Why volatiles are required for cratonic flood basalt volcanism:

Two examples from the Siberian craton

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Abstract

The Siberian Craton was affected by flood basalt volcanism at least twice during Devonian (Yakutsk-Vilyui province) and Permian-Triassic (Siberian province) periods. In both cases volcanism appeared as brief pulses of flood basalt eruptions, followed by kimberlitic (and lamproitic) emplacement. Pressure estimations for the kimberlite-entrained mantle xenoliths reflect that lithosphere was 190-230 km thick at the time of the Devonian flood basalt volcanism. Differently from Devonian kimberlites, the majority of Triassic kimberlites are diamond-free, but at least one Triassic kimberlite pipe and some lamproites are diamondiferous, suggesting that the Siberian lithosphere remained thick during the Permian-Triassic flood basalt volcanic activity. If both the lithosphere and the asthenosphere were volatile-poor, thick cratonic lithosphere prevented melting even at elevated geotherm. During Paleozoic, Siberia was surrounded by subduction systems. The water deep cycle in association with fast subduction and slab stagnation in the mantle transition zone is proposed to cause fluxing of the asthenosphere by water plus other fluids via wet diapir formation in the mantle transition zone. Such diapirs started to melt in the asthenosphere beneath thick cratonic lithosphere producing voluminous melts. Probably mafic melts accumulated beneath

cratonic lithosphere and rapidly erupted on the surface in response to stress-induced drainage events as assumed for some other cratonic flood basalts.

Keywords: flood basalts, subduction, mantle transition zone, Siberian craton

INTRODUCTION

The origin of large igneous provinces (LIPs) is one of the most intriguing and unresolved geological problems. The essential features of LIPs are the particularly large size of volcanically active regions and the significant volume of magma erupted at higher than normal rate. Of course, such definition includes some poorly defined categories such as 'large', 'significant' and 'normal', whereas there is no sharp threshold value distinguishing LIPs from other types of volcanic provinces (e.g., Foulger, 2010). However, a consensus is that a LIP is characterized by size of $>0.5\text{-}1\times10^5$ km^2 and the volume in excess of $0.5\text{-}1\times10^5$ km^3 emplaced during short duration volcanic pulses with overall duration of order of one million year or even shorter (e.g., Coffin and Eldholm, 1994; Sheth, 2007; Bryan and Ernst, 2008; Ivanov et al., 2013; Konstantinov et al., 2014; Ernst, 2014). The pulsing nature of volcanism distinguishes LIPs from other long-lasting provinces such as associated with oceanic spreading ridges and active continental margins.

It is generally believed that some abnormal geological processes either in the Earth's core, mantle and/or crust caused the LIP origin. Numerous models have been proposed and they can be subdivided into those considering terrestrial and extraterrestrial (meteorite impact) causes of the LIP volcanism. Extraterrestrial models are not considered here, but interested readers can check the abundant literature existing with pro- (Jones et al., 2002; Ingle and Coffin, 2004; Hagstrum, 2005) and counterarguments (Ivanov and Melosh, 2003; Korenaga, 2005; Ivanov et al., 2013). As for the terrestrial models, they can be separated by the depth of origin (from deep to shallow) to a number of different type models; lower mantle plume (e.g., Campbell, 2005), transition zone wet plume/diapir (Ivanov and Litasov, 2014), upper mantle heat redistribution (e.g., King and Anderson, 1998), lithospheric delamination (e.g., Elkins-Tanton, 2005) and intralithospheric tectonic processes (Devès et al., 2014).

Cratonic flood basalt volcanism, an important subclass of the LIPs, is of particular interest because it is typically emplaced through thick and cold lithosphere. A cold cratonic geothermal gradient is expected to prevent melting of the lithosphere, whereas thick lithosphere hampers melting at sub-lithospheric depths due to the pressure effect, which suppress magma generation (here lithosphere is defined as a volume from the surface down to the base of the thermal boundary layer (McKenzie and Bickle, 1988)). The melt production preventing is especially true if the source of melting is volatile-free. At a normal mantle geothermal gradient, assuming that we know what is the norm (e.g., Turcotte and Schubert, 2002), dry peridotite starts melting if lithosphere is thinned to about 60 km depth or shallower (Fig. 1). For melting of dry pyroxenite and eclogite the reduction of the lithospheric thickness could be to deeper levels; about 80 and 130 km, respectively (Fig. 1). If not thinned, the cratonic lithosphere at such depth is too cold for generation of magma unless it is volatile-rich (Fig. 1). Devès et al. (2014) provided a model of lithospheric heating due to shearing, but obviously, such a model predicts highly localized melting, whereas continental flood basalts are often characterized by enormous spatial extent, even considering that the lava field is larger than the source of melting. In addition to that, significant displacement of lithospheric block is not expected for intracratonic areas.

High temperature within an upwelling mantle plume, which is up to 300 degrees higher compared to a normal mantle (e.g., Campbell, 2005), or increasing temperature due to upper mantle internal warming (radioactivity, reorganization of convective flow, thermal blanketing by supercontinents etc.) to about the same high temperature (e.g., King and Anderson, 1998; Coltice et al., 2007; Anderson, 2011) may produce melting of the eclogitic part of a composite thermochemical plume (Yasuda and Fujii, 1998; Sobolev et al., 2011), floating eclogitic blob (Korenaga, 2004; Anderson, 2007) at the base of thick cratonic lithosphere (Fig. 1). However, generation of a high volume flood basalt province would still require thinning of the lithosphere either via rifting or delamination and further decompression melting irrespective of the assumed source of the eclogite, delaminated continental crust (e.g., Anderson, 2005; Lustrino, 2005) or recycled oceanic crust (e.g., Korenaga, 2004; Sobolev et al., 2011). Melting of within lithospheric metasomatic veins (carbonated or micaceous) is possible with slightly elevated temperature from the cratonic geotherm, but there should be a reasonable explanation for the increase of the temperature. Usually, heating from a plume is invoked (e.g., Gallagher and Hawkesworth, 1992).

In other words, assuming the cratonic geotherm as illustrated in Fig. 1 and dry mantle, melting of the cratonic lithospheric or sub-lithosheric mantle is possible only for the eclogitic portions possibly present within the peridotite matrix. Due to this fact, there are only two potential options to originate cratonic flood basalts without requiring temperature excesses. Cratonic flood basalts can be generated only if the depth of melting decreases via rifting and/or lithospheric delamination or if the source of melting is fluxed by volatiles such as H_2O, CO_2 (Fig. 1) and probably F (Brey et al., 2009). Combination of the two processes (thinning and fluxing) is also possible. Another alternative is that magma forms along thinner cratonic lithosphere edges, i.e. the loci of ancient suture zones (e.g., King and Anderson, 1998). In such case a lateral emplacement of radiating dyke swarms for distance of 1000 km or even 2000 km into the interior of the craton from its margin is required (e.g., Ernst et al. 2005; Ernst, 2014).

In this paper, I consider two flood basalt provinces; Yakutsk-Vilyui and Siberian, which were emplaced onto the Siberian Craton in Devonian and Permian-Triassic, respectively. I provide evidence that the lithosphere remained thick during the flood basalt volcanism. I suggest the mechanism of volatile fluxing in association with subduction and transition zone slab stagnation processes beneath Siberia.

WHAT IS THE PLUME?

According to fluid dynamic definition, a plume is any self-buoyancy driven mass flow (Korenaga, 2005). In this sense practically any motion within the Earth's mantle should be considered as plume including subducting slabs and sinking delaminated portions of lithosphere. These lead to the situation when scientific literature is overwhelmed by the term 'plume' used in variously possible meanings. There are super- and secondary plumes, Morganian and Andersonian plumes, lower mantle, transition zone and wedge plumes, thermal, thermochemical, hot, cold and wet plumes among others (e.g., Courtillot et al., 2003; Gerya and Yuen, 2003; Campbell, 2005; Gerya et al., 2006; Zorin et al., 2006). Overuse of the term plume often results in misunderstanding what is actually meant. To overcome this problem, the term plume should be specifically defined. Here I use the term plume in its original sense (Morgan, 1971; Campbell, 2005), that the plume is a thermally- and buoyancy-driven solid mass flow, which originates in the lowermost mantle. Usually such plumes are thought to contain two major parts; a large head (~1000 km in diameter) followed and feed by a thin tail (<100 km in diameter) (Campbell, 2005). Plumes are not, at least directly, related to the plate tectonic processes and probably provide driving forces for the plates (Morgan, 1971). If the plume contains subducted lithologies, it can be referred to as thermochemical plume. Any other buoyancy-driven solid mass flows associated with slabs, delaminated lithosphere, upper mantle convective motions are not considered here plumes. They are plate tectonics-related phenomena. I suggest using for the plate-tectonic related upwelling flows the traditional term 'mantle diapir'. Location of a mantle plume and associated volcanic province is unpredictable from the plate tectonic reconstructions (but see the contrary view by Burke et al., 2008).

SIBERIAN CRATON AND ITS FLOOD BASALT PROVINCES

The Siberian Craton is a structure composed of the Paleoarchean to Paleoproterozoic blocks, which were welded together during a Paleoproterozoic episode of magmatism and metamorphism that peaked at about 1.87 Ga (Gladkochub et al. 2006; Rojas-Agramonte et al. 2011). After that time, the Siberian Craton represented a single, large tectonic block, which was a constituent part of the supercontinents Nuna/Columbia (Paleoproterozoic), Rodinia (Mesoproterozoic) and Pangea (Paleozoic) and between the supercontinent cycles it acted as core of a separate continental block referred to as Siberian continent (Cocks and Torsvik, 2007; Li et al., 2008; Domeier and Torsvik, 2014).

In the Phanerozic, the Siberian Craton experienced flood basalt volcanism at least twice, during Devonian (e.g., Kiselev et al., 2012) and Permian-Triassic (e.g., Ivanov et al., 2013). The Precambrian flood basalt provinces on the Siberian Craton are inferred from dyke records. Ernst (2007) argues that mafic/ultramafic dykes with average width of >10 m are indicators for being feeders of a flood basalt province. The Paleoproterozoic dykes are well known within the Siberian Craton (Gladkochub et al., 2010; Ernst et al. 2013; Ernst et al. 2014). Recent data indicates on a ~1 Ga mafic/ultramafic event within the Siberian craton (Ivanov et al., 2012; Savel'eva et al., in press). Here I focus only on the Phanerozoic flood basalt provinces.

The Devonian Yakutsk-Vilyui flood basalt (YVFB) province is located mainly within a large Vilyui rift (Fig. 2). Numerous Devonian dykes and sills are known over a larger portion of the eastern Siberian Craton and they are considered as constituent part of the YVFB province with eroded lava (Kiselev et al., 2012). Volcanic and intrusive rocks of the YVFB are mainly low magnesium basalts of both the high- and low-Ti series (Fig. 3). High-Ti basalts are predominant. Typical rock compositions are provided in the Table 1, whose primitive mantle (McDonough and Sun, 1995) normalized incompatible element spectra are shown in Fig. 4. Evolved rocks such as mugearites and benmoreites are also present. The limited compositional variability of the YVFB could be an artefact of the insufficient analytical data, however.

The YVFB province is poorly dated, but available geochronological data, mainly $^{40}Ar/^{39}Ar$ ages (Courtillot et al., 2010; Ricci et al., 2013; Kiselev et al., 2014; Ivanov et al., submitted) and only two U-Pb ages (Powerman et al., 2013), suggest that volcanism appeared in pulses (Fig. 5). U-Pb ages of kimberlites (Kinny et al., 1997), combined with geologic constraints (Kiselev et al., 2014) indicate that emplacement of the kimberlitic magma followed mafic volcanism (Fig. 5, Ivanov et al., submitted), not the vice versa as would be expected from lithoshperic thinning model of the flood basalt formation. Many of the Devonian kimberlites of the Siberian Craton are diamondiferous, suggesting that lithosphere was thick enough for preservation of diamonds.

PT estimates for the Devonian kimberlite-entrained mantle xenoliths show that the lithosphere within the Siberian Craton was 190-230 km thick depending on location (Griffin et al., 1999).

The Siberian flood basalt (SFB) province is much larger than the YVFB province (Fig. 2). About a half of the SFB province was emplaced onto the Siberian Craton. Another half was emplaced within the younger lithosphere of the rifted West Siberian Basin. Unlike the YVFB, the SFB is characterized by much larger chemical variation of magma compositions (Fig. 3). The volcanic and intrusive rocks of both provinces belong to both high- and low-Ti rock series. The high-Ti rock series is characterized by large variations in MgO content from low magnesium basalts to high magnesium meimechites and dunites. The latter are likely cumulates of the primary meimechite magmas. Low-Ti basalts are uniform in composition and represent up to 80% of the total volume of the SFB (Ivanov, 2007). Typical low- and high-Ti basalt compositions are listed in the Table 1 and their primitive mantle (McDonough and Sun, 1995) normalized incompatible element spectra are shown in Fig. 4. Felsic volcanic rocks are also present, but they are rare. Granites are abundant on the southern and northern periphery of the SFB (Vernikovsky et al., 2003; Dobretsov et al., 2005). In addition to the felsic magma, within the SFB there were eruptions of carbonatites, which are spatially associated with meimechites (Fedorenko and Czamanske, 1997; Kogarko and Zartman, 2007).

Similarly to the YVFB province, the SFB province formed in pulses (Ivanov et al., 2013). The major volcanic pulse was coeval with the Permo-Triassic boundary (Renne and Basu, 1991; Reichow et al., 2002; 2009; Kamo et al., 2003; Svensen et al., 2009; Paton et al., 2010). A later Middle Triassic pulse is also recognized (Ivanov et al., 2005; 2009; 2013; Reichow et al., 2009). During Triassic, the Siberian Craton was affected by kimberlitic (Kinny et al., 1997) and lamproitic (Ivanov et al., 2013; Letnikova et al., 2013) volcanism. The kimberlitic and lamproitic volcanism followed the mafic volcanism (Fig. 5). The majority of Triassic kimberlites are diamond-free, which, in combination with relatively shallow pressure estimates for mantle xenoliths entrained in the kimberlites, was taken as evidence for thinning of the lithosphere to 150 km (above the diamond stability field) due to the Siberian flood basalt volcanism (Howarth et al., 2014a). However, all known diamond-free Triassic kimberlites are located either outside the Siberian flood basalts or near its marginal parts (Fig. 2). Exception is the Malokuonapsky kimberlite pipe, which contains industrially-grade diamond deposit (Khar'kiv et al., 1998; Sobolev et al., 2013).

A diamond-rich deposit of lower Carnian age tuffs (237 Ma for the Carnian/Landian boundary according to the latest Triassic timescale of Ogg et al., 2014) is known in the north-eastern Siberian Craton with non-kimberlitic diamonds of so-called morphologic variety V; octahedral habit diamonds with syngenetic graphitic inclusions (e.g., Zinchuk et al., 1999;

Ragozin et al., 2009; Grakhanov and Koptil', 2003; Shatsky et al., 2014). The composition of the tuffs suggests that their sources (and probably the source of diamonds) were lamproitic pipes and dykes (Letnikova et al., 2013).

Until recently, Carnian age lamproite intrusions were not known for the Siberian Craton. Ivanov et al. (2013) obtained 238 to 235 Ma $^{40}Ar/^{39}Ar$ ages for a lamproite sample, thus straddling Ladinian-Carnian (Ogg et al., 2014). The dated lamproite is located within the Noril'sk section of the Siberian flood basalt province (Fig. 2). Potential connection of lamproites and diamonds of the variety V suggests that at least some parts of the Siberian lithosphere remained thick (base of the lithosphere was below the diamond stability field) in the Triassic. Peridotitic source for the dominant low-Ti tholeiites of the SFB province was inferred from geochemical data, whereas volumetrically less abundant high-Ti basalts could be derived from eclogite-bearing source (Ivanov, 2007; Sobolev et al., 2009a). Seismic data on the present-day lithospheric structure of the Siberian Craton suggests that it is thicker than 180 km everywhere (Priestley and McKenzie, 2006; Pasyanos, 2010), which is another evidence for thick and unperturbed Siberian lithosphere.

These two examples of cratonic flood basalts return us to the question how can the large volume of mafic magma be produced within cold or under Siberian thick cratonic lithosphere? Eclogitic source alone is not enough. Although the Siberian mantle contains metasomatic zones with phlogopite (e.g., Solov'eva et al., 2012) and the nominally anhydrous upper mantle minerals from xenoliths contain up to ~300 ppm of H_2O (Doucet et al., 2014), these xenoliths may represent material located near melt conduits and thus may not represent the bulk of the cratonic lithosphere (Doucet et al., 2014). Apparently, the flux of volatiles like H_2O, CO_2 and probably F, all species that can significantly decrease the temperature of melting of upper mantle rocks (Wallace and Green, 1988; Sato et al., 1997; Brey et al., 2009), could solve the problem.

GEODYNAMIC SETTING OF THE YVFB AND SFB PROVINCES

Fig. 6 shows paleogeographic reconstructions (Domeier and Torsvik, 2014) in Devonian (390 Ma) and Permian (270 Ma) times, thus ~20 Ma before the major pulses of the YVFB and SFB provinces, respectively. In Devonian, the Siberian continent was bounded to the paleo-South by a subduction system involving recycling of the Rheic Ocean and to the paleo-North and to the paleo-West by spreading systems of the Mongolia-Okhotsk Ocean (Fig. 6). However, further to the paleo-North and to the paleo-West there were other subduction systems and thus, in a broader sense, the Siberian continent was surrounded by Devonian subduction systems from every side. Distances from the reconstructed paleo-arcs to the margins of the YVFB province varied from about 1 to 2 thousand km. In the late Devonian, the Siberian continent was characterized by another volcanic activity, named the Altai-Sayan Province (Vorontsov and Sandimirov, 2010; Vorontsov et al., 2013) located closer to the continental boundary (Fig. 2, 6). Although the Altai-Sayan Province was considered as plume-related, the volcanic rocks show subduction-related incompatible element patterns in primitive mantle-normalized plots (Vorontsov et al., 2013) similar to some other flood basalt provinces (e.g., Puffer, 2001).

In the late Permian, Siberia was at the northern end of the supercontinent Pangea (Fig. 6). Two subduction systems influenced this part of the Pangea: the Mongolia-Okhotsk slab and Paleotethys slab subducting towards paleo-Southwest and paleo-Northwest, respectively. The SFB province covered an enormous territory of ~7×10^6 km^2 (Ivanov, 2007). The closest distance from the reconstructed Mongolia-Okhotsk subduction system to the margin of the SFB province was ~700-800 km (Ivanov and Litasov, 2014) and from the Paleotethys it was more than 2000 km (Fig. 6). In Triassic, there was another volcanic province formed closer to the Paleotethys subduction in Fore-Caucasus region (Chalot-Prat et al., 2007). The Fore-Caucasus volcanism was considered as within-plate, but it is characterized by clear subduction-related incompatible element signature (Chalot-Prat et al., 2007). The Emeishan flood basalt province, which formed in the late Permian was located even in closer association with subduction systems (Fig. 6). Despite this province is considered by many as plume-related (e.g., Campbell, 2005; Shellnutt, 2014), it also has subduction-like incompatible element signatures (e.g., Zhu et al., 2005; Shellnutt, 2014).

The low-Ti ‘subduction’ signature is commonly considered as either lithospheric contribution or crustal contamination (e.g., Lightfoot et al., 1993; Wooden et al., 1993; Hawkesworth et al., 1995; Reichow et al., 2005; Jourdan et al., 2007; Kiselev et al., 2012). From the Table 1 it may be seen that there is no significant difference between high-Ti (conventionally considered uncontaminated) and low-Ti (often considered contaminated) basalts in terms of Nd and Sr isotopes. Thus, similar isotopic ratios for the high- and low-Ti basalts are ruling out crustal contamination for the low-Ti basalt, which exhibit ‘subduction’ signatures (relative depletion of Nb and relative enrichment of Sr and Pb) in the incompatible element normalized plots (Fig. 4). It should be emphasized, that lithospheric and crustal contamination is likely for the flood basalts, but it cannot explain the voluminous low-Ti basalt series as a whole, at least in case of the SFBB (Ivanov, 2007; Ivanov et al., 2008).

Considering the paleotectonic settings for the Siberian Craton flood basalt provinces, it should be stated that both flood basalt provinces under consideration were located in far back-arc setting. Such a connection between the Permian-Triassic subduction systems and the SFB province was noted previously by Cox (1978) and Nikishin et al. (2002). The association of the YVFB province to subduction systems is made here for the first time.

Ernst (2014) noted that most of the continental flood basalt provinces are located in back-arc regions within 1000-2000 km from the corresponding arc systems. However, he questioned that such large distances should allow consideration of such setting as 'back-arc'. However, modern examples show that the great majority of slabs stagnate in the mantle transition zone and propagate hundreds to thousands km under continents (Fukao et al., 2001; 2009). Thus, there is a physical way to explain how flood basalts could be linked to subduction; this link was originally suggested many years ago by Cox (1978) without knowing the stagnant slab phenomenon.

DISCUSSION

In a previous section, I have shown that diamondiferous kimberlites and lamproites emplaced after (not before) flood basalt volcanism, suggesting that lithosphere was thick before, during and after generation of the flood basalt magma. Unless the cratonic flood basalts were not fed laterally through crustal dykes, whose original source was located in a region of thin lithosphere, the large thickness of the cratonic lithosphere requires the solidus of melting to be depressed by volatile fluxing; H_2O, CO_2 and may be F. Such a requirement for depressing mantle solidus for origin of flood basalts (and 'hot spots') was noted by many others (e.g., Bonatti, 1990; Gallagher and Hawkesworth, 1992; Anderson, 1995; Puffer, 2001; Silver et al., 2006).

Gallagher and Hawkesworth (1992) assumed that the lower portion of the cratonic lithosphere is wet and weak, but proposed the presence of a hot upwelling mantle plume as a source of heat for production of the continental flood basalts. Anderson (1995) argued that such weak layer, referred to as perisphere, cannot be locally stable due to its rheology and should spread laterally. According to Anderson (1995), the perisphere is located between strong lithosphere defined by ~650 ^{o}C isotherm and convecting asthenosphere. It is deeper beneath continents and shallower beneath oceans. Deformed peridotites (Fig. 1) are the candidates for the perisphere beneath the cratons, though the equilibration temperature of the deformed peridotites is >1100 ^{o}C. Anderson (1995) assumed that perisphere is wet and could be created at mantle wedges above subducting slabs in the geologic past. Later this concept was transformed to the Laminated Lithology with Aligned Melt Accumulations (LLAMA) model, which was applied for oceanic regions (Anderson, 2011). In terms of continents, flood basalts are expected along cratonic boundaries where lithosphere is thin, not beneath the central parts of thick cratons.

Puffer (2001) used a concept, which resembles both the wet lithosphere (Gallagher and Hawkesworth, 1992) and the perisphere (Anderson, 1995) models. He showed that some flood basalt provinces (e.g., Karoo, Siberian and Central Atlantic Magmatic Province) are characterized by incompatible element patterns similar to the island arc basalts, whose origin is indisputably associated with water fluxing from subducting slabs (e.g., Stern, 2002). Further, Puffer (2001) assumed that the magma source of such subduction-type flood basalt provinces is located within paleo-subduction mantle wedges, which underplated the cratonic lithosphere in some geologic past. According to Puffer (2001), the reactivation of such low solidus lithospheric/perispheric mantle can produce flood basalt provinces. Many interpretations of continental flood basalts as being sourced from lithospheric mantle (e.g. Jourdan et al., 2007; Kamenetsky et al., 2012) are conceptually the same as Puffer’s (2001) model.

Silver et al. (2006) noted the importance of subduction-derived fluids in the cratonic flood basalts too. The new idea introduced by Silver et al. (2006) was that the rate of magma production at depth and the rate of magma eruption on the surface are not the same, with the latter being much faster compared to the former. Magma is probably accumulated for prolonged geological time in mantle and erupted quickly due to a tectonic venting event. This is of particular importance for the extremely short (Konstantinov et al., 2014) and pulsing (Ivanov et al., 2013) nature of flood basalts.

The concept, which is preferred in this paper (Fig. 7), is reviving the idea of direct and genetic connection between the flood basalt volcanism and subduction (Cox, 1978). Cox (1978) noted that some flood basalt provinces, including those later identified by Puffer (2001) as subduction-type, were located in back-arc setting and could be directly linked to subduction processes.

Subduction is a long-lasting process, which continues for tens and hundreds of Ma. Flood basalt volcanism is pulsing and of short duration (few Ma). In the frame of the subduction concept, the flood basalt phenomenon is related to unusual mode of ultrafast subduction. The faster subduction the colder slab penetrates to deep levels without degassing beneath volcanic arcs (Ivanov and Litasov, 2014). If the rate of subduction is in the order of 20 cm/yr (similar to convergence rate at Tonga trench; Bevis et al., 1995), most of the water budget can probably be subducted down to the mantle transition zone in form of solid ice VII (Bina and Navrotsky, 2000) or other hydrous phases stable at high T (e.g., antigorite, phase 10-Å, phase A; Litasov and Ohtani, 2013; Schmidt and Poli, 2014; Fig. 6).

Slabs tend to stagnate in the mantle transition zone and can probably move horizontally for large distances underplating convecting upper mantle beneath continents. After some period of time, the stagnant slabs are inevitably heated up to the ambient mantle temperature and rapidly dehydrate. Dehydration induces hydrous melting in the mantle transition zone and a composite mantle diapir can rise up creating flood basalt province on the surface (Fig. 7). Alternatively, the released water is incorporated into ringwoodite and/or wadsleyite, the major water-bearing minerals of the mantle transition zone (Ohtani, 2005). Hydrated ringwoodite and wadsleyite are characterized by increasing volume; for example, in experiments at zero pressure 0.5 % of water in wadsleyite produces the same effect on crystal volume as about heating by 240 degrees and the effect for ringwoodite volume increase is about double of that for wadsleyite (Smyth and Jacobsen, 2006). Despite that there is no experiments on effect of water on the thermal expansivity at high pressure, it is generally agreed that a hot plume is not the only reason for creation of buoyant diapirs, water can do the work instead of temperature (Gerya and Yuen, 2003; Gerya et al., 2006; Faccenda, 2014).

Experiments show that subducted carbonates are expected to melt at T and P conditions occurring in the mantle transition zone producing carbonatites (Litasov et al., 2013). Carbonatitic melts can probably rise up from the mantle transition zone, providing source of CO_2, oxidize upper mantle, and provoking voluminous melting (Litasov et al., 2013).

Another possibility for deep sub-lithospheric melting is redox melting, which involve recycling of carbon, hydrogen or methane from the stagnant slab in the transition zone mantle and their oxidation due to reaction with ambient mantle in the sub-cratonic asthenosphere (Foley, 2011; Rohrbach and Schmidt, 2011).

Such subduction model referred to as the deep water cycle (Ivanov and Litasov, 2014), was tested for the SFB (e.g. Ivanov, 2007; Ivanov et al., 2008; Ivanov and Litasov, 2014). The supporting observations are the following:

(1) Peculiar location of the SFB province with respect to subducting systems (Nikishin et al., 2002; Ivanov, 2007; Fig. 6);

(2) Abundant water-bearing (mica and amphibole) minerals in mafic intrusions (Ivanov, 2007; Ivanov et al., 2008);

(3) Subduction-like incompatible element patterns with depletion of Nb and Ta and enrichment of Sr and Pb relative neighboring elements in dominant low-Ti basalt rock series (Puffer, 2001; Ivanov et al., 2008) (Fig. 4).

Another debatable, but on my view also supporting observation, is availability of carbonatitic complexes temporally associated with the SFB (Kogarko and Zartman, 2007). Ernst and Bell (2010) suggest that the association of carbonatites with flood basalts is evidence of a plume, when carbonatites are produced from a more volatile rich-area along the plume margins with melting of the main portion of the plume causing normal basalts. Ernst and Bell (2010) argues that carbonatites never associate with subduction environment. However, recent interpretations of carbonatites (and kimberlites) link this type of magmatism with slabs penetrating into the transition zone mantle (e.g., Duke et al., 2014) in agreement with experimental results on the deep origin of, at least, some types of carbonatites (Litasov et al., 2013).

The deep water cycle model can be easily combined with the tectonic venting idea of Silver et al. (2006) to explain the extremely short duration of individual volcanic pulses within the SFB province (Konstantinov et al., 2014). Indeed, the shorter volcanic pulses the harder to reconcile them with thermal (plume) anomalies, because thermal processes are inert and cannot produce short-lived volcanism. In other words, the association of the SFB province with a lower mantle plume, superplume or similar is a tradition, but it is not required by evidence.

The deep water cycle model was not tested for the Devonian YVFB province. So far, a plume model is considered as most promising by other authors (Kiselev et al., 2012). They noted radial distribution of dykes and rifts (Fig. 2) and assumed that they can be traced to the same plume centre located outside the Siberian Craton (Fig. 6). According to that idea, melting occurred at shallow depth outside (or in between) the cratonic area and then magma flowed via dyke conduits, reaching crustal levels.

To test this idea, a counterpart of the YVFB province has to be found within other Devonian terranes adjacent to Siberia (Fig. 6). It should also explain significant difference in chemistry between the low-Ti and high-Ti basalts (Fig. 3, 4), considering their isotopic similarity (Table 1).

By the way, the model of radiating dykes from the same plume center do not require either hot or lower mantle plume. This model may work similarly with a wet transition zone generated diapir as suggested in the deep water cycle model.

CONCLUSIONS

If volatile-poor, cratonic lithosphere is too cold and thick to allow melting within or under the lithosphere and thus it prevents generation of voluminous flood basalts. In order to explain cratonic flood basalts it is necessarily to assume that either (1) the magma emplaced outside the cratonic area or magma was produced in correspondence of lithospheric thickness reduction along ancient suture zones separating different cratonic portions and propagated laterally to the thick cratonic areas via long dykes, or (2) cratonic lithosphere was thinned via delamination and/or rifting, or (3) sub-lithospheric mantle was H_2O-, CO_2-, F-rich for lowering temperature of melting. In this paper I show that there was no pre-volcanic thinning of the lithosphere. There is no evidence on lithospheric reduction along ancient zones either. Among the two remaining explanations, I consider that the volatile fluxing at sub-lithospheric or lithospheric depths is more plausible explanation. The volatile-rich source could be formed by various processes, but the most probable is the wetting of the sub-cratonic mantle via fast subduction, transition zone slab stagnation, slab warming, slab degassing and generating wet diapirs not long before the flood basalt volcanism. This process is referred to as the deep water cycle.

ACKNOWLEDGEMENTS

I thank informal 'Platonics' (plate tectonics) group for stimulating discussions over the years. Especially I express my deep respects to Don L. Anderson for his influence on these discussions and his positive attitude to the role of volatiles and the mantle transition zone in the origin of continental flood basalts. Critical comments, questions and useful suggestions of Vadim Kamenetsky, Robert Stern, Richard Ernst and Michele Lustrino are greatly appreciated.

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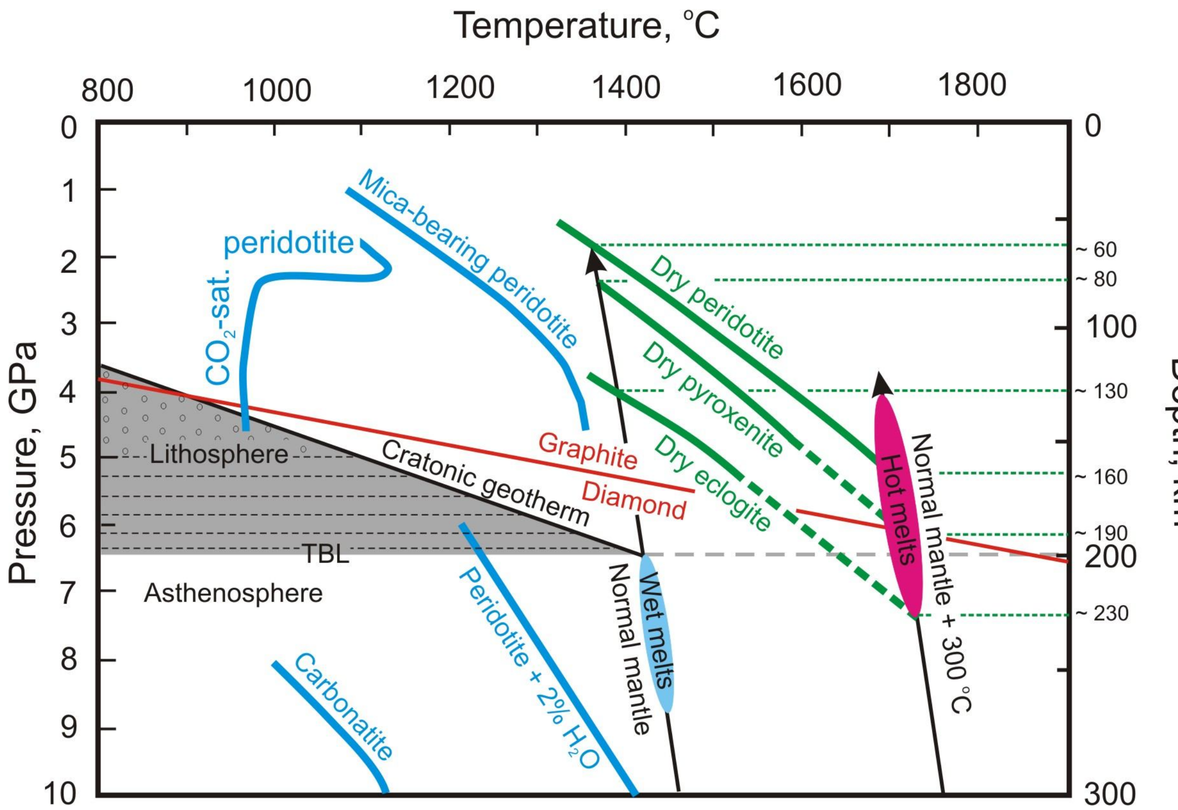

Figure 1. Solidi of dry and wet mantle lithologies (Wallace and Green, 1988; Sato et al., 1997; Hirschmann, 2000; Kogiso et al., 2003; Litasov and Ohtani, 2003; Litasov et al., 2013) and carbonatites (Litasov et al., 2013). Solid lines are based on experimental data and hatched lines are extrapolations. Red line marks graphite-diamond equilibrium (Pal'yanov et al., 2002). TBL – thermal boundary layer. Lithosphere is defined as the volume between the surface and the TBL. Mantle geotherm is from (Turcotte and Schubert, 2002). Cratonic geotherm is from typical PT values of cratonic xenoliths including those of the Siberian Craton and corresponds to surface heat flux of about 45 mW/m^2 (Rudnick and Nyblade, 1999; Lee et al., 2011). Dashed thin lines and open circles within the grey field of the cratonic lithosphere mark predominant localization of deformed and coarse-granular peridotites, respectively (e.g., Rudnick and Nyblade, 1999; Solov'eva et al., 2008). The figure shows that fluxing by H_2O and CO_2 can produce mantle melting beneath thick cratonic lithosphere, whereas a high temperature plume (plus 300 ^oC over normal mantle geotherm) can produce melting of dry eclogite, but is not able to melt other types of dry mantle lithologies.

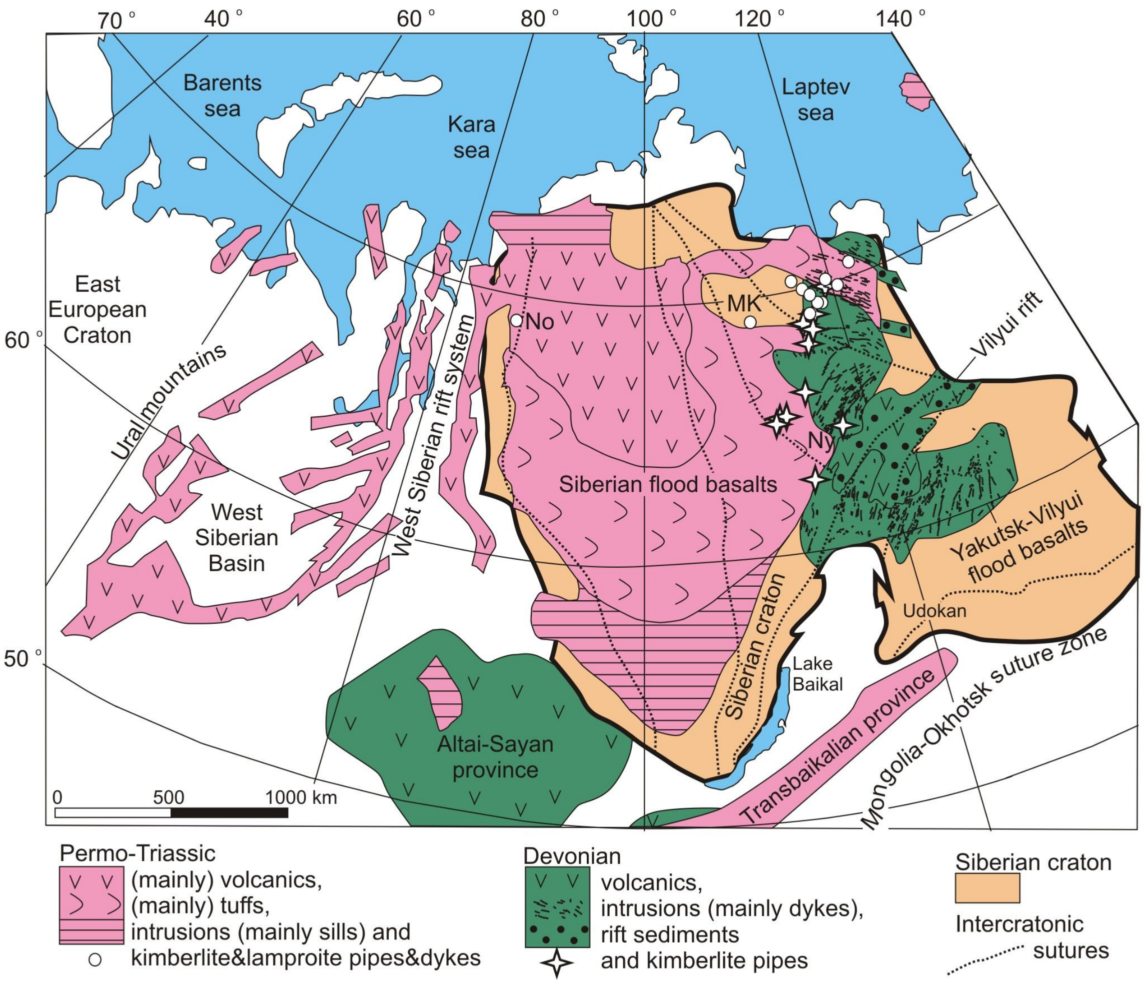

Figure 2. Distribution of Devonian YVFB and Permo-Triassic SFB provinces. The outlines are from Kiselev et al. (2012) and Ivanov et al. (2013), respectively. Devonian dykes are after Kiselev et al. (2012). Boundary of the Siberian craton is from Rosen et al. (1994) and Smelov and Timofeev (2007). Bold dashed curves are intercratonic sutures after Rosen et al. (1994). No – marks position of a dated lamproite dyke, which cuts the Noril'sk section of the Siberian flood basalts (Ivanov et al., 2013). Ny – marks position of Nyurba kimberlite with evidence of emplacement between the two Devonian flood basalt pulses (Kiselev et al., 2014). MK – marks position of the Malokuonapskaya diamondiferous Triassic kimberlite pipe (Khar'kiv et al., 1998; Sobolev et al., 2013). The Altai-Sayan volcanic and Transbaikalian provinces are shown after Vorontsov et al. (2013) and Yarmolyuk et al. (2001), respectively. Late Cenozoic volcanic fields, which probably are related to the Pacific stagnant slab are also shown (Ivanov et al., 2011).

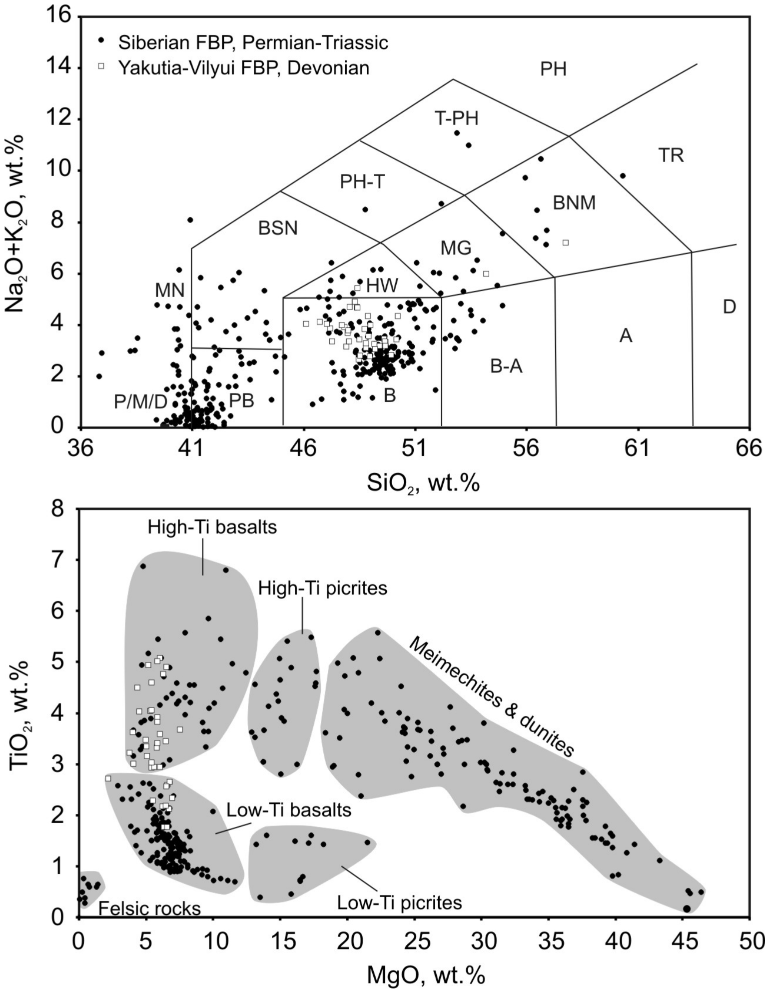

Figure 3. SiO_2-Na_2O+K_2O and MgO-TiO_2 diagrams for the YVFB (Kiselev et al., 2012) and SFB (Zolotukhin and Al'Mukhamedov, 1991; Sobolev et al., 1992; Lightfoot et al., 1993; Wooden et al., 1993; Arndt et al., 1995; Hawkesworth et al., 1995; Fedorenko and Czamanske, 1997; Kogarko and Ryabchikov, 2000; Ryabchikov et al., 2001; Medvedev et al., 2003; Reichow et al., 2005; Carlson et al., 2006; Ivanov et al., 2008; Panina and Usoltseva, 2008; Sobolev et al., 2009; Black et al., 2012) provinces. Rock names and dividers in the SiO_2-Na_2O+K_2O are after Le Bas and Streckeisen (1991); PB – picrobasalt, B – basalt, B-A – basaltic andesite, A – andesite, D – dacite, HW – hawaiite, MG – mugearite, BNM – benmoreite, TR – trachyte, BSN – basanite, PH-T – phonolitic tephrite, T- PH – tephritic phonolite, PH – phonolite. Position of the high magnesium rocks (P – picrite, M – meimechite, D – dunite, MN - melanonephelinite) is shown approximately, because the high magnesium rocks shall not be classified with this diagram.

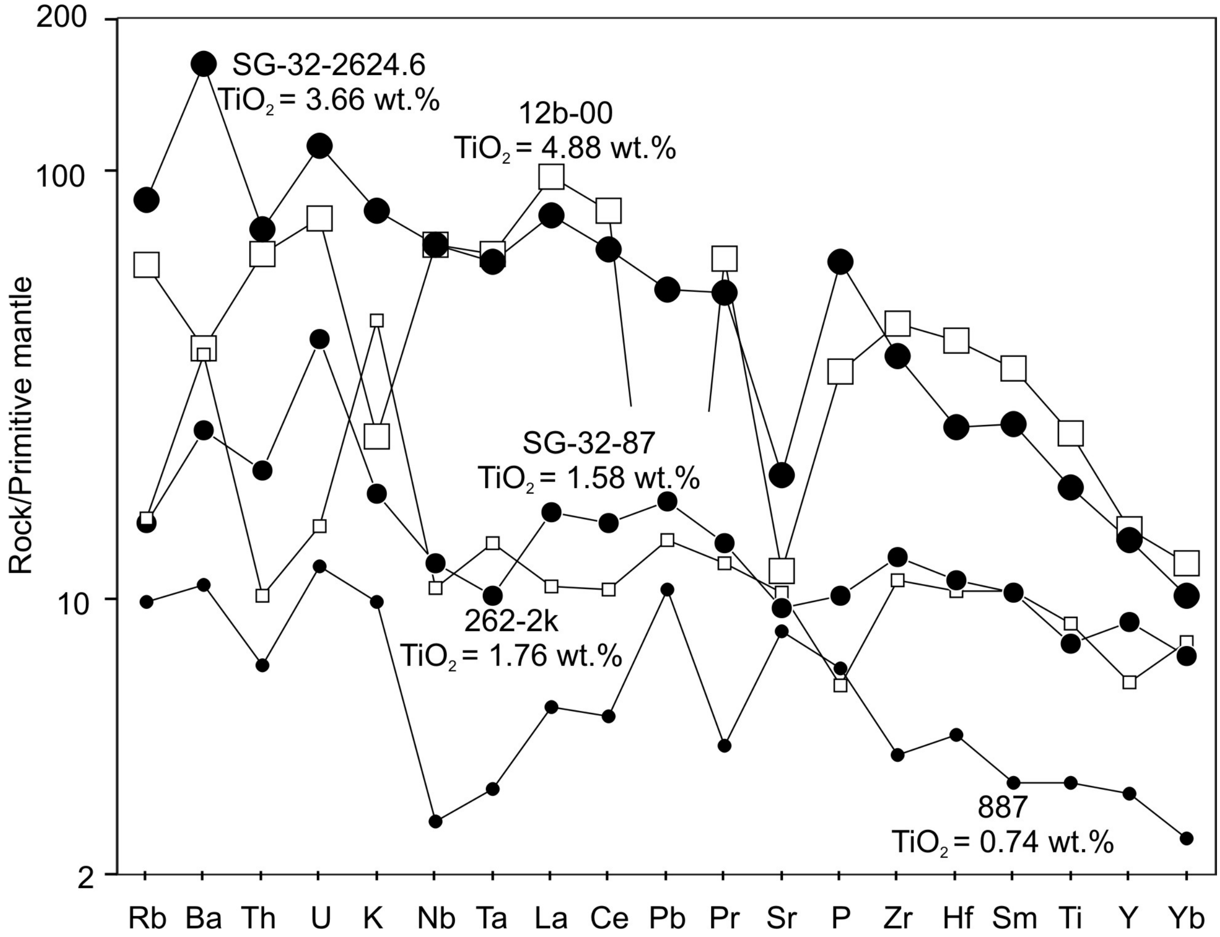

Fig. 4. Primitive mantle (McDonough and Sun, 1995) normalized diagram for incompatible elements with selected basalt samples from the YVFB (open squares) and SFB (filled circles) provinces. Sample numbers and TiO_2 concentrations are shown close to the corresponding spectra (see Table 1). Pb normalized concentrations are not shown for the sample 12b-00, because of suspicion for the analytical problem in Kiselev et al. (2012). Original data are after Wooden et al. (1993), Ivanov et al. (2008) and Kiselev et al. (2012).

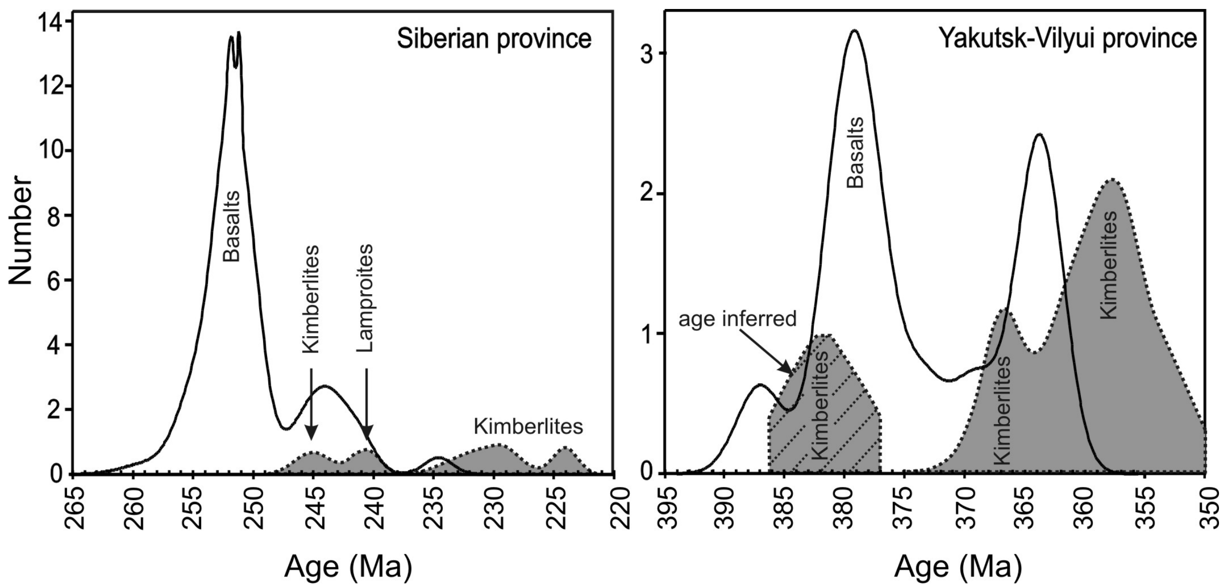

Figure 5. Temporal relation between mafic and kimberlitic (and lamproitic) volcanism of the YVFB and SFB provinces. Data for the YVFB and SFB is represented by $^{40}Ar/^{39}Ar$ and U-Pb age compilations provided by Ivanov et al. (submitted) and Ivanov et al. (2013), respectively. Whereas data for kimberlites are SHRIMP U-Pb ages on perovskites from Kinny et al. (1997). One $^{40}Ar/^{39}Ar$ age for a lamproite is from Ivanov et al. (2013). All $^{40}Ar/^{39}Ar$ ages are recalculated according to calibration of Renne et al. (2010; 2011), which allows direct comparison between $^{40}Ar/^{39}Ar$ and U-Pb ages. To plot age probability histogram, errors for the $^{40}Ar/^{39}Ar$ ages were set to 1% unless the analytical error exceeds this value.

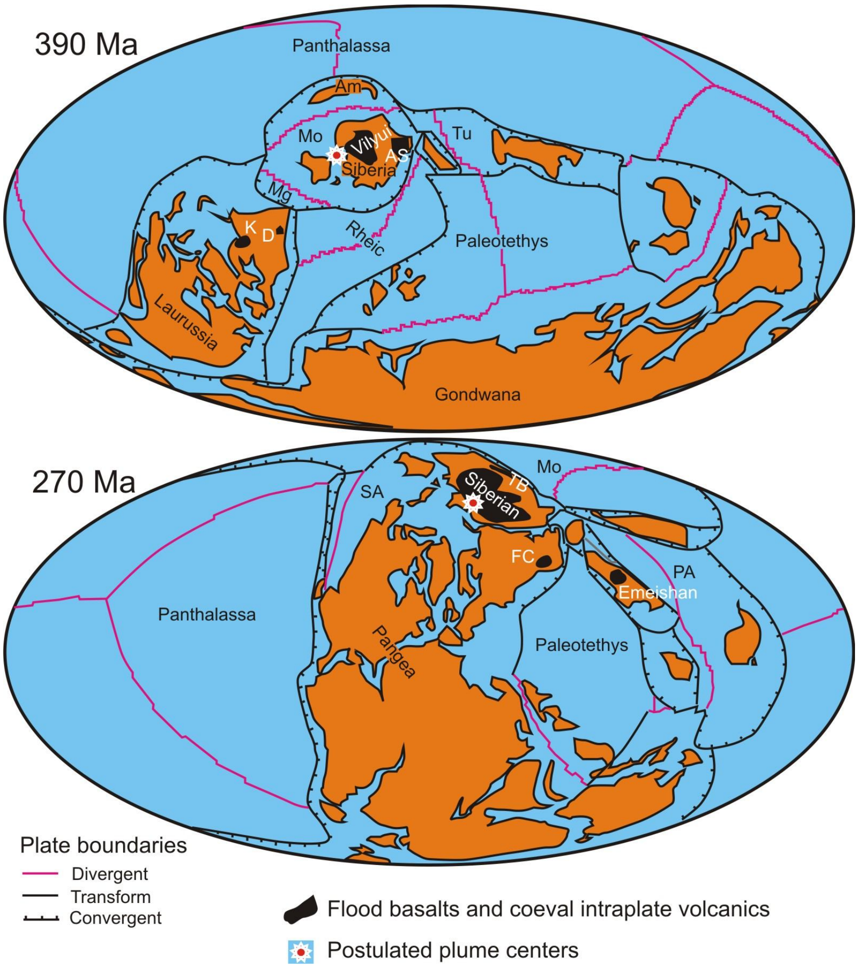

Figure 6. Paleogeographic reconstructions for 390 and 270 Ma (Domeier and Torvik, 2014), thus about 20 Ma before major phases of the YVFB and SFB volcanism, respectively. Other flood basalt and intracontinental volcanic provinces are Donbass (also named as Kola-Dneipr) (D, ~370 Ma), Kola (K, ~370 Ma), Altai-Sayan (AS, ~360 Ma), Emeishan (E, ~260 Ma), Transbaikalian (TB, ~260 Ma), and Fore-Caucasus (FC, ~230 Ma). Acronyms: Am – Amurian superterrane, Mg – Magnitogorsk arc, Mo – Mongolia-Okhotsk Ocean, Tu – Turkestan Ocean, SA – Slide Mountain – Angayucham Ocean, PA – Paleo-Asian Ocean. Names in white and black are for the volcanic provinces and plates, respectively. Postulated plume centers for the YVFB and SFB provinces are after (Kiselev et al., 2012) and Sobolev et al. (2011), respectively. According to Donskaya et al. (2013) subduction of the Mongolia-Okhotsk oceanic slab could start in the Middle Devonian, however Domeier and Torvik (2014) consider that subduction of this slab started later in the Early Carboniferous.

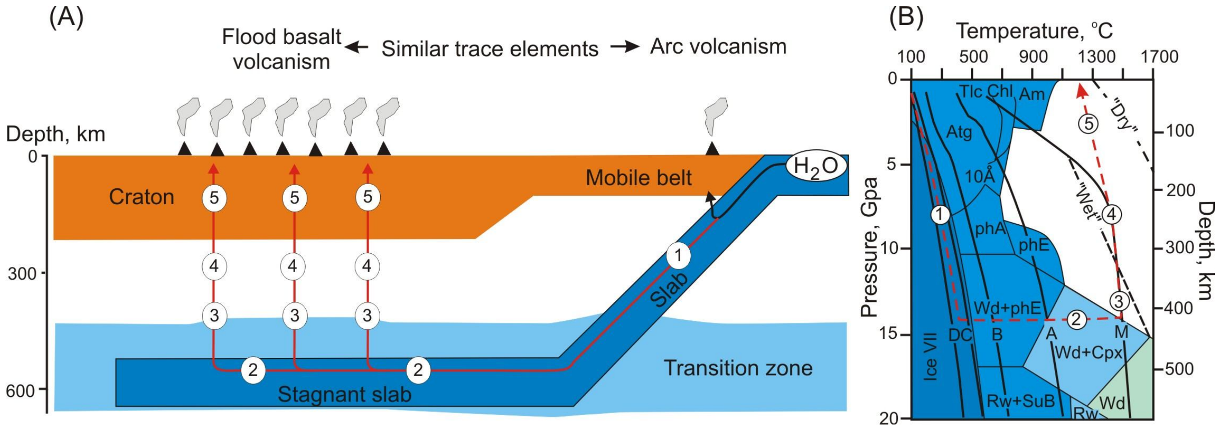

Figure 7. The deep water cycle model of origin of continental flood basalts with some simplifications after (Ivanov and Litasov, 2014). Numbered stages in (A) and (B) are the same. Acronyms in (B) are the following: Tlc – talc, Chl – chlorite, Am – amphibole, Atg – antigorite, 10A – 10 angstrom phase, phA – phase A, phE – phase E, Wd – wadsleyite, Rw – ringwoodite, Cpx – clinopyroxene. The stability field of ice VII is superimposed. Solid curves marked A, B, C, and D are for the coldest PT paths of the A, B, C, D type slabs of Kirby et al. (1996) calculated by Bina and Navrotsky (2000). M is for a 'normal' mantle geotherm (Turcotte and Schubert, 2002). Dashed curves marked 'wet' and 'dry' are 2 wt.% H_2O and dry peridotite solidi after (Hirschmann, 2000) and (Litasov and Ohtani, 2003), respectively. Intensity of blue is decreasing in decreasing order of water content. References for mineral stability fields see (Ivanov and Litasov, 2014).

Table 1. Selected basalt compositions from the YVFB and SFB provinces.

Province	YVFB		SFB		
Series	High-Ti	Low-Ti	High-Ti	Low-Ti	Low-Ti
Sample no.	12b-00	262-2k	SG-32-2624.6	SG-32-87	887
Reference	[1]	[1]	[2]	[2]	[3]
SiO_2	46.7	48.25	47.20	49.31	47.60
TiO_2	4.88	1.76	3.66	1.58	0.74
Al_2O_3	11.7	14.61	14.92	15.30	14.25
Fe_2O_3	-	6.32	-	-	2.52
FeO	-	7.12	-	-	7.72
FeO_t	16.11	-	13.50	12.55	-
MnO	0.16	0.20	0.24	0.19	0.16
MgO	5.66	6.51	4.03	6.77	10.38
CaO	7.71	11.74	9.45	11.27	12.76
Na_2O	2.15	2.41	3.41	2.32	1.76
K_2O	1.54	0.38	2.33	0.51	0.28
P_2O_5	0.7	0.13	1.26	0.21	0.14
LOI	1.02	1.21	4.86	2.89	1.34
Total	98.33	100.64	104.86	102.90	99.65
Rb	36.1	9.28	51.0	9.0	5.90
Sr	232	207	386	190	168
Y	62.7	27.6	59.0	38.0	15.2
Zr	460	116	386	132	45.4
Nb	44.0	6.98	44.0	8.0	2.00
Ba	253	246	1164	163	71.2
La	62.7	6.94	50.7	10.34	3.64
Ce	135	17.6	109	25.2	8.95
Pr	15.8	3.07	n.d.	n.d.	1.16
Nd	63.6	14.5	52.6	15.4	6.17
Sm	14.0	4.24	10.4	4.21	1.51
Eu	3.74	1.69	2.95	1.37	0.69
Gd	14.1	5.35	9.51	5.25	2.25
Tb	2.18	0.93	1.49	0.88	0.41
Dy	12.6	6.24	n.d.	n.d.	2.45
Ho	2.50	1.35	1.99	1.24	0.50
Er	6.23	3.91	n.d.	n.d.	1.52
Tm	0.92	0.54	0.75	0.52	0.23
Yb	5.33	3.52	4.48	3.25	1.2
Lu	0.74	0.47	0.65	0.48	0.20
Hf	11.3	2.95	7.1	3.14	1.37
Ta	2.35	0.50	2.25	0.38	0.13
Pb	0.53	2.06	7.88	2.53	1.6
Th	5.05	0.81	5.79	1.58	0.56
U	1.56	0.30	2.32	0.82	0.24
$^{87}Sr/^{86}Sr_T$	0.70452	0.70504	0.70575	0.70472	0.70587
$^{\varepsilon}Nd_T$	4.8	6.0	-0.2	1.9	n.d.

Refrences: [1] Kiselev et al., 2012; [2] Wooden et al., 1993; [3] Ivanov et al., 2008.