Geochemical heterogeneity and element mobility in deeply subducted oceanic crust; insights from high-pressure mafic rocks from New Caledonia

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Abstract

Bulk-rock major and trace element geochemistry of a range of eclogite, garnet blueschist and garnet amphibolite rocks from northern New Caledonia has been determined in order to geochemically characterise subducted oceanic crust. The rocks experienced peak metamorphic conditions of 1.9 GPa and 600 °C and represent excellent samples of oceanic crust that was subducted to depths of approximately 60 km. The rocks can be divided into seven rock types that respectively have geochemical characteristics of enriched and normal mid-ocean ridge basalt, back-arc basin basalt, alkaline basalt, plagioclase-rich cumulate, seafloor-altered basalt and Fe–Ti basalt. All of the samples studied represent a single slice of oceanic crust interpreted to have formed in a back-arc or marginal basin setting. Examination of modern oceanic crust suggests that most subducting crust also contains a diverse range of mafic rock-types. The presence of minor amounts of alkaline and seafloor-altered basalts in the slab can greatly influence the recycling of incompatible elements and the depth of fluid release during subduction.

Comparison of the high-grade metamorphic rocks with equivalent igneous rocks from western New Caledonia demonstrates that the main chemical variations of the rocks are related to differences in their magmatic history and to different degrees of seafloor alteration, whereas high-pressure metamorphism produced only minor changes. There is evidence for some depletion of LILE and B during subduction in a few of the analysed samples. However, most of the blueschists and eclogites with protoliths showing high LILE contents such as the back-arc basin and alkaline basalts still contain high LILE contents of 10–100 times the amount found in normal mid ocean ridge basalts. Therefore, even fluid mobile elements (B, LILE) may be efficiently subducted to sub-arc depths. Trace elements are most likely to be removed from the slab in regions of elevated temperature or in zones of intense fluid–rock interaction or partial melting. In contrast to the trace elements, large volumes of fluid are liberated from mafic rocks prior to eclogite-facies metamorphism, providing evidence for a decoupling of fluid and trace element release in subducted oceanic crust.
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Keywords: Subduction; New Caledonia; Oceanic crust; Trace elements; Metamorphism; Eclogite

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1. Introduction

The recycling of elements through convergent plate boundaries strongly influences the chemical differentiation of Earth. Subduction zones are probably the most enigmatic and least understood of tectonic environments, but there is a growing consensus that a significant component of arc magmas is derived from subducting oceanic crust. Recent isotopic and geochemical work (Hawkesworth et al., 1993; Elliott et al., 1997; Turner et al., 1997; Taylor and Nesbitt, 1998; Class et al., 2000; Macdonald et al., 2000; Clift et al., 2001) has focussed on identifying the relative contributions to arc magmatism from both mafic and sedimentary components of subducting slabs. Most of these studies have concluded that elements from both components are found in arc magmas. Extensive sampling of the seafloor has enabled researchers to characterise in detail the nature and composition of subducting sediments (Ben Othman et al., 1989; Plank and Langmuir, 1993, 1998; Rea and Ruff, 1996; Elliott et al., 1997; Kilian and Behrmann, 2003). By contrast, our knowledge of the composition of the mafic rock of subducting crust is generally quite poor. It is often assumed that the majority of the downgoing crust is of MORB or altered MORB composition (Hawkesworth et al., 1993; Becker et al., 2000; Hacker et al., 2003). However, the spectrum of oceanic mafic rocks covers a wide range of trace element compositions (Wilson, 1989). Therefore, the common assumption that oceanic crust has a MORB bulk composition may be invalid. Here, we investigate the geochemical implications of subducting heterogenous oceanic crust into the mantle.

Experimental (Pawley and Holloway, 1993; Peacock, 1993; Liu et al., 1996; Schmidt and Poli, 1998) and petrologic studies (Bebout et al., 1999; Scambelluri and Philippot, 2001; Spandler et al., 2003) have shown that subducting crust undergoes extensive dehydration and fluid loss during eclogite-facies metamorphism. It is clear that trace elements are also liberated from the subducting plate, but the degree and mechanisms of element removal are highly contentious. Blueschist and low-temperature (T) eclogite terranes often represent exhumed fragments of previously subducted crust (Schliestedt, 1990). Therefore, studies of these terranes may provide fundamental information concerning the nature of the slab prior to subduction and the effects of high-pressure (P) metamorphism and deformation during subduction. Studies comparing the composition of high-P metamorphic rocks to their estimated protoliths suggest that significant amounts of Sr, K, Rb, Ba and U are removed from mafic rocks during subduction up to eclogite facies (Arculus et al., 1999; Becker et al., 2000). However, these estimates are highly speculative, as the compositions of the protoliths for these high-P rocks were not precisely constrained.

In this paper we describe the petrology and geochemistry of a range of garnet blueschist, garnet amphibolite and eclogite-facies mafic rocks from northeastern New Caledonia. These rocks are part of an extensive high-P, low-T metamorphic terrane that is an excellent analogue of oceanic crust that was subducted to depths of up to 60 km (Aitchison et al., 1995; Clarke et al., 1997). We will show that these mafic rocks span a great compositional range. Our data are compared with weakly metamorphosed equivalent rocks from New Caledonia in order to constrain the geochemistry and lithology of the mafic rocks of the slab and evaluate the degree of trace-element remobilisation by subduction-zone metamorphism. The results have important implications for the recycling of elements through subduction zones and models of arc magma genesis.

2. Geological setting

A detailed synopsis of the geology of New Caledonia has been presented by Aitchison et al. (1995) and Cluzel et al. (2001). Here we focus on the geological units that are relevant to this study. New Caledonia represents one of the largest exposures of continental crust in the southwest Pacific. It is composed of three late Palaeozoic to Mesozoic basement terranes that are overlain by a number of Cretaceous to Eocene sedimentary sequences and ophiolite nappes (Fig. 1). The extensive (~ 2200 km^2) high-P, low-T metamorphic belt in the northeast of the island comprises lawsonite blueschist to eclogite-facies rocks that experienced maximum metamorphic conditions of approximately 1.9 GPa and 600 °C (Black, 1977; Clarke et al., 1997; Carson et al., 1999). The belt is interpreted to have formed during Eocene subduction of an oceanic basin, causing
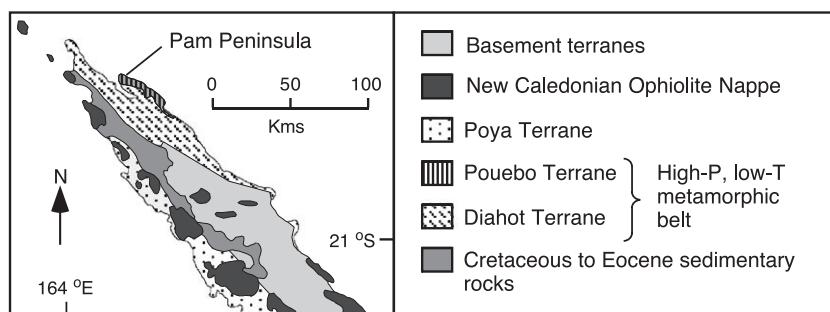
Fig. 1. Simplified geological map of northern New Caledonia (adapted from Aitchison et al., 1995).

recrystallisation of sedimentary and volcanic proto-liths under blueschist and eclogite-facies conditions (Aitchison et al., 1995; Clarke et al., 1997). There is a clear progressive increase in metamorphic grade through the belt from the south and west to the northeast (Black, 1977; Yokoyama et al., 1986; Spandler et al., 2003). However, the apparent rapid increase in metamorphic grade is largely due to extensive disruption of the original sequence by exhumation-controlled normal faulting (Aitchison et al., 1995; Clarke et al., 1997; Rawling and Lister, 1999). The belt has been subdivided into two terranes on the basis of metamorphic grade and lithology (Clarke et al., 1997). The sediment-dominated Diahot Terrane comprises most of the belt and includes almost all of the blueschist-facies rocks. The Pouébo Terrane outcrops along the northeastern coast and contains garnet blueschists and eclogite and amphibolite gneisses of mafic and pelitic composition. These rock-types comprise blocks within a highly sheared and deformed pelite and serpentinite mélange matrix (Maurizot et al., 1989; Rawling and Lister, 1999, 2002). Nonetheless, many of the mafic rocks of the Pouébo Terrane preserve peak metamorphic mineral assemblages and have not experienced significant retrogression or deformation. Clarke et al. (1997) have identified three high-grade mafic rock-types based on field relations and petrography. These are Types I and II eclogite and garnet glaucophanite. Preliminary geochemical work also indicates that there is considerable compositional variation in these mafic rocks (Cluzel et al., 2001).

The western side of New Caledonia is largely composed of a sequence of Late Cretaceous to Early Eocene mafic rocks known as the Poya Terrane (Cluzel et al., 1997, 2001; Eissen et al., 1998). Poya Terrane rocks are within 20 km of the high-P, low-T metamorphic belt in some places. The Poya Terrane consists of pillowed and massive basalts, dolerites, and gabbros with associated cherts and volcaniclastic rocks (Eissen et al., 1998) and is interpreted to be a segment of oceanic crust that was thrust over New Caledonia during the Eocene. Metamorphism is largely sub-greenschist facies and is related to seafloor hydrothermal alteration (Cluzel et al., 2001). Many samples have well-preserved igneous textures and minerals. Geochemical studies suggest that most of the mafic rocks are of enriched (E) MORB composition, but normal (N) MORB, alkaline basalts and back-arc basin basalts (BABB) have been identified (Eissen et al., 1998; Nicholson et al., 2000; Cluzel et al., 2001).

A direct link between the Poya Terrane and the mafic rocks of the Pouébo Terrane has been made based on tectonic, paleontological, isotopic and geochemical data (Aitchison et al., 1995; Cluzel et al., 2001). Direct correlation of the terranes is also supported by recent U–Pb SHRIMP dating of igneous zircon cores from a range of high-P rocks of the Pouébo Terrane and will be presented in detail elsewhere. Ages obtained are ~ 85 and ~ 55 Ma, both of which fall within the age of the Poya Terrane (Cluzel et al., 2001). Both terranes are suggested to be slices of the same oceanic crust prior to Eocene tectonic and metamorphic events. On this basis, geochemical comparison of the Poya and Pouébo Terranes allows for the qualitative determination of bulk-rock geochemical changes associated with prograde metamorphism up to eclogite facies.

3. Petrology of the mafic rock of the Pouébo Terrane

We have identified seven high-grade mafic rock-types from the Pouébo Terrane, on the basis of field relations, petrography, mineralogy and bulk-rock chemistry. Although most of the rock-types have been identified throughout the terrane and are often closely associated, many have not been previously described. Features of these rock-types are presented in Table 1 and are outlined below. Detailed petrographic descriptions and the metamorphic evolution of Types I and II eclogite and garnet glaucophanite have been presented by previous researchers (Clarke et al., 1997; Carson et al., 1999, 2000). Estimates of peak metamorphic conditions for the other rock-types have also been calculated using well-calibrated models of phase equilibria and exchange reactions (Ellis and Green,

1979; Holland, 1979, 1983; Green and Hellman, 1982; Graham and Powell, 1984). The thermobarometry is consistent with field relationships as it yields similar P and T values for all rock-types. Our estimate of peak metamorphism for the Pouébo Terrane is in line with previous suggestions (Carson et al., 1999) of approximately 1.9 GPa and 600 °C.

3.1. Types I and II eclogite and garnet amphibolite

Type I and II rock types have been described in detail elsewhere (Clarke et al., 1997; Carson et al., 1999), so only a brief description of these rock types is given here. Type I eclogite and garnet amphibolite is the most abundant and widespread mafic rock type of the Pouébo Terrane. It is usually well foliated and consists mainly of Na–Ca hornblende, idioblastic garnet and zoisite. Omphacite, rutile, quartz, apatite

Table 1
Petrology and mineralogy of high-P, low-T mafic rock-types of the Pouébo Terrane

Rock-type	Type I	Type II	Type III	Type IV	Type V	Type VI	Type VII
Metamorphic facies	eclogite, garnet amphibolite	eclogite	eclogite	garnet blueschist	eclogite	garnet glaucophanite (blueschist)	garnet amphibolite
Grainsize	medium	coarse	fine	medium–very coarse	medium	medium–very coarse	medium–coarse
Mineralogy							
Garnet (%)	12–18	15–20	18–25	~10	12	10–15	8–12
Hornblende (%)	50–55	10–40	~5–10	2–55	–	0–5	70–85
Omphacite (%)	2–12	5–40	~30	0–5	42	–	1–2
Epidote (%)	18–22	10–20	10–15	5–12	14	<1	3–6
Glaucophane (%)	0–5	0–5	~25	10–60	–	70–80	1–2
Phengite (%)	<2	<1	<1	15–25	4	1–4	<1
Paragonite (%)	–	~5	<1	0–5	28	1–4	–
Quartz (%)	1–7	>5	>5	>5	>5	0–1	1–2
Apatite (%)	<1	–	<1	1–3	–	<1	<1
Rutile (%)	<1.5	<1	<2	<3	–	<1	1–3
Titanite (%)	<1	<1	<1	0–6	trace	3–4	0–3
Ilmenite (%)	–	–	–	–	–	–	<1
Albite (%)	–	–	–	–	–	0–5	–
Trace minerals	zc	zc	zc, py, cpy	zc, all	–	zc	zc, all, mzt
Peak P (GPa)	1.9	1.9	1.3–2.0	1.4–2.0	1.4–2.0	1.6–1.8	1.4–2.3
Peak T (°C)	530–680	500–650	500–580	550–650	580–650	480–600	530–620
P, T source	1, 2, 3	1, 2, 3	3, 4	3, 4	3, 4	2, 3	3, 4
Geochemical affinity	EMORB	BABB	NMORB	alkaline basalt or OIB	plagioclase cumulate	altered MORB	Fe–Ti basalt

Thermobarometry sources are; 1—Clarke et al. (1997); 2—Carson et al. (1999); 3—thermometry calculated from coexisting mineral pairs and equations from Ellis and Green (1979), Graham and Powell (1984) and Green and Hellman (1982); 4—barometric ranges calculated from mineral stability relations from Holland (1979, 1983) and Poli and Schmidt (1995). Mineral abbreviations; zc—zircon, py—pyrite, cpy—chalcopyrite, all—allanite, mzt—monazite.

and phengite may be minor constituents and fine-grained zircon is always present in trace amounts. Greenschist retrogression of these rocks is very limited, being restricted to occasional and incomplete replacement of garnet by chlorite, rutile by titanite and amphibole overgrowths on omphacite. Type II eclogites are light in colour and are distinguished by their large splays of clinozoisite, omphacite, amphibole and paragonite. Idioblastic garnets contain inclusions of glaucophane and lawsonite indicating prior metamorphism under blueschist-facies conditions. Again trace amounts of zircon are ubiquitous in these rocks.

3.2. Type III eclogite

Type III eclogites are common in the south of the Pouébo Terrane and have been observed as boulders in streams draining the Pam Peninsula to the north. They are easily identified in the field by their dark colour, fine grainsize and high sulfide content. Although idioblastic garnets and zoisite laths may reach 0.5 mm in size, most grainsizes are less than 100 μm. Zoisite, glaucophane, omphacite and garnet comprise the bulk of the rocks. Hornblende, quartz, phengite and paragonite may be minor phases. Garnet compositions are variable but cores tend to be relatively spessartine-rich and contain an assemblage of mineral inclusions that is the same as the groundmass assemblage. Type III have relatively high rutile contents but similar amphibole compositions to Type II amphiboles. Pyrite, chalcopyrite and zircon are present in minor amounts in the groundmass and as inclusions in garnet. Retrogression is restricted to thin titanite overgrowths on rutile.

Clarke et al. (1997) has described a Type III eclogite from the Pouébo Terrane. However, this rock-type lacks omphacite and is not mafic in composition, so we do not regard the rock type as an eclogite. Instead, the rock-type is likely to be part of a sedimentary sequence that comprises a significant proportion of the Pouébo Terrane (Black, 1977; Spandler et al., 2003).

3.3. Type IV garnet blueschist

Highly foliated, high-grade mafic rocks that contain large amounts of phengite occur throughout the Pouébo Terrane. These rocks are labelled Type IV garnet blueschists. They are often very coarse-grained with some samples containing inclusion-rich phengite crystals up to 5 cm in length. Some samples are petrographically and mineralogically similar to Type I garnet amphibolites, but contain higher contents of phengite, glaucophane, zircon and rutile. Most samples are dominated by phengite and glaucophane reflecting their alkaline composition. Garnet, apatite, rutile, titanite and epidote-group minerals comprise a minor part of these samples. Epidote is usually strongly zoned from allanite to zoisite (Spandler et al., 2003). Garnet is relatively grossular-poor and may contain inclusions of epidote, rutile, titanite, quartz, albite, chlorite, and hornblende. Most samples have been retrogressed to some degree, but this is usually limited to minor chlorite replacement of garnet and titanite replacement of rutile.

3.4. Type V eclogite

We have only recognised one sample of the Type V eclogite, which looks very similar to the Type II eclogite in hand sample. However, the Type V eclogite is petrographically unlike any other rock-type known from the Pouébo Terrane. The sample consists entirely of omphacite, epidote, garnet, white mica, quartz and trace amounts of titanite. In contrast to other eclogite types, amphibole, zircon and rutile are not present. Zoisite and clinozoisite occur as well-formed tabular crystals. Garnets are euhedral idioblastic grains and are inclusion-free with the exception of rare epidote inclusions. Fine-grained intergrowths of paragonite and phengite often mantle garnet but can be found throughout the rock. There is no evidence of late-stage alteration or retrogression of any kind, hence the peak metamorphic mineralogy and texture of this rock is exceptionally well preserved.

3.5. Type VI garnet glaucophanite

Garnet glaucophanite is one of the most common rock-types of Pouébo Terrane and has been suggested to represent the hydrated and retrogressed equivalent of Type I eclogite (Carson et al., 2000). However, our chemical analyses presented below demonstrate that at least some of the glaucophanites have significantly different compositions with respect to Type I rocks.

Table 2
Bulk rock geochemistry of high-P, low-T mafic rock-types from the Pouébo Terrane

Rock type	Type I					Type II				Type III
	Garnet amphibolite			Eclogite	Garnet amphibolite	Eclogite				Eclogite
Sample	1001	703	704	705	3106	707	706	3102	3101	803
Latitude (S)	20° 17′ 00″	20° 15′ 23″	20° 15′ 23″	20° 14′ 11″	20° 15′ 00″	20° 15′ 41″	20° 14′ 56″	20° 16′ 41″	20° 16′ 41″	20° 29′ 30″
Longitude (E)	164° 26′ 59″	164° 18′ 31″	164° 18′ 31″	164° 19′ 10″	164° 18′ 23″	164° 19′ 32″	164° 20′ 00″	164° 18′ 42″	164° 18′ 42″	164° 46′ 52″
SiO_2	50.46	46.07	46.60	48.74	49.31	49.09	50.83	51.31	52.24	50.36
TiO_2	1.46	1.33	1.50	1.66	1.60	0.64	0.83	1.07	1.16	1.90
Al_2O_3	14.15	14.90	15.12	14.45	14.54	18.45	15.97	15.97	15.29	13.59
Fe_2O_3	12.00	12.51	12.54	12.56	12.32	6.48	7.73	8.76	9.67	14.00
MnO	0.20	0.22	0.20	0.20	0.18	0.11	0.13	0.15	0.17	0.21
MgO	6.81	9.16	7.50	7.31	8.47	8.40	8.30	7.18	7.57	7.13
CaO	10.55	12.46	12.95	10.92	11.13	13.40	13.16	11.87	10.96	9.23
Na_2O	2.44	2.17	2.50	2.71	2.16	1.87	2.26	2.50	2.54	3.50
K_2O	0.11	0.12	0.11	0.19	0.16	0.12	0.10	0.20	0.54	0.02
P_2O_5	0.16	0.22	0.18	0.14	0.15	0.07	0.07	0.11	0.11	0.14
S	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.08	0.16
LOI	1.44	1.21	0.75	1.02	1.23	1.64	1.36	1.25	1.18	0.51
Total	99.78	100.37	99.95	99.90	101.26	100.27	100.75	100.37	101.51	100.75
trace elements										
Sc	44.5	45.4	42.2	44.1	51.1	30.8	39.9	41.7	40.9	46.8
V	331	355	324	340	408	186	230	323	331	394
Cu	94	92	80	63	–	4	69	–	–	65
Zn	95	97	94	98	–	46	56	–	–	127
Co	43.3	58.8	46.3	41.4	68.3	26.3	25.1	53.1	75.2	46.5
Ni	79.2	109	102	91.6	86.6	113	54.1	61.1	139	55.1
Cr	185	387	312	245	351	860	204	544	377	88
Cs	0.02	0.03	0.01	0.05	0.01	0.07	0.09	0.11	0.24	0.01
Rb	1.56	0.91	0.70	2.57	1.40	1.50	1.59	5.36	11.28	0.09
Ba	8.82	8.09	6.12	30.2	28.9	40.9	24.2	98.6	222	2.48
Th	0.36	0.33	0.40	0.42	0.46	0.44	0.48	0.82	1.18	0.18
U	0.13	0.19	0.13	0.12	0.14	0.11	0.11	0.20	0.27	0.08
Nb	4.78	4.52	4.70	4.93	5.46	1.72	1.87	2.68	3.13	2.63
Ta	0.29	0.30	0.29	0.32	0.42	0.10	0.11	0.25	0.29	0.15
La	4.29	4.91	5.02	7.07	5.54	2.60	2.37	4.80	8.26	3.30
Ce	11.11	12.72	13.05	16.07	14.44	6.86	6.56	12.13	18.58	10.23
Pr	1.75	1.96	2.01	2.60	2.28	1.06	1.04	1.88	2.69	1.92
Mo	1.35	1.08	1.34	1.34	0.49	1.38	1.21	0.60	0.74	1.74
Sr	237	153	202	119	185	113	101	145	181	83.2
P	731	960	814	580	883	279	336	625	683	609
Nd	9.01	10.24	10.40	13.16	11.67	5.40	5.44	9.79	13.07	11.00
Sm	2.87	3.59	3.26	4.07	3.79	2.00	1.89	3.29	4.12	4.26
Zr	80.4	76.5	97.3	99.7	104.3	42.2	49.1	87.8	101	117
Hf	2.05	2.05	2.44	2.62	2.73	1.14	1.34	2.37	2.69	3.20
Eu	1.13	1.32	1.23	1.46	1.40	0.72	0.75	1.12	1.38	1.40
Ti	7986	7440	8773	9268	11 798	3672	4534	6824	7384	10 420
Gd	3.86	4.55	4.07	5.12	4.92	2.65	2.63	4.67	5.95	6.15
Tb	0.64	0.85	0.71	0.92	0.84	0.48	0.52	0.82	1.07	1.14
Dy	4.40	5.64	4.60	5.82	5.38	3.22	3.41	5.52	7.16	7.92
Y	26.3	31.9	26.9	37.9	31.3	20.2	21.2	35.0	49.2	51.5
Ho	0.90	1.19	0.96	1.25	1.12	0.70	0.74	1.23	1.58	1.73

Type III	Type IV				Type V	Type VI		Type VII		
	Garnet blueschist				Eclogite	Garnet blueschist		Garnet amphibolite		
805	1005	1101	2810	2818	3009	922	1003	80 105	80 104	72 811
20° 29′ 30″	20° 17′ 00″	20° 18′ 40″	20° 16′ 26″	20° 16′ 02″	20° 18′ 37″	20° 19′ 08″	20° 17′ 00″	20° 26′ 00″	20° 26′ 00″	20° 16′ 15″
164° 46′ 52″	164° 26′ 59″	164° 25′ 27″	164° 23′ 06″	164° 23′ 12″	164° 26′ 42″	164° 25′ 23″	164° 26′ 59″	164° 38′ 59″	164° 38′ 59″	164° 23′ 08″
50.00	47.74	49.60	49.95	50.16	48.87	53.23	53.65	45.10	46.63	47.14
1.88	1.87	2.89	2.36	2.00	0.27	1.71	1.79	2.90	3.02	3.85
13.48	16.18	15.01	15.62	14.52	23.01	11.77	12.21	13.06	12.74	11.06
14.26	10.48	10.64	11.27	11.88	7.57	11.70	12.70	18.66	20.18	19.04
0.23	0.21	0.37	0.42	0.36	0.08	0.55	0.46	0.71	0.28	0.26
6.72	8.71	7.25	7.59	7.63	3.70	9.99	9.28	7.37	5.43	4.81
10.13	7.70	5.61	6.91	6.78	9.69	3.76	3.21	8.43	9.45	8.27
2.89	2.58	3.75	1.79	4.47	5.17	5.36	5.50	3.40	2.56	4.49
0.04	2.13	2.29	2.17	0.29	0.80	0.13	0.38	0.28	0.13	0.04
0.14	0.33	0.85	0.51	0.44	0.11	0.40	0.01	0.03	0.16	0.42
0.22	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.50
0.43	2.29	2.07	2.46	2.07	1.90	1.46	1.37	1.25	0.25	1.35
100.42	100.23	100.33	101.05	100.60	101.17	100.06	100.56	101.19	100.84	101.23
46.0	35.6	25.6	28.2	28.6	17.7	38.9	42.8	54.4	53.5	49.0
392	324	193	274	293	474	246	292	637	1144	729
73	22	29	–	–	–	60	47	–	–	–
119	89	88	–	–	–	59	88	–	–	–
48.7	38.3	35.5	49.8	50.3	39.1	47.5	51.5	114	94.5	70.6
62.1	89.9	84.7	67.8	55.4	55.9	93.2	97.6	89.4	45.1	19.82
75	208	208	191	135	13.66	212	273	20.4	8.9	11.8
0.03	0.77	0.87	0.72	0.26	0.26	0.03	0.08	0.06	0.13	0.02
0.56	28.83	74.1	66.9	7.39	13.46	4.17	12.50	2.85	0.98	0.41
19.94	163	691	670	51.6	107	30.1	99.2	96.9	24.5	4.00
0.17	2.13	4.13	3.33	3.46	0.27	0.22	0.99	1.11	0.50	0.53
0.06	0.70	1.67	1.04	1.14	0.20	0.53	0.51	0.44	0.14	0.24
2.63	29.28	50.67	40.52	34.72	0.66	6.05	7.96	9.02	6.22	8.02
0.16	1.75	3.36	2.55	2.40	0.09	0.22	0.42	0.70	0.49	0.67
3.07	17.33	40.96	28.10	26.82	4.88	5.08	11.12	18.49	28.70	11.47
10.00	36.14	82.20	54.38	51.33	7.60	13.53	22.22	86.73	47.52	33.46
1.80	4.54	10.02	6.85	6.39	1.07	2.27	3.48	6.33	7.91	5.72
1.69	1.03	1.09	1.07	0.95	0.39	1.51	1.60	1.51	0.74	1.49
89.2	229	303	200	284	303	31.41	15.27	465	184	164
685	1458	3603	3127	2764	664	1945	68.9	159	957	2580
10.46	20.06	42.04	29.12	27.03	4.10	12.22	16.61	30.70	34.62	30.30
4.00	4.97	9.51	7.12	6.73	0.93	4.35	4.67	9.26	10.01	10.34
117	162	268	229	253	30.62	111	104	218	121	336
3.13	3.59	5.80	5.16	5.55	0.80	3.04	2.73	5.47	3.25	8.47
1.36	1.63	3.18	3.29	2.23	0.92	2.21	1.85	3.31	3.22	3.22
10 580	10 580	17 250	16 990	14 540	1659	9831	9713	21 040	21 770	27 860
5.64	5.43	9.01	7.27	7.01	1.16	6.26	5.29	11.90	12.27	13.83
1.06	0.98	1.30	1.11	1.13	0.20	1.16	0.89	2.15	2.29	2.50
7.43	6.51	7.33	6.08	6.50	1.36	7.83	6.06	13.58	14.48	15.97
46.9	36.9	41.1	32.6	35.7	8.27	51.0	37.8	79.3	88.4	98.81
1.60	1.36	1.37	1.13	1.28	0.28	1.71	1.27	2.90	3.02	3.51

Table 2 (*continued*)

Rock type	Type I					Type II				Type III
	Garnet amphibolite			Eclogite	Garnet amphibolite	Eclogite				Eclogite
Sample	1001	703	704	705	3106	707	706	3102	3101	803
Er	2.66	3.19	2.74	3.61	3.17	2.11	2.16	3.57	4.67	5.27
Tm	0.36	0.44	0.39	0.51	0.43	0.30	0.31	0.51	0.66	0.75
Yb	2.49	3.20	2.62	3.29	2.88	1.97	2.14	3.41	4.45	5.05
Lu	0.38	0.45	0.38	0.50	0.45	0.30	0.31	0.53	0.72	0.79
Li	14.20	5.61	9.74	14.31	6.27	8.18	6.21	7.41	9.67	–
Be	1.11	0.67	0.62	0.51	0.45	0.25	0.24	0.37	0.50	0.44
B	5.74	–	6.01	6.08	4.94	5.50	5.15	4.68	4.69	5.44
As	1.31	0.52	0.59	0.79	0.18	0.50	0.48	0.45	0.24	
Sb	0.32	0.24	0.24	0.14	0.02	0.05	0.23	–	0.10	0.14

Moreover, we have found no evidence of pervasive retrogression of the samples examined in this study. All of our samples are dominated by glaucophane and have variable phengite contents, but contain significantly less phengite and epidote than the Type IV garnet blueschists. Hornblende is minor and occurs as intergrowths with glaucophane. Titanite is the major Ti-bearing phase and albite porphyroblasts may be present. Omphacite is not observed and clinozoisite is only found as inclusions in titanite. By contrast, Clarke et al. (1997) report both clinozoisite and omphacite as important constituents, indicating that there is considerable variation in mineral assemblages

(Carson et al., 2000). Garnet contains a variety of inclusions (hornblende, glaucophane, quartz, epidote, rutile, chlorite, albite, zircon and phengite) and may have thin spessartine-rich rims formed during minor retrogression. Apatite and zircon may be present as accessory minerals.

3.6. Type VII garnet amphibolite

Along the northeastern shoreline of the Pam Peninsula are a number of outcrops of black garnet amphibolite. These rocks are mostly comprised of dark Fe-rich hornblende and almandine-rich garnet.
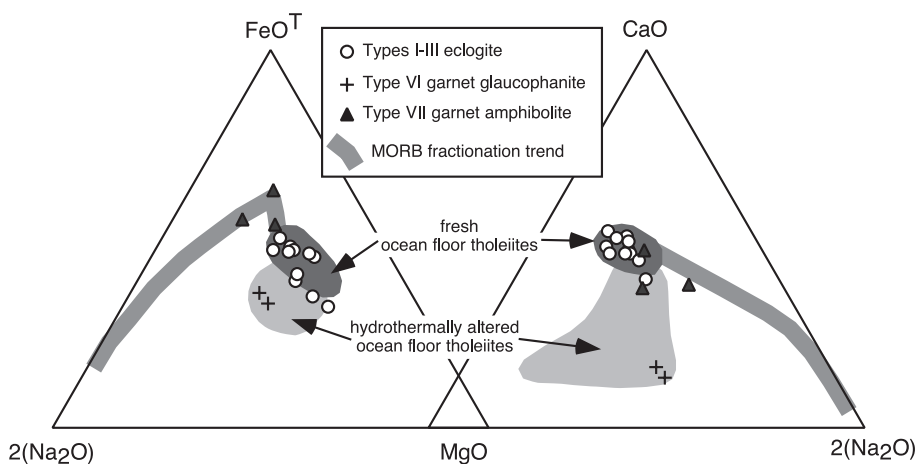
Fig. 2. FeO^T–CaO–Na_2O–MgO characteristics of high-*P*, low-*T* subalkaline mafic rocks of the Pouébo Terrane. All Fe as FeO for FeO^T. Data for the MORB fractionation trend is from Carmichael (1964) and Wood (1978). Field of hydrothermally altered ocean floor tholeiites is extrapolated from Mottl and Holland (1978) and data from Humphris and Thompson (1978a) and Alt and Emmermann (1985). Data for the field of fresh ocean floor tholeiites is from Carmichael (1964), Wood (1978), Humphris and Thompson (1978a) and Alt and Emmermann (1985).

Type III	Type IV				Type V	Type VI		Type VII		
	Garnet blueschist				Eclogite	Garnet blueschist		Garnet amphibolite		
805	1005	1101	2810	2818	3009	922	1003	80 105	80 104	72 811
4.87	3.72	3.77	2.91	3.39	0.75	5.10	3.83	8.49	8.62	10.02
0.68	0.54	0.50	0.37	0.45	0.12	0.68	0.52	1.18	1.19	1.44
4.83	3.66	3.09	2.33	2.93	0.75	4.61	3.57	8.30	7.86	9.64
0.73	0.55	0.46	0.34	0.42	0.12	0.69	0.54	1.23	1.18	1.49
–	12.32	18.02	10.37	40.34	–	–	–	–	–	–
0.48	1.08	0.81	0.80	0.99	–	0.43	0.81	–	–	–
5.51	7.17	9.13	7.74	6.24	–	6.50	6.04	–	–	–
2.28	1.39	0.47	0.64	1.04	–	0.36	0.68	–	–	–
0.36	0.23	0.47	0.41	0.36	–	0.18	0.33	–	–	–

Ti-bearing minerals are relatively abundant and include rutile, ilmenite and titanite, although titanite is most likely retrogressive. Other minor phases include zoisite, quartz, phengite and omphacite. Pyrite, zircon, allanite and monazite may be present as accessory phases. Omphacite is only found as inclusions in garnet and as highly corroded grains surrounded by hornblende. Hornblende is not found as inclusions in garnet indicating that the hornblende formed from retrogression of omphacite. However, as the garnet grains are unaltered, retrogression is likely to have occurred under high-P amphibolite facies conditions.

4. Bulk-rock geochemistry

Bulk-rock major and trace element geochemistry was conducted on 21 samples covering all of the rock types described above. Weathered material was completely removed prior to preparation of the samples for analysis. In order to minimise the effects of geochemical changes associated with late-stage alteration and retrogression, samples with the best preservation of peak metamorphic mineral assemblages and least alteration were chosen for analysis. The samples were analysed for major elements Cu, and Zn by X-ray fluorescence (XRF) spectrometer at the Department of Geology, Australian National University (ANU) and all other trace elements by laser ablation, inductively coupled plasma mass spectrometry (LA ICP-MS) at the Research School of Earth Sciences, ANU. XRF and LA ICP-MS instrument specifications

are outlined in Spandler et al. (2003) and Eggins et al. (1998), respectively. For LA ICP-MS analyses, a spot size of 100 μm was used and the counting time was 40 s for the background and 80 s for sample analysis. ^{43}Ca was employed as the internal standard isotope, based on CaO concentrations previously measured by XRF. Instrument calibration was against NIST 612 glass using the reference values presented in Spandler et al. (2003). Standard reference values for B, Be, Li, As and Sb were taken from Pearce et al. (1997). Most trace element concentrations were determined on glasses made from rock powders fused with lithium metaborate flux (1:3 ratio) at 1010 °C for 1 h. B, Be, Li, As and Sb concentrations were determined from rock powders that were fused to glass on 0.3-mm-thick molybdenum strips under 5 bars of argon pressure, using similar methods to those reported by Nicholls (1974). Fusion of the rock powders was obtained by passing current of up to 8 amps through the molybdenum strips. Rapid quenching was achieved by simultaneously turning off the current and dropping the argon pressure to 1 bar. NIST 610 and NIST 612 glasses fused on molybdenum strips were compared to unfused NIST glasses to test for trace element loss during fusion. The fused and unfused glasses were comparable to within 5% for a range of trace elements, indicating that trace element loss during fusion is minimal. Loss-on-ignition (LOI) values were calculated from the mass difference to 2 g of powdered sample after heating to 1010 °C for 1 h. All analytical data are presented in Table 2 and Figs. 2–5. We use multiple aspects of the chemical data combined with the petrographic features to distinguish

between the rock-types and determine their likely origin.

All of the rock-types are mafic in composition, but there are distinct differences between each rock-type in terms of major and trace element geochemistry. Type I rock-types generally have very little variation in composition and have major-element characteristics that are typical of ocean floor tholeiites (Figs. 2 and 3). They are slightly enriched in incompatible elements compared to NMORB (Fig. 4A), but have variable Cs, Rb, and Ba contents. Type II eclogites have lower Fe, Nb, Ta and Ti and higher Al contents than the Type I rocks (Fig. 3) and have trace element patterns that are characterised by enrichments in large-ion lithophile elements (LILE; K, Rb, Cs, Ba, Sr), Th and U and slight negative Nb, Ta and Ti anomalies (Fig. 4B). While these features are typical of arc lavas (Pearce and Peate, 1995), the Nb and Ta contents of the Type II rocks are close to typical MORB values. Magmas with these major and trace element characteristics are commonly found in back arc or marginal basins and are referred to as back-arc basin basalts (BABB; Saunders and Tarney, 1984; Sinton and Fryer, 1987). Type III eclogites have MORB-like compositions (Fig. 4C) but have slight negative Eu and Sr anomalies, high Ti and Fe contents and low K and Rb contents. They are distinguished from all other rock-types by their LREE-depleted REE patterns. Type III eclogites have lower Nb, U and Th contents than the Type I rocks and have lower U/Nb ratios than the Type II eclogites. Type IV garnet blueschists are relatively low in Ca, but have high contents of Ti and incompatible-elements; affinities which are typical of alkaline or ocean island basalts (OIB; Sun and McDonough, 1989; Wilson, 1989). It is important to note that this rock type has elevated LILE contents of up to 100 times the NMORB value (Fig. 4).

The Type V eclogite is high in Al, Na and LILE but is very depleted in rare earth elements (REE) and high field strength elements (HFSE; Nb, Ta, Ti, Hf, Zr). The REE pattern is LREE-enriched and has a large positive Eu spike (Fig. 4E). Most of these chemical features are unique to igneous plagioclase (Rollinson, 1993; Jang and Naslund, 2001). Therefore, we sug-
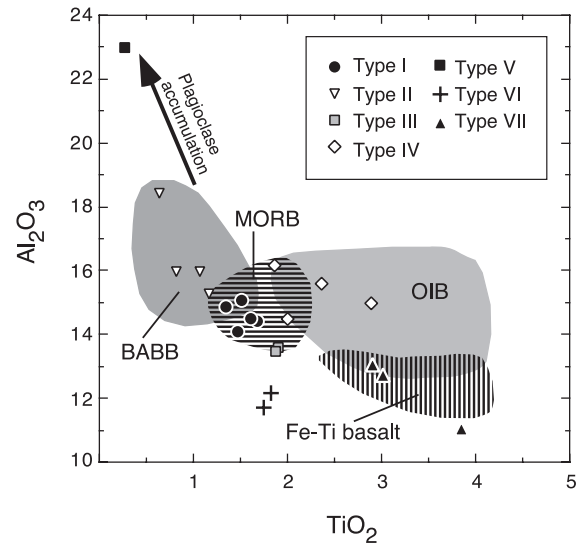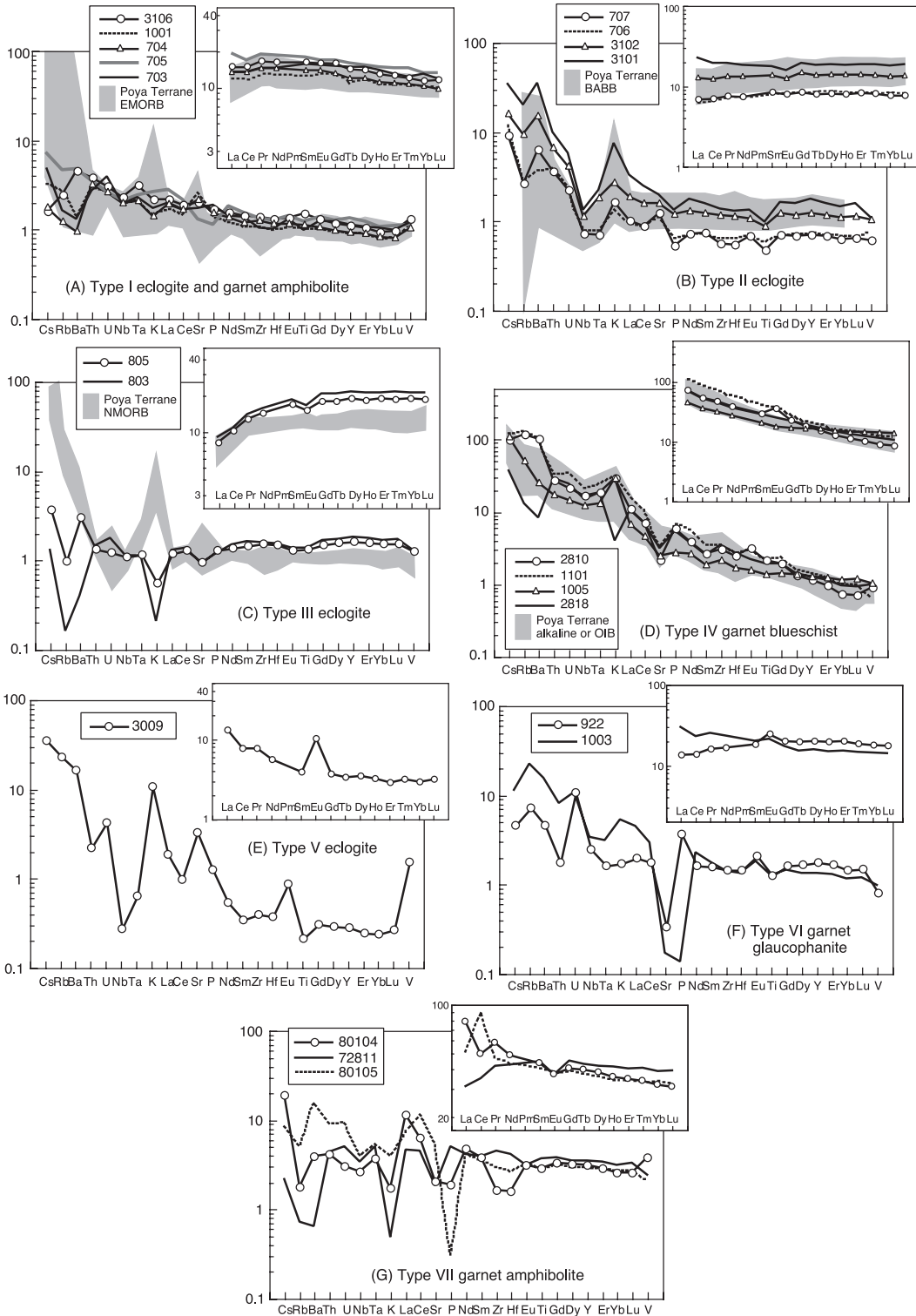

Fig. 3. $Al_2O_3 - TiO_2$ characteristics of high-P, low-T mafic rocks of the Pouébo Terrane. Data for the fields of various basalt types is taken from Carmichael (1964), Clague and Bunch (1976), Wood (1978), Basaltic Volcanism Study Project (1981), Sinton and Fryer (1987) and Wilson (1989).

gest that this rock is a high-grade equivalent of a plagioclase-rich, Fe−Ti oxide-poor cumulate (troctolite or leucogabbro). Cumulates of this nature are believed to comprise an important part of oceanic crust formed at fast-spreading ridges (Pearce, 2002; Poli and Schmidt, 2002).

Carson et al. (2000) regarded the garnet glaucophanites from the Pouébo Terrane as isochemically retrogressed Type I eclogites. However, our samples of (Type VI) garnet glaucophanite contain higher Na and Mg contents and much lower Ca contents than Type 1 rocks (Fig. 2). The garnet glaucophanites are geochemically distinct from all other rock-types of the Pouébo Terrane. Their trace-element geochemistry resembles either EMORB (1003) or NMORB (922) except for relatively high U contents, variable LILE enrichment and very low Sr contents (Fig. 4F). These chemical features are typical of basaltic rocks that have been affected by seafloor hydrothermal alteration (Fig. 2; Humphris and Thompson, 1978a,b; Mottl and Holland, 1978; Alt and Emmer-

Fig. 4. NMORB-normalised multi-element diagrams and chondrite-normalised REE diagrams (inset) for high-P, low-T mafic rock-types of the Pouébo Terrane. NMORB and chondrite normalising values are from Sun and McDonough (1989) and Taylor and McLennan (1985), respectively. Data for Poya Terrane rocks from Eissen et al. (1998), Nicholson et al. (2000) and Cluzel et al. (2001).

Fig. (A) Type I eclogite and garnet amphibolite

Fig. (B) Type II eclogite

Fig. (C) Type III eclogite

Fig. (D) Type IV garnet blueschist

Fig. (E) Type V eclogite

Fig. (F) Type VI garnet glaucophanite

Fig. (G) Type VII garnet amphibolite

mann, 1985; Staudigel et al., 1996). The abundance of Na, Mg and K-rich minerals (glaucophane, phengite, chlorite, albite) as inclusions in garnet indicates that these components were present before high-P metamorphism. As there is little evidence that the garnet glaucophanite have been significantly affected by retrogression, they are suggested to be high-P metamorphic equivalents of hydrothermally altered basalts.

The Type VII garnet amphibolites are distinctly enriched in Fe and Ti (Figs. 2 and 3) and have trace element contents that are elevated with respect to NMORB (Fig. 4G). The compositions of these rocks are comparable to ferrobasalts (or Fe–Ti basalts) that were formed by differentiation of ocean floor tholeiites (Carmichael, 1964; Clague and Bunch, 1976; Wood, 1978; Fornari et al., 1983). In addition, Type VII samples plot close to or on the differentiation trend for MORB in Fig. 2.

Although we have demonstrated that the Type VII garnet amphibolites represent metamorphosed equivalents of fractionated MORB, there are some significant chemical differences between the samples. Protolith variations and pre-metamorphic processes are invoked to explain these differences. The protoliths of samples 80104 and 80105 are suggested to be differentiates of the Type I protolith: EMORB. Samples 72811 has remarkably similar trace element characteristics to the Type III eclogites, and hence

is likely to be a metamorphosed equivalent of fractionated NMORB. We suggest that fractionation of apatite prior to igneous crystallisation caused the phosphorus depletion in samples 80105. The high Mg content and positive Ce anomaly of 80105 is attributed to hydrothermal alteration by seawater prior to metamorphism (Mottl and Holland, 1978; De Baar et al., 1983).

We have plotted the concentrations of Be, B, Li, As, and Sb against elements of similar compatibility during mantle melting for a range of the mafic rocktypes of the Pouébo Terrane (Fig. 5). Be, Li, As, and Sb contents are within or at the high end of the range of values for oceanic basalts (MORB and OIB). B concentrations are higher than fresh MORB or OIB. However, as B is strongly affected by even low degrees seafloor alteration, oceanic crust may have highly elevated B contents (Leeman and Sisson, 1996). Therefore, rocks of the Pouébo Terrane are likely to have been affected by at least mild seafloor alteration.

5. Comparison with equivalent protolith rocks

It is clear that the high-P metamorphic rocks from the Pouébo Terrane are comparable with a variety of ocean floor mafic igneous rocks. However, a direct comparison with weakly or unmetamorphosed equiv-
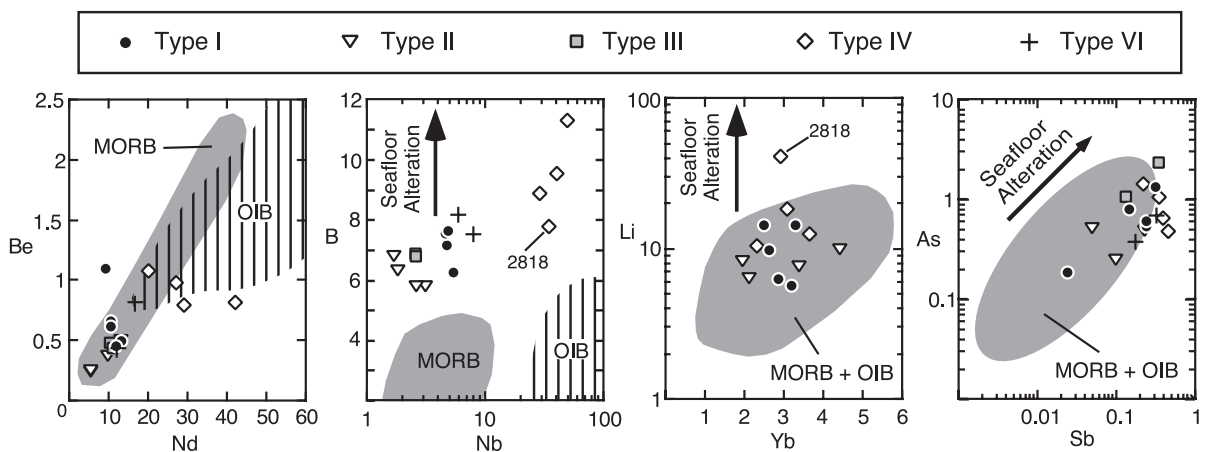

Fig. 5. Be, B, Li, As, and Sb systematics of high-P, low-T mafic rocks of the Pouébo Terrane. Data for the MORB and OIB fields is from Ludden and Thompson (1979), Chan et al. (1992), Dostal et al. (1996), Leeman (1996), Leeman and Sisson (1996), Noll et al. (1996), Teagle et al. (1996), Kamenetsky et al. (2000), Neal et al. (2002) and Ryan (2002).

alent rocks is required to qualitatively evaluate element mobility during subduction. Various igneous rock-types of the Poya Terrane are directly comparable to many of the mafic rock-types of the Pouébo Terrane. As a first step, we compare the trace element patterns of elements that are generally regarded to be immobile during metamorphism, such as REE and HFSE, in order to assign a probable protolith to each high-grade rock-type. On this basis, we will subsequently discuss the extent of depletion of fluid-mobile elements that is likely to have occurred during subduction zone metamorphism. The HFSE and REE composition of all the Type I eclogites and garnet amphibolites fall within the range of compositions of EMORBs from the Poya Terrane (Fig. 4A). EMORB and the Type I rocks are similarly widespread and abundant in the Poya and Pouébo Terranes, respectively, further supporting a genetic link between the terranes. Type II eclogites have BABB-like compositions and also closely compare to BABB from the Poya Terrane (Fig. 4B). HFSE and REE of the Type III eclogites compare with NMORB from the Poya Terrane (Fig. 4C). Alkaline mafic rocks are a minor component of both the Poya and Pouébo terranes and are also directly comparable in composition (Fig. 4D).

Type V–VII rocks of the Pouébo Terrane are not comparable to any rocks that have been described from the Poya Terrane. Therefore, the geochemical effects of high-*P*, low-*T* metamorphism on these rocks cannot be evaluated. Nonetheless, the chemical features of these rock types can be explained through processes of igneous fractionation and seafloor hydrothermal alteration, as described above.

The mafic rocks of the Pouébo Terrane represent the metamorphosed equivalents of a variety of igneous rock types. Many of these rock types may have belonged to a single igneous suite, but some are genetically unrelated. Interpretations of previous workers (Aitchison et al., 1995; Clarke et al., 1997; Cluzel et al., 2001) coupled with our estimates of peak metamorphic conditions (Table 1) indicate that the Pouébo Terrane represents a coherent slice of oceanic crust. A back-arc or marginal basin is the only tectonic environment in which igneous equivalents of all of the mafic rocks described above may be found (Saunders and Tarney, 1984; Eissen et al., 1994). This interpretation is also consistent with suggestions of a marginal basin origin for the Poya Terrane (Ali and Aitchison, 2000; Cluzel

et al., 2001). As the southwest Pacific region has undergone numerous episodes of opening and closing of marginal basins during the Cenozoic (Hall, 2002), a back-arc or marginal basin origin for the Pouébo Terrane is not unexpected.

6. Discussion

The geochemical data presented above demonstrate that a variety of mafic rocks comprise the Pouébo Terrane. The geochemical variation observed in the rock-types is attributed to four main processes: (1) variation in the original mantle sources (Types I, II, III, IV); (2) shallow level fractionation processes (Types V, VII); (3) hydrothermal alteration at the seafloor (Type VI); and (4) modifications during subduction zone metamorphism. As the Pouébo Terrane represents a slice of deeply subducted oceanic crust, these results have important implications for subduction-zone processes as discussed in detail below.

6.1. Heterogeneity of subducting oceanic crust

Many petrological (Liu et al., 1996; Kerrick and Connolly, 2001; Poli and Schmidt, 2002), geochemical (Hawkesworth et al., 1993; Becker et al., 2000), and geophysical (Hacker et al., 2003) studies of generic subduction-zone processes include the assumption that mafic rocks of subducted oceanic crust have a bulk chemical composition of MORB or altered MORB. By contrast, our data indicate that subducted oceanic crust may comprise a variety of mafic rock types. Many ophiolite sequences that have been obducted in collisional tectonic settings also contain a diverse range of mafic rocks (e.g. Ahmed and Ernst, 1999; Harper, 2003; Harper et al., 2003). Knowledge of the composition of currently subducting oceanic crust is limited to samples recovered from ocean drilling or dredging programs. Nonetheless, these programs have also revealed that there is significant chemical heterogeneity in the oceanic crust. Subducting back-arc basins such as the Sulu Sea Plate (Spadea et al., 1991), the Philippine Sea Plate (Wood et al., 1980; Tokuyama, 1985a,b), the Woodlark Basin (Perfit et al., 1987; Dril et al., 1997), the South China Sea (Wang et al., 1992), and the Molucca Sea Plate (Hall et al., 1991) all

contain a variety of mafic rock-types including MORB, BABB, island-arc tholeiites and alkaline basalts. Oceanic crust subducted beneath the New Britain and Solomon–Vanuatu arcs is expected to have complex tectonic origin (Hall, 2002) and therefore may also contain a variety of mafic rock-types. The proportion of alkaline basalt in oceanic crust formed at normal mid-ocean ridges is also much greater than commonly recognised. Alkaline basalts have been identified in ocean floor sampling of the Nazca Plate (Scheidegger et al., 1978; Scheidegger and Corliss, 1981), the Cocos Plate (Maury et al., 1982), the Indian Plate (Hekinian, 1974; Robinson and Whitford, 1974), and the Pacific Plate near the Marianas arc (Bogdanov et al., 1979; Plank et al., 2000). In addition, alkaline seamounts may locally comprise between 5% and 25% of oceanic crust (Wilson, 1989). Therefore, it is clear that the oceanic crust may contain a great diversity of mafic rocks.

Subduction of chemically (and isotopically) heterogenous oceanic crust has important consequences for models of arc magma genesis. Overall, MORB is likely to be the dominant mafic rock-type of oceanic crust, but alkaline basalts may contain orders of magnitude higher concentration of many incompatible elements relative to NMORB (Fig. 4D). Therefore, even minor amounts of alkaline basalt in the slab may significantly affect the budget of incompatible-element recycling from the crust to the mantle. The source of LILE enrichment in arc magmas is commonly regarded to be the sedimentary portion of the subducting slab (Ben Othman et al., 1989; Elliott et al., 1997; Plank and Langmuir, 1998). Alkaline basalts contain up to 100 times the amount of incompatible elements found in NMORB and thus even a small proportion of alkaline basalts in the slab should also be considered as a major source of incompatible elements to arc magmas. Calculations of the relative contributions from the downgoing slab to arc magmas is usually made assuming that subducting mafic rocks are of MORB composition (e.g. Hawkesworth et al., 1993; Class et al., 2000). We stress that these estimates may be grossly inaccurate if heterogeneity of the slab is not considered.

All of the mafic rock-types examined in this paper experienced similar conditions of peak metamorphism (Table 1), but have large variations in mineral assemblages and metamorphic facies. These variations are primarily controlled by bulk-rock composition and highlight the importance of using precise bulk-rock compositions for conducting detailed thermobarometric modelling (e.g. Marmo et al., 2002; Fitzherbert et al., 2003). Mafic rocks with high Mg/Fe^{2+} transform from blueschist to eclogite at significantly higher P and T conditions than rocks with lower Mg/Fe^{2+} (Evans, 1990). Therefore, it is expected that the Mg-rich Type VI garnet blueschists coexist with relatively Fe-rich eclogites and garnet amphibolites in the Pouébo Terrane. The blueschist–eclogite transition and associated fluid release (Peacock, 1993) will occur over a protracted range of P and T conditions in subducting oceanic crust that contains a range of mafic protoliths. Mg-rich hydrothermally altered mafic rocks may be particularly important for transporting volatiles (H_2O, Cl) to great depths (>70 km) in subduction zones (Ito et al., 1983; Peacock, 1993; Philippot et al., 1998).

6.2. Chemical modification during high-P, low-T metamorphism

It is widely accepted that material from the subducting slab is transferred to the source regions of arc magmas (Tera et al., 1986; Hawkesworth et al., 1993; Pearce and Peate, 1995). However, the processes involved in removing elements from the slab are poorly understood. It is commonly assumed that the fluid released by the breakdown of hydrous mineral in the slab also transfers 'fluid mobile' elements from the slab to the overlying mantle wedge (Hawkesworth et al., 1993; Pearce and Peate, 1995). Comparison between rocks of the Pouébo Terrane and weakly or unmetamorphosed equivalents indicates that in all samples key element ratios such as Th/U, U/Nb and Ce/Nb are unfractionated relative to their likely protoliths. The extremely good match between HFSE and REE of the high-grade rocks with their protoliths indicates that no significant mobility of these elements occurred during subduction zone metamorphism. Using this correlation between protoliths and high-grade metamorphic rocks, we can now evaluate possible removal of fluid-mobile elements, such as LILE, Be, B, Li, As and Sb, that are relatively enriched in arc magmas (Tera et al., 1986; Pearce and Peate, 1995; Dostal et al., 1996; Leeman and Sisson, 1996; Noll et al., 1996; Zack et al., 2003). It has been proposed that

deeply subducted rocks that have undergone dehydration are strongly depleted in these elements (Arculus et al., 1999; Bebout et al., 1993; Becker et al., 2000; Zack et al., 2003). The best evidence for element mobilisation is documented in sample 2818 which is significantly lower in LILE (70–90% depleted) and B (~ 20% depleted) and higher in Li (~ 300% enriched) compared to other Type IV rocks (Figs. 4D and 5). These features cannot be attributed to protolith variation and are most likely related to fluid alteration during subduction. The normalised pattern of this rock shows an accentuated negative K anomaly and a depression of the Cs, Rb and Ba trend by about a factor of 2–3 (Fig. 4D). Similar features are also present in some samples of Type I rocks (703, 704, 1001), in Type III eclogite and in sample 72811 of the Type VII garnet amphibolites. Although it is likely that at least parts of these features are related to metamorphism, it is extremely difficult to quantify. We have previously shown that the K and Rb anomalies for the Type III eclogites may be inherited from the protolith (Spandler et al., 2003). The large variation in LILE content in the Poya Terrane EMORBs is attributed to hydrothermal alteration and weathering (Cluzel et al., 2001). The range of LILE values of Type I eclogites remains within the range of Poya Terrane EMORB values (Fig. 4), so we cannot definitively attribute this variation to subduction-zone metamorphism.

The most important point of the presented data set is that most samples with originally high LILE contents such as the BABB and the alkaline basalts still have high LILE contents at eclogite-facies conditions (Fig. 4). Moreover, other fluid-mobile elements such as Be, B, Li, As and Sb are still present in high concentrations with respect to unaltered mafic protoliths (Fig. 5). If the samples have gained some of these elements during seafloor alteration, the subsequent loss during metamorphism must have been to a similar or smaller degree, as highlighted by B systematics (Fig. 5). The investigated rocks reached a depth of about 60 km and were situated in the fore-arc region at peak metamorphic conditions. Our data indicate that significant amounts of fluid-mobile elements are still present at these conditions and can be transported to subarc depths. Furthermore, the data suggests that the significant dehydration of mafic crust at the blueschist to eclogite-facies transition is not necessarily coupled

with a significant trace element release (e.g. Spandler et al., 2003).

The presented bulk-rock data of mafic oceanic crust shows that the chemical variation in eclogite-facies rocks is mainly determined by variations in the mantle source, the degree of differentiation and the amount of seafloor alteration. The effect of dehydration during prograde subduction-zone metamorphism results in much smaller chemical variations. Similar results were obtained from geochemical studies of high-P rocks from the Western Alps (Chalot-Prat et al., 2003).

Previous work on the geochemistry of high-P metamorphic rocks has concluded that subducting mafic rocks undergo extensive (50–98%) and ubiquitous loss of some incompatible trace elements (Sr, Rb, Ba, K, U) during metamorphism (Arculus et al., 1999; Becker et al., 2000). This is not supported by the findings of this study and Chalot-Prat et al. (2003). We emphasise that the element depletions calculated by Arculus et al. (1999) and Becker et al. (2000) were made using an assumed protolith composition. Our work benefits from having well-constrained igneous protoliths which allows for a more meaningful evaluation of chemical modification due to metamorphism. It is clear that the composition of the mafic protoliths of the Pouébo Terrane cover a large range of compositions, particularly for the incompatible elements. Therefore, we stress the importance of accurately defining the pre-metamorphic composition of high-grade rocks in order to evaluate chemical modification with metamorphism.

6.3. Towards a holistic model for element recycling in subduction zones

In this paper we have investigated the geochemical variation of subducting oceanic crust and the processes responsible for removing elements from the slab under high-P conditions. Other research has emphasised the critical role of subducted sediments in providing elements to arc magmas (Tera et al., 1986; You et al., 1996; Plank and Langmuir, 1998; Hermann and Green, 2001). It is clear that thermal conditions in the slab strongly influence the degree of element mobility in subducted rocks. Crustal rocks subducted in a high temperature subduction-

zone environment tend to undergo extensive depletion of incompatible trace elements at fore-arc depths (Bebout et al., 1999; Becker et al., 2000). By contrast, exhumed fore-arc rocks from relatively cold subduction zones (Bebout et al., 1999; Chalot-Prat et al., 2003; this study) are relatively undepleted in trace elements. Based on the growing body of literature on subduction-zone processes, we propose the following generic model for fluid and element release from slabs in a relatively cold (Marianas-type) subduction zone. The model is shown schematically in Fig. 6.

(1) On arrival at a convergent plate margin, cold oceanic crust is likely to contain a range of mafic rock-types that have been variably altered and hydrated by seafloor hydrothermal processes. A considerable thickness of sediment may overlie these mafic rocks. As the plate becomes dragged beneath the accretionary wedge, sediments may be accreted to or eroded from the overlying plate (von Huene and Scholl, 1991), leading to further lithological heterogeneity and structural complexity at the top of the subducting slab (Cloos and Shreve, 1988; Bebout, 1989). Compaction and diagenesis induce high pore-fluid pressures and upward fluid flow along fault zones in the accretionary wedge (Cochrane et al., 1996).

Fig. 6. Schematic cross-section of a generic cold subduction zone. The numbers refer to points discussed in the text.

(2) As the plate continues to descend into the mantle, mafic and sedimentary rocks recrystallise to lawsonite blueschist-facies mineral assemblages. However, some igneous and detrital minerals may persist to much greater depths due to the low temperatures of metamorphism (<400 °C; Spandler et al., 2003). Under these conditions, mafic rocks may contain in excess of 5 wt.% H_2O (Peacock, 1993; Poli and Schmidt, 1995). With increasing P and T, low-grade hydrous minerals break down and release fluid from the slab. B and LILE are highly mobile under hydrothermal conditions (You et al., 1996) and are enriched in pore fluids that are currently upwelling in active forearcs (Fryer et al., 1999; Benton et al., 2001; Mottl et al., 2002; Savov et al., 2002). It is clear that these elements are sourced from the subducting slab. You et al. (1996) have shown that low-T fluids are capable of stripping appreciable amounts of B and LILE from sedimentary rocks. By contrast, we propose that element liberation from mafic rocks in the slab is controlled by partitioning and hence is most effective in zones of elevated temperature and/or high fluid–rock interaction, such as the slab-mantle wedge interface (Bebout and Barton, 1993). Therefore, element depletion is likely to only occur in localised regions of the slab, whereas other portions of subducted rocks, especially those situated in low strain regions, are able to retain the majority of fluid mobile elements. Such a process creates further heterogeneity in deeply subducted crust.

(3) Fluid is continually released from the slab as it undergoes progressively higher grades of metamorphism. The most important dehydration and fluid release event in the slab occurs at the blueschist to eclogite transition (Peacock, 1993). Over this transition, mafic rocks may release over 2 wt.% H_2O (Peacock, 1993; Poli and Schmidt, 1995). However, in most slabs the transition will extend over a large depth range (40–100 km) due to thermal variations in the slab (Peacock, 1993), prolonged or continuous mineral dehydration reactions (Schmidt and Poli, 1998) and chemical heterogeneity in the slab. Furthermore, fluid flow under these conditions is likely to be highly restricted (Philippot and Selverstone, 1991; Philippot and Rumble, 2000; Scambelluri and Philippot, 2001). Our data show that the bulk of the mafic rocks do not suffer extensive element depletion at this stage. Therefore, significant amounts of the slab-
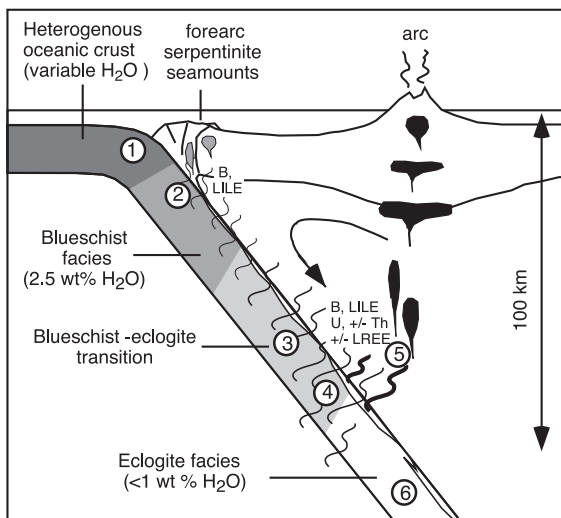
derived elements in arc magmas (e.g. LILE, Th, U, and LREE) may be transported down to sub-arc depths (70–150 km).

(4) At depths of 70–150 km, fluids in the slab may be derived from the breakdown of hydrous minerals in mafic rocks or antigorite in ultramafic rocks (Ulmer and Trommsdorff, 1995; Scambelluri et al., 2001). Only small volumes of fluid are sourced from pelitic rocks (Spandler et al., 2003; Hermann and Green, 2001). In zones of high T (>600 °C) these hydrous fluid may be capable of dissolving large quantities of elements from the mafic rocks (Scambelluri and Philippot, 2001; Spandler et al., 2003) or may induce partial melting of sedimentary rocks (Hermann and Green, 2001). As most rocks have not undergone extensive prior element depletion, the resulting fluids or melts are likely to contain high B, LILE, Th, U, and LREE contents and may cause extensive metasomatism during channelled migration through the slab (Sorensen and Grossman, 1989; Bebout and Barton, 1993, 2002).

(5) Melts and solute-rich fluids that escape the slab migrate across the overlying depleted mantle lithosphere as it convects downwards with the subducting plate (Pearce and Peate, 1995). At temperatures below the water-saturated mantle solidus, metasomatism may fix many elements transferred from the slab to the mantle. Much of this material may be recycled to deeper levels of the mantle and may be important for the genesis of back-arc magmatism. Infiltration of slab fluids into mantle domains that are above the water-saturated mantle solidus will cause partial melting and the formation of basaltic magmas that are enriched in B, LILE, Th, U and LREE.

(6) Once fully converted to eclogite facies, the slab has a higher density than the surrounding mantle and will descend to at least the 660 km discontinuity (Tatsumi and Eggins, 1995). Despite extensive dehydration, the slab can retain small amounts of H_2O in high-P minerals such as phengite, lawsonite and phase A (Bose and Ganguly, 1995; Poli and Schmidt, 1995; Ono, 1998). Breakdown of these phases at high P may contribute to magma generation in back arc regions. The slab is likely to be chemically and structurally heterogenous due to initial protolith variations, high-P deformation and metasomatism, partial melting and fluid-induced depletion. This heterogeneity may be transferred to mantle plumes during recycling of slab materials in the lower mantle and may help to explain the large variations in isotopic compositions observed in OIB (Wilson, 1989).

7. Conclusions

We used the high-P, low-T metamorphic rocks (1.9 GPa and 600 °C) of northern New Caledonia as an analogue of subducted oceanic crust to investigate the geochemical characteristics of subducting slabs. The geochemical affinities of the mafic rock-types include EMORB, BABB, NMORB, alkaline basalt, plagioclase-rich cumulate, seafloor-altered basalt and Fe–Ti basalt. This variety of rock types may only be found in a back arc or marginal basin tectonic setting. Nonetheless, we propose that subducting oceanic crust generally contains a greater variety of mafic rock-types than is commonly assumed. This heterogeneity has important implications for the recycling of fluid and trace element through subduction zones.

Direct comparison between the high-grade mafic rocks and equivalent protolith rocks from the nearby Poya Terrane or other igneous mafic rock-types indicate that most of the high-grade rocks have not experienced significant geochemical modification by high-P, low-T metamorphism. The lack of extensive element depletion in the slab at depths below the fore-arc of cold subduction zones indicates that key elements (e.g. LILE) may be efficiently transferred to depths below the arc before migrating into the mantle wedge. This indicates that the transfer of elements from the subducting plate to mantle wedge does not simply occur in response to mineral dehydration and fluid release from the slab. We suggest that elements are removed from the slab in regions of elevated temperature and intense fluid–rock interaction.

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