Iceland is fertile: The geochemistry of Icelandic lavas indicates extensive melting of subducted Iapetus crust in the Caledonian suture

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Abstract

The geochemistry and large melt volume at the Iceland volcanic province may be explained by extensive melting of subducted Iapetus crust trapped in the Caledonian suture that was recycled locally in the upper mantle beneath the Mid-Atlantic Ridge after continental rifting. Fractional remelting of abyssal gabbro explains the major-, trace- and rare-earth-element compositions, and the isotopic characteristics of primitive Icelandic tholeiite. An enriched component already present in the recycled crustal section in the form of enriched basalt and related intrusive material contributes to the diversity of Icelandic basalts. Basalt compositions ranging from ferrobasalt to olivine tholeiite are produced by various degrees of partial melt in eclogite, and the crystallization of ferrobasalt as oxide gabbro provides an explanation for the anomalously high densities in the Icelandic lower crust. We suggest that the deeper part of re-rifted suture zones delaminates from the rigid portion of the separating cratons and stays in the shallow mantle beneath new ocean basins, where it is later tapped along the spreading ridge. The melt-extraction anomaly is persistent at the Mid-Atlantic Ridge in the neighborhood of Iceland because the delaminated remnants of the suture there ~~runs~~ transversely to the Mid-Atlantic Ridge, which thus cannot have migrated laterally away from the trapped ocean crust in the suture. The large volume of basalt at Iceland is accounted for by the high melt fraction possible to obtain from eclogite plus the steep dip, imbrication or deformational thickening of the crust in the suture. This great diversity of basalts in the North Atlantic indicates that profoundly different mantle sources supply different regions. They cannot be explained by mere temperature variations acting only on homogenous source. A thermal plume is not supported by observations in the Iceland region, nor is it required by basalt geochemistry. Continental breakup often occurs along old suture zones. Thus similar processes may explain igneous provinces elsewhere that are traditionally attributed to plumes.

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1. Introduction

The Iceland volcanic province comprises a belt of anomalously thick crust that varies from ~ 250 - 600 km in north-south extent and crosses the entire width of the North Atlantic. It formed as a result of chronic, locally enhanced magmatism at the Mid-Atlantic ridge (MAR) where it crosses a transverse branch of the ~ 440-Myr-old Caledonian suture. The seismic crustal thickness of ~ 30 km suggests melt extraction at a rate up to 3 times greater than on the neighboring Kolbeinsey and Reykjanes ridges, where the crust is only ~ 10 km thick (Foulger et al., 2003).

Foulger (2002) and Foulger et al. (2003) suggest that enhanced melt production resulted from remelting eclogitised subducted Iapetus crust that was trapped in the shallow mantle beneath the Caledonian suture when the continents of Laurasia and Baltica collided at ~ 440 Ma. The continent has since been rifted, but much of the suture material that was in the mantle within and below the lithosphere~~,~~ is still present in the center of the North Atlantic and continues to influence the compositions of basaltic magmas along the MAR. Because the MAR crossed the suture branch in the future neighborhood of Iceland approximately orthogonally, buoyant masses of oceanic crust trapped in the upper mantle have been available at the ridge long after the continents were transported away, and much longer than to the north and south, where the proto-MAR formed longitudinally along other branches of the Caledonian suture.

Calculated compositions of parental melts, concentrations of trace- and rare-earth elements (REE) and ratios of radiogenic isotope provide evidence that remelted ocean crust of Caledonian age contributed to the compositions of basalts of from east Greenland, Iceland and Britain (Chauvel and Hemond, 2000; Korenaga and Kelemen, 2000; Breddam, 2002; Lesher et al., 2003). Subducted slabs comprise sediments, altered basaltic upper crust, gabbroic lower crust, and depleted lithospheric mantle. The crustal part combines contributions from normal mid-ocean-ridge basalt (N-MORB), enriched MORB (E-MORB), alkalic olivine basalt (AOB) and related differentiates such as occur on spreading ridges and nearby seamounts today. Such a variety of source material, combined with processes associated with remelting Icelandic crust (Oskarsson et al., 1982) accounts for both the petrological and geochemical variability of Icelandic basalts, and their exceptionally large volume.

The depth from which recycled crustal material might be derived is dependent on the depth to which subducted slabs finally sink. This is unknown and may be variable. One model suggests that all slabs sink to the core-mantle boundary from where the crustal material is transported back to the surface in plumes, to provide the enriched component observed in ocean-island basalt (OIB) (e.g., Hofmann and White, 1982; Kellogg and Wasserburg, 1990). There is little support from seismic tomography for such deep subduction. Maximum depths of subduction in the mid-lower mantle, and at 1,000 and 650 km have also been suggested on the basis of global tomography, geoid modeling and kinematics (van der Hilst et al., 1997; Wen and Anderson, 1997; Hamilton, 2002). Alternatively, hot, young slabs such as are subducted in the final stages of ocean basin closure may be trapped in the continental mantle lithosphere in the sutures of collided continents. Beneath the lithosphere they may achieve neutral buoyancy in the upper few hundred kilometers of the mantle (Oxburgh and Parmentier, 1977; Wortel, 1980; Meibom and Anderson, 2003).

The North-Atlantic geochemical anomaly within which the Iceland volcanic province lies was attributed by Schilling (1973) to the deep mantle plume that had been proposed to underlie Iceland (Morgan, 1971). Schilling (1973) proposed that the plume was hotter than beneath ridges to account for the large volume of Icelandic basalts. This hypothesis subsequently experienced complications. The widths and magnitudes of some of the geochemical anomalies compared with those of the Azores, which were also proposed to be underlain by a plume, are the reverse of what is expected for the much larger Iceland melt anomaly (Schilling et al., 1983). The compositions of Icelandic lavas cannot be explained by the original two source "components" (mantle peridotite and an enriched "plume" component) suggested by Schilling (1973). Different studies advocate different numbers and combinations of the suite of components proposed (e.g., Stracke et al., 2003), including North Atlantic depleted mantle, fertile peridotite, both "enriched" and "depleted" plume components, and an additional component to explain the helium isotope ratios. The expected radial symmetry about central Iceland of geochemical signatures attributed to a plume (e.g., Condomines et al., 1983; Schilling et al., 1983) is not observed, geochemical discontinuities occur across relatively minor structures such as the 120-km-long Tjornes Fracture Zone in north Iceland, and the temperatures of the most primitive Icelandic melts are similar to those calculated for mid-ocean ridge basalts (Breddam, 2002).

We propose that the geochemistry of both primitive and differentiated Icelandic tholeiites can be explained by extensive melting of a complete section of subducted oceanic crust in the eclogite facies, and in particular that primitive Icelandic basalts are derived from remelted abyssal gabbro (Chauvel and Hemond, 2000; Breddam, 2002). Chemical comparisons show that the "depleted plume component" (Anderson, 1994; Kempton et al., 2000) in Icelandic basalts is similar to the geochemistry of abyssal olivine gabbro, and that Icelandic ferrobasalts are similar to abyssal oxide gabbros. The "enriched plume component" may be derived from remelting axial or seamount E-MORB, alkalic olivine basalt, and associated intrusive rocks, from melting of sedimentary materials of the subducted crust, or possibly small amounts of ancient continental crust that may, as at Jan Mayen, still underlie portions of Iceland (Amundsen et al., 2002; Foulger and Anderson, 2003).

High temperatures and a deep mantle source, neither of which are observed, are unnecessary. The derivation of up to 30 km of melt from remelting of ~ 7-km-thick crust suggests that the subducted slab dips steeply or is imbricated, and by this means derived melts are concentrated beneath a smaller area than if the slab were flatter. We discuss possible slab geometry in the context of the late stages of closure of the Iapetus Ocean.

2. The end game of plate tectonics

The formation of ocean crust with normal thickness, uniform spreading at ridges and the subduction of old, dense lithosphere to at least the base of the upper mantle are aspects of steady-state plate tectonics that may not occur during the opening and closing stages of oceans and the formation of collisional sutures. The opening stage is often associated with bursts of magmatism that build seaward-dipping reflectors, thick sequences of flood basalts and large igneous provinces. Lateral temperature gradients and edge-driven convection may be important (King and Anderson, 1995; King and Anderson, 1998; Boutilier and Keen, 1999). When continental break-up occurs along old sutures, as it did in the North Atlantic, magmatism may be further enhanced as a result of mantle made unusually fertile and fusible by eclogitised subducted oceanic crust trapped in the rifting lithosphere. Ridge-trench collision in the final stages of coalescence also introduces sediments and water into the shallow mantle at the suture.

The final oceanic lithosphere consumed when oceans close will be young, thin, and hot near ridges and beneath back-arc basins (Meibom and Anderson, 2003). Such lithosphere is buoyant, evidence for which is found in the obduction of ophiolites during the terminal stages of continental collision, and flat subduction of young lithosphere. Both low-angle subduction observed tomographically for young lithosphere, and thermal modelling, suggest that if oceanic lithosphere is younger than ~ 50 Myr it may sink no deeper than a few 100 km (Oxburgh and Parmentier, 1977). At a half-spreading rate of 1 cm/a, this would amount to 500 km of plate. Much young subducted lithosphere, including ocean crust, could thus be retained in the shallow mantle, along with mantle wedge material and dehydration fluids. Old, thick, cold lithosphere probably sinks to much greater depths.

Where might this material reside? A length of the final subducting lithosphere equivalent to the thickness of the colliding cratons, or up to ~ 150 - 200 km (Polet and Anderson, 1995), would be trapped between them and retained in the continental lithosphere. The remainder of the late-subducting, buoyant oceanic lithosphere below this, perhaps up to several 100 km, might be retained in the asthenosphere as flat slabs beneath the sutured cratons.

The formation of the Caledonian suture at ~ 440 Ma involved the unusual collision of three continents. Laurentia and Baltica collided in an east-west direction, closing the Iapetus Ocean between them, and northerly drifting Avalonia docked against their southern borders when Tornquist's sea closed (Dewey and Shackleton, 1984; Soper et al., 1992) (Figure 1). This supercontinent broke up again at ~ 54 Ma with the formation of the new MAR. The northern part of the new MAR ran longitudinally along the Caledonian suture, but in the area where the Iceland volcanic region later formed, the MAR crossed the southwesterly branch of the suture, which extends from east Greenland to Britain, where a southerly dipping slab had been subducted. High rates of magmatism for the first few Myr built a volcanic margin with sequences of basalt including seaward-dipping reflectors up to ~ 25 km thick (e.g., Keen and Potter, 1995). At ~ 64˚N, magmatism did not wane but continued at a high level to the present day, and built a band of crust ~ 30 km thick traversing the Atlantic from east Greenland to the Faeroe Islands (Foulger and Anderson, 2003).

These geological observations are consistent with the hypothesis that the source of excess melt lies in Iapetus crust trapped at shallow depth in the Caledonian suture. The thick continental crust was ruptured and carried away to either side of the North Atlantic, to Greenland and northwestern Europe, but much of the underlying sub-cratonal lithospheric mantle and suture zone material on both sides still remains beneath the North Atlantic. The Iceland volcanic region itself began to form at a location where 150 - 200 km of slab was trapped between the cratons, and modern Iceland overlies the remnants of this material. In this geometry, the North Atlantic to the south of the Iceland region may also contain Iapetus slab material abandoned in the upper mantle, which might explain the continuation of geochemical anomalies south down the

Reykjanes ridge from Iceland. This, along with a moderate temperature anomaly of 50-100˚C resulting from continental insulation prior to breakup (Anderson, 2000b), may explain the geochemical, bathymetric and geoid highs there (Foulger and Anderson, 2003).

3. Geochemistry

Most Icelandic basalts are tholeiites similar to N-MORB, and differ from them only slightly in their isotopic ratios, trace-element concentrations, and major-oxide abundances (Meyer et al., 1985; Hanan and Schilling, 1997; Kempton et al., 2000). Most compositions are higher in fractionation-corrected parental Na_2O (Na_8) (Figure 2a) and TiO_2 than those from the adjacent ridges. Partial melting of a homogeneous mantle to produce thick crust (Figure 2b) would predict less Na_2O and TiO_2 because of the greater degree of partial melting and larger depth range of melting required (Langmuir et al., 1992). Nevertheless, even above the thickest (~ 40-km) crust in central Iceland (Foulger et al., 2003), the primitive lavas found at Kistufell volcano (Breddam, 2002) have Na_8 similar to basalt glasses from the East Pacific Rise (EPR), where the crust is only 7 km thick. Below, we use the primitive Kistufell basalts as type examples of primitive Icelandic tholeiite in comparisons with oceanic crust.

Eruptive temperatures calculated for Kistufell lavas are only ~ 1240˚C (Breddam, 2002), close to those of similarly magnesian N-MORB (Ford et al., 1983). Heat flow in the ocean surrounding Iceland is no higher than elsewhere in sea floor of the same age, which also indicates normal temperatures (Stein and Stein, 2003). Petrologic indicators of high temperatures, e.g., very high Mg numbers, forsteritic olivine, and picritic glass, are absent. The attenuation of seismic waves and Vp/Vs ratios in the lower crust are low, indicating subsolidus conditions and temperatures lower than at equivalent depths beneath the EPR (Menke and Levin, 1994; Menke et al., 1995; Menke et al., 1996). A theory for Icelandic petrogenesis is thus required that explains the geochemical similarities to typical MORB and excessive crustal thickness with eruptive temperatures similar to normal mid-ocean ridges.

As suggested by Chauvel and Hemond (2000) and Breddam (2002), primitive Icelandic tholeiite is similar to gabbroic ocean crust. We extend this comparison using large data sets from Hess Deep in the eastern Pacific, and Atlantis Bank on the Southwest Indian ridge. At the latter, a very long section of gabbroic oceanic crust was cored at ODP Hole 735B. At both locations almost all gabbros are adcumulates with little trapped melt (Natland et al., 1991; Natland and Dick, 1996; Natland and Dick, 2001; Natland and Dick, 2002). They span all stages of differentiation including, with decreasing temperature, troctolite, olivine gabbro, gabbronorite, oxide gabbro, and minor residual granitic material. The oxide gabbros from Hole 735B contain up to 11% TiO_2 and 30% cumulus magmatic oxides. About 20% of the section is gabbro with 1-30% of ilmenite and magnetite. At pressures > 1.5 – 3.0 GPa these would be rutile-rich eclogite (e.g., Yoder and Tilley, 1962).

The diverse gabbroic facies at Hole 735B experienced differentiation by deformation (Bowen, 1920) during which the diverse lithologies were intimately juxtaposed in complex fashion while melt was present. Crystsallization temperatures ranged from ~1200˚C to ~700˚C (Natland et al., 1991). A similarly wide range of remelting temperatures would be expected for granitic to troctolitic rocks transformed to the eclogite facies. Experimental studies on natural eclogites (Yoder and Tilley, 1962; Ito and Kennedy, 1974) indicate a crystallization/melting interval of 80 - 90 K at pressures of > 2 GPa, compared with ~200 K for gabbro of the same composition at < 1 GPa. Most experiments, however, have been done on bimineralic, garnet-clinopyroxene eclogite rather than either iron-rich ferroeclogite with significant proportions of rutile ± amphibole or granitic material in the eclogite facies. These undoubtedly would extend the temperature interval of crystallization/melting for a bulk assemblage of ocean crust in the eclogite facies.

Trace-element concentrations in gabbroic cumulates are controlled by the compositions of precipitated minerals rather than the liquids. Concentrations in individual rocks may be variable, but the ratios of elements and their average concentrations in the complete crustal section are determined mainly by partitioning into clinopyroxene and plagioclase, and crystal separation from the residual liquid fraction. The bulk gabbro assemblages at Hess Deep and Hole 735B thus have low concentrations of, e.g., Y, Zr, and TiO_2, in proportions set primarily by clinopyroxene, compared with their parental liquids (Table 1).

We calculate an average bulk composition for 1508 m of gabbro from Hole 735B using all analyses for which major oxides, REE and many trace elements have been obtained (Natland and Dick, 2002), and weighting the analyses by the density-corrected mass proportion of each lithology in the hole (Dick et al., 2000). The section probably represents about 60% of the gabbroic layer, including its central, most typical, portion. Complete remelting of all lithologies within it would probably leave only about 10% of the most refractory crustal materials behind (which were not cored at Hole 735B) (Natland and Dick, 2002). This composition (Table 1) would provide a liquid that is magnesian but non-picritic, with lower concentrations of Y, Zr, and TiO_2 than typical primitive N-MORB. In all these respects such a liquid is similar to primitive Kistufell tholeiite (Table 1).

The REE pattern of this average Hole 735B gabbro is also similar to those of Kistufell basalts, being flat and with only a small Eu anomaly (Figure 3). The low overall REE concentrations calculated are a consequence of partitioning into cumulus minerals, and the expulsion of intercumulus melt. Average Hole 735B gabbro is, however, slightly depleted in light REE compared with Kistufell tholeiites. Icelandic tholeiites are also lower in Zr compared with REE than in N-MORB, one of the attributes assigned to the so-called "depleted plume component" (Kempton et al., 2000). However, this attribute is also matched by most troctolitic and olivine-gabbro cumulates at Hole 735B (Figure 4).

Average Hole 735B gabbro has lower Zr/Y and Nb than either Kistufell tholeiite or typical N-MORB (Figure 5). The lower Zr/Y again is a consequence of the rocks being cumulates in which concentrations of these elements are controlled mainly by clinopyroxene. Both the higher Nb and light REE in the Kistufell tholeiite and N-MORB may be explained by the addition of or mixing with a small fraction of an enriched (E-MORB or AOB) component, or perhaps silicic material such as trondhjemite or rhyolite, with liquid derived from melting abyssal gabbro. All of these are already present in recycled ocean crust by analogy with rocks from the EPR and nearby seamounts (Natland, 1989; Niu et al., 2002). The high Nb in basalt glasses from the EPR reveals an E-MORB influence among many N-MORB glasses (Figure 5) and we expect a comparable geochemical range to be retained during remelting of subducted oceanic crust. The enriched component is distributed as lavas, dikes, veins, cumulates and reaction zones in ocean crust and adjacent abyssal peridotite. It need not reside in a physically separate reservoir, nor come from the deep mantle. Light REE flattening and Nb enrichment in Kistufell tholeiite can be explained by addition of 4.8% of seamount-type AOB to average Hole 735B gabbro (Table 1; Figures 3-5). This is slightly less than the 6.2% of E-MORB with > 0.5% K_2O dredged on the EPR and nearby seamounts between Siqueiros and Clipperton Fracture Zones, according to the Lamont Petrology Database. It can also be explained by mixing with a similarly small amount of rhyolite which is present at many central volcanoes in Iceland.

Additional similarities between average Hole 735B gabbro and the Kistufell basalts include positive Sr and negative Pb anomalies, low-^{18}O (Breddam, 2002), and elevated $^{87}Sr/^{86}Sr$, which results from seawater alteration. This is associated with elevated Rb in altered gabbros (Hart et al., 1999). By radioactive decay of Rb over several 100 Myr, the average $^{87}Sr/^{86}Sr$ of trapped Iapetus ocean crust would increase somewhat from the original values in abyssal gabbro. Other isotopic ratios suggest an age for sources of basalts of the Iceland volcanic province of several hundred Myr, consistent with an origin by remelting of ocean crust dating from the Caledonian suture (Korenaga and Kelemen, 2000; Lesher et al., 2003).

Great abundances of differentiated ferrobasalts are observed in Iceland, particularly at central volcanoes (Carmichael and McDonald, 1961; Walker, 1963), and the diverse rock compositions there have been attributed to remelting of the thick Icelandic crust (Oskarsson et al., 1982). Ferrobasaltic liquid cannot be produced by partial melting of mantle peridotite, as liquids so produced must be in equilibrium with magnesian olivine and pyroxenes. Primitive olivine tholeiite, however, occurs mainly in small volumes along the rift zones of Iceland, always away from central volcanoes. The ferrobasalts of both the rifts and central volcanoes could alternatively be derived from partial melting of the *eclogitic* mantle source beneath Iceland, in which case a great deal of differentiated basalt could cross the mantle into the crust. Thus, significant quantities of oxide gabbro, carrying several percent or more of the dense magmatic oxide minerals, ilmenite and magnetite, may have crystallized in the Icelandic lower crust. This may explain the anomalously high density of the Icelandic lower crust. Isostasy indicates that average Icelandic lower crust has an unusually high density only ~ 90 kg/m^3 lower than that of the underlying mantle (Menke, 1999), despite having a crust-like average compressional-wave velocity of ~ 7.2 km/s. This may be compared with a density contrast of ~ 300 kg/m^3 expected if the lower crust were olivine gabbro and the mantle peridotite. On the other hand, because abyssal oxide gabbros have so much ilmenite and magnetite, they have very high densities of up to 3,200 kg/m^3, but crust-like seismic velocities (Itturino et al., 1991).

Partial to nearly complete remelting of eclogitic ocean crust, transformed from a typically diverse suite of abyssal gabbros and basalts, can thus account for both primitive and differentiated Icelandic tholeiites. During transformation to eclogite in subduction zones, trace-element concentrations and ratios in abyssal gabbros plausibly are preserved and survive later remelting. Radiogenic isotopes that evolved following subduction add to those in E-MORB, dikes and reaction zones already present in the subducted crust, as well as altered ocean crust. Recycled abyssal gabbro can explain the high Y/Zr, low Zr/La, low concentrations of REE and high-field-strength incompatible elements such as Zr, Y, Ti, Hf and Ta, positive Sr and negative Pb anomalies, flat REE patterns, MORB-like general characteristics, non-picritic MgO and Mg#

of most primitive Icelandic basalts (Chauvel and Hemond, 2000), and eruptive temperatures of ~ 1240˚C (Breddam, 2002). More enriched and differentiated Icelandic basalts result from a smaller extent of eclogite partial melting, that taps the iron-rich oxide-gabbro in the suture, and incorporation of AOB and related rock, which must also reside in the subducted crust and adjacent abyssal peridotite. Some characteristics of primitive Icelandic basalts previously attributed to alteration and sub-arc melt extraction from ocean crust (Breddam, 2002), e.g., Y, Nb, Zr, and TiO_2 proportions, positive Sr, and negative Pb anomalies, instead mimic the same features in abyssal gabbro cumulates. Those features thus do not amount to proof that ocean crust in a mantle source has been subducted past a line of arc volcanoes and then recycled back from the lower mantle.

4. Helium isotope ratios

The highest $^3He/^4He$ isotope ratios at a currently active hotspot, up to ~ 42 Ra, (where Ra is the atmospheric $^3He/^4He$ ratio) are found in Iceland (Hilton et al., 1999; Breddam and Kurz, 2001). (Recently, ratios of up to 49.5 ± 1.5 Ra have been reported from basalts in Baffin Island (Stuart et al., 2003).) Such observations are widely considered to be an unambiguous indicator of a lower mantle component in surface rocks, partly because it is assumed that this results from high ^3He content, and partly because other regions assumed to be underlain by plumes, such as Yellowstone and Hawaii, also have high ratios. However, this interpretation is non-unique. Other models for derivation of high $^3He/^4He$ from the upper mantle have been proposed (Anderson, 1998b; Anderson, 1998a; Foulger and Pearson, 2001; Meibom et al., 2003).

The presumption that high $^3He/^4He$ is derived from the lower mantle assumes that the lower mantle is essentially undegassed. A high absolute concentration of ^3He would have prevented $^3He/^4He$ ratios from reducing with time in response to the production of ^4He from the decay of U+Th over the lifetime of the Earth. The value of $^3He/^4He$ in the lower mantle is thus predicted to have remained at a few 10s of Ra, and to provide a source for the high $^3He/^4He$ observed at Iceland and elsewhere. This theory has the following problems:

1. Mass balance calculations show that the concentration of ^3He predicted in the lower mantle is comparable with that in chondritic meteorites (Kellogg and Wasserburg, 1990). This is at odds with high-temperature models of planetary accretion. The Earth is thought to have accreted at high temperature because of the energy of accretion, segregation of the core, and impact with a Mars-sized body that caused the Moon to form. Extensive degassing must have occurred at these times, which is supported by the observation that the Earth is strongly depleted in cosmochemically volatile species such as Na, K, Cl, and Rb, which are even less volatile than He. A large part of the Earth is thus not likely to have remained undegassed (Anderson, 1989).

2. If high-$^3He/^4He$ arises from a component with a high absolute He concentration then the helium concentrations in OIB, where such high ratios are commonly observed, are expected to be much higher than in MORB. However, the helium abundances in OIB are 2 – 3 orders of magnitude *lower* than in MORB. It is generally assumed that this results from preferential degassing of OIB as a result of their shallow depth of eruption. This can be tested by looking at

the relative concentrations of heavier noble gases. The heavy noble gasses have a greater tendency to degas upon eruption than helium, so He/Ne and He/Ar in OIB should be higher the more degassed a magma is. It is found, on the contrary, that these ratios are higher in MORB than in OIB, suggesting that MORB is more degassed than OIB, not less (Anderson, 1998b; Anderson, 2000a; Moreira and Sarda, 2000; Ozima and Igarashi, 2000). OIB therefore do not have high $^3He/^4He$ because they arise from a source intrinsically high in helium (Anderson, 1998b; Anderson, 1998a).

High-$^3He/^4He$ can also result from a deficit of ^4He as a result of storage of old helium in a low time-integrated U+Th host rock or mineral (Anderson, 1998b; Anderson, 1998a). The controlling parameters are then U/^3He, U/Th and time.

Possible low-U+Th host materials are the residuum left after basalt melt is extracted from mantle peridotite (Anderson, 1998b; Anderson, 1998a) and any ultramafic cumulate (Natland, 2003). Either contains only traces of U+Th and thus helium stored in such rocks for hundreds of millions or even billions of years would preserve its older, higher $^3He/^4He$ ratio with little change. Olivine-rich cumulates of various ages are present in both sub-cratonic and oceanic mantle lithosphere, and both are therefore likely hosts for volatiles released during remelting in the complex mantle beneath Iceland. Olivine crystals in cumulates contain encapsulated gas bubbles with He, but they contain essentially no U+Th because melt fractions are excluded by the mechanism of incorporation of inclusions during crystal growth (Natland, 2003). The minerals will aggregate as cumulates where intercumulus melts are expelled during compaction and subsequent adcumulus growth. Cumulates thus are time capsules that preserve unchanged any initial $^3He/^4He$ ratio, and they are not restricted to reservoirs derived from the deep lower mantle. They could readily reside in ancient subcontinental or oceanic lithosphere.

The volatiles in olivine-rich cumulates probably would be the *only* materials scavenged from them during later partial remelting, and thus high $^3He/^4He$ would not necessarily correlate with other indices of isotopic enrichment in Icelandic tholeiites, a continuing conundrum of Icelandic isotope geochemistry (Stracke et al., 2003). Indeed, depending on how volatiles entrained in this way concentrate in ascending magmas or temporary storage areas in their progress toward eruption, there is no formal reason why even the absolute concentration of ^3He should be low in any given lava.

Our model specifies that residual mantle and lowermost oceanic crust of Caledonian age are trapped between delaminated remnants of thick blocks of ancient subcratonal mantle in the shallow mantle beneath Iceland. Any of the ultramafic materials plausibly could contribute moderate- to high $^3He/^4He$ to Icelandic tholeiite depending on the age and distribution of the sources. Quantitative calculations of the $^3He/^4He$ values in such rocks are difficult because the details of degassing, metasomatism and radiogenic ingrowth cannot be accurately determined. The strength of the helium argument for an Icelandic deep mantle plume has previously been the necessity to tap an ancient, long-isolated, undegassed reservoir that is clearly not the same as the modern MORB reservoir. We suggest here that the ancient reservoir resides in mantle near the surface of the Earth. This model does not require high ^3He or a deep mantle plume to explain high $^3He/^4He$.

5. Physical models for melting subducted crust

The extent to which subducted crust trapped at shallow levels in the mantle re-homogenizes with its peridotite host is not known. The retention of essentially pristine blocks of crust of the order of kilometers in thickness, and complete homogenization with mantle peridotite, represent end-member scenarios. We have focused on the former possibility, and shown that the geochemistry of lavas from the Iceland volcanic province can be matched by large-scale remelting of pristine, eclogitized, subducted Caledonian crust, with contributions from adjacent peridotite. There is evidence that the necessary km-scale lengths of pristine blocks are maintained, even in a convecting mantle (Kellogg et al., 2002; Meibom et al., 2003).

How much melting of an eclogitic source is required? Experiments on eclogite and in simple four- or five-component systems (without TiO_2) show that initial small-degree melts are similar to andesite (Yoder and Tilley, 1962). As the degree of batch melting increases, the melt compositions progressively approach the starting composition of the sample being melted. At 10-30% partial melting, liquids are ferrobasaltic with low SiO_2, and at 80% they are olivine tholeiite similar to those of Iceland (Ito and Kennedy, 1974). The effect of higher pressure is to reduce SiO_2 in the basaltic partial melts and the proportion of normative hypersthene in liquids (Ito and Kennedy, 1974). These effects are the same as differences observed between the average bulk composition of abyssal gabbro and that of primitive Icelandic tholeiite. Even at the extremes of eclogite partial melting, however, when contributions from adjacent peridotite might be most significant, the geochemical signature of abyssal gabbro still predominates. Strict comparison of Icelandic tholeiite to natural gabbro compositions, and the experiments of Ito and Kennedy (1974), suggest that 60-80% melting of the original bulk gabbroic assemblage is required.

How might such high degrees of melting be attained? A subducted eclogite slab will thermally re-equilibrate after residing in the mantle for ~ 10 thermal relaxation times (e.g., Stein and Stein, 1997). The thermal relaxation time of subducted lithosphere is 10 – 20 Myr, and thus 440 Myr is ample for Caledonian slabs to have attained nearly the same temperature as the surrounding mantle. Thus, eclogite upwelling beneath the MAR in the Iceland region will be much closer to its solidus than peridotite, or it may even be partially molten. Experimental work shows that, considering only the principal silicate phases (clinopyroxene, orthopyroxene, plagioclase and olivine at low pressure, and garnet and clinopyroxene in bimineralic eclogite at high pressure) the crystallization/melting interval for eclogite (80 – 90 K) is smaller than for gabbro (> 200 K), and much smaller than for peridotite (~ 400 K) (Yoder and Tilley, 1962; Ito and Kennedy, 1974). No melting experiments have been conducted on the eclogitic equivalent of a very oxide-rich ferrogabbro (rutile-rich eclogite) or on high-pressure metasediments in the eclogite facies. However, we conclude that a melt equivalent to a moderately differentiated abyssal tholeiite can readily be generated from an average eclogitic abyssal gabbro, and partial melting beyond a few tens of percent will not greatly affect the generally basaltic composition.

Differentiated oxide-rich, amphibole-bearing or granitic lithologies present in the ocean crust may add significantly to the melting interval and would diversify the geochemistry. Nevertheless, the large degrees of melt suggested by the geochemistry could be attained by heating eclogite to approximately the same temperature as is required to produce 20% of partial melt in peridotite (Yaxley, 2000). The difference between 20% partial melting, producing a liquid approximately equivalent to ferrobasalt, and 80% partial melting, producing olivine tholeiite (Ito and Kennedy, 1974), may only correspond to a few tens of ˚C temperature difference at a given pressure.

How could such large degrees of melt be retained prior to eruption? Melt extraction from partially molten rock can begin at degrees of melting of less than one percent (e.g., McKenzie, 1984). Consequently, progressively extracted melt increments derived from eclogite must pond and re-homogenize in some reservoir, probably below the base of the crust, prior to eruption. A similar process of fractional melting, aggregation, and homogenization is also required beneath normal spreading ridges since MORB is thought to be formed by up to ~ 20% partial melting of peridotite integrated over a melt column. Incremental melt extraction, however, may not operate efficiently over a large melting domain, thus the extent to which ideal fractional melting can operate is uncertain (Natland, 1989).

Another possibility, raised by the experiments of Yaxley et al. (1998), is that the subducted slab completely homogenizes with surrounding peridotite and that most or all remelting takes place in a particularly fertile peridotite assemblage. The homogenization could occur by progressive incorporation of early, small melt increments leaked from an inclined mass of eclogite into overlying peridotite. The hybrid can then be considered as peridotite with a shift in bulk composition toward basalt, or what Anderson (1989) has termed "piclogite". However, much homogenization of this type may not be important for the slab material trapped in the cold cratonic lithosphere.

Nevertheless, the thermal and mass-balance aspects of this case are useful to consider. At temperatures normal for extraction of MORB along other parts of the Mid-Atlantic Ridge, a homogenous mixture of peridotote completely combined with a few tens of percent of subducted crust can satisfy the volumetric requirements for Iceland (Yaxley, 2000; Foulger and Anderson, 2003). The basaltic products of partial melting would be similar to those derived from normal peridotite but larger in volume.

Whether passive, isentropic upwelling of eclogite, an eclogite/peridotite mixture, or a wet piclogite/peridotite assemblage would release sufficient energy to provide the required latent heat of melting to produce the thickness of crust at Iceland is not known. This cannot be calculated yet because the entropy values for the relevant minerals at the relevant temperatures and pressures are not known. Nor do we know exactly how much melt must be explained. Although the seismic crust (i.e., the layer with compressional-wave-speed $Vp < \sim 7.2$ km/s) is three times thicker than that beneath the Reykjanes ridge, isostatic considerations indicate that the Icelandic lower crust has a density very close to that of mantle peridotite, which might thus comprise a significant component of the lower crust through entrainment (Foulger and Anderson, 2003). The amount of melt that needs to be explained is somewhere in the range 15 – 30 km, or 1.5 – 3 times the crustal thickness on the Reykjanes and Kolbeinsey ridges. The ambient temperature of the upper mantle throughout the whole North Atlantic may be slightly higher than that of the average mantle as a result of the long period of continental insulation prior to opening of the North Atlantic at ~ 54 Ma, and lack of recent subduction cooling.

6. The geometry of slabs in a suture

In the extreme case where 100% of the 15 – 30 km of crust at Iceland results from remelting subducted crust, at least this amount of crust is required from the uppermost mantle. Remelting of a single average 7-km thickness of ocean crust is thus insufficient. Thickening of slab material, whether by increased slab dip or imbrication, is also required. In the simplest case, where a single, intact dipping slab is trapped, continental collision will steepen it to a high angle (Figure 6a). Archaean cratons may be ~ 150 - 200 km thick, which is a measure of the depth extent of the sutures that form when they collide. A slab caught in such a steep orientation will produce much more than a single thickness of crust. If, in addition, the slab bends, then the amount of melt rising beneath a particular location will increase down dip. The sense of thrust fronts in Greenland and Scandinavia suggests that the slab in the suture over which the Greenland-Iceland-Faeroe ridge formed dipped to the south. The north-south asymmetry of this structure may thus be responsible for the north-south tectonic and geochemical asymmetry observed in the Iceland volcanic province at the surface (Foulger and Anderson, 2003).

The trapped slab may also be contorted, deformed, faulted and perhaps broken into segments and imbricated, in a style similar to that of the sinistral Caledonian strike slip faults in northern Britain (Soper et al., 1992) (Figures 1 and 6b). Such imbrication could also increase the total thickness of underlying subducted crust at a given place, thereby increasing the potential amount of additional basalt locally (Carswell et al., 1999). Structure in the mantle~~,~~ inherited from the Caledonian suture~~,~~ may be the source of the surface tectonic and crustal structural complexities observed, such as the persistent spreading about a parallel pair of ridges, the distribution of volcanism in Iceland, and the abrupt changes in crustal thickness and geochemistry that occur across the Icelandic shelf edge (Foulger and Anderson, 2003).

Magma focusing and other three-dimensional effects might also occur. The MAR plate boundary at Iceland is diffuse (Zatman et al., 2001; Foulger and Anderson, 2003), and fusible material is expected to be widespread throughout the suture at depth. Melt may thus be extracted from a broader region than directly beneath the ridge, increasing melt production. Whether or not such a process is necessary will depend on eventual calculation of the heat budget of isentropic upwelling of eclogite.

7. Summary and extension of the model

The slab trapped in the suture is expected to be the last ~ 150 - 200 km of oceanic lithosphere that was subducted. Upon final collision, the descent of this material stops. The trapped slab will contain marine pelagic and biogenic sedimentary material, possibly along with hemipelagic or mature continental sediment, portions of island arcs, backarc basins, and the depleted abyssal peridotite underlying the slab. Mantle wedge material partially hydrated by fluids from the subducting slab~~,~~ may accrete to the base of the craton. Mantle wedge volatiles may be needed to explain some widespread OIB geochemical signatures (Smith, 2003).

Continental breakup and rifting frequently occur along pre-existing sutures, where they are accompanied by high rates of magmatism (Smith, 1993). The remelting of subducted oceanic crust trapped in these sutures contributes significantly to melt production. The question then arises, how does the deeper part of the suture remain in the region of melting as the ocean basin widens? We propose that when rifting commences, only the cooler, shallower, most rigid and generally granitic portion of the sutured lithosphere remains intact and is transported away with the cratons. Much of the hotter, ductile, lower portion of the slab and adjacent peridotite delaminates, stretches, and thins, staying in the mantle beneath the new spreading ridge. Pressure-release melting, resulting both from unroofing and thinning, affects all exhumed lithologies, whether they are ultramafic or eclogitic. The new crust that is produced is oceanic in the sense that it is basaltic rather than granitic, and it is unusually thick, but the source of the basalts is not depleted MORB mantle.

Where rifting occurs longitudinally along a suture, the most common situation, the early phase of eclogite-fuelled intense magmatism may last for up to ~ 10 Myr. Where rifting occurs obliquely or transversely to a suture, however, enhanced magmatism may persist for much longer. This may have occurred at Iceland, the New England seamounts, and Tristan da Cunha (Smith, 1993). Long-lasting, intense magmatism at such places is generally attributed to mantle plumes, but it clearly may also be explained by long-term access to variably fertile and enriched material in sutures. Such access might be maintained in a number of ways after the cratons have drifted apart. Lateral ridge migration over underlying asthenosphere, would prolong access to delaminated crustal material in the suture. That such migration must occur is an inevitable consequence of the world's ridges not being fixed with respect to one another. Where a suture is oblique to the direction of ridge migration, the geometry of the suture will be such that the locus of melt extraction will migrate along the ridge, such as occurs at the Walvis-Tristan da Cunha Ridge.

Another possibility is that sublithospheric suture material resides beneath supercontinents for long periods and is only involved again in mantle convection after continental breakup occurs. The compositional variation of North Atlantic mantle revealed by regional variations in basalt geochemistry suggests that bulk lithological heterogeneities in the mantle are maintained for long periods. The sharp changes in the compositions of basalts away from Iceland and along Reykjanes and Kolbeinsey Ridges (Figure 1) suggest that the sources of the latter are mainly peridotitic, not eclogitic. If the eclogitic source of Icelandic basalts is indeed a delaminated, stretched, steeply dipping slab of Palaezoic ocean crust, then the peridotite sources of basalts from the adjacent ridges may be combinations of even older refractory subcratonic mantle and depleted Paleozoic abyssal peridotite, and not simply the comparatively fertile MORB mantle that supplies most other ridges. The nearest such mantle is, as Schilling (1973) proposed, south of the Charlie-Gibbs Fracture Zone (Kempton et al., 2000). Still further south, the Azores platform has basalts with still different, more typically alkalic, compositions. How sublithospheric Iapetus slab material could have traveled with Laurasia as it drifted through several tens of degrees of latitude in the period ~ 440 – 54 Ma is not known. However, the mantle that supplies the North Atlantic clearly is not a homogeneous entity that produces variable parental basalts simply as a result of differences in mantle temperature (Langmuir and Bender, 1984; Klein and Langmuir, 1987; McKenzie and Bickle, 1988).

The Iceland melt anomaly is derived from shallow structures in the mantle and processes consequential to plate tectonics. This is consistent with seismic tomography, which shows that the mantle anomaly beneath the Iceland region is by far the strongest in the upper 250 km, with a weak tail that extends down no deeper than the transition zone (Ritsema et al., 1999; Foulger et al., 2000; Foulger et al., 2001). It is also consistent with the north-south tectonic and compositional asymmetry, which is expected if rifting occurs above a heterogeneous mantle harboring north-south asymmetry inherited from an embedded dipping slab that has variable melt fertility (Foulger, 2002). Above such variable mantle, small structures, e.g., across the Tjornes Fracture Zone in north Iceland, are marked by abrupt changes in geochemistry. The location of the Iceland volcanic province over an easterly trending branch of the Caledonian suture is no coincidence, but a cause of the melt anomaly.

In general, sutures result from continental collision and ocean basin closure that annihilates ridges and traps young oceanic lithosphere, island arcs, and backarc basins in the upper mantle. Many continental flood basalts, e.g., the Deccan Traps and the Siberian Traps, erupted at sutures (Smith, 1993). Thermal anomalies and major precursory uplift at these places are usually absent (Czamannske, 1998; Anderson, 2000b; Stein and Stein, 2002). Enhanced mantle fertility can explain large melt anomalies without the need for high temperatures, and thus constitutes a viable alternative explanation to plumes for igneous provinces in general.

8. Conclusions

1. The geochemistry, melt volumes and lack of elevated source temperatures of basalts at the Icelandic volcanic province are explained by extensive melting of subducted Iapetus ocean crust trapped in the Caledonian suture that is retained locally in the upper mantle beneath the Mid-Atlantic ridge. The crust involved is probably that which subducted latest, prior to closure of the Iapetus Ocean, and was therefore hot and buoyant, and did not subduct deeper than the shallow upper mantle.

2. Fractional remelting of abyssal gabbro as eclogite, along with a component of E-MORB, explains the major-, trace- and rare-earth-element compositions, and the isotopic characteristics of primitive Icelandic tholeiite. Basalts ranging from ferrobasalt to olivine tholeiite result from various degrees of partial melting of eclogite. The occurrence of substantial oxide gabbro crystallized from ferrobasalt in the lower crust explains the anomalously high densities there.

3. High $^3He/^4He$ ratios result from the preservation of old helium either in U+Th-poor residual mantle of Caledonian or Archaean age still in the melt region, or in olivine-rich cumulates from which late-stage, U+Th-bearing melt was both excluded during crystal growth and expelled during compaction.

4. The subducted Iapetus ocean crust beneath Iceland must lie at a high angle, or be thickened by imbrication or other deformation, in order to be able to supply the thick crust of Iceland.

5. High rates of melt-extraction have persisted for 54 Myr at the mid-Atlantic ridge in the Iceland region because the Caledonian suture there is orientated transverse to the ridge. Lateral ridge migrations have thus not transported the ridge away from the fertile source.

6. The diversity of basalts throughout the North Atlantic requires source variations that cannot be explained simply by temperature variations in a homogenous source.

7. Continental breakup commonly occurs along old suture zones, and thus similar processes may explain igneous provinces elsewhere that are traditionally attributed to plumes.

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Figure captions

Figure 1 (a) Closure of the Iapetus Ocean at 400 Ma by convergence of Laurentia, Baltica and Avalonia. Arrows: convergence directions; thick lines: faults and orogenic fronts. Black triangles indicate sense of thrust faults. Slabs were subducted beneath Greenland, Baltica and Britain. Thick red dashed line: inferred line of opening of Mid-Atlantic Ridge at ~ 54 Ma. (b) Bathymetry of the North Atlantic, showing the Greenland-Iceland-Faeroe bathymetric ridge that which is underlain by crust ~ 30 km thick. Other shallow areas are blocks of stretched continental crust. Thin purple line: MAR; thin dashed black lines: extinct ridges; thick lines: faults of the Caledonian suture (Soper et al., 1992); thick dashed line: inferred trend of suture crossing the Atlantic Ocean (Bott, 1987). RR: Reykjanes Ridge, KR: Kolbeinsey Ridge (Adapted from Foulger (2003).)

Figure 2 (a) Parental soda (Na_8) in basalt glass v. latitude. Data from Kolbeinsey and Reykjanes ridges from Lamont petrological database (PetDB). Icelandic compositions from Meyer (1985) and Breddam (2002). Range for Pacific-Antarctic EPR is from Castillo (1998). (b) Crustal thickness vs. latitude, from a compilation of seismic experiments in Iceland and the North Atlantic (adapted from Foulger (2003)).

Figure 3 Chondrite-normalized REE patterns for average Hole 735B gabbro lithologies (solid black lines) compared with the range (black) and average (white line) of basalts from Kistufell (Breddam, 2002) and the range of basalts from Iceland rifts (MacLennan et al., 2002) (gray field). Bold dashed line: average of two Ne-normative AOB from seamounts near the EPR (Table 1), two red lines: weighted average of all samples analyzed for REE from Hole 735B (lower), and the same composition plus 4.8% AOB (upper). The latter closely matches average Kistufell olivine tholeiite. In general, the flat to enriched REE patterns of Icelandic basalts can be produced by mixing melt derived from abyssal gabbro and either an enriched material such as AOB, or silicic material, such as trondhjemite. Data are given in Table 1.

Figure 4 Zr vs. La showing how reduced Zr/La in Icelandic tholeiites can result from extensive melting of abyssal gabbro. Blue symbols: averages of olivine gabbro and troctolite (lower left) and differentiated gabbro through trondhjemite (upper right). Basalt compositions plot to upper right, whereas gabbro cumulate compositions plot to lower left. Extreme differentiates are at the extreme upper right. The weighted average Hole 735B gabbro (lower left bold open triangle) combines primitive gabbro cumulates with differentiated seams and veins of oxide gabbro and trondhjemite, and resembles primitive Icelandic tholeiite. Strip: average of 17 strip samples of the upper 500 m of Hole 735B (Table 1; Hart et al., 1999), is similar to moderately differentiated Icelandic tholeiite. A model composition (upper right bold open triangle) combining average Hole 735B gabbro plus 4.8% AOB (Niu et al., 2002) resembles primitive Kistufell tholeiite. Icelandic basalts are shifted to lower Zr at given La than MORB from the Pacific-Antarctic EPR. The Iceland trend could be produced either by crystallization differentiation of primitive Icelandic tholeiite inheriting low Zr/La from a gabbroic melt source, or direct partial melting of such a source, mixed with a small amount of AOB. Mixing with small amounts of either enriched (e.g., AOB) or silicic material (e.g., rhyolite, trondhjmeite) does not shift Zr/La. (Data from Table 1; Breddam, 2002; MacLennan et al., 2002).

Figure 5 Ternary diagram comparing proportions of Zr, 10*Nb, and and 3*Y. The latter two are expanded for clarity. Symbols and labels are as in Figure 4. Data fields for the Pacific-Antarctic EPR and eastern Pacific seamounts (Niu et al., 2002) are shown for comparison. Also shown are estimated compositions of clinopyroxene (gray dots) calculated to be in equilibrium with seamount liquids, using partition coefficients of Hart (1993). Cumulates rich in clinopyroxene that crystallized from N-MORB to AOB thus should have higher Y/Zr than the liquids themselves. Nb enrichment at given Y/Zr is explained by mixing with AOB. Icelandic basalts thus plot toward higher Y at given Nb and Zr than basalts from the Pacific-Antarctic-EPR, indicating derivation by melting from abyssal-gabbro precursors, and they have similar proportional enrichment in Nb as many N-MORB from the Pacific-Antarctic-EPR. This figure distinguishes two mixing trends, between: (1) primitive tholeiite derived from melting gabbro, and trondhjemite (high Zr, low Nb), and (2) primitive tholeiite and AOB (high Zr, high Nb). Basalts from both Iceland and the Pacific-Antarctic-EPR demonstrate at least the latter, and perhaps both types of mixing in some samples. Some Icelandic tholeiites with higher Y/Zr than those from Kistufell closely resemble Hole 735B troctolite, and thus may have been derived from a less differentiated, average gabbroic precursor.

Figure 6 Schematic diagram illustrating how anomalously large amounts of melt may be obtained from remelting a subducted crustal slab of normal thickness. a) The slab is emplaced at a high angle in the mantle. In this particular example, two thicknesses of melt might be derived from remelting eclogite in trapped oceanic crust, and one thickness is derived from melting mantle peridotite, yielding triple the amount of melt normally observed at MORs. b) Slab material is thickened by imbrication.

Table caption

Table 1 Data pertaining to lithologies in ODP Hole 735B, unless otherwise stated. Major elements are given in weight %, trace elements in parts per million. MgNo = Mg/(Mg+Fe2+), where Fe2+/(Fe2++Fe3+) = 0.90, the average for all lithologies at Hole 735B. Columns 1-6: average gabbros, computed for all samples for which REE measurements were made, using chemical definitions of Natland (2002) and their Appendix B; Column 7: average of 17 strip samples of the upper 500 m of the hole (Hart et al., 1999); Columns 8-12 similar averages for Hess Deep; Column 13; weighted average gabbro from Hess Deep; Column 14: weighted average for Hole 735B for major oxides and selected trace elements using all chemical analyses, from Dick (2000); Column 15: average of all Hole 735B samples analyzed for REE, using weighted percentages given below each average of columns 1-6; Column 16: weighted average olivine gabbro plus troctolite from Columns 1 and 2; Column 17: weighted average of differentiated gabbro, oxide gabbro, felsic veins and trondhjemite, from columns 3-6; Column 18: average of two Ne-normative alkalic olivine basalts from seamounts near the EPR; Column 19: average olivine tholeiite glass from Kistufell (Breddam, 2002); Column 20: same as column 19 except for whole rock; Column 21: 95.2% of column 16 plus 4.8% of column 18.

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Table 1 Average Compositions

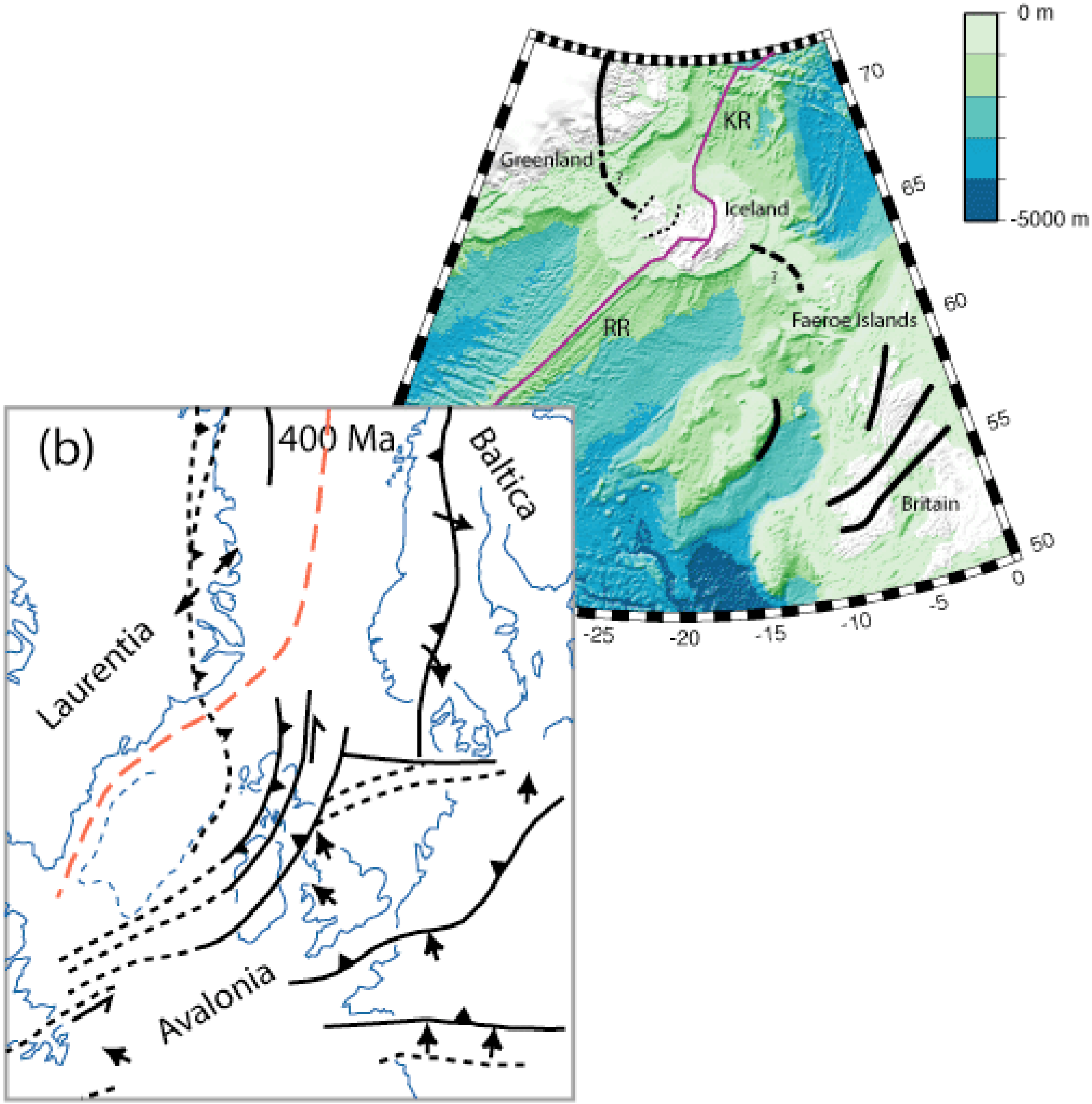
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
SAMPLE	735B troctolite	735B olivine gabbro	735B differ gabbro	735B oxide gabbro	735B Felsic	735B Trond	735B strip average	Hess Troctolite	Hess Olivine Gabbro	Hess difer. gabbro	Hess Oxide Gabbro	Hess Tonalite	Hess weighted average	Dick et al (2000) 735b Average	735B Ave all REE	735B Troct+ Oliv. Gab.	735B DOOG+	EPR AOB	Kistufell average glass	Kistufell aver oliv. tholeiite	All REE + 4.8% AOB
SiO2	47.11	50.56	50.98	46.35	58.98	69.54	50.74	46.60	50.24	51.11	50.45	67.30	50.45	50.60	50.11	50.46	48.65	49.18	48.34	48.48	50.07
TiO2	0.25	0.38	0.62	4.23	0.54	0.24	1.34	0.33	0.33	0.68	1.71	0.70	0.56	0.87	0.81	0.38	2.59	2.44	0.99	0.95	0.89
Al2O3	15.29	17.17	16.62	13.21	16.81	16.67	16.08	19.16	19.18	15.24	11.64	15.52	18.05	16.10	16.68	17.13	14.82	17.7	15.87	15.84	16.73
Fe2O3	0.81	1.24	1.92	3.86	0.43	0.31	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.37	1.58	1.23	2.98	0	0.00	0.00	1.5
FeO	6.89	4.25	6.40	12.16	4.16	1.10	8.34	6.26	5.22	8.49	12.92	3.56	6.53	6.19	5.30	4.27	9.55	8.86	9.18	9.22	5.47
FeOT	7.62	5.37	8.13	15.63	4.55	1.38	8.34	6.26	5.22	8.49	12.92	3.56	7.25	7.42	6.72	5.38	12.22	8.86	9.18	9.22	6.82
Fe2O3T	8.46	5.96	9.03	17.37	5.05	1.53	9.27	6.95	5.80	9.43	14.35	3.96	9.05	8.25	7.46	5.98	13.58	9.84	10.20	10.24	7.57
MnO	0.14	0.12	0.15	0.24	0.09	0.02	0.16	0.10	0.11	0.15	0.22	0.05	0.14	0.14	0.14	0.12	0.22	0.19	0.16	0.16	0.14
MgO	15.88	9.42	7.77	6.44	4.61	0.37	8.84	14.20	8.75	8.23	9.10	0.97	9.31	9.21	9.01	9.48	7.08	6.17	9.71	10.66	8.87
CaO	10.72	13.03	11.51	9.44	7.66	2.07	11.11	12.27	13.66	12.06	11.37	6.46	12.74	12.50	12.49	12.99	10.42	9.07	13.80	12.52	12.33
Na2O	1.91	2.76	3.32	3.30	5.51	6.92	3.19	1.00	2.46	2.34	2.16	4.56	2.52	2.80	2.87	2.75	3.36	3.94	1.78	1.73	2.92
K2O	0.06	0.05	0.11	0.11	0.24	0.65	0.11	0.12	0.13	0.02	0.03	0.19	0.03	0.05	0.07	0.05	0.14	1.34	0.07	0.07	0.13
P2O5	0.01	0.02	0.06	0.15	0.07	0.01	0.09	0.03	0.02	0.06	0.24	0.46	0.04	0.05	0.04	0.02	0.15	0.59	0.05	0.09	0.07
Total	99.07	99.00	99.46	99.49	99.10	97.90	100.00	100.07	100.10	98.38	99.84	99.77	100.11	99.88	99.05	98.88	99.77	99.66	99.95	99.71	99.72
MgNo	0.812	0.784	0.664	0.460	0.677	0.357	0.687	0.824	0.776	0.667	0.593	0.360	0.741	0.720	0.705	0.758	0.596	0.590	0.686	0.705	0.699
# in ave	27	175	52	69	34	3	17	3	21	8	6	1			360	158	202	2	10	16	
Weighted %	0.85	79.56	8.84	10.59	0.11	0.05															

Trace Elements

Ba		4.85	6.44	8.55	20.52		9.5	7.24			23.93	33.7						354.9		14.77	
Rb	1		1	1	1	1	0.562	1		1	1						1.06	26		1	
Sr	130	162	172	151	152	144	158	74	118	93	82	119	114	163	161	161.50	161	526		123	178.9
Y	7	12	20	37	59	46	27	10	11	22	47	129	15	13	15.4	11.93	29.6	24		17	15.9
Zr	11	24	51	168	297	107	78	18	14	43	173	469	23	30	41.8	23.83	116.2	255		44	52.1
Nb	0.5	0.4	0.5	3.7	5.6	4.7	1.7	0.6		1.1	4.4	17			0.78	0.41	2.33	44.7		2.9	2.9
Th		0.2	0.3	0.3	1.2	2	0.2										0.37	2.9		0.1	
Pb	0.5	0.8	1.2	1.5	1.9	2.5	0.6								0.92	0.80	1.44	2.04		0.2	0.97
Zn	47.3	37.1	51.7	107.1	31.2	10	54.2	38.3	36.6	62	74.5	0.3		43	45.9	37.20	81.9	99.4		69.6	48.4
Cu	66	59.3	49.8	88.5	22	10.8		60.7	52.1	67	44.7	5		51	61.6	59.30	70.9	44.8		112.6	60.8
Ni	284.4	107.7	74.6	56.9	42.7	31.2	138.1	390	150	103	93.5	9	154	106	100.8	109.40	65.1	95.8		250.7	100.6
Cr	551	250	105	94	61	41		516	388	267	181			295	223	252.90	99.1	120		431	218
Co	65.7	36.8	41.3	60.5	19.8	7.1	42.2	37.5	36	45.1	51.5				39.9	37.10	51.8	33.5		50.8	39.6
V	91	142	186	535	85	32	197	182		2387	455	49		240	187	141.30	376	176		237	186
Hf					8.7		2.1		12.1									5.5		1.3	
Sc	24.4	36.8	37.1	42.7	20.4	5		11.5	30.4						37.3	36.60	40.2	21.1		38.6	36.5
Ta		0.1		0.3	0.5	0.7												2.5		0.2	
U	0.02	0.02	0.12	0.07	0.38	0.54	0.035								0.02	0.02	0.19	0.98		0.03	0.098

Rare Earth Elements

La	0.63	1.04	1.99	3.84	15.38	24.2	4.79	0.68	0.56	1.54	4.76	13.7			1.45	1.030	3.18	30.34		2.5	2.84
Ce	2.15	3.16	6.22	12.76	42.77	58.84	14.89	2.12	1.86	4.72	14.58	41.73			4.5	3.150	10.1	65.11		6.7	7.41
Pr	0.49	0.59	1.39	2.26	6.91	7.78				0.92	3.18				0.86	0.590	2	8.38		1.01	1.22
Nd	2.06	3.12	5.54	11.77	25.13	29.72	10.22	1.77	1.81	4.51	11.93	38.85			4.29	3.110	9.16	32.2		5.38	5.62
Sm	0.7	1.1	2.08	3.91	6.68	6.48	3.09	0.61	0.7	1.96	4.32	12.47			1.51	1.090	3.22	6.68		1.83	1.76
Eu	0.38	0.58	1.08	1.58	1.92	1.48	1.14	0.33	0.41	0.76	1.21	3.22			0.75	0.580	1.47	2.19		0.75	0.82
Gd	1.28	1.64	3.01	4.87	8.01	6.16		0.97	1.11	2.77	5.83	15.36			2.13	1.630	4.17	6.35		2.51	2.33
Tb	0.17	0.3	0.54	1	1.4	1.2	0.73		0.2	0.48	1.15				0.42	0.300	0.91	0.9		0.43	0.44
Dy	1.59	1.99	3.62	5.69	9.95	7.38				3.37	6.69	17.22			2.55	1.980	4.91	5.11		2.83	2.68
Ho	0.34	0.43	0.78	1.31	2.26	1.65			0.27	0.67	1.6				0.58	0.430	1.2	1		0.64	0.6
Er	0.99	1.22	2.24	3.66	6.95	5.21				2.26	4.14	10.79			1.6	1.220	3.17	2.7		1.68	1.64
Tm	0.15	0.19	0.35	0.63	1.12	0.87			0.13						0.28	0.190	0.64	0.36		0.26	0.28
Yb	0.72	1.14	2.07	3.7	7.17	6.59	2.77	0.71	0.86		3.02	9.96			1.52	1.130	3.12	2.35		1.71	1.56
Lu	0.11	0.17	0.32	0.56	1.09	1.06	0.4	0.11	0.14	0.3	0.61	1.69			0.25	0.170	0.59	0.35		0.28	0.26
	115.15	110.33	116.62	132.49	108.70	100.81	117.61	113.28	111.12	116.30	127.11	107.29									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70
65
60
55
50
KR
Greenland
Iceland
RR
Faeroe Islands
Britain
-25
-20
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-10
-5
0
(b)
400 Ma
Baltica
Laurentia
Avalonia

Figure 1

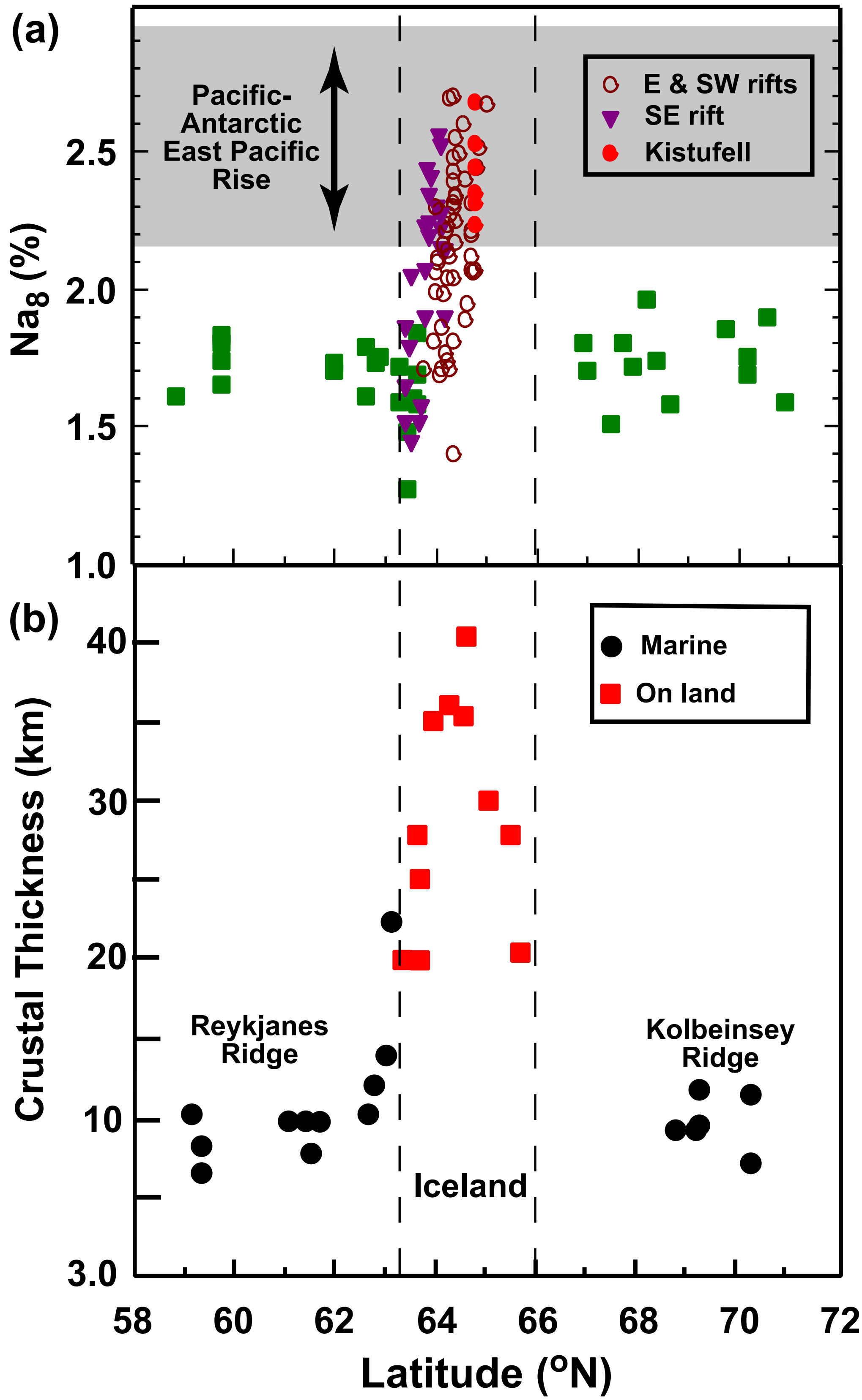
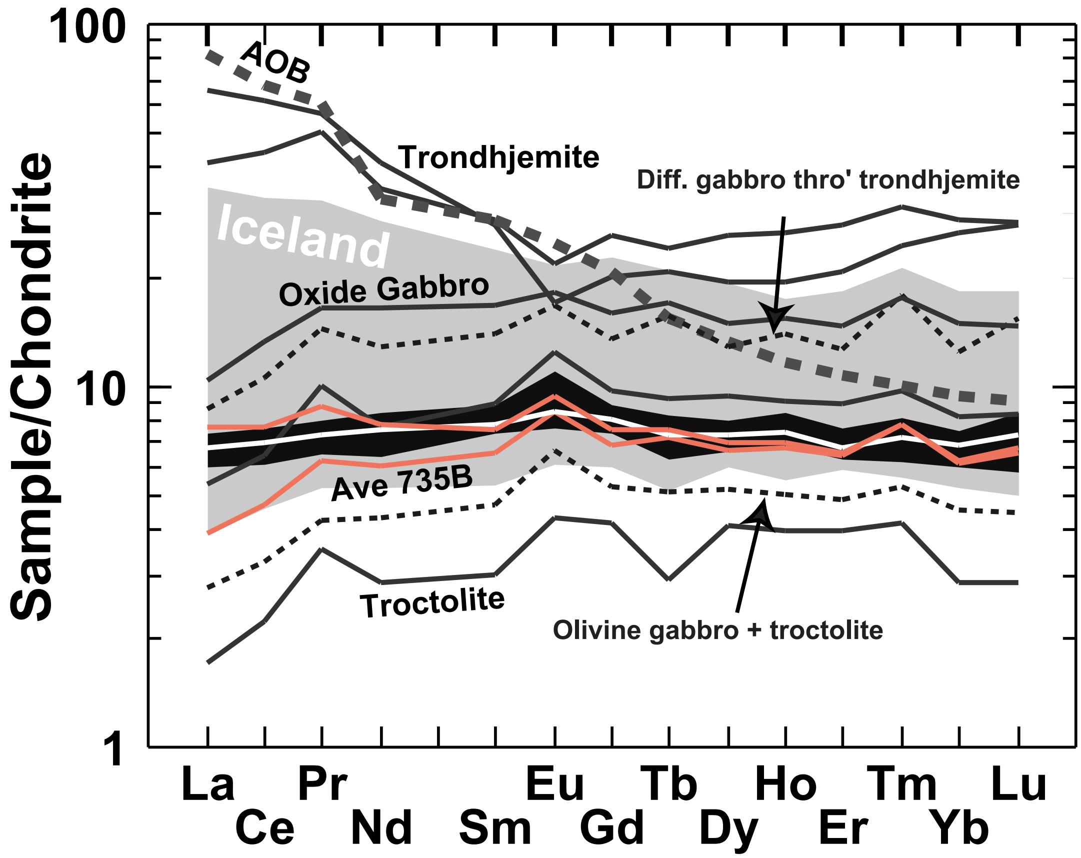

(a)
Pacific-
Antarctic
East Pacific
Rise
E & SW rifts
SE rift
Kistufell
Na_8 (%)
2.5
2.0
1.5
1.0
(b)
Marine
On land
Crustal Thickness (km)
40
30
20
10
3.0
Reykjanes
Ridge
Kolbeinsey
Ridge
Iceland
58
60
62
64
66
68
70
72
Latitude (°N)

Figure 2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10
1
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Trondhjemite
Diff. gabbro thro' trondhjemite
Iceland
Oxide Gabbro
Ave 735B
Troctolite
Olivine gabbro + troctolite
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Ce
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Tb
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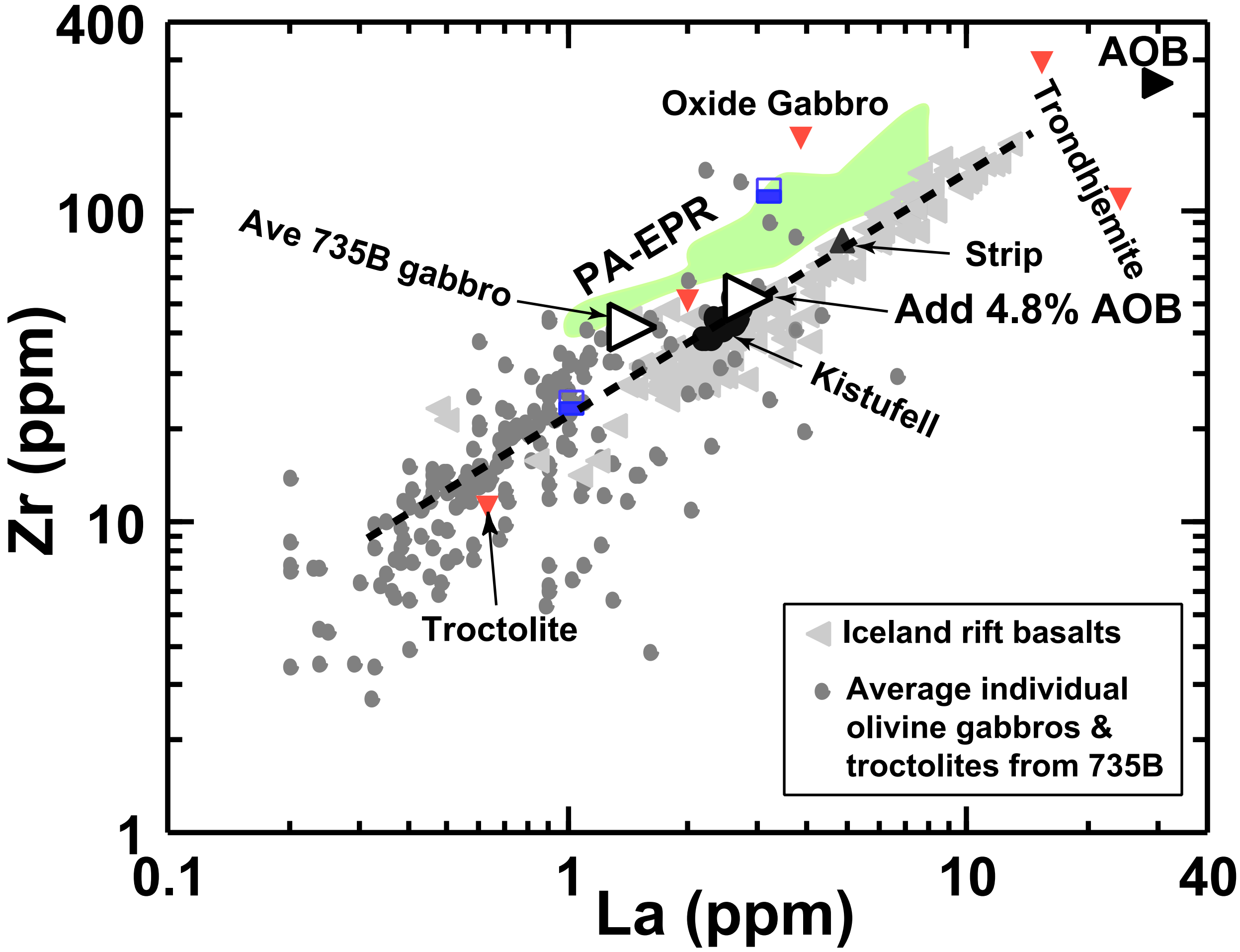
Yb
Lu

Figure 3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10
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10
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Zr (ppm)
La (ppm)
AOB
Oxide Gabbro
Trondhjemite
PA-EPR
Ave 735B gabbro
Strip
Add 4.8% AOB
Kistufell
Troctolite
Iceland rift basalts
Average individual
olivine gabbros &
troctolites from 735B

Figure 4

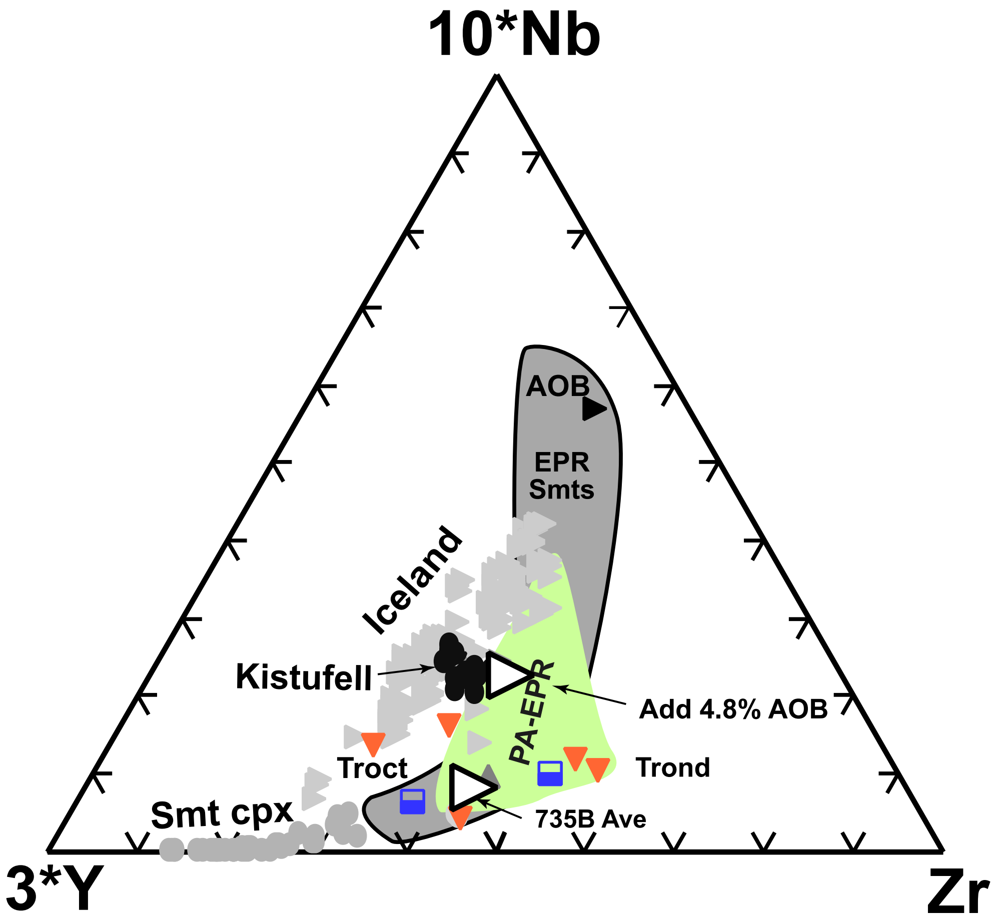
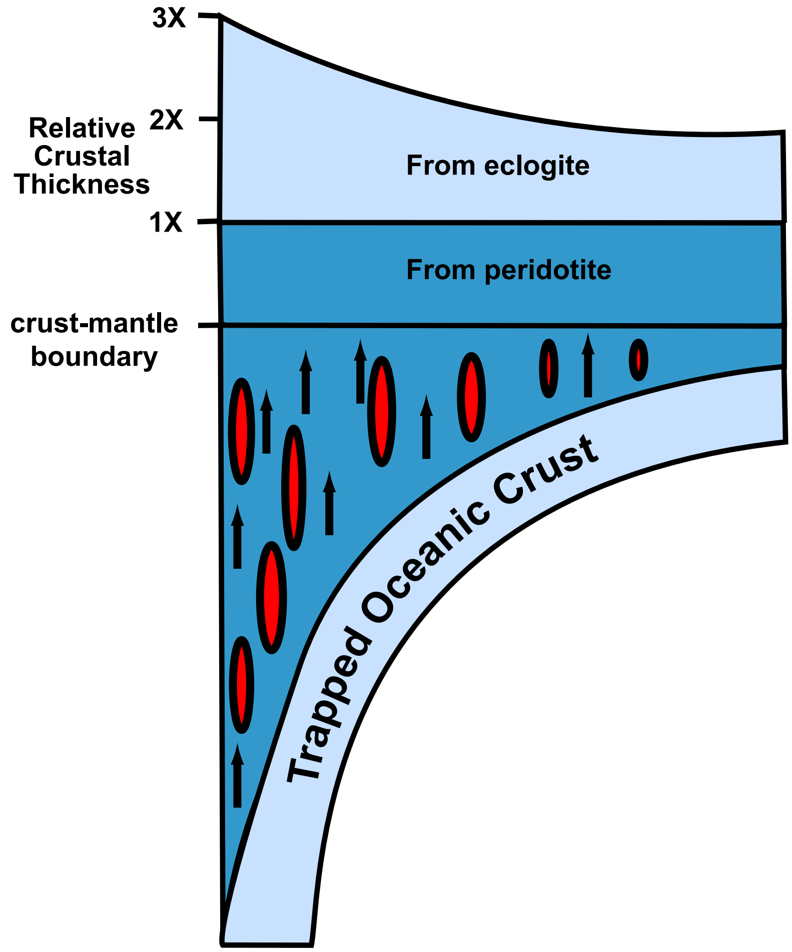

10*Nb
3*Y
Zr
AOB
EPR
Smts
Iceland
Kistufell
Add 4.8% AOB
PA-EPR
Troct
Trond
Smt cpx
735B Ave

Figure 5

a)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2X
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Crustal
Thickness
1X
crust-mantle
boundary
From eclogite
From peridotite
Trapped Oceanic Crust

b)

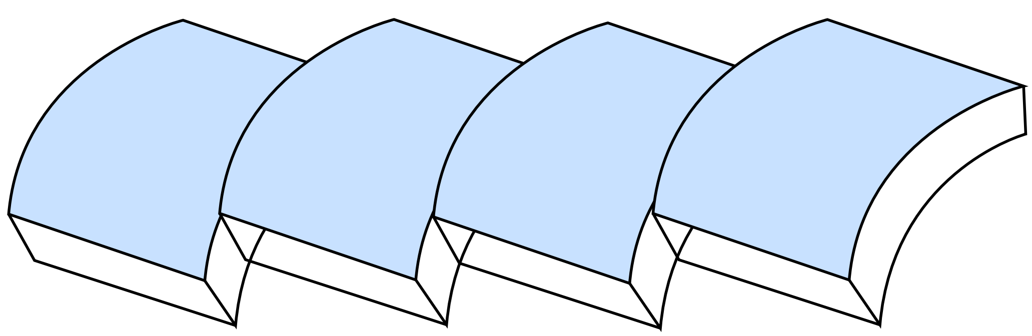

Figure 6