrison et al., 1998; Murphy & Harrison, 1999].

The above is an alternative interpretation (see Fig. 3b) on collision-zone associated leucogranite genesis. This interpretation complements the prevailing ideas of granitic magma genesis. That is, heating-induced crustal melting is the very process that is responsible for most granites on Earth. The observation that most granites or granitoids occur in belts along active continental margins like the present-day Andes suggest that the most important heat responsible for crustal melting comes from the mantle rapidly carried up by mafic magmas. Therefore, most, if not all, granites are genetically associated with mantle melting and basaltic magma generation, at least in terms of the heat source. During episodes of voluminous continental flood basalt volcanisms, the mafic magmas can also assimilate the crust and lead to localized crustal melting and granitoid magma genesis. Anatexis may also occur to produce granitic magmas at deep crustal levels if the rocks there are rich in heat-producing elements (e.g., K, U, Th), if there are overlaying heat-insulating rocks to prevent heat loss, and if they may also experience burial compression (see Fig. 3b).

3.2.6. Young alkalic volcanisms associated with “dead” and active collision zones. There is widespread, though volumetrically small, alkalic volcanism in tectonically active Tibet and some old orogenic belts (e.g., some Paleozoic orogenic belts such as the west Kunlun, Tianshan, and west Qingling, China). The origin of these volcanisms is poorly understood, but cannot be caused by collision. The ideas of localized lithospheric extension, lithospheric delamination etc. are interesting, but these cannot pass simple physical tests. Given the fact that much of the western China consists of multiple Paleozoic collision zones or orogenic belts, it is more like “regions” of diffuse plate boundaries within the giant Eurasia plate. Frequent earthquakes in the vast regions, and in particular along these paleo-orogenic belts suggest reactivation of lithosphere-cutting faults or fault zones. This may trigger highly localized decompression melting of deep lithosphere that is enriched in easily-melted metasomatic dikes/veins with elevated abundances of volatiles and incompatible elements. This may very well explain some volumetrically small yet unusually enriched alkalic volcanics [see Yu et al., 2004], but more detailed work is needed.

3.3. Magma generation at transform fault systems.

Volcanism is either absent or very rare in transform fault systems. This is consistent with the fact that (1) there is no excess heat there available for asthenospheric mantle melting; (2) there is no excess water available to trigger melting; and (3) the strike-slip dominated plate motions does not create gravitational void for passive asthenospheric upwelling and decompression. However, limited volcanism has indeed been found in some **leaky transforms** such as the Garrett transform [Hékinian et al., 1995; Niu & Hékinian, 1997b], Raitt transform [Castillo et al., 1998], and Siqueiros transform [Perfit et al., 1997] in the Pacific. These transforms are tectonically transtensional in nature where such extension induced passive upwelling and melting is evident not only from the erupted basalts but also the morphology and orientations of the intra-transform volcanic ridges [Niu & Hékinian, 1997b; Wendt et al., 1999].

3.4. Intraplate volcanisms.

Magmatism occurring within plates away from plate boundaries cannot be readily explained by plate tectonics because plates are defined as being “rigid” and essentially nothing should take place within plates. Many of the voluminous intraplate volcanisms making within-plate ocean islands such as Hawaii, the Society, and Samoa in the Pacific, the Azores, St. Helena, and Tristan in the Atlantic, and the Kerguelen, Bouvet and Marion in the Indian Oceans are called **hotspot** volcanisms. These hotspots are implicitly interpreted as surface expressions of deep rooted mantle plumes, perhaps originated from the lower mantle. Belonging to this category are also voluminous volcanisms responsible for many continental flood basalts and oceanic plateaus (also called large igneous provinces or LIPs). Magmas associated with mantle plumes are interpreted as resulting from decompression melting of ascending plumes. These plumes may be originated from the deep mantle, but melting begins in the upper mantle at depths greater than melting beneath ocean ridges (Figs. 1,4,6). Different from decompression melting of sub-ridge mantle where mantle upwelling is a passive response to plate separation, the upwelling of mantle plumes is considered active or dynamic because of positive thermal buoyancy – hot and less dense.

Ever since the inception of deep-rooted mantle plume hypothesis [Morgan, 1971, 1981], and the successful prediction [Morgan, 1971] and elegant explanation of the volcanic age progression along the Hawaii-Emperor seamount chains in the Pacific [see Clague & Dalrymple, 1989], mantle plumes, or

hotspots, have been widely invoked to explain mantle melting anomalies within the interiors of lithospheric plates. There have been thousands of research papers dealing with intraplate volcanisms and there are also volumes on hotspots, mantle plumes, and large igneous provinces in the last decades. However, the mantle plume hypothesis has now been seriously challenged by many using various arguments [Anderson, 2002, 2004; Hamilton, 2003; Foulger & Natland, 2003]. The debate becomes progressively more intense, but it is the debate that can expedite the revelation of the truth. The mantle plume website¹, updated weekly, can provide interested readers with the always current views whether they favour or disfavor mantle plume hypothesis. My personal view is that if subducting slabs (at least some) can indeed penetrate the 660 km seismic discontinuity into the lower mantle, then to balance the mass, materials from the lower mantle must rise to the upper mantle [Niu & O'Hara, 2003]. The latter, if developed to show surface expressions as hotspots, could very well come from the Earth's deep hot thermal boundary layer, i.e., the seismic D" region or regions near the core-mantle boundary (CMB) as the most popular plume theory prefers. The CMB origin is possible, but it is not required. Subducted oceanic lithosphere with abundant hydrous phases may, with deep mantle heating, experience phase changes (possibly forming a melt phase too), develop into compositionally buoyant domains to ascend. If they ascend adiabatically, they should be hotter than the surroundings in the upper mantle and then further develop into "plumes" feeding intraplate volcanisms. The plumes of the kind may indeed exist in the Earth's mantle, but not all the so-called plumes are plumes, but localized melting anomalies. In summary, if "plume-like" domains do exist to rise in the mantle, they are most likely derived from regions of compositional anomalies. They could very well be hotter when rising into the upper mantle, but they do not need to come from the core-mantle boundary and they do not and perhaps cannot take the cylindrical form in the deep mantle (viscosity is too high) [Anderson, 2004] until reaching some rather shallow levels in the upper mantle. This argument does not say mantle plumes come from upper mantle, but emphasizes that "plume" materials may ascend from the lower mantle, but only develop into narrow geometries of "plumes" at shallow levels where pressure-dependent mantle viscosity is reduced.

Numerous seamounts both in short "chains" and randomly scattered throughout world ocean floors away from plate boundaries [Batiza, 1982] are obviously mantle melting anomalies best explained by lithosphere cracking and localized melting of geochemically-enriched and physically easy-to-melt compositional heterogeneities [Niu & Hékinian, 2004] or mantle wet-spots [Green & Falloon, 1998].

4. Magma Evolution

Basaltic magmas (e.g., volcanic lavas) we sample on the Earth are not primary magmas; all have evolved to various extents from their "primary" status [see O'Hara, 1968a, 1968b, 2001; O'Hara & Herzberg, 2002]. **Primary** magmas are those produced by one of the mechanisms (Figs. 2-3) in the mantle source regions in equilibrium with source minerals without undergoing any modification. **Primitive** magmas are those that after being extracted from their source regions have experienced relatively less modification than the more **evolved** ones. **Parental** magmas are any magma from which more evolved ones are derived. Conceptually, primary magmas are only produced experimentally.

Magma evolution refers to changes the magmas in question are experiencing after being extracted from the source regions and emplaced in a magma chamber or some shallow levels during ascent. The changes are mostly the result of magma cooling. Cooling leads to magma crystallization. Density contrast between the forming/formed crystals and cooling magmas leads to crystal-melt separation if viscosity permits, and formation of cumulate crystalline rocks. This simple description refers to the familiar process of "crystal fractionation" or "magma differentiation" [Bowen, 1928]. Magma evolution is, in simplest sense, synonymous to magma differentiation. For example, more evolved magmas can also be called more differentiated magmas, which depart more from relatively primitive magmas due to crystal separation. In practice, the process of magma evolution/differentiation can be quite complex. The magmas may interact both thermally and compositionally with the wall-rock, i.e., magma assimilation, resulting in changes of the residual magmas as well as fractionating mineral assemblages. Furthermore, most magma chambers are very likely open systems with crystallization and mineral accumulation being accompanied both temporally and spatially by replenishment of new batches of the more primitive magmas. The concept "geochemical evolution in an advancing, periodically replenished, periodically tapped, continuously fractionated magma chamber" [O'Hara, 1977; O'Hara & Mathews, 1981] well summarizes the complexity of natural magma chambers. Magma eruption may well be triggered by replenishment of new batches of magmas or as a result

¹<http://www.mantleplumes.org/>,

of external forces such as possible earthquakes etc. There are many quantitative treatments of magma chamber processes in the literature in terms of petrology, geochemistry and fluid dynamics [see O'Hare, 1977; Sparks et al., 1984; Campbell & Turner, 1986], and there are also numerous trace element modeling efforts summarized in many textbooks [see Rollinson, 1993; Albarède, 1995]. In this section, I emphasize basic concepts of cooling-induced basaltic magma evolution and the petrologic consequences.

4.1. Liquidus temperatures and the concept of liquid lines of descent

It has been well-established in the past decades that the liquidus temperature (T_{liquidus}) of a basaltic melt is proportional to its MgO content (Fig. 8a) [see Niu et al., 2002b]. This MgO- T_{liquidus} relationship is largely based on "dry" basaltic systems, but the first-order relationship remains valid for water-saturated basaltic melts although the T_{liquidus} becomes lower at a given MgO (see below). Because of this simple T_{liquidus} -MgO relationship (Fig. 8a), MgO of basaltic melts (or whole-rock compositions of aphyric basalts and diabases) are often used as proxy for the liquidus or eruption temperatures. MgO variation diagrams of major elements (e.g., SiO₂, TiO₂, Al₂O₃, FeO, CaO, Na₂O etc. against MgO) are used to understand how magma changes in composition with cooling (decreasing MgO). Such melt compositional change with cooling results from separation of liquidus minerals that have thus far crystallized. For a suite of basalts genetically related to crystal fractionation, the compositional trends on the MgO variation diagrams are called **liquid lines of descent (LLDs)**, a term coined by Bowen [1928], the father of modern igneous petrology, to describe magma evolution in terms of his crystallization-differentiation theory.

There is a close equilibrium relationship between $Mg^{\#}$ (defined as atomic Mg/[Mg+Fe²⁺]) of the melt and the composition of liquidus olivine (i.e., forsterite content, Fo): $Mg^{\#}(\text{melt}) = 1/([1/\text{Fo}-1]/K_d+1)$, where $K_d = 0.30 \pm 0.03$ [Roeder & Emslie, 1970] at low pressures. Both $Mg^{\#}$ of the melt and Fo content (Mg/[Mg+Fe]) of the liquidus olivine are also proportional to the liquidus temperature although the relationship is not as well defined for MgO in the melt (see below). For this reason, Fo of olivine is a useful index for the liquidus temperature of the melt from which this olivine has crystallized. Note that the T_{liquidus} -Fo_{olivine} relationship can indeed be used to estimate the liquidus temperatures of the basalts under study, even if the rocks may be porphyric (cumulate minerals) or perhaps even altered as long as the composition of the olivine can be properly determined. Figure 8b shows this useful relationship. It states that if the olivine has the composition falling on the upper band, this olivine must be in equilibrium with the melt of the corresponding $Mg^{\#}$ (vertical axis) at the same temperature (horizontal axis). The two equations in Figure 8b are adequate for calculating T_{liquidus} of basaltic melts if the olivine composition is known. Caution: whole-rock compositions are NOT compositions of basaltic melts unless the rocks are (1) aphyric (i.e., aphyric basalts and diabases, but **NOT** gabbros) and (2) entirely fresh. Also note that the T_{liquidus} -MgO relationship in Figure 8a is valid in most cases for relatively dry basalts (e.g., MORB and most OIB). However, T_{liquidus} -Fo_{olivine} is not always true. For example, in general (not always), "mantle plume" associated primitive melts have both high MgO and FeO, whereas primitive MORB melts have both low MgO and FeO [see Herzberg & O'Hara, 2002]. Thus, both "plume" melts and MORB melts could have the same or similar or overlapping $Mg^{\#}$, but "plume melts" have higher MgO at a given $Mg^{\#}$, thus have higher T_{liquidus} because, again, T_{liquidus} is more directly proportional to MgO, not $Mg^{\#}$, in melts! Caution is still necessary when using olivine composition to infer T_{liquidus} of the melts from which the olivine under study crystallized.

Figure 9 shows LLDs of anhydrous basaltic melts, in terms of SiO₂, Al₂O₃, FeO, CaO and P₂O₅, derived from > 400 MORB glasses from the East Pacific Rise [Batiza & Niu, 1992; Niu & Batiza, 1997; Niu et al., 1999, 2002a; Regelous et al., 1999] plotted as a function of $Mg^{\#}$. As the magma cools (decreasing $Mg^{\#}$ from right to the left), the compositional variation trends of the remaining liquid reflects what minerals have been crystallized and separated. This is readily seen from the bottom two panels. For example, from this set of the data we see that olivine (oliv) begins to crystallize at $Mg^{\#} \sim 0.72$, $\sim 1240 \pm 10^\circ\text{C}$. The olivine only mineral crystallized along with chromite (chmt) would be dunite. This is the very reason why chromites and chromite deposits are always associated with dunite (e.g., the Luobusha chromite deposit). At $Mg^{\#} \sim 0.69$, $\sim 1222 \pm 10^\circ\text{C}$, plagioclase (plag) joins olivine to form plag + oliv rock, called troctolite. At $Mg^{\#} \sim 0.58$, $\sim 1180 \pm 10^\circ\text{C}$, clinopyroxene (cpx) joins plag and oliv to form plag + cpx + oliv rock, called gabbros. At an even lower temperature, $\sim 1115 \pm 10^\circ\text{C}$ with $Mg^{\#} = \sim 0.37$ in the melt, Ti-Fe oxides begin to crystallize (with or without orthopyroxene and fayalitic olivine) forming Fe-Ti gabbros, gabbro-norite etc. Note that apatite, and perhaps even zircons may crystallize at this late stage evolution of tholeiitic basaltic magmas. Importantly, magmatic V-Ti-Fe ore deposits are the consequences of this late stage basaltic magma evolution (e.g., the Panzhihua V-Ti-Fe deposit).

It should be clear by now that dunites, troctolites, gabbros etc. are **NOT** melts, but cumulate rocks (i.e., consisting of accumulated crystals separated from the cooling melts before solidified)! Therefore, the terminology such as dunite magmas, gabbro magmas etc. are **WRONG** in principle and **DO NOT** exist in practice. Also, note that the statement “gabbros” are intrusive equivalent of basalts in many textbooks is **WONG**. They are **NOT** equivalent. Gabbros are cumulates, but basalts, in particular, the aphyric ones are cooled, evolved, erupted, and then solidified basaltic liquids [see Niu et al., 2002b]. On the other hand, it is valid to say “dunite fractionation”, “troctolite fractionation”, “gabbro fractionation” etc. to describe the process as these *specify* what minerals are on the liquidus.

4.2. Liquidus temperatures of “wet” basaltic magmas

The above MgO- T_{liquidus} relationship (Fig. 8a) is based on anhydrous (dry) experiments and works well for volatile-poor MORB and perhaps also for most OIB. However, island arc basalts (IAB) are quite wet, and it is important to know if the MgO- T_{liquidus} relationship also exists in hydrous basaltic melts. Experiments on hydrous basaltic system are rare, and limited run numbers (data points) are often inadequate to derive rigorous relationships between melt compositions and physical parameters. The experimental work by Gaetani & Grove [1998] may be the best attempt of the kind. Figure 10 plots their experimentally determined melt compositions in terms of H₂O and MgO as a function of run temperatures under three isobaric pressures (i.e., 1.2, 1.6 and 2.0 GPa respectively). Figure 10a shows that there is no obvious correlation between run temperatures and H₂O contents in the melts. However, if we consider only 1.2 GPa runs (solid circles), an inverse correlation becomes apparent. The interpretation of such an H₂O- T_{liquidus} relationship is not straightforward. It could mean that the capacity of the melt dissolving H₂O declines with increasing temperature (?). So the exsolved H₂O could be a vapor phase at elevated temperatures, in which case such vapor phase would be in equilibrium with the melt. It is also possible however, the declining H₂O with increasing temperature could be due to H₂O loss out of the experimental capsules through H diffusion. The two runs with low-H₂O is also peculiar because the T_{liquidus} would be higher than that of dry melts at a given MgO as seen in Figure 10b. In Figure 10b, if we exclude the two low-H₂O runs (we do not have justified reason in doing so yet, but they are at odds with T_{liquidus} determined for dry basaltic melts), all the experimental runs show a positive MgO- T_{liquidus} , which gives $R = 0.951$, statistically significant at > 98% confidence level. This indicates that a positive MgO- T_{liquidus} exists for hydrous basaltic melts. Note that the trend defined by the experimental data has slope very similar to that derived from the anhydrous experiments (the thick line, also see Fig. 8a). This means that while H₂O can lower T_{liquidus} , the latter is independent of the actual H₂O contents in the melts, at least for melts with > 3.0 wt % H₂O. As the composition of melts produced at 3 pressures lie on the same trend, the pressure effect is insignificant. Considering most observed magmas have MgO < 12 wt%, a simple regression is done on experimental runs with MgO < 12 wt % as indicated by the thin line. This regression line is essentially parallel to the slope of the anhydrous MgO- T_{liquidus} , with $T_{\text{liquidus}} \sim 36 \pm 9^\circ\text{C}$ lower than anhydrous melts at a given MgO. This suggests that we can still use the equation in Figure 8a to evaluate T_{liquidus} of hydrous melts (e.g., arc magmas) with some confidence, but considering the relationship: $T_{\text{wet liquidus}} = T_{\text{dry liquidus}} - 40^\circ\text{C}$ (Fig. 10b). This approximation is adequately precise considering the experimental and analytical uncertainties.

4.3. Distinctive evolution paths of anhydrous vs. hydrous basaltic magma evolution

The reason that MORB are essentially anhydrous basaltic liquids is because MORB mantle sources are depleted in incompatible elements as well as volatiles like water. Therefore, MORB melts follow an evolution/fractionation path (see Fig. 9): **dunite-troctolite-gabbro** in terms of cumulate mineral assemblages. In contrast, hydrous basaltic melts such as IAB with abundant water take a different path in that the cumulate equivalent to troctolite in anhydrous system becomes wehrlite in hydrous system. That is, differentiation of hydrous basaltic melts takes a path of **dunite-wehrlite-gabbro** in terms of mineral assemblages. This is well established by distinct LLDs. For example, Figure 11 shows that on CaO, Al₂O₃ and CaO/Al₂O₃ vs. MgO plots, the anhydrous MORB melts define LLD that differs from LLD of an IAB suite from the Vanuatu Arc. This difference due to different crystallization sequence of cpx and plag is controlled by the amount of water in the melts. This has been understood experimentally using simple binary phase diagram diopside (Di)-anorthite (An), which is analogous to basalt [Yoder, 1965]. For example, a basalt with ~ 60% normative plagioclase (An) and ~ 40% normative diopside (Di) (Fig. 11) will crystallize An before Di begins to crystallize at the “dry” eutectic. However, the same basalt with abundant water (i.e.,

high water pressures) will crystallize Di before An joins in at the “wet” eutectic. This has been experimentally confirmed for multicomponent systems [Gaetani et al., 1993].

The above has significance in inferring tectonic settings for field geologists. The presence of troctolite in a mafic cumulate sequence would suggest an anhydrous (water poor) parental magma [Niu et al., 2003], which, in combination with the coexisting lithologies, may point to a tectonic setting of ocean ridges. On the other hand, the presence of wehrlite in a mafic cumulate sequence would suggest a hydrous (water rich) parental magma, which, in combination with the coexisting lithologies, may point to a tectonic setting associated with island arcs, forearcs or perhaps back arcs. Note that cooling of anhydrous basaltic magmas at elevated pressure may lead to early crystallization of augite (perhaps before plagioclase), which can lead to wehrlite (vs. troctolite) fractionation. However, this requires pressures well in excess of 8 kbars [see Langmuir et al., 1992], which is equivalent to a magma chamber at depths in excess of > 25 km. In general, the effect of water, say > 2-3 wt %, has greater effect than pressure in advancing augite (vs. plagioclase) crystallization. Note that in the absence of aphyric basalts, the above cumulate assemblage is a more effective discriminant for tectonic settings – much more so than trace element geochemistry.

The trace element approach applies only to aphyric basalts (including aphyric diorites), and **does not** apply to cumulate rocks to infer tectonic settings. Using cumulate bulk-rock trace element geochemistry to discriminate tectonic setting is dangerous. For example, on a commonly used “spidergrams”, whether the bulk gabbroic rocks show Nb-Ta-Ti anomalies or not depends on the abundances of titanomagnetite (ilmenite, magnetite and/or solid solutions) in the rock. If a modern oceanic lower crustal gabbro sample [Niu et al., 2002b] is deficient in these oxides, the bulk-rock would have negative Nb-Ta-Ti anomalies, and could be interpreted otherwise as genetically associated with an arc setting, which is known to be wrong. The challenge comes from studying rocks along orogenic belts in the geological record. It can not be over emphasized that field observations and lithological assemblages are fundamental to any geochemical interpretations. It is also worth to emphasize that modern analytical instruments allow acquisition of unprecedented high quality geochemical data, but any meaningful interpretations of these data cannot be divorced from basic petrological understanding and field observations.

5. A hypothesis for the origin of Mesozoic-Cenozoic volcanism in eastern China

Focusing our discussion on basaltic volcanism, we must ask ourselves what may have caused the asthenospheric mantle melting to produce the volcanism in “plate” interiors far from plate boundaries. This is not simply an issue of Chinese regional geology, but a scientific problem of global significance on Earth’s dynamics. The latter interests me profoundly, and justifies that I present a personal view here. My view is perhaps rather simplistic, but I maintain the personal philosophy that “simple models are preferred over complex models, and complex models become necessary if no simple models work”. The reality is necessarily complex, but we must understand first order problems before we can logically get into details. My view presented here may not necessarily be **perfectly** correct, but this is not the purpose. The purpose here is three-fold: (1) I would like to share and discuss my view with our Chinese colleagues; (2) we should not make simple things complex, but rather try to make apparently complex matters as simple as practically possible without neglecting the observations; and (3) I hope my view will provide some food for thought. A more comprehensive and substantial account of Mesozoic/Cenozoic young volcanism in eastern China is under consideration (Y. Niu, S.-G. Song, L.-F. Zhang, X.-X. Mo, J.-F. Deng, S.-G. Su, Z.-F. Guo and J.-Q. Liu).

5.1. Hints from topography, gravity, crustal thickness and mantle seismic tomography

Figure 12a, portion of the world topography map, shows a sharp altitude contrast in continental China between the high plateaus to the west and hilly plains to the east along a straight line marked by the arrowed NNE-SSW dash line. This line, which represents the steepest altitude gradient from the east to the west of China, also coincides with the steepest gradients in gravity anomalies (Fig. 12c) and crustal thickness (Fig. 12d). Interestingly, this line also marks the steepest gradient in mantle seismic velocity clearly seen at depths of 100 km and 150 km (Fig. 12b). The sudden seismic velocity (V_p) decrease across the line from the west to the east is consistent with the interpretation that at such depths the mantle beneath the plateaus in the west is “cold” and “fast” lithosphere whereas beneath the east it is the “hot” and “slow” asthenosphere. The latter is consistent with the recognition that the lithosphere beneath eastern China is **anomalously** thin, considering the geologically perceived cratonic nature in the North China – the North China Craton (NCC). Petrologically, the existence of Paleozoic diamondiferous kimberlites in the NCC (e.g., Fuxian in Liaoning

Province, Mengyin in Shandong Province) and elsewhere in eastern China (e.g., Jilin, Jiangsu, Anhui, Henan, Shanxi provinces, and Inner Mongolia) [Sun et al., 1993; Chi et al., 1992; Lu et al., 1991, 1995, 2000; Griffin et al., 1998] all indicates that the eastern China lithosphere must have been ~ 200 km thick back in the Paleozoic. However, recent studies of mantle xenoliths [see Song & Frey, 1989, 1990; Zhi et al., 1990; Xu et al., 1998, 2000; Xu, 2001] indicates a much thinner present-day lithosphere, perhaps no more than 80 km thick beneath eastern China, which is confirmed by seismic studies [Chen et al., 1991] and mantle tomography (Fig. 12b). Hence, the lithosphere beneath eastern China, in particular beneath the NCC, must have lost > 120 km thick bottom portion [Menzies et al., 1993], probably in the Mesozoic [see Menzies et al., 1993; Deng et al., 1998, 2004; Griffin et al., 1998; Zheng, 1999; Wu & Sun, 1999; Xu, 2001; Gao et al., 2002; Zhang et al., 2002; Yan et al., 2003].

In this context, it is noteworthy that the steep altitude gradient line (Fig. 12a) is parallel to the Indian-Eurasian collision vector (NNE) and is more or less perpendicular to the vector of present-day Pacific plate motion (NW). For convenience, we may call the line, Great Gradient Line or **GGL**, which applies to (1) topography, (2) gravity anomaly, (3) crustal thickness, (4) mantle seismic velocity, and thus (5) lithosphere thickness variation as well. Note that the **GGL** is likely to be a relatively young phenomenon as a result of Indian-Eurasian collision, which began no earlier than the Tertiary, and is likely younger than the major episodes of the eastern China lithosphere thinning (see below). “Are lithospheres forever?” [O’Reilly et al., 2001] is a good question to which the authors have already provided an answer. Lithospheres are not forever. The question is what may have caused the lithosphere thinning.

5.2. What may have caused the lithosphere thinning beneath eastern China?

The conventional view holds that the subcontinental lithospheric mantle (SCLM) represents partial melting residues, and is thus compositionally depleted (i.e., more refractory with high Mg/Fe) [Ringwood, 1975] and physically buoyant with respect to the underlying asthenosphere [Jordan, 1988; Abbott et al., 1997; Niu et al., 2003]. As a result, the buoyant SCLM, ever since its inception, has been physically isolated from the convective asthenospheric mantle. This explains why the SCLM is generally old, on average > 2.5 Ga, in contrast with the young (< 200 Ma) oceanic lithosphere. It is also the buoyant nature of the SCLM that allows it to protect the overlying crust from being destroyed [Jordan, 1975, 1988; Anderson, 1995]. The similar age of the SCLM and the overlying crust [Pearson, 1999] in some cratonic regions not only points to their physical association, but probably the genetic link as well. Given all these widely perceived physical properties of the SCLM, lithosphere thinning is not a norm, but an exception. The fact that lithosphere thinning did not seem to happen beneath African cratons but has indeed happened beneath the NCC gives us some hints. The African plate (or lithosphere) is characterized by passive continental margins and is related to adjacent plates through ocean ridges. By contrast, the NCC is surrounded by zones of diffuse plate boundaries [Gordon, 1998] to the west as a result of Indian-Eurasian collision, and westward Pacific/Philippine plate subduction to the east. From all these, we can reason that the lithospheric thinning and the Mesozoic/Cenozoic volcanism in eastern China in general and the NCC in particular may be somehow related to the western Pacific subduction beneath east Asia. The Indian-Eurasian collision represents a rather young event (since tertiary), and it may not contribute to the lithosphere thinning beneath eastern China, but may contribute indirectly to the Cenozoic eastern China volcanism (see below). Let us explore these scenarios or possible/probable cause/effects relationships further in terms of existing observations and more recent data.

5.2.1. Possible mechanisms of lithosphere thinning. To better understand the cause of the lithosphere thinning, it is instructive to review how a lithosphere is generated and thickened. The simplest example would be the origin and accretion of oceanic lithosphere as illustrated in Figure 13. We know that the thickness (T) of oceanic lithosphere is proportional to the square-root of its age (t), i.e., $T_{\text{lithosphere}} \propto t^{1/2}$. This is consistent with, and can be perfectly explained by, the simple half-space cooling model [see Stein & Stein, 1996]. That is, the lithosphere thickening results from thermal contraction [see Davies & Richards, 1992] due to conductive mantle heat loss to the seafloor. Consequently, the lithosphere accretion is achieved through addition of the asthenospheric materials from below, very much like the thickening of the ice sheet on a lake in a cold winter. To put it even simpler, the lithosphere thickening is to “convert” some asthenospheric material into lithospheric material added to the bottom of the lithosphere. With this simple concept in mind, we might ask ourselves the question if we can in fact “convert” bottom part of the lithosphere into asthenosphere, a process just opposite to the accretion of oceanic lithosphere (Fig. 13), which may explain the lithosphere thinning beneath eastern China.

Recognizing that (1) the lithosphere is defined as being “cold”, rigid, strong, highly viscous, whereas the asthenosphere as being “hot”, soft, weak and less viscous, (2) the lithosphere temperature increases with depth, and (3) the lithosphere-asthenosphere boundary is physically not sharp, but gradational, then it should be physically straightforward to “convert” the bottom part of the lithosphere into the convective asthenosphere if some aspects of physical requirement can be satisfied (Fig. 14). This process equals lithosphere thinning. Figure 14 shows schematically that there exist two probable mechanisms that can convert the deep portion of the lithosphere into convective asthenosphere: (1) heating and (2) addition of water. Both can make the lithosphere “soft”, “weak”, “less viscous”, and “easy to flow”. We show below that (1) heating is not an option, but (2) addition of water is readily achieved beneath eastern China.

5.2.2. No need for “hot” mantle plumes. Figure 15 (from Kárasón & van der Hilst, 2000) shows vertical mantle sections of seismic tomography across some subduction zones with “cold” (geophysicists’ interpretations) and fast subducting slabs clearly seen in blue. Many subducting slabs go down to the lower mantle across the 660-km seismic discontinuity (660-D). However, section B-B’ across the Southern Kurile arc and section C-C’ across the Izu-Bonin arc show subducting slabs that fail to penetrate the 660-D, but lie horizontally extending towards the west within the transition zone between 410-D and 660-D. This means that beneath eastern China, there is a layer of “cold” slabs lying horizontally in the transition zone. Recently, Zhao et al. [2004] and Pei et al. [2004] have obtained similar tomographic images beneath eastern China and showed that such flat, within-transition zone, slabs actually distributes widely throughout eastern China. Whether the transition zone slabs are subducted portions of the present-day Pacific plate (locally the Philippine Plate) or the older predecessors of the Pacific plate are unknown, but they must be relatively old, in particular the parts at the western most ends because of the likely slow horizontal motion of the slabs (vs. vertical motion under gravity). It is also possible that the oldest portions of the slabs may no longer be seismically observable because of the thermal equilibration (i.e., no longer in blue). In any case, all these observations suggest that eastern China is unique, and it is such uniqueness that may have led to the exception – the lithospheric thinning. As subduction-zones and subducted/subducting slabs with anomalously high seismic velocity represent coldest regions in the mantle (Fig. 7), the upper mantle beneath eastern China should be cooler (if not hotter) than elsewhere without such horizontally lain “cold” slabs. The thermal gradient in this part of the upper mantle may not be perfectly adiabatic because these cold slabs tend to suck heat from below and above to reach thermal equilibrium. As a result, there should not be excess heat available beneath eastern China that could lead to the lithosphere thinning, the first scenario in Figure 14. Therefore, we can rule out with confidence that the upper mantle beneath eastern China should be thermally “cold” rather than “hot” relative to normal upper mantle elsewhere. Explicitly, the horizontal “cold” slabs in the transition zone beneath eastern China function as a cold, *not* hot, thermal boundary layer that (1) prevents any thermal mantle plumes coming up from the lower mantle, and (2) does not favour the initiation of hot thermal plumes from within-upper mantle. In summary, invoking mantle plumes to explain the Mesozoic/Cenozoic geology in eastern China is not a sound option.

5.2.3. Basal hydration and thinning of the lithosphere beneath eastern China. The ocean crust atop the subducting lithosphere is likely to experience greatest extents of dehydration for arc volcanism (see section 3.2 above), but its dehydration is arguably incomplete because of the formation of hydrous lawsonite, which can contain ~ 11 wt % H₂O, and is stable up to 11 GPa (~ 350 km depth) [see Williams & Hemley, 2001]. Importantly, water transported into the deep mantle through serpentinized peridotites within subducting/subducted slabs is perhaps far more significant than what we thought. Experimental and modeling studies suggest that the presence of water and phase changes of hydrous minerals within the subducting slab explain most of the subduction zone earthquakes from ~ 50 to ~ 650 km [Meade & Jeanloz, 1991; Kirby et al., 1996]. Highly serpentinized peridotites, which are likely the major lithology atop the lithospheric mantle beneath the crust [Dick, 1989; Niu & Hékinian, 1997b; Niu, 1997, 2004], can contain up to 13 wt % water. Because of the thermal structure of the subducting slab, it is less likely that the mantle peridotite section will experience dehydration as efficient as the crustal lithologies atop the slab. Therefore, the water should be largely maintained in the rock as long as the host minerals are stable. For example, serpentine contains up to 13 wt % H₂O, and is stable up to 7 GPa (at ~ 600°C; Ulmer & Trommsdorff, 1995) before transformed to dense hydrous magnesium silicate phases (DHMS: A, B, D-F-G and G) at greater pressures (~ 5 to 50 GPa) [Frost, 1999; Williams & Hemley, 2001] to carry the water (~ 3 to 20 wt%) deep into the mantle [Kuroda & Irifune, 1998]. The thickness of the serpentinized mantle peridotites atop the oceanic lithosphere is unknown, depending on the maximum depth of seawater penetration, but it is predictably thicker when it is generated beneath slow-spreading ridges than beneath fast spreading ridges

because of spreading rate dependent sub-ridge thermal structure [Niu & Hékinian, 1997a]. Perhaps, a layer of up to 2-3 km thick is not an unreasonable estimate for the lithosphere formed at slow-spreading ridges [Muller et al., 1997]. However, the lower zone of the double seismicity within some subducting slabs requires the presence and transformation of hydrous phases at within-slab top-to-bottom depth of ~ 50 km [Seno & Yamanaka, 1996; Dobson et al., 2002; Kerrick, 2002]. It is unlikely that near-ridge serpentinization can reach ~ 50 km depth, but trench-outer rises may be the very sites where seawater penetrates into that depth via extensional fractures, causing trench-outer-rise seismic events [Seno & Yamanaka, 1996], and importantly deep level serpentinization of oceanic lithosphere prior to subduction [Dobson et al., 2002; Kerrick, 2002; Phipps Morgan et al., 2002].

All these hydrous phases, while can be stable up to very high pressures, they tend to decompose and form new and less hydrous phases (e.g., Wadsleyite, < 3.0 wt% H₂O, Ringwoodite, < ~ 2.2 wt% H₂O, and Mg₂SiO₄-spinel is essentially anhydrous) as the temperature increase [Williams & Hemley, 2001]. The horizontal slabs in the transition zone beneath eastern China (Fig. 15) experience isobaric (horizontal movement) heating with time, and will thus lose water accordingly. Assuming the horizontal slabs in the transition zone are ~ 150 km thick (minimum) with a mean density of ~ 3.8 g/cc, and can release 0.5 wt % water (rather conservative), then for a 150-km column per m² map-view dimension, we expect ~ 2850 kg of water release. This is a tremendous amount. If the water so released rises through the upper asthenospheric mantle and reaches the lithosphere (see below for complexities), it will hydrate the basal portion of the existing lithosphere to reduce its rigidity, strength, and viscosity. This is in fact the very process that converts a significant deep portion of the lithosphere into asthenosphere – the thinning of the lithosphere. The newly created asthenosphere becomes part of the convective mantle system (Fig. 14), and can be carried away through asthenospheric flow. This is a physically straightforward mechanism. This mechanism (1) does not require hot mantle plumes that may not exist, (2) does not require lithospheric “delamination”, which describes part of the lithosphere that sinks into the asthenosphere without considering the density difficulty, (3) does not require extension/stretching whose existence and scale in the Mesozoic remains elusive (see below), (4) explains the lithosphere thinning in the entire eastern China, not just the NCC, and thus (5) questions the significance of South China continental subduction as a cause of lithosphere thinning beneath the NCC.

Note that the idea and mechanism presented here (Fig. 14) partly came out of my attempt to understand the geochemical consequence of subduction zone metamorphism of oceanic crust [Niu & O'Hara, 2003] and serpentinized abyssal peridotites [Niu, 2004], but was mostly enlightened by the unique seismic tomography beneath East Asia (Fig. 15). Figure 15, however, represents the present-day snapshot, and does not automatically translate back to the Mesozoic during which, I agree, much of the eastern China lithosphere thinning was taking place [see Menzies et al., 1993; Deng et al., 1998, 2004; Griffin et al., 1998; Zheng, 1999; Wu & Sun, 1999; Xu, 2001; Gao et al., 2002; Zhang et al., 2002; Yan et al., 2003]. It is thus unknown if the subducted Pacific (or the predecessor) plate did move westwards horizontally in the transition zone back in the Mesozoic. However, *absence of evidence is not evidence of absence*. Also, the age of the slabs since the subduction at a given longitude is unknown. Let's consider a E-W tomographic profile at latitude of 30°N with the Izu trench at ~ 142°E and the western tip of the horizontal slab reaching 110°E [Pei et al., 2004] in the transition zone, this would represent a Great Circle distance of ~ 3000 km (neglecting the radial effect). The present-day Pacific plate spreads north-westwards at ~ 100 mm/yr (or 100 km/Myr), and may have a westward component of ~ 80 mm/yr (or 80 km/Myr). It is however unlikely that the horizontal motion of the slabs in the transition zone would be that fast because there is no gravitational pull applied to it as an effective driving force. Also, the mature plate is ~ 90 km thick [Stein & Stein, 1996], but the detected thickness of the slabs within the transition zone is up to 200 km thick [Pei et al., 2004] (a consequence of resistive accumulation). Hence, the maximum speed of westward motion of the within-transition zone slabs is probably < ~ 40 mm/yr (or 40 km/Myr). Assuming this is the case, then the maximum age of the western tip reaching 110°E would be ~ 75 Ma since the subduction. This is a late Cretaceous (Mesozoic) age. Perhaps, nothing older than this age could be seismically detected because of the thermal equilibration of such slabs with the ambient mantle. This suggests that the slabs beneath eastern China during the Cretaceous, if not earlier, may be similar to the present-day situation – horizontally within the transition zone. The inference given here is reasonable, but more evidence is needed to prove so or improve/reject it.

5.3. Mesozoic basaltic volcanism in East China

Xu [2001] summarized nicely that Mesozoic basalts (> 90 Ma) from the NCC is geochemically enriched with $\epsilon_{\text{Nd}} < 0$, which varies from about zero to as low as -20 . This is in fact expected because this reflects an isotopically enriched “ancient” lithospheric (vs. asthenospheric) source – the hydrated, newly “transformed” asthenosphere (Fig. 14). We know for many years that deep portions of the continental lithosphere, despite depleted in terms of major elements, are enriched in incompatible elements as a result of metasomatism over its long histories (isolated from mantle convection). The enriched trace element signatures such as high Rb/Sr, Nd/Sm etc. will produce radiogenic Sr and unradiogenic Nd isotopes with time [see Menzies & Hawkesworth, 1987; O’Reilly & Griffin, 1988; McDonough, 1990, 1992; Hawkesworth et al., 1990; Niu & O’Hara, 2003].

The process may be narrated as follows. The isobaric heating of horizontal slabs in the transition zone (Fig. 15) beneath eastern China releases water as a result of thermal equilibration with time. The water so released reduces both the viscosity and density of the asthenospheric mantle above, forming hydrous melts. The latter, when rising at the base of the lithosphere, further hydrates the lithosphere and “converts” the basal portions into convective asthenosphere (Fig. 14) while producing enriched basaltic melts with $\epsilon_{\text{Nd}} < 0$. Hence, the lithospheric thinning and enriched Mesozoic basaltic volcanism are both the response to the hydration of the basal lithosphere with the water ultimately derived from “ancient” subducted oceanic lithosphere (slabs) lying within the transition zone. The latter is in manner similar to what is seen today (Fig. 15). That “ancient” (pre-Cretaceous) subducted lithosphere/slabs may no longer “exist” in terms of seismic images because of the sufficient time for thermal equilibration. It is noteworthy that the 73 Ma basalts from Shandong has $\epsilon_{\text{Nd}} > 0$, $\sim +7.55$ [Yan et al., 2003], suggesting an isotopically depleted asthenospheric source at this time. This further implies that the major episode of the lithospheric thinning may have ended before this time.

5.4. Cenozoic basaltic volcanism in eastern China

5.4.1. Basic observations and inferences. Xu [2001] also stressed that the Cenozoic basaltic volcanism in the NCC is isotopically depleted with most studied samples having $\epsilon_{\text{Nd}} > 0$, up to $+8$, reflecting a young and asthenospheric mantle source. In fact, such isotopically depleted Cenozoic basalts are not limited to the NCC, but also widespread in Southeast ($\epsilon_{\text{Nd}}(t) \sim +2$ to $+7$) [Zhao, 1990; Tu et al., 1991; Zhang et al., 1998; Zou et al., 2000] and Northeast China [Zhang et al., 1998; Zou et al., 2000]. Despite being isotopically depleted, all these Cenozoic basalts are enriched in incompatible elements and there is an obvious decoupling (lack of correlation) between radiogenic isotopes and incompatible trace elements [see Zhao, 1990; Zou et al., 2000]. The isotopically depleted nature is indeed consistent with a “typical” asthenospheric mantle source, but the decoupled and incompatible element enriched features reflect a more recent low-degree melt “metasomatic” effect, which may very well take place in the depths of the asthenospheric mantle [Niu et al., 1996]. All these observations, along with xenolith studies [see Xu et al., 1998, 2000; Griffin et al., 1998], emphasize:

1. The lithosphere beneath eastern China had already been thinned in the Cenozoic (~ 80 km or less) [Xu, 2001; Yan et al., 2003];
2. The lithosphere thinning is not limited to the NCC, but the entire eastern China, which is further supported by other observations (Fig. 12a-d). Therefore, there is no logical reason why the NCC should be isolated when discussing lithosphere thinning in eastern China.
3. A within-asthenosphere low-degree melt “infiltration” or “metasomatism” is required to explained the incompatible element enriched, yet isotopically depleted (e.g., $\epsilon_{\text{Nd}} > 0$), nature of the Cenozoic basaltic volcanism.

All these constraints may help us to understand the physical mechanisms of Cenozoic basaltic magma genesis in eastern China. One or more of the first three mechanisms (Section 2.3 and Fig. 2) may apply: (1) decompression, (2) heating, and (3) addition of water (plus other volatiles and alkalis). Considering the discussion in Section 5.2.3 in association with Figure 15, the “cold” horizontal slabs within the transition zone would prevent thermal plumes rising from below, and will not favour the development of hot thermal plumes in the upper mantle either. Therefore, “mantle plumes” in the traditional sense are not a logic option for explaining the Cenozoic volcanisms.

As water released from the slabs in the transition zone will certainly lower the solidus of the overlying asthenospheric mantle and induce melting at some depths. The situation differs from subduction related mantle wedge melting (Fig. 7) in fundamental ways. The water released from the slabs in the transition zone (Fig. 15) will have rather complex histories through forming very low-degree hydrous melts that may ‘infiltrate’ and ‘metasomatize’ the overlying asthenospheric mantle [Niu et al., 1996]. The asthenospheric mantle may be so modified with lowered solidus, but it remains largely under “subsolidus” condition (note 410 km depth of the top of the transition zone). This is suggested by many mantle tomographic models that the low-velocity layer beneath eastern China extends no more than ~ 200 km deep (see Harvard University Website¹). It follows that the depth of magma generation beneath eastern China is quite shallow, perhaps, within ~ 50 km beneath the already thinned lithosphere, which is in fact consistent with petrologic and geochemical studies of basalts and associated spinel lherzolite xenoliths [Griffin et al., 1998; Xu et al., 1998; Zhang et al., 1998].

Passive upwelling and related decompression melting have been invoked by some researchers, but why should there be passive upwelling beneath eastern China? Beneath ocean ridges, passive upwelling is driven by the plate separation (Fig. 6), but the situation beneath eastern China does not at all in any way resemble ocean ridge scenario. Continental rifting as a result of lithospheric extension would have similar effects as beneath ocean ridges except that the “separation rate” is essentially zero. This suggests that the extension or rifting beneath eastern China cannot in any physically straightforward way to induce significant passive asthenospheric upwelling.

Cenozoic lithospheric extension, rifting, rift systems etc. both as terminologies and as concepts appear numerous in the literature concerning within-continent volcanism in general and eastern China volcanism in particular. Extension and rifting can indeed cause localized mantle melting and associated volcanism. However, there is no explicit evidence for significant Mesozoic/Cenozoic extension or rifting in eastern China. Of course, **absence of evidence is not evidence of absence**, but the commonly invoked strong evidence for eastern China extension is the existence of a number of basins. Indeed, lithospheric extension and rifting can cause lithosphere thinning and basin formation, but basin formation does not necessarily require lithosphere extension and rifting. From the isostatic view point, basins could reflect high density materials beneath – maybe shallow or perhaps quite deep, depending on whether isostatic equilibrium may have been attained. One possibility is that basins in eastern China may be genetically associated with the “cold” and dense slabs in the transition zone (410-660 km; Fig. 15). If we choose 660-km as the compensation depth, then the ~ 150 km thick dense slab materials in the transition zone would allow less dense materials at shallow depths or basin formation on the surface. For simplicity, let’s assume the density in the upper mantle (also within the crust) is more-or-less uniform, and the mean density of the transition-zone slabs (Fig. 16) (say, 3.876 g/cc) is about 2% denser than the mean density of the transition-zone elsewhere without such slabs (say, 3.800 g/cc), then the isostasy requires a surface depression of ~ 2% of the 150 km, which is ~ 3 km deeper in regions with a dense transition-zone slabs than elsewhere. This suggests that the presence or absence of subducted slabs within the transition zone can be a possible cause of the basins in eastern China. Note that this simple analysis only suggests the possibility of the genetic link between the transition-zone dense slabs and the surface basins, but does not prove that it is the case. The point is that basins are not necessarily the evidence for extension or rifting, and to use basins of unknown origin as evidence for extension or rifting is dangerous.

The above analysis suggests that the widely perceived plume models, rifting models or slab dehydration scenario could all be possible causes for the Cenozoic volcanism in eastern China, but none stands well without problems. Below I discuss another possibility that (1) requires asthenospheric flow, (2) the flow is driven by the pressure gradient, and (3) the melting is ultimately accomplished by decompression of the flowing asthenospheric material.

5.4.2. Pressure gradient drives asthenospheric flow. Asthenospheric flow in the context of mantle convection is familiar to everyone. That flow occurs as a result of pressure gradient is also familiar to all. However, when constructing conceptual models, many of us do not seem to consider the necessity of pressure gradient – the very driving force for flows. For example, the concept of “passive upwelling” that takes place beneath seafloor spreading centers has been invoked by some to explain intra-plate volcanism without questioning that such “passive upwelling” is physically difficult because of lacking pressure gradient. Flows cannot begin or continue if the “flowing material” has ultimately nowhere to go! Oceanic lithosphere can flow (subduct) into the asthenosphere because it is dense and driven by its negative

¹ <http://cfavvcs5.cfa.harvard.edu/lana/rem/slice.htm>

buoyancy (one type of pressure gradient). Mantle plumes can rise because the plume materials are thought to be hot and less dense than the surrounding mantle, which is driven by the positive buoyancy. Therefore, it is necessary to provide an example of asthenospheric mantle flow and its mechanism before we see how asthenospheric flow may explain the Cenozoic volcanism in eastern China.

Figure 16 (after Niu & Hékinian, 2004) shows that there is a strong asthenospheric material need beneath ocean ridges and the material need increases with increasing spreading rate. This is because > 99% mass of ocean crust is formed at ocean ridges (excluding the addition of oceanic plateaus, and ocean islands), which requires asthenospheric material flow toward ridges. Also, about ~ 50% total thickness of oceanic lithosphere is formed in the first ~ 17-18 Myr (note the oldest oceanic lithosphere is ~ 200 Ma before subduction in the western Pacific), which also demands asthenospheric material flow toward ridges. Therefore, the asthenosphere beneath ocean ridges represents the regions of the lowest pressure in the entire asthenospheric mantle. There is thus a ridge suction force that drives asthenospheric material to flow toward ridges – the concept of pressure gradient as a flow-driving force. The question then is whether the needed material comes from great depths or laterally. As the viscosity increases exponentially with depth (Fig. 16b), the asthenospheric material that can easily move would lie within the seismic low velocity zone (LVZ, also standing for Low Viscosity Zone) right beneath the lithosphere atop the asthenosphere. It follows that vertical flow from great depths is physically difficult because of the elevated viscosity, but horizontal flow within the LVZ becomes easy because of the low viscosity. Hence, it is physically straightforward that the material needs at ocean ridges come laterally through LVZ because of the ridge suction and because of the low viscosity within the LVZ. Such ridge-ward flow is seemingly difficult as it moves in direction against the overlying oceanic plate (Fig. 16c), and it requires significant decoupling and shearing at the interface between the growing lithosphere (Fig. 13) and the asthenosphere. Such asthenospheric flow may be inconsistent with the mainstream view, but mass conservation requires that this must be the case [Niu & Hékinian, 2004]. The point here is that asthenospheric mantle flow is necessarily driven by pressure gradient.

5.4.3. Decompression of flowing asthenospheric mantle for Cenozoic volcanism in eastern China.

Figure 7 shows that the mantle wedge overlying the subduction zone must be replenished with new asthenospheric materials, as required physically by the subduction-induced corner flow, and chemically by the observed temporal IAB compositional variation. This indicates that mantle wedges represent another type of low pressure regions in the asthenosphere that suck asthenospheric materials to move laterally towards subducting slabs. Perhaps, we may call it “**wedge suction**”. While **wedge suction** may not be as strong as ridge suction, it necessarily drives asthenospheric flow (Fig. 7). Western Pacific subduction zones are arguably the most dynamic subduction systems on Earth in terms of the rate of oceanic lithosphere subduction and both frequency and volumes of arc volcanic eruptions. These require asthenospheric flow from the west beneath Chinese continent to the western Pacific subduction systems (see Fig. 7 for concept).

Figure 17 illustrates the concept. Western Pacific **wedge suction** draws asthenospheric material from the west, i.e., beneath eastern China. The eastward asthenospheric flow from beneath eastern China, in turn, requires material replenishment further from the west again. Because the thickened lithosphere beneath the plateaus in western China (see Fig. 12 and discussion above) contrasts the thinned lithosphere beneath the hilly plains in eastern China, the asthenospheric replenishment from beneath the plateaus to beneath eastern China must experience an upwelling or decompression (Fig. 17). It is this decompression (Fig. 2) that causes the flowing asthenosphere to melt and to give rise to the Cenozoic volcanism in eastern China. As the source materials are asthenospheric, the basalts are expected to be isotopically depleted, i.e., $\epsilon_{Nd} > 0$, yet enriched in alkalis and incompatible elements [see Zhao, 1990; Tu et al., 1991; Zhang et al., 1998; Zou et al., 2000]. The isotopic and trace element decoupling must result from more recent asthenospheric metasomatism of low-degree melts genetically associated with dehydration of subducted oceanic lithosphere lying within the transition zone since the Mesozoic.

5.4.4. Accretion of young lithosphere. Recent studies suggest that there seem to be new lithosphere that has been formed or forming since the major lithospheric thinning in the Mesozoic beneath eastern China [see Menzies & Xu, 1998; Xu et al., 2000; Gao et al., 2002; Xu et al., 2003; Zhang et al., 2004]. This is important recognition, but the process of such accretion should be physically straightforward. Cooling of the mantle through conductive heat loss to the surface is natural and is driven by the thermal gradient. The mantle cooling would be faster if the lithosphere is thin because the thermal gradient would be steep, i.e., temperature drops more rapidly with decreasing depth. The most relevant question would be the rate of new material addition from below to the lithosphere (see Fig. 13). The rule of $T_{\text{lithosphere}} \propto t^{1/2}$ (see Section 5.2.1)

should apply because of conductive heat loss, but the actual rate also depends on possible heat supply carried by the asthenosphere as it moves and replenishes from west to the east (see Section 5.4.3). The message is that the lithosphere thickening again should not be a surprise, but a natural consequence of mantle heat loss. Also, the statement that “new lithosphere replaces old lithosphere” is acceptable only if it is meant to stress a temporal sequence, but it should be abandoned if it is meant to emphasize the processes or physical mechanism because it is misleading.

6. Summary

1. This paper reviews some basic concepts in basaltic magma generation and evolution. Compression as a probable mechanism for the genesis of some granitic magmas is highlighted because of the likelihood of the mechanism and because of its poor realization.
2. Magma evolution in reality can be quite complex, but the basic cooling-induced “crystallization-differentiation” process needs better understanding before invoking some elusive possibilities.
3. MgO wt % in basaltic liquids (quenched glasses or aphyric basalts) is a good indicator of the liquidus temperature, more precise than liquidus phase based geothermometers. The almost linearly positive $T_{\text{liquidus}}\text{-MgO}$ relationship derived from anhydrous experimental data works well for MORB, and perhaps also for most OIB. The $T_{\text{liquidus}}\text{-MgO}$ relationship also exists in hydrous melts such as IAB with > 2-3 wt % H₂O, but the wet T_{liquidus} is ~ 40°C lower than for anhydrous systems at a given MgO value.
4. Anhydrous basaltic melts have the fractionation path of **dunite** → **troctolite** → **gabbro**, whereas hydrous basaltic melts have the path of **dunite** → **wehrlite** → **gabbro**. Whether troctolite or wehrlite cumulate lithologies may be found in an ophiolitic sequence is effective to tell mid-ocean ridge settings (e.g., dry, troctolite) from subduction or supra-subduction settings (e.g., wet, wehrlite).
5. Dunite, troctolite, wehrlite, gabbro, gabbronorite etc. are cumulate. They form primarily as a result of crystal accumulation during basaltic magma cooling. They, as rocks, have never been in the state of being melt or magma. Therefore, they are **NOT** intrusive equivalent of basalts. Basalts, in particular aphyric ones, are cooled, evolved, and quenched melts. Hence, geochemical discrimination diagrams derived from basalts cannot be used so for cumulate rocks such as gabbros.
6. A hypothesis for the Mesozoic lithosphere thinning and Mesozoic/Cenozoic volcanism in eastern China is presented. The hypothesis is rather simplistic, but is consistent with important observations. The correctness of the hypothesis needs testing. Alternative interpretations are welcome, so long as they are consistent with observations and simple physics.
7. While the Mesozoic/Cenozoic volcanism in eastern China can be considered as “intra-plate” volcanism, it is in fact a special consequence of plate tectonics. The Mesozoic lithosphere thinning in eastern China is best explained by a process that “transformed” the lower portion of the lithosphere into convective asthenosphere. This “unusual” process is achieved by basal hydration of the lithosphere. The water that did so may come from dehydration of subducted Pacific (or predecessor) oceanic lithosphere that is presently lying horizontally in the transition zone beneath eastern Chinese continent as detected by seismic tomographic models.
8. Isotopically enriched Mesozoic basalts ($\epsilon_{\text{Nd}} < 0$) may be genetically associated with the lithosphere thinning of the kind because the basaltic source materials are ancient lithospheric materials being converted into the asthenosphere.
9. The Mesozoic lithosphere thinning may not be limited to the east of the Great Gradient Line (**GGL**), but may extend more to the west. The older transition-zone slabs may not be seismically detected because they are old enough and may have reached thermal equilibrium with their ambience. The **GGL** is likely a young feature as a result of Indian-Eurasian collision.
10. The **GGL** also marks clearly the present-day steep gradient of lithosphere thickness variation from probably > 150-200 km beneath the plateaus in the west to the thin, probably < 80 km beneath eastern China. The “remote” western Pacific subduction systems induce asthenospheric flow (**wedge suction**) from beneath eastern China towards the subduction zones, which, in turn, requires asthenospheric material replenishment from beneath the plateaus to the eastern China. As a result, such eastward

asthenospheric flow experiences an upwelling and decompression, which causes asthenospheric partial melting and Cenozoic eastern China basaltic volcanism, giving asthenospheric isotopic source signature, e.g., $\epsilon_{\text{Nd}} > 0$. Such volcanism may have already begun at the late Cretaceous.

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Figure 1. Illustration of experimental data on fertile mantle solidus and subsolidus lherzolite stability fields of plagioclase (shallow), spinel (deep) and garnet (deepest) in P-T space. The shaded band above the solidus represents P-T conditions under which mantle melts of varying compositions could be generated based on static experimental studies. K – komatiite melt, P – picrite melt, AOB – alkali olivine basalt, OTh – olivine tholeiites, and QTh – quartz tholeiite. (After Wyllie [1992])

图 1. 地幔固相线及地幔二辉杆栏岩在不同温压条件下的稳定范围：斜长石二辉杆栏岩(浅)，尖晶石二辉杆栏岩(深) 及石榴石二辉杆栏岩(更深)。固相线之上的阴影部分示意理想状况下各种原始岩浆的形成条：K-科马提岩，P-苦杆岩，AOB-硷性杆栏玄武岩，OTh-杆栏拉斑玄武岩，及 QTh-石英拉斑玄武岩 (引自 Wyllie [1992])。

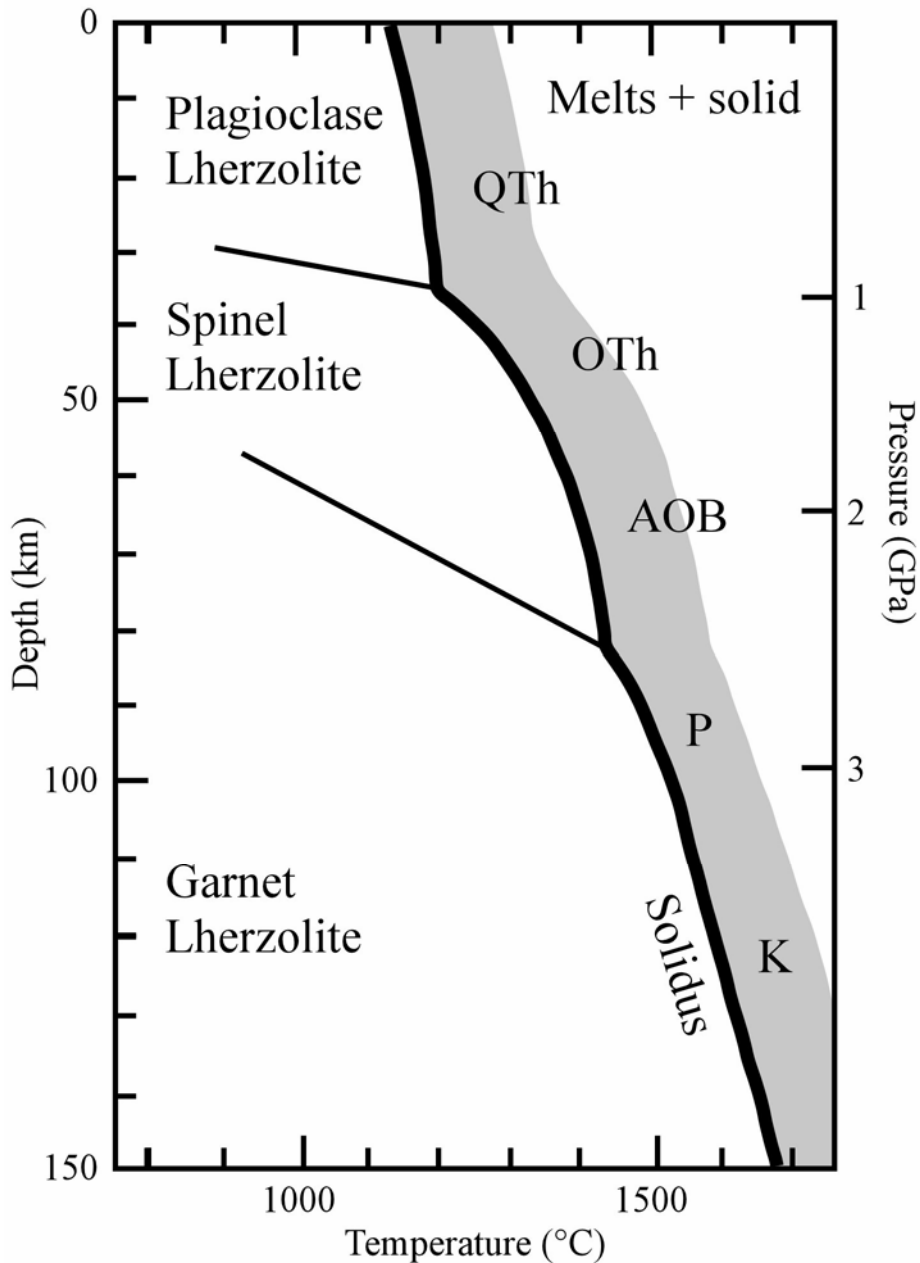


Figure 2. Schematic representations of 3 endmember mechanisms of magma generation in P-T space. See text for details.

图 2. 温压图示意玄武岩浆起源的三个端员机制。

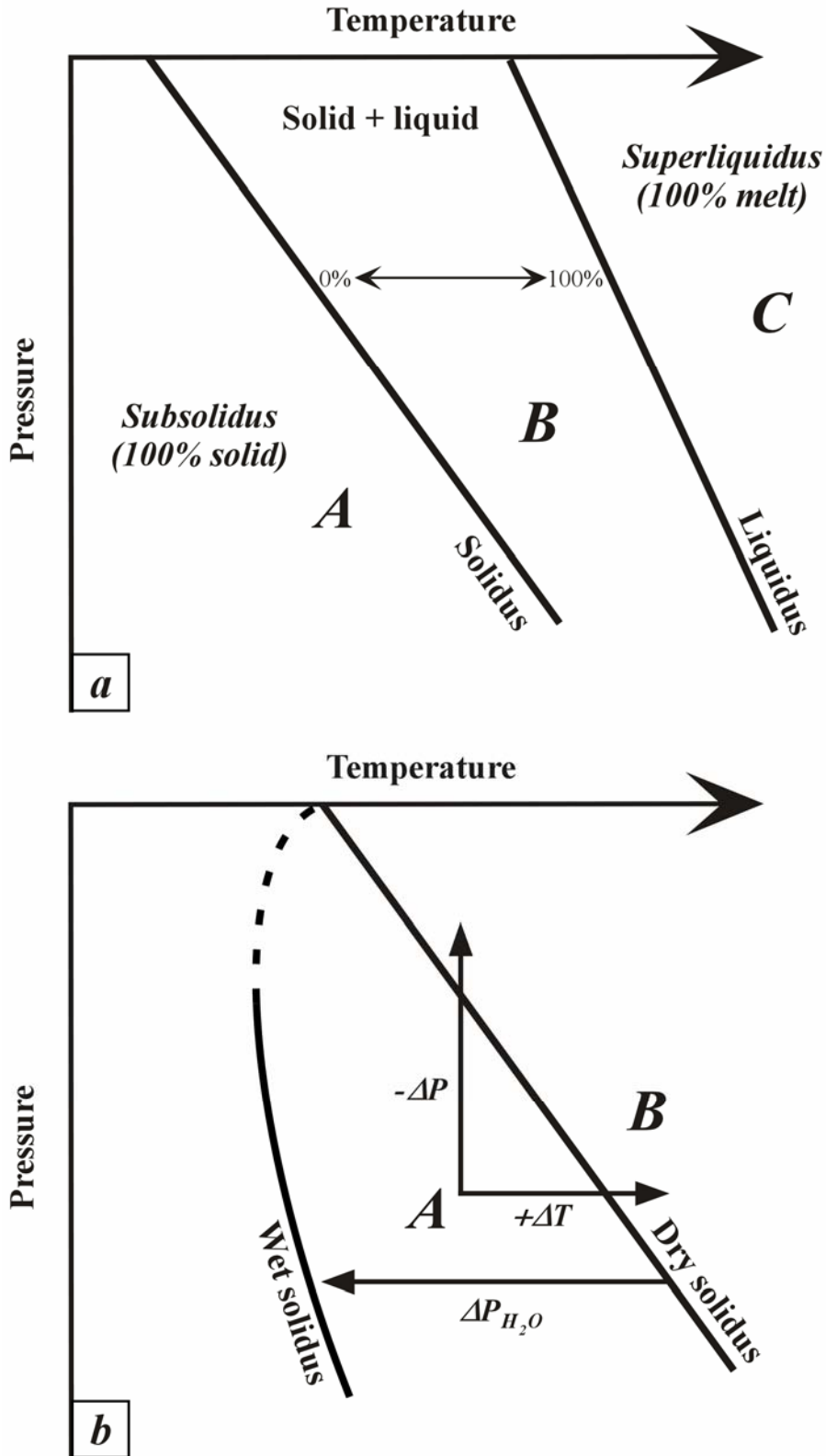


Figure 3. Quantitative illustration of dry and wet solidus of mantle peridotite systems (a) (after Wyllie [1978, 1979]; Green [1991] Takahashi & Kushiro [1983]); and crustal materials (b) (after Stern et al. [1975], Yoder [1975]).

图 3. (a) 地幔二辉杆岩在无水(或微量水)及 H_2O-CO_2 饱和时固相线在温压条件下的形态和位置 (引自 Wyllie [1978, 1979], Green [1991], Takahashi & Kushiro [1983])。 (b) 陆壳岩(玄武质-花岗质岩包括陆源碎屑岩) 水饱和时固相线在温压条件下的形态和位置 (引自 Stern *et al.* [1975], Yoder [1975])。

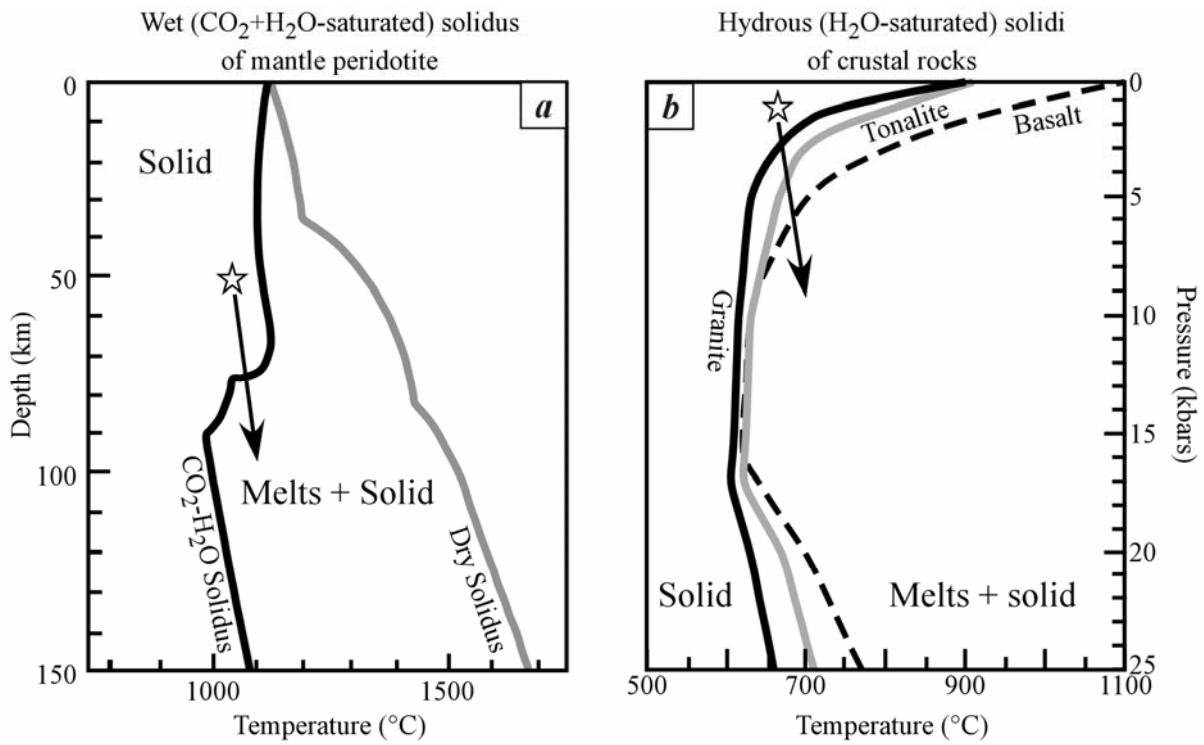


Figure 4. Quantitative illustration of *dry* mantle solidus (after McKenzie and Bickle [1988]), and adiabatic thermal gradients for mantle upwelling beneath ocean ridges and assumed mantle plumes. See text for details.

图 4. 地幔二辉杆栏岩无水(或微量水)固相线 (引自 McKenzie and Bickle [1988]) 及洋脊地幔和地幔柱地幔上涌绝热梯度。

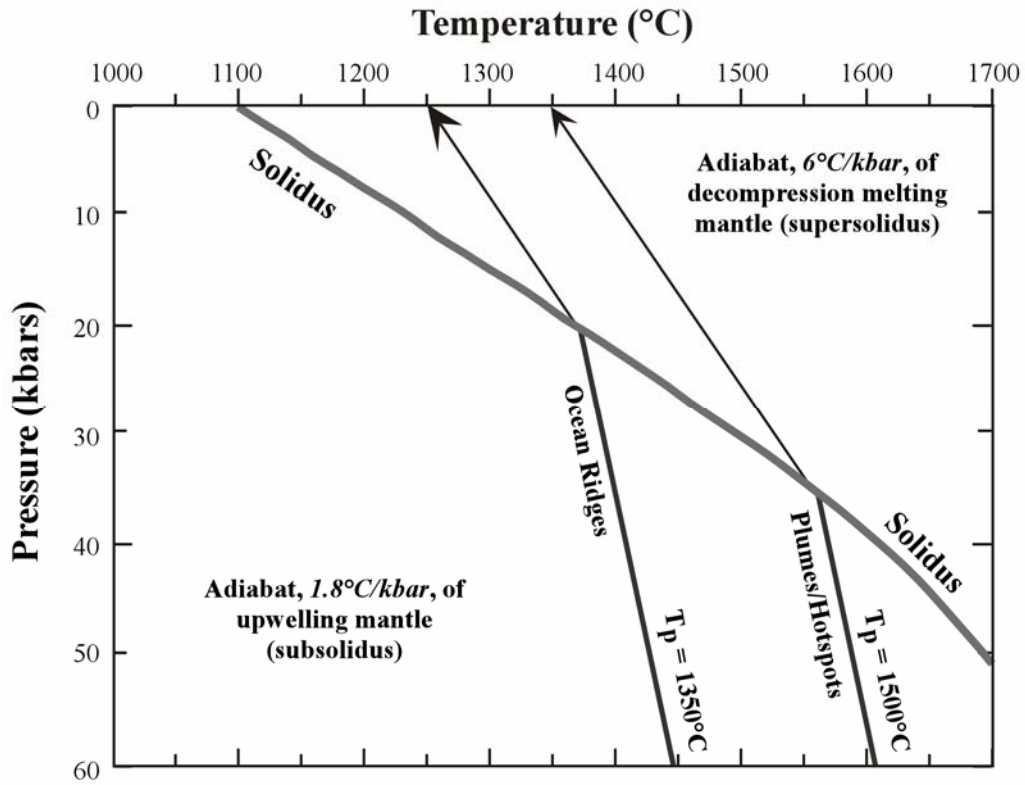


Figure 5. Model compositions (selected oxides SiO₂, Al₂O₃, FeO, MgO and CaO) of mantle melts by decompression melting (left columns) and corresponding melting residues (right columns) to illustrate that compositions of primary mantle melts, so are the residues, vary as a function of initial melting depths (e.g., 25, 20 and 15 kbars) and the extent of melting. The decompression melting models use the parameterization of Niu [1997] and an ideal fertile lherzolite composition given in Niu [1997]. The decompression melting is arbitrarily stopped at ~ 8 kbars for all the three paths chosen. Note that all these chosen oxides in primary mantle melts vary with extent of decompression melting, but SiO₂, FeO and MgO also strongly depend on initial pressures of melting.

图 5. 模型地幔二辉杆岩(无水或微量水) 减压分熔产物的化学组成(如 SiO₂, Al₂O₃, FeO, MgO, CaO) 沿 3 条绝热梯度分别与固相线相交于 25, 20 及 15 千巴。左边指原始分熔岩浆组成, 右边指分熔残余组成(Niu [1997] 模型计算结果)。

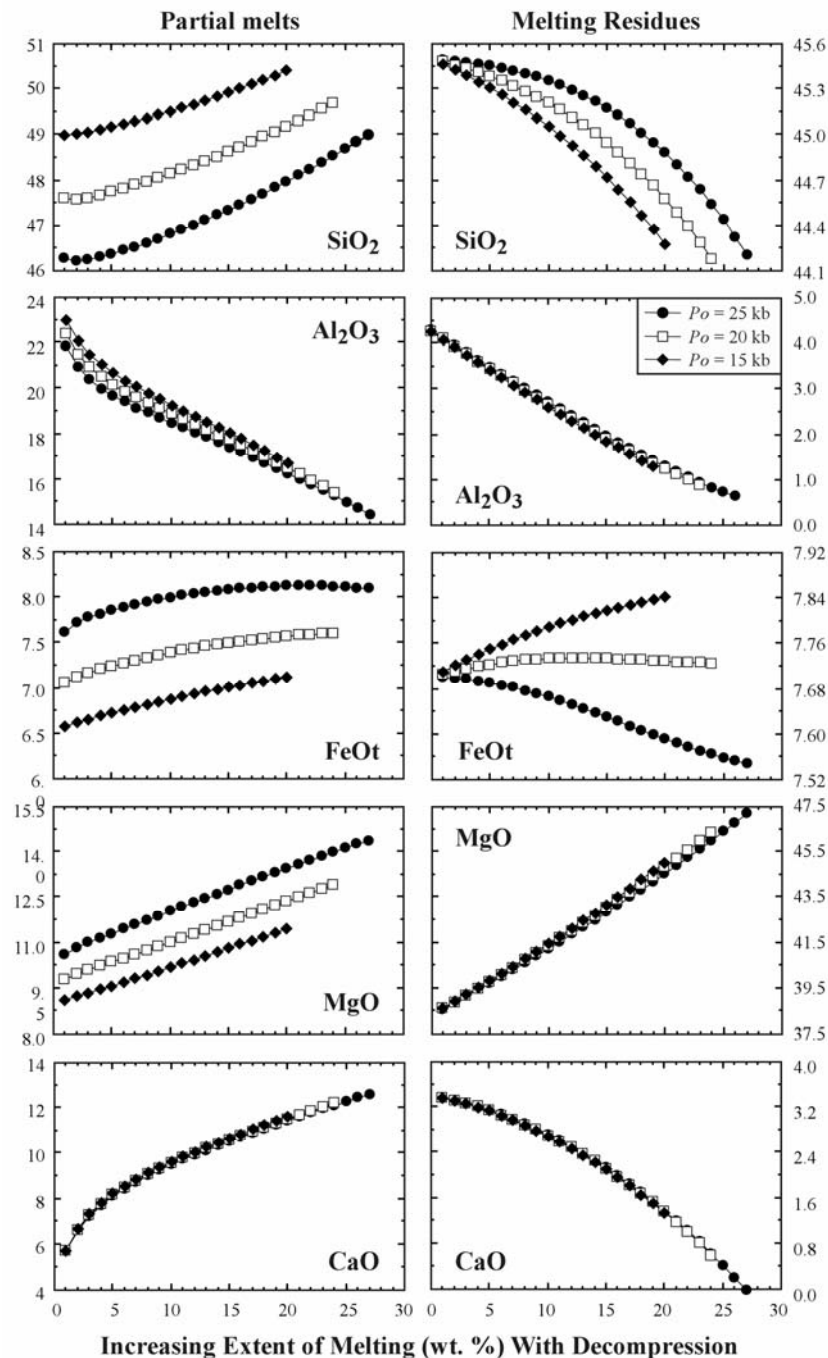


Figure 6. (a) Cartoon showing plate separation-induced passive upwelling of the asthenospheric mantle and the resultant decompression melting beneath ocean ridges. (b) Schematic illustration of the processes described in (a) in P-T space. Passive upwelling brings hot mantle material to rise adiabatically and melt when encountering the solidus. Melting continues with upwelling (depth range 1) until the upwelling mantle reaches the cold thermal boundary layer (TBL, depth range 2) as a result of conductive heat loss to the surface [Niu, 1997; Niu & Hékinian, 1997a].

图 6. (a) 板块扩张引起的洋脊软流圈地幔被动上涌及其减压分熔。(b) 温压条件下(a) 的示意图解。(1) 减压分熔深度范围；(2) 近洋底热损失而形成的冷的“热边界层” (见 Niu, [1997] , Niu & Hékinian [1997]) 。

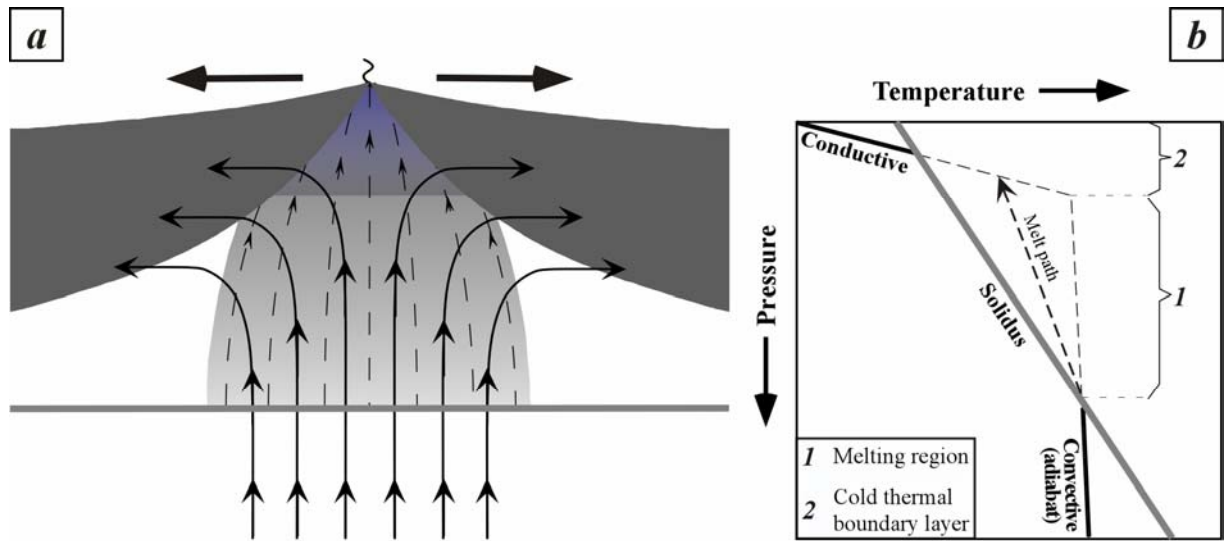


Figure 7. Cartoon showing subducting slab dehydration induced mantle melting that produce island arc basaltic (IAB) magmas (see Fig. 2b). Note that subducting slab induced mantle wedge “corner flow” (solid lines with arrows) and replenishment of new asthenospheric material due to “wedge suction” (see text for details) as indicated by the large hollow arrow.

图 7. 示意俯冲板块脱水引起地幔楔杆岩分熔过程及岛弧玄武岩浆的形成(见图 2 b)。注意俯冲板块引起地幔楔软流层的“拐角”流动(带箭头实线)。该“拐角”流(亦即,“地幔楔吸力”)会引起软流层物质的侧向补给(图左侧空心大箭头)。

Arc volcanism and “characteristic” asthenospheric mantle flow

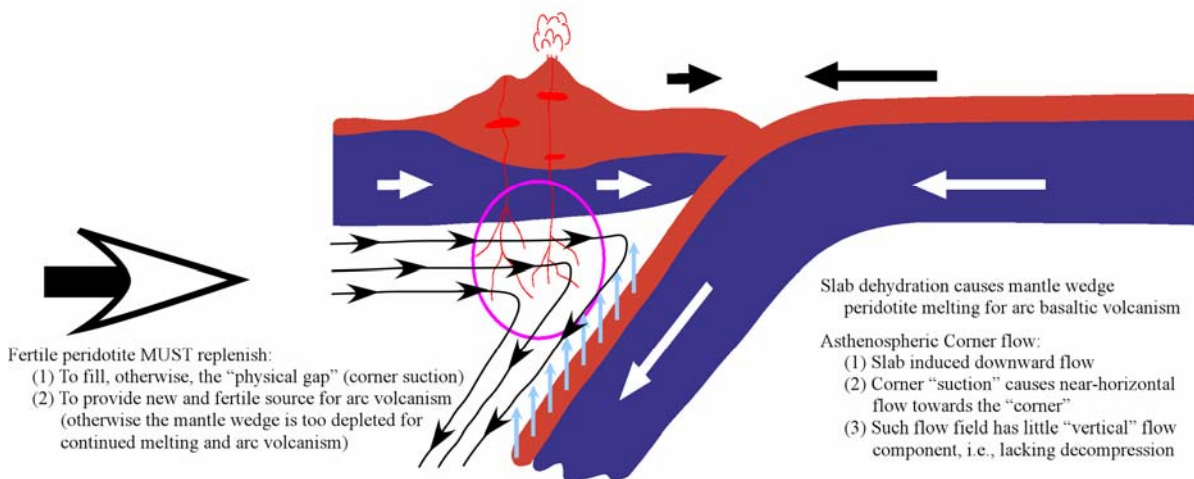


Figure 8. (a) Illustration of simple relationship between liquidus temperature ($T_{liquidus}$) and MgO content of “dry” basaltic magmas (e.g., MORB melts). (b) $Mg^{\#}$ of “dry” basaltic melts exhibits an equilibrium relationship with the liquidus olivine. This is after Niu et al. [2002b] derived from various experimental data [Bender et al., 1978; Walker et al., 1979; Langmuir & Hanson, 1981; Nielsen & Dungan, 1983; Weaver & Langmuir, 1990; Grove et al., 1992].

图 8. (a) 示意无水(或少水)玄武岩浆(熔体) MgO 与液相线温度($T_{liquidus}$) 的线性(或近线性) 关系。(b) 玄武岩浆(熔体) 中 $Mg^{\#}$, 与之平衡的杆栏石成分和液相线温度($T_{liquidus}$) 的关系。该图源自 Niu *et al.* [2002b] (其原始实验数据来源: Bender et al. [1978], Walker et al. [1979], Langmuir & Hanson [1981], Nielsen & Dungan [1983], Weaver & Langmuir [1990], Grove et al. [1992])。

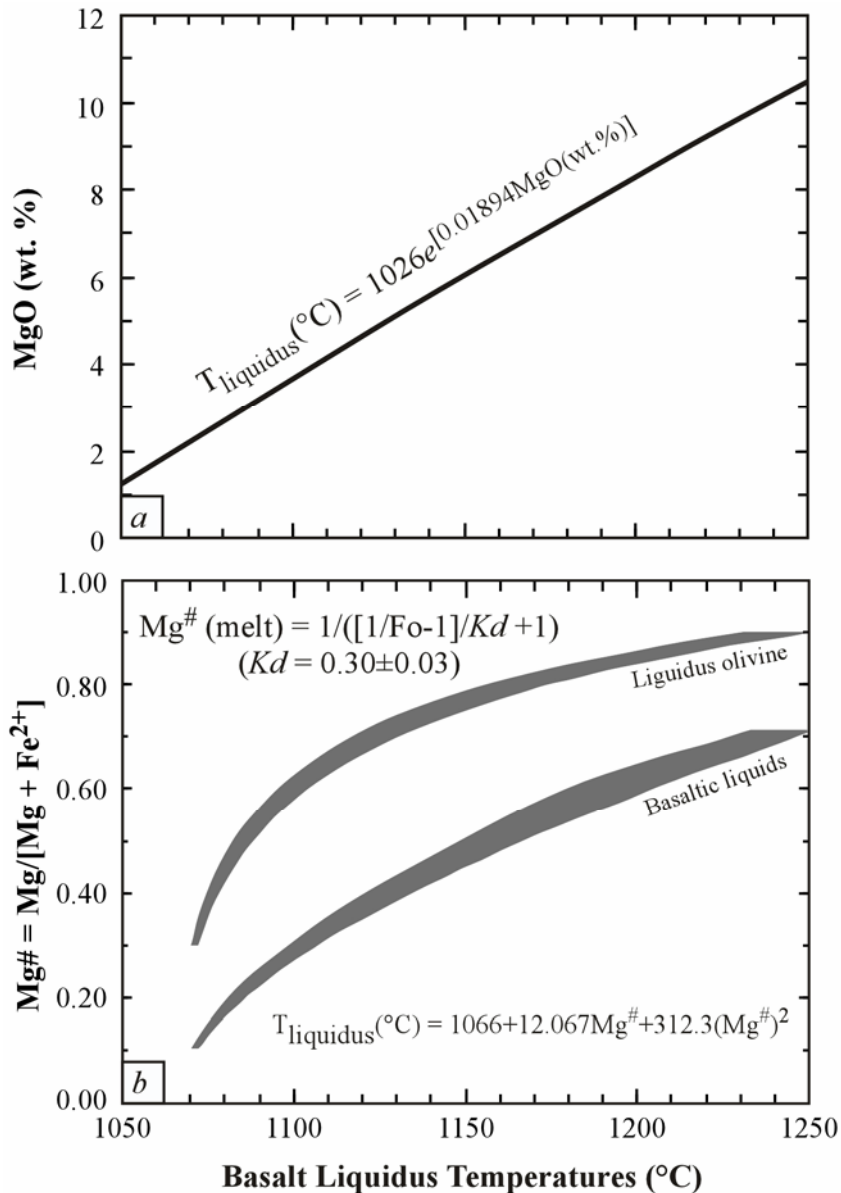


Figure 9. Liquid lines of descents (LLDs) for anhydrous basaltic systems derived from > 400 MORB glasses from the East Pacific Rise axis and nearby seamounts. The bottom two panels are model results obtained using the algorithms of Weaver & Langmuir [1990] in combination with the petrography.

图 9. 无水(或少水) 玄武岩浆(熔体) 演化线。该熔体演化线代表 >400 东太平洋洋脊玄武岩玻璃的平均组成。底部两副图系计算(Weaver & Langmuir [1990])和岩相学观察所得。

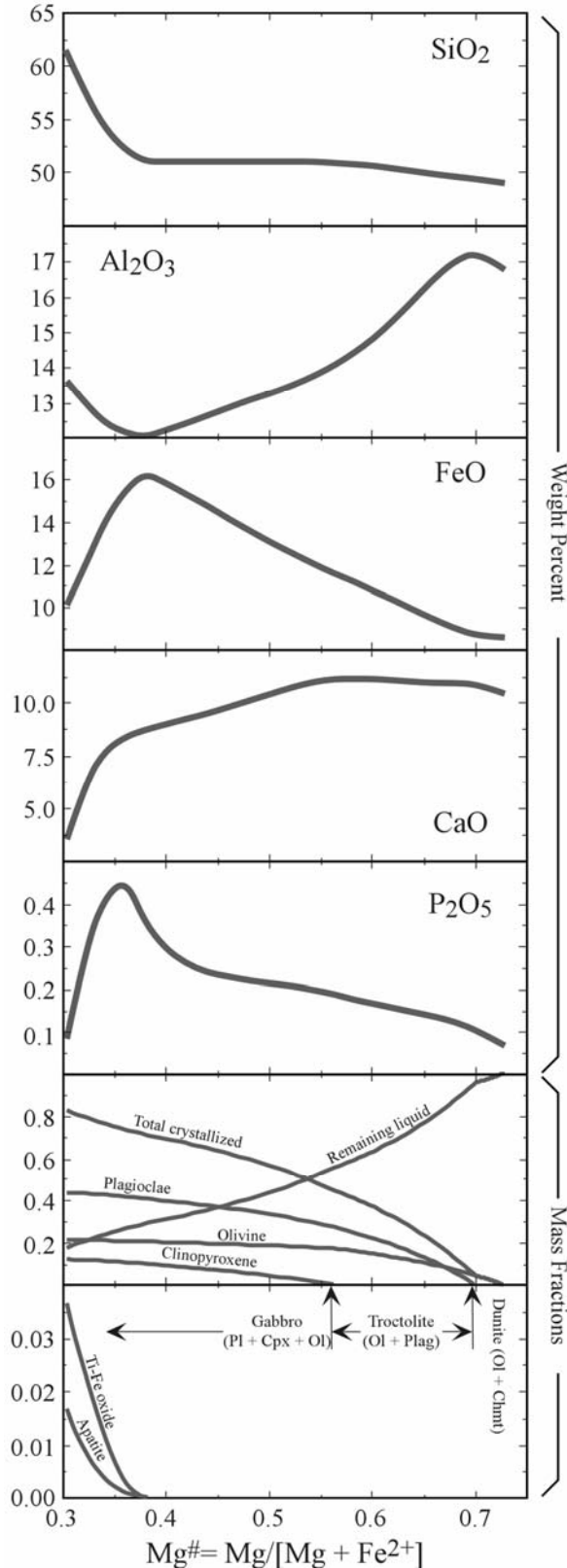


Figure 10. Hydrated peridotite-basalt melting experimental data [Gaetani & Grove, 1998] showing melt compositions in terms of H₂O wt % (a) and MgO wt % (b) as a function of liquidus temperature ($T_{liquidus}$). For comparison, the “dry” MgO- $T_{liquidus}$ relationship (solid line in b as seen in Fig. 8a) is also plotted. Despite some complications, the MgO- $T_{liquidus}$ relationship for hydrated basaltic system resembles that of the dry basaltic system, but is in general about 40°C lower for a given MgO (wt %).

图 10. 水饱和及过饱和(多水, 湿)玄武岩浆(熔体)- 杆栏岩实验数据[Gaetani & Grove, 1998]。(a) 多水(湿)H₂O- $T_{liquidus}$ 关系。(b) 多水 MgO- $T_{liquidus}$ 关系; 少水(干)MgO- $T_{liquidus}$ 关系(图 8a) 也示意做比较。尽管比较复杂, 无水(或少水) MgO- $T_{liquidus}$ 关系仍然适用于水饱和及过饱和玄武岩熔体, 只是在给定 MgO 值, 含水 $T_{liquidus}$ 约 40°C 底于无水 $T_{liquidus}$ 。

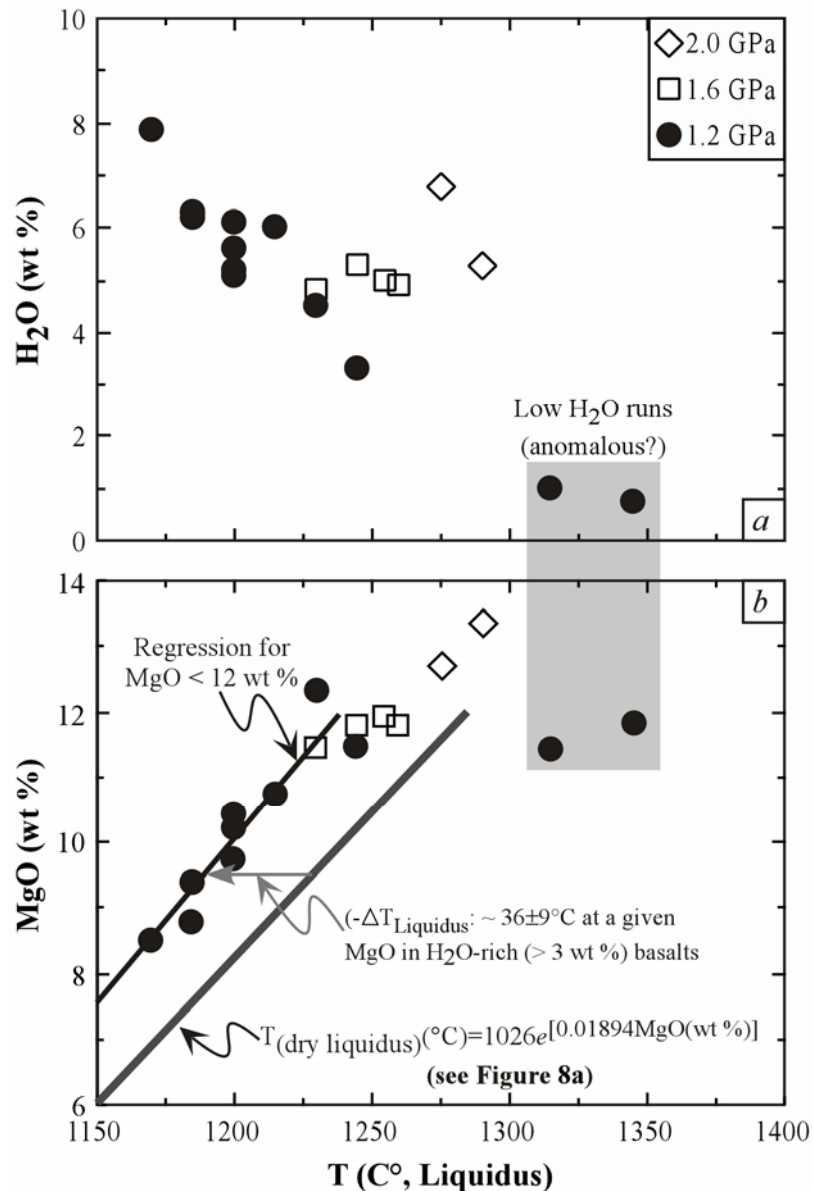


Figure 11. Illustration of two distinct liquid lines of descent between anhydrous melts (MORB) and hydrous melts (IAB). The simple Di-An binary phase diagram of Yoder [1965] is used to illustrate the principle, which also applies to multicomponent systems [Gaetani et al., 1993].

图 11. 示意少水 (如洋脊玄武岩浆) 和多水 (如岛弧玄武岩浆) 两个截然不同的岩浆演化线或矿物结晶顺序。简单的透辉石-钙长石二元岩浆体系[Yoder, 1965] 和天然多组分体系[Gaetani et al., 1993]均表明少水岩浆中斜长石结晶早于单斜辉石, 而多水岩浆中单斜辉石结晶早于斜长石。

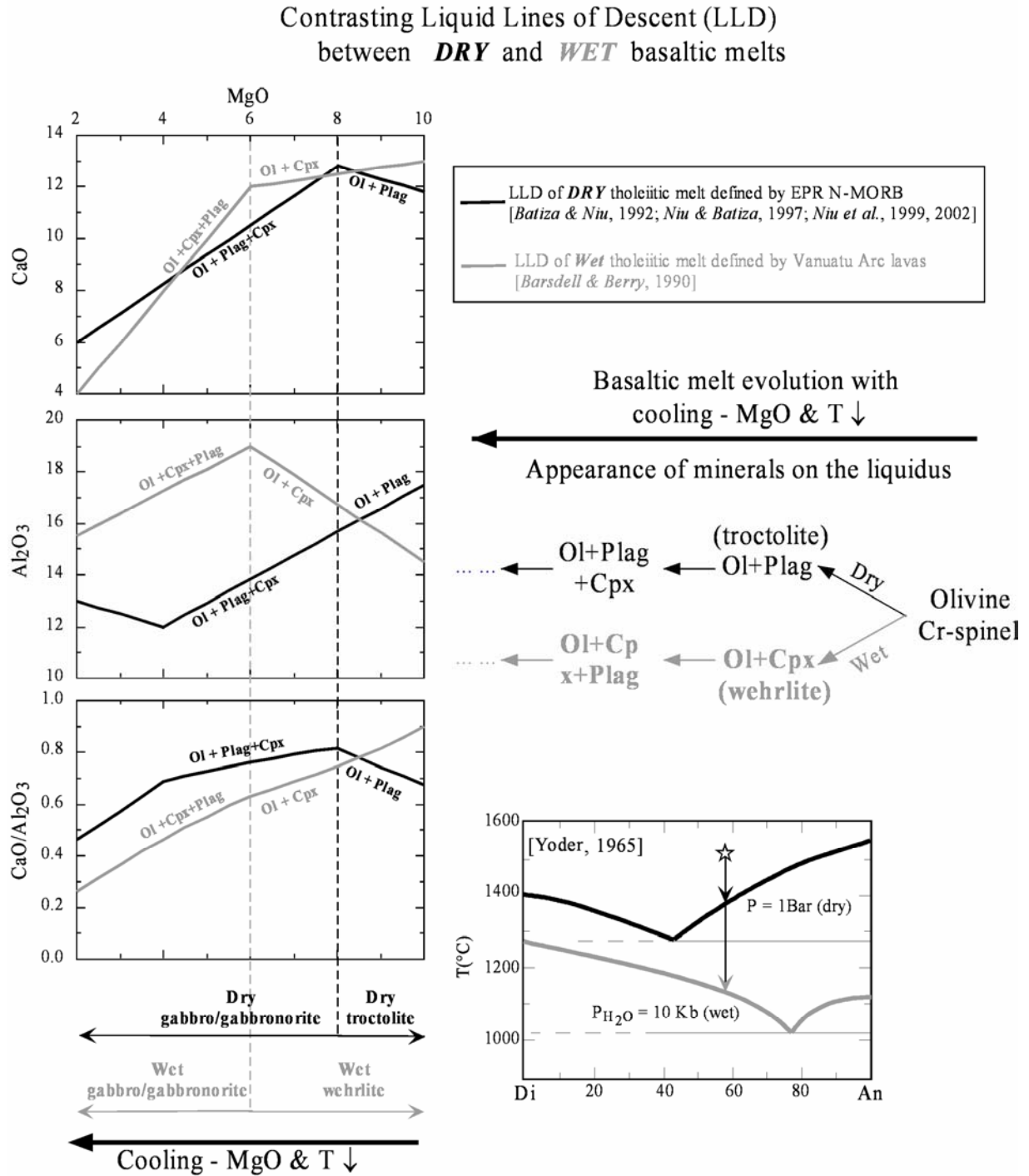


Figure 12. (a) Portion of the world topography map [Smith & Sandwell, 1997] showing a sharp altitude contrast in continental China as indicated by the arrowed dash line. (b) Harvard 1998 seismic tomography at depths of 100 km and 150 km to show that the surface altitude contrast in (a) coincides with the steep gradient in seismic velocity [Ekström & Dziewonski, 1998] (on line: <http://cfauvcs5.cfa.harvard.edu/lana/rem/slice.htm>). (c) Bouguer gravity anomaly on continental China [Tang, 1996] also shows sharp contrast that coincides with the steep gradients in (a) and (b). (d) Crustal thickness variation on continental China [Li & Mooney, 2001] (on line: <http://quake.wr.usgs.gov/research/structure/CrustalStructure/china/index.html>).

图 12. (a) 全球地形图[Smith & Sandwell, 1997]的一部分, 其“东-西梯度界”(红色箭头虚线)显示中国大陆东-西海拔的截然差异。(b) 全球 100-150 公里深度地震波速异常分布[Ekström & Dziewonski, 1998]显示中国大陆地幔地震波速异常梯度与海拔“东-西梯度界”吻合 (on line: <http://cfauvcs5.cfa.harvard.edu/lana/rem/slice.htm>)。 (c) 中国大陆东-西重力布格异常梯度也与海拔“东-西梯度界”吻合[Tang, 1996]。 (d) 中国大陆东-西地壳厚度变化梯度也与海拔“东-西梯度界”吻合[Li & Mooney, 2001] (on line: <http://quake.wr.usgs.gov/research/structure/CrustalStructure/china/index.html>)。

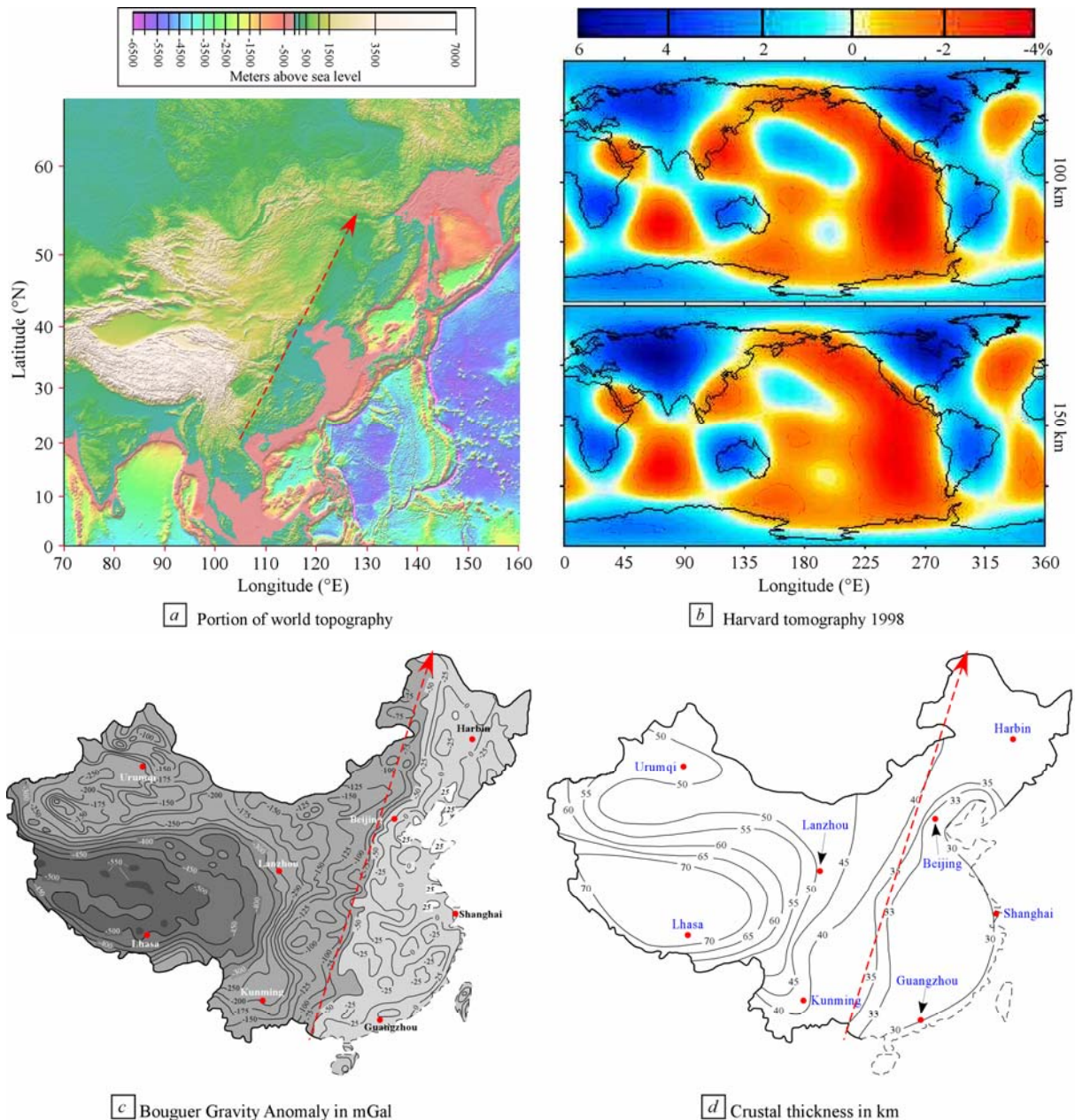


Figure 13. Cartoon illustrating the concept of oceanic lithosphere accretion. The lithosphere thickens as it moves away from the ocean ridge because of heat loss of the mantle, which results in thermal contraction of the topmost mantle, forming cold lithosphere through basal addition, i.e., by “converting” the asthenospheric mantle into the lithospheric mantle [Niu & O’Hara, 2003].

图 13. 示意大洋岩石圈增生的概念。因为地幔热损失，大洋岩石圈随其年龄的增长而增厚。简单地讲，其增厚机制是通过把软流圈顶部物质转化为新的岩石圈底部，亦即，底部迭加[Niu & O’Hara, 2003]。

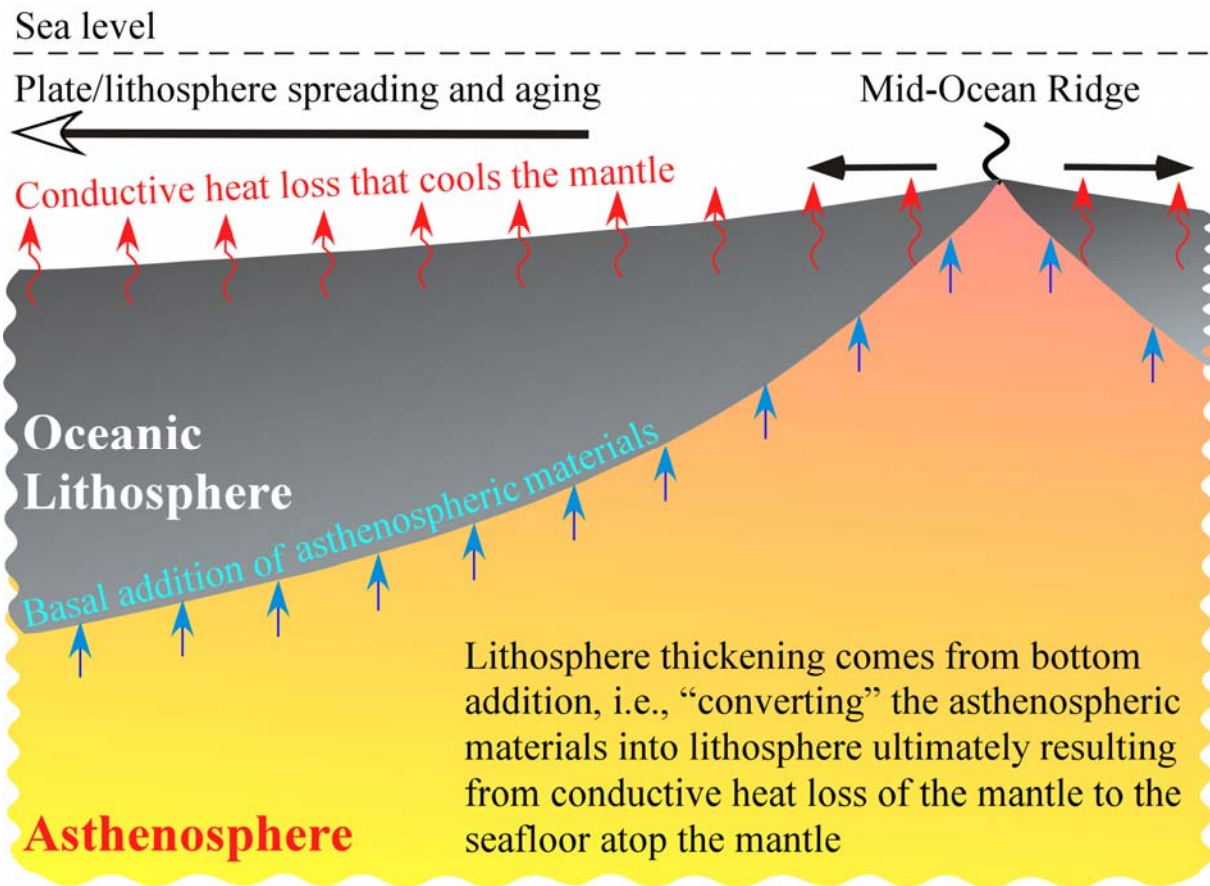


Figure 14. Cartoon illustrating the concept that the lithosphere can be thinned in ways just opposite to oceanic lithosphere accretion (see Fig. 13). This can be achieved by “converting” the basal portion of the lithosphere into the convective asthenosphere by means of heating if there is excess heat available or by adding water which can hydrate the lithosphere. See text for details.

图 14. 示意岩石圈减薄可通过与石圈增生相反的机制。亦即，把岩石圈底部的物质转化为软流圈顶部物质通过加热或加水。

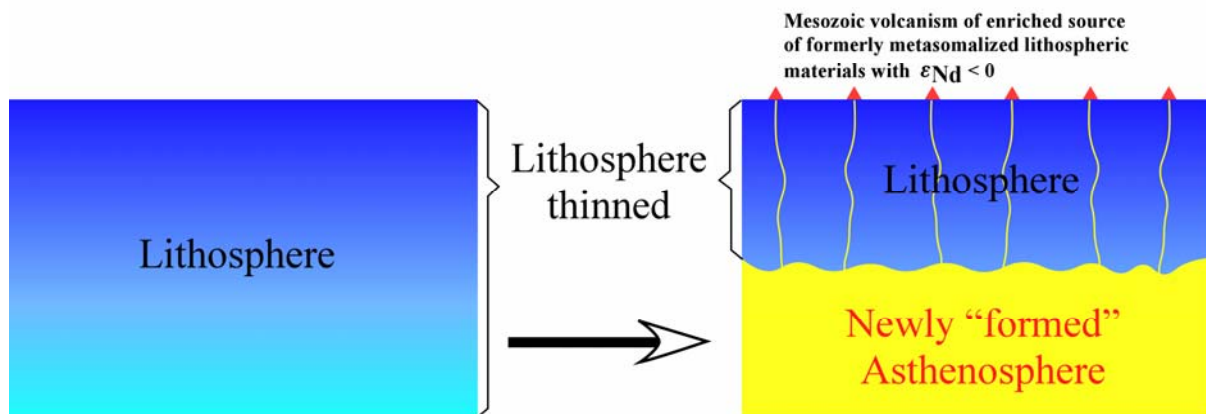
Definitions:

Subcontinental lithospheric mantle:

“Cold”, rigid, strong, highly viscous; isolated from convection

Asthenosphere:

“Hot”, soft, weak, less viscous; part of convection system



Two possibilities to “transform” lithosphere into asthenosphere:

(1) Heating and (2) Adding water

Both can make lithosphere “soft”, “weak”, “less viscous”, and “easily flow”, i.e., the convective asthenosphere

Beneath eastern China, however, (1) Inadequate heat, but (2) abundant water!

Figure 15. Subducting slab structure illustrated by vertical mantle sections across several subduction zones: A-A': the Hellenic (or Aegean) arc; B-B': the southern Kurile arc; C-C': Izu Bonin arc; D-D': the Sunda arc (Java); E-E': the northern Tonga arc; and F-F': central America (after Kárasón & van der Hilst, 2000). Note Sections B-B' and C-C', which show that the subducting slabs do not penetrate 660-km seismic discontinuity (vs. all other sections), but lie horizontally in the transition zone (410 km to 660 km) towards west beneath eastern China continent. This is recently confirmed by the more detailed work of Pei *et al* [2004].

图 15. 俯冲大洋板块地震波速异常的模型影象截面图。虽然许多俯冲板块透过 660 公里地震波不连续面，截面 B-B' and C-C' 表明西太平洋俯冲板块在这里并未穿透 660 公里不连续，而是滞待于 410-660 公里过渡带并西延至中国大陆 (引自 Kárasón & van der Hilst [2000])。

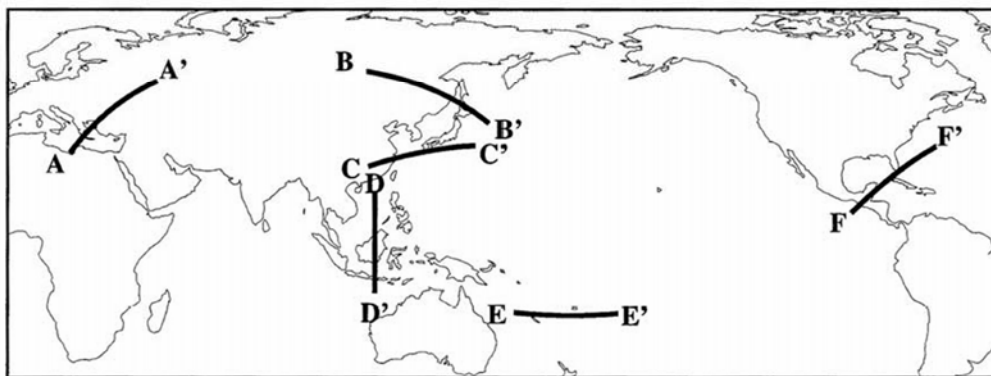
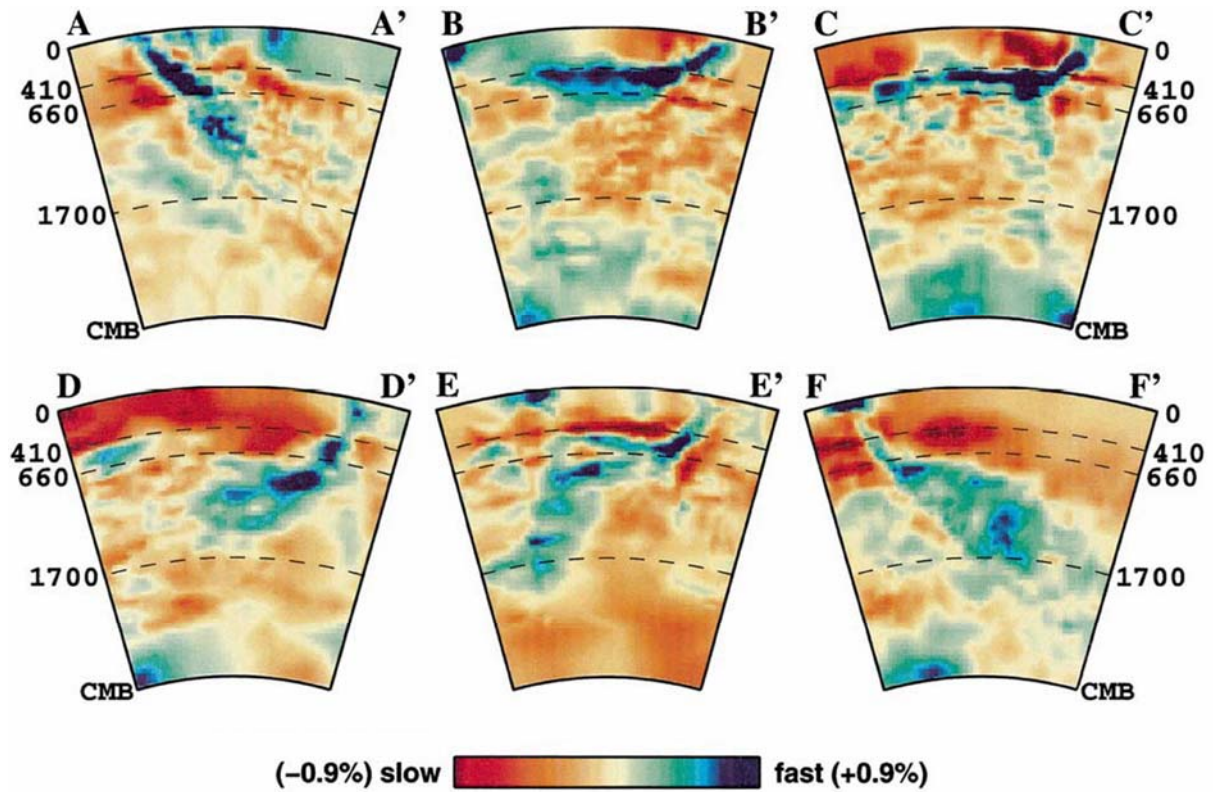


Figure 16. (a) showing that the asthenospheric material flow towards the ridge to form the lithosphere is linearly related to spreading rate. Plotted is total material flux (in km^3) requirement across the ridge in the first one million years against half spreading rate. The total mass required to form the lithospheric mantle as a result of conductive cooling is $\Phi (\text{km}^3) = 15R_{1/2} (\text{km Myr}^{-1})$, and it becomes $\Phi (\text{km}^3) = 25R_{1/2} (\text{km Myr}^{-1})$ if an average of 5 km thick crust is also considered. The important conclusion is that ridge suction force or ridge-ward material flux is significant and increases with increasing spreading rate (the calculations neglect the effect of density changes on the volume, but this effect is $\ll 2\%$ and will not affect the conclusion here). (b) showing that mantle viscosity is the lowest in the asthenosphere, in particular the seismic low velocity zone (LVZ), at depths $< \sim 250$ km, and increases exponentially with depth [see Phipps Morgan et al., 1995; Lambeck & Johnson, 1998], so it is physically straightforward that the ridge-ward material flows must be mostly asthenospheric and horizontal. (c) showing that regions of asthenosphere beneath ocean ridges have the lowest pressure that drives asthenospheric flows (i.e., ridge suction). The square at the ridge axis approximates the size of Fig. 6a. All these are after Niu & Hékinian [2004].

图 16. (a) 洋壳在洋中脊的形成，大洋岩石圈的相当部分也形成于近洋脊附近区域，这要求软流圈物质向洋脊的流动。其流量与板块扩张速度成正比。(b) 地幔黏度随深度的增加而成指数倍增长，因此洋脊区地幔所需要的物质由低速带侧向补给而非源于洋脊下更深处。(c) 示意洋脊下软流层代表整个软流圈内压力最低的区域，因此驱动软流层物质向洋脊的流动，构成巨大的洋脊吸力。

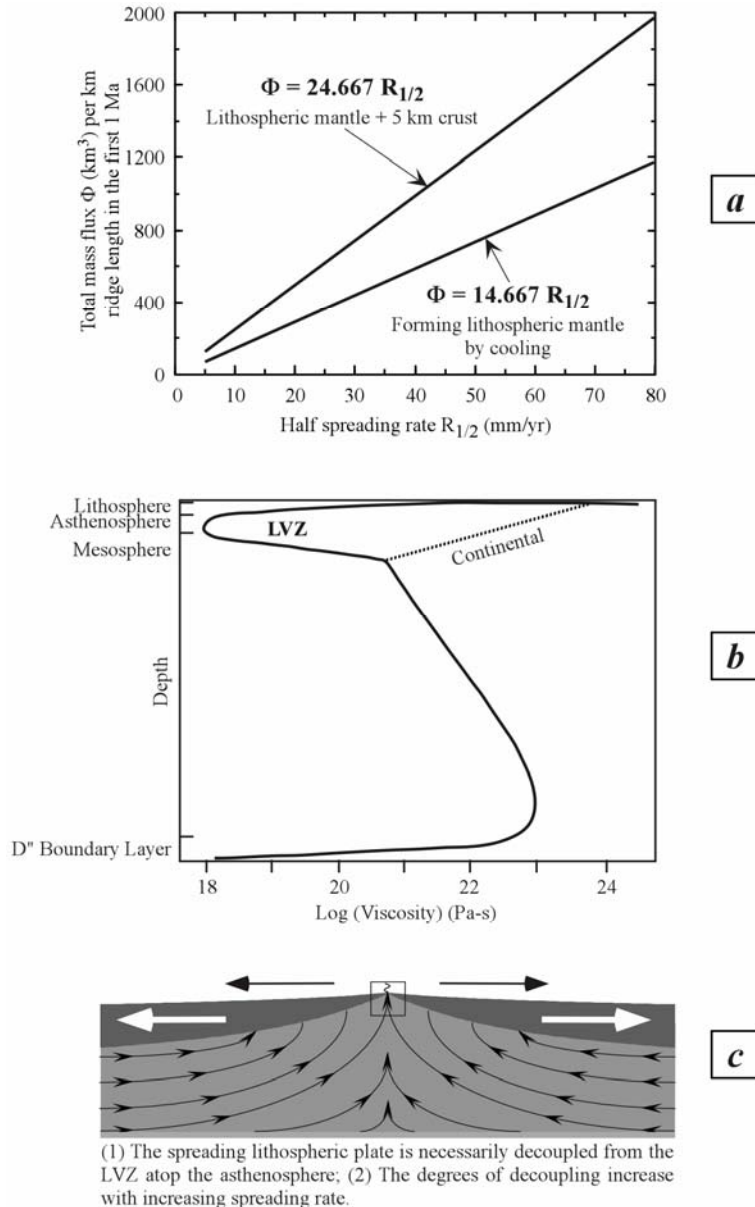


Figure 17. Cartoon showing that the lithosphere is thick beneath the high plateaus in western China, but thin in eastern China. The “transition” is a sharp gradient (Great Gradient Line or *GGL*) in altitude, crustal thickness, and gravity anomaly all due to the lithospheric thickness change (i.e., isostatic effect) as apparent in Fig. 12b. Western Pacific subduction induced corner-flow requires asthenospheric material replenishment from the west (see Fig. 7). In response, the eastward flow of the asthenospheric material beneath eastern China, in turn, requires material replenishment from the west beneath the plateaus. The latter flow experiences a decompression of the order of > 50 km vertically, which results in decompression melting and Cenozoic volcanism in eastern China. This model is rather simplistic, but the concept is physically straightforward.

图 17. 示意中国大陆东-西梯度界不仅表现在东-西海拔高度差别, 而且也反映在地壳厚度差别, 重力异常, 以及岩石圈厚度之差别(亦即, >150 公里在西部高原区, ~ 80 公里在东部丘陵平原区)。与“遥远”的西太平洋俯冲带有关地幔楔吸引作用会引起中国东部软流层东流。该软流层东流必然要引起西部高原底部软流层的东向补给(流动)。这一过程必然会导致东向软流层的减压及减压分熔, 从而导致中国东部新生代基性火山活动。

