The Island Arc (2005) **14**, 272–294

Research Article

Complete mantle section of a slow-spreading ridge-derived ophiolite: An example from the Isabela ophiolite in the Philippines

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Abstract The Isabela ophiolite shows a complete ophiolite sequence exposed along the eastern coast of northern Luzon, the Philippines. It forms the Cretaceous basement complex for the northeastern Luzon block. This ophiolite is located at the northern end of a trail of ophiolites and ophiolitic bodies along the eastern margin of the Philippine Mobile Belt. This paper presents new findings regarding the nature and characteristics of the Isabela ophiolite. Peridotites from the Isabela ophiolite are relatively fresh and are composed of spinel lherzolites, clinopyroxene-rich harzburgites, depleted harzburgites and dunites. The modal composition, especially the pyroxene content, defines a northward depletion trend from fertile lherzolite to clinopyroxene-rich harzburgites and more refractory harzburgites. Variation in modal composition is accompanied by petrographic textural variations. The chromium number of spinel, an indicator of the degree of partial melting, concurs with petrographic observations. Furthermore, the Isabela ophiolite peridotites are similar in spinel and olivine major-element geochemistry and clinopyroxene rare earth-element composition to abyssal peridotites from modern mid-oceanic ridges. Petrological and mineral compositions suggest that the Isabela ophiolite is a transitional ophiolite subtype, with the fertile lherzolites representing lower sections of the mantle column that are usually absent in most ophiolitic massifs. The occurrence of the fertile peridotite presents a rare opportunity to document the lower sections of the ophiolitic mantle. The variability in composition of the peridotites in one continuous mantle section may also represent a good analogy of the melting column in the present-day mid-oceanic ridges.

Key words: mantle column, mantle peridotite, ophiolite, partial melting.

INTRODUCTION

In Southeast Asia, multiple basin spreading and closing, plate collision, subduction and large-scale faulting have led to major tectonic rearrangement, at least during the last 50 Ma (Hall 1996). As a result, blocks of various ages and origins are believed to have been brought together in a complex belt of collision zones and island arcs along the western Pacific rim, from the southwestern Japanese islands down to New Zealand (Yumul *et al.* 2003). Located near the central portion of the western Pacific rim, the Philippine Islands are a perfect example of a complex plate boundary, composed of allochthonous terranes and hosting numerous ophiolites and ophiolitic rocks (Gervasio 1967; Balce *et al.* 1979; Wolfe 1983; McCabe *et al.* 1985; Karig *et al.* 1986; Mitchell *et al.* 1986; Sarewitz & Karig 1986; Geary *et al.* 1988; Yumul *et al.* 1997; Yumul *et al.* 2003). However, Encarnación (2004) suggested that these terranes in the north-

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Received 12 January 2005; accepted for publication 24 May 2005.

ern Philippines are autochthonous and were formed from episodic generation of oceanic crust within the Philippine island arc system. Ophiolite complexes, often regarded as the basement of island arcs, provide important information about the original tectonic setting of each terrane. The controversy regarding the continuity or discontinuity of the basement of the Philippines, the presence or absence of correlation between basement rocks of adjacent terranes and the age and origin of each ophiolite (Karig *et al.* 1986; Geary *et al.* 1988; Deschamps *et al.* 2000; Encarnación 2004) suggests that the study of ophiolites in the Philippines is still at a youthful stage. Advances in the study of ophiolites in the Philippines can contribute to the overall understanding of the western Pacific evolution. This paper provides new data from, and interpretations of, the Isabela ophiolite, a Cretaceous ophiolite exposed in the eastern margin of northern Luzon that is one of the key elements in understanding the evolution of the Philippines Mobile Belt (Figs 1,2).

A train of ophiolites and dismembered and metamorphosed ophiolitic fragments line the eastern margin of the Philippines (Fig. 1). They are grouped into two belts; namely, the Casiguran Ophiolite Belt and the Eastern Bicol–Eastern Mindanao Ophiolite Belt (Balce *et al.* 1979, quoted in Yumul *et al.* 1997) (Fig. 1). Together, the two belts cover the entire length of the eastern margin of the Philippine Mobile Belt. The Isabela ophiolite is the northern termination of this line of ophiolites and ophiolitic fragments (Fig. 1). Aside from being the northernmost exposed basement rock in the region, it also possesses the most pristine and extensive exposure of mantle peridotites within the belt. It is one of the few (three at present) mid-oceanic ridge (MOR)-type ultramafic bodies in a supra-subduction zone (SSZ)-dominated ophiolitic basement in the Philippine archipelago (Yumul *et al.* 1997). Contrary to its importance, it has been one of the least studied ophiolite sequences in the Philippines.

Among the earliest work on the Isabela ophiolite is the 1976 master's thesis of G. Balce who coined the name Casiguran Ophiolitic Belt, which refers to the chain of ultramafic rock bodies exposed from Divilacan to Baler (Billedo *et al.* 1996) (localities indicated in Fig. 3). Later studies by Aurelio and Billedo (1987, unpubl.), RPJP (1987) and Billedo *et al.* (1996) recognized that the ultramafic rocks (Isabela ultramafic complex), along with associated lithologies (massive and layered gabbros, dike and pillow basalt complex and capping sedimentary units), define a complete ophiolite sequence (Billedo *et al.* 1996). This upper mantle–crust sequence was named the Casiguran ophiolite.

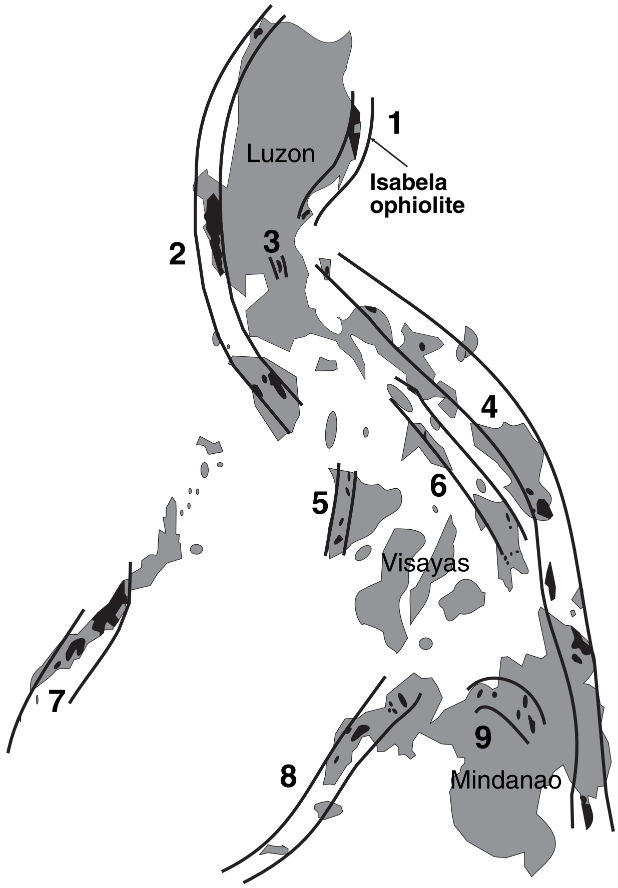

Fig. 1 Map of the Philippine archipelago showing the major ophiolite belts (reproduced from Yumul *et al.* 1997, adopted from Balce *et al.* 1979 and BMG 1982). (1) Casiguran Ophiolite Belt; (2) Zambales Ophiolite Belt; (3) Angat Ophiolite Belt; (4) Eastern Bicol–Eastern Mindanao Ophiolite Belt; (5) Antique Ophiolite Belt; (6) Western Bicol–Eastern Leyte Ophiolite Belt; (7) Palawan Ophiolite Belt; (8) Zamboanga–Sulu Ophiolite Belt; and (9) Central Mindanao Ophiolte Belt. Black areas represent ophiolite/ophiolitic bodies.

The following discussion describes the geology, petrography and geochemistry of the crust–upper mantle rock sequence exposed from Dinapigui to Divilacan, all within the province of Isabela, the Philippines (hereafter referred to as the Isabela ophiolite). Special emphasis will be given to the petrography and geochemistry of the mantle peridotites. The discussion on the geology, petrology and geochemistry is based on the results of fieldwork conducted in December 2002, January 2003 and March 2004, with input from published reports. This is the first of a series of reports dealing with the detailed study of the Isabela ophiolite.

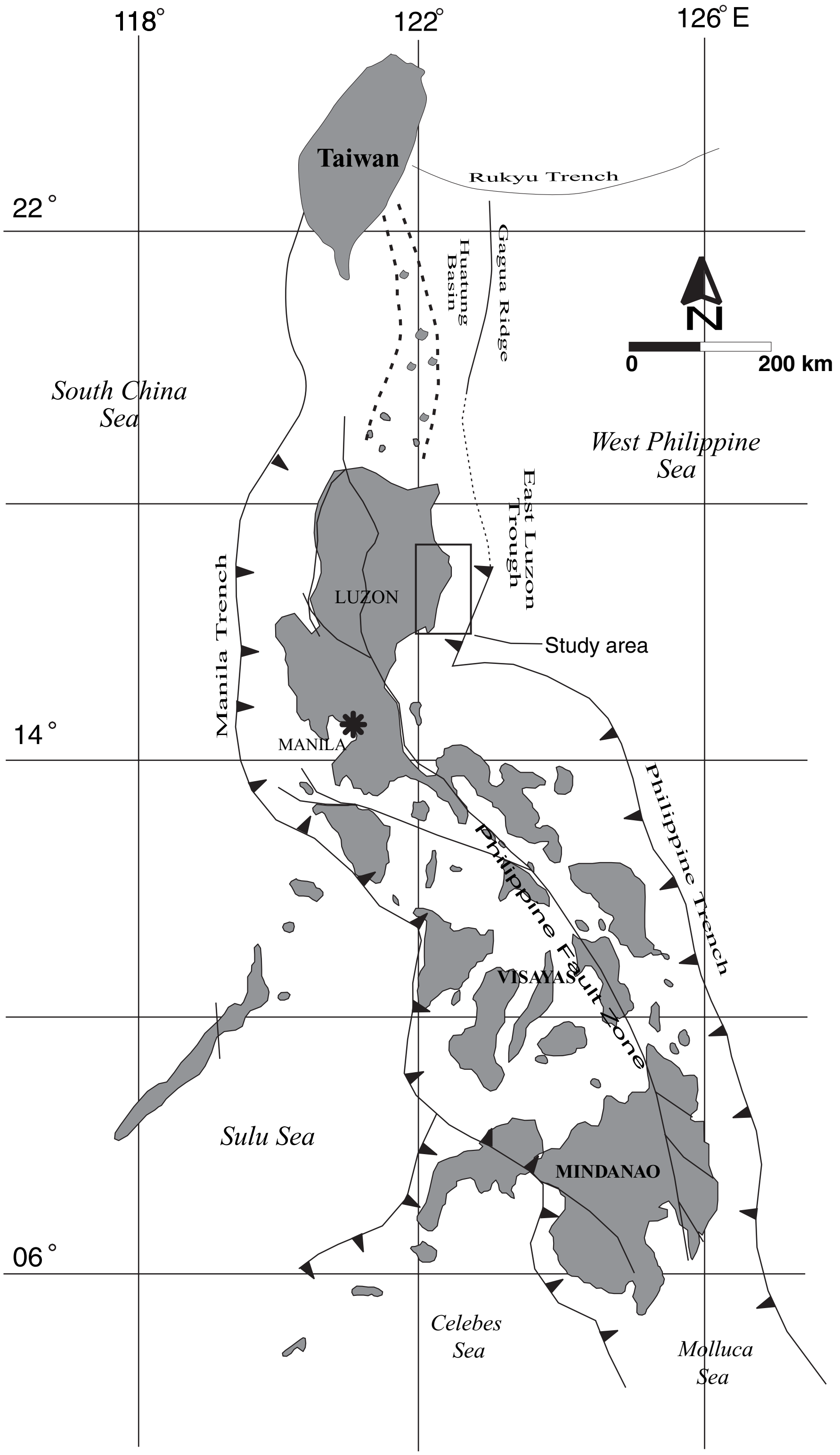

Fig. 2 Map showing the Philippine archipelago bounded on both sides by two subduction zone systems and marginal basins. It is divided into two parts by the sinistral Philippine Fault. The heavy broken line is the approximate trace of the Luzon Arc. The study area is indicated by the box. The location of Manila is marked by an asterisk for reference.

GEOLOGY

REGIONAL TECTONIC SETTING

The study area is located in the eastern boundary of northern Luzon (Fig. 1). Northern Luzon is composed of at least four major morphological subdivisions emplaced against each other in a parallel to subparallel fashion. From west to east, they are the Ilocos Basin, Central Cordillera Mountain Range, Cagayan Valley Basin and the Northern Sierra Madre Mountain Range (NSM).

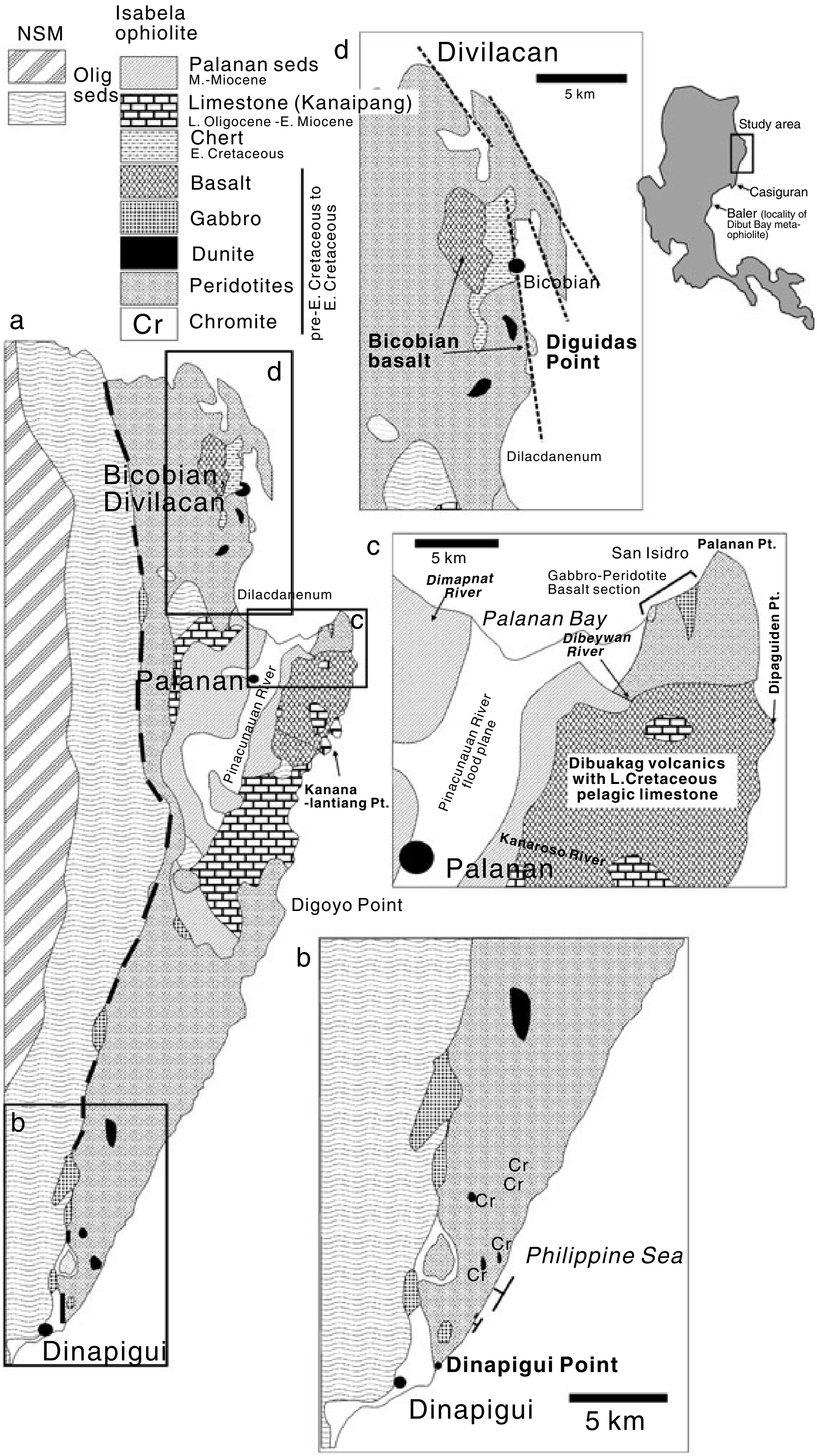

Fig. 3 Simplified geology of the Isabela ophiolite and adjacent Northern Sierra Madre Mountain Range (NSM) (modified from RPJP (1987), including input from Billedo *et al.* (1996) and field survey data). Surveyed areas are indicated by the boxes. (a) Isabela ophiolite; (b) Dinapigui area; (c) Palanan area; and (d) Divilacan area. The heavy broken line in (a) is the approximate trace of the Divilacan Fault. The broken lines in (d) represent the approximate trace of faults and the filled circle marks the location of Sitio Tagbak. Olig, Oligocene; seds, sediments.

Two subduction zones of opposite polarities sandwich the whole of northern Luzon; the Manila Trench located to the west and the East Luzon Trough to the east. The East Luzon Trough is approximately 75 km east of the study area (Fig. 2).

The Isabela ophiolite is exposed in the ear-like landmass protruding along the eastern flank of the NSM (Fig. 2). The rock unit is elongated and oriented approximately north–south along the eastern coast of northeastern Luzon, within the enclosure of ~16°30′ to 17°25′N and ~122°15′ to 122°32′E (Figs 2,3a). Approximately 90 km in length, the main mass of the ophiolite body occupies the eastern boundary of Isabela province, from the towns of Divilacan and Palanan to Dinapigui (Fig. 3). Metamorphosed equivalents of the Isabela ophiolite are exposed in Dibut Bay (Dibut Bay meta-ophiolite) near Baler; small fragments are also exposed in the Casiguran area (BMG 1982; Billedo *et al.* 1996) (locations indicated in Fig. 3). All of the above constitute the Casiguran Ophiolitic Belt (Balce *et al.* 1979) (Fig. 1). Ringenbach (1992) suggested that the trail of metamorphic rocks along the Casiguran Ophiolitic Belt might be a remnant of a major shear zone.

The Isabela ophiolite is bound by the NSM to the west. The NSM is interpreted to be in thrust contact with the Isabela ophiolite through the Divilacan Fault (Billedo *et al.* 1996). It is made up of volcaniclastics of Middle to Late Eocene age, intruded by several pulses of arc-related magmas from the Middle to Late Eocene, Oligocene and Miocene periods (Aurelio 2000). Essentially, the Isabela ophiolite serves as a basement to a mature island arc.

GEOLOGY OF THE ISABELA OPHIOLITE

A large portion of the Isabela ophiolite is made up of the Isabela ultramafic complex (Fig. 3a). It is well exposed from Dinapigui to Divilacan. Extensive coastline exposures with high walls of moderate to densely faulted ultramafic rocks reaching more than 50 m characterize the Dinapigui area (Fig. 3b). In Palanan, the peridotites, together with basalt and gabbros, have limited exposure along coastlines in the base of low-lying ridges (Fig. 3c). In Divilacan, mantle peridotites are also well exposed from Dilacdanenum to Bicobian (Fig. 3d). Pillow lavas and bedded chert are also exposed in Bicobian.

DINAPIGUI AREA

From field observations, the peridotites exposed in Dinapigui can be divided mainly into two varieties, homogeneous lherzolites and layered peridotites (Fig. 4a,b, respectively). A very homogeneous lherzolite body occurs in the southernmost tip of the study area (Dinapigui Point) (Figs 3b,4a). Resistant pyroxene crystals protrude from the surface of the rocks prominently, giving an estimate of how much richer this unit is in terms of two-pyroxene compared to other peridotites in Dinapigui (Fig. 4a). This feature is also highlighted by the absence of mineral foliation, showing no preferred crystal elongation direction and layering. Cross-cutting veins are also very scarce. This lithology is estimated to extend approximately 0.5–0.7 km northward (referred to as Dinapigui Point lherzolites hereafter; petrological and geochemical characteristics will be discussed later).

The more dominant peridotite variety (referred to as Dinapigui peridotites hereafter) is characterized by prominent mineral foliation. Well-formed pyroxene-rich (lherzolitic) and pyroxene-poor (harzburgitic) bands can be traced continuously in the extensive exposures (Fig. 4b). The foliation plane generally has a northeast strike, dipping northwest at variable angles (from nearly horizontal to nearly vertical, mostly about 30°) (Fig. 3b). Some reversal in the orientation of foliation can be observed in the landslide-disrupted sections. The estimated thickness of the exposed foliated peridotites, at least for the surveyed area in the southern part, is 400–450 m.

Very limited occurrences of plagioclase peridotite were identified. White-colored plagioclase psuedomorphs occur in narrow zones/bands (1–3 cm) cutting the foliation of the peridotites.

Billedo *et al.* (1996) reported on the samples collected from Digoyo (Fig. 3a), located between Palanan Point and Dinapigui Point. The petrography of the samples yielded lherzolites and olivine websterites.

Both concordant and discordant dunite and websterite veins cross-cut the peridotites. Dunite veins concordant to mineral foliation are seen either as layers with uniform thickness or lensoid-shaped pods. The thickness of the veins varies from a few centimeters to 20–30 cm. Discordant dunite veins vary in thickness from a few centimeters to a few meters (3–5 m) and show a replacive nature. This is especially evident where relatively thick dunite veins cut the concordant pyroxenite layers, and

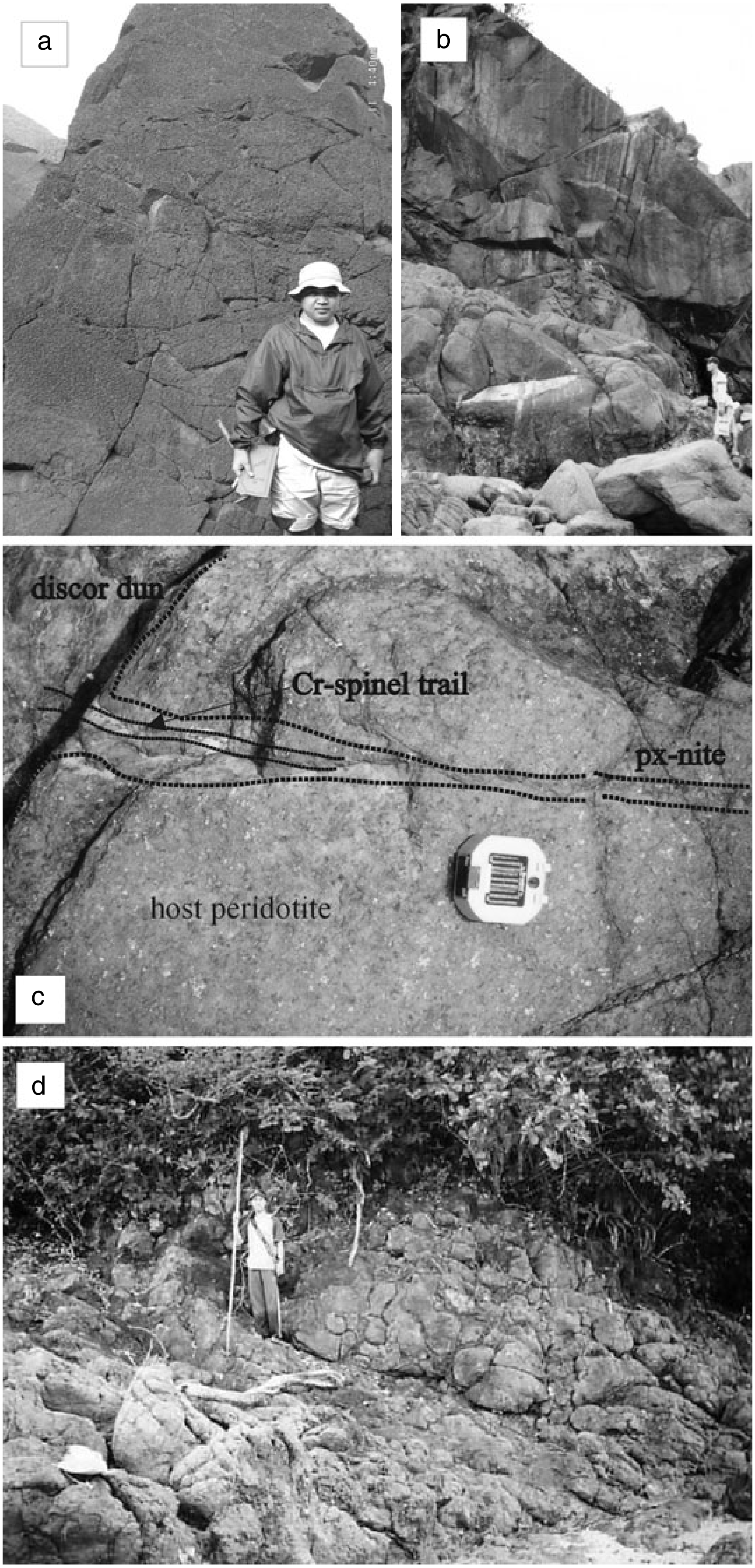

Fig. 4 (a) A section of the Dinapigui Point lherzolites showing homogeneous texture. (b) Typical layered peridotites in Dinapigui. Dip angle of foliation varies from nearly horizontal to nearly vertical. (c) Replacive discordant dunite (discor. dun.) cutting the host peridotite spreading laterally towards the pyroxenite (px-nite) vein. A trail of chromian spinel (Cr-spinel) is left behind by the reaction of the melt and the pyroxenite vein. (d) Pillow basalt exposed in San Isidro, Palanan. The pillows are undeformed and in an upright position.

the dunite branches outward through the pyroxenite. Trails of Cr–spinel nodules were left behind by a reaction along the trace of the original pyroxenite layer (Fig. 4c) (Kelemen *et al.* 1995). Occasional pods of chromitite (maximum thickness of 10–15 cm) can also be observed within thick (meters-wide) discordant dunites. These are late-stage dunites as they cut across the foliation and earlier discordant pyroxenite veins. Occasional gabbroic dikes were also observed cutting Dinapigui peridotites. The dike thickness varies from more than 1 cm to 5 cm.

Aside from the extensive faulting and jointing along the entire length of the coastal exposure, plastic deformation features, such as sinusoidal to overturned folds of peridotite foliation and boudinage structures, were observed near the boundary between the Dinapigui Point lherzolites and Dinapigui peridotites. Along the ridge crest above the same area, mylonitized peridotite boulders can be observed.

Layered gabbros were observed as boulder-sized floats embedded in the thick soil (lateritic) profile which cap the ultramafic bodies in Dinapigui. Boulders and cobbles of gabbros were also observed as a significant component of the beach sediments in the area, strongly suggesting that they were derived from eroded materials from above the ridge. Unfortunately, there was no direct field evidence showing how the peridotites and the gabbros are related. RPJP (1987) also reported gabbro bodies along the suture zone, sandwiched between the NSM-folded sedimentary rock and the peridotite massif of the Isabela ophiolite, along the trace of the Divilacan Fault.

Five chromite deposits are known to exist in the southern part of the Isabela ophiolite (Fig. 3b). Outcrops show that the dunite bodies hosting the chromite deposits have intruded the pyroxene-rich harzburgite host discordantly. It is not possible to estimate the size of the chromitite pods as they have already been mined out; however, the largest mine visited during the survey extends not more than 300 m. This estimate includes the dunite envelope. Chromitite is found mainly as the nodular variety.

PALANAN AREA

Palanan approximates the geographical center of the Isabela ophiolite (Fig. 3c). RPJP (1987) and Billedo *et al.* (1996) both produced a geological map of the area; however, their versions have significant differences, especially in the area where the Dibuakag volcanics and Kanaipang limestone are exposed. The map by Billedo *et al.* (1996) was also referenced in the review of northern Philippine ophiolites by Encarnación (2004). A section of the Isabela ophiolite exposed along the southern coast of Palanan Bay is highlighted in this study (Fig. 3c).

Along the southern coast of Palanan Bay, mantle peridotites, isotropic gabbro and pillow basalts are found side by side in fault contact with each other (Fig. 3c). Foliated lherzolites, pyroxene-rich harzburgites and harzburgites, with occasional dunite and pyroxenite layers, make up the mantle section. The boundary between the Isabela ultramafic complex and the overlying basalts is interpreted to be a thrust fault contact (Billedo *et al.* 1996).

In Palanan, the gabbros and pillow lavas represent the crust section of the ophiolite. Isotropic, very coarse-grained to pegmatitic gabbros were encountered in San Isidro, Palanan (Fig. 3c). Grain-size variation is clearly shown in the outcrops. They are exposed along the base of low-lying north-northwest ridges terminating along the coastline.

From the sudden change in lithology over a very short distance, it is inferred that the gabbros, peridotites and pillow basalts are in fault contact with each other (Fig. 3c). Streams/rivers mark the boundaries between lithologies. Although proximal to the Dibuakag volcanics (discussed in the next paragraph), the pillow lava unit in San Isidro, Palanan, is mapped as part of the main ophiolite body and not as part of the former.

The Dibuakag volcanics unit is exposed in Dipaguiden and Kananalatiang Point, east of Palanan. They are found interstratified with calcareous and siliceous limestone of Late Cretaceous age (Billedo *et al.* 1996). Along the banks of Kanaroso River, weathered outcrops of this unit are found interrupted by another set of limestone unit. Billedo *et al.* (1996) suggested that this younger limestone unit (Kanaipang limestone; Early to Late Miocene period) unconformably overlies the pillow lavas. The Dibuakag volcanics unit is younger (K–Ar age of 87.15 ± 5.82 Ma) than the Bicobian basalt and unconformably overlies the Isabela ophiolite (Billedo *et al.* 1996). Selected samples showed mid-oceanic ridge basalt–island arc tholeiite (MORB–IAT) affinity (Billedo *et al.* 1996). Furthermore, Billedo *et al.* 1996 suggested that the fault contact between the Isabela ultramafic complex and the Dibuakag volcanics is covered under the Kanaipang limestone and the Palanan sediments (Early to Late Miocene sediments).

DIVILACAN AREA

In Divilacan, extensive outcrops of mantle peridotites are uncovered along the stretch from Dilacdanenum to Bicobian (Fig. 3d). The peridotites are mainly harzburgites and clinopyroxene-rich harzburgites with less lherzolites. More than half of the surveyed outcrops in the area are massive and nearly homogeneous. Foliated peridotites occur in some areas. The mineral foliations measured in Bicobian Peninsula are mainly oriented N46–76°E; however, the dip differs on opposite sides of the peninsula.

An increase in the amount and thickness of dunite pods relative to the southern part was noted in Divilacan. Thick (4–5 m wide) discordant dunite pods were observed (Fig. 5a). Although in some outcrops, the relationship between the thick dunite and the host peridotites is not clear, these dunite bodies occur discordant to the foliation of the host peridotites in several areas. Associated with these thick dunite pods are websterite veins (approximately 25 cm thickness) with coarse crystals of two-pyroxenes. These websterite veins cut both the dunite pods and the host peridotites.

An increase in the abundance and thickness of gabbroic veins was also noted in Divilacan. The cross-cutting relationship suggests a multiple intrusion of gabbroic veins as they cut both the host peridotites and older gabbroic veins. A series of parallel gabbroic veins (approximately 3–7 cm thickness) intrude into the massive host harzburgites in Dilacdanenum, Divilacan (Fig. 5b).

A patch of pillow basalt crops out in Diguidas Point, Divilacan (Figs 3d,5c). This pillow lava unit is probably in fault contact with the adjacent peridotites because of the sudden change in rock type and the linearity of the peridotite ridge (Fig. 3d). This inferred fault can be traced seaward through a series of linear islets. Basalt

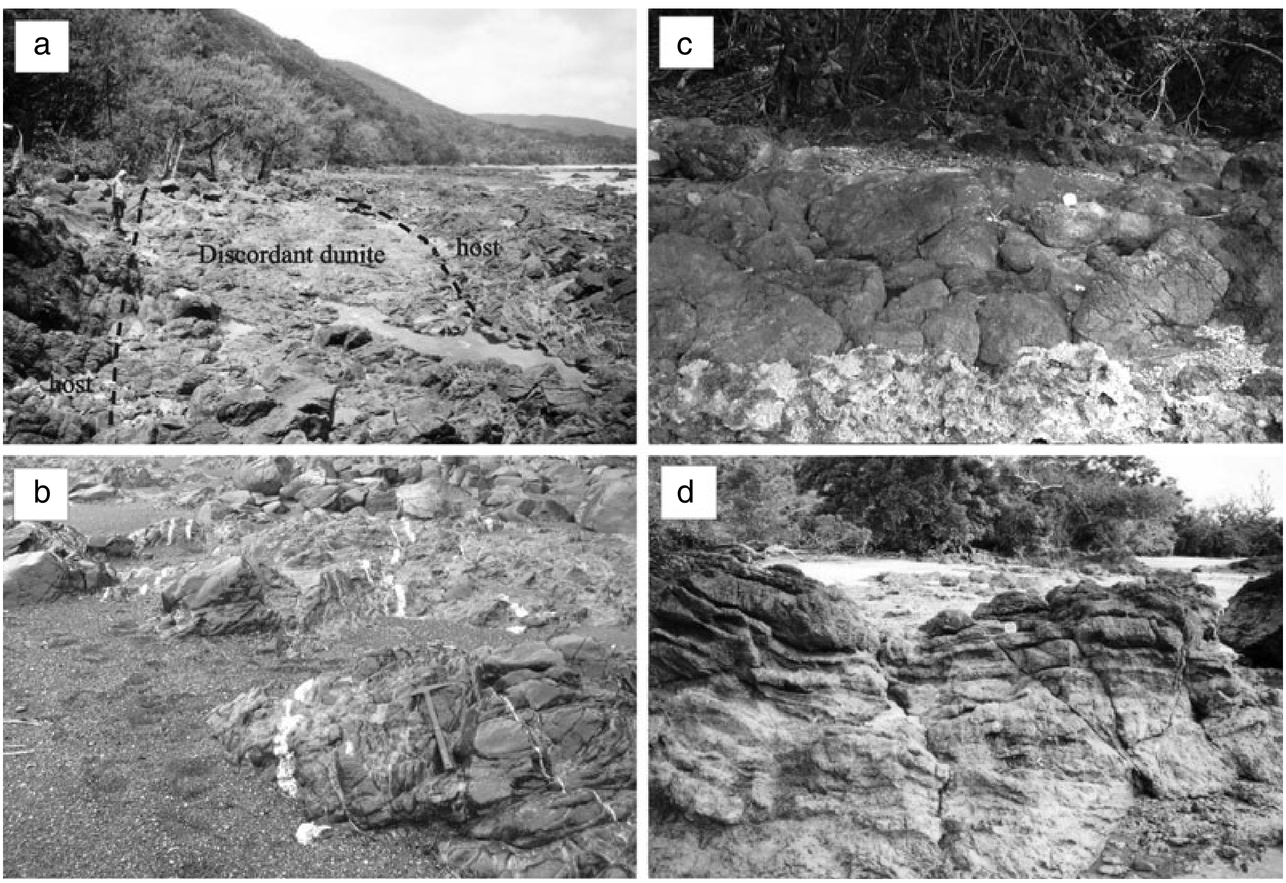

Fig. 5 (a) Thick (4–5 m wide) discordant dunite cutting mantle peridotites in Divilacan. Similar-sized dunite veins are more common in Divilacan than in Dinapigui and Palanan. (b) Parallel gabbroic veins cutting mantle peridotites in Divilacan. Gabbroic veins are more common and thicker in Divilacan than in the southern part of the ophiolite. (c) Undeformed pillow basalt (part of the Bicobian basalt) exposed in Diguidas Point. (d) Bedded red chert extensively exposed along Port Bicobian, Divilacan, interpreted as the first sedimentary cover of the Isabela ophiolite.

occurs as brecciated lavas and well-formed pillow lavas (Fig. 5c). RPJP (1987) reported pillow lavas in Bicobian, Divilacan. Aurelio and Billedo (1987, unpubl.) referred to this basalt unit as the Bicobian basalt. In the same report, they mentioned the occurrence of a dike complex associated with the Bicobian basalt in Bicobian, Isabela.

An extensive outcrop of bedded reddish and brown chert is exposed along the coastline in Sitio Tagbak, Bicobian, Divilacan (Figs 3d,5d). The exposure is approximately 750 m long and the beds are oriented N44°E, dipping 40°NW. RPJP (1987) geological maps show sedimentary units exposed in Port Bicobian and in the upstream portion of Dikinamaran River. This unit is interpreted as the first sedimentary cover of the Isabela ophiolite and was shown to be of Early Cretaceous age by fossil dating (Aurelio & Billedo 1987, unpubl.). The sedimentary package was assigned the name 'Dikinamaran River pelagics' (Billedo *et al.* 1996). Linear features evident in topographical maps and the highly sheared and serpentinized character of peridotites found in Port Bicobian provide evidence of a fault boundary between the chert and peridotite (Fig. 3d). Aurelio and Billedo (1987, unpubl.) observed the Bicobian basalt in thrust contact with the chert in Bicobian, Isabela (the locality was not specified by the authors).

Chromite mineralization occurs at two localities in Divilacan. Although the actual outcrops are inaccessible, a stockpile of samples from these deposits was available in Port Bicobian. The collected samples contained nodular-type chromitite, similar to the chromite deposits found in Dinapigui.

PETROGRAPHY

PERIDOTITES

Samples from the mantle section of the Isabela ophiolite collected from Dinapigui, Palanan and Divilacan represent the southern, middle and northern sections of the ophiolite, respectively (Fig. 3a). Figure 6 shows the modal composition of peridotites from the three localities. Note the trend in rock type from fertile lherzolites to residual harzburgites and dunite depicting a gradual decrease of the modal amount of pyroxenes (especially clinopyroxenes).

The Dinapigui peridotites include lherzolites, pyroxene-rich harzburgites, clinopyroxene-poor and depleted harzburgites, dunites, plagioclase peridotites and wehrlites, in decreasing order of

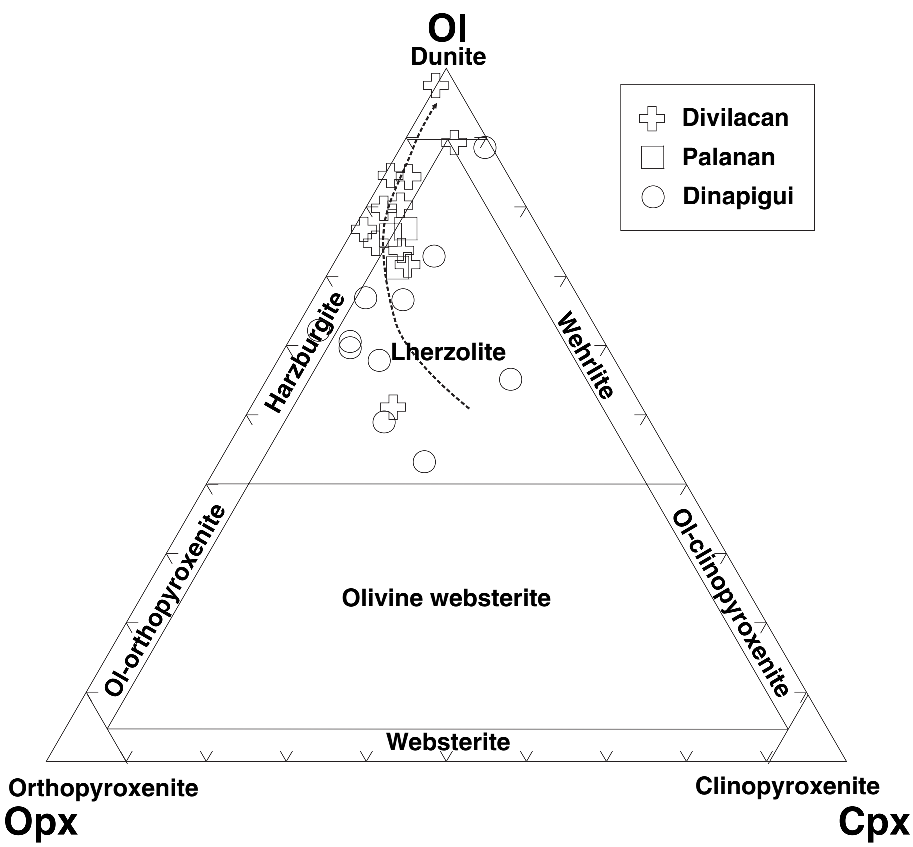

Fig. 6 Modal composition of olivine (Ol)–orthopyroxene (Opx)–clinopyroxene (Cpx). The broken arrow approximates the transition path from fertile lherzolite to residual harzburgite and dunite.

abundance. Dinapigui Point lherzolites have a clinopyroxene modal composition greater than 15%, which differentiates them from other lherzolites in Dinapigui. Other than the difference in modal clinopyroxene, their petrographic characteristics are nearly identical. Lherzolites exposed in Dinapigui are relatively fresh and mainly exhibit primary porphyroclastic texture (Fig. 7a). Olivine crystals show typical un-oriented fractures and reach more than 4 mm in size. They show selective alteration into serpentine and magnetite along grain boundaries. Neoblastic olivine also occurs as interstitial grains between coarse pyroxene porphyroclasts (Fig. 7a). The porphyroclasts of pyroxenes reach a maximum of 5–6 mm across in size and are often subhedral to anhedral with irregular grain boundaries. Signs of plastic deformation are evident in the pyroxene porphyroclasts showing bent lamellae and kink banding. Chromian spinel is present with less than 4% modal abundance, and is pale red to yellowish green in thin sections. Spinel is subhedral, anhedral or vermicular, and can be as coarse as approximately 2 mm. Spinel crystals are closely associated with pyroxenes (Fig. 7b), and in some instances completely surrounded by orthopyroxene porphyroclasts. Olivine and clinopyroxene inclusions occur in some spinel. The presence of Ti-rich pargasite (1.5–3.7 wt% of TiO_2) was noted in the Dinapigui Point lherzolites. It makes up less than 2% of the modal composition and shows a pale brown to green pleochroic color. Ti-rich pargasite occurs as discrete crystals and does not show any replacement textures, which suggests that it is a primary component of this peridotite.

Plagioclase lherzolite occurs in Dinapigui with very limited distribution. The rocks are moderately serpentinized and only pseudomorphs of plagioclase remain. Plagioclase pseudomorphs occur in limited narrow bands (about 1–3 cm). Individual grains range from about 0.5 to 2 mm in size, showing elongated shapes with irregular boundaries. Spinel grains within the narrow zones are reddish in color and are commonly surrounded by plagioclase. Plagioclase can also be found associated with olivine and clinopyroxene. Clinopyroxene crystals range in size from approximately 5 to 7 mm outside the plagioclase rich-zone and from 1 to 3 mm within the zone. Petrographic evidence strongly suggests melt infiltration and impregnation as the origination of the plagioclase lherzolites.

The peridotites exposed in San Isidro, Palanan, are mainly pyroxene-rich harzburgite, lherzolites, harzburgites and dunite, in decreasing order of abundance. Lherzolites from Palanan have a lower

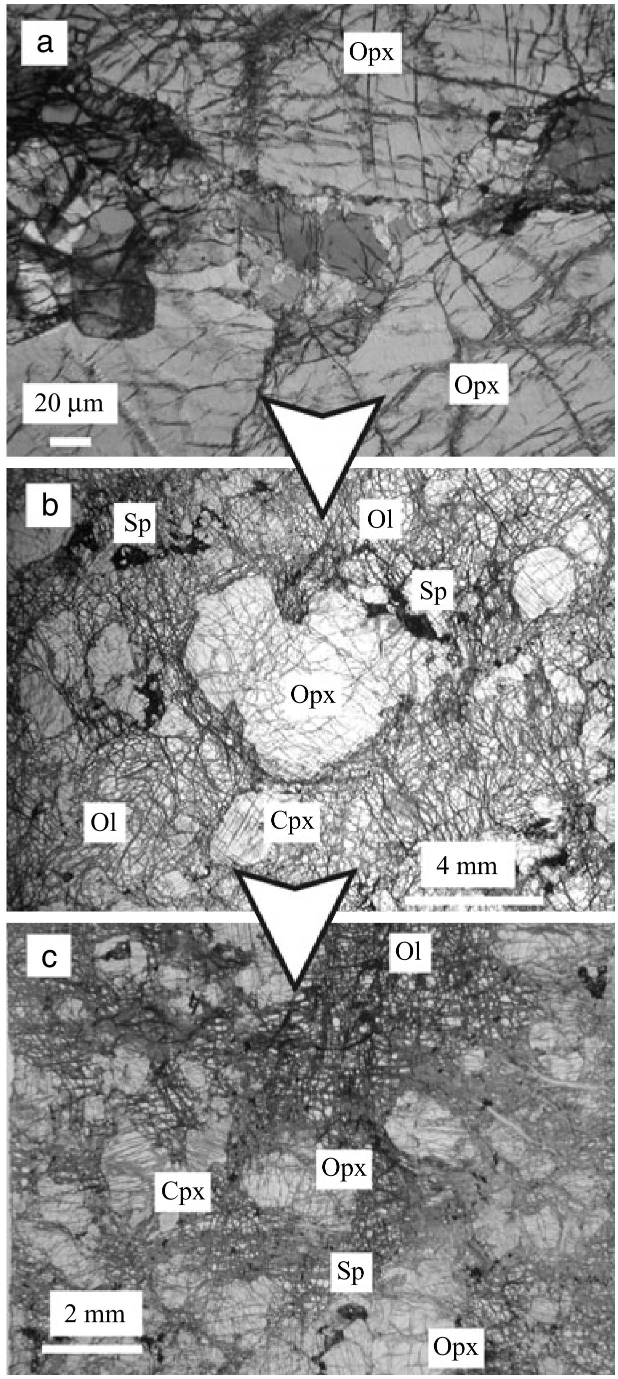

Fig. 7 (a) Photomicrograph of a fertile lherzolite from Dinapigui showing primary porphyroclastic orthopyroxene (Opx) with interstitial olivine (Ol). Smaller neoblastic silicate minerals are also shown. (b) Photomicrograph of a fertile lherzolite from Dinapigui showing coarse-grained porphyroclastic orthopyroxene, clinopyroxene (Cpx) and granular olivine. Cr-spinel (Sp) are usually associated with the two-pyroxene. (c) Photomicrograph of a clinopyroxene-rich harzburgite from Divilacan showing the secondary porphyroclastic to granular texture of silicate minerals. Notice the considerable change in grain size from (a) to (b). Parts (b) and (c) show the evolution from primary to secondary textures.

clinopyroxene modal composition (5.1–7.3%) compared to those from Dinapigui. The weathering effects are also more severe compared to the peridotites in the Dinapigui area. The clinopyroxene-rich harzburgite has slightly less modal clinopyroxene than the lherzolites, but has nearly the same secondary porphyroclastic texture. The orthopyroxene porphyroclasts reach a maximum size of 1–2 cm. Clinopyroxene is less altered than the orthopyroxene and olivine crystals and is mainly observed as subhedral to anhedral crystals varying from 1 mm to less than 1 mm in size. Bent lamellae are commonly observed in the pyroxene crystals. Like the Dinapigui lherzolites, the samples exhibited an alignment of orthopyroxene and clinopyroxene crystals to form bands.

Chromian spinel (0.8–4.1% modal composition) varies in size from micrometers to as coarse as 5 mm across, the former being more common. Most of the observed crystals have subhedral to anhedral shapes; however, some smaller ones are euhedral with well-defined boundaries. The color under a plane-polarized view varies from reddish brown to dark red. Some of the coarse-grained spinels have small pyroxene and olivine inclusions. Coarse Cr-spinel grains in harzburgites have an elongation parallel to the mineral banding shown by the pyroxenes.

Along the stretch of Divilacan, most of the peridotites are clinopyroxene-rich harzburgites and harzburgites with fewer lherzolites. They can be found in both undifferentiated and foliated forms. Thick discordant dunites and clinopyroxenite to websterite veins cut these host rocks. The degree of serpentinization is variable depending on the rock type, with the harzburgite and dunite being highly to completely serpentinized. The lherzolites and clinopyroxene-rich harzburgites are relatively well preserved, and the texture varies from secondary porphyroclastic to equigranular (Fig. 7). Coarse porphyroclasts of pyroxenes are still observed, especially in the lherzolites and clinopyroxene-rich harzburgites. However, peridotites with smaller porphyroclasts (<2 mm) are more common. Spinel grains also vary in size and form, and are dominantly dark reddish in color.

GABBRO

The layered gabbro exposed in Dinapigui is composed of plagioclase, olivine and clinopyroxene, with minor orthopyroxene and amphibole. The sample was fresh orthopyroxene-bearing olivine gabbro with mesocumulate texture. Olivine and plagioclase were the earliest to crystallize among the cumulus phases, followed by pyroxenes. The grain size is about 2 mm on average, and can reach 5 mm. Clinopyroxenes are mainly augite and plagioclase feldspars are labradorite. Some of the plagioclase crystals are partly altered.

The isotropic gabbros exposed in Palanan have highly variable composition and grain size. The composition ranges from orthopyroxene-bearing olivine gabbro (<1 wt% of orthopyroxene) to gabbronorite. The grain size of cumulus plagioclase, clinopyroxene and orthopyroxene ranges from less than 1 mm to 2–3 cm in pegmatitic sections. The pegmatitic part is mainly gabbronorite. The very coarse-grained sections also show little or no intercumulus minerals (adcumulate texture). Furthermore, the plagioclase show pyroxene inclusion, which suggests that pyroxene crystallized ahead of plagioclase. The isotropic gabbros show evidence of low-temperature metamorphism. Clinopyroxene is partly to completely replaced by hornblende and fibrous amphibole. Although some thin sections show fresh olivine, it is often completely replaced by chlorite–actinolite and only a relict texture remains. Plagioclase is also partly altered with associated prehnite. This altered mineral assemblage is similar to that of gabbros drilled from the southwest Indian Ridge (Meyer *et al.* 1989). Billedo *et al.* (1996) reported cumulate gabbro exposures along the Pinacunauan and Dimapnat rivers (Fig. 3c). The petrology of the olivine gabbro suggests that they can be related to the Bicobian basalt (discussed separately).

BASALT

The pillow lavas observed in San Isidro, Palanan, are mostly porphyritic basalt. Phenocrysts of plagioclase feldspars are euhedral to subhedral in shape and range from 0.3 mm up to 0.5 mm in length, with larger, less abundant crystals reaching 1.25 mm. Groundmass is composed mostly of fine-grained plagioclase laths, pyroxene and magnetite with interstitial glass in an intergranular texture.

Doleritic basalts, in contrast, are composed of interlocking crystals of plagioclase laths, subhedral clinopyroxene, magnetite and amphiboles, showing an intersertal to hyaloophitic or subophitic texture. The plagioclase crystals have labradorite composition and range in size from 0.3 to 0.7 mm, with occasional coarse crystals reaching

1.5 mm in length. Clinopyroxene crystals are euhedral to subhedral in shape. The crystals range in size from 0.3 to 0.5 mm and are clustered with the plagioclase laths. The magnetite occurs in the form of subhedral to euhedral crystals ranging in size from 0.2 to 0.3 mm. Brown glass occupies the spaces between the coarse-grained crystals. The amount of interstitial glass decreases with the overall increase in the grain size of the primary minerals. Relict olivine phenocrysts with preserved Cr-spinel inclusions are also present in these basalts.

As mentioned earlier, the basalts exposed in Diguidas Point occur as well-preserved pillow lavas and brecciated lavas. The pillow lavas are similar in petrography to those exposed in San Isidro, Palanan, which were described previously. In contrast, brecciated lavas show a glomeroporphyritic texture, with clusters of coarse (1–3 mm) subhedral to euhedral plagioclase crystals in aphanitic groundmass. These plagioclase phenocrysts contain rounded to subhedral Cr-spinel inclusions.

MINERAL CHEMISTRY

The major-element mineral chemistry was analyzed using a JEOL electron microprobe (JEOL JXA8800) at the Center for Cooperative Research in Kanazawa University (acceleration voltage of 20 kV, probe current of 2.0×10^{-8} A and probe diameter of 3.0 μm). Only core compositions of minerals were used in this study. The magnesium number (Mg#) is calculated using Mg/(Mg + Fe_{total}) and Mg/(Mg + Fe^{2+}), which are the atomic ratios for silicates and chromian spinel, respectively. The calculation of the ferrous and ferric iron composition of spinel is based on chromian spinel stoichiometry. The chromium number (Cr#) corresponds to Cr/(Cr + Al), which is the atomic ratio in spinel. The representative analysis of samples is presented in Table 1.

Clinopyroxene rare earth element (REE) compositions were determined using a quadrupole inductively coupled plasma–mass spectrometer (Agilent 7500s) outfitted with a MicroLas GeoLas Q-Plus laser microprobe (Ishida *et al.* 2004) at the Center for Cooperative Research in Kanazawa University. Diamond-polished thin sections of the peridotites were used for the analysis. The ablation time was set at 50 s with an ablation spot diameter of 50 μm, and National Institute of Standards and Technology (NIST) standard reference materials (SRM) 612 and 614 glasses were used as standards. A detailed description of the instrumentation is given by Ishida *et al.* (2004).

MAJOR-ELEMENT MINERAL CHEMISTRY

Olivine displays a narrow forsterite (Fo) composition from 88 to 92 for all peridotites (Fig. 8). Although the majority of the samples have an Fo composition of 90–92 (Fo_{90-92}), Dinapigui Point lherzolites contain the lowest Fo content (Fo_{88-90}). The spinel Cr# varies widely, from as low as ~0.08 to ~0.7. Dinapigui Point lherzolites contain spinels with very low Cr#, ranging from ~0.08 to ~0.16. The bulk of the peridotites have spinel Cr# ranging from ~0.17 to ~0.43. The spinel Cr# increases abruptly with a small change in olivine Fo, depicting a residual trend in the olivine–spinel mantle array (OSMA) (Arai 1994). The spinel chemistry of the Isabela ophiolite peridotites covers the entire range of abyssal peridotites (Fig. 9) defined by Dick and Bullen (1984). Some dunite from Divilacan, chromitites and a representative chromitite-hosting discordant dunite have a Cr# between ~0.56 and ~0.7. The Fe^{3+} content of spinel in all peridotites is generally low (Fig. 10).

Clinopyroxene major elements (especially Na) classify the Isabela ophiolite mantle rocks into Na-rich (1.4–2.0 wt%) and Na-poor (0.0–0.8 wt%) peridotites. Some samples from Dinapigui plot between the two major groups. Na-rich clinopyroxenes have a higher Ti content compared to the Na-poor group (Fig. 11a). Coexisting spinel of the Na-rich clinopyroxenes is aluminous whereas Na-poor clinopyroxenes are associated with more chromian spinel (Fig. 11b).

Primary amphiboles in the fertile peridotites are mainly pargasitic in composition (Table 1). A relatively high Ti content was noted (between 1 and 4 wt%), thus classifying them as Ti-rich pargasite.

RARE-EARTH ELEMENT GEOCHEMISTRY OF CLINOPYROXENE

The chondrite-normalized REE spider diagram for clinopyroxene shows nearly flat patterns for the heavy rare earth elements (HREEs), and progressive depletion of the light rare earth elements (LREEs) (Fig. 12). In general, the REE content of clinopyroxene decreases constantly from Dinapigui toward Divilacan. The spider diagrams are nearly parallel to each other and encompass

Table 1 Representative analysis of minerals from the Isabela ophiolite peridotites

Locality	Dinapigui									Palanan			
Rock type	Spinel lherzolite					Spinel lherzolite				Spinel lherzolite			
Sample no.	DP-02121702					DP-02121809				PL-03012103A			
	Sp	Cpx	Opx	Ol	Amp	Sp	Cpx	Opx	Ol	Sp	Cpx	Opx	Ol
SiO_2	0.00	51.65	55.28	41.21	43.24	0.00	51.99	54.57	40.69	0.07	52.56	56.38	40.89
TiO_2	0.04	0.51	0.08	0.00	2.42	0.05	0.56	0.11	0.01	0.02	0.08	0.04	0.00
Al_2O_3	57.98	7.53	4.27	0.00	13.93	59.65	7.04	5.31	0.00	46.55	4.51	2.79	0.02
Cr_2O_3	9.05	0.86	0.33	0.00	0.86	7.75	0.65	0.39	0.06	20.75	1.10	0.44	0.04
FeO*	11.15	2.45	6.32	9.97	4.02	11.66	2.75	6.76	10.23	14.17	2.37	6.00	9.30
MnO	0.09	0.12	0.15	0.14	0.07	0.10	0.10	0.13	0.14	0.21	0.11	0.15	0.14
MgO	21.17	13.96	33.17	49.30	16.97	20.11	15.03	31.72	48.51	18.02	16.33	32.94	49.36
CaO	0.00	20.52	0.43	0.02	11.43	0.00	20.02	0.53	0.03	0.02	23.70	0.44	0.01
Na_2O	0.00	1.67	0.05	0.00	3.77	0.01	1.62	0.07	0.02	0.00	0.08	0.01	0.01
K_2O	0.02	0.01	0.01	0.00	0.17	0.03	0.02	0.03	0.00	0.01	0.00	0.04	0.01
NiO	0.37	0.02	0.07	0.36	0.10	0.39	0.05	0.08	0.37	0.21	0.05	0.10	0.39
Total	99.87	99.29	100.15	101.00	96.98	99.75	99.82	99.69	100.05	100.03	100.89	99.31	100.17
O =	4	6	6	4	23	4	6	6	4	4	6	6	4
Si	0.000	1.878	1.906	1.001	6.194	0.000	1.880	1.894	1.000	0.002	1.895	1.954	0.999
Ti	0.001	0.014	0.002	0.000	0.261	0.001	0.015	0.003	0.000	0.000	0.002	0.001	0.000
Al	1.757	0.323	0.173	0.000	2.351	1.811	0.300	0.217	0.000	1.491	0.192	0.114	0.000
Cr	0.184	0.025	0.009	0.000	0.097	0.158	0.019	0.011	0.001	0.446	0.031	0.012	0.001
Fe^{2+}	0.187	0.075	0.182	0.203	0.482	0.227	0.083	0.196	0.210	0.267	0.071	0.174	0.190
Fe^{3+}	0.053	NA	NA	NA	NA	0.025	NA	NA	NA	0.055	NA	NA	NA
Mn	0.002	0.004	0.004	0.003	0.008	0.002	0.003	0.004	0.003	0.005	0.003	0.004	0.003
Mg	0.811	0.757	1.705	1.785	3.623	0.772	0.810	1.641	1.777	0.730	0.877	1.702	1.798
Ca	0.000	0.800	0.016	0.000	1.754	0.000	0.776	0.020	0.001	0.001	0.915	0.016	0.000
Na	0.000	0.117	0.003	0.000	1.048	0.000	0.114	0.005	0.001	0.000	0.005	0.001	0.001
K	0.000	0.001	0.000	0.000	0.030	0.000	0.001	0.001	0.000	0.000	0.000	0.002	0.000
Ni	0.008	0.001	0.002	0.007	0.012	0.008	0.001	0.002	0.007	0.005	0.001	0.003	0.008
Total	3.002	3.993	4.003	2.999	15.860	3.003	4.002	3.993	3.000	3.001	3.994	3.983	3.000
Mg#	0.813	0.910	0.903	0.898	0.883	0.773	0.907	0.893	0.894	0.732	0.925	0.907	0.904
Cr#	0.095	NA	NA	NA	NA	0.080	NA	NA	NA	0.230	NA	NA	NA
Fo	NA	NA	NA	89.81	NA	NA	NA	NA	89.42	NA	NA	NA	90.44

Elements Ca, Na and K were not included in the calculation of spinel stoichiometry.
Amp, amphibole; Cpx, clinopyroxene; NA, not applicable; Ol, olivine; Opx, orthopyroxene; Sp, spinel.

the abyssal peridotite field defined by Kelemen *et al.* (1995).

DISCUSSION

LITHOLOGIC DISTRIBUTION

Various studies have considered Cr-spinel as a good indicator of host rock petrogenesis, and it is a common understanding that Cr# of spinel in residual peridotite increases with an increase in the degree of partial melting (Dick & Bullen 1984; Arai 1994). Similarly, silicate minerals, such as olivine, also become refractory with an increase in the degree of partial melting. Stratification of the oceanic upper mantle, from fertile lherzolite to depleted harzburgite, mainly as a result of partial melting, has long been established (Ringwood 1975; Spray 1989; Ishikawa *et al.* 2004).

Figure 13 shows a correlation between the modal composition of clinopyroxene and Cr# of coexisting spinel in the peridotites. This indicates that the degree of depletion of the peridotite generally increases from the south toward the north. Peridotites from the Isabela ophiolite represent a column of oceanic lithospheric mantle emplaced right-side up toward the north. The systematic northward depletion of the peridotites corresponds to an ideal mantle column (Ringwood 1975; Spray 1989; Ishikawa *et al.* 2004). The trend observed in the modal composition is corroborated by the gradual textural evolution of the peridotites, from a primary porphyroclastic texture in the lherzolites to secondary textures in the harzburgites (Fig. 7b,c), indicating progressive mantle-related deformation. The systematic northward increase in the occurrence and size of gabbroic, pyroxenite and dunite veins also suggests that the northern part of the ophiolite represents the upper sections of the mantle column. This explanation is consistent with the fact that the pillow lavas, dike complex and sedimentary carapace are only exposed in the central to the northern parts of the ophiolite.

It is noteworthy that a peridotite believed to be host to a chromitite body in Bicobian contains

Table 1 *Continued*

								Divilacan						Dinapigui Chromitite Dp-03110454
Spinel lherzolite PL-03012110				Spinel lherzolite PL-03012205				Spinel lherzolite Dv-03190407				Dunite PL-03012203		
Sp	Cpx	Opx	Ol	Sp	Cpx	Opx	Ol	Sp	Cpx	Opx	Ol	Sp	Ol	Sp
0.00	51.32	54.60	40.60	0.00	52.03	55.22	40.53	0.00	52.02	55.05	40.70	0.09	40.67	0.00
0.08	0.26	0.06	0.00	0.07	0.07	0.00	0.00	0.01	0.17	0.07	0.00	0.08	0.00	0.27
51.16	5.55	5.02	0.00	38.29	3.97	3.58	0.02	51.84	5.04	4.15	0.00	14.09	0.00	19.10
15.92	1.00	0.69	0.01	28.91	1.20	0.82	0.00	15.79	0.98	0.56	0.00	54.34	0.00	50.50
12.25	2.47	5.70	8.50	15.35	2.32	5.83	8.74	12.50	2.63	5.91	9.19	19.50	7.99	16.91
0.17	0.08	0.12	0.14	0.41	0.02	0.14	0.13	0.09	0.07	0.10	0.10	0.82	0.15	0.31
19.37	15.70	32.84	50.43	16.18	16.57	33.51	50.50	19.10	16.55	33.21	50.47	11.48	50.82	13.44
0.02	23.43	0.96	0.02	0.01	23.18	0.70	0.01	0.00	22.78	1.29	0.02	0.05	0.01	0.01
0.01	0.42	0.02	0.01	0.01	0.05	0.01	0.00	0.00	0.24	0.03	0.00	0.00	0.00	0.04
0.00	0.00	0.02	0.00	0.02	0.02	0.00	0.02	0.01	0.02	0.01	0.01	0.01	0.02	0.01
0.32	0.04	0.09	0.38	0.18	0.06	0.09	0.40	0.31	0.03	0.07	0.41	0.04	0.38	0.08
99.29	100.27	100.11	100.09	99.43	99.49	99.91	100.34	99.64	100.53	100.45	100.91	100.51	100.03	100.68
4	6	6	4	4	6	6	4	4	6	6	4	4	4	4
0.000	1.865	1.884	0.990	0.000	1.901	1.909	0.988	0.000	1.881	1.896	0.988	0.003	0.991	0.000
0.002	0.007	0.002	0.000	0.001	0.002	0.000	0.000	0.000	0.005	0.002	0.000	0.002	0.000	0.006
1.610	0.238	0.204	0.000	1.285	0.171	0.146	0.001	1.626	0.215	0.168	0.000	0.531	0.000	0.696
0.336	0.029	0.019	0.000	0.651	0.035	0.023	0.000	0.332	0.028	0.015	0.000	1.375	0.000	1.234
0.228	0.075	0.164	0.173	0.311	0.071	0.169	0.178	0.241	0.080	0.170	0.187	0.446	0.163	0.383
0.046	NA	NA	NA	0.055	NA	NA	NA	0.037	NA	NA	NA	0.076	NA	0.054
0.004	0.002	0.003	0.003	0.010	0.001	0.004	0.003	0.002	0.002	0.003	0.002	0.022	0.003	0.008
0.771	0.850	1.689	1.834	0.687	0.902	1.727	1.835	0.757	0.892	1.705	1.826	0.548	1.845	0.619
0.000	0.912	0.035	0.001	0.000	0.907	0.026	0.000	0.000	0.882	0.047	0.001	0.002	0.000	0.000
0.001	0.030	0.002	0.001	0.001	0.003	0.001	0.000	0.000	0.017	0.002	0.000	0.000	0.000	0.002
0.000	0.000	0.001	0.000	0.000	0.001	0.000	0.001	0.000	0.001	0.001	0.000	0.000	0.001	0.000
0.007	0.001	0.002	0.007	0.004	0.002	0.003	0.008	0.007	0.001	0.002	0.008	0.001	0.007	0.002
3.002	4.009	4.005	3.010	3.003	3.996	4.007	3.012	3.002	4.002	4.012	3.012	3.004	3.010	3.002
0.772	0.919	0.911	0.914	0.688	0.927	0.911	0.912	0.759	0.918	0.909	0.907	0.551	0.919	0.618
0.173	NA	NA	NA	0.336	NA	NA	NA	0.170	NA	NA	NA	0.721	NA	0.639
NA	NA	NA	91.36	NA	NA	NA	91.15	NA	NA	NA	90.73	NA	91.90	NA

spinel with Cr# of 0.11. However, little is known about the location of this sample. It was collected from a stockpile of rocks quarried from a chromite mine located some 6 km southeast of Bicobian. This highlights the possibility that there are other fertile lherzolite bodies exposed in the northern part that have not been documented yet.

ORIGIN OF THE ISABELA OPHIOLITE PERIDOTITES

The main part of the mantle section of the Isabela ophiolite has a spinel, clinopyroxene and olivine mineralogy similar to abyssal peridotites. The spinel Cr# provides evidence of this similarity (Figs 8,9).

This similarity is further supported by the REE composition of clinopyroxene (Fig. 12), which encompasses the entire field of abyssal peridotites defined by Kelemen *et al.* (1995). The spider diagram shows the typical REE pattern of abyssal peridotites from a MOR. The REE content of the clinopyroxenes steadily decreases from the fertile lherzolites to the residual harzburgites.

Deschamps *et al.* (2000) suggested the continuity of the oceanic plate from the Huatung Basin to the Luzon Arc (Fig. 2). Dating of the Huatung Basin revealed an Early Cretaceous age, similar to the age of the Isabela ophiolite. If indeed they are continuous, the Isabela ophiolite may be correlated as an exposed cross-section of the Huatung Basin lithosphere. It is not within the scope of this paper to correlate the Huatung Basin with the Isabela ophiolite. However, this is a feasible concept, and hence will be the subject of future research.

Geological facts about the Isabela ophiolite suggest the probability of a slow-spreading ridge derivation. The dominance of lherzolites and clinopyroxene-rich harzburgites is consistent with this idea (Boudier & Nicolas 1985; Niu & Hëkinian 1997). Slow-spreading ridge ophiolites are usually lherzolite-dominated in contrast to fast-spreading ridge-derived ophiolites, which are usually harzburgite-dominated (Boudier & Nicolas 1985). A fast-spreading ridge ophiolite usually has a well-formed dike complex, such as the perfect dike

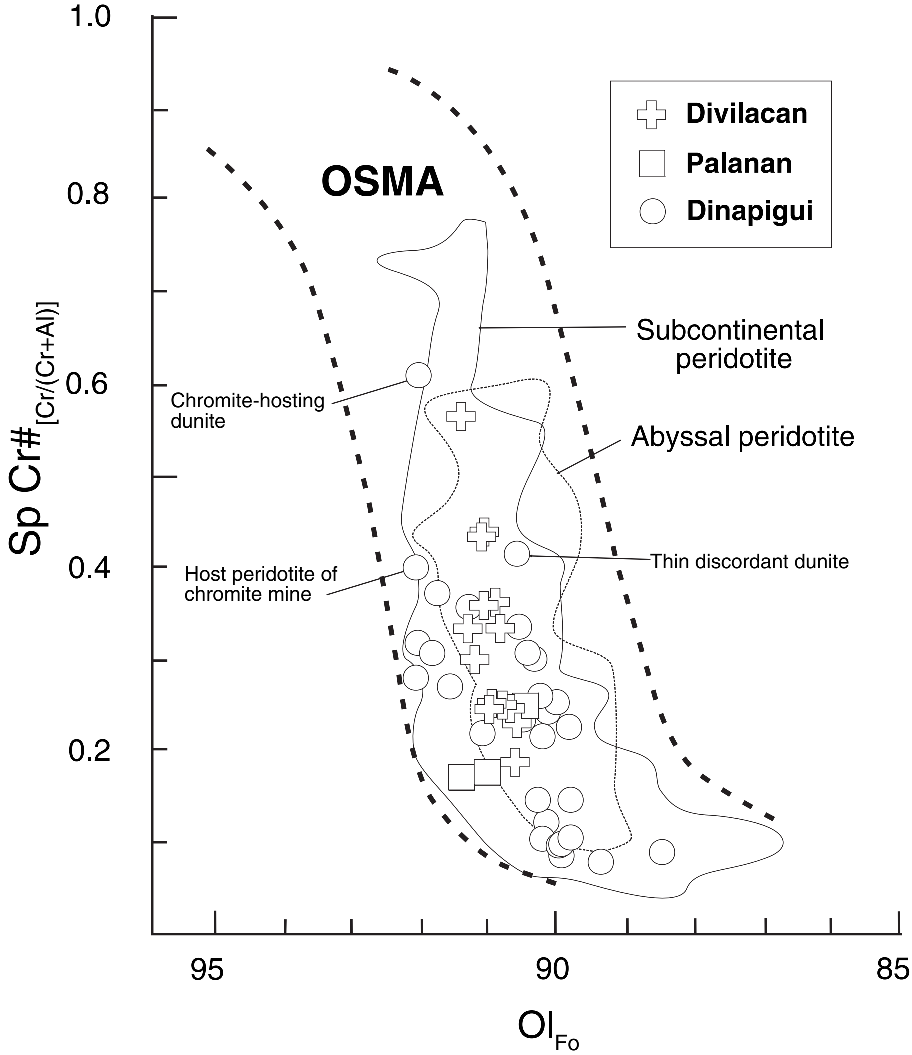

Fig. 8 Mantle peridotites from the Isabela ophiolite define a residual trend in the olivine (Ol)–spinel (Sp) mantle array (OSMA) (Arai 1994). They cover most of the abyssal peridotite field, with the exception of some low-Cr# spinels which plot within the subcontinental field. Fo, forsterite.

complex of the Oman ophiolite (Nicolas 1989). The Isabela ophiolite only has a thin dike complex exposed in Bicobian, Isabela, along with the Bicobian basalt, as reported by Billedo *et al.* (1996). Submersible observations of the ocean floor in the St Paul Fracture Zone in the Mid-Atlantic Ridge, classified as a slow-spreading ridge, recorded a series of lithological changes (Hekinian *et al.* 2000). Alternating occurrences of peridotites, isotropic gabbro and volcanic rocks (within 4 km) were all found to be in fault contact with each other, perpendicular to the ridge axis. This sequence is strikingly similar to the geology of San Isidro, Palanan, where isotropic pegmatitic gabbro, foliated mantle peridotites and pillow lavas are in fault contact with one another and were observed within a ~2.5-km traverse (Fig. 3c). It is difficult to prove if the correlation is true or just a coincidence. Tamayo *et al.* (2001; pers. comm.) compared the olivine Mg# (0.89–0.90) with the Al_2O_3 content (2.98–6.24 wt%) of coexisting orthopyroxene from two lherzolite samples from Dinapigui and a harzburgite (olivine Mg# of 0.90 and orthopyroxene Al_2O_3 content of 2.67 wt%) from Tariktik Point (about 20 km south of Dinapigui, near Casiguran). They found that the lherzolites had a composition similar to that of slow-spreading ridge peridotites.

Mineral chemistry suggests that the late-stage (based on cross-cutting relationships) discordant dunites and associated chromitites with high-Cr# spinels may not be related to ocean floor products, but arc-related processes (Fig. 10). This implies a switch in the tectonic environment, from abyssal to arc region, during the emplacement (Arai *et al.* 2004). A review of the mineralization potential of arc basement rocks in the Philippines suggests that SSZ-related ophiolites are more likely to hold economically viable deposits (Yumul *et al.* 1997). This holds true for the Isabela ophiolite. Although

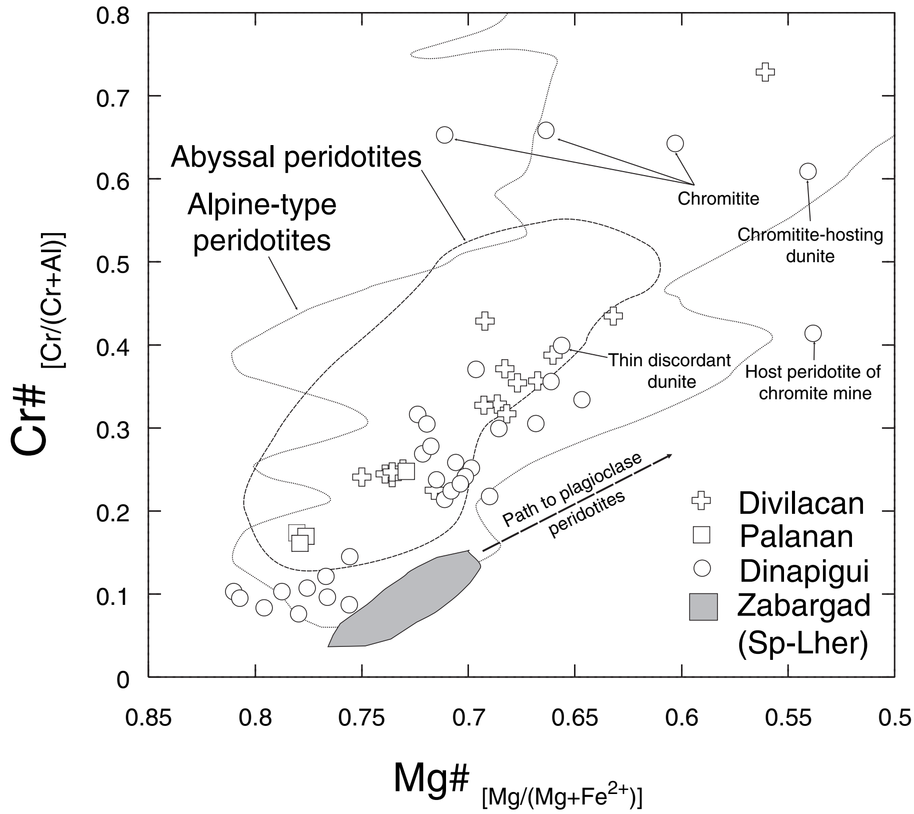

Fig. 9 Diagram showing the similarity in spinel chemistry between Isabela ophiolite peridotites and abyssal peridotites (Dick & Bullen 1984). The shaded area represents the approximate composition of spinel from the spinel lherzolites (Sp-Lher) of Zabargad. The broken arrow shows the direction of transition to plagioclase lherzolites of Zabargad and St Peter–Paul peridotites (Bonatti 1990). Aluminian spinels from Isabela ophiolite spinel have a higher Mg# compared with spinels from Zabargad with the same Cr#. Apart from the chromitites and chromitite-hosting dunite, note that the Isabela ophiolite spinels define a continuous series of compositions from a Cr# of less than 0.1 to a Cr# of 0.4, mainly in equilibrium with the lherzolitic composition.

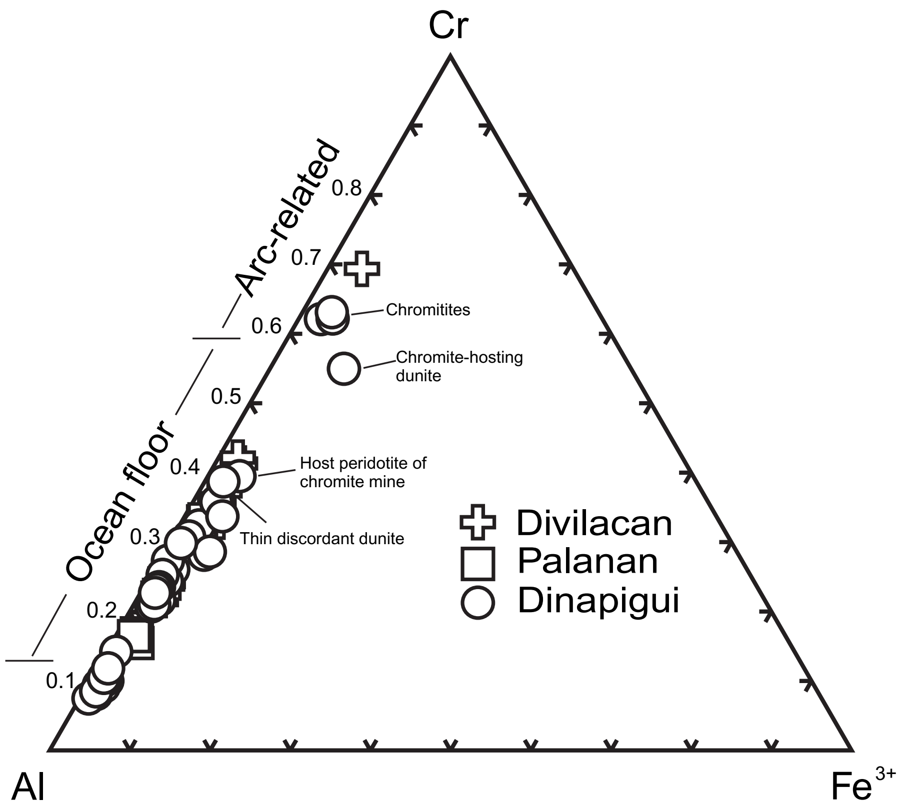

Fig. 10 Diagram showing the difference in spinel Al–Cr composition between the ocean floor and an arc-related setting. The bulk of the Isabela ophiolite peridotites show a continuous geochemical transition in the ocean floor compositional range. Chromitite and late dunites show an affinity toward an arc-related formation. This difference in composition is also evident in Figure 9.

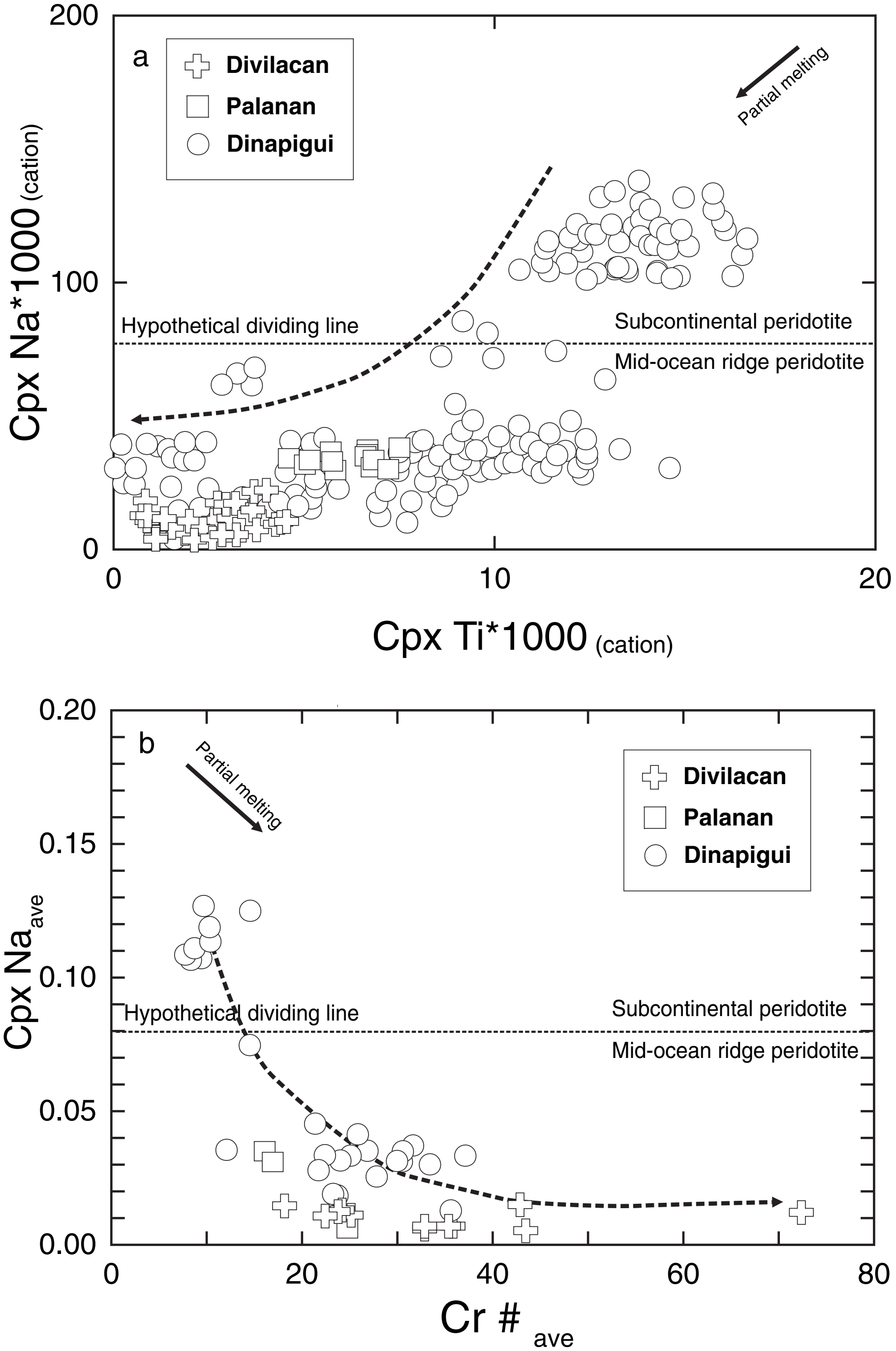

Fig. 11 Differentiation of subcontinental peridotites from suboceanic peridotites using clinopyroxene Na content. (a) Clinopyroxene (Cpx) Na content versus Ti content. (b) Clinopyroxene Na content versus Cr#. Dinapigui Point lherzolites show an elevated Na content compared to the rest of the Isabela ophiolite peridotites. The figure also shows that the Dinapigui Point lherzolites generally have a higher Ti content and a lower Cr#. Broken arrows show the hypothetical melting trend that can explain the behavior of Na in clinopyroxene during partial melting, assuming an initial composition similar to the Dinapigui Point lherzolites. Note: In (a), each plot is equivalent to one calculation. In (b), each plot is the average calculation for one rock.

several chromite mines were previously operated in the area, the deposits were very limited in size (a few hundred meters) and confined only to dunite pods, as compared to the SSZ portion of the Zambales ophiolite, which hosts extensive mineral deposits (Yumul *et al.* 1997).

ORIGIN OF THE DINAPIGUI POINT LHERZOLITES

In the spinel Cr# versus Mg# diagram (Fig. 9), Dinapigui Point lherzolites plot on and beyond the lower end of the abyssal peridotite field of Dick and Bullen (1984). This is also true in the OSMA diagram (Arai 1994), wherein the Dinapigui Point lherzolites plot inside the lower end of the subcontinental field, whereas the rest of the data set reflects abyssal peridotite values (Fig. 8). This is suggestive of an exotic block of fertile lherzolites of subcontinental affinity within the oceanic mantle section. The Dinapigui Point lherzolites also have some similarities (high modal clinopyroxene, low spinel Cr# and occurrence of Ti-rich pargasite)

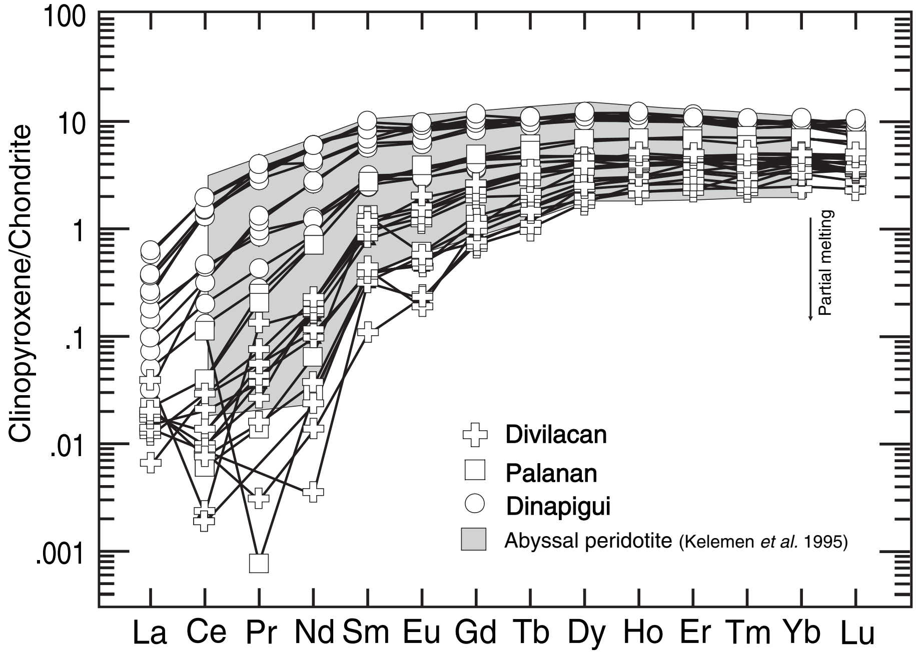

Fig. 12 Chondrite-normalized rare earth element (REE) composition of clinopyroxene in the Isabela ophiolite peridotites. The typical abyssal peridotite REE pattern encompasses the entire range of abyssal peridotite compositions defined by Kelemen *et al.* (1995). The clinopyroxene REE composition systematically decreases from the fertile lherzolites in Dinapigui toward the Palanan and Divilacan peridotites. Data are from the laser ablation inductively coupled plasma–mass spectrometry (ICP–MS) analysis, and chondrite values are from Sun and McDonough (1989). Detection limit for most LREEs is approximately 0.1; data plotted below this is only for reference.

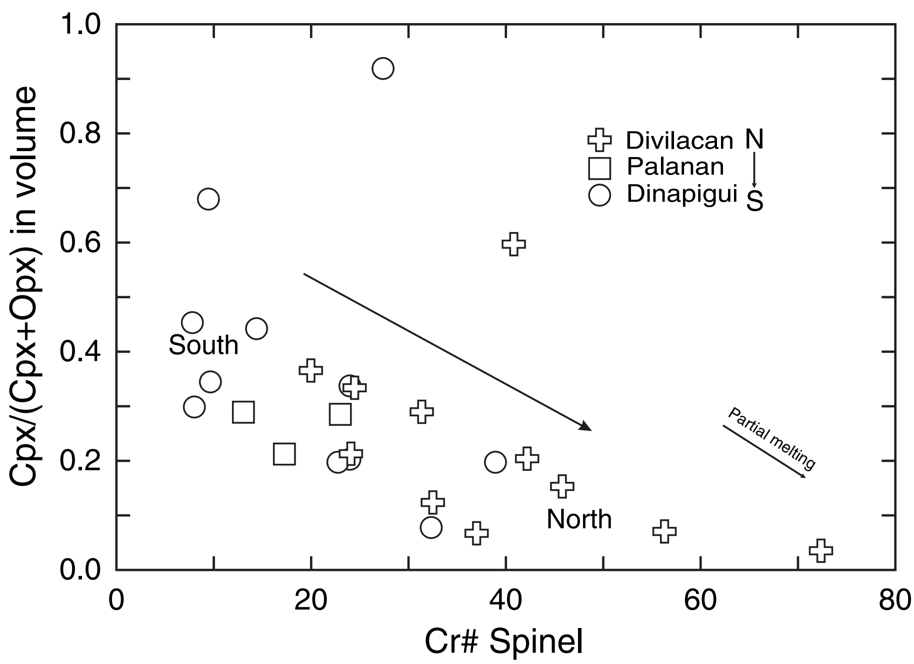

Fig. 13 Diagram showing the correlation between modal composition clinopyroxene (Cpx) and Cr# of coexisting spinel. The arrow represents the approximate path of the clinopyroxene modal change from fertile peridotites (Dinapigui [south]) to residual peridotites (Divilacan [north]). Opx, orthopyroxene.

with subcontinental-derived lherzolites from St Peter–Paul rocks (Bonatti 1990) and some Alpine–Apennine peridotites (Ishiwatari 1985; Müntener & Piccardo 2003). Ozawa (1988) and Takahashi (1991) pointed out the occasional occurrence of exotic ultramafic blocks within the mantle tectonites. In comparison, the compositional range of spinel Cr# in the Isabela ophiolite is very similar to the range of spinel Cr# (0.1–0.4) from the aluminous spinel peridotite and pyroxenite (ASPP) unit of an island arc ophiolite described by Ozawa (1988). Lherzolite-dominated ophiolite massifs can be regarded as either undepleted or less depleted subcontinental mantle or the deeper part of sub-oceanic mantle (quoted in Boudier & Nicolas 1985).

Some studies have suggested that variations in the Na content of clinopyroxene could be used to distinguish MOR peridotites from subcontinental peridotites (Kornprobst *et al.* 1981; Seyler & Bonatti 1994). Figure 11a,b clearly shows the difference in the Na content of clinopyroxene from Dinapigui Point lherzolites compared to the rest of the Isabela ophiolite peridotites. The question now at hand is whether the difference in clinopyroxene Na content and the very low Cr# of spinel are true characteristics of a subcontinental lithosphere or a consequence of other factors (e.g. source region character, mantle refertilization or metasomatism).

The validity of using Na content in clinopyroxene as a criterion for tectonic differentiation has long been debated. Partitioning of Na in the mantle is mainly controlled by pressure and temperature. Kornprobst *et al.* (1981) revealed that in the garnet and spinel stability field (>8–10 kbar), clinopyroxene contains the bulk of the Na, but Na is shared by clinopyroxene and plagioclase in the plagioclase stability field. They recognized that clinopyroxene contains less Na in suboceanic spinel peridotites than in subcontinental spinel peridotites. This difference in composition was attributed to the more extensive partial melting experienced by suboceanic mantle. From this, two important deductions can be made:

- By accepting that partial melting is responsible for the difference in clinopyroxene composition, it is assumed that the initial composition of both subcontinental and suboceanic mantle is the same
- The Na–Cr system in clinopyroxene might not be able to distinguish slightly depleted suboceanic peridotites from subcontinental peridotites

A later study by Seyler and Bonatti (1994) argued that the premelting Na content of clinopyroxene between the two tectonic settings differs by the enrichment of the jadeite component in the subcontinental peridotites. Contrary to the conclusions reached by Seyler and Bonatti (1994), a later study by Rivalenti *et al.* (1996) using mantle xenoliths from both continental and oceanic environments supports the view that the premetasomatic lithosphere in both environments does not differ. They further concluded that the clinopyroxene composition could not be used as a basis to distinguish suboceanic and subcontinental mantle, and that the differences are brought about by the different petrological processes the peridotites have undergone. On a similar note, Plove and Allègre (1980) suggested that subcontinental and suboceanic mantle could be subjected to the same kind of evolution.

Having made these arguments, the existence of a uniform subcontinental and suboceanic premelting composition cannot be denied. The Dinapigui Point lherzolites could represent the fertile end of the suboceanic mantle column, and not a subcontinental lithosphere. The clinopyroxene Na content of the Dinapigui Point lherzolite is not comparable to the rest of the peridotites as it has clearly undergone a lower degree of partial melting compared to the other peridotites (Fig. 11a,b). This is evidenced by the higher Ti content of the clinopyroxene and the lower Cr# of coexisting spinel in the Dinapigui Point lherzolites. The samples plotted on Figure 11a,b may actually define a partial melting trend, from the fertile lherzolites to harzburgites. The difference in the degree of partial melting accounts for the difference in the Na content of clinopyroxene (Kornprobst *et al.* 1981). The abrupt decrease in the Na content of clinopyroxene may suggest a sudden increase in the amount of melt extraction because of decompression or its partitioning into plagioclase; both are expected in the shallower sections of the mantle column. Since plagioclase lherzolite is absent in the Dinapigui Point lherzolites, the former hypothesis is favored. The continuity of the partial melting trend is supported by the presence of clinopyroxenes with intermediate Na composition in the area where there is an overlap in Ti and Cr# values between the fertile and the residual peridotites (Fig. 11a,b).

Spinels from the Dinapigui Point lherzolites have a very low spinel Cr# (~0.07 to ~0.15). Arai (1994) showed that fertile lherzolites from subcontinental lithospheres have a distinctively low Cr# (<0.2) compared to peridotites from other settings. For comparison, the Zabargad peridotites and St Peter–Paul's rocks, exposed in the Red Sea and near the Mid-Atlantic Ridge, respectively, are known examples of subcontinental peridotites. Figure 9 shows that the spinels from the Zabargad and St Peter–Paul's peridotites have a slightly lower Mg#. Bonatti (1990) explained that the aluminous spinel from Zabargad spinel peridotite is enriched in Fe^{2+}. Aluminum-rich spinels from the Dinapigui plot directly below the abyssal peridotite field, in line with the melting trend defined by the samples. The Ti-rich pargasite observed in the Dinapigui Point lherzolites shows no significant Fe activity as their Fe/(Mg + Fe) remains low (~0.1), and the Ti-rich pargasite has not been changed to ferro-pargasite. This suggests that the spinel chemistry of the Dinapigui Point lherzolites is comparable to that of abyssal peridotites rather than subcontinental peridotites. However, this proposal is constrained by data limitations and should be further investigated using additional geochemical data (e.g. stable isotope data).

Another explanation for the occurrence of fertile lherzolites is mantle refertilization by ascending melts. Refertilization in MORs often occurs at low pressures (less than 5 kbar) (Elthon 1992) within the plagioclase stability field. This is consistent with the idea that refertilization of residual peridotites in spreading ridges occur above the 'cold' thermal boundary layer (TBL), the upper part of the mantle column (Niu 2004). According to Niu (2004), the TBL in slow-spreading ridges is equivalent to the plagioclase stability field. He postulated that above the TBL, partial melting ceases because the lithosphere is too cold to melt, but the cold residual rocks react with ascending melts instead. The Dinapigui Point lherzolites have no plagioclase and therefore have not been in equilibrium with the plagioclase stability field. Furthermore, the clinopyroxenes in these lherzolites are primary porphyroclasts. This is inconsistent with their crystallization from refertilizing melts. The spinels in the Dinapigui Point lherzolites have a low Cr# and TiO_2 content (less than 0.1 wt%), suggesting a fertile and non-residual composition, respectively. If indeed refertilization in this peridotite unit occurred, the spinel may retain its original depleted composition, similar to the case of the Othris ophiolite in Greece (Barth *et al.* 2003), or it could have reacted with the melt, increasing its Cr# as a result of the partitioning of Al into plagioclase. Furthermore, magmatic veins can modify spinel compositions and cause the heterogeneity of spinel within one rock (Hellebrand *et al.* 2002). All these possibilities were not observed in the Isabela ophiolite.

Mantle metasomatism is also a probable explanation for the origin of the fertile Dinapigui Point lherzolites. However, available data on the chemistry of primary mineral phases do not show evidence of such an episode. The clinopyroxene does not show selective enrichment or depletion of LREEs or HREEs as in most metasomatized mantle peridotites; for example, those from Lherz Massif in France (Bodinier *et al.* 2004). No metasomatic mineral was observed in thin sections of the lherzolites. The presence of Ti-rich pargasite could, however, suggest the presence of hydrous fluids.

The continuity of the spinel and clinopyroxene compositional spectra supports the correlation between the fertile and less fertile sections of the Isabela ophiolite. Both minerals reflect the original magmatic conditions, unless modified by later processes (e.g. metasomatism). The clinopyroxenes in abyssal peridotites with abundant clinopyroxenes are more likely to be residual (Niu *et al.* 1997; Niu 2004). All available data show that the Dinapigui Point lherzolite is a continuity of the Isabela ophiolite lithosphere and represents the deeper section of the column or the deeper part of an oceanic lithosphere. In normal fast-spreading ridges, melting starts somewhere in the spinel–garnet transition zone (~60–70 km in depth) (Shen & Forsyth 1995). In this region, undepleted or slightly depleted peridotites can be expected. Considering that the Isabela ophiolite is derived from a slow-spreading ridge, fertile lherzolites could have been brought to the surface from a great depth with very little melting. Recent studies have proven that in slow-spreading ridges, mantle materials can be brought to the surface along transform faults or detachment faults (Karson 1990; Cannat 1993; Blackman *et al.* 1998; Escartin & Cannat 1999; MacLeod *et al.* 2002).

The field of abyssal peridotites, defined by Dick and Bullen (1984) from spinel mineral chemistry, does not include abyssal peridotites as fertile as the Dinapigui Point lherzolites. It is highly probable that ordinary sampling methods for abyssal peridotites are unable to reach the depth where fertile lherzolites like the Dinapigui Point lherzolites exist. Recognition of the Dinapigui lherzolites may be a rare acknowledgement of such fertile abyssal peridotites. For this reason, the Isabela ophiolite is a very important area for the study of oceanic lithosphere processes.

OPHIOLITE CLASSIFICATION

The Isabela ophiolite is mainly composed of clinopyroxene-rich harzburgites and lherzolites. Furthermore, the range of spinel Cr# from the analyzed samples falls in the low end of the abyssal peridotite field (<0.1 to <0.4). However, the exposures of massive depleted harzburgites in the northern part (~0.25–0.43) and the pyroxene-rich harzburgites (Cr# ≤ 0.4) hosting the chromitite pods should also be taken into account. Such characteristics suggest that the Isabela ophiolite does not strictly belong to the lherzolite ophiolite subtype (LOS). It is an intermediate type between LOS and the harzburgite ophiolite subtype (HOS), thus belonging to the transitional ophiolite suites (Boudier & Nicolas 1985).

CONCLUSIONS

The systematic change in terms of lithology, supported by petrography and mineral chemistry, shows that the peridotites from the Isabela ophiolite belong to one continuous lithosphere, from the lower to the upper sections of the mantle column. There is evidence of the exposure of a stratified upper mantle–crust column emplaced right-side up from Dinapigui to Divilacan. Available data strongly suggest that the fertile lherzolites in Dinapigui may represent the lower part of the upper mantle that is absent in most ophiolites.

Geological data also support the possibility that this ophiolite is a fragment of a Cretaceous basin formed in a slow-spreading ridge. In this regard, the mantle section of the Isabela ophiolite is a perfect analog for the entire mantle column of a slow-spreading ridge lithosphere. On the basis of its overall characteristics, a transitional ophiolite subtype classification is proposed for the Isabela ophiolite.

The slow-spreading MOR origin of the Isabela ophiolite may further complicate the scenario of Philippine geology as most of the ophiolites in the Philippines are believed to be from SSZs. In addition, oceanic basins within the vicinity of the Philippine archipelago have rocks that are much younger than the Isabela ophiolite and which were formed in fast-spreading ridges. This is in contrast with the petrological evidence that implies that an Early Cretaceous oceanic crust from a slow-spreading ridge underlies northeastern Luzon Island. The extent of this block is still subject to further investigation.

ACKNOWLEDGEMENTS

We are greatly indebted to Dr Tomoaki Morishita for his invaluable assistance during one of the field surveys and for sharing his expertise both in the field and in the laboratory. We are also grateful to the following Arai laboratory members for their assistance during our electron probe microanalysis (EPMA) and inductively coupled plasma–mass spectrometry (ICP–MS) analysis: Dr Jiro Uesugi, Dr Yohei Shimizu, Ms Satoko Ishimaru, Dr Miki Shirasaka and Mr Yoshito Ishida. Critical comments from and discussions with Dr Akihiro Tamura-Hasebe and Dr Rodolfo Tamayo, Jr. were also helpful. Dr Victor B. Maglambayan is also thanked for the National Institute of Geological Sciences (NIGS) contribution to the fieldwork. We also thank members of the Rushurgent Working Group (RWG) for their assistance in fieldwork-related matters. We would like to express our special thanks to Dr Petri Peltonen and Dr Hirokazu Maekawa for their critical reviews, and Professor Akira Ishiwatari and Dr Yoshio Watanabe for their handling of our manuscript.

In Isabela, the municipal government officials of Dinapigui, Palanan and Bicobian, along with their kind and warm-hearted people, provided us with invaluable assistance that made the series of fieldwork expeditions successful. We express our deepest gratitude to them.

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