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The Taitao ophiolite-granite complex, Chile: Emplacement of ridge-trench intersection oceanic lithosphere on land and the origin of calc-alkaline I-type granites

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The late Miocene – early Pliocene Taitao ophiolite is exposed ~30 km southeast of the Chile triple junction, where a spreading center of the Chile ridge system is subducting underneath the South America plate. This unique tectonic setting provides an excellent opportunity to study the emplacement mechanism of a ridge-trench intersection ophiolite and the complex magmatic interactions between the subducting ridge, overlying crust and sediments, and the mantle wedge. As a result of these interactions, several granitic plutons were formed contemporaneously with emplacement of the ophiolite. We review previous studies of the Taitao ophiolite and use new geochemical data to discuss the mechanism that formed juvenile magma of calc-alkaline I-type granites during ridge subduction. Our model implies that the magmas of the Taitao granites formed due to partial melting of hot oceanic crust adjacent to the subducting mid-oceanic ridge that has been contaminated by deep crustal material and/or metasomatized sub-arc mantle. The partial melting took place under garnet-free amphibolite conditions. The juvenile magmas then incorporated different amounts of subducted sediments and/or continental material to form the I-type granites with various compositions. Subduction of fracture zones played an important role in these processes.

Introduction

Ophiolites are regarded as pieces of ancient oceanic lithosphere

emplaced onto continental crust (Anonymous, 1972). Once it was presumed that this allochthonous material invariably formed at a mid-ocean spreading ridge. Today, most ophiolites are thought to form in supra-subduction zone (SSZ) settings in an island-arc or back-arc spreading environment (Miyashiro, 1973; Pearce and Robinson, 2010; Dilek and Furnes, 2011, 2014). However, even for extensively exposed and intensively studied examples like the Semail ophiolite, Oman, disagreement persists between ridge-origin scholars (Boudier *et al.*, 1996; MacLeod and Yaouancq, 2000; Miyashita *et al.*, 2003; Adachi and Miyashita, 2003; Boudier and Nicolas, 2011) and SSZ-origin scholars (Dilek and Flower, 2003; Dilek and Furnes, 2009; Dilek and Thy, 2009; Pearce and Robinson, 2010). One confounding factor is that all ophiolites must have experienced modification in SSZ environments just before their emplacement that may have obscured or obliterated primary features formed during their seafloor spreading origin (Tsuchiya *et al.*, 2013; Kanke and Takazawa, 2014; Yoshikawa *et al.*, 2015). Another factor may be that tectonic settings at the times of ophiolite formation and emplacement are ambiguous (Dilek and Robinson, 2003; Warren *et al.*, 2005).

The Taitao ophiolite, southern Chile (Fig. 1), is one of the few ophiolites for which the tectonic setting is clearly understood, because of its young age (< 6 Ma: Anma *et al.*, 2006) and the nearby presence of a subducting spreading-ridge segment of the Chile ridge system (Cande *et al.*, 1982, 1987; Cande and Leslie, 1986; Behrmann *et al.*, 1994). The plate configuration and plate kinematics before and after the oceanic lithosphere formed and was emplaced as an ophiolite can be well reconstructed from various data.

The thermal history of an ophiolite can be prolonged, especially one that formed at a spreading ridge far from land in a large ocean, migrated toward a continental margin and emplaced in a SSZ environment. However, in the case of the Semail ophiolite, its generation and emplacement took place in a short period: trondhjemitic in the Semail ophiolite yielded zircon U-Pb ages of ~95.3 Ma, whereas amphibolite in the metamorphic sole yielded zircon U-Pb ages of ~94.5 Ma (Warren *et al.*, 2005). Tsuchiya *et al.* (2013)

also reported similar ages (~100 Ma) in both the quartz diorite that formed at the ridge axis and the late-stage tonalite that represents arc-like magmatism. However, progressive metamorphism continued after emplacement as obduction and subduction proceeded, and the peak metamorphism that resulted in eclogite formation took place at ~79 Ma (Warren *et al.*, 2005). Exhumation of blueschist in the southern part of the Semail ophiolite resulted in cooling of these rocks to the closing temperature of the zircon fission-track system (~250°C) at ~66–70 Ma (Yamato *et al.*, 2007). Jacobs *et al.* (2015) reported apatite fission-track and zircon and apatite (U-Th)/He ages of igneous rocks and the metamorphic sole from the section of the Semail ophiolite exposed in the UAE. They showed that the plutonic rocks were cooled to the closing temperatures of two of these three systems (zircon (U-Th)/He and apatite fission track) within 10 to 30 m (84.5–64.7 Ma and 87–62 Ma, respectively) of crust formation, whereas exhumation of metamorphic rocks took a longer period of time (until 63–31 Ma according to apatite fission-track ages). Both the plutonic and metamorphic rocks were cooled to the closing temperature of the apatite (U-Th)/He system in Miocene time (Jacobs *et al.*, 2015). Thus, during its prolonged thermal history, various modification processes could have affected the Semail ophiolite. This history can be compared with that, which is much more precisely determined for the Taitao ophiolite. It took only 1 my from the formation of the oceanic lithosphere through its emplacement and exhumation, as detailed in this paper (Fig. 1d).

The formation, emplacement and exhumation processes of ophiolites generate a wide range of geochemical variations (Flower and Dilek, 2003; Balestro *et al.*, 2015; Fareeduddin and Dilek, 2015, Saccabu *et al.*, 2015). In the case of the Semail ophiolite, the plutonic rocks range in chemical composition from ~48 wt.% SiO₂ in oceanic massive gabbro to ~75 wt.% SiO₂ in late-stage tonalite (Tsuchiya *et al.*, 2013). The chemical variation of the Taitao ophiolite-granite complex is similar, ranging from 48 to 74 wt.% SiO₂ for the plutonic rocks. However, in the Semail ophiolite, the late-stage felsic plutonic rocks are minor components. In contrast, in the Taitao ophiolite these rocks occupy as much as 10% of the exposure (Fig. 1c).

The advantage of studying the Taitao ophiolite and its related igneous rocks is that this area has a complete record of the magmatic history and the full range of chemical variation seen in ophiolites, with better tectonic and geochronological constraints than any other example in the world. Evidence of every process, from crustal formation to ophiolite emplacement, that took place over the span of 1 my was “frozen” into the geological record of this area. The Taitao ophiolite provides an excellent opportunity to study the emplacement mechanism of a ridge-trench intersection ophiolite, with its complex magmatic interactions between the subducting ridge, the overlying crust and sediments, and the mantle. Among these interactions, the mechanism that created voluminous intermediate to felsic magma in this area is of fundamental importance for understanding the processes that create continental crust (Kon *et al.*, 2013). In this study, we used chemical compositions and the Sr-Nd isotope system to ascertain the magmatic interactions during the formation and emplacement of the Taitao ophiolite.

Tectonic setting of the Taitao ophiolite and related rocks

The Taitao ophiolite is ~30 km southeast of the Chile triple

junction, where a segment of the spreading Chile ridge system is subducting underneath the South America plate off the Taitao Peninsula, the westernmost promontory of the Chilean coast. The Chile ridge system, which separates the Nazca plate to the north and the Antarctica plate to the south, has a central axis trending north-northwest that consists of several segments separated by transform fracture zones trending east-northeast (Fig. 1a). The convergent rates of the Nazca and Antarctica plates with respect to the South America plate are 9 cm/y and 2 cm/y, respectively (Herron *et al.*, 1981; Cande *et al.* 1982; Tebbens and Cande, 1997; Tebbens *et al.*, 1997). The current plate configuration and plate kinematic data (Cande and Leslie 1986; Breitspecher and Thorkelson, 2009) allow us to deduce that short segments of this spreading center (< 100 km long) began subducting off the Taitao Peninsula at ~6 Ma and ~3 Ma. The formation and emplacement of the Taitao ophiolite are related to the first of these subduction events (Anma *et al.*, 2006).

The study area, in the westernmost part of the Taitao Peninsula, exposes four major rock suites (Fig. 1b): the Chonos metamorphic complex, the Chile Margin Unit (CMU), the Taitao ophiolite, and the Taitao granites (Forsythe *et al.*, 1986; Nelson *et al.*, 1993; Bourgois *et al.*, 1992; 1993; Guivel *et al.*, 1999). The Chonos metamorphic rocks, which mostly originated from siliciclastic and volcanoclastic protoliths, form the basement of the study area. Thomson and Herve (2002) determined Late Permian to Jurassic (258 to 207 Ma) fission-track and U-Pb ages from zircons in the basement rocks in the forearc region.

Chile Margin Unit

The CMU is a mixture of lava, pyroclastic rocks and semi-consolidated sediments (Fig. 1b, c). It was first defined by Bourgois *et al.* (1993) to distinguish it from the Main Volcanic Unit (see next section). The CMU directly overlies the pre-Jurassic basement, is of lower metamorphic grade, and appears to have deposited in the forearc area of the Chilean margin. Whole-rock K-Ar ages for the CMU range from 4.4 to 2.5 Ma (Mpodozis *et al.*, 1985). Volcanic rocks of the CMU have rather enriched compositions comparing with normal mid-ocean ridge basalt (N-MORB) (Lagabriele *et al.*, 1994; Le Moigne *et al.*, 1996; Guivel *et al.*, 1999).

Anma and Orihashi (2013) dredged pyroclastic rocks resembling those of the CMU from the South Taitao Ridge, offshore from the Taitao ophiolite (Fig. 1b), and argued that the magmatic activity of the CMU started there at 5.2 Ma, migrated eastward at a rate of 2 to 5 cm/y, forming the CMU (Fig. 1c, d), and continued to the Pan de Azucar region (PA in Fig. 1b) where volcanic activity took place at 4.4 Ma based on zircon U-Pb data. All of these rocks contained inherited zircons that originated from subducted sediments. Anma and Orihashi (2013) suggested that the magmatic activity of the CMU is related to the subduction of a fracture zone separating segments of the Chile ridge system.

Taitao ophiolite

The Taitao ophiolite exposes the complete sequence of a Penrose-type oceanic lithosphere (Anonymous, 1972) (Fig. 1c). The top part, in the north, is represented by pillow lava and sheet flows interbedded with clastic deposits, which we refer to as the Main Volcanic Unit (MVU) after Bourgois *et al.* (1993). To the south are sheeted dikes, gabbros and ultramafic rocks (Forsythe *et al.* 1986; Kaeding *et al.*,

Figure 1. (a) Tectonics of the Chile Ridge subduction zone. (b) Distribution of the Taitao ophiolite and related igneous rocks together with detailed bathymetry. (c) Bedrock geology of the study area, modified from Anma et al. (2006), showing distribution of sampling sites for this study. MUV: Main Volcanic Unit, CMU: Chile Margin Unit, EC: Estero Cono Pluton, SH: Seno Hoppner pluton, BB: Bahia Barrientos intrusion, CR: Cabo Raper pluton, TM: Tres Montes pluton. (d) U-Pb age distribution of samples in this study.

1990; Nelson *et al.*, 1993; Bourgois *et al.*, 1993). The ophiolite shows evidence of hydrothermal alteration that is typical of ocean-floor metamorphism (Shibuya *et al.*, 2007). The ultramafic rocks provide evidence of at least two stages of melting, one that took place at ~1.6 Ga and the other of recent age related to magmatism in the Chile ridge (Schulte *et al.*, 2009). The gabbros have compositions similar to N-MORB, whereas the pillow basalts of the MVU have enriched compositions (Lagabriele *et al.*, 1994; Le Moigne *et al.*, 1996; Guivel *et al.*, 1999) and isotopic compositions (Kaeding *et al.*, 1990) that suggest crustal contamination. Because of this compositional difference, the plutonic section has been referred to separately as the Bahia Barrientos ophiolite (Lagabriele *et al.*, 1994; Le Moigne *et al.*, 1996; Guivel *et al.*, 1999). However, Schulte *et al.* (2009) reported that the ultramafic rocks and gabbros of the plutonic section also have a range of chemical and isotopic compositions that could be explained in terms of seawater contamination (for Sr isotopic ratios) and incorporation of sediment and basement rocks into the melt source of the Taitao ophiolite. Anma *et al.* (2006) explained the enriched composition of the MVU in terms of alteration along fracture zones or sub-segment boundaries, as reported previously from the Chile ridge (Klein and Karsten, 1995; Karsten *et al.*, 1996; Sherman *et al.*, 1997; Strum *et al.*, 1999, 2000), and considered the MVU to be a member of the Taitao ophiolite.

Previously reported U-Pb ages of the gabbros and sheeted dikes are ~5.6 Ma and ~5.2 Ma, respectively (Fig. 1c, d; Anma *et al.*, 2006). The sheeted dikes were thermally metamorphosed by the intrusion of the 5.1 Ma Seno Hoppner pluton (SH in Fig. 1c) (Shibuya *et al.*, 2007). Thus, the center of magmatic activities that formed the ophiolite migrated from south to north as a segment of the Chile ridge subducted obliquely against the Chilean continental margin (Anma *et al.*, 2006, 2009). Anma *et al.* (2009) reported U-Pb ages of zircons in clastic sediments in the MVU, of which 19 grains out of 24 yielded ages younger than 5.6 Ma, and concluded that their deposition postdated ~4.9 Ma. Veloso *et al.* (2007) argued that these clastic materials were derived from the south-southeast based on sedimentary structures and anisotropy of magnetic susceptibility fabrics. Thus, the post-5.6 Ma clastic zircons must have been derived from the plutonic section of the Taitao ophiolite and the Taitao granites, and by the time of deposition, most of these plutonic rocks must have been exposed subaerially (Anma *et al.*, 2009). Zircon fission-track ages from gabbros of the ophiolite that coincided with the U-Pb ages also attest to rapid exhumation and cooling of the plutonic section (Anma *et al.*, 2006).

The gabbros continued to fold even below their Curie temperature, as evidenced by macroscopic folding and Z-shaped demagnetization paths in the layered gabbros (Anma *et al.*, 2005; Veloso *et al.*, 2005), whereas the sheeted dike complex and the MVU rotated as rigid blocks as deformation proceeded (Veloso *et al.*, 2005, 2009). It was the eastern part of the subducted segment of the Chile ridge that was emplaced as the plutonic section of the Taitao ophiolite (Veloso *et al.*, 2005; Anma *et al.*, 2005, 2006). As the spreading center subducted obliquely, the center of magmatic activities migrated northward. During the time the midsection of the MVU was deposited, the plutonic section must have been subaerially exposed, supplying sediment to the subducting central valley (Anma *et al.*, 2006, 2009).

Taitao granites

The term “Taitao granites” collectively denotes the felsic plutonic rocks distributed around the Taitao ophiolite. These four plutons and

a small body of granitic rocks exposed along the Bahia Barrientos (Fig. 1c) have zircon U-Pb ages that range from 5.7 Ma to 3.9 Ma (Fig. 1d) and compositions that range from trondhjemitic to granitic and tonalitic to granodioritic (Fig. 2b; Anma *et al.*, 2009). The plutons are the Tres Montes pluton (TM in Fig. 1c; 5.7 Ma) and Cabo Raper pluton (CR; 3.9 Ma) to the south of the Taitao ophiolite and the Seno Hoppner (SH) and Estero Cono (EC) plutons (5.2–5.1 Ma) to the northeast (Fig. 1). The Bahia Barrientos granite (BB in Fig. 1c; 4.9 Ma), intruded along the edge of the CMU, is thought to be related to CMU magmatism on the basis of its age (Anma and Orihashi, 2013).

The origin of the Taitao granites is controversial. Kaeding *et al.* (1990) were the first to recognize that the Cabo Raper pluton has a tonalitic to granodioritic composition with calc-alkaline I-type affinity, whereas the Seno Hoppner pluton has characteristics of S-type affinity. In contrast, Bourgois *et al.* (1996) suggested that the Cabo Raper pluton is a heavy rare earth element (HREE)-depleted trondhjemitic-tonalite-granodiorite (TTG) that formed by slab melting under conditions of the amphibolite-eclogite transition. Guivel *et al.* (1999) reported that the Seno Hoppner pluton differs from the Cabo Raper and Seno Hoppner plutons in having a trondhjemitic-granite composition and characteristics, typical of the calc-alkaline series. They also suggested that the Cabo Raper and Tres Montes plutons formed by slab melting under amphibolite-eclogite transition conditions.

Kon *et al.* (2013) recognized hornblende-bearing and hornblende-free members in the Seno Hoppner and Cabo Raper plutons. The hornblende-free members have trondhjemitic to granitic compositions whereas the hornblende-bearing members plot in the tonalitic to granodioritic compositional field along with the hornblende-bearing Cabo Raper pluton. The hornblende-free members have a pronounced negative Eu anomaly whereas the hornblende-bearing members have none or a small Eu anomaly. On the basis of trace-element compositions and tectonic constraints, they suggested that these granites were generated by partial melting of subducted oceanic crust with garnet-free-amphibolite conditions at depths shallower than 30 km.

The Tres Montes granodiorite includes numerous inherited zircons (7 of 15) with an age distribution similar to those from the CMU, whereas inherited zircons are rare in the Seno Hoppner and Cabo Raper granites (Herve *et al.*, 2003; Anma *et al.*, 2009). Thus, it has been proposed that the Tres Montes pluton originated with a strong influence from subducted sediments, whereas the Seno Hoppner and Cabo Raper plutons formed from juvenile granitic magmas (Anma *et al.*, 2009; Kon *et al.*, 2013). Kaeding *et al.* (1990) argued from Sr-Nd isotopic compositions that the Seno Hoppner and Cabo Raper granites formed by assimilation of basaltic magmas with 5–15% forearc sediments or metamorphic basement rocks.

Scope and analytical methods

To investigate the complicated magmatic processes resulting from the interaction between the spreading ridge and the subduction zone, and to test previous hypotheses, we conducted Sr-Nd isotopic analyses of this ridge-trench intersection ophiolite (Wakabayashi and Dilek, 2003; Dilek and Furnes, 2011, 2014) and its related rocks. The analyzed samples are described in Table 1 and their locations shown by circles in Figure 1c. Apart from 5 ophiolite and 10 granite samples, we measured two mafic volcanic rocks from the CMU and two mafic

Table 1. Geological unit and rock description of this study

Geological unit		Unit description	Sample ID	Age (Ma)*	Rock description
Mafic dikes in granite plutons		Estero Cono (EC) dike	TPD185	<5.2	basaltic dike intruded into the Seno Hoppner pluton
		Tres Montes (TM) dike	TPA009	<5.7	syn-plutonic basaltic dike intruded into the Tres Montes pluton
Chile Margin Unit (CMU)		Lavas and clastic rocks	TPB174	4.8 ~ 4.6	sheet flow: doleritic basalt
			TPB175		glass-rich hyaloclastite
Taitao Ophiolite	Main Volcanic Unit (MVU)	Lavas and clastic rocks	TPB345	< 5.2	basaltic pillow lava
			TPB347		basaltic massive sheet flow
	Estero Cono dike complex	Sheeted dike complex	TPD208	< 5.2	dolerite to basalt, metamorphosed by the intrusion of the Seno Hoppner pluton
	Plutonic complex	Gabbros and associated doleritic dikes	TPB020	5.6	medium-grained gabbro (massive)
			TPB024		fine-grained gabbro (mesocumulate)
	Taitao Granites	Tres Montes (TM) pluton	TPA007	5.7	hybrid rocks formed due to syn-plutonic mafic dike intrusion
			TPA014		fine to medium-grained biotite hornblende tonalite and granodiorite
		Seno Hoppner (SH) pluton	TPD110	5.2 ~ 5.1	medium-grained hornblende biotite granite, thermal aureole contact in the ophiolite, mafic dike in the northern part.
			TPD115		
			TPD172		
		Estero Cono (EC) pluton	TPD168	5.1	fine to medium-grained biotite (hornblende) granodiorite
			TPD169		
		Bahia Barrientos (BB) pluton	TPB243	4.9	fine to medium-grained biotite tonalite and trondhjemite
			TPB246		
Others	Sediments	trench-slope sediments	TPB338	< 4.9	turbiditic sediment in the MVU
			TPD095		felsic igneous rock (ryolitic)?
	Basement	meta-igneous rocks	TPB240	Jurassic**	sandstone/mudstone alternation
			TPB335		consolidated silici-clastic sediments
			TPD099		pelitic schist

* Age data from Anma et al. (2006, 2009) and Anma and Orihashi (2013), ** from Thomson et al. (2002)

dikes intruded into the granite plutons. One of the granite samples (TPA007) was a hybrid rock mixed with a syn-plutonic basaltic dike. We also measured samples of basement rocks and turbiditic sediments to estimate the respective influences of contamination from Jurassic wall rocks and sediments.

Major- and trace-element concentrations were measured using an X-ray fluorescence spectrometer and a laser ablation inductively coupled plasma mass spectrometer at the Earthquake Research Institute using conventional methods (Tani *et al.*, 2002; Orihashi and Hirata, 2003) (Table 2). The analytical results in Table 2 also include data for some granites reported in Anma *et al.* (2009). For whole-rock Sr and Nd isotopic analyses, approximately 100 mg of powdered sample was completely dissolved in a mixture of 38% HF (1.0 ml), 68% HNO₃ (0.7 ml), and 70% HClO₄ (0.3 ml) at 140°C for 24 h using a hot plate. After evaporation, the residues were dissolved in 2 ml of 6M HCl and heated at 100°C for 8 h. Sr and REEs were separated by conventional ion-exchange chromatography. Separation of Nd from REEs was facilitated using Ln Resin (Eichrom Tec. Inc).

The purified Sr was loaded with a Ta activator on preconditioned W filaments and was measured in single filament mode. Nd isotope ratios were measured in Ta-Re double filaments mode. Sr and Nd isotopic ratios were determined using a TRITON mass spectrometer

(Thermo Fisher Scientific Co.) at the Research Institute for Humanity and Nature. Analytical precisions of isotopic ratio measurements are given as $\pm 2\sigma$ standard errors. The measured $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, respectively. The $^{87}\text{Sr}/^{88}\text{Sr}$ value of the NIST SRM 987 reference material throughout the analyses was 0.710264 ± 0.000014 (2σ , $n = 14$). The $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of the La Jolla reference material was 0.511863 ± 0.000015 (2σ , $n = 14$). All samples were recalculated with reference to the values of reference materials. The results are listed in Table 3. We calculated the initial value of Sr and Nd isotope ratios (designated by the suffix *i*) using zircon U-Pb ages reported by Anma *et al.* (2006, 2009). The basement rock isotopic ratios were calculated assuming the age of incorporation to be 5.5 Ma.

Analytical results

The analytical results for major and trace elements are shown in Figs. 2 to 6, and results for isotopic compositions are shown in Figs. 7 and 8. Our analyses in general are in good agreement with previously reported results.

The mafic rocks had a basaltic to basaltic-andesitic composition (Fig. 2a) that could be divided into N-type to E-type (enriched) MORB

Table 2. Major and trace element compositions for the Taitao Ophiolite, the Taitao granites, sediments and basement rocks

Locality	Taitao Ophiolite						CMU		mafic dikes in granites		Taitao Granites										Sediment		Basement			
	MVU		EC	gabbro		TPB	TPB	TPD	TPA	TM		SH		EC		BB		CR	TPB	TPB	TPD	TPD	TPB	TPB	TPD	TPD
	TPB	TPB		TPB	TPB					TPA	TPA	TPD	TPD	TPD	TPD	TPD	TPD									
Sample No.	345	347	208	020	024	174	175	185	009	007	014	110	115	172	168	169	243	246	236	338	335	240	095	099		
SiO ₂ *	55.03	51.08	50.99	48.01	53.33	53.81	52.03	51.08	56.38	59.25	65.84	73.96	74.30	73.93	69.10	68.11	69.31	69.33	67.92	62.87	69.68	69.08	72.31	70.30		
TiO ₂	0.80	1.21	1.11	0.66	0.63	1.07	1.09	1.16	1.08	0.89	0.89	0.23	0.20	0.20	0.44	0.49	0.40	0.43	0.45	0.70	0.60	0.46	0.51	0.72		
Al ₂ O ₃	17.65	19.00	17.41	20.70	15.30	16.25	15.64	15.88	17.23	16.51	15.96	13.16	13.04	13.06	14.82	14.94	16.04	15.43	15.48	14.67	14.11	10.79	13.51	14.33		
Fe ₂ O ₃	5.67	6.54	8.34	6.46	8.70	7.55	8.02	9.41	7.84	6.71	5.64	2.25	2.16	2.29	2.78	3.21	2.51	3.05	3.01	5.42	4.55	6.78	3.55	4.96		
MnO	0.10	0.10	0.08	0.10	0.14	0.12	0.14	0.11	0.14	0.12	0.08	0.03	0.03	0.03	0.05	0.05	0.04	0.11	0.05	0.10	0.07	0.04	0.06	0.07		
MgO	6.41	6.55	6.47	11.46	7.41	6.21	9.43	7.74	4.50	4.27	2.18	0.34	0.27	0.20	1.61	1.83	1.78	1.54	1.96	2.42	1.70	1.64	1.30	1.85		
CaO	8.64	9.22	11.67	10.16	10.96	7.62	5.72	11.18	8.37	6.60	3.54	1.14	0.97	1.08	3.44	3.58	3.00	2.75	3.64	5.18	2.92	1.69	2.34	0.98		
Na ₂ O	4.15	4.31	3.07	2.24	3.65	5.77	3.04	2.81	3.58	4.10	3.94	4.59	4.62	4.71	3.95	4.09	4.59	4.40	4.20	3.51	2.71	2.55	3.57	2.57		
K ₂ O	0.52	0.38	0.10	0.04	0.04	0.37	1.16	0.52	0.52	1.05	2.09	3.12	3.26	3.11	2.28	2.05	1.51	1.75	2.06	1.55	2.33	1.19	2.05	2.42		
P ₂ O ₅	0.11	0.17	0.12	0.04	0.05	0.13	0.12	0.11	0.14	0.11	0.20	0.04	0.03	0.04	0.08	0.09	0.10	0.09	0.11	0.17	0.14	0.09	0.14	0.17		
Total	99.07	98.55	99.34	99.86	100.20	98.88	96.38	100.00	99.78	99.61	100.35	98.87	98.87	98.64	98.55	98.44	99.28	98.87	98.88	96.58	98.80	94.30	99.32	98.35		
Sc**	26	30	33	15	44	30	30	39	28	27	16	5	6	6	10	10	7	7	10	17	13	9	9	10		
V	143	196	203	108	209	194	188	223	186	162	128	15	9	11	51	54	43	44	59	119	97	67	68	86		
Co	38	37	49	64	72	42	40	66	61	60	74	107	112	102	84	76	73	69	79	44	44	88	61	37		
Zn	51	52	109	43	31	56	63	16	101	72	68	35	31	29	32	33	40	161	42	64	37	27	68	74		
Ga	16	16	15	13	16	15	15	15	18	16	20	16	15	15	15	16	18	18	18	18	17	14	17	19		
Rb	14	10	3.9	0.4	1.0	14	55	25	19	38	89	132	141	137	72	68	50	41	77	45	77	43	86	120		
Sr	228	239	238	125	122	170	295	142	235	271	212	55	51	53	169	169	268	188	271	166	220	174	337	112		
Y	17	25	25	12	25	26	26	25	22	29	29	36	37	39	20	23	13	19	17	23	26	17	28	32		
Zr	83	105	83	35	42	105	107	77	100	155	215	217	219	223	161	168	129	160	157	170	157	153	235	286		
Nb	4	6	4	0.8	0.6	5	5	3	6	7	12	10	8	9	6	7	4	7	8	7	9	8	11	16		
Cs	1.6	1.4	2.9	1.6	1.8	1.6	11.5	3.9	1.8	1.0	5.2	9.2	10.6	6.4	8.1	3.6	3.5	1.4	3.4	3.7	8.3	4.0	5.6	6.4		
Ba	95	90	41	10	16	66	109	60	151	229	424	475	470	459	525	465	285	489	377	204	479	203	441	422		
La	6.25	7.45	3.45	0.55	1.0	7.69	6.76	3.43	11.6	15.5	25.3	31.9	39.1	37.4	14.7	12.1	13.0	8.66	18.0	18.9	22.7	20.2	38.3	44.3		
Ce	13.4	17.1	9.16	2.68	3.8	16.9	17.6	9.76	23.4	34.8	48.9	71.3	84.6	71.5	34.0	27.8	31.0	20.1	37.7	41.1	46.0	40.6	80.4	89.5		
Pr	1.70	2.20	1.34	0.49	0.79	2.15	2.28	1.47	2.97	4.05	5.68	7.66	8.25	7.67	3.55	3.04	3.13	2.02	4.01	4.66	5.20	4.39	8.27	9.91		
Nd	7.48	10.1	6.83	2.74	4.9	11.0	10.2	8.06	13.3	17.5	24.5	28.1	29.4	28.2	13.6	12.7	12.1	7.51	14.8	18.8	19.7	17.3	30.2	35.0		
Sm	1.89	3.09	2.33	0.98	2.21	3.03	2.97	2.74	3.07	4.34	5.06	6.21	5.52	5.75	2.94	2.77	2.54	1.41	3.03	4.01	4.08	3.52	5.65	7.28		
Eu	0.95	1.25	0.89	0.67	0.94	1.09	1.06	1.14	1.38	1.25	1.30	0.68	0.77	0.78	0.66	0.67	0.77	0.34	1.00	1.05	0.90	0.88	1.10	1.55		
Gd	2.28	3.05	2.89	1.41	2.55	3.45	3.35	3.73	3.43	4.04	4.52	5.86	6.33	5.71	3.26	2.67	1.91	1.44	2.33	3.32	4.07	3.06	5.22	5.70		
Tb	0.41	0.56	0.50	0.23	0.50	0.55	0.62	0.60	0.58	0.65	0.81	0.98	0.99	0.95	0.44	0.46	0.30	0.26	0.39	0.56	0.59	0.42	0.74	0.97		
Dy	2.55	3.21	3.27	1.59	3.53	3.78	3.73	3.47	3.52	4.64	4.99	6.18	5.54	6.06	2.80	3.07	1.81	1.31	2.41	3.31	2.94	2.75	3.87	5.50		
Ho	0.50	0.74	0.71	0.34	0.82	0.88	0.80	0.80	0.73	0.93	1.05	1.25	1.23	1.43	0.55	0.65	0.43	0.30	0.57	0.70	0.72	0.51	0.82	1.12		
Er	1.47	1.98	2.06	0.83	2.12	2.30	2.24	2.27	1.99	2.64	2.78	3.45	3.20	3.96	1.63	1.77	1.09	0.78	1.51	1.85	2.07	1.45	1.98	3.35		
Tm	0.22	0.27	0.30	0.14	0.33	0.36	0.32	0.33	0.32	0.39	0.41	0.55	0.41	0.65	0.27	0.29	0.17	0.12	0.23	0.29	0.30	0.24	0.32	0.39		
Yb	1.62	2.08	2.10	1.08	2.26	2.24	2.27	2.32	2.04	2.77	2.93	4.06	3.56	4.37	1.88	1.87	1.21	0.82	1.32	1.92	2.27	1.60	2.27	3.04		
Lu	0.21	0.29	0.28	0.16	0.30	0.32	0.34	0.33	0.28	0.40	0.47	0.56	0.50	0.66	0.25	0.35	0.15	0.14	0.21	0.31	0.43	0.25	0.32	0.46		
Hf	2.02	2.21	1.73	0.79	1.30	2.43	2.36	1.88	2.35	3.66	6.35	6.71	6.64	6.43	4.11	3.79	3.40	2.04	3.58	4.18	3.58	3.90	5.39	8.29		
Ta	0.24	0.28	0.21	0.03	0.04	0.25	0.26	0.15	0.34	0.43	0.76	0.74	0.81	0.81	0.44	0.46	0.45	0.25	0.57	0.43	0.61	0.53	0.70	0.99		
Pb	2.98	1.46	1.89	n.d.	0.26	2.19	2.11	0.37	17.5	5.10	15.7	11.2	13.6	12.1	12.4	10.7	10.8	9.17	10.7	10.3	7.34	6.35	18.5	17.7		
Th	1.01	0.90	0.53	0.01	<0.04	1.61	1.37	0.51	3.53	4.74	8.41	11.9	11.3	15.4	7.28	5.94	5.39	3.81	6.56	6.17	7.00	5.94	12.3	13.1		
U	0.42	0.37	0.23	0.01	<0.01	0.52	0.45	0.19	0.64	1.26	1.98	3.52	4.67	3.56	3.18	3.11	2.54	1.47	2.04	1.89	2.35	1.64	4.21	3.91		
Mg#***	69.14	66.49	60.59	77.85	62.79	61.97	69.97	61.97	53.21	55.77	43.37	23.04	19.85	14.75	53.43	53.04	58.42	50.01	56.34	46.94	42.54	32.40	42.05	42.50		
ASI****	0.76	0.78	0.66	0.93	0.59	0.68	0.94	0.62	0.80	0.83	1.05	1.01	1.01	1.00	0.97	0.97	1.10	1.09	0.98	0.87	1.15	1.26	1.09	1.66		
T(Zr)	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	801	811	813	814	774	775	767	786	770	n.d.	n.d.	n.d.	n.d.	n.d.		

*Compositions are quoted in weight %. Total Fe as Fe₂O₃. **Concentrations are in weight ppm. ***Mg#: 100× molar [Mg/(Mg+Fe)]. ****ASI: molar Al₂O₃/(Na₂O+K₂O+CaO). The sample localities are shown in Fig. 1.

Table 3. Sr and Nd isotopic ratios for the Taitao Ophiolite, the Taitao granites, sediments, and basement rocks

Geological Unit			Sample ID	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}^*$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}_i$	ϵNd_i^{**}
Mafic igneous rocks	Taitao Ophiolite	MVU	TPB345	0.703523 (± 5)	0.173	0.703510	0.512957 (± 4)	0.1526	0.512952	6.25
			TPB347	0.703408 (± 5)	0.127	0.703399	0.513038 (± 4)	0.1859	0.513031	7.81
		EC-SDC	TPD208	0.704236 (± 5)	0.047	0.704233	0.51306 (± 3)	0.2067	0.513053	8.22
		Gabbro	TPB020	0.702808 (± 5)	0.009	0.702807	0.513162 (± 5)	0.2168	0.513154	10.21
			TPB024	0.703175 (± 5)	0.024	0.703173	0.513156 (± 3)	0.2699	0.513146	10.05
	Mafic Diike	CMU	TPB174	0.705274 (± 6)	0.230	0.705257	0.512909 (± 3)	0.1661	0.512903	5.31
			TPB175	0.706713 (± 4)	0.541	0.706674	0.512934 (± 5)	0.1754	0.512928	5.79
		EC-dike	TPD185	0.703181 (± 4)	0.503	0.703145	0.513119 (± 4)	0.2054	0.513112	9.38
Felsic igneous rocks	Taitao Granites	TM	TPA007	0.708212 (± 3)	0.402	0.708179	0.512448 (± 3)	0.1500	0.512443	-3.68
			TPA014	0.707458 (± 5)	1.214	0.707360	0.512705 (± 19)	0.1251	0.512701	1.36
		SH	TPD110	0.704587 (± 4)	6.992	0.704081	0.512839 (± 3)	0.1339	0.512834	3.96
			TPD115	0.704795 (± 3)	7.998	0.704216	0.512828 (± 3)	0.1136	0.512824	3.76
			TPD172	0.704821 (± 3)	7.493	0.704278	0.512797 (± 5)	0.1233	0.512793	3.15
		EC	TPD168	0.705430 (± 3)	1.242	0.705340	0.512661 (± 4)	0.1305	0.512656	0.49
			TPD169	0.705440 (± 3)	1.164	0.705356	0.512676 (± 6)	0.1323	0.512671	0.79
		BB	TPB243	0.705155 (± 3)	0.542	0.705117	0.512756 (± 4)	0.1273	0.512751	2.35
			TPB246	0.705404 (± 4)	0.630	0.705360	0.512667 (± 9)	0.1135	0.512663	0.62
		CR	TPB236	0.704972 (± 5)	0.820	0.704927	0.512714 (± 4)	0.1238	0.512710	1.53
Contaminants	Sediment		TPB338	0.706549 (± 3)	0.781	0.706495	0.512102 (± 3)	0.1291	0.512097	-10.41
	Basement		TPB240	0.707882 (± 5)	0.723	0.707830	0.512436 (± 4)	0.1235	0.512432	-3.89
			TPB335	0.707408 (± 5)	1.014	0.707335	0.512469 (± 4)	0.1250	0.512465	-3.25
			TPD095	0.713521 (± 5)	0.741	0.713467	0.512175 (± 3)	0.1129	0.512171	-8.97
			TPD099	0.720999 (± 5)	3.103	0.720774	0.512186 (± 3)	0.1260	0.512181	-8.77

* Initial ratios (i) for the Sr and Nd isotopic ratios are calculated with the zircon age data from Anma *et al.*, (2006, 2009). ** ϵNd_i : deviation in parts per 10^4 from bulk earth values at their initial age (zircon age). Basement rocks are calculated with 5.1 Ma for comparison

(Figs. 4a and 5a). The gabbros had lower TiO_2 , K_2O and P_2O_5 contents (Fig. 3) than other mafic rocks and had N-MORB signatures, being depleted in large ion lithophile elements (LILEs) such as Rb, Th and U (Fig. 4a). The other mafic rocks were relatively enriched in LILEs and slightly depleted in Ba, Nb and Ta. The mafic dike sample from the Tres Montes pluton (TM-dike in the figures and TPA009 in the tables) showed notable enrichment in Pb and Th, possibly due to the contamination of crustal rocks (Fig. 4a). Gabbros (Fig. 5a) had depleted light REE (LREE) contents with a positive Eu anomaly that implies accumulation of Ca-plagioclase. MVU samples and the TM-dike had less distinct positive Eu anomalies. Together with CMU samples, they had slightly LREE-enriched features. In the Estero Cono pluton, the samples from an isolated mafic dike (EC-dike; TPD185) and the sheeted dike complex (EC-SDC; TPD208) had a rather flat REE pattern. However, EC-SDC was slightly depleted in Rb, Ba and K and enriched in Pb compared to EC-dike, most likely due to alteration by the intrusion of the Seno Hoppner–Estero Cono pluton (Fig. 4a).

The gabbros and the EC-dike had the most primitive isotopic ratios among the mafic rocks we analyzed, ranging from 0.70281 to 0.70715 for $^{87}\text{Sr}/^{86}\text{Sr}_i$ and from 10.21 to -0.49 for ϵNd_i values (Figs. 7 and 8). Compared to the gabbros, the MVU samples had slightly lower Nd isotopic ratios, whereas EC-SDC had a slightly higher $^{87}\text{Sr}/^{86}\text{Sr}_i$. The samples from CMU and TM-dike had higher Sr and lower Nd isotopic values than the other mafic rocks.

The Taitao granites plot in three distinct fields, except for one hybrid rock with a tonalitic composition (Fig. 2b). Each pluton has

relatively homogeneous major-element compositions. The samples from the Seno Hoppner pluton plot in the field for true granite (Fig. 2b) and coincide with the hornblende-free granitoids of the Seno Hoppner–Estero Cono pluton reported by Kon *et al.* (2013) (Fig. 4b, 5b). They have the highest silica contents among the Taitao granites (Fig. 3) and low MgO and CaO contents with low $\text{Na}_2\text{O}/\text{K}_2\text{O}$ and $\text{CaO}/\text{Na}_2\text{O}$ ratios (1.4–1.5 and 0.2, respectively) (Table 2). The samples from the Estero Cono and Cabo Raper plutons plot on the boundary between the tonalite and granodiorite fields, coinciding with the hornblende-bearing granitoids of the Seno Hoppner–Estero Cono and Cabo Raper plutons reported by Kon *et al.* (2013) (Fig. 4b, 5b). The Tres Montes granodiorite plot in the same position as the hornblende-bearing granitoids (Fig. 2b), but has lower silica contents (Fig. 3). The samples from the Bahia Barrientos pluton have distinctive compositions, plotting on the boundary between the tonalite and trondhjemite fields (Fig. 2b). These samples have lower K_2O contents and higher aluminum saturation index (ASI) values (Table 2) than other granites.

In other respects, the Estero Cono, Cabo Raper and Bahia Barrientos plutons have similar major-element compositions with slight variations in Na_2O and K_2O contents. The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios range from 1.7 to 3.0, and the $\text{CaO}/\text{Na}_2\text{O}$ ratios range from 0.6 to 0.9. The ASI values (Table 2) indicate that the granites are metaluminous to slightly peraluminous and are classified as I-type granite (Chappell and White, 1974). All of the Taitao granites fall into the medium-K classification (Fig. 3). On the Sr/Y vs. Y diagram, all Taitao granites fell within the non-adakitic field (Fig. 6a). The Taitao granites range

Figure 2. Chemical classification of the Taitao ophiolite and Taitao granites. (a) Total alkali vs. silica diagram after Le Maitre et al. (2005). (b) Normative anorthite (An)-albite (Ab)-orthoclase (Or) ternary diagram after Barker (1979).

Figure 3. Harker variation diagrams for the Taitao ophiolite and Taitao granites. Symbols are the same as in Fig. 2.

Figure 4. Trace-element variation diagrams for (a) the Taitao ophiolite and (b) the Taitao granites normalized with respect to primitive mantle (McDonough and Sun, 1995).

in Y content from 17 ppm to 39 ppm, but the Seno Hoppner pluton has distinctly high Y contents among them.

In their trace-element profiles, the Taitao granites appear enriched in LILEs and Pb with respect to primitive mantle (Fig. 4b). The granites are depleted in Ba, Nb, Sr and Eu and showed wide variation in REEs. The Seno Hoppner samples are distinguished by higher contents of most trace elements and greater depletions in Sr and Eu. The Tres Montes samples do not show negative anomalies of Sr and Eu. Sample TPB246 from the Bahia Barrientos granite differs from the others in its enrichment in Ba, Zr and Y and depletion in Th, Ta and REEs. The chondrite-normalized REE patterns (Fig. 5b) confirm these geochemical features, showing relatively enriched and inclined LREE and flattened HREE patterns. The Eu anomaly is large in the Seno Hoppner pluton, but slight or absent in the other granites.

The Taitao granites also fall into three groups with restricted Sr and Nd isotopic values (Figs. 7 and 8). The Seno Hoppner granite, with the highest SiO_2 content of all felsic rocks in this area, has the most primitive isotopic composition: three $^{87}\text{Sr}/^{86}\text{Sr}_i$ values range from 0.7045 to 0.7048, and ϵNdi values range from 3.15 to 3.96. The Tres Montes granodiorite, with the lowest SiO_2 content among the Taitao granites, has the highest $^{87}\text{Sr}/^{86}\text{Sr}_i$ value (0.7073) and a relatively low ϵNdi value (1.36). Other plutons has intermediate $^{87}\text{Sr}/^{86}\text{Sr}_i$ values (0.7049–0.7053) and ϵNdi values ranging from 0.49 to 2.35. The hybrid granite in the Tres Montes pluton has an intermediate composition between those of the mafic dike and the host granodiorite; however, it has the highest $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.7073) and the lowest ϵNdi value (–3.68) among the Taitao granites.

Figure 5. REE variation diagrams for (a) the Taitao ophiolite and (b) the Taitao granites normalized with respect to chondritic values (Taylor and McLennan, 1985).

The sediments in the MVU has the lowest Nd isotopic ratios (–10.41 for $\epsilon\text{Nd}_{5.5}$) of all our samples. The basement rocks were divided into two groups by their $\epsilon\text{Nd}_{5.5}$ values, one with values around –3 and the other with values around –9. The latter group of basement rocks has the highest $^{87}\text{Sr}/^{86}\text{Sr}_i$ values of our samples.

Magmatic interaction in the ridge-trench intersection environment

Our data indicate that the compositions of igneous rocks in the Taitao Peninsula area vary chemically and isotopically. Schulte *et al.* (2009) have demonstrated that gabbros and ultramafic rocks, which display rather restricted isotopic compositions in our data, have wider isotopic variations outside our study area (“reference data” in Fig. 8); most ultramafic rocks have $^{87}\text{Sr}/^{86}\text{Sr}_i$ values higher than those of N-MORB gabbros. The highest $^{87}\text{Sr}/^{86}\text{Sr}_i$ value was 0.708, and six of our eight ultramafic samples had high ϵNdi values (9–10). Few gabbro samples plotted in that field (solid circles with $\epsilon\text{Ndi} > 9$ in Fig. 8; Schulte *et al.*, 2009). This trend implies alteration by seawater (indicated by arrow labeled “seawater alteration” in Fig. 8) with a $^{87}\text{Sr}/^{86}\text{Sr}$ value around 0.709.

Our data from the MVU basalts have lower ϵNdi and slightly higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ than the gabbros and the EC-dike with N-MORB

Figure 6. (a) Y vs. Sr/Y ratio diagram for the Taitao granites, and (b) Mg# vs. Rb/Sr plot for all analyzed rocks.

Figure 7. (a) SiO_2 vs. $^{87}\text{Sr}/^{86}\text{Sr}_i$ diagram and (b) K_2O vs. ϵNdi diagram for the Taitao ophiolite, Taitao granites, and basement rocks. Reference data sources: Klein and Karsten (1995) and Strum and Klein (1999) for Chile Ridge basalts, Kaeding *et al.* (1990) and Schulte *et al.* (2009) for Taitao ophiolite and granitic rocks.

composition. Some volcanic rocks from the Taitao ophiolite have been reported to have similar compositions (Kaeding *et al.*, 1990; solid circles with $\epsilon\text{Ndi} = 4-5$ in Fig. 8). The distribution of these rocks nearly coincides with that of the basalts dredged from the Chile ridge segments (gray field in Fig. 8; Klein and Karsten, 1995; Strum *et al.*, 1999). Thus, these rocks were likely formed in a segment of the Chile ridge that had already subducted (segment -2 in Fig. 9a).

The volcanic rocks from the Chile ridge and from the Taitao ophiolite both plotted in a wider field that extends to lower ϵNdi values than the N-MORB field (Fig. 8). This isotopic variation implies that the source material of these rocks in the upper mantle was contaminated by crustal material with lower ϵNdi value. The most probable candidate for such material is lower crust that has experienced

partial melting (indicated by arrow labeled “deep crustal contamination” in Fig. 8). To introduce such material into the sub-ridge area, Strum *et al.* (2000) suggested a mechanism in which suboceanic lithospheric mantle material foundered during subduction and was brought beneath the spreading ridge entering the subduction zone by mantle upwelling (Fig. 10a), or one in which old continental crust or metasomatized mantle beneath the arc communicated with mantle beneath the ridge through a slab window (Fig. 10b). The contaminant must have had an isotopic composition similar to the sediment in the MVU (TPB338; green cross in Fig. 8). However, this sample contained clastic zircons of the same age as the Taitao granites, and there is no rational explanation for the extremely low ϵNdi value of this sediment.

Figure 8. $^{87}Sr/^{86}Sr$ vs. $\epsilon_{Nd}i$ diagram for the Taitao ophiolite, Taitao granites, and basement rocks and sediments. The thin solid line is a mixing line that ties isotopic composition of average gabbro and average basement rocks. Reference data sources: Kilian and Behrmann (2003) for the Chile trench sediments, Klein and Karsten (1995) and Strum and Klein (1999) for Chile Ridge basalts, Kaeding *et al.* (1990) and Schulte *et al.* (2009) for Taitao ophiolite and granitic rocks. CHUR(0) = 0.512638 (Wasserburg *et al.*, 1981) was used for the calculation of $\epsilon_{Nd}i$.

Alteration of the source mantle by deep crustal material and seawater must have been proceeding when segments of the Chile ridge were entering the subduction zone. The ultramafic section of the ophiolite underwent severe alteration by seawater, but no basaltic rocks from the Chile ridge and only a few gabbros of the Taitao ophiolite exhibited such evidence. In contrast, evidence for contamination of the sub-ridge mantle by deep crustal material has been observed in both the ridge and the ophiolite. Thus, pervasive contamination by deep crustal material must have been taking place before seawater alteration started as the ridge approached the subduction zone.

Anma *et al.* (2013) have shown that the magmatism that formed the 4.6 Ma CMU started in the South Taitao Ridge (Fig. 1b) at around 5.3 Ma and migrated eastward. They attributed this phenomenon to the subduction of the Tres Montes fracture zone. They also have reported evidence for the incorporation of trench-fill sediment that was brought to the upper mantle with the subducting fracture zone, then returned to the surface by volcanic activities of the CMU. Our new isotopic data support this model in that the mafic volcanic rocks of the CMU and TM-dike have higher $^{87}Sr/^{86}Sr$ values than the Chile ridge basalts. The isotopic composition of the CMU may be explained in terms of marine alteration of basaltic rocks that were contaminated by deep crustal material (trajectory 1 in Fig. 8), contamination by trench-fill sediment of mantle that was altered by seawater (trajectory

3 in Fig. 8), or the combined influence of seawater alteration and sediment contamination of N-MORB magmas (trajectory 2 in Fig. 8). Alteration of mantle by seawater was most likely to occur along a fracture zone (Fig. 9a–d). As the fracture zone approached the subduction zone, sediment from the forearc area was carried onto it through submarine canyons. This fracture-filling sediment, accompanied by seawater, may have penetrated the crust during shear along the fracture zone and altered the mantle (Fig. 10c). The influence of slab tearing that resulted from buoyancy differences between the subducting slab segments may have enhanced this penetration.

Origin of calc-alkaline I-type granites

The fact that the Taitao granites are restricted to the vicinity of the Taitao ophiolite implies that the plutonism was related to the subduction of the hot ridge. The Taitao granites are calc-alkaline, meta-aluminous to slightly peraluminous ($0.83 < ASI < 1.1$), I-type granites. Such granites are generally considered to form by partial melting of an igneous source material. In the Taitao Peninsula region, possible source materials are limited to the subducting oceanic crust or igneous crustal material in the forearc region, if any. The low Sr ratios and high ϵ_{Nd} ratios support the oceanic crustal origin of the granitic melts. However, contamination by sediment or other continental crustal material did occur, especially in the Tres Montes granodiorite, which contains over 40% recycled zircon of Cretaceous to Late Proterozoic ages (Anma *et al.*, 2009).

The mixing line in Figure 8 represents a petrogenetic model based on Sr–Nd isotopic compositions. Most Taitao granites plot on the mixing line between the average composition of gabbros, with primitive N-MORB isotopic compositions, and the average composition of metasedimentary basement rocks with the lowest ϵ_{Nd} values, chosen as an extreme end-member. Taking N-MORB as the source material of the granite melts, our calculations indicate that the Seno Hoppner granite (with the most depleted isotopic composition) requires incorporation of no more than 5–10% of contaminant. Other plutons, except the Tres Montes pluton, would require slightly more contamination to yield a result concordant with the calculation by Kaeding *et al.* (1990). Because recycled zircon is rare in these plutons, the contaminant could also be lower crustal material originating from a foundering slab or from sub-arc mantle that had experienced partial melting (Strum *et al.* 2000). Conversely, the major- and trace-element compositions of our samples favor contamination by sediment or basement of the upper crust. To produce the Tres Montes granodiorite requires a significant alteration by seawater or incorporation of sediment, most probably related to the subduction of the Esmeralda fracture zone (Fig. 9b, c).

The Taitao ophiolite has undergone low-pressure metamorphism at zeolite to amphibolite facies conditions (Shibuya *et al.*, 2007). Experimental studies at less than 1 GPa suggest that decomposition of amphiboles in hot oceanic crust could result in K-poor partial melts

Figure 9. (a) Migration and subduction of the Chile ridge at 8 Ma. (b) Emplacement of the Taitao ophiolite and Taitao granites at ~6 Ma. (c) Emplacement of the Taitao ophiolite and Taitao granites at 5.7–4.6 Ma. (d) Migration and subduction of the Chile ridge at 4 Ma. (e) Cross section of ophiolite and granite emplacement at 5.7–5.1 Ma; location in (b). Figures are modified from Anma *et al.* (2006, 2013).

with tonalitic, trondhjemitic or granodioritic compositions (Beard and Lofgren, 1990; Rapp *et al.*, 1991; Patino Douce, 1995; Rapp and Watson, 1995; Springer and Seck, 1997; Sisson *et al.*, 2005). The juvenile magmas that constitute the principal component of the Taitao granites, with unfractionated HREE patterns and low La/Yb ratios, could be explained by partial melting of amphibolite in the absence of garnet (Kon *et al.*, 2013).

The high K content and lower MgO and CaO contents in the Seno Hoppner granite imply partial melting of oceanic crust at a higher temperature than other plutons. This temperature contrast is reasonable because the 5.2 Ma Seno Hoppner pluton is farther from the trench, hence the subducting slab underneath it was deeper and hotter than was the case for the 5.2 Ma Estero Cono pluton and the 5.7 Ma Tres Montes pluton near the trench. Zircon saturation temperatures (Hanchar and Watson, 2003) calculated for the granites support the higher temperatures (over 810°C) for the Seno Hoppner granite compared to the others (down to 770°C). Partial melting of K-rich basaltic rocks of the Chile Ridge (Klein and Karsten, 1995; Sherman *et al.*, 1997) could have formed such granitic melts (Sisson *et al.*, 2005). Nevertheless, producing high-K granitic melts with 74% SiO₂ content is not easy and may require further processing. The Seno Hoppner granite has a prominent negative Eu anomaly that implies fractional crystallization by settlement of Ca-plagioclase, whereas the hypothesized source gabbros with N-MORB composition have positive Eu anomalies. Considering that REE concentrations in our samples were about 10 times greater in granite than in gabbro, the granitic melts must have collected material from a vast amount (10 times their volume) of gabbro that was continually brought into the subduction zone with the moving plate.

Lagabriele *et al.* (2003) emphasized the importance of repeated passages of several ridge segments to develop high thermal gradients that allow partial melting of oceanic lithosphere at temperatures of 800–900°C and low pressures corresponding to depths of 10–20 km, and to form Quaternary calc-alkaline acidic magmatism in the Chile triple junction area. It is noteworthy that the I-type calc-alkaline magmatism took place during the first ridge subduction event (Fig. 9a, b).

Implication for ophiolite tectonomagmatism

The tectonomagmatic processes we report here, especially the formation of the calc-alkaline granites at this ridge-trench intersection environment, have important implications for constructing tectonomagmatic models of ophiolites. Presence of the calc-alkaline granites is usually interpreted as a supra-subduction zone signature. Our study shows that the

Figure 10. Models for the contamination of oceanic mantle by continental material (a) through incorporation of foundered slab material (Strum et al., 2000), (b) through a slab window connecting sub-arc mantle and sub-ridge mantle (Strum et al., 2000), and (c) through incorporation of seawater and sediment penetrating a fracture zone (this study).

calc-alkaline magmas might be generated at mid-ocean ridge segments that are about to be subducted.

As exemplified in introduction, the Taitao ophiolite-granite complex has a full range of chemical variations that is comparable to the Semail ophiolite, which had a rather prolonged tectonomagmatic history. In the case of the Semail ophiolite, the history from the generation of ophiolitic rocks to magmatism and modification at an arc environment is supported by age constraints. However, if such history relies only on geochemical signatures, then it may result in misinterpretation of the rock record of many ophiolites by mistaking what are essentially of mid-oceanic segment origin for an apparent supra-subduction zone signature.

Compared to the Semail ophiolite, the Taitao ophiolite-granite complex has a significantly greater fraction of granitic rocks. Previous have argued for heating of the forearc region as evidenced by high geothermal gradient metamorphism in this region (Lagabrielle et al., 2000; Shibuya et al., 2007; Kon et al., 2013). Such evidence for a high geothermal gradient is lacking in the generally cold settings

of emplacement for ophiolites such as Semail (Wakabayashi et al., 2010), with minor exception of the metamorphic sole that indicates a ridge subduction origin. Perhaps, the amount of calc-alkaline granites incorporated into ophiolites depends on whether the concerned oceanic lithosphere was young and hot: in the case of Semail with a prolonged history, the place of origin of ophiolitic rocks and the obducted margin were far enough such that the Cretaceous oceanic lithosphere was cooled sufficiently before obduction. This process might have resulted in a distinctive arc-type magmatism and a less amount of granitic rocks.

Summary

The late Miocene – early Pliocene Taitao ophiolite provides an excellent opportunity to study the emplacement mechanism of a ridge-trench intersection ophiolite and the complex magmatic interactions between the subducting ridge, overlying crust and sediments, and mantle. The MORB erupted near the ridge subduction zone appears to have been contaminated by a lower crustal material that has experienced partial melting. Such material could have been delivered to the sub-ridge mantle through a slab window from the sub-arc mantle, or it could have originated from sub-oceanic lithospheric mantle that had foundered during subduction and been brought to the spreading ridge by mantle upwelling. Alteration of mantle material by seawater along fracture zones and contamination by fracture zone-fill sediments are other important factors that may control magmatic interactions in a ridge-trench intersection environment. As a result of ridge subduction, juvenile magmas of calc-alkaline I-type granites were formed by partial melting of young and hot oceanic ridge lithosphere under garnet-free-amphibolite conditions. The magmas were then contaminated by subducted sediment or upper crustal components. Fractional crystallization of Ca-plagioclase favored production of granitic melt with high K_2O and SiO_2 contents.

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