Geochemical characterization and petrogenesis of intermediate to silicic rocks in ophiolites: A global synthesis

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ABSTRACT

We present a global synthesis of the geochemistry of intermediate to silicic intrusive and volcanic rocks from 149 Phanerozoic to Archean ophiolites, and evaluate models for their genesis during the development of oceanic crust in different tectonic environments. Our synthesis shows that the evolved rocks in subduction-unrelated, *Rift/Continental Margin* ophiolites are predominantly basaltic andesite and andesite, whereas *MOR type* (mid-ocean ridge) ophiolites exhibit nearly equal proportions of basaltic andesite/andesite and rhyodacite and *Plume/MOR* type ophiolites are characterized by rhyolites. Intermediate to silicic volcanic units in the *Backarc* sub-group of subduction-related ophiolites are characterized by similar amounts of basaltic andesite/andesite and rhyodacite, whereas in the *Backarc to Forearc*, *Forearc*, and *Volcanic Arc* sub-groups they are mainly basaltic andesite/andesite. The Zr, Nb and La contents of the intermediate to silicic rocks in subduction-related ophiolites are significantly lower than those of their counterparts in subduction-unrelated ophiolites. Intermediate to silicic rocks in Rift/Continental Margin and Plume/MOR type ophiolites are generally LREE-enriched, whereas those in the MOR type vary from LREE-depleted to LREE-enriched. The Backarc and Backarc to Forearc types are similar to the MOR type; silicic rocks of the Forearc and Volcanic Arc types are generally LREE-enriched. Nb, Sr, Eu, and Ti exhibit negative anomalies of varying magnitudes in multi-element diagrams. Our modeling based on the La-SiO₂ proxies of Brophy (2009) suggests that the main process in the formation of the majority of the intermediate to silicic rocks in both subduction-unrelated and subduction-related ophiolites is partial melting of basaltic and/or gabbroic rocks beneath the spreading centers, whereas a minor volume in subduction-related ophiolites are adakites that resulted from partial melting of the subducting slab. Silicic to intermediate rocks in Plume/MOR type ophiolites appear to be generated by fractional crystallization of basaltic melt. The incompatible, non-conservative elements, such as Ba and Th, are weakly to strongly enriched in subduction-related ophiolites as a result of shallow to deep enrichment associated with subduction zone processes.

Key words: Archean to Phanerozoic ophiolites, greenstone belts, intermediate to silicic volcanic rocks, plagiogranite, geochemistry, petrogenesis of ophiolites

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107 **1. Introduction**

108 Magmatic rocks of ophiolites are predominantly basaltic in composition, but
 109 intermediate to silicic intrusive and extrusive units are also present in subordinate amounts in
 110 most ophiolite complexes. Throughout the nearly 200 years of ophiolite research around the
 111 world (Brongniart, 1821; Dilek and Newcomb, 2003; Dilek and Robinson, 2003), the
 112 principal focus has been on the petrogenesis of mafic rock assemblages in these ancient
 113 oceanic crustal units (e.g., Miyashiro, 1973, 1975; Dilek, 2003a; Pearce, 2003, 2008) and
 114 relatively little attention has been paid to the origin of intermediate to silicic rocks. Since the
 115 expansion of the ophiolite concept from mid-ocean ridge generated oceanic crust (e.g.,
 116 Moores and Vine, 1971; Anonymous, 1972) to subduction-unrelated and subduction-related
 117 ophiolites (Dilek and Furnes, 2011, 2014), focus on the generation of the intermediate to
 118 silicic rocks in the various types of ophiolites has become more pertinent.

119 In this paper we examine the occurrence and geochemistry of intermediate to silicic rocks
 120 in different ophiolite types, and discuss their petrogenesis during the formation of ancient oceanic
 121 crust in various tectonic settings. These rocks, although not in large volumes in comparison to
 122 mafic–ultramafic units in ophiolites, reveal significant clues about melt evolution, magmatic
 123 accretion processes, and partial melting of pre-existing oceanic crust during ophiolite evolution;
 124 they also yield ample accessory minerals, such as zircon and monazite, which are widely used to
 125 date the timing of oceanic crust generation. To better understand the diversity in the distribution

and geochemical makeup of intermediate to silicic rock types in ophiolites, we have compiled their geochemistry in 156 ophiolites and Precambrian greenstone belts in a global synthesis. We first define these ophiolites by using geochemical discriminants in order to classify them in line with the new ophiolite classification criteria (Table 1; Dilek and Furnes, 2011, 2014). We then evaluate the geochemical compositions of intermediate to silicic rocks occurring in each of the ophiolite types separately in order to find out whether these leucocratic assemblages have specific geochemical identities. Following their geochemical characterization, we discuss their petrogenesis and magmatic evolution in different ophiolite types, which developed in distinct tectonic environments with varied mantle compositions, melt regimes, and fluid types. The field occurrence and the geochemical character of leucocratic rocks in ophiolites show considerable variations, providing additional constraints on the petrogenesis of ophiolites in different tectonic settings.

2. Tectonic environment of formation and anatomy of ophiolites

Dilek and Furnes (2011) have defined ophiolites as “suites of temporally and spatially associated ultramafic to felsic rocks evolved from separate melting episodes and processes of magmatic differentiation in particular oceanic environments”. Figure 1 depicts the major tectonic settings in which subduction-unrelated and subduction-related oceanic crust undergoes its magmatic construction before emplacement into continental margins as ophiolites.

2.1. Subduction-unrelated ophiolites

Tectonic extension, continental rifting and rift-drift processes lead to lithospheric-scale thinning accompanied by asthenospheric upwelling and its decompression melting. The result is the development of a transition zone crust between the continental and oceanic lithospheric domains, known as Ocean – Continent Transition (OCT). OCTs represent the embryonic stage in the formation of oceanic crust and generally contain volcanic rocks directly overlying subcontinental mantle peridotites, limited gabbroic rocks and associated silicic intrusions and basaltic dikes and dike swarms (Fig. 1). An in-situ, modern OCT example is represented by the early Cretaceous Iberian rifted margin (Seyfert et al., 1997). On-land, fossil remnants of OCT lithospheric fragments constitute continental margin ophiolites (CMO; Dilek and Furnes, 2014),

and some of the best CMOs occur in the Western Alps (Balestro et al., 2015; Festa et al., 2015), the Northern Apennines (Saccani et al., 2015, and the references therein), and the Yarlung-Zangbo suture zone in SW Tibet (Liu et al., 2015).

With the onset of seafloor spreading after continental breakup, a mid-ocean ridge system gets established and new oceanic lithosphere develops (Fig. 1). The internal structure and the igneous stratigraphy of this mid-ocean ridge (MOR) generated oceanic lithosphere are largely dependent on the spreading rate, the melt-magma budget, and the magnitude and the rate of tectonic deformation keeping pace with seafloor spreading (Dilek et al., 1998). At fast spreading ridges, for example, the high rates of melt production creates a “hot environment” of oceanic crust generation in which a typical Penrose-type, crustal stratigraphy forms, complete with sheeted dikes and a transitional Moho (Fig. 1). Silicic intrusions commonly occur within the uppermost gabbros and at any structural level within the sheeted dike complex, in fast-spread MOR oceanic crust. At slow-spreading ridges, melt production is low and magma chambers are generally ephemeral, resulting in a “cold environment” of oceanic crust generation (Dilek et al., 1998). Tectonic extension commonly overpaces magmatic construction rates in this type of cold MOR setting and thus widespread high- to low-angle normal faulting and development of lithospheric-scale detachment faults occur, significantly attenuating the newly formed oceanic crust. Attendant hydrothermal alteration causes low-temperature metamorphism of the lower and upper crustal units and serpentinization of the upper mantle peridotites. The Moho becomes an alteration front with advanced serpentinization.

Plume-type ophiolites may be part of oceanic plateaus or plume-proximal oceanic ridges (e.g., Iceland). The volcanic sequence, consisting predominantly of massive lava flows and subordinate pillow lavas constitute a thick sequence that may reach several tens of km, crosscut by gabbroic to ultramafic intrusions (Fig. 1). Plume-type ophiolites are best represented by the Cretaceous oceanic crust exposed in the peri-Caribbean region (Kerr et al. 1998).

2.2. Subduction-related ophiolites

Fig. 1 shows a backarc spreading system in which suprasubduction zone (SSZ) ophiolites form as part of the upper plate of a subduction zone. In this tectonic environment backarc

spreading centers may evolve as trench-proximal versus trench-distal (Dilek and Furnes, 2011). Tectonic settings within subduction-accretion systems may be highly variable in terms of plate kinematics, tectonic architecture, and chemical geodynamics (Doglioni et al., 2007). We distinguish four main tectonic settings (Backarc, Backarc to Forearc, Forearc and Volcanic arc). The backarc- and MOR-generated oceanic lithosphere may structurally look alike, but their crustal and upper mantle units have different geochemical fingerprints.

Forearc oceanic crust may also be structurally and lithologically similar to that of the MOR oceanic crust, particularly in its lower sections. However, it differs from the subduction-unrelated ophiolites in its upper crustal sections, by the presence of intermediate (andesitic) to silicic (dacite and rhyolite) extrusive and intrusive rocks, and with boninitic extrusive and intrusive rocks (Fig. 1; Dilek and Thy, 2009). The Backarc to Forearc oceanic crust makes up a hybrid type between the Backarc- and Forearc-generated oceanic crust.

The Volcanic Arc oceanic crust represents more than 20-30 million-years of magmatic construction in a suprasubduction zone setting (Dilek et al., 1991; Dilek and Furnes, 2011). The middle to lower crust of this type is dominated by abundant intrusions of diorite, tonalite and granodiorite (Fig. 1). The thickness of this type may, therefore, be considerably thicker than the other subduction-related ophiolite types.

3. Field relationships of intermediate to silicic rocks in ophiolites

Intermediate to silicic rocks occur in ophiolite complexes of all ages as intrusive bodies at different pseudostratigraphic levels and/or in lavas and pyroclastic deposits within their upper volcanic parts (Fig. 2) (see Fig. 14 of Furnes et al., 2015). Plagiogranite intrusions are common in modern oceanic crust and ophiolites, and generally occur near the boundary between the gabbroic rocks and the dike complex, but most commonly as pods and late-stage dikes within the sheeted dike complex. These plagiogranite types have been widely reported from many ophiolites around the world (e.g., Dilek and Eddy, 1992; Dilek and Thy, 1998, 2009; Robinson et al. 2008; Sarifakioglu et al., 2015; Dilek and Furnes, 2014), and from the oldest reported ophiolite in the Isua Supracrustal Belt in SW Greenland (Furnes et al., 2009). They also exist in the upper mantle units as dikes, crosscutting different peridotites (Fig. 2; Dilek et al., 1998), and have been

reported from the Oman (Rollinson, 2014), Kizildag (Dilek and Delaloye, 1992), and Xigaze (Dai et al., 2013) ophiolites.

Andesitic, dacitic and rhyolitic extrusive rocks form the stratigraphically uppermost parts in some ophiolites, although they are less common than their intrusive equivalents (Fig. 2). Some of the best Phanerozoic ophiolites with intermediate to silicic volcanic rocks include the early Ordovician Karmøy Ophiolite in Western Norway (Pedersen and Hertogen, 1990), the Jurassic Pindos (Greece) and Mirdita (Albania) ophiolites, and the Cretaceous Oman Ophiolite (Dilek and Furnes, 2009). A Proterozoic example in which these extrusive rocks occur is the 1.73 Ga Payson Ophiolite (Arizona, USA). The 3.51 Ga Iron Ore Group in India (Mukhopadhyay et al., 2012), the 3.12 Ga Whundo in W. Australia (Smithies et al., 2005), and the 2.75-2.65 Ga Wawa in Canada (Polat et al., 1999) are some of the notable Archean greenstone belts (Furnes et al., 2015), in which intermediate to silicic volcanic rocks (including adakites) occur in the uppermost sections of ophiolitic sequences.

4. Intermediate to silicic rock types

In the ophiolite complexes we have investigated, intrusive and extrusive rocks show a wide range in composition (Table 1). Intermediate rocks (with SiO₂ between 52 and 63 wt. %) include andesite and diorite, whereas the silicic rocks (with SiO₂ more than 63 wt. %) include dacite, trondhjemite, tonalite and granite. It should be noted that the term “trondhjemite” is not a name recognized in the IUGS systematics of igneous rocks (Le Bas and Streckeisen, 1991), but *leucotonalite* is. However, in the ophiolite literature, trondhjemite is a well-known and commonly used term, and hence we also adopt it in this paper. The term “*plagiogranite*”, which was originally introduced as *oceanic plagiogranite* (Thayer, 1977), is commonly used in reference to *trondhjemite*, *tonalite*, *quartz diorite*, and even *diorite* (Coleman and Peterman, 1975; Koepke et al., 2007; Rollinson, 2009). Plagiogranite generally consists of quartz, plagioclase, hornblende, Fe-Ti oxide, and locally biotite (e.g., Brown et al., 1979). Figure 3 displays the feldspar variation of Anorthite (An), Albite (Ab) and Orthoclase (Or) in a ternary diagram, displaying various compositional fields of plagiogranite types from a selection of Phanerozoic ophiolites. Crustally-derived plagiogranites display a wide compositional range and straddle the fields for tonalite, trondhjemite and granite, whereas those that crystallized from mantle – derived melts (e.g. those

in the Oman ophiolite) cover a much more restricted field across the tonalite/granodiorite boundary (Fig. 3). *Keratophyre*, an altered and albitized silicic volcanic rock, has been reported from two of the ophiolite complexes (no. 49. Fidalgo Island, and no. 65. Chamrousse, Table 1). *Plagiogranite* and *keratophyre* are geochemically similar but represent intrusive and volcanic rocks, respectively Brown et al. (1979).

For subduction-related and subduction-unrelated MORB-type ophiolites in the Phanerozoic orogenic belts, the most commonly reported intermediate to silicic intrusive rocks, independent of their original tectonic setting, are *trondhjemite*, *plagiogranite* and *tonalite*, and less commonly *quartz-diorite* and *diorite*. These rock types have been documented from 80 percent of the 104 ophiolites included in this study (Table 1). Intermediate to silicic lavas and/or pyroclastic rocks (*basaltic andesite*, *andesite*, *dacite* and *rhyolite*) are less abundant, and have been reported from only 35 percent of the 104 ophiolites. *Trachyte* occurs in some of the Rift/Continental Margin type, subduction-unrelated ophiolites (e.g., Kaczawa, Poland, no 62 of Table 1).

Of the 42 Precambrian ophiolites included in this study (Table 1), 83 percent displays intermediate to silicic lavas (*andesite*, *dacite* and *rhyolite*), whereas only 21 percent of them contain silicic intrusive rocks (*plagiogranite*, *trondhjemite*, *tonalite*). We have included in Fig. 3 the fields of Archean TTG (*tonalite-trondhjemite-granodiorite*) suites and the compositions of experimental melts. A few occurrences of *adakite*, an intermediate to silicic rock type characterized by high Sr, Na and Eu, and low Y and HREE (Defant and Kepezhinskis, 2001), have been documented from Precambrian ophiolites in India and Russia (Table 1).

Dacitic to rhyolitic extrusive rocks are most common in ophiolites occurring in non-orogenic tectonic settings (nos. 147-149) and in *in-situ* oceanic crust (nos. 150 – 156 of Table 1). Oceanic plagiogranites have been recovered as veins and dikes in the upper mantle peridotites and gabbros drilled at several oceanic core complexes along some segments of the slow-spreading Mid-Atlantic Ridge and Central Indian Ridge (Silantyev et al., 2010; 2011). Andesites and dacites are recovered from IODP drill sites along the East Pacific Rise, the Juan de Fuca Rise and the Galapagos Spreading Centre and may represent significant volumes of the uppermost in-situ oceanic crust at a local scale (Wanless et al., 2010; Erdmann et al., 2015). Significant volumes of silicic volcanic rocks have also been recovered in two of the many ODP drill holes in the

submarine large igneous province of the Kerguelen Plateau in the southern Indian Ocean (Frey et al., 2000).

A large number of terms have been used in the literature in regard to the terminology of the intermediate to silicic rocks (Table 1). This is potentially confusing for some readers. Therefore, we simplify the terminology in this study and use instead the rock names as shown in Fig. 4A (i.e. andesite, basaltic andesite, rhyodacite, trachyandesite, trachyte, and rhyolite).

5. Selected Phanerozoic and Precambrian ophiolites

Phanerozoic orogenic belts that host the ophiolites examined in this study represent both collisional and accretionary types (Isozaki, 1997; Dilek, 2006; Condie, 2007; Windley et al., 2007; Cawood et al., 2009; Wilhem et al., 2011). The Caledonian-Appalachian, Hercynian, Uralian, Maghrebian-Alpine-Himalayan, and Qinling/Qilian/Kunlun orogenic belts are the best examples of the collisional orogenic type. The peri-Caribbean and Central Asian Orogenic belts, Gondwanide-Tasmanide, Andes, and Western Pacific and Cordilleran belts represent the most salient examples of the accretionary type orogenic belts. The modern Indonesian-Myanmar orogenic belt is a hybrid one with both active subduction – accretion processes and collision (e.g., Timor) at work.

The term *greenstone belt* has been generally used to describe these terranes consisting of spatially and temporally related, Precambrian intrusive – extrusive ultramafic, mafic to silicic rock suites that are associated with various types of metasedimentary rocks and intruded by granitoid plutons (de Wit, 2004). Whether greenstone belts contain ophiolites has been a long-standing and unresolved question. Some authors (e.g., Bickle et al., 1994; Hamilton, 1998, 2011) argue that true oceanic lithosphere is not represented in the Archean greenstone belts. Many others, however, present different lines of evidence supporting its presence in the Archean rock record (e.g., Dann, 1991; de Wit and Ashwal, 1997; Dann and Bowring, 1997; St-Onge et al., 1997; Sylvester et al., 1997; Dilek and Polat, 2008; Furnes et al., 2007, 2009, 2012a).

5.1. Phanerozoic orogens

In this study we have investigated the igneous pseudostratigraphy and geochemical makeup of 104 ophiolite complexes in 13 Phanerozoic orogenic belts around the world with a special focus on their intermediate to silicic rocks. A comprehensive description of these ophiolites and their classification based on their tectonic and geochemical fingerprints (Dilek and Furnes, 2011, 2014) are given in Furnes et al. (2014). Here, we present a synopsis of each orogenic belt with respect to its ophiolites (Table 1).

5.1.1. Indonesian-Myanmar

The Mesozoic - Cenozoic Indonesian-Myanmar orogenic belt (Hall, 2012; Fareeduddin and Dilek, 2015) has had a complex magmatic and tectonic history. It evolved largely during the closure of the Meso-Tethyan Ocean. Subduction rollback since the Eocene has resulted in the opening up back-arc basins (e.g., the Celebes, Philippine and Banda Seas) in the east, subduction – accretion events in the west (NE India and Indo-Myanmar Ranges) and an arc-continent collision in the SE (Timor – Australia). We have included two ophiolites from this orogenic belt in this synthesis (Table 1): Andaman (Andaman Sea, India) and Myitkyina (Myanmar).

5.1.2. Maghrebian-Alpine-Himalayan

The Maghrebian-Alpine-Himalayan orogenic belt extends from the Atlas Mountains in Morocco eastwards through the European Alps, Dinarides-Hellenides in the Balkan Peninsula, Anatolides-Taurides in the eastern Mediterranean region, Zagros and Makran belt in the Persian Gulf region, to Tibet – Himalaya in south-central Asia. The development of this belt occurred during a protracted evolution of various Tethyan seaways that overlapped in time and space (Stampfli and Hochard, 2009). We have examined a total of 36 ophiolites in this orogenic belt, representing continental margin, MOR, and SSZ types that were associated with the opening, seafloor spreading and subsequent closure stages of a series of Neotethyan branches (Dilek et al., 1999, 2008; Robertson, 2000; Manatschal and Müntener, 2009). The Tethyan ophiolites in this belt are principally of two age groups (Table 1): an older group in the western realm with ages around 170 -140 Ma and a younger group (east of the Aegean Sea) with Cretaceous (125-90 Ma) ages.

5.1.3. Caribbean Region

The Mesozoic evolution of the Caribbean region is characterized by the development and tectonic accretion of a Cretaceous large igneous province (LIP). This Caribbean LIP was constructed during three main igneous episodes in the time interval of 125 - 70 Ma (e.g., Escuder-Virue et al., 2011). In this study we have compiled geochemical data from five ophiolites ranging in age from ca. 126 Ma to 82 Ma (Table 1).

5.1.4. Cordilleran and Western Pacific

The Circum-Pacific orogenic system, including the Cordilleran belt of western North America and the Western Pacific belt, encompasses Mesozoic-Cenozoic subduction-accretion complexes along the Pacific Rim, and parts of it are still tectonically active. The characteristic features of the Circum-Pacific orogenic system have resulted largely from the prolonged subduction of oceanic lithosphere from Triassic time onwards along the active continental margins and island arc systems fringing the North and South American, and Asian continents (e.g., Godfrey and Dilek, 2000; Dickinson, 2004). Early models for the North American Cordillera invoke a Jurassic-Cretaceous eastward-dipping subduction zone, with oceanic crust being subducted underneath the continental margin (e.g., Saleeby, 1983; Harper and Wright, 1984), whereas more recent interpretations envision a series of fringing Triassic-Jurassic island arc systems and back-arc basins with opposing subduction polarities off the coast of North America that collapsed into the continental margin through arc-trench collisions in the latest Jurassic (e.g., Dilek et al., 1991; Ingersoll, 2000; Hildebrand, 2013). In Japan, the subduction history has been shown to go back to ca. 500 Ma (Isozaki et al., 2010). We have included 18 ophiolite complexes from this vast, peri-Pacific orogenic system in this study (Table 1).

5.1.5. Hercynian/Variscan

The Hercynian (or Variscan) orogenic belt (e.g. Zwart, 1967) extends from Western Iberia to Anatolia in the eastern Mediterranean region (see Fig. 1 of Furnes et al. 2014), and represents a complex subduction-accretion-collision belt. It developed during the closure of a series of Paleozoic ocean basins during the Ordovician through early Carboniferous as a result of the

convergence of the supercontinents of Gondwana and Laurussia (e.g., Matte, 1991; Kroner and Romer, 2013). Four ophiolite complexes from this Paleozoic orogenic system are included in this study (Table 1), i.e. Kaczawa (Poland), Beja-Acebuches and Morais (Portugal), and Chamrousse (France).

5.1.6. Uralian

The Uralides represent an arc–continent collision orogenic belt (e.g., Brown and Spadea, 1999; Ryazantsev et al., 2008) that extends from the Kara Sea in the north to Kazakhstan in the south. The remnants of early Paleozoic subduction-accretion complexes occur along a suture zone between the East European platform and the west Siberian craton (Spadea and D’Antonio, 2006; Puchkov, 2009). In this study we have included two of the Uralian ophiolites, i.e., the Nurali and Magnitogorsk (Table 1).

5.1.7. Central Asian

The Central Asian orogenic belt, situated between the Siberian and North China Cratons (Tarim and Sino-Korean Cratons), developed by the accretion of island arcs, ophiolites and continental blocks along the bounding cratons during the evolution of the palæo-Asian Ocean from the late Mesoproterozoic through the Mesozoic (Kröner et al., 2008; Jian et al., 2010b; Wakita et al., 2013; Zheng et al., 2013). We have included 11 ophiolite complexes in our synthesis, mostly from China with several from Mongolia (Table 1).

5.1.8. Qingling-Qilian-Kunlun

The Qingling-Qilian-Kunlun fold belt, also known as the Central China orogenic belt, extends in a NW-SE direction between the North China and Tarim Cratons to the North and the Yangtze Craton to the southeast (e.g., Dong et al., 2011; Song et al., 2013; Li et al., 2015). This orogenic belt developed when a series of ocean basins closed during the Paleozoic–Mesozoic, following the Late Proterozoic break-up of Rodinia. Four ophiolites from this belt are included here (Table 1).

374 5.1.9. Bangong-Nujian Suture

375 The E-W-trending Bangong-Nujiang Suture Zone occurs between the southern Lhasa and
 376 the northern Qingtang blocks in central Tibet, and includes discontinuous exposures of Paleozoic
 377 mélanges and ophiolites (Jian et al., 2009). We have examined five ophiolites from this suture
 378 zone (Table 1).

379 5.1.10. Andean

380 The Andean belt is the longest active orogenic system and is divided into Northern,
 381 Central, and Southern segments. Its tectonic evolution involved nearly 600 Ma of oceanic
 382 subduction along the western continental margin of South America, beginning shortly after the
 383 break-up of Rodinia in the late Proterozoic (Oncken et al., 2006; Ramos, 2009). The Andean
 384 ophiolites (Table 1) included in this study are the Sarmiento ophiolite (Chile) and the Famatinian
 385 magmatic arc (Argentina).

386 5.1.11. Appalachian

387 The evolution of the Appalachian composite orogenic belt involved discrete episodes of
 388 arc – continent and continent–continent collisions during the Paleozoic. Although it differs in
 389 some respects from the Scandinavian Caledonides (e.g., van Staal et al., 2009; Zagorevski and
 390 van Staal, 2011), the timing of oceanic crust and island arc construction during the development
 391 of the Appalachians was almost the same as in Ireland (e.g. Hollis et al., 2013), Scotland (Sawaki
 392 et al., 2010), and Scandinavia (e.g., Dunning and Pedersen, 1988; and Fig. 13B of Pedersen and
 393 Hertogen, 1990). Nine representative ophiolites from this orogenic belt are included in this study
 394 (Table 1).

395 5.1.12. Caledonides

396 The tectonic history of the Scandinavian Caledonides (e.g., Roberts et al., 2007; Gee et al.,
 397 2008; Andersen et al., 2012; Hollocher et al., 2012) lasted for about 200 million years (Gee et al.,
 398 2008), starting around 600 Ma with continental rifting and rift-drift, seafloor spreading tectonics

between Baltica and Rodinia. The main oceanic crust-building episodes occurred between 500 – 470 Ma, and later between 445 – 435 Ma (Dunning and Pedersen, 1988; Pedersen et al., 1991; Dilek et al., 1997; Furnes et al., 2012b). Five ophiolite complexes from this orogenic belt are included in this study (Table 1).

5.1.13. *Tasmanides*

The Tasmanides occupy approximately one third of the eastern Australian continent and include a collage of mountain belts that formed during a period of nearly 500 million years. The break-up of the Rodinia supercontinent between 750-525 Ma was followed by ca. 300 million years of mainly convergent-margin tectonics and magmatic activity (e.g., Spaggiari et al., 2003; Glen, 2005). We have included the 530 Ma Weraera terrane of the Tasmanides in this study (Table 1).

5.2. *Precambrian Complexes*

In this paper we present geochemical data from 10 greenstone belts and 42 complexes that we regard as ophiolitic in origin (Furnes et al., 2015), encompassing a time-span of nearly 3000 million years of Earth history. For detailed information the reader is referred to the comprehensive descriptions of these complexes, including regional geological maps and relevant references, presented in Furnes et al. (2015). Below we give a brief account of mafic-ultramafic complexes investigated in each of the ten greenstone belts (see also Table 1). When referring to the subdivisions of the Archean and Proterozoic eons, we have used the boundaries of Walker et al. (2013): Hadean (> 4.0 Ga); Eoarchean (4.0 - 3.6 Ga); Paleoarchean (3.6 - 3.2 Ga); Mesoarchean (3.2 - 2.8 Ga); and Neoarchean (2.8 - 2.5 Ga); Paleoproterozoic (2.5 - 1.6 Ga); Mesoproterozoic (1.6 - 1.0 Ga), and Neoproterozoic (1.0 - 0.541 Ga).

5.2.1. *Cadomian*

The Cadomian orogeny developed along the northern margin of the Gondwana supercontinent during the late Neoproterozoic (650-550 Ma) and involved the development of back-arc basin and volcanic arc ophiolites (e.g., Rossetti et al., 2015; Linnemann et al., 2008). Only one

425 Cadomian ophiolite, the Frolosh/Struma complex in Bulgaria (Kounov et al., 2012) is included in
426 this study (Table 1).

427 *5.2.2. Arabian-Nubian Shield*

428 The Arabian–Nubian Shield comprises several, ENE-WSW-trending Neoproterozoic
429 greenstone belts (Dilek and Ahmed, 2003; Yibas et al., 2003; Abd El-Rahman et al., 2009a) with
430 ages of 0.6 Ga – 0.8 Ga (Farahat, 2010). Six of these greenstone belts are examined in this study
431 (Table 1 and Fig. 11 of Furnes et al., 2015), two from the Arabian Shield and four from the
432 Egyptian Nubian Shield.

433 *5.2.3. India*

434 We have examined five greenstone belts from the Indian Craton (Table 1). Apart from the
435 Neoproterozoic Manamedu greenstone belt, the other four are of Archean age. The Paleoproterozoic
436 (3510 Ma) Iron Ore Group is among the oldest greenstone belts in the world, and is dominated by
437 basaltic and komatiitic rock units (Mukhopadhyay et al., 2012) (see Fig. 12 of Furnes et al.,
438 2015). The Manamedu greenstone sequence includes basaltic lavas, sheeted dikes, gabbros, and
439 peridotites (Yellappa et al., 2010), forming a typical Penrose-type ophiolite (Dilek, 2003a).

440 *5.2.4. Baltica and Siberia*

441 We have investigated eight greenstone belts from Baltica and Siberia (Table 1), two from
442 Finland: the Paleoproterozoic (Jormua) and Neoarchean (Suomussalmi) ophiolites (Heilimo et al.,
443 2010). The Russian greenstone belts (Skrufin and Theart, 2005) we have examined include
444 Neoproterozoic (Enganepe), Paleoproterozoic (Pilgijärvi, Kholodnikan, Avarench), and
445 Neoarchean (Tikshozero, Khizovaara) examples (see Figs. 5 and 6 of Furnes et al., 2015).

446 *5.2.5. Australia*

447 We have examined five greenstone occurrences in Australia with ages ranging from
448 Neoarchean (Gindalbie and Kurnalpi, Meekatharra-Cue), to Mesoarchean (the Whunda Group)

and Paleoarchean (the Kelly and Warrawoona Groups of the Pilbara Supergroup) (Table 1, and Fig. 13 of Furnes et al., 2015). The Pilbara Craton is one of the hotly disputed cases with disparate views concerning the origin of its Paleoarchean greenstone belts (Van Kranendonk and Pirajno, 2004; Van Kranendonk et al., 2007).

5.2.6. *China*

We have included in this study two Neoproterozoic ophiolite complexes from Chinese greenstone belts in China (Table 1, and see Fig. 7 of Furnes et al. (2015) and references therein). These are the Anhui and Jiangxi ophiolites in the Cathaysia Block (Zhou, 1989; Li et al., 1997) and the Miowan ophiolite in the Yangtze Block (Peng et al., 2012) of the South China Craton.

5.2.7. *North America*

Six greenstone belts from the North American continent have been included in this study. The youngest of these greenstone belts is the Proterozoic (0.76 Ga) Burin Group (Murphy et al., 2008) in the Avalon Zone of the southeast Newfoundland Appalachians (Table 1, and Fig. 3 in Furnes et al., 2015). The Paleoproterozoic (1.73 Ga) Payson ophiolite (Table 1) within the Mazatzal block in Arizona (USA) contains all of the mafic volcanic and plutonic components of a Penrose-type ophiolite (Dann, 1992, 1997; Dann and Bowring, 1997). The other four examples, all in the Canadian Shield (Elu Link, Abitibi, Yellowknife, Wawa) are of Neoarchean age (e.g., Cousens, 2000; Polat et al., 1999; Polat and Kerrich, 2000; Dostal and Mueller, 2013; Mvondo et al., 2013).

5.2.8. *South America*

We have examined two South American greenstone belts (Table 1). The oldest (Rio das Velhas) is Neoarchean (2.77 Ga in age), whereas the Mazaruni and Barama are Paleoproterozoic (2.25 Ga in age) and belong to the Amazonian and Sao Francisco Cratons, respectively (Cordani et al. 2009) (see also Fig. 9 of Furnes et al., 2015).

5.2.9. *Greenland*

The Precambrian rocks of southwest Greenland constitute several greenstone belts ranging in age from ca. 3.0 to 3.8 Ga (Henriksen et al., 2009). The oldest of these is in the 3.7-3.8 Ga Isua supracrustal belt (Nutman et al., 1997), which includes an ophiolite (Isua ophiolite) containing plagiogranite pods within its sheeted dike complex (Furnes et al., 2007, 2009). We have included the Isua ophiolite in this study (Table 1).

5.2.10. Africa

We have examined six of the many greenstone belts existing in the African continent (Table 1, and for location, see Fig. 10 of Furnes et al., 2015). Three of these, the Leerkranz Formation, Bas Draa, and the Chewore sequences, are all of Mesoproterozoic age (Johnson and Oliver, 2000; Bailie et al., 2011; El Bahat et al., 2013). The Birimian terrane (e.g., Vidal and Alric, 1994) in the southern part of the West African Shield is of Paleoproterozoic age, and the remaining two, i.e., the Belingwe (Zimbabwe) and the Noisy Complex of the Barberton Greenstone Belt (South Africa), are of Mesoarchean and Paleoarchean ages, respectively.

5.3. Non-orogenic Complexes

We have included three young ophiolite complexes that are not related to orogenesis, i.e. Taitao (Chile), Tihama Asir (Red Sea, Arabia), and Iceland (Table 1). The ca. 6 Ma Taitao ophiolite formed by seafloor spreading at the Chile Ridge and was accreted into the fore-arc region of the South American Plate due to ridge-trench interaction (Veloso et al., 2005, 2009; Dilek and Furnes, 2014; Shin et al., 2015). The 22 Ma Tihama Asir complex is an in-situ continental margin ophiolite, formed during the initial stages of rifting between the African and Arabian plates in the Miocene (Dilek et al., 2009). It thus forms a typical OCT crust and transitions into the young oceanic lithosphere of the Red Sea to the west. The island of Iceland has a unique geological history in that its oceanic rocks record a complex plume/mid-ocean-ridge evolution of more than 16 million-years (Foulger, 2006). Intermediate and silicic extrusive and intrusive rocks are significant components (up to 10%) of the in-situ oceanic crust in Iceland (Walker, 1966; Saemundsson, 1979; Carreter et al. 2013). The deeper parts of the Icelandic upper crust include dike swarms comparable to the sheeted dike complex in the Troodos (Cyprus) ophiolite (Walker, 1975). As Iceland is part of the Mid-Atlantic spreading system and located

above a hotspot (or plume), we regard its crustal units as part of an *in-situ*, plume-proximal MOR oceanic lithosphere (Table 1; Furnes et al., 2014).

5.4. *In-situ oceanic crust*

We have also included seven examples of *in-situ* oceanic crust in this study (Table 1) that range in age from recent to about 48 million years. These examples represent different tectonic environments of oceanic crust formation, including mid-ocean ridges, fore-arcs and back-arcs.

6. Element mobility

In this study we have used the major oxides SiO₂, TiO₂, Al₂O₃, FeO^t, MgO, CaO, Na₂O and K₂O, the incompatible trace elements Sr, Y, Zr, Nb, V, Ba, Th, and the rare earth elements (REE) in order to discriminate between different types of ophiolites (on the basis of the compositions of their basaltic rocks), and characterize the intermediate to silicic rock assemblages in them. The behaviour of major and trace elements during low-temperature alteration and low to medium grade metamorphism of oceanic rocks (mainly mafic rocks), has been evaluated previously in several other studies. The element mobility is a highly complicated process, but is in general affected and controlled by factors such as the composition and stability of the mineral phases in unaltered protoliths and in the alteration products. The compositions, temperatures and volumes of fluid phases circulating through the lithospheric system also play a major role in element mobility. Some of the major elements, in particular Ca, Na and K, have been reported (e.g., Masters and Ague, 2005) to be rather mobile during metamorphism, with fluid-driven gain of Na and Ca (and Sr), and loss of K (and Rb and Ba). However, a recent study by Ghatak et al. (2011) of mafic-ultramafic rock units that were affected by subduction-metamorphism (from unmetamorphosed to prehnite-pumpellyite to eclogite to amphibolite/garnet amphibolite) has demonstrated no systematic variations with increasing metamorphic grades. These authors have concluded that the most likely variability in Na₂O and K₂O contents reflects variations in the protolith compositions rather than their mobilization during subduction metamorphism. The general consensus is that Ti, Al, V, Y, Zr, Nb, REE (particularly HREE) and Th are relatively immobile (e.g., Scott and Hajash, 1976; Staudigel and Hart, 1983; Seyfried et al., 1988; Komiya et al., 2004; Hofmann and Wilson, 2007; Dilek et al., 2008; Furnes et al., 2012a). Seven of these

elements (Ti, V, Zr, Y, Nb, Th and Yb) have been employed for various discrimination diagrams in order to differentiate the ophiolite types, and are therefore crucial in this connection. The elements of Ba and Sr, used in the petrogenetic discussion below, are generally regarded as rather mobile. However, in our study they show patterns comparable with those of the immobile incompatible elements (e.g. Figs. 7 and 8). We hence regard them as useful in our characterization of the intermediate to felsic rocks in ophiolites (as discussed below).

7. Geochemical classification of ophiolitic basalts

In addition to distinguishing ophiolites based on their geological characteristics and tectonic environment of their igneous development (Fig. 1), we can evaluate the geochemical fingerprints of their mafic lavas and dikes in order to classify them thoroughly. In the geochemical classification of ophiolitic basalts we have used combinations of seven relatively immobile elements (Ti, V, Zr, Y, Nb, Yb, Th) in four discrimination diagrams (Zr/Ti – Nb/Yb, Th/Yb – Nb/Yb, V – Ti/1000, and $\text{TiO}_2/\text{Yb} - \text{Nb/Yb}$). This classification involves four steps (Fig. 4A-D), following the method of Pearce (2014). As the first step, we have used the Zr/Ti – Nb/Y diagram of Floyd and Winchester (1975) in order to select only the data from rocks with basaltic compositions. The second step employs the Th/Yb – Nb/Yb diagram of Pearce (2008) to discriminate subduction-related versus subduction-unrelated basalts. All basalts that plot within the OIB-MORB array (also known as the mantle array) have been considered as unrelated to subduction, whereas those plotting above it are interpreted to be influenced by subduction processes. The subduction-related basalts encompass oceanic and continental arc types, or plot within the area between the mantle array and the arc type fields. Step 3 involves further discrimination of basaltic rocks using the V/Ti ratio as a proxy for supra-subduction zone (SSZ) derivation (Shervais, 1982). The modified V-Ti diagram (Pearce, 2014) discriminates between boninite, island-arc tholeiite (IAT), and MORB fields, the latter representing the tectonic environment most distant from a trench and the associated slab of subducting oceanic lithosphere. Basalts that plot within the OIB-MORB array have been further analyzed in the $\text{TiO}_2/\text{Yb} - \text{Nb/Yb}$ diagram (Step 4), as the TiO_2/Yb ratio is a good indicator of the depth of mantle melting.

Ophiolitic basalts are first divided, following this classification procedure, as well as characteristic geological features, into two main groups: subduction-unrelated and subduction-

related (Dilek and Furnes, 2011, 2014). The subduction-unrelated group can be further discriminated into Rift/Continental Margin, MOR, and Plume/MOR types, a subdivision that is defined by their grouping in the $\text{TiO}_2/\text{Yb-Nb/Yb}$ diagram. MOR type basalts plot within the MORB array while plume-related basalts plot within the OIB array and the Rift/Continental Margin types fall into the alkaline field of the OIB array (Fig. 4). The subduction-related basalts are subdivided into back-arc, back-arc to fore-arc, fore-arc, and volcanic arc types. The back-arc, back-arc to fore-arc and fore-arc types are distinguished when the majority of their compositions plot in the oceanic arc field of the Th/Yb-Nb/Yb diagram and in the MORB, MORB-IAT-boninite and boninite fields in the V-Ti diagram. For the volcanic arc basalts the data plot mainly in the continental arc field of the Th/Yb-Nb/Yb diagram and within the IAT and boninite fields in the V-Ti diagram. A systematic classification of the majority of the ophiolites listed in Table 1 is provided in Furnes et al. (2014, 2015).

Table 2 presents a geochemical summary of 168 ophiolites based on the compilation of Furnes et al. (2014, 2015), encompassing 89 Phanerozoic, 41 Proterozoic, and 40 Archean examples. These ophiolites are divided into seven types (three types of subduction-unrelated, and 4 types of subduction-related). The various proxies used to define these different ophiolite types show a wide range and a large overlap between the various types. The most useful values are the averages (shown in bold in Table 2). The subduction-unrelated ophiolite class mostly exhibits zero subduction influence (the Th/Yb -proxy), although several Phanerozoic ophiolites reflect possible subduction fingerprint. In such cases, the ophiolites are mostly of the MORB-type (see V-Ti proxy for SSZ melting). The Rift/Continental Margin and Plume type ophiolites mostly reflect formation by shallow melting E-MORB to deep melting tholeiitic to alkaline basalt (in the Ti/Yb proxy for plume melting), whereas the MORB type indicates formation dominantly of shallow melting N-MORB.

The vast majority of ophiolites fall into the category of subduction-related types. Subduction influence in the global ophiolite record seems to increase with age into the deep time (73% Phanerozoic, 83% Proterozoic, and 95% Archean). In general, the Backarc Type exhibits the least degree of subduction influence (average around 50%), whereas the Backarc to Forearc, Forearc and Forearc/Volcanic Arc Types displays considerably higher values of subduction influence (around 70-90%). As defined by the V-Ti proxy, the Phanerozoic and Proterozoic

Backarc Type ophiolites mostly plot in the MORB field, whereas the Archean examples mostly have an island arc tholeiitic (IAT) character. The Forearc Type (all ages) falls mostly in the boninite (Bon) and IAT fields, whereas the Backarc to Forearc Type shows values intermediate between the Backarc and Forearc Types. The Volcanic Arc Type is defined predominantly by IAT and MORB in the V-Ti proxy (Table 2). The majority of the subduction-related ophiolite types are the products of shallow melting, similar to the subduction-unrelated ones.

All the ophiolites presented in Table 1 have been classified according to the geochemical signatures of their basalts employing the four steps outlined above and in Fig. 4. The intermediate (basaltic andesite, andesite and trachyandesite) and silicic (dacite, rhyodacite, rhyolite and trachyte) rocks from each of the seven types have been isolated and treated separately. In this context, we have used only the extrusive rock terms but it should be emphasized that a large component of many ophiolites is represented by intrusive rocks. Thus we apply the following names for the corresponding rocks: intermediate rocks are transitional gabbro/diorite, diorite and transitional syenite/diorite; for the silicic rocks granodiorite, transitional granite/granodiorite, granite and syenite.

In the following section, we treat first the subduction-unrelated group in the order of Rift/Continental Margin (Rift/CM), MOR, and Plume/MOR (P/MOR) types, and then the subduction-related group, starting with Backarc (BA), Backarc to Forearc (BA-FA), Forearc (FA), and Volcanic Arc (VA) types. This order follows the general evolutionary development of oceanic lithosphere in an ocean basin from its opening and widening (via sea-floor spreading) to closure stages (via intra-oceanic subduction) (Dilek et al., 2005).

8. Geochemical classification of the intermediate to silicic igneous rocks

We have plotted the intermediate to silicic rocks of the seven types of ophiolites in the Zr/Ti – Nb/Y classification diagram (Fig. 5). For each ophiolite type, the percentage of examples that plot within the different fields of Fig. 5 have been calculated and plotted in Fig. 6. We recognize that some sampling bias may exist, and that the anticipated variations in standards among the analytical facilities in different laboratories where the compiled data have been produced may have certain effects on our synthesis here. However, we think that our systematic

approach to fingerprinting and classifying intermediate to silicic rock units in ophiolites provide a realistic, first-order geochemical characterization of their petrogenesis.

The intermediate and silicic rocks in the Rift/CM type ophiolites are predominantly basaltic andesite and andesite (67%), although trachytes and rhyodacites also occur in significant and similar proportions (14% for both). The Kaczawa ophiolite (no. 62 in Table 1) is a representative example of the Rift/CM type among the ophiolites we have evaluated in this study. Basaltic andesite, andesite and rhyodacite occur in comparable proportions (49% and 41%, respectively) among MOR-type ophiolites; trachyandesite and trachyte are the minor components. The Corsica ophiolite (no. 6 of Table 1) is a representative example of MOR-type ophiolites (Saccani et al., 2008). The predominant, evolved rock types for the P/MOR type are rhyolite (67%) followed by trachyte (22%). Basaltic andesite/andesite, rhyodacite and trachyandesite (8, 6, 6%, respectively) make up only minor proportions in this ophiolite type. The most representative occurrence of the P/MOR type is Iceland (no. 149 of Table 1). We emphasize that there are significant geochemical differences between tholeiitic andesite (icelandite) and orogenic andesite, specifically demonstrated by higher TiO_2 and FeO^t , and lower Al_2O_3 and K_2O contents in the former (e.g., Jakobsson et al., 2008; Ridley, 2012).

The subduction-related BA type of ophiolite is characterized by approximately equal proportions of basaltic andesite and andesite, and rhyodacite (45 and 41%, respectively), accompanied by 9% trachyandesite and minor proportions of trachyte and rhyolite (4 and 1%, respectively) (Fig. 6B). The Troodos ophiolite is a representative example of the BA type (no. 16 of Table 1). The BA-FA, FA, and VA types all show similar proportions of basaltic andesite and andesite as the predominant composition (72, 73 and 74%, respectively). The BA-FA and FA types also exhibit similar proportions of rhyodacite (21 and 22%, respectively), minor proportions of trachyandesite (4.5 and 5%, respectively), and insignificant proportions of trachyte and rhyolite. Rhyodacite and trachyte occur in similar proportions (12%), with insignificant proportions of trachyte (2%), in the VA type ophiolites. The Mirdita, Betts Cove and Yakuno ophiolites (nos. 11, 97, 54 of Table 1) are good examples of BA-FA, FA, and VA, respectively.

8.1. Harker diagrams

In order to provide an overview of the compositional variations among the examples from the subduction-unrelated and subduction-related ophiolites, we have plotted a series of Harker diagrams (Figs. 7 and 8). The data for the mafic rocks are from Furnes et al. (2014, 2015 and references therein), and include 5620 analyzed samples from a total of 228 ophiolites. Of these, 51 ophiolites are subduction-unrelated (MOR=25, P/MOR=16, and Rift/CM=10) and contribute 1257 analyses (653, 451, and 153, respectively). The subduction-related group includes 177 ophiolites (BA=96, BA-FA=46, FA=15, and VA=20) and contributes 4363 analyses (1797, 1623, 513, and 430, respectively).

8.1.1. Subduction-unrelated

The Harker diagrams for subduction-unrelated ophiolites are shown in Fig. 7. The mafic and intermediate rocks define some overlap with respect to their SiO₂ contents, an artifact of the presence of boninites with high SiO₂ contents. In the mafic rocks the spread of some of the major elements, particularly TiO₂, Al₂O₃, MgO and CaO, is higher than that observed for the intermediate to silicic rocks. This phenomenon may stem from several different factors: some of the analyzed samples may be cumulates; alteration or metamorphism may have changed the compositions of some samples; and, some of the samples with very high MgO values may be komatiite, boninite, or cumulates. The most likely candidate samples representing cumulates, as reflected by their anomalously high contents of Al₂O₃, MgO, Cr and Ni, have been deleted from the data base, and no longer considered in our synthesis. The major element compositions (TiO₂, Al₂O₃, FeO^t, MgO, CaO and Na₂O) of the intermediate to silicic rocks, together with the mafic rocks, all define broad trends with increasing SiO₂ contents. At a particular SiO₂ concentration, the TiO₂ variation is 1wt.%, for K₂O about 4-5wt.%, for CaO and Na₂O about 5wt.%, for Al₂O₃ and MgO about 6wt.%, and for FeO^t about 8wt.%. The FeO^t, MgO and CaO contents all decrease continuously as the SiO₂ values increase. The Al₂O₃ and Na₂O trends show inflections (at around 65% SiO₂) from increasing to decreasing concentrations. In the case of trace element behavior, the Rift/CM and P/MOR ophiolites show large spreads in concentrations. The MOR type shows large variations in Sr, Y and Zr, and low concentrations of Nb, Ba and Th.

8.1.2. Subduction-related

Harker diagrams for subduction-related ophiolites are shown in Fig. 8. The data for all of the subgroups, i.e., BA, BA-FA, FA and VA, define large spreads in all the elements at any given SiO_2 (considerably more than for the subduction-unrelated ophiolites). They define rather similar trends with increasing SiO_2 contents and hence have not been distinguished in the figure. The average concentrations for each of the subgroups display similarities. For the mafic rocks the concentrations of some incompatible trace elements increase to about 50% SiO_2 , and then decrease. The peak concentrations of the incompatible elements Sr, Ba and La reach ~70 wt. % SiO_2 , with a marked, saddle pattern of maximum concentrations between 50 and 70 wt.% SiO_2 . The highest concentrations of Y, Zr and Nb of the intermediate to silicic rocks are defined by two maxima at ca. 65wt. % and ca. 75wt. % SiO_2 . The concentrations of Zr, Nb and La in the intermediate to silicic rocks (Fig. 8) are in general considerably lower than those in the subduction-unrelated ophiolites (Fig. 7).

8.2. Rare Earth Elements

We present chondrite-normalized REE data from intermediate to silicic rocks from 108 ophiolites, with a total of 833 analyzed samples. Of these ophiolites the 28 are subduction-unrelated (Rift/CM = 6, MOR = 19, P/MOR = 3) with a total of 188 analyses, and 80 ophiolites are subduction-related (BA = 39, BA-FA = 24, FA = 5, VA = 12) with a total of 645 analyses. In Figs. 9 through 15 we show a selection of characteristic REE patterns for each type and from the youngest to the oldest ophiolites.

8.2.1. Subduction-unrelated

The REE patterns of the intermediate to silicic rocks of the Rift/CM type ophiolites, ranging in age from 450 Ma to 2429 Ma, are shown in Fig. 9. The $(\text{La/Lu})_N$ ratios of these ophiolites are all positive, with the majority ranging from 10 to 20.

The REE patterns of the MOR-type ophiolites are shown in Fig. 10. The majority (ca. 70%) of these samples exhibit rather flat patterns. Nearly 34% of the samples have $(\text{La/Lu})_N$ ratios < 1, and 36% have $(\text{La/Lu})_N$ ratios between 1 and 2. However, the tonalities from the Aruba Lava Formation define slightly downward-convex REE patterns. Such patterns are comparable to

subduction-originated rocks, but the close temporal relationships between the MOR-type basalts and the tonalities make such an origin implausible (White et al., 1999).

The REE patterns of the P/MOR types (Fig. 11) are all LREE-enriched, with the majority of them (ca. 75%) having $(\text{La/Lu})_N$ ratios between 1 and 10 (and mainly 5-10).

8.2.2. Subduction-related

The REE patterns of intermediate to silicic rocks in BA type ophiolites are shown in Fig. 12. The majority of this sub-group is characterized by relatively flat to moderately LREE-enriched patterns, with ca. 60% and 25% having $(\text{La/Lu})_N$ ratios between 1 and 10 (mostly 1 to 3), and < 1 , respectively. A few samples from the oldest complex (Elu Link, northeast Slave Craton, Canada; no.134 of Table 1) show exceptionally high $(\text{La/Lu})_N$ ratios.

Intermediate to silicic rocks from the BA-FA type ophiolites (Fig. 13) show REE patterns highly similar to those from BA ophiolites. However, the former have a considerably higher proportion of REE patterns with $(\text{La/Lu})_N$ ratios of 10 to 20 than the latter. As with the BA type, the majority of the oldest complexes (Hutti and Southern Iron Ore Group, India; nos. 114 and 116 of Table 1, respectively) show steep REE patterns with low HREE.

The REE patterns of intermediate to silicic rocks from the FA type ophiolites (Fig. 14) are similar to those from the BA-FA type ophiolites, but with a higher proportion of $(\text{La/Lu})_N$ ratios between 1 and 10. However, there are pronounced differences between the rather flat patterns of the silicic rocks of the Izu-Bonin-Mariana samples and the saddle-shaped patterns of the samples from the Pekulney Ridge.

The REE patterns of intermediate to silicic rocks from the VA type ophiolites are shown in Fig. 15. The majority of the samples are characterized by high $(\text{La/Lu})_N$ ratios. The oldest VA type ophiolite, the 3.4 Ga Noisy Complex of the Barberton Greenstone Belt (South Africa; no. 146 of Table 1), displays particularly high $(\text{La/Lu})_N$ ratios (16-30).

8.3. Multi-element diagrams

We present MORB-normalized multi-element diagrams for intermediate to silicic rocks from 43 ophiolites and a total of 213 samples. Of these 43 ophiolites, 11 complexes are subduction-unrelated (Rift/CM = 4, MOR = 5, P/MOR = 2) with a total of 94 analyses, 32 are subduction-related complexes (BA = 15, BA-FA = 4, FA = 4, VA = 9) with a total of 119 analyses. Figs. 16-19 show a selection of multi-element patterns for each type, organized according to the age of the ophiolites.

8.3.1. Subduction-unrelated

The multi-element patterns of intermediate to silicic rocks from Rift/CM (ranging in age from 450 Ma to 1970 Ma), MOR (with ages ranging from < 1 Ma to 175 Ma) and P/MOR (with ages ranging from < 1 to 21 Ma) ophiolites are shown in Figs. 16 and 17. The Rift/CM and P/MOR types display similar patterns, characterized by relative enrichment in the most incompatible elements (Th, Nb, La), pronounced negative anomalies in Ti and Sr, and minor negative Eu anomalies. The MOR type ophiolites are, on the other hand, much less enriched in incompatible elements and display negative Ti, Sr and Eu anomalies of varying sizes. Some samples show positive Sr anomalies. An unexpected feature is that most of the MOR samples exhibit negative Nb anomalies (Fig. 16).

8.3.2. Subduction-related

The multi-element patterns of intermediate to silicic rocks from BA (30 Ma to 2690 Ma), BA-FA (38 Ma to 3510 Ma), FA (47 Ma to 480 Ma), and VA (late Cambrian-Middle Ordovician to 3455 Ma) ophiolites are shown in Figs. 18 and 19. A common feature of the intermediate to silicic rocks in all of these ophiolite types is a pronounced negative Nb anomaly, though in the case of the FA type Lac Brompton ophiolite (no. 95 of Table 1) this feature is weak (Fig. 19). Ti and Sr display negative anomalies in the majority of the samples, although a small number of samples exhibit positive Ti and Sr anomalies. Apart from Ti, Sr and Nb, the MORB-normalized ratios increase towards the more incompatible elements (with Th being the most incompatible). However, some of the examples show relatively flat patterns (i.e., those from the Izu-Bonin-Mariana and Troodos, nos. 156 and 16, respectively), whereas some others show very steep patterns (e.g. Northland and Annieopsquotch, nos. 61 and 94, respectively).

8.4. *Nb, Eu and Ti anomalies*

The Nb, Eu and Ti anomalies in the multi-element diagrams are characteristic of subduction-unrelated and subduction-related ophiolites in general (Figs. 16-19). They are mostly negative, although there are also some positive anomalies. Figs. 20-22 summarize the Nb, Eu and Ti anomalies in histograms. The Nb-anomalies in these samples from subduction-unrelated ophiolites (Fig. 20A) range from strongly negative to positive, with a peak in the -50 to -40 interval. All samples from subduction-related ophiolites show highly negative anomalies in the -100 to -70 interval (Fig. 20B-E); no samples show positive anomalies. The largest negative Nb-anomaly in samples from BA-FA ophiolites is in the -100 to -90 interval, whereas samples from the other types (BA, FA, and VA), it is in the -90 to -80 interval. The Eu-anomalies for the subduction-unrelated and subduction-related ophiolites are summarized in Fig. 21. The Eu-anomalies in samples from subduction-unrelated ophiolites display a rather uneven pattern with two slight peaks at the -60 to -50 and -30 to -20 intervals, whereas samples from the subduction-related ophiolites show a symmetrical distribution around a peak at -20 to -10. The Ti-anomalies for the subduction-unrelated and subduction-related ophiolites are also highly different (Fig. 22). The majority of the samples from subduction-unrelated ophiolites exhibit negative Ti-anomalies with the largest in the -100 to -80 interval, while the samples from subduction-related ophiolites show the largest negative anomalies in the -50 to -40 interval. Furthermore, the samples from subduction-related ophiolites also display a higher proportion of positive Ti anomalies than those from the subduction-related ophiolites (Fig. 22).

9. Petrogenesis

It is appropriate to point out at the outset that the geochemical data used in this study are compiled from numerous papers published over a considerable period of time. Thus, some uncertainties regarding the accuracy and the consistency of the geochemical data may exist, such as sampling bias, inter-laboratory analytical discrepancies, and the completeness and quality of the data. However, we are confident that our systematic approach to analyzing and synthesizing the extant geochemical data on the global ophiolite record provides an internally coherent discussion for the petrogenesis of intermediate to silicic rock units in ophiolites.

The results of experimental and modeling studies (e.g., Spulber and Rutherford, 1983; Koepke et al., 2004; 2007; Annen et al., 2006; Grimes et al., 2013; Scaillet et al., 2016) and field studies (e.g., Gunnarson et al., 1998; Rollinson, 2009; 2014; Beier et al., 2015; Lundstrom and Glazner 2016) have shown that intermediate to silicic rocks can form through several different processes, including: (1) fractional crystallization of basaltic magma; (2) partial melting of hydrated basalt or gabbros in the roof zone of magma chambers; and (3) liquid immiscibility in evolved basalt. The major and trace element compositions of intermediate and silicic rocks in ophiolites, as illustrated in the Harker diagrams (Figs. 7 and 8), exhibit large variations at any SiO_2 content, suggesting that different petrogenetic processes were involved in their production. This inference is supported by the large variations shown both in the REE patterns of the silicic rocks (Figs. 9-15) and the multi-element diagrams (Figs. 16-19). The principal origin of intermediate to silicic rocks is largely a result of fractional crystallization or partial melting (e.g., Gillis and Coogan, 2002; Koepke et al. 2004, 2007; Brophy, 2009; Rollinson, 2009; Grimes et al., 2013; Beier et al. 2015), although there is evidence that silicic magmas may also result from liquid immiscibility (e.g., Philpotts, 1982; Jakobsen et al., 2005), particularly in gabbroic complexes, such as the Skaergaard Layered Series (Veksler et al., 2007). Charlier et al. (2011) have also suggested that liquid-liquid unmixing producing Fe-rich and Si-rich liquids may contribute significantly to the "Daly Gap" along the liquid line of descent. The "Daly Gap" describes the lack of intermediate compositions such as andesites and trachyandesites in various volcanic provinces, in which only bimodal basalt-rhyolite suites are found. In our database, however, there is no indication of such a compositional gap (see Figs. 5, 7 and 8), and hence we regard the potential contribution of immiscible liquids as insignificant in this context. Hence, even though we do not exclude the presence of immiscible liquids, we do not consider this process further in this study.

In the following section, we discuss the importance of fractional crystallization and partial melting in the evolution of the intermediate to silicic rock suites in the ophiolites we have examined in this study.

9.1. Partial melting and fractional crystallization

Using the results of an experimental study of hydrous partial melting of cumulate gabbros (Koepke et al., 2004) and applying the extant D-values (partition coefficients) in the literature, Brophy (2009) has modeled the behavior of REEs and SiO_2 during melting and fractional crystallization of mid-ocean ridge basalt and gabbro. His modeling work has shown that hydrous melting (equilibrium and fractional) can yield melts with $\text{SiO}_2 > \sim 62$ wt. % with decreasing concentration of REE as SiO_2 values increase. On the other hand, fractional crystallization of basaltic melt results in increasing concentrations of REE. This model, in which gabbroic parental material is used as the starting material for both partial melts and fractional crystallization, we regard as most appropriate for the generation of intermediate to silicic melts in ophiolite complexes. In the structurally lower sections of many ophiolites, cumulate as well as varitextured and isotropic gabbros are abundant (Fig. 2). We infer that the magmatic mush that produced these lower crustal rocks provided the parental material for intermediate to silicic liquids higher up in the oceanic crust. We are aware of the numerous experimental studies related to the production of silicic magmas, but wish to make it clear that in this first-order attempt to explain the production of silicic melts in ophiolites, the details of such studies are beyond the scope of the current study. We have applied the results of the modeling of Brophy (2009) in an attempt to explain the principal behavior of incompatible and compatible elements with progressive melting and fractional crystallization, using La and SiO_2 . In modeling of trace-element enrichment with increasing SiO_2 we use C_1/C_0 (where C_0 = original concentration of La in a starting liquid, and C_1 = concentration of La in the evolved liquid). The enrichment factor C_1/C_0 is dependent on relevant bulk distribution coefficients for partial melting and fractional crystallization. We have chosen to employ the following initial values (C_0) for the melting and fractional crystallization models: 2 and 10. For the melting model we have chosen batch melting, and for the fractional crystallization model, we have chosen the experimental data produced under the QFM buffer conditions (see Fig. 6 of Brophy, 2009). The enrichment factors (C_1/C_0) as a function of increasing SiO_2 are shown in Fig. 23A and B.

9.1.1. Subduction-unrelated

The La – SiO_2 relationships in the subduction-unrelated ophiolite types are shown in Fig. 23C. In the MOR type ophiolites the relationships are consistent with melting of a

gabbroic/basaltic parent within the limits chosen for C_0 . The majority of the La – SiO₂ data for the P/MOR type oceanic crust (predominantly from Icelandic rocks) is compatible with fractional crystallization of basaltic parent melt. This inference is in agreement with the test performed on Icelandic silicic rocks by Brophy (2009). A few samples, however, are more in accord with partial melting of a gabbroic/basaltic parent, in agreement with the suggestion that some Icelandic silicic magmas are produced by partial melting of hydrothermally altered basaltic or gabbroic rocks (e.g., Gunnarson et al., 1998). Silicic rocks in the Rift/CM type ophiolites show distributions that can be explained by either partial melting or fractional crystallization.

9.1.2. Subduction-related

The La – SiO₂ relationships of intermediate to silicic rocks in the four types of subduction-related ophiolites are shown in Fig. 24. There exists a large geochemical database for the BA and BA-FA types due to the predominance of ophiolites, which evolved in these tectonic settings (in the order of 70 – 80%) as reflected in the global ophiolite record (Furnes et al., 2014, 2015).

The majority of the La concentrations in the BA type ophiolites define a dense cluster between ca. 2 – 15 ppm at all SiO₂ concentrations (Fig. 24A), compatible with melting model 2 (i.e. $C_0 = 2$). The extremely low concentration of incompatible trace elements, TiO₂ and K₂O in the majority of the silicic rocks (Fig. 8) is compatible with the generation of their magmas from depleted mafic sources such as gabbroic cumulates. This interpretation has been proposed as the most common origin of silicic magmas in oceanic crust and ophiolites (e.g., Koepke et al., 2007; Rollinson, 2009). However, there is a large scatter in the La concentrations and some samples have very high La values (ca. 70 ppm) consistent with fractional crystallization of basaltic parent melt. A good, modern analogue may be the silicic rocks recovered from the Manus backarc basin (no. 152 of Table 1) that were produced by closed-system fractional crystallization from a basaltic parent melt (Beier et al., 2015). Subsequent fractional crystallization of magma produced by partial melting may also take place, taking La concentrations away from the partial melt trend, as indicated by black arrows (Fig. 24A).

The data from the BA-FA ophiolite type are very similar to those for the BA ophiolites. However, the dense cluster of the La concentrations is wider than that of the BA, between ca. 2 – 20 ppm (Fig. 24B).

Although the data from the FA ophiolites are limited, the majority of the La concentrations from these trench-proximal suprasubduction zone ophiolites fall along the trend for melting model 2 (Fig. 24C).

The data for the VA ophiolites are scattered, analogous to those from the BA and BA-FA types. However, a substantial amount of the data is clustered around the trends of the two melting models (Fig. 24D), suggesting a more varied genesis of the VA silicic rocks compared to the other types of ophiolites.

In Fig. 24 we have indicated the average La and SiO₂ concentrations of intermediate to silicic rocks in Phanerozoic, Proterozoic and Archean ophiolites. In all the datasets (except for the FA for which the data is limited), the average La values for the Archean samples are higher than for those displayed by the Proterozoic and Phanerozoic samples.

Adakites make up an important group of intermediate to silicic rocks that is mainly restricted to subduction-related tectonic settings, and are characterized by particularly high Sr and low Y, and hence high Sr/Y ratios (e.g., Martin, 1999). In Fig. 25A-C all the data (1312 samples) (subdivided into Phanerozoic, Proterozoic and Archean) have been plotted in Sr/Y – Y diagrams, discriminating adakites from a typical, subduction-related calc-alkaline andesite-dacite-rhyolite (ADR) association. We have used the discrimination diagram of Castillo (2006), and have accepted rocks with Sr/Y > 37 as adakitic. These diagrams show that in the Phanerozoic (870 samples), Proterozoic (105 samples) and Archean (337 samples) subduction-related ophiolites contain ca. 15%, 33% and 35% adakites, respectively. Figure 25D-F demonstrates that the majority of the samples plot in the high-SiO₂ field (Martin et al., 2005). None of the samples from the Rift/CM and P/MOR ophiolites plot in the adakite field, whereas ~ 5% of the samples from MOR ophiolites plot in the adakite field (not shown).

Although adakites may form in different ways (e.g., Castillo, 2012), it is widely accepted that high-SiO₂ adakites are mainly a result of melting of altered – hydrated basalts in a subducted slab. Low-SiO₂ adakites, on the other hand, appear to be the partial melting products of upper

mantle peridotites, which were modified by silicic melts (Defant and Drummond, 1990; Martin et al., 2005; Yin et al., 2015). Under appropriate geothermal gradients along the Benioff zone (25-30°C/km), adakitic magmas can be produced at temperatures of 650-700°C at pressures of 8-18 kbar (Martin, 1999). During ascent, these melts may be modified by reaction with the surrounding mantle. The high Sr can be accounted for by ~ 20% melting of a MOR-type basalt containing 100-200 ppm Sr (Rollinson, 1993). In addition, the high Sr indicates that plagioclase was not separated, as also indicated by the lack of negative Eu-anomalies in ~ 20% of the data (Fig. 21). The low Y (and HREE) contents in the same rocks (Fig. 25A-C) indicate that garnet and/or hornblende were residual phases in the source rocks (e.g., Martin, 1999).

The La – SiO₂ relationships of the intermediate to silicic rocks from subduction-unrelated and subduction-related ophiolites (Figs. 23 and 24) indicate that partial melting of a gabbroic or basaltic source rock was the most important petrogenetic process. However, fractional crystallization from basaltic parental melt appears to be more important for the subduction-unrelated than for the subduction-related ophiolites. This is particularly the case for the P/MOR types, and to some extent for the Rift/CM ophiolites. This interpretation is supported by the histograms showing the distribution of Eu-anomalies and TiO₂-depletion (Figs. 21 and 22). Both figures show a considerably higher percentage of negative values for the subduction-unrelated ophiolites than for the subduction-related ones, attesting to increased importance of fractional crystallization in the former. Indeed, nearly 27% of the subduction-related ophiolites have positive Eu-anomalies, indicating accumulation of plagioclase in the magma, compared to ~18% for the subduction-unrelated ophiolites (Fig. 21).

9.2. Enrichment of non-conservative elements; proxies for subduction processes

Enrichment of non-conservative elements (i.e., Cs, Ba, Th, U, K, La, Ce, Pb, Sr, P, Nd) in the mantle wedge above subduction zones occurs when fluids and silicic melts are released from subducted, altered oceanic crust and associated sediments move into and react with upper mantle peridotites (e.g., Pearce and Parkinson, 1993; Hawkesworth et al., 1997; Elburg et al., 2002; Pearce et al., 2005). The transport of non-conservative incompatible elements (i.e., Ba and Th, see below) in hydrous fluids differs significantly. While Ba is highly mobile and released over a wide range of temperatures, Th is much less mobile in a fluid phase (e.g., Hawkesworth et al., 1997).

Therefore, fluids released from subducted material and their interaction with melt source regions result in high Ba/Th values in the magmas generated. Th-enrichment is enhanced, on the other hand, by its incorporation into melts from subducted sediments (Plank and Langmuir, 1998). High Th concentrations in silicic rocks are thus consistent with magmas that originated from sources that were enriched in Th by melts derived from subducted sediments (e.g., Plank and Langmuir, 1998; Plank, 2005).

In contrast to the non-conservative elements, conservative elements (e.g., Nb, Ta, Zr, Hf, Ti, HREE) are not transported in fluids, resulting in negative anomalies in MORB-normalized, multi-element diagrams (see Figs. 16-19). One of the most conspicuous conservative element negative anomalies is the “Nb-anomaly”, commonly displayed by subduction-related magmatic rocks. Several models have been proposed (e.g., Kelemen et al., 2004) to explain negative Nb anomalies. A recent study of Baier et al. (2008) suggests that the main reason for Nb-depletion may be a consequence of the low solubility of Nb in a water-rich environment in the shallow parts of subduction zones.

In a comprehensive study, the behavior of non-conservative versus conservative elements and their ratios in the Mariana arc-basin system, Pearce et al. (2005) developed some proxies for shallow, deep, and total subduction contributions. High Ba/Th characterizes the lower-temperature fluid component because both elements are mobilized in melts, but only Ba is mobilized in fluids. Hence, the Ba/Th ratio is a proxy for shallow subduction enrichment. High Th/Nb characterizes the high-temperature portion of a subduction zone, and thus the Th/Nb ratio is a proxy for deep subduction enrichment. The Ba/Nb ratio is a proxy for the total subduction enrichment.

Here, we compare the distribution of these ratios, plotted against Th, for mafic and associated intermediate to silicic rocks from subduction-related ophiolites. The data are presented in Fig. 26. In all the plots in Fig. 26 there is a pronounced scatter, and at any given Th-value the elemental ratios (Ba/Th, Th/Nb, Ba/Nb) vary by 2 to 4 orders of magnitude. There is a substantial overlap between the mafic and silicic rocks, but the data for the silicic rocks are displaced to higher Th-values than for the mafic rocks. Since the bulk distribution coefficients for Ba, Th, Nb are rather similar (Pearce et al., 2005), the ratios of these elements should not change significantly during partial melting and/or fractional crystallization. Thus, the scatter in the ratios as shown in

Fig. 26 can be attributed mainly to enrichment of Ba and Th during shallow and deep subduction. The average values of BA and BA-FA rocks are very similar, though the Ba/Th ratio (shallow subduction component) of the BA is slightly higher than that of the BA-FA. The largest shallow subduction component is shown by the FA generated silicic rocks. We note, however, that the amount of data is much smaller for FA ophiolites than other SSZ environments. The highest average value of the deep subduction contribution (Th/Nb ratio) is shown by VA rocks, which also show the highest Th-values.

The substantial variations in the ratios shown in Fig. 26 may be a result of a combination of several factors, including: (1) the duration of subduction, (2) subduction rate, and (3) character of the subducted material. The duration of subduction in the igneous development of an ophiolite is difficult to constrain and is largely unknown. The longer the subduction lasts at a given site, the higher the enrichment of non-conservative elements is in a potential melt source. The subduction rate and hence the volume of crust subducted per unit time may differ significantly along subduction zones, for example from about 20 to 170 mm/year at the South Sandwich and the Tonga Trenches, respectively (Plank and Langmuir, 1998). As to the third factor, the sedimentary cover on the subducting oceanic crust also differs considerably both in terms of its lithology and geochemistry. Lithologically, the sedimentary cover comprises siliceous ooze, chert/radiolarite, clay, carbonate, silt/sand and volcanoclastic sediments. The proportions of these components may vary considerably, and the variations in their compositions (e.g., Ba, Th, Sr and REE) may be more than an order of magnitude (Plank and Langmuir, 1998). Furthermore, the sedimentary cover on an oceanic plate with abundant detritus derived from an old continental source may exhibit considerably higher concentrations of Th than those in an oceanic setting. For example, the sedimentary cover on the oceanic plate subducting at the East Sunda, Java and Sumatra Trenches has had a long history of sedimentation from the Australian continent, and it thus has high Th (about 8-10 ppm). In contrast, the sedimentary cover on the Pacific Plate subducting at the Mariana and Izu-Bonin Trenches contains around 2.5 ppm Th. The Nb contents, on the other hand, show rather minor variations (Plank and Langmuir, 1998).

10. Summary and geodynamic models

In order to explain the geochemical character of the intermediate to silicic rocks of the various ophiolite types, it is relevant to review the most significant aspects of the mafic rocks in the same ophiolite complexes. This is because the geochemical characteristics of the mafic rocks are inherited by the intermediate and silicic rocks, as a result, of either fractional crystallization or partial melting processes.

Fig. 27 summarizes the tectonic settings of subduction-unrelated (A), and subduction-related (B) ophiolites, and their igneous processes. In the continuous progression from rifting to drifting (e.g., Dilek et al., 2005; Liao and Gerya, 2015) the Rift/CM ophiolites represent the initial generation of oceanic crust (Fig. 27A1), forming a rather simple magmatic and sedimentary assembly (e.g., Dilek and Furnes, 2011, 2014) compared to the MOR ophiolites, with their mature oceanic crust (Fig. 27A2). P/MOR ophiolites (Fig. 27A3) are special cases of crustal growth at plume-proximal oceanic ridges (e.g. Iceland) or near oceanic plateaus (Dilek, 2003b). The magmatic evolution of these types of oceanic crust is dominated by ultramafic to mafic intrusions and lavas of E- to N-MORB composition (e.g., Dilek and Furnes, 2011; Saccani et al. 2015), produced by decompression melting of fertile lherzolitic mantle. However, the Rift/CM, MOR and P/MOR ophiolites in this compilation all contain minor to significant volumes of intermediate to silicic rocks (Fig. 5; Table 1). Geochemically the majority of P/MOR rocks, and a subordinate part of Rift/CM rocks are high in incompatible elements (Fig. 23). Most of the intermediate to silicic rocks of the MOR ophiolites, a substantial part of the Rift/CM type, and a minor part of the P/MOR ophiolites have low to very low contents of incompatible elements (e.g., La) at all SiO_2 concentrations (Fig. 23).

The tectonic settings and physical constraints of the structurally and geochemically more complex subduction-related FA, BA and BA-FA ophiolites (see Fig. 2) are shown in Fig. 27B. A fourth category, the VA ophiolites (see Fig. 2 of Dilek and Furnes, 2014), differ from these three in that they have a thicker, more arc-like crustal architecture with a well-developed middle crust (Dilek and Furnes, 2011). The crustal structure and main melting processes at the spreading centers, yielding mafic rocks of the subduction-related ophiolites, are comparable with those of the subduction-unrelated MOR ophiolites. However, processes taking place at the interface between the subducting slab and the overlying mantle result to varying extent in changes in the solidus temperature and geochemistry of the mantle. Thus, the geochemical signatures of the

basaltic magmas, and associated gabbroic rocks (isotropic and layered gabbro), generated in subduction-related environments may differ profoundly from those generated in subduction-unrelated settings. The mantle wedge above the subduction zone becomes enriched in fluid-mobilized elements (Ba and several other non-conservative elements), whereas Th can be transferred into the mantle by melts derived from subducted sediments. As H₂O is released from subducting metabasalt and overlying sediments, the solidus of the mantle peridotite is lowered by as much as 500-600°C compared to the dry solidus at depths of 75 – 120 km (Mysen and Bietcher, 1975; Grove et al., 2006), and the degree of melting increases upward and away from the subduction zone as the temperature increases (Grove et al., 2006; 2009). The Th concentrations in the mafic melts and subsequently the intermediate and silicic rocks derived from the mafic rocks may vary by two orders of magnitude (Fig. 26), controlled by several factors such as the original Th concentration in the mantle, the Th concentration in the subducted sediments (e.g. Plank and Langmuir, 1998), the volume of sediment, and the subduction rate. Furthermore, sediments may detach from the slab/mantle interface to become a sediment reservoir in the back-arc mantle (Currie et al., 2007; Gonzalez et al., 2015) and get incorporated into melts distant from the subduction zone (Fig. 27B). The basaltic and gabbroic rocks, with highly variable concentrations of non-conservative elements (e.g. Ba and Th), may in turn be the source of intermediate and silicic melts, generated at interfaces between basaltic magma chambers and the overlying upper crustal rocks.

In Fig. 27C we summarize the main processes inferred as responsible for the formation of the intermediate to silicic rocks, and propose a tentative quantification of the importance of these processes in different ophiolite types. Fractional crystallization appears to have been an important process in the formation of the intermediate to silicic rocks in Rift/CM and P/MOR ophiolites in the subduction-unrelated class, as well as VA ophiolites in the subduction-related class. This inference suggests that magma chambers might have operated as closed-systems for a sufficient amount of time for fractionation and compositional zoning of the magma to take place, as illustrated in Fig. 27C1. A closed-system magma chamber operation appears to be a common phenomenon reported from numerous volcanic systems (e.g., Hildreth, 1981; Widom et al., 1992; Watanabe et al., 2006). Fractional crystallization was likely to be of subordinate importance, on the other hand, during the igneous construction of the MOR, BA, FA, and BA-FA ophiolites. Partial melting of hydrothermally altered crust was the most common process resulting in the

majority of the intermediate to felsic rocks of the Rift/CM (estimated to 60%), MOR (estimated to 95%), and the BA, FA and BA-FA (estimated to 75%) ophiolites, as well as a high proportion of the VA ophiolites (estimated to 40%). This process, in which an upward-migrating melt lens impinges upon, reheats and melts the basal part of a previously crystallized magma chamber, has been proposed by several investigators for the origin of plagiogranites (e.g., Flagler and Spray, 1991; France et al., 2010; Grimes et al., 2013). Experimental partial melting of hydrothermally altered crust shows that melting starts at a temperature exceeding 850°C (Gillis and Coogan, 2002; France et al., 2010).

In Fig. 27C2 we illustrate a scenario whereby various, hydrothermally altered components of an ophiolite, i.e. the sheeted dike complex, isotropic gabbro, and cumulate gabbro, are in contact with a melt lens (magma front), and undergo partial melting. This scenario, in which the parental rocks range from solidified liquid (dikes) to cumulate gabbro, may result in intermediate to silicic melts with highly variable concentrations of incompatible trace elements. We show this development by numbers (1-3) in Fig. 27C2 and correspondingly in the La-SiO₂ diagram, in which partial melts from the dikes (marked 1) acquire the highest concentration of incompatible trace elements (e.g., La), and those from the cumulates (marked 3) obtain the lowest concentration. When a melt lens solidifies (producing dikes + gabbro) and is hydrothermally altered, impingement of a younger melt lens from below may result in the production of a new generation of intermediate to silicic melts, and this process may be repeated several times. This model is consistent with field observations in ophiolites, where plagiogranite bodies crosscut plutonic gabbros and basalt dike swarms, and are in turn crosscut by younger basalt dikes (Fig. 2). Melts derived from these three different components (dikes, isotropic and cumulate gabbro) may also have a profound influence on the concentration of Th, and thus be partly responsible for the large concentration range observed in Fig. 26.

Slab-generated melts that result in intermediate to silicic rocks of adakitic character, described above and illustrated in Fig. 25, represent a relatively small fraction (~15% in Phanerozoic ophiolites) of subduction-related ophiolites (Fig. 27C3). Some adakites are very low in incompatible elements (e.g., La, see Fig. 27C3), suggesting that these melts were generated from highly-depleted MORB. It is interesting to note, however, that Precambrian ophiolites seem to contain a higher percentage of adakites (ca. 35%). This observation indicates that physical

conditions during the Precambrian were more favorable for slab melting than during the Phanerozoic. This phenomenon may have resulted from higher geothermal gradient along relatively shallow-dipping subduction zones during the Precambrian (e.g., Yin et al., 2015; Palin and White, 2016). This inference is also consistent with the observation that the rare low-SiO₂ adakites (Fig. 25D) are of Phanerozoic age, i.e. they formed within the mantle (rather than at the slab/mantle interface) at a lower geothermal gradient (e.g., Martin, 1999).

11. Conclusions

Geochemical characterization of intermediate to silicic rock sequences of 149 Phanerozoic to Archean ophiolites and *in-situ* oceanic crust at 7 sites has revealed that:

- Rift/Continental Margin and MOR ophiolites are composed mainly of basaltic andesite/andesite, and basaltic andesite/andesite and rhyodacite, respectively, whereas Plume/MOR ophiolites mainly contain rhyolite. Backarc ophiolites are characterized by similar amounts of basaltic andesite/andesite and rhyodacite, whereas the BA-FA, FA, and VA ophiolites are predominantly basaltic and andesitic in composition.
- The REE and multi-element patterns of the Rift/Continental Margin and Plume/MOR ophiolites show enrichment in the most incompatible elements (La, Ce, Nd), whereas the MOR ophiolites are depleted to enriched in these elements. The BA and BA-FA ophiolites are similar to the MOR type; the FA ophiolites are generally enriched, whereas VA ophiolites are generally enriched in incompatible elements.
- La-SiO₂ relationships suggest that the principal process behind the generation of the intermediate to silicic magmas in most of the subduction-unrelated and subduction-related ophiolites was partial melting of basaltic and/or gabbroic source rocks at the spreading centers. In P/MOR ophiolites the principal process was fractional crystallization of basaltic melt.
- Subduction-related ophiolites are weakly to strongly enriched in incompatible, non-conservative elements (i.e. Ba and Th) due to shallow- to deep-level enrichment associated with subduction zone processes.

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Figure and Table Captions

Fig. 1. Schematic plate tectonic diagram (not to scale), depicting the major settings of magmatic construction of the oceanic lithosphere (above) and the structural architecture of the main types of oceanic lithosphere, developed in these settings (below). Modified from Dilek and Furnes (2011, 2014). See text for further discussion.

Fig. 2. Columnar section showing the upper mantle and crustal components of a generalized subduction-related ophiolite (modified from Dilek and Furnes, 2014). Photo 1: Massive andesite flows, Mirdita ophiolite, Albania; Photo 2: Silicic intrusions crosscutting the plutonic gabbro and dike swarms, Karmøy ophiolite, Western Norway; Photo 3: Silicic dike crosscutting a coarse-grained, brecciated gabbro, Eldivan ophiolite, Turkey; Photo 4: Silicic dike crosscutting mafic and ultramafic cumulates, Mirdita ophiolite, Albania; Photo 5: Silicic intrusions (strongly deformed) in the upper mantle harzburgite, Xigaze ophiolite, Southern Tibet.

Fig. 3. A ternary An-Ab-Or diagram depicting the feldspar compositional variations among various silicic rock types in different ophiolites (Barker, 1979), including plagiogranites of crustal origin (e.g., Oman ophiolite: Rollinson, 2014; Nidar ophiolite: Rao et al., 2004; Mawat ophiolite: Mirza and Ismail, 2007; the Central Dinaride Ophiolite Belt (CDOB) and the Vardar Zone Ophiolite Belt (VZOB): Majer and Garašič, 2001) and those of mantle origin (e.g., Oman: Rollinson, 2014). Fields of experimental felsic melts produced by partial melting of gabbro (Koelpe et al., 2004) and Archean TTGs (after Rollinson, 2014) are also shown.

Fig. 4. Various tectonic discrimination diagrams used in this study for the first-order classification and discrimination of the tectonic settings of formation of ophiolitic crustal units (after Pearce, 2014). The Zr/Ti – Nb/Y diagram is originally after Floyd and Winchester (1975), the Th/Yb – Nb/Yb and TiO₂/Yb – Nb/Yb diagrams after Pearce (2008), and the V – Ti/1000 diagram after Shervais (1982).

Fig. 5. All intermediate and silicic rocks of the subduction-unrelated and subduction-related ophiolites plotted in a Zr/Ti – Nb/Y diagram.

- 1698 **Fig. 6.** Percentage intermediate (basaltic andesite, andesite, and trachyandesite) and silicic
 1699 (rhyolite and trachyte) rocks in subduction-unrelated (A), and subduction-related (B)
 1700 ophiolites.
- 1701 **Fig. 7.** Harker diagrams for subduction-unrelated ophiolites.
- 1702 **Fig. 8.** Harker diagrams for subduction-related ophiolites.
- 1703 **Fig. 9.** Chondrite-normalized REE diagrams and the distribution of $(La/Lu)_N$ ratios of
 1704 intermediate to silicic rocks of Rift/Continental margin ophiolites. Chondrite data from
 1705 McDonough and Sun (1995).
- 1706 **Fig. 10.** Chondrite-normalized REE diagrams and the distribution of $(La/Lu)_N$ ratios of
 1707 intermediate to silicic rocks of MOR ophiolites. Chondrite data as in Fig. 9.
- 1708 **Fig. 11.** Chondrite-normalized REE diagrams and the distribution of $(La/Lu)_N$ ratios of
 1709 intermediate to silicic rocks of Plume/MOR ophiolites. Chondrite data as in Fig. 9.
- 1710 **Fig. 12.** Chondrite-normalized REE diagrams and the distribution of $(La/Lu)_N$ ratios of
 1711 intermediate to silicic rocks of back-arc ophiolites. Black lines have been used to
 1712 emphasize those patterns that crosscut the majority of lines. Chondrite data as in Fig. 9.
- 1713 **Fig. 13.** Chondrite-normalized REE diagrams and the distribution of $(La/Lu)_N$ ratios of
 1714 intermediate to silicic rocks of backarc to fore-arc ophiolites. Black lines have been used
 1715 to emphasize those patterns that crosscut the majority of lines. Chondrite data as in Fig. 9.
- 1716 **Fig. 14.** Chondrite-normalized REE diagrams and the distribution of $(La/Lu)_N$ ratios of
 1717 intermediate to silicic rocks of fore-arc ophiolites. Black lines have been used to
 1718 emphasize those patterns that crosscut the majority of lines. Chondrite data as in Fig. 9.
- 1719 **Fig. 15.** Chondrite-normalized REE diagrams and the distribution of $(La/Lu)_N$ ratios of
 1720 intermediate to silicic rocks of volcanic-arc ophiolites. Chondrite data as in Fig. 9.
- 1721 **Fig. 16.** N-MORB-normalized multi-element diagrams of intermediate to silicic rocks of
 1722 Rift/Continental margin and MOR ophiolites. N-MORB values are from Pearce and
 1723 Parkinson (1993).

- 1724 **Fig. 17.** N-MORB-normalized multi-element diagrams of intermediate to silicic rocks of
 1725 Plume/MOR ophiolites. N-MORB values as in Fig. 16.
- 1726 **Fig. 18.** N-MORB-normalized multi-element diagrams of intermediate to silicic rocks of backarc
 1727 and backarc to fore-arc ophiolites. N-MORB values as in Fig. 16.
- 1728 **Fig. 19.** N-MORB-normalized multi-element diagrams of intermediate to silicic rocks of fore-arc
 1729 and volcanic-arc ophiolites. N-MORB values as in Fig. 16.
- 1730 **Fig. 20.** Histograms (in percentage) showing Nb-anomalies in subduction-unrelated and
 1731 subduction-related ophiolites. The anomalies have been calculated on the basis of MORB-
 1732 normalized values of the adjacent elements Th, Nb, La (Figs. 15-18), using the formula
 1733 $[(\text{Th}+\text{La})/2 - \text{Nb}]_N$.
- 1734 **Fig. 21.** Histograms (in percentage) showing Eu-anomalies in subduction-unrelated and
 1735 subduction-related ophiolites. The anomalies have been calculated on the basis of
 1736 chondrite-normalized values of adjacent elements Sm, Eu, Gd (Figs. 8-14), using the
 1737 formula $[(\text{Sm}+\text{Gd})/2 - \text{Eu}]_N$.
- 1738 **Fig. 22.** Histograms (in percentage) showing Ti-anomalies in subduction-unrelated and
 1739 subduction-related ophiolites. The anomalies have been calculated on the basis of MORB-
 1740 normalized values of adjacent elements Gd, Ti, Dy (Figs. 15-18), using the formula $[(\text{TiO}_2$
 1741 $- (\text{Gd}+\text{Dy})/2)]_N$.
- 1742 **Fig. 23.** Modeled variations in the enrichment factor (C_l/C_0) for La with an increasing SiO_2
 1743 content during (A) batch melting, and (B) fractional crystallization (after Brophy, 2009).
 1744 (C) La – SiO_2 diagram of mafic to felsic rocks of subduction-unrelated ophiolites. The
 1745 diagram includes models curves for partial melting and fractional crystallization (modified
 1746 from Brophy, 2009). See text for further details.
- 1747 **Fig. 24.** La – SiO_2 diagram of mafic to felsic rocks in subduction-related ophiolites. The
 1748 superimposed curves for partial melting and fractional crystallization are modified from
 1749 Brophy (2009). The black arrows indicate fractional crystallization of the produced melts
 1750 See text for further details.

Fig. 25. Intermediate to silicic rocks plotted in Sr/Y – Y diagrams (A, B, C) showing adakite and ADR (andesite, dacite, rhyolite) fields. Modified after Castillo (2006) and Naqvi et al. (2006). Diagrams (D, E, F) that separate high-SiO₂ and low-SiO₂ adakites are from Martin et al. (2005).

Fig. 26. Proxies for shallow (Ba/Th), deep (Th/Nb), and total (Ba/Th) subduction, plotted against Th. All values are MORB-normalized. For the last three diagrams showing “summary of averages”, “Felsic” refers to the intermediate and silicic rocks.

Fig. 27. Summary of different tectonic settings of the igneous construction of oceanic crust / ophiolites and the related processes. Data in panels A, B and C are from the following sources: A1: Saccani et al. (2015); A2: Zeng et al. (2015); A3: Xu et al. (2004); B (general): Haase et al. (2015); Palin and White (2016); for element transport, mantle hydration and melting during subduction: Tatsumi and Kogiso (1997); Grove et al. 2006; for the metamorphic transition in the subducted slab: Tatsumi and Kogiso (1997), Schmidt and Poli (1998), Angiboust et al. (2014), Bebout and Penniston-Dorland (2016); C1: Hildreth (1981), Watanabe et al. (2006); C2: Grimes et al. (2013), France et al. (2010), C3: Martin (1999). Abbreviations: BABB – Back-arc basin basalt; CC - Continental crust; LS - Lithosphere; OLS - Oceanic lithosphere; AS - Asthenosphere; OC - Oceanic crust; L - Lava; Dk - Dike; Gb - Gabbro; Isot – Isotropic gabbro; DM – Depleted mantle; EM – Enriched mantle; Cum - Cumulate; Um – Ultramafic rocks; Gr.sch - Greenschist; Amph - Amphibolite; Ecl - Eclogite; Intm – Intermediate; EM - enriched mantle; DM - depleted mantle; MDM - multiply depleted mantle; BABB - backarc basin basalt.

Table 1. Summary of ophiolite complexes. Full references are provided in Supplementary File-1.

Abbreviations: Column A: East Pacific Rise: EPR; East Scotia Ridge: ESR; Izu-Bonin-Mariana: IBM; Kyushi-Palau Ridge: K-PR; Manus Spreading Centre: MSC; Melange: Mel. **Column B:** California: Cal. **Column C:** Cambrian: Camb., Ordovician: Ord., Permian: Perm., Triassic: Trias., Jurassic: Jur., Cretaceous: Cret. **Column D:** Albite granite: Alb.gr.; Andesite: And.; Basaltic andesite: Bas. and.; Dacite: Dac.; Dike: Dk.; Diorite: Di.; Diorite porphyry: Di.porph.; Felsic: Fel.; Granodiorite: Gr.di.; Granite: Gr.; Granophyre: Gr.phy.; Hornblende-plagioclase

1780 porphyry: Hbl-pl.porph.; Icelandite: Isl.di.; Intermediate: interm.; Keratophyre: Ker.;
 1781 Leucodiorite: Leucodi.; Leucogranite: Leucogr.; Monzogranite: Mon.gr.; Plagiogranite: Pl.gr.;
 1782 Quartz diorite: Quartz monzodiorite: Q-mon.di.; Q-di.; Quartz porphyry: Q-porph.; Rhyodacite:
 1783 Rhy.dac.; Rhyolite: Rhy.; Tonalite: Ton.; Trachydacite: Try.dac.; Trondhjemite: Tr.hj.ite.;
 1784 Volcanics: Volc. **Column F:** Shafaii Moghadam: Sh.Mogh.; Van Kranendonk & Pirajno: Van Kr.
 1785 & Pir. **Column G:** Lithosphere: Li.sph.; Mantle plume: MP; Marginal basin: Marg.bas.; Oceanic:
 1786 Oc.; Seafloor spreading: Sfs.; Subducted: Subd.; Suprasubduction zone: SSZ. **Column H:**
 1787 Rift/Continental Margin: R/CM; Mid-ocean-ridge: MOR~~B~~; Plume/MOR: P/MOR; Backarc: BA;
 1788 Backarc to Forearc: BA-FA; Forearc: FA; Volcanic Arc: VA. In italics: ophiolite type suggested
 1789 from literature.

1790

1791 **Table 2.** Summary of geochemical proxies for the classification of ophiolite types.

ABSTRACT

We present a global synthesis of the geochemistry of intermediate to silicic intrusive and volcanic rocks from 149 Phanerozoic to Archean ophiolites, and evaluate models for their genesis during the development of oceanic crust in different tectonic environments. Our synthesis shows that the evolved rocks in subduction-unrelated, *Rift/Continental Margin* ophiolites are predominantly basaltic andesite and andesite, whereas *MOR type* (mid-ocean ridge) ophiolites exhibit nearly equal proportions of basaltic andesite/andesite and rhyodacite and *Plume/MOR type* ophiolites are characterized by rhyolites. Intermediate to silicic volcanic units in the *Backarc* sub-group of subduction-related ophiolites are characterized by similar amounts of basaltic andesite/andesite and rhyodacite, whereas in the *Backarc to Forearc*, *Forearc*, and *Volcanic Arc* sub-groups they are mainly basaltic andesite/andesite. The Zr, Nb and La contents of the intermediate to silicic rocks in subduction-related ophiolites are significantly lower than those of their counterparts in subduction-unrelated ophiolites. Intermediate to silicic rocks in Rift/Continental Margin and Plume/MOR type ophiolites are generally LREE-enriched, whereas those in the MOR type vary from LREE-depleted to LREE-enriched. The Backarc and Backarc to Forearc types are similar to the MOR type; silicic rocks of the Forearc and Volcanic Arc types are generally LREE-enriched. Nb, Sr, Eu, and Ti exhibit negative anomalies of varying magnitudes in multi-element diagrams. Our modeling based on the La-SiO₂ proxies of Brophy (2009) suggests that the main process in the formation of the majority of the intermediate to silicic rocks in both subduction-unrelated and subduction-related ophiolites is partial melting of basaltic and/or gabbroic rocks beneath the spreading centers, whereas a minor volume in subduction-related ophiolites are adakites that resulted from partial melting of the subducting slab. Silicic to intermediate rocks in Plume/MOR type ophiolites appear to be generated by fractional crystallization of basaltic melt. The incompatible, non-conservative elements, such as Ba and Th, are weakly to strongly enriched in subduction-related ophiolites as a result of shallow to deep enrichment associated with subduction zone processes.

Table 1

Table 1. Summary of ophiolitic sequences, their felsic rock types and suggested tectonic enviromment and ophiolite types

A. Complex	B. Phanerozoic Orogenic belt/ Precambrian regions/ In-situ oceanic/ Location	C. Age (Ma)	D. Intermediate and felsic rocks	E. Main references to the intermediate and felsic rock data	F. Main references to the basalt and ultramafic data and general geology	G. Suggested tectonic setting (from literature)	H. Proposed ophiolite type (this paper)
Phanerozoic Orogens							
1. Indonesian- Myanmar							
1. Andaman	Andaman Sea, India	95	Tr.hj.ite., pl.gr.	Jafri et al., 1995; Pedersen et al., 2010	Pedersen et al., 2010	Mid-ocean ridge	MOR
2. Myitkyina	Myanmar	176	Pl.gr.	Yang et al., 2012	Yang et al., 2012	Subduction-related	BA-FA
2. Maghrebien- Alpine- Himalayan							
3. Platta	E Switzerland	Jurassic	Di.	Desmurs et al., 2002	Desmurs et al., 2002	Mid-ocean ridge	MOR
4. Ligurian Alps and Montgenèvre	France	Ca.200	Leucodi.	Costa & Caby, 2001	Lewis & Smewing, 1980; Costa & Caby, 2001	Oceanic domain	MOR
5. Northern Apennines	Italy	156-151	Pl.gr.	Borsi et al., 1996	Montanini et al., 2008	Mid-ocean ridge	MOR
6. Corsica ophiolites	Italy	Mid Cret.	Pl.gr.	Beccaluva et al., 1997	Saccani et al., 2008	Rift stage to ocean opening	MOR
7. "Sjenica"	Serbia	Middle Jur.	Pl.gr.	Milovanovic et al., 2012	Segvic et al., 2014; Ustaszewski et al., 2009	Oceanic domain	MOR
8. North Kozara	Bosnia-Herzegovina	Upper Cret.	Q-di., gr., rhy-dac.	Cvetkovic et al., 2014	Cvetkovic et al., 2014	Mid-ocean ridge	BA
9. Mandrica	Bulgaria	Jurassic	Bas. and., and.	Bonev & Stampfli, 2008	Bonev & Stampfli, 2008	Island arc	VA
10. Sebes Mtn. S. Carpathians	Romania	?	Tr.hj.ite	Dobrescu & Stoian, 1994	?	?	VA
11. Mirdita	Albania	150	Tr.hj.ite, and., dac., rhy.	Dilek et al., 2009; Dilek & Furnes (unpubl. data)	Dilek et al. (2008)	SSZ	BA-FA
12. Evros	Greece	M. Jur.	Tr.hj.ite	Bonev & Stampfli, 2009	Bonev & Stampfli, 2009	SSZ	FA
13. Vourinos	Greece	M. Jur.	And., dac., rhy.	Beccaluva et al., 1984	Beccaluva et al., 1984; Kapsiotis, 2013	Forearc	FA
14. Koziakas	Greece	Jurassic	Pl.gr.	Pomonis et al., 2012	Pomonis et al., 2007; Chiari et al., 2012	SSZ; backarc basin	MOR
15. Othris	Greece	Triassic	And. to rhy.	Koutsovitis et al., 2012	Koutsovitis et al., 2012	Front-arc setting	BA-FA
16. Troodos	Cyprus	92	Pl.gr., tr.hj.ite, ton.	Pantazis, 1980; Gillis & Coogan, 2002; Freund et al., 2014	Rautenschlein et al., 1985	Suprasubduction zone	BA
17. Oman	Oman	96	Ton., tr.hj.ite, pl.gr., gr.di., leucogr., mon.gr.	Rollinson, 2009, 2014; Cox et al., 1999; Stakes & Taylor, 2003	Goodenough et al., 2010	Suprasubduction zone	BA-FA
18. Eldivan	Turkey	179	Pl.gr.	Üner et al., 2014	Çelik et al. 2013	Above subducted oc. li.sph.	MOR
19. Ankara Melange	Turkey	179	Pl.gr.	Dilek & Thy, 2006	Dilek & Thy, 2006	Backarc	BA

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20. Sarikaraman	Turkey	ca. 92	Pl.gr., tr.hj.ite, rhy.	Floyd et al., 1998; Yaliniz et al., 2000	Yaliniz et al. 1996	SSZ	BA
21. Kizildag	Turkey	ca. 92	Pl.gr.	Karaoğlu et al., 2013	Dilek & Thy, 2009	SSZ	BA
22. Guleman	Turkey	ca. 92	Pl.gr.	Kiliç, 2009	Kiliç, 2009, 2012	SSZ	BA-FA
23. Sabsevar	Iran	100-78	Pl.gr.	Sh. Mogh. et al., 2014	Sh. Mogh. et al., 2014	Mid-ocean ridge	BA
24. Neyriz	Iran	96-82	Pl.gr., rhy.dac.	Babaie et al., 2006	Babaie et al., 2006	SSZ	BA
25. Band-e-Zeyarat/ Dar Anar	Iran	142-141	Tr.hj.ite	Ghazi et al., 2004	Ghazi et al., 2004	Mid-ocean ridge	BA
26. Haji-Abad	Iran	Late Cret.	And., dac., rhy., ton., pl.gr.	Sh. Mogh. et al., 2013a	Sh. Mogh. et al., 2013a	Forearc	BA-FA
27. Gogher-Baft	Iran	Late Cret.	Felsic volcanics	Sh. Mogh. et al., 2013b	Sh. Mogh. et al., 2013b	Subduction initiation setting	BA-FA
28. Dehshir	Iran	Late Cret.	Dac., gr.	Sh. Mogh. et al., 2010	Sh. Mogh. et al., 2010	SSZ	BA
29. Mawat	NE Iraq	ca. 100	Pl.gr.	Mirza & Ismail, 2007	Mirza, 2008; Azizi et al., 2013	Oceanic crust	MOR
30. Mingora ophiolitic melange	Pakistan	Mid. Jur - Early Cret.	Pl.gr.	Barbieri et al., 1994	Barbieri et al., 1994	?	MOR
31. Waziristan	NW Pakistan	Late Jur.- Early Cret.	Tr.hj.ite	Khan et al., 2003	Khan et al., 2007	Island arc/mid-ocean ridge	BA
32. Muslim Bagh	NW Pakistan	Trias. - Cret.	Pl.gr.	Kakar et al., 2014	Kakar et al., 2014	Island arc/mid-ocean ridge	BA
Yarlung-Zangbo							
33. Dras	India	ca. 135	And.	Clift et al., 2002	Clift et al., 2002	Intraoceanic arc	VA
34. Nidar	India	130-110	Pl.gr.	Rao et al., 2004	Mahéo et al., 2004	Intraoceanic arc	BA
35. Spontang	India	130-110	Di., and.	Pedersen et al., 2001	Mahéo et al., 2004	Intraoceanic arc	BA
36. Sangsang	Tibet	155-130	Felsic lava	Bedard et al., 2009	Bedard et al., 2009	Backarc-arc	BA
37. Jinlu	Tibet	125	Bas. and.	Dubois-Cote et al., 2005	Dubois-Cote et al., 2005	Intraoceanic arc	BA-FA
38. Xigaze	Tibet	130-120	Q-di.	Dai et al., 2013	Dai et al., 2013	Suprasubduction zone	BA
3. Caribbean Region							
39. Mayari-Baracoa	Eastern Cuba	Cret.	And., dac.	Marchesi et al., 2007	Frei et al., 2006; Marchesi et al., 2007	Backarc spreading to front arc	BA-FA
40. Aruba Lava Fm	Dutch Caribbean	85-82	Ton., hbl-pl. porph., gr.	White et al., 1999	White et al., 1999	MORB	MOR
41. Nicoya	Costa Rica, Caribbean	95-86	Pl.gr.	Wildberg, 1987	Hauff et al., 2000	Oceanic plateau	MOR
42. Central Hispaniola	Caribbean	115	Dac., rhy.	Escuder Viruete et al., 2008	Escuder Viruete et al., 2008	Subduction initiation	BA
43. Puerto Plata	Dominican Republic	126	Bas. and., and., dac.	Escuder Viruete et al., 2014	Escuder Viruete et al., 2014	Island arc/backarc system	BA
4. Cordilleran and Western Pacific							
44. Elder Creek	California, USA	Mid-Jur.	Tr.hj.ite, ton., di., q-di.	Shervais, 2008	Shervais, 2008	SSZ	BA
45. Ingalls	California, USA	Mid-Late Jur.	Tr.hj.ite, ton.	MacDonald et al., 2008	MacDonald et al., 2008	SSZ	BA
46. Smartville	California, USA	Late Jur.	Ton., gr.di.	Beard, 1998	Beard, 1998	Arc magmatism	BA-FA
47. Josephine	Cal.-Oregon, USA	165	Pl.gr.	Harper, 2003	Harper, 2003	SSZ	BA
48. Canyon Mtn.	Oregon, USA	Early Perm.	Di., ton., tr.hj.ite., alb.gr.	Gerlach et al., 1981a, b	Gerlach et al., 1981a, b	Arc magmatism	VA
49. Fidalgo Island	Washington State, USA	Jurassic	Tr.hj.ite., ker., alb-gr.	Brown et al., 1979	Brown et al., 1979	Ocean floor	MOR
50. Ganychalan terrane	NE Russia	Paleozoic	Ton., pl.gr.	Sokolov et al., 2003	Sokolov et al., 2003	SSZ	BA-FA

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51. Main melange zone	NE Russia	Early Mes.	Pl.gr.	Sokolov et al., 2003	Sokolov et al. 2003	Arc system	VA
52. Pekulney Ridge island arc	NE Russia	L.Jur-E. Cret.	Pl.gr., ton.	Luchitskaya et al., 2005	Luchitskaya et al. 2005	SSZ	FA
53. E. Kamchatka oph.	Russia	Upper Cret.	Pl.gr.	Tsukanov et al., 2007	Tsukanov et al., 2007	Volcanic arc	BA-FA
54. Yakuno	Japan	280	Ton., q-di., gr.di., and., rhy., gr., q-mon.di.	Ishiwatari et al., 1990; Suda, 2004	Ichiyama and Ishiwatari, 2004	Backarc basin	VA
55. Horokanai	Japan	Late Jur.	And.	Takashima et al., 2002	Takashima et al., 2002	Marginal sea above SSZ	BA
56. Mineoka	Japan	66	Di.	Hirano et al., 2003	Hirano et al., 2003	Subduction and within-plate	BA
57. Darvel Bay	Malaysia	ca. 140	Pl.gr.	Omang & Barber, 1996	Omang & Barber, 1996	Suprasubduction zone	BA
58. Cyclops	New Guinea Isl.	43.2	And., dac.	Monnier et al., 1999	Monnier et al., 1999	Forearc to backarc	BA-FA
59. New Caledonia	New Caledonia	Late Cret. - Early Eocene	Felsic volc., pl.gr., gr., and., dac., leucodi.	Cameron, 1989; Meffre et al., 1996; Cluzel et al., 2006	Ulrich et al., 2010	Backarc basin	BA
60. Dun Mtn.	New Zealand	280	Pl.gr.	Sano et al., 1997	Sano et al., 1997	Backarc basin	BA
61. Northland	New Zealand	ca. 30	Pl.gr., and.	Thompson et al., 1997; Whattam et al., 2005, 2006	Thompson et al., 1997; Whattam et al., 2006	Backarc basin	BA
5. Hercynian/Variscan							
62. Kaczawa	Poland	Lw.Pal.	Trachyte, rhy.dac.	Furnes et al., 1994	Furnes et al., 1994	Rift type	Rift/CM
63. Beja-Acebuches	SE Portugal	350	Leucodi., felsic rocks	Pin et al., 2008	Pin et al., 2008	Slab break-off that followed subduction	MOR
64. Morais (group II)	North Portugal	405	Felsic rocks	Pin et al., 2006	Pin et al., 2006	Introceanic subduction	BA
65. Chamrousse	France	496	Pl.gr., ton., ker.	Ménot et al., 1988; Ménot, 1987	Bodinier et al., 1981; Ménot, 1987	Mid-ocean ridge	MOR
6. Uralian							
66. Magnitogorsk	Russia	400-385	And., dac.	Spadea & Scarrow, 2000	Spadea & Scarrow, 2000	Forearc region	BA-FA
67. Nurali	Russia	468-483	And., dac., gr.di., ton., gr.	Gaggero et al., 1997	Gaggero et al., 1997	Intraoceanic arc/continental margin	VA
7. Central Asian							
68. Dananhu island arc	NW China	360-320	Pl.gr., and., q-porph.	Zhang et al., 2006	Zhang et al., 2006	Island arc	VA
69. Kekesentao-Qiaoxiahala-Kuerti	NW China	372	Pl.gr.	Zhang et al., 2003	Wang et al., 2012a, b	Mid-ocean ridge/SSZ	BA
70. Tangbale	NW China	Middle Ord.	Felsic volc.	Buckman & Aitchison, 2004	Chi et al., 1993; Buckman & Aitchison, 2004	Intra-oceanic setting	BA
71. Zhaheba	NW China	495-422	Pl. gr., di.porph.	Luo et al., 2015; Zeng et al., 2015	Luo et al., 2015	Oceanic island arc	BA-FA
72. Chagantaolegai	NW China	518	And.	Zhao & He, 2014	Zhao & He, 2014	Mid-ocean ridge	MOR
73. Yueyashan-Xichangjing (Beishan Collage)	NW China	533	Pl.gr., ton., and., dac.	Ao et al., 2012	Ao et al., 2012	Subduction-related	BA
74. Solonker	China and Mongolia	Permian	Ton., and., dac., rhy.	Jian et al., 2010b	Jian et al., 2010b	Arc-trench system	BA-FA
75. Manlay	South Mongolia	509-482	And., dac.	Zhu et al., 2014	Zhu et al., 2014	SSZ to island arc	BA-FA
76. Agardagh Tes-Chem	Mongolia	569	And., dac.	Pfänder et al., 2002	Pfänder et al., 2002	Intra-oceanic island arc system	BA

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77. Bayannur	Western Mongolia	572	Ton., tr.hj.ite	Kovalenko et al., 2014	Dijkstra et al., 2006; Kovalenko et al., 2014	and associated backarc basin SSZ	BA-FA
78. Bayankhongor	Central Mongolia	636	Pl.gr.	Jian et al., 2010a	Jian et al., 2010a	Mid-ocean ridge	MOR
8. Qingling-Qilian-Kunlun							
Jinshajiang oph. Mel.							
79. Jiyidu	SW China	283	Ton.	Zi et al., 2012	Jian et al., 2009	Subduction setting	BA
80. Dongzhulin	SW China	347	Tr.hj.ite	Zi et al., 2012	Jian et al., 2009	Rift setting	Rift/CM
81. Oyttag	NW China	338-328	Tr.hj.ite., ton.	Jiang et al., 2008	Jiang et al., 2008	Island arc	BA
82. Shangdan	East China	517	Dac.	Dong et al., 2011	Dong et al., 2011	Backarc basin	BA-FA
9. Bangong-Nujian Suture							
83. Changning-Menglian (Gicha)	SW China	311-277	Ton.	Jian et al., 2009	Jian et al., 2009	Mid-ocean ridge	BA
84. Changning-Menglian(Banpo)	SW China	288-284	Ton.	Jian et al., 2009	Jian et al., 2009	Mid-ocean ridge	BA
85. Qiantang	North Tibet	356	Pl.gr.	Zhai et al., 2013	Zhai et al., 2013	Mid-ocean ridge	MOR
86. Ailaochan	SW China	387-374	Pl.gr.	Jian et al., 2009	Jian et al., 2009	Mid-ocean ridge	MOR
87. W Gagma Co	Tibet	501-437	Pl.gr.	Zhai et al., 2016	Zhai et al., 2016	SSZ	MOR
10. Andean							
88. Sarmiento	Chile	150	Tr.hj.ite, gr.phy., pl.gr.	Saunders et al., 1979	Saunders et al., 1979	Marginal basin	BA
89. Famatinian magmatic arc	Argentina	Late Camb. - Middle Ord.	Leucogr., leucoton, ton.	Otamendi et al., 2009	Otamendi et al., 2009	Magmatic arc	VA
11. Appalachian							
90. Tyrone Ign. Complex	Ireland	490-475	Ton., pl.gr., and., dac., rhy.	Draut et al., 2009; Hollis et al., 2013;	Draut et al., 2009; Hollis et al., 2013	SSZ	BA-FA
91. Fournier	Quebec, Canada	461	Pl.gr.	Spray et al., 1990; Brophy & Pu, 2012; Flagler & Spray, 1991	Brophy & Pu, 2012	Mid-ocean ridge	BA
92. Thetford Mines	Quebec, Canada	470	Pl.gr.	Olive et al., 1997; Whitehead et al., 2000	Olive et al., 1997; Page et al., 2009	Spreading in forearc	BA-FA
93. Boil Mountain	Maine, USA	477	Ton.	Van Vleck & Beane, 2001	Kusky et al., 1997; Van Vleck & Beane, 2001	backarc	BA-FA
94. Annieopsquotch	Newfoundland	480	Tr.hj.ite, fel. dk, tuff	Cutts et al., 2012	Lissenberg et al., 2005	SSZ - backarc	BA
95. Lac Brompton	Quebec, Canada	480	Q-di., pl.gr.	De Souza et al., 2008	De Souza et al., 2008	Forearc seafloor	FA
96. Bay of Islands	Newfoundland	485	Tr.hj.ite.	Malpas, 1979; Jenner et al., 1991	Kurth-Velz et al., 2004	Marginal basin - arc	BA-FA
97. Betts Cove	Newfoundland	488	Dac., rhy.	Bédard et al., 2000	Bédard, 1999	SSZ	FA
98. Little Port	Newfoundland	505	Tr.hj.ite.	Malpas, 1979; Jenner et al., 1991	Malpas, 1979; Jenner et al., 1991	Backarc-related	BA

Table 1. Summary of ophiolitic sequences, their felsic rock types and suggested tectonic environment and ophiolite types

12. Caledonides							
99. Lyngen	North Norway	c. 480	Ton.	Selbekk et al., 1998	Kvassnes et al. 2004; H.Furnes (unpubl. data)	Incipient oceanic arc	VA
100. Leka	Central Norway	497	Pl.gr.	Furnes et al., 1992	Furnes et al., 1992	Introceanic subduction	BA-FA
101. Trondheim area	Central Norway	493-481	Pl.gr., tr.hj.ite	Grenne, 1989; Slagstad, 2003	Grenne, 1989; Slagstad, 2003	Island arc/backarc	BA
102. Karmøy	Western Norway	497-470	Pl.gr.	Pedersen & Malpas, 1984	Pedersen & Hertogen, 1990	Island arc/arc basin	BA-FA
103. Solund-Stavfjord	Western Norway	443	Pl.gr.	Skjerlie, 1992; Furnes (unpubl. data)	Furnes et al., 1990, 2006, 2012b	Backarc basin	BA
13. Tasmanide							
104. Weraera terrane	SE Australia	530	Pl.gr.	Aitchison & Ireland, 1995	Aitchison & Ireland, 1995	SSZ	BA
Precambrian regions							
14. Cadomian							
105. Frolosh/Struma	Bulgaria	560	Interm. to fel. rocks	Kounov et al., 2012	Kounov et al., 2012	Magmatic arc	BA
15. Arabian-Nubian Shield							
106. Janbu	Saudi Arabia	789	And.	Ahmed & Hariri, 2008	Ahmed & Hariri, 2008	Small ocean basin	VA
107. Darb Zubaydah	Saudi Arabia	830	And., rhy.	Quick, 1990	Quick, 1990	Intra-arc rifting	BA
108. Wadi Ghadir	Egypt	750	Interm. volc. rocks	Abd El-Rahman et al., 2009a	Abd El-Rahman et al., 2009a; Basta et al., 2011	backarc basin	BA
109. Wadi Kareim	Egypt	750	Felsic volc. rocks	Ali et al., 2009	Ali et al., 2009	backarc basin	BA
110. Wadi El Dabbah	Egypt	750	Felsic volc. rocks	Ali et al., 2009	Ali et al., 2009	Volcanic arc	VA
111. Fawakhir	Egypt	800-700	Interm. volc. rocks	Abd El-Rahman et al., 2009b	Abd El-Rahman et al., 2009b	Intra-oceanic SSZ, (incipient arc-forearc)	BA
16. India							
112. Manamendu	India	800	Pl.gr.	Yellappa et al., 2010	Yellappa et al., 2010	SSZ	FA
113. Kolar	India (Dharwar Craton)	2550	Fel. Volc.	Chardon et al., 2002; Ugarkar et al., 2013	Rajamani et al., 1985	Island arc	VA
114. Hutti	India	2600	Mg-And., adakite	Manikyamba et al., 2009	Manikyamba et al., 2009	Intraoceanic subduction	BA-FA
115. Kushtagi-Hungund	Southern India	2746	Adakite	Naqvi et al., 2006	Naqvi et al., 2006	Plume-fed oceanic slab, subd. in intraoceanic setting	BA-FA
116. Southern Iron Ore Group	Northeast India	3510	Dac. Volc.	Mukhopadhyay et al., 2012	Mukhopadhyay et al., 2012	Suprasubduction zone	BA-FA
17. Baltica and Siberia							
117. Enganepe	Polar Urals, Russia	670	Adakite	Scarrow et al., 2001	Scarrow et al., 2001	Oceanic island arc	BA-FA
118. Jormua	Baltica, Finland	1950	Tr.hj.ite	Kontinen, 1987	Peltonen et al., 1996	Red Sea type	Rift/CM
119. Pilgujärvi	Pechenga, Kola, Russia	1970	Rhy.	Skruf'in & Theart, 2005	Skruf'in & Theart, 2005	Red Sea type	Rift/CM
120. Kholodnikan	Siberia, Russia	2410	And., dac., rhy.	Lavrik & Miskin, 2010	Lavrik & Miskin, 2010	Mantle plume	P/MOR
121. Avarench	Kola, Russia	2429	And., dac., rhy.	Vrevsky, 2011	Vrevsky, 2011	Intracratonic rifting	Rift/CM
122. Suomussalmi	Baltica, Finland	2650	Interm. - fel. volc.	Jahn et al., 1980	Jahn et al., 1980	No suggestion	BA

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123. Tikshozero	Karelia, Russia	2785	Interm. - fel. volc.	Mil'kevich et al., 2007	Mil'kevich et al., 2007	Backarc and initial back arc	BA-FA
124. Khizovaara - Iringora	Karelia, Russia	2800	Interm. - fel. volc.	Shchipansky et al., 2004	Shchipansky et al., 2004	SSZ	BA-FA
18. Australia							
125. Gindalbie & Kurnalpi	SW Australia	2700	And., dac., rhy.	Barley et al., 2008	Barley et al., 2008	Intra-arc to mature arc-rift	VA
126. Meekatharra-Cue	SW Australia	2800-2760	And., fel. vol.	Wyman & Kerrich, 2012	Wyman & Kerrich, 2012	Subduction-related magm. from depleted plume	BA-FA
127. Whundo Group	NW Australia	3120	And., dac., rhy.	Smithies et al., 2005	Smithies et al., 2005	Intracoeanic arc setting	BA-FA
Pilbara Supergroup							
128. Kelly Group	NW Australia	3350-3290	Rhy.	Jahn et al., 1981	Van Krd. & Pir., 2004	Plume matatism	FA
129. Warrawoona Group	NW Australia	3530-3430	And., dac., rhy.	Smithies et al., 2007	Van Krd. & Pir., 2004; Smithies et al., 2007	Plume magmatism	BA-FA
19. China							
130. Anhui & Jiangxi	South China	970	Pl.gr.	Li & Li, 2003; Gao et al., 2009	Zhou, 1989; Li et al., 1997	Anhui: Oc. crust in marg. bas. Jiangxi: Oc. crust in interarc bas.	BA
131. Miowan	South China (Yangtze Craton)	1000	Tr.hj.ite	Peng et al., 2012	Peng et al., 2012	Arc/forearc	BA
20. North America/Canada							
132. Burin Group	Newfoundland	760	Fel. rocks	Murphy et al., 2008	Murphy et al., 2008	Oceanic setting, ensim. arc	VA
133. Payson	Central Arizona	1730	Dac., ton.	Dann, 1992	Dann, 1992	Intra-arc basin, sfs	BA
134. Elu Link	North Canada	2716-2663	Fel. volc.	Mvondo et al., 2013	Mvondo et al., 2013	Backarc	BA
135. Abitibi	SE Canada	2735-2670	Fel. volc.	Gaboury & Pearson, 2008	Xie et al., 1993; Wyman, 1999; Kerrich et al., 1998; Sprole et al., 2002; Mueller et al., 2009; Dostal & Mueller, 2013	Oceanic arc and plume magmatism	BA
136. Yellowknife	NW Canada	2720-2670	Fel. Tuff	Cousens, 2000	Cousens, 2000	Backarc	VA
137. Wawa	Canada	2750-2650	Fel. volc. & dk.	Polat & Kerrich, 2000	Polat & Kerrich, 2000	Island arc	VA
21. South America							
138. Mazaruni & Barama	North Guyana, Brazil	2250	Fel. rocks	Renner & Gibbs, 1987	Renner & Gibbs, 1987	Incipient arc or marg. bas.	BA
139. Rio das Velhas	SE Brazil	2772	Dac.	Da Silva et al., 2000	Zucchetti et al., 2000; Baltazar & Zucchetti, 2007; Noce et al., 2007	Submarine plateau (plume); and island arc/backarc basin	VA
22. Greenland							
140. Isua	SW Greenland	3800	Pl.gr., fel. rocks	Polat & Hofmann, 2003; Komiya et al., 2004;	Polat & Hofmann, 2003; Komiya et al., 2004;	Plume to SSZ	BA-FA

Table 1. Summary of ophiolitic sequences, their felsic rock types and suggested tectonic environment and ophiolite types

				Furnes et al., 2007	Furnes et al., 2007		
23. Africa							
141. Leerkrans Fm	South Africa	1300	Pyroclast	Bailie et al., 2011	Bailie et al., 2011	Continental backarc basin	BA
142. Bas Draa	Morocco	1380	Rhy.	El Bahat et al., 2013	El Bahat et al., 2013	Rift magmatism	R/CM
143. Chewore	Kalahari - Congo, Africa	1400	And., dac., rhy.	Johnson & Oliver, 2000	Johnson & Oliver, 2000	Marginal basin	BA
144. Birimian terrane	Western Africa	2100	And., dac., rhy.	Vidal & Alric, 1994	Abouchami & Boher, 1990; Sylvester & Attoh, 1992; Vidal & Alric, 1994	Backarc immature arc on oceanic crust	BA
145. Belingwe	Zimbabwe	2900-2700	And.	Bolhar et al., 2003	Kusky & Kidd, 1992; Bolhar et al., 2003; Hofmann & Kusky, 2004	1. Intra-oceanic MP; 2. Ensialic rift above MP 3. Backarc oceanic crust	BA-FA
146. Noisy Complex, Barberton	South Africa	3455	Ton.; rhy.dac.	de Wit et al., 1987; Louzada, 2003; Furnes et al., 2013		Island arc	VA
24. Ophiolitic sequences unrelated to orogenic belts							
147. Taitao	Chile	6	Dac., try.dac., rhy.	Guivel et al., 1999	Le Moigne et al., 1996; Guivel et al., 1999; Veloso et al., 2005, 2009	Mid-ocean ridge	MOR
148. Tihama Asir	Red Sea, Saudi Arabia	23-19	Rhy., gr.phy.	Dilek & Furnes (unpubl. data)	Dilek & Furnes (unpubl.data)	Rift stage, incipient oc. crust	P/MOR
149. Iceland	Mid-Atlantic Ridge	<1	Ice.di., rhy. (mainly); gr., gr.di.	MacDonald et al., 1990; Furman et al., 1992; Hemond et al., 1993; Gunnarson et al., 1998; Martin & Sigmarsson, 2007	Hemond et al., 1993	Plume/mid-ocean ridge	P/MOR
25. In-situ oceanic crust							
150. EPR	Pacific Ocean	< 0.03	Dac.	Wanless et al., 2010	Pandey et al., 2013	Mid-ocean ridge	MOR
151. K-PR	Philippine Sea	38	Ton.	Haragushi et al., 2003	Haragushi et al., 2012	Early stage in oceanic island arc formation	BA-FA
152. MSC	Manus Basin	<1	Dac., rhy.	Beier et al., 2015	Beier et al., 2015	Backarc basin	BA
153. ESR	E Scotia Sea	<1	Dac.	Leat et al., 2000	Leat et al., 2000	Backarc basin	BA
154. Kermadec arc	Lau Basin	<1	Dac., rhy.	Smith et al., 2003a,b	Ewart et al., 1977	Oceanic arc	BA-FA
155. Spreading axes	Lau Basin	<1	And.	Pearce et al., 1994	Pearce et al., 1994	Backarc basin	BA
156. IBM	Bonin Isl., Japan	46-48	Fel. rocks	Ishizuka et al., 2006; Reagan et al., 2010	Ishizuka et al., 2006; Reagan et al., 2010	Forearc	FA

Abbreviations: **Column A:** East Pacific Rise: EPR; East Scotia Ridge: ESR; Izu-Bonin-Mariana: IBM; Kyushi-Palau Ridge: K-PR; Manus Spreading Centre: MSC; Melange: Mel. **Column B:** California: Cal. **Column C:** Mesozoic: Mes., Cambrian: Camb., Ordovician: Ord., Permian: Perm., Triassic: Trias., Jurassic: Jur., Cretaceous: Cret. **Column D:** Albite granite: Alb.gr.; Andesite: And.; Basaltic andesite: Bas. and.; Dacite: Dac.; Dike: Dk.; Diorite: Di.; Diorite porphyry: Di.porph.; Felsic: Fel.; Granodiorite: Gr.di.; Granite: Gr.; Granophyre: Gr.phy.; Hornblende-plagioclase porphyry: Hbl-pl.porph.; Icelandite: Isl.di.; Intermediate: interm.; Keratophyre: Ker.; Leucodiorite: Leucodi.; Leucogranite: Leucogr.; Monzogranite: Mon.gr.; Plagiogranite: Pl.gr.; Quartz diorite: Quartz monzodiorite: Q-mon.di.; Q-di.; Quartz porphyry: Q-porph.; Rhyodacite: Rhy.dac.; Rhyolite: Rhy.; Tonalite: Ton.; Trachydacite: Try.dac.; Trondhjemite: Tr.hj.ite.; Volcanics: Volc. **Column F:** Shafaii Moghadam: Sh.Mogh.; Van Kranendonk & Pirajno: Van Kr. & Pir. **Column G:** Ensimatic: Ensim., Lithosphere: Li.sph.; Magmatism: Magm., Mantle plume: MP; Marginal basin: Marg.bas.; Oceanic: Oc.; Seafloor spreading: Sfs.; Subducted: Subd.; Suprasubduction zone: SSZ. **Column H:** Rift/Continental Margin: Rift/CM; Mid-ocean-ridge: MOR; Plume/MOR: P/MOR; Backarc: BA; Backarc to Forearc: BA-FA; Forearc: FA; Volcanic Arc: VA. In italics: ophiolite type suggested from literature.

Table 2

Table 2. Proxies for subduction and melting for Phanerozoic, Proterozoic and Archean ophiolites

Ophiolite Type	Percent subduction influence	Th/Yb-proxy for subduction (%)					V - Ti proxy for SSZ melting (%)			Ti/Yb proxy for plume melting (%)			
		Mantle array	oceanic		Cont. arc (ca)	Joint oa/ca	for SSZ melting (%)			MORB - shallow		MORB - deep	
			weak	arc (oa)			Bon	IAT	MORB	N-MORB	E-MORB	Thol.	Alk.
Phanerozoic Complexes													
Subduction-unrelated													
Rift/Continental margin (1)	0	100	0	0	0	0	0	0	0	0	0	100	0
MORB (18)	0-10 (2)	90-100 (98)	0-8 (1)	0-6 (0.3)	0	0-10 (0.7)	0	0-100 (40)	0-100 (60)	0-100 (66)	0-100 (23)	0-67 (10)	0-8 (1)
Plume (4)	0-10 (4)	90-100 (96)	0-10 (4)	0	0	0	0	0	100	0-30 (8)	0-29 (7)	31-100 (77)	0-33 (8)
Subduction-related													
Backarc (35)	6-100 (50)	0-94 (50)	0-78 (31)	0-75 (13)	0	0-73 (6)	0-100 (1)	0-100 (15)	0-100 (84)	0-100 (75)	0-100 (18)	0-70 (6)	0-18 (1)
Backarc to forearc (22)	25-100 (74)	0-75 (26)	0-64 (19)	0-92 (35)	0-16 (3)	0-57 (17)	0-75 (16)	0-100 (53)	0-75 (31)	0-100 (84)	0-50 (8)	0-100 (8)	0
Forearc (4)	29-100 (78)	0-71 (22)	0-29 (10)	0-90 (43)	0-14 (8)	0-29 (17)	75-100 (88)	0-25 (12)	0	0-100 (75)	0-50 (25)	0	0
Forearc/Volcanic arc (3)	55-100 (78)	0-45 (22)	0-20 (12)	0-20 (7)	31-40 (37)	0-60 (22)	0-66 (29)	17-40 (27)	17-60 (44)	0-100 (50)	0-92 (46)	0-8 (4)	0
Proterozoic Complexes													
Subduction-unrelated													
Rift/Continental margin (3)	0	100								0-25 (8)	25-100 (67)	0	0-75 (25)
MORB (4)	0	100								0-100 (59)	0-100 (33)	0-25 (8)	0
Plume (2)	0	100								0	0-50 (25)	50-100 (75)	0
Subduction-related													
Backarc (13)	0-100 (39)	0-100 (61)	0-100 (33)	0-96 (13)	0-14 (3)	0-88 (11)	0	0-100 (28)	0-100 (72)	0-100 (42)	0-100 (37)	0-100 (21)	0
Backarc to forearc (9)	70-98 (82)	2-30 (18)	0-30 (11)	0-64 (32)	0-29 (11)	0-60 (28)	0-45 (21)	6-59 (33)	6-81 (46)	0-100 (50)	0-100 (25)	0-100 (25)	0
Forearc (3)	33-100 (72)	0-67 (28)	0-17 (6)	0-100 (39)	0-8 (3)	0-40 (24)	36-100 (79)	0-64 (21)	0	100	0	0	0
Forearc/Volcanic arc (7)	46-100 (80)	0-54 (20)	0-8 (1)	0	0-100 (36)	0-100 (43)	0	50-100 (68)	0-50 (32)	0-100 (25)	0-100 (61)	0-30 (8)	0-25 (6)
Archean Complexes													
Subduction-unrelated													
Rift/Continental margin (0)													
MORB (1)	50	50											
Plume (1)	0	100								100			
Subduction-related													
Backarc (10)	10-79 (57)	21-90 (43)	10-65 (45)	5-82 (17)	0	0-18 (5)	0-27 (5)	0-70 (63)	19-100 (32)	0-100 (49)	0-75 (21)	0-50 (28)	0-13 (2)
Backarc to forearc (10)	21-100 (78)	0-79 (22)	15-94 (48)	0-60 (10)	0-31 (3)	0-40 (17)	0-56 (11)	16-100 (56)	0-82 (33)	0-100 (80)	0-100 (12)	0-50 (8)	0
Forearc (4)	63-100 (83)	0-37 (17)	0-100 (44)	0-50 (12)	0-22 (6)	0-50 (21)	0-26 (7)	94-100 (93)	0	0-30 (15)	0	70-100 (85)	0
Forearc/Volcanic arc (14)	62-100 (91)	0-38 (9)	0-28 (6)	0-33 (5)	0-90 (17)	10-100 (63)	0-35 (5)	0-100 (41)	0-100 (54)	0-100 (47)	0-100 (30)	0-70 (23)	0

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Subduction-unrelated
ophiolite types

Subduction-related
ophiolites types

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

Figure 13

Figure 14

Figure 15

Figure 16

Figure 17

Figure 18

Figure 19

Figure 20

Figure 21

Figure 22

Figure 23

Figure 24

Figure 25

Figure 26

Figure 27

A: Environment of subduction-unrelated ophiolite types

B: Environment of subduction-related ophiolite types

C: Summary and models of intramagmatic silicic magma generation