

Transfer of melts in the sub-arc mantle:
Insights from high-pressure experiments
and from the New Caledonia ophiolite.

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PART. B

B.1. Ophiolites

B.1.1. General description of ophiolites

Ophiolites were first described by Alexandre Brogniart in 1813 as an assemblage of green rocks present in the Alps (Brogniart, 1813). This definition evolved slowly through the 20th century and became known as a sequence of serpentine, diabase, pillow lavas and chert until the early 60's. With the precise description of large ophiolitic complexes in Cyprus and Oman and the beginning of ocean drilling programmes, the true nature of ophiolitic complexes was unravelled as remnant of oceanic floors preserved on the continental crust. An idealised but formal description of ophiolites was given during the Penrose field conference in California and Oregon in 1972. Since these times, many ophiolites have been found all around the world and the definition of an ophiolitic massif has evolved in complexity [Fig.B.1]. Ophiolites are still highly studied as they remain unique natural opportunities for insights on the composition and the structure of the oceanic crust.

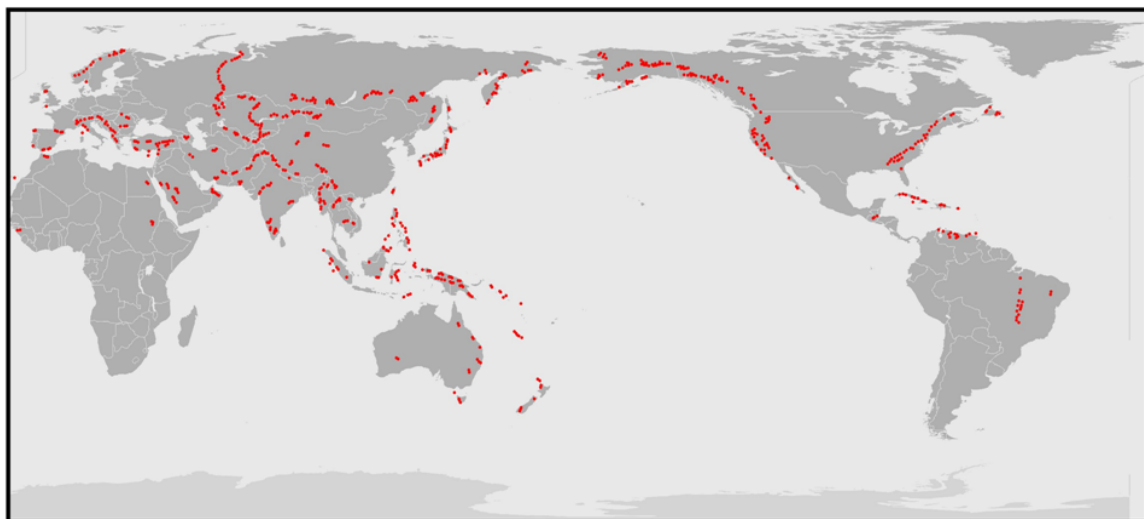


Fig.B.1. Distribution of world's ophiolites with their approximate position. (Inspired from Coleman, 1977; Kelemen et al., 1998; Matter & Kelemen, 2009).

Two main groups have to be distinguished in terms of emplacement. On one hand, orogenic ophiolites are elongated terranes which typically occur in collisional zones (e.g. Alps, Himalayas, Rockies, Appalachian, etc.). Although these ophiolitic bodies are very common, their structure is often strongly affected by orogenic processes and many of these remnants of oceanic crusts provide poor quality information on large scale features and variation in the petrology of the mantle. On the other hand, a few places around the world have experienced a short lived and rare geodynamic process called non-collisional obduction (Coleman, 1977). This event occurs on convergent margins when continental crust is subducted in the continuation of its attached oceanic crust. The sinking of continental crust into the mantle causes a jamming of the subduction process due to the high buoyancy of continental rocks. Such ophiolite nappes can be found in places such as Semail and Wadi Tayin (Oman), Massif du Sud (New Caledonia); Troodos (Cyprus) or Bay of Islands (Newfoundland) where large scale studies can be done.

Our knowledge of the oceanic crust has considerably evolved since the days of the Penrose conference and early deep sea drilling allowing a refinement of our views on ophiolites. A typical sequence is composed of basal peridotites (harzburgite, lherzolite), a transition zone of ultramafic rocks (dunite, wehrlite, pyroxenite), lower layered gabbros, upper isotropic gabbros and plagiogranites, the sheeted dyke complex, basaltic pillow lavas and finally sediments (pelite, red clays, carbonates) [Fig.B.2].

Although early studies were generalising the structure of ophiolitic bodies at mid-ocean ridges, more detailed structural and geochemical work on ophiolitic massifs has shown that different kind of environments are conserved in these units. Because of their geodynamic settings, fore-arc peridotites (Horokanai, Massif du Sud, Betts Cove, Coastal Range) are the most likely to be obducted. Arcs (Kohistan, Talkeetna), back-arcs (Trinity, Betts Cove, Troodos) and marginal basins (Josephine, Oman, Bay of Islands) are also very likely to be thrust on a continent whereas true mid-ocean ridge ophiolites remain orogenic or in abnormal settings (Argolis, Macquarie Island) (Metcalf & Shervais, 2008). A large number of other ophiolites can be placed in these groups but a strong bias will appear in the exposed section of each massifs. If fore-arc and back-arc ophiolites can often display a large sequence

of ultramafics rocks present at the root of the terrane, the large thickness of arc ophiolites make them relatively rare and arcs are more likely to be preserved as accretionary complexes of sediments and volcanoclastic rocks with no outcropping ultramafic roots.

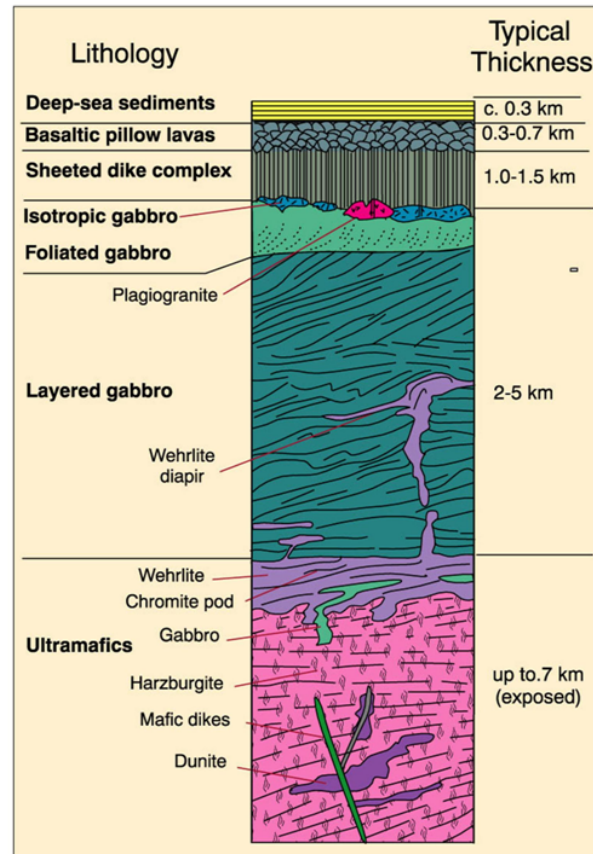


Fig.B.2. Stratigraphic column of a typical ophiolitic sequence (Nicolas, 1989)

B.1.2. Generation of replacive dunites

Many ophiolites are characterised by the presence of large dunite bodies and channels which cross cut the entire ultramafic section and can form large layers in the lower part of the crust-mantle transition (Prinzhofer et al., 1980; Lago et al., 1982; Nicolas & Prinzhofer, 1983; Nicolas, 1989; Edwards, 1995; Godard et al., 2000; Suhr et al., 2003; Tamura & Arai, 2006; Morgan et al., 2008; Pagé et al., 2008). These dunite units occur as replacive features within the mantle and show several characteristics such as anastomosed conduits with fingers in country rocks, residual spinel trails, an

absence of offset in mantle discontinuities and a lack of zonation or vein crystallisation features (Prinzhofer et al., 1980; Lago et al., 1982; Kelemen, 1990; Daines & Kohlstedt, 1994; Godard et al., 2000; Morgan & Liang, 2003; Suhr et al., 2003; Batanova & Savelieva, 2009). These characteristics suggest that replacive dunites are formed by the consumption of pyroxenes and the precipitation of additional olivine by the percolation of a basaltic melt in the peridotite (Kelemen, 1990; Kelemen et al., 1995; Suhr, 1999; Morgan & Liang, 2003).

Replacive dunites and their mechanism of formation benefited of a particular interest as it bears the ability to explain the lack of equilibrium which exist between ocean basalt and the uppermost mantle. One of these specific characteristic is the constant undersaturation in orthopyroxene which typically characterise MORB (Stolper, 1980). In a simple system, where melts were produced by an adiabatic melting of a mantle column, the percolation of melts to the surface would modify the uppermost peridotite and the ascending melts so that MORB and upper mantle would be in equilibrium. However, this model did not conform to natural observations and transport in chemical isolation by fractures (Nicolas, 1986) or channels (Kelemen, 1990) were proposed as the main mode of basaltic melt transfer under a mid-ocean ridge (Kelemen et al., 1995; Lambart et al., 2008). The mechanism of formation of replacive dunite involves an initial heterogeneity which triggers the focusing of melts. As melts rises in the porosity of the rocks, it becomes saturated only in olivine, consuming other mineral phases from the surrounding as it will be developed in detail in chapter 5.

4. The New Caledonia obduction complex

4.1. Introduction

The New Caledonia archipelago hosts an important ultramafic ophiolite complex preserved without excessive tectonic overprint. The southern massif (Massif du Sud) is one of the largest coherent ophiolitic bodies in the world, comparable with some ophiolites in Oman (Semail, Wadi Tayin, Aswad, Fizh). It also presents the largest surface area of exposed mantle in the world and has the specificity to still remain attached to its non-obducted oceanic crust (Prinzhofer et al., 1980; Paris, 1981). The Massif du Sud stands as a wonderful natural case study of large scale features for melt transport in the upper part of the mantle.

The New Caledonia ophiolite is also well known to be extremely depleted geochemically (Prinzhofer & Allègre, 1985; Marchesi et al., 2009). This feature implies that large amount of melts have been produced and the peridotite could preserve pathways for melt transfer into the lower crust. The description of several features such as dykes, channels and magmatic intrusions in previous studies (Rodgers, 1973b; Paris, 1981; Prinzhofer et al., 1981 and references therein) gives good ground to start a detailed study of such features and attempt to decipher the magmatic history of the Massif du Sud mantle.

Finally, despite being one of the largest ophiolites, New Caledonia has not been extensively studied like its counterparts in Oman, the Mediterranean Sea, Newfoundland or the west coast of North America. Most studies have been economically orientated and have not provided a model for the formation of the ophiolite. In recent years, advances in modern analytical techniques have generated a renewed interest in the New Caledonian ophiolite which should lead to new information about this region of the Pacific Ocean (Marchesi et al., 2009; Ulrich et al., 2010).

4.2. Geological framework

4.2.1. General historical introduction of New Caledonia

New Caledonia is an archipelago in the south-west Pacific Ocean composed of several islands (Loyalty Is., Belep Is., Ile des Pins, Ile Ouen, Grande Terre), reefs and lagoons which are a territorial collectivity of France. The archipelago lies approximately 1500 km east of Australia, 2000 km north of New Zealand and south of Vanuatu [Fig.4.1]. The main island (Grande Terre) forms 80% of the landmass; it is an elongated island 400 km long and 40 km wide, relatively mountainous with a few peaks above 1600 m. New Caledonia lies astride the Tropic of Capricorn and possesses an oceanic tropical climate with highly seasonal rainfall. The unusual geological history of New Caledonia has allowed the archipelago to maintain a unique flora and fauna preserving characteristics of an ancient Gondwanian ecosystem (Pirard, 2007 and references therein).

Human settlement in the area started approximately 3000 years ago with the succession of Lapita, Naia Oundjo and Kanak cultures. The island was discovered by Europeans in 1774 by James Cook and named for its supposed resemblance with Scotland, something which has baffled visitors ever since. In the middle of the 19th century, the archipelago was proclaimed a French colony and used successively as a detention island, a mining centre and more recently a tourist destination. The population of New Caledonia is currently around 250,000 inhabitants with two thirds concentrated in the capital, Nouméa. In the last two decades, a self-governance movement have risen in influence and a referendum in the next few years could proclaim the independence of the archipelago.

The geological and mining potential of New Caledonia was truly exposed a few years after the arrival of the first settlers when garnierite, a Ni-serpentine mineral (Dana, 1873), was discovered by the French geologist and explorer Jules Garnier (Garnier, 1864). Industrial mining of nickel started in 1876 and has kept expanding for a century during which New Caledonia was the largest producer in the world of nickel, cobalt and ferronickel alloys. At the

present time, New Caledonia is still in the top five producers and has the largest tonnage of nickel saprolite exploitation (Pirard, 2007).

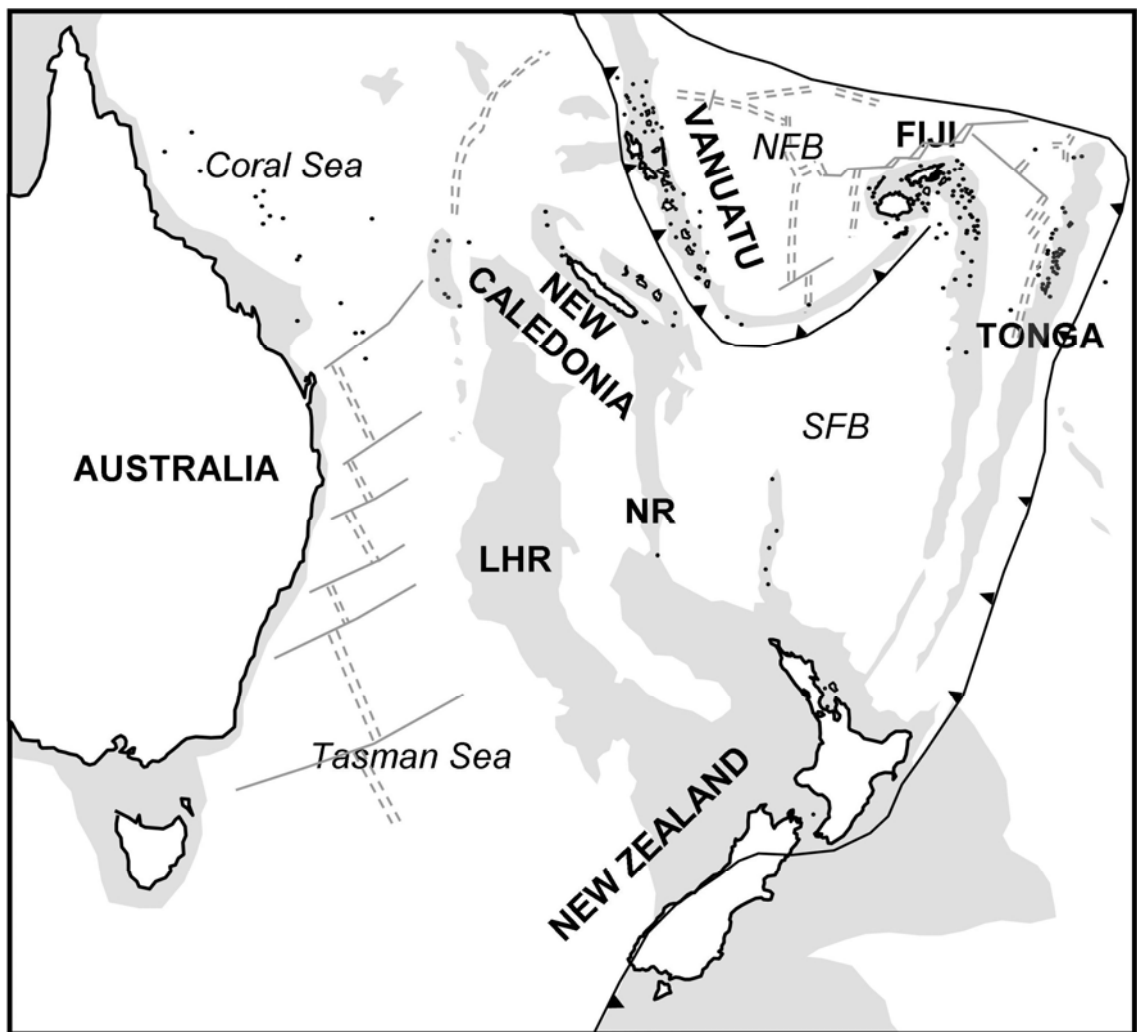


Fig.4.1. Present-day tectonic configuration of the south-west Pacific showing the distribution of major tectonic features (Hall, 2002; Whattam, 2009) Abbreviations: LHR : Lord Howe Rise; NR : Norfolk Ridge; SFB : South Fiji Basin; NFB : North Fiji Basin.

Most of the geological research on New Caledonia is supported by the mining industry and focused mainly on lateritic alteration processes, formation of chromitite bodies (Leach & Rodgers, 1977; Leblanc et al., 1980; Cassard et al, 1981; Lago et al., 1982) and post-mining rehabilitation. In the 1970's, the confirmation of plate tectonics as a major model to understand the Earth provided new views on the terrane relationships between New Caledonia, New Zealand, Australia and Antarctica (Brothers & Blake, 1973; Parrot & Dugas, 1980; Kroenke, 1984). That decade also produced a coherent model for the interpretation of ophiolites as remnants of oceanic crust preserved on the

continents (Coleman, 1977). This new view about large ultramafic bodies initiated the reconsideration of the New Caledonian peridotites as an obducted oceanic floor (Prinzhofer, 1981; Paris, 1981) and no longer as an ultramafic - mafic layered intrusion (Avias, 1955; Guillon & Lawrence, 1973). Some research on the petrology and the geochemistry of the ultramafic rocks was done in the following years (Ohnenstetter & Ohnenstetter, 1981; Prinzhofer & Berger, 1981; Nicolas et al., 1981; Nicolas & Prinzhofer, 1983; Hamelin et al., 1984; Prinzhofer & Allègre, 1985) but it is only recently that the New Caledonian peridotites have been the subject of new publications (Marchesi et al., 2009; Ulrich et al., 2010).

4.2.2. Geological settings of New Caledonia

4.2.2.1. Regional geological settings

New Caledonia bears unusual characteristics for an oceanic island of the southern Pacific. Most pacific islands are relatively young and formed by Cenozoic volcanic activity or carbonate platforms. Grande Terre, by contrasts, has a basement made of a Palaeozoic and Mesozoic accretionary complexes. The island is an emerged part of the Norfolk Ridge which, with the Lord Howe Rise and the Fairway Ridge, is part of a small continental plate called Zealandia which drifted away from the Australian continent as a result of the Late Cretaceous break-up of Gondwana (Dubois et al., 1976; Crawford et al., 2003; Lafoy et al., 2005) [Fig.4.1]. The Norfolk ridge has been investigated through deep sea drilling and seismic studies which have revealed a 32 km thick sialic crust (Dubois, 1969) covered by Oligocene to Recent sedimentary rocks (Dupont et al., 1975; Dubois et al., 1976; Daniel et al., 1976).

In terms of general geodynamics, the history of New Caledonia can be divided into three parts which represents major changes in the geological position of Grande Terre. During the Palaeozoic and most of the Mesozoic, New Caledonia was attached to the eastern coast of Australia in an area near the Gympie terrane (Cluzel & Meffre, 2002). The primary basement of Grande Terre was created by several accreting terranes making a complex mosaic of volcanic, sedimentary and metamorphic units (Aitchison & Meffre, 1992). The

tectonic collage of the terranes itself occurred from the Late Jurassic to Early Cretaceous and is related to the New Zealand Rangitata orogeny (Bradshaw, 1979; Paris, 1981). At the end of the Cretaceous, the Tasman Sea started to open and several ridges forming Zealandia drifted apart forming narrow marginal basins (Davies & Smith, 1971).

The movement of New Caledonia away from Gondwana was compensated by an eastward dipping subduction of the Pacific plate which has formed the Loyalty ridge. The Loyalty Islands are interpreted as the remnant of the Eocene island arc and the Loyalty Basin as a remnant of the fore-arc basin (Cluzel et al., 1999; 2001).

At the very end of the Eocene, a part of the Loyalty Basin was obducted on to Grande Terre due to the jamming of the subduction zone by the arrival of the New Caledonian continental crust. This led to metamorphism of terranes attached to Grande Terre and the obduction of Loyalty oceanic crust onto the continental crust. Since at least the middle Miocene, the northern Zealandia continent has been passively carried along by the Australian plate while its eastern margin is being subducted beneath the Pacific Plate along the Vanuatu Trench (Auzende et al., 1995).

4.2.2.2. General tectonics

The two main tectonic features which characterise New Caledonia are the thrust faults limiting the obducted terranes and the main transcurrent faults which affect these same terranes by separating the Massif du Sud from basalts and other ultramafic massifs on the west coast of Grande Terre (Brothers, 1974). All terranes have faulted boundaries related to their accretion process onto the Norfolk ridge [Fig.4.2].

Thrust faulting is easily visible by the low dipping angle and the intense serpentinisation and mylonitisation at the contact between mafic - ultramafic terranes and the basement. Within the ultramafic terrane, small dip-slip faults occur and are highlighted by small serpentinised zones but no large displacement is recorded regarding post-obduction faults (Chardon & Chevillotte, 2006).

Present day New Caledonia is mainly affected by extension faulting with a NW-NNW trending direction. This faulting has been occurring since the Early Miocene (Lagabrielle et al., 2005; Chardon & Chevillotte, 2006) and these recent tectonic movements are responsible for many fluvial formations, drowned valleys in the western lagoon and uplifted coral reefs on the east coast. This extension process seems to be controlled by post-obduction isostatic readjustments and more recently by the possible flexure of the oceanic lithosphere currently subducting under the Vanuatu trench at a rate of 12cm/year (Lagabrielle et al., 2005).

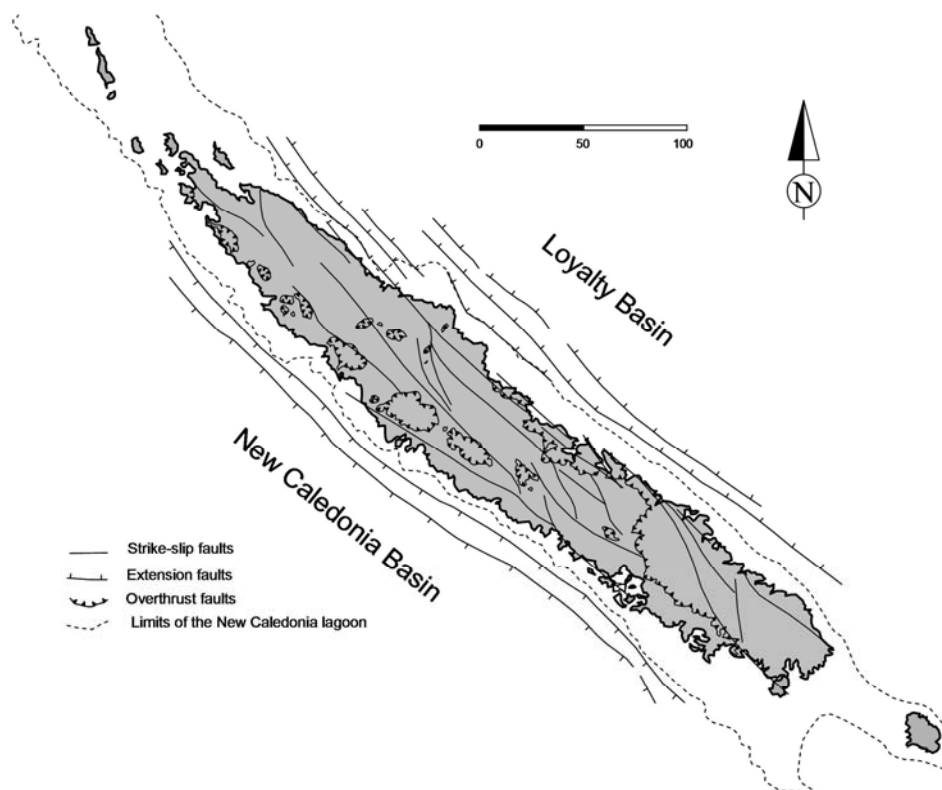


Fig.4.2. Simplified tectonic map of New Caledonia (Pirard, 2007, modified from Chardon & Chevillotte, 2006).

4.2.2.3. Terranes of New Caledonia

The terranes of New Caledonia can be grouped into three super-terranes which are related to the different position of New Caledonia since the Palaeozoic [Fig.4.3]. A general review of all terranes is given in Pirard (2007) where references for each group of rocks can be found.

Ante-Senonian terranes

The Ante-Senonian (pre-Upper Cretaceous) terranes are a group of four Late Palaeozoic – Mesozoic terranes (Koh, Térémba, Central Chain, Boghen) forming the basement of Grande Terre. These terranes show evidence of convergent margin settings and were developed along or accreted to the eastern Gondwana margin (Cluzel & Meffre, 2002). Many features present in these terranes have particularities which are shared by terranes in New Zealand, Antarctica and Queensland (Waterhouse & Sivell, 1987; Aitchison et al., 1998) [Tab.4.1].

<u>Ante Senonian Terranes</u>		
Koh	Late Carboniferous Permian	Harzburgite, gabbro, basalt (tholeiite, boninite), plagiogranite, sandstone, siltstone, chert
Central chain	Permian Upper Jurassic	Volcanoclastic sandstone, black and tuffaceous siltstone, conglomerate
Boghen	Pre-Lias	Schist, metagreywacke
Térémba-Moindou	Permian Late Jurassic	Calc-alkaline volcanoclastics (greenschist)

Tab.4.1. Terranes of New Caledonia of Permian to Late Jurassic age.

High-Pressure terranes

In the north-east part of Grande Terre, a well preserved high-pressure low-temperature metamorphic sequence is exposed. Four units are described (Formation à charbons, Koumac, Diahot and Pouébo) which show an increasing and roughly continuous metamorphism from diagenetic conditions to eclogite facies (Cluzel et al., 1999; Fitzherbert et al., 2003; Spandler et al., 2005; Potel et al., 2006). Metamorphism recorded in these rocks reach peak conditions of 24 kbar and 600°C and is related to the subduction of New Caledonia under the Loyalty Basin from the end of the Cretaceous until the Middle Eocene (Spandler et al., 2005) [Tab.4.2].

High Pressure Terranes

Formation à Charbons	Upper Cretaceous	Coal, conglomerate, arkose, sandstone, siltstone
Koumac	Upper Cretaceous Basal Eocene	Chert, limestone, marl, flysch
Diahot	Upper Cretaceous Middle Eocene	Metasediment, metabasalt, metarhyolite (Blue schists)
Pouébo	Upper Cretaceous Middle Eocene	Metasediment, metachert, metabasalt (Eclogite)

Tab.4.2. Terranes of New Caledonia showing a High Pressure - Low Temperature metamorphism

Obduction nappes

Contemporaneous to the formation and metamorphism of the high pressure terranes, a series of oceanic terranes (Bourail, Poya, and later, Ultramafic) were obducted onto the New Caledonia continent. These units jammed the subduction process on the eastern margin of Zealandia in the Late Eocene. The Bourail and Poya nappes were scraped off the New Caledonian margin (Cluzel et al., 1998; 2001; Ulrich et al., 2010) whereas the ultramafic terrane was emplaced as an obducted fore-arc from the South Loyalty Basin (Paris, 1981; Prinzhofer et al., 1980; Collot et al., 1987; Auzende et al, 2000; Cluzel et al., 2001) [Tab.4.3].

Obducted Terranes

Bourail	Palaeocene	Flysch
Poya	Upper Cretaceous Basal Eocene	Basalt (tholeiite, boninite, alkali basalt) hyaloclastite, argillite, chert
Ultramafic	Pre-Eocene	Peridotite, chromitite, gabbro, granitoid

Tab.4.3. Obducted terranes of New Caledonia with age between Paleocene and Eocene

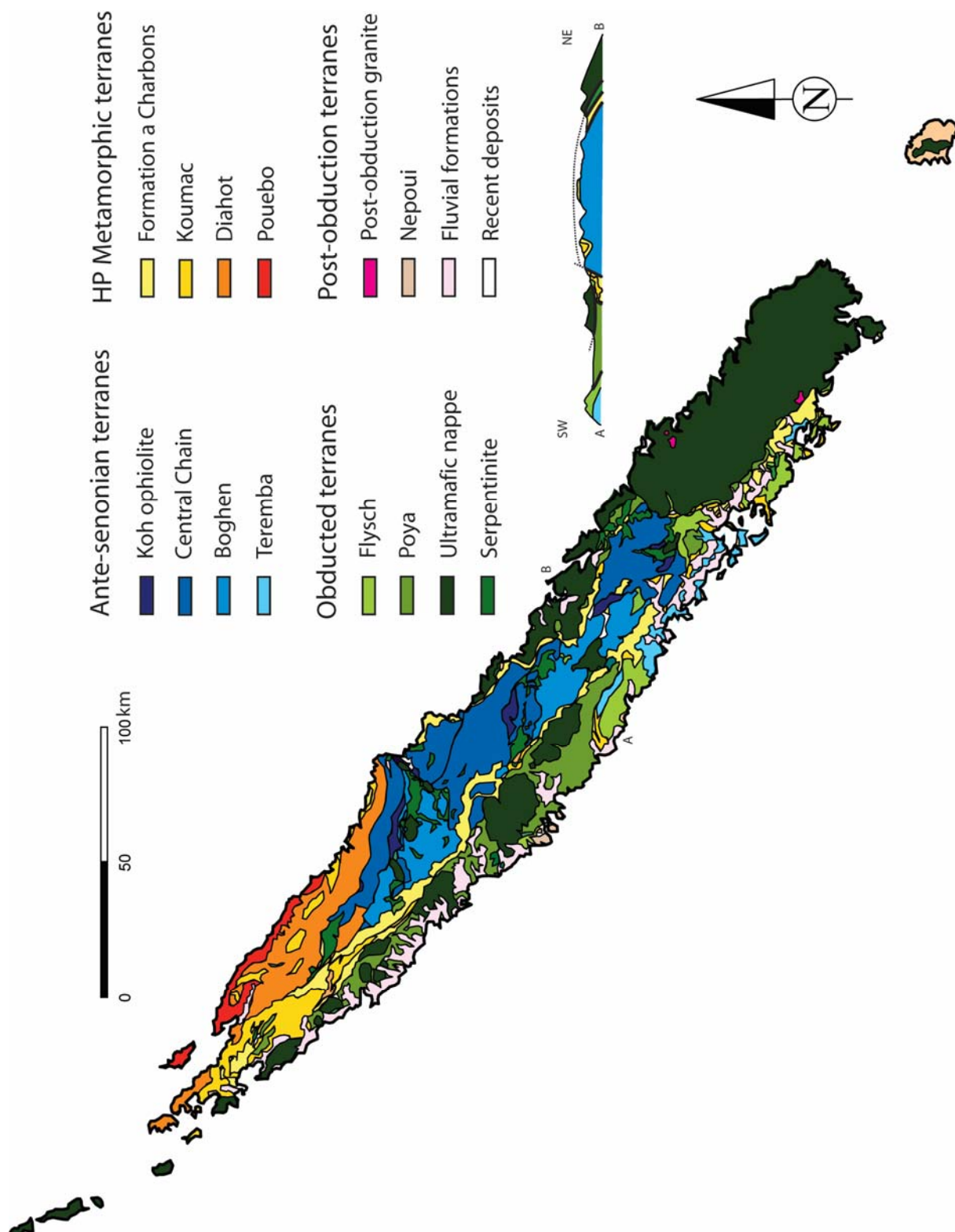


Fig.4.3. Simplified geological map of New Caledonia (Pirard, 2007 from data of the BRGM and DIMENC).

Post-obduction formations

After the emplacement of the main nappes, minor magmatism occurred in New Caledonia in the form of small intrusions of granitoids (Cluzel et al., 2005) or basalt. Other formations are sedimentary units formed by the erosion of the continent, and the surface alteration in a tropical environment, in relation with post-obductive tectonic events [Tab.4.4].

<u>Post-obduction formations</u>		
St-Louis	Late Oligocene	Granitoids
Koum-Borindi		
Goa N'Doro Fm.	Late Oligocene Miocene	Breccia, Conglomerate, sandstone, siltstone, paleosol
Nepoui Fm.	Lower Miocene	Conglomerate, calcarenite (Deltaic)
Laterite	Miocene Present	Garnierite, asbolane, saprolite, limonite, pisolite, ferricrete
Alluvial and Recifal deposits	Post-Pliocene	Sand, silt, limestone

Tab.4.4. Cenozoic terranes of New Caledonia.

4.2.3. The ultramafic terrane

The ultramafic nappes are a major feature of New Caledonian geology. It covers 8000 km² of Grande Terre (41%) but is thought to have once covered the whole island (Paris & Lille, 1977). The distribution of the terrane includes several massifs on the west coast, some small nappes in the central chain and the “Grand Massif Minier du Sud” which dominates the whole southern part of the island [Fig.4.3]. Peridotites also occur on the smaller islands of the Norfolk Ridge in altered outcrops on Belep Is., Yandé Is., Ile des Pins and Ouen Is. and have been identified in seismic studies as far as 300 km north of Grande Terre (Rodgers, 1975; Guillon, 1975; Cassard et al., 1981).

The general structure of the ophiolite shows that it is between a 1000 to 4000 m thick and is bounded at its base by a serpentinite sole which highlights the obductive thrust fault. Most of the massifs have similar characteristics and

show an evolution in chemistry from north to south (Paris & Lille, 1977; Cassard et al., 1981) [Fig.4.22]. With the exception of the northernmost part, the ophiolite is a depleted mantle of dunite-harzburgite composition. In the Tiébaghi massif, a lherzolite zone is known (Moutte, 1979, 1982) and further north, the Poum massif has the least depleted composition and is formed of spinel and plagioclase lherzolites (Moutte, 1979; Sécher, 1981).

The general stratigraphy shows that the highest levels of the ophiolite are in the southern part of the Massif du Sud where the palaeo-Moho can be identified in several places. The absence of upper and middle oceanic crust is probably due to deep erosion or tectonic detachment (Cluzel et al., 2001). It is considered that around 8 km of additional rocks would have been present above the crust-mantle transition (Prinzhofer et al., 1980). Within the Massif du Sud, the deepest part of the ophiolite is on the north-east coast where the ultramafic nappe is considered as being still attached to the South Loyalty Basin oceanic crust (Paris, 1981; Prinzhofer 1981, Collot et al., 1987).

The Massif du Sud is the only part of the nappe which is sufficiently large to show kilometre-scale features and allow establishment of an accurate stratigraphy for most of the ophiolite. The interpretation of this ultramafic sequence is relatively straightforward as no major faults disrupt the overthrusting nappe. Above the serpentinite sole, harzburgite forms around three quarters of the stratigraphic sequence. It has a tectonite texture where foliation and lineation can be determined from the orthopyroxene porphyroclast orientation (Guillon et al., 1974; Guillon & Trescases, 1976; Trescases & Guillon, 1977; Prinzhofer et al., 1980). These studies show that the general structure of the ophiolite is conserved with a shallow dipping orientation which is only steeper in localised areas such as Mt Humboldt (Prinzhofer et al., 1980). Within the harzburgite, channels of dunite and pyroxenite dykes are common, but more likely to be observed in the higher part of the stratigraphy. In the areas surrounding regions where the palaeo-Moho has been identified, dunite dominates over harzburgite and forms large coalescing pods occasionally associated with chromitite deposits. In the zone above the crust-mantle transition, mafic lithologies appear and a variety of rocks is formed in the transition zone (dunite, wehrlite, clinopyroxenite, websterite, troctolitic gabbros, gabbro-norite) (Dupuy et al., 1981a). Finally, the

uppermost part of the ophiolite is mostly composed of gabbro-norite which occurs in five areas in the southern part of the Massif du Sud [Fig.4.4].

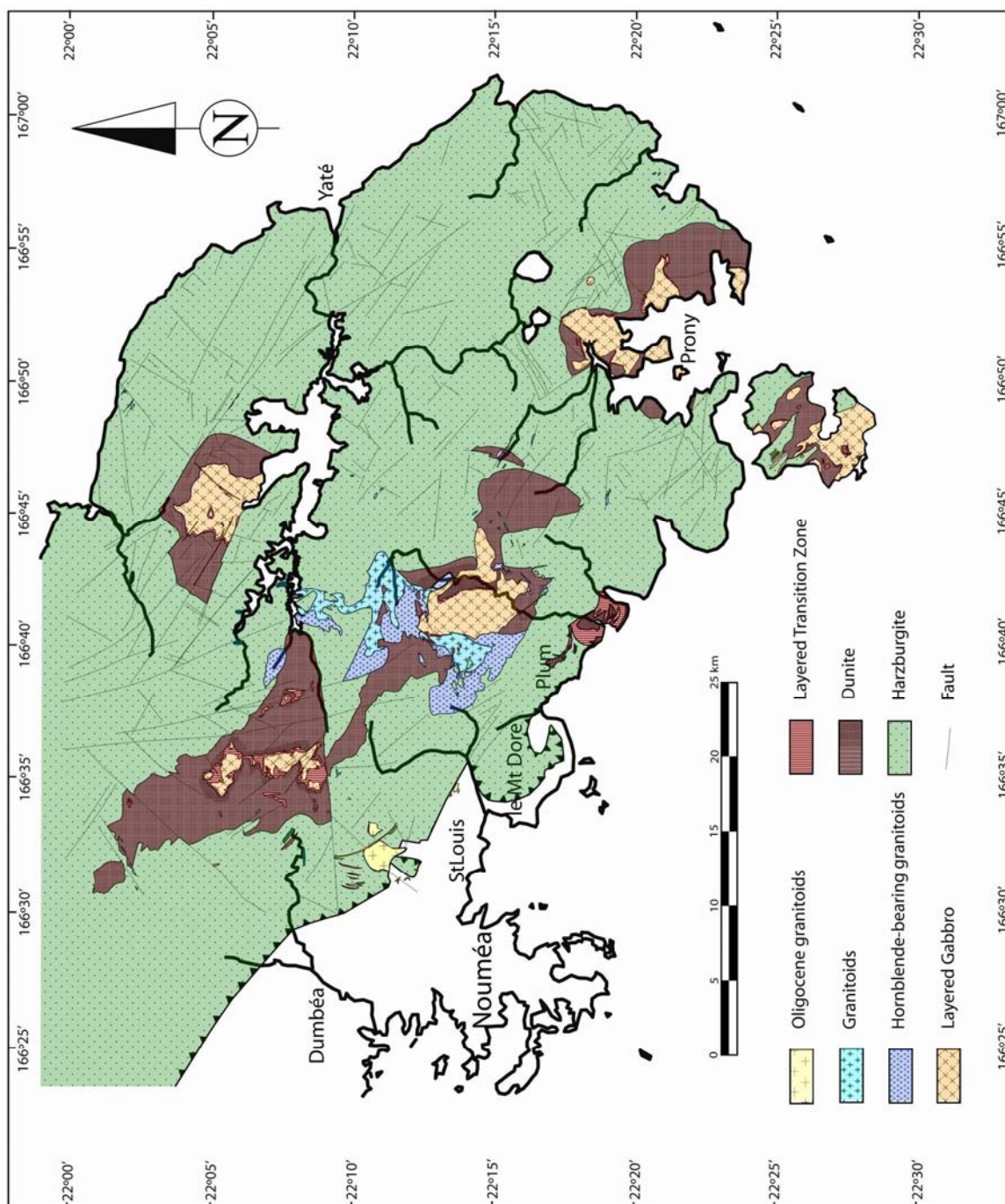


Fig.4.4. Simplified geological map of the southern part of the Massif du Sud. The map is a compilation of data from Noesmoen, 1971; Guillon et al., 1974; Guillon & Trescases, 1976; Trescases & Guillon, 1977; Prinzhofer et al., 1980 and my own field observations.

4.3. Results

4.3.1. Field strategy

Reconnaissance sampling of ophiolitic rocks in New Caledonia (G.Mallmann, H.O'Neill, J.Hermann) allowed some preliminary analyses of the wide variety of rocks which occur in the ultramafic nappe. This set of initial data was of considerable help in formulating the strategy for the field campaign and the collection of samples in the ophiolite.

The first objective was to collect a series of peridotites along a stratigraphic transect which would provide information on the petrological and geochemical variations in relation to palaeo-depth. To attain this aim, the region going from Yaté (east coast) to the Montagne des Sources (centre) was sampled in places where rocks are representative of the area. Other areas were intensively investigated such as Rivière des Pirogues and some additional samples were collected from the north-east coast. This sampling covers the harzburgitic evolution, the transition from harzburgite to dunite and the composition of the transition zone and the crustal mafic rocks according to the B.R.G.M. maps [Fig.4.4; 4.5, Tab.S11].

The second objective was to sample discordant ultramafic rocks (dunites, pyroxenites) which crosscut the ophiolite in many places. Previous studies on the New Caledonia ophiolite consider these dykes essentially as segregation out of the harzburgite. However, preliminary geochemical analyses had shown clear geochemical differences and that more careful research has to be carried out on these rocks. Zones such as the Yaté dam or the Parc de la Rivière bleue are rich in such rocks. Due to the many variables observed among all these discordant rocks (channelling, dyking, folding, composition, texture, structure) during field work, a large number of these rocks were sampled to sort out the different mechanisms that have been responsible for such segregates [Tab.S11].

Finally, the last objective was to investigate abnormalities regarding to a standard ophiolitic sequence. Felsic intrusions are mapped as an important component of the Massif du Sud but lack proper geochemical characterisation

(Rodgers, 1973a). The felsic rocks were sampled in the zone south-east of Montagne des Sources, Parc de la Rivière Bleue, Col de Prony and other small localities scattered in the Massif du Sud. Chromitite complexes have been sampled on some occasions in mines (GR2H, Alain-André, Peut-être) in the dunite zone. The lithological unit mentioned as “euphotide” in Baie des Pirogues (Noesmoen, 1971) is presented as a specific unit which contrast with the rest of the ophiolite and was sampled for further investigation [Fig.4.4; 4.5]. Finally, the Ile des Pins ophiolite, 50km south-east of the Massif du Sud, was sampled as this ophiolite has never been adequately described petrologically [Tab.S11].

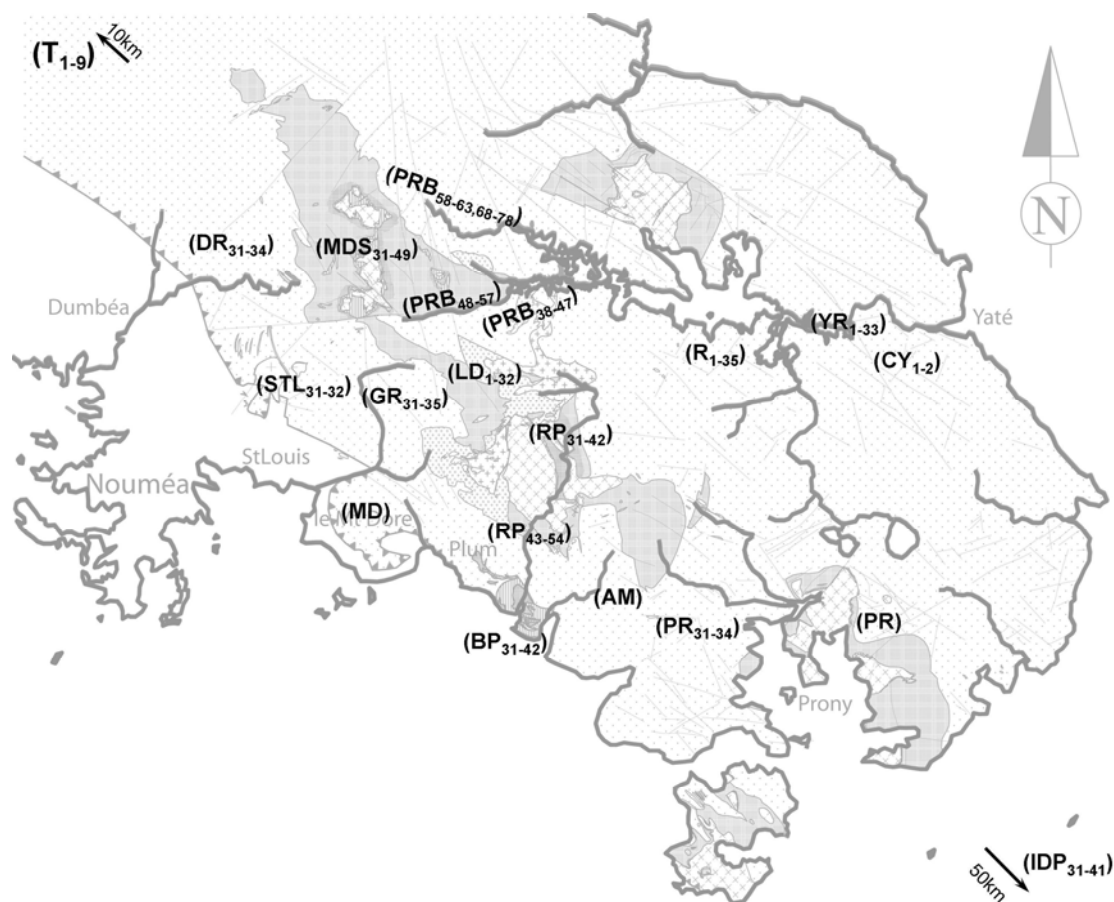


Fig.4.5. Location of the samples collected and studied in part B. **T₁₋₉**: Vallée de la Tontouta (Tontouta valley); **DR₃₁₋₃₄**: Rivière de la Dumbéa (Dumbea River); **MDS₃₁₋₄₉**: Montagne des Sources (Spring mountain); **PRB_{58-63, 68-78}**: Rivière bleue (Blue River); **PRB₄₈₋₅₇**: Rivière blanche (White River); **PRB₃₈₋₄₇**: Rivière de la locomobile (Locomobile River); **R₁₋₃₅**: Route Mont-Dore-Yaté (Road); **YR₁₋₃₃** (and **YA₁₋₁₃**), Barrage de Yaté (Yate Dam); **CY₁₋₂**: Col de Yaté (Yate Pass); **PR**: Prony/Goro; **PR₃₁₋₃₄**: Col de Prony (Prony Pass); **AM**: Plaine du champ de bataille (Battlefield Plain) (Mine de l’Aiglon (Aiglon Mine)); **IDP₃₁₋₄₁**: Ile des Pins (Pine Island); **BP₃₁₋₄₂**: Baie des Pirogues (Pirogue Bay); **RP₄₃₋₅₄**: Rivière des Pirogues Sud (Southern Pirogue River); **RP₃₁₋₄₂**: Rivière des Pirogues Nord (Northern Pirogue River); **MD**: Mont Dore; **LD₁₋₃₂**: Col des Dalmates (Dalmatian Pass); **GR₃₁₋₃₅**: Mine GR2H (GR2H mine); **STL₃₁₋₃₂**: Saint Louis.

The topography, the tropical vegetation and the alteration in New Caledonia limit the collection of fresh samples. Specific areas were selected such as river beds, steep valley sides and the top of some mountains. Field routes were chosen to cover key areas and transitional facies within the ophiolite. In some regions however, the collection of fresh samples is impossible due to laterite cover which can reach thickness as much as 60 m in areas like Goro (Horn & Bacon, 2002; Wells et al., 2009).

4.3.2. Analytical techniques

Mineral major elements chemical analyses were performed on polished thin sections (30 μm thick) with a CAMECA SX100 electron microprobe of RSES, ANU operating in wavelength-dispersion mode. Acceleration voltage and beam current were 15 kV – 20 nA with a focused beam diameter of 1 μm . Counting time per element ranged from 20 to 50 s depending on concentration and time optimisation for each spectrometer. Several analyses on each phase were performed to obtain representative compositions and zoning patterns through grains and samples. Standards were natural (Mg, Si, Na, Al, K, Cr, Mn, Fe) and synthetic minerals (Ca,Ti) as well as pure metals (V, Co, Ni), all analysed on K α emission peak.

Trace elements in minerals were analysed by LA-ICP-MS at RSES, ANU using a pulsed 193 nm ArF Excimer laser with 100 mJ of output energy at a repetition rate of 5 Hz (Eggins et al., 1998) coupled to an Agilent HP7500 quadrupole ICP-MS. Laser sampling was performed in a He-Ar atmosphere with various beam diameter (30 to 250 μm) on several grains to evidence cryptic zoning through samples if present. Analyses were standardised on ^{29}Si (silicates) and ^{27}Al (spinel) calibration as an internal standard isotope based on SiO_2 and Al_2O_3 concentrations measured by micro-analyses techniques. NIST-612 and BCR-2G glasses were used as external standards every 10 to 15 spot analyses.

Whole-rock major and trace elements were determined by standard X-ray fluorescence (XRF) and LA-ICP-MS at RSES, ANU. Rock powders were fused with a flux of lithium tetraborate and lithium metaborate in the ratio 4:25 for major elements and 1:3 for trace elements at 1200°C for 20 minutes. Major

elements were determined with a PW2400 wavelength dispersive X-ray fluorescence spectrometer housed at RSES. Loss-on-ignition was calculated as mass difference of powdered sample after heating to 1090°C for 90 minutes and correcting for weight gain due to ferrous iron oxidation.

4.3.3. Rock descriptions

4.3.3.1. Harzburgite

The lower and main part of the ophiolite is a depleted magnesian spinel harzburgite which is approximately 3000 m thick and covers 85% of the Massif du Sud [Fig.4.4, 4.6]. Flow textures and shearing in the harzburgite classify the rock as a tectonite [Fig.4.7]. This texture is present throughout the ophiolite but can be lost locally when olivine – orthopyroxene segregation is important, forming a secondary layering parallel to the foliation [Fig.4.7c] (Guillon et al., 1974; Guillon & Trescases, 1976; Trescases & Guillon, 1977). The general stratigraphy shows an intensification of foliation up-section and an increasing fraction of olivine, such that the rock changes from orthopyroxene-rich harzburgite to dunite with some orthopyroxene grains.

The primary mineralogy of most harzburgites is dominated by Mg-rich olivine which comprises 75-85% of the rock. Grain size can vary from 1 cm to more than 5 cm but grains are always fractured and serpentinised and show many subgrain domains. Orthopyroxene forms 10-25% of the rock and occurs as large cm-sized elongated porphyroclasts [Fig.4.7b]. Spinel is Cr-rich and forms between 2 to <10% of the rock. Crystals can occur as symplectitic porphyroclasts of pyroxenes and spinels (20% of spinels in samples) in some areas (e.g. Yaté Dam) and contrasts with xenomorphic chromite blebs which normally grow between olivine or orthopyroxene crystals [Fig.4.7d; 4.26]. Clinopyroxene does not occur as a primary phase and is only present as minute exsolution laths in orthopyroxene or as larger zones in pyroxene + spinel symplectites. In all cases, its modal concentration in harzburgite is less than 0.5%. Accessory primary phase are mainly sulfides and graphite (Augé et al., 1999) and occur only in the lower part of the ophiolite.

The base of the harzburgite section is formed by a sole of serpentinite resting on the previously obducted terranes or the Ante-Senonian basement. Ultramafic outcrops found in more distal position from this basal contact, and far from any major fault going through the massif, show less intense serpentinisation and 30% of olivine is altered in average in the peridotite. However, relatively fresh samples can be easily found with less than 10% serpentine and in some rare occasions, almost none [Tab.S11]. In altered samples, serpentine (lizardite, antigorite) is common and occasionally associated with brucite, carbonates, magnetite and some Ni-Fe-Cu sulfides and alloys. Bastite, which is the variety of serpentine altering orthopyroxene, only occurs in the most altered samples.

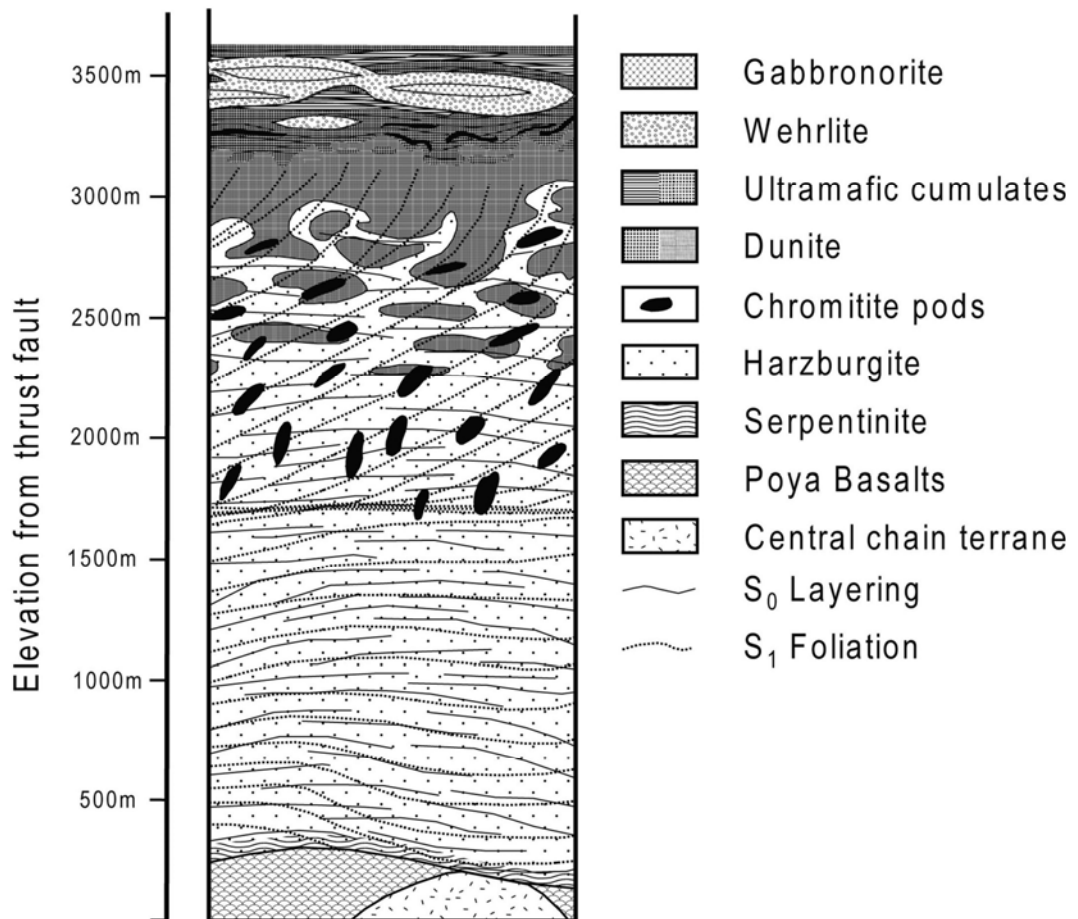


Fig.4.6. Schematic representation of the stratigraphical column in the Massif du Sud.

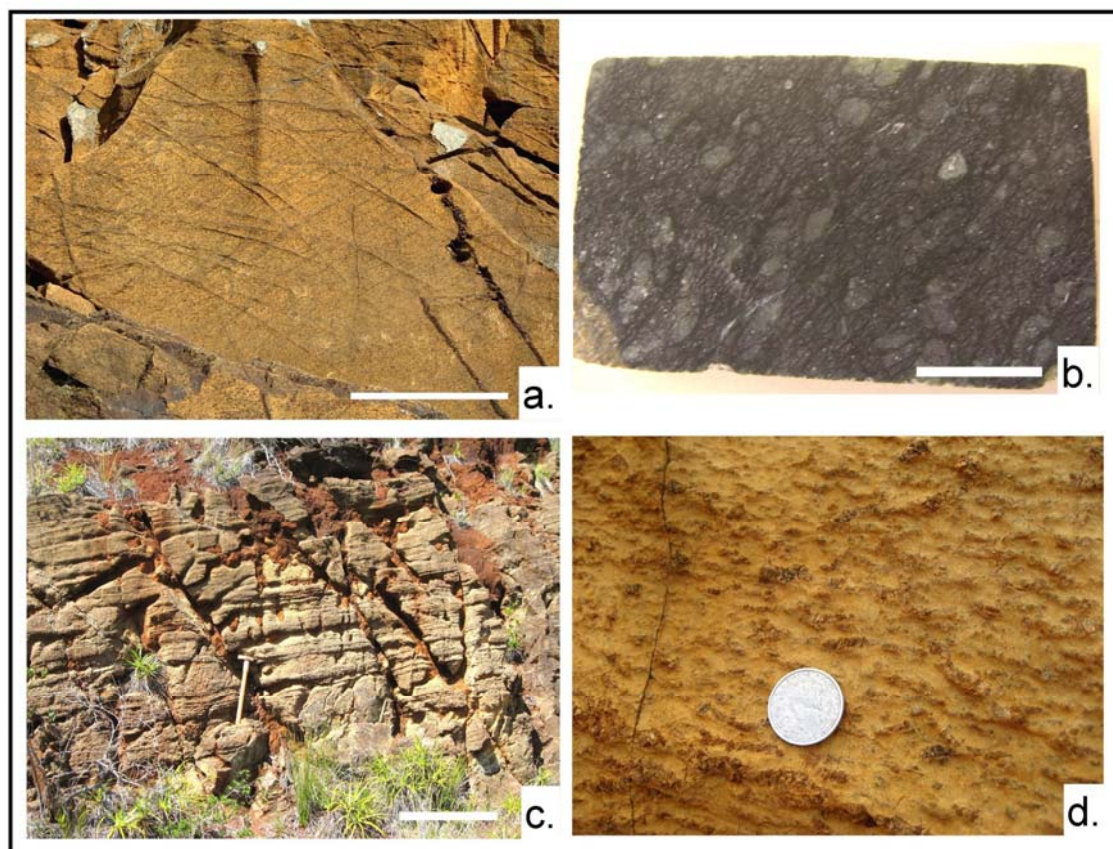


Fig.4.7. Peridotites from the middle and lower part of the Massif du Sud.

a. Harzburgite outcrop crosscut by numerous generations of orthopyroxenite dykes. The foliation is given by the earliest stage of pyroxenite veins dipping 45° left. (Yaté dam (YR-series)) (Scale bar is ~1 m).

b. Thin section block showing the tectonite texture underlined by the orientation of orthopyroxene porphyroclasts (light grey) in the olivine matrix (dark grey). (Rivière bleue) (Scale bar is 1cm).

c. Strongly foliated harzburgite giving a layered appearance of orthopyroxene-rich and -poor layers (GR2H mine) (Scale bar is ~1 m).

d. Harzburgite in the Yaté area with orthopyroxene porphyroclasts more resistant to alteration than olivine giving this rugged surface texture. Some porphyroclasts display spinels in symplectitic textures (Coin in 3.5 cm).

4.3.3.2. Dunite

Higher in the ultramafic sequence, orthopyroxene modal content in the harzburgite decreases progressively and finally disappears in the dunite zone. The transition between the two rocks is not always obvious and is highly variable. Some places show a progressive drift from olivine-rich harzburgite to dunite whereas other places display a standard opx-poor harzburgite in a sharp contact with a dunite [Fig.4.8a]. In addition, this transition occurs several times along the stratigraphy and it suggests that dunite pods and layers, which are common in the upper harzburgite, define the harzburgite/dunite transition

at the mapping scale when pods greatly outnumber the host harzburgite and become the main lithology [Fig.4.6]. The dunite zone is around 600 m thick in the Montagne des Sources area and is very homogeneous in its lower part (-600 m to -200 m).

The composition of the dunite is mostly restricted to olivine which forms 90-95% (rarely up to 98%) of the rock in the lower dunite zone. Chromite is the other magmatic phase and forms between 5 and 10% modal in the dunite zone. In some areas below the crust-mantle transition, chromite can show monomineralic accumulation as layers and schlierens of chromitite (>50% spinel) [Fig.4.8b]. The tectonite texture, which was still strongly visible in the olivine-rich harzburgite and in the lowest dunite, becomes macroscopically absent in the upper dunite zone. Textures tend to be very homogeneous or have rare cumulate characteristics (layering) in the upper part of the dunite zone. However, crystals remain large and olivine can reach decimetre size. Chromite is subidiomorphic and can still sometimes occur as holly leaf shaped grains but xenomorphic blebs or large symplectite are absent.

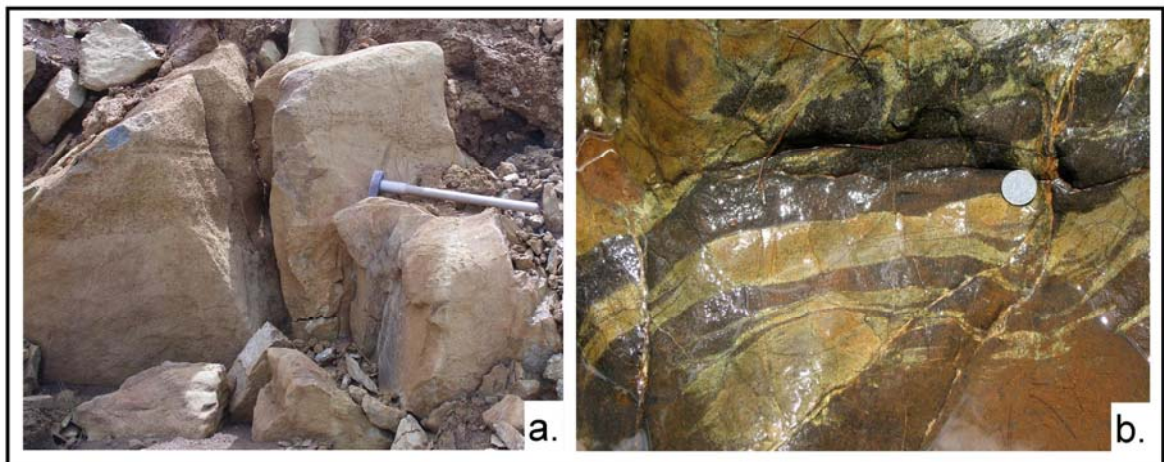


Fig.4.8. Rocks from the dunite transition zone

a. Dunite pod in the Pass of Prony. Dunite is the clear homogeneous texture at the bottom whereas harzburgite is more rugged and has a sharp contact with the pod (PR31) (hammer is 70 cm long).

b. Chromitite schlierens in the replacive dunite of Rivière des Pirogues (RP31). Despite pseudo-layered textures, these rocks are not cumulates (Coin is 3.5 cm).

4.3.3.3. Transition zone

The dunite zone ends with the reappearance of other silicates. Hundreds of meters of olivine $\pm$ chromite which only display very subtle and microscopic changes contrasts strongly with the first appearance of cumulus pyroxenes. The crystallisation of thick layers of orthopyroxene and clinopyroxene, and the presence of interstitial clinopyroxene and plagioclase between the dunite grains classically mark the transition zone [Fig.4.9]. The structure of the transition zone starts to display more obvious cumulative character on a large scale of observation. Layering is also visible at smaller scales but it is more localised [Fig.4.12a,d]. The texture of the rock generally shows a rounded olivine matrix going from pure dunite (<5% chromite) to rocks where olivine keep similar shape but form less than 30% of the rock. Orthopyroxene crystallises as relatively large grains of sizes similar or larger than olivine [Fig.4.12a]. In some areas, massive layers of pyroxenites or olivine wehrlite are also observed and present cumulative textures [Fig.4.11]. When olivine is subordinate, clinopyroxene $\pm$ plagioclase start to form larger crystals and tend to include olivine grains in larger clinopyroxene oikocrysts [Fig.4.9]. Spinels are present in all these transitional rocks but tend to disappear in the same ratio as olivine. Mapping showed that although cpx- or plag- bearing dunites are abundant in the lower levels, wehrlite or plagioclase-wehrlite are the most common rocks in the area surrounding gabbros [Fig.4.9b]. This typical sequence has been interpreted as cumulative by previous authors (Prinzhofer et al., 1980) or as impregnation (Marchesi et al., 2009). Anorthosites and troctolite have been observed but are marginal compositions in the transition zone and more representative of local variations.

While the transition zone is broadly similar in all localities which have been studied, some local variations are observed. In Montagne des Sources, the sequence is relatively simple, as described above. In Rivière des Pirogues, the palaeo-Moho was examined from downstream (north) and upstream (south). The upstream sequence is similar to Montagne des Sources although more chromitite bodies and late intrusions are present. In the downstream sequence, the transition zone is sharper with clear cumulates and shows a large variety of rocks as well as the presence of late intrusions [Fig.4.11]. In

the Prony area, the transition zone is broadly similar but plagioclase-rich rocks are more common (C.Gineste, pers. comm.). Finally, in Baie des Pirogues, the transition zone shares similarities with other areas such as poikilitic wehrlites and dunites. However, later magmatism has modified the composition of the rocks and this subject is treated separately in chapter 6.

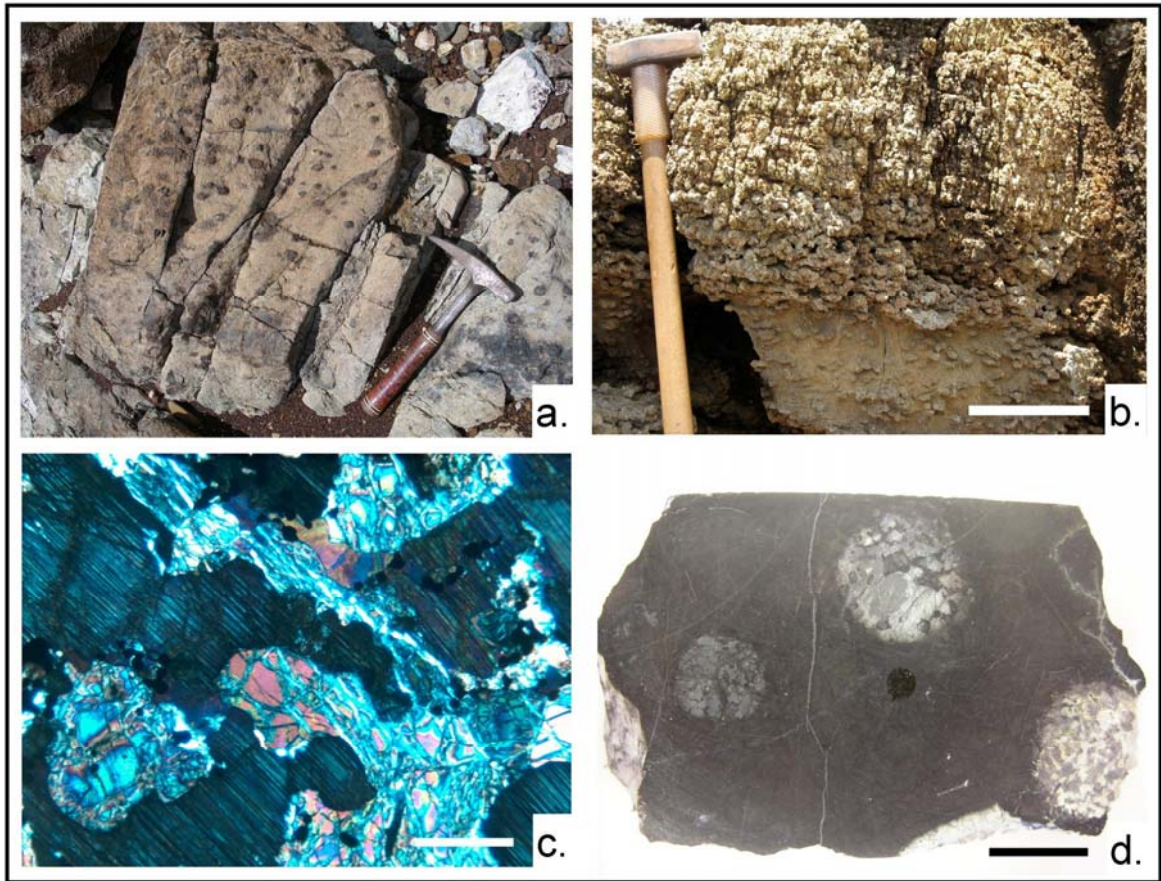


Fig.4.9. Wehrlitic replacement textures in dunites.

- a.** Bigpox texture of clinopyroxene-bearing dunite. Clinopyroxene occurs as large centimetre-sized oikocrysts in the dunite giving a spotted appearance (RP 38) (Hammer is 40 cm long).
- b.** Transition from cpx-dunite to clinopyroxenite with minor olivine and plagioclase. This feature occurs commonly in the area surrounding gabbro sills (MDS38) (Scale bar is ~20 cm).
- c.** Poikilitic texture viewed in crossed polars. Olivine is visible as isolated grains and clinopyroxene is at an extinction angle (MDS35) (Scale bar is 100 μ m).
- d.** Thin section block of a dunite with three small sized clinopyroxene oikocrysts. (Scale bar is 1 cm).

4.3.3.4. Gabbro

Gabbros are represented on the geological map as uniform and consistent bodies. However, field observations show that gabbros occur as large thick sills in rocks of the upper transition zone and do not form a continuous formation on their own [Fig.4.10]. If a higher stratigraphic level of the ophiolite were observed in New Caledonia, it might be expected that gabbros would become the main lithology as in other ophiolites. Two out of the five gabbroic massifs have been studied and sampled during fieldwork. The structure observed at Montagne des Sources and Rivière des Pirogues is generally a succession of lenses of several meters thick with diffuse boundaries evolving in plagioclase-bearing wehrlites [Fig.4.9]. Layering is visible in some places [Fig.4.12d] but is not widespread and evident in all outcrops or in thin sections.

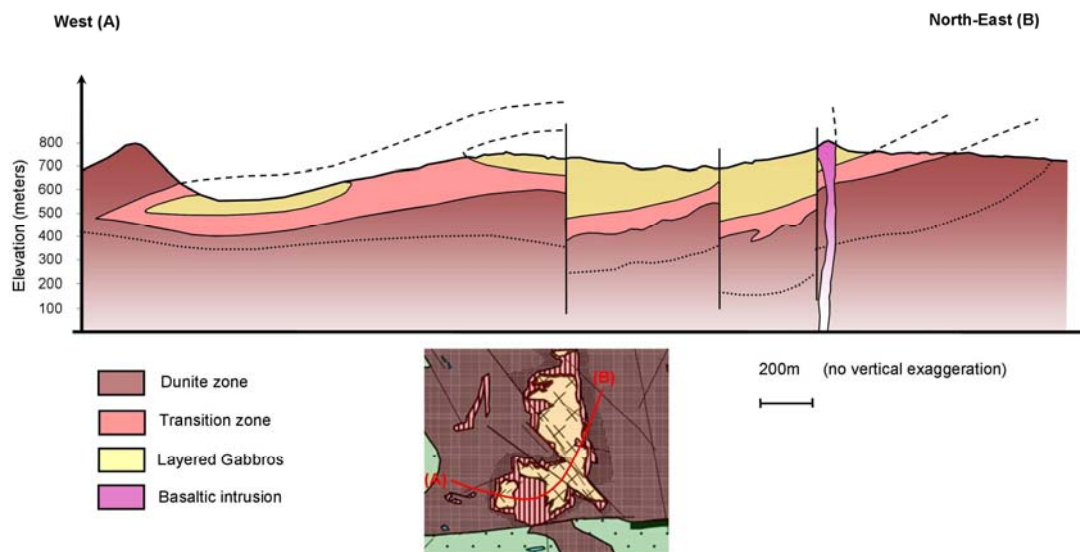


Fig.4.10. Cross-section through the southern Montagne des Sources gabbro massif to display evidence of more than a single gabbroic body on the basis of stratigraphy and structural data on the area.

Gabbroic rocks have a range of composition going from mela- to leuco-gabbro with a noritic component (5-20%) which has been noticed in every rock sampled. Olivine grains are present as an accessory mineral. They form rounded and fractured grains in a matrix of large crystals of plagioclase, clinopyroxene and orthopyroxene [Fig.4.13a-c]. Spinels are very rarely present

and are not Cr-rich. Alteration phases show iddingsite mixtures in olivine grains and hydrogrossular on the rim of plagioclase. Uralitisation and saussuritisation are frequent in some areas and form tremolite, nephrite, epidote and smaragdite.

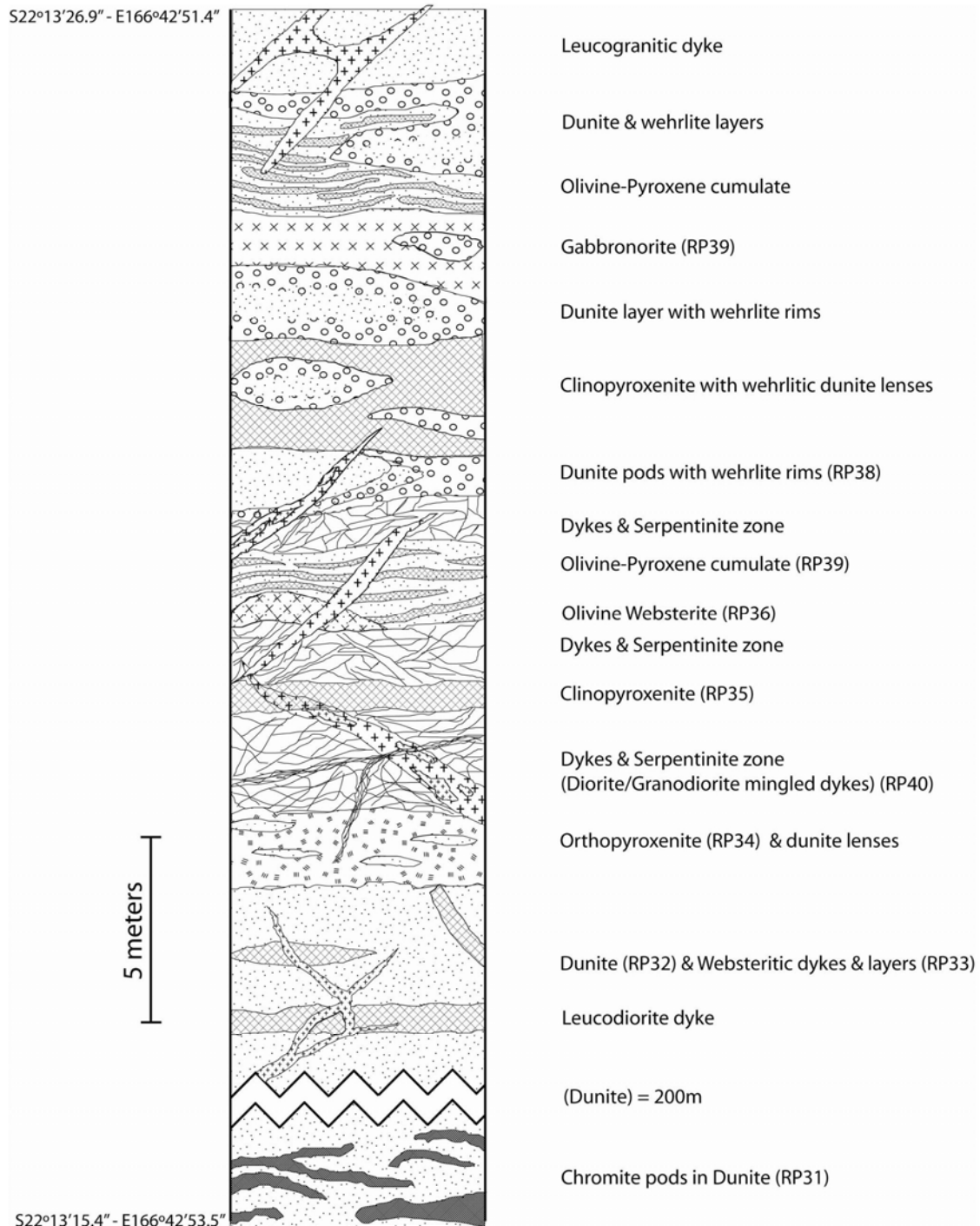


Fig.4.11. Field sketch of the transition zone in the downstream sequence of Rivière des Pirogues.

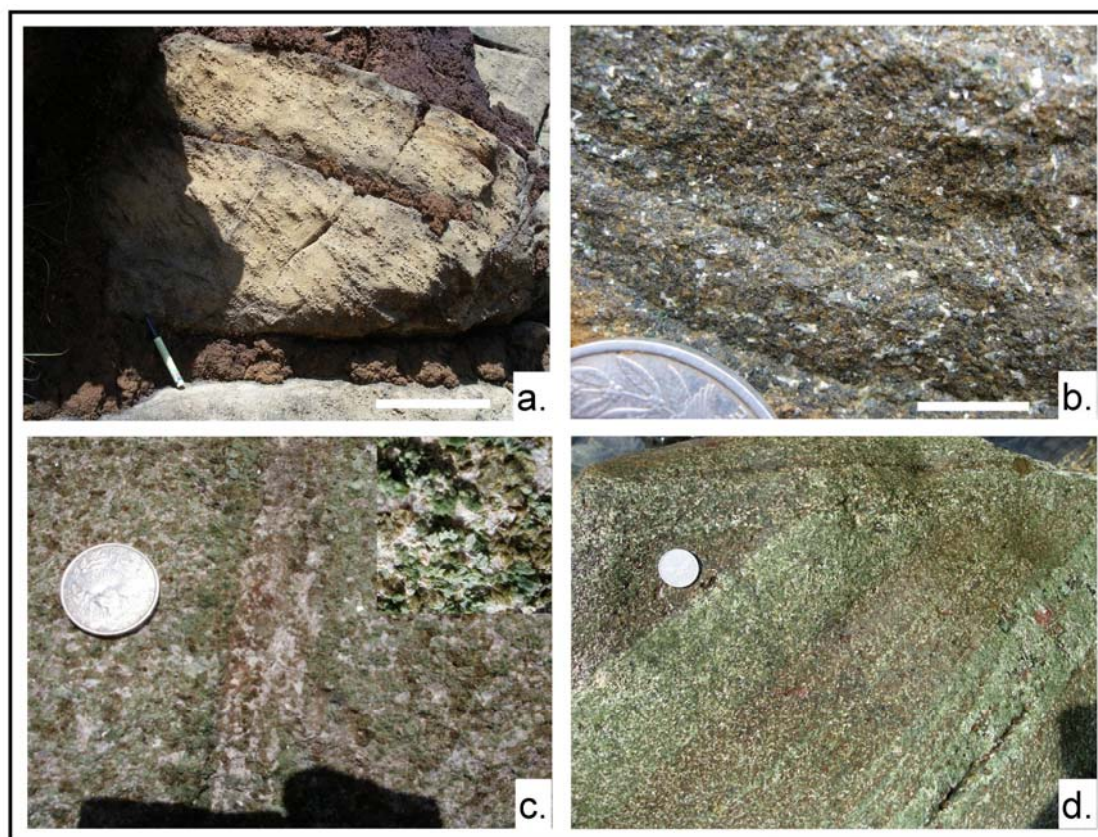


Fig.4.12. Crustal cumulates in Montagne des Sources area.

a. Olivine-orthopyroxene cumulate with minor impregnation showing a layered texture in the lower part of the transition zone (MDS49) (Scale bar is ~30 cm).

b. Impregnated dunite with olivine (dark green), pyroxenes (orange or greenish) and plagioclase (white) (MDS35) (Scale bar is 1 cm).

c. Gabbronorite with plagioclase (white) clinopyroxene (green) and orthopyroxene (brown) in the lower Sill of the Casse-Cou mountain (MDS37) (Coin is 3.5 cm).

d. Layered olivine gabbro boulder in Rivière bleue showing alternation of gabbronorite (green white) and olivine gabbro (brownish) and discordant layering (top) related to magma stoping events in the oceanic crust (coin is 3.5 cm).

4.3.3.5. Other rock types

Felsic rocks

In addition to the ultramafic series, felsic rocks are present in the southern part of the Massif du Sud. These rocks occur as dykes, pods or larger intrusions with a large range of compositions and are discussed in chapter 6.

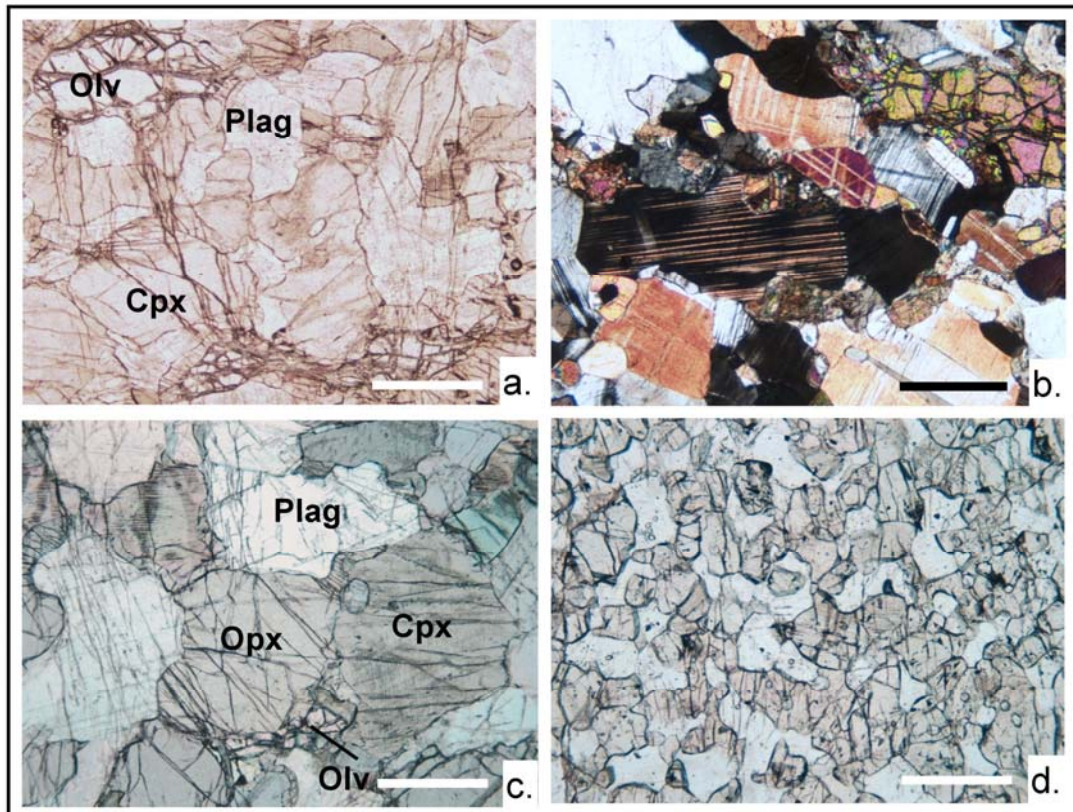


Fig.4.13. Microscopic observations in gabbros (Scale bar is 250 μm).

- a.** Gabbronorite with accessory olivine (RP48).
- b.** Crossed polar view showing slight layering textures for plagioclase, clinopyroxenes and olivine in gabbronorite (RP51).
- c.** Coarse grained gabbronorite with minute olivine grains (MDS37).
- d.** Microgabbro dyke with plagioclase and clinopyroxene (MDS48).

Ore bodies

In addition to nickel mining, ultramafic rocks are known for their high chromium content. In New Caledonia, chromite exploitation has been going on more than a century and only stopped in 1990's in Tiébaghi. Chromite deposits occur as chromitite pods or layers and have been mined from the upper harzburgite and dunite zone. The general structure of a chromitite pod is a zoned discordant intrusion in the upper harzburgite. The outer rim is a heterogeneous harzburgite where chromite and sulfides are enclosed in orthopyroxene trails chemically and optically different from orthopyroxene porphyroclasts present in normal harzburgite [Fig.4.14a]. The inner zone is formed of dunite similar to the dunite zone and contains long schlierens of chromite ore [Fig.4.14c]. Modal proportions vary from 10-95% chromite with a large range of textures. Low chromite content pods generally show textures similar to the dunite zone with small idiomorphic spinel crystals scattered

homogeneously. At higher grade, a whole range of textures such as leopard, nodular, anti-nodular, orbicular, layered (Cassard et al., 1981) exists between olivine and chromite [Fig.4.14b,c]. For modal composition with more than 75% of chromite, olivine is a serpentinised interstitial phase but can remain as individual hexagonal shaped grains enclosed in a chromitite matrix [Fig.4.26]. Chromitite pods also have the particularity to contain clinopyroxene as inclusions in spinel crystals in a stratigraphic level where clinopyroxene is barely present, even as a secondary phase.

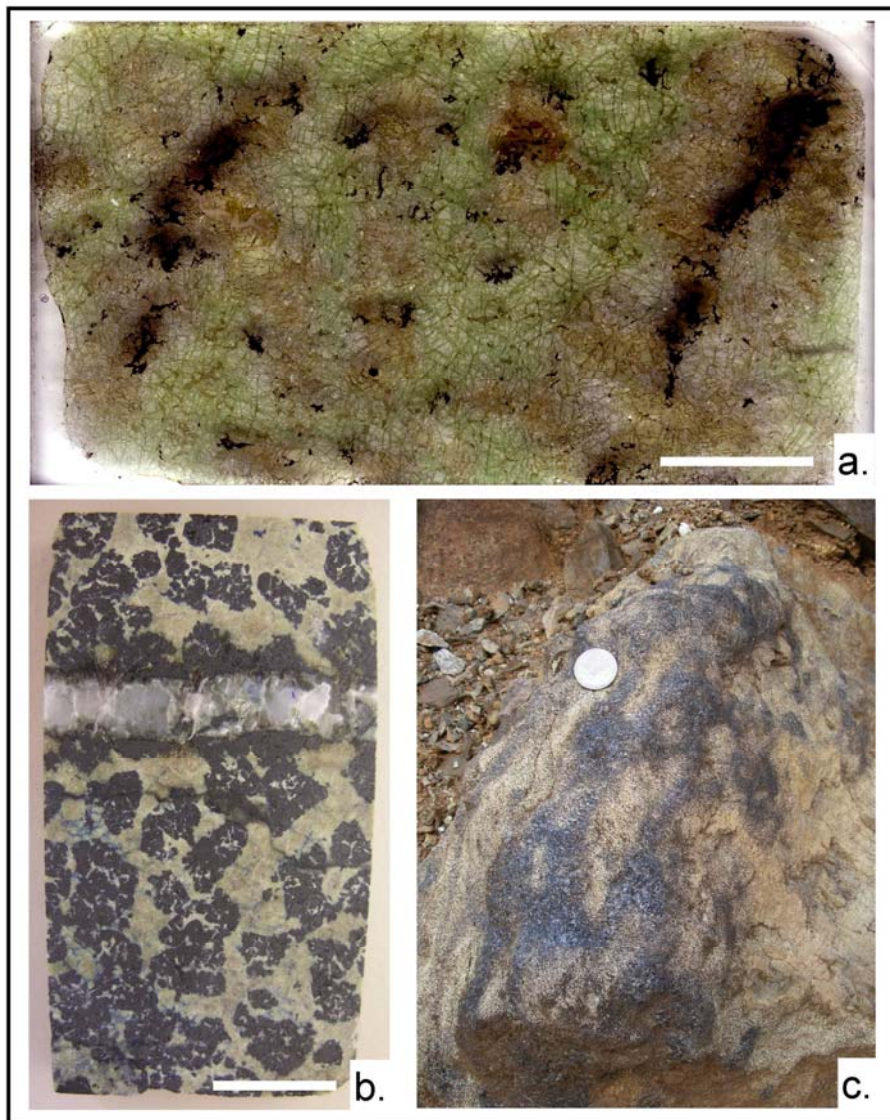


Fig.4.14. Chromitite deposits south-east of Montagne des Sources.

- a.** Thin section of a metasomatised peridotite surrounding chromitite deposits, green is olivine, white is orthopyroxene and black is chromite (PRB1) (Scale bar is 1 cm).
- b.** Thin section block of a chromitite deposit (olivine, yellowish; chromite, black) crosscut by a tardive gabbro vein (white). (RP43) (Scale bar is 1 cm).
- c.** Chromium ore with olivine (beige) and chromite (dark) in a sub-antinodular ore type in GR2H mine (GR32) (coin is 3.5 cm).

Cumulative pods

A few samples were collected because their textures or composition were unusual for their position in the ophiolitic sequence. Such rocks occur in the harzburgite, the dunite zone and the transition zone and seem to bear similarities. These rocks can have clear cumulus textures (olivine, pyroxenes) [Fig.4.15b] or impregnation textures in olivine-rich rocks which are very different from surrounding rocks. The mineralogy shows an assemblage of orthopyroxene, clinopyroxene and olivine (in harzburgite) [Fig.4.15b; 4.23] or diopside and olivine (in gabbros) [Fig.4.15a,c]. However, the strong modal layering of some samples gives composition alternating from dunite to websterite in hand specimens. Plagioclase is occasionally seen in clinopyroxene-bearing samples. Olivine-orthopyroxenite cumulates are particularly remarkable for their high content of sulfides of nickel and iron, plus chromite and rutile [Fig.4.23].

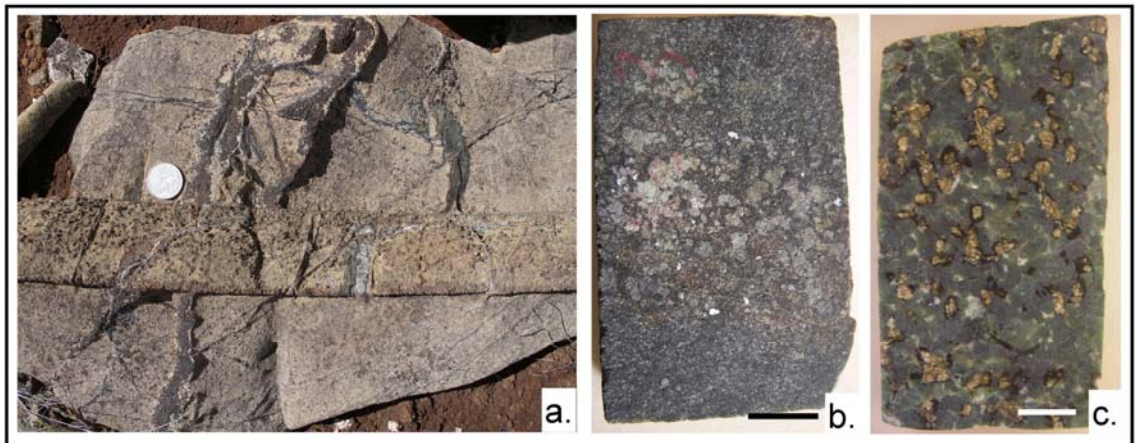


Fig.4.15. Cumulate pods and related rocks.

a. Clinopyroxene-bearing dunite dyke with clinopyroxenite walls crosscutting cumulus dunite and layers of chromitites (black) (MDS44-45) (Coin is 3.5 cm).

b. Cumulus texture in a pod showing alternating dunite (dark grey) and olivine websterite (coarse grained light grey layers) (RBL3) (Scale bar is 1 cm).

c. Diopside (green) with olivine crystals (orange) and rare interstitial plagioclase (whitish) (MDS42) (scale bar is 1 cm).

Dykes & Channels

A large range of discordant rocks are present in the Massif du Sud ophiolite. Detailed descriptions, analyses and interpretations of these rocks are developed in chapter 5 and 6. In the area below the dunite zone, dunite channels are an abundant feature and are easily recognisable from the surrounding harzburgite by the total lack of pyroxenes and smooth meteoric

alteration textures. In many areas, pyroxenites dykes have also been described and display a wide range of features [Fig.4.7a]. Most of these dykes are orthopyroxenites but clinopyroxene-rich and amphibole-bearing veins have also been observed. In most cases, these dykes are exempt of olivine and spinels.

Some basaltic dykes have also been sampled in the dunite zone and in the gabbro. These rocks are clearly volcanic with fine-grained textures and millimetre-size amphibole phenocrysts. Their volcanic textures and the relationship with the host rock suggests that these are a later production of magma and are likely to be related to the post-obductive magmatism (St. Louis; Koum-Borindi granitoids) and hydrothermalism (quartz veins) which affected the Massif du Sud during Oligocene.

4.3.4. Major element mineral chemistry

4.3.4.1. Olivine

Olivine is the most common mineral in the ultramafic terrane. With the exception of some pyroxenite dykes and felsic intrusions, grains of olivine can be found at all stratigraphic levels of the ophiolite. Olivine shows a chemical composition which only varies over long distances in the stratigraphy. The geological mapping of the area, coupled with the structural studies of Prinzhofer et al. (1980) provides an approximate stratigraphic sequence of the Massif du Sud. This stratigraphy is also confirmed in this study by the variation of Mg-number of the olivine which ranges from 0.895 to 0.927 for harzburgites [Fig.4.16]. The lowest values in Mg-number are found on the east coast where the peridotite is supposed to come from the deepest. At Yaté and Thio, olivine has Fo_{89-91} with local variations which are related to segregation processes or a later event such as pyroxenite dyke emplacement. Along a transect towards the middle of the island, Mg-number remains relatively stable around Lake Yaté and starts to increase in the area surrounding the gabbro massifs. In the river valleys extending east of Montagne des Sources, Mg-number shows an increase from 0.915 to 0.93 and stabilises around the value of 0.925(5) in the lower dunite zone. In the transition zone, dunite has olivine which has a lower

Mg-number (Fo_{90}) clearly distinct from olivine analysed in the lower dunite zone. The Mg-number of olivine drops in mafic cumulates to Fo_{84} in gabbros [Fig.4.17, Tab.S12]. Dupuy et al. (1981a) mentioned values as low as Fo_{80} (and En_{60}) in that same region but the provenance of the samples was not clearly described and this might apply to post-obductive basaltic dykes.

NiO and MnO contents show small differences between lithologies [Fig.4.18]. In harzburgites, no particular change in composition is seen with stratigraphy (0.41 ± 0.03 wt.% NiO; 0.12 ± 0.02 wt.% MnO). Dunites have a slightly lower NiO content than harzburgite but MnO does not show any variation. In the cumulate series, NiO content drops from mantle values (~ 0.37 wt.%) to low values in gabbros (0.15 wt.%). MnO has the opposite trend and increase from 0.12 to 0.25 wt.% above the crust-mantle transition [Tab.S12].

Discordant rocks such as chromitite pods occur at different levels in the upper harzburgite and in the dunite zone. These rocks often display a trend with higher Mg-number for olivine than their respective host rock [Fig.4.16]. In the case of chromitite bodies, the high modal content in chromite has a strong influence on subsolidus re-equilibration of Mg-Fe in the olivine-spinel system (Irvine, 1967). This rotating tie-line effect results in Mg-number values as high as 0.96 [Tab.S12].

However, heterogeneous harzburgite surrounding chromitite pods show a similar enrichment in magnesium ($\text{Mg\#} = 0.940(5)$) where strong subsolidus Mg-Fe exchange cannot be involved. These series of rocks also show enrichment in nickel and depletion in manganese [Fig.4.18]. The cumulative pods previously mentioned have Mg-numbers which are similar to the surrounding rocks. The NiO content of the olivine is in most cases similar to gabbroic rocks, which clearly have a different origin [Fig.4.18a, Tab.S12].

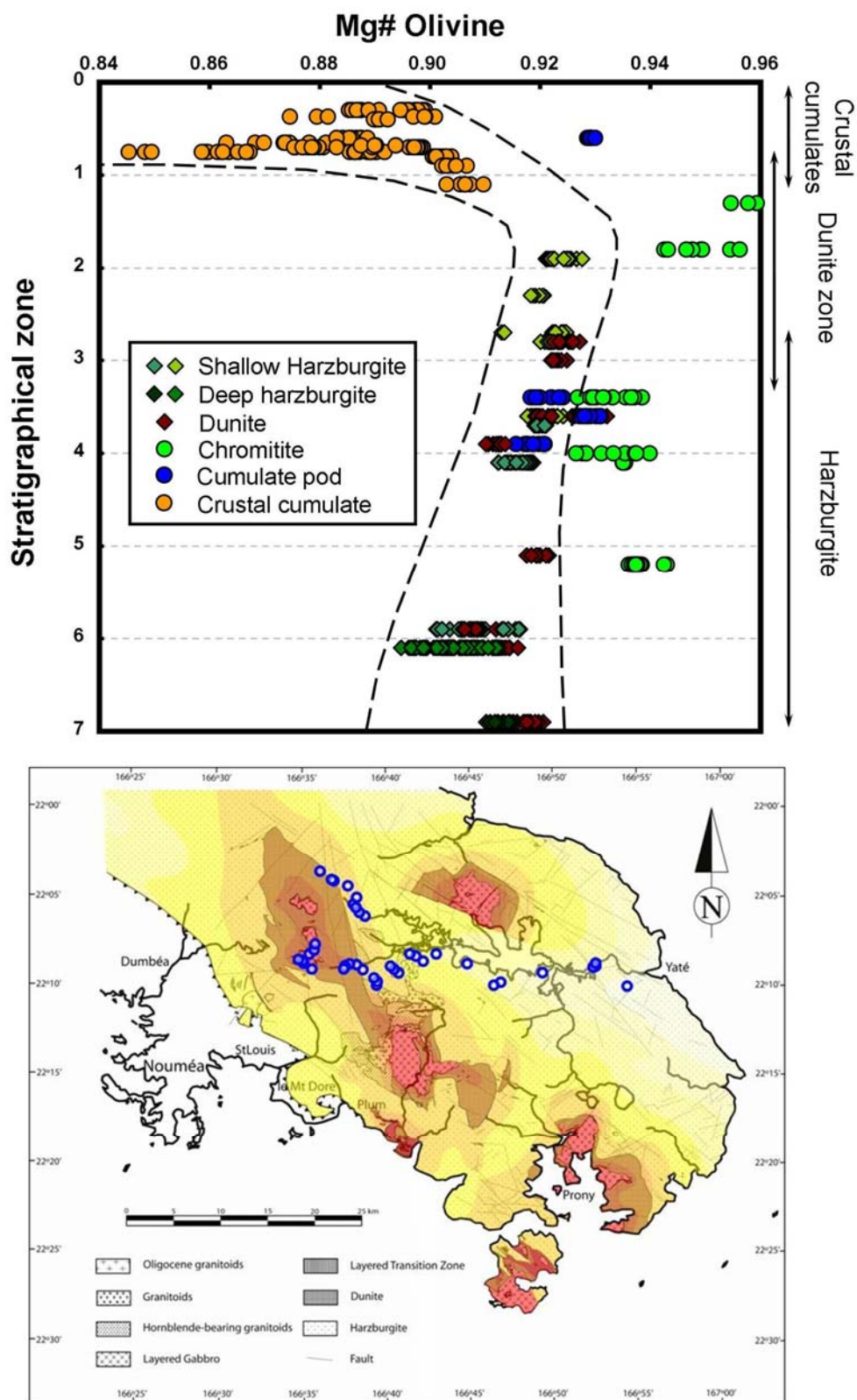


Fig.4.16. Mg# of olivine versus approximate stratigraphy in the Massif du Sud ophiolite. The stratigraphical zones are assigned in relation with the map for the Yaté-Montagne des Sources area (samples are indicated in blue). The palaeo-depth colour gradient is established on modal petrology and structural studies. An increase in Mg-number is noticed in olivine, particularly in region closer to the dunite zone. When ultramafic cumulates start to occur, the Mg-number drop to lower values.

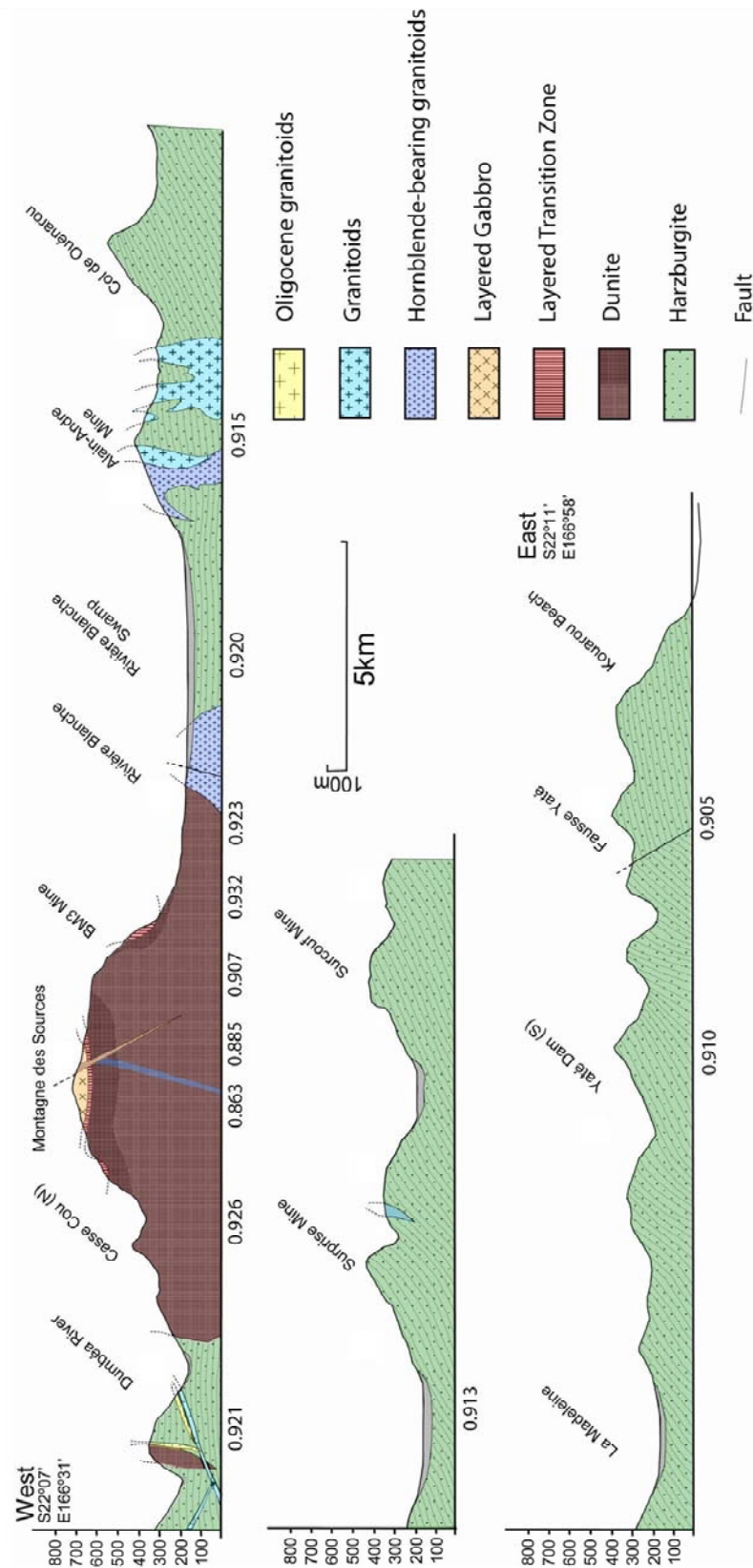


Fig.4.17. Cross section from Yaté to Montagne des Sources and Dumbéa River showing the stratigraphical sequence and the associated stratigraphical planes. Mg-number of olivine for specified area along the transect are indicated on the x-axis.

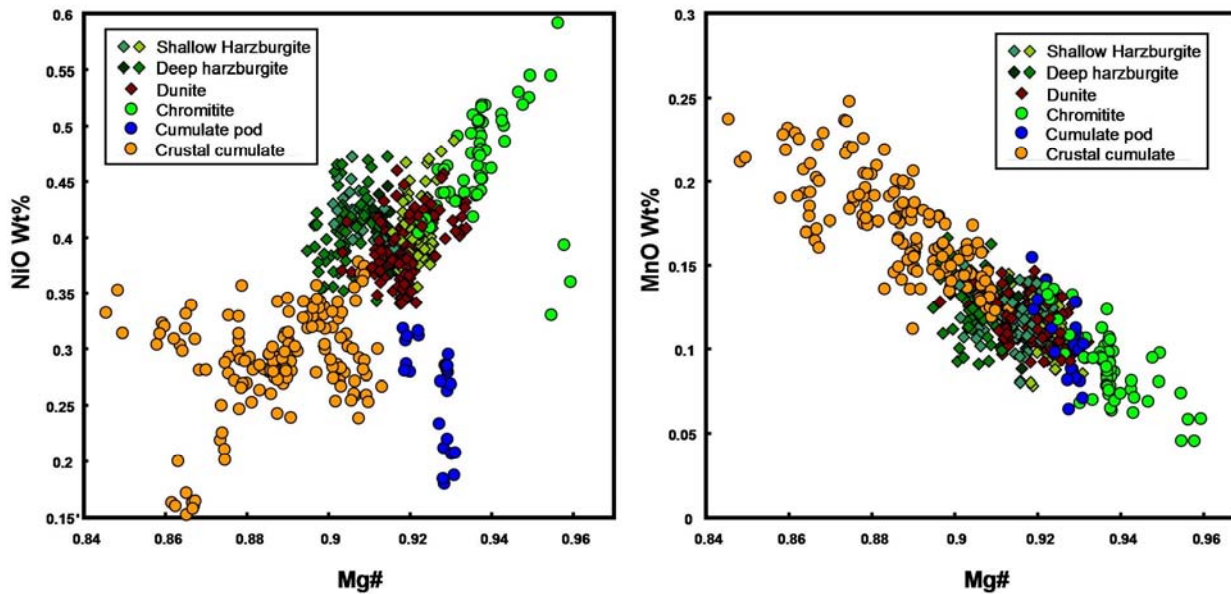


Fig.4.18. Major element data for olivine.

- a.** Mg# versus NiO content. Three main magmatic trends emerge from the harzburgite field. Chromitites (bright green), crustal cumulates (orange) and cumulates pods (blue).
b. Mg# versus MnO content. Trends are still present but develop in an opposite way than NiO for chromitite and crustal cumulates whereas cumulate pods are still poor in MnO

4.3.4.2. Orthopyroxene

The chemical composition of orthopyroxene in the harzburgite mirrors the behaviour of olivine. From the east coast to the dunite zone, Mg# increases from 0.899 to 0.932. Calcium contents show a weak positive correlation with Mg# and varies from ~0.35 to 1.05 wt.% CaO [Fig.4.19]. Al_2O_3 and more particularly Cr_2O_3 are highly related to the stratigraphic position. Alumina varies from ~0.65 to ~1.5 wt.% Al_2O_3 and Cr_2O_3 from 0.2 to 0.85 wt.% in the uppermost harzburgite [Fig.4.20]. In the mafic cumulates, orthopyroxenes are high in alumina (2 ± 0.5 wt.% Al_2O_3) and have high CaO in the early cumulates (1.5 to 2.5 wt.%) but drops to less than 1 wt.% CaO in mafic cumulates where plagioclase accumulation is important. Cr_2O_3 decreases from 0.8 in the ultramafic cumulate to 0.5 wt.% in gabbro-norites [Tab.S14].

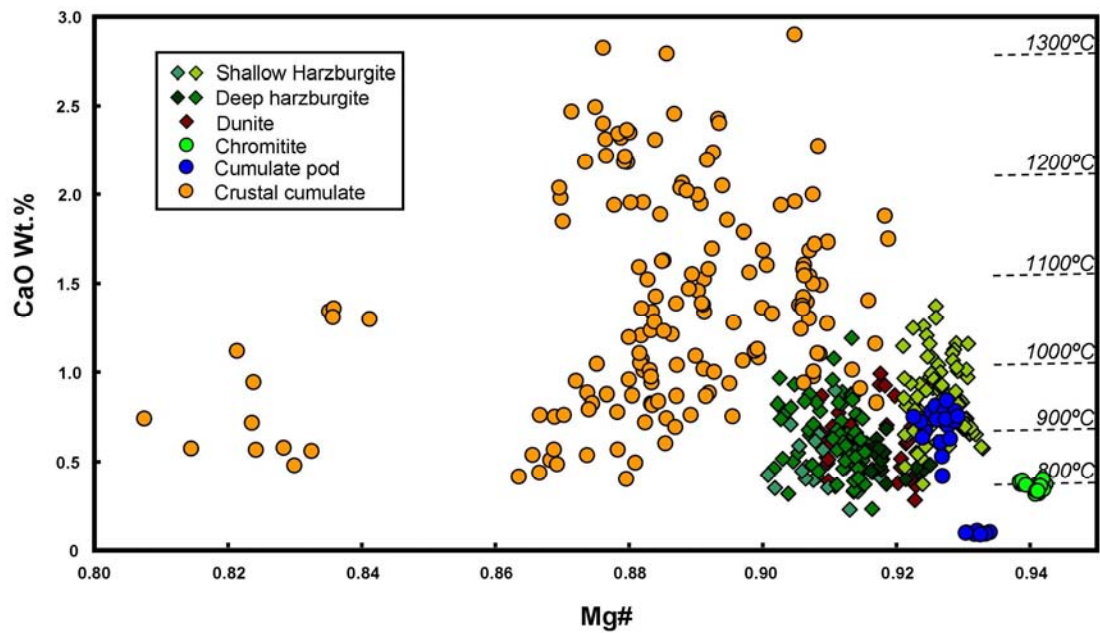


Fig.4.19. Mg-number versus CaO in orthopyroxenes shows a slight increase in CaO in the harzburgite (green diamonds) with stratigraphy (dark to light colours). However, orthopyroxene contains frequent clinopyroxene exsolutions formed during low temperature re-equilibration and CaO content are not representative of initial magmatic content. Orthopyroxenes in gabbros are Ca-rich but highly variable as a function of clinopyroxene and plagioclase present in the assemblage. Lines are approximate clinopyroxene saturation in a Mg-rich system (Brey & Kohler, 1990)

Chromitite bodies lack of orthopyroxene but the surrounding harzburgite shows high Mg# (0.94 ± 0.02) with high content in Al_2O_3 (2 wt.%), Cr_2O_3 (0.9 wt.%) [Fig.4.20] and low content in CaO (0.37 wt.%) [Fig.4.19]. These compositions confirm the metasomatic origin of these harzburgites as they clearly contrast the normal mantle harzburgite. Cumulative pods have Mg-number similar to their host rocks or slightly higher. Al_2O_3 , CaO and Cr_2O_3 content in orthopyroxene is highly variable and depends on the mineralogy of single layers within those pods [Tab.S14].

4.3.4.3. Clinopyroxene

Primary clinopyroxene is absent from the peridotites of the Massif du Sud. As a consequence, clinopyroxene analyses obtained for these rocks are mostly extracted from exsolutions in orthopyroxene porphyroclasts or from spinel-pyroxene symplectites. Mg-number is higher than for orthopyroxene (0.938 to 0.956) and Al_2O_3 and Cr_2O_3 have analogous behaviour to the orthopyroxene host [Fig.4.31]. Alumina is 1 ± 0.2 wt.% and barely increases in

higher stratigraphic levels whereas Cr_2O_3 ranges from 0.4 to 1.2 wt.% with decreasing palaeo-depth. CaO has an opposite behaviour to the concentration in orthopyroxene. Clinopyroxenes analyses in Yat  (east coast) have 26 wt.% CaO and this value drops to 22.5 wt.% in the olivine-rich harzburgite of the uppermost mantle [Fig.4.21]. In the cumulate series, clinopyroxene is a primary and dominant phase with decreasing Mg# (0.947 to 0.871) and variable concentration in Al_2O_3 (1.7 to 3.1 wt.%). Cr_2O_3 is constant at 1.1 ± 0.3 wt.% and CaO decreases steadily from 25 to 20 wt.% with differentiation [Tab.S16].

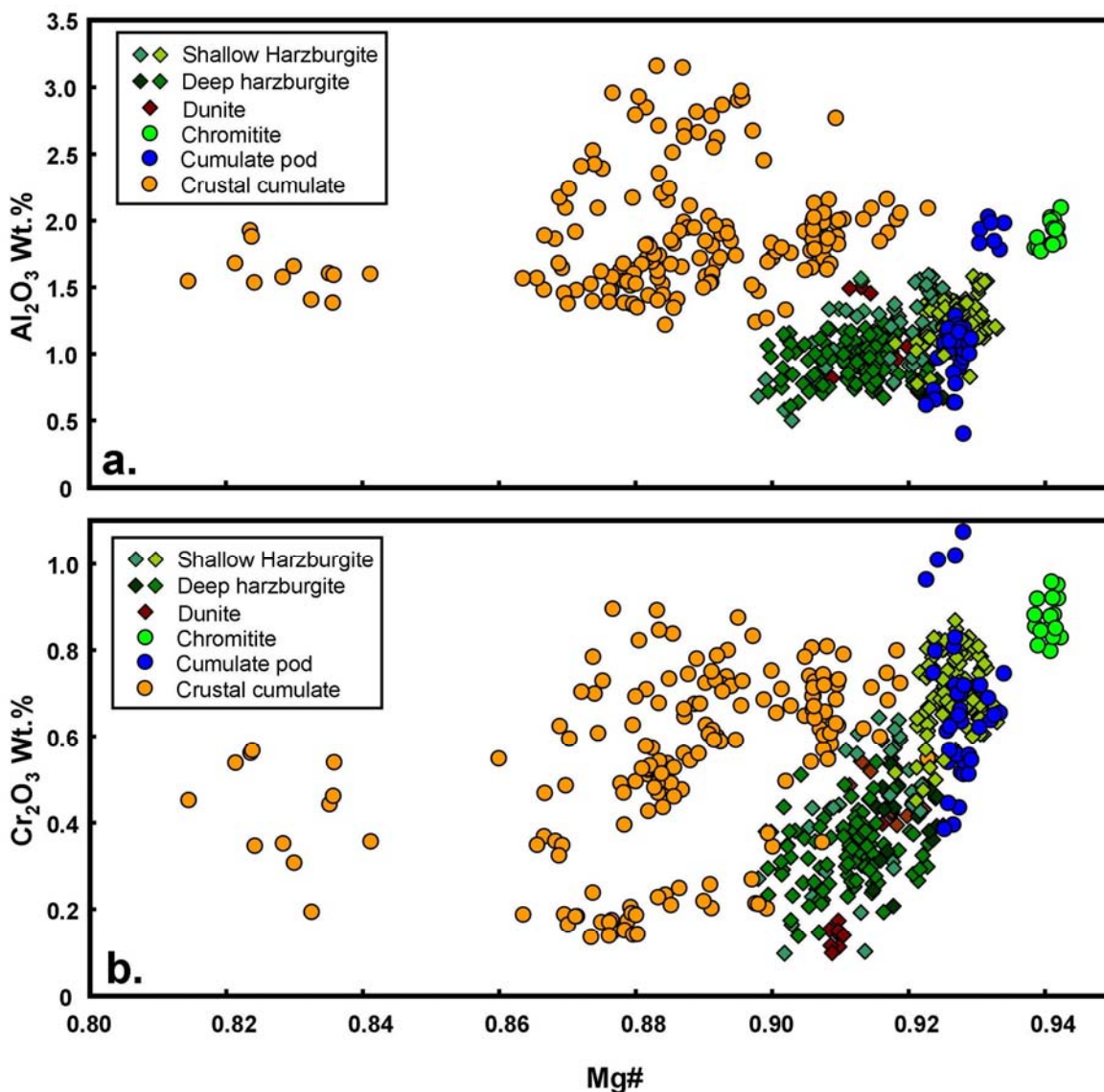


Fig.4.20. Orthopyroxenes major element contents.

a. Mg-number versus alumina shows higher content in alumina for chromitites and gabbros as well as one spinel-rich cumulate pod.

b. Mg-number versus chromium shows a strong increase in Cr_2O_3 with stratigraphy in the harzburgite. Metasomatised harzburgite are highly enriched in chromium. Early crustal cumulates are rich but the concentration drops for latter mafic cumulates.

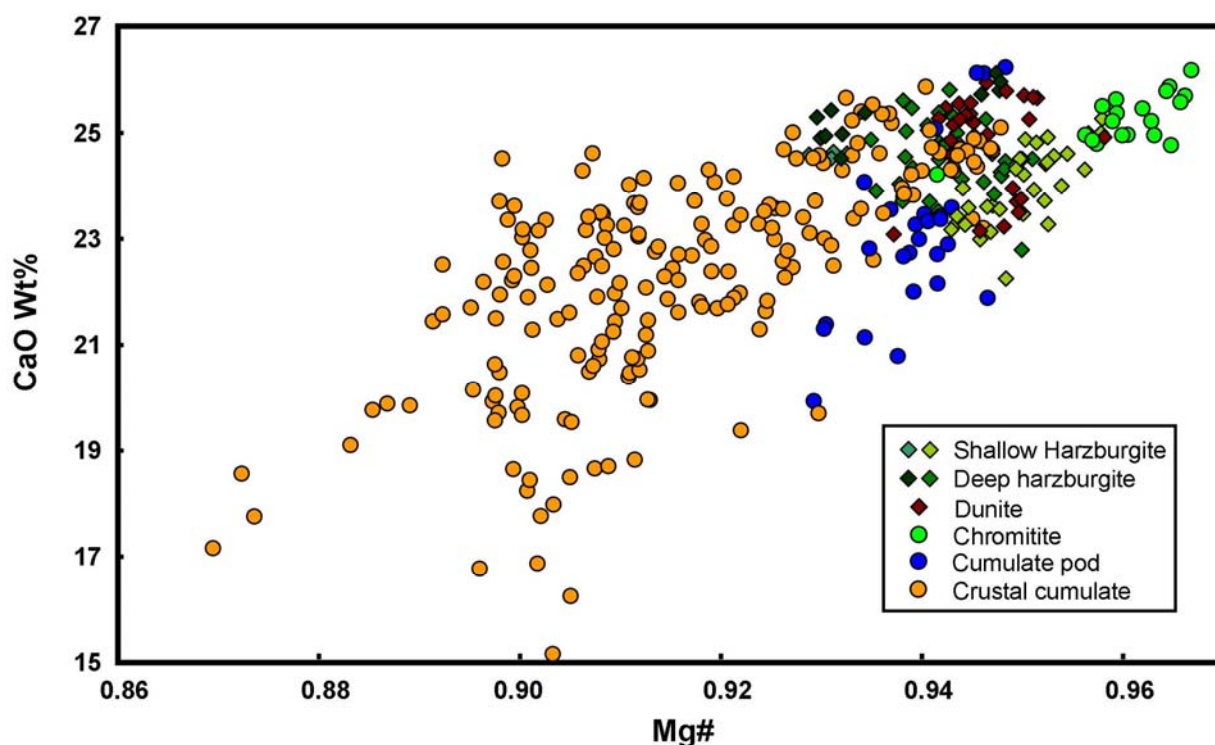


Fig.4.21. Mg-number versus calcium in clinopyroxene shows high content in CaO for clinopyroxene exsolutions in harzburgite. Gabbros (orange) steadily decrease in CaO with differentiation and the cumulative pods (blue) also show similar trends with a steeper gradient.

Clinopyroxene inclusions in chromitite pods have very high Mg-number (0.963(3)) and display high CaO and Cr_2O_3 concentrations. Interestingly, cumulate pods have low Mg# (0.929-0.948) relative to harzburgite and Al_2O_3 and Cr_2O_3 are higher than gabbro contents. CaO has quite a range from 25 to 20 wt.% in relation with differentiation and massive crystallisation of Ca-rich pyroxenes [Tab.S16].

4.3.4.4. Spinel

The chemical composition of spinels is difficult to characterise as variation can occur from grain to grain within a single sample. Such observations on the scattering of spinels composition have been mentioned for Bay of Islands and Oman ophiolites (Bédard & Hébert, 1998; Arai, 2006). These variations are related to morphology (blebs, symplectite), serpentinisation (magnetite and iron-rich spinels) and natural variations within grains. Harzburgite spinels are magnesiochromite with $\text{Mg\#} = 0.5 \pm 0.1$ and a high $\text{Cr\#}$ ($=\text{Cr}/(\text{Cr}+\text{Al}+\text{Fe}^{3+})$) of 0.68 ± 0.02 [Fig.4.22]. TiO_2 content is slightly

higher in deep harzburgite spinels (>0.08 wt.% TiO_2) than in olivine-rich harzburgite (<0.05 wt.% TiO_2) [Fig.4.29, Tab.S18].

In the lower dunite zone, the Mg# of chromite is unmodified but Cr# is slightly higher (0.72 ± 0.02 wt.%) [Fig.4.22]. Higher in the stratigraphy, dunites have higher Mg# (~ 0.6) and then progressively drop in the gabbroic rocks to values of ~ 0.4 . Cr# and Cr_2O_3 content are lower than the mantle rocks (Cr# ~ 0.55) and relatively constant throughout the upper dunites and the transition zone [Tab.S18].

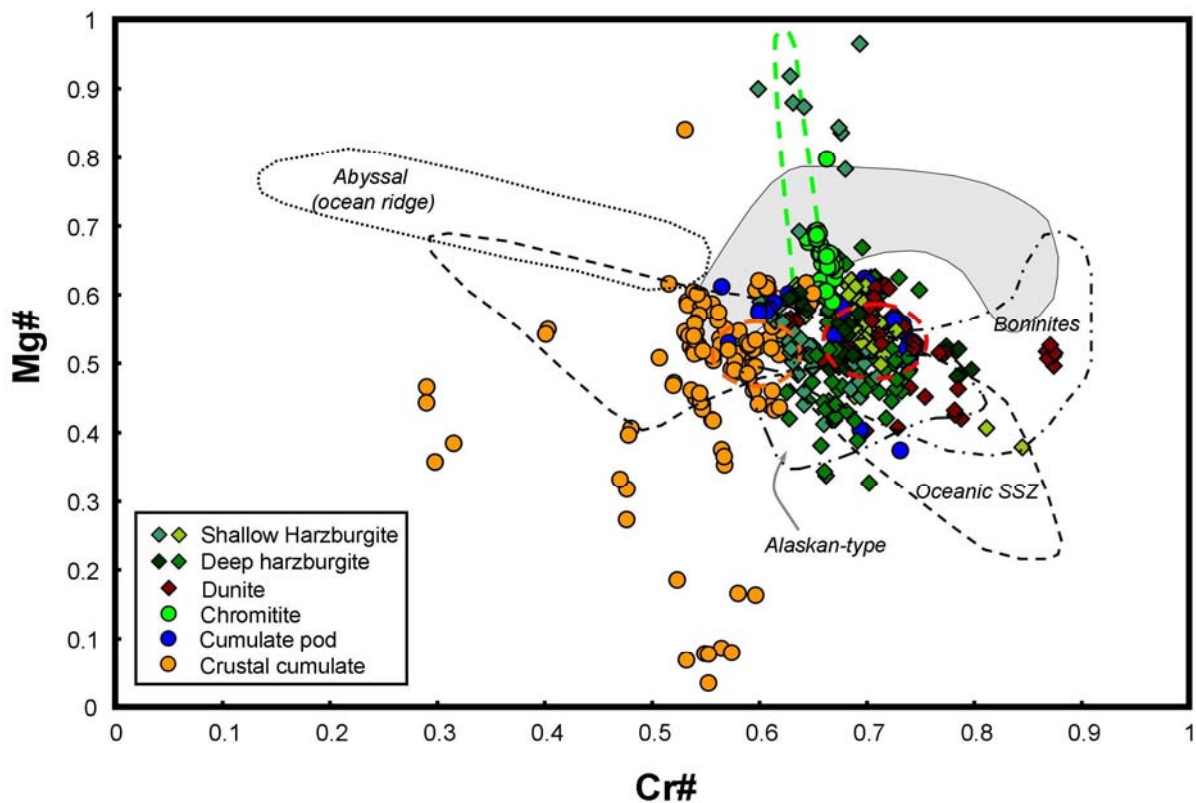


Fig.4.22. Ellipses are previous data on chromite in harzburgite (red), dunite (orange) and chromitite (green) of the Massif du Sud (Rodgers, 1973, 1977a, 1977b, Leach & Rodgers, 1977). The grey field covers all chromites of chromitite bodies from the northernmost part of New Caledonia (High Cr#) to the Massif du Sud (low Cr#) (Cassard et al., 1981). The dashed field is SSZ peridotites composition, dotted field abyssal peridotites and dash-double dot are Alaskan-type ultramafic rocks (Dick & Bullen, 1984). The dash-dot field are boninites (Metcalf & Shervais, 2008). Here, $\text{Cr\#} = (\text{Cr} + \text{V}) / (\text{Cr} + \text{V} + \text{Al} + \text{Fe}^{3+})$

Chromitite bodies and related rocks are characterised by a high Mg# ($\sim 0.65 \pm 0.05$) and Cr_2O_3 around 54 wt.% for chromitites and 52 wt.% for metasomatic harzburgites. Spinels in the latter are also characterised by a high TiO_2 (0.3 ± 0.05 wt.%) whereas chromitites have normal mantle values (0.05-0.10) [Fig.4.29]. Vanadium contents also show some strong variations

for discordant rocks compared to host rocks. In the ophiolite section, V_2O_3 stands around 0.20 ± 0.08 wt.% for all harzburgites, crustal cumulates and chromitites. The only exception concerns the harzburgite surrounding chromitite pods and some cumulate pods, which show very high coulsonite component with a solid solution including 3 to 7 wt.% V_2O_3 in chromite [Tab.S18].

4.3.4.5. Plagioclase

In the main ophiolite sequence, plagioclase only occurs in the cumulate pile topping the dunite zone in rocks which already contains clinopyroxene as an interstitial phase. The composition of the plagioclase is calcium-rich and ranges from An_{74-85} in some ultramafic cumulates to An_{87-98} in other cumulates and crustal gabbro-norites [Fig.6.14, Tab.S19].

4.3.4.6. Other phases

Phyllosilicates

The presence of sheet silicates in the ultramafic terrane is a sign of alteration by hydrothermal or meteoric fluids. Serpentine is the dominant hydrous phase in mafic and ultramafic rocks. Although some rocks are completely serpentinised by lizardite in fractured areas and in exposed outcrops, antigorite is the main alteration phase in the studied samples and generally does not exceed 30% replacement in most peridotites. Alteration mainly affects olivine and is strongly dependent on the texture of the rock. Dunites are generally fresher than harzburgite whereas chromitites are so altered that many samples are devoid of olivine. However, pyroxenes and spinels remain unaffected by the dissolution of olivine up to high levels of alteration. The composition of antigorite and lizardite are similar and both have Mg-numbers higher than the parental olivine. The Mg-Fe balance is maintained during serpentinisation, by the formation of magnetite trails in the centre of serpentine veins (Rodgers, 1974a; Barnes et al., 1978).

Orthopyroxene-rich rocks are occasionally altered and show veins or a coating of bastite (serpentine). Talc and magnetite are occasionally seen during this alteration process. At the contact between felsic rocks and peridotite, the introduction of large amount of aluminium in hydration reactions can form zones rich in clinochlore. Finally, the presence of Ni-rich olivine as well as abundant nickel sulfides promotes the formation of a twin nickelian serpentinisation. Nepouite, “pimelite”, willemseite Ni-nontronite and eventually pecoraite have been identified in some samples as a group of Ni-phyllosilicates which are commonly gathered under the name garnierite (Wells et al., 2009).

In felsic rocks, kaolinite and pyrophyllite are frequent alteration products whereas micas (biotites) are a primary phase. Finally, beside the abundance of sheet silicates, brucite and calcite can also be precipitated in the rocks of the ultramafic terrane during hydrothermal events and meteoric and pedologic alteration (Rodgers, 1974a; Barnes et al., 1978).

Amphiboles

Amphibole is not a common phase as a primary mineral in the mantle rocks of New Caledonia. It occurs as minute exsolutions of pargasite in orthopyroxenes porphyroclasts of Dumbéa harzburgite as well as some areas south of Lake Yaté. Pargasitic amphiboles in the Massif du Sud mostly occur in dykes and later intrusions and are discussed in details in chapter 6. In all other cases, amphibole is a secondary phase formed by hydrothermal alteration. These amphiboles are recognisable by their fibrous textures and their compositions which range between anthophyllite and tremolite when a source of calcium is available. Massive pods of nephrite are also occasionally observed as an alteration product of the transition zone.

Garnets

In gabbroic rocks, the alteration of plagioclase often produces a mixture of sericite and hydrogrossular. In cases where gabbroic veins are crosscutting chromitite deposits, minute uvarovite crystals form at the contact but this is rare.

Zeolites s.l.

In layered cumulates, the alteration of calcic feldspar also produces large pockets of prehnite ($\text{Ca}_2\text{Al}_2\text{Si}_3\text{O}_{10}(\text{OH})_2$) and more rarely Ca-thompsonite ($\text{NaCa}_2\text{Al}_2\text{Si}_5\text{O}_{20} \cdot 5\text{H}_2\text{O}$) as a coating phase around altered plagioclase ($\text{An}_{<90}$).

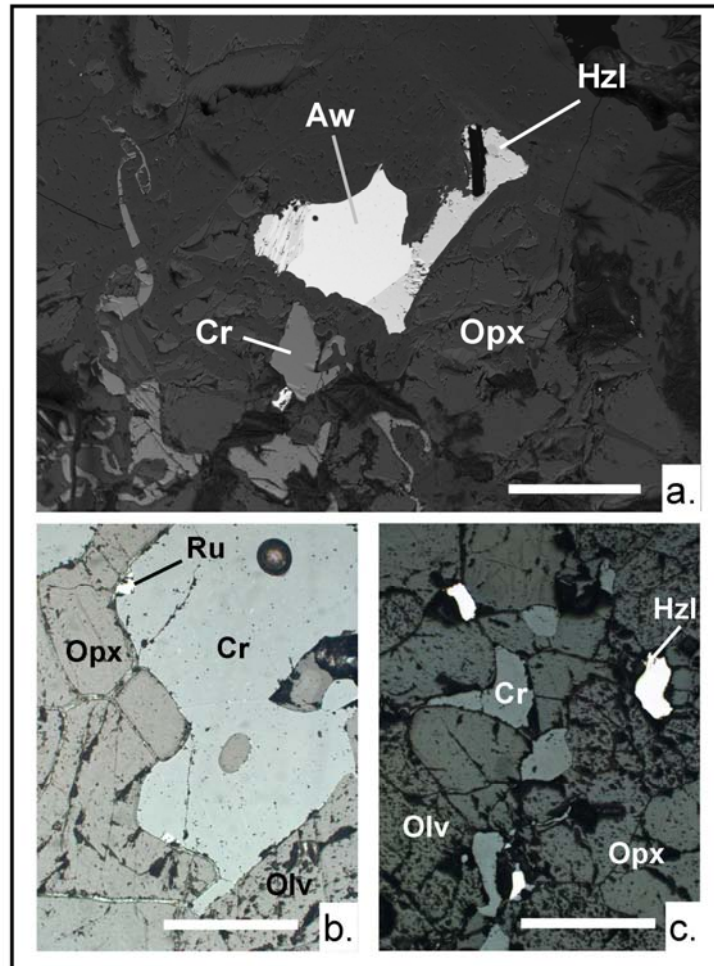


Fig.4.23. Accessory minerals in harzburgite and cumulate pods.

a. Backscattered electron mode picture of an assemblage of awaruite (white), heazlewoodite (light grey) and chromite (dark grey) in a matrix of olivine and serpentine (PRB1). Scale bar is 250 μm .

b. Reflected light picture of a vanadium-rich chromite (light grey) with rutile inclusions (white) in orthopyroxenes of RP42. Scale bar is 100 μm .

c. Reflected light picture of a vanadium-rich chromite associated with heazlewoodite in a matrix of olivine and orthopyroxene (RP42) Scale bar is 100 μm .

Sulfides, Alloys and Oxides

Sulfides are frequent in ultramafic samples and more likely occur in areas where dykes are abundant. Pentlandite ((Ni,Fe)₉S₈) is considered as a primary phase but is rarely observed in our samples. Crystals are rich in nickel (Ni/Fe = 1.2) and cobalt (2 wt.% CoO). However, in most cases, Fe-Ni aggregates occur as a fine intergrowth of heazlewoodite (Ni₃S₂) and awaruite (Ni₂₋₃Fe) [Fig.4.23]. Although these two phases can occur separately, associations in fine lamellae are common and suggest that these phases are a breakdown product of a primary pentlandite. In addition to these sulfides and native minerals, graphite is mentioned by Augé et al. (1999) in harzburgite as well as a long list of sulfides, oxides and alloys of platinoids associated with chromite and awaruite (Augé & Legendre, 1994; Augé & Maurizot, 1995; Augé et al., 1999).

In some cumulate pods, an exotic ultramafic mineralogy is observed. Orthopyroxene-olivine cumulates, in addition of hosting V-rich chromite and an important amount of heazlewoodite + awaruite, are also saturated in apatite which is present as minute grains. Finally, rutile is also observed as an exsolution or an overgrowth of chromite and is sometimes associated with rare laths of ilmenite in these cumulate pods [Fig.4.23].

4.3.5. Trace element mineral chemistry

4.3.5.1. Olivine

One of the main reasons for the lack of data available for the New Caledonia ophiolite is the high depletion observed in the rocks. This problem is partly avoided by studying bulk rock composition, which are sensitive to secondary alteration or concentrating on clinopyroxenes which is of course absent as a primary phase in the harzburgite (Prinzhofer & Allègre, 1985; Marchesi et al., 2009). In this study, selected trace elements were determined in olivine, orthopyroxene and spinels, the primary phase of the peridotite [Tab.S13].

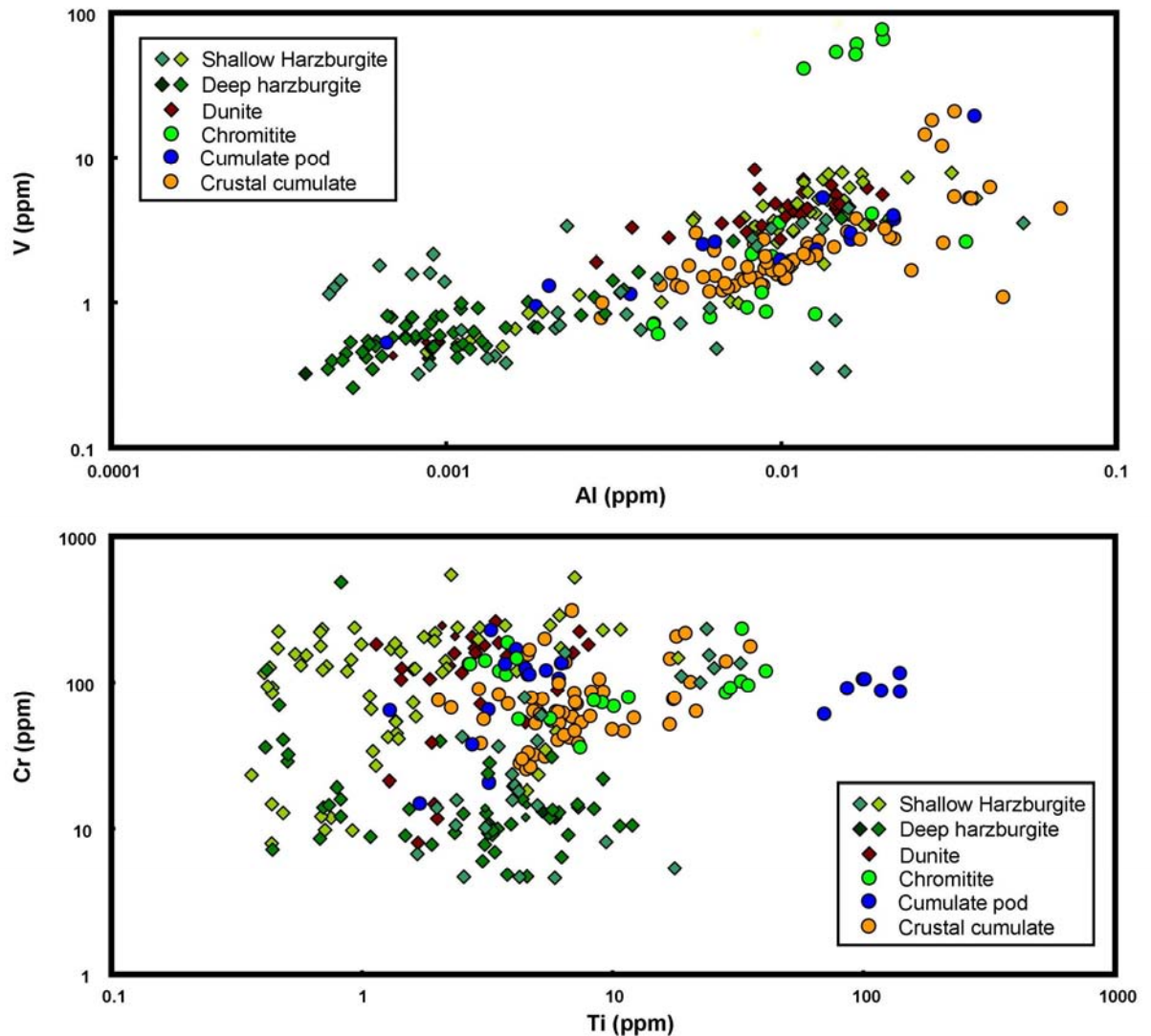


Fig.4.24. Trace element data on olivine.

- a. Aluminium versus vanadium shows a correlation with stratigraphy, increasing in both components.
- b. Chromium versus titanium shows an increasing amount of Cr in correlation with stratigraphy in the harzburgite. No such trend is visible for TiO_2 content but crustal cumulates are higher in titanium.

Olivine in harzburgite show increasing content in Al, Cr and V that correlates with stratigraphy [Fig.4.24]. Values increase around 100x between the lower harzburgite and peridotites of the higher part of the mantle section. However, in this uppermost zone, the presence of dunite and pyroxenites veins seems to have some influence on the trace element concentration in the surrounding harzburgite which introduce some variation to the data. Other measurable trace elements such as Y, Ca, Ti, do not show any detectable variation in harzburgites [Tab.S13].

Dunites lie at the end of the harzburgite enrichment trend for Al, Cr and V. Higher calcium contents are also observed but are probably related to the lack of pyroxenes available for subsolidus re-equilibration, keeping Ca values

in olivine closer to the primary magmatic concentrations. Yttrium is only slightly enriched in dunitic rocks. In the transition zone, cumulative rocks have lower V than the dunite zone and this effect is also observed to a lesser extent for Cr, Al and Ca. Ti and Y are enriched compared to the mantle rocks. Mineral trace element data on the dunite zone and the transition zone for olivine and pyroxene is described in greater detail in the chapter 5 [Tab.S13].

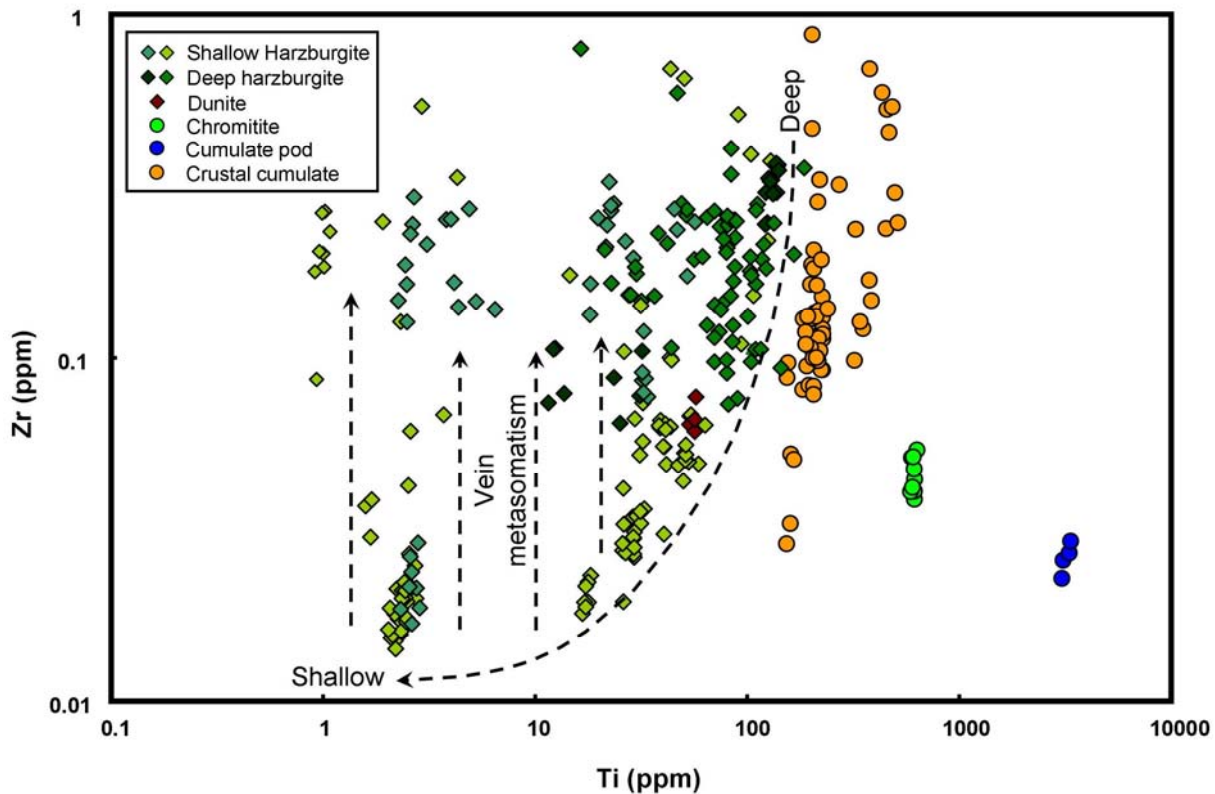


Fig.4.25. Zirconium versus titanium content in orthopyroxene shows an increasing trend from depleted shallow harzburgite to deep harzburgite. Values with low zirconium and relatively high titanium for shallow harzburgite are peridotites in the surrounding of vein metasomatism. Crustal gabbros are clearly enriched in titanium as well as chromitite and cumulate pods.

4.3.5.2. Orthopyroxene

Orthopyroxene porphyroclasts in the harzburgite have large variations within single samples due to local conditions and serpentinisation reactions which affect some trace elements (LILE, Pb, Sr). Therefore, these trace elements show no clear changes in concentration with stratigraphy. Titanium and zirconium show a decrease by 100x between harzburgite from the east coast and olivine-rich harzburgite in the uppermost mantle [Fig.4.25]. However, it has been observed that harzburgite collected in the

neighbourhood of major melt pathways (dykes) do not have depletion in some elements and can occupy an intermediary position between harzburgite and gabbros for elements such as Ti or V. Chromium in orthopyroxene show a slight increase from 2500 ppm in deep harzburgite to 5000 ppm in harzburgite pods in the dunite zone [Tab.S15].

Crustal cumulates have orthopyroxenes which have high content in titanium and vanadium but other trace elements remain mostly unchanged in comparison to mantle rock values. Orthopyroxene in discordant pods of chromitite and other mantle cumulates have very high enrichment in Ti and V and are particularly low in Sr and Y [Tab.S15].

4.3.5.3. Other minerals

Clinopyroxene

Clinopyroxenes analysed in peridotite only represent exsolutions of orthopyroxene and have trace element pattern reflecting this low temperature process. Ca-rich pyroxenes provide interesting data in the cumulate series where they occur as a truly magmatic phase [Fig.6.17]. Patterns are similar to Marchesi et al. (2009) and show no europium anomalies in upward REE patterns ($\text{La/Lu}^N = 0.02$ to 0.3). Strontium and titanium have a similar behaviour and do not show anomalies in cumulative rocks. Clinopyroxene occurring in gabbros and in mafic intrusions are covered in more detail in the following chapters [Tab.S17].

Plagioclase

Plagioclase in crustal cumulates have a flat REE pattern ($\text{La/Lu}^N = 0.4$ to 1.08) with high content in Eu as well as for Sr and Be. Concentrations are relatively high in V and Mn. The trace element chemistry of plagioclase is developed in more detail when compared to felsic rocks in chapter 6 [Tab.S20].

Spinel

Spinel provide data analogous to major elements and have strong variations within samples. Most information provided by trace elements does not give additional details than the major chemical composition. Trace element contents are characterised by high contents of transition metals, High HFSE are related to interferences and LILE and REE concentrations are below quantification limits.

4.4. Discussion

4.4.1. Stratigraphic scale and ophiolitic sequence

Previous studies based on the geological mapping of the area (Noesmoen, 1971; Guillon et al., 1974; Guillon & Trescases, 1976; Trescases & Guillon, 1977; Prinzhofer et al., 1980; Augé & Maurizot, 1995) have given an overview of the stratigraphic sequence in New Caledonia [Fig.4.4]. The general aspect of the ophiolite is similar to other ultramafic-mafic sections studied in Oman, Cyprus, Philippines, Newfoundland or Hokkaido (Violette, 1980; Smewing, 1981; Pallister & Hopson, 1981; Girardeau & Nicolas, 1981; Takashima et al., 2002). It consists of an ultramafic root composed of harzburgite, a highly depleted zone of dunite in the upper part of the mantle section, a transition zone with various ultramafic rocks and layered gabbros in the lower part of the oceanic crust [Fig.4.6].

4.4.1.1. The ultramafic sequence

The Massif du Sud is extremely depleted. No lherzolite has been reported in the ultramafic part and primary clinopyroxene is almost nonexistent below the palaeo-Moho. Other harzburgitic ophiolites which show thicknesses similar to the Massif du Sud are known to have a lherzolitic base in the deepest part of the section. Such stratigraphy is known in Oman (Smewing, 1981) and sometimes interpreted as a frozen melt - mantle interaction (Godard et al., 2000). In the Massif du Sud, no evident trend towards lherzolitic

composition is seen. The lherzolitic northern massifs of Tiébaghi and Poum, which have been sometimes considered as deeper part of the ultramafic nappe are known to be the result of refertilisation of a depleted mantle rather than low amount of partial melting (Ulrich et al., 2010).

The harzburgite is a tectonite with metamorphic porphyroclastic textures. At depth, the peridotite is relatively homogeneous but parallel alternation of harzburgite and dunite as a result of deformation with subsolidus recrystallisation is observed in the higher zones (Nicolas, 1989). These features are common for many ophiolites as the peridotite lies nearer the crust-mantle transition (Boudier & Coleman, 1981; Lippard et al., 1986). The high temperature foliation and mineral lineation is mostly parallel to the palaeo-Moho and the thickness of the peridotite zone is estimated between 2000 m (Dupuy et al., 1981) and 3500 m (Guillon, 1975).

Although general petrological observations have been made on the peridotite of New Caledonia, geochemical data on the stratigraphy of the ophiolite has not been reported or had no constraints on palaeo-depth and led to genetic misinterpretation of the data (Dupuy et al., 1981a; Prinzhofer & Allègre, 1985). The chemical composition of olivine and orthopyroxene in the harzburgite confirms the general stratigraphy of the area with relatively steady variations from the east coast towards the gabbroic units [Fig.4.16, 4.17, 4.19]. This evolution is the result of an increasing partial melting in the mantle rocks, leaving residues with more refractory composition in the uppermost mantle. A study from Dupuy et al. (1981b) on the distribution of Cu and Au in peridotite also shows the residual character up section as incompatible Cu and Au concentrations decrease. The geochemical trends observed in the rocks sampled in the Massif du Sud extend to the dunite zone but remain relatively constant within that zone until the palaeo-Moho is reached.

4.4.1.2. The dunite complex

The dunite zone in the Massif du Sud ophiolite is strongly developed and reaches a thickness of 600 ± 100 m (Prinzhofer et al., 1980). The transition from harzburgite to dunite is progressive with an increasing amount of dunite pods within the upper harzburgite. The change in mapping unit is placed in the area where dunite lithologies become dominant over harzburgite. This very diffuse transition is observed in several ophiolites in Oman, Coto (Philippines) and Bay of Islands (Newfoundland) (Girardeau & Nicolas, 1981; Nicolas & Prinzhofer, 1983). The texture of the dunite zone is not fundamentally different from harzburgite but olivine crystals are larger due to the monomineralic nature of the rock (Hiraga et al., 2010). According to the fabric classification of Mercier & Nicolas (1975) dunites remain porphyroclastic with a syntectonic recrystallisation leading to equigranular textures. Plastic strain is strong in the lowest dunite and fades upward to disappear after a few hundred meters (Nicolas & Prinzhofer, 1983).

4.4.1.3. The crustal sequence

The sections along four transition zones were studied during fieldwork which provides a comparison between areas relatively close to each other. In the area surrounding Montagne des Sources, the crust-mantle transition is relatively clear and shows a prograde sequence from dunite to gabbro characterised by increasing modal component in clinopyroxene and plagioclase. Previous mapping studies suggested that the layered gabbros present in the Montagne des Sources are topping the entire ultramafic sequence and the transitional rocks [Fig.4.4] (Prinzhofer et al., 1980; Marchesi et al., 2009). However, detailed mapping of the crustal sequence has revealed that the layered gabbros are not a continuous and thick gabbroic unit but is comprised of several thick sills separated by wehrlitic zones [Fig.4.12]. Such features in the upper part of transition zones have been observed in other crustal sections such as Oman where a series of mafic sills are intruding ultramafic rocks (Smewing, 1981; Boudier et al., 1996). The presence of basalt at the top of Montagne des Sources has been wrongly interpreted by Dupuy et

al. (1981a) as a part of the ophiolitic sequence. Field observations show that such basaltic outcrops can occur in the dunite zone or in the harzburgite and are likely to be related to post-obductive magmatism.

In the Rivière des Pirogues transition zone, the mafic cumulates are reached from the North and the South and show a more acute transition zone than in Montagne des Sources [Fig.4.11] and shares similarities with the transition zone of Nurali ophiolite in the Urals (Pertsev et al., 1997). However, a microscopic study of the rocks just below the palaeo-Moho shows that olivine is cumulus and that small amounts of interstitial clinopyroxenes are present. Despite this observation, the mafic part of the crust-mantle transition in Rivière des Pirogues is more complex and several layers of contrasted cumulates are observed in the area.

In Baie des Pirogues, the transition zone is more pyroxenitic and mafic intrusions are only observed in the area of Mt Ngo. Clinopyroxenite and websterite dominate the area by forming large sills or dykes in an olivine-rich rock.

In all four sections, the transition zone shares similarities. Dunite is the host rock and there is a gradient in modal olivine towards mafic sills or lenses where clinopyroxenes and minor plagioclase are of increasing abundance. In locations closer to the gabbros, poikilitic dunite becomes wehrlitic or even plagioclase clinopyroxenite [Fig.4.9]. When olivine becomes a minor component, the rocks can generally be classified as gabbros [Fig.4.13]. This feature is observed in the footwall and the hanging wall of gabbro layers which give additional evidence that gabbros are emplaced as intrusive sills with local impregnation of the surrounding rocks and that cpx-dunite—wehrlite—gabbro zonation is not the product of large cumulative processes. Such progressive and sequential mantle-crust sections have been described in ophiolites of the Arabian peninsula (Benn et al., 1988; Boudier & Nicolas, 1995).

4.4.2. Origin of variations within single members of the ophiolite suite

The geological maps of the B.R.G.M. (Guillon et al., 1974; Guillon & Trescases, 1976; Trescases & Guillon, 1977) are based on the major lithological differences which occur between rocks. Most of the boundaries and the names attributed to the different units, although useful for field observations, are based on arbitrary changes which do not reflect the official nomenclature of plutonic rocks (Streckeisen, 1972). In addition to a simplified representation of the crustal and transitional rocks, the initial mapping presents homogeneous ultramafic formations which lack of any stratigraphic features for levels below the palaeo-Moho. A careful and sometimes microscopic observation of samples shows that variations exists in single units and can reflect different magmatic processes.

4.4.2.1. Ultramafic zone

The stratigraphic variation of Mg# and several other major elements in silicate phases provide a rough estimation of the position of a sample with respect to the palaeo-Moho [Fig.4.16, 4.17]. However, at palaeo-depth of more than 500m below the dunite zone, local variations affect the chemistry of single peridotite samples and although average values for an area are still usable for stratigraphic purposes, single sample analyses can be misleading. Several other observations such as modal abundances, chemistry and the textures of rocks are potentially useful as indicator of stratigraphy.

In the lowest part of the ophiolite sequence, the only variation which is noticeable for major phases on an outcrop scale is the amount of chromite present in the harzburgite, which typically increase with stratigraphical level. Accessory phases such as sulfides and alloys (mainly pentlandite, heazlewoodite, awaruite, millerite, copper, chalcocite and many other phases) are particularly abundant in the bottom 1000 m of the ultramafic terrane (Guillon & Lawrence, 1973). Our field study has shown that sulfides also occur in higher levels but enrichment in Ni-Fe minerals is often restricted to heterogeneities in the mantle such as metasomatic harzburgite of the chromitite trend and cumulative pods [Fig.4.23]. The presence of these

minerals is not as widespread as in lower harzburgites and concentrations never reach values mentioned by Augé et al. (1999) who describes peridotites that contain up to 10% of sulfides.

Although the entire harzburgitic mantle show porphyroclastic textures, a change of the shear direction based on the orientation of orthopyroxene seem to indicate a line of approximate palaeo-depth in the ophiolite (Prinzhofer et al., 1980). In higher levels of the peridotite unit, plastic strain increases and reaches a maximum in the uppermost part of the harzburgite and the dunite zone [Fig.4.6].

The mineral chemistry of the spinel shows that this aluminous phase is not a very good proxy of palaeo-depth in the mantle. However, the morphology of the spinels *s.l.* is highly dependant on the position and the composition of the rock in the mantle column. In most of the harzburgite, but particularly in the lower part, large symplectites of chromite and pyroxenes are observed [Fig.4.26]. Similar assemblages have been observed in high-pressure peridotites as kelyphites surrounding garnet porphyroclasts (Godard & Martin, 2000). The classic interpretation of these textures is a metamorphic reaction $Grt + Olv \rightarrow Opx + Sp + Cpx \pm Amph$, which involves the breakdown of garnet (Green & Burnley, 1988). The total breakdown of Cr-pyrope is known to produce kelyphites with pyroxene and chromian spinels (Wallace, 1975; Field & Haggerty, 1994). However, the aluminous phase analysed in New Caledonia ophiolite is a magnesiochromite and a reconstruction of the symplectite porphyroclasts on a garnet basis is incompatible with common knowledge about garnet peridotite. Several analyses of porphyroclasts in Yaté area were calculated on 60% opx + 10% cpx + 30% spinel and produce a garnet composition of $(Mg,Fe,Ca)_3Cr(Al,Fe)[SiO_4]_3$. Such garnet with 47% knorringite, 42% pyrope and 7% uvarovite are only comparable to some rare inclusions in kimberlitic xenoliths and garnet reconstruction out of similar assemblage in the hypothetical lunar mantle (Bell & Mao, 1975). Although symplectites are probably formed by the breakdown of garnet, the most likely explanation for this exotic composition is that spinel is affected by later fluids as it has been observed in the Pannonian basin (A.Toth, pers.comm.). Similar calculations based on the model of Mercier & Nicolas (1975) with a Al(Cr)-Ca orthopyroxene base provide similar exotic result with an extreme kosmochlor

component. Most of the other spinels in the porphyroclastic harzburgite occur as blebs or holly leaves shape crystals with a homogeneous distribution in the peridotite. Their composition is similar to symplectite spinels which would favour the fluid re-equilibration hypothesis.

In the uppermost part of the ultramafic section, the modal decrease in orthopyroxene is accompanied by the growth of spinels included in the pyroxene with small vermicular overgrowth [Fig.4.26]. This feature is explained by the increase in partial melting in the upper part of the peridotite, leading to the consumption of orthopyroxene. The melting reaction promoted the precipitation of spinel, which grew on the edge of rounded orthopyroxene crystals (Mercier & Nicolas, 1975; Leblanc et al., 1980; Niida, 1984; Boudier & Nicolas, 1995).

4.4.2.2. Dunite zone

Since the early days when ophiolites were described as remnants of oceanic crust in New Caledonia and other localities (Auboin et al., 1977), the nature of the dunite zone has remained enigmatic. These highly refractory rocks are interpreted as cumulates (Coleman, 1977, George, 1978; Pallister & Hopson, 1981; Elthon et al., 1982, 1984; Smewing et al., 1984; Lippard et al., 1986; Coogan et al., 2003; Clénet et al, 2010) or as residual rocks (Sinton, 1977; Nicolas et al, 1981, Nicolas & Prinzhofer, 1983, Benn et al., 1988; Boudier & Nicolas, 1995; Jousselin & Nicolas, 2000; Yumul, 2004; Zhou et al., 2005). Most geological mapping works on ophiolites have chosen to neglect the petrogenetic variations of the dunite zones and these units are mapped as homogeneous bodies (Prinzhofer et al., 1980) [Fig.4.4]. The boundary between harzburgite and dunite clearly shows that replacive processes are playing a major role in this transition. The clear interface between dunite and harzburgite at an outcrop scale is identical to the emplacement of dunite channels [Fig.4.8]. A residual origin for the lowest dunite is unlikely as the disappearance of orthopyroxene would occur more progressively without the observation of clear chemical and modal gradients (Suhr et al., 2003; Morgan et al., 2008).

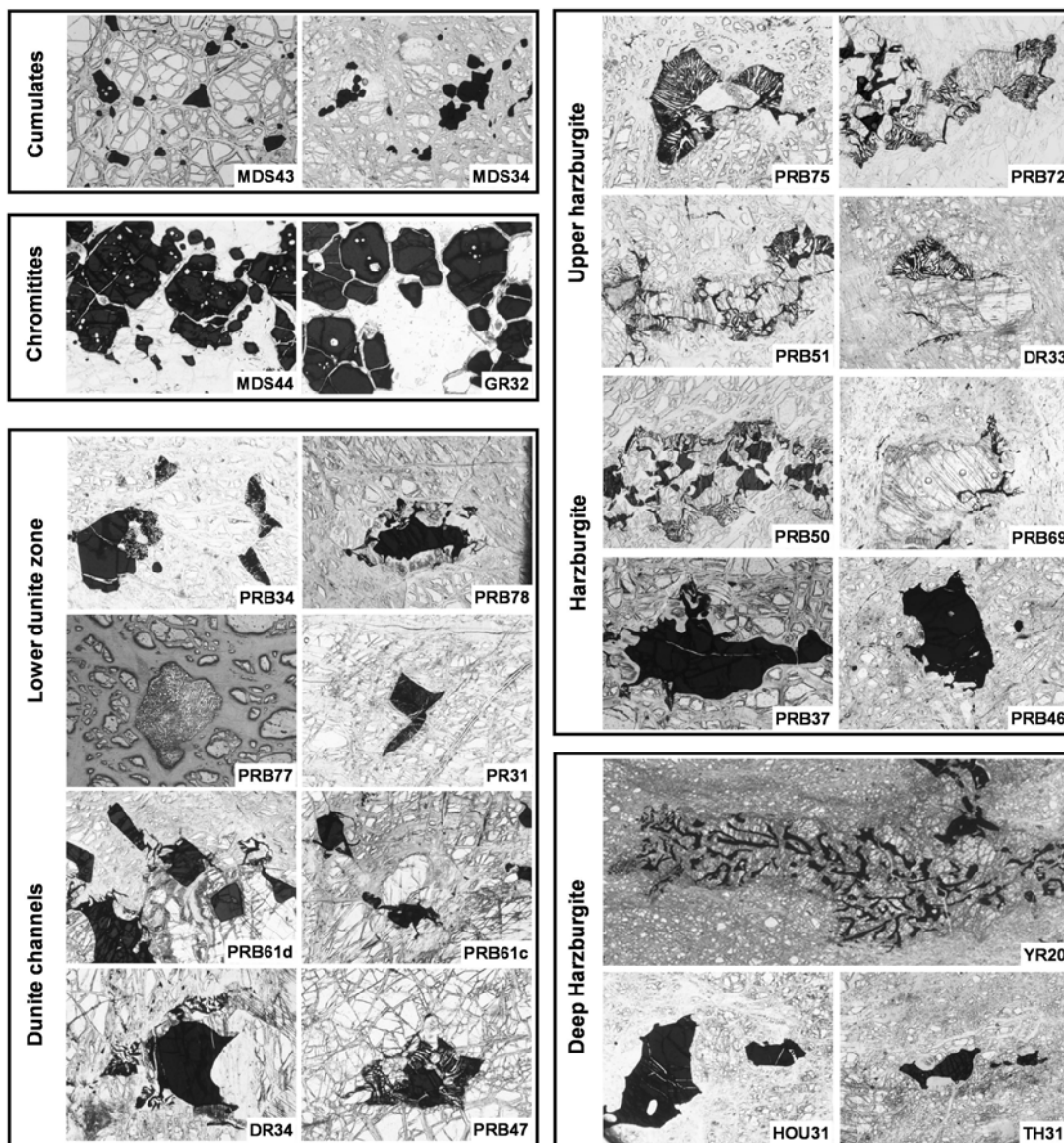


Fig.4.26. Spinels from different stratigraphic units.

Deep harzburgite are spinels from the east coast (Houailou, Thio, Yaté) showing holly leaves shape and Opx + Cpx + Sp symplectite.

Harzburgite are spinels in peridotite 1000 to 500 m below the dunite zone showing small amount of orthopyroxene dissolution and the associated precipitation of spinel.

Olivine harzburgite come from the harzburgite – dunite transition where orthopyroxene can be strongly dissolved, leading to large vermicular outgrowth on the rim.

Dunite channels are spinel observed in channel cross-cutting the harzburgite. Morphologies are similar to harzburgite with a clear outgrowth of aluminous phase.

Lower dunite zone share similarities with microvermicular textures present in channels and massive dunites.

Chromitites show rounded spinels with no overgrowth and silicate inclusions.

Cumulates are two dunite showing smaller rounded grains of spinels with silicate inclusions and no overgrowth. The length of pictures is 500 μm except for YR20 which is 1.2 cm long and PRB77 (reflected light) & PR31 which are 200 μm .

Although geochemistry is not necessary to identify the transition between harzburgite and replacive dunite, it has an appropriate use for the stratigraphy of the dunite zone itself. The composition of the olivine shows that a distinction can be made between residual, replacive (Kelemen, 1990) [Chap.B.1.2]) and cumulus olivine dunites. Residual processes should leave an olivine with more refractory composition than the underlying harzburgite whereas replacive process crystallise a homogeneous dunitic body with composition very similar to the harzburgite. The identical Mg# and slightly lower NiO in olivine and slightly higher Cr# in spinel support the replacive hypothesis for the entire lower dunite zone [Fig.4.18, 4.22]. In addition, it has been shown that residual dunite is unlikely as such rocks necessitate unreasonable amounts of partial melting (Nicolas & Prinzhofer, 1983; Pertzev et al., 1997).

The lack of differences in textures and chemistry between dunite channels and the dunite zone give additional support to the idea that the lower part of the dunite zone is formed by an orthopyroxene consuming reaction. This assumption is also proven by the careful observation of the aluminous phase in dunites [Fig.4.26]. It has been argued in New Caledonia and other ophiolites that the consumption of orthopyroxene during partial melting leaves behind an outer rim of spinel vermicules (Leblanc et al., 1980; Boudier & Nicolas, 1995). Similar observations can be made in some dunite channels and in the dunite zone where chromite forming the core of dissolving orthopyroxenes are left behind with a microsymplectitic rim of spinel and eventually interstitial clinopyroxenes. It seems that partial melting or replacive processes that consume orthopyroxene lead to texture for the spinels similar to the dunite channels.

In the upper part of Montagne des Sources and near the cumulate series of Rivière des Pirogues, dunite s.s. (98% olv $\pm$ 2% sp) have been sampled and shows compositions which are clearly different from the replacive dunite at the bottom of the unit. The Mg-number of olivine is 1 to 2% lower, NiO, V and Cr are identical or lower whereas MnO, Y and TiO₂ are higher [Fig.4.16, 4.18]. The aluminous phase has a Cr# which is 10% lower and V is also lower [Fig.4.22]. These observations suggest that these upper dunites are formed by fractional crystallisation and not by replacive processes. Spinel

with associated vermicular outgrowth which were common in the lower dunites are not present in the upper dunites. Chromites are rounded grains or idiomorphic and frequently contain inclusions of olivine which were completely absent in lower dunites.

Although the sampling in the area of Montagne des Sources is discontinuous within the dunite zone, it is expected that cumulus dunite starts to form around 200 m below the appearance of the first cumulus pyroxenes. In Rivière des Pirogues, dunite with replacive features has been sampled at -200 m and cumulus olivine 20 m below the first pyroxenite cumulates but similar to the harzburgite-dunite interface, this transition might be highly convoluted. This estimation on the thickness of the cumulus olivine is supported by structural observations. The plastic deformation, which reached a maximum in the bottom part of the dunite zone, is last observed 200 to 300 m below the crustal gabbros (Prinzhofer et al., 1980; Nicolas & Prinzhofer, 1983). The absence of plastic strain in the upper dunite zone would be consistent with the formation of olivine by magmatic fractionation. The combination of fabrics and geochemical data provide a solid discrimination on the origin of massive dunites in the Massif du Sud [Fig.4.4, 4.6]. It also shows that idiomorphic spinels, layered textures and monomineralic lithologies taken as a proof of cumulative process (Cassard et al., 1981; Dupuy et al., 1981; Nicolas & Prinzhofer, 1983; Augé et al., 1999; Cluzel et al., 2001; Picard et al., 2004) as well as impregnation, thickness and composition to support replacive processes (Nicolas & Prinzhofer, 1983; Marchesi et al., 2009) are not effective tools to infer the origin for dunites [Fig.4.26].

4.4.2.3. Crustal rocks

With the distinction made between cumulative (crust-related) and replacive (mantle-related) dunites as argued in the previous paragraph, this transition would not correspond to the geophysical Moho and represents a cryptic petrological discontinuity. The unit labelled as transition zone in geological maps [Fig.4.4] correspond to the appearance of orthopyroxene which form clear layered textures [Fig.4.12a]. It can also correspond to the interstitial precipitation of clinopyroxene which does not show accumulation

fabrics. The crystallisation of clinopyroxene as a cumulus phase appears slightly higher than orthopyroxene in the Rivière des Pirogues section [Fig.4.11]. Plagioclase has not been observed in this fractional crystallisation sequence and only occurs in gabbronorite sills or in the immediate surrounding as an interstitial phase [Fig.4.12, 4.13].

The morphology of the gabbronorite intrusions is not compatible with models involving large magma chambers (Smewing, 1981) and the sharp interface between sills and the surrounding cpx-plag-dunite is not entirely consistent with infiltration models [Fig.4.10]. In the latter, wehrlites are the passageway of basaltic melt (Benn et al., 1988) and gabbro sills are the result of the compaction of the wehrlitic mush leading to local accumulation of melts (Boudier & Nicolas, 1995; Godard et al., 2000; Marchesi et al., 2009). The presence of olivine with mafic composition [Fig.4.16] in the gabbronorite argues against the idea that sills are the result of infiltration where the host phase is entirely consumed (Benoit et al., 1996) as olivine seems in equilibrium with clinopyroxene and plagioclase in the wehrlite rims surrounding sills.

The emplacement of sills through injection events due to permeability barriers (Kelemen et al., 1996) or the episodic compaction of the olivine-melt mush present in the lower dunite zone (Rabinowicz et al., 1987; Clénet et al., 2010) could explain the morphologies of the sills and the rhythmic layering observed for some gabbros [Fig.4.12, 4.13d]. The presence of gabbroic dykes with identical composition [Fig.4.13d] [Chap.5] to the sills in the transition zone and within the gabbros [Fig.4.13c] suggests that the percolation of melt through the dunite pile forming wehrlite zoning is a consequence of sill intrusion and not a cause. The emplacement of gabbronorite as a melt in the crystal mush of the transition zone would induce a percolation of silicate liquids below and above the sills. This model where dykes would be episodically feeding gabbro sills is supported by observations in the Oman ophiolite where dykes and sills are branching into each other (Iddefonse et al., 1993).

4.4.3. General interpretation of the magmatic environment

4.4.3.1. Ultra-depleted mantle and SSZ environment

When the New Caledonia ophiolite was recognised as a fragment of oceanic crust and rather than a layered intrusion (Aubouin et al., 1977), an early interpretation claimed that the ultramafic nappe had the characteristics of a slow-spreading ridge producing MOR basalt (Dupuy et al., 1981). This model was supported by structural studies which suggested the presence of palaeo-transform faults with a N-S orientation (Prinzhofer & Nicolas, 1980). The measurement of foliation and lineation as well as the orientation of dykes in the sub-Moho region was taken to imply the presence of an E-W orientated ridge where the Massif du Sud would be a remnant of the northern flank (Prinzhofer et al., 1980). Petrologically, the idea was supported by Prinzhofer & Allègre (1985) where a complex disequilibrium melting of the mantle between garnet and plagioclase field was developed to explain the composition of the peridotites in New Caledonia.

However, as more ophiolites were studied in the 80's and an increasing amount of knowledge was gathered from deep sea drilling programs and volcanic arcs, New Caledonia did not seem to present all the characteristics of a mid-ocean ridge. Comparisons of geochemical data, comparison with supra-subduction zones and palaeogeography reconstructions indicated that the ultramafic terrane was formed in a convergent setting (Aitchison et al., 1995; Cluzel et al., 2001). By comparison to other ophiolites and due to the geodynamical setting of the Loyalty arc, the Massif du Sud became classically considered as an ultra-depleted fore-arc mantle (Eissen et al., 1998; Crawford et al., 2003; Whattam et al., 2008; Whattam, 2009) whereas the northern lherzolitic massifs represent a refertilised mantle wedge (Ulrich et al., 2010).

melting. Batch melting alone cannot reach such values and would not produce U-shaped REE pattern (Yamasaki et al., 2006; Aldanmaz et al., 2009). A process similar to other highly depleted peridotites is hypothesised where an initial stage would have produced basalts similar to the Poya terrane with <20% of partial melting (Cluzel et al., 2001) followed by fore-arc hydrous magmatism which has led to refertilisation and the generation of boninitic rocks (Ulrich et al., 2010).

The position of peridotites in the olivine-spinel mantle array shows that the entire ultramafic section of New Caledonia carries supra-subduction zone signature [Fig.4.28] and an eventual boninitic trend [Fig.4.22]. The harzburgites and dunites display values which differ from abyssal peridotites and would suggest that the final partial melting event has occurred between 5 and 10 kbar for 25 to 30% of partial melting (Arai, 1994; Sobolev & Batanova, 1995), whereas in other ophiolites related to ridges the peridotites plot in the abyssal field (Aldanmaz et al., 2009). The chemistry of spinels also shows that high Cr# and extremely low TiO₂ content is in agreement with a sub-arc environment (Arai, 2006). Replacive dunites have spinel composition slightly higher than the surrounding rocks which make them more likely to be the source of boninitic magmatism [Fig.4.29, Chap.5] (Suhr et al., 2003). TiO₂ in clinopyroxene is very low but inconclusive to determine if the fractionation trend is boninitic or tholeiitic [Fig.4.30].

A supra-subduction zone is strongly characterised by high water content in the magmatic system (Parkinson & Arculus, 1999). As mentioned before, the early appearance of pyroxenes in cumulus dunite and the extensive delay in the saturation of plagioclase support a water-rich magmatic system. Recent field observations in the Oman ophiolite and experimental petrology on wehrlite pods (Koepke et al., 2009) have confirmed that conditions near water saturation are necessary to produce such rocks (Feig et al., 2006). In addition, the trend observed in clinopyroxene-bearing rocks in the New Caledonia crustal section is analogous to experimental studies on primitive arc melts (Müntener et al., 2001) and the lower part of the layered complex of Chilas in the Kohistan arc (Jagoutz et al., 2007). Alumina present in clinopyroxene suggests that early cumulates are crystallised in hydrous conditions where saturation of plagioclase is not reached [Fig.4.31].

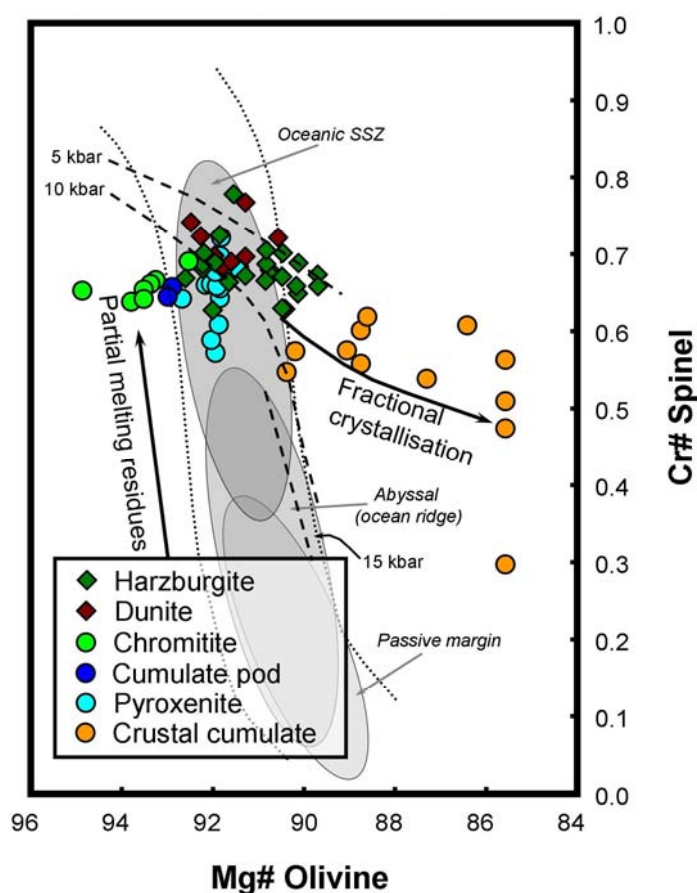


Fig.4.28. Average Mg# of olivine and Cr# of spinel in a variety of rocks from the Massif du Sud. The olivine-spinel mantle array is given in dotted line (OSMA) with supra-subduction field (top grey ellipse), abyssal peridotites (middle grey ellipse) and passive margin peridotites (bottom light grey) (Dick & Bullen, 1984; Arai, 1994; Pearce et al., 2000) and boninites (Metcalf & Shervais, 2008). The arrow underline the fractional crystallisation present in the crustal cumulate from cumulate dunite (high Mg#) to gabbronorite (low Mg#). Pressure curves give approximate values on the depth of melting of the peridotites. 5 and 10 kbar curves are from Sobolev & Batanova (1995) and 15 kbar from Jaques & Green (1980). The distinction between the replacive dunites (brown diamond) and cumulative dunites (high Mg# orange circles) is clearly visible and confirms notably that dunites from Marchesi et al (2009) are cumulative.

In the gabbros, the presence of water is supported by the high anorthite content of plagioclase. The liquidus of the albite - anorthite – H₂O system is considerably depressed near the calcium-rich pole which allows the crystallisation of high anorthite plagioclase at relatively low temperatures. The orthopyroxene content in the gabbros is also a specific feature which suggests that gabbronorites were initially a hydrous basaltic melt (Juster et al., 1989; Boudier et al., 2000). Similar gabbronorites are described in other ophiolites, often containing magmatic amphiboles confirming the hydrous nature in these oceanic crusts. In New Caledonia, the absence of primary hydrous minerals in the mafic rocks is linked to the crystallisation temperature of these gabbros.

Two pyroxenes geothermometers applied on these gabbronorites (Brey & Kohler, 1990) have provided minimum crystallisation temperatures in the range 1070-1180°C which is too high to crystallise amphibole in the lower part of the crust (Holloway & Burnham, 1972).

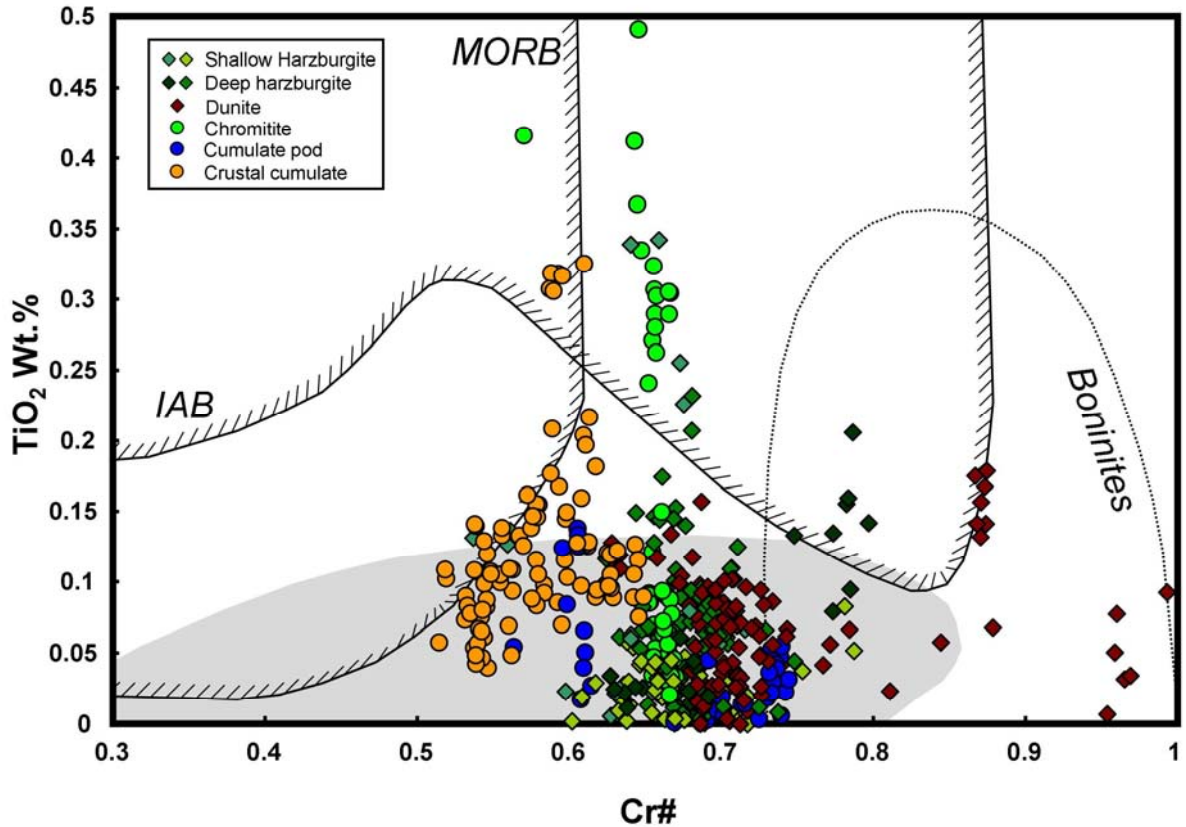


Fig.4.29. TiO_2 content versus Cr# in spinel of a variety of rocks in the Massif du Sud. Spinel compositions in MORB, island arc basalts (IAB) and boninites are from (Arai (1992), Kelemen et al. (1995) and Dick & Natland (1996)). SSZ peridotites from Oman are in the grey field (Arai, 2006). Cr# for replacive dunites (brown) and cumulate dunite (orange) are clearly distinctive. Abyssal harzburgite have spinels with $\text{Cr\#} < 0.6$ (LeMée et al., 2004).

Finally, the mantle wedge present over a subducted slab is known to have relatively oxidised peridotites compared to the rest of the depleted mantle (Brandon & Drapper, 1996; Blatter & Carmichael, 1998; Parkinson & Arculus, 1999). In the Massif du Sud, this feature is suggested by the presence of gabbronorites (Grove & Baker, 1984) and confirmed by the calculation of oxygen fugacity based on olivine and spinel (Ballhaus et al., 1991). Most harzburgites and dunites lie in the range of FMQ-0 to FMQ+4 which is in agreement with sub-arc peridotites collected in Japan, New Ireland and Kamtchaka (Arai & Ishimaru, 2008) and clearly different to the $f\text{O}_2$ of MORB (FMQ-2.5 to FMQ-0; Aldanmaz et al., 2009) [Fig.4.32]. High oxygen

fugacity values (FMQ+4 to +5) seem to be related to pre-obduction dykes which are discussed in chapter 6. Increasing fO_2 in relation with the presence of veins has been observed in Avacha xenoliths (Kamchatka) and is thought to be related to the percolation of hydrous fluids (Arai et al., 2003).

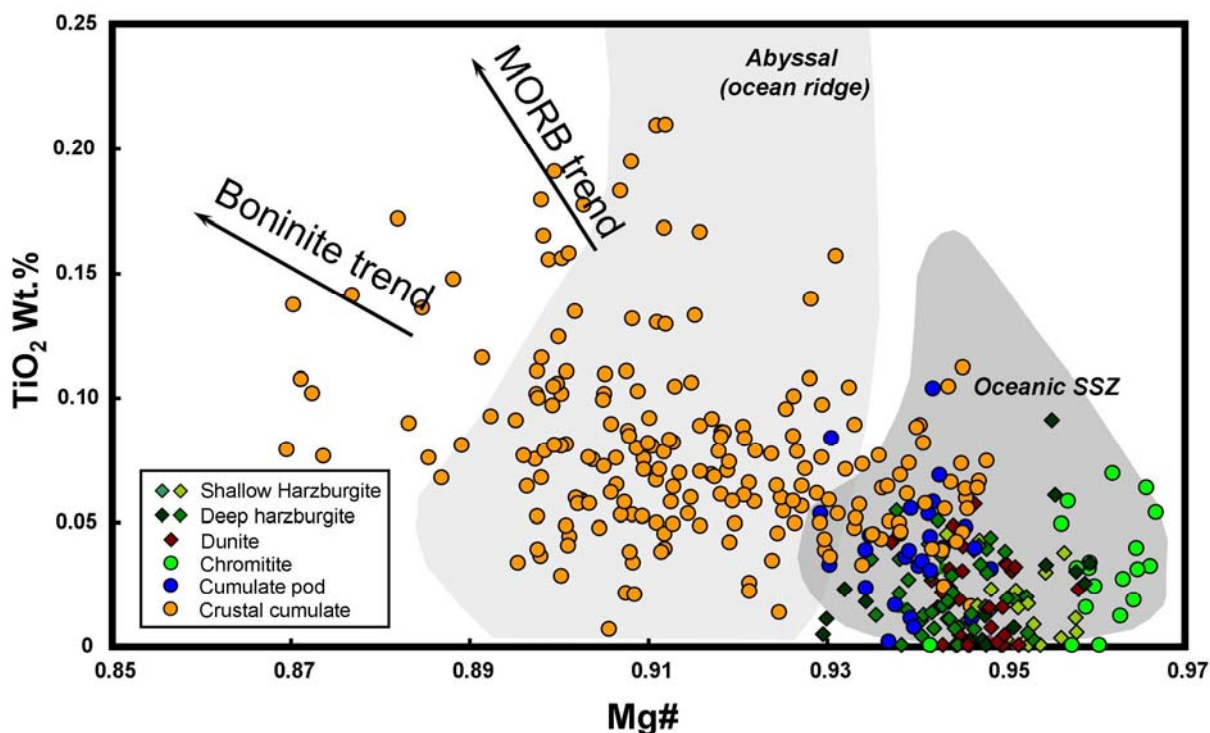


Fig.4.30. TiO_2 in clinopyroxene as a function of Mg# from the Massif du Sud. The gabbroic trend is compared to the general trend of MORB melts and boninitic melts. Trends are extracted from Benoit et al. (1999) and Cl  net et al. (2010). Fields are for supra-subduction zone peridotites (Ishii et al., 1992) and mid-ocean ridge peridotites (Johnson et al., 1990).

The study of sulfides, despite bearing magmatic textures, suggests that they are considerably affected by fluid percolation. The assemblage heazlewoodite – awaruite – pentlandite is classically interpreted as the result of serpentinisation reaction (Krishnarao, 1964; Sinton 1976; Lorand, 1989) but these sulfides aggregates have also been considered as primary magmatic phase in the Lanzo Massif (Lorand et al., 1993). From our observations, it is possible that a Ni-rich pentlandite was the initial phase that broke down into alloys and sulfides during a reducing hydrothermal event which also led to the crystallisation of graphite (Aug   et al, 1999).

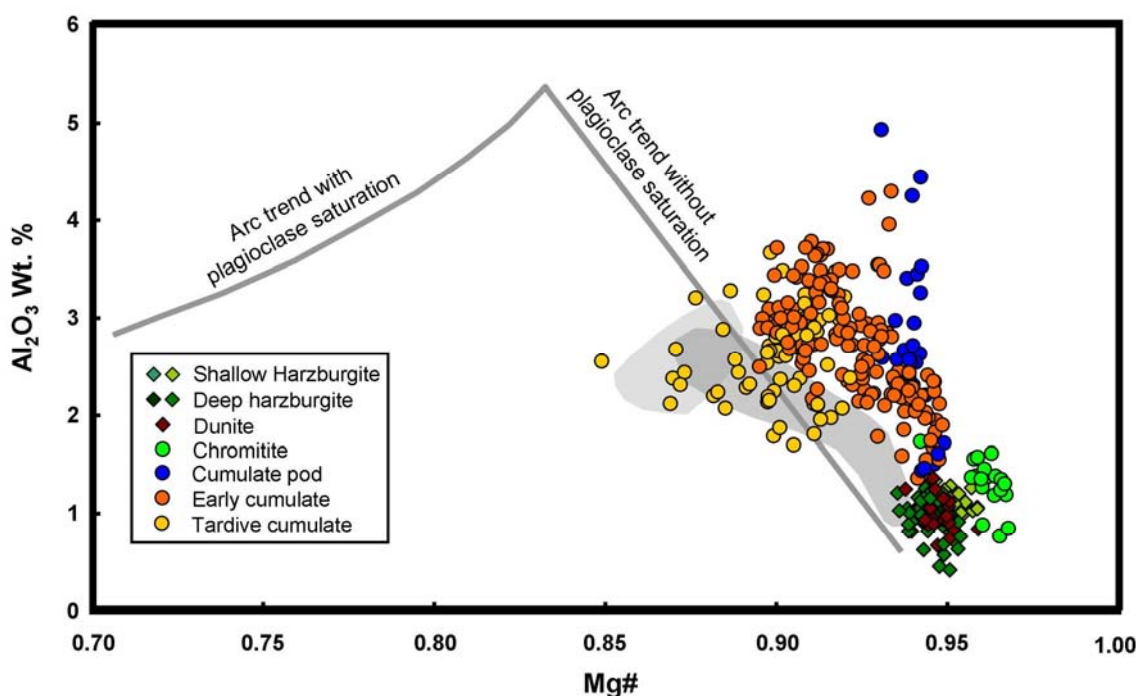


Fig.4.31. Al_2O_3 versus Mg# of clinopyroxene in the Massif du Sud. Dark orange circles are crustal cumulates without obvious plagioclase saturation whereas light yellow circles are gabbro-norite and ultramafic rocks with abundant plagioclase. Included is the trend for typical arc cumulates from the Chilas complex, Kohistan (Jagoutz et al., 2007) and Oman wehrlites and gabbro-norites without (dark grey) and with (light grey) plagioclase saturation (Koepke et al., 2009).

4.4.3.2. Chromitites and cumulate pods

The mining potential of New Caledonia for chromium ore in the last century led to important work on the genesis of chromitite pods and indicators of economic deposits (Cassard et al., 1981; Lago et al., 1982). Our data concerning the composition of chromite deposits is similar to what was obtained by previous studies (Leach & Rodgers, 1977; Cassard et al., 1981) [Fig.4.22].

The chromitite pods lie in a large but reliable stratigraphic horizon which is in the upper harzburgite (not more than 1000 m from the dunite zone) and in the lower dunite zone. In consequence, chromite deposits are only present in the area surrounding the gabbro massifs and in the north-west part of the Massif du Sud (Mt Dzumac, Cassard et al., 1981). Chromitite bodies are enclosed in a dunite shell and occasionally display a reaction zone with the surrounding harzburgite. Previous descriptions of these pods mention finger like projection of dunite into the harzburgite, following the foliation of the host

rock (Cassard et al., 1981). The metasomatic harzburgites analysed in this chapter are probably these reaction zones as they fit with these field descriptions. It has also been shown by field study that such metasomatised harzburgite do not always occur in relation with a chromitite deposit. These harzburgites probably represent a marginal reaction in relation with the percolation of melts capable of forming chromitite pods.

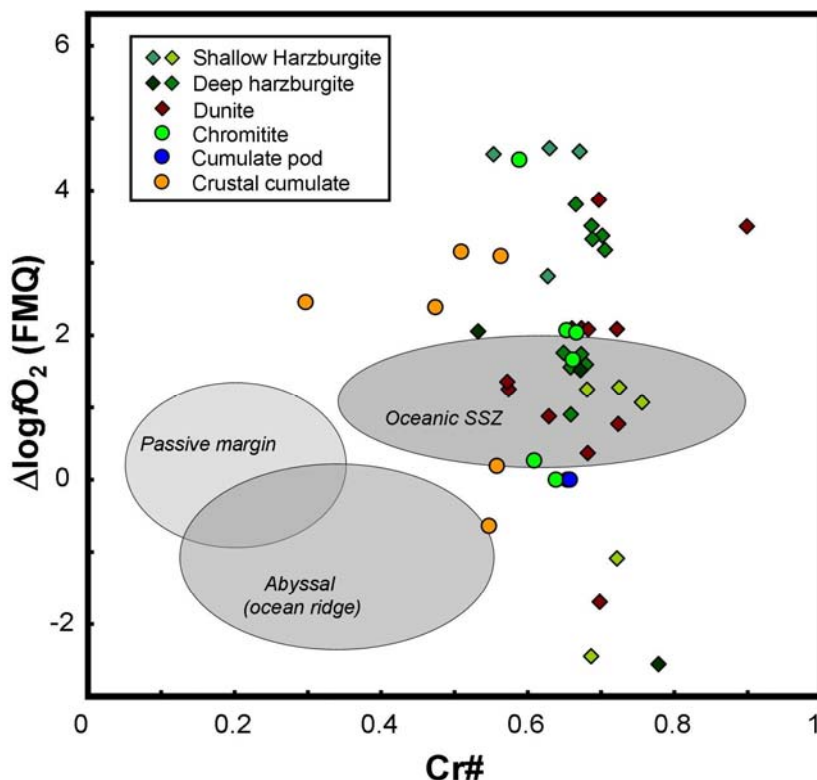


Fig.4.32. Oxygen fugacity calculated in some of the Massif du Sud rocks fO_2 is relative to fayalite-magnetite-quartz buffer and calculated using the olivine-spinel equation of Ballhaus et al. (1991). Fields for passive margin, abyssal and SSZ peridotites are from Parkinson & Pearce (1998). Most rocks with no local metasomatism are near or in the supra-subduction field.

The classical model to explain the formation of chromitite pods involves the percolation of hydrous picritic melts lying on the olivine + spinel cotectic in a convective (Lago et al., 1982; Lorand & Ceuleneer, 1989) or turbulent environment (Leblanc et al., 1981, Leblanc, 1985). The presence of olivine and clinopyroxene as inclusions with negative spinel crystal shape in chromite grains suggest that these silicates are entrapped during convection within the chromitite pocket [Fig.4.26]. However, a comparison with other chromitite bodies shows that no sulfides and hydrous minerals occur in the pods of New Caledonia (Lorand & Ceuleneer, 1989; Matveev & Ballhaus, 2002). This

observation suggests that no hydrothermal event has occurred after the emplacement (Johan et al., 1983; Hurlbert & von Grunewaldt, 1985; Lorand & Cottin, 1987) or that temperatures were too high to allow the crystallisation of hydrous phases (Talkington et al., 1984; Augé, 1987; Peng et al., 1995). Sulfides have been observed in the metasomatic harzburgite surrounding the pods. The composition of the sulfides suggests low fS for this part of the New Caledonia mantle. The models of Ballhaus (1998) and Matveev & Ballhaus (2002) which opposed the idea of Irvine (1977) do not seem to be applicable in New Caledonia where the Lago et al. (1982) model was partly designed. Ballhaus (1998) suggests the mixing of large amount of silicic melt with a fayalitic melt within the lithospheric mantle but no field observation support this idea. The magmatic accumulation of spinel by fractional crystallisation (Irvine, 1977) or variable densities (Matveev & Ballhaus, 2002) is probably applicable for the layered chromitite deposits present in the transition zone but chromitite pods do not show such features. Although the composition of the silicates enclosed in the spinel of chromitite schlierens is similar to the discordant pods, some features relates more to crustal cumulates such as low NiO in olivine or low CaO in clinopyroxenes than silicate compositions in chromitite pods.

The small cumulate pods which have been sampled at various places in the Massif du Sud (dunite zone, transition zone) remain an enigmatic series of rock. This is due to the small number of outcrops discovered and the variety of rocks unearthed (olivine clinopyroxenite, olivine orthopyroxenite, wehrlite). In these three cases, indication of strong cumulative textures have been observed and are confirmed by low NiO content in olivine and high alumina in clinopyroxene. However, unusual features are associated and all rocks have very high Mg# and can have abnormal enrichment in sulfides, titanium and depletion in calcium. The high Mg# and some particular trends suggest a relationship with chromite deposits but additional information has to be collected to develop this hypothesis.

4.5. Conclusions

The compilation of previous observations and the large amount of geochemical data collected in the Massif du Sud ophiolite in New Caledonia provides new insights on the general features of the ultramafic massif. The stratigraphy shows that the mantle unit is ultra-depleted and form a layer of mantle peridotite between 2 and 3.5 km thick. General structural observations supported by geochemical data show that the massif is a continuous body and is initially formed by an oceanic ridge north-east of the archipelago.

The upper zone of the ophiolite shows an increasing content in dunitic lithologies which dominate the stratigraphic sequence for up to 600 m until pyroxenes cumulates or interstitial phases start to crystallise within the dunite. The transition zone is formed of various cumulates with rare plagioclase and the intrusion of sills into the cumulate pile to form gabbro-norite. An impregnation halo is commonly observed surrounding these mafic sills.

The detailed observations in the field, in thin section and from geochemical data refine considerably the previous data available for the Massif du Sud. The lower part is rich in sulfides and has porphyroclasts of pyroxene – spinel symplectite. Higher levels display an evolution in the geochemical features as well as the presence of chromitite pods and a drop in orthopyroxene modal content in the uppermost harzburgite. In the dunite zone, the careful observation of spinel and olivine compositions, allow clarification of the controversial origin of the dunite zone, and show that the lower part is formed by replacive processes whereas the upper part is formed of cumulus olivine which represents the initial precipitation of magma crystallisation.

Finally, several features present in the mantle and the crustal sequence suggest that the New Caledonia ophiolite is mostly formed in a convergent environment. The high level of partial melting (20-30%) at low pressures (5-10 kbar) suggests that the crust was thickened by a secondary melting event. High oxygen fugacity in the mantle and hydrous magmas forming chromitites and delaying the precipitation of plagioclase in the crustal cumulates indicate that an arc environment is applicable when considering the geodynamical position of New Caledonia at the end of the Mesozoic.

5. Focused melt transport in the sub-arc mantle

5.1. Introduction

At mid ocean ridges, new basaltic crust is produced by peridotite melting in the upper mantle. The transfer of silicate liquids in the mantle from their generation at depth to the crust can be divided into two end-member processes: (1) reactive porous flow with continuous equilibration and (2) focused channelized flow with limited interaction of the melt with the surrounding peridotite. Detailed studies on basaltic glasses occurring at the ocean floor have shown that these melts are not in equilibrium with the peridotites occurring at the base of the newly formed crust (Spiegelman & Kenyon, 1992). This absence of equilibrium is inconsistent with reactive porous flow and supports focused flow as the main transport process (Kelemen et al., 1995). The presence of discordant features in mantle peridotites, such as replacive dunite channels and dykes in ophiolites, provides further evidence for focused melt transport (Nicolas, 1986; Kelemen, 1990).

In an arc setting, the composition of the mantle, partial melting products, the crust and the temperature at which the crust-mantle transition occurs is significantly different to a MOR-setting and therefore the melt transport processes are less well known in this environment. Both focused flow and reactive porous flow (Hopson et al., 2008; Bouilhol et al., 2009) have been considered as transport mechanisms of melt in the sub-arc mantle wedge. The higher water content of primitive arc lavas with respect to MORB (Parkinson & Arculus, 1999) is likely to have an influence on melt transport. As hydrous melts reach lower pressure and lower temperature, the crystallisation sequence can be modified and propagation mechanisms such as hydraulic fracturing are more likely to occur than continuous focused porous flow (Rabinowicz & Ceuleneer, 2005; Lambart et al., 2008). Therefore, melt-filled fractures are expected to be present in supra-subduction environments.

The important differences which exist between an arc complex and an ocean ridge (crustal cumulates, melt composition, thickness) (Pearce et al., 1984) reflects different sources of magmas but also a different petrological history for the melts present in these two environments. The exact nature of the processes

happening at the bottom of an island-arc are still debated (Greene et al., 2006; Garrido et al., 2007) in order to explain the bulk composition of the arc crust and the incoming melts. This controversy is known in the literature as the arc paradox (Kelemen et al., 2003).

This chapter will describe channels and dykes that are discordant to the harzburgite tectonite in the upper mantle section of the Massif du Sud. The exceptional conservation of 3 to 4 kilometres of a sub-arc mantle and a thick section of the crust-mantle transition zone provides a unique view on the features of the magma plumbing system under an arc. We will show that dunite channels are also widespread in the sub-arc environment and that these conduits evolve into pyroxenite channels and then dykes below the crust-mantle transition. Direct field and chemical evidence is gathered to establish the link between dunite channels, pyroxenites and the lower-crustal gabbros of the New Caledonia ophiolite. This chapter will also use major and trace element compositions of olivine, orthopyroxene and clinopyroxene to show a genetic relationship between the different channels, dykes and gabbros. The evidence of strong melt-mantle interaction and melt fractionation at sub-Moho levels has implications on the differentiation processes and the composition of the liquids forming the nascent arc crust.

5.2. Analytical techniques

Analytical techniques used in this chapter are identical to the methods described in chapter 4. Analytical procedures are adapted to the study of dykes and channels and a specific interest is given to zoning within grains and across discordant features cutting through ultramafic and mafic rocks.

5.3. Results

5.3.1 Sampling and field relations

Sampling for this study was done in parallel to the collection of rocks used to establish the stratigraphical sequence of the Massif du Sud. The selection of samples was done from the bottom to the top of the ophiolite, focusing on fresh rocks that show evidence for melt-rock interaction [Fig.5.1, 5.2]. Dunite channels were sampled from the deepest roots of the ophiolite to the bottom part of the dunite zone where dunite pods are more abundant. Different generations of pyroxenites occur in the Massif du Sud and only one series is mentioned here [Chap.6], which is sampled in the upper harzburgite zone and within the dunite transition zone. In the uppermost part of the mantle section, mafic dykes occur and were sampled in several locations within the harzburgite, the dunite zone and the cumulate series. To provide a constant monitoring on the surrounding rocks of the discordant dykes, the host harzburgite, dunite and transitional ultramafic and mafic rocks (dunite, wehrlite, pyroxenite and olivine gabbro-norite) were sampled and this dataset is discussed in chapter 4 [Tab.S11].

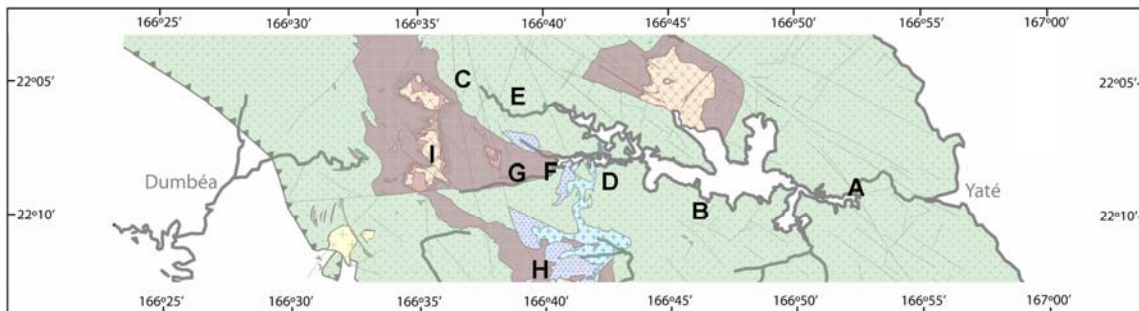


Fig.5.1. Location of representative samples of this chapter in the geological section from Yaté to Montagne des Sources and Dumbéa.

- A.** Yaté dam area (YR and YA series): dunite channels, harzburgite
- B.** Yaté Lake (R32): dunite channel, harzburgite
- C.** Upper Rivière bleue (PRB75): olivine harzburgite – (PRB74-77b): dunite channels
- D.** Rivière de la locomobile (PRB34-36): dunite channel – (PRB35-37): harzburgite
- E.** Upper Rivière Bleue (PRB60-61): proto-pyroxenite series – (PRB63-68-69): pyroxenite dykes
- F.** Rivière Blanche (RBL2): websterite dyke
- G.** Upper Rivière Blanche (PRB49): composite mafic dyke series – (PRB50-53): dunite, olv harzburgite
- H.** Yaté Road (R2) - olivine cumulate
- I.** Montagne des sources (MDS series): crustal cumulates and mafic dykes.

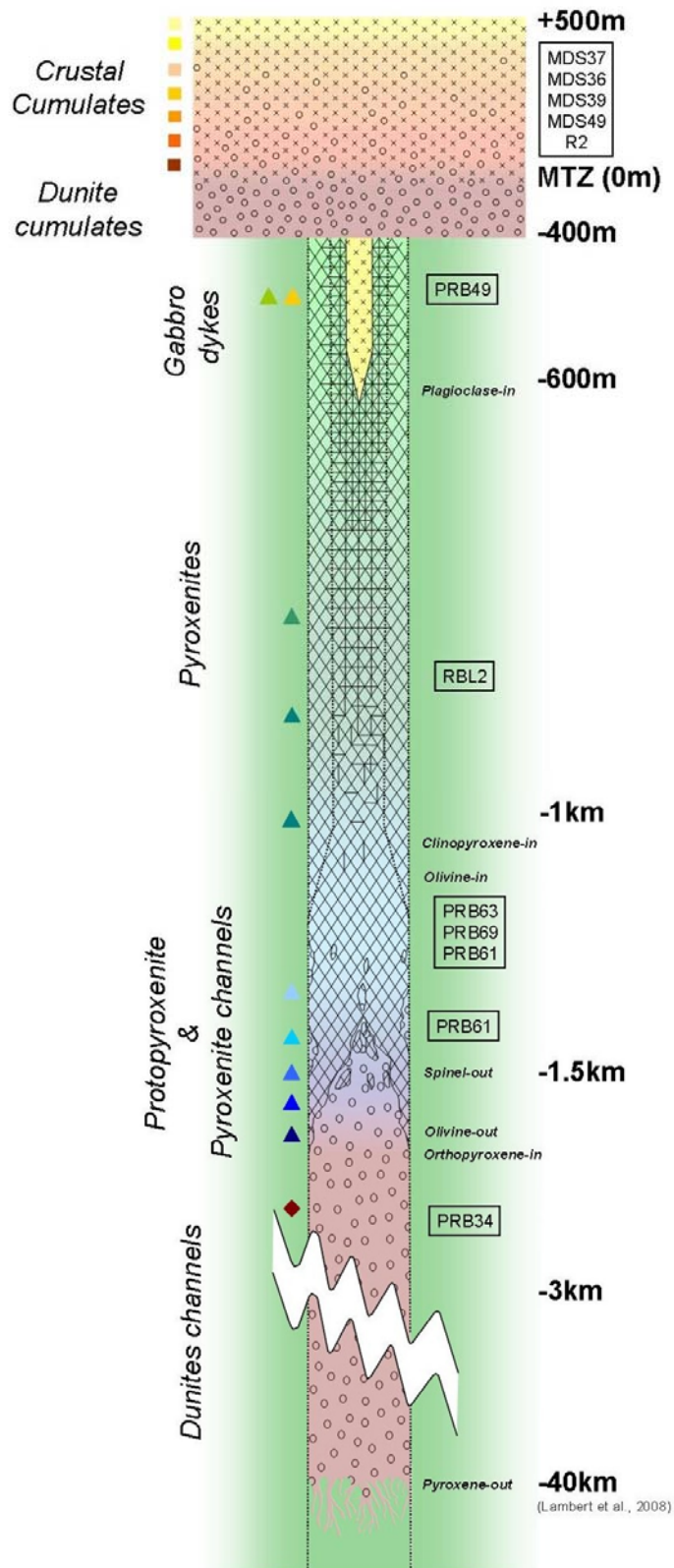


Fig.5.2. Schematic evolution of the discordant features in the Massif du Sud with depth (from bottom to top). Lithologies are indicated on the left as well as the icons used in the geochemical plots. On the right side, representative sample names and an approximate depth for the transition between lithologies is given.

5.3.1.1 Dunite to pyroxenites

Dunite channels

Dunite channels are observed throughout the Massif du Sud ophiolite. They cross-cut the main tectonite harzburgite foliation at an angle from a few degrees to 30-40° and can be followed for several meters [Fig.5.3]. Channels have a width ranging between a few centimetres to a few metres. Their abundance as well as their size increase up section to the point where channels start to coalesce forming large decametre sized pods and tabular dunite bodies which have been partly developed in chapter 4 [Fig.4.8a] where they finally become the main lithology of the lower transition zone.

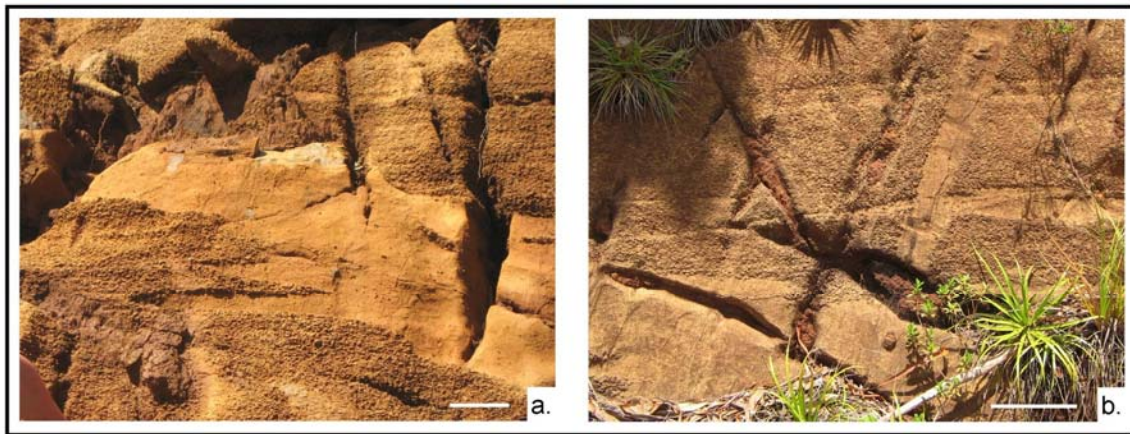


Fig.5.3. Dunite channels in the upper harzburgite. The scale bar is 10 cm long.

a. Dunite channel PRB34 showing fingers of dunite penetrating the surrounding harzburgite as well as the presence of peridotitic restites within the channel.

b. Dunite channels in the Rivière bleue area showing cross-cutting relationship and the high angle of discordance regarding to the harzburgite foliation.

Dunite channels display sharp boundaries with respect to the harzburgitic country rock, marked by the disappearance of orthopyroxene porphyroclasts present in the harzburgite. Boundaries are not straight and occasionally “fingers” of dunite replacing the surrounding harzburgite occur [Fig.5.3a]. The modal composition is similar to the dunite zone and dominated by olivine with only a few percent of chromite present. In contrast to the tectonite harzburgite, clearly discordant channels have no highly strained olivine crystals and the grain size is often larger than in the surrounding rocks. Chromite grains can show idiomorphic recrystallisation or microsymplectites [Fig.4.26] although initial structures, such as trails, are preserved from the harzburgite through the channel without modification, suggesting a replacive nature for these rocks. In the upper part of the mantle

section, dunite pods and the lower dunite zone display the same microscopic textures than the channels. However, it has been shown that the upper part of the dunite zone has a cumulate origin and differs from replacive processes [Chap.4.4.2.1].

Proto-pyroxenites

Proto-pyroxenites have many textural features in common with dunite channels such as their size ranging from a few centimetres to a few decimetres or their relationship with the surrounding harzburgite, notably the clear but curved discordant boundaries and the presence of pyroxenite fingers in the host rock. They occur in the first 1 km below the transition zone and occasionally show a complete transition from dunite to orthopyroxenite within that zone [Fig.5.4]. Proto-pyroxenites are technically a harzburgite (Streckeisen, 1972) but the channels display a very heterogeneous composition in many cases [Fig.5.4, 5.5]. Along the width of a channel, the composition ranges from dunite to orthopyroxenite randomly or as a suite of parallel layers. Often, a greater amount of orthopyroxenes occurs on the walls than in the middle of the channel, which can remain olivine-rich. Variations are also observed when the channel is followed upwards. First, the precipitation of some large orthopyroxene crystals within a dunite channel occurs, followed by an increasing modal content in orthopyroxene grading into an olivine-bearing orthopyroxenite and finally into a pure orthopyroxenite channel [Fig.5.4e]. In these orthopyroxenite channels, spinels are typically absent and orthopyroxene contains rare clinopyroxene or amphibole exsolutions. In many cases, dunite walls, dunite pockets or residual olivine crystals can remain in the orthopyroxenite channels [Fig.5.5]. The field relationships, therefore, suggest that the proto-pyroxenites and pyroxenite channels represent an upward evolution of the dunite channels in the mantle section. Apart from the more common “inward growth” of these proto-pyroxenites, less frequent “outward growth” or cyclic orthopyroxene precipitation was also observed [Fig.5.5, 5.6].

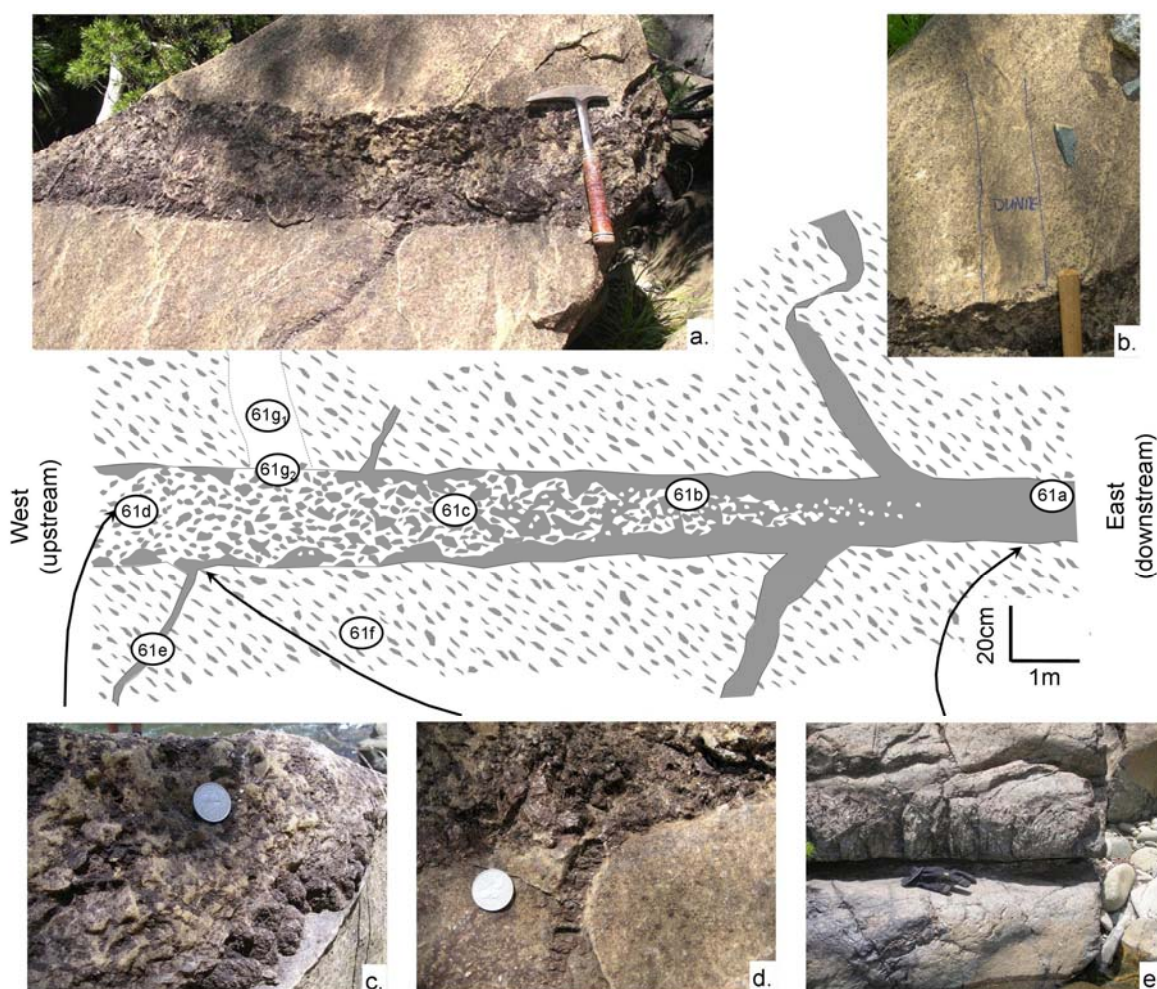


Fig.5.4. Proto-pyroxenite transition PRB61. The schematic representation show olivine (white) and pyroxenes (grey) along the channel. The position of the collected samples is given by the ellipses as well as the location of the pictures.

a. PRB61d, PRB61e and PRB61g show the general view of a proto-pyroxenite. The main channel is crosscutting the harzburgite and dunite channel (b) (PRB61g) which is hardly seen in the upper wall whereas a small pyroxenite dykelet (d) (PRB61e) is on the bottom wall.

b. PRB61g show a dunite channel, highlighted with a pen, crosscut by the proto-pyroxenite channel.

c. Closer view of PRB61d showing the large orthopyroxene crystals (brownish) embedded in a dunitic matrix (yellowish).

d. Pyroxenite dykelets PRB61e at its origin on the wall of the channel.

e. Pyroxenite channel around 20 meters downstream of PRB61d (a) which show an orthopyroxenite composition.



Fig.5.5. Proto-pyroxenite channel PRB68 with cyclic replacement. Segments of orthopyroxenite are seen on the left within a dunite channel whereas a dunite core is seen on the right. This channel can be followed for several tens of meters and become a websterite dyke [Fig.5.7a]. The coin is 3.5 cm in diameter.

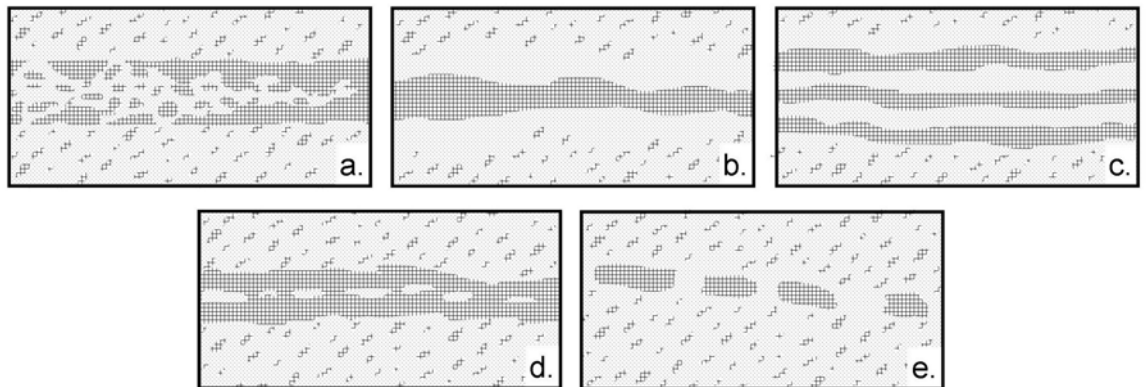


Fig.5.6. Various schematic representation of proto-pyroxenite transition.

- a. Standard transition with inward growth of orthopyroxene (PRB60, PRB61) [Fig.5.4].
- b. Orthopyroxenite with dunite walls, with slow outward growth (PRB71).
- c. Layered proto-pyroxenites alternating locally orthopyroxene and olivine layers (PRB69).
- d. Proto-pyroxenite with dunite restite enclosed in the channel (PRB68) [Fig.5.5].
- e. Cyclic precipitation of pyroxenes within a dunite channel (PRB68) [Fig.5.5].

5.3.1.2 Pyroxenites

Pyroxenite dykes occur in the upper part of the mantle [Fig.5.2] and can range from a few millimetres to a few decimetres. They cross-cut the harzburgite and the dunite zone with highly discordant angles and show a large grain size and common harrisitic growth on their walls [Fig.5.7]. Some of these dykes are attached to proto-pyroxenite channels, shooting off from the channel wall with a sharp angle into the surrounding host-rock [Fig.5.4a, 5.4d]. The cross-cutting relationship such as sharp contacts and high angle junctions between dykes indicates an emplacement by hydraulic fracturing of the surrounding harzburgite and dunite.

Pyroxenite dykelets directly related to channels and primitive dykes consist mainly of orthopyroxene but occasionally there is a significant increase in

clinopyroxene modal content along the dykes which become websterite and clinopyroxenite [Fig.5.7]. Minor amounts of olivine crystals do occur in these dykes, often as inclusions within orthopyroxene grains. Spinel is very rare and occur as minute inclusions.



Fig.5.7. Pyroxenite dykes.

a. Websterite dyke cross-cutting olivine harzburgite and an older stirred orthopyroxenite. This dyke is directly related to the proto-pyroxenite in Fig.5.5.

b. Orthopyroxenite dyke crosscutting harzburgite and dunite channel in Dumbéa valley (DR32).

c. Clinopyroxenite dyke RP33 in the cumulus dunite of Rivière des Pirogues. The coin is 3.5 cm in diameter.

5.3.1.3. Composite dykes and cumulative rocks

Composite gabbro dykes

The uppermost part of the mantle sequence (olivine harzburgite and dunite) is crosscut by many dunite channels, pyroxenites and gabbro dykes. The latter are particularly recognisable in the field by the presence of plagioclase. The size of the dykes varies from a few centimetres to a few decimetres. Dykes often display a gabbro core surrounded by a reaction zone consisting of pyroxenite walls [Fig.5.8a]. One dyke with an exceptionally well-preserved zoning has been sampled and analysed in detail [Fig.5.8b]. Ideal composite dykes show from core to host rock: microgabbro – gabbro – clinopyroxenite – websterite – orthopyroxenite – dunite – harzburgite. Contacts between those different zones are relatively sharp, notably between gabbros and pyroxenites. The pyroxenite sequence is segregated but lacks clear boundaries whereas the pyroxenite-dunite or pyroxenite-host rock contact is less defined and shows an inward modal increase in orthopyroxene regarding to the host rock. These observations link the formation of discordant orthopyroxenites and websterite dykes to the formation of gabbro dykes in the uppermost part of the mantle section.

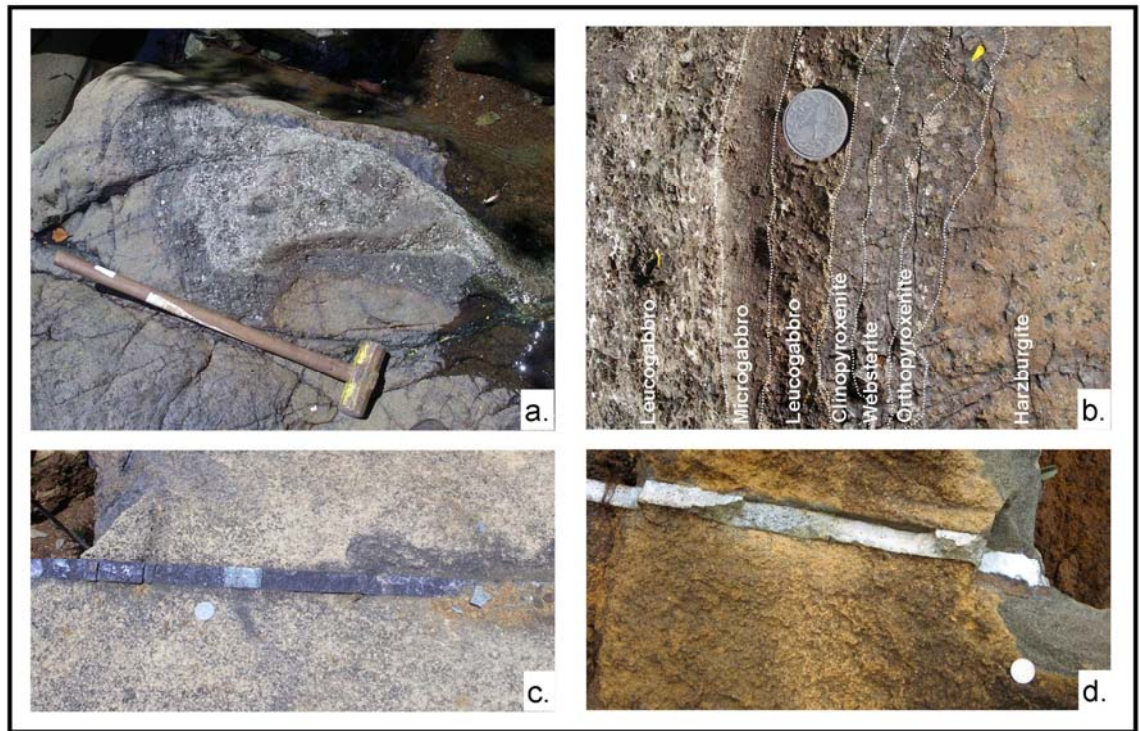


Fig.5.8. Gabbro dykes.

- a.** Composite gabbro dykes PRB49 cropping out along its dipping. The light grey zone is gabbrobronic, the dark zone is pyroxenites and the outer brownish zone is an olivine harzburgite.
- b.** Close view of the composite dyke PRB49 going from harzburgite (right) to the microgabbrobronic core (left).
- c.** Simple microgabbrobronic dyke (MDS40) in cumulus dunite.
- d.** Leucogabbro dyke (MDS32) crosscutting an impregnated cpx-plag dunite.

The microgabbrobronic has an equigranular texture of millimetre size grains of clinopyroxene and plagioclase with some larger orthopyroxene. Minor amounts of olivine are present but spinel is mostly absent. The surrounding coarse-grained gabbrobronic has a similar modal composition to the central microgabbrobronic but displays a cumulate texture. The wall in contact with the gabbrobronic is a coarse grained clinopyroxenite with minor amount of orthopyroxene. Orthopyroxene modal content increases progressively outwards to the websterite and orthopyroxenite layers. Olivine and spinels are barely present and clinopyroxene only occurs in trace amounts in the outer orthopyroxenite part of the dyke. The surrounding olivine-rich harzburgite bears residual mantle features with a strong tectonite foliation and orthopyroxene porphyroclasts. The outermost zone can occasionally be a dunite layer. This layer is undifferentiated from the surrounding when it occurs in the dunite transition zone but is visible when composite gabbro dykes occur in the olivine-rich harzburgite. The described sequence is the most complete example of a composite mafic dyke in the mantle section. In many

cases, it is common to see gabbro dykes displaying sharp contacts with the ultramafic host rock, particularly in the upper parts of the transition zone and within the oceanic crust [Fig.5.8c,d].

Crustal cumulates

The crustal cumulates have been developed in detail in the previous part [Chap.4.4.2.3]. In the scope of this chapter, it is worth highlighting the similarities between the gabbro-norite sills of Montagne des Sources [Fig.4.12, 4.13] and the modal composition of the mafic dykes present from the uppermost part of the mantle to the layered crustal rocks. Crustal pyroxenite layers present in Rivière des Pirogues [Fig.4.11] also share similarities with some of the pyroxenite dykes and pyroxenite walls of composite dykes. However, the cumulus olivine which has distinctive features compared to the lower dunite zone, has no equivalent in dunite channels or in the walls of composite mafic dykes, suggesting that dunite walls have the same origin as the dunite channels and are formed by replacement reactions.

5.3.2. Major elements mineral chemistry

5.3.2.1. Olivine

The major element composition of olivine in dunite channels is directly related to the composition of the surrounding rocks. In consequence, there is a correlation between the Mg# of olivine and stratigraphy within the channels which slightly increase from 0.903 to 0.925. This increase is particularly visible when the modal concentration in olivine increases in the harzburgite of the uppermost mantle [Fig.5.15]. Dunite channels and pods of this area tend to be more magnesian than dunite channels of the east coast, which lie at deeper levels [Tab.S12].

In the proto-pyroxenite channels, the Mg# (0.91-0.92) is in the upper range of values found for dunite channels. It reflects the composition of the olivine-rich harzburgite of the uppermost mantle whereas in pyroxenite dykes, the rare olivine grains typically show a small decrease in Mg# (0.91-0.89), lower than the surrounding rocks. These values are similar to what is observed in the pyroxenite

walls of the composite mafic dykes [Fig.5.9] but contrasts with olivine grains in the gabbronorite core of the dyke, where the Mg-number of olivine drops sharply to 0.84 [Fig.5.16]. Finally, as mentioned earlier [Chap.4.3.4.1], olivine crystals in the crustal section have a Mg-number ranging from 0.91 in dunites to 0.85 for the accessory olivine in the leucogabbronorite sills [Tab.S12].

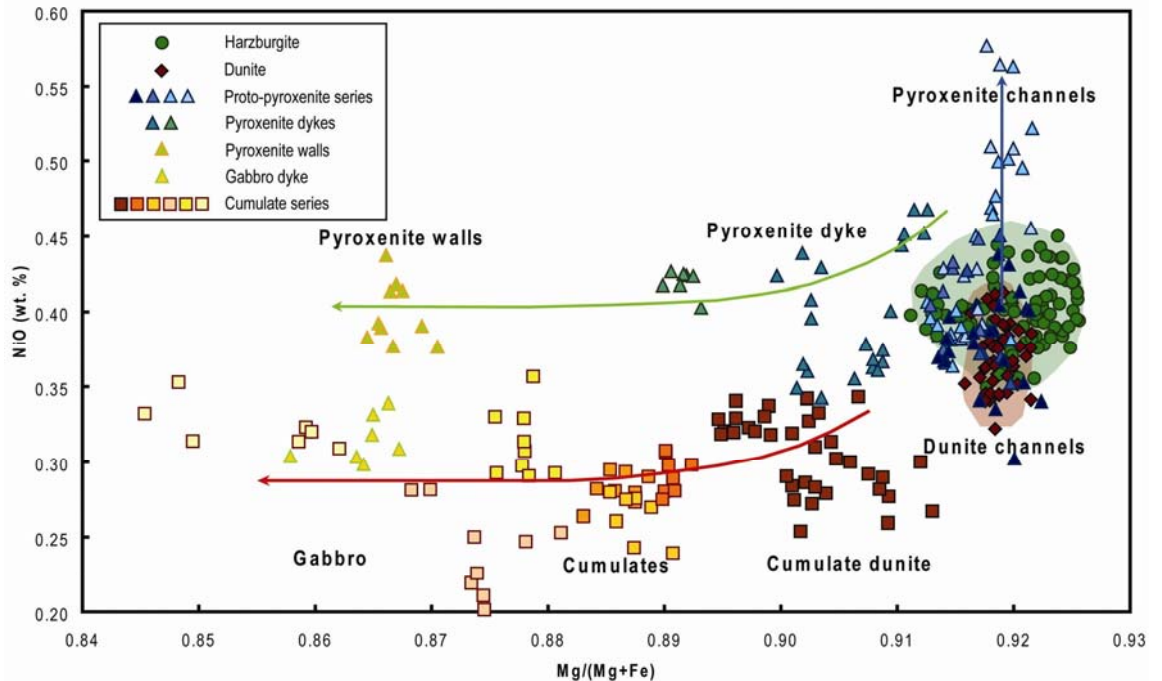


Fig.5.9. Mg-number ($\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$) versus nickel oxide content in olivine. Harzburgite are dark green circles in the green field. Dunite channels are brown diamond covered by the brown field. Proto-pyroxenites and pyroxenite channels are blue triangles. Pyroxenite dykes are blue-green triangles with decreasing mg-number values. Squares with low nickel content are crustal cumulates with different colours related to rocks with increasing differentiation. Arrows underline the evolution trends of the different rocks (blue = proto-pyroxenite transition; green = pyroxenite dykes; red = crustal rocks).

In dunite channels, olivine generally displays a lower average Ni concentration with respect to the host rock [Fig.5.9] with small variations noticeable through the width of the channel. In the proto-pyroxenite, the composition of the olivine shows a clear increase in NiO from 0.33 to 0.58 wt.% in direct correlation with the increasing modal abundance of orthopyroxene. This high content in nickel is not observed in the pyroxenite dykes and the olivine cumulates where values are back to lower levels (0.40-0.30 wt.%). In the zoned gabbro dykes, olivine from the outer zone contains ~0.4 wt.% NiO in the outer walls and drops to 0.3 wt.% in the cpx-bearing rocks. Manganese show the same behaviour as presented in chapter 4 [Fig.4.18b] and has an opposite behaviour with respect

to NiO in the discordant rocks and in the crustal cumulates where it increases from 0.10 to 0.25 wt.% MnO as part of a differentiation trend [Tab.S12].

5.3.2.2. Orthopyroxene

Similarly to olivine, the orthopyroxene Mg-number in harzburgite slowly increases with stratigraphy from 0.90 to 0.93 [Fig.4.19] and results in the orthopyroxene precipitating in the proto-pyroxenite of the upper mantle have the same Mg-number as the surrounding harzburgite (~0.92) [Fig.5.10]. In pyroxenite dykes, the Mg-number decreases from 0.91 to 0.88 and is consistent with the zoned dykes where harzburgite and orthopyroxenite are at mantle values (~0.93) whereas inner cpx-bearing layers decrease from 0.92 to 0.83 [Fig.5.16]. In crustal cumulates, the Mg# decreases as differentiation progresses to values as low as 0.80 [Fig.5.10, Tab.S14].

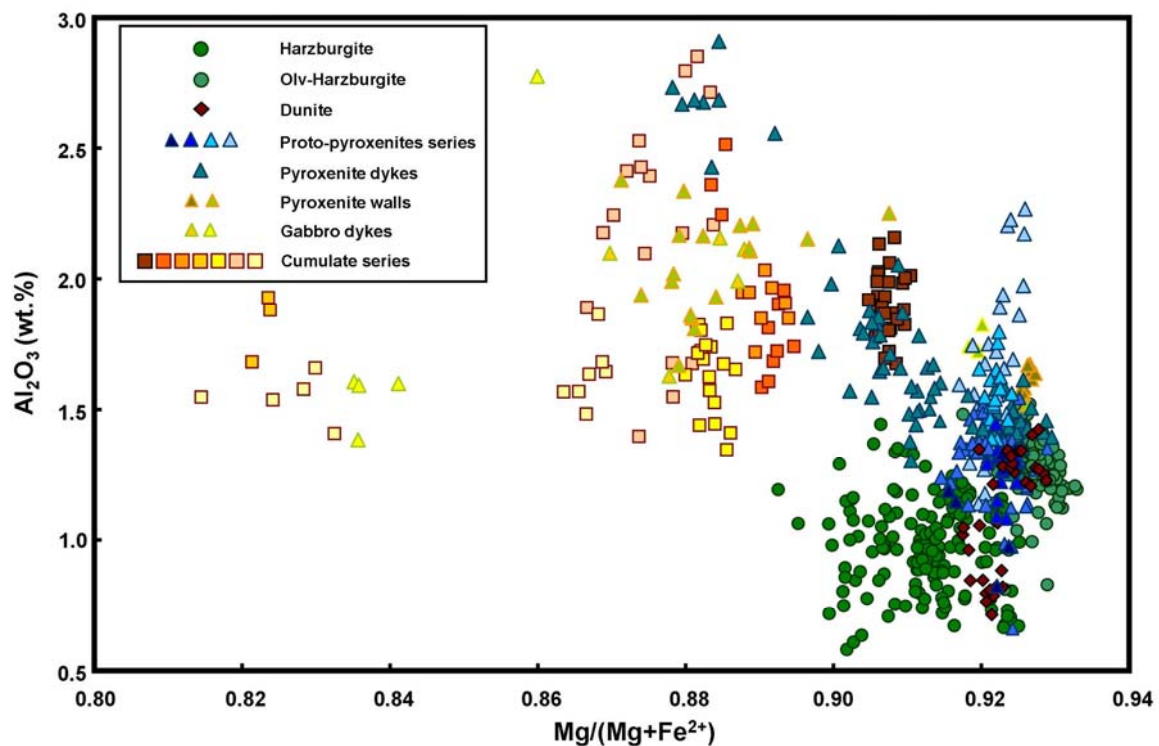


Fig.5.10. Mg-number ($\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$) versus alumina content in orthopyroxene. Harzburgite are dark green circles and olivine harzburgite are light green circles with higher Mg# and Al_2O_3 . Dunite channels with orthopyroxene residues are brown diamond. Proto-pyroxenites and pyroxenite channels are blue triangles with a range from dark blue to light blue as a function of orthopyroxene abundance. Pyroxenite dykes are blue-green triangles with decreasing mg-number values. Squares are crustal cumulates with different colours related to rocks with increasing differentiation.

Al_2O_3 strongly increases between mantle orthopyroxene porphyroclasts (~1.3 wt.%) and crystals within the pyroxenite channels (up to 2.3 wt.%) [Fig.5.10]. A similar behaviour is seen for chromium. In the dykes, Al_2O_3 remains high in the orthopyroxene (1.5-2.5 wt.%) whereas Cr_2O_3 drops from 1.0 to 0.5 wt.% in relation with the appearance of Ca-rich pyroxenes. In the zoned dykes, alumina content increases progressively between harzburgite, the different pyroxenites layers and the gabbro zone. However, a small decrease is noticed between the gabbro norite and the fine-grained gabbro norite constituting the core of the zoned dykes. In the crustal cumulates, the alumina content remains steady at 2.0 wt.% [Tab.S14].

The calcium content also shows a strong increase in the proto-pyroxenites (up to 2.5 wt.%) in comparison with the mantle porphyroclasts (~0.7 wt.% CaO) and remains high in pyroxenite dykes. In the zoned mafic dykes, an increase from ~1.0 to 2.0 wt.% CaO is noticed from outside to inside the dyke [Tab.S14].

5.3.2.3. Clinopyroxene

In the harzburgite, proto-pyroxenites and other opx-dominated rocks, clinopyroxene occurs mainly as exsolutions within orthopyroxene grains. As a consequence, such clinopyroxene composition mainly reflects subsolidus equilibration of the surrounding phase. In websterite dykes, primary Ca-rich pyroxene crystallises with a high Mg#, which decreases as clinopyroxene is more abundant (0.95 to 0.90). In the zoned mafic dykes, there is a slight decrease in Mg# within the clinopyroxene-rich walls. Then, the Mg-number drops sharply from 0.94 to 0.88 towards the central gabbro zone [Fig.5.16]. In the crustal gabbro cumulates, Mg-number varies from 0.95 to 0.85 in correlation with the stratigraphical position of the cumulate layers [Tab.S16].

Alumina content in clinopyroxene of cpx-bearing pyroxenite dykes is ranging from 2.0 to 3.5 wt.% Al_2O_3 . In the pyroxenite walls of the composite dykes, Al_2O_3 is increasing towards the central gabbro zone. A similar increase up to 3.0 wt.% Al_2O_3 is noticed in the ultramafic cumulates of the crust and in the gabbro norite sills. Cr_2O_3 rises up to 1.6 wt.% in the websterite dykes and drops back down to ~0.8 wt.% in the plagioclase-bearing rocks. CaO content decreases constantly from 25 wt.% in the poikilitic clinopyroxenes and from 24 to 22 wt.% CaO in gabbro norite sills [Tab.S16].

5.3.2.4. Spinel

Spinel is mostly magnesiochromite in the dunite channels and pods where the Cr-number is between 0.67 and 0.79 which is slightly higher than the chromite grains occurring in the harzburgite, ranging from 0.60 to 0.71 [Fig.4.22]. In proto-pyroxenites, spinels display the same chemical characteristics as in the dunite channels and remain unchanged until their dissolution. In the harzburgite and dunite surrounding the mafic dykes, spinels are more sensitive to the presence of percolating melt and can show a lower Cr-number towards the centre of the dyke [Fig.5.16]. In the crustal cumulates, as it was discussed earlier [Chap.4.3.4.4], idiomorphic spinels have a decreasing Cr# from 0.63 in the earliest dunite cumulates to 0.51 in the gabbro-norites [Fig.4.22]. In addition, cumulative spinels have a slightly higher TiO₂ content than the mantle (dunite and harzburgite) spinels [Fig.4.29, Tab.S18].

5.3.2.5. Plagioclase

Plagioclase only occurs in evolved mafic dykes close to the crust-mantle transition and within the cumulate series. In most cases, these plagioclases are calcium-rich and show an anorthite content between An₈₈ and An₉₂ [Fig.6.14, Tab.S19].

5.3.3. Trace elements mineral chemistry

5.3.3.1. Olivine

Olivine grains in dunite show larger variations in trace element concentrations than the major elements. These variations are noticeable for Ca, Ti, Cr, V, Al, Y, P and Li between olivine from harzburgite, residual olivine-rich harzburgite and dunite channels [Fig.5.11]. Most trace elements and notably Al, Y, Cr, P and Sc are enriched in channels with respect to harzburgite. Ti and Al in olivine are efficiently discriminating residual olivine harzburgite from enriched replacive dunites as well as Ca and V in dunite channels of the uppermost

harzburgite. However, in some cases, the enrichment is related to the modal composition of the rocks (Al, Ti, V, Cr) and eventual subsolidus re-equilibration (Ca). Li, Ni and Co show a depletion in the dunite channels when concentrations are compared to the olivine from the host rocks [Tab.S13].

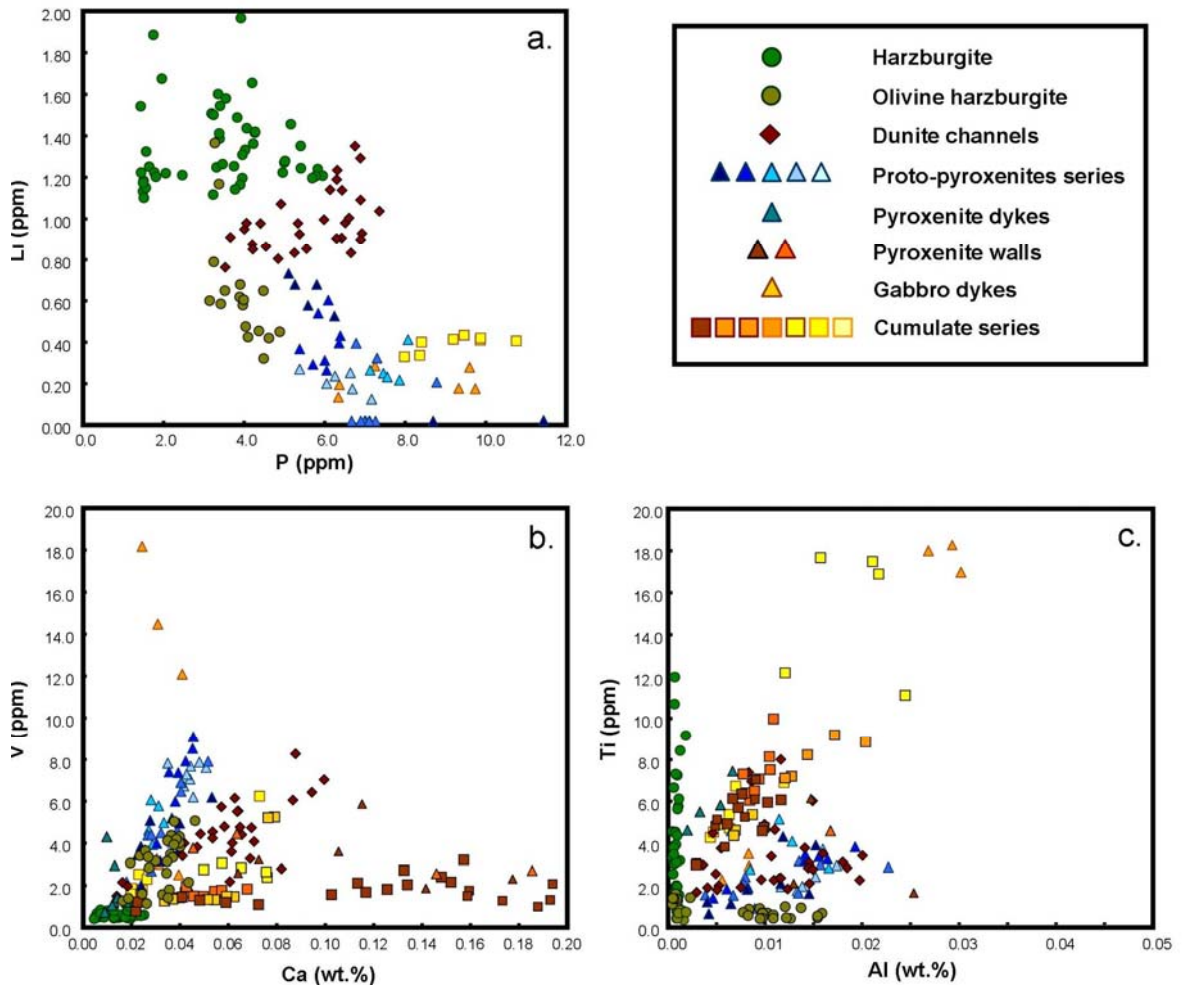


Fig.5.11. Trace elements in olivine.

a. Phosphorus versus lithium shows a trend with high-Li and low-P for harzburgite to low-Li and high-P for the crustal cumulates.

b. Calcium versus vanadium shows very low values in the harzburgite whereas olivine harzburgite and dunites have higher contents. Pyroxenite channels, dykes and walls are high in vanadium whereas crustal cumulates are relatively low but higher in calcium.

c. Aluminium versus titanium shows very low values in aluminium in harzburgite and decreasing content in titanium to shallower peridotites. Dunite channels and pyroxenites show an intermediary position between olivine-harzburgite and crustal cumulates.

In the protopyroxenites channels, Ni, Co, V, Sc and P increase and are enriched compared to the surrounding harzburgites and dunites. Alternatively, elements which are compatible in pyroxenes such as Ca, Al and Mn, decrease in both pyroxene-bearing channels and dykes. Finally, lithium is also decreasing and is stabilised at low values in the pyroxenites [Tab.S13].

Above the crust-mantle transition, olivine grains keep mostly the features acquired in the pyroxenite dykes with noticeable depletion in Ni, Cr, Al, Ca, V and Li and enrichment in Y, Mn and P. However, the variation of concentration in the different trace elements is highly dependent on the presence and amount of pyroxenes and oxides in those cumulate rocks [Tab.S13].

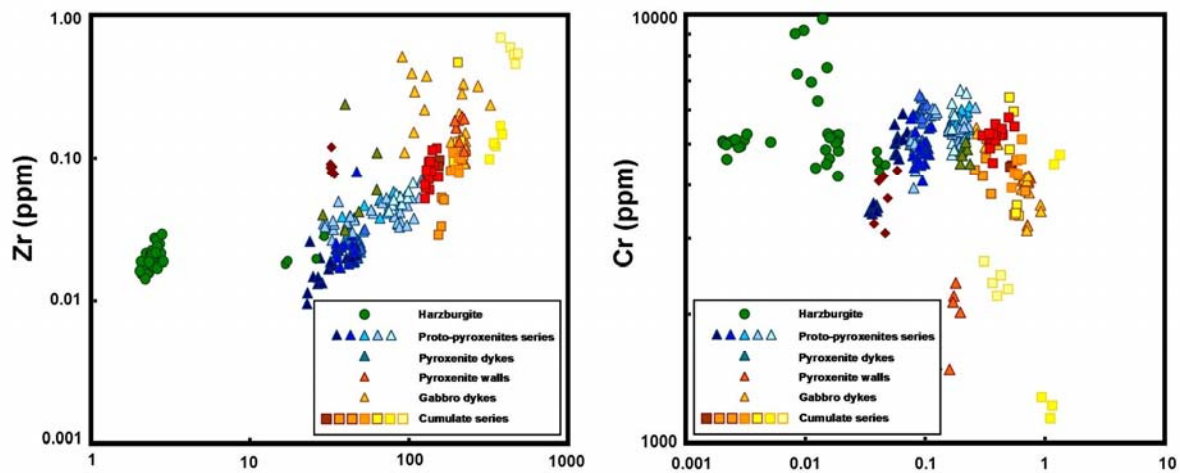


Fig.5.12. Trace element in orthopyroxene. Blue triangles are the proto-pyroxenite sequence with lighter colours along with increasing orthopyroxene modal content. Orange triangle are orthopyroxenes from walls and gabbro dykes and red to yellow squares are grains in crustal cumulates.

- a.** Titanium versus zirconium show a strong increase in both elements during the proto-pyroxenite transition and a steeper enrichment for zirconium in the crustal cumulates.
- b.** Yttrium versus chromium show that yttrium has a behaviour similar to titanium whereas chromium is stable in pyroxenite channels and early cumulates but drops considerably in mafic cumulates

5.3.3.2. Orthopyroxene

Orthopyroxene in the protopyroxenites have a clearly distinct trace element signature when compared with porphyroclasts of the surrounding harzburgite. This signature is generally characterised by a greater abundance in trace elements. In addition, as the proportion of orthopyroxene increases in the channels, clear evolution trends for Ti, Cr, V, Zr, Nb, Y, HREE and Sr underline an enrichment going from dunite channels to orthopyroxenite dykes [Fig.5.12]. The appearance of clinopyroxene in the dykes left the trends unchanged although some decrease such as Ti/Zr and Cr could be related to the presence of accessory spinels or clinopyroxenes. In the crustal cumulates, concentrations of the incompatible elements Ti, Y, Nb, Zr and Sr are still increasing whereas V, Cr and Ni now show a depletion trend, in relation with the precipitation of oxides [Tab.S15].

In the composite dykes, the same stepped structure seen in modal and major element composition is also visible in the trace elements [Fig.5.13]. HREE, LILE and TM get slowly enriched inward whereas LREE and HFSE display the opposite behaviour and tend to decrease toward the core of the dyke. In the gabbronorite zone, however, and particularly in the microgabbro, most trace elements are slightly enriched, particularly in HREE and Ti [Tab.S15].

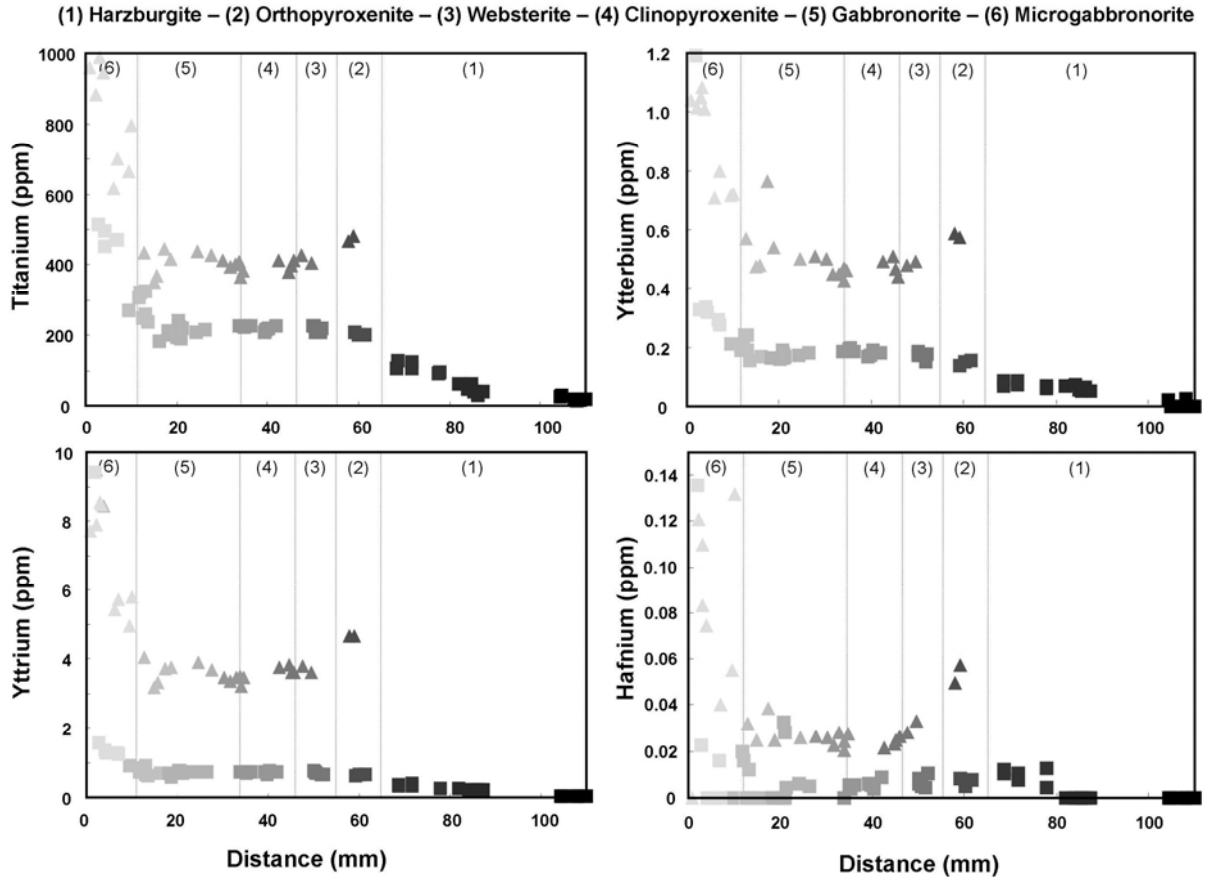


Fig.5.13. Trace element content in a transect from host rock (1) to the microgabbronorite core (6) of PRB49. Zones are in relation with Fig.5.8b. Squares are orthopyroxene grains and triangles are clinopyroxene showing a concentration gap between the host rock (1) and the cumulate walls (2-5) and a strong increase in the fine-grained core (6) of the dyke.

The comparison of the rare earth elements patterns for each rock shows a progressive increase in concentration for medium and heavy REE in the proto-pyroxenite transformation. As the channels and dykes evolve in composition, the REE pattern is getting slightly more enriched and more fractionated between the LREE and HREE ($Nd/Lu_{Hzb} : 0.07$; $Nd/Lu_{D.Channel} : 0.11-0.06$; $Nd/Lu_{Dyke} : 0.05-0.03$; $Nd/Lu_{Gabbro} : 0.02-0.01$), approaching the pattern of the cumulate gabbros [Fig.5.14, Tab.S15].

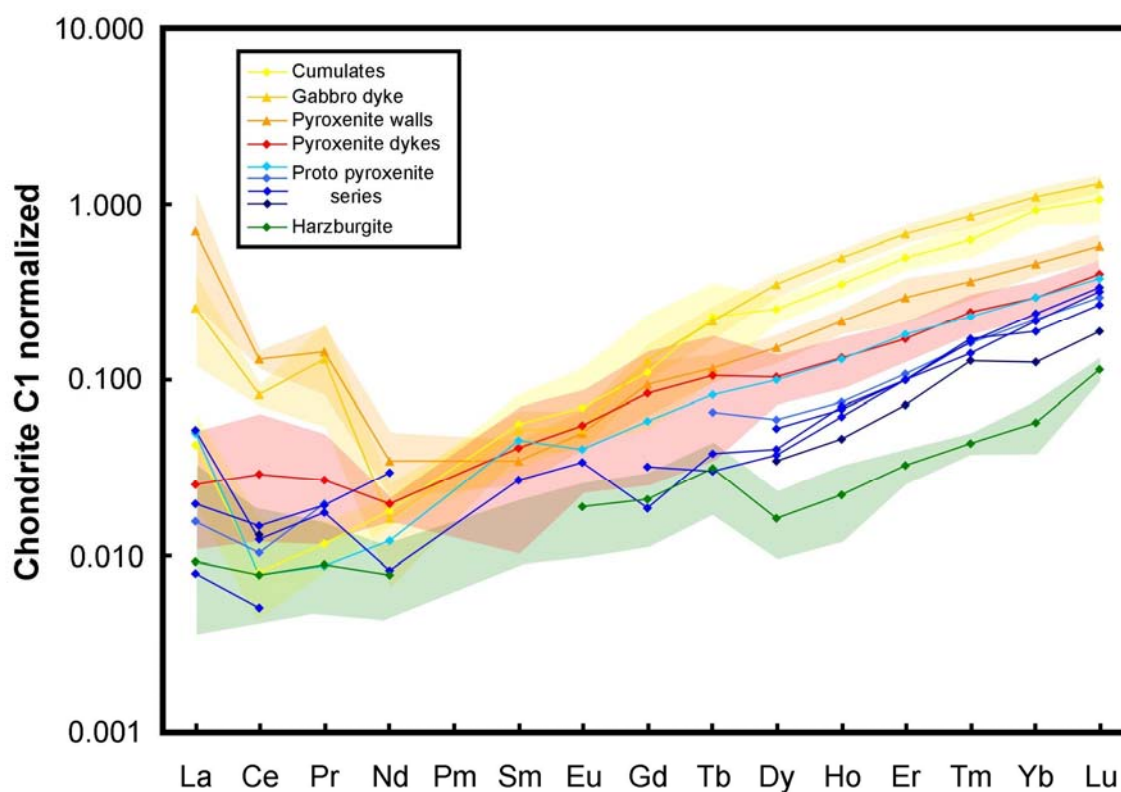


Fig.5.14. Rare earth elements spidergram of orthopyroxenes normalised on C1 chondrite (McDonough & Sun, 1995). The dark to light blue colours represent the increasing orthopyroxene modal content in proto-pyroxenites and pyroxenite dykes. Average values are given by the dots; 2σ standard deviation is given by respective shaded areas.

5.3.3.3. Clinopyroxene

Concentrations of trace elements in rocks with primary clinopyroxene seem to be partly related to the modal content of clinopyroxene in the samples. Early websterite and clinopyroxenite dykes as well as early cpx-dunite cumulate have a high content in REE. Gabbronorite sills have the lowest concentration of the cumulate series. Fine grained gabbronoritic dykes, on the other hand, contain the highest concentrations in REE and are the only rocks that display a negative Eu anomaly in the clinopyroxenes [Fig.5.19, Tab.S17].

Clinopyroxenes in the walls of the zoned gabbro dykes generally show a decreasing content of LILE, HFSE and HREE from the outside towards the contact with the gabbro. Only TM such as Co, Cr, Mn or Zn show an opposite trend. However, within the gabbro core, and particularly for the microgabbronorite, most

elements are enriched, in many cases, more than 10x the concentration of the adjacent clinopyroxenite [Fig.5.13, Tab.S17].

At a larger scale, a similar behaviour is observed in the crustal cumulates where early clinopyroxene in the cumulus dunite is more enriched than clinopyroxene in the gabbro norite, whereas fine grained gabbroic veins cutting the cumulate pile are again highly enriched in trace elements [Tab.S17].

5.3.3.4. Spinel

Trace elements in spinels seem less constrained by magma evolution. Spinel from dunite channels do not differ very much from chromite present in the harzburgite. In the proto-pyroxenites however, spinels are generally depleted in HFSE and TM in comparison to dunites. As orthopyroxene get more abundant in the channels, only a few elements such as vanadium, increase in the chromite. In the crustal cumulates, spinels have a similar pattern to the oxides present in the proto-pyroxenite although enrichment in Ti is noticeable.

5.4. Discussion

5.4.1. Sub-arc environment

In chapter 4, it has been shown that on the basis of the composition of the mineral phases, the Massif du Sud is related to a convergent margin. The extreme depletion observed in the harzburgitic mantle of New Caledonia, its range of values in the OSMA and the existence of boninitic magmatism support the idea that most of the ophiolite was formed in an arc environment.

With the basic consideration that the Massif du Sud ophiolite is formed in a convergent margin environment, several comparisons with other arc-related ophiolites show interesting similarities which are interpreted as typical features for ultramafic and mafic terranes of convergent margins. The crust-mantle transition of New Caledonia bears a very prominent ~600 m thick sequence of dunitic rocks previously described as partly cumulative and replacive [Chap.4.4.2.2, Fig.4.6].

Such large dunite zones are restricted to arc ophiolites such as Sapat, Kohistan palaeo-arc (~450 m, Bouilhol et al., 2009), Tonsina, Talkeetna palaeo-arc (~3 km, Burns, 1985), Darb Zubaydah, Saudi Arabia (~1 km, Quick, 1990) or Kamuikotan, Hokkaido (~500 m, Ishizuka, 1980). The specific textures and chemistry of these dunites are very similar in all sub-arc rock formations. An extensive debate also exists on the presence of an important pyroxenite cumulate zone in the lower crust of mature arc systems. Calculated thicknesses of such cumulates is estimated to be between 15 and 35 km (Kelemen et al., 2003; Garrido et al., 2007) but arc ophiolites lack such large amounts of cumulates, leading to the arc paradox (Garrido et al., 2006). Cumulus pyroxenes remain however a fairly typical lithology of the crust-mantle transition in supra-subduction zones (Jan & Howie, 1981; Burns, 1985; Jan & Windly, 1990; Garrido et al., 2007).

In New Caledonia, most of the cumulate section above the massive dunite is of wehrlitic or websteritic composition and the upper limit of these cumulates is unknown. Orthopyroxene is the first cumulus phase to appear in the olivine-chromite layers whereas clinopyroxene and plagioclase are generally more tardive providing a crystallisation sequence compatible with an arc environment (Pearce et al., 1984) [Chap.4.4.3.1]. Gabbro-norite sills are common in mature arc magmatism (Burg et al., 2006; Greene et al., 2006; Jagoutz et al., 2007). In addition, the composition of the mineral phases in the gabbro-norite provides constraints on the characteristic of the melts which is relatively oxidised and hydrous during its crystallisation [Fig.4.31, 4.32]. These features would be typical of cumulates crystallising from a melt with arc affinities and give additional evidence that the ophiolitic sequence of the southern part of New Caledonia is related to a sub-arc environment.

However, despite the fact that the Massif du Sud ophiolite shares some strong analogies with other sub-arc terranes, there is evidence that a full development to a mature arc as it is seen in Alaska (Greene et al., 2006) or Kohistan (Burg et al., 1998) was never reached in New Caledonia. The absence of intense arc magmatism in New Caledonia gives new insights on the early basaltic-dominated magmatic network of a nascent arc which is generally inaccessible in volcanic island arc and obliterated in mature palaeo-arc terranes.

5.4.2. Relationships between channels, dykes and cumulates

The well-preserved section of the Massif du Sud ophiolite provides a clear non-disrupted view of the magmatic processes occurring in the mantle of a nascent arc. The field study combined with mineral and geochemical analyses constrains focused melt transport in the sub-arc upper mantle from 3 to 4 km below the crust-mantle transition. The relationships between the different discordant features and the position within the mantle section are schematically shown in [Fig.5.2]

Dunite channels are a distinctive characteristic of the Massif du Sud ophiolite. The discordant and non-brittle features associated with these channels are easily recognisable in the field and are present from the deepest zone of the ophiolite to the crust-mantle transition. Modal compositions are mostly olivine with accessory spinels bearing specific textural features such as the grain shape and the persistence of chromite trails from the harzburgite through the channels [Fig.4.26]. Major element composition of olivine show little variations between channels and is related to the composition of the surrounding rocks [Fig.5.9]. Mg-numbers in olivine vary in a small range between dunite and surrounding harzburgite and are in agreement with the models of P.Kelemen (1990) to explain the replacive features of dunite channels [Fig.5.15]. Mg# and Cr# in chromite grains are identical to the surrounding harzburgite but a drift is observed for MgO (10 to 13.5 wt.%) and Cr₂O₃ (48 to 57 wt.%) between the deep dunite channels of Yaté and dunite channels near the dunite zone. The abundance and size of the channels are increasing at shallower depth but the general features remains essentially the same. Trace elements in olivine show slight variations for trivalent cations (Al, Cr, V) which are generally low in the deep dunite channels and increase by 10x in the shallower dunite channels and the dunite zone [Tab.S13].

In the uppermost harzburgite, dunite channels start to display minute crystals or small clusters of orthopyroxenes whereas olivine major and trace element geochemistry remains identical [Fig.5.9, 5.11]. Field observations show that this gradation from dunite channels s.s. to orthopyroxene-bearing channels starts to occur about 1 km below the crust-mantle transition [Fig.5.2]. In a few key continuous outcrops of proto-pyroxenites [Fig.5.4], it can be observed how orthopyroxene replaces olivine and how dunite channel transforms into an

orthopyroxenite channel. This transformation leave the Mg-number unchanged while it enriches the remaining olivine phase in compatible elements such as Ni, Co and Sc and depletes them in Ca, Al and Mn which are taken by the neo-formed pyroxenes [Fig.5.9, 5.11]. The dissolution of chromite grains in the latest stages of the proto-pyroxenite transition is likely to be the reason for the enrichment in Ti, Cr, Zr, Nb and V in the pyroxenes [Fig.5.12]. In the best outcrops, a couple of tens of metres are enough for a complete modification of the channel mode [Fig.5.4]. Despite the fundamental difference in the composition of the channels, the relationship with the surrounding rocks, the size and the growth of the minerals is analogous to the formation of dunite channels and indicate that orthopyroxenites are formed by a focused porous flow process in the upper sub-arc mantle. In many cases, the transition occurs over a wider section of mantle. Intermediary stages of replacement are conserved and dunite walls or dunite restites are a common feature in the New Caledonia ophiolite and in the literature (Prinzhofer et al., 1980; Varfalvy et al., 1997; Python & Ceuleneer, 2003; Savelieva et al., 2008) [Fig.5.6].

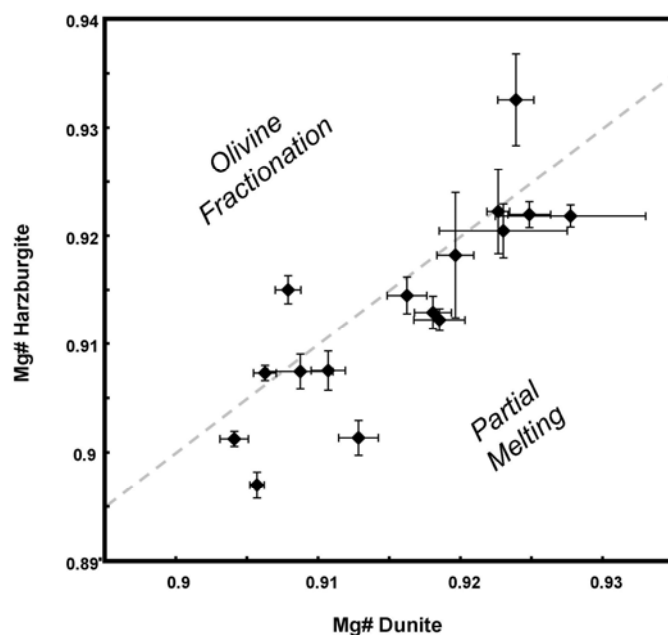


Fig.5.15. Mg-number in olivine of dunite channels in comparison to the olivine present in adjacent (less than 5m) harzburgite. Although Mg-number in the channels can be slightly higher than in the host-rock, the variations through the ophiolite give inconclusive data on a specific process of formation and the formation by metasomatic replacement (Kelemen, 1990) is the most likely mechanism. The grey line (1:1) represents identical Mg-number for dunite and harzburgite. Error bars are 1σ values for samples

The transition between orthopyroxenite channels which clearly display replacive features and the widespread pyroxenites dykes emplaced through fractures occurs in the uppermost kilometre of the harzburgite section. Most of these transitions are not seen at an outcrop scale and are visible as a general trend in the peridotite section [Fig.5.5, 5.7a]. However, in some cases, this transition is observed on the side of proto-pyroxenite channels where dykelets of

orthopyroxenite shoot off into the harzburgite with a clearly brittle rheological behaviour [Fig.5.4d]. These dykelets are identical to the widespread pyroxenite dykes and differentiate themselves from channels by their harrisitic growth, the absence of olivine and spinels and a geochemical signature with more evolved trace element patterns than the adjacent channels [Fig.5.14]. In contrast to the orthopyroxenite channels, fractional crystallisation within the dykes allows the precipitation of clinopyroxenes and an evolution from orthopyroxenite to websterite and clinopyroxenite compositions, associated with a decreasing Mg-number and increasing alumina content [Fig.5.7, 5.10]. Pyroxenite dykes are a very common feature in supra-subduction zone ophiolites and similar dykes have been considered as the pathways for melts feeding the crust (Varfalvy et al., 1997; Beccaluva et al., 2004; Berly et al., 2006).

The final step of melt transport within the mantle section is the appearance of plagioclase and the presence of gabbro-norite lenses in the core of some pyroxenite dykes. These veins range from simple gabbro-norite dykes to well-developed zoned dykes conserving gabbro-noritic melts in the core of a sequence of cumulates and reaction zones [Fig.5.8]. These composite dykes are an excellent case study to establish the link between pyroxenites observed in deeper parts of the mantle section and the gabbro-norite cumulates present in the crust-mantle transition zone. The outer part of the composite dykes, composed of dunite and orthopyroxenite, have similar geochemical features, notably REE patterns, to those of dunite and pyroxenite channels. The inner cumulate layers of websterite and clinopyroxenite surrounding the mafic dykes have compositions and textures identical to pyroxenites and other unzoned mafic dykes whereas the fine-grained gabbro-norite cores only share similarities with the mafic rocks from the crustal cumulate pile [Fig.5.14, 5.16, 5.19]. The inward sequence observed in the composite mafic dykes is analogous to the vertical evolution which can be seen in the discordant channels and dykes of the mantle section. Therefore, it is thought that the small scale zoned features surrounding mafic dykes establish the ontogeny for the pathways of basaltic melt through the upper mantle, evolving from dunite to orthopyroxenite channels and then precipitating pyroxenes within dykes from a hydrous basaltic melt represented by the microgabbro-norite [Fig.5.16].

The composition of the microgabbro norite dyke and the associated gabbro noritic cumulative walls show strong similarities with the mafic rocks present in the sills of the cumulate pile of the crustal section [Fig.5.19]. With the exception of the fine-grained mafic dykes, the crustal gabbro norite is the most differentiated rock of the Massif du Sud ophiolite. The close geochemical relationship that oceanic crust shares with the mafic dykes shows that gabbro sills are genetically directly related to the discordant rocks of the uppermost mantle (composite mafic and pyroxenite dykes) and thus to the dunite channels which are widespread in the deeper harzburgite [Fig.5.2].

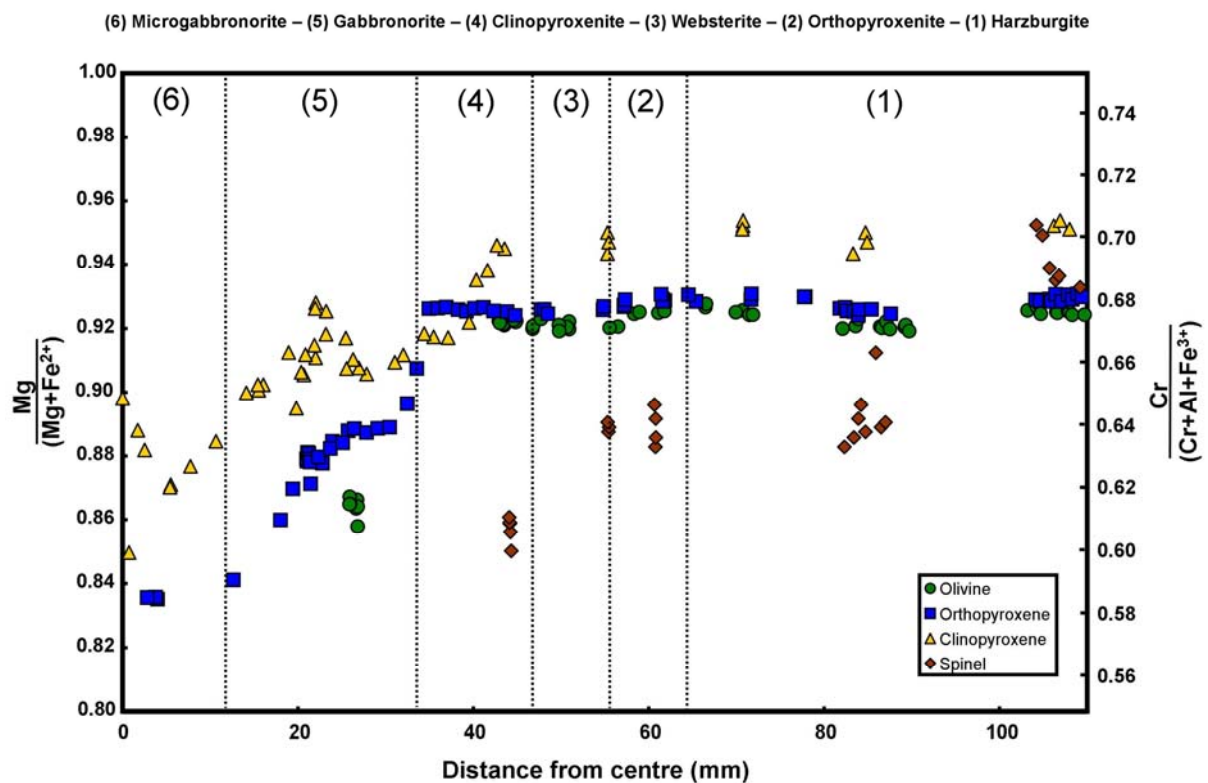


Fig.5.16. Mg-number in olivine (circle), orthopyroxene (square), clinopyroxene (triangle) and Cr-number in spinel (diamond) in a transect from host rock (1) to microgabbro norite core (6) of PRB49. Silicates show a very small decrease in Mg# between the harzburgite and the pyroxenite cumulates. Cr# of spinels is more sensitive in that zone and already decrease to <0.65 in the harzburgite. In the gabbro norite wall, values drop for the silicate and reach the lowest values in the microgabbro norite core. Zones are in relation with Fig.5.8b.

5.4.3. Evolution of channels and dykes in the sub-arc mantle

5.4.3.1. Dunite channels

The dunite channels occurring in New Caledonia share similar features to discordant dunite rocks in other ophiolites. This suggests that the sub-arc position does not create drastic changes in melt transport features and topological characteristics of the channels which are similar to MORB-related channels. During mantle anatexis, the accumulation of melt within the peridotite reaches a threshold at a few percent of partial melting where melts segregate from the solid phase into channels (Suhr, 1999; Godard et al., 2000; Morgan & Liang, 2003; Lambart et al., 2008) [Fig.5.2]. The focusing event is a consequence of the instability of the porous media and can initiate on heterogeneities present within the mantle (Daines & Kohlstedt, 1994; Morgan & Liang, 2003). Focused flow occurs in almost isothermal conditions in the mantle and allows the propagation of the initial channel without fractional crystallisation [Fig.5.15], permitting the silicate melt to rise in the peridotite, aggregating additional melt by lateral input. The rise of multi-saturated basaltic melts formed by peridotite melting in the mantle leads to decompression and changes the properties of the magma which becomes only saturated in olivine as a result of the expansion of the primary phase field of olivine [Fig.5.17]. Such melt has the ability to consume the surrounding peridotite by assimilating pyroxenes from the host rock, forming an evolved melt and leaving a metasomatic olivine residue (Kelemen, 1990; Kelemen et al., 1995). In these conditions, melt transport in channels is a sustainable process over long distances which requires minimum effort and restricts melt modification. The small amount of interactions between the melt percolating through the channels and the host rock explain the absence of equilibrium between the ultra-depleted New Caledonia harzburgite residue and the melts which are preserved in mafic dykes and the crustal cumulates. Crustal gabbros of the Massif du Sud are essentially formed by an assemblage of clinopyroxene and plagioclase without extreme depletion and cannot be considered as a direct product of harzburgite partial melting. Moreover, rare clinopyroxene present in uppermost dunite channels and trace element compositions of olivine [Fig.5.11] show that melts which used the channels as pathways are not extremely depleted and are likely to be produced in less

refractory zones of the sub-arc mantle. The disequilibrium which exists between adjacent rocks could partly explain the difficulties of previous workers to link the New Caledonia ophiolite with the basalts of Poya terrane despite the possibility of a parental link (Eissen et al., 1998; Ulrich et al., 2010) as the ultramafics rocks of the Massif du Sud cannot be considered as a residue of the oceanic crust.

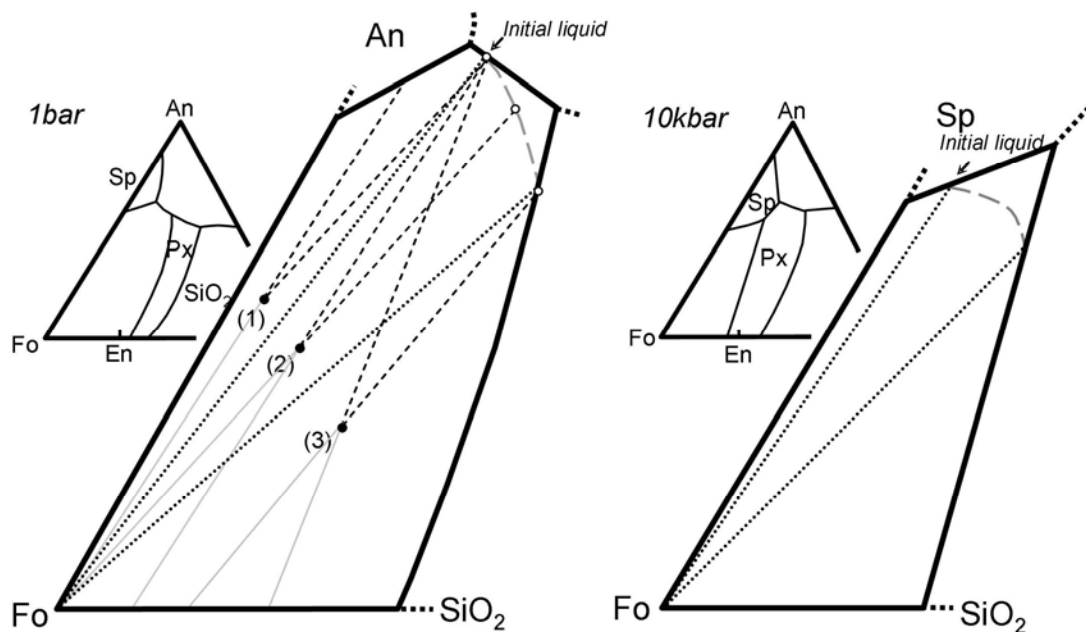


Fig.5.17. Phase diagram near the olivine apex in a forsterite-anorthite-quartz ternary diagram. (modified from Kelemen, 1990). The olivine primary phase field is expanding considerably between 10kbar and 1 bar and partially explains the mono-saturation in the olivine phase. The two dotted lines of the left diagram position the limit for different starting peridotites.

(1) Lherzolite will be affected by pyroxenes dissolution and crystallisation of olivine and a minor aluminous phase.

(2) Cpx-poor lherzolite will have pyroxene dissolution and an evolution of the liquid along an isotherm (grey dashed line) until the Fo-peridotite (2) tie line is reached where the melt composition is fixed despite further assimilation and olivine crystallisation.

(3) Harzburgite host rock will behave similarly to (2) until the Fo-En field boundary is reached and where the liquid composition is fixed.

Reaction (2) and (3) are compatible with the formation of large dunitic bodies with a non-picritic basalt and (3) can potentially produce low-Ca pyroxene by consumption of all other phases at low pressure whereas high pressure (anhydrous conditions) will cease the assimilation process.

Petrological models and simulations have shown that although olivine composition in the dunite channels remains relatively unchanged as a solid reaction product of the replacive process, melts which are going through the channels are becoming more and more abundant and show compositional

changes notably by an increase in silica content (Kelemen, 1990; Suhr et al., 2003) [Fig.5.17]. The dunite channels of New Caledonia show that the trace element contents of the olivine can be used to infer the origin of olivine-rich rocks. The increase of incompatible elements such as Ti, P [Fig.5.11] and Y in olivine of the dunite channels with respect to olivine in the surrounding harzburgite provides evidence for the interaction with a trace element-enriched agent such as a mafic melt. More enigmatic is the observed decrease in Li, which suggests that the passing melt was undersaturated in Li with respect to the host rock. In an arc environment, the melting of a depleted mantle will produce hydrous basaltic melts which will evolve within the channel. Such melt rising in the mantle wedge would be relatively different from standard tholeiitic basalt from mid-ocean ridges and would be closer to boninitic compositions [Fig.5.17]. The involvement of non-MORB melts for the formation of dunite channels has been proposed in some ophiolites as an explanation for the presence of pyroxenite lenses in dying channel networks (Tamura et al., 1999; Suhr et al., 2003; Kelemen et al., 2003).

5.4.3.2. Orthopyroxenite channels

The transport of large amounts of boninite-like melt through dunite channels will eventually reach the highly depleted part of the upper mantle. In the vicinity of the crust-mantle transition, the potential for olivine precipitation due to decompression is being exhausted and other factors start to dominate. In the New Caledonia mantle section, in the top 1 km of the ultramafic column, the conjunction of silica-enrichment with the start of cooling allow the melt to reach the forsterite-enstatite peritectic which initiate the precipitation of orthopyroxene while consuming olivine in the channels [Fig.5.17, 5.18]. Such a reverse reaction in the replacive channels was theoretically expected when evolved melts would reach the uppermost mantle and multi-saturation could re-occur (Kelemen, 1990; Kelemen et al., 1995; Suhr et al., 2003). Although multi-saturation event is known to be a theoretical end-member of the replacive process, it was never observed as such in dunite channels of other ophiolites and multi-saturation is only considered as the result of a cooling process (Muntener & Piccardo, 2003) or when melts are extracted to form large crustal magma chambers. The arc-related melt transport system of New Caledonia suggests that a depleted (non-Iherzolitic) mantle melting

in supra-subduction zone conditions will allow the saturation of orthopyroxene with respect to a MORB system, leading to the formation of proto-pyroxenite and orthopyroxenite channels. At a small scale of observations, the development of orthopyroxene replacement on the edge of the channels [Fig.5.4, 5.5, 5.6] suggests that temperature gradients have a role in the orthopyroxene saturation.

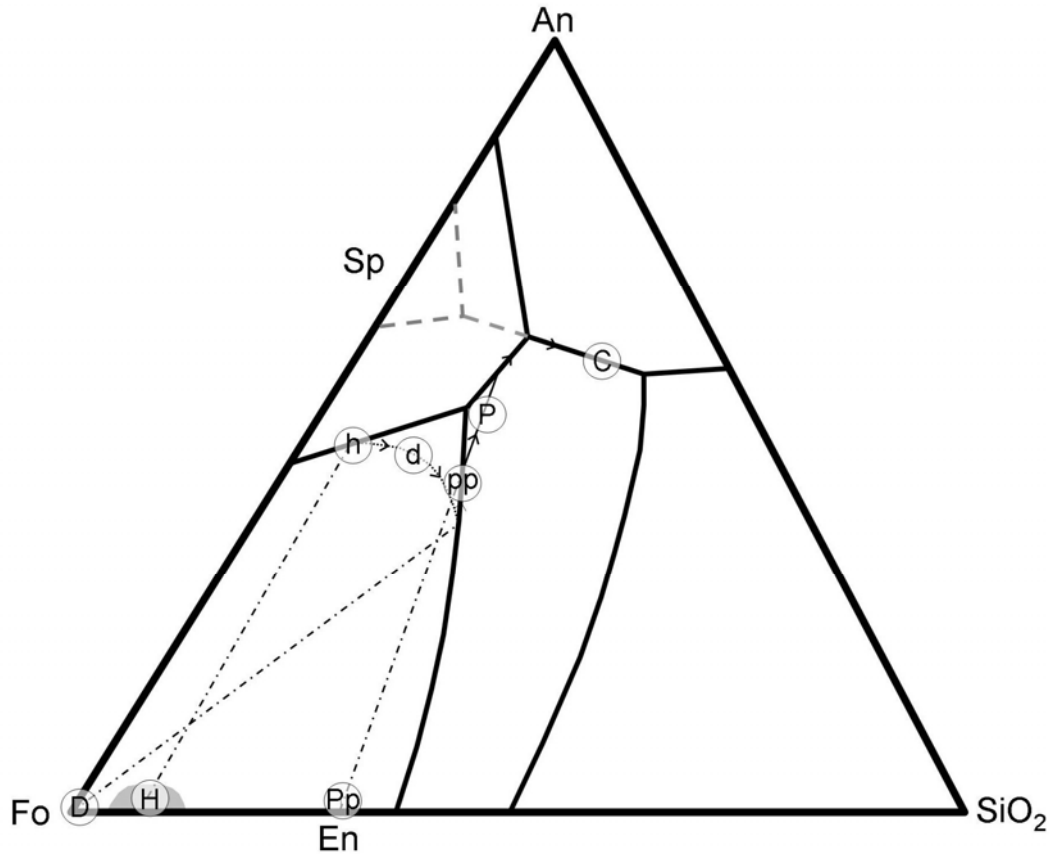


Fig.5.18. Ternary diagram Forsterite-Anorthite-Quartz showing the evolution of the melts during their ascent through channels and dykes.

D : dunite (channels), H : harzburgite (the grey field represents the composition in the Massif du Sud), Pp : Proto-pyroxenite / orthopyroxenite channels, P : Pyroxenite dykes, C : Mafic cumulates and dykes, h : melt composition at the time of the channel aggregation, d : melt evolution (arrows along the isotherm) in the dunite channels, pp : position of the melt during the proto-pyroxenite transition.

The initial melt (h) is formed by fractional melting of harzburgite (H). Melt-rock interaction modifies the melt (d) towards SiO₂ along an isotherm by consuming pyroxenes and precipitating olivine (D). When the Fo-En peritectic is reached (pp), enstatite precipitation starts by consumption of olivine (D→Pp) until the reaction is complete (Pp) or that the kinetic effect associated with lower temperatures allows the formation of dykes (P). The extraction of melt without olivine moves the liquid composition within the enstatite primary phase field where it will form pyroxenite dykes. Fractional crystallisation within that field and on the cotectic lines will allow the crystallisation of clinopyroxene and plagioclase (C) and form the core of composite dykes.

In addition, cryptic geochemical heterogeneities are known to occur in dunite channels (Godard et al., 2000; Suhr et al., 2003; Morgan et al., 2008; Lambart et al., 2008) and are interpreted as a lateral contribution of small amount

of melt produced by the partial melting of the harzburgite. Such melts are orthopyroxene-saturated and could enhance the saturation in pyroxenes on the edge of the dunite channels.

Pyroxenite veins geometrically related to dunite channels (dunite walls) have been described in many supra-subduction zone ophiolites. Diverse explanations are given for these observations such as an exhaustion of phases other than olivine in the walls surrounding the dykes (Prinzhofer et al., 1980); porous reaction around melt-filled fractures (Kelemen et al., 1995; Tamura et al., 1999; Python & Ceuleneer, 2003); late evolved melt using pathways made by dunite channels (Suhr et al., 2003; Morgan et al., 2008); local accumulation of melts within focused porous flow (Muntener and Piccardo, 2003) or an evolution and discrete replacement of dunite channels (Varfalvy et al., 1997; Suhr et al., 2003; Savelieva et al., 2008). In New Caledonia, we have a comprehensive set of field, petrological and geochemical evidences that pyroxenite veins are formed by an evolution of the melt percolating through the dunite channels, leading to the formation of orthopyroxenite channels [Fig.5.6, 5.18].

5.4.3.3. Pyroxenite dykes

As pyroxenite channels occur in an area close to the crust-mantle transition, the rheological behaviour of the surrounding rocks in a young arc system are different and could display more brittle features than under a mid-ocean ridge (Sparks & Parmentier, 1991; Spiegelman, 1993; VanderWal & Bodinier, 1996). The transition from pyroxenite channels to dykes occurs in such conditions where melt is extracted from the channel to crystallise directly into fractures. Natural evidence (Muntener & Piccardo, 2003), experiments (Lambart et al., 2008) and models (Rabinowicz & Ceuleneer, 2005) of focused porous flow in a cooling, low pressure and less plastic mantle, show that the mechanism forming replacive channels cannot be sustained indefinitely. Melt-solid segregation occurs in those conditions, causing hydraulic fractures and the apparition of dykes. Some rare cases show minute grains of low-Ni low-Mg# olivine which suggest that extraction can occasionally occur before the proto-pyroxenite stage however, in most dykes, the extraction of melts standing already on the peritectic line between olivine and enstatite, as they do in proto-pyroxenite, will crystallise only enstatite [Fig.5.18]. As

these melts remain in the orthopyroxene primary phase field, it will form orthopyroxenite until the melt reaches the two pyroxenes cotectic line, precipitating also clinopyroxene. The abundance of pyroxenite veins (with or without dunite) in many arc-related ophiolites (Prinzhofer et al., 1980; Kelemen et al., 1995; Varfalvy et al., 1997; Python & Ceuleneer, 2003; Suhr et al., 2003) and the continuous mineral chemistry evolution observed in New Caledonia support this interpretation.

When the ascending melt is segregated from the solid phase by hydraulic fracturing, interactions with solids are minimised and only cumulus phases in equilibrium with the melt, are crystallised in the dykes. The liquid composition evolves through a standard liquid line of descent, subsequently forming websterite and clinopyroxenite dykes. As cumulus clinopyroxene is strongly enriched in chromium, its precipitation tends to inhibit the formation of Cr-spinels (Feig et al., 2006) which are absent within most pyroxenite dykes.

The crystallisation sequence observed in New Caledonia is highly dependant on several factors related to the arc environment of the ophiolite. Initially, the formation of dunite channels is behaving similarly to mid-ocean ridge and produce olivine-saturated melts. However, as the melt, enriched in silica, reaches the upper part of the mantle, only a low pressure environment (<2.3 kbar (Boyd et al., 1964; Chen & Presnall, 1975) could allow peritectical reaction leading to the consumption of olivine by enstatite. At higher pressures such as those encountered in the Massif du Sud [Chap.6.4.1], the forsterite-enstatite tie line would cause saturation in both minerals and assimilation/precipitation would cease (Kelemen, 1990) [Fig.5.17]. The demonstration that crustal gabbros of New Caledonia are formed by relatively aluminium-rich hydrous basaltic melts [Chap.4.4.3.1] explains that the peritectic behaviour of enstatite is kept at higher pressures (Kushiro et al., 1968b; Taylor, 1973; Presnall et al., 1979). In such conditions, the reaction $\text{Olv} + \text{melt}_1 \rightarrow \text{Opx} + \text{melt}_2$ is conserved and applicable to pressures up to 15 kbar which covers the entire pressure range of the Massif du Sud ophiolite. The transfer of hydrous basaltic melt provided by the arc environment through a focused porous flow network is subject to an evolution as a function of focussing factors (Michael & Chase, 1987; Lambart et al., 2008) and replacive process which lead to a silica and water enrichment. The short range over which this transition occurs (sometimes less than twenty meters, [Fig.5.4] and observations of cyclic proto-pyroxenite precipitation [Fig.5.6] within channels give

additional support that the water content in the ascending magma is an important component which has a strong effect on the crystallisation of orthopyroxene (Tamura et al., 1999; Müntener et al., 2001).

5.4.3.4. Gabbro dykes

The final stage of fractional crystallisation within the dykes is well observed in the composite dykes and is characterised by the appearance of plagioclase. Mineral trace element concentrations in the different zones of the dykes show that the coarse grained gabbro norite has features similar to the pyroxenite walls. This suggests that the gabbro norite is also a cumulate where plagioclase is present [Fig. 5.13]. On the other hand, microgabbro norite cores and some gabbro norite dykes without wall interactions have geochemical features of pristine frozen melts with more enriched trace element contents. The size of the reaction zone (pyroxenites) surrounding the gabbro norite seems to be related to the thermal state of the surrounding rocks. Reaction zones are more likely to occur in the harzburgite and in the lower dunite zone and will be less common in the cumulate pile where sharp homogeneous microgabbro norite dykes have been observed [Fig. 5.8c,d]. This is in agreement with observations in the uppermost mantle and the transition zone of other ophiolites (Prinzhofer et al., 1980; Kelemen et al., 1995; Bédard & Hébert, 1998; Savelieva et al., 2008) where mafic rocks are described as distinct homogeneous dykes (Python & Ceuleneer, 2003) or as lenses in ultramafic veins (Lago et al., 1982; Suhr, 1999; Savelieva et al., 2008). Crystallisation sequence and textures in the gabbro norite dykes give further confirmation that a certain amount of water is present in these melts as plagioclase is clearly a late crystallising phase. Moreover, the presence of stable orthopyroxene in these hydrous conditions would require a high oxygen fugacity (Grove & Juster, 1984; Feig et al., 2006) which is confirmed by the $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio of the rare spinel grains.

Oxides are uncommon in the gabbro dykes and the cumulate pile which was sampled in the Massif du Sud. However, the modification in Ti and V compared to Zr suggest that Ti-rich spinel saturation occurs during fractional crystallisation [Fig. 5.12]. The evolution of HFSE in the layered gabbros and dykes is consistent with the assumption that phases such as apatite, ilmenite and zircons

are relatively rare in the lower oceanic crust (Coogan et al., 2000) or within the gabbro dykes below the Moho. The decreasing trends in Ti/Zr and V/Ti are used as a geochemical tool in arc gabbro to identify the presence of cumulus ilmenite (Pearce & Norry, 1979; Nielsen et al., 1994). In New Caledonia, in the absence of zircon saturation, the evolution of these trends is mainly related to the crystallisation of Ti-rich spinels. In addition, as zirconium content is exclusively controlled by fractionation, the absence of increasing Zr/Sm in the New Caledonia cumulates provides additional confirmation that amphibole did not appear as a cumulus phase (Thirlwall et al., 1994).

5.4.4. Morphology of the plumbing system under the arc crust

The full understanding of the processes occurring in the sub-arc mantle cannot be inferred from erupted volcanic product as melt interaction and crustal evolution strongly modify primitive arc basalts (Leeman, 1983). Juvenile arc systems are rarely dissected enough to sample their roots and mature arcs are affected by intense felsic magmatism which obliterates the previous proto-arc structure (Treloar et al., 1996; Greene et al., 2006). In the Massif du Sud ophiolite, the regional and geochemical features show that this arc-related terrane provides new range of information on the morphology and the potential evolution of the plumbing system under a nascent oceanic arc.

The study of oceanic spreading ridges shows that from depth, basaltic melts aggregate into dunite channels (Daines & Kohlstedt, 1994). In an arc environment, some authors (Hopson et al., 2008) have suggested that such focused processes do not occur as melt production is occurring at a shallower level. However, there is ample evidence that focused channel transport is a common feature of supra-subduction zones as it is seen in New Caledonia and other ophiolites with similar settings (Quick, 1981; Edwards, 1995; Tamura et al., 1999). In the uppermost part of the mantle lying right under ridges, the heat flow is the highest and dunite channels are known to coalesce, forming pods and finally a dunite transition zone where melts of the different channels are fully aggregated without further modification (Godard et al., 2000; Spiegelman & Kelemen, 2003). In a sub-arc environment, such coalescence is also occurring although extremely large channels have not been observed in New Caledonia. Very thick dunite zones are a

characteristic of fore-arc and proto-arc ophiolitic sections (Nicolas & Prinzhofer, 1983; Kelemen et al., 1995; Bédard et al., 1998; Braun & Kelemen, 2002; Yumul, 2004; Savelieva et al., 2008) and melt transport models (Sparks & Parmentier, 1991; Vander Wal & Bodinier, 1996; Spiegelman & Kelemen, 2003) consider such structures as an accumulation of melts in the sub-Moho region by the presence of a permeability barrier to the ascension of the melt in the earliest stage of arc formation [Fig.5.20].

The dunite zone has been described in detail in chapter 4. In the past, it has been successively interpreted as entirely cumulus (Prinzhofer et al., 1980; Prinzhofer & Allègre, 1985; Cluzel et al., 2001) or fully replacive (Marchesi et al., 2009). Our data show that the lower part of the dunite zone is replacive whereas the upper part has textural and geochemical features in agreement with a cumulative origin. The cryptic differences between both types of dunites seem to be a characteristic of arc systems as the absence of clear delimitation between replacive (or residual) dunite and olivine cumulate is mentioned in the Talkeetna palaeo-arc (Debari & Coleman, 1989; Debari & Sleep, 1991; Greene et al., 2006), the Kamuikotan ophiolite (Tamura et al., 1999) and the Zambales ophiolite (Yumul, 2004).

Although dunite channels s.s. show little variation in relation with the composition of the melt carried, the evolution of the channels when fractional crystallisation occurs can differ considerably. In a water-poor system, such as MORB and back-arc basin basalts, the growth of off-axis dykes typically follow troctolitic series (Kelemen et al., 1995; Python & Ceuleneer, 2003; Rabinowicz & Ceuleneer, 2005) whereas hydrous basaltic magmas tend to follow a sequence of crystallisation which involves early pyroxene saturation. As the silicate melts percolating through the New Caledonia mantle are hydrous [Chap.4.4.3] the formation of troctolite or anorthosite dykes is very rare and pyroxenite dykes dominate the ophiolite section. Because the strongly depleted surrounding mantle has little influence on the composition of the melt conduits (Varfalvy et al., 1997; Beccaluva et al., 2004; Berly et al., 2006), pyroxenite dykes in SSZ ophiolites are generally considered as exotic in comparison to their host rock (Edwards, 1995; Rabinowicz & Ceuleneer, 2005). Strong linkages between dykes and the surrounding mantle are only found in rare cases where pyroxenite cross-cut a metasomatised peridotite as it is the case in Finero, Italy (Zanetti et al., 1999).

In this process of focused transport, the development of crystallisation sequences in the mantle only starts when melt propagation switches from channelized porous flow to dykes. In a mid-ocean ridge setting, hydraulic fracturing can only occur outside the ridge axis, where the thermal gradient modifies the rheological properties of the ultramafic rocks (Dantas et al., 2007; Lambart et al., 2008). Dykes are formed but only carry a marginal amount of magma in the oceanic crust. The thermal structure under an arc is considerably different and a pre-arc crust is likely to have a brittle behaviour at deeper levels than the crust-mantle transition (Sparks & Parmentier, 1991; Spiegelman, 1993; Vander Wal & Bodinier, 1996). Therefore, melt evolution can occur at deeper levels and as a function of the rheological characteristics, melts can be tapped out of the channels at relatively deep levels in the uppermost mantle. These melts ultimately end up in the crust-mantle transition zone where they are mixed to form large mafic sills which contribute to the formation of the arc lower crust. At a regional scale, it is possible that the contributions of the dykes and the channels in the formation of magma chambers varies from place to place, leading to some heterogeneity between melt ponds. In the Massif du Sud, bulk rock differences between gabbro sills of different crustal sequences (Montagne des Sources & Prony) have been described (Marchesi et al., 2009). These features could be the result of an incomplete mixing or local variations in the melt transport where mantle contribution or fractional crystallisation would be more important [Fig.5.20].

5.4.5. Implications of melt-mantle interactions and fractionation in the sub-arc mantle

New Caledonia ophiolite magmatism is dominated by melts which have an ultra-depleted low-Ti composition which show boninitic and arc tholeiite affinities (Marchesi et al., 2009). Pyroxenite dykes, gabbro dykes and crustal cumulates contain clinopyroxene which allows the calculation of parental melt composition on the basis of REE pattern (Hart & Dunn, 1993). Melts which formed the mafic rocks show strong similarities with arc boninites of the Kohistan arc [Fig.5.19]. Such low-Ti, high-Mg boninite-like melts, carried by focused flow, are known to be typical of a proto-arc stage when spontaneous subduction is initiated 5 to 10 millions years before reaching a mature size (Treloar et al., 1996; Burg et al., 1998; Dhuime et

al., 2007) and have recently been observed in the present day Tonga island arc (Rubin et al., 2009). The hydrous and silica-rich composition of the boninitic melts allows the early precipitation of orthopyroxenes at pressures equivalent to the crust-mantle transition of arcs (Kelemen et al., 1995). In the Chilas (Burg et al., 1998), Ladakh (Dhuime et al., 2007) and Bay of Islands (Varfalvy et al., 1997) ophiolites, pyroxenite dykes with boninitic affinities are known in what is assumed to be a fore-arc or proto-arc crust [Fig.5.20].

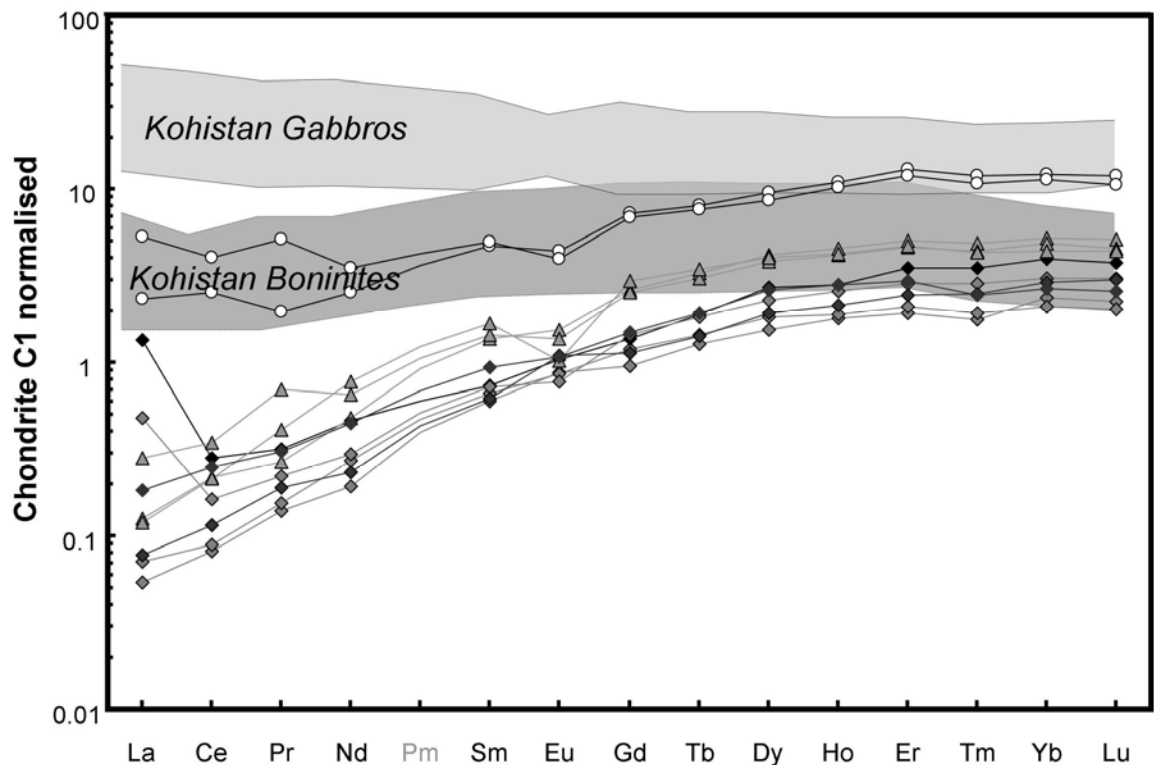


Fig.5.19. Rare earth element spidergram of clinopyroxenes normalised on C1 chondrite. (McDonough & Sun, 1995). Filled symbols are crustal cumulates (diamonds), darker are olivine-rich, light colours are cumulate gabbro. Filled triangles with higher concentrations are microgabbro dykes. Open circles are calculated melts (Hart & Dunn, 1993) for microgabbro. The dark shaded area represents the field covered by Kohistan boninites (Garrido et al., 2007) and lighter shade is the Kohistan gabbros (Garrido et al., 2006).

This study shows that in the initial stages of arc crust formation in the Massif du Sud, melts are percolating through dunites which were previously formed and emplaced at the bottom of the oceanic crust. These liquids can evolve and precipitate orthopyroxene by the consumption of relict olivine in the channels and later on, directly precipitate pyroxenes from the ascending melt. Therefore, an evolved melt, not primitive arc basalt, is feeding the proto-arc crust. When these melt-mantle interactions are transferred to a full-grown arc system,

the peritectic reaction of orthopyroxene formation within the mantle could give critical information in solving the arc paradox. In a mature arc system, the evolved composition of the upper arc crust cannot be explained if primitive arc basalts do not undergo extensive intracrustal fractionation of ultramafic cumulates. This process should leave a complementary ultramafic keel under island arcs, extending a few kilometres below the arc (Garrido et al., 2007). The growth of large zones of ultramafic cumulates of pyroxenite are modelled by two opposite processes. The crystallisation of one batch of melt involving pure pyroxenite cumulates has been supported by experimental works (Debari & Coleman, 1989; Treloar et al., 1996; Müntener et al., 2001; Greene et al., 2006) whereas melt-rock interactions where pyroxenites are formed by replacement of the uppermost residual mantle has been proposed in Kohistan (Garrido et al., 2007; Dhuime et al., 2007). As such massive cumulates are not seen in arc sections (Greene et al., 2006; Garrido et al., 2007), this leads to an observational paradox regarding to the general petrological understanding of the arc system.

In New Caledonia, we have evidence that melt-mantle interaction and fractionation occur prior to the injection of magma into lower crustal magma chambers. These processes lead to the formation of pyroxenite in the uppermost mantle or at the transition level. Although the pyroxenites of New Caledonia are quantitatively insufficient to explain the arc paradox, an extrapolation to a mature arc system would produce an important amount of pyroxenite dykes and channels within the uppermost mantle leading to the formation of a keel by replacement and not by cumulative processes [Fig.5.20]. The replacement process could also occur in the thick dunite zone formed in the infant stage of the arc where large pods of pyroxenite could be developed (Burg et al., 1998). In New Caledonia, such pyroxenite layers and pods are scarce but the impregnation of the cumulus dunite by gabbro-noritic melts to form wehrlite, websterite and clinopyroxenite is indicative that such processes exists [Chap.4.4.1.3] next to the presence of pyroxenes cumulates [Fig.4.10, 4.11].

The residual melt produced by these reactions is not entirely primitive and will be enriched in H₂O and aluminium. A comparison with mature arcs shows that in Talkeetna palaeo-arc, such enrichment is occurring at the base of the arc crust (Debari & Sleep, 1991). The confirmation that the ultramafic unit and the crustal gabbros of Kohistan are not the result of a single fractionation stage (Garrido et

al., 2007) is also supporting the idea of a replacive ultramafic keel and more standard fractionating processes within the crust forming amph-grt gabbros. The nascent arc crust of New Caledonia has an estimated thickness of 12 to 15km [Chap.6.4.1] which is in the range of values found for young arc such as Tonga (Gill, 1981). In an aging arc, increasing thickness and lower temperatures allow some changes in the crystallisation of an Al-rich hydrous magma. Amphiboles which are absent from gabbros in the Massif du Sud, are more likely to appear and can precipitate massively as it is the case in Kohistan (Garrido et al., 2006) and Talkeetna (Debari & Sleep, 1991). Higher pressures will also allow primitive magmas to saturate magma chambers in garnet and could also be a product of replacive processes in the crust-mantle transition. In the Kohistan arc, such garnet-bearing gabbros have been described as an important component of the lower crust (Müntener et al., 2001; Garrido et al., 2006) [Fig.5.20].

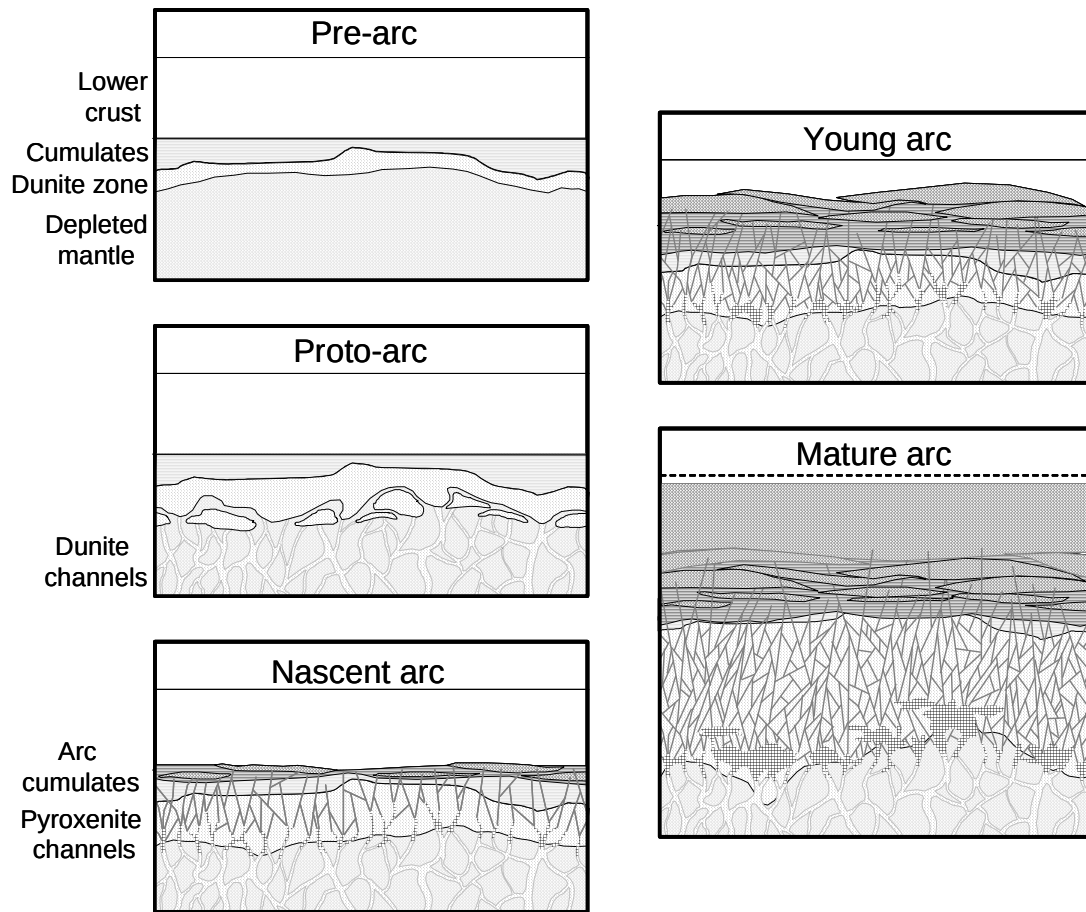


Fig.5.20. Development of an arc in the crust-mantle transition zone.

Pre-arc stage is the oceanic crust present prior to the emplacement next to a convergent margin. The structure of the crust is a function of the previous history but contains a mafic lower crust, some ultramafic cumulates, a small dunite zone and a depleted mantle (cpx-poor lherzolite or harzburgite). The mantle contains sub-concordant dunite channels and pyroxenites from the previous magmatic history (not represented).

Proto-arc stage shows the initiation of subduction and the percolation of the first melts transferred through dunite channels forming dunite pods below the transition zone and thickening the dunite zone. This stage can be analogue to Darb-al Zubaydah (Quick, 1990), the Sapat ophiolite in Kohistan (Jan et al., 1993; Bouilhol et al., 2008) and probably some part of the Kamuikotan belt in Hokkaido whereas some other section are more evolved (Takashima et al., 2002).

Nascent arc stage results in the appearance of replacive orthopyroxenite in the dunite channels and the transfer of melts through pyroxenite and mafic dykes in the dunite zone. This stage is reached in New Caledonia, Bay of Islands and Voykar Sin'ya (Edwards, 1995; Varfalvy et al., 1997; Suhr et al., 2003; Savelieva et al., 2008).

Young arc stage has further development of the orthopyroxenite replacement in the dunite zone and an increasing amount of pyroxenite dykes feeding the lower crust in large amount of melts forming gabbro intrusions. This early stage is partially conserved in the Jilal complex in Kohistan (Treloar et al., 1996) and in the Ladakh ophiolites (Mahéo et al., 2004). The Tonsina complex in Alaska seems transitional to a mature arc (Debari & Sleep, 1991).

Mature arc stage display an important replacement of the dunite zone by the orthopyroxenite (lower part) and by the large amount of pyroxenites dykes feeding the arc crust. The crust itself is formed of lower pyroxenite $\pm$ garnet cumulates, gabbro sills and garnet gabbros which form most of the lower arc crust. This mature stage is reached in Talkeetna (Greene et al., 2006) and in the Chilas complex of the Kohistan arc (Burg et al., 1998; Garrido et al., 2007).

5.5. Conclusions

A wide range of discordant features have been studied in the Massif du Sud ophiolite through field work and major and trace elements analyses. Direct evidence of a link between dunite channels, replacive pyroxenites, pyroxenites and gabbro dykes and finally the lower crustal gabbros has been found. The relationship between these discordant rocks provides new insights on the transfer and the evolution of melts in a sub-arc mantle.

The trace element chemistry and the evolution of dunite channels show that replacive conduits can carry silicate melts different in composition from normal MORB. The melting of the mantle wedge in a supra-subduction environment produces hydrous boninitic melts with a strong arc affinity. These melts become easily saturated in orthopyroxene when they evolve and initiate a reverse reaction where enstatite replaces forsterite leading to the formation of pyroxenite channels.

In the crust-mantle transition and the uppermost mantle, rheological conditions cannot sustain the presence of focused porous flow. Channels are segregated between solid and melts and the latter forms dykes cross-cutting the ultramafic rocks. These dykes go through fractional crystallisation and carry a melt identical to the crustal gabbro norite sills. The precipitation of large amount of pyroxenes at sub-Moho levels and the evolution of melts prior to the incorporation to the crust provides new views on the arc paradox. Primitive arc basalts which were once thought to be the parental melts for fractional crystallisation within the crust are here considerably modified when they enter and feed the arc crust. The extrapolation of pyroxene replacement and pyroxene precipitation reactions to a mature arc system would indicate that large pyroxenite cumulates in crustal magma chambers are not the only process capable of explaining the arc paradox.

6. Evidence of a fore-arc magmatic event in the New Caledonia ophiolite

6.1. Introduction

The oceanic environment of a typical ophiolite prior to its obduction is commonly referred to a SSZ setting and more particularly a fore-arc. This geodynamic position is generally characterised by an intense hydrothermal serpentinisation (Fryer et al., 1985) and more rarely by the presence of magmatic intrusions emplaced at different levels in the lithosphere.

Fore-arc magmatism is mainly known from xenoliths dragged from the ocean floor above subduction zones. It is also identified in a few ophiolites where late magmatic features are interpreted as events occurring in a fore-arc position. However, the identification of this type of igneous activity is mainly based on the interpretation of geochemical data obtained on dredged samples or xenoliths for which there is little information on geometric relationships with the country rocks. To the best of my knowledge no direct field observations coupled with detailed petrology of magmatic rocks occurring in fore-arc lower crust have been described.

Here we give a preliminary interpretation of a series of mafic to felsic rocks observed in the mantle section or crust-mantle boundary of the New Caledonia ophiolite. The presence of a large variety of felsic rocks (trondhjemites, granodiorites, diorites) in an ultramafic environment has been described in some rare ophiolite settings. However, previous studies on the Massif du Sud have shown that these felsic plutons are intruded prior to the obduction of the oceanic crust on the New Caledonian continental crust. The felsic magmatism present in the Massif du Sud has to be related to a fore-arc volcanic activity and provides a unique case study to investigate the emplacement of magmas in such an environment. The large size of the Massif du Sud ophiolite gives an unparalleled opportunity to characterise felsic plutons in the upper part of the mantle with a study of lateral variations and in some places, insights are given on the melt transport process which leads to the accumulation of slab melts in the fore-arc

lithosphere. The large variety of rocks which are preserved in the fore-arc setting of the Massif du Sud ophiolite reflects the wide range of processes and interactions which occur between felsic melts and ultramafic rocks. The accessibility to rock outcrops which are representative of lateral and depth variations is of considerable importance to understand the transfer of melts in a sub-arc mantle.

Finally, the relationship of this series of rocks with other magmatic features in the mantle section of New Caledonia and other terranes provides important information on the geodynamic history of the southwest Pacific. The ability to constrain ages for these rocks allows the history of New Caledonia at the beginning of the Cenozoic era to be deciphered.

6.2. Magmatism in a fore-arc environment

The fore-arc is classically considered as a convergent margin environment with relatively minimal magmatic activity. Several features such as serpentinite seamounts and rare volcanic products are known in such systems and thought to reflect the initial dehydration of the subducting oceanic plate as fluids present in the slab are released in the coldest part of the mantle wedge (Fryer et al., 1985; Fryer & Fryer, 1987).

However, when the fore-arc oceanic crust goes into extension small amounts of new basaltic melt can be created. Boninitic magmatism has been described from several ophiolites as the result of fore-arc extensional activity (Varfalvy et al., 1997; Burg et al., 1998; Dhuime et al., 2007, Pagé et al., 2009). This record is however scarce and is generally limited to dykes of pyroxenites where melt compositions are inferred from pyroxene-melt partitioning.

Felsic intrusions such as veins are generally related to late magmatism but in some xenoliths, veins containing felsic melts have been described as contemporaneous with an arc setting (Kepezhinskis et al., 1995; Arai et al., 2003). A few examples of large felsic intrusions within ophiolites are also known but in most cases, these plutons are post-obductional or collisional and are not representative of a true fore-arc magmatism (Whitehead et al., 2000; Saric et al., 2009; Tani et al., 2010).

6.3. Results

6.3.1. Field observations

In the Massif du Sud, a large number of discordant rocks occur in the harzburgite. Although many of these were discussed in the chapter 5 as sub-concordant early dykes or as an evolution of dunite channels to dykes, a last group of intrusive bodies is represented by highly discordant features which show relationships of late emplacement in the ophiolite. This series of rocks is represented by intrusions of granitoids and intermediate rocks, amphibole-rich dykes and pods and pyroxenite dykes [Fig.6.1, Tab.S11].

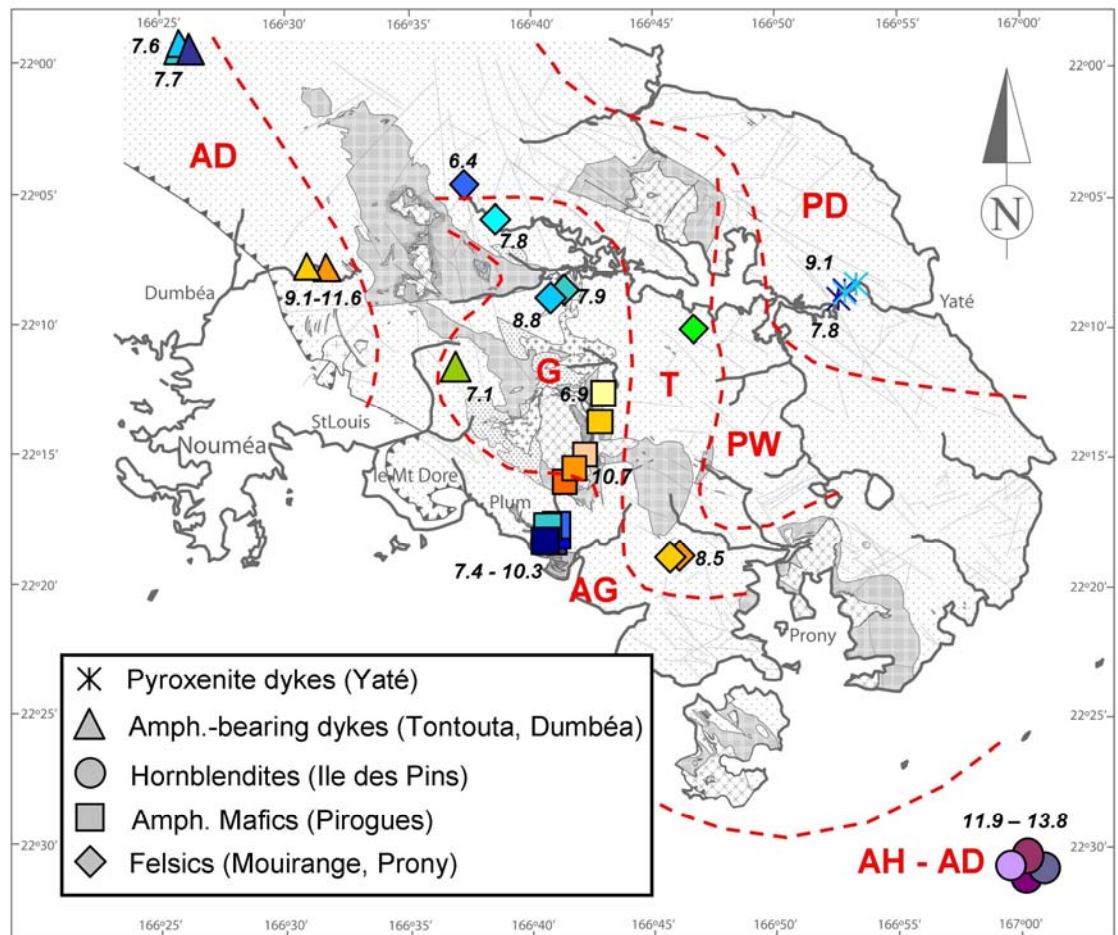


Fig.6.1. Map of the southern part of the Massif du Sud with the location and symbols of the different samples used in the plots. The red line represents the approximate extension of zones where dominant lithologies were found or start to appear. PD: Pyroxenite dykes, PW: Pyroxenite walls with felsic core, T : Trondhjemite pods dominate, G : Granitoids dominate, AD : zone rich in amphibole-bearing dykes, AG : Amphibole mafic and amphibole-rich ultramafic rocks, AH : amphibole-rich harzburgites and gabbros in Ile des Pins. Numbers next to symbols indicate Al_2O_3 -contents in amphibole of felsic or intermediate intrusions and can provide information on the pressure of emplacement for these rocks (Schmidt, 1992).

6.3.1.1. Granitoids

Granitoid intrusions are the only abundant lithology in this series of fore-arc rocks. They occur in an area between Rivière des Pirogues and Montagne des Sources and form a mappable unit ($\sim 35\text{km}^2$) in the geological map of the Massif du Sud [Fig.4.4]. These felsic intrusions are sometimes present surrounding the gabbro massifs, but are mostly emplaced at the limit between harzburgite and dunite. Field observations show that these granitoids form large hectometric intrusions. According to geological maps, intrusions surrounding Rivière des Pirogues can reach kilometre size plutons and are gathered under the name of Mouirange Pluton (Rodgers, 1973a). In this study, a clear distinction is made between the Mouirange amphibole granitoids and the bojites (amphibole-rich gabbros) which are developed in a following paragraph. Such distinction was not always done in previous studies and led to confusion between the intermediate-mafic rocks and their ultramafic host (Lacroix, 1942; Guillon, 1969; Guillon & Routhier, 1971; Rodgers, 1973a, Maurizot & Vendé-Leclerc, 2009; Maurizot & Collot, 2009).

The granitoids which were collected fall into two main modal compositions of amphibole meladiorite and amphibole quartz-leucodiorite. The contact between these two rocks shows a mingling [Fig.6.2d,e], which is observed over a few meters. K.Rodgers (1973a), however, describes a continuous trend from meladiorite to granodiorite from the outside of the intrusion to the inner part which was not observed during this field work. Our granitoids have an hypautomorphic granular texture with orientated prisms of amphibole (25-35%) and plagioclase (55-65%) parallel to the slivers of mingling textures suggesting that it is indicative of flow rather than accumulation [Fig.6.2a,b,d]. It contains minor amounts of quartz (<10%) and ilmenite and zircon are present as common accessory phases. The dioritic rocks described by Rodgers (1973a) are slightly more mafic than our samples with 45% amphibole and 50% plagioclase. Rodgers (1973a) also established modal variation transect through the Mouirange pluton which show that most of the rocks sampled in this thesis have been collected in the transition zone between diorites and pure granodiorite. The core of the intrusion labelled by Rodgers (1973a) as a granodiorite is in fact a quartz-monzodiorite (Streckeisen,

1976) which contains accessory amphibole and 80% plagioclase, the remainder being quartz (10%) and orthoclase (10%).

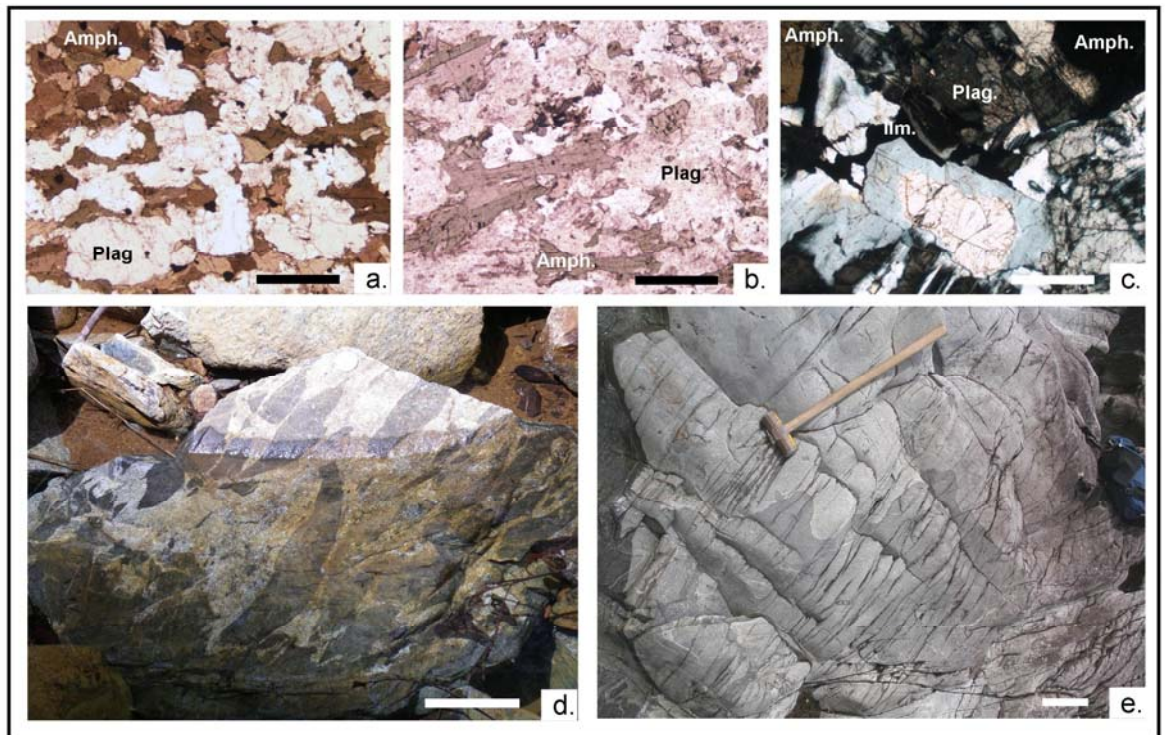


Fig.6.2. Granitoids of Mouirange area.

a. PRB38. Plane polarised view of the meladiorite of 2d showing flow textures.

b. PRB70. Plane polarized view of a quartz-bearing alkali-diorite of Rivière bleue of 2e.

c. RP45. Crossed polarised view of the amphibole gabbrodiorite in southern Rivière des Pirogues. Amphibole is present in both upper corners, ilmenite is an interstitial phase between zoned plagioclases (core : An_{84} – rim : An_{51}).

d. PRB38-39 mingled outcrop in Rivière de la Locomobile. PRB38 (2a) are meladiorite slivers (dark grey) in a quartz-bearing leucodiorite (PRB39) (white grey) (the horizontal limit is the water surface).

e. PRB70 in Rivière bleue showing a contact between a quartz alkali-diorite (light grey) and a diorite (grey).

Scale bar for 2a and 2b is 250 µm, 100 µm for 2c., 20 cm for 2d and 2e.

The contact zone with the surrounding rocks shows the presence of peridotitic enclaves within the granitoids. The interface between the two rocks is also characterised by the development of an orthopyroxenite reaction zone which can reach a thickness of a few centimetres. In addition, some areas are characterised by anastomosed veins of orthopyroxenite within the harzburgite orientated perpendicular to the contact of the intrusion [Fig.6.6c]. Alteration phases show a late hydrothermal event characterised by the deposition of sulfides (cobalto-pentlandite, sphalerite) and veins of laumontite.

Other granitoid-like rocks are observed in Rivière des Pirogues as dykes intruding the harzburgite or the transition zone [Fig.6.8b]. These rocks seem to

show a fractionation trend from the south to the north. Northern granitoids are amphibole diorite with zoned plagioclases [Fig.6.2c] and clinopyroxene-bearing diorite can occur further south, in a mingled contact with more felsic magmas. In Col de Prony, an amphibole gabbro containing 50% plagioclase and two pyroxenes is crosscutting dunite-harzburgite pods with no reaction zone with the surrounding rocks. This intrusion forms a mafic end-member in the granitoid suite of Mouirange [Tab.S11].

Finally, the granites of St Louis area have been sampled as part of this study but their field relationship (intruding the basement and the ophiolite), their composition (biotite-rich granites) and their age (27 and 24.4 Ma) (Cluzel et al., 2005) show that these intrusions as well as the intrusions of Koum-Borindi (NE of the Massif du Sud) are post-obductive and unrelated to the felsic rocks previously mentioned [Tab.S11].

6.3.1.2. Trondhjemites

Several small intrusions of sodium-rich felsic rocks are observed scattered in the harzburgite and dunite rocks. These intrusions have been recorded at Col de Prony, on the southern side of Lake Yat  , in Col des Dalmates (SW of Montagne des Sources) and are also mentioned in several places according to the previous studies in the southern part of the Massif du Sud (Noesmoen, 1971; Guillon & Trescases, 1976; Trescases & Guillon, 1977; Cluzel et al., 2006).

These intrusions are always restricted to a small size and do not reach more than a few tens of meters [Fig.6.3a,b,e]. They can occur as pods or as thick dykes and generally show a low-temperature reaction rim with the surrounding rocks (chlorite-amphibole, serpentine or none) and more rarely orthopyroxenite [Fig.6.3c,d]. Their emplacement is clearly brittle as they often contain xenoliths of the surrounding peridotite with a partial reaction zone [Fig.6.3c]. Amphibole is present in variable amounts and is more likely to occur on the rims of the intrusions [Fig.6.3b]. These felsic rocks have a chalky white appearance due to tropical alteration which forms large amounts

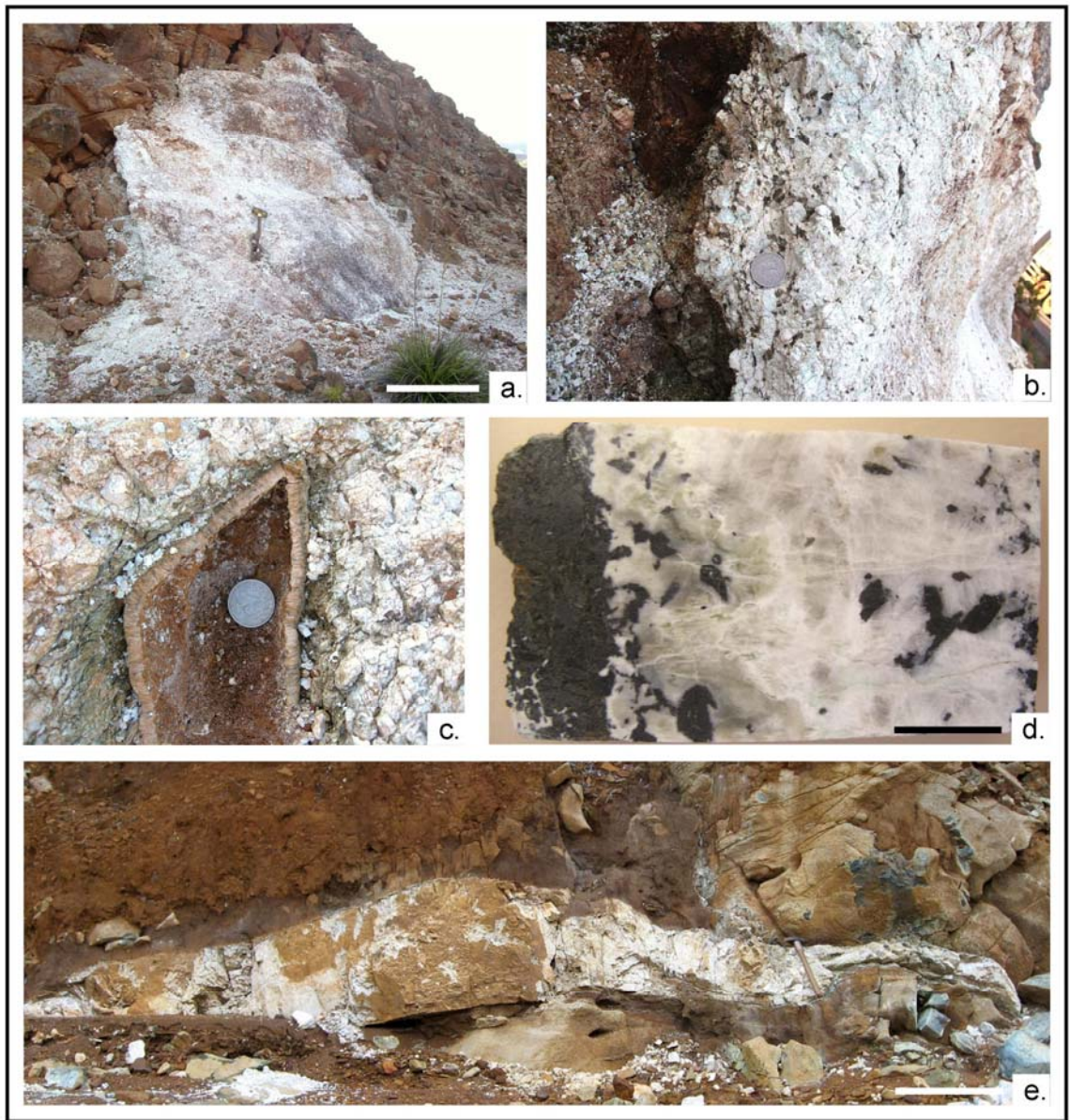


Fig.6.3. Trondhjemite outcrops.

a. R1. Trondhjemite pod intruded into harzburgite at Lake Yaté.

b. R1. Detail of trondhjemite contact with the surrounding harzburgite. Dark crystals are large mafic minerals absent in the core of the intrusion. Green patches are nickel-rich plagioclase.

c. R1. Altered peridotite enclave with an anthophyllite zone (brown orange) and an external hornblende (greenish) reaction zone.

d. PR31c. Trondhjemite dyke (right) with orthopyroxene crystals forming at the contact (left) and amphibole/opx growing into the dyke.

e. RP40. Trondhjemite dyke in the upper part of Rivière des Pirogues intruded into replacive dunites.

Coin is 3.5 cm wide. Scale bar is 2 m for 3a, 1 cm for 3d and 1 m for 3e.

of kaolinite in these alumina-rich rocks. Plagioclase is generally a few centimetres long and can have a light to strong green hue [Fig.6.3b]. K-feldspar and micas only forms a few percent of the rock in some intrusions and is generally only present as exsolutions or an accessory phase in other trondhjemites. Accessory minerals in these rocks can be very abundant and include orthopyroxene, rutile, titanite, zircon

and rare monazite. The main alteration phase is kaolinite, while at the interface between the intrusion and the surrounding ultramafic rocks, pyrophyllite, antigorite, clinochlore and various amphiboles (anthophyllite, tremolite and hornblende) form as reaction/alteration products with the surrounding ultramafic rocks [Tab.S11].

6.3.1.3. Hornblendites s.l.

In this chapter, the term hornblendite covers a large range of rocks which are mafic or ultramafic and rich in amphibole (>20%) by comparison to other rocks and contain variable amounts of pyroxene or plagioclase. The broad use of this term in this chapter is done to cover a group of rocks which have a relatively wide range of mafic compositions but similar textures and locations. For particular cases, the official terminology (LeMaitre et al., 2002) is employed to facilitate comparison and a distinction is made between amphibole-impregnated mafic rocks and hornblendite s.s.. The official definition does not impose any constraints on the composition of the amphibole (LeMaitre et al., 2002) which is not always hornblende in the Massif du Sud.

In Baie des Pirogues area, amphibole is a minor component in most rocks which are rich in pyroxenes and olivine and form dykes and pods in a transitional dunite zone. Generally, dykes of pyroxenite occur and contain 5-20% of amphibole as an interstitial phase which cements the various anhydrous crystals present in the dyke or intrusion [Fig.6.4d-f]. It is also very common to have amphibole-bearing zoned websterite or clinopyroxenite dykes which show a late veinlet of amphibole crystallised in the core of the dyke [Fig.6.4c]. A few outcrops also show amphibole gabbro-norite present in the core of some pyroxenitic dykes [Fig.6.4b].

Rarer hornblendite s.s. are recorded in the Baie des Pirogues complex as well as in Mt Ngo, just to the south (Rodgers, 1974b). These rocks occur in the dunite zone as large veins with features identical to the core of some pyroxenites previously mentioned. However, these hornblendite dykes lack a reaction zone with the surrounding rocks.

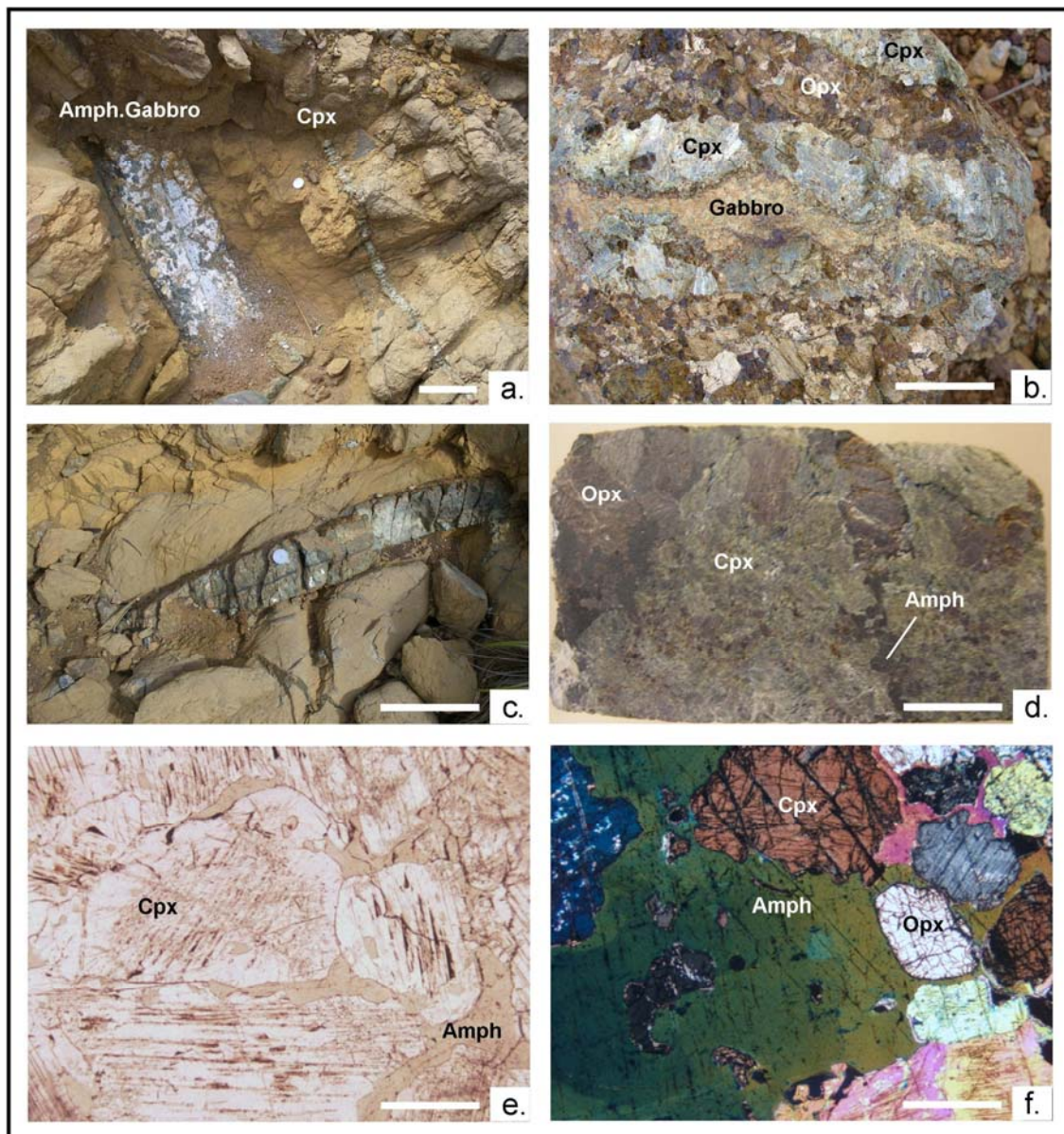


Fig.6.4. Late amphibole metasomatism in the lower part of Rivière des Pirogues.

a. BP33. Large dyke of heterogeneous amphibole gabbro parallel to a clinopyroxenite dyke cross-cutting the dunite zone of Baie des Pirogues.

b. BP34. Intrusion of gabbro in an orthopyroxenite forming a clinopyroxenite wall and amphibole infiltration into the orthopyroxenite.

c. Clinopyroxenite dyke (greenish) in the dunite zone (brown) with harrisitic growth and a hornblende veinlet (black) in the middle.

d. BP40. Thin section block showing orthopyroxene porphyroclasts (brownish) in a matrix of large crystals of clinopyroxene (greenish) with interstitial amphibole (black).

e. BP35. Impregnated gabbro with clinopyroxene crystals cemented in late amphibole.

f. RP46. Zonation of the interstitial amphibole embedding pyroxenes. Amphibole rim is 0.58 wt.% Na₂O and core is 1.49 wt.% Na₂O.

Scale bar is 10 cm in 4a and 4c, 1 cm in 4b and 4d and 250 µm in 4e and 4f.

The area of Baie des Pirogues is also characterised by the presence of large intrusions of amphibole leucogabbros which are historically known under the name of euphotide (Noesmoen, 1971) or bojite (Rodgers, 1972). These rocks occur as large (plurimetric) dykes or pockets of gabbro containing a large amount

of plagioclase and decimetric crystals of amphibole [Fig.6.4a]. Within dykes, a modal zonation is common and amphibole is more likely to occur on the edge of the vein. A reaction zone with the surrounding rock has been observed and is generally composed of zoned pyroxenite.

South of the Massif du Sud, several outcrops of highly metasomatised harzburgite crosscut by hornblendite s.s. veins and amphibole gabbro dykes have been observed at Ile des Pins. Hornblendite veins contain minor amount of plagioclase and olivine. Their relationship with the host rock are nebular and the surrounding peridotite is generally rich in amphibole (~50%) the remaining being formed by olivine and orthopyroxene. The texture shows recrystallization of peridotite in a fine grained mixture of slightly foliated hornblendite. Veins are generally coarser grained [Fig.6.5] than the surrounding hydrous peridotite [Tab.S11].

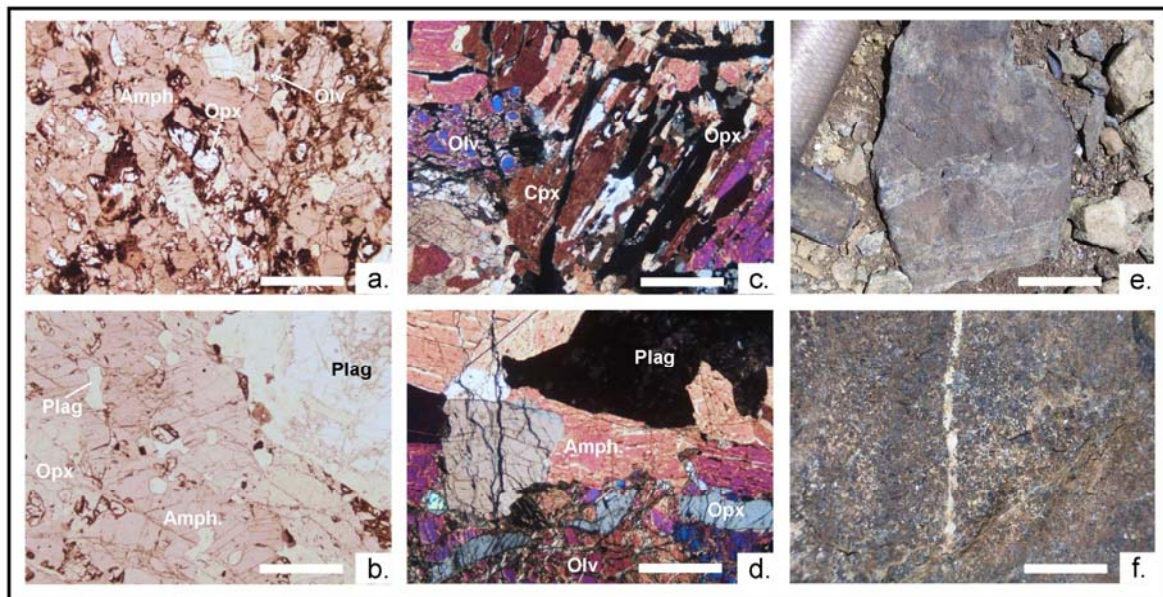


Fig.6.5. Hornblendites of Ile des Pins.

- a.** IDP31. Hornblendite (5e) with residual orthopyroxenes and olivines. Dark zones are iron oxhydroxides.
 - b.** IDP33. Hornblendite with residual orthopyroxene crosscut by a plagioclase veinlet of 5f.
 - c.** IDP34. Complex textures in amphibole gabbro where olivine and orthopyroxene (at extinction) are early phases and clinopyroxene (orange brown) is replacing orthopyroxene.
 - d.** IDP37. Amphibole gabbro vein (upper part) in contact with harzburgite (lower part).
 - e.** IDP31. Hornblendite with slivers of residual harzburgite.
 - f.** IDP33. Hornblendite with plagioclase veinlet.
- Scale bar of 5a to 5d is 250 μ m. 5e is 4 cm long and 5f is 1cm.

6.3.1.4. Pyroxenites

Pyroxenite dykes are particularly visible in the lowest part of the Montagne des Sources – Yaté section where they form a network of randomly orientated dykes ranging from a few millimetres thick to 50 cm wide. One dyke 6 cm thick was traceable for 70 metres through the harzburgite before it sharply changed direction. The latter feature is a main characteristic for this generation of dykes [Fig.6.6a]. By comparison with other discordant pyroxenites, which have sub-parallel orientations, boudinage, folding or replacive textures, the large pyroxenites dykes in Yaté area are highly discordant and have sharp angles associated with their propagation through the mantle.

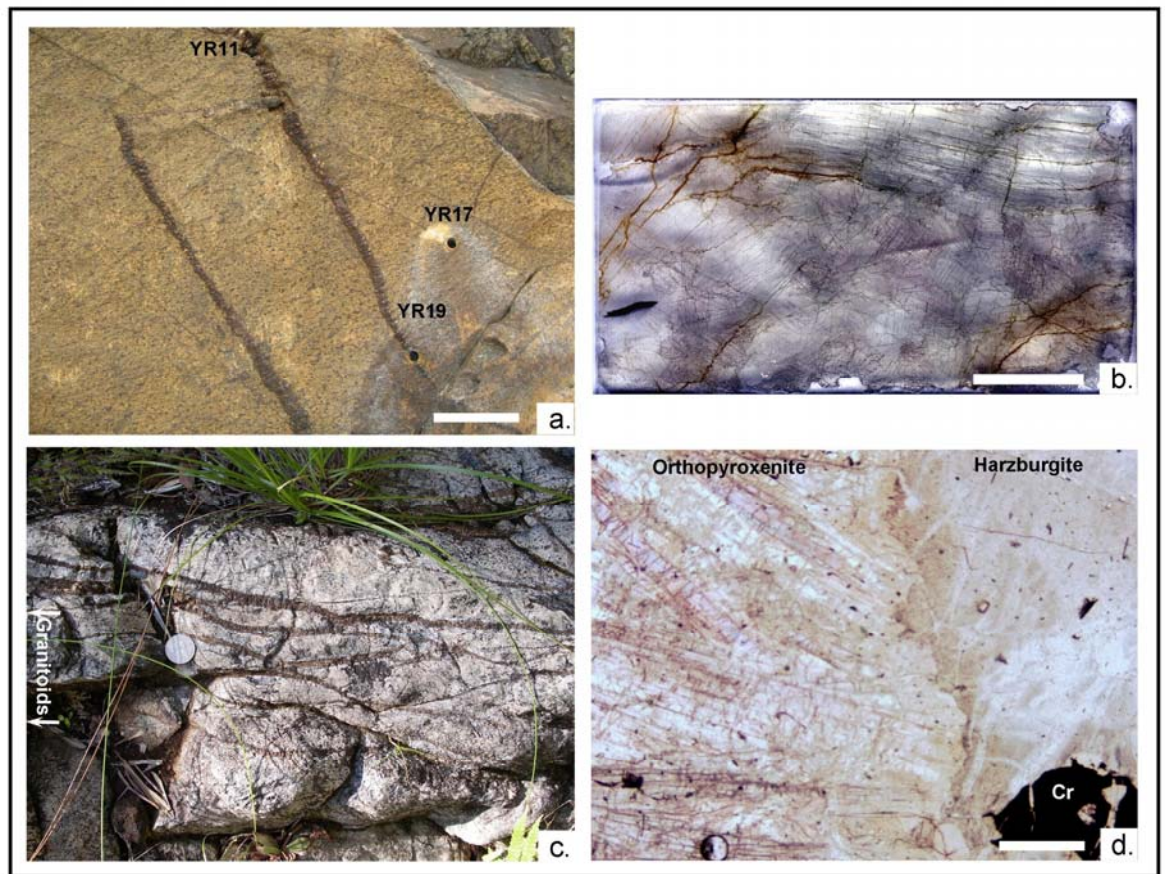


Fig.6.6. Pyroxenite dykes from Yaté and Rivière de la Locomobile.

a. Pyroxenite dykes at the Yaté dam with the location of sample YR11, YR17 and YR19 extracted by drilling. The dyke clearly show brittle emplacement with sharp angles and cross-cutting previous harzburgitic features.

b. YR1. Thin section of orthopyroxenite of Yaté dam. The coarse grain size of the orthopyroxene is clearly visible.

c. PRB41. Orthopyroxenite dikelets of Rivière de la Locomobile at less than 1 metre from the intrusive felsic rocks (left, Fig.6.2d).

d. PRB41. Thin section showing the fibrous crystal habit of the orthopyroxenes (left) in contact with a strongly serpentinised harzburgite (right).

The coin in 6c is 3.5 cm diameter. Scale bar for 6a is 20 cm, 6b is 1cm and 6d is 100 μ m.

Their mineralogical composition varies from pure orthopyroxenite to websterite with absent or very rare olivine and chromite grains. Exsolutions of clinopyroxenes or amphibole in orthopyroxene is a common feature in some of these dykes. Textures are generally equigranular with no particular growth orientation. Grain size is generally large and can reach several centimetres [Fig.6.6b].

On the western flank of the Massif du Sud, pyroxenite dykes are found in Dumbéa valley and Tontouta area in the north-east. Both localities have similar discordant dykes to the Yaté river occurrence showing identical sizes and structures of emplacement. These dykes have a wider range of modal composition and extend from orthopyroxenite to opx-bearing clinopyroxenite with an important amount of amphibole present as primary grains and displaying complex textures [Fig.6.7a].

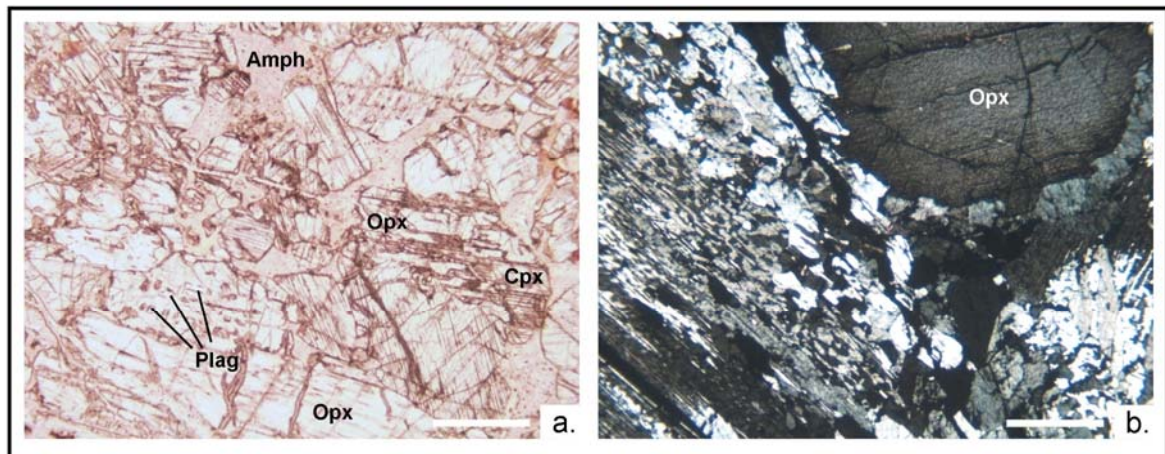


Fig.6.7. Amphibole-bearing pyroxenite dykes from Tontouta.

a. T6. Orthopyroxenite with crystals replaced by clinopyroxene and containing plagioclase inclusions. The grains are cemented in an amphibole matrix.

b. T7. Very thin section of an orthopyroxenite with large orthopyroxene crystal (top right) and complex intergrowth of orthopyroxene (dark) and clinopyroxene (white) and interstitial amphibole crystals (grey).

Pyroxenite dykelets are also observed in the area between Baie des Pirogues and Col de Prony. The harzburgite of Aiglon mine shows an important network of pyroxenite veinlets which range from ~0.5-10 cm in size. Many dykes which terminate as pyroxenites are characterised by plagioclase-rich cores and the accumulation of pyroxene crystals in their wider parts, similar to trondjemite pods. At the intersection of different dykes, larger pockets of melts (100-200 cm²) can be formed within the harzburgite [Fig.6.8, Tab.S11].

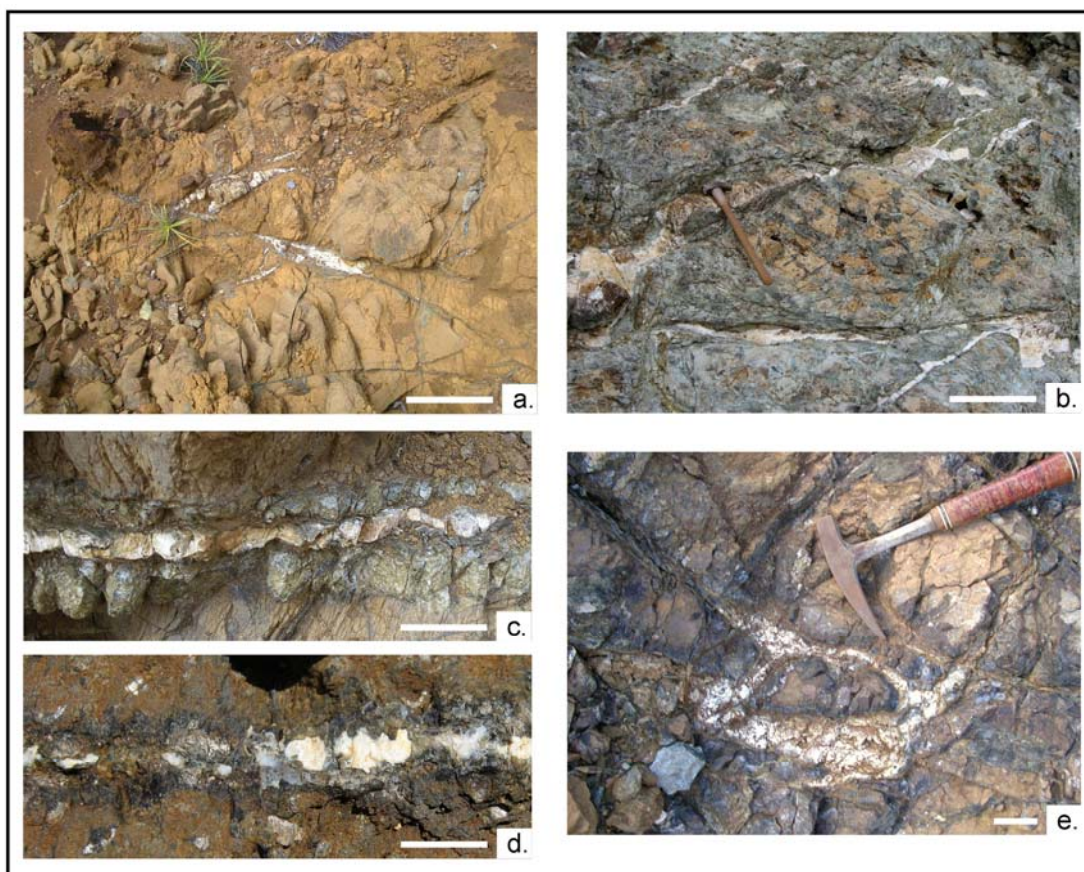


Fig.6.8. Plagioclase cores of pyroxenites.

a. Trondhjemite pocket formation in dunite at the intersection of dykes in Baie des Pirogues.

b. RP40. Network of granitic veins in the wehrlite of the transition zone of Rivière des Pirogues.

c. Plagioclase vein with clinopyroxenite walls in the dunite of Baie des Pirogues.

d. Altered plagioclase dykelets with pyroxenite walls in the harzburgite at the Aiglon Mine.

e. Aggregation of felsic melts by pyroxenite dykelets at the Aiglon Mine.

Scale bar is 50 cm for 8a and 8b, 5 cm for 8c and 8e and 1 cm for 8d.

6.3.2. Analytical techniques

Analytical techniques for EPMA and LA-ICP-MS have been previously described in chapter 4.

U-Pb dating was performed on zircon crystals from 6 samples. The zircons were separated after rock crushing using conventional heavy liquid and magnetic properties. The grains were mounted in epoxy resin and polished down to expose the near equatorial section. Cathodoluminescence (CL) investigation was carried out on a HITACHI S2250N scanning electron microscope (SEM) with an ellipsoidal mirror for CL at the Electron Microscopy Unit at the ANU. Operating conditions for the SEM were 15 kV, 60 μ A and 20 mm working distance.

Zircons were analysed for U, Th and Pb using the sensitive high resolution ion microprobes (SHRIMP II and SHRIMP RG) at RSES, ANU. Instrumental conditions and data acquisition were generally as described by Compston et al. (1992) and Williams (1998). The data were collected in sets of six scans throughout the masses and a reference zircon was analysed each fourth analysis. The measured $^{206}\text{Pb}/^{238}\text{U}$ ratio was corrected using reference values for zircon from the Temora granodiorite (TEM, Black et al., 2003). Analyses were corrected for common Pb based on the measured $^{207}\text{Pb}/^{206}\text{Pb}$ according to Williams (1998). Data evaluation and age calculation were done using the software Squid 1 and Isoplot/Ex (Ludwig, 2003), respectively. U-Pb data were collected over three analytical sessions using the same standard, with the three analytical sessions having calibration errors of 1.1, 2.0 and 2.8% (2σ), which was propagated to single analyses. Average ages are quoted at the 95% confidence level.

Hafnium isotopes analyses were carried out on zircons and rutile separates mounted on epoxy resin and analyzed for $^{176}\text{Hf}/^{177}\text{Hf}$ on LA-MC-ICP-MS at RSES, ANU. Laser ablation sampling was conducted in a He atmosphere to suppress surface deposition (Eggins et al., 1998), and a small amount of N_2 (2 cc/min) was introduced downstream of the ablation cell to increase sensitivity (Durrant, 1994). The laser was pulsed at 5 Hz, with an applied laser flux of 5 J/cm^2 . Zircon analyses used spots with diameters of 47, 62 and 81 μm whereas a 233 μm spot was employed for rutile to maximize the signal from this low-Hf phase. Further details on analytical techniques are given by Ewing et al. (2010).

6.3.3. Major elements mineral chemistry

6.3.3.1. Olivine

Olivine is an accessory phase in the mafic rocks of this series although generally absent in most samples. Olivine grains in pyroxenite dykes have a major element composition which is undistinguishable from their host harzburgite and no specific trend is observed in these ultramafic dykes other than the change associated with the host rock. In Ile des Pins hornblendites, olivine grains are more abundant (15 to 90%) and have a chemical composition similar to pristine harzburgite present in the area [Tab.S12].

Some rare amphibole-bearing dykes show minute grains of olivine with lower Mg# and NiO and slightly higher MnO which follow a trend similar to the one described in chapter 4 for crustal cumulates. Finally, in amphibole-bearing gabbros of Baie des Pirogues, similar trends are also observed for the rounded olivine grains which can abound in some samples. [Fig.6.9, Tab.S12].

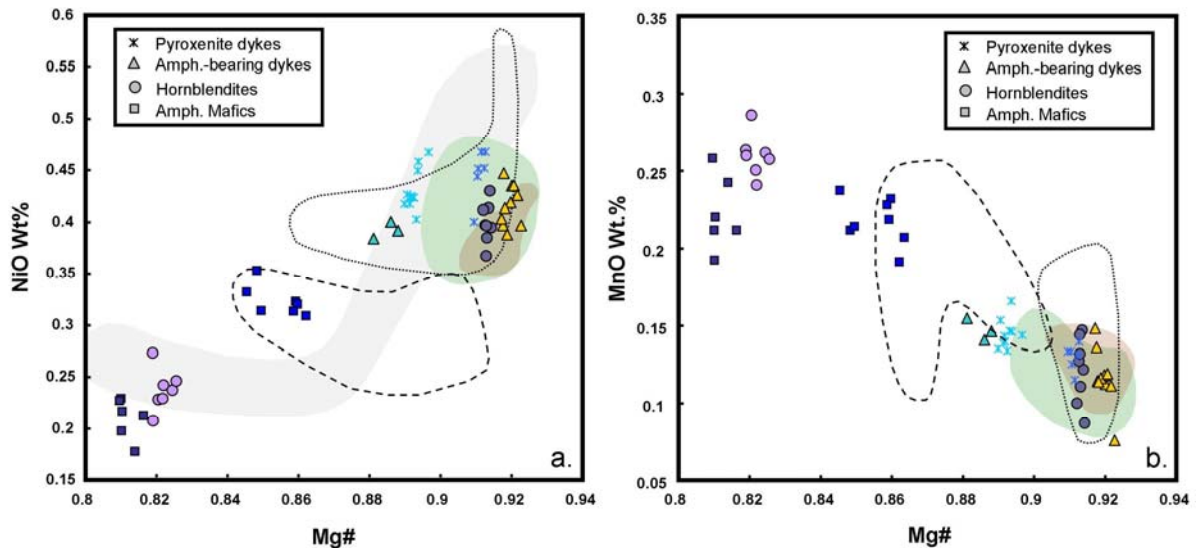


Fig.6.9. Major elements in olivine. Green field is the harzburgite and brown field are dunites. The dotted area covers the olivine from arc-related channels and dykes [Chap. 5] and dashed field are crustal cumulates. The grey field represents olivine phenocrysts in high magnesian andesite of Setouchi (Tsuchiya et al., 2005). Symbols colours represent separate samples from a same region.

a. Nickel content versus Mg-number shows that NiO decrease steadily with decreasing Mg# from the harzburgite field.

b. MnO content versus Mg-number show an increase from the harzburgite field to lower Mg#.

6.3.3.2. Orthopyroxene

As discussed in previous chapters, orthopyroxene is a key mineral in this series of rocks as it occurs in all lithologies, from pyroxenite dykes to trondhjemite. The widespread presence in this series can be divided into two modes of occurrence, as a primary phase in the melt (ultramafic and mafic rocks) or as a reaction product with the host rock (felsic rocks).

Mg-number is characteristic of the chemical evolution of the rock series. It ranges from 0.92 to 0.69 along a well defined trend. In amphibole-bearing harzburgite of Ile des Pins and in the least evolved orthopyroxenite dykes of Yaté and Tontouta area, high Mg-number similar to the surrounding peridotite are obtained. In websterite and clinopyroxenite dykes of the latter two areas, Mg-number decreases from 0.90 to 0.83 in relation with the increasing clinopyroxene

content in these rocks. Such values are also present in orthopyroxenite dykelets in Rivière de la Locomobile where the dykes are formed as a reaction zone between granitoids and the dunite-harzburgite complex [Fig.6.6c]. In Baie des Pirogues, orthopyroxene crystals have a range of Mg-number going from 0.85 to 0.71 which is in correlation with the decreasing abundance of modal orthopyroxene in the amphibole-impregnated mafic pyroxenites. The nearby amphibole gabbro-norite of Col de Prony is also associated with that range of values and bears a Mg-number of 0.86. Hornblendite veins from the Ile des Pins contain small amounts of orthopyroxene which possess a Mg-number between 0.78 and 0.80. Finally, orthopyroxene grains obtained by mineral separation from the core of trondhjemite pods provide a range of values going from 0.76 to 0.69 [Fig.6.10, Tab.S14].

Alumina is fairly constant in all orthopyroxenes, averaging around 1 ± 0.2 wt.%. Noticeable exceptions are impregnated gabbros in Baie des Pirogues and Col de Prony, as well as hornblendite dykes of Ile des Pins which have values of 1.5 to 2 wt.% and are the only rocks which overlap with crustal gabbro-norites discussed in chapter 4 [Fig.6.10a, Tab.S14].

Chromium content is similar to mantle values in amphibole-bearing harzburgite and drops to lower contents in orthopyroxenite dykes. In websterite and clinopyroxenites, Cr_2O_3 ranges from lower mantle values ($\text{Cr}_2\text{O}_3 = 0.3$ wt.%) to very low amounts ($\text{Cr}_2\text{O}_3 < 0.04$ wt.%). Hornblendite dykes and trondhjemites share similar low values whereas hornblendites *s.l.* of Baie des Pirogues have a decreasing content from 0.35 wt.% to <0.05 wt.% Cr_2O_3 along with decreasing Mg# [Fig.6.10d, Tab.S14].

Calcium is generally low from 0.25 to 0.78 wt.% CaO with a slight increasing trend from pyroxenite where small clinopyroxene exsolutions can occur to hornblendites. Some meladiorites and impregnated rocks from Baie des Pirogues and Ile des Pins are the only rocks with slightly higher values (0.5 to 0.9 wt.%) and do not contain any clinopyroxene exsolutions. Orthopyroxene in trondhjemite have similar features and show low values from 0.5 to 0.1 wt.% [Fig.6.10b, Tab.S14].

Nickel content is generally in a similar range to the cumulate gabbros (0.01 to 0.08 wt.% NiO) for all rock-forming crystals. Reaction zones of granitoids and some orthopyroxenes grains in felsic and intermediate plutons have higher values

(0.05 to 0.16 wt.% NiO) with the highest values for orthopyroxenite offshoots in the surrounding ultramafic rocks (up to 0.27 wt.% NiO) [Fig.6.6c, 6.10b, Tab.S14].

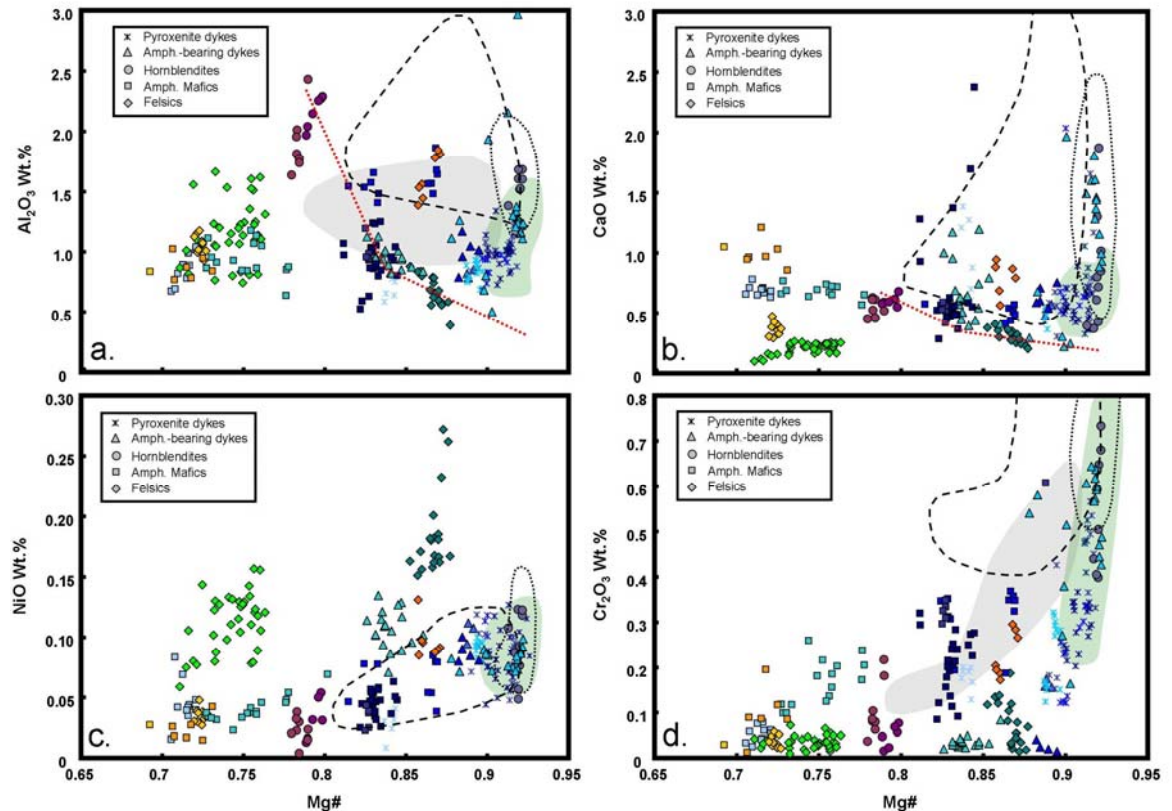


Fig.6.10. Major elements in orthopyroxene as a function of Mg-number. Green field represents harzburgite composition. The dotted area are orthopyroxenes from arc-related channels and dykes [Chap. 5] and dashed field are crustal cumulates. The red line represent the orthopyroxene composition in the experimental work of Rapp et al. (1999, 2010) for increasing melt:rock ratio (to lower Mg#). The grey field are pyroxenite dykes from Solomon Island ophiolite (Berly et al., 2006).

a. Alumina shows that it is essentially constant in all rocks and remains low except in some amphibole-rich rocks.

b. CaO shows a slight rise from very low values in pyroxenite dykes (~0.3 wt.%) to amphibole-bearing felsic rocks which have concentrations at least twice higher. Orthopyroxene obtained from trondhjemite pods are very low in CaO.

c. NiO is decreasing steadily from mantle values to felsic rocks. However, pyroxenites formed by interaction with the surrounding rock can have a very high content in NiO in pyroxenes as well as pyroxenes present in some hornblendite dykes and trondhjemites.

d. Cr₂O₃ decreases strongly from orthopyroxenite to websterite dykes and remains mostly low in amphibole-bearing dykes, hornblendites and felsic rocks. However some amphibole gabbros formed by infiltration of melts have higher content. Pyroxenites walls are not particularly enriched.

6.3.3.3. Clinopyroxene

Clinopyroxene is present in most rocks with the exception of orthopyroxenite reaction zones and in felsic intrusions. In orthopyroxenite dykes, it occurs as exsolutions and does not represent a primary phase.

Mg-number ranges from 0.94 to 0.92 in most pyroxenite dykes. Amphibole-bearing harzburgite also has a similar composition at ~0.93. A few exceptions are websterite dykes with lower Mg-numbers and can have values as low as 0.88. Amphibole-impregnated rocks of Baie des Pirogues have a continuous range from 0.91 to 0.79 in Mg-number whereas the meladiorite of Col de Prony ranges from 0.91 to 0.88 [Fig.6.11, Tab.S16].

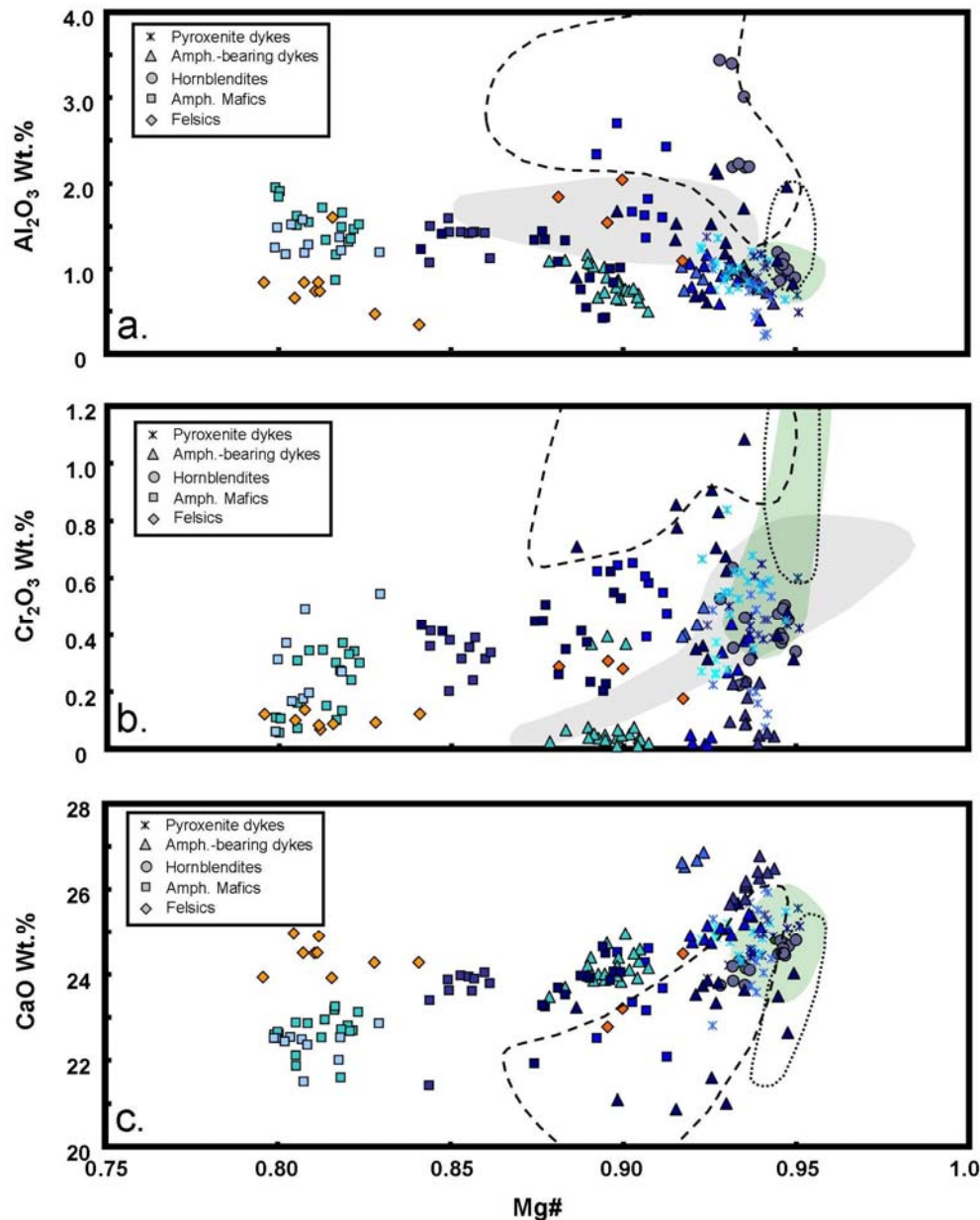


Fig.6.11. Major elements in clinopyroxene as a function of Mg-number. Green field covers harzburgite clinopyroxene exsolutions. The dotted area shows clinopyroxene from arc-related channels and dykes [Chap. 5] and dashed field are crustal cumulates. The grey field are pyroxenites dykes from Solomon Island ophiolite (Berly et al., 2006).

a. Alumina show steady values generally below 2.0 wt. % with a very slight increase with differentiation.

b. Cr_2O_3 in clinopyroxenes behaves similarly to orthopyroxenes and decreases steadily with differentiation or is very low in amphibole-bearing dykes.

c. CaO is decreasing steadily from the values found in the harzburgitic mantle as exsolutions to ~22.5 wt.% for amphibole gabbros. Clinopyroxene-bearing felsic rocks can have higher CaO values.

Alumina content is low and slightly increasing from 0.85 ± 0.2 to 1.50 ± 0.3 wt.% Al_2O_3 from pyroxenites to granitoids. Amphibole-bearing harzburgite in Ile des Pins has high content up to 3.4 wt.% Al_2O_3 and a couple of samples with amphibole-impregnated mafic lithologies have around 2 wt.% of alumina [Fig.6.11a, Tab.S16].

Chromium content is similar to the surrounding harzburgite porphyroclasts for orthopyroxenite dykes ($\sim 0.5\pm0.15$ wt.% Cr_2O_3). For some websterite dykes of Tontouta, the Cr_2O_3 content is extremely low (<0.07 wt.%). Amphibole-bearing gabbros of Baie des Pirogues have decreasing values from 0.6 ± 0.05 to 0.2 ± 0.1 wt.% Cr_2O_3 with decreasing Mg-number [Fig.6.11b, Tab.S16].

Calcium content in pyroxenite dykes is slightly higher than the host harzburgite (24.5 to 26.7 wt.% CaO) as it lacks of orthopyroxene exsolutions. A well defined trend from these values to 22.5 wt.% is observed for decreasing Mg-number in dykes and amphibole gabbros [Fig.6.11c, Tab.S16].

6.3.3.4. Amphibole

Primary magmatic amphiboles have a restricted range of composition in Na-Ca amphiboles. Amphiboles from Ile des Pins in both hornblendites s.s. veins and amphibole-rich harzburgite are pargasite with a similar composition is found for amphibole-gabbros of Baie des Pirogues. The grains of amphibole present in websterite dykes of Tontouta are edenite or hornblende. In granitoids related to Mouirange, amphiboles are hornblende and similar composition are found in the felsic dykes cross-cutting the harzburgite, chromitites and the transition zone. Finally, in pyroxenite dykes of the Yaté dam, exsolutions of tremolite are observed in some orthopyroxene of some samples but these are grown by re-equilibration processes. Anthophyllite is a common phase in trondhjemite as veins or reaction zones [Fig.6.3c] and ferro-actinolite occurs as rims around hornblende grains in some granitoids [Fig.6.2b]. Both phases are interpreted as late stage and are likely to be related to hydrothermal fluids.

The Mg-number is highly variable and reflects both the composition of the rock and notably the content of anhydrous mafic minerals, and the size of the intrusion or the type of interaction with the host rock. Amphibole-bearing

harzburgite has a Mg-number of 0.9, a value which is also found for some primary amphibole in some dykes of pyroxenites. Other dykes hosting amphiboles have Mg-number between 0.9 and 0.75, the latter being for hornblendite s.s. dykes. Amphibole gabbros of Baie des Pirogues have variable contents between 0.85 and 0.75 and trondhjemite pods contain amphibole with Mg# of 0.70, increasing to 0.80 at the contact with the surrounding ultramafic rocks. Finally, primary amphiboles in granitoids have a Mg-number between 0.5 and 0.6 whereas secondary hydrothermal rims are around 0.2 [Fig.6.12, Tab.S21].

Calcium content has a positive correlation with Mg-number and range from 1.5 in granitoids to 2.0 Ca p.f.u. in rocks with a Mg-number around 0.9. When nickel and chromium content are detectable, a similar correlation with Mg-number of amphiboles is observed from 0.7 to 0.9 [Fig.6.12b]. With the exception of pyroxenite dykes in Yaté dam area which are low in alumina, most of the amphibole-bearing rocks have between 3 and 10 wt.% Al_2O_3 . Hornblendites of Ile des Pins have between 11 and 15 wt.% Al_2O_3 [Fig.6.13a]. Titanium contents follow the same evolution and range between 2 wt.% TiO_2 in hornblendites to very low amounts in some alkali-rich felsic rocks [Fig.6.13b]. Silica follows an opposite trend and is around 6 Si p.f.u. in hornblendites and increase from 6.5 to 7 Si p.f.u. in the amphibole-impregnated rocks of Baie des Pirogues and reach 8 Si p.f.u. in the mafic end-members of the Mouirange intrusions [Fig.6.13b, Tab.S21].

Sodium content is high (0.55 to 0.75 Na p.f.u.) in hornblendite veins and amphibole-bearing harzburgite. It ranges between 0.3 and 0.55 Na p.f.u. for amphibole-bearing ultramafic rocks of Baie des Pirogues and some amphibole-bearing dykes of the northwest region where it show a steep correlation with decreasing Mg-number. Meladiorites to granitoids and the inner reaction zone of trondhjemite pods goes from 0.05 to 0.35 Na p.f.u. with decreasing Mg-number, along a trend which displays a lower gradient from the orthopyroxenite dykes field [Fig.6.12a, 6.13a, Tab.S21]. A strong positive correlation exists between sodium (or $\text{Na}/(\text{Na}+\text{Ca})$) and aluminium ($\text{Na} = 0.32 \text{ Al} - 0.03$ ($r^2 = 0.89$)) which is only valid for magmatic and primary amphiboles. This variation is mainly related to amphibole modal abundance in rocks but there is also a trend which show an increase from northern amphibole-bearing rocks to southern outcrops (Ile des Pins) where sodium and alumina content are particularly high [Fig.6.1, 6.13a].

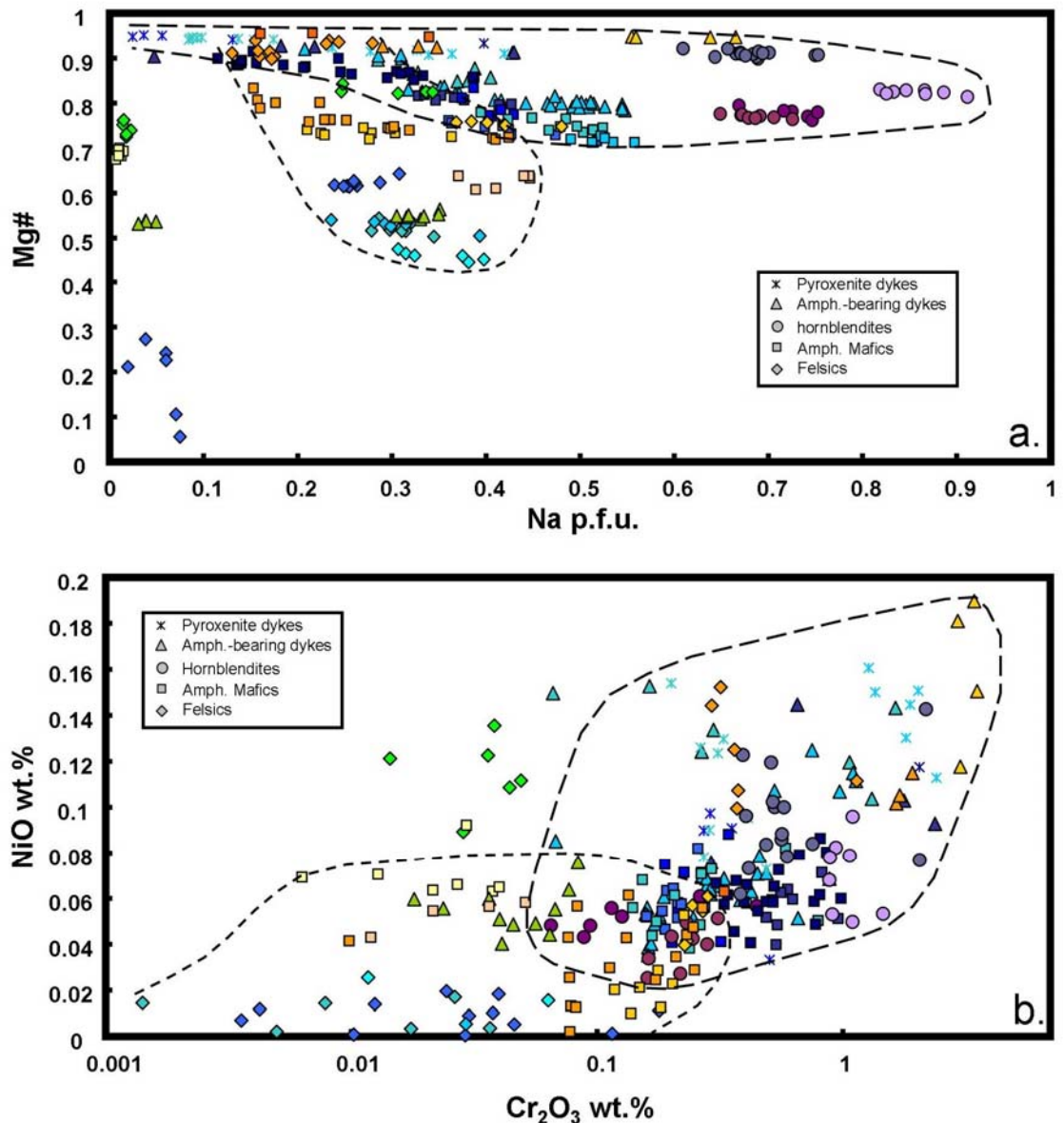


Fig.6.12. Amphibole major element composition. The two fields highlight two different trends among the amphiboles.

a. Na content per amphibole formula unit (23 oxygens) versus Mg#. Very low sodium content with low Mg# (<0.9) are secondary hydrothermal amphiboles. Pyroxenites have high Mg# and low sodium. Amphibole-bearing dykes are variable whereas hornblendites from Ile des Pins are very rich in sodium. Amphibole mafic rocks of Baie des Pirogues and granitoids have similar Na content but the Mg-number varies as a function of differentiation.

b. Amount of NiO and Cr₂O₃ in amphiboles show that high content are present in pyroxenite dykes and amphibole-rich rocks. Amphibole mafic rocks are lower in Ni and Cr and granitoids have very low content.

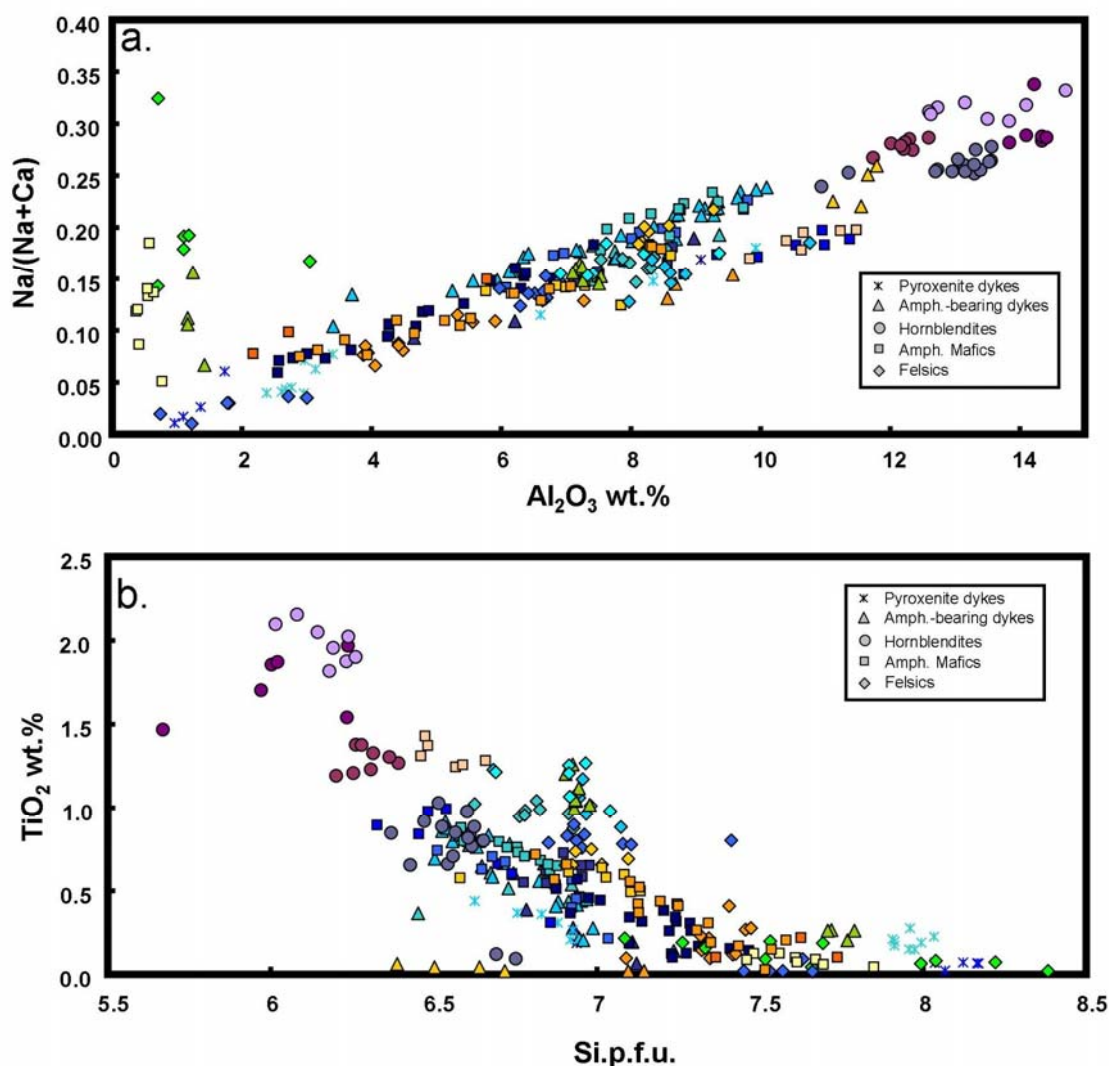


Fig.6.13. Amphibole major element composition.

a. Alumina content as a function of the Na/(Na+Ca) ratio in amphibole show a strong correlation between sodium and aluminium. Contents increase from pyroxenites to amphibole-bearing gabbros, intermediate rocks and granitoids. Hornblendites are particularly high in sodium and aluminium content.

b. Titanium content versus silica per formula unit in amphiboles (23 oxygens) shows an inverse correlation. Pyroxenites and amphibole gabbros have high silica content and negligible titanium. Granitoids have intermediate values and hornblendites have low silica and high titanium.

6.3.3.5. Feldspars

Feldspars are present in all granitoids, trondhjemitic rocks and related veins of mafic rocks with the exception of pyroxenites in Yaté. Anorthite content in plagioclase is widely variable. In amphibole-impregnated cumulates, the albitic component is barely present and plagioclase is An₉₇₋₉₈.

In hornblendites of Ile des Pins, calcium content is similar to the crustal gabbronorites and range from An₈₆₋₉₅. Diorite pods and dykes of Col de Prony are intermediary (An₆₁₋₈₈) with the granitoids which have values between 35 and 66% Ca/(Na+Ca). Although most granitoids show normal zoning of the plagioclase crystals with variations within 10-15%, some intermediate rocks show plagioclase core with An₈₅ and rims at An₅₀ and lack of transitional composition [Fig.6.2c]. Towards more sodic compositions, a gap exists between granitoids and trondhjemite which goes from An₀ to An₁₇. Several of the trondhjemite and alkali-diorite also have zoned plagioclases and display an albitisation event at a late stage of crystallisation, probably related to the growth of secondary iron-rich amphiboles. K-feldspars are present in trondhjemites as an accessory phase which only reaches ~6-7% in one dyke and is generally below 5% modal abundance [Fig.6.14, Tab.S19].

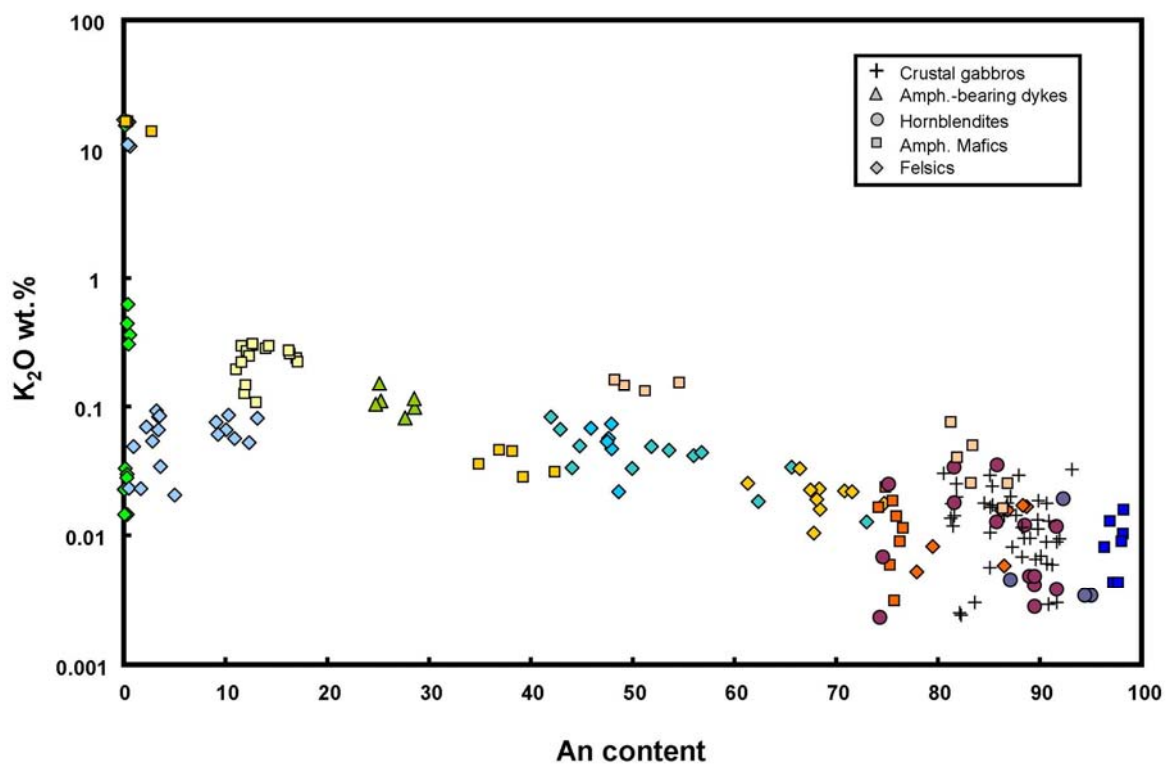


Fig.6.14. Anorthite content (Ca/(Na+Ca)) in feldspars in function of K₂O concentration. Cumulate crustal gabbros are restricted to An₈₀₋₉₃. Feldspar with high K₂O (>10 wt.%) are orthoclase present in some alkali granitoids.

6.3.3.6. Other phases

Spinel

Spinel is absent from pyroxenites and occurs rarely in websterite and mafic rocks which contain minor amounts of amphibole. The composition of the spinel phase in these rocks is chromite with iron as the dominant divalent cation. Magnetite occurs in most rocks of this series as a secondary phase and is only absent in trondhjemite. In hornblendite s.s. dykes, trevorite (NiFe_2O_4) has been observed.

Zircons

Zircons have been observed in all felsic rocks. They are scattered in the rock mass and form idiomorphic crystals of 80 to 400 μm in size. Felsic dykes generally have rounded zircons with less visible zonation. Quartz and nickel-rich gedrite are occasional inclusions within crystals [Fig.6.21].

Ti-bearing phases

Rutile is abundant in some trondhjemites where it accumulates between plagioclase grains as small eroded prisms [Fig.6.15a].

Ilmenite can occur as a filling in cracks or as a coating phase of some rutile [Fig.6.15b] and seems to be a late phase in trondhjemite. In granitoids, ilmenite can be an important component of meladiorite. Compositions only vary in a small range and $\text{Fe}/(\text{Fe}+\text{Mn})$ is between 0.94 and 0.98.

Titanite is commonly observed in trondhjemite pods where it is particularly abundant on the edge of the intrusions where rutile becomes less ubiquitous [Fig.6.15d]. Titanite crystals in some trondhjemite can reach sizes of 500 μm but are frequently between 50 and 150 μm . Some samples show a rutile core in idiomorphic titanite crystals [Fig.6.15c] suggesting that titanite is a late phase formed on the side of the pods by interaction with the surrounding rock. Some amphibole-bearing websterite dykes of Tontouta also carry rare grains of titanite.

Phosphates

A few rare grains of monazite have been obtained from mineral separation on some trondhjemite pods and have an average chemical composition

$(\text{Ce}_{0.46}\text{La}_{0.26}\text{Nd}_{0.18}\text{Ca}_{0.04})_{\Sigma=0.94}(\text{PO}_4)_{1.08}$. Apatite has been detected in granitoids as minute prisms but is relatively ubiquitous throughout the Mouirange intrusion.

Quartz

Quartz is an accessory phase in trondhjemites but can form around 10% of the rock in some granitoids of the Mouirange pluton where it often crystallise as granophyric clusters of 1-2 mm between plagioclase crystals.

Sulfides

Sulfides are rare in most lithologies except some amphibole websterite and amphibole gabbro containing small amount of pentlandite. Granitoids are cross-cut by several veins which can be enriched in opaque material. Although these phases were not studied in detail, iron, zinc and copper sulfides as well as cobalto-pentlandite have been detected.

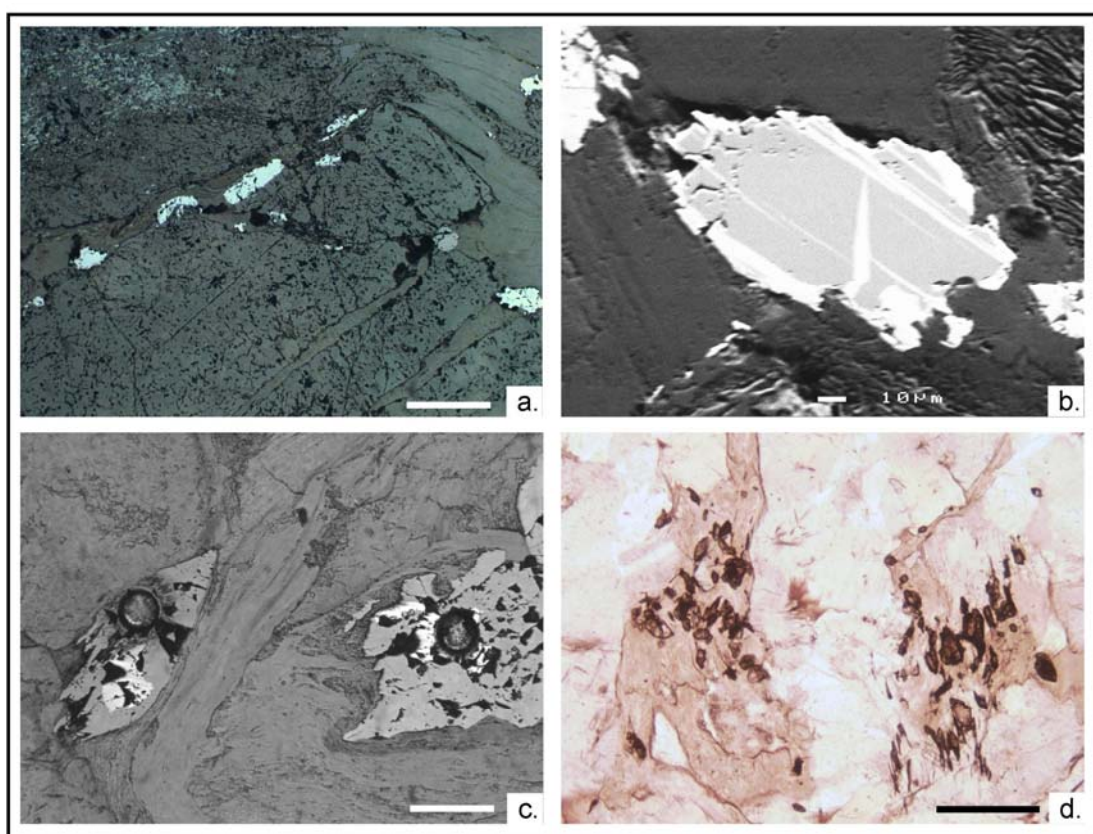


Fig.6.15. Accessory minerals in trondhjemites.

a. R1. Trondhjemite pod [Fig.6.3a] in reflected light showing rutile (light grey) and titanite (grey) in contact with albite crystals (dark grey).

b. R1. Back scattered electron image of a rutile grain showing ilmenite exsolutions as laths, rims and blebs.

c. R33. Titanite crystals with rutile core in trondhjemite dyke. Circular holes are laser spots.

d. R33. Titanite accumulation in kaolinitised zone of trondhjemite.

Scale bar is 150 μm for 15a, 10 μm for 15b, 50 μm for 15c and 250 μm for 15d.

Alteration phases

Late hydrothermal assemblage have been observed in the felsic rocks where veins of laumontite [$\text{CaAl}_2\text{Si}_4\text{O}_{12} \cdot 4\text{H}_2\text{O}$] have been analysed. Albitisation seems to occur in trondhjemite pods but mostly leads to an assemblage of albite and donbassite [$\text{Al}_2\text{Al}_{2.33}(\text{Si}_3\text{AlO}_{10})(\text{OH})_8$]. At the contact of felsic and ultramafic rocks, anthophyllite, talc and chlorites are common phases. Most of the post-hydrothermal assemblage involves the oxidation of minerals and the formation of kaolinite. In mafic rocks, alteration products are similar to the one previously mentioned [Chap.4.3.4.6] and are mainly serpentines, magnetite and prehnite.

6.3.4. Trace elements mineral chemistry

6.3.4.1. Orthopyroxene

Trace elements concentrations in orthopyroxene are highly consistent within each group of rocks and show a progressive evolution in relation with the variation in the lithology of the samples. Titanium is a good discriminating trace element and is generally 10 to 100x more enriched in these rocks than in the surrounding harzburgite. A few exceptions are noticed such as some hornblendite dykes of Tontouta, hornblendite in Ile des Pins and the orthopyroxenite dykes and walls formed around granitoids [Fig.6.16]. Their trace element patterns suggest that these grains have a relationship with the country rock. Zirconium has similar concentration than harzburgite but a strong enrichment (100x) occurs within the series from orthopyroxenite dykes in Yaté to meladiorites of Mouirange to amphibole diorites of Baie des Pirogues, pyroxenites walls of granitoids and finally trondhjemite pods [Fig.6.16b]. Chromium content decreases steadily. Orthopyroxenite dykes have similar values to the harzburgite, while amphibole meladiorite has an intermediary position and amphibole-bearing felsic rocks and trondhjemites have the lowest values [Fig.6.16a]. Barium has generally higher values in orthopyroxenites than in the surrounding harzburgite and is particularly high for orthopyroxenes related to felsic intrusions. Niobium also has high values in these “felsic” orthopyroxenes whereas amphibole diorites and hornblendites have lower values than the mantle orthopyroxene porphyroclasts. Orthopyroxenite dykes have similar content than harzburgite [Tab.S15].

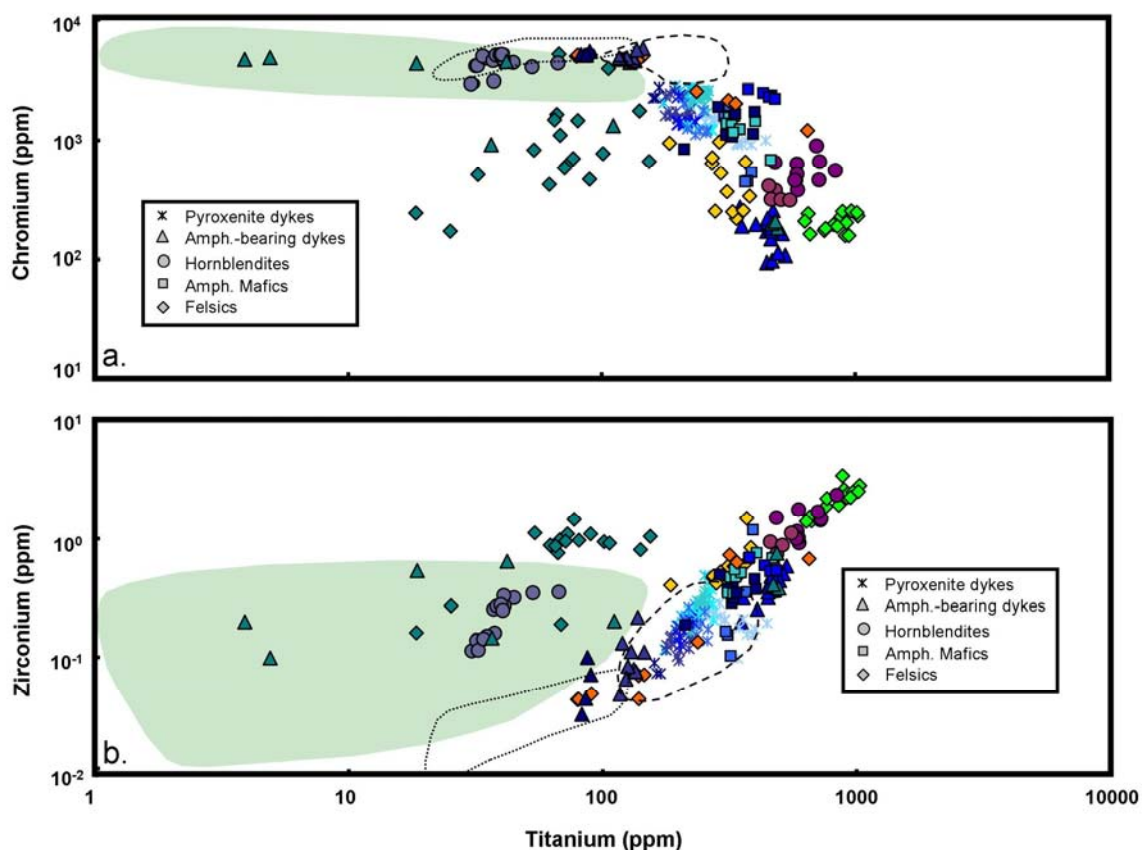


Fig.6.16. Trace elements in orthopyroxenes. The green field represents orthopyroxene porphyroclasts of harzburgite, the dotted field covers the composition of pyroxenites channels and dykes [Chap.5] and the dashed field are crustal cumulates [Chap.4].

a. Chromium versus titanium show increasing titanium content with differentiation and a constant decrease in chromium. From their chromium content, this series is clearly distinctive from the proto-pyroxenites and gabbronorites.

b. Zirconium versus titanium shows a positive correlation between both trace elements. Zirconium content of Yaté pyroxenites is around 10x higher than the pyroxenite dykes developed in chapter 5. Mafic and felsic rocks are clearly distinctive from the gabbronorites although they share the same enrichment gradient.

6.3.4.2. Clinopyroxene

Trace element contents in clinopyroxene provide similar figures for all the rocks in the series. Amphibole gabbros of Baie des Pirogues has content generally 2 to 3x higher than pyroxenites and only show significantly different patterns for U-Th, Mn and Zn. Metasomatised harzburgite of Ile des Pins and amphibole-bearing dykes of Tontouta have concentrations which are generally 5 to 8x higher than pyroxenites, particularly the LREE. Alternately, HFSE are strongly depleted compared to other ophiolitic rocks.

Rare earth element patterns show similar figures but concentrations increase in relationship with higher mafic modal contents in the rocks. La/Sm varies from 0.14 in amphibole bearing rocks to 0.51 in pyroxenites and Sm/Lu is

generally near the unity or slightly higher for pyroxenite (0.85-1.22), above these values for amphibole-impregnated mafics (1.49-1.84), and very high for amphibole-bearing dykes and hornblendites (1.76-3.36). A light negative europium anomaly is present in all REE diagrams. These features contrast with the clinopyroxene REE patterns which are obtained on cumulate gabbro norites [Chap.4, Fig.6.17, Tab.S17].

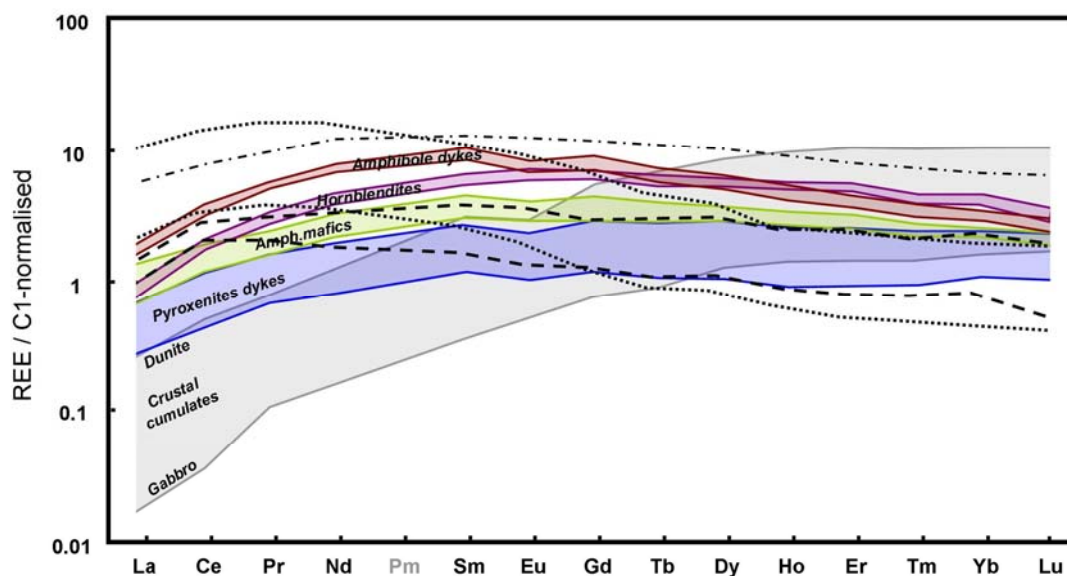


Fig.6.17. Clinopyroxenes REE content normalised to C1 chondrite (Sun & McDonough, 1989). All patterns share similar figures with a convex upward pattern centred on Sm. Fractionation between REE increases with differentiation. The grey field represent cumulate or interstitial clinopyroxene of the crustal sequence [Chap.4] with a decreasing content in REE with differentiation. The pattern for these clinopyroxenes is clearly different from the fore-arc series and has only increasing concentrations for lighter REE. The two couples of lines are the chemical variations for the trace elements obtained on clinopyroxene phenocrysts of setouchiites from Setouchi (dashed) and Kitakami (dotted) (Tsuchiya et al., 2005). Dash-dot line is interstitial clinopyroxene in dunite xenoliths of Monglo (Philippines) (Grégoire et al., 2008).

6.3.4.3. Amphibole

With the exception of amphibole formed by the hydrothermal alteration of rocks (mostly anthophyllite, sometimes tremolite), amphibole grains have very similar trace element patterns. Despite the large variety of rocks where amphibole was analysed, all grains show a strong positive anomaly in Ba, high LREE/MREE ratios and relatively flat or slightly decreasing MREE/HREE. Transition metals are consistent between samples but show wide variations between trivalent (Sc, V, Cr) and divalent metals (Ni, Co, Cu, Zn), the latter having low concentrations [Fig.6.18, Tab.S22].

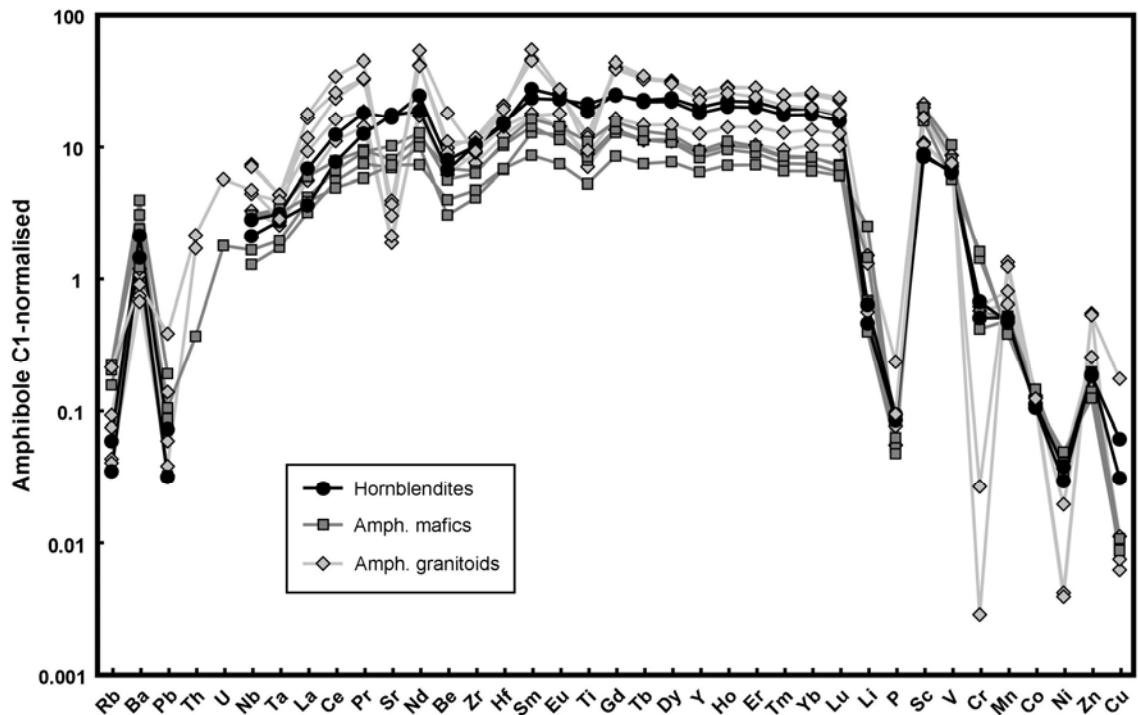


Fig.6.18. Amphibole trace element pattern normalised on C1 chondrite (Sun & McDonough, 1989). Patterns show similar figures for all rocks, with high Ba, high LREE/MREE and flat MREE-HREE pattern. The main differences concern low Sr and Eu* in felsic rocks as well as low Ti, Cr and Ni in granitoids.

Beside the general variations which are related to the evolved and felsic character of the Mouirange plutons regarding to amphibole-bearing mafics and ultramafics, a few secondary differences in the patterns appear between groups of rocks. Amphibole gabbros of Baie des Pirogues and hornblendites of Ile des Pins are particularly enriched in barium as well as chromium and nickel whereas other elements (HFSE, REE, TM) are in lower concentration than other amphiboles [Fig.6.18]. Amphiboles in granitoids have higher content in HFSE and REE but slightly lower values in LILE. These rocks also display clear negative anomalies for Sr and Eu which are both absent in amphibole-bearing mafic rocks which contain little or no plagioclase. This observation suggests that amphibole formed later than plagioclase. Trace element patterns of granitoids also display a negative titanium anomaly absent in mafic rocks and strong depletion in Cr and Ni whereas such variations are almost unnoticed in other rocks. Finally, the Nb/Ta ratio is higher than 1 in the granitoids of Mouirange and Rivière des Pirogues and contrasts with amphibole-rich mafic rocks and hornblende which have a low Nb/Ta [Fig.6.18, Tab.S22].

6.3.4.4. Plagioclase

Trace element contents of plagioclase in the suite of rocks investigated in this chapter display higher concentrations than anorthite-rich plagioclase which crystallised in the cumulate gabbros [Chap.4.3.4.5] [Fig.6.19]. The strong enrichment is particularly visible for barium and LREE in granitoids and is also present in other amphibole mafic rocks where it has an intermediary position between cumulate gabbros and granitoids [Fig.6.20]. Amphibole-impregnated mafic rocks in Baie des Pirogues are characterised by an europium positive anomaly which is slightly less pronounced than for the felsic rocks. Similar observations can be made for beryllium, but not strontium which shows similar values in all plagioclase-bearing rocks excluding the crustal cumulates [Fig.6.20]. The REE pattern of plagioclase in trondhjemites show relatively high values for HREE which is a feature not shared by cumulus plagioclase and is below detection limits in other granitoids [Fig.6.19, Tab.S20].

The green plagioclase present in some trondhjemite pods [Fig.6.3b] is characterised by very high contents in nickel (650ppm) and cobalt (50ppm). Although felsic rocks with green plagioclase show an extreme enrichment, high values are also observed in some amphibole mafic rocks whereas large intrusions of granitoids have values for transition metals which are lower than the crustal cumulates [Tab.S20].

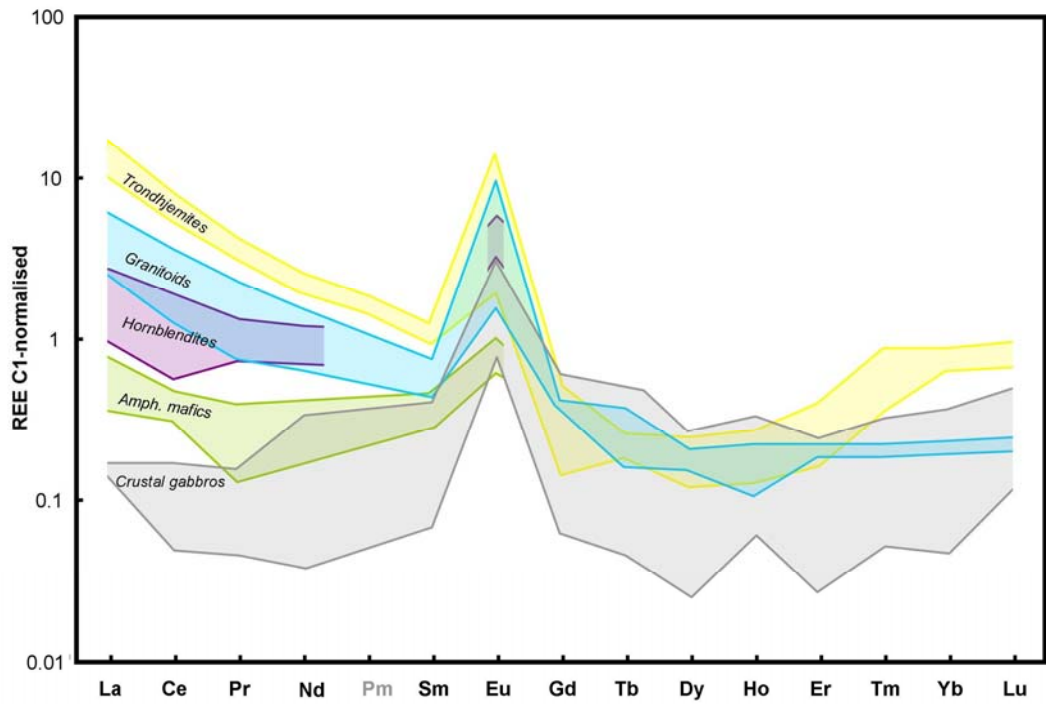


Fig.6.19. Plagioclase trace element pattern for REE show increasing LREE concentration with increasing plagioclase modal content and sodium content.

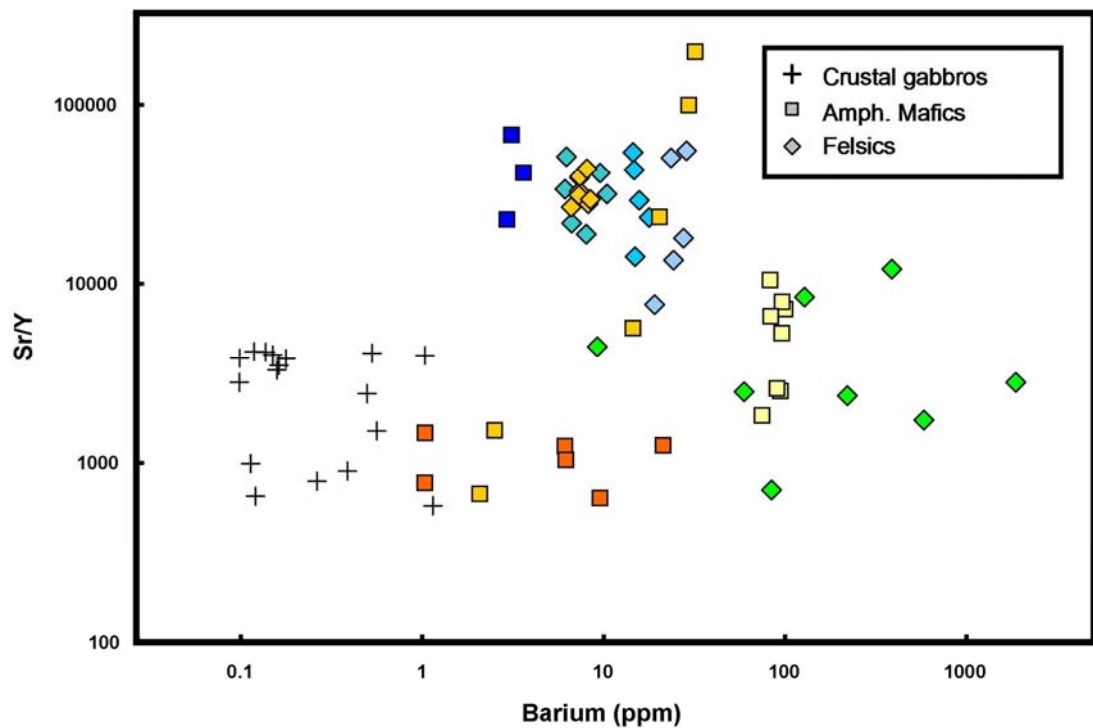


Fig.6.20. Sr/Y ratio in plagioclase as a function of barium content. Crustal cumulates have low content in Ba and a low Sr/Y ratio. Some transitional gabbro dykes have higher Ba concentration but it is in the granitoids and amphibole-rich mafic rocks that high Ba and high Sr/Y are clearly visible.

6.3.4.5. Other Phases

Accessory phases play an important role as the host for several highly incompatible elements. Most mafic and ultramafic rocks lack any accessory phases with the exception of some spinels whose chemical composition is suggestive that they are inherited from the host rock. With the exception of Ti, V, Cr and Zn, spinels *s.l.* are generally not enriched in other components. In some amphibole-bearing pyroxenite dykes, rare titanite crystals have been observed but no trace element data is currently available on this phase.

In amphibole-bearing mafic intrusions and granitoids, the main accessory phase is ilmenite. This oxide is an important host for HFSE, Sc, V and Mn. Magnetite also occurs in these rocks and can be a potential host phase for elements similar to the spinels mentioned above. However, these two oxides were not covered in this study and need further work to separate primary magnetite from ilmenite and alteration products. In the granitoid intrusions, apatite and zircon have been observed by cathodoluminescence but no trace element data has been obtained from these phases.

Trondhjemite pods are rich in accessory phases which are easily observed in thin sections [Fig.6.15]. Zircons are an important host for HFSE and MHREE. The large amount of rutile also make this mineral an important potential host for HFSE such as Nb and Ta and to a lesser extent, V and Mn which might be hosted as exsolution of ilmenite laths within rutile crystals [Fig.6.15b]. Some rare grains of monazite have been collected during mineral separation. Monazite is the main host phase for U and Th, LREE and MREE as well as P in these alkali rocks where apatite is not present. Finally, titanite is also a common mineral which is the major host for titanium on the edge of the intrusion.

6.3.5. Isotopic data

6.3.5.1. U-Pb dating

Mineral separation for zircons was successfully completed for 6 samples: two trondhjemites, one meladiorite, two granitoids (amph-qtz diorite) and one amphibole trondhjemite. All samples are relatively zircon-rich. Crystals are mostly euhedral and yellowish or pinkish but have variable size and shapes depending of their occurrence. Zoning observed by cathodoluminescence does not show any inherited cores and suggest that these zircons are magmatic [Fig.6.21]. Zircons extracted from trondhjemite pod (R1) are large 200-400 μm grains which are well-developed and show oscillatory and sector zoning. In the trondhjemite dyke of Les Dalmates (LD1), zircons are thinner and prismatic and have a length varying between 50 and 150 μm . These zircons generally have a large homogeneous core and oscillatory zoning only appear on the edge of the crystal. In granitoids, zircons range from 100 to 400 μm in size, being slightly smaller in the darker ilmenite-rich rock. These grains have clear euhedral shape and show rare amphibole inclusions. Most grains of the leucodiorite (PRB39) show a homogeneous core with an oscillatory zoning on the edge of the grain whereas the meladiorite (PRB38) is lacking such feature. Both groups of zircons have also a significant proportion of grains (5-10%) which have complex convoluted zoning which suggests the presence of some hydrothermal agent which has interacted with these rocks. Zircons collected in a large pod of alkali diorite (PRB70) show elongated prismatic grains between 50 and 200 μm in size, generally relatively dark and homogeneous in CL, with a bright thin rim occurring in many grains. Finally, the zircons collected in the granitoid dyke (RP40) crosscutting the transition zone of Rivière des Pirogues are rounded grains with rare sector zoning and have diameter between 50 and 150 μm [Fig.6.21].

U-Pb ages obtained on these zircons were calculated on an average of 12 to 18 spots collected in core, rims and convoluted areas. All ages are consistent within samples and zoning or convoluted internal features do not significantly affect isotopic ratios. Trondhjemite pod (R1) and dyke (LD1) give ages of 53.9 ± 0.5 Ma [Fig.6.3a] and 54.5 ± 0.4 Ma, respectively.

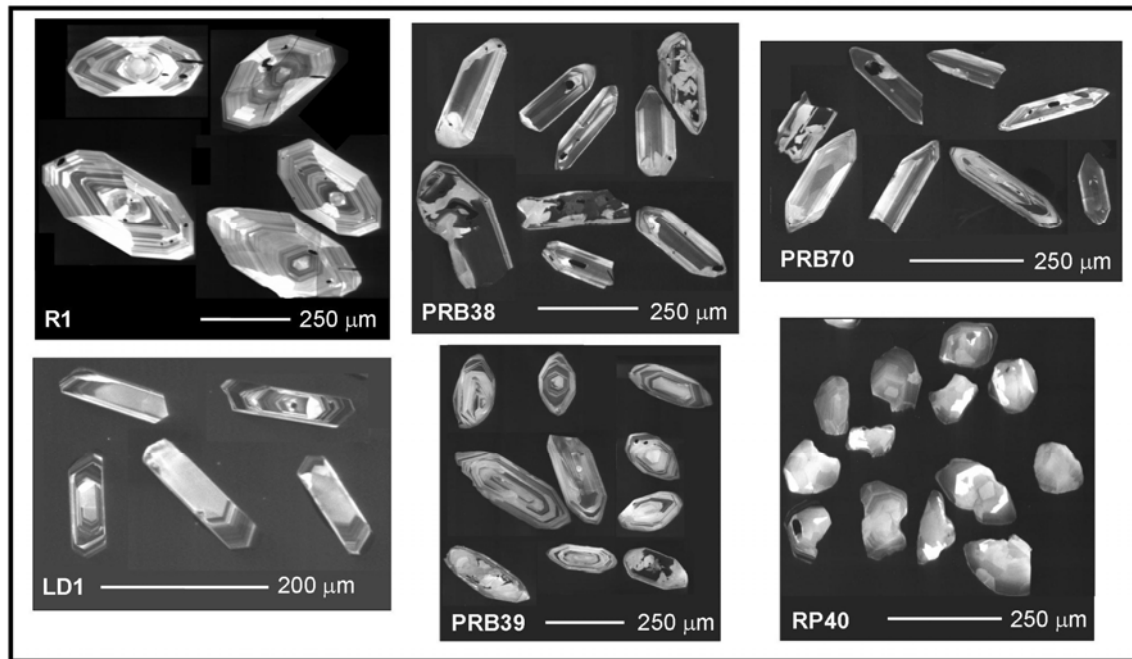


Fig.6.21. Cathodoluminescence (CL) images of zircon crystals from sample separated for zircons. R1 & LD1: trondhjemite [Fig.6.3a], PRB38 & PRB39 mingled meladiorite and leucodiorite [Fig.6.2d], PRB70 alkali quartz diorite [Fig.6.2e] and RP40 quartz diorite dyke in the transition zone [Fig.6.8b].

The latter, as well as the meladiorite (PRB39) show a large mean square weight deviation (MSWD) which could indicate the presence of a mixed zircon population separated by a less than a couple of million years. Granitoids which have mingled boundaries gives ages of 54.1 ± 0.7 Ma (PRB39) and 54.7 ± 0.5 Ma (PRB38) [Fig.6.2d]. The alkali diorite pod of Rivière bleue (PRB70) [Fig.6.2e] is dated at 54.1 ± 0.6 Ma and the amphibole gabbro dyke crosscutting the transition zone (RP40) [Fig.6.8b] provides an age of 55.0 ± 0.6 Ma [Tab.6.1, Fig.6.22].

Sample	Lithology	(n)	U (ppm)	Th (ppm)	Th/U	$^{238}\text{U}/^{206}\text{Pb} \pm 2\sigma$	$^{207}\text{Pb}/^{206}\text{Pb} \pm 2\sigma$	Age $\pm 2\sigma$ (Ma)
PRB38	Meladiorite	14	579	303	0.46	117.41 ± 2.07	0.048 ± 0.002	54.7 ± 0.5
PRB39	Qtz-Leucodiorite	17	243	173	0.56	117.87 ± 2.93	0.051 ± 0.006	54.1 ± 0.7
PRB70	Na-amph diorite	13	480	284	0.57	118.67 ± 1.87	0.05 ± 0.003	54.1 ± 0.6
RP40	Leucodiorite	12	218	105	0.36	116.87 ± 2.31	0.052 ± 0.009	55 ± 0.6
R1	Trondhjemite	20	119	29	0.27	116.48 ± 3.37	0.069 ± 0.013	53.9 ± 0.5
LD1	Trondhjemite	21	1138	246	0.26	115.41 ± 1.66	0.05 ± 0.004	54.5 ± 0.4

Tab.6.1. Isotopic data on zircons of felsic rocks of the Massif du Sud.

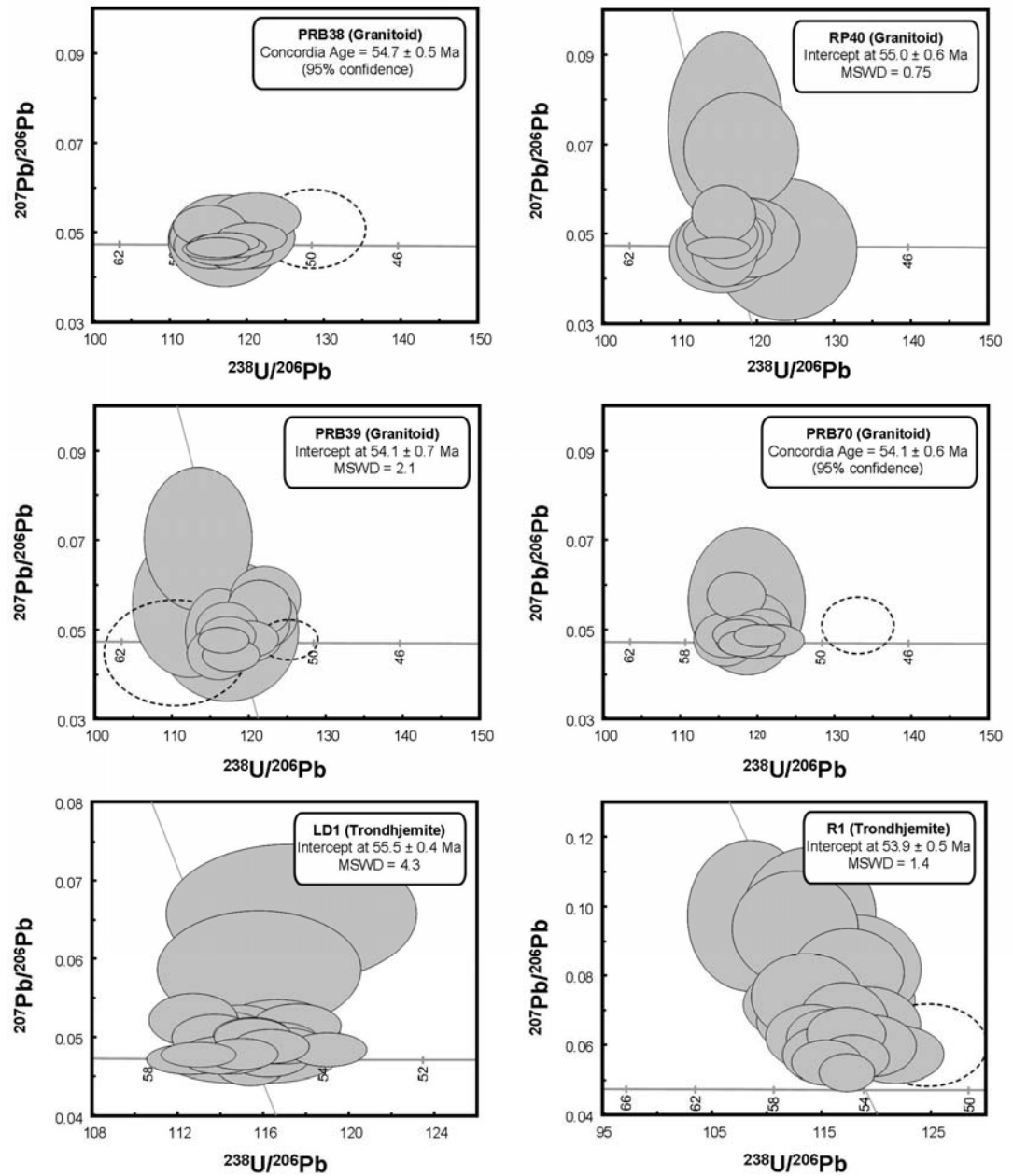


Fig.6.22. Tera-Wasserburg diagrams for U-Pb data. Ages are defined by their location on the concordia or by the intercept of concordia with the regression line to common Pb. Error ellipses represent 2σ . Data plotted as dashed ellipses are excluded from the age calculation. Numbers on the concordia lines are in millions of years.

6.3.5.2. Hafnium isotopes

Zircons and rutiles from a trondhjemite pod (R1) were analysed by T.Ewing to obtain $^{176}\text{Hf}/^{177}\text{Hf}$ as part of a pioneering analytical technique for in situ analysis in rutile (Ewing et al., 2010). Rutiles generally do not show compositional zoning and/or multiple generations of growth were seldom observed in the grains. However, it is common to see exsolution by laths or blobs of ilmenite and this was

avoided during analyses by examination of BSE images [Fig.6.15b]. The average content of hafnium in rutile is 49ppm and 1.1% in zircons. With the exception of the spot size which was smaller for zircon than rutile, an identical analytical setup was used and provides results in a similar range to values which confirm the idea that these two accessory phases are formed during the same magmatic event. ϵ_{Hf} which was obtained on zircons is $+11 \pm 1$ whereas rutile gave a value of $+12 \pm 6$ calculated assuming an age of 54 Ma.

6.4. Discussion

6.4.1 Evidence of a late magmatic event

There are several lines of evidence that the igneous rocks described above were emplaced during a magmatic event which occurred after the main building phase of the ophiolite. This hypothesis is based on textural, mineralogical and isotopic data.

The determination of U-Pb isotopes from zircons of several felsic intrusions provide an early Eocene age (55.5 – 53.9 Ma) for these plutons. Previous isotopic ages on some felsic dykes of the Massif du Sud have been obtained by Cluzel et al., (2006) and give slighter younger ages (53.2 – 49.6 Ma). The differences between the two age groups and the possibility that mixed population of zircons can exist could indicate that large intrusions were emplaced slightly before the felsic dykes analysed by Cluzel et al. (2006).

Although, absolute timing on geodynamic events in New Caledonia is scarce, there is a large amount of relative dating based on tectonic features, field observations and the palaeontological record which confirms that the series of intrusive rocks present in the Massif du Sud have a pre-obductive origin. The age of the formation of the oceanic crust which forms most of New Caledonia remains undetermined but is likely to be in the second part of the Mesozoic era (Aitchison et al., 1995; Cluzel et al., 2001). At the time of the magmatic events studied in this chapter, contemporaneous isotopic ages exist for eclogites of the Pouébo terrane in the north (Spandler et al., 2005) and an Eocene age was also calculated for the Loyalty-d'Entrecasteaux volcanic arc (Eissen et al., 1998; Cluzel et al., 2001;

Crawford et al., 2003) which is now covered by Neogene coral reefs. The obduction itself, which marks the end of an oceanic environment for the Massif du Sud, is around 34 Ma (Cluzel et al., 2001), clearly post-dating the intrusions of Mouirange.

The absolute age which is obtained for this series of rock is in agreement with field observations. All the rocks related to this magmatic event are clearly independent of obduction features and bear no relationship with the basement. A clear sequence from Yaté to the Montagne des Sources - Mouirange area shows that pyroxenite dykes are present in the deepest parts of the ophiolite [Fig.6.6]. Higher in the section, plagioclase starts to appear in the core of some of these pyroxenite dykes [Fig.6.8] and in the same area, large trondhjemite pods start to occur, with similar pyroxenite and hornblendite wall-rock reactions [Fig.6.8, 6.3]. Just below the crust-mantle transition, large intrusions of granitoids are emplaced [Fig.6.2] whereas above the palaeo-Moho, only dykes of granitoids cross-cut through the ultramafic cumulates [Fig.6.8b]. On the western flank of the ophiolite, a variant sequence occurs where pyroxenite dykes evolve to amphibole-rich dykes and eventually to hornblendites *s.l.* in Baie des Pirogues and Ile des Pins [Fig.6.4, 6.5]. Such lateral variations related to the ophiolite column are the result of an emplacement in an oceanic environment, when the ophiolite was a coherent oceanic crust with a constant thickness.

The granitoids possess the right mineral assemblage to use the Al-geobarometer for amphiboles (Schmidt, 1992) which is inapplicable for trondhjemites and south-west K-feldspar-absent granitoids. Temperatures have been calculated at $800\pm45^{\circ}\text{C}$ (Blundy & Holland, 1990) to correct pressure data (Anderson & Smith, 1995). Mouirange granitoids provide pressures from 1.9 ± 0.4 kbar for alkali diorites in the dunite zone to 3.5 ± 0.6 kbar for leucodiorite in the uppermost harzburgite. Granitoidic dykes in the transition zone of Rivière des Pirogues give pressures between 2.6 ± 0.1 and 3.1 ± 0.4 kbar. The position of the granitoids provides an approximate depth for the palaeo-Moho in the fore-arc and shows that the upper ophiolitic sequence (gabbros, basalts, sediments) was not scrapped off at the time of emplacement. This hypothesis is consistent with U-Pb ages as the denudation of the 12 to 15 km thick oceanic crust is likely to have started with the obduction of the crust on New Caledonia during the Oligocene.

Pyroxenite dykes of Yaté area clearly show that they can crosscut previous dunite channels and earlier pyroxenites which are related to the building of the oceanic crust [Chap.5]. In addition to the lack of foliation textures in these dykes, the characteristics of emplacement such as high discordant angle, sharp changes in the direction of propagation [Fig.6.6a] and the absence of preferred orientation (Prinzhofer et al., 1980) suggest that pyroxenite dykes were emplaced in a brittle harzburgitic mantle. Such rheological behaviour is compatible with a fore-arc mantle which has cooled down since the formation of its crust a few tens of millions years earlier. Similar observations can be made for amphibole-bearing dykes of Tontouta, Dumbéa, GR2H and Baie des Pirogues. Felsic rocks also provide evidence that the mantle and the crust were cold during their emplacement by showing numerous dykes of granitoids in the transition zone [Fig.6.8b], crosscutting cumulate layers without any strong wall-rock reaction with the exception of serpentinisation. Finally, in opposition to the St. Louis and Koum-Borindi granitic intrusions which are post-obductive events (Cluzel et al., 2005), all the features in the Massif du Sud ophiolite do not cross-cut the basement and have to be pre or syn-obductive as it was reported for some plagiogranites of Oman (Briqueu et al., 1991).

Interactions between the large felsic intrusions of Mouirange and the surrounding ultramafic rocks are sharp and have a limited depth of metasomatism penetration. It is generally restricted to a band of pyroxenite at the interface between the two rocks and in some cases, the injection of a fluid into the host rock, leaving dykelets of orthopyroxenite [Fig.6.6c]. Trondhjemites and granitoids also display peridotite xenoliths and magmatic breccia which show pyroxene or amphibole reaction zone in small intrusions [Fig.6.3c] or on the edge of larger plutons (Rodgers, 1973a). The reaction zones and the angular aspect of the enclaves are also indicative that the mantle was not in a plastic state at the time of emplacement. In areas where amphibole is a dominant phase (Baie des Pirogues, Ile des Pins), more diffuse boundaries occur, leading to a metasomatic replacement in the area surrounding some veins [Fig.6.4b, 6.5b]. Additionally, the presence of successive events of magmatic injection in those zones [Fig.6.2c, 6.4a,e,f, 6.5f] gives a more complex picture for the relative timing of the events and could be related to the age scattering in the early Eocene. However, the coherent topology between all these magmatic pulses show that these rocks are

posterior to the final emplacement of the mantle (Ile des Pins) or the dunite zone (Baie des Pirogues).

The presence of amphibole as a common primary phase in mafic and ultramafic environment, added to the brittle emplacement features suggests that the peridotite was relatively cold during the emplacement of this series of rocks. The formation of hornblende in the reaction zone of trondhjemite pods and the extensive amphibole metasomatism in Ile des Pins support the idea of a lithospheric mantle significantly colder than during previous magmatic events which have led to dunite or pyroxenite reaction zones. Amphibole is also a common primary phase in the mafic and felsic intrusions of Mouirange. The fore-arc environment coupled with the composition of these intrusions also suggests that this magmatic event produced hydrous magmas which are likely to be colder than typical mantle melts. Colder temperature allowed these rocks to crystallise amphibole in contrast to the layered crustal gabbros which, despite being hydrous, are unable to saturate in hydrous phases [Chap.4, 5].

6.4.2 Relationship between pyroxenites, hornblendites and granitoids.

The series of rocks described in this chapter represents the development of a distinct magmatic sequence in the mantle section of an ophiolite. The presence of large amounts of evolved melt is an unusual feature illustrated by the size and the variety of interactions which occur in the Massif du Sud fore-arc. Such series of rocks are rarely mentioned in other ophiolites (Briqueu et al., 1991; Peters & Kamber, 1994; Cox et al., 1999) and the New Caledonia ophiolite stands as an exception where a felsic magmatic event is conserved in a convergent margin oceanic environment. The relationship between the discordant rocks in the ultramafic rocks is supported by field observations, microscopic textures, geochemical data and isotopic dating. This rock series provides new understanding on the processes that lead to crust formation as all these magmas, from gabbros to granites, are present in the mantle and are not the result of crustal processes.

A main feature common to all the rocks intruded during the fore-arc history is that the preservation of the series is completely unaffected by the foliation of the harzburgite and the late plastic deformation which has some effect on the mantle

rocks previously mentioned [Chap.4, 5]. The Eocene magmatic event has generally pristine microscopic textures and crosscuts most of the previous magmatic activity developed during the nascent arc stage such as dunite channels, chromitite pods or transition zone cumulates. The modal mineralogy gradient from the east to the west suggests a relationship between pyroxenites, trondhjemites, granitoids and finally the amphibole-bearing mafic rocks of the west coast.

The major element composition of the mineral phases present in these rocks is a key argument linking these various lithologies. Orthopyroxene is particularly useful as it is a mineral present in all rocks, being a major, accessory or a reaction phase with the host rock. Chemical trends observed for orthopyroxene are clearly different from the dunite-pyroxenite-gabbro trend presented in the chapter 5. With the exception of rare samples which have an ambiguous position, most discordant rocks can be assigned clearly to the crustal gabbro trend [Chap.5] or to the fore-arc trend on the basis of alumina and chromium content as a function of Mg# [Fig.6.10]. The trend observed in clinopyroxene for the same elements as well as calcium also shows distinct behaviour from the rocks related to the arc environment [Fig.6.11]. Similarly, Mg-number also shows a trend related to the felsic character of the rocks. Ultramafic dykes tend to have relatively high values whereas granitoids and trondhjemites have lower values. Hornblendites *s.l.* mainly cover the gap between the two groups. The observation that Mg-number is relatively unrelated to the modal composition of the rock and that different groups overlap to a great extent suggests that although fractional crystallisation can have a local effect, the whole series is not the result of a single magmatic process [Fig.6.24].

This variation and overlap of compositions between lithologies is also mirrored by the anorthite content in the plagioclase which extend up to An₉₈ in some amphibole-rich ultramafic rocks and has a more progressive variation to hornblendites (An₉₅₋₇₄), amphibole gabbros (An₈₉₋₆₂), granitoids (An₇₃₋₉) and trondhjemites *s.l.* (An₁₇₋₀) which also contain small amounts of K-feldspar [Fig.6.13]. The low amount of potassium in this series of rock is a key feature to distinguish fore-arc magmatism from the post-obduction intrusions of St. Louis and Koum-Borindi which are rich in biotite and orthoclase (Cluzel et al., 2005).

In terms of modal composition and major elements, this series of rocks is also easily discriminated from other discordant rock by the presence of amphibole. With the exception of pyroxenite dykes s.s. which only contain amphibole as exsolutions or minute grains, this hydrous phase is a major component in most of the other rocks. Variation in Mg-number follows the same trend as for other minerals decreasing in more felsic rocks. Calcium content decreases steadily with decreasing Mg-number whereas Na_2O and Al_2O_3 content show an increase which suggests the presence of two trends within the series [Fig.6.12, 6.13]. In pyroxenites, amphibole-bearing ultramafic rocks and hornblendites, sodium content increases strongly with changing amphibole mode. In large intrusions such as amphibole gabbros and granitoids, an increase is also noticeable but is less pronounced and reflects a different magmatic process and the more evolved character of the granitoids.

Finally, trace element contents provide additional proof of a magmatic event which is clearly different from the other discordant features. The concentrations of Ti, Zr and Cr in orthopyroxene follow a trend which is not observed in any other mantle rocks [Fig.6.16]. Rare earth element patterns in clinopyroxene are distinct, showing a convex upward pattern for all the rocks of this series whereas crustal cumulates only display a continuous increase from LREE to HREE [Fig.6.17]. Fractionation of MREE with respect to the other REE is a typical feature which relates pyroxenite dykes of Yat  to the amphibole-bearing dykes and pods of the western coast. Plagioclases also provide support for a genetic link between amphibole-rich rocks and felsic intrusions with similar REE patterns which show progressive LREE fractionation and high Sr and Ba contents [Fig.6.19, 6.20].

6.4.3. Melt compositions and evolution

Intermediate and felsic intrusions

The Mouirange area and the surroundings (Lake Yat , Riv re des Pirogues and Col de Prony) show a large variety of magmatic intrusions which are misleading as to the range of initial parental melts in the area. Granitoids and trondhjemites provide the best idea of the composition of the percolating melt. Their bulk composition shows a low-K series of rocks with variable enrichment in sodium and frequent but accessory quartz saturation. In amphibole gabbrodiorite,

which can be interpreted as the most primitive melt composition sampled, the presence of considerable amounts of clinopyroxene associated with anorthite-rich plagioclase shows that a calcium-bearing source or enrichment has to be suspected in the genesis of these rocks. The observation of strongly zoned plagioclase with An_{85} core surrounded by An_{50} overgrowth and interstitial plagioclase [Fig.6.2c] suggests that the sodium-rich composition is a secondary injection of magma in a partially crystallised mush which has made most of the granitoids in Mouirange.

Several indicators in terms of stratigraphy and cross-cutting relationship clearly show that these rocks are felsic intrusions emplaced in the mantle of a fore-arc environment. The geodynamical context prior to the obduction phase limits the source of these rocks to slab-melt or oceanic plagiogranites (Coleman & Peterman, 1975). However, the latter hypothesis which involves the formation of large felsic intrusions by immiscibility of a basaltic melt (Lippard et al., 1986), the melting of the mafic crust or the melting of mantle peridotite by the deep penetration of seawater-derived hydrothermal fluids (Koepke et al., 2004; Amri et al., 2007) are not supported by the observations made in New Caledonia. Plagiogranite pods generally occur at higher stratigraphical levels (isotropic gabbros) [Fig.B.2] and would not be conserved or accumulated under the dunite zone. Moreover, although plagiogranites are frequently sodium-rich magmas, they do not show complex association of different melts and are generally quartz-saturated even for the most primitive melts (Rollinson, 2009). Trace element content in the felsic rocks of New Caledonia also show that Mouirange granitoids are at least 10x more enriched than plagiogranites and have a high LREE enrichment which is absent from flat REE pattern in oceanic granites (Gerlach et al., 1981).

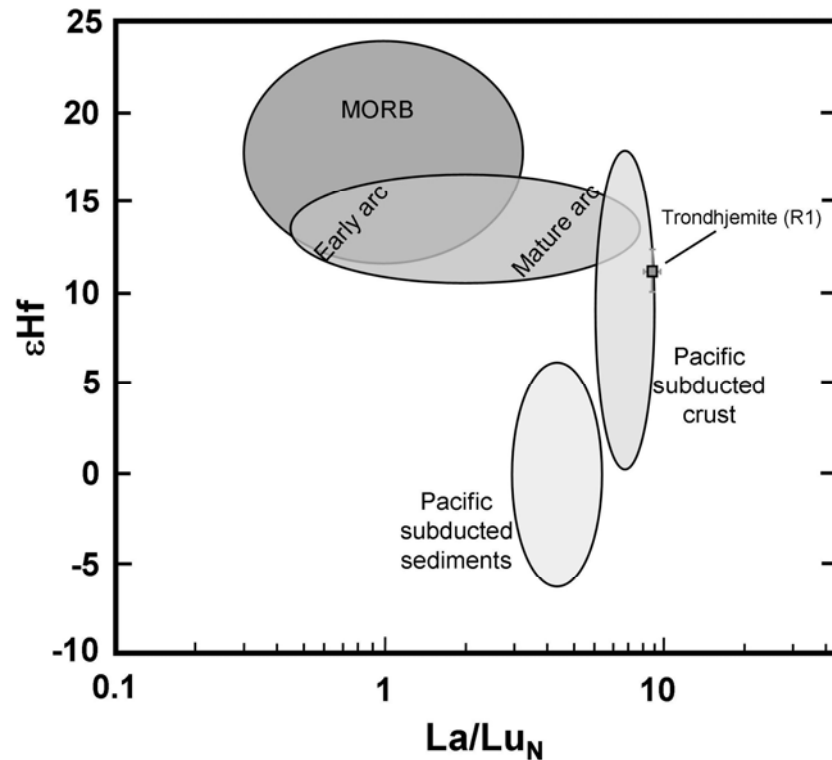


Fig.6.23. Hafnium isotopes of a trondhjemite pod (R1) and its normalised La/Lu content (C1-chondrite, Sun & McDonough, 1989). A comparison with the general composition for igneous and sedimentary rocks for the Pacific region is given (Plank & Langmuir, 1998; Pearce et al., 1999).

The modal composition of the felsic rocks with Na-rich plagioclase and amphiboles and the presence of numerous accessory phases suggest that this series of intrusions is directly related to a subducting slab and contrast with rocks previously described in the Massif du Sud [Chap.4, 5] which are strongly interacting with the host rock. As mentioned earlier, the strong fractionation in trace elements ($\text{La/Yb} = 8.5$), generally low HFSE and enrichment in LILE is in support for a slab signature as is also suggested by a study on mantle granitoids of Oman (Rollinson, 2009). Hafnium isotopes obtained on a trondhjemite pod (R1) ($\epsilon_{\text{Hf}} = 11$) are also efficiently discriminating these rocks from the MORB field ($\epsilon_{\text{Hf}} > 13.5$) and are in the lower range of arc-related isotopic compositions ($\epsilon_{\text{Hf}} = 16-11$). This value suggests a mantle melt strongly contaminated by Pacific subducted sediments [Fig.6.23] (Plank & Langmuir, 1998; Pearce et al., 1999).

A study of the Mourange area by K.A. Rodgers (1973a) reported that the pluton is zoned and characterised by an inward evolution from a hornblende-rich meladiorite to quartz leucodiorite and a granodiorite core. According to this author, the annular structure is related to increasing peridotite assimilation by a calc-

alkaline melt on the edge of the intrusion. However, although peridotite xenoliths are observed on the margin of some diapirs, the mingling textures observed between mela- and leuco-diorite [Fig.6.2d,e] do not support the idea that Mouirange granitoids are the result of a single magmatic injection event with extensive margin interactions. Nevertheless, the high Mg-number and the high Ni, Co, Cr and V contents clearly indicate that all felsic melts have interacted with the sub-arc mantle at some point during their ascent into the fore-arc mantle [Fig.6.24, 6.26].

Although whole-rock analyses are necessary to obtain a better understanding of the origin and the formation process of these rocks, an adakitic trend is clearly noticeable in many rocks of this series. Adakites (Defant & Drummond, 1990) are generally characterised as arc magmas with the composition of hornblende andesite, dacite or rhyolite where plagioclase and hornblende are the main phases, and pyroxenes are less frequent than in other arc-related rocks. The low K_2O/Na_2O of the Mouirange plutons is indicated by the rarity of K-feldspar and the composition of the amphibole. Mg-number between 0.45 and 0.8 can also be estimated on amphibole and pyroxenes crystals. These two chemical features are typical characteristics of adakite-related rocks such as Archaean TTG, sanukitoids & setouchiites and bajaites where slab melts have considerably interacted with the surrounding mantle to produce hybrid magmas (Martin et al., 2005; Tsuchiya et al., 2005; Xiong et al., 2006; Rapp et al., 2010).

The genetic theory of adakitic melts (Martin et al., 2005) suggests that high-silica adakites (HSA: 60-75% SiO_2) are melts which are transferred directly from the slab with a significant proportion of assimilated mantle material. On the other hand, low-silica adakite (LSA: 50-60% SiO_2) are interpreted as the melting of a metasomatised mantle producing rocks which are particularly characterised by higher Mg-numbers and high Sr, Ti and K/Rb. The features present in the granitoids and amphibole gabbros of Mouirange indicate that LSA is a likely origin for the formation of these rocks. The presence of pyroxenes, high Mg-number and high ilmenite content are all giving support to this hypothesis. A comparison done for NiO in olivine and REE in clinopyroxenes of phenocrysts present in the high-Mg andesite lavas of Setouchi, Japan (Tsuchiya et al., 2005) show similarities in trends which would confirm that Mouirange granitoids belong to the group of low-silica adakites (Tatsumi et al., 2003) [Fig.6.9, 6.17].

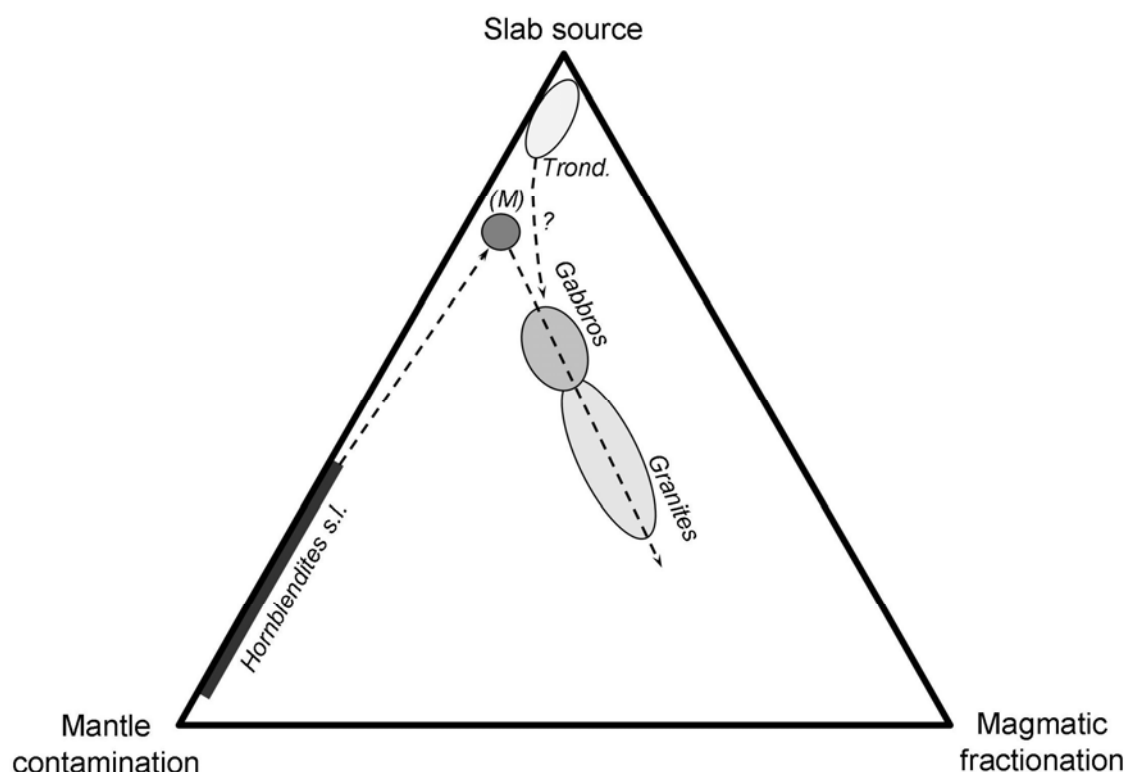


Fig.6.24. Schematic representation of the three main processes occurring in fore-arc magmatism.

Slab source are melts directly produced by the melting of the subducted slab. Mantle contamination can indicate assimilation of mantle component during the ascent of the melts or could indicate the melting of a metasomatised mantle. Magmatic fractionation covers all fractionation processes which occur during magmatic differentiation.

The composition of the trondhjemites is interpreted here as a pristine slab melt with various contamination by the surrounding mantle. Hornblendites *s.l.* are considered as the interaction of this slab melt with the surrounding mantle forming amphibole-rich ultramafics. Gabbro dykes and pods (M) occurring in hornblendites *s.l.* share similarities with the hydrous melts produced by the melting of metasomatised peridotite in piston-cylinder experiments [Chap.2]. From these parental mafic melts, fractionation could produce the felsic series. However, the influence of mantle assimilation and the mixing with trondhjemite might produce similar trend and this preliminary dataset is unable to provide such information distinguish these processes.

Highly sodic melts accumulated in the mantle are thought to be only achieved by the direct transfer of melts from the slab to the uppermost mantle (Martin et al., 2005). However, experimental work has shown that alkali melts can be formed in the mantle in water-saturated conditions (Prouteau et al., 2001) [Chap.2]. Most of the slab melt chemical features are conserved but a strong ultramafic imprint is also noticeable for temperatures above the wet solidus of a metasomatised peridotite [Chap.3]. In New Caledonia, it remains unclear if trondhjemite pods are related to HSA or are the result of metasomatised mantle melting. The highly discriminative SiO_2/MgO and K/Rb ratios strongly suggest that trondhjemite are emplaced by the direct transfer of slab melts (Martin et al., 2005)

[Fig.6.25]. However, excessively high nickel content (~650ppm in plagioclase) suggests intense interaction with the mantle, supported by the strong modal metasomatism on the edge of the pods and the presence of partially reacted peridotite xenoliths within [Fig.6.3, 6.24].

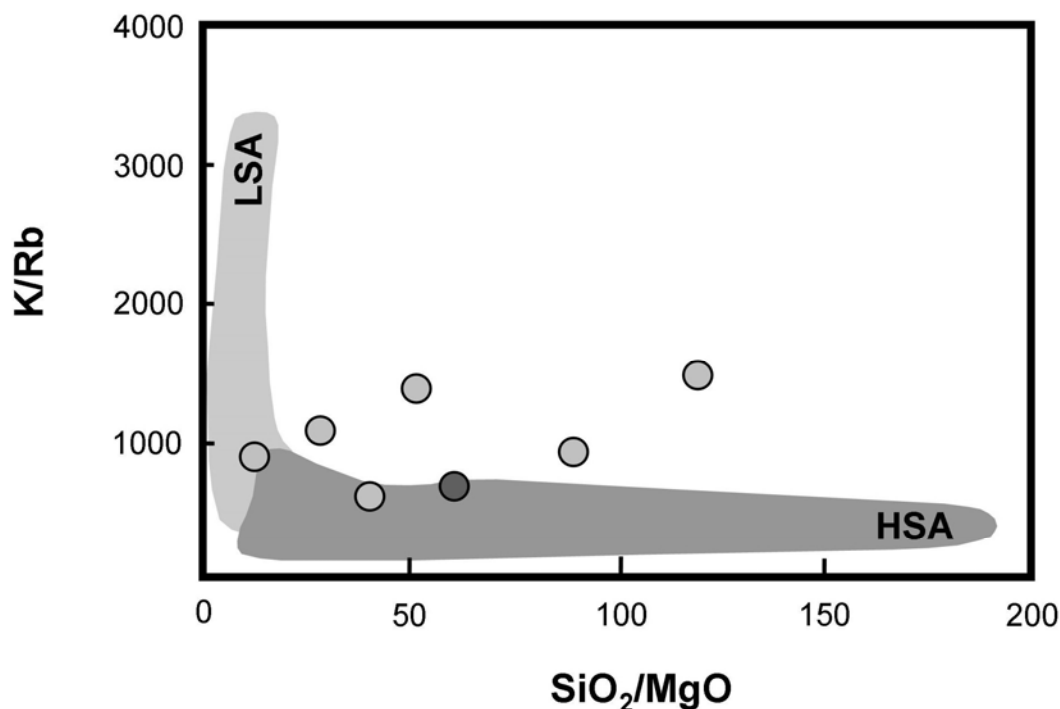


Fig.6.25. K/Rb versus SiO₂/MgO in felsic rocks of the Massif du Sud. The dark grey circle is a trondhjemite pod from Col des Dalmates (LD1) and light grey circles are whole-rock analyses of granitoids in the Massif du Sud (Cluzel et al., 2006).

Experimental studies (Rapp et al., 1999; 2010) [Part A] help to understand the interactions between felsic slab melts and the surrounding depleted mantle. At the contact between the Mouirange pluton and the harzburgite, it is common to observe fibrous orthopyroxene growing perpendicular to the contact. This unusual crystal habit for orthopyroxene is also seen in veins intruding the host rock and has been observed in xenoliths at Lihir and Tubur, Papua New Guinea (McInnes et al., 2001; Grégoire et al., 2001), Monglo and Iraya, Philippines (Arai et al., 2004; Grégoire et al., 2008) and in southern Kamchatka (Ishimaru et al., 2007). Such orthogonal growth for orthopyroxene is also described in layered experiments of chapter 3 where a fibrous orthopyroxenite layer is formed at the contact between the slab melt and the olivine [Fig.3.3]. The major element composition of the orthopyroxenite veins present at Mouirange are very similar to orthopyroxenite

layers formed during experiments with particularly low Al_2O_3 as a consequence of felsic melt interaction with the depleted mantle (Rapp et al., 2010) [Chap.3]. REE patterns which are obtained on the fibrous orthopyroxenite veins of Mouirange are extremely fractionated and show a general LREE depletion with a slight La enrichment compared to Ce. Although the enrichment in MREE in volcanic xenoliths is not as extreme (Grégoire et al., 2001, 2008), features are mostly identical with the exception of Eu negative anomaly which only occurs in Mouirange due to plagioclase saturation. Layered experiments with or without the formation of garnet also display MREE enrichment with low LREE and no europium anomaly [Chap.3].

Among the pyroxenes which are formed during metasomatic reactions in the New Caledonia ophiolite, a few samples have low- Al_2O_3 and CaO compositions which are off the main trend. A comparison with experimental studies on interaction between granitic melt and a depleted mantle where emphasis is put upon melt:rock ratio (Rapp et al., 1999) shows a perfect correlation between experimental and natural pyroxenes. For low melt:rock ratios, such as the contact surface between a dyke or an intrusion and the surrounding mantle, Al_2O_3 and CaO will remain low in the orthopyroxenes. However, in Baie des Pirogues and more obviously in Ile des Pins, we have evidence that melt has percolated through the surrounding rocks leading to a larger surface of interaction, effectively increasing the melt:rock ratio. The metasomatic orthopyroxenes which are formed in these pervasively metasomatised rocks have higher alumina and calcium as predicted by the experimental data [Fig.6.10a,b].

Amphibole-rich mafic rocks

The general field characteristics of Baie des Pirogues show that cumulate dunite and other ultramafic rocks are cross-cut by pyroxenites, hornblendites and sodium-rich amphibole diorites and gabbros. Close field and microscopic examination [Fig.6.4d-f] of the samples shows that the amphibole present in mafic and ultramafic rocks crystallised as a late phase (e.g. a late injection of magma in an already partially crystallised mush). This hypothesis is also confirmed by the injection of hornblendite dykelets in the core of pyroxenite dykes [Fig.6.4c] and the textures of amphibole-bearing websterites pods. Textures clearly suggest that an initial melt had crystallised pyroxenes as an early phase and started to accumulate

in pods or dykes where zoned interstitial amphibole has cemented the subidiomorphic anhydrous crystals [Fig.6.4f]. Although trace element geochemistry lacks in our dataset to support the presence of two episodes of melt injection in the area, the textures are very similar to some xenoliths of the adakitite intrusion of Monglo where interstitial clinopyroxenes are infiltrating a dunite (Grégoire et al., 2008). The melt composition which precipitated clinopyroxenes in these xenoliths is inferred as being a calc-alkaline basaltic melt. It is possible that lower temperature conditions or higher dissolved water contents in New Caledonia may have formed amphibole rather than clinopyroxene as an interstitial phase.

The Baie des Pirogues area is also known to host Al-Ti-rich chromite-bearing pyroxenite dykes which are thought to be formed by the percolation of an evolved melt in oxidising conditions. These dykes can contain up to 11.5 ppm of Pt and 0.8 ppm of Pd as sulfides, alloys and sulfosalts associated with the spinel phase (Augé & Maurizot, 1995). The early interpretation to explain the massive enrichment in PGE involves the mixing of two magmas leading to increased oxygen fugacity and lower temperatures. Comparison with the Leka ophiolite in Norway has suggested that the PGE-Cr-bearing magma has a boninitic affinity (Pedersen et al., 1993). Ore-geology studies associated the deposit of Baie des Pirogues to an Alaskan-type mineralisation (Irving, 1974; Debari et al., 1987; DeBari & Coleman, 1989) based on the striking similarities for the assemblage and the crystallisation sequence of PGE-phases in the spinels (Johan et al., 1989; Nixon et al., 1990). These observations and the interpretation of their origin seem to support the hypothesis that amphibole-rich mafic rocks are formed by two slightly different magmas which also led to the PGE mineralisation.

Finally, in Ile des Pins, the late metasomatism which has affected Baie des Pirogues cumulate sequence and intrusions by forming an interstitial phase of hornblende composition is here the only metasomatic event. Amphiboles formed in these ultramafic rocks are also clearly subduction-related amphiboles and have relatively high sodium and aluminium contents (Coltorti et al., 2007). The harzburgite host rock as well as the composition of pyroxenes and rare olivine in the hornblende surrounding dykes of hydrous gabbros suggests that the depleted mantle which forms most of the New Caledonia ophiolite is strongly metasomatised by localised porous flow in Ile des Pins. The small amount of olivine in these wall-rocks illustrates that a reaction developed in chapter 2 between slab melts and a

depleted mantle (melt + olv => opx + amph + phl + aq.fluid) is taking place in the natural environment. However, the highly sodic composition of the metasomatic agent in New Caledonia does not produce biotite as a metasomatic phase.

Pyroxenites s.s. and amphibole-bearing pyroxenite dykes

The modal composition and the chemistry of the minerals present in pyroxenite dykes show a typical low- Al_2O_3 - Cr_2O_3 -CaO trend. Such pyroxenites have been observed in several peridotite xenoliths where it is classically interpreted as the remnant of melt or fluid transfer in a sub-arc mantle (McInnes et al., 2001; Berly et al., 2006). Emplacement at low-temperature which is suggested for these dykes is supported by previous studies, which suggest temperatures around 1000°C based on pyroxene geothermometers (Berly et al., 2006). The formation of these pyroxenites is considered to be induced by the percolation of aqueous fluid in a cold mantle (Kelemen et al., 1992; Morishita et al., 2003; Berly et al., 2006). However, experimental work shows that a hydrous silica-rich melt can be directly related to the formation of low- Al_2O_3 pyroxenite veins (Rapp et al., 2010) [Chap.3]. Experimental runs show that a slab-melt in contact with a depleted peridotite forms pyroxenites with similar characteristics to those of Yaté dykes over a large range of temperatures (800-1000°C). In New Caledonia, this hypothesis is confirmed by the presence of plagioclase-cored pyroxenite dykelets [Fig.6.8c,d] and evidence is also found in peridotite xenoliths of Tallante, Spain where quartz diorite veinlets have pyroxenite walls (Arai et al., 2003).

The Massif du Sud ophiolite shows a dichotomy between the eastern flank of the ultramafic nappe where orthopyroxenite dykes evolve into segregated rocks of felsic compositions and the western side where amphibole-bearing veins are more likely to be found. Concordant ages and chemical compositions suggest that these magmatic trends originate from a single magmatic source. Similar evidence is found in adakite lavas from Monglo, Philippines (Grégoire et al., 2008) where dunite xenoliths contain orthopyroxenite and hornblendite veins in separate blocks from the same intrusion. The interpretation given to this xenolith suite is based on an evolution of melt-mantle interaction related to host-rock assimilation, despite having the same initial metasomatic slab melt [Fig.6.26]. The amphibole present in the pyroxenite dykes of the western coast in New Caledonia have a composition which also lies in the field of Coltorti et al. (2007) and Prouteau & Scaillet (2003)

for subduction-related mantle amphibole which confirm the relationship with a subducting slab for the fore-arc series.

6.4.4. Spatial distribution of magmatic activity in the ophiolite

The large size of the Massif du Sud ophiolite provides the opportunity to study metasomatism in a relatively cold fore-arc mantle in three dimensions. The melt-peridotite interactions can be followed for 3 km in depth and over a large surface of $\sim 500 \text{ km}^2$. Hypotheses which are often based on the study of xenoliths (Grégoire et al., 2008) become observational facts when large coherent portions of the mantle can be sampled.

The distribution of the pre-obductive discordant rocks [Fig.6.1] shows that pyroxenite dykes are present in deeper levels of the ophiolite, away from the transition zone. A progressive development of felsic rocks within the dykes to form large felsic plutons is noticeable and the latter are accumulated below the crust-mantle transition. This sequence indicates that hydrous melts are probably transferred from deeper zones of the mantle to the fore-arc through pyroxenite dykes [Fig.6.6], which can develop by hydraulic fractures in a cold and brittle mantle. As the uppermost mantle is reached, the transferred melt is not always fully extracted and can remain in the core of some dykes [Fig.6.8]. The rheological and density change created by the crust-mantle transition or the dunite zone prevents melts being propagated any further and these felsic magmas tend to accumulate at sub-crustal levels where they form large intrusions such as Mouirange [Fig.6.2]. In this uppermost part, the melt raises mainly by magma stopping as it is shown by the presence of peridotitic enclaves [Fig.6.3c], magmatic breccia (Rodgers, 1973a) and the presence of pyroxenite dykelets fracturing the host rock [Fig.6.6c]. Rare cases where adakitic magmas reached the surface in the fore-arc zone, show that the serpentinised mantle is fragmented in a similar fashion, by pyroxenite or hornblendite veins to form xenoliths within the magma (Grégoire et al., 2008).

On the western flank of the ophiolite, melts are not segregated efficiently from the surrounding rocks. This results in the formation of amphibole-metasomatised dunite (Baie des Pirogues) and harzburgite (Ile des Pins) which suggest lower temperature confirmed by experimental data on amphibole-bearing

peridotite [Chap.2]. As this reaction consumes the hydrous melt, only aqueous fluids could escape such system and could be the source of hydrothermal activity in the fore-arc environments [Fig.6.26].

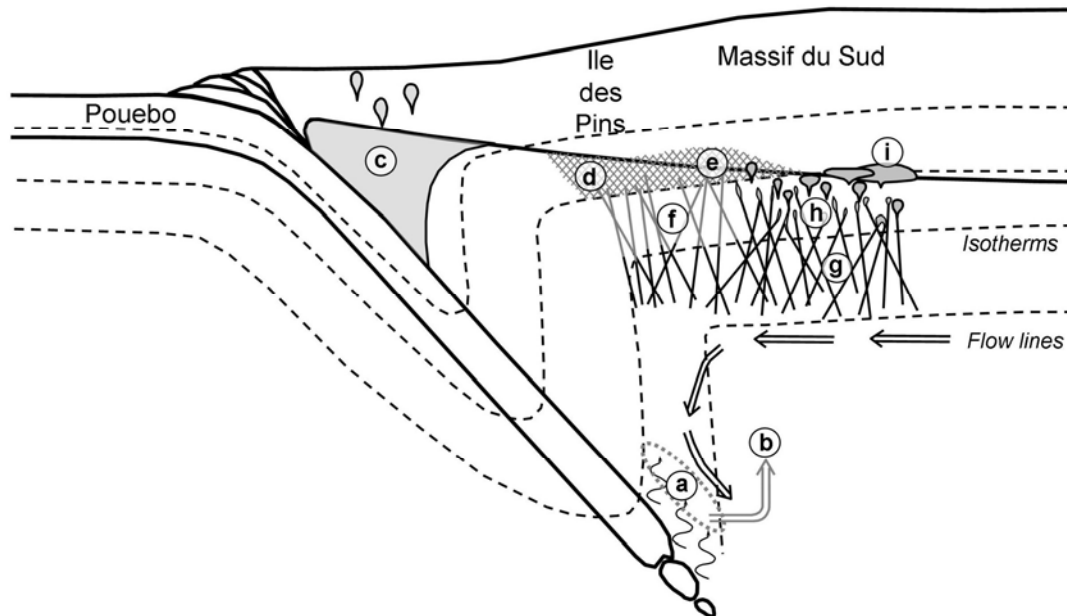


Fig.6.26. Schematic representation of the Early Eocene fore-arc zone in the New Caledonia subduction zone.

a. Slab melting of the young hot subducted oceanic crust forming a metasomatised mantle.
b. The metasomatised mantle is dragged to hotter part of the fore-arc mantle wedge where melting occurs.
c. Serpentinised mantle and serpentine diapirs initiated by the low-pressure dehydration of the subducted crust (not observed in New Caledonia).
d. Ile des Pins metasomatised harzburgite in hornblende.
e. Baie des Pirogues amphibole impregnation of the fore-arc transition zone.
f. Amphibole-bearing dikes of Tontouta, Dumbéa, Baie des Pirogues and Ile des Pins.
g. Pyroxenite dykes of Yaté.
h. Pyroxenite with plagioclase core around Mouirange pluton.
i. Mouirange plutons emplaced below or in the transition zone.

Structural and geophysical studies suggest that the obduction of the ophiolite on the New Caledonia basement occurred from the North to the South as a result of the opening of the Loyalty basin and the presence of the Norfolk ridge in the subduction zone (Prinzhofer et al., 1980; Collot et al., 1987, 1988). Reconstruction of the fore-arc area prior to the obduction would indicate that the south-west part of the Massif du Sud contains rocks which are closer to the subduction trench than rocks on the east coast [Fig.6.26, 6.27]. Here, we interpret the general distribution of the metasomatised rocks as a consequence of temperature gradients in the fore-arc. Dykes and intrusions from Yaté to Mouirange represent an area of the fore-arc where the mantle wedge is at a temperature above the solidus of slab melt + harzburgite system. In consequence,

melts aggregate and form large intrusions with a high temperature reaction zone ($\geq 800^{\circ}\text{C}$) with the surrounding ultramafics. Closer to the trench, water could be more abundant and lower temperature helps to form amphibole in dykes. However, the mantle wedge hosting the rising melts is at a temperature below the solidus of the metasomatised peridotite and melt-mantle reaction consumes the ascending magma by forming large amounts of amphibole. If the ultramafic nappe was conserved further west, it could be expected that intense serpentinisation would be observed and melt-mantle interaction could be similar to the amphibole-metasomatised serpentinised peridotites which have been described in Monglo (Grégoire et al., 2008).

The presence of small pockets of melt in all these metasomatised zones and further away from the trench as large intrusions is recorded in New Caledonia and provides new information for changes in the composition of hydrous melts emplaced within the fore-arc. The general evolution shows that plagioclase becomes more calcic on the western coast, closer to the trench, whereas amphibole shows the opposite trend and is more sodic in Baie des Pirogues and Ile des Pins. Potassium is only present as K-feldspar in rocks of Mouirange and seems to be absent in the near-trench environment.

6.4.5. Timing of magmatic events in New Caledonia

Fore-arc magmatism in New Caledonia is characterised by a relatively well constrained age for the crystallisation of granitoids. This work shows that U-Pb ages on zircon are between 53.9 to 55.5 Ma similar to the 53 Ma pre-obductive event measured on the same suite of intrusions by Cluzel et al. (2006). This collection of ages over a relatively small time range suggests that intense magmatic activity was occurring between 55 and 52 Ma in an oceanic environment between the Norfolk ridge and the South Loyalty basin.

Reconstruction of the position of the ophiolite considering that obduction was from north to south suggests that the peridotites in the Massif du Sud were north-east of New Caledonia, in a position which was above the subducted oceanic plate now preserved as the HP-LT complex of Diahot and Pouébo [Fig.4.3, Fig.6.27, Tab.4.2]. Early Eocene ages of 55 Ma have been obtained on zircon cores of the Pouébo terrane in eclogite metapelites (Spandler et al., 2005).

A local volcanic origin is suspected to explain their highly constrained age variation considering that sediments from the Norfolk ridge give a much larger time span (Aronson & Tilton, 1971; Aitchison et al., 1998). The presence of the Massif du Sud ophiolite in close proximity at this time and the abundance of magmatic rocks in sub-crustal position with similar ages provide a very likely source for the volcanogenic sediments which have been dated in the eclogite.

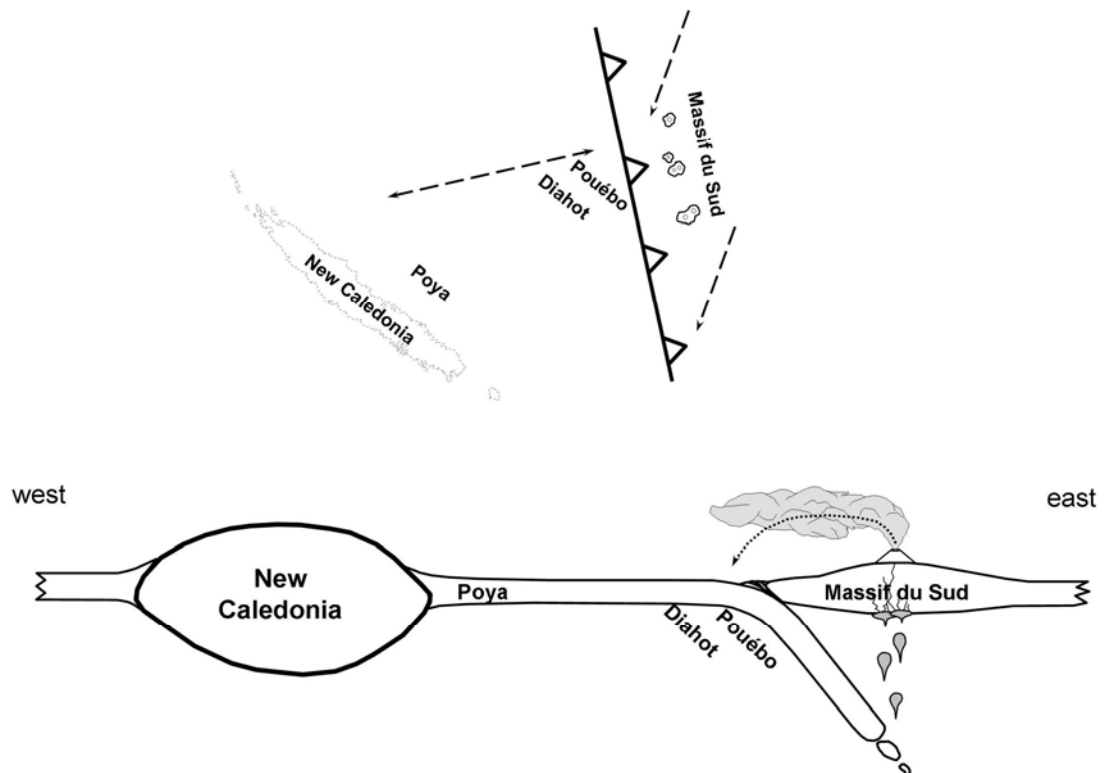


Fig.6.27. Local palaeogeographic reconstruction of the eastern basins of New Caledonia during the early Eocene (55Ma). Poya-Diahot-Pouébo terranes form different parts of the South Loyalty basin. The initiation of subduction in the New Caledonia subduction zone creates slab melting in the fore-arc and a volcanic event above the Massif du Sud. Igneous zircons are ejected and accumulated in the sediment pile of the oceanic trench, incorporated to the future Pouébo terrane. Relative motion of the different terranes is given in the upper figure. The South Loyalty basin obduction (Poya) and exhumation (Diahot, Pouébo) is mainly a reversal of the plate movement in the Eocene. The oceanic crust hosting the ultramafic terrane has a north to south motion which makes the obduction relative to the Norfolk ridge more southerly.

The geochemical data presented in this chapter show that granitoids are related to the melting of the subducted oceanic plate. The classical view for the production of slab melts involves the subduction of a young and hot oceanic plate into a relatively hot mantle (Defant & Drumond, 1990, Peacock, 1996; Kincaid & Sacks, 1997). Short-lived subduction is suggested by Spandler et al. (2005) to combine the theoretical length of the subducted slab with the delay between peak

eclogite metamorphism (44 Ma) and the entrance of the protolith into the subduction zone (55 Ma). However, field observations in the Massif du Sud show that the emplacement of the granitoids imposes a relatively cold overlying mantle which is not compatible with the standard conditions for adakite formation. In the light of the recent studies on the subducted slab, it is known that melting conditions for the oceanic crust are generally reached and that partial slab melting will occur even if all the conditions mentioned above were not met (Hermann & Spandler, 2008; Hermann & Rubatto, 2009; Plank et al., 2009).

Comparisons with palaeogeographic reconstructions of the south-west Pacific during the Early Cenozoic suggests that the 55 Ma magmatic event is induced by the initiation of subduction in the South Loyalty Basin (Schellart et al., 2006). This model would explain the presence of the “Massif du Sud” oceanic crust in a geodynamical environment where subduction is initiated on its western margin [Fig.B.4, B.5]. Moreover, it would also explain that volcanism in this fore-arc environment could produce zircons which would be transported westward and be deposited in the sediment of the future Pouébo terrane [Fig.6.27]. Other models suggesting the subduction of an oceanic ridge within the South Loyalty basin (Whattam et al., 2008; Whattam, 2009) fail to explain the observations made about the emplacement of the granitoids. Although asthenospheric flow combined with a subducting ridge could produce slab melting, such a setting would create a proto-fore-arc crust and magmatism in an oceanic crust which had no time to cool down and to form a relatively thick lithospheric mantle as it is observed for the Massif du Sud ophiolitic section in the early Eocene.

6.5. Conclusions

This preliminary study of the late intrusives in the ophiolite section of New Caledonia shows that a wide variety of intermediate and felsic rocks occur in the mantle. The general consensus which suggests that arc granites are the result of crustal fractionation is confronted here with observations of granitoids within the mantle. The geodynamics of the south-west Pacific during the Tertiary and the U-Pb ages obtained for these rocks show that the Early Eocene magmatic events of the Massif du Sud predate obduction and are unrelated to the main building phase of the oceanic crust.

The large size of the ophiolite provides new insights on melt transport in the mantle and its spatial distribution relative to depth and temperature gradient. Such information was lacking from the study of xenoliths where topological relations are often missing. The Massif du Sud ophiolite shows that pyroxenite dykes are the main feeding conduits for melts and that slab-related melts tend to accumulate under the crust-mantle transition. The lateral variations show that in colder part of the fore-arc, felsic melts interact with the host ultramafic rocks to form an amphibole-metasomatised mantle.

The composition of the granitoids and the associated rocks shows a strong relationship with adakites and slab-melting petrology. Preliminary data remain unclear if the Early Eocene magmatic event in New Caledonia is directly a product of slab-melting or if a metasomatised peridotite is involved in the process. Additional studies will be needed to decipher the true origin and the melt transport mechanisms which played a role in this event. Several aspects will have to be developed in further research such as hafnium isotopes, mineral trace element analyses and whole-rock analyses in order to develop a consistent geological model for fore-arc magmatism in New Caledonia.

B.2. Conclusions of Part B

Study of the magmatic rocks present in the Massif du Sud of New Caledonia has provided the first detailed petrologic and geochemical description of this ophiolite. Careful description of field relationships, trends in the mineral compositions of different rocks and trace element and isotopic compositions of specific rock types show that three main events have occurred during the oceanic history of the ophiolite. These three steps in the building of the oceanic crust are an initial stage in a spreading environment, the development of a nascent arc which has formed most of the ophiolite crustal sequence and finally, an important fore-arc magmatic event.

The initial formation of the oceanic crust is supported by the highly refractory character of the mantle section. The subsequent magmatic events cannot be fully responsible for the depletion of the harzburgite which necessitates an unreasonable amount of melting for a single event. The structural studies of Prinzhofer et al. (1980) coupled with the presence of sub-concordant channels and dykes within the ophiolite support the existence of a ridge with an E-W orientation [Chap.4]. Although little information is available at this stage on the features of this earliest event, it is likely that the Massif du Sud primary crust was formed between Gondwana and a Cretaceous subduction zone which lay around 1000 km east of the continental margin (Matthews et al., 2010) [Fig.B.4]. A precise age for this event is unavailable at this time and Rb-Sr age obtained on a harzburgite of Dumbéa (730 Ma, De Roé, 1964 *in* Paris, 1981) seems unrealistic in the geodynamic environment of the south-west Pacific.

A large part of the Massif du Sud is related to formation in an arc environment. The general structure of the ophiolite containing an ultra-depleted mantle, an extensive dunite zone, and the early appearance of pyroxenes and gabbro-norite in the cumulate sequence indicates a supra-subduction zone setting [Chap.4]. In addition, the study of discordant channels, pods and dykes show for the first time that in the uppermost part of a sub-arc environment, primitive arc melts are transferred through channelized flow. Field relationships and geochemistry show that hydrous boninitic basalts percolating in the dunite channels evolve and precipitate pyroxenes prior to their intrusion into the lower arc

crust. The clear relationship between these discordant rocks and the crustal mafic cumulates provides evidence that primitive arc melts produced in the locus of partial melting in the mantle wedge, are not pristine melts when they accumulate as magma chambers into the lower crust [Chap.5].

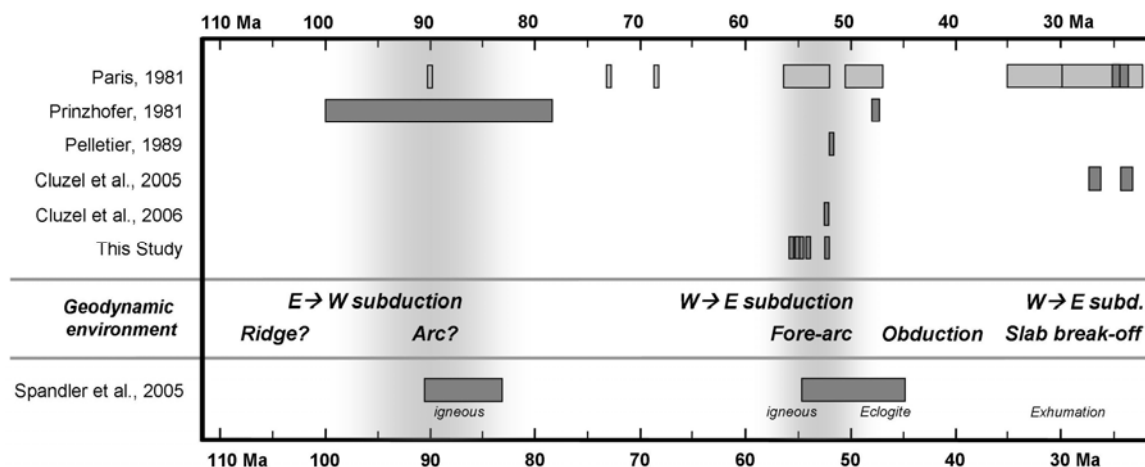


Fig.B.3. Comparative table of age obtained on rocks of the Massif du Sud.

Data from Paris (1981) are a collection of mostly unpublished data (U.Auckland, B.R.G.M, U.Cleveland) which are presented in this work. Data were collected during the sixties and seventies and are subject to scrutiny. Data from A.Prinzhofer (1981) and B.Pelletier (1989) are K-Ar data on mafic dykes and Cluzel et al (2005, 2006) are U-Pb data on zircons and titanite of felsic dykes and intrusions. The data collected by Spandler et al. (2005) on zircons of the Pouébo eclogite is given for comparison.

The field observations and geochemical data indicate that the nascent arc was formed in a non-rigid mantle but where plastic deformation is not intense. These observations suggest that a subduction zone was initiated near the Massif du Sud oceanic crust in a relatively close time-frame after its ridge history. Geodynamic reconstructions (Schellart et al., 2006) suggests that the Massif du Sud was near the Chatham subduction zone during the Cretaceous and was affected by subduction-related melts for a short time in its history before slab roll-back of the subducting Phoenix-Pacific margin [Fig.B.4, B.5]. A Cretaceous age would coincide with the data obtained on igneous zircon cores which are found in the Pouébo eclogite (Spandler et al., 2005). It also supports the poorly described K-Ar age data of A.Prinzhofer (1981) which ranges from 100 to 77 Ma and a 90 Ma age for mafic dykes by H.Bellon included in Paris (1981) [Fig.B.3]. However, this dataset remain incomplete and new information on the different pre-Cenozoic events of the ophiolite remains essential to position the Massif du Sud in the south-west Pacific during the Cretaceous.

The preliminary study of the granitoids of Mouirange and the associated rocks has provided data for a new view on fore-arc magmatism and the behaviour of slab-melts in the mantle wedge. The description of the series of intrusions and the associated mafic and ultramafic rocks (hornblendites *s.l.* & pyroxenites) show that deep slab-melt transport is ensured by the numerous pyroxenite dykes. At shallower depths, the temperature and the rheology of the peridotites induce the accumulation of magma diapirs leading to the Mouirange granitoids. Low-temperature conditions on the western side of the ophiolite have produced metasomatised ultramafic rocks whereas at higher temperature, a large variety of magmas (gabbros to trondhjemites) is emplaced at sub-Moho levels. The fore-arc environment of New Caledonia provides the first three-dimensional evidence that complex assemblages of melts can be produced by several petrological processes in the sub-arc mantle and are not restricted to crustal levels [Chap.6].

The presence of granitoids allows accurate U-Pb dating and produce similar ages (55-52 Ma) to previous studies (Cluzel et al., 2006). Similar K-Ar ages were obtained on undetermined gabbros (Paris, 1981; Pelletier, 1989) [Fig.B.3]. The Early Eocene igneous activity corresponds to the formation of the Loyalty arc by the subduction of the New Caledonia oceanic plate. This subduction occurred a few millions years earlier than generally considered by geodynamic reconstruction models (Schellart et al., 2006; Whattam et al., 2008; Whattam, 2009). After this short lived subduction which seeded zircons into the protolith of the Pouébo terrane, the New Caledonia basement was subducted, leading to a jamming of the subduction zone and the obduction of the Massif du Sud around 34 Ma (Cluzel et al., 2001). Subduction was then re-initiated behind the Norfolk ridge and temporarily subducted the New Caledonia basin before the final slab break-off which led to the St. Louis and Koum-Borindi magmatism during the Oligocene [Fig.B.3, B.4, B.5].

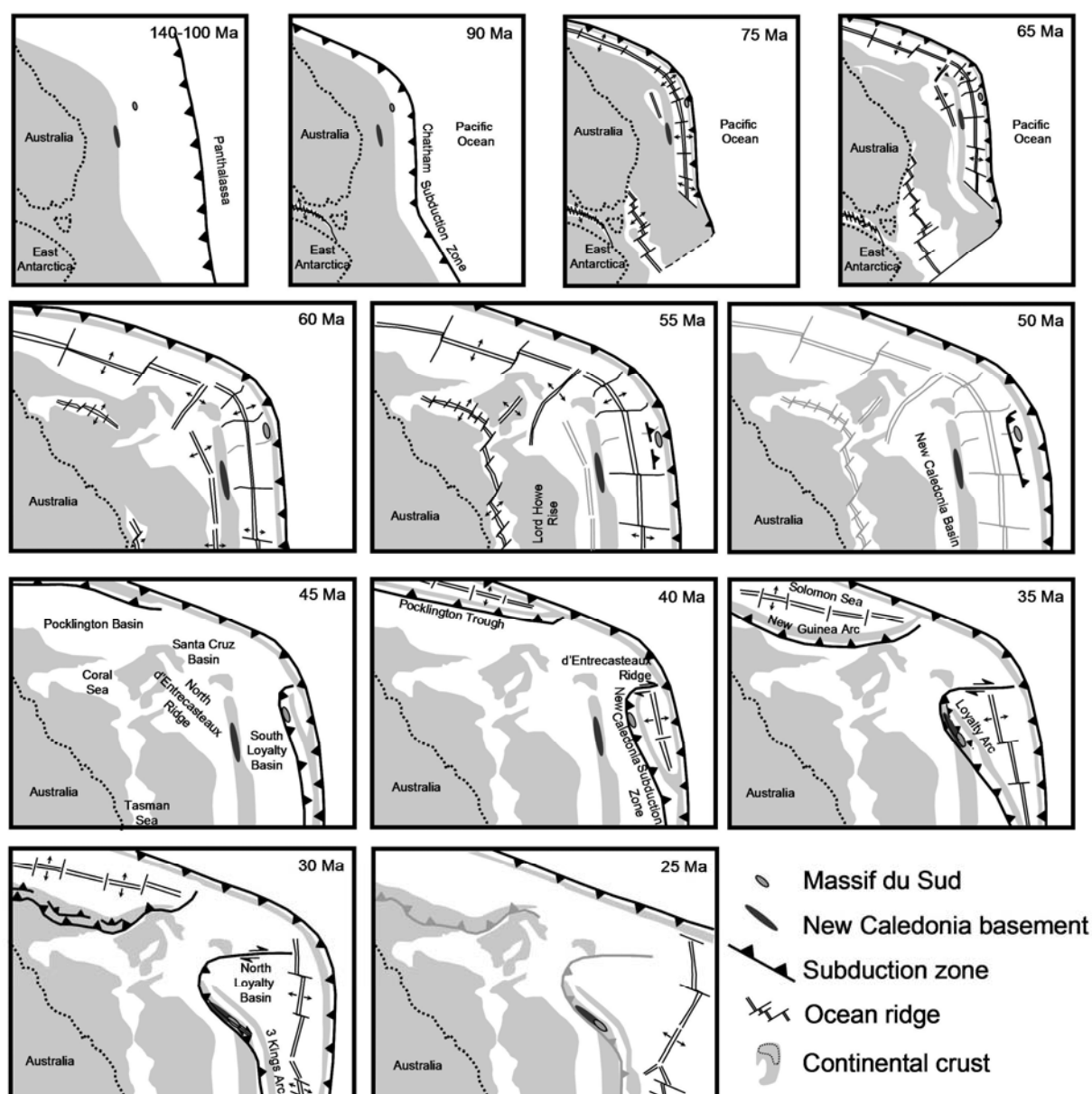


Fig.B.4. Reconstruction of the south-west Pacific region from the Cretaceous period to the Oligocene epoch. New Caledonia ante-Senonian basement is represented by the dark grey ellipse whereas the inferred position of the Massif du Sud oceanic crust is given by a light grey ellipse. (Modified from Schellart et al., 2006; Matthews et al., 2010).

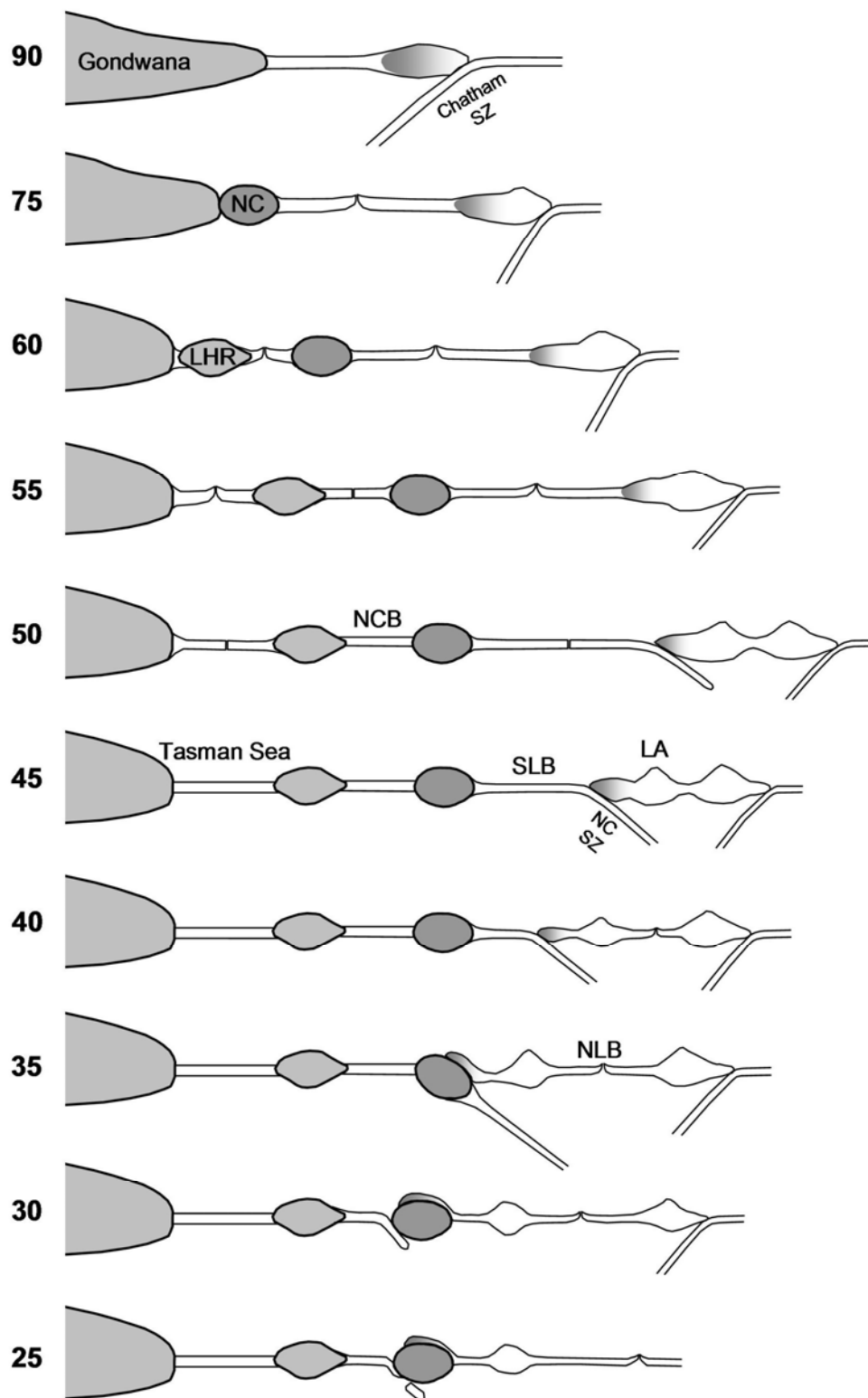


Fig.B.5. East-West cross-sections illustrating the evolution of the south-west Pacific region from the Cretaceous to the Oligocene. Chatham SZ : Chatham Subduction zone; NC : New Caledonia; LHR : Lord Howe Rise; NCB : New Caledonia Basin; SLB : South Loyalty Basin; NCSZ : New Caledonia Subduction zone; LA : Loyalty Arc; NLB : North Loyalty Basin. The dark grey ellipse is New Caledonia ante-Senonian basement; the shaded oceanic/arc crust is the inferred position of the Massif du Sud (modified from Schellart et al., 2006).

7. General conclusions

The underlying theme of this thesis is the melt transport processes and the transfer of slab melt or primitive arc basalts through the mantle wedge. Although the transportation of melt is relatively well-understood at mid-ocean ridges and spreading oceanic environments, the modalities of this transfer remain elusive in the sub-arc mantle. The mantle wedge remains the least understood part of the subduction factory and slab to crust transfer is generally simplified as an ultramafic enrichment of slab-related melts under an arc. The transfer of melt between the slab and the locus of partial melting has been investigated with an experimental study of different types of interactions between slab melts and olivine whereas transport of partial melts in the sub-arc mantle has been studied in the New Caledonia ophiolite.

Experimental petrology and investigations of natural rocks provide evidence that a large range of interactions between melts and peridotite can occur in the mantle wedge. These melt-mantle interactions are divided into two main processes of focused flow and porous flow. Despite major differences such as melt composition (slab melt and primitive arc basalts), pressure (35 kbar and 3-5 kbar) and host rock (dunite – various ultramafic and mafic rocks), many common features are observed between experimental work and field observations in the Massif du Sud.

Focused flow is the main mechanism of melt transport in the uppermost mantle. The hot-stage history of the Massif du Sud ophiolite [Chap.5] does not show any evidence of widespread pervasive porous reactive flow. Focused transport is achieved through a network of channels and dykes. Beside the initial replacement stage, which leads to channel propagation or wall growth, further melt-mantle interactions are insignificant. Arc melts are efficiently transferred from the locus of partial melting to the crust-mantle transition using this network. In the region just overlying the subducted slab, similar observations have been made using high-pressure experiments [Chap.3]. The evidence presented shows that shielding is also occurring and that focused flow is the most efficient process to transfer the slab signature from the subducted oceanic plate to the locus of partial melting. The silica-saturated trondhjemite dykes of New Caledonia show that the

shielding by pyroxenite walls can allow the melts to travel a long distance in the mantle, despite the strong disequilibrium with ultramafic rocks. The transfer times which are expected in such focused melt transport from the slab to the arc are expected to be short and in agreement with previous models and isotope disequilibrium data.

The percolating melts remain relatively unaffected by this process of transfer as minimal interaction is taking place between the ultramafic host-rock and the silica-rich melt. Specific trace elements that are abundant in the mantle wedge (Ni, Co, Cr, potentially Os) and major element ratios (Mg#) variations are taking place but generally have a small impact on the incompatible element ratios of the melt. Both environments of study have also shown that modification of P-T conditions can affect the melt itself and the crystallisation of solid phase can occur as phenocrysts or as cumulates within a melt conduit cross-cutting the mantle [Chap.3, Chap.5]. However, the evolution of the melt does not alter incompatible element ratios and K_2O/H_2O and H_2O/Ce slab thermometers can still be applied in those cases of focused melt transport.

The implication of porous reactive flow within the mantle was studied in detail in the 25-45 kbar range. Experimental studies have shown that the infiltration of slab-melt through a refractory peridotite leads to the formation of a stable alkali-rich amphibole as well as phlogopite in a range of pressures where such phases were not expected. The formation of a metasomatised peridotite has a strong influence on the output fluid which loses most of its slab-related features [Chap.2]. A multi-stage process, which could transfer the slab signature would considerably lengthen the transfer time between the slab and arc volcanoes and is hardly compatible with U-Th disequilibrium series. In the Massif du Sud fore-arc environment, porous reactive flow leading to the formation of amphibole is observed in cold areas where the melt infiltrating the harzburgite or the dunite crystallises a sodium-rich amphibole [Chap.6]. These observations confirm that amphibole can potentially play a fundamental role in hosting slab signature in the sub-arc mantle from the deepest part to sub-crustal peridotites. The metasomatic agent which has produced amphibole-bearing peridotites is confirmed in both cases as a hydrous melt as it has been shown that aqueous fluid would only have a minor influence in this system. Geochemistry and field evidence also show that the melting of a metasomatised peridotite produces a mafic melt, which is similar

to the initial slab/arc melt in terms of incompatible trace elements. However, such a melt has an alkaline character and thus it seems not likely that this is a major and unique process to produce primitive arc basalts [Chap.2, Chap.6]. In addition, such melt would not reflect the initial slab trace element/water ratio and slab thermometers cannot be applied in the case of porous reactive flow.

The range of pressure and temperature conditions in the sub-arc mantle and the position relative to the slab bring some variation in melt transport and the composition of the slab/arc melts. Experimental study has shown that an amphibole-phlogopite-bearing peridotite is formed at lower temperature whereas at higher temperature, amphibole and then phlogopite break down [Chap.2]. In the convergent environment of the New Caledonia ophiolite, regional zoning is also visible as focussed flow seems to occur at depth and relatively high temperature whereas lower thermal conditions will induce the formation of metasomatised peridotites [Chap.6]. The impact that these variations have on the composition and the evolution of the melts are still poorly known and several geochemical gradients observed in convergent margins can be related to such P-T variations in the mantle. Although we have evidence that primitive arc melts are transported through focused and replacive channel flow [Chap.5], additional experimental work is required to claim that such process also occur between the slab and the locus of partial melting. Field observations in New Caledonia have shown that a large variety of melts can potentially reach the surface in an arc environment [Chap.6]. The combination of pervasive melt-mantle interactions and focused transport can provide a base to explain natural observations. Variation within slab melt compositions were already studied by Hermann & Spandler (2008) but end-member processes such as partial melting, mantle assimilation, fractional crystallisation, melt mixing and melt segregation have to be considered in order to understand the composition of arc magmas.

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Appendices

Part A

Table S1. Composition of olivine

Exp	C2933 LHK	C2945 LHK	C2951 LHK	D953 LHK	C2910 MHK	C2940 MHK	D891 MHK	D979 MHK	C3082 LLK	C3076 LLK	D965
P (Gpa)	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
T (°C)	950	1000	1050	1100	950	1000	1050	1100	800	850	900
n	3	3	3	3	3	3	5	3	3	4	8
SiO ₂	41.26 (23)	41.85 (82)	41.80 (77)	41.08 (8)	40.75 (39)	39.85 (27)	41.66 (10)	41.70 (30)	41.21 (34)	41.38 (26)	41.68
TiO ₂	tr.	tr.	tr.	tr.	0.06 (10)	tr.	tr.	0.01 (2)	0.10 (14)	0.03 (6)	0.02
Al ₂ O ₃	tr.	tr.	tr.	tr.	tr.	tr.	tr.	0.04 (11)	0.06 (3)	tr.	0.04
FeO	8.36 (44)	9.20 (85)	9.13 (33)	8.74 (97)	9.50 (13)	10.69 (32)	10.58 (86)	9.09 (19)	9.42 (96)	9.95 (66)	9.78
MnO	-0.06 (14)	0.15 (34)	0.23 (21)	0.18 (16)	0.14 (28)	0.18 (3)	0.22 (23)	0.10 (12)	0.07 (24)	0.09 (5)	0.04
MgO	49.80 (22)	50.25 (48)	50.04 (47)	50.02 (56)	49.06 (40)	47.22 (28)	50.74 (13)	50.29 (34)	49.54 (7)	49.11 (26)	49.54
CaO	tr.	0.06 (11)	0.00 (16)	0.04 (7)	0.03 (3)	0.06 (13)	tr. (3)	0.03 (7)	0.08 (15)	0.07 (4)	0.07
Na ₂ O	-	-	-	-	-	-	-	-	-	-	-
K ₂ O	-	-	-	-	-	-	-	-	-	-	-
Sum	99.25 (67)	101.39 (89)	101.09 (168)	99.95 (28)	99.46 (37)	97.82 (102)	103.04 (101)	101.16 (54)	100.25 (44)	100.50 (77)	101.10
Si	1.01 (0)	1.00 (1)	1.01 (1)	1.00 (0)	1.00 (1)	0.98 (0)	0.99 (2)	1.00 (1)	1.00 (1)	1.01 (0)	1.01
Ti	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Al	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Fe	0.17 (1)	0.19 (1)	0.18 (0)	0.18 (2)	0.20 (0)	0.20 (0)	0.18 (2)	0.18 (0)	0.19 (2)	0.20 (1)	0.20
Mn	0.00 (0)	0.00 (1)	0.00 (0)	0.00 (0)	0.00 (1)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (1)	0.00 (0)	0.00
Mg	1.82 (2)	1.80 (2)	1.80 (1)	1.81 (2)	1.80 (1)	1.74 (1)	1.79 (4)	1.80 (1)	1.79 (0)	1.78 (1)	1.78
Ca	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00
Na	-	-	-	-	-	-	-	-	-	-	-
K	-	-	-	-	-	-	-	-	-	-	-
Σcat	2.99	3.00	2.99	2.98	3.00	2.92	2.96	2.99	2.99	2.99	2.99
Mg#	0.914 (4)	0.904 (4)	0.907 (2)	0.911 (10)	0.902 (2)	0.897 (3)	0.909 (9)	0.908 (2)	0.904 (9)	0.898 (6)	0.900

Table S1. Composition of olivine (continued)

LLK	D963 LLK	C3083 MLK	D971 MLK	D969 MLK	C3085 MLK	D1078 MLK	D1064 MLK	D1077 MLK	D1073 UM	D1052 UM	D1058 UM
	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
	950	800	850	900	950	975	1000	1050	650	700	800
	3	3	2	3	3	7	3	6	3	7	3
(23)	41.12 (22)	40.89 (53)	42.46 (142)	41.26 (47)	40.38 (29)	41.58 (22)	43.53 (298)	40.70 (77)	39.25 (83)	40.86 (130)	41.61 (162)
(13)	0.01 (7)	tr.	0.12 (9)	0.14 (11)	tr.	tr.	0.02 (13)	-	0.17 (10)	0.09 (10)	0.04 (9)
(5)	tr.	tr.	0.08 (3)	0.04 (15)	tr.	tr.	0.24 (19)	tr.	0.07 (7)	1.32 (196)	0.56 (83)
(16)	9.94 (41)	10.33 (69)	9.41 (30)	9.54 (49)	10.09 (46)	8.12 (46)	9.04 (138)	6.97 (65)	13.03 (200)	11.34 (96)	9.96 (24)
(12)	-	0.19 (23)	0.05 (2)	0.11 (18)	0.17 (7)	0.12 (8)	0.21 (10)	0.13 (5)	0.32 (8)	0.22 (9)	0.09 (8)
(37)	49.14 (38)	47.61 (115)	49.22 (227)	48.68 (28)	47.83 (46)	50.17 (90)	45.44 (262)	49.61 (68)	43.91 (93)	44.65 (226)	47.20 (122)
(3)	0.03 (5)	0.05 (7)	0.03 (3)	0.07 (10)	0.03 (3)	0.01 (3)	0.06 (4)	0.07 (4)	0.31 (46)	0.70 (68)	0.10 (19)
	-	-	-	-	0.08 (6)	0.18 (6)	0.33 (14)	0.12 (2)	-	0.23 (4)	0.30 (16)
	-	-	-	-	-	-	-	-	0.02 (2)	-	0.08 (5)
(27)	100.42 (57)	98.91 (149)	101.40 (85)	99.76 (71)	98.53 (39)	100.04 (88)	98.87 (93)	97.52 (201)	97.07 (45)	99.42 (33)	99.95 (89)
(1)	1.00 (0)	0.00	1.01 (0)	1.01 (1)	1.00 (0)	1.01 (1)	1.06 (6)	1.01 (1)	1.01 (1)	1.01 (2)	1.02 (3)
	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
	tr.	tr.	tr.	tr.	tr.	tr.	0.01 (1)	tr.	tr.	0.04 (6)	0.02 (2)
(0)	0.20 (1)	0.01	0.21 (1)	0.20 (1)	0.21 (1)	0.18 (1)	0.19 (3)	0.16 (1)	0.28 (5)	0.23 (2)	0.20 (0)
(0)	-	0.00	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.01 (0)	0.00 (0)	0.00 (0)
(1)	1.78 (1)	0.02	1.75 (2)	1.78 (1)	1.77 (1)	1.81 (1)	1.66 (11)	1.83 (1)	1.69 (2)	1.65 (9)	1.72 (6)
(0)	0.00 (0)	0.00	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.01 (1)	0.02 (2)	0.00 (0)
	-	-	-	-	tr.	-	0.02 (1)	-	-	0.01 (0)	0.01 (1)
	-	-	-	-	-	-	-	-	tr.	-	tr.
	2.99	0.04	2.98	2.99	2.99	3.00	2.94	3.00	3.00	2.96	2.98
(1)	0.898 (4)	0.891 (7)	0.903 (7)	0.901 (5)	0.894 (5)	0.912 (6)	0.900 (9)	0.922 (7)	0.8571 21.34	0.8753 7.307	0.894 4.559

Table S2. Composition of olivine (trace elements)

Exp	C2933 LHK	D953 LHK	C3082 LLK	C3076 LLK	C3076 LLK	D965 LLK	D963 LLK	C3085 MLK	D1064 MLK	D1077 MLK
P (Gpa)	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
T (°C)	950	1100	800	850	850	900	950	950	1000	1050
Nb	1.48 0.63	0.09 0.03	0.24	0.06 0.05	0.06	0.05	0.21 0.06	3.99 0.33		
P	23 8.38	17	15 8.68	18 4.34	11 3.26	16 6.80	21 4.20	69 32.8	64 29.0	53 22.4
Li	2.22 0.46	3.16 0.50	2.14 0.48	2.08 0.39	2.09 0.12	2.78 0.95	3.23 0.73	4.48 0.66		
Sr	2.87 3.02	0.12 0.10	0.18	0.15 0.10	0.06 0.00		0.13 0.06	94.81 10.4	0.13	
Zr	2.30 1.05	0.23 0.15	0.14	0.19 0.11	0.13	0.14 0.04	0.33 0.07	79.17 13.2	0.25	
Ti	156 126	48 49.3	18 3.96	19 3.33	11 1.98	23 6.28	30 15.1	2346 110	5 4.82	4 0.34
Y	0.91 0.66	0.43	0.09 0.01	0.19 0.11	1.03 0.89	0.07 0.02	0.06	22.64 0.90		
Sc	5.31 1.48	5.03 2.26	3.42 0.06	3.67 0.14	3.29 0.20	3.12 0.11	3.28 0.21	28.96 3.42	3.78 0.57	2.40 0.20
V	6.71 4.29	20.48 25.0	3.14 0.21	3.40 0.23	3.11 0.10	2.84 0.21	3.22 0.79	54.72 5.45	3.04 0.37	3.64 0.57
Mn	1081 34.9	910 257	1191 0.65	1227 41.3	1162 105	1232 74.3	1053 60.1	558 29.1	1068 53.1	816 5.75
Ni	2805 181	1843 1542	3027 71.0	2868 87.4	3196 201	3051 331	2947 139	959 144	2585 617	510 15.6
Cu	0.14 0.03	0.17 0.00	0.25 0.03	0.60 0.55	0.24	0.22 0.04	0.35	3.89 0.28	2.83	
Zn	61 3.55	49 28.3	73 4.9	66 2.69	75 5.28	78 0.84	70 0.71	22 4.42	72 5.94	38 0.32

Table S3. Composition of orthopyroxene

Exp	C2933 LHK	C2945 LHK	C2951 LHK	D953 LHK	C3058 LHK	C2910 MHK	C2940 MHK	D891 MHK	D979 MHK
P (Gpa)	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
T (°C)	950	1000	1050	1100	1100	950	1000	1050	1100
n	10	2	6	10	6	5	6	5	8
SiO ₂	57.73 (40)	56.96 (5)	58.61 (78)	56.17 (73)	57.75 (24)	52.59 (658)	54.99 (81)	57.50 (63)	57.40 (60)
TiO ₂	0.12 (17)	0.18 (8)	0.19 (17)	0.12 (14)	0.07 (11)	0.28 (34)	0.10 (9)	0.07 (19)	0.02 (13)
Al ₂ O ₃	1.41 (65)	2.41 (93)	1.62 (23)	2.32 (146)	0.85 (21)	6.25 (937)	1.55 (6)	1.83 (56)	1.97 (41)
FeO	5.79 (63)	5.53 (49)	4.28 (18)	5.09 (100)	4.25 (21)	6.78 (169)	6.57 (42)	6.89 (20)	5.79 (32)
MnO	0.22 (12)	0.10 (16)	0.01 (13)	0.18 (14)	0.19 (12)	0.06 (9)	0.16 (8)	0.15 (13)	0.15 (8)
MgO	35.20 (84)	34.56 (40)	36.46 (46)	34.07 (151)	35.31 (69)	31.79 (590)	33.11 (52)	34.69 (62)	34.58 (16)
CaO	0.17 (17)	0.20 (0)	0.09 (5)	0.18 (13)	0.48 (23)	1.00 (139)	0.24 (9)	0.24 (7)	0.22 (13)
Na ₂ O	0.14 (8)	0.10 (11)	tr.	1.11 (27)	1.14 (17)	0.25 (7)	tr.	0.22 (8)	tr.
K ₂ O	-	-	-	-	-	-	-	-	-
Sum	100.81 (70)	100.15 (94)	101.23 (143)	99.25 (45)	100.00 (205)	99.01 (132)	96.80 (131)	101.61 (109)	100.10 (69)
Si	1.97 (2)	1.95 (2)	1.97 (1)	1.95 (3)	1.98 (1)	1.84 (21)	1.94 (1)	1.93 (1)	1.97 (2)
Ti	tr.	tr.	tr.	tr.	tr.	0.01 (1)	tr.	tr.	tr.
Al	0.05 (2)	0.10 (4)	0.06 (1)	0.09 (6)	0.03 (1)	0.26 (39)	0.06 (0)	0.08 (2)	0.08 (2)
Fe	0.18 (4)	0.16 (1)	0.12 (0)	0.15 (3)	0.12 (1)	0.20 (5)	0.17 (1)	0.17 (0)	0.17 (1)
Mn	0.01 (0)	tr.	tr.	0.01 (0)	0.01 (0)	tr.	tr.	tr.	tr.
Mg	1.78 (5)	1.77 (3)	1.83 (1)	1.76 (8)	1.81 (4)	1.66 (30)	1.74 (2)	1.74 (2)	1.77 (1)
Ca	0.01 (1)	0.01 (0)	0.00 (0)	0.01 (0)	0.02 (1)	0.04 (5)	0.01 (0)	0.01 (0)	0.01 (0)
Na	tr.	tr.	tr.	0.07 (2)	0.08 (1)	0.02 (1)	tr.	tr.	tr.
K	-	-	-	-	-	-	-	-	-
Σcat	3.99	3.98	3.99	4.04	4.04	4.03	3.93	3.94	3.98
Mg#	0.910 (22)	0.918 (8)	0.938 (2)	0.922 (17)	0.937 (4)	0.887 (51)	0.909 (5)	0.909 (2)	0.914 (4)

Table S3. Composition of orthopyroxene (continued)

C3082 LLK	C3076 LLK	D965 LLK	D963 LLK	C3083 MLK	D971 MLK	D969 MLK	C3085 MLK	D1078 MLK	D1064 MLK
3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
800	850	900	950	800	850	900	950	975	1000
6	6	13	6	6	4	6		6	3
58.07 (50)	57.62 (95)	57.44 (115)	57.83 (52)	56.62 (79)	56.02 (180)	57.11 (98)	56.34 (2)	57.42 (20)	57.18 (27)
0.14 (9)	0.01 (8)	0.06 (9)	0.03 (8)	0.09 (9)	0.09 (2)	0.12 (10)	0.04 (2)	tr.	0.11 (7)
0.58 (43)	1.31 (71)	1.61 (114)	0.87 (22)	1.29 (65)	2.34 (87)	1.98 (119)	1.46 (38)	1.14 (16)	1.32 (21)
5.77 (137)	6.42 (101)	5.60 (70)	5.57 (131)	6.59 (58)	6.41 (40)	6.46 (53)	5.87 (26)	5.69 (25)	6.25 (56)
0.05 (12)	0.19 (20)	0.21 (11)	0.16 (11)	0.09 (6)	0.18 (9)	0.15 (16)	0.14 (11)	0.08 (9)	0.17 (1)
34.38 (147)	34.40 (175)	34.55 (143)	34.02 (170)	33.62 (137)	31.69 (115)	32.72 (167)	33.42 (60)	34.37 (29)	34.21 (32)
0.01 (5)	0.11 (13)	0.25 (16)	0.32 (18)	0.14 (12)	0.36 (15)	0.53 (48)	0.19 (11)	0.41 (5)	0.21 (12)
tr.	tr.	tr.	tr.	tr.	0.70 (24)	0.75 (71)	0.27 (12)	0.30 (4)	0.47 (15)
-	-	-	-	-	-	-	-	-	-
99.03 (65)	100.00 (142)	99.93 (113)	98.90 (59)	98.70 (103)	97.81 (328)	99.82 (96)	97.73 (42)	99.40 (44)	99.89 (41)
2.01 (1)	1.98 (2)	1.97 (2)	2.00 (1)	1.98 (1)	1.97 (1)	1.97 (2)	1.98 (1)	1.98 (0)	1.97 (1)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
0.02 (2)	0.05 (3)	0.07 (5)	0.04 (1)	0.05 (3)	0.10 (3)	0.08 (5)	0.06 (2)	0.05 (1)	0.05 (1)
0.17 (4)	0.18 (3)	0.16 (2)	0.16 (4)	0.19 (2)	0.19 (1)	0.19 (2)	0.17 (1)	0.16 (2)	0.18 (2)
tr.	0.01 (1)	0.01 (0)	tr.	tr.	0.01 (0)	tr.	tr.	tr.	tr.
1.77 (7)	1.76 (7)	1.77 (5)	1.76 (7)	1.75 (6)	1.66 (5)	1.68 (8)	1.75 (3)	1.77 (3)	1.76 (2)
0.00 (0)	0.00 (0)	0.01 (1)	0.01 (1)	0.01 (0)	0.01 (1)	0.02 (2)	0.01 (0)	0.02 (0)	0.01 (0)
tr.	tr.	tr.	tr.	tr.	0.05 (1)	tr.	0.02 (1)	0.02 (0)	0.03 (1)
-	-	-	-	-	-	-	-	-	-
3.97	3.99	3.98	3.97	3.98	3.99	3.94	3.99	4.00	4.00
0.914 (22)	0.905 (17)	0.916 (13)	0.915 (22)	0.901 (11)	0.898 (4)	0.900 (9)	0.910 (2)	0.918 (14)	0.907 (8)

Table S3. Composition of orthopyroxene (continued)

D1077 MLK	D1073 UM	D1052 UM	D1058 UM	C2951 Fl.traps	D963 Fl.traps	D1077 Fl.traps
3.5	3.5	3.5	3.5	3.5	3.5	3.5
1050	650	700	800	1050	950	1050
13	3	3	2			
55.04 (613)	53.12 (26)	53.94 (255)	52.73 (434)	58.18 (1)	55.23 (205)	57.33 (34)
0.02 (15)	tr.	0.08 (9)	0.10 (7)	0.16 (5)	0.07 (15)	0.03 (12)
1.13 (55)	2.86 (45)	1.78 (49)	3.16 (142)	1.67 (27)	4.46 (60)	2.39 (42)
5.70 (136)	5.82 (23)	8.19 (278)	5.29 (85)	4.81 (94)	6.68 (6)	4.62 (18)
0.12 (10)	0.07 (9)	0.04 (17)	0.03 (10)	0.02 (3)		0.19 (11)
36.86 (577)	31.44 (60)	32.53 (236)	31.87 (232)	35.54 (117)	0.22 (4)	34.83 (25)
0.28 (17)	0.48 (22)	0.76 (73)	0.70 (21)	0.19 (11)	0.95 (14)	0.14 (5)
tr.	tr.	0.39 (2)	tr.	tr.	0.99 (11)	0.26 (7)
-	-	-	-	-	-	-
98.70 (126)	93.18 (96)	97.92 (126)	95.34 (445)	100.59 (39)	68.62 (94)	99.79 (60)
1.95 (13)	1.96 (1)	1.93 (6)	1.89 (8)	1.98 (1)	1.95 (1)	1.96 (1)
tr.	tr.	tr.	tr.	tr.	tr.	tr.
0.06 (2)	0.12 (2)	0.08 (2)	0.14 (6)	0.07 (0)	0.19 (3)	0.10 (2)
0.15 (3)	0.18 (1)	0.24 (8)	0.16 (2)	0.15 (0)	0.20 (1)	0.13 (1)
tr.	tr.	tr.	tr.	tr.	0.01 (0)	0.01 (0)
1.87 (26)	1.73 (2)	1.73 (15)	1.71 (6)	1.76 (0)	1.53 (6)	1.78 (0)
0.01 (0)	0.02 (1)	0.03 (3)	0.03 (1)	0.01 (0)	0.04 (1)	0.01 (0)
tr.	tr.	0.03 (0)	0.06 (3)	0.00 (1)	0.07 (1)	0.02 (0)
-	-	-	-	-	-	-
4.03	4.01	4.04	3.97	3.98	3.98	4.00
0.926 (6)	0.90586 2.431	0.87608 43.529	0.91511 6.922	0.921 (2)	0.885 (8)	0.931 (3)

Table S4. Composition of orthopyroxene (Trace elements)

Exp	C2933 LHK	C2945 LHK	C2951 LHK	D953 LHK	C2910 MHK	C2940 MHK	C3082 LLK	C3076 LLK	D965 LLK	D963 LLK
P (Gpa)	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
T (°C)	950	1000	1050	1100	950	1000	800	850	900	950
Rb	1.71 0.14	0.77 0.60		0.64 1.06		0.73 0.32	5.39 2.40	2.96	0.40	0.46 0.51
Cs	0.78 0.32	0.24		0.60 0.89		0.75 0.20	2.51 0.30	0.60 0.61	0.54	0.49 0.55
Ba	12.5 6.10	10.6 15.8	0.8	8.5 9.90	2.3 1.95	7.3 3.09	9.1 7.87	11.9 14.3	3.5 3.17	5.3 4.69
Pb	0.70 0.71	0.91 0.22	0.71 0.63	0.60 0.39	1.46	43.43	1.17		1.27 1.20	0.48 0.33
Th	0.83 0.29	0.85 0.22		0.22 0.32	2.36	2.36	0.57 0.37		0.24	0.35 0.20
U	0.80 0.11	0.75 0.08	0.08	0.29 0.46	0.90 1.20	9.02 12.5	0.58 0.11	0.21	0.43	0.31 0.26
Nb	6.89 10.2	4.01 2.76	0.21 0.08	0.51 0.88	2.26 3.19	6.43 8.71	0.70 0.31	1.23 0.33	0.36 0.31	0.59 0.56
Ta	2.65 3.85	2.60 1.32	0.14 0.08	0.37 0.44	1.47 1.80	8.55 11.8	0.28 0.09	0.43 0.11	0.26	0.40 0.27
P	117 11.1	160 65.6	52 8.74	60 21.0	81 43.5	552 754	138 61.9	103 64.5	122 86.7	78 47.9
Li	12.0 2.43	36.6		8.1 6.68			3.9 1.40	5.8 0.56	5.9 1.09	5.6 2.29
Be	8.03 0.84	6.13 2.15	2.34 0.55	3.01 1.60	7.65	2.76 2.45	1.27 0.72		1.54 0.17	1.72 0.31
La	1.07 0.25	1.66 0.85	0.11	0.42 0.58	4.85	0.62 0.60	0.54 0.34	0.27 0.08	0.33 0.24	0.46 0.37
Ce	2.35 0.80	4.54 3.86	0.15 0.07	0.69 1.02	3.31 5.19	3.42 4.31	1.22 0.72	0.47 0.19	0.61 0.65	0.89 0.62
Pr	2.03 0.99	3.82 4.21	0.13 0.00	0.39 0.52	2.87 3.73	6.63 8.91	0.61 0.43	0.26 0.16	0.47 0.47	0.47 0.32
Sr	3.25 1.90	6.37 9.23	0.42 0.10	4.60 4.44	7.28 11.6	3.07 2.13	3.56 2.07	1.17 1.34	1.14 1.97	2.15 2.53
Nd	2.79 1.63	6.33 7.69		0.75 0.63	5.10	17.91 24.6	1.11 0.00	0.55	1.13	0.69 0.31
Zr	13.9 7.32	25.3 28.2	1.3 0.81	5.3 4.73	6.1 9.65	151.2 210	1.1 0.95	0.5 0.25	1.2 1.18	1.9 1.53
Hf	2.94 0.33	8.03 9.51	0.66 0.35	1.34 1.22	3.42	19.72 27.3	0.53	0.48		0.49 0.23
Sm	5.74 3.74	11.27 15.4	0.56	0.81 0.62	5.17	33.65 46.6	1.25		2.04	0.91 0.37
Eu	4.08 2.67	7.00 9.99	0.22 0.00	0.85 0.65	4.36	38.04 52.7	0.68 0.45	0.35	1.07 0.90	0.68 0.33
Ti	1332 255	2025 1648	691 53.8	485 87.7	243 318	820 650	232 66.2	107 21.8	262 146	331 43.8
Gd	9.38 6.44	14.11 20.5	0.72	1.35 0.84	4.05	36.30 49.7	1.26 0.20		1.90 0.96	1.03 0.39
Dy	11.6 8.85	17.0 22.9	1.2 0.30	2.5 1.55	3.1	35.6 48.6	1.3 0.31		2.0 1.42	1.3 0.50
Y	6.25 4.80	8.11 10.8	0.83 0.26	1.81 1.02	1.48	18.13 24.8	0.46 0.29	0.20 0.08	0.86 0.64	0.81 0.24
Er	6.84 5.03	8.72 11.2	0.85 0.27	1.55 0.95	1.68	15.67 21.6	0.68 0.18		1.10 0.57	0.90 0.23
Yb	6.76 5.17	9.22 10.9	1.00 0.32	1.67 1.04	1.11	14.70 20.1	0.82 0.24		1.08 0.56	1.32 0.18
Lu	6.34 4.55	8.65 9.87	1.01 0.45	1.66 1.05	1.12	13.03 17.8	0.54 0.42	0.26 0.06	0.91 0.71	1.31 0.21
Sc	22 5.64	24 19.1	10 1.75	9 1.15	6 1.63	15 12.2	4 0.89	3 0.64	6 1.44	10 1.39
V	84 22.6	88 53.4	63 8.96	60 20.1	11 13.6	63 65.4	16 8.78	5 1.29	16 10.0	58 24.1
Mn	1853 294	1859 406	936 124	963 220	1105 154	1448 715	921 164	1385 227	1205 117	1124 261
Ni	1327 296	1097 420	1368 152	1005 147	2985 2089	750 378	3445 803	4186 430	3608 1103	1695 465
Cu	2.82 2.39	2.72 0.62	2.52 1.75	0.34 0.19		5.81	1.15 1.08	0.50 0.22	1.92 2.81	1.78 1.91
Zn	47 5.05	62 11.7	47 7.61	42 4.89	98 22.6	64 16.7	164 18.4	184 27.9	108 24.1	68 12.1

Table S4. Composition of orthopyroxene (Trace elements) (continued)

D969 MLK	C3085	MLK	D1078	MLK	D1064	MLK	D1077
3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
900	950	975	1000	1000	1050	1050	1050
0.13	0.24	0.12	0.79 1.11	0.24			
	4.0	0.98	0.98	1.2			
		10.2					
0.84	0.78 0.42			0.19			
0.86	0.53 0.10			0.18 0.16			
0.90	1.25 0.94	0.14		0.18			
0.61	0.86						
76 5.20	57 28.3	31 3.99		40 5.25	29 2.13		
6.4 2.25	5.2 0.50			0.0 0.00	0.0 0.00		
2.48	2.51 0.43	1.27 0.01		2.07 0.43	0.77 0.04		
1.11	1.14 0.97	0.08 0.03		0.18 0.15			
1.84	1.90 1.59	0.23 0.06		0.37 0.26			
0.82	1.04 0.87	0.17 0.01		0.23 0.11	0.07 0.00		
1.01 1.24	3.14 3.56	0.74 0.49		1.33 1.30	0.14 0.04		
1.12	1.81	0.39		0.55 0.13			
4.5	2.7 2.60	1.0 0.16		1.3 0.63			
0.93	0.75	0.32		0.38 0.06	0.3 0.05		
1.09	1.47 0.59	0.52 0.06		0.77 0.07			
1.12	1.09 0.39	0.80 0.00		0.90 0.07	0.17 0.03		
248 292	309 28.4	410 14.6		327 19.9	264 8.14		
1.73	1.96 1.32	1.13 0.09		1.10 0.07	0.39		
2.5	2.1 0.67	2.2 0.10		2.3 0.15	0.7 0.11		
1.68	1.31 0.30	1.29 0.08		1.35 0.12	0.43 0.08		
1.19	1.29 0.59	1.61 0.12		1.42 0.16	0.53 0.08		
2.25	2.09 0.47	2.06 0.09		2.21 0.34	0.77 0.16		
2.03	2.05 0.44	2.30 0.20		2.11 0.23	0.78 0.06		
	7 0.48	10 0.22		9 0.36	6 0.42		
26 29.5	21 1.59	24 0.16		28 0.52	29 2.15		
1344 308	1007 16.2	1079 22.9		1029 8.41	818 13.9		
2812 1970	824 8.44	762 267		536 196	352 253		

Table S5. Composition of clinopyroxene

Exp	C2945	LHK	C3082	LLK	C3076	LLK	D965	LLK	D1073	UM
P (Gpa)	3.5		3.5		3.5		3.5		3.5	
T (°C)	1000		800		850		900		650	
n	7		5		5		18		5	
SiO ₂	58.68 (0)		58.60 (236)		58.52 (115)		56.50 (65)		54.06 (49)	
TiO ₂	0.83 (0)		0.71 (19)		0.71 (13)		0.52 (12)		0.17 (2)	
Al ₂ O ₃	20.40 (1)		16.80 (82)		15.56 (83)		11.69 (87)		3.40 (147)	
FeO	1.95 (1)		2.84 (41)		2.88 (35)		2.81 (26)		3.26 (70)	
MnO	0.24 (0)		0.12 (18)		0.17 (7)		0.21 (10)		0.05 (18)	
MgO	3.61 (1)		4.73 (137)		5.26 (116)		9.47 (155)		15.77 (129)	
CaO	3.10 (1)		4.83 (46)		5.53 (38)		9.52 (152)		21.39 (136)	
Na ₂ O	11.66 (1)		10.23 (54)		9.68 (40)		7.35 (45)		1.32 (55)	
K ₂ O	-0.02 (0)		0.16 (9)		0.43 (53)		0.05 (14)		0.06 (6)	
Sum	100.459322 (1)		99.01 (125)		98.75 (108)		98.13 (212)		99.47 (70)	
Si	2.42 (0)		2.04 (6)		2.04 (5)		2.01 (3)		1.97 (1)	
Ti	0.01 (0)		0.02 (0)		0.02 (0)		0.01 (0)		0.00 (0)	
Al	0.55 (0)		0.69 (4)		0.64 (3)		0.49 (3)		0.15 (5)	
Fe	0.01 (0)		0.08 (1)		0.08 (1)		0.08 (1)		0.10 (2)	
Mn	0.00 (0)		0.00 (1)		0.00 (0)		0.01 (0)		0.00 (0)	
Mg	0.02 (0)		0.25 (7)		0.27 (6)		0.50 (8)		0.86 (6)	
Ca	0.01 (0)		0.18 (2)		0.21 (1)		0.36 (6)		0.83 (4)	
Na	0.31 (0)		0.69 (4)		0.66 (2)		0.51 (3)		0.09 (3)	
K	0.23 (0)		0.01 (0)		0.02 (2)		0.00 (1)		0.00 (0)	
Σcat	3.56		3.96		3.95		3.99		4.01	
Mg#	0.78 (0)		0.743 (30)		0.762 (31)		0.855 (18)		0.895 (28)	

Table S6. Composition of clinopyroxene

Exp	C2945	LHK	D953	LHK	C3082	LLK	C3076	LLK	D965	LLK
P (Gpa)	3.5		3.5		3.5		3.5		3.5	
T (°C)	1000		1100		800		850		900	
Rb			0.17 (0.07)		0.57 (0.19)		4.65 (7.63)		6.98 (4.89)	
Cs			0.10 (0.07)		0.51 (0.23)		4.72 (7.73)		7.60 (5.56)	
Ba	2.6 (1.8)		10.0 (5.1)		10.1 (3.9)		57.2 (92.7)		59.8 (74.8)	
Pb			0.46 (0.23)		0.22 (0.08)		0.59 (0.65)		0.55 (0.38)	
Th	0.22		0.13		1.00 (0.54)		2.27 (3.51)		2.08 (2.44)	
U	0.25 (0.06)		0.09 (0.04)		1.09 (0.52)		2.43 (3.73)		2.49 (2.81)	
Nb	2.06 (2.47)		0.33 (0.13)		3.42 (1.35)		6.56 (10.21)		6.19 (7.19)	
Ta	0.98 (0.61)		0.27 (0.10)		1.22 (0.46)		3.01 (4.55)		2.63 (3.06)	
P	348 (143)		111 (20)		38 (11)		129 (133)		155 (110)	
Li	90.5 (12.0)		40.9 (4.8)		7.0 (4.2)		19.6 (17.5)		9.4 (3.5)	
Be	4.71 (2.91)		1.83 (0.03)		1.63 (0.50)		5.63 (7.29)		5.42 (4.82)	
La	1.74 (1.29)		1.17 (0.60)		1.20 (0.44)		4.84 (6.15)		3.84 (4.46)	
Ce	6.78 (4.56)		4.14 (2.01)		2.20 (0.80)		10.16 (10.36)		7.17 (8.32)	
Pr	7.45 (4.74)		4.21 (2.13)		1.46 (0.62)		7.74 (6.96)		4.10 (4.93)	
Sr	18.89 (8.10)		16.69 (7.43)		7.28 (3.53)		42.10 (43.87)		31.48 (40.45)	
Nd	14.91 (9.81)		7.73 (3.87)		2.09 (1.08)		12.70 (9.85)		4.47 (5.34)	
Zr	64.7 (26.3)		17.8 (8.5)		11.4 (5.2)		28.7 (27.9)		16.9 (20.1)	
Hf	17.26 (6.24)		5.32 (2.44)		2.10 (1.00)		5.86 (5.06)		2.69 (3.05)	
Sm	27.79 (14.40)		15.76 (7.64)		2.76 (1.64)		19.11 (18.17)		5.71 (7.43)	
Eu	16.44 (8.37)		14.38 (6.87)		2.82 (1.61)		19.71 (19.15)		5.65 (7.36)	
Ti	7991 (4124)		1763 (756)		606 (347)		1642 (1392)		561 (454)	
Gd	35.42 (14.34)		23.28 (11.04)		3.57 (2.06)		23.90 (23.39)		6.42 (8.27)	
Dy	39.1 (11.8)		28.6 (13.1)		4.7 (2.9)		30.7 (31.3)		7.5 (9.4)	
Y	17.63 (6.12)		15.90 (7.06)		2.33 (1.41)		15.25 (15.64)		4.22 (4.97)	
Er	18.59 (6.45)		13.62 (6.03)		2.45 (1.60)		15.99 (16.50)		4.09 (4.87)	
Yb	19.02 (6.06)		11.93 (4.95)		2.72 (1.79)		15.71 (16.35)		4.19 (4.64)	
Lu	17.99 (6.99)		10.68 (4.54)		2.49 (1.54)		14.22 (14.49)		4.11 (4.53)	
Sc	84 (65.43)		60 (18.09)		13 (7.74)		29 (20.85)		11 (6.71)	
V	290 (198)		631 (126)		29 (18)		45 (32)		47 (36)	
Mn	1816 (542)		2232 (99)		140 (83)		724 (772)		1106 (143)	
Ni	40 (34)		20 (8)		6 (4)		203 (289)		2149 (1692)	
Cu	6.91 (0.09)		4.15 (1.65)		1.48 (0.76)		3.34 (2.72)		1.34 (1.21)	
Zn	31 (20)		26 (12)		15 (10)		45 (43)		77 (30)	

Table S7. Composition of garnet

Exp	C2933 LHK	C2945 LHK	C2951 LHK	D953 LHK	C2910 MHK	C2940 MHK	D891 MHK	D979 MHK	D1073 UM	D1052 UM	D1058 UM
P (Gpa)	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
T (°C)	950	1000	1050	1100	950	1000	1050	1100	650	700	800
n	8	8	9	8	5	8	6	6	4	9	9
SiO ₂	41.45 (92)	41.13 (77)	43.37 (54)	42.80 (33)	42.07 (100)	40.43 (173)	42.93 (106)	43.63 (92)	40.72 (60)	41.88 (83)	42.54 (70)
TiO ₂	0.62 (22)	0.87 (24)	0.17 (24)	0.45 (15)	0.38 (26)	0.27 (17)	0.18 (14)	0.15 (11)	0.53 (13)	0.49 (20)	0.40 (10)
Al ₂ O ₃	23.10 (50)	22.49 (58)	24.04 (41)	23.57 (26)	23.94 (67)	21.15 (236)	23.09 (112)	24.17 (64)	21.47 (111)	19.96 (217)	19.63 (157)
FeO	13.63 (185)	13.98 (151)	6.99 (118)	10.51 (45)	7.97 (103)	10.70 (326)	9.01 (46)	7.95 (38)	12.36 (77)	8.66 (75)	8.01 (55)
MnO	0.54 (63)	1.17 (26)	0.58 (16)	0.64 (16)	0.22 (15)	0.36 (25)	0.26 (12)	0.31 (7)	0.77 (28)	0.41 (14)	0.42 (12)
MgO	18.16 (292)	17.81 (134)	24.76 (78)	22.22 (39)	23.16 (158)	20.57 (289)	24.24 (143)	24.28 (50)	14.65 (163)	19.85 (234)	19.54 (160)
CaO	2.43 (137)	2.40 (28)	0.88 (12)	1.35 (32)	1.81 (79)	1.87 (36)	1.58 (48)	0.86 (11)	8.33 (142)	7.51 (112)	7.50 (108)
Na ₂ O	0.26 (7)	0.18 (5)	0.10 (9)	0.94 (9)	0.27 (11)	0.11 (5)	0.09 (9)	-0.06 (12)	0.04 (4)	0.20 (6)	0.30 (13)
K ₂ O	-	-	-	-	-	-	-	-	-	-	-
Sum	100.17 (133)	100.08 (117)	100.90 (55)	102.47 (37)	99.84 (139)	95.49 (249)	101.41 (82)	101.29 (232)	98.89 (86)	98.98 (51)	98.55 (79)
Si	2.99 (2)	2.99 (2)	2.99 (3)	2.97 (2)	2.96 (4)	2.96 (7)	2.93 (5)	3.00 (1)	3.01 (3)	3.01 (6)	3.08 (6)
Ti	0.03 (1)	0.05 (1)	0.01 (1)	0.02 (1)	0.02 (1)	0.02 (1)	0.01 (1)	0.01 (1)	0.03 (1)	0.03 (1)	0.02 (1)
Al	1.96 (2)	1.92 (2)	1.96 (2)	1.93 (2)	1.98 (5)	1.82 (15)	1.87 (11)	1.96 (2)	1.87 (9)	1.87 (18)	1.69 (14)
Fe	0.82 (14)	0.85 (10)	0.40 (7)	0.61 (3)	0.47 (7)	0.60 (21)	0.47 (2)	0.46 (1)	0.77 (5)	0.77 (5)	0.49 (3)
Mn	0.03 (4)	0.07 (2)	0.03 (1)	0.04 (1)	0.01 (1)	0.02 (2)	0.02 (1)	0.02 (0)	0.05 (2)	0.05 (1)	0.03 (1)
Mg	1.95 (28)	1.93 (12)	2.55 (6)	2.30 (4)	2.43 (12)	2.24 (25)	2.49 (12)	2.49 (3)	1.62 (17)	1.62 (25)	2.13 (16)
Ca	0.19 (12)	0.19 (2)	0.06 (1)	0.10 (2)	0.14 (6)	0.15 (3)	0.10 (4)	0.06 (1)	0.66 (12)	0.66 (9)	0.59 (9)
Na	0.03 (1)	0.03 (1)	tr.	0.13 (1)	0.04 (1)	0.02 (1)	tr.	-	tr.	tr.	0.04 (2)
K	-	-	-	-	-	-	-	-	-	-	-
Σcat	8.01	8.02	8.01	8.09	8.05	7.82	7.89	8.00	8.01	8.01	8.07
Mg#	0.699 (72)	0.694 (38)	0.863 (23)	0.790 (9)	0.838 (25)	0.789 (77)	0.841 (12)	0.845 (4)	0.677 (39)	0.802 (27)	0.813 (11)

Table S8. Composition of garnet

Exp	C2945 LHK	C2945 LHK	C2951 LHK	D953 LHK	C2910 MHK	C2940 MHK
P (Gpa)	3.5	3.5	3.5	3.5	3.5	3.5
T (°C)	950	1000	1050	1100	950	1000
Rb				0.32 0.03	0.45	0.58
Cs				0.11 0.01	0.43	
Ba				7.6 9.37	6.4 4.15	4.4
Pb				0.11 0.27	7.62 7.54	3.11 3.81
Th	0.75 0.15	1.91		0.00 0.00	6.51 3.32	2.12 1.83
U	1.26 0.34	1.25 1.05	0.26	0.51 0.12	20.50 11.5	5.33 4.29
Nb	5.63 7.75	2.98 1.70	0.24 0.07	0.64 0.29	10.80 4.99	3.63 3.16
Ta	2.76 2.68	1.70 1.18		0.69 0.26	9.09 4.26	2.31 1.42
P	1322 61.4	1056 213	524 157	749 250	374 108	397 170
Li				5.5 0.25		
Be		1.36		0.00 0.00	1.74 0.52	0.91
La	0.52 0.50	2.27 1.28		0.13 0.06	3.48 1.23	1.70 1.41
Ce	1.79 1.10	4.49 2.74	0.28 0.05	0.43 0.11	18.22 10.1	7.90 9.73
Pr	2.66 0.55	4.74 2.49	0.40 0.06	0.91 0.11	24.08 14.6	11.17 13.0
Sr	1.05 1.00	3.44 3.59	0.52 0.34	1.48 1.61	11.38 9.61	2.14 0.63
Nd	8.32 0.47	10.94 4.52	1.32 0.09	3.39 0.24	47.91 28.2	26.58 28.7
Zr	112 11.7	94 18.4	38 16.7	116 38.1	311 88.0	188 58.0
Hf	11 0.65	10 1.95	4 1.81	12 3.94	37 10.0	21 5.49
Sm	48 6.48	43 13.4	8 2.60	20 6.82	77 39.6	59 39.1
Eu	48 11.1	37 9.76	8 1.69	20 7.24	91 47.3	82 52.2
Ti	3300 265	3335 1222	1236 375	2177 569	3126 1178	1486 851
Gd	138 21.0	103 48.94	33 6.89	65 19.3	77 39.1	104 54.1
Dy	271 17.4	178 111	96 13.2	159 43.2	80 51.3	161 82.0
Y	159 10.9	101 67.9	76 7.27	115 34.1	40 27.2	90 48.1
Er	184 14.8	115 84.0	85 6.28	113 33.7	35 24.0	92 54.8
Yb	214 26.5	133 105	116 4.35	123 38.0	32 23.9	92 65.2
Lu	211 26.6	133 108	123 5.34	118 38.2	30 22.1	91 68.9
Sc	209 19.8	149 111	149 14.0	106 12.9	25 6.31	63 21.6
V	295 112	251 163	267 36.9	210 50.4	25 2.46	46 28.0
Mn	9232 1208	6754 3458	4184 626	4223 862	1563 80.08	2121 330
Ni	181 5.50	390 423	112 86.3	161 110	358 334	264 134
Cu		1.07 0.55		0.28 0.00	3.67 2.21	1.47
Zn	127 20.0	93 39.6	35 12.3	61 33.7	34 9.37	35 15.0

Table S9. Composition of micas

Exp	C2910 MHK	C2940 MHK	D891 MHK	D979 MHK	C3083 MLK	D971 MLK	D967 MLK	D969 MLK	C3085 MLK	D1078 MLK	D1064 MLK
P (Gpa)	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
T (°C)	950	1000	1050	1100	800	850	900	900	950	975	1000
n	6	7	5	9	5	7	4	14	6	5	7
SiO ₂	40.77 (75)	39.13 (62)	40.93 (47)	41.50 (106)	42.55 (161)	41.76 (55)	41.42 (44)	41.54 (74)	40.83 (67)	40.94 (54)	40.73 (33)
TiO ₂	0.57 (22)	0.46 (9)	0.45 (10)	0.44 (11)	0.38 (7)	0.39 (8)	0.52 (8)	0.50 (13)	0.49 (6)	0.67 (8)	0.49 (11)
Al ₂ O ₃	13.52 (15)	13.37 (41)	14.48 (29)	14.30 (51)	12.26 (123)	13.15 (49)	13.17 (85)	13.36 (49)	13.36 (48)	13.96 (25)	13.71 (37)
FeO	3.39 (26)	3.85 (26)	3.63 (27)	3.71 (26)	3.72 (32)	3.46 (10)	3.66 (14)	3.48 (34)	3.44 (13)	2.54 (21)	3.91 (16)
MnO	0.22 (16)	0.02 (11)	0.07 (9)	-0.03 (12)	0.07 (9)	0.06 (7)	tr.	0.09 (11)	tr.	0.05 (6)	0.08 (8)
MgO	24.70 (20)	23.64 (61)	24.89 (29)	24.45 (71)	24.63 (55)	25.00 (62)	24.55 (28)	24.88 (51)	24.55 (33)	24.57 (41)	23.78 (30)
CaO	0.01 (11)	-0.04 (7)	-0.03 (8)	0.00 (6)	0.19 (30)	0.04 (7)	0.09 (7)	0.03 (4)	-0.01 (6)	0.04 (3)	0.04 (5)
Na ₂ O	1.17 (17)	0.87 (18)	0.78 (21)	0.74 (56)	1.93 (14)	1.89 (18)	1.41 (57)	1.98 (30)	1.74 (20)	1.31 (9)	1.63 (5)
K ₂ O	8.79 (16)	8.98 (16)	9.63 (25)	9.72 (96)	6.23 (11)	6.76 (28)	8.35 (73)	7.20 (41)	7.51 (49)	8.22 (20)	7.81 (13)
Sum	93.15 (122)	90.28 (167)	94.83 (71)	94.84 (261)	91.96 (117)	92.51 (94)	93.17 (86)	93.06 (153)	91.92 (64)	92.30 (94)	92.17 (138)
Si	2.93 (2)	2.89 (1)	2.88 (1)	2.93 (3)	3.05 (9)	2.98 (4)	2.97 (5)	2.97 (2)	2.94 (4)	2.97 (2)	2.95 (2)
Ti	0.03 (1)	0.03 (0)	0.03 (1)	0.02 (1)	0.02 (0)	0.02 (0)	0.03 (0)	0.03 (1)	0.03 (0)	0.03 (1)	0.03 (1)
Al	1.15 (1)	1.17 (2)	1.20 (3)	1.19 (2)	1.04 (11)	1.11 (4)	1.11 (4)	1.13 (6)	1.15 (5)	1.19 (2)	1.17 (2)
Fe	0.20 (2)	0.21 (2)	0.20 (1)	0.22 (1)	0.22 (2)	0.21 (1)	0.22 (1)	0.21 (2)	0.21 (1)	0.15 (1)	0.24 (1)
Mn	0.01 (1)	0.00 (1)	0.00 (1)	0.00 (1)	tr.	tr.	tr.	0.01 (1)	tr.	tr.	tr.
Mg	2.65 (2)	2.61 (3)	2.61 (3)	2.58 (3)	2.63 (8)	2.66 (5)	2.62 (2)	2.65 (3)	2.64 (3)	2.65 (2)	2.57 (1)
Ca	0.00 (1)	0.00 (1)	0.00 (1)	0.00 (0)	0.01 (2)	tr.	0.01 (1)	tr.	tr.	tr.	tr.
Na	0.16 (2)	0.13 (2)	0.10 (3)	0.10 (8)	0.27 (2)	0.26 (2)	0.20 (8)	0.27 (4)	0.25 (3)	0.18 (1)	0.23 (1)
K	0.81 (3)	0.85 (2)	0.88 (2)	0.88 (7)	0.57 (1)	0.62 (2)	0.76 (7)	0.66 (4)	0.70 (5)	0.76 (2)	0.72 (1)
Σcat	7.95	7.88	7.89	7.92	7.81	7.86	7.92	7.92	7.91	7.94	7.90
Mg#	0.929 (5)	0.924 (5)	0.930 (5)	0.921 (4)	0.922 (4)	0.928 (2)	0.923 (3)	0.927 (6)	0.926 (3)	0.945 (4)	0.916 (3)
K/(K+Na)	0.83 (2)	0.87 (2)	0.89 (3)	0.90 (8)	0.68 (1)	0.70 (2)	0.80 (8)	0.70 (4)	0.74 (3)	0.80 (1)	0.76 (1)

Table S9. Composition of micas (continued)

D1077 MLK	D1073 UM	D1052 UM	D1058 UM	Fluid traps C2945 LHK	Fluid traps C2910 MHK	Fluid traps C2940 MHK	Layer D891 MHK	Fluid traps D979 MLK	Fluid traps D969 MLK	Fluid traps D1078 MLK	Fluid traps D1064 MLK
3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
1050	650	700	800	1000	950	1000	1050	1100	900	975	1000
5	3	6	3	3	4	4	4	4	2	4	4
38.44 (522)	43.42 (32)	43.44 (165)	41.77 (185)	29.45 (397)	39.71 (500)	38.84 (33)	40.96 (77)	40.84 (63)	33.27 (6)	40.57 (91)	41.00 (62)
1.13 (35)	0.42 (9)	0.31 (8)	0.36 (10)	0.10 (6)	0.57 (26)	0.50 (2)	0.74 (5)	0.40 (16)	0.24 (15)	(0)	0.54 (17)
10.14 (142)	11.01 (24)	10.10 (125)	11.46 (119)	8.56 (116)	10.69 (408)	13.10 (29)	12.26 (129)	13.26 (60)	10.81 (20)	13.80 (40)	14.01 (14)
5.04 (85)	4.20 (31)	3.46 (48)	3.49 (39)	2.66 (62)	3.02 (19)	3.64 (33)	5.07 (138)	2.51 (22)	2.70 (19)	2.00 (13)	4.07 (14)
0.11 (7)	0.02 (8)	0.06 (12)	0.03 (9)	0.15 (11)	0.03 (113)	0.03 (12)	0.14 (22)	0.10 (20)	-0.13 (3)	0.08 (11)	-0.02 (9)
18.84 (309)	25.13 (45)	27.34 (200)	25.49 (100)	17.53 (216)	23.94 (323)	23.57 (25)	21.78 (265)	24.85 (78)	19.93 (28)	24.60 (49)	24.31 (24)
0.08 (9)	0.27 (42)	0.13 (5)	0.21 (16)	0.06 (15)	0.02 (13)	-0.01 (7)	-0.01 (12)	0.02 (8)	0.09 (4)	0.08 (9)	0.01 (7)
1.60 (42)	0.47 (3)	0.79 (6)	1.02 (17)	0.25 (15)	1.35 (300)	1.26 (23)	1.09 (24)	1.01 (5)	1.63 (12)	1.15 (7)	1.97 (27)
5.26 (27)	6.91 (17)	6.29 (69)	7.35 (106)	6.88 (86)	6.76 (32)	8.46 (26)	8.51 (109)	9.25 (54)	5.90 (6)	7.94 (25)	7.84 (10)
80.65 1065	91.84 (58)	91.93 (120)	91.17 (271)	65.65 (831)	86.11 (774)	89.39 (73)	90.55 (332)	92.23 (250)	74.43 (35)	90.21 (190)	93.73 (58)
3.20 (2)	3.12 (2)	3.10 (8)	3.04 (7)	3.03 (5)	2.95 (39)	2.90 (3)	3.03 (11)	2.96 (5)	2.96 (1)	2.97 (1)	2.92 (2)
0.07 (1)	0.02 (0)	0.02 (0)	0.02 (1)	0.01 (0)	0.04 (2)	0.03 (0)	0.04 (1)	0.02 (1)	0.02 (1)	(0)	0.03 (1)
0.99 (2)	0.93 (2)	0.85 (12)	0.97 (11)	1.02 (7)	1.14 (36)	1.15 (2)	1.11 (10)	1.13 (3)	1.13 (3)	1.19 (1)	1.18 (2)
0.35 (2)	0.25 (2)	0.21 (3)	0.21 (2)	0.22 (3)	0.16 (7)	0.20 (2)	0.28 (8)	0.15 (1)	0.20 (1)	0.12 (1)	0.24 (1)
0.01 (0)	0.00 (0)	0.00 (1)	0.00 (1)	0.01 (1)	0.00 (1)	0.00 (1)	0.01 (0)	0.01 (1)	-0.01 (0)	0.00 (1)	0.00 (1)
2.33 (7)	2.69 (3)	2.91 (19)	2.76 (6)	2.69 (8)	2.65 (38)	2.62 (2)	2.40 (22)	2.68 (2)	2.64 (2)	2.68 (2)	2.58 (2)
0.01 (1)	0.02 (3)	0.01 (0)	0.02 (1)	0.00 (1)	0.00 (1)	0.00 (1)	-0.01 (1)	0.00 (1)	0.01 (0)	0.01 (1)	0.00 (1)
0.26 (4)	0.07 (0)	0.11 (1)	0.14 (3)	0.03 (1)	0.19 (5)	0.18 (3)	0.13 (5)	0.14 (1)	0.28 (2)	0.16 (1)	0.27 (4)
0.57 (8)	0.63 (2)	0.57 (7)	0.68 (10)	0.88 (8)	0.80 (30)	0.80 (3)	0.81 (7)	0.85 (4)	0.67 (0)	0.74 (1)	0.71 (0)
7.78	7.74	7.79	7.85	7.89	7.93	7.89	7.80	7.95	7.90	7.87	7.94
0.869 (6)	0.914 (5)	0.934 (4)	0.929 (6)	0.923 (9)	0.941 (8)	0.928 (6)	0.891 (28)	0.947 (4)	0.929 (4)	0.956 (3)	0.914 (3)
0.69 (6)	0.91 (1)	0.84 (1)	0.82 (3)	0.96 (1)	0.81 (5)	0.82 (3)	0.85 (6)	0.86 (1)	0.70 (2)	0.82 (1)	0.72 (3)

Table S10. Composition of micas (trace elements)

Exp	C2910 MHK	C2940 MHK	D891 MHK	D891 MHK	D891 MHK	D971 MLK	D969 MLK	C3063 MLK	D1078 MLK	D1064 MLK
P (Gpa)	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5	3.5
T (°C)	950	1000	1050	1050	1050	850	900	950	975	1000
n	6	7	5	5	5	7	14	6	5	7

Rb	87.5 (5.7)	37.9 (24.2)	106.8 (10.6)	1.5 (0.5)	26.0 (1.3)	24.8 (18.1)	35.1 (6.9)	91.5 (19.0)	118.1 (27.1)	124.9 (8.2)
Cs	75.1 (4.5)	33.1 (22.6)	70.1 (7.7)	4.2 (0.8)	46.0 (4.3)	20.5 (12.8)	29.3 (7.3)	69.0 (13.8)	73.1 (14.9)	96.5 (3.1)
Ba	1029.5 (84.2)	406.7 (261.8)	848.7 (107.1)	42.2 (31.6)	417.3 (7.1)	232.2 (138.8)	422.7 (98.5)	884.2 (117.5)	958.3 (228.6)	1311.5 (82.5)
Pb	8.37 (5.12)	20.36 (6.06)	4.62 (2.71)	21.31 (13.14)	87.07 (6.84)	6.64 (6.82)	2.08 (0.45)	4.44 (3.03)	3.98 (2.54)	3.48 (3.55)
Th	1.51 (0.48)	1.44 (1.16)	0.77	2.95 (1.14)	45.86 (5.34)	5.51 (1.10)	4.09 (1.32)	1.19 (0.44)	1.03 (0.87)	0.61 (0.20)
U	1.17 (0.33)	1.63 (1.09)	0.55	4.95 (2.76)	48.83 (3.67)	7.56 (3.41)	4.59 (1.54)	0.95 (0.17)	0.88 (0.82)	0.67 (0.18)
Nb	11.18 (0.34)	5.27 (0.09)	3.06 (2.29)	7.74 (3.00)	127.91 (9.15)	6.62 (1.67)	9.47 (3.64)	3.45 (0.83)	5.36 (1.09)	4.51 (0.35)
Ta	5.33 (0.03)	2.62 (0.02)	2.27 (0.96)	3.35 (1.21)	56.54 (4.79)	2.63 (0.71)	3.89 (1.76)	1.67 (0.41)	2.48 (0.60)	2.12 (0.38)
P		102 (49)	54 (38)	278 (35)	1135 (373)	115 (28)	226 (38)	128 (72)	39 (2)	63 (39)
Li				8.06 (1.01)	21.28 (0.96)	5.76 (1.04)	7.53 (0.52)	5.79 (0.41)		
Be	13.66 (2.29)	6.48 (0.35)	3.71 (2.23)	4.86 (1.89)	94.61 (13.34)	6.49 (2.51)	7.38 (1.56)	4.85 (0.58)	3.21 (0.69)	6.21 (0.58)
La		1.81 (1.64)	0.67 (0.30)	6.40 (2.11)	83.33 (13.79)	5.72 (0.84)	6.21 (3.07)	1.55 (0.63)	0.99 (1.19)	0.39 (0.22)
Ce		3.45 (3.21)	0.99 (0.70)	10.96 (4.05)	136.09 (15.70)	7.95 (1.56)	9.74 (4.76)	1.73 (1.14)	1.71 (2.01)	0.65 (0.41)
Pr		1.88 (1.71)	0.58 (0.01)	5.31 (1.90)	75.93 (10.58)	2.57 (0.61)	3.65 (2.41)	1.00 (0.59)	0.84 (0.98)	0.38 (0.29)
Sr	131.2 (17.5)	51.2 (20.1)	30.5 (23.2)	87.1 (90.7)	384.8 (83.5)	30.2 (15.7)	63.8 (14.5)	65.5 (4.8)	43.1 (7.3)	84.9 (9.6)
Nd		2.42 (1.94)	1.16 (0.56)	5.80 (2.12)	85.30 (12.18)	2.25 (0.62)	3.12 (1.96)		1.45 (1.09)	
Zr	2.73 (0.17)	8.44 (5.69)	12.73 (18.41)	20.02 (7.31)	330.39 (42.88)	7.24 (2.68)	16.39 (11.03)	3.79 (2.54)	5.28 (4.59)	2.85 (1.25)
Hf		1.55 (0.93)	2.86 (2.56)	3.14 (1.08)	50.45 (5.89)	1.10 (0.47)	2.03 (1.37)	0.87	1.22 (0.89)	0.49 (0.10)
Sm		2.76	7.04	4.89 (1.81)	77.36 (10.14)	2.21 (1.55)	2.57 (1.85)		2.37 (1.52)	
Eu	1.31 (0.09)	2.52 (1.42)	4.00 (5.38)	4.55 (1.78)	68.63 (10.81)	1.75 (0.89)	3.38 (2.09)	1.85 (0.54)	2.48 (0.84)	2.32 (0.30)
Dy		1412 (714)	2050 (1091)	266 (65)	3547 (103)	615 (247)	961 (147)	1846 (178)	3375 (873)	2817 (242)
Ti	4045 (14)	1.95 (1.47)	23.85	4.24 (1.58)	61.82 (10.22)	1.64 (1.42)	2.12 (2.53)	1.02	1.68 (1.18)	
Gd		2.20	71.98	2.87 (0.98)	42.47 (7.04)	2.02 (1.06)	2.77 (2.02)	1.49	1.97 (1.14)	0.79
Y		0.66 (0.48)	17.41 (29.81)	1.41 (0.52)	18.98 (3.23)	1.11 (0.50)	1.57 (1.06)	0.54 (0.26)	0.68 (0.68)	0.20 (0.10)
Er		0.69 (0.41)	60.11	1.07 (0.36)	15.02 (2.48)	1.14 (0.36)	1.59 (1.01)		0.96 (0.53)	
Yb		0.91	84.25	0.91 (0.28)	12.45 (2.11)	1.42 (0.33)	1.82 (1.10)		1.50 (0.90)	
Lu		0.53 (0.37)	42.18 (59.35)	0.87 (0.24)	11.03 (1.72)	1.43 (0.30)	1.81 (0.85)	0.67 (0.26)	0.85 (0.77)	0.24 (0.05)
Sc	2.94 (0.15)	4.24 (1.17)	22.27 (38.17)	4.51 (0.13)	15.02 (0.99)	4.13 (0.60)	4.74 (0.92)	3.19 (0.66)	4.68 (0.72)	2.81 (0.49)
V	42.1 (0.2)	29.1 (8.0)	39.4 (5.2)	5.6 (1.6)	25.2 (1.8)	12.7 (3.5)	18.3 (1.8)	26.6 (8.2)	40.1 (6.7)	29.3 (7.0)
Mn	136 (1)	464 (182)	595 (810)	948 (23)	710 (16)	667 (119)	554 (48)	297 (58)	251 (113)	177 (16)
Ni	1473 (334)	1379 (802)	584 (374)	949 (200)	44 (36)	1624 (277)	710 (151)	1324 (398)	738 (375)	1001 (636)
Cu	0.89	2.99 (2.39)	2.39 (1.20)	9.48 (7.28)	9.19 (4.33)					
Zn	43 (9)	37 (3)	24 (3)	54 (9)	38 (7)					

Part B

Table S11. List of samples

Location	Name	Lithology	Type	Alteration	EPMA	XRF	LA-ICP-MS	Isotopes	Chapter	RSES Archives	Size
Houailou	HOU31	Hzb		IV	×				4	11-040	2-2 (AM)
	HOU32	Dun	Channel	V	×				4-5	11-041	1
	HOU33	Opx in Dun	Dyke & Channel	IV	×				4-5		0
Thio											
	TH31	Hzb		IV	×				4	11-042	3 (AM)
	TH32	Dun	Channel	IV	×				4-5	11-043	3
Camp des Sapins	TH33	Opx	Dyke	III	×		×		4-5	11-044	2
Col de Yaté											
	CY1	Dun	Channel	III	×	×	×		4-5	11-045	3 (AM)
	CY2	Hzb		III	×	×	×		4	11-046	1-2 (AM)
	CY3	Opx in Hzb	Dyke						4-5		
Barrage de Yaté											
	YR1	Opx	Dyke	I	×	×			6	11-047	3
	YR2	Opx in Hzb	Dyke	IV	×				4-6	11-048	1
	YR3	Dun in Hzb	Channel	IV	×	×			4-5	11-049	2
	YR4	Dun	Channel						5	11-050	2
	YR5	Opx in Hzb	Dykelet	V	×				4-6	11-051	1
	YR6a	Opx in Hzb	Dyke			×			4-6	11-052	1
	YR6b	Hzb		IV					4	11-053	1
	YR7a	Dun+Opx in Hzb	Channel & dyke	IV	×				4-5-6	11-054	1
	YR7b	Opx	Dyke						6		
	YR8	Hzb		IV					4	11-055	1
	YR9a	Opx in Dun	Dyke & Channel	IV	×				5-6		0
	YR9b	Dun	Channel	III	×				5	11-056	3
	YR10	Opx	Dyke						6	11-057	1
	YR11	Opx in Dun	Dyke & Channel	III	×				5-6	11-058	1
	YR12	Hzb		III		×			4	11-059	1-3 (AM)
	YR13	Opx	Dyke						6	11-060	2
	YR14	Hzb		I					4	11-061	1
	YR15	Opx in Hzb	Dyke						4-6	11-062	1
	YR17	Opx in Hzb	Dyke	II	×				4-6	11-063	2
	YR18	Opx in Hzb	Dyke	II	×				4-6	11-064	2
	YR18b	Opx in Hzb	Dyke						4-6		
	YR19	Opx in Hzb	Dyke	III	×				4-6	11-065	2
	YR20	Dun	Channel	II	×	×			4-5	11-066	2
	YR21	Hzb							4	11-067	4
	YR22	Dun		I					4	11-068	1
	YR31	Hzb							4	11-069	4 (AM)

	YR32	Opx	Dyke						4	11-070	2
	YA1	Opx in Hzb	Dyke						4-6		0
	YA2	Opx in Hzb	Dyke						4-6		0
	YA3	Opx in Dun	Dyke & Channel						5-6		0
	YA4	Hzb							4		0
	YA5	Opx in Hzb	Dyke	III	×		×		4-6		0
	YA7	Opx in Hzb	Dyke		×				4-6		0
	YA8										0
	YA9	Opx in Hzb	Dykelet						4-6		0
	YA10										0
	YA11	Opx in Hzb	Dyke	II	×		×		4-6		0
	YA12	Opx in Dun	Dyke & Channel	IV	×		×		5-6		0
	YA13	Opx in Hzb	Dyke	III	×				4-6		0
	YB1	Opx in Hzb	Dyke						4-6		0
	YB2	Opx in Hzb	Dyke						4-6		0
	YB3	Opx	Cumulate pod		×		×		4		0
	YB4	Hzb							4		0
	YC	Opx	Cumulate						6		0
Route de Yaté											
	R1 core	TTG	Pod		×	×	×	×	6	11-071	2-5 (TE)
	R1 rim	TTG	Pod		×	×	×	×	6	11-072	5 (TE)
	R2	Dun	Channel	III	×				5	11-073	1
	R31	Hbl	Wall-rock	II	×				4-6	11-074	1
	R32	Hbl	Wall-rock	II	×				4-6	11-075	1
	R33	TTG	Pod	IV	×		×		6	11-076	1
	R34	Hzb		III	×		×		4	11-077	2
	R35	TTG	Dyke	III					6	11-078	3
Parc de la Rivière Bleue											
Piste sud	RB1	Wbst in Dun	Dyke & Channel	I					5	11-079	1
Piste sud	RB2	Srp in Dun	Vein & Channel	V					5	11-080	1
Piste sud	RB3	Olv Wbst	Cumulate pod	O	×				4	11-081	1
Piste sud	RBL1	Cr	Pod	III	×				4	11-082	1
Piste sud	RBL2	Opx	Cumulate	II	×		×		4	11-083	1
Piste sud	RBL3	Olv Wbst	Cumulate	III	×				4	11-084	4
Piste sud	RBL4	Amph Wbst	Dyke	IV	×				6	11-085	3
Piste sud	PRB1	Hzb	Pod	II	×	×			4	11-086	3
Piste sud	PRB2	Dun	Pod	III	×				4-5	11-087	1
Piste sud	PRB3										
Piste sud	PRB31	Opx	Dyke	III	×					11-088	1
Piste sud	PRB31b	Dun in Hzb	Channel	III					4-5	11-089	2
Piste sud	PRB32a	PPx	Channel	III					5	11-090	2
Piste sud	PRB32b	Hzb		IV					4	11-091	1
Piste sud	PRB33	Hzb	Pod	II	×				4	11-092	2-4 (AM)
Piste sud	PRB34	Dun	Channel	IV	×		×		5	11-093	3 (AM)
Piste sud	PRB35	Hzb		IV					4-5		0
Piste sud	PRB36	Dun	Channel	III					5	11-094	1
Piste sud	PRB37	Hzb		III					4-5	11-095	3
Riv. Locomobile	PRB38	TTG	Pod	I	×		×	×	6	11-096	3
Riv. Locomobile	PRB39	TTG	Pod	II	×		×	×	6	11-097	3
Riv. Locomobile	PRB40	Olv Hzb		III					4	11-098	1
Riv. Locomobile	PRB41	Opx in Hzb	Dykelet	IV	×		×		6	11-099	1
Riv. Locomobile	PRB42	Opx	Dyke	I					6	11-100	1
Riv. Locomobile	PRB43	Hzb		III					4	11-101	1

Riv. Locomobile	PRB44a	TTG		II				6	11-102	3
Riv. Locomobile	PRB44b	Opx	Wall-rock	V				6	11-103	2
Riv. Locomobile	PRB45a	Dun	Pod	V				4-5	11-104	2
Riv. Locomobile	PRB45b	Opx in Dun	Dyke & Pod	III				4-5	11-105	2
Riv. Locomobile	PRB46	Hzb		IV				4	11-106	2
Riv. Locomobile	PRB47	Hzb	Pod	II				4	11-107	2
Riv. Blanche	PRB48	Dun in Hzb	Pod	III				4-5		0
Riv. Blanche	PRB49a	Gab in Opx	Dyke	IV	×	×		5		0
Riv. Blanche	PRB49b	Wbst in Hzb	Wall-rock	III	×	×		5	11-108	2
Riv. Blanche	PRB49c	Hzb	Wall-rock	IV	×	×		4-5	11-109	4
Riv. Blanche	PRB49d	Gab	Dyke	IV	×			5		0
Riv. Blanche	PRB49e	Hzb		III	×	×		4-5		0
Riv. Blanche	PRB50	Dun	Pod	III				5	11-110	1 (AM)
Riv. Blanche	PRB51	Opx Dun	Pod	III				5	11-111	1
Riv. Blanche	PRB52a	Hzb		IV				4	11-112	2
Riv. Blanche	PRB52b	Opx Dun	Pod	II				5	11-113	3
Riv. Blanche	PRB53	Dun	Pod	II				5	11-114	3 (AM)
Riv. Blanche	PRB54a	Basalt	Dyke	III	×			4	11-115	2
Riv. Blanche	PRB54b	Basalt	Dyke	IV				4	11-116	2
Riv. Blanche	PRB55	Olv Gab	Cumulate	0				4	11-117	2
Riv. Blanche	PRB56	Wehrlite	Cumulate	III	×			4-5	11-118	3
Riv. Blanche	PRB57a	Olv Gab	Cumulate	I				4	11-119	2
Riv. Blanche	PRB57b	Wbst	Cumulate	II				4	11-120	2
Riv. Blanche	PRB57c	Dun	Cumulate	II				4	11-121	2
Riv. Bleue	PRB58a	Opx in Dun	Dyke	II	×	×		5	11-122	3 (AM)
Riv. Bleue	PRB58b	Opx in Dun	Dyke	III	×	×		5		0
Riv. Bleue	PRB59	TTG	Pod	I				6	11-123	3
Riv. Bleue	PRB60a	PPX	Channel	III				5	11-124	3
Riv. Bleue	PRB60b	PPX	Channel	I				5	11-125	2
Riv. Bleue	PRB61a	PPX	Channel	II	×	×		5	11-126	2
Riv. Bleue	PRB61b	PPX	Channel	III	×	×		5	11-127	2
Riv. Bleue	PRB61c	PPX	Channel	II	×	×		5	11-128	2
Riv. Bleue	PRB61d	PPX	Channel	II	×	×		5	11-129	2
Riv. Bleue	PRB61e	Opx in Hzb	Dykelet	III	×	×		5	11-130	2
Riv. Bleue	PRB61f	Hzb		II	×	×		4-5	11-131	2
Riv. Bleue	PRB61g1	Dun	Channel	II	×	×		5	11-132	0
Riv. Bleue	PRB61g2	Dun in PPX	Channel	II	×	×		5	11-133	2
Riv. Bleue	PRB62a	TTG	Pod	II	×			6	11-134	3
Riv. Bleue	PRB62b	Srp	Wall-rock	V				6	11-135	1
Riv. Bleue	PRB63	Wbst in Dun	Channel	III	×	×		5	11-136	2
Riv. Electrique	PRB64	Srp	Pod	V				4	11-137	3
Riv. Electrique	PRB65	Dun	Pod	II				5	11-138	4
Riv. Electrique	PRB66	Cr	Pod	II				4	11-139	3
Palmetum	PRB67	Opx in Hzb	Dyke	III				4-5		
Riv. Bleue	PRB68	Dun in Opx	Channel & dyke	III	×	×		5	11-140	2
Riv. Bleue	PRB69	Opx in Hzb	Dyke	IV	×	×		5	11-141	2
Riv. Bleue	PRB70	TTG	Pod	II	×	×	×	6	11-142	2
Riv. Bleue	PRB71	Opx in Dun	Dyke & pod	III	×	×		5	11-143	2
Riv. Bleue	PRB72a	Dun	Pod	III				5	11-144	2
Riv. Bleue	PRB72b	Dun	Pod	III	×	×		5	11-145	3

Riv.Bleue	PRB73a	Olv Gab	Cumulate	III				4	11-146	1
Riv.Bleue	PRB73b	Olv Gab	Cumulate	III				4	11-147	1
Riv.Bleue	PRB73c	Olv Gab	Cumulate	III				4	11-148	1
Riv.Bleue	PRB73d	Gab	Cumulate	III				4	11-149	1
Riv.Bleue	PRB73e	Gab	Cumulate	II				4	11-150	1
Riv.Bleue	PRB73f	Olv Gab	Cumulate	III				4	11-151	1
Riv.Bleue	PRB74	Dun in Hzb	Pod	III				4-5	11-152	3
Riv.Bleue	PRB75	Olv Hzb		III	×		×	4	11-153	3 (AM)
Riv.Bleue	PRB76	Olv Hzb		III				4	11-154	3
Riv.Bleue	PRB77a	Olv Hzb		III	×			4	11-155	2
Riv.Bleue	PRB77b	Dun	Pod	IV	×			4-5	11-156	3
Riv.Bleue	PRB78	Cr Dun in Hzb	Channel	IV				5	11-157	4
Mine Alain-André	PRB79	Cr	Pod	II				4	11-158	1
Vallée de la Tontouta										
	T2	Hzb						4		0
	T3	Wbst in Hzb	Dyke	I	×		×	4-6		0
	T4	Wbst in Hzb	Dyke	II	×			4-6		0
	T5	Opx in Dun	Dyke & Channel		×		×	6		0
	T6	Wbst in Dun	Dyke & Channel		×		×	6		0
	T7	Wbst in Dun	Dyke & Channel		×			6		0
	T8	Wbst in Hzb	Dyke	II	×		×	4		0
	NC13	Hzb			×			4		0
Les Dalmates										
	LD1	TTG	Dyke	IV		×		6	11-160	2
	LD2	Dun	Channel					5	11-161	1
	LD3	Hzb						4	11-162	1
	LD31	Chl in Dun	Wall-rock	V				4-6	11-163	2
	LD32	Srp	Wall-rock					4-6		
	DA1	Opx in Dun	Dyke & Channel					5		0
	DA3	Opx in Dun	Dyke & Channel					5		0
Rivière de la Dumbéa										
	DR31	Opx in Hzb	Dyke	II	×		×	4-5	11-164	2
	DR32	Dun in Hzb	Channel	III	×		×	4-5	11-165	3 (AM)
	DR33	Hzb		III	×			4	11-166	2 (AM)
	DR34	Olv Cpx	Cumulate	IV				4	11-167	2
Montagne des Sources										
Piste des Crêtes	MDS31	Opx Dun	Cumulate	II				4	11-168	4
Piste des Crêtes	MDS32	Olv Gab in Dun	Dyke & cumulate	II	×		×	4-5	11-169	1
Piste des Crêtes	MDS33	Wbst	Cumulate	II				4	11-170	4
Piste des Crêtes	MDS34	Dun	Cumulate	II	×		×	4	11-171	3-3 (AM)
Piste des Crêtes	MDS35	Olv Gab	Cumulate	II				4	11-172	4
Piste du Casse-Cou	MDS36	Whr in Dun	Pod & cumulate	II	×		×	4-5	11-173	2
Piste du Casse-Cou	MDS37	Olv Gab	Cumulate	O	×		×	4-5	11-174	1 (AM)
Piste du Casse-Cou	MDS38	Olv Gab	Cumulate	O				4	11-175	3 (AM)
Piste du Casse-Cou	MDS39	Whr in Dun	Pod & cumulate	II	×			4	11-176	4
Piste du Casse-Cou	MDS40	Olv Gab	Dyke & cumulate	I				4-5	11-177	2
Piste du Casse-Cou	MDS41	Olv Gab	Cumulate	I				4	11-178	2
Piste du Casse-Cou	MDS42	Olv Cpx	Pod & cumulate	III	×		×	4	11-179	3
Piste du Casse-Cou	MDS43	Dun	Cumulate	I	×		×	4	11-180	2
Piste du Casse-Cou	MDS44	Cr	Cumulate	IV	×			4	11-181	2

Piste du Casse-Cou	MDS45	Cr Whr in Dun	Dyke & pod	IV	×			4	11-182	4
Piste de la Corniche	MDS46	Olv Gab	Cumulate	I				4	11-183	2-2 (HON-AM)
Piste de la Corniche	MDS47	Dun	Cumulate	II		×		4	11-184	3
Piste de la Corniche	MDS48	Gab	Dyke	O				4-5	11-185	2
Piste de la Corniche	MDS49	Opx Dun	Cumulate	II	×	×		4	11-186	2 (AM)
Mine GR2H										
	GR31	Hzb		III				4	11-187	2
	GR32	Cr	Pod	V	×	×		4	11-188	2
	GR33	TTG	Dyke	II	×			6	11-189	2
	GR34	Dun	Pod					4	11-190	2
	GR35	Cr	Pod	V				4	11-191	4 (AM)
Saint Louis										
	STL31	Granite	Pod	I				4-6	11-192	2
	STL32	Granite	Pod	I				4-6	11-193	2
Rivière des Pirogues										
Nord	RP31	Cr	Cumulate	V	×			4	11-194	3
Nord	RP32	Dun	Cumulate	V	×			4	11-195	2
Nord	RP33	Olv Wbst	Cumulate & Dyke	II	×			4	11-196	3
Nord	RP34	Whr in Dun	Pod	IV	×			4	11-197	2
Nord	RP35	Plag Cpx	Cumulate	IV				4	11-198	2
Nord	RP36	Olv Wbst	Cumulate	III				4	11-199	3
Nord	RP37	Wbst	Cumulate	III				4	11-200	2
Nord	RP38	Whr in Dun	Pod	IV				4	11-201	2
Nord	RP39	Gab	Cumulate	III				4	11-202	2
Nord	RP40	TTG	Dyke	III	×	×	×	6	11-203	2
Nord	RP41	TTG	Dyke	III	×	×		6	11-204	3
Nord	RP42a	Olv Opx	Pod	O				4	11-205	2 (AM)
Nord	RP42b	Olv Opx	Pod	O	×	×		4		0
Nord	RP43a	Cr	Pod	III	×	×		4	11-206	2
Sud	RP43b	Gab in Cr	Dyke & Pod	III	×	×		4-5	11-207	2
Sud	RP44	Whr in Dun	Pod	IV				4-5	11-208	2
Sud	RP45	Amph Gab	Pod	I	×			6	11-209	2
Sud	RP46	Amph Gab	Pod	I	×			6	11-210	2
Sud	RP47	TTG in Olv Gab	Dyke & cumulate	O				6	11-211	2
Sud	RP48	Olv Gab	Cumulate	O				4	11-212	2
Sud	RP49	Whr in Dun	Pod	V				4	11-213	1
Sud	RP50	Olv Gab	Cumulate	IV				4	11-214	3
Sud	RP51	Plag Cpx	Cumulate	O	×			4	11-215	2
Sud	RP52	Dun	Cumulate	III	×			4	11-216	2
Sud	RP53	Cr	Sand	I				4	11-217	1
Sud	RP54	Cr	Sand	I				4	11-218	3
Baie des Pirogues										
	BP31	Opx in Dun	Channel & Pod	III	×			6	11-219	2
	BP32	Wbst	Dyke	I	×	×		4-6	11-220	3
	BP33	Amph Gab	Pod	III	×			6	11-221	2
	BP34	Amph Wbst	Dyke	III	×	×		4	11-222	2
	BP35	Amph Gab	Pod	II	×	×		6	11-223	2
	BP36	Opx in Dun	Dyke & Pod	IV				6	11-224	2
	BP37	Dun	Pod	IV	×			4	11-225	2
	BP38	Wbst	Dyke	II				6	11-226	2
	BP39a	Wbst in Dun	Dyke & Pod	IV				6	11-227	2

Col de Prony	BP39b	Olv Opx	Dyke	IV			6	11-228	2
	BP40	Amph Gab	Pod	II	×	×	6	11-229	2
	BP41	Olv Wbst	Pod	II	×	×	6	11-230	2
	BP42	Amph Opx	Dyke	II			6	11-231	2
	PR31a	Dun	Pod	III	×	×	4-5	11-232	2 (AM)
	PR31b	Dun	Pod	III			4-5	11-232	2
	PR31c	Amph Gab	Dyke	II			6	11-233	2
	PR31d	Amph Plag	Dyke	IV			6		0
	PR32	Gab in Hzb	Dyke	III	×	×	4-5	11-234	2
	PR33	Amph Gab	Pod	I	×	×	6	11-235	1
Ile des Pins (Mont Nga)	PR34	TTG	Pod	I			6	11-236	3
	IDP31	Hbl in Amph Hzb	Dyke	II	×	×	6	11-237	2
	IDP32a	Amph Hzb		II			4-6	11-238	2
	IDP32b	Hzb		III	×	×	4	11-239	2
	IDP33a	TTG in Amph Hzb	Dyke	II	×		6	11-240	1
	IDP33b	Amph Hzb in Dun	Pod	III	×	×	6	11-241	1
	IDP34	Gab in Dun	Dyke	II			6	11-242	1
	IDP35	Hzb		III			4	11-243	2 (AM)
	IDP36	Olv Amph Gab	Dyke	III			4-6	11-244	2
	IDP37	TTG in Hzb	Dyke	II			6	11-245	2
	IDP38	Amph Gab	Dyke	III			6	11-246	1
	IDP39a	Hzb		IV			6	11-247	2
	IDP39b	Amph Gab in Dun	Dyke	V			6	11-248	2
	IDP40	TTG in Hzb	Dyke	IV			6	11-249	1
	IDP41	Amph Gab	Dyke	III	×		6	11-250	2

in the following tables, several acronyms are used :

Hzb : Harzburgite
 Dun : Dunite
 Ppx : Proto-pyroxenite
 Opxite : Orthopyroxenite
 Wbst : Websterite
 Cpxite : Clinopyroxenite
 Hblite : Hornblendite
 Tdj : Trondhjemite
 4 : metasomatised

Thin section 0
 Thin section block 1
 ~1/4 dm³ 2
 1/2 dm³ 3
 ~ dm³ 4
 > dm³ 5

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Table S12. Composition of olivine

Sample Type	CY1 Dun channel n	CY2 Hzb 10	YR20 Hzb 11	R2 Hzb 9	R34 Hzb 6	PRB34 Dun channel 8	PRB1 Hzb pod 3	GR32 Chromitite pod 10	PR31a Dun pod 8	PRB61f Hzb 8
SiO ₂	42.14 (17)	42.00 (23)	41.82 (133)	42.05 (17)	40.81 (33)	40.76 (66)	42.81 (13)	41.15 (35)	41.78 (31)	40.83 (39)
TiO ₂		tr.	tr.	tr.		tr.	tr.	tr.	tr.	tr.
FeO	7.90 (13)	8.55 (14)	8.48 (11)	9.43 (16)	7.45 (24)	8.04 (7)	6.42 (0)	6.81 (51)	8.21 (17)	7.84 (22)
MnO	0.12 (2)	0.12 (1)	0.12 (1)	0.14 (1)	0.11 (2)	0.13 (1)	0.10 (0)	0.10 (2)	0.12 (2)	0.11 (1)
NiO	0.38 (2)	0.41 (2)	0.39 (3)	0.28 (2)	0.43 (3)	0.37 (1)	0.49 (2)	0.44 (2)	0.40 (2)	0.49 (10)
CoO	0.02 (2)	0.03 (2)	0.03 (1)	0.03 (2)	0.02 (2)	0.02 (2)		tr.	tr.	0.02 (2)
MgO	49.63 (37)	50.28 (25)	49.81 (53)	48.61 (46)	52.32 (17)	51.64 (67)	51.92 (20)	52.77 (62)	51.91 (21)	52.19 (34)
CaO	0.06 (2)	0.01 (2)	0.05 (14)	0.03 (2)	0.03 (1)	0.10 (2)	0.03 (1)	0.39 (36)	0.11 (1)	0.02 (1)
Sum	100.27 (36)	101.45 (22)	100.73 (98)	100.64 (60)	100.54 (27)	102.09 (64)	101.88 (26)	101.73 (60)	102.58 (40)	101.57 (63)
Si	1.02 (0)	1.01 (0)	1.02 (2)	1.02 (1)	0.98 (0)	0.98 (0)	1.01 (0)	0.98 (1)	0.99 (0)	0.98 (0)
Ti		tr.	tr.	tr.		tr.	tr.	tr.	tr.	tr.
Fe	0.16 (0)	0.17 (0)	0.17 (0)	0.19 (0)	0.15 (0)	0.16 (0)	0.13 (0)	0.14 (1)	0.16 (0)	0.16 (0)
Mn	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Ni	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)
Co	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Mg	1.79 (1)	1.80 (1)	1.78 (4)	1.76 (1)	1.88 (1)	1.86 (1)	1.83 (0)	1.88 (2)	1.84 (1)	1.87 (1)
Ca	tr.	tr.	tr.	tr.	tr.	tr.	tr.	0.01 (1)	tr.	tr.
Σcat	2.98 (0)	2.99 (0)	2.98 (2)	2.98 (1)	3.02 (0)	3.02 (0)	2.99 (0)	3.02 (1)	3.01 (0)	3.02 (0)
Mg#	0.917 (0)	0.913 (0)	0.913 (0)	0.902 (0)	0.926 (0)	0.920 (0)	0.935 (0)	0.933 (0)	0.919 (0)	0.922 (0)

Table S12. Composition of olivine (continued)

PRB61a PPx channel 8	PRB61b PPx channel 10	PRB61c PPx channel 8	PRB61d PPx channel 25	PRB61e Opxite dyke 10	PRB63 Opxite dyke 13	PRB69 Wbst dyke 5	PRB75 Hzb 14	PRB49 Hzb 10	PRB49 Opxite dyke 10	PRB49 Wbst dyke 8
40.88 (72)	41.63 (37)	40.49 (44)	41.77 (44)	41.15 (23)	40.78 (100)	39.73 (85)	41.00 (69)	39.84 (51)	39.68 (58)	42.30 (47)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
8.20 (9)	8.37 (11)	8.20 (13)	8.07 (14)	7.95 (20)	7.87 (57)	7.75 (9)	7.80 (33)	7.42 (11)	7.76 (11)	7.76 (6)
0.10 (2)	0.13 (3)	0.11 (1)	0.11 (3)	0.11 (2)	0.12 (3)	0.12 (1)	0.11 (2)	0.11 (1)	0.12 (2)	0.12 (1)
0.46 (4)	0.40 (3)	0.38 (4)	0.40 (5)	0.39 (3)	0.40 (4)	0.40 (1)	0.48 (16)	0.41 (3)	0.40 (2)	0.41 (3)
0.03 (1)	0.02 (2)	0.02 (1)	0.02 (2)	0.02 (2)	tr.	0.03 (2)	0.03 (2)	0.02 (2)	0.03 (1)	0.02 (1)
51.80 (39)	50.18 (115)	51.48 (22)	51.38 (65)	51.60 (37)	50.34 (100)	51.12 (132)	51.90 (82)	51.55 (32)	50.54 (36)	51.37 (27)
0.02 (1)	0.04 (1)	0.03 (2)	0.03 (1)	0.05 (1)	0.06 (9)	0.01 (1)	0.03 (2)	0.03 (2)	0.04 (2)	0.04 (2)
101.54 (71)	100.80 (153)	100.76 (60)	101.82 (102)	101.33 (66)	99.72 (211)	99.21 (215)	101.43 (124)	99.43 (79)	98.62 (92)	102.09 (64)
0.98 (1)	1.01 (1)	0.98 (0)	1.00 (0)	0.99 (0)	1.00 (1)	0.98 (0)	0.99 (1)	0.98 (0)	0.98 (0)	1.01 (1)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
0.16 (0)	0.17 (0)	0.17 (0)	0.16 (0)	0.16 (0)	0.16 (1)	0.16 (0)	0.16 (1)	0.15 (0)	0.16 (0)	0.15 (0)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
1.86 (2)	1.81 (2)	1.86 (1)	1.83 (1)	1.85 (1)	1.83 (2)	1.87 (1)	1.86 (1)	1.88 (1)	1.86 (1)	1.82 (1)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
3.02 (1)	2.99 (1)	3.02 (0)	3.00 (0)	3.01 (0)	3.00 (1)	3.02 (0)	3.01 (1)	3.02 (0)	3.02 (0)	2.99 (1)
0.918 (0)	0.914 (0)	0.918 (0)	0.920 (0)	0.920 (0)	0.919 (1)	0.922 (0)	0.922 (0)	0.925 (0)	0.921 (0)	0.922 (0)

Table S12. Composition of olivine (continued)

PRB49 Cpxite dyke 6	MDS32 Gabbro dyke 8	MDS37 Gabbro cumulate 9	MDS43 Wehrlite cumulate 11	MDS49 Dun cumulate 7	RP33 Wbst cumulate 12	IDP32b Hzb μ 8	BP34 Mafic μ 8	T7 Wbst dyke 6	PRB41 Opxite dyke 7
41.22 (28)	40.20 (46)	39.48 (26)	40.32 (41)	40.49 (37)	40.98 (135)	41.39 (30)	41.06 (620)	41.40 (17)	39.66 (70)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
13.08 (8)	9.96 (11)	11.90 (14)	10.04 (14)	9.44 (13)	11.04 (17)	8.73 (7)	13.45 (177)	8.37 (12)	8.18 (55)
0.18 (1)	0.15 (1)	0.19 (1)	0.16 (1)	0.16 (1)	0.16 (1)	0.12 (2)	0.22 (1)	0.12 (0)	0.11 (1)
0.32 (2)	0.34 (1)	0.31 (2)	0.33 (1)	0.32 (2)	0.30 (1)	0.40 (2)	0.29 (10)	0.40 (1)	0.44 (3)
0.04 (1)	0.03 (0)	0.03 (1)	0.02 (1)	tr.	0.03 (1)	tr.	0.04 (2)	0.02 (0)	0.02 (1)
47.07 (43)	48.97 (24)	47.97 (23)	48.93 (27)	49.76 (34)	49.32 (94)	51.50 (33)	44.65 (473)	50.55 (22)	51.53 (61)
0.02 (1)	0.08 (1)	0.04 (1)	0.20 (3)	0.04 (1)	0.03 (2)	0.03 (1)	0.06 (14)	0.03 (1)	0.02 (1)
101.97 (66)	101.56 (68)	99.97 (35)	100.05 (57)	100.26 (60)	101.90 (234)	102.22 (39)	100.06 (92)	100.90 (32)	99.98 (81)
0.94 (0)	1.98 (1)	0.98 (0)	0.99 (0)	0.99 (0)	0.99 (1)	0.99 (1)	1.02 (11)	1.00 (0)	0.97 (1)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
0.18 (0)	0.41 (0)	0.25 (0)	0.21 (0)	0.19 (0)	0.22 (0)	0.17 (0)	0.28 (4)	0.17 (0)	0.17 (1)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)
tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
1.66 (1)	3.60 (2)	1.78 (1)	1.79 (1)	1.81 (1)	1.78 (1)	1.84 (1)	1.66 (21)	1.82 (0)	1.88 (1)
tr.	tr.	tr.	0.01 (0)	tr.	tr.	tr.	tr.	tr.	tr.
2.80 (0)	6.02 (1)	3.02 (0)	3.01 (0)	3.01 (0)	3.01 (1)	3.01 (1)	2.98 (12)	3.00 (0)	3.03 (1)
0.842 (0)	0.898 (0)	0.878 (0)	0.897 (0)	0.904 (0)	0.888 (0)	0.913 (0)	0.856 (1)	0.915 (0)	0.918 (1)

Table S13. Composition of olivine (trace elements)

Sample Type	CY1 Dun channel	CY2 Hzb	R34 Hzb	PRB34 Dun channel	PRB1 Hzb pod	GR32 Chromitite pod	PR31a Dun pod	PRB61f Hzb	PRB61a PPx channel
n	8	9	9	8	6	8	7	8	6
Li	1.18 0.05	1.56 0.14	0.77 0.14	1.03 0.21	1.31 0.11	1.39 1.38	1.09 0.13	0.61 0.07	0.32 0.13
Al	0.006 0.008	0.001 0.000	0.006 0.006	0.009 0.004	0.017 0.003	0.016 0.010	0.005 0.004	0.012 0.002	0.011 0.004
P	20.6 1.2	21.3 0.6	20.9 0.5	6.6 0.2	104.2 5.9	15.7 8.1	18.4 1.0	4.0 0.3	7.4 0.6
Ca	0.057 0.012	0.014 0.002	0.018 0.007	0.084 0.011	0.024 0.004	0.199 0.009	0.100 0.018	0.031 0.008	0.029 0.003
Sc	4.45 0.30	3.02 0.11	2.88 0.62	4.95 0.60	3.96 0.20	6.31 2.15	4.56 0.51	4.07 0.32	3.75 0.54
V	0.8 0.7	0.4 0.0	1.6 1.0	5.4 2.1	58.2 12.3	2.9 0.9	2.0 1.6	3.9 0.6	4.8 1.0
Ti	4.50 1.14	1.58 0.19	0.51 0.13	5.66 1.93	33.14 4.48	3.52 0.83	2.87 1.04	0.46 0.04	4.04 1.03
Cr	36 44	11 2	102 158	133 63	129 60	138 27	96 83	136 52	160 51
Mn	607 251	767 271	862 10	1066 18	780 17	840 69	1003 6	1012 19	1034 8
Co	138 1	148 1	134 1	140 3		110 29	138 1	135 2	151 1
Ni	3107 38	3427 24	3215 57	3106 234	3628 123	3645 113	3033 20	2919 116	3065 27
Zn	49 1	52 1	47 4	48 5		24 4	52 2	43 2	45 1

Table S13. Composition of olivine (trace elements) (continued)

PRB61b PPx channel	PRB61c PPx channel	PRB61d PPx channel	PRB61d PPx channel	PRB63 Opxite dyke	PRB69 Wbst dyke	PRB75 Hzb	PRB49 Hzb	PRB49 Opxite dyke	PRB49 Wbst dyke
10	8	8	8	7	5	5	6	7	4
0.31 0.10	0.40 0.12	0.56 0.16	0.65 0.19	0.87 0.30	0.93 0.10	0.41 0.05	0.62 0.19	0.20 0.06	0.75 0.25
0.013 0.009	0.012 0.005	0.010 0.004	0.011 0.004	0.007 0.004	0.005 0.004	0.011 0.003	1.166 0.156	0.020 0.012	0.031 0.026
7.2 0.6	6.0 0.3	12.8 15.6	16.8 0.7	21.1 1.0	5.3 0.2	4.5 0.3	58.3 78.9	9.9 4.9	28.7 3.1
0.034 0.010	0.038 0.006	0.033 0.011	0.039 0.020	0.025 0.007	0.019 0.013	0.033 0.005	1.328 0.214	0.033 0.006	0.110 0.067
2.69 0.54	3.94 0.71	2.77 0.42	3.62 0.56	3.45 0.66	3.34 0.24	4.26 0.40	25.28 0.86	5.70 2.86	3.46 0.53
4.8 1.8	6.7 2.1	4.3 1.3	4.9 1.0	2.6 1.0	2.0 1.3	3.1 0.5	169.0 12.9	10.7 7.7	3.4 0.9
2.15 0.74	2.78 0.92	2.10 1.33	1.71 0.65	4.54 1.76	1.84 0.12	0.65 0.06	330.79 44.05	15.37 13.23	7.93 8.44
142 67	180 55	129 67	136 48	139 148	98 94	162 40	1906 435	130 52	114 39
988 13	1083 8	950 21	969 35	941 10	971 22	1041 13	1193 42	1389 323	854 12
143 4	151 2	140 2	135 2	139 2	137 2	139 1	66 1	195 48	136 2
3163 203	2882 40	3145 147	3016 59	3066 46	3024 63	3011 56	397 12	3032 261	2721 38
54 26	51 2	43 5	49 3	45 3	44 2	47 1	72 3	84 24	70 28

Table S13. Composition of olivine (trace elements) (continued)

PRB49 Gabbro dyke	MDS32 Gabbro dyke	MDS37 Gabbro cumulate	MDS43 Wehrlite cumulate	MDS49 Dun cumulate	IDP32b Hzb μ	T3 Wbst dyke	T8 Wbst dyke	BP34 Mafic μ	PRB41 Opxite dyke
6	7	8	9	7	6	6	9	4	4
0.72 0.19	0.74 0.17	0.39 0.04	1.14 0.08	1.28 0.47	1.75 0.19	1.39 0.15	1.20 0.06	0.80 1.13	0.96 0.08
0.029 0.014	0.018 0.005	0.011 0.013	0.010 0.010	0.012 0.015	0.001 0.000	0.005 0.006	0.001 0.001	0.010 0.015	0.023 0.021
30.1 2.6	33.4 1.1	9.2 0.9	27.1 1.6	8.3 0.9	5.7 1.8	30.1 1.2	1.6 0.2	52.0 14.9	0.2
0.113 0.043	0.061 0.014	0.029 0.018	0.184 0.018	0.053 0.029	0.024 0.003	0.035 0.032	0.007 0.001	0.004 0.009	0.048
3.66 0.49	6.34 0.64	3.77 0.40	5.89 1.95	4.97 0.61	3.12 0.09	2.86 0.27	3.41 0.11	2.59 0.18	2.31 0.13
3.2 1.4	2.6 0.5	2.4 1.6	2.0 1.4	1.2 0.3	0.8 0.1	0.4 0.1	0.6 0.1	0.5 0.1	2.1 1.3
6.10 6.30	16.14 3.87	5.53 1.12	6.85 5.27	5.09 1.14	1.62 0.38	5.16 2.68	1.03 0.47	6.76 0.96	
222 175	62 13	69 98	91 53	69 39	17 6	32 63	16 9	4	247 136
830 7	1168 18	1684 14	1145 10	1407 12	1110 37	1063 138		1724 84	776 233
134 1	130 1	198 3	143 2	148 2	152 2	155 10	139 2	236 4	142 4
2731 35	2509 25	2627 72	2555 16	2597 48	3278 110	2925 41	2709 33	2672 153	3384 244
58 4	46 4	136 163	40 3	42 1	55 1	75 10		123 42	229 254

Table S14. Composition of orthopyroxene

Sample Type	CY2 Hzb	YR20 Hzb	R2 Hzb	R34 Hzb	PRB1 Hzb	PRB61f Hzb	PRB61a PPx channel 19	PRB61b PPx channel 16	PRB61c PPx channel 15	PRB61d PPx channel 18
n	16	16	10	7	7	20				
SiO ₂	58.50 (34)	58.29 (156)	57.44 (53)	58.60 (13)	58.30 (33)	59.43 (21)	58.31 (36)	58.28 (70)	58.08 (84)	58.50 (56)
TiO ₂	0.01 (1)	0.01 (1)	0.02 (1)	0.01 (2)	0.09 (2)	0.00 (2)	0.01 (2)	0.01 (2)	0.01 (1)	0.00 (2)
Al ₂ O ₃	1.10 (12)	0.89 (11)	1.97 (10)	1.53 (4)	1.81 (3)	1.34 (10)	1.56 (11)	1.31 (8)	1.26 (10)	1.32 (7)
Cr ₂ O ₃	0.44 (11)	0.37 (6)	0.69 (6)	0.75 (4)	0.88 (5)	0.74 (7)	0.78 (9)	0.67 (12)	0.66 (5)	0.69 (8)
FeO	5.64 (7)	5.40 (16)	5.92 (22)	4.82 (3)	4.09 (12)	5.04 (10)	5.31 (10)	5.48 (15)	5.22 (11)	5.00 (16)
MnO	0.13 (2)	0.13 (1)	0.15 (1)	0.12 (2)	0.09 (1)	0.11 (1)	0.12 (2)	0.15 (3)	0.14 (2)	0.11 (3)
NiO	0.09 (2)	0.08 (2)	0.08 (2)	0.11 (1)	0.15 (2)	0.10 (2)	0.11 (2)	0.11 (3)	0.09 (2)	0.09 (2)
MgO	34.99 (32)	34.96 (83)	32.56 (81)	36.16 (21)	35.65 (39)	36.04 (28)	35.16 (39)	34.65 (50)	35.01 (26)	35.29 (40)
CaO	0.64 (34)	0.88 (94)	2.29 (107)	0.72 (3)	0.37 (2)	0.01 (1)	1.36 (41)	1.33 (27)	1.47 (18)	1.16 (13)
Na ₂ O	0.00 (1)	0.01 (2)	0.03 (1)	0.01 (2)	0.02 (1)	0.00 (1)	0.02 (2)	0.00 (2)	0.01 (1)	0.01 (1)
Sum	101.58 (30)	101.06 (190)	101.20 (92)	102.84 (28)	101.66 (85)	103.61 (35)	102.78 (34)	101.44 (132)	102.00 (106)	102.53 (94)
Si	1.98 (1)	1.98 (2)	1.96 (1)	1.96 (0)	1.96 (0)	1.98 (0)	1.96 (1)	1.97 (1)	1.96 (1)	1.97 (1)
Ti	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)
Al	0.04 (0)	0.03 (0)	0.08 (0)	0.06 (0)	0.07 (0)	0.05 (0)	0.06 (0)	0.05 (0)	0.05 (0)	0.05 (0)
Cr	0.01 (0)	0.01 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)
Fe	0.16 (0)	0.15 (1)	0.17 (1)	0.13 (0)	0.12 (0)	0.14 (0)	0.15 (0)	0.15 (0)	0.15 (0)	0.14 (0)
Mn	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)
Ni	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)
Mg	1.77 (1)	1.78 (5)	1.66 (3)	1.80 (1)	1.79 (1)	1.79 (1)	1.76 (2)	1.75 (1)	1.76 (2)	1.76 (1)
Ca	0.02 (1)	0.03 (4)	0.08 (4)	0.03 (0)	0.01 (0)	0.00 (0)	0.05 (1)	0.05 (1)	0.05 (1)	0.04 (0)
Na	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)
Σcat	3.99 (0)	4.00 (2)	3.99 (1)	4.00 (0)	3.99 (0)	4.01 (0)	4.00 (1)	4.00 (1)	4.00 (1)	3.99 (1)
Mg#	0.92 (0)	0.92 (0)	0.91 (0)	0.93 (0)	0.94 (0)	0.93 (0)	0.92 (0)	0.92 (0)	0.92 (0)	0.93 (0)

Table S14. Composition of orthopyroxene (continued)

PRB61e Opxite dyke 28	PRB63 Opxite dyke 20	PRB69 Wbst dyke 21	PRB49 Hzb 10	PRB49 Opxite dyke 8	PRB49 Wbst dyke 4	PRB49 Cpxite dyke 4	PRB49 Gabbro dyke 18	PRB49 Mgabbro dyke 10	PRB75 Hzb 10	MDS32 Gabbro dyke 22	MDS49 Dun cumulate 11
57.86 (33)	56.23 (46)	58.43 (75)	57.21 (79)	57.41 (67)	60.78 (46)	60.51 (16)	59.61 (62)	58.38 (120)	58.87 (35)	56.08 (125)	56.25
0.00 (2)	0.02 (2)	0.01 (2)	0.00 (2)	0.00 (2)	0.01 (1)	0.01 (1)	0.03 (2)	0.05 (3)	-0.01 (2)	0.05 (1)	0.03
1.43 (9)	1.42 (16)	1.25 (19)	1.24 (15)	1.48 (4)	1.62 (3)	1.62 (6)	2.03 (22)	1.89 (41)	1.23 (7)	1.57 (17)	1.85
0.73 (8)	0.69 (14)	0.64 (14)	0.66 (8)	0.72 (5)	0.76 (3)	0.75 (3)	0.57 (7)	0.50 (6)	0.69 (6)	0.31 (21)	0.64
5.19 (15)	5.27 (14)	5.16 (15)	4.76 (6)	4.97 (7)	4.93 (5)	4.99 (5)	7.46 (85)	9.03 (140)	4.93 (5)	7.50 (76)	6.20
0.11 (2)	0.15 (2)	0.12 (2)	0.12 (2)	0.12 (2)	0.12 (2)	0.13 (1)	0.18 (2)	0.20 (3)	0.11 (2)	0.18 (2)	0.17
0.09 (2)	0.10 (3)	0.10 (2)	0.10 (2)	0.10 (2)	0.09 (2)	0.09 (1)	0.09 (2)	0.07 (2)	0.10 (2)	0.08 (1)	0.07
35.72 (30)	34.25 (35)	35.36 (45)	35.19 (14)	34.65 (36)	34.97 (14)	35.05 (8)	32.49 (78)	31.56 (130)	36.08 (16)	32.13 (77)	34.00
1.21 (23)	1.31 (33)	1.33 (48)	1.11 (13)	1.15 (16)	1.52 (26)	1.44 (9)	2.00 (75)	1.79 (64)	0.82 (4)	2.20 (47)	1.26
0.01 (1)	0.02 (2)	0.02 (1)	0.01 (1)	0.01 (1)	0.02 (1)	0.01 (1)	0.02 (2)	0.01 (1)	0.00 (1)	0.01 (1)	0.01
102.40 (48)	99.48 (58)	102.46 (83)	100.44 (103)	100.67 (94)	104.85 (41)	104.62 (18)	104.54 (72)	103.56 (149)	102.88 (43)	100.14 (213)	100.51
1.95 (1)	1.95 (1)	1.97 (1)	1.96 (1)	1.96 (1)	1.99 (1)	1.98 (0)	1.98 (1)	1.97 (1)	1.97 (0)	1.95 (1)	1.94
0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00
0.06 (0)	0.06 (1)	0.05 (1)	0.05 (1)	0.06 (0)	0.06 (0)	0.06 (0)	0.08 (1)	0.08 (2)	0.05 (0)	0.06 (1)	0.08
0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.02 (0)	0.01 (0)	0.02 (0)	0.01 (1)	0.02
0.15 (0)	0.15 (0)	0.14 (0)	0.14 (0)	0.14 (0)	0.13 (0)	0.14 (0)	0.21 (2)	0.26 (4)	0.14 (0)	0.22 (2)	0.18
0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.01 (0)	0.00 (0)	0.01 (0)	0.01
0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00
1.79 (1)	1.77 (1)	1.77 (2)	1.80 (2)	1.77 (1)	1.71 (1)	1.71 (1)	1.61 (3)	1.59 (4)	1.80 (1)	1.67 (3)	1.75
0.04 (1)	0.05 (1)	0.05 (2)	0.04 (0)	0.04 (1)	0.05 (1)	0.05 (0)	0.07 (3)	0.06 (2)	0.03 (0)	0.08 (2)	0.05
0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00
4.01 (0)	4.01 (0)	4.00 (1)	4.01 (1)	4.00 (1)	3.97 (1)	3.97 (0)	3.97 (1)	3.98 (1)	4.00 (0)	4.01 (1)	4.01
0.92 (0)	0.92 (0)	0.92 (0)	0.93 (0)	0.93 (0)	0.93 (0)	0.93 (0)	0.89 (1)	0.86 (2)	0.93 (0)	0.88 (1)	0.91

Table S14. Composition of orthopyroxene (continued)

	MDS37 Gabbro cumulate 19	RP33 Wbst cumulate 12	YR1 Opxite dyke 7	T7 Wbst dyke 9	IDP32b Hzb μ 10	IDP31 Hblite 6	PRB41 Hzb 11	PRB41 Opxite dyke 19	BP34 Mafic μ 5	R1 Tdj walls 36	PR33 Gabbro 8
(18)	57.60 (59)	56.29 (140)	56.18 (29)	56.30 (62)	57.38 (62)	54.00 (55)	58.22 (34)	59.94 (67)	56.11 (120)	57.13 (50)	54.61 (16)
(1)	0.03 (2)	0.02 (1)	0.03 (2)	0.03 (2)	0.01 (2)	0.12 (3)	0.00 (2)	0.01 (2)	0.07 (2)	0.14 (3)	0.06 (1)
(12)	1.64 (14)	2.85 (19)	0.73 (10)	1.76 (67)	1.48 (18)	2.16 (13)	1.19 (4)	0.67 (13)	1.55 (9)	1.16 (26)	1.06 (8)
(6)	0.50 (4)	0.76 (15)	0.18 (3)	0.57 (5)	0.51 (19)	0.05 (2)	0.65 (3)	0.08 (5)	0.33 (9)	0.03 (1)	0.04 (1)
(10)	7.75 (21)	7.51 (35)	10.76 (18)	6.06 (89)	5.60 (21)	13.33 (23)	4.79 (9)	8.22 (35)	11.60 (66)	15.11 (74)	18.19 (20)
(2)	0.21 (2)	0.17 (1)	0.14 (5)	0.14 (3)	0.13 (2)	0.34 (2)	0.12 (2)	0.19 (2)	0.26 (2)	0.47 (17)	0.37 (2)
(1)	0.07 (2)	0.06 (1)	0.02 (2)	0.08 (1)	0.09 (3)	0.03 (2)	0.10 (2)	0.18 (3)	0.06 (1)	0.11 (2)	0.04 (1)
(31)	32.90 (53)	33.44 (75)	31.74 (34)	33.78 (118)	35.69 (38)	28.80 (35)	36.24 (16)	30.05 (40)	30.84 (39)	24.55 (72)	26.68 (16)
(27)	1.55 (82)	0.96 (22)	0.93 (36)	1.88 (128)	0.84 (47)	0.60 (4)	0.68 (7)	0.31 (5)	0.55 (4)	0.21 (5)	0.37 (6)
(2)	0.01 (1)	0.01 (1)	0.02 (1)	0.02 (2)	0.01 (2)	0.00 (1)	0.00 (1)	0.09 (2)	0.00 (1)	0.11 (3)	0.01 (1)
(38)	102.30 (66)	102.11 (199)	100.75 (33)	100.67 (31)	101.77 (71)	99.47 (89)	102.05 (44)	99.80 (58)	101.41 (219)	99.05 (65)	101.46 (38)
(0)	1.96 (1)	1.92 (1)	1.97 (0)	1.94 (1)	1.95 (1)	1.94 (0)	1.96 (1)	2.07 (1)	1.96 (1)	2.06 (1)	1.96 (0)
(0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)
(0)	0.07 (1)	0.11 (1)	0.03 (0)	0.07 (3)	0.06 (1)	0.09 (0)	0.05 (0)	0.03 (1)	0.06 (0)	0.05 (1)	0.05 (0)
(0)	0.01 (0)	0.02 (0)	0.00 (0)	0.02 (0)	0.01 (1)	0.00 (0)	0.02 (0)	0.00 (0)	0.01 (0)	0.00 (0)	0.00 (0)
(0)	0.22 (1)	0.21 (1)	0.32 (0)	0.17 (3)	0.16 (1)	0.40 (1)	0.13 (0)	0.24 (1)	0.34 (1)	0.46 (2)	0.55 (1)
(0)	0.01 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.01 (0)	0.00 (0)	0.01 (0)	0.01 (0)	0.01 (1)	0.01 (0)
(0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.01 (0)	0.00 (0)	0.00 (0)	0.00 (0)
(1)	1.67 (3)	1.70 (3)	1.66 (2)	1.74 (5)	1.80 (2)	1.54 (0)	1.82 (1)	1.55 (2)	1.60 (2)	1.32 (3)	1.43 (1)
(1)	0.06 (3)	0.04 (1)	0.03 (1)	0.07 (5)	0.03 (2)	0.02 (0)	0.02 (0)	0.01 (0)	0.02 (0)	0.01 (0)	0.01 (0)
(0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.01 (0)	0.00 (0)	0.01 (0)	0.00 (0)
(1)	4.00 (1)	4.01 (1)	4.02 (0)	4.02 (0)	4.02 (1)	4.01 (0)	4.01 (0)	3.92 (1)	4.00 (0)	3.92 (1)	4.01 (0)
(0)	0.88 (0)	0.89 (1)	0.84 (0)	0.91 (1)	0.92 (0)	0.79 (0)	0.93 (0)	0.87 (1)	0.83 (1)	0.74 (1)	0.72 (0)

Table S15. Composition of orthopyroxene (trace elements)

Sample Type	CY2 Hzb	R34 Hzb	PRB1 Hzb	PRB61f Hzb	PRB61a PPx channel	PRB61b PPx channel	PRB61c PPx channel	PRB61d PPx channel	PRB61e Hzb	PRB61e Opxite dyke
n	8	7	9	12	34	16	15	17	4	13
Li	1.37 0.23	1.58 0.43	0.83 0.13	3.36 0.72	0.84 0.34	1.70 0.75	1.89 0.61	2.55 1.26	2.34 0.37	2.26 1.59
P	23.1 0.9	22.2 0.8	15.5 2.4	6.4 0.5	6.9 2.4	7.0 0.2	6.0 3.3	7.2 2.0	18.4 0.2	18.7 1.0
Sc	16.4 0.9	15.8 1.4	15.7 0.6	19.3 1.1	16.3 1.1	15.7 0.8	16.9 1.1	16.6 1.5	19.4 1.0	17.4 1.0
Ti	20 9	3 0	610 12	2 0	62 17	48 2	45 4	33 8	30 1	46 14
V	41 2	39 1	1080 35	58 2	97 6	97 3	96 5	73 11	75 1	83 16
Cr	3804 390	7544 1762	5773 522	4908 356	5629 338	5818 412	4985 421	4471 869	4874 172	5170 651
Mn	964 109	922 25	760 20	1052 43	1049 13	1025 20	1077 37	896 123	1004 25	995 22
Co	58.4 0.9	59.3 2.6		57.7 3.0	63.7 2.8	62.2 2.2	61.1 2.5	52.3 9.7	57.7 1.3	59.8 3.4
Ni	790 24	834 85	898 76	719 60	815 71	822 77	746 47	701 163	731 36	818 91
Cu		0.82 0.46		0.94 0.46	0.31 0.46	0.16 0.14	0.17 0.28	0.19 0.26	0.07 0.03	0.22 0.17
Zn	41.99 0.85	42.00 0.99		37.50 2.10	46.67 2.90	47.82 4.79	46.88 4.39	40.55 24.80	42.57 0.52	42.27 1.45
Sr	0.063 0.027	0.031	0.008 0.002	0.017 0.017	0.040 0.036	0.070 0.084	0.254 0.725	0.061 0.088	0.018 0.011	0.056 0.089
Y	0.119 0.011	0.011 0.003	0.005 0.001	0.017 0.002	0.200 0.022	0.092 0.008	0.119 0.066	0.051 0.012	0.085 0.004	0.092 0.020
Zr	0.087 0.016	0.022 0.004	0.046 0.006	0.017 0.002	0.059 0.043	0.023 0.003	0.436 1.349	0.018 0.005	0.033 0.002	0.039 0.008
Nb	0.017 0.003	0.023 0.002	0.049 0.006	0.018 0.002	0.023 0.005	0.022 0.005	0.030 0.022	0.017 0.007	0.028 0.003	0.032 0.009
Cs					0.004 0.003		0.018			
Ba			0.011	0.011 0.007	0.014 0.009	0.091 0.086	0.238 0.732	0.333 0.493		0.089 #DIV/0!
La		0.028	0.002 0.002		0.011 0.025	0.004 0.003	0.042 0.096			
Ce			0.005 0.005		0.005 0.004	0.006 0.007	0.050 0.128	0.008 0.008		
Pr			0.001	0.001	0.001 0.001	0.002	0.007 0.019			
Nd				0.004	0.006 0.001	0.054 0.070	0.084 0.138			
Sm					0.007 0.001		0.018 0.023			
Eu		0.010		0.001	0.002 0.000		0.003 0.004			
Gd				0.004	0.012 0.003		0.025 0.030			
Tb				0.001	0.003 0.001	0.002	0.002 0.002			0.004 0.001
Dy	0.027 0.006			0.003 0.000	0.025 0.005	0.015 0.002	0.013 0.011	0.008 0.003	0.021	0.019 0.003
Ho	0.006 0.001		0.001	0.001 0.000	0.007 0.001	0.004 0.001	0.004 0.002	0.003 0.001	0.006	0.005 0.001
Er	0.024 0.004		0.002 0.001	0.004 0.001	0.029 0.003	0.017 0.004	0.018 0.008	0.011 0.003		0.018 0.001
Tm	0.005 0.000		0.001 0.000	0.001 0.000	0.006 0.001	0.004 0.001	0.004 0.001	0.003 0.000		0.004 0.001
Yb	0.041 0.008		0.012 0.001	0.011 0.002	0.048 0.006	0.036 0.005	0.041 0.011	0.020 0.005	0.028 0.006	0.032 0.005
Lu	0.010		0.004 0.000	0.003 0.001	0.009 0.001	0.007 0.002	0.009 0.002	0.005 0.001	0.005 0.000	0.006 0.001
Hf			0.002		0.005 0.002		0.036 0.072			
Ta			0.001	0.001	0.002 0.001		0.003 0.003			
Pb			0.075 0.024	0.004 0.002	0.006 0.004	0.212 0.239	0.085 0.208	0.168 0.230	0.025 0.011	0.022
Th						0.004	0.060 0.101	0.014 0.009		
U				0.002 0.002	0.001 0.001	0.009	0.004 0.012			

Table S15. Composition of orthopyroxene (trace elements) (continued)

PRB63 Opxite dyke	PRB69 Wbst dyke	PRB75 Hzb	PRB49a Cpxite dyke	PRB49a Gabbro dyke	PRB49a Mgabbro dyke	MDS32 Gabbro dyke	MDS37 Gabbro cumulate	MDS49 Dun cumulate	PRB41 Hzb	PRB41 Opxite dyke
12	22	13	7	7	3	6	8	2	10	18
2.54 0.63	1.99 1.24	2.51 1.50	1.35 1.06	0.70 0.17	0.75 0.07	0.52 0.26	0.26 0.23	1.77 0.35	1.52 0.83	0.58 0.35
22.6 1.7	8.8 0.7	5.8 0.5	33.5 2.1	33.7 4.2	40.2 5.4	33.8 0.3	3.7 0.7	5.6 0.2	7.9 4.7	18.2 3.9
18.0 2.0	17.9 2.0	19.4 1.1	31.3 1.1	34.1 2.0	36.2 1.9	39.2 4.5	37.8 1.5	36.6 0.0	16.1 4.3	5.1 1.4
93 11	46 9	3 0	208 18	330 100	488 33	355 26	207 9	155 1	7 15	75 36
79 10	83 10	45 2	221 9	254 9	227 9	133 12	121 3	74 2	33 2	45 17
5375 297	4769 844	5015 161	4093 103	2970 603	2746 444	2566 1856	4450 1216	4433 25	4141 1364	1238 1344
975 16	1062 24	1054 30	1308 21	1526 114	1739 36	1431 209	1687 16	1443 31	1098 252	1606 280
62.0 2.8	61.5 6.1	60.8 1.0	78.7 1.8	86.1 3.6	94.4 0.6	64.5 11.9	80.5 0.8	55.8 0.4	56.3 1.8	64.6 3.9
804 54	813 90	790 36	784 45	594 43	562 35	677 48	564 4	556 6	888 356	1367 309
0.43 0.38	0.39 0.87	0.15 0.22	0.83 0.92	35.22 59.66	2.13 2.70	0.39 0.51	0.16 0.15	0.30 0.38	0.28 0.31	0.83 1.80
43.43 1.25	46.20 6.85	38.78 0.58	73.18 5.60	123.32 93.37	108.03 13.12	55.67 13.56	52.53 1.14	31.61 0.27	44.59 13.46	87.83 31.57
0.028 0.008	0.050 0.057	0.034 0.053	2.560 5.223	1.278 0.874	1.047 0.973	0.084 0.082	0.083 0.066	0.049 0.007	0.296 0.220	0.347 0.197
0.202 0.027	0.138 0.025	0.003 0.001	0.696 0.060	0.881 0.268	1.388 0.151	1.145 0.144	0.557 0.035	0.521 0.013	0.158 0.465	1.577 0.886
0.041 0.008	0.059 0.008	0.024 0.012	0.247 0.278	0.242 0.150	0.262 0.035	0.133 0.027	0.163 0.125	0.092 0.006	0.212 0.227	0.803 0.385
0.024 0.003	0.051 0.009	0.021 0.002	0.103 0.213	0.018 0.007	0.022 0.018	0.108 0.181	0.022 0.008	0.019 0.003	0.035 0.015	0.080 0.053
	0.007 0.010	0.005 0.002	0.103 0.213	0.018 0.007	0.022 0.018	0.108 0.181	0.175		0.019 0.018	0.035 0.041
0.150	0.041 0.012	0.018 0.015	0.356 0.366	0.540 0.333	1.163 1.196	0.309 0.427	0.004		0.171 0.259	0.255 0.212
	0.006 0.004	0.002 0.001	0.028 0.038	0.008 0.004	0.016 0.008	0.008 0.002	0.014 0.014	0.006 0.002	0.041 0.034	0.272 0.799
0.054	0.009	0.005 0.003	0.054 0.057	0.076 0.155	0.046 0.043	0.039 0.037	0.008 0.008	0.001	0.029 0.019	0.084 0.030
		0.001 0.000	0.015	0.011 0.006		0.009 0.007	0.001 0.001	0.001	0.005 0.004	0.014 0.006
	0.008		0.040 0.015				0.009 0.005	0.007	0.043 0.065	0.103 0.043
			0.023			0.021 0.007	0.008 0.002		0.030 0.056	0.135 0.068
			0.010			0.008	0.004 0.001		0.003 0.001	0.004 0.002
0.038	0.014			0.038 0.020	0.052 0.011	0.040 0.002	0.022 0.005		0.062 0.125	0.325 0.173
0.008	0.002 0.001		0.010 0.004	0.010 0.007	0.014 0.002	0.011 0.002	0.007 0.002	0.005	0.011 0.021	0.059 0.032
0.041	0.020 0.004	0.003	0.093 0.018	0.107 0.039	0.153 0.028	0.133 0.014	0.073 0.008	0.052 0.004	0.168 0.223	0.349 0.192
0.011 0.001	0.005 0.001		0.028 0.004	0.032 0.013	0.048 0.003	0.041 0.007	0.023 0.004	0.017 0.002	0.012 0.023	0.054 0.031
0.037 0.006	0.023 0.005		0.101 0.018	0.139 0.038	0.197 0.015	0.177 0.022	0.086 0.010	0.087 0.003	0.026 0.049	0.132 0.064
0.008 0.001	0.005 0.001	0.001 0.000	0.021 0.004	0.028 0.007	0.035 0.003	0.033 0.006	0.015 0.001	0.019 0.000	0.003 0.004	0.015 0.008
0.065 0.016	0.042 0.011	0.007 0.001	0.169 0.011	0.222 0.050	0.329 0.008	0.296 0.038	0.131 0.012	0.179 0.005	0.016 0.025	0.090 0.030
0.012 0.001	0.008 0.001	0.002 0.001	0.032 0.004	0.040 0.010	0.055 0.004	0.055 0.007	0.024 0.002	0.034 0.000	0.004 0.002	0.011 0.005
		0.002	0.030 0.003	0.017 0.002	0.022	0.014	0.006 0.003	0.006	0.013 0.016	0.060 0.034
							0.002 0.001		0.005 0.003	0.012 0.007
0.939	0.033 0.019	0.008 0.009	7.879 18.368	0.870 1.036	2.478 2.036	0.431 0.782	0.045 0.060		0.190 0.186	0.522 0.794
		0.002			0.010	0.005			0.014 0.024	0.088 0.052
		0.002 0.001		0.020 0.001	0.036		0.003 0.002		0.017 0.027	0.104 0.059

Table S15. Composition of orthopyroxene (trace elements) (continued)

YR1 Opxite dyke 8	T3 Wbst dyke 9	T8 Wbst dyke 19	T8 Hzb 23	IDP31 Hblite 11	IDP32b Hzb 17	IDP33b Hzb 6	BP34 Mafic 5	BP40 Mafic 9	PR33 Gabbro 12	R1 Tdj wall 15
1.90 0.31	0.84 0.29	0.34 0.23	0.76 0.20	1.19 0.33	1.40 0.48	1.09 0.28	0.57 0.06	1.08 0.26	0.58 0.17	2.46 1.32
6.1 1.4	33.2 0.7	4.8 0.8	3.1 0.5	30.1 1.5	7.1 0.5	29.6 2.3	37.1 2.6	28.6 9.0	32.7 1.2	40.9 23.0
17.6 1.5	21.1 6.5	22.9 1.8	21.7 1.2	12.1 2.3	19.2 1.3	11.5 1.5	21.7 2.8	27.8 1.4	12.4 1.6	11.0 5.2
289 20	185 226	455 46	15 19	640 103	84 154	578 208	444 41	355 53	308 54	706 345
50 8	61 15	52 6	52 8	63 11	66 8	60 8	70 7	70 13	58 12	39 19
1516 53	2351 2154	177 50	3141 790	567 148	4436 682	367 63	2395 172	1299 289	511 271	171 85
1828 26	1297 447			2650 77	1111 67	2716 95	1735 35	2403 202	2884 67	3161 1944
95.3 1.3	64.9 16.5	64.9 0.9	53.5 2.0	81.8 2.5	55.9 1.6	85.4 10.2	86.2 1.4	98.3 1.8	90.1 1.0	40.0 28.4
213 16	715 81	674 30	640 20	333 77	706 57	327 135	542 6	327 21	278 16	637 310
	0.64 0.69			7.10 7.12	0.07 0.08	56.73 67.59	1.77 1.24	0.10 0.09	0.19 0.14	0.13 0.19
	156.78 275.23			142.38 2.28	42.27 2.42	140.24 4.82	77.80 7.29	128.83 6.30	200.59 7.04	113.40 78.93
4.777 2.224	0.533 0.548	0.168 0.109	0.209 0.246	3.843 7.568	0.111 0.046	4.094 3.457	0.639 0.417	0.734 0.360	0.137 0.140	0.126 0.092
0.394 0.063	0.201 0.182	0.537 0.083	0.232 0.084	0.961 0.358	0.241 0.185	0.840 0.735	0.400 0.068	0.518 0.049	0.283 0.165	2.732 1.325
0.237 0.030	0.373 0.232	0.410 0.114	0.204 0.051	1.429 0.436	0.253 0.097	1.303 1.105	0.560 0.089	0.517 0.145	0.613 0.298	1.856 0.987
0.007 0.000	0.018 0.009	0.002 0.001	0.035 0.014	0.009 0.005	0.043 0.007	0.062 0.087	0.018 0.011	0.010 0.008	0.011 0.008	0.081 0.049
	0.018 0.009					0.019 0.011	0.018 0.011			0.036 0.049
0.163 0.069	0.139 0.078	0.080 0.093	0.030 0.041	1.044 1.824		5.639 11.269	0.452 0.440	0.147 0.162	0.062 0.020	0.499 1.177
0.002 0.001	0.014 0.009	0.003 0.001	0.002 0.002	0.018 0.016	0.006 0.002	0.048 0.054	0.017 0.020	0.010 0.009	0.022 0.023	0.096 0.194
0.006 0.002	0.047 0.050	0.011 0.006	0.007 0.007	0.050 0.060	0.009 0.004	0.116 0.220	0.094 0.098	0.031 0.024	0.048 0.088	0.104 0.196
0.002 0.001	0.064 0.068	0.002 0.001	0.002 0.001	0.013 0.013		0.026 0.045	0.009 0.008	0.006 0.003	0.039	0.015 0.017
0.006 0.002	0.040 0.011	0.017 0.008	0.014 0.009	0.094 0.086	0.020 0.006	0.162 0.277	0.037 0.004	0.041 0.026	0.117 0.123	0.078 0.083
0.005 0.002	0.022 0.003	0.014 0.005	0.009 0.005	0.047 0.034		0.115 0.135		0.023 0.007	0.072	0.077 0.061
0.003 0.001		0.006 0.002	0.004 0.002	0.017 0.013	0.006 0.001	0.028 0.038	0.007	0.008 0.003	0.012 0.009	0.004 0.003
0.018 0.004	0.030 0.010	0.032 0.008	0.018 0.007	0.065 0.051	0.035 0.010	0.185 0.210	0.025 0.005	0.040 0.013	0.089	0.176 0.120
0.006 0.001	0.007	0.008 0.002	0.004 0.002	0.015 0.009	0.006 0.004	0.016 0.021	0.006 0.001	0.007 0.002	0.008 0.004	0.041 0.023
0.052 0.010	0.046 0.038	0.076 0.014	0.035 0.012	0.132 0.060	0.041 0.024	0.132 0.153	0.048 0.009	0.075 0.011	0.044 0.029	0.357 0.175
0.016 0.003	0.010 0.006	0.021 0.004	0.009 0.003	0.038 0.014	0.011 0.007	0.032 0.028	0.015 0.004	0.020 0.003	0.012 0.008	0.092 0.048
0.059 0.009	0.035 0.029	0.082 0.012	0.031 0.012	0.146 0.048	0.041 0.033	0.123 0.082	0.057 0.010	0.079 0.010	0.042 0.023	0.386 0.190
0.012 0.002	0.011 0.005	0.016 0.002	0.006 0.002	0.027 0.008	0.010 0.006	0.022 0.011	0.011 0.002	0.014 0.001	0.010 0.003	0.068 0.035
0.096 0.015	0.069 0.051	0.135 0.015	0.047 0.020	0.255 0.065	0.063 0.051	0.191 0.073	0.109 0.018	0.122 0.006	0.088 0.019	0.606 0.303
0.018 0.003	0.012 0.007	0.025 0.002	0.009 0.003	0.046 0.012	0.013 0.009	0.035 0.008	0.018 0.002	0.022 0.002	0.017 0.004	0.100 0.051
0.012 0.002	0.020 0.007	0.021 0.007	0.005 0.002	0.062 0.020	0.019 0.007	0.050 0.035	0.029 0.003	0.017 0.006	0.029 0.013	0.123 0.067
0.001 0.000		0.000 0.000	0.001 0.000	0.003 0.001		0.024				0.003 0.003
0.373 0.134	0.533 0.696	0.009 0.008	0.060 0.096	0.018 0.003	0.027	0.585 0.246	1.293 1.802	0.022 0.016	0.098 0.080	0.050 0.053
0.003 0.001	0.008	0.002 0.001	0.003 0.002	0.008		0.012	0.010 0.001	0.008 0.002	0.036	0.049 0.114
0.003 0.001		0.001 0.000	0.001 0.000	0.004		0.006	0.009	0.004 0.001		0.003 0.005

Table S16. Composition of clinopyroxene

Sample Type	CY2 Hzb	YR20 Hzb	R2 Hzb	PR31a Dun pod 7	GR32 Chromitite pod 7	PRB61a PPx channel 8	PRB61b PPx channel 7	PRB61c PPx channel 3	PRB61d PPx channel 10
n	4	11	19	7	7	8	7	3	10
SiO ₂	54.78 (18)	55.33 (26)	54.50 (29)	54.52 (66)	53.81 (34)	55.18 (46)	54.92 (44)	55.69 (48)	54.77 (92)
TiO ₂	tr.	tr.	0.06 (2)	tr.	tr.	tr.	tr.	tr.	tr.
Al ₂ O ₃	1.09 (4)	0.96 (8)	2.87 (30)	1.11 (4)	1.37 (10)	1.47 (28)	1.09 (29)	1.21 (2)	1.24 (23)
Cr ₂ O ₃	0.81 (4)	0.68 (11)	1.30 (15)	1.31 (16)	1.49 (22)	1.01 (28)	0.77 (9)	0.83 (6)	0.85 (20)
FeO	1.75 (3)	1.82 (12)	2.66 (36)	1.51 (9)	1.25 (10)	1.60 (7)	1.83 (15)	1.78 (7)	1.81 (16)
MnO	0.06 (2)	0.07 (1)	0.09 (1)	0.03 (2)	0.03 (2)	0.05 (2)	0.07 (3)	0.07 (1)	0.08 (4)
NiO	0.07 (3)	0.05 (2)	0.04 (2)	0.03 (2)	0.03 (3)	0.06 (2)	0.07 (3)	0.04 (1)	0.06 (2)
MgO	17.64 (10)	17.96 (19)	17.61 (72)	17.96 (20)	17.69 (15)	18.07 (23)	17.98 (30)	18.74 (17)	18.51 (67)
CaO	25.90 (18)	25.34 (39)	23.27 (124)	24.88 (24)	25.12 (36)	24.61 (36)	24.08 (39)	24.01 (36)	23.59 (80)
Na ₂ O	0.19 (2)	0.30 (3)	0.29 (4)	0.36 (2)	0.55 (9)	0.28 (7)	0.17 (4)	0.19 (2)	0.17 (3)
Sum	102.33 (28)	102.56 (40)	102.56 (27)	101.75 (90)	101.41 (49)	102.39 (48)	102.24 (89)	102.62 (48)	101.13 (112)
Si	1.95 (0)	1.96 (0)	1.94 (1)	1.95 (1)	1.94 (1)	1.96 (1)	1.97 (1)	1.97 (1)	1.96 (1)
Ti	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Al	0.05 (0)	0.04 (0)	0.10 (1)	0.05 (0)	0.06 (0)	0.06 (1)	0.05 (1)	0.05 (0)	0.05 (1)
Cr	0.02 (0)	0.02 (0)	0.03 (0)	0.04 (0)	0.04 (1)	0.03 (1)	0.02 (0)	0.02 (0)	0.02 (1)
Fe	0.05 (0)	0.05 (0)	0.07 (1)	0.05 (0)	0.04 (0)	0.05 (0)	0.05 (0)	0.05 (0)	0.05 (0)
Mn	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Ni	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Mg	0.94 (1)	0.95 (1)	0.92 (4)	0.96 (1)	0.95 (1)	0.95 (1)	0.96 (1)	0.99 (1)	0.99 (3)
Ca	0.99 (1)	0.96 (1)	0.92 (5)	0.95 (1)	0.97 (1)	0.93 (2)	0.93 (2)	0.91 (1)	0.91 (4)
Na	0.01 (0)	0.02 (0)	0.02 (0)	0.03 (0)	0.04 (1)	0.02 (0)	0.01 (0)	0.01 (0)	0.01 (0)
Σcat	4.02 (0)	4.02 (0)	4.01 (0)	4.02 (1)	4.03 (0)	4.01 (1)	4.00 (0)	4.00 (1)	4.01 (1)
Mg#	0.95 (0)	0.95 (0)	0.93 (1)	0.95 (0)	0.96 (0)	0.95 (0)	0.95 (0)	0.95 (0)	0.95 (0)

Table S16. Composition of clinopyroxene (continued)

PRB61e Opxite dyke 3	PRB63 Opxite dyke 9	PRB69 Wbst dyke 8	PRB49 Hzb 3	PRB49 Wbst dyke 3	PRB49 Cpxite dyke 4	PRB49 Gabbro dyke 14	PRB49 Mgabbro dyke 20	MDS32 Gabbro dyke 18	MDS49 Dun cumulate 12
54.88 (28) tr. 1.23 (9) 0.77 (6) 1.65 (12) 0.03 (3) 0.04 (2) 18.74 (13) 23.51 (19) 0.21 (2)	53.16 (83) tr. 1.37 (28) 1.17 (53) 1.68 (17) 0.07 (3) 0.05 (3) 17.76 (62) 23.52 (82) 0.30 (8)	53.70 (155) tr. 1.52 (37) 1.19 (35) 1.64 (35) 0.06 (1) 0.04 (1) 17.87 (44) 24.07 (62) 0.22 (5)	54.95 (50) tr. 0.99 (9) 0.72 (1) 1.68 (6) 0.08 (2) 0.06 (1) 18.88 (31) 23.88 (14) 0.15 (3)	54.73 (6) 0.03 (2) 1.28 (7) 1.12 (21) 1.83 (12) 0.06 (2) 0.05 (3) 18.30 (17) 23.46 (17) 0.21 (2)	57.23 (56) 0.05 (3) 2.16 (35) 0.98 (4) 2.08 (18) 0.07 (1) 0.06 (3) 18.66 (56) 23.24 (49) 0.25 (3)	56.15 (87) 0.08 (2) 2.68 (34) 0.92 (15) 3.04 (34) 0.11 (1) 0.04 (2) 17.93 (50) 22.79 (88) 0.19 (4)	55.70 (139) 0.10 (4) 2.77 (34) 0.81 (15) 3.66 (73) 0.13 (2) 0.05 (2) 17.83 (58) 22.57 (71) 0.18 (3)	53.32 (120) 0.11 (2) 2.45 (37) 0.71 (61) 4.10 (81) 0.13 (2) 0.06 (1) 19.64 (179) 19.50 (263) 0.20 (4)	52.39 (68) 0.06 (2) 2.72 (42) 1.25 (17) 2.46 (30) 0.10 (2) 0.05 (2) 17.83 (53) 23.09 (100) 0.34 (3)
101.09 (23)	99.15 (103)	100.37 (188)	101.42 (79)	101.10 (19)	104.84 (78)	104.01 (107)	103.89 (189)	100.41 (192)	100.33 (64)
1.96 (1) tr. 0.05 (0) 0.02 (0) 0.05 (0) tr. tr. 1.00 (1) 0.90 (1) 0.01 (0)	1.95 (1) tr. 0.06 (1) 0.03 (2) 0.05 (1) tr. tr. 0.97 (4) 0.92 (3) 0.02 (1)	1.94 (2) tr. 0.07 (2) 0.03 (1) 0.05 (1) tr. tr. 0.96 (2) 0.93 (1) 0.02 (0)	1.96 (1) tr. 0.04 (0) 0.02 (0) 0.05 (0) tr. tr. 1.01 (1) 0.91 (0) 0.01 (0)	1.96 (0) tr. 0.05 (0) 0.03 (1) 0.05 (0) tr. tr. 0.98 (1) 0.90 (1) 0.01 (0)	1.97 (1) tr. 0.09 (1) 0.03 (0) 0.06 (1) tr. tr. 0.96 (2) 0.86 (2) 0.02 (0)	1.95 (1) tr. 0.11 (1) 0.03 (0) 0.09 (1) tr. tr. 0.93 (3) 0.85 (3) 0.01 (0)	1.95 (1) tr. 0.11 (1) 0.02 (0) 0.11 (2) tr. tr. 0.93 (4) 0.85 (2) 0.01 (0)	1.93 (2) tr. 0.10 (2) 0.02 (2) 0.12 (2) tr. tr. 1.06 (9) 0.76 (11) 0.01 (0)	1.90 (1) tr. 0.12 (2) 0.04 (0) 0.07 (1) tr. tr. 0.97 (3) 0.90 (4) 0.02 (0)
4.01 (0) 0.95 (0)	4.01 (1) 0.95 (0)	4.01 (1) 0.95 (1)	4.01 (1) 0.95 (0)	4.00 (0) 0.95 (0)	3.98 (0) 0.94 (1)	3.98 (1) 0.91 (1)	3.99 (1) 0.90 (2)	4.01 (1) 0.90 (1)	4.03 (1) 0.93 (1)

Table S16. Composition of clinopyroxene (continued)

MDS43 Wehrlite cumulate 11	MDS37 Gabbro cumulate 8	RP33 Wbst cumulate 15	YR1 Opxite dyke 3	T7 Wbst dyke 15	IDP32b Hzb μ 15	BP34 Mafic μ 8	BP40 Mafic μ 16
52.07 (55) 0.06 (2) 2.17 (11) 1.30 (9) 1.79 (13) 0.04 (2) 0.02 (1) 16.13 (18) 24.26 (36) 0.63 (3)	54.80 (60) 0.09 (4) 2.61 (12) 0.85 (7) 3.69 (36) 0.15 (2) 0.05 (2) 18.47 (89) 21.53 (125) 0.18 (3)	53.13 (149) 0.07 (2) 3.53 (19) 1.24 (9) 3.19 (28) 0.11 (2) 0.04 (1) 18.35 (98) 21.09 (143) 0.17 (3)	53.92 (14) 0.05 (2) 0.91 (15) 0.38 (2) 3.59 (18) 0.09 (10) 0.02 (1) 17.27 (6) 24.70 (28) 0.26 (3)	53.46 (62) 0.06 (4) 1.34 (51) 0.74 (51) 2.69 (62) 0.09 (2) 0.05 (1) 18.60 (123) 22.43 (160) 0.19 (5)	53.45 (124) 0.15 (21) 1.79 (94) 0.41 (11) 2.06 (23) 0.06 (1) 0.05 (2) 18.22 (42) 24.33 (35) 0.21 (5)	54.15 (73) 0.13 (4) 1.94 (48) 0.57 (9) 3.36 (26) 0.12 (2) 0.05 (1) 17.86 (61) 23.53 (93) 0.12 (8)	53.10 (24) 0.12 (3) 1.52 (27) 0.22 (11) 6.50 (34) 0.16 (2) 0.02 (1) 15.75 (26) 22.66 (44) 0.31 (5)
98.51 (81)	102.49 (49)	100.97 (280)	101.21 (23)	99.70 (31)	100.77 (105)	101.85 (61)	100.40 (45)
1.93 (1) tr. 0.09 (1) 0.04 (0) 0.06 (0) tr. tr. 0.89 (1) 0.96 (1) 0.05 (0)	1.94 (1) tr. 0.11 (1) 0.02 (0) 0.11 (1) tr. tr. 0.97 (4) 0.82 (5) 0.01 (0)	1.91 (0) tr. 0.15 (0) 0.04 (0) 0.10 (1) tr. tr. 0.98 (4) 0.81 (5) 0.01 (0)	1.95 (0) tr. 0.04 (1) 0.01 (0) 0.11 (1) tr. tr. 0.93 (0) 0.96 (1) 0.02 (0)	1.95 (2) tr. 0.06 (2) 0.02 (1) 0.07 (2) tr. tr. 1.01 (6) 0.89 (7) 0.01 (0)	1.93 (3) tr. 0.08 (4) 0.01 (0) 0.06 (1) tr. tr. 0.98 (2) 0.94 (1) 0.01 (0)	1.94 (1) tr. 0.08 (2) 0.02 (0) 0.10 (1) tr. tr. 0.95 (3) 0.90 (4) 0.01 (1)	1.95 (1) tr. 0.07 (1) 0.01 (0) 0.20 (1) 0.01 (0) tr. tr. 0.86 (1) 0.89 (2) 0.02 (0)
4.02 (0) 0.94 (0)	4.00 (1) 0.90 (1)	4.00 (0) 0.91 (0)	4.03 (0) 0.90 (1)	4.02 (1) 0.93 (2)	4.03 (1) 0.94 (1)	4.01 (1) 0.90 (1)	4.02 (1) 0.81 (1)

Table S17. Composition of clinopyroxene (trace elements)

Sample Type	GR32 Crie Pod	PRB49b Cpxite dyke	PRB49a Mgabb dyke	MDS32 Gabbro dyke	MDS37 Gabbro cumulate	MDS49 Dun cumulate	YR1 Opxite dyke	T8 Wbst dyke	T3 Wbst dyke	IDP32b Hzb μ	BP32 Mafic μ	BP40 Mafic μ	PR32 Gabbro
n	3	16	14	10	8	4	3	22	3	5	6	8	7
Li		2.83 3.08	2.27 0.85	2.09 0.47	0.89 0.18	4.26 0.75	3.59 2.40	0.75 0.18	1.56 0.14	1.65 0.25	1.54 0.13	2.28 0.16	2.13 0.39
Be		0.006	0.019		0.002	0.009 0.007			0.085 0.010	0.096	0.038 0.006	0.050 0.007	
P	38.4 14.8	6.6 6.3	55.8 38.8	30.3 0.7	4.2 1.2	10.3 6.7	6.2 1.4	5.1 0.8	32.7 4.6	13.6 2.2	21.9 0.5	25.0 1.9	27.6 0.8
Sc	135.2 35.0	64.8 4.0	72.8 9.7	74.8 3.8	71.5 1.7	81.7 7.9	68.1 8.8	83.9 9.6	121.1 2.9	75.3 10.9	70.7 6.5	85.7 16.0	61.7 4.3
Ti	368 44	413 31	686 252	687 127	364 33	340 26	471 102	832 115	676 44	1775 679	599 163	749 283	711 370
V	213 20	440 50	461 68	273 19	242 7	171 15	137 7	125 19	233 12	263 65	149 32	188 55	317 43
Cr	7996 700	6929 355	5800 1596	6288 3555	7325 1377	9424 461	2696 497	168 99	256 107	3606 1497	2682 264	2422 511	3644 1828
Mn	428 186	1125 51	1144 98	973 118	1220 39	1167 283	839 198		943 50	627 17	1178 70	1255 76	897 168
Co	28.6 10.0	42.3 2.3	46.1 4.8	31.1 4.9	42.0 2.2	26.6 3.2	47.0 11.3	22.2 1.6	32.7 1.5	20.6 0.7	37.1 2.1	41.5 2.2	32.7 4.3
Ni	158 5	453 41	428 81	403 24	351 12	313 25	175 73	343 52	449 17	308 3	214 9	204 9	481 92
Cu	4.43 5.61	1.22 2.12	9.49 13.38	0.33 0.30	17.00 47.09	2.30 1.75			0.11 0.01	0.29 0.09	0.17 0.03	0.41 0.50	0.23 0.20
Zn	7.8 1.6	28.3 2.9	42.1 8.8	20.1 4.3	21.3 1.5	11.0 3.1			18.1 1.0	11.7 1.1	29.6 1.8	31.9 3.3	22.3 7.2
Rb		0.04 0.04			0.01 0.01	0.19 0.28	0.98				0.03 0.02		
Sr	12.11 8.86	1.28 0.93	5.35 4.52	1.25 0.52	1.70 0.16	6.80 4.06	25.08 6.86	43.38 1.71	41.37 1.86	14.73 0.79	18.18 2.49	22.35 2.54	0.84 0.29
Y	5.06 1.08	3.71 0.42	5.78 2.06	6.40 0.68	2.69 0.15	4.34 0.79	3.65 0.17	6.27 0.75	4.81 0.71	6.87 1.56	3.57 0.86	4.43 1.21	5.92 1.84
Zr	1.87 1.30	0.64 0.42	1.26 0.73	0.78 0.22	0.47 0.07	0.96 0.38	2.97 0.60	6.61 1.21	7.38 0.69	9.06 2.58	2.33 1.66	3.94 1.74	2.39 2.39
Nb	0.03 0.00	0.04 0.01	0.05 0.02	0.04 0.02	0.04 0.00	0.07 0.03	0.02 0.00	0.00 0.00	0.03	0.04 0.01	0.01 0.00	0.02 0.01	0.03 0.01
Cs	0.035 0.004	0.011 0.008	0.047 0.021	0.040 0.021		0.047			0.031				0.028 0.009
Ba	0.159 0.023	0.255 0.315	1.391 1.255	0.457 0.332	0.028 0.037	1.008 1.218	1.403 1.688	0.112 0.146	0.114 0.063	0.197	0.035	0.223 0.212	0.250 0.084
La	0.114 0.090	0.113 0.095	0.067 0.087	0.029 0.038	0.017 0.004	0.317 0.297	0.165 0.050	0.442 0.042	0.338 0.029	0.204 0.020	0.167 0.050	0.230 0.059	0.028 0.016
Ce	0.323 0.253	0.099 0.016	0.210 0.127	0.132 0.115	0.054 0.006	0.172 0.034	0.517 0.225	2.153 0.217	1.535 0.193	1.175 0.093	0.739 0.285	0.989 0.266	0.130 0.112
Pr	0.066 0.046	0.021 0.010	0.065 0.089	0.025 0.009	0.014 0.002	0.029 0.005	0.083 0.036	0.506 0.053	0.325 0.037	0.274 0.037	0.152 0.063	0.212 0.060	0.038 0.029
Nd	1.305 0.991	0.134 0.070	0.300 0.148	0.217 0.088	0.123 0.014	0.211 0.059	0.529 0.250	3.446 0.390	2.231 0.295	1.925 0.281	0.997 0.410	1.427 0.387	0.351 0.247
Sm	0.331 0.205	0.107 0.028	0.213 0.113	0.203 0.047	0.098 0.010	0.108 0.005	0.283 0.119	1.417 0.172	0.981 0.133	0.867 0.234	0.442 0.169	0.638 0.185	0.249 0.123
Eu	0.148 0.071	0.043 0.009	0.077 0.047	0.087 0.013	0.048 0.004	0.058 0.004	0.089 0.026	0.442 0.056	0.322 0.044	0.361 0.051	0.162 0.049	0.223 0.058	0.058 0.021
Gd	0.716 0.327	0.283 0.051	0.500 0.219	0.524 0.099	0.235 0.021	0.273 0.014	0.511 0.144	1.629 0.206	1.295 0.111	1.245 0.270	0.618 0.171	0.861 0.244	0.591 0.216
Tb	0.125 0.050	0.066 0.010	0.110 0.045	0.117 0.017	0.052 0.003	0.068 0.006	0.094 0.024	0.236 0.029	0.181 0.016	1.201 0.058	0.104 0.029	0.137 0.039	0.124 0.043
Dy	0.985 0.323	0.565 0.079	0.939 0.372	1.027 0.129	0.446 0.030	0.661 0.115	0.665 0.076	1.429 0.180	1.130 0.166	1.403 0.381	0.708 0.189	0.907 0.258	0.994 0.314
Ho	0.198 0.047	0.139 0.020	0.229 0.087	0.247 0.028	0.103 0.007	0.154 0.017	0.144 0.008	0.264 0.033	0.214 0.031	0.305 0.087	0.145 0.033	0.180 0.050	0.231 0.064
Er	0.527 0.113	0.459 0.055	0.736 0.250	0.800 0.081	0.333 0.019	0.563 0.093	0.422 0.020	0.686 0.086	0.565 0.102	0.826 0.225	0.415 0.100	0.503 0.141	0.753 0.237
Tm	0.063 0.003	0.070 0.007	0.109 0.034	0.120 0.012	0.048 0.003	0.087 0.015	0.057 0.004	0.088 0.012	0.069 0.011	0.103 0.027	0.054 0.012	0.066 0.019	0.107 0.036
Yb	0.382 0.033	0.488 0.044	0.784 0.222	0.839 0.071	0.335 0.025	0.632 0.112	0.390 0.029	0.522 0.071	0.440 0.066	0.668 0.115	0.361 0.082	0.434 0.115	0.707 0.240
Lu	0.061 0.012	0.075 0.007	0.112 0.034	0.126 0.010	0.050 0.003	0.092 0.012	0.055 0.005	0.070 0.009	0.060 0.010	0.082 0.020	0.049 0.010	0.060 0.015	0.108 0.039
Hf	0.184 0.123		0.067 0.040	0.054 0.019	0.022 0.005	0.048 0.021	0.130 0.042	0.313 0.061	0.325 0.043	0.475 0.171	0.092 0.069	0.165 0.063	0.122 0.087
Ta			0.045 0.041	0.048	0.002 0.001	0.003							
Pb	0.105 0.072		1.879 2.288	0.087 0.055	0.019 0.026	0.790 0.678	0.303 0.038	0.023 0.004	0.136 0.101	0.226 0.065	0.028 0.002	0.060 0.014	0.027
Th		0.002 0.002	0.036 0.010		0.002 0.001	0.091 0.116	0.012	0.002 0.001			0.007 0.000	0.014 0.006	
U		0.004 0.003	0.052 0.030	0.021	0.002 0.001	0.037 0.025	0.001 0.000	0.001 0.000			0.005 0.000	0.008 0.002	

Table S18. Composition of spinels

Sample Type	CY1 Dun channel	CY2 Hzb	YR20 Hzb	PRB34 Dun channel	PRB1 Hzb pod	GR32 Crite pod	PRB61f Hzb	PRB61a PPx channel	PRB61b PPx channel	PRB61c PPx channel	PRB61d PPx channel
n	10	11	10	8	6	7	6	7	8	6	5
TiO ₂	0.11 (2)	0.02 (1)	0.03 (2)	0.06 (3)	0.28 (2)	0.03 (1)	tr.	0.04 (2)	0.03 (2)	0.03 (4)	0.03
Al ₂ O ₃	15.94 (72)	16.64 (56)	15.38 (56)	15.98 (66)	18.55 (24)	18.13 (25)	17.55 (36)	19.21 (44)	16.56 (52)	16.92 (172)	16.38
Cr ₂ O ₃	51.14 (213)	50.82 (65)	52.71 (84)	55.05 (26)	50.89 (85)	53.97 (40)	56.25 (52)	53.10 (104)	52.27 (142)	54.24 (70)	53.14
V ₂ O ₅	0.21 (3)	0.16 (2)	0.19 (3)	0.41 (3)	2.98 (13)	0.17 (2)	0.21 (3)	0.28 (2)	0.32 (6)	0.36 (7)	0.23
FeO	22.22 (164)	21.73 (39)	19.77 (55)	17.71 (45)	13.83 (86)	16.45 (44)	17.30 (90)	14.31 (105)	17.79 (250)	17.06 (200)	17.51
MnO	0.18 (3)	0.18 (1)	0.15 (3)	0.23 (2)	0.10 (2)	0.18 (2)	0.22 (1)	0.17 (3)	0.26 (4)	0.22 (3)	0.26
NiO	0.08 (3)	0.06 (2)	0.07 (2)	0.05 (1)	0.07 (2)	0.07 (2)	0.06 (2)	0.07 (3)	0.05 (2)	0.05 (3)	0.08
CoO	0.06 (2)	0.07 (2)	0.06 (2)	0.09 (1)	tr.	tr.	0.08 (2)	0.06 (1)	0.06 (4)	0.08 (2)	0.05
MgO	10.87 (41)	10.27 (24)	11.45 (36)	11.01 (45)	13.19 (68)	14.03 (29)	11.40 (48)	13.13 (69)	10.80 (143)	10.88 (161)	10.81
Sum	100.86 (44)	99.99 (34)	99.86 (48)	100.65 (64)	99.94 (37)	103.10 (62)	103.12 (28)	100.45 (44)	99.28 (82)	99.91 (190)	98.54
Ti	tr.	tr.	tr.	tr.	0.01 (0)	tr.	tr.	tr.	tr.	tr.	tr.
Al	0.60 (3)	0.63 (2)	0.58 (2)	0.60 (2)	0.68 (1)	0.65 (1)	0.64 (1)	0.70 (2)	0.63 (2)	0.62 (5)	0.62
Cr	1.28 (5)	1.29 (2)	1.33 (2)	1.38 (2)	1.26 (1)	1.29 (0)	1.37 (1)	1.30 (2)	1.34 (2)	1.38 (3)	1.36
V	0.01 (0)	tr.	tr.	0.01 (0)	0.07 (0)	tr.	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01
Fe	0.59 (5)	0.58 (1)	0.53 (2)	0.47 (2)	0.36 (2)	0.42 (1)	0.45 (3)	0.37 (3)	0.47 (7)	0.49 (7)	0.47
Mn	tr.	tr.	tr.	0.01 (0)	tr.	tr.	0.01 (0)	tr.	0.01 (0)	0.01 (0)	0.01
Ni	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Co	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Mg	0.51 (2)	0.49 (1)	0.55 (2)	0.52 (2)	0.61 (3)	0.63 (1)	0.52 (2)	0.60 (3)	0.53 (6)	0.49 (6)	0.52
Σcat	7.88 (3)	7.91 (1)	7.91 (1)	7.99 (1)	8.02 (1)	7.94 (1)	8.01 (1)	8.00 (2)	7.98 (2)	8.00 (1)	7.99
Cr#	0.68 (2)	0.67 (1)	0.70 (1)	0.70 (1)	0.65 (1)	0.67 (0)	0.68 (0)	0.65 (1)	0.68 (1)	0.69 (2)	0.69
Mg#	0.52 (2)	0.50 (1)	0.55 (2)	0.53 (1)	0.63 (3)	0.64 (1)	0.54 (2)	0.62 (3)	0.54 (6)	0.50 (7)	0.53

Table S18. Composition of spinels (continued)

	PRB49 Hzb	PRB49 Opxite dyke	PRB49 Wbst dyke	PRB75 Hzb	PR31a Dun pod	MDS32 Gabbro dyke	MDS43 Wehrlite cumulate	MDS49 Dun cumulate	IDP32b Hzb μ	T7 Wbst dyke	BP34 Mafic μ
	6	8	6	13	16	6	12	7	12	5	4
(3)	0.01 (2)	0.02 (3)	0.04 (1)	tr.	0.03 (2)	0.32 (1)	0.11 (1)	0.10 (2)	0.17 (17)	0.12 (13)	0.48 (9)
(49)	16.88 (58)	19.78 (61)	21.44 (32)	16.52 (32)	14.59 (109)	21.06 (42)	19.75 (60)	25.75 (83)	25.26 (464)	15.28 (149)	22.38 (40)
(70)	56.19 (31)	52.89 (60)	49.43 (15)	53.88 (64)	57.27 (201)	45.67 (77)	47.76 (79)	46.37 (79)	44.33 (420)	46.33 (405)	30.05 (87)
(3)	0.14 (3)	0.18 (3)	0.18 (3)	0.17 (2)	0.19 (3)	0.34 (1)	0.22 (1)	0.20 (2)	0.20 (5)	0.26 (5)	0.23 (5)
(158)	15.12 (38)	14.85 (98)	14.30 (8)	16.09 (120)	19.64 (117)	21.19 (37)	20.79 (47)	17.76 (27)	20.28 (132)	27.25 (365)	35.27 (189)
(4)	0.21 (2)	0.20 (4)	0.20 (3)	0.21 (4)	0.24 (2)	0.45 (2)	0.24 (3)	0.22 (2)	0.21 (3)	0.28 (2)	0.29 (4)
(5)	0.04 (2)	0.03 (2)	0.05 (2)	0.08 (3)	0.06 (2)	0.17 (1)	0.08 (1)	0.07 (2)	0.11 (3)	0.11 (4)	0.22 (2)
(2)	0.08 (1)	0.07 (1)	0.06 (1)	0.08 (2)	tr.	0.14 (1)	0.06 (1)	tr.	tr.	0.07 (1)	0.10 (3)
(117)	11.57 (58)	11.93 (51)	12.56 (18)	11.57 (82)	11.31 (44)	11.27 (14)	11.04 (36)	12.96 (32)	12.78 (115)	9.43 (103)	7.30 (105)
(80)	100.37 (67)	100.02 (98)	98.34 (44)	98.65 (48)	103.49 (80)	102.06 (59)	100.13 (33)	103.49 (87)	103.37 (86)	103.66 (74)	96.35 (50)
(1)	tr.	tr.	tr.	tr.	tr.	0.01 (0)	tr.	tr.	tr.	tr.	0.01 (0)
(1)	0.63 (1)	0.73 (2)	0.79 (1)	0.63 (1)	0.54 (4)	0.77 (1)	0.73 (2)	0.90 (2)	0.88 (13)	0.59 (6)	0.86 (1)
(0)	1.41 (2)	1.31 (2)	1.23 (1)	1.37 (1)	1.41 (5)	1.12 (2)	1.19 (2)	1.08 (3)	1.04 (12)	1.19 (10)	0.78 (1)
(0)	tr.	tr.	tr.	tr.	tr.	0.01 (0)	0.01 (0)	tr.	tr.	0.01 (0)	0.01 (0)
(5)	0.40 (1)	0.39 (2)	0.38 (0)	0.43 (4)	0.51 (3)	0.55 (1)	0.55 (1)	0.44 (1)	0.50 (4)	0.74 (11)	0.97 (6)
(0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)
(0)	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	0.01 (0)
(5)	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
(5)	0.55 (2)	0.56 (2)	0.59 (1)	0.55 (3)	0.52 (2)	0.52 (1)	0.52 (1)	0.57 (1)	0.56 (4)	0.46 (5)	0.36 (5)
(1)	8.03 (2)	8.04 (1)	8.02 (1)	7.99 (1)	7.94 (3)	7.93 (1)	7.92 (1)	7.98 (1)	7.93 (1)	7.78 (6)	7.67 (3)
(0)	0.69 (1)	0.64 (1)	0.61 (0)	0.69 (0)	0.72 (2)	0.59 (1)	0.62 (1)	0.55 (1)	0.54 (7)	0.67 (3)	0.47 (0)
(5)	0.58 (2)	0.59 (3)	0.61 (0)	0.57 (4)	0.54 (2)	0.52 (1)	0.53 (1)	0.58 (1)	0.57 (4)	0.47 (5)	0.36 (4)

Table S19. Composition of feldspars

Sample Type	MDS32 Gabbro dyke	MDS37 Gabbro cumulate	IDP32b Hzb μ	BP34 Mafic μ	PR33 Gabbro	PRB38 Meladiorite	PRB39 Leucodiorite	RP40 Diorite	PRB70 Leucodiorite	PRB70 Leucodiorite	RP40 Diorite
n	16	15	8	8	10	12	6	5	13	2	3
SiO ₂	46.24 (112)	46.02 (45)	44.22 (74)	42.96 (57)	50.59 (75)	54.35 (244)	57.05 (65)	59.15 (96)	68.23 (193)	48.74 (66)	66.22 (56)
Al ₂ O ₃	32.62 (197)	36.86 (59)	37.60 (65)	38.27 (48)	32.22 (58)	30.08 (177)	28.81 (16)	26.41 (25)	20.75 (77)	36.23 (220)	19.06 (26)
CaO	16.93 (51)	18.45 (32)	18.93 (53)	20.26 (32)	14.56 (68)	11.29 (198)	10.01 (21)	8.09 (59)	1.43 (96)	0.07 (2)	0.22 (29)
Na ₂ O	1.84 (26)	1.19 (20)	0.96 (31)	0.27 (9)	3.66 (41)	5.40 (116)	6.02 (12)	7.12 (32)	11.21 (68)	0.07 (1)	0.65 (96)
K ₂ O	0.01 (1)	0.01 (1)	0.00 (1)	0.01 (1)	0.02 (1)	0.04 (2)	0.04 (2)	0.04 (1)	0.06 (2)	10.77 (22)	15.64 (154)
Sum	101.56 (68)	102.85 (77)	101.83 (82)	101.93 (79)	101.25 (55)	101.30 (52)	101.95 (70)	100.92 (70)	101.83 (152)	97.36 (75)	101.85 (84)
Si	2.17 (3)	2.06 (2)	2.01 (3)	1.96 (1)	2.28 (3)	2.42 (10)	2.51 (1)	2.62 (2)	2.94 (4)	2.29 (5)	2.99 (1)
Al	1.81 (4)	1.95 (2)	2.01 (3)	2.05 (1)	1.71 (3)	1.58 (10)	1.50 (1)	1.38 (2)	1.05 (4)	2.01 (10)	1.01 (1)
Ca	0.85 (3)	0.89 (1)	0.92 (3)	0.99 (2)	0.70 (3)	0.54 (10)	0.47 (1)	0.38 (3)	0.07 (4)	0.00 (0)	0.01 (1)
Na	0.17 (2)	0.10 (2)	0.08 (3)	0.02 (1)	0.32 (4)	0.47 (10)	0.51 (1)	0.61 (3)	0.94 (5)	0.01 (0)	0.06 (8)
K	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.65 (1)	0.90 (9)
Σcat	5.01 (1)	5.01 (1)	5.03 (0)	5.03 (1)	5.02 (0)	5.02 (0)	5.00 (1)	5.00 (1)	5.00 (1)	5.03 (0)	4.98 (1)
An	0.84 (2)	0.90 (2)	91.63 (268)	0.98 (1)	0.69 (3)	0.54 (10)	0.48 (1)	0.38 (3)	0.07 (4)	0.01 (0)	0.01 (1)

Table S20. Composition of plagioclase (trace elements)

Sample Type	MDS32 Gabbro dyke	MDS37 Gabbro cumulate	IDP32b Hzb μ	IDP33 Hzb μ	BP34 Mafic μ	PR33 Gabbro	PR32 Gabbro	PRB38 Meladiorite	PRB39 Leucodiorite	RP40 Diorite	PRB70 Leucodiorite	RP41 Tdj dyke	R33 Tdj pod
n	10	9	6	6	5	8	4	8	6	4	7	9	6
Li	2.07 1.09	0.38 0.19			1.18		1.73 0.55	1.15 0.67		0.43 0.09	1.31 1.09	0.20 0.06	1.53 1.17
Be	0.12	0.01 0.00	0.16		0.14 0.01	0.42 0.06		0.64 0.14	0.83 0.04	0.60 0.18	0.83 0.24	0.85 0.19	1.29 0.62
P	38.1 3.6	9.6 1.6	30.9 7.6	58.4 2.6	55.3 20.1	60.3 14.0	34.3 1.4	58.5 5.0	57.3 14.5	44.2 5.0	37.5 7.5	72.4 19.7	245.0 210.9
Sc	1.81 0.08	0.66 0.21	0.73 0.07	0.57 0.03	1.92 0.05	0.99 0.04	0.78 0.01	1.50 0.15	1.38 0.16	1.25 0.05	3.60 0.16	1.37 0.07	1.61 0.12
Ti	42.9 5.8	31.3 20.2	7.0 2.1	12.5 2.4	6.5 4.9	19.6 1.9	11.6 7.0	48.6 47.8	23.3 2.6	16.4 1.4	13.0 11.0	37.6 6.8	30.9 26.8
V	3.32 3.12	2.47 1.18	0.20 0.05		2.07 2.32	0.09 0.02	0.92 0.61	0.39 0.55	0.08 0.02	0.08 0.09	2.19 2.19	0.17 0.05	0.04 0.02
Mn	11.2 2.3	17.5 2.1	7.6 0.5	21.2 0.7	23.9 9.7	16.8 3.0	9.3 0.2	24.6 11.0	14.2 3.8	7.0 0.5	15.1 14.7	1.8 1.0	4.8 3.9
Co	0.32 0.09	0.35 0.06	0.20		0.28 0.20	0.04 0.00	0.08 0.01	0.17 0.19	0.06 0.02	0.04 0.02	1.42 1.53	0.04 0.01	83.9 85.3
Ni	0.73 0.58	0.35 0.31	1.47 0.65		11.43 9.77	0.27 0.02	0.41	0.19 0.09	0.86 1.25	8.08 14.74	4.85 5.50	31.37 34.61	1118 1020
Cu	15.35 42.76	0.76 0.07	1.16 0.40	0.84 0.14	3.70 4.71	2.66 0.54	1.02 0.12	2.81 0.45	7.95 10.83	5.21 1.42	64.3 139.2	5.92 0.21	5.68 1.60
Zn	4.19 4.83	1.81 0.29		0.61	4.02 4.95	0.84 0.10	0.81 0.86	3.40 1.51	2.54 0.56	0.66 0.23	99.8 256.3	1.06 1.20	1.93 1.59
Rb		0.01 0.00				0.30 0.13				1.78 2.50		0.34 0.29	28.89 48.53
Sr	26 12	44 1	399 15	701 54	721 17	539 30	20 6	733 35	708 10	689 96	764 228	331 51	329 184
Y	0.09 0.07	0.03 0.00			0.05 0.03	0.04 0.01		0.06 0.02	0.06 0.03	0.03 0.02	0.12 0.11	0.20 0.14	0.17 0.10
Cs	0.24	0.00 0.00			0.01					0.39 0.44	0.02 0.02		0.03 0.01
Ba	1.0 1.4	0.1 0.0	4.3 3.2	8.4 13.9	3.1 0.3	7.8 0.7	3.9	7.8 1.6	15.2 1.5	24.1 8.1	20.3 8.4	88.3 9.3	585.7 646.4
La	0.04 0.03	0.03 0.06	0.22 0.04	0.64 0.04	0.15 0.02	1.23 0.09		1.27 0.17	1.45 0.19	1.43 0.44	0.68 0.86	3.46 0.82	1.06 1.07
Ce	0.10 0.12	0.03 0.01	0.35 0.05	1.15 0.04	0.25 0.22	1.86 0.14	0.04	2.16 0.21	2.26 0.36	1.65 0.51	0.84 1.02	4.76 1.20	1.76 1.78
Pr	0.014	0.004 0.001	0.069	0.123 0.006	0.014	0.153 0.015		0.207 0.016	0.198 0.046	0.114 0.037	0.075 0.075	0.407 0.106	0.155 0.160
Nd	0.150 0.078	0.018 0.008		0.565		0.453 0.023		0.717 0.101	0.649 0.187	0.306 0.098	0.312 0.225	1.243 0.344	0.534 0.527
Sm		0.011 0.005			0.053			0.111 0.032	0.101 0.009	0.067	0.068	0.151 0.039	0.111 0.059
Eu	0.071 0.020	0.045 0.006		0.205 0.019	0.042 0.004	0.229 0.020	0.095 0.013	0.549 0.059	0.532 0.046	0.266 0.053	0.139 0.128	0.678 0.048	0.296 0.267
Pb	0.686 0.757	0.060 0.131	0.396 0.036		1.276 1.234	0.783 0.047		0.930 0.822	1.032 0.090	0.828 0.433	0.661 0.652	3.753 0.284	7.913 5.055

Table S21. Composition of amphibole

Sample Type	T7 Websterite dyke	IDP32b Harzburgite μ	IDP31 Hornblende 7	BP40 Mafic μ	BP34 Mafic μ	PR33 Gabbro 6	RP40 Diorite 7	PRB38 Meladiorite 7	PRB39 Leucodiorite 7	PRB70 Leucodiorite (core) 8	PRB70 Leucodiorite (rim) 6
n	32	15	7	11	7						
SiO ₂	48.71 (196)	46.01 (62)	42.04 (96)	47.92 (78)	47.18 (140)	49.11 (56)	50.24 (106)	46.55 (49)	47.68 (91)	49.05 (55)	49.55 (110)
TiO ₂	0.62 (22)	0.76 (29)	1.66 (24)	0.72 (7)	0.73 (24)	0.69 (5)	0.54 (7)	0.97 (5)	1.06 (12)	0.80 (4)	0.03 (3)
Al ₂ O ₃	7.66 (178)	13.30 (80)	13.85 (44)	8.44 (70)	10.01 (86)	8.47 (53)	6.96 (90)	7.97 (33)	8.81 (90)	6.33 (37)	2.03 (86)
Cr ₂ O ₃	0.38 (27)	0.75 (59)	0.17 (14)	0.29 (20)	0.25 (6)	0.39 (33)	0.22 (16)	0.00 (2)	-0.01 (2)	0.01 (4)	0.04 (6)
V ₂ O ₃	0.08 (2)	0.06 (2)	0.07 (2)	0.08 (2)	0.07 (3)	0.07 (1)	0.06 (2)	0.07 (2)	0.08 (1)	0.04 (1)	0.16 (10)
FeO	7.44 (176)	3.44 (23)	7.78 (71)	9.84 (58)	7.91 (92)	10.06 (17)	11.03 (49)	18.98 (62)	18.51 (24)	15.56 (33)	31.05 (257)
MnO	0.10 (1)	0.05 (2)	0.12 (1)	0.12 (1)	0.11 (2)	0.15 (2)	0.19 (3)	0.33 (3)	0.33 (3)	0.28 (2)	0.26 (12)
NiO	0.07 (2)	0.10 (2)	0.05 (1)	0.06 (1)	0.06 (2)	0.05 (1)	0.02 (1)	0.01 (1)	0.01 (4)	0.01 (1)	0.00 (1)
CoO	0.01 (0)	0.00 (0)	0.01 (1)	0.01 (1)	0.01 (1)	0.02 (1)	0.01 (1)	0.02 (1)	0.01 (3)	0.01 (1)	0.01 (1)
MgO	18.78 (133)	19.44 (28)	15.84 (38)	16.73 (38)	17.38 (88)	17.29 (28)	16.96 (69)	11.51 (29)	11.67 (48)	14.00 (30)	3.60 (186)
CaO	11.84 (32)	12.77 (11)	11.81 (29)	11.77 (22)	12.06 (24)	10.98 (14)	10.62 (78)	10.18 (37)	10.46 (30)	10.87 (25)	11.86 (16)
Na ₂ O	1.58 (34)	2.47 (12)	2.72 (23)	1.75 (14)	1.45 (11)	1.50 (13)	0.99 (17)	1.08 (7)	1.08 (17)	0.91 (7)	0.20 (7)
K ₂ O	0.12 (3)	0.03 (1)	0.10 (1)	0.13 (3)	0.13 (2)	0.05 (1)	0.07 (1)	0.06 (2)	0.12 (4)	0.14 (2)	0.13 (4)
Sum	97.38 (44)	99.17 (51)	96.21 (216)	97.86 (37)	97.35 (62)	98.83 (55)	97.92 (103)	97.74 (56)	99.82 (36)	98.03 (105)	98.92 (40)
Si	6.95 (22)	6.39 (8)	6.16 (4)	6.89 (8)	6.76 (13)	6.96 (8)	7.18 (10)	6.95 (5)	6.94 (12)	7.17 (4)	7.73 (10)
Ti	0.07 (2)	0.08 (3)	0.18 (2)	0.08 (1)	0.08 (3)	0.07 (1)	0.06 (1)	0.11 (1)	0.12 (1)	0.09 (0)	0.00 (0)
Al	1.29 (31)	2.18 (13)	2.39 (6)	1.43 (12)	1.69 (16)	1.41 (9)	1.17 (15)	1.40 (6)	1.51 (16)	1.09 (6)	0.38 (16)
Cr	0.04 (3)	0.08 (6)	0.02 (2)	0.03 (2)	0.03 (1)	0.04 (4)	0.02 (2)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (1)
v	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.00 (0)	0.02 (1)
Fe	0.89 (22)	0.40 (3)	0.95 (7)	1.18 (7)	0.95 (12)	1.19 (2)	1.32 (6)	2.37 (8)	2.25 (3)	1.90 (6)	4.05 (38)
Mn	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.02 (0)	0.02 (0)	0.04 (0)	0.04 (0)	0.04 (0)	0.03 (2)
Ni	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)
Co	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)	0.00 (0)
Mg	4.00 (25)	4.02 (5)	3.46 (11)	3.58 (7)	3.71 (15)	3.65 (6)	3.61 (14)	2.56 (6)	2.53 (10)	3.05 (4)	0.83 (42)
Ca	1.81 (4)	1.90 (2)	1.85 (2)	1.81 (3)	1.85 (3)	1.67 (2)	1.63 (12)	1.63 (6)	1.63 (5)	1.70 (3)	1.98 (1)
Na	0.44 (10)	0.66 (3)	0.77 (7)	0.49 (4)	0.40 (3)	0.41 (3)	0.27 (5)	0.31 (2)	0.31 (5)	0.26 (2)	0.06 (2)
K	0.02 (1)	0.00 (0)	0.02 (0)	0.02 (1)	0.02 (0)	0.01 (0)	0.01 (0)	0.01 (0)	0.02 (1)	0.03 (0)	0.03 (1)
Σcat	15.54 (11)	15.74 (3)	15.84 (5)	15.55 (5)	15.52 (5)	15.45 (4)	15.31 (4)	15.40 (4)	15.35 (6)	15.34 (2)	15.12 (3)
Mg#	0.88 (4)	0.91 (1)	0.81 (1)	0.83 (3)	0.80 (3)	0.85 (1)	0.85 (1)	0.59 (2)	(0)	0.71 (1)	0.21 (10)

Table S22. Composition of amphibole (trace elements)

Sample Type	IDP33 Hbz μ	IDP31 Hblite 8	BP32 Mafic μ	BP40 Mafic μ	BP41 Mafic μ	PR33 Gabbro 9	PRB38 Meladiorite 8	PRB39 Leucodiorite 8	RP40 Diorite 8	PRB70 Leucodiorite 8	RP41 Tdj 5
n	5	8	6	7	4						
Li	0.96 0.24	0.70 0.23	0.60 0.21	1.03 0.34	3.78 0.57	1.96 0.27	2.27 0.60	0.18 0.03	0.85 0.15	0.93 0.44	4.32 0.68
Be	0.200 0.010	0.167 0.015	0.077 0.016	0.142 0.017	0.100 0.030	0.154 0.022	0.246 0.034	0.275 0.036	0.227 0.023	0.454 0.083	0.064 0.023
P	101 9	92 14	59 3	52 2	68 2	84 6	256 591	60 9	60 5	102 122	38 4
Sc	50 4	52 6	110 5	119 11	94 16	62 5	120 36	127 26	64 8	100 26	47 1
V	367 50	355 49	440 35	580 66	321 95	396 38	406 21	485 106	362 25	424 286	331 40
Ti	8173 594	9379 1242	3502 415	4845 617	2329 950	3912 347	5504 762	5226 584	3120 348	4162 1193	484 36
Cr	1353 789	1793 861	3799 322	1115 209	4276 1327	1534 372		7 6 2 6	1654 984	71 57	176 67
Mn	986 15	920 43	766 30	941 45	739 34	1245 16	2580 107	2401 271	1551 46	2385 103	2620 147
Co	52.7 1.0	55.9 1.6	60.1 2.3	72.9 1.6	58.7 2.0	65.4 1.8	63.8 0.8	57.4 15.9	62.4 1.8	62.0 10.1	62.2 1.5
Ni	309.7 3.4	397.0 71.3	427.9 16.6	360.1 17.6	517.2 40.2	439.5 8.1	44.3 1.0	41.5 16.9	206.2 9.9	60.5 9.8	633.1 26.9
Cu	3.7 2.8	7.4 13.6	1.2 0.6	1.3 0.1	1.1 0.4	0.9 0.2	0.8 0.3	21.1 17.8	1.3 0.7	7.6 4.3	0.5 0.4
Zn	58.8 1.0	57.2 1.5	43.4 1.3	58.7 1.9	38.7 2.6	79.8 2.8	171.9 8.9	166.5 26.6	79.6 2.8	174.1 14.9	215.3 5.4
Rb	0.14 0.01	0.08 0.01	0.36 0.03	0.47 0.04	0.52 0.29	0.09 0.02			0.10 0.03		0.09 0.01
Sr	121.64 11.96	128.06 13.48	50.49 7.20	74.04 11.86	52.89 13.28	15.44 1.25	28.66 7.11	26.89 7.40	13.70 2.03	22.07 11.56	0.35 0.20
Y	28.47 2.81	31.51 2.75	13.97 0.91	14.71 1.29	10.17 3.55	19.88 1.35	40.64 5.54	40.35 4.79	14.84 2.72	35.97 10.50	9.07 0.54
Zr	38.86 4.92	39.96 4.27	15.91 2.02	24.21 3.70	18.19 3.95	34.61 2.21	44.98 2.79	41.15 7.66	28.88 3.78	35.58 6.36	1.37 0.26
Nb	0.68 0.15	0.51 0.05	0.31 0.10	0.74 0.11	0.40 0.19	0.80 0.06	1.77 0.13	1.70 0.50	1.07 0.19	1.14 0.33	0.08 0.02
Ba	5.14 0.31	3.50 0.33	2.96 0.34	7.43 1.20	5.81 3.70	1.63 0.25	2.59 0.77	2.87 0.37	1.79 0.50	2.18 0.57	0.54 0.46
La	1.62 0.15	0.87 0.05	0.76 0.07	1.00 0.10	0.92 0.16	1.33 0.08	2.79 0.98	3.96 3.05	2.20 0.28	4.22 0.65	0.25 0.10
Ce	7.64 0.99	4.74 0.26	3.38 0.31	4.06 0.43	3.02 0.58	6.81 0.48	14.40 3.47	16.04 2.61	10.01 1.67	21.22 4.21	0.10 0.06
Pr	1.69 0.32	1.17 0.07	0.70 0.07	0.85 0.11	0.54 0.14	1.33 0.09	3.06 0.56	3.11 0.17	1.74 0.37	4.20 1.01	0.07 0.03
Nd	11.27 2.76	8.55 0.61	4.61 0.46	5.85 0.80	3.36 1.19	7.91 0.48	19.31 2.87	19.07 1.23	8.92 2.04	24.61 6.66	0.34 0.12
Sm	4.10 1.05	3.48 0.26	1.90 0.15	2.43 0.29	1.28 0.52	2.66 0.20	6.87 0.88	6.74 0.63	2.57 0.56	8.09 2.44	0.16 0.02
Eu	1.38 0.16	1.30 0.12	0.70 0.08	0.81 0.08	0.42 0.12	0.81 0.05	1.44 0.21	1.47 0.16	1.01 0.19	1.56 0.34	0.04 0.02
Gd	5.01 1.01	4.93 0.43	2.56 0.17	3.10 0.36	1.70 0.71	3.29 0.22	8.06 1.30	7.81 0.95	2.68 0.55	8.76 2.63	0.34 0.02
Tb	0.80 0.11	0.82 0.08	0.41 0.03	0.48 0.05	0.27 0.11	0.53 0.04	1.22 0.21	1.19 0.15	0.41 0.08	1.26 0.38	0.10 0.01
Dy	5.49 0.64	5.81 0.52	2.80 0.21	3.06 0.26	1.90 0.74	3.65 0.28	7.91 1.35	7.74 0.90	2.72 0.50	7.52 2.27	1.08 0.05
Ho	1.11 0.10	1.23 0.12	0.56 0.04	0.60 0.06	0.40 0.14	0.77 0.06	1.58 0.26	1.55 0.20	0.56 0.10	1.41 0.43	0.31 0.01
Er	3.21 0.27	3.54 0.34	1.57 0.11	1.62 0.12	1.17 0.40	2.28 0.15	4.55 0.63	4.57 0.58	1.66 0.31	3.87 1.09	1.24 0.11
Tm	0.434 0.037	0.478 0.047	0.206 0.013	0.211 0.018	0.163 0.048	0.319 0.020	0.614 0.081	0.620 0.077	0.236 0.037	0.511 0.134	0.234 0.023
Yb	2.845 0.201	3.163 0.288	1.342 0.117	1.349 0.107	1.059 0.284	2.185 0.138	4.027 0.465	4.192 0.495	1.667 0.294	3.227 0.805	1.922 0.211
Lu	0.389 0.028	0.431 0.033	0.180 0.019	0.177 0.016	0.150 0.038	0.312 0.023	0.561 0.051	0.588 0.065	0.253 0.042	0.442 0.104	0.310 0.036
Hf	1.525 0.280	1.576 0.161	0.696 0.084	1.111 0.188	0.706 0.271	1.481 0.083	2.135 0.131	2.116 0.364	1.256 0.279	1.996 0.408	0.104 0.010
Ta	0.043 0.007	0.038 0.007	0.024 0.009	0.043 0.006	0.027 0.012	0.035 0.006	0.060 0.006	0.059 0.023	0.053 0.012	0.039 0.012	
Pb	0.180 0.137	0.078 0.009	0.080 0.014	0.215 0.047	0.260 0.074	0.345 0.849	0.081 0.016	0.147 0.058	0.095 0.017	0.956 0.949	0.256 0.284
Th	0.018 0.003			0.011 0.003	0.011 0.005		0.026 0.015	0.177 0.325	0.050 0.019	0.063 0.030	0.016
U				0.012 0.003	0.013 0.006		0.032 0.006	0.156 0.279	0.032 0.008	0.042 0.010	

Notes

