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Geochemistry and Petrogenesis of the Yishak Volcanic Sequence, Kudi ophiolite, West Kunlun (NW China): Implications for the magmatic evolution in a subduction zone environment

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Abstract The Yishak Volcanic Series (YVS) is part of the Late Cambrian Kudi ophiolite in the western Kunlun Mountains, NW China, which marks the oldest suture zone in the Tibetan Plateau. The YVS is a well-preserved sequence comprising five conformable units (A to E) of basalts to andesites, with mafic dykes in Units A to D. These rocks are sequentially characterized upward by geochemical features of N-MORB (Unit A), E-MORB (Unit B), island-arc basalt (Unit C), BABB (Unit D and mafic dykes) and boninite series rocks (Unit E), reflecting complex interactions among various source components. The evolution of magma compositions from MORB-dominated to boninite series rocks clearly indicates temporally increasing subduction signatures, which, in combination with the extensional tectonic background, suggests that the Kudi ophiolite most likely formed in a spontaneous nucleation of subduction zone.

Key words: Geochemical variation, Subduction initiation, Kudi ophiolite, West Kunlun, Tibetan Plateau

Introduction

Ophiolites are obducted fragments of oceanic lithosphere believed to have formed predominantly in supra-subduction zone environments, where old crust returns to the mantle and new crust is generated (e.g. Hofmann 1988; Elthon 1991; Bloomer et al. 1995; Tatsumi and Eggins 1995). Subduction-related magmas show wide variations in composition, which are generally attributed to variable source compositions and contributions from multiple components (e.g. Hickey-Vargas 1992; Tatsumi and Eggins 1995; Wendt et al. 1997). Compositional variations in the source region may be closely related to the complicated geodynamic evolution of subduction systems (Flower and Levine 1987; Stern et al. 1990; Bloomer et al. 1995; Gribble et al. 1998), and tracing the compositional evolution of magmatic rocks in ophiolites should shed light on the geodynamic evolution of subduction processes and the nature of crustal-mantle interaction. Temporal variations in lithology and composition of volcanic sequences have been documented in some ophiolites (e.g. Photiades et al. 2000; Bortolotti et al. 2002; Einaudi et al. 2003) and these can provide insights into the tectonomagmatic history of subduction zone systems.

The Kudi ophiolite in the West Kunlun Mountains, NW Tibet, is a key element for understanding the earliest accretionary history of the plateau (Pan et al. 1994; Matte et al. 1996; Wang, 2004). Its volcanic succession, known as the Yishak Volcanic Sequence (YVS), is well-preserved and contains a continuous and relatively intact magmatic record of a supra-subduction environment (Deng, 1995; Yang et al. 1996). However, previous studies of the Kudi ophiolite mainly focused on its geology and petrology (e.g. Deng, 1995; Wang et al. 2002), and temporal variations of magma compositions in the YVS have not been previously investigated.

For this study, we conducted detailed field observations of the Kudi Ophiolite and sampled the YVS along a traverse with well-exposed outcrops. New geochemical analyses for major oxides, trace elements and Sm-Nd isotopic compositions revealed significant temporal variations along the profile that can be linked to the geodynamic evolution of the paleo-subduction zone. This work, in conjunction with other case studies (e.g. Piercey and Murphy 2001; Pfänder et al. 2002), should improve our understanding of subduction zone processes.

Geological background

Being a part of Tethyside orogenic collage (Sengör and Natal'in 1996), the Tibetan Plateau formed by continuously northward accretion of various terranes, as evidenced by southward younging ophiolitic belts (Dewey et al., 1988) (Fig.1). The oldest ophiolite belt in the region, the Kudi-Oytag belt, occurs on the northwestern margin of the plateau and extends westwards into northern Pamir and Pakistan (Pan et al., 1994; Schwab et al., 2004). However, its eastward extension is uncertain and connections with ophiolites in both

Altyn Tagh (Pan et al. 1996; Sobel and Arnaud 1999) and East Kunlun (Yin and Harrison 2000) have been proposed (Fig.1). The tectonic setting of the Kudi and Oytag Ophiolites has been hotly debated. Wang et al. (2001; 2002) considered that these ophiolites were produced in a marginal backarc basin, whereas Xiao et al. (2002) believed that they were arc-related.

The western Kunlun Mountains can be divided into the North Kunlun and South Kunlun Blocks separated by the Early Paleozoic Kudi ophiolite. The North Kunlun Block is generally believed to be the extension of the Tarim Craton because of its 2.4-Ga granitic basement (Zhang et al. 2003), whereas the South Kunlun Block possesses relatively young Nd model ages (up to 1.5 Ga) and accreted to the south margin of the North Kunlun Block in the Silurian (Matte et al. 1996; Yuan et al. 2003). The Kudi ophiolite is composed of several slices distributed mainly within the northern margin of the South Kunlun Block and represents a dismembered fragment of Proto-Tethyan lithosphere that once existed between Laurasia and Gondwana (Wang 1983; Pan et al. 1994; Deng 1995; Matte et al. 1996; Mattern and Schneider 2000; Wang et al. 2002) (Fig. 2). These slices contain a variety of rocks including dunite, peridotite, gabbroic cumulates, volcanic rocks and turbidites comparable to an ophiolite assemblage as defined by the Penrose conference in 1972 but lack a sheeted dyke swarm (Coleman, 1977). Both the dunite and peridotite in the mantle sequence are variably serpentinized and contain spinel with high Cr# (> 60), indicative of a supra-subduction environment (Wang et al., 2001). Abundant pyroxenite veins within the peridotite consist of clinopyroxene (60-80 modal%) with subordinate orthopyroxene and range in width from a few millimeters to ten centimeters.

Radiolarian fossils in chert overlying the YVS indicate an equivocal Ordovician age for the Kudi ophiolite (Fang 1998). More recently, a SHRIMP U-Pb zircon age of 510±4 Ma has been obtained for a gabbroic sample (Xiao et al. 2003), indicating that the Kudi ophiolite formed in the Late Cambrian.

Field relationships and petrography of the YVS

The YVS is located around the 137-km marker of the Yecheng-Mazar Highway (Fig. 2). It is thrust over a Precambrian metamorphic complex and is in fault contact with the Yirba pluton to the north (Mattern and Schneider 2000). The Yirba pluton is dated at 471 Ma and represents the active continental margin of the Tarim Craton (Pan et al. 1994; Matte et al. 1996; Yuan et al. 2002). The YVS and the overlying turbidite constitute the Yishak Group, which represents the uppermost part of the Kudi ophiolite (Wang, 1983). Pillow structures of the volcanic rocks and bedding of the overlying turbidite indicate that the volcanic strata dip to the WSW (Deng 1995). Paleocurrent data from the turbidites suggest deposition in a S- to SW-dipping forearc basin (present direction) (Fang 1998).

The well-preserved YVS has a total thickness of over 1,000 meters and consists of five major units, labeled A to E from the base upward (Fig. 3). In addition, a few lenses of chert, shale and tuffite are present locally between units C and D. These units are easily distinguished in the field on the basis of lithology and grain size, although the boundary between units D and E is somewhat obscure. Unit A consists of massive basalts that grade upward into pillow basalts of Unit B. Unit C also consists of massive basalts, which are overlain by amygdaloidal pillow basalts of Unit D. The uppermost Unit E is composed dominantly of amygdaloidal andesitic rocks, including a 300-meter sequence of andesitic breccia and tuff (Wang 1983; Fang 1998) (Fig. 3). Mafic dykes cut rocks of Units A to D, but are absent in Unit E. Pillows in the YVS are generally less than 1 meter in diameter and sometimes display radiating fractures and vesicular rims, containing secondary phases, such as chlorite, epidote, calcite and quartz. Although fine fractures can be locally observed on outcrops, the YVS did not undergo ductile deformation, as indicated by the well-preserved pillow structures.

Units A, B and C of the YVS are dominated by fine- to medium-grained basalt, with crystal sizes ranging from 50 to 200 μm. Rocks in the three units generally show ophitic to subophitic textures and are composed of plagioclase (30-40 modal%), clinopyroxene (20-40 modal%), Fe-Ti-oxides (< 5 modal%), and very fine-grained interstitial material (20-30 modal%). Olivine is very rare in the YVS and was only identified in Unit A. Units D and E are dominated by fine-grained, amygdaloidal basalts and andesites with abundant (~ 50 modal%), needle-like plagioclase (less than 100 μm in length). Some samples in Unit E contain even higher proportions of plagioclase (~ 60 modal%). The amygdules in these rocks range up to 5 cm in diameter and are generally filled with carbonate or silica. The mafic dykes are relatively coarse-grained with ophitic textures and are composed mainly of plagioclase (40-50 modal%) and clinopyroxene (30-40 modal%).

Rocks in the YVS have undergone low greenschist facies metamorphism, resulting in cloudy plagioclase, replacement of mafic minerals by chlorite and formation of very fine-grained, unidentifiable interstitial material. In addition, epidote, chlorite, quartz and calcite locally fill veins and vesicles.

Sample preparation and analytical methods

The Yishak Group is well exposed in an elongate triangle along the Yecheng-Mazar Road. A traverse, along the Yixiek Valley, cuts through the YVS in an E-W direction nearly perpendicular to its strike and provides a well-exposed profile for sampling. Samples in this study were collected both in the valley and along the road. After removal of weathered surfaces and amygdules and veins, the samples were crushed into small pieces and cleaned with deionized water in an ultrasonic vessel for 15 minutes. The dried rock chips were then powdered in an agate mill.

Major oxides were measured on fused glass discs, whereas Sc, V, Cr, Cu and Ni, were analyzed on pressed powder pellets using standard X-ray fluorescence (XRF) techniques at the University of Hong Kong. Rare earth elements (REE), Nb, Ta, Zr, Hf, Th and U were analyzed on a VG Elemental Plasma Quad Excell ICP-MS also at The University of Hong Kong, following the analytical procedures described by Qi et al. (2000). Precision of the XRF analyses is estimated to be 2% for major oxides present in concentrations greater than 0.5%, whereas precision for the ICP-MS analyses is generally better than 5%. Forty-nine samples were analyzed for major oxides and trace elements and the results are listed in Table 1.

Samples for Sm-Nd isotopic analyses were digested using mixed HNO_3 and HF acid. Nd was separated from other elements using a two-step exchange procedure described in Li et al. (2004). $^{143}Nd/^{144}Nd$ ratios were measured on a Micromass IsoProbe multi-collector mass spectrometer (MC-ICP-MS) at the Guangzhou Institute of Geochemistry (GIG), following the procedure of Liang et al. (2003). Repeat analyses for the Shin Estu JNdi-1 standard during the course of this study gave an average $^{143}Nd/^{144}Nd$ ratio of 0.512120±8 (2σ). Accordingly, the $^{143}Nd/^{144}Nd$ ratios reported in this study were adjusted relative to the Shin Estu JNdi-1 reference value of 0.512115 (Tanaka et al. 2000), corresponding to the La Jolla standard value (0.511860). The $^{147}Sm/^{144}Nd$ ratios were calculated from the Sm/Nd concentrations determined by the isotope dilution ICP-MS method on an ELAN 6000 quadrupole mass spectrometer also at the GIG. The initial Nd isotope compositions were calculated to 510 Ma, and the data are presented in Table 2.

Results

Major and trace elements

Major elements

Because alkaline elements are generally mobile during alteration, K_2O and Na_2O are not considered for the classification of the YVS. The YVS rocks can be classified as basalts, basaltic andesites, andesites and dacites, based on their anhydrous normalized SiO_2 concentrations (47.8 - 65.2 wt%). Rocks of Units A to C are mainly basalts, and those of Units D and E are dominantly basalts and andesites.

All of the units show relatively wide ranges of Cr and Ni (Table 1), suggesting extensive fractional crystallization. Samples from Units A and B contain intermediate TiO_2 (0.86 – 1.44 wt%), whereas those from Unit C have relatively low TiO_2 (< 1.0 wt%). Samples of Unit D are characterized by highly variable TiO_2 (0.91 - 2.8 wt%) and $Fe_2O_3^T$ (12.94 – 18.69 wt%) contents, some of which are significantly higher than those for other units (Table 1; Fig. 4). Rocks of Unit E are the most compositionally variable of the sequence and more than half of the analyzed samples have very low TiO_2 (< 0.5 wt%) and high Al_2O_3/TiO_2 ratios (23-102) (Table 1), features characteristic of boninite series rocks (Meijer 1980; Hickey and Frey 1982). However, two low-Ti samples of Unit E (96KL125 and 126) contain significantly high Al_2O_3 (18 - 20 wt%), suggesting that their depletion in HFSE was caused by accumulation of plagioclase. Therefore, the two samples are not classified as boninite series rocks although they possess boninite-like geochemical characteristics.

Trace elements

Rocks of Unit A are relatively depleted in light REE (LREE) (La/Yb_{Ch} = 0.7 - 0.8) and have positive Nb anomalies (Nb/Nb* = 1.2 - 2.7) (Table 1), features characteristic of N-MORB (Fig. 5). In contrast, samples from Unit B are characterized by high Nb contents (10.4 - 12.2 ppm) and are LREE-enriched (La/Yb_{Ch} = 1.7 - 2.2). These features, as well as their incompatible element ratios (e.g. Zr/Nb = 7.1 - 8.2; Th/Nb = 0.07 - 0.09) are very similar to those of E-MORB (Sun and McDonough 1989) (Fig. 5).

Samples of Unit C are moderately enriched in LREE (La/Yb_{Ch} = 1.40 - 3.13) and have relatively high La/Ta (20 - 60) and Th/Ta (2.57 - 11.37) ratios, with chondritic Zr/Nb (15 - 17) and Nb/Yb (1.28 - 1.68) ratios. Unlike the rocks of Units A and B, samples from Unit C are depleted in HFSE (Nb, Zr, Hf and Ti) relative to Th and REE, and have negative Nb anomalies (Nb/Nb*= 0.8 - 0.4) (Fig. 5), exhibiting typical island-arc signatures.

Samples from Unit D are LREE-depleted (La/Yb_{Ch} = 0.47 - 0.68) and are characterized by relatively high total REE contents (ΣREE = 39-92 ppm), with chondritic $(Gd/Yb)_{Ch}$ ratios (0.83 - 1.08) (Fig. 5). In addition, they have high Zr/Nb ratios (34 - 60) and variably negative Eu anomalies (Eu/Eu* = 0.79-0.97), as well as slight to intermediate Nb-depletion (Nb/Nb* = 0.55 - 0.89). The geochemical characteristics of these samples are very similar to those of backarc basin basalts (BABB) (Hochstaedter et al. 1990; Hawkins and Allan 1994).

Samples of Unit E are characterized by relatively high Cr contents (110 - 940 ppm) and possess high Th/La (0.10 - 0.47) and Th/Nb (0.14 - 1.00) ratios. Despite their wide range of compositions, all the samples of Unit E show negative Nb anomalies (Nb/Nb*= 0.20 - 0.86). These samples can be further divided into three subgroups: E-1, E-2 and E-3. The E-1 samples are LREE-depleted (La/Yb_{Ch} = 0.56 - 0.62) and contain relatively high total REE (ΣREE = 32-76 ppm). The E-2 and E-3 samples are all LREE-enriched (La/Yb_{Ch} = 1.14 - 2.36). The E-2 samples have intermediate Zr/Nb ratios (16 - 28), whereas the E-3 samples contain very low total REE (6.63 - 8.54 ppm) and show positive Zr-Hf anomalies (Zr/Sm_{PM} = 1.9 - 2.3) (Fig. 5). Two Al_2O_3-rich samples (96KL125 and 126) in the E-3 subgroup possess positive Eu anomalies (Eu/Eu* = 1.13 - 1.18), consistent with their high proportions of plagioclase.

The mafic dykes intruding the volcanic sequence are relatively depleted in LREE and have significant negative Nb anomalies (Nb/Nb* = 0.36 - 0.53), exhibiting BABB-like characteristics (Fig. 5). These rocks have incompatible element ratios (e.g. Zr/Nb, Zr/Sm, Nb/La and Zr/Yb), very similar to those of Unit D, but with generally lower total REE contents (Table 1).

Nd isotopic compositions

Rocks of the YVS exhibit a wide variation in initial Nd isotope compositions (εNd_T = +8.8 - +1.9), but they generally fall within the range of depleted mantle. Rocks of Units A (εNd_T =+6.9 - +8.8) and D (εNd_T =+7.3 - +8.0) have similar Nd isotope compositions, which are considerably higher than those of Units B (εNd_T =+5.9 - +6.3) and C (εNd_T =+5.6 - +5.7). Only the E-2 samples of Unit E were analyzed for Nd isotope compositions. Although all the E-2 samples are LREE-enriched and have similar trace element patterns (Fig. 5), they possess the most heterogeneous Nd isotopic compositions of the YVS (εNd_T = +1.9 - +8.7). The mafic dykes have εNd_T values (+5.1 - +7.4) slightly lower than those of Unit D, although they are otherwise geochemically similar.

Discussion

Evaluation of secondary alteration

Petrographic observations and relatively high volatile contents indicate that all the YVS rocks have undergone low-temperature seafloor hydrothermal alteration, which may result in significant uptake of H_2O, CO_2, Na, Cs, U and loss of Mg and the substantial redistribution of most major elements in basalts (e.g. Staudigel et al. 1996). Nevertheless, Th, REE, HSFE (e.g. Zr, Nb, Ti and Hf) and Nd isotopic compositions are generally unchanged during alteration of the oceanic crust (Kerr et al. 1997; Bédard 1999), and thus the following discussion of the geochemical and petrogenetic features of the YVS is based largely on the immobile elements and Nd isotope compositions.

Fractional crystallization versus multiple sources

Large compositional variations are observed both within and between the five units of the YVS. From the base upward the units of the YVS exhibit a regular alternation of LREE-enriched and LREE-depleted samples (Fig. 5). Likewise, the TiO_2 contents, the incompatible element concentrations and incompatible element ratios exhibit significant variations through the section. In particular, the Nb anomalies change from positive in Units A and B to negative in Units C, D, E and in the mafic dykes, whereas the Th/Nb and Th/La ratios increase from Unit A to Unit E.

Possible mechanisms to explain these variations include fractional crystallization and partial melting of heterogeneous source regions. Rocks from the YVS display wide ranges of Cr (5.57 - 939 ppm) and Ni (8.0 – 162 ppm) contents, and there are good correlations of some elements with TiO_2 and Yb (Fig. 7). These features, in combination with the variable negative Eu anomalies, argue for some degree of fractional crystallization. However, fractional crystallization alone cannot explain the observed geochemical variations, particularly, the wide ranges of La/Sm and Zr/Nb ratios and the Nd isotope compositions (Figs. 6, 8), because these features are relatively insensitive to magmatic differentiation (Sun and McDonough 1989; Pearce and Peate 1995). Thus, the observed geochemical variations of the YVS can best be explained by derivation from different mantle sources.

Identification of the components in the source region

Less-depleted N-MORB source for Unit A

The relative depletion of LREE and the absence of negative Nb anomalies in the rocks of Unit A are typical of N-MORB lavas (Fig. 5). However, their Zr/Nb ratios are lower than those of typical N-MORB (Sun and McDonough 1989) (Fig. 9), suggesting that Unit A was derived either from a “less depleted” MORB source or from low degrees of partial melting of a N-MORB mantle. Because these rocks are LREE-depleted and have relatively low Nb (2.44 - 4.08 ppm) and total REE contents (32 - 41 ppm), they cannot have been derived by low-degrees of partial melting. Therefore, rocks of Unit A were probably derived from a “less depleted” N-MORB source.

E-MORB source for Unit B

Rocks from Unit B may have originated from a relatively enriched mantle source, because they have LREE-enriched patterns, relatively high Nb contents, low Zr/Nb ratios and low εNd_T values (Fig. 5, 8). Enrichment of Nb and low Zr/Nb ratios have been ascribed to plume-related sources (Hofmann 1988; Gill and Whelan 1989), or involvement of partial melting of subducted slab material (Pearce and Peate 1995; Schiano et al. 1995) including sediments (Hoogewerff et al. 1997; Peat et al. 1997). Unit B possesses Zr/Y (< 3.3), Ta/Yb (< 0.27) and Th/Yb (< 0.36) ratios significantly lower than those of OIB (Sun and McDonough 1989). In particular, the low degree of LREE enrichment (La/Yb<3.2 vs. 17) and nearly flat HREE patterns (Gd/Yb<1.5 vs. 3.5) do not support an OIB origin for Unit B. It has been suggested that melting of relatively young, hot, subducted lithosphere can generate adakitic melts (Defant and Drummond 1990; Peacock et al. 1994), which may fertilize the upper mantle during melt-rock interaction and produce Nb-enriched basalts by partial melting (Kelemen et al. 1990; Sajona et al. 1996). However, Nb-rich basalts generated in this way possess lower Nb/La and higher Gd/Yb than plume-related rocks or E-MORB. In the Gd/Yb vs. Nb/La diagram, Unit B samples plot in the oceanic plateau field, as do most plume- and E-MORB-related samples (Fig. 10). Therefore, Unit B cannot be ascribed to partial melting of subducted lithosphere.

That subducted sediments can lead to Nb-enrichment in arc magmas has been documented in high-K rocks from Indonesia, which are characterized by low Nb/La and abnormally high Nb/Ta (>20) (Stolz et al. 1996; Hoogewerff et al. 1997). However, the rocks of Unit B have high $(Nb/La)_{PM}$ (1.18-1.53) and chondritic Nb/Ta (16 - 18), and are characterized by homogeneous compositions almost identical to those of E-MORB (Fig. 5 and 9). These geochemical characteristics preclude contributions from subducted sediments and imply an enriched (E-MORB) mantle source for Unit B.

Subduction-modified E-MORB residual source for Unit C

The rocks of Unit C are characterized by the depletion of HFSE (Nb, Zr, Hf and Ti) relative to REE (Fig. 5), suggesting the involvement of a subduction component. Unit C samples have a wide range of Ce/Y but relatively constant Zr/Nb, which show a horizontal array (Fig. 9). This feature is consistent with the addition of slab-derived fluid to the mantle source, because the dehydration of slab material generally mobilizes LILE and LREE and results in an increase in Ce/Y. This process would not affect the Zr/Nb ratios of the melts because both Zr and Nb are generally immobile during dehydration (e.g. Saunders et al. 1991; Thirlwall et al. 1994; Pearce and Peate 1995).

Because they are little affected by slab dehydration, the Zr/Nb ratios of Unit C rocks may reflect the nature of their mantle source. Because Nb is more incompatible than Zr, partial melting of a mantle source will leave a residuum with higher Zr/Nb. The Zr/Nb ratios (15 - 17) of Unit C rocks are lower than those of Unit A (20 - 21) (Fig. 9), suggesting the mantle source of Unit C was less depleted than that of Unit A. This precludes the possibility that Unit C rocks were derived from residual mantle formed by extraction of Unit A melts. However, the mantle source of Unit C could be the residual mantle produced by extraction of Unit B melts, because during the extraction of these melts the residuum would evolve toward higher Zr/Nb and lower Ce/Y. Sample KL42 has the lowest Ce/Y ratio (0.51) and represents the least subduction-modified sample in Unit C. Its Ce/Y ratio is lower than those of Unit B rocks (0.57 -

0.71), which is consistent with its derivation by aqueous partial melting of residuum Unit B mantle, although other possibilities may also exist.

Subduction-modified, depleted mantle source for Unit D and the mafic dykes

The BABB-like rocks of Unit D imply that they were derived by partial melting of a depleted mantle source with involvement of a subduction component. The relatively high TiO_2 content of the Unit D rocks probably reflects limited fractionation of Fe-Ti oxides under low oxygen fugacity (Juster and Grove 1989; Harper 2003). Their high REE concentration may reflect low-pressure fractional crystallization. The negative Eu anomalies (Eu/Eu* = 0.79 - 0.97) and low Cr (5.57 - 88.2 ppm), Ni (13.1 –50.5 ppm) and Sc (4.87 - 55.5 ppm) contents certainly indicate that plagioclase, chromite, olivine and perhaps pyroxene were fractionated from the Unit D melts. The good correlations shown by Unit D samples (Fig. 7) also supports the interpretation that the relatively high REE contents are due to crystal fractionation. Because low-pressure fractional crystallization, either in a closed or open system, rarely changes the REE patterns (Saunders 1984), the relatively low $(La/Yb)_{Ch}$ ratios, in conjunction with the high Zr/Nb ratios and consistent εNd_T values (Fig. 8), suggest that the Unit D magmas came from a depleted mantle source. Rocks of Unit D possess nearly flat HREE patterns (Gd/Yb_{Ch} = 0.83 - 1.08), suggesting that there is no significant garnet left in the residual source to fractionate the HREE during partial melting. The low $(La/Yb)_{Ch}$ and Th/Nb ratios and relatively consistent εNd_T values probably reflect a relatively small input of slab-derived fluid.

The mafic dykes are broadly similar in composition to the lavas of Unit D. Their high Zr/Nb ratios, low $(La/Yb)_{Ch}$ ratios and variably negative Nb-anomalies indicate that the dykes formed by partial melting of a depleted mantle source modified by slab-derived fluid. One sample (KL32) has relatively high total REE and TiO_2 but low Cr contents and a small Nb depletion (Nb/Nb*= 0.9). Its incompatible element ratios and εNd_T values are similar to those of Unit D samples (Fig. 8, 9), suggesting that it may have formed by a similar process. Other samples show small Eu anomalies (Eu/Eu*= 0.88 - 1.07) and considerable Nb-depletion (Nb/Nb*= 0.2 - 0.8), implying that they experienced less fractional crystallization of plagioclase.

Variable arc-related mantle source for Unit E

The LREE-depleted characteristics and chondritic $(Gd/Yb)_{Ch}$ ratios of E-1 samples indicate that they were produced by partial melting of a depleted mantle source. The variably negative Nb, Zr-Hf and Ti anomalies (Fig. 5) generally reflect the retention of these elements by rutile or other residual titanium-bearing phases in subduction zones (Foley et al. 2000). One REE-rich sample (KL60) contains lower Sc, V, TiO_2 and possesses smaller Eu/Eu* values than the other E-1 sample (Table 1; Fig. 5), indicating that it may have experienced more fractionation of clinopyroxene, Ti-magnetite and plagioclase.

The E-2 samples are LREE-enriched with relatively low Zr/Nb ratios. Their Zr/Nb and Ce/Y ratios co-vary on a hyperbolic curve connecting the depleted mantle component and the global average oceanic sediments (Fig. 9), suggesting that they resulted from mixing of N-MORB and sediment-derived melts. The involvement of subducted sediments may have resulted in a significant increase in their Th/La and Th/Nb ratios and a decrease in their Nb/La ratios. However, precise estimation of the contribution of subducted sediments is difficult and largely depends on the mobility and compatibility of elements used in calculation. It was estimated, for example, that more than 99 wt% LIL elements in Shasta lavas has been contributed from a fluid-rich component, whereas only 80-95 wt% Nd and 60-80 wt% Sm were associated with such a component (Grove et al. 2002). Simple mixing calculation based on Nd isotopic composition and Ce/Y ratios suggests that the contribution of sediments to the rocks of sub-unit E-2 was smaller than 15%. However, the contribution was up to 60% if the calculation is done using Ce/Y and Zr/Nb ratios. The calculated proportion represents only the contribution of elements to the magma and is not the proportion of sediment melts. Sample KL59 plots close to the N-MORB component on the mixing line and its compositional characteristics could reflect the nature of the N-MORB source. Its εNd_T value (+8.7) is the highest among the E-2 samples and is very close to those of Unit A samples, suggesting that it may have been derived from a depleted mantle source similar to that of Unit A.

The E-3 samples are characterized by high Zr/Nb and Zr/Sm ratios and very low incompatible element concentrations. The very low HREE contents (Yb = 0.30 - 0.46 ppm) of the E-3 samples imply a garnet-rich residue. High Zr/Sm ratios in island arc magmas have been ascribed to involvement of an OIB-like component (e.g. Hickey-Vargas 1989; Stern et al. 1991), addition of slab melts (Taylor et al. 1994), or partial melting of a refractory source with significant residual garnet (Smithies et al. 2004). The E-3 samples contain very low Nb (0.30 - 0.45 ppm) contents and have pronounced Nb-depletion (Nb/Nb*= 0.2 - 0.3), precluding the involvement of an OIB component. Partial melting of a subducting slab would generate adakitic magma, which generally possesses Zr/Nb ratios lower than those of N-MORB (e.g. Stern and Kilian 1996). Because Nb is more enriched in partial melts than Zr, an influx of slab- or sediment-derived melts into the mantle will substantially decrease Zr/Nb ratios in the hybrid magmas (Münker 2000). The E-3 samples have Zr/Nb ratios (53 - 65) much higher than those of N-MORB (ca. 32, Sun and McDonough 1989), inconsistent with the addition of slab-derived melts. The high Zr/Sm (47 - 57) ratios may indicate a severely depleted, clinopyroxene-poor source, because orthopyroxene possesses higher HFSE/REE ratios than clinopyroxene and partial melting of a clinopyroxene-poor source will lead to liquids with high Zr/Sm ratios (Sun and Kerrich 1995; Wolde et al. 1996). Since the high Zr/Nb ratios of the E-3 samples do not support the involvement of slab- or sediment-derived melt, the most likely subduction component is aqueous fluid generated by dehydration of a subducting slab (Fig. 9). Two Al_2O_3-rich (18 - 20 wt%) E-3 samples contain abundant plagioclase, probably reflecting the relatively late crystallization of plagioclase under high-H_2O contents (Müntener et al. 2001). Samples from the three subunits are interbedded with each other (Fig. 3). A similar stratigraphic relation has been documented in Guam, which was ascribed to alternately tapping differently depleted, hydrated mantle sources from distinct levels (Reagan and Meijer 1984), or alternatively to partial melting of a hydrous mantle diapir with thermal and compositional zoning (Tamura 1994).

Implications for initiation of subduction zones

The evolution of magma compositions in the YVS from MORB-dominated to boninite series rocks clearly indicates temporally increasing subduction signatures, which, in combination with the extensional tectonic background, suggests that the Kudi ophiolite most likely formed in a spontaneous nucleation of subduction zone (SNSZ) (Stern 2004). SNSZ generally occurs along a line of weakness where lithospheres with different ages and densities are juxtaposed, and subduction is initiated when the older and denser lithosphere sinks. Continuous subsidence of the cold lithosphere would lead to hinge retreating and extension above the sinking lithosphere (Kincaid and Olson 1987), and increasing material supply from the sinking slab would endow the magmas with increasingly evident subduction signatures.

N-MORB and E-MORB components commonly exist in oceanic mantle, with the latter being enclosed in the deeper N-MORB mantle, possibly in the form of blebs or streaks (e.g. Zindler and Hart 1986; Hickey-Vargas 1992; Woodhead et al. 1993). The successive eruption of N-MORB and E-MORB in Units A and B suggests upwelling and decompression melting of asthenosphere either in a spreading center, or during an early stage of lithosphere sinking (Stern 2004 and references therein). In some cases (e.g. Lau Basin), MORB-type rocks can also occur in the early stage of back-arc opening, but they actually formed in a forearc setting and usually show negative Nb-anomalies (Hawkins and Allan 1994). Unlike those of Lau Bain, the MORB-type rocks in Units A and B do not show Nb-depletion, indicating that their mantle sources were not significantly affected by fluids from subsiding slab.

With the subsidence of the lithosphere, increasing pressure and temperature on the sinking slab will result in extensive expulsion of water and mantle metasomatism. Rocks of Units C ($Nb/Nb^*<0.8$; $La/Yb_{Ch}>1.4$) and D ($Nb/Nb^*<0.9$; $La/Yb_{Ch}<0.7$) are compositionally different, which may be ascribed to different degrees of hydrous metasomatism in their mantle sources. The mantle upwelling and the presence of mafic dykes within the YVS indicate an extensional environment, which could be closely related to the hinge retreating of the subsiding lithosphere.

Unit E consists chiefly of andesites with rather heterogeneous compositions. E-2 and E-3 samples, the boninite series rocks contain relatively low HFSE and HREE but possess high Cr, indicating a mantle source more depleted than those for other units. The severely depleted nature of these rocks may be related to extensive melt extraction of the former four units. Genesis of boninite series rocks requires abnormally hot heat flux (Crawford et al. 1989), and probable scenarios include initial subduction along a transform fault (Meijer 1980), heat conduction from an ascending MORB mantle diapir (Crawford et al. 1981), or ridge subduction (Crawford et al. 1989). More recently, Macpherson and Hall (2001) invoked subduction initiation on a mantle plume to explain the predominant Eocene boninites in the Izu-Bonin-Mariana forearcs. Rocks of Units A and B do not possess plume-related characteristics, and the boninite series lavas are volumetrically subordinate in the YVS, therefore a mantle plume influence seems unlikely. Ridge subduction can also be excluded for the YVS boninite series rocks because of the absence of slab melts and Nb-enriched basalts (Macpherson and Hall 2001 and references therein). The mantle upwelling and extensional background revealed in the YVS suggest that the boninite series lavas were most likely generated by partial melting of hydrated mantle under conductive heating of an ascending mantle diapir that may occur in early subsidence of the lithosphere (Stern 2004), or in the earliest stage of backarc spreading (Crawford et al. 1981). Because the mantle upwelling and mafic dykes all occurred prior to Unit E, the boninite series rocks in Unit E cannot represent the magmas of the earliest backarc spreading. Instead, they most likely formed in the infant stage of a subduction zone.

Comparison to other ophiolites with boninite series rocks

The compositions of SSZ magmas generally reflect complex interactions among various source components and the geodynamics of subduction zones. Although volcanic rocks in ophiolites generally show variable compositions, multiple units coexisting within a continuous volcanic sequence as exhibited in the YVS is not common. Like the YVS, some ophiolites show increasing subduction signature in erupted sequences. Lavas from the Dun Mountain ophiolite, New Zealand evolved from E-MORB, BABB to IAT and boninite (Sivell and McCulloch 2000); whereas the Oman ophiolite, the largest ophiolite complex in the world, simply shows a compositional transition from MORB-like to boninitic rocks (Ishikawa et al. 2002). These volcanic sequences were interpreted to be generated in response to the initiation of subduction. The Troodos ophiolite, Cyprus shows a compositional transition from arc tholeiite to boninite in its volcanic sequence, which was interpreted to reflect the splitting of a volcanic arc or propagation of backarc spreading into an arc lithosphere (Flower and Levine 1987; Portnyagin et al. 1997). However, geochemical data from the volcanic sequence of the Troodos ophiolite indicated that there was an increasing depletion from the basal Group A to the upmost Group C (Flower and Levine 1987). This may suggest that the seriously depleted shallow source was not replenished by fertile deep mantle. Lack of replenishment implied that there was no steady regime of mantle upwelling as usually occurred during spreading of a backarc basin. Instead, the increasing depletion and subduction signatures of the volcanic sequence suggest that the Troodos ophiolite was most likely generated during the initial stage of subduction (Shervais 2001).

Some other ophiolites possess progressively enriched lava sequences. One example is the Betts Cove ophiolite, Newfoundland, the volcanic sequence of which evolved from extremely depleted boninite, upward to less depleted boninite and tholeiitic rocks (Coish et al. 1982). The Koh ophiolite, New Caledonia, however, evolved from low-Ti tholeiitic rocks to boninite and BABB-like tholeiite (Meffre et al. 1996). The increasing enrichment in the lava composition may be related to progressive involvement of less depleted mantle sources (Bédard 1999), and the sequential eruption of these lavas has been ascribed to seafloor spreading in a forearc environment or the propagation of a spreading center into a proto-arc crust (Meffre et al. 1996; Bédard 1999). Most of the aforementioned ophiolitic volcanic sequences can be generated in forearc environment during the early stage of a subduction zone, where mantle upwelling, extension and severely depleted mantle source material for boninite genesis can be provided (Stern 2004).

Conclusions

The Yishak Volcanic Sequence can be divided into five units based on their compositional characteristics, recording a complete spectrum of magmatic evolution in a subduction-related environment. Unit A has N-MORB-like characteristics and the most depleted Nd isotope compositions, whereas the overlying Unit B possesses E-MORB-like compositions, characterized by relatively high Nb contents and relatively low εNd_T values. Both units lack significant subduction-related signature and formed by sequential partial melting of upwelling mantle. Units C to E and the mafic dykes exhibit subduction-related characteristics and were produced from partial melting of different mantle sources, induced by various slab-derived components. The slab-derived components changed from aqueous fluid in Units C, D and the mafic dykes to partial melts of subducting sediments in the E-2 subunit. The progressively increasing subduction signatures, existence of boninite series rocks and extensional tectonic background of the YVS suggest that the Kudi ophiolite most likely formed in an initial stage of a subduction zone.

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Figure Legends

Fig. 1 Tectonic framework of the Kunlun Mountains and adjacent areas

Fig. 2 Geological Map of the Kudi area, West Kunlun, showing the distribution of the Kudi ophiolite

Age data of the granitoids are from Yuan et al. (2002).

Fig. 3 Stratigraphic column of the Yishak Volcanic Sequence of the Kudi ophiolite.

The thickness was estimated according to Wang (1983) and Xiao WJ et al. (2002).

Fig. 4 Correlation diagrams of Al_2O_3 vs. TiO_2 and $F_2O_3^T$ vs. TiO_2 for the YVS rocks.

Major element data have been recalculated on anhydrous basis.

Fig. 5 Primitive mantle normalized trace elemental spider diagrams of the volcanic rocks from the YVS

Data for Primitive mantle are from Sun and McDonough (1989).

Fig. **6** Yb vs. La/Sm diagram for the YVS rocks

Fig. 7 Major and trace element correlation diagram for the YVS racks

Major element data have been recalculated on anhydrous basis. Symbols as in Fig. 4.

Fig. 8 Zr/Nb vs. εNd_T diagram for the YVS rocks

Fig. 9 Ce/Y vs. Zr/Nb diagram illustrating the source components and their relationship during the genesis of the YVS.

Data for N-MORB, E-MORB and OIB are from Sun and McDonough (1989). Data for the global average oceanic sediments are from Plank and Langmuir (1998).

Fig. 10 Gd/Yb Vs. Nb/La diagram discriminating Nb-enriched basalts and basalts from enriched mantle.

Oceanic plateau basalt data from Floyd (1989); Mahoney et al. (1993); Kerr et al. (1997); Condie et al. (2002); Tejada et al. (2002). Nb-enriched basalt data from Sajona et al. (1996); Wyman et al. (2000); Polat and Kerrich (2001); Hollings (2002).

Table 1. Major oxides and trace element abundance of representative samples from the YVS

Sample	**Unit A**								**Unit B**								**Unit C**						**Unit D**		
	KL18	KL20	KL21	KL23	KL24	KL25	KL26	KL27	KL29	KL30	KL31	KL33	KL34	KL38	KL39	96KL 112	KL42	KL44	KL74	KL75	KL76	96KL 60	KL46	KL47	KL48
SiO_2	45.21	47.08	47.53	48.39	46.44	47.34	47.86	47.02	49.71	55.45	56.15	50.54	48.32	51.56	47.13	52.16	49.79	46.62	43.22	51.25	46.32	50.85	47.52	47.89	47.41
TiO_2	0.92	0.92	0.91	1.05	1.15	0.94	0.86	1.00	1.05	1.40	1.40	1.40	1.41	1.39	1.32	1.44	0.56	0.65	0.62	0.95	0.96	0.59	1.67	1.68	2.2
Al_2O_3	14.17	14.89	14.34	14.67	13.05	13.78	13.01	13.43	14.24	12.58	12.39	12.56	12.16	12.74	11.89	13.99	13.5	16.76	11.47	14.63	12.93	14.72	12.78	12.71	12.19
$Fe_2O_3^T$	11.21	12.06	11.94	11.64	13.9	12.06	11.08	13.05	12.02	12.62	12.72	13.63	17.86	14.69	14.24	13.04	14.19	10.64	11.58	14.67	13.36	11.73	17.29	17.38	18.36
MnO	0.15	0.20	0.18	0.19	0.19	0.18	0.17	0.21	0.18	0.19	0.19	0.20	0.27	0.22	0.17	0.10	0.18	0.14	0.29	0.18	0.44	0.14	0.23	0.24	0.23
MgO	7.25	7.97	8.12	6.90	7.19	7.95	7.09	9.69	7.60	4.50	4.63	5.38	5.53	5.24	3.69	3.62	3.92	7.02	11.55	3.90	2.85	7.44	5.94	5.93	4.97
CaO	6.55	10.9	10.07	7.25	8.91	10.05	12.03	9.25	5.86	4.61	4.64	6.92	7.71	5.79	8.75	5.76	6.61	4.66	9.94	3.64	9.49	6.65	7.51	7.25	7.21
Na_2O	5.07	3.07	2.86	5.69	4.42	3.65	2.56	2.70	5.04	6.06	5.07	3.81	2.95	5.02	3.10	5.13	3.49	5.87	2.77	6.99	2.80	3.95	4.38	4.47	3.97
K_2O	0.16	0.41	0.49	0.24	0.11	0.58	0.07	0.08	0.14	0.10	0.10	0.20	0.42	0.15	0.36	0.03	0.06	0.09	0.03	0.04	0.81	0.52	0.24	0.21	0.04
P_2O_5	0.07	0.07	0.07	0.12	0.08	0.07	0.07	0.08	0.14	0.17	0.18	0.18	0.19	0.17	0.17	0.17	0.24	0.09	0.13	0.19	0.24	0.09	0.14	0.15	0.21
LOI	8.73	2.86	2.95	3.54	3.22	2.82	4.77	3.44	3.39	2.82	2.44	4.69	2.62	2.74	8.85	4.97	6.96	6.91	8.66	2.59	10.05	3.80	2.38	2.18	2.62
Total	99.47	100.44	99.46	99.68	98.66	99.42	99.57	99.93	99.37	100.49	99.9	99.52	99.45	99.72	99.69	100.41	99.49	99.44	100.25	99.04	100.25	100.48	100.09	100.07	99.42
Sc	49.9	59.6	51.6	54.4	54.8	49.7	44.5	50.6	43.9	44.3	40.9	42.0	37.9	41.7	38.0	37.4	36.1	42.9	57.5	48.5	42.6	41.1	4.87	15.1	49.1
V	314	356	340	362	407	322	326	353	288	325	363	393	355	380	319	418	221	405	274	300	269	259	417	394	481
Cr	405	372	265	81.4	90.3	274	285	245	251	73.0	60.2	69.2	100	82.0	53.9	20.0	345	53.5				25.8	77.5	54.7	68.3
Ni	134	129	114	72.7	69.5	97.0	89.9	101	90.9	27.6	38.7	38.8	39.1	35.6	38.8	28.8	99.4	21.2				62.8	50.5	14.5	35.9
Cu	147	148	149	170	193	155	143	161	102	103	97.4	85.2	90.3	94.0	97.9	58.5	95.8	79.8				110	98.6	38.6	81.4
Y	21.4	25.5	21.3	27.5	26.7	22.8	21.8	23.9	22.1	28.6	28.9	29.4	28.0	29.9	30.1	28.1	14.1	30.4	15.2	29.2	29.7	17.9	37.9	40.5	59.3
Zr	51.2	66.0	53.4	85.0	62.1	53.3	49.2	55.5	73.6	86.8	90.4	90.7	84.3	92.3	84.8	88.9	34.2	69.1	33.2	70.7	70.4	38.5	120	119	163
Nb	2.53	3.25	2.60	4.08	3.09	2.69	2.44	2.73	9.84	12.2	11.2	11.0	10.4	11.2	10.6	11.1	2.22	4.15	2.27	4.11	4.23	2.55	2.00	1.98	3.35
La	2.45	3.29	2.53	3.42	3.08	2.70	2.59	2.74	6.87	8.95	9.13	8.27	8.16	7.98	7.64	6.99	2.81	9.11	6.25	9.65	11.0	6.45	3.03	3.06	5.78
Ce	6.89	9.08	6.99	9.95	8.41	7.28	6.93	7.46	15.8	19.5	19.8	18.7	18.2	18.5	18.4	16.1	7.20	19.9	13.8	21.5	23.1	13.2	10.2	10.3	17.2
Pr	1.17	1.50	1.16	1.67	1.41	1.22	1.16	1.25	2.21	2.73	2.73	2.67	2.52	2.68	2.70	2.24	1.00	2.74	1.91	2.95	3.18	1.71	1.92	1.94	3.04
Nd	6.30	8.06	6.33	8.86	7.52	6.55	6.27	6.75	10.1	12.8	12.9	12.4	11.7	12.7	12.8	10.7	5.00	12.4	8.89	13.0	14.4	7.51	11.1	11.0	16.6
Sm	2.14	2.72	2.21	3.02	2.57	2.26	2.06	2.29	2.72	3.52	3.54	3.39	3.20	3.50	3.62	3.05	1.58	3.33	2.50	3.41	3.54	1.95	4.05	4.04	5.71
Eu	0.70	0.77	0.79	1.03	0.95	0.85	0.77	0.77	0.81	1.22	1.05	1.11	0.99	1.09	1.20	1.08	0.52	1.02	0.66	1.00	1.06	0.73	1.34	1.33	1.72
Gd	2.79	3.51	2.87	3.92	3.41	2.98	2.89	3.07	3.15	4.15	4.14	3.95	3.81	4.24	4.25	3.81	1.94	3.97	2.79	3.77	4.10	2.37	5.43	5.37	7.32

Table 1. (Continued)

Sample	Unit A								Unit B								Unit C						Unit D		
	KL18	KL20	KL21	KL23	KL24	KL25	KL26	KL27	KL29	KL30	KL31	KL33	KL34	KL38	KL39	96KL 112	KL42	KL44	KL74	KL75	KL76	96KL 60	KL46	KL47	KL48
Tb	0.49	0.62	0.50	0.67	0.61	0.53	0.50	0.54	0.51	0.68	0.69	0.66	0.64	0.70	0.72	0.70	0.33	0.68	0.42	0.63	0.67	0.43	0.97	0.98	1.30
Dy	3.33	4.27	3.38	4.59	4.10	3.63	3.51	3.72	3.45	4.50	4.65	4.49	4.28	4.69	4.81	4.56	2.25	4.70	2.61	4.30	4.53	2.80	6.65	6.68	8.82
Ho	0.72	0.91	0.72	0.97	0.89	0.79	0.77	0.81	0.73	0.95	1.00	0.97	0.93	1.01	1.02	0.99	0.49	1.04	0.53	0.94	0.99	0.62	1.48	1.48	1.95
Er	2.12	2.68	2.15	2.77	2.69	2.34	2.28	2.47	2.17	2.80	2.91	2.86	2.76	3.02	3.05	2.97	1.41	3.12	1.44	2.79	2.93	1.91	4.39	4.41	5.72
Tm	0.29	0.41	0.31	0.41	0.38	0.33	0.35	0.37	0.31	0.41	0.43	0.40	0.38	0.45	0.42	0.41	0.20	0.49	0.21	0.40	0.43	0.27	0.62	0.64	0.82
Yb	2.05	2.68	2.14	2.65	2.69	2.27	2.38	2.39	2.17	2.77	2.87	2.81	2.77	2.98	2.96	2.79	1.36	3.25	1.35	2.82	2.89	1.85	4.38	4.38	5.80
Lu	0.32	0.42	0.33	0.40	0.40	0.33	0.38	0.36	0.33	0.44	0.44	0.42	0.42	0.46	0.44	0.43	0.21	0.51	0.22	0.44	0.46	0.29	0.67	0.67	0.90
Hf	1.37	1.78	1.44	2.23	1.60	1.40	1.34	1.50	1.77	2.09	2.20	2.19	2.03	2.25	2.19	2.14	0.89	1.72	0.87	1.65	1.61	0.97	2.94	2.90	3.99
Ta	0.17	0.21	0.17	0.26	0.19	0.17	0.16	0.17	0.59	0.71	0.69	0.67	0.62	0.67	0.66	0.60	0.14	0.22	0.10	0.19	0.19	0.12	0.14	0.12	0.21
Th	0.16	0.24	0.20	0.33	0.20	0.19	0.17	0.21	0.79	0.90	0.92	0.76	0.84	0.91	0.91	0.98	0.36	1.15	1.18	1.79	1.25	0.83	0.40	0.41	0.56
U	0.11	0.12	0.19	0.29	0.09	0.08	0.14	0.11	0.35	0.05	0.34	0.34	0.32	0.26	0.29	0.22	0.11	0.36	0.53	0.66	0.33	0.27	0.69	0.71	0.45
ΣREE	31.8	40.9	32.4	44.3	39.1	34.1	32.8	35.0	51.3	65.4	66.2	63.1	60.8	63.9	64.0	56.8	26.3	66.2	43.6	67.5	73.2	42.1	56.2	56.2	82.6
Nb/Nb*	1.37	1.24	1.24	1.30	1.33	1.27	1.24	1.22	1.43	1.46	1.30	1.49	1.34	1.41	1.36	1.44	0.75	0.43	0.28	0.33	0.39	0.37	0.61	0.60	0.63

Nb/Nb*= 0.3618*Nb/(Th*La)^$^{0.5}$; LOI = loss on ignition; Chondrite data are from Taylor and McLennan 1985; Primitive Mantle data are from Sun and McDonough 1989.

Table 1. (Continued)

Sample	Unit D							Unit E													Mafic Dyke			
	KL50	KL78	KL79	96KL 58	96KL 113	96KL 115	96KL 118	KL59†	KL60*	KL62†	KL65†	KL66*	KL69‡	96KL 122†	96KL 124†	96KL 125‡	96KL 126‡	96KL 128†	96KL 129†	96KL 130†	KL32	KL81	KL83	96KL 108
SiO_2	53.23	49.75	48.83	45.82	48.13	44.60	51.3	54.75	53.93	59.26	46.11	56.37	48.16	46.17	56.57	51.38	49.6	57.98	58.33	61.25	47.93	47.02	45.09	45.12
TiO_2	0.91	1.66	1.14	2.80	2.15	1.03	1.23	0.83	0.81	0.75	0.72	0.89	0.14	0.67	0.46	0.23	0.21	0.51	0.47	0.46	2.16	0.76	0.66	0.67
Al_2O_3	12.51	14.78	13.7	13.40	12.33	13.79	15.98	10.32	10.18	9.96	15.38	12.19	14.28	15.30	11.60	20.00	18.00	12.55	12.37	11.61	12.38	15.05	14.05	15.1
$Fe_2O_3^T$	13.06	14.58	14.96	18.69	18.28	14.75	12.94	17.26	16.97	14.42	8.85	13.07	6.88	11.63	9.63	6.80	4.03	10.82	11.29	9.39	17.67	8.96	8.79	10.11
MnO	0.16	0.08	0.10	0.26	0.20	0.10	0.15	0.19	0.19	0.12	0.16	0.21	0.11	0.17	0.15	0.05	0.10	0.08	0.09	0.08	0.24	0.12	0.15	0.12
MgO	2.13	3.18	3.66	5.25	5.26	3.66	2.61	2.75	2.68	2.94	8.00	2.11	8.29	8.56	3.40	3.90	7.33	3.82	4.25	3.84	5.08	7.87	7.25	9.14
CaO	6.40	5.40	5.89	5.49	6.38	6.51	4.59	5.86	5.80	3.13	6.67	5.68	8.88	4.47	6.33	4.19	4.46	3.84	3.67	3.51	7.43	7.07	9.28	7.49
Na_2O	2.40	2.70	2.33	3.73	4.53	3.63	4.10	1.23	1.62	1.99	2.55	2.22	0.99	2.78	2.91	3.97	4.95	2.28	2.67	2.88	4.10	3.38	4.30	3.27
K_2O	0.31	1.07	0.91	0.11	0.32	0.15	0.47	0.07	0.07	0.16	0.75	0.28	0.23	0.68	0.70	1.72	0.05	0.93	0.87	0.76	0.40	0.04	0.08	0.07
P_2O_5	0.09	0.29	0.18	0.26	0.17	0.09	0.25	0.26	0.26	0.08	0.13	0.12	0.01	0.17	0.27	0.21	0.19	0.16	0.15	0.11	0.21	0.07	0.07	0.19
LOI	8.23	7.36	8.00	4.56	2.76	12.40	7.15	5.86	7.38	6.79	11.04	7.34	12.34	9.91	7.95	7.53	11.00	7.16	6.37	6.14	1.75	10.13	10.6	9.27
Total	99.43	100.85	99.70	100.37	100.51	100.71	100.77	99.37	99.90	99.61	100.35	100.49	100.32	100.51	99.97	99.98	99.92	100.13	100.52	100.03	99.36	100.48	100.32	100.55
Sc	33.0	14.7	38.5	55.5	47.9	32.6	34.3	14.3	25.1	16.6	18.2	30.6	31.9	40.8	27.9	40.3	40.4	72.2	75.8	92.0	51.1	44.8	43.8	43.4
V	329	196	279	523	475	358	278	244	206	182	212	299	113	209	139	128	94.2	164	131	140	495	249	206	211
Cr	65.7			88.2	48.2	5.57		353	115	110	580	112	486	733	535		463	939	472	377	79.4			133
Ni	13.1			51.0	43.6	13.3		75.7	10.0	9.25	149	8.00	73.0	162	66.4		60.6	81.9	78.1	69.9	39.8			66.9
Cu	58.9			72.6	85.3	28.1		93.1	25.0	18.3	24.4	6.57	10.1	47.3	22.1		3.30	14.6	13.6	19.6	77.7			107
Y	31.2	41.3	32.6	64.1	49.0	30.2	31.6	24.7	58.2	21.8	18.5	24.3	3.15	15.9	14.4	4.50	4.00	14.8	14.5	14.7	55.3	20.1	18.2	17.3
Zr	59.2	145	81.0	188	142	60.1	94.3	87.9	63.0	43.9	56.7	50.7	19.4	48.5	53.8	24.8	22.6	59.3	62.2	57.2	167	53.6	45.2	47.3
Nb	1.67	2.82	1.61	4.14	3.37	1.77	2.65	3.67	1.99	1.58	3.21	1.81	0.30	2.90	3.14	0.45	0.43	3.62	3.46	3.28	3.43	0.91	0.76	0.86
La	2.74	4.45	2.76	6.04	5.03	2.80	3.32	4.03	5.33	2.98	6.15	2.01	0.83	5.93	5.17	1.02	0.85	4.03	4.56	4.97	5.21	1.87	1.30	2.14
Ce	7.95	13.7	8.79	18.6	14.50	7.49	9.78	11.3	15.3	8.08	14.6	6.02	2.04	14.6	11.7	2.50	2.28	10.1	10.5	11.6	16.2	5.78	4.29	6.49
Pr	1.34	2.52	1.65	2.95	2.29	1.13	1.65	1.85	2.59	1.31	2.04	1.06	0.25	1.89	1.55	0.34	0.31	1.35	1.43	1.53	2.89	1.02	0.82	0.98
Nd	7.24	13.7	9.58	17.5	13.11	6.10	9.02	9.66	14.2	6.94	9.69	6.12	1.20	9.06	7.23	1.71	1.49	6.65	6.83	7.24	15.7	5.58	4.72	5.65
Sm	2.55	4.61	3.31	6.25	4.56	2.18	3.45	3.09	5.04	2.35	2.56	2.37	0.34	2.40	1.91	0.48	0.48	1.93	1.92	1.97	5.53	1.97	1.73	1.89
Eu	0.84	1.44	1.05	2.13	1.63	0.84	1.03	1.19	1.42	0.78	0.74	0.72	0.13	0.78	0.70	0.19	0.21	0.59	0.63	0.66	1.53	0.65	0.61	0.77
Gd	3.42	5.92	4.26	8.42	6.20	3.21	4.55	3.74	7.02	3.08	2.77	3.18	0.45	2.81	2.35	0.55	0.62	2.40	2.39	2.45	7.30	2.60	2.29	2.56

Sample names with symbols *, † and ‡ stand for E-1, E-2 and E-3 rocks of Unit E, respectively.

Table 1. (Continued)

Sample	**Unit D**							**Unit E**													**Mafic Dyke**			
	KL50	KL78	KL79	96KL 58	96KL 113	96KL 115	96KL 118	KL59†	KL60*	KL62†	KL65†	KL66*	KL69‡	96KL 122†	96KL 124†	96KL 125‡	96KL 126‡	96KL 128†	96KL 129†	96KL 130†	KL32	KL81	KL83	96KL 108
Tb	0.63	1.00	0.76	1.61	1.19	0.64	0.83	0.64	1.27	0.52	0.45	0.54	0.07	0.50	0.43	0.10	0.12	0.44	0.44	0.45	1.28	0.46	0.41	0.49
Dy	4.43	6.88	5.22	10.5	7.89	4.52	5.59	4.21	8.78	3.40	2.94	3.67	0.51	2.95	2.73	0.67	0.76	2.86	2.79	2.83	8.85	3.18	2.84	3.16
Ho	1.00	1.54	1.14	2.31	1.73	1.02	1.25	0.88	1.95	0.71	0.63	0.81	0.11	0.62	0.58	0.15	0.17	0.61	0.61	0.62	1.92	0.70	0.62	0.68
Er	3.00	4.57	3.41	6.91	5.22	3.20	3.74	2.53	5.81	2.00	1.84	2.46	0.31	1.85	1.73	0.41	0.49	1.87	1.84	1.86	5.74	2.08	1.84	2.07
Tm	0.44	0.66	0.48	0.95	0.73	0.45	0.57	0.35	0.84	0.27	0.26	0.37	0.04	0.26	0.25	0.06	0.07	0.27	0.26	0.27	0.87	0.31	0.26	0.29
Yb	3.11	4.42	3.31	6.52	4.99	3.13	3.50	2.39	5.85	1.76	1.76	2.45	0.30	1.76	1.74	0.32	0.46	1.85	1.85	1.88	5.76	2.05	1.80	2.03
Lu	0.49	0.67	0.51	0.99	0.77	0.49	0.57	0.35	0.90	0.26	0.26	0.36	0.04	0.27	0.27	0.04	0.07	0.29	0.29	0.29	0.89	0.33	0.26	0.31
Hf	1.63	3.48	1.97	4.48	3.32	1.62	2.67	2.11	1.59	1.26	1.43	1.33	0.45	1.47	1.58	0.74	0.67	1.79	1.91	1.80	4.04	1.68	1.12	1.36
Ta	0.10	0.19	0.10	0.25	0.20	0.10	0.17	0.24	0.14	0.10	0.19	0.10	0.02	0.18	0.21	0.03	0.03	0.24	0.24	0.23	0.21	0.07	0.05	0.07
Th	0.31	0.27	0.20	0.41	0.35	0.42	0.54	0.52	0.60	0.31	1.16	0.44	0.30	1.24	0.96	0.36	0.40	1.02	1.06	1.01	0.32	1.27	0.08	0.14
U	0.53	0.36	0.23	0.22	0.15	0.23	0.89	0.33	0.43	0.24	0.46	1.10	0.22	0.40	0.47	0.37	1.09	0.69	0.48	0.48	0.20	0.68	0.22	0.07
ΣREE	39.2	66.1	46.2	91.7	69.8	37.2	48.9	46.2	76.4	34.4	46.7	32.1	6.63	45.7	38.4	8.54	8.38	35.3	36.3	38.6	79.7	28.6	23.8	29.5
Nb/Nb*	0.61	0.87	0.73	0.89	0.86	0.55	0.67	0.86	0.38	0.56	0.41	0.65	0.20	0.36	0.48	0.25	0.25	0.60	0.53	0.50	0.90	0.20	0.80	0.53

Table 2. Nd isotope composition of representative samples from the YVS

Sample	**Unit A**			**Unit B**			**Unit C**		**Unit D**			
	KL20	KL23	KL26	KL30	KL38	96KL112	KL44	96KL60	KL47	96KL58	96KL113	96KL115
$^{147}Sm/^{144}Nd$	0.204	0.206	0.199	0.166	0.167	0.162	0.163	0.152	0.223	0.203	0.203	0.199
$^{143}Nd/^{144}Nd$	0.513043±09	0.513025±06	0.513096±10	0.512837±09	0.512857±12	0.512843±12	0.512816±09	0.512775±08	0.513125±11	0.513070±06	0.513066±13	0.513020±09
$(^{143}Nd/^{144}Nd)i$	0.512361	0.512336	0.512432	0.512283	0.512298	0.512302	0.512271	0.512267	0.5123803	0.512392	0.512388	0.512355
εNd_T	7.4	6.9	8.8	5.9	6.2	6.3	5.7	5.6	7.8	8.0	7.9	7.3

Table 2 (Continued)

Sample	**Unit E （E-2）**							**Mafic Dyke**	
	KL59	KL65	96KL122	96KL124	96KL128	96KL129	96KL130	KL32	96KL108
$^{147}Sm/^{144}Nd$	0.193	0.160	0.161	0.157	0.169	0.169	0.160	0.213	0.204
$^{143}Nd/^{144}Nd$	0.513071±10	0.512748±15	0.512658±10	0.512604±08	0.512645±09	0.512675±09	0.512637±09	0.513071±10	0.512926±09
$(^{143}Nd/^{144}Nd)_i$	0.512425	0.512214	0.51212	0.512079	0.51208	0.51211	0.512102	0.512358	0.512244
εNd_T	8.7	4.6	2.7	1.9	1.9	2.5	2.4	7.4	5.1

εNd_T values are calculated at 510 Ma and reported relative to the evolution of the chondrite uniform reservoir (CHUR, today: $^{143}Nd/^{144}Nd = 0.512638$; $^{147}Sm/^{144}Nd = 0.1967$).
$\lambda_{Sm} = 6.54\times10^{-12}.yr^{-1}$, $\lambda_{Rb} = 1.42\times10^{-11}.yr^{-1}$.