Regional-scale Grenvillian-age UHT metamorphism in the Mollendo–Camana block (basement of the Peruvian Andes)

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ABSTRACT The Mollendo–Camana Block (MCB) is a 50×150 km Precambrian inlier of the Andean belt that outcrops along the Pacific coast of southern Peru. It consists of stromatic migmatites of Paleoproterozoic heritage intensely metamorphosed during the Grenville event (c. 1 Ga; U-Pb and U-Th-Pb ages on zircon and monazite). In the migmatites, aluminous mesosomes (FMAS) and quartzofeldspathic leucosomes (KFMASH), contain various amounts of K-feldspar (Kfs), orthopyroxene (X_{Mg} Opx = 0.86), plagioclase (Pl), sillimanite (Sil; exceptionally kyanite, Ky) ilmenite (Ilm), magnetite (Mag), quartz (Qtz), and minor amounts of garnet (X_{Mg} Grt = 0.60), sapphirine (X_{Mg} Spr = 0.87), cordierite (X_{Mg} Crd = 0.92) and biotite (X_{Mg} Bt = 0.83). The ubiquitous peak mineral assemblage is Opx-Sil-Kfs-Qtz-(± Grt) in most of the MCB, which, together with the high Al content of orthopyroxene (10% Al₂O₃) and the local coexistence of sapphirine-quartz, attest to regional UHT metamorphism (> 900 °C) at pressures in excess of 1.0 GPa. Fluid-absent melting of biotite is responsible for the massive production of orthopyroxene that proceeded until exhaustion of biotite (and most of the garnet) in the southern part of the MCB (Mollendo-Cocachacra areas). In this area, a first stage of decompression from 1.1-1.2 to 0.8-0.9 GPa at temperatures in excess of 950 °C, is marked by the breakdown of Sil-Opx to Spr-Opx-Crd assemblages according to several bivariant FMAS reactions. High-T decompression is also shown by Mg-rich garnet being replaced by Crd-Spr- and Crd-Opxbearing symplectites, and reacting with quartz to produce low-Al-Opx-Sil symplectites in quartz-rich migmatites. Neither osumilite nor spinel-quartz assemblages being formed, isobaric cooling at about 0.9 GPa probably followed the initial decompression and proceeded with massive precipitation of melts towards the (Os) invariant point, as demonstrated by Bt-Qtz-(± pl) symplectites in quartz-rich migmatites (melt + Opx + Sil = Bt + Grt + Kfs + Qtz). Finally, Opx rims around secondary biotite attest to late fluid-absent melting, compatible with a second stage of decompression below 900 °C. The two stages of decompression are interpreted as due to rapid tectonic denudation whereas the regional extent of UHT metamorphism in the area, probably results from large-scale penetration of hot asthenospheric mantle at the base of an over-thickened crust.

Key words: Arequipa; Grenville; Peru; sapphirine; UHT metamorphism.

INTRODUCTION AND GEOLOGICAL SETTING

Precambrian inliers form the backbone of the South American cordilleras (Fig. 1) from Columbia (Irwin, 1975; Restrepo-Pace *et al.*, 1997), to northern Chile (Pacci *et al.*, 1981). A connection with the western part of the Amazon craton has been suggested for some of these inliers, based on the similarity of plutonic suites (e.g. Sadowski & Bettencourt, 1996) and isotopic signatures (e.g. Litherland *et al.*, 1989; Tosdal, 1996). In Peru, Precambrian rocks are found in the oriental cordillera (Audebaud *et al.*, 1971), in the occidental cordillera, north-west of Arequipa (Charcani gneisses) and near the towns of Omate, Moquegua and Tacna. Along the Pacific coast, Precambrian rocks outcrop from Paracas to Ilo (Dalmayrac *et al.*, 1977; Shackleton *et al.*, 1979). Shackleton *et al.* (1979)

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coined the term Arequipa Massif for a block of Precambrian gneisses that outcrops along the Pacific coast for 800 km and extends inland for c. 100 km. They subdivided this massif into (nonlimited) domains, three of which are more or less similar in their high metamorphic grade, namely Mollendo, Quilca and Camana (Fig. 2). They noted the ubiquitous presence of sillimanite, the occurrence of the assemblage orthopyroxene-sillimanite-quartz, the exceptional occurrence of sapphirine near Mollendo, in rocks that are otherwise described as migmatites, those from the Quilca area being more pegmatite-rich. Shackleton et al. (1979) attributed these parageneses to a M1 (Precambrian) metamorphism. These high-grade areas are separated from nonmigmatitic, muscovite-rich schists and migmatitic amphibolites of the Ocoña - Marcona segment (Fig. 1) by a phyllonite zone.

Several attempts at dating rocks of the Arequipa massif have been made since the preliminary Rb/Sr isochron of 1012 ± 52 Ma (Charcani gneisses of

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Fig. 1. Precambrian inliers in the Andean belt of Peru (modified from Shackleton *et al.*, 1979 and the tectonic map of South America, Almaida *et al.*, 1978); (1):Undifferentiated Precambrian rocks; (2): Proterozoic granulites; (3): Proterozoic greenschist to amphibolite facies rocks; (4): Phanerozoic sediments; (5): Mid-Palaeozoic to Triassic batholiths; (6): Andean batholiths; (7): Upper Cretaceous to Quaternary effusive rocks.

James & Brooks, 1976). A Rb/Sr whole-rock K-feldspar isochron on a granulite from the Mollendo area yielded an age of 1811 ± 39 Ma with an initial ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ of 0.7086 ± 0.0009 , an age interpreted as that of high-grade metamorphism (Cobbing et al., 1977). In the same area, another Rb/Sr isochron based on whole rock gave 1918 \pm 33 Ma with an 87 Sr/ 86 Sr initial ratio of 0.704 \pm 0.002 (Shackleton *et al.*, 1979). A preliminary U-Pb zircon plot from a granulite of the Mollendo area gave a ${}^{207}Pb/{}^{206}Pb$ age of c. 1950 Ma with a potential episode of lead loss at c. 600 Ma (Dalmayrac et al., 1977). More recently Wasteneys et al. (1995) based on U-Pb determinations from zircon concluded that two domains with different metamorphic ages exist within high-grade rocks along the coast of Peru: the first one located north of Quilca and dated at c.1198 \pm 4 Ma, the second one located north of Mollendo and dated at c. 970 Ma. Inherited zircon in both domains suggests a c. 1900 Ma age for the protolith of the Arequipa massif.

The aim of this contribution is to present new geochronological, structural, mineralogical and petrological data on granulite-grade rocks, ultra-high temperature assemblages (UHT; T > 900 °C) and associated migmatites that occupy the portion of the Arequipa massif located between the towns of



Fig. 2. Foliation trajectories in the Mollendo–Camana Block. Stereonets (lower hemisphere projection): poles to metamorphic foliations (black dots); mineral lineations (white stars): (1) Precambrian migmatites; (2): Precambrian (undated) granites; (3): Devonian to Carboniferous sediments; (4): Phanerozoic sediments; (5): Phanerozoic intrusive rocks (adapted from geological maps of the 'Servicio de Geología y Minería' (scale 1 : 100 000; Yesira, Aplao, Arequipa, Camana, Mollendo, La Joya, Punta de Bonbón).

Mollendo and Camana (Fig. 2). This area will be referred to here as the Mollendo–Camana block (MCB) and will be subdivided into the Camana, Quilca, Arantas and Pampa Blanca areas (northern part of MCB) and the Mollendo and Cocachacra areas (southern part of the MCB). This will add to the database for the few localities around the world where UHT metamorphic rocks are well exposed such as in Antarctica (e.g. Harley, 1998a and reference therein), Eastern Ghats (e.g. Grew, 1982), Hoggar (e.g. Kienast & Ouzegane, 1987), Italian Alps (Droop & Bucher-Nurminen, 1984) and Scotland (Baba, 1999). Quantitative pressure and temperature estimates and *P*-*T* trajectories of Peruvian UHT migmatites will be dealt with in a separate paper.

STRUCTURE

The high-grade migmatitic Mollendo–Camana Block (MCB) outcrops mainly along the shore of the Pacific, but also in deep V-shaped valleys that dissect the pampa, and on some hills rising a few hundred metres above the pampa (Fig. 2). However, even in areas of continuous outcrop, the highly monotonous lithological succession makes difficult the identification and mapping of large-scale structural elements.

Most rocks of the MCB are foliated migmatites with interlayered, more homogeneous gneisses. In the migmatites, the foliation is marked by the separation of irregular, alternating cm-scale light-coloured quartzofeldspathic (leucosomes) and dark sillimaniteorthopyroxene-(\pm oxides) layers (mesosomes). Quartzofeldspathic layers may be discontinuous but rarely anastomosing although they may constitute as much as 50% of the rock (stromatitic migmatites; Fig. 3). Locally, layering is more conspicuous with thinner layers, suggesting higher strain. Foliations are dominantly oriented E-W to NW–SE with varying dips (Fig. 2). The most common linear fabric is defined by the alignment of sillimanite prisms and is generally striking E-W in the Mollendo area and N-S in the Quilca area.

Cm- to m-scale folds are found in most outcrops including several occurrences of superimposed folds. Fold axes are dominantly E-W in the Mollendo area, namely parallel to the mineral lineation. Variations in foliation strikes and dips and in foliation trends at map scale suggest the presence of regional-scale folds that cannot be confidently traced due to the absence of marker horizons.

PETROGRAPHY

The main rock types in the MCB are migmatites (Fig. 3), aluminous gneisses, rare mafic granulites and granites. Pegmatite pods, aplitic sills and dykes, are common in the Quilca and Camana areas but rare in the Mollendo area. Most outcrops of the MCB can be described as genuine migmatites with well-segregated leucosomes and aluminous pods or layers that can be described as mesosomes. More homogeneous outcrops that lack the stromatitic structure of most migmatites are referred to as aluminous gneisses due to the ubiquitous presence of sillimanite. Undated muscovite pegmatites, Phanerozoic granites, and Mesozoic diabase dykes intersect the lithological units of the MCB. This contribution will be concerned essentially with the migmatites and the intercalated aluminous gneisses.

Migmatites

Leucosomes

Leucosomes consist of light, coarse-grained aggregates (Fig. 4) of mesoperthite, perthitic orthoclase or microcline (up to 60%), quartz (up to 50%), plagioclase (10–20%), with minor and variable amounts of orthopyroxene, sillimanite, oxides, biotite and traces of sapphirine. Grain size is usually larger than 1 mm especially concerning the orthopyroxene that is much larger than that in the mesosomes.

Mesosomes

The mesosomes are comprised of the following minerals: sillimanite (55-75%), orthopyroxene (0-15%), garnet (0-10%), cordierite (0-20%), magnetite-ilmenite (2-10%), biotite (0-10%), sapphirine and spinel (0-10%). Grain size ranges from 100 to 1500 μ m. Isolated prismatic sillimanite occurs within subinterstitial oxide grains (Fig. 5; sample 15, see Fig. 6 for location). Monomineralic sillimanite aggregates containing several tens of grains, with interstitial oxides, may reach cm-scale. Nests of strongly pleochroic pink to light green orthopyroxene surround sillimanite aggregates whereas isolated grains are interstitial or associated with oxides. Small idiomorphic, inclusionfree garnet and trails of inclusion-rich garnet occur within the matrix or at the interface between the mesosome and the leucosome. Oxides (mainly magnetite) constitute up to 10% of the mesosome especially in sillimanite-oxide layers.

Aluminous and semialuminous gneisses

Aluminous gneisses are devoid of the migmatitic habit although their mineral composition is similar to that of the migmatites. Grain size is intermediate between that of leucosome and mesosome and it is likely that some layers of aluminous gneisses are in fact thick mesosomes. Semi-aluminous (sillimanite-free) gneisses are less common and preferentially located in the Camana area. They consist of quartz (15–30%), feldspars (orthoclase and plagioclase; 50–60%), orthopyroxene (0–10%), garnet (0–10%), magnetite-ilmenite (0–5%), biotite (0–2%), spinel (0–10%), with accessory zircon and monazite.

Arrested reactions features are common in both the mesosomes and the aluminous gneisses. They involve garnet, cordierite, orthopyroxene, sapphirine, sillimanite, oxides, quartz and plagioclase. The most common reaction textures are represented by cordierite rims around garnet and sapphirine-bearing pseudomorphs after garnet or sapphirine-cordierite intergrowth at the contact between sillimanite and orthopyroxene. Reaction textures involving melts are also common in the leucosomes or at the leucosomemesosome interface.



Fig. 3. UHT migmatite showing at least 50% of quartzofeldspathic leucosome with centimetre-scale oxide clots; Quilca area.



Fig. 4. Sodium cobaltinitrite stained slab showing large amounts of K-feldspar in the leucosome; Cocachacra area.

Granites

Rocks that can be described as granites, to the exclusion of Palaeozoic and Mesozoic plutons, occur as dm-scale sills, less commonly as m-scale dykes or stocks. They are relatively common in Quilca and Camana areas. They are pink, hololeucocratic, aphyric, medium-grained (3 mm) rocks with a faint flow fabric defined by the alignment of sillimanite needles. Additional minerals are sparse garnet and biotite.

ROCK CHEMISTRY

In spite of the heterogeneity of the migmatites, attempts were made to obtain analyses from the leucosomes, the mesosomes and also from the aluminous gneisses and the granites. Some analyses were carried out by X-ray fluorescence while others were obtained by electron microprobe scanning of thin sections.

Analytical results are reported in Table 1 and have been projected from feldspar onto AFM and AS(FM) diagrams (Fig 7a,b).



Fig. 5. Texture of an orthopyroxene-sillimanite-quartz migmatite: note isolated prismatic sillimanite grains within large, interstitial, oxides in the lower left part of the picture; leucosomes made of orthoclase, quartz, plagioclase and large orthopyroxene (e.g. upper half of picture); mesosomes are more narrow and less abundant than the leucosomes (sample 15; Camana area).

All rocks are Ca-poor and Al-rich. The mesosomes are Al-rich (up to 37 wt% Al₂O₃) and Si-poor (37 wt% SiO₂), with a relatively high Ti and low but variable K (up to 2.4 wt% K₂O). The leucosomes are K-rich (up to 6.8 wt% K₂O) and their Al-content is such that they can be classified as peraluminous granites (Fig. 8). In AFM plots, most rocks are silica-saturated and have an $X_{Mg} < 0.50$. Such Fe-rich compositions are not compatible with the occurrence of Mg-rich minerals like sapphirine- and cordierite-bearing UHT granulites, much of which are silica- undersaturated and Mg-rich (e.g. Harley *et al.*, 1990; Raith *et al.*, 1997). The role of magnetite and the formation of mm-scale subassemblages will be discussed later in an attempt to reconcile bulk compositions and mineral assemblages.

U-Th-Pb AGE DETERMINATIONS

Given the poorly constrained age of high-grade metamorphism, Th-U-Pb chemical age determinations on monazite were attempted by *in situ* microprobe measurements using calibration and statistic treatment of Montel *et al.* (1996). Analyses were performed on a Cameca SX100 electron microprobe with four-wavelength dispersion spectrometer detectors at the laboratoire Magmas et Volcans of Clermont-Ferrand, France, according to the procedures followed by Goncalves, 2002). An accelerating voltage of 15 kV and a beam current of 150 nA were adopted. U and Th were analyzed successively with a PET crystal in the same WDS detector with a counting time of 225 s and 75 s on peak, respectively. Pb was analyzed with a LPET crystal during 300 s on peak. Standards were UO₂, ThO₂, vanadinite and a synthetic glass for Pb. Beam current used for standards is 100 nA. Counting time is 50 s on peak and 20 s on background for UO₂ and ThO₂, and 300 s on peak and 100 s on



Fig. 6. Sample locations and occurrences of selected minerals and mineral assemblages (for legend see Fig. 2).

background for PbO. Individual ages were calculated using the decay scheme of Montel *et al.* (1996) from the U, Th and Pb concentrations assuming that all lead in monazite is radiogenic. The 2 σ errors given on individual ages are calculated by propagating the uncertainties on U, Th and Pb concentrations (with 95% confidence level). Calculation of mean ages and associated errors (95% confidence level) is

based on least-squares modelling, which discriminates multiple age populations. The quality of modelling can be assessed from the mean square weighted deviation (MSWD).

Small monazite is found in most leucocratic aluminous gneisses from the Camana area (samples 10, 20, 28; Fig. 6). Sample 10 contains eight small monazite (10 μ m), on which 17 analyses were made. In sample 20, a garnet-orthopyroxene aluminous gneiss, 12 measurements were made on two monazite 50 μ m in size. For sample 28, an orthopyroxene-sillimanite gneiss, 22 measurements were made on five monazite ranging in size from 20 µm to 70 µm. Calculated individual ages from U. Th and Pb concentrations range from 1064 to 956 \pm 50 Ma (Table 2). The age distribution for sample 28 allows statistic calculation (Montel et al., 1996) that give meaningful ages attributable to a single metamorphic event at 998 \pm 11 Ma (MSWD: 1.07; Fig. 9). Likewise, statistical ages for samples 10 and 20 are, respectively, 994 ± 14 Ma (MSWD: 1.53), and 1011 ± 18 Ma (MSWD: 1.07). The same procedure was applied to monazite from Quilca, Mollendo, Pampa Blanca and Cocachacra area but the results are not as straightforward. For example, 52 analyses on 12 monazites (10–50 μ m) from a migmatite of the Mollendo area (sample 114) give dispersed ages with a maximum around 919 Ma. Although the dispersion of ages may have some significance, the geological knowledge of the area is not sufficient to provide a sound interpretation. Moreover, severe weathering may also contribute to Pb- and/or U-loss, hampering data interpretation. Based upon the already published isotopic determinations and the above chemical ages, the Mollendo-Camana block appears to be dominated by a quartzofeldspathic protolith dated at about 1.9 ± 0.1 Ga that was rejuvenated around 1.0 Ga during a regional high-grade metamorphic event.

MAJOR MINERALS: HABIT AND CHEMISTRY

Minerals were analyzed with a JEOL JXA-8900 L electron microprobe at McGill University with LiFH, PET, PETj and TAP analyzing crystals, operating at 20nA specimen current, 15 kV accelerating potential, and counting time of 20 s for major elements. Stoechiometry was calculated using THEBA 7.4 software (Martignole *et al.*, unpublished) from which Fe^{3+} estimates are obtained assuming charge balance. Mineral analyses are for the main phases present in migmatites, aluminous and semialuminous gneisses; they are plotted in an AFM diagram (Fig. 10; see Fig. 6 for sample locations). Abbreviations for minerals are from Kretz (1983).

Garnet

Garnet is found preferentially at the leucosome-mesosome interface although isolated grains are found in the leucosomes. It occurs, as mm-scale porphyroblasts (Grt I) with quartz, plagioclase, sillimanite and biotite inclusions. In some cases, porphyroblasts are corroded, and even strongly resorbed residual relicts of peak assemblages. Less frequently, garnet is inclusion-free, not resorbed and could represent a late phase. Garnet was analyzed from rim to rim with an average spacing of 10 μ m for some well-shaped round to elongated grains, whereas a single analysis for core and rim was performed in most cases. Analyzed garnet (Table 3) are almandinepyrope solid solution (Figs 11 & 17). Pyrope varies from 22 to 60 mol%, almandine from 32 to 60 mol%. The less magnesian garnet is from the Camana area and the most magnesian from the Mollendo area (Figs 11 & 17). Grossularite and andradite contents are < 5 mol%, the latter being in many cases more abundant than grossularite. Spessartine content is highly variable (usually 1-7 mol%), with very high values (> 20 mol%) in resorbed garnet. Most garnet is virtually unzoned except for minor rimward increase in almandine and decrease in pyrope. A spessartine-rich garnet from the Pampa Blanca area (sample 103) shows a plateau-type zoning with a rimward almandine zoning from 46 to 55%, correlated with a decrease in the pyrope content from 36 to 22%. Similarly, spessartine increases from 13 in the core to 22% at the

Table 1. Major and trace elements of leucosomes, mesosomes and granitic rocks from Quilca, Mollendo, and Cocachacra areas of the MCB.

Area	QU	QU	QU	МО	MO	МО	MO	МО	МО	МО	СО	СО	СО	СО	СО	СО
Sample	242*	245*	246*	51	53a(m)	53b(m)	55	63	81	220*	93	95	95*	95(m)	95(m)*	209(m)*
SiO ₂	73.92	73.30	73.04	72.15	48.99	61.15	69.04	77.37	75.74	76.57	78.75	73.13	87.63	52.45	36.67	67.58
TiO_2	0.04	0.08	0.06	0.77	1.20	0.81	0.67	0.47	0.60	0.14	0.60	0.26	0.05	1.45	1.97	0.89
Al ₂ O ₃	14.93	15.21	14.69	14.97	36.86	25.56	16.64	10.41	12.29	13.03	13.64	14.88	6.27	31.74	37.54	16.05
FeO	0.35	0.76	0.53	5.29	10.24	7.13	5.67	3.66	3.76	1.12	4.12	1.75	0.55	11.81	16.33	7.22
MnO	0.018	0.071	0.025	0.00	0.11	0.12	0.14	0.12	0.08	0.009	0.10	0.00	0.042	0.00	0.130	0.133
MgO	0.18	0.29	0.24	1.25	3.08	2.59	1.70	0.84	1.03	0.17	1.07	0.88	0.39	3.81	3.47	2.05
CaO	0.92	0.78	0.77	1.28	0.07	0.12	2.83	0.91	1.32	0.27	0.49	0.57	0.18	0.61	0.14	0.96
Na ₂ O	2.91	2.57	2.56	2.32	0.45	0.62	3.28	2.02	1.75	3.55	0.91	1.35	0.66	0.55	0.37	0.95
K_2O	5.46	5.79	6.85	3.05	2.44	3.12	0.86	1.66	2.77	4.26	1.84	6.27	3.34	0.79	1.83	2.08
Total	98.73	98.85	98.77	101.08	103.44	101.21	100.83	97.46	99.33	99.12	101.51	99.08	99.11	103.22	98.45	97.91
XMg	47.8	40.6	44.6	29.6	34.9	39.3	34.8	29.1	32.7	21.2	31.7	47.5	55.9	36.5	27.5	33.6

Quilca: QU, Mollendo: MO and Cocachacra: CO areas of the MCB; samples 242, 245 & 246: metre-scale anastomosing leucosomes containing sillimanite and rare orthopyroxene (granitic pods or sills); sample 53a: messosme (m); sample 53b (m): mixed mesosome-leucosome; samples 51, 55, 63 & 81: leucosomes; sample 220: stock of muscovite granite; sample 93: aluminous gneiss with evidence for solid-melt reaction; sample 95(m): mesosome from a stromatic migmatite (see Fig. 3); sample 95: leucosome from the same migmatite; sample 209(m): aluminous gneiss, mostly comprised of mesosome from a migmatite; samples 95(m)*, 95*, 209(m)*, 242*, 245*, 246* & 220*: analysed by X-ray fluorescence others by micro-probe scanning (courtesy A. Indares).



Fig. 7. (a) AFM and (b): S(FM)A plots of bulk compositions for analyzed rocks of the MCB (filled symbols: XRF analyses; white symbols, electron probe analyses; $F = FeO^{tot}$); sample numbers refer to those in Table 1.



Fig. 8. Metaluminous-peraluminous Al-(K + Na + 2 Ca) vs. Fe + Mg + Ti diagram of Debon & Le Fort (1988).

rim whereas the grossularite-andradite content is constant at about 3%. This type of zoning suggests garnet resorption consecutive to the formation of cordierite rims. An extreme case of resorption is found in the Cocachacra area (sample 93; Fig. 20c,f) where residual garnet (20 μ m) occupying the centre of a cordierite-orthopyroxene symplectite has spessartine up to 19 mol% (Fig. 11).

 Table 2. Monazite ages calculated from U, Th, Pb concentration in monazite from sample 28 (Opx-Sil-Qtz leucocratic gneiss).

Analytical spot	Th	U	Pb	Age (Ma)
1	93520 ± 774	1230 ± 167	$4410~\pm~104$	991 ± 36
2	93950 ± 777	1050 ± 167	$4380~\pm~103$	987 ± 36
3	44850 ± 573	2930 ± 171	$2570~\pm~83$	1027 ± 52
4	48730 ± 593	2840 ± 170	$2640~\pm~83$	992 ± 49
5	109460 ± 830	1640 ± 170	5080 ± 112	970 ± 32
6	96570 ± 785	$1310~\pm~168$	$4400~\pm~104$	957 ± 34
7	$49660~\pm~596$	$4050~\pm~174$	$2970~\pm~87$	1025 ± 47
8	51050 ± 604	$4220~\pm~175$	$3090~\pm~88$	1034 ± 46
9	$62680~\pm~656$	$3040~\pm~172$	$3320~\pm~91$	997 ± 42
10	52490 ± 609	$4210~\pm~176$	$3120~\pm~88$	1023 ± 45
11	53220 ± 613	$4000~\pm~175$	$3040~\pm~88$	997 ± 45
12	$50050~\pm~598$	$4200~\pm~175$	$3010~\pm~87$	1024 ± 46
13	90570 ± 765	$970~\pm~167$	$4190~\pm~101$	981 ± 36
14	$84300~\pm~743$	$1180~\pm~166$	$3840~\pm~98$	956 ± 37
15	81870 ± 733	$1130~\pm~165$	$3860~\pm~97$	989 ± 39
16	52700 ± 610	$4190~\pm~176$	$3050~\pm~88$	998 ± 45
17	48700 ± 594	$3180~\pm~172$	$2790~\pm~85$	1027 ± 49
18	$48280~\pm~590$	$2440~\pm~169$	$2620~\pm~83$	1015 ± 51
19	61970 ± 652	$2310~\pm~170$	$3210~\pm~89$	1009 ± 44
20	$87990~\pm~756$	$2410~\pm~172$	$4320~\pm~103$	986 ± 36
21	$45890~\pm~578$	$2300~\pm~168$	$2500~\pm~82$	1020 ± 53
22	52560 ± 611	$5000~\pm~179$	$3390~\pm~92$	$1064~\pm~45$

Th, U, Pb in p.p.m.; error limits at 95% confidence level.



Fig. 9. Weighted histogram of data for Sample 28. Each small bell-shaped curve represents the probability density function for one measurement (age and standard deviation). The dashed curve 'weighted histogram' is the sum of all individual bell-shaped curves. The central curve represents the ages calculated by the statistical procedure of Montel *et al.* (1996). *Y* axis is dimensionless.



Fig. 10. AFM plot of mafic minerals from the MCB.

Orthopyroxene

Orthopyroxene occurs in three main habits. The first type, termed Opx I, are large, mm-scale, strongly pleochroic porphyroblasts with magnetite or quartz inclusions and is found preferentially in the

Table 3. Representative compositions of minerals from Camana, Quilca, Arantas, Pampa Blanca, Mollendo and Cocachacra areas.

	Garnet											Orthopyroxene					
Sample Fig. in text	2 CA	2 CA	132 QU	139 AR	140 AR	103 PB 19c	62 MO	93 CO 20a	93 CO 20c	Sample Fig. in text	28 CA	28 CA 4	15 CA 19f	78 MO			
SiO ₂	38.03	39.23	38.54	39.19	39.04	37.80	40.03	38.20	37.61	SiO ₂	47.13	47.80	48.32	51.79			
TiO ₂	0.01	0.00	0.02	0.00	0.00	0.00	0.01	0.00	0.00	TiO ₂	0.10	0.09	0.08	0.00			
Al ₂ O ₃	21.33	22.40	22.13	22.24	22.40	21.64	22.67	21.87	21.82	Al_2O_3	10.33	9.80	9.48	5.95			
Cr ₂ O ₃	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	Cr_2O_3	0.04	0.02	0.00	0.01			
FeO	28.02	20.74	20.54	22.68	20.55	23.41	18.05	18.20	17.87	FeO	16.87	17.39	16.85	13.32			
MnO	1.54	3.06	4.38	1.62	3.31	7.19	1.01	8.91	12.50	MnO	0.98	1.00	0.82	0.40			
MgO	8.23	13.36	12.52	13.21	13.29	8.44	16.55	10.88	8.12	MgO	24.08	23.35	24.49	28.33			
CaO	1.92	1.31	1.06	0.98	1.05	1.21	1.62	1.05	1.08	CaO	0.06	0.06	0.06	0.05			
Na ₂ O K ₂ O F										Na ₂ O	0.00	0.02	0.02	0.00			
Total	99.08	100.10	99.19	99.92	99.64	99.69	99.95	99.12	99.00	Total	99.59	99.53	100.12	99.85			
Oxygen	12	12	12	12	12	12	12	12	12	Oxygen	3	3	3	3			
Si	2.99	2.95	2.95	2.96	2.95	2.96	2.96	2.95	2.96	Si	0.85	0.87	0.87	0.92			
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Ti	0.00	0.00	0.00	0.00			
Al	1.97	1.99	1.99	1.98	2.00	2.00	1.97	1.99	2.02	Al	0.22	0.21	0.20	0.12			
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	Cr	0.00	0.00	0.00	0.00			
Fe ³⁺	0.04	0.06	0.06	0.06	0.05	0.05	0.07	0.05	0.02	Fe ³⁺	0.07	0.05	0.06	0.04			
Fe ²⁺	1.80	1.25	1.26	1.38	1.25	1.48	1.04	1.12	1.16	Fe ²⁺	0.19	0.22	0.20	0.16			
Mn	0.10	0.20	0.28	0.10	0.21	0.48	0.06	0.58	0.83	Mn	0.02	0.02	0.01	0.01			
Mg	0.96	1.50	1.43	1.49	1.50	0.98	1.82	1.25	0.95	Mg	0.65	0.63	0.66	0.75			
Ca	0.16	0.11	0.09	0.08	0.09	0.10	0.13	0.09	0.09	Ca	0.00	0.00	0.00	0.00			
Na K										Na K	0.00	0.00	0.00	0.00			
X _{Alm}	0.59	0.41	0.41	0.45	0.41	0.49	0.34	0.37	0.38	X_{Mg}	0.65	0.69	0.69	0.77			
X _{Prp}	0.32	0.49	0.47	0.49	0.49	0.32	0.60	0.41	0.31								
X_{Sps}	0.03	0.06	0.09	0.03	0.07	0.16	0.02	0.19	0.27								

Camana: CA; Quilca: QU; Arantas: AR; Pampa Blanca: PB; Mollendo: MO; Cocachacra: CO areas. Garnet: 2 CA: high-Mg garnet; secondary low-Mg garnet; 132 QU: garnet surrounded by plagioclase; 139, 140 AR: garnet from leucosome; 103 PB: garnet surrounded by cordierite (see Fig. 19c); 62 MO: garnet from leucosome; 93 CO: pseudomorphs of a spessartine-rich garnet (see Fig. 20c). Orthopyroxene: 28 CA: porphyroblast of high-Al orthopyroxene; 15 CA: high-Al orthopyroxene porphyroblast from a leucosome; 78 MO: low-Al orthopyroxene rim around large secondary biotite (see Fig. 19c); 82 CO: saphirine pseudomorphs after sillimanite. Cordierite: 103 PB: cordierite moats around garnet (see Fig. 19c); 93 CO: cordierite in a cordierite-sapphirine symplectite (see Fig. 20b). Biotite: 140 AR: large biotite in contact with garnet; 78 MO: biotite surrounded by a rim of orthopyroxene (see matching Opx analysis and Fig. 19f). Spinel: 2 CA, 28 CA, 85 MO: small spinel in inclusion or in contaccut with oxide.

Table	3. ((Cont'd)	
1 ante	J. 1	Com a	

S	apphirine			Cordier	rite			Bioti	te		Spin	el		
Sample Fig. in text	28 CA	206 CO	Sample Fig. in text	140 AR 20c	103 PB 20c	93 CO	Sample Fig. in text	140 AR	78 MO 19f	78 MO 19f	Sample Fig. in text	2 CA	28 CA	85 MO
SiO ₂	11.59	13.46	SiO ₂	48.40	49.65	46.35	SiO ₂	38.12	37.90	38.28	SiO ₂	0.00	0.00	0.23
TiO ₂	0.00	0.04	TiO ₂	0.00	0.00	0.00	TiO ₂	2.58	2.53	3.25	TiO ₂	0.28	0.01	0.03
Al ₂ O ₃	62.01	59.12	Al_2O_3	31.17	31.68	34.05	Al_2O_3	16.71	16.28	15.61	Al_2O_3	60.26	63.49	60.87
Cr ₂ O ₃	0.22	0.09	Cr_2O_3				Cr_2O_3				Cr_2O_3	0.27	0.50	0.87
FeO	9.39	10.14	FeO	2.45	2.43	2.43	FeO	8.49	6.57	6.87	FeO	14.61	17.28	19.27
MnO	0.38	1.09	MnO	0.16	0.17	0.52	MnO	0.08	0.01	0.00	MnO	0.04	1.13	0.07
MgO	15.16	15.85	MgO	12.03	11.91	11.12	MgO	19.14	21.07	20.48	MgO	5.49	13.66	11.07
CaO			CaO				CaO	0.00	0.01	0.01	CaO	0.01	0.00	0.00
Na ₂ O			Na ₂ O	0.10	0.03	0.08	Na ₂ O	0.15	0.13	0.14	ZnO	19.65	4.03	7.10
							K_2O	9.78	9.83	10.22				
							F	0.787	1.24	1.17				
Total	98.75	99.79	Total	94.31	95.87	94.55	Total	95.84	95.57	96.03	Total	100.61	100.10	99.51
Oxygen			Oxygen	18	18	18	Oxygen	11	11	11	Oxygen	4	4	4
Si	0.71	0.82	Si	5.06	5.10	4.85	Si	2.76	2.75	2.77	Si	0.00	0.00	0.01
Ti			Ti	0.00	0.00	0.00	Ti	0.14	0.14	0.18	Ti			
Al	4.46	4.23	Al	3.84	3.83	4.20	Al	1.43	1.39	1.33	Al	2.00	1.97	1.95
Cr	0.01	0.00	Cr				Cr				Cr	0.01	0.01	0.02
Fe ³⁺	0.16	0.18	Fe ^{tot}	0.21	0.21	0.21	Fetot	0.51	0.40	0.42	Fe ³⁺	0.00	0.03	0.04
Fe ²⁺	0.31	0.33	Fe ²⁺	0.34	0.36	0.40								
Mn	0.02	0.06	Mn	0.01	0.01	0.05	Mn	0.00	0.00	0.00	Mn	0.00	0.03	0.00
Mg	1.38	1.44	Mg	1.88	1.82	1.73	Mg	2.07	2.28	2.21	Mg	0.23	0.54	0.45
Ca			Ca				Ca				Ca	0.00	0.00	0.00
Na			Na	0.02	0.01	0.02	Na	0.905	0.909	0.944	Zn	0.41	0.08	0.14
			K				K							
X_{Mg}	0.81	0.81	X_{Mg}	0.90	0.90	0.89	X_{Phl}	71.03	77.26	76.24	X_{spl}	0.23	0.54	0.45
											X _{her}	0.35	0.36	0.40
											X_{gah}	0.42	0.08	0.14



Fig. 11. Almandine-pyrope-(spessartine + grossularite) plot of garnet from the MCB.

leucosomes. The second type, termed Opx II, are small, a few tenths of a mm, inclusion-free grains, either isolated in the leucosome where they are not fundamentally different from the previous type except for the size, or rims around biotite, cordierite or garnet (Fig. 19d,f). The third type, also termed Opx II without any implication of contemporaneity with the previous type occurs as symplectites, usually with cordierite. Orthopyroxene (Opx I & II; Table 3) are enstatiterich (en60 to en82; Fig. 10). On a regional basis, the enstatite content (or $X_{Mg} = Mg/Mg + Fe^{2+}$) of orthopyroxene increases from Camana to the Mollendo area (Fig. 12), a variation that correlates with that of pyrope in garnet (Figs 11 & 17). Most



Fig. 12. X_{AIM1} vs. X_{Mg} plot for orthopyroxene of the MCB; all but sample 2 are sillimanite bearing.

orthopyroxene from the MCB is highly aluminous, Al₂O₃ content ranging from 4.27 to 10.33%, which results in Mg-tschermak content ranging from 17 to 35%. Some large orthopyroxene (Opx I) are zoned in Al₂O₃, from about 10% ($X_{A1} > 0.19$) in the core to about 7.5% ($X_{A1} < 0.18$) towards the margins (Fig. 13). The highest Al₂O₃ values are found in porphyroblastic orthopyroxene I from the leucosomes in the Camana area (sample 28: 10.33%), the Mollendo area (sample 85: 10.06%) and the Cocachacra area (sample 95, 10.13%). The lowest Al₂O₃ values are from small secondary pyroxenes growing at the expense of biotite (e.g. Mollendo area, sample 78: $X_{Mg} = 0.82$; Al₂O₃ = 5.65–6.95 wt%) or from pyroxene



Fig. 13. Al and Mg zoning in an orthopyroxene porphyroblast (sample 132).

from the Camana area (sample 2: $X_{Mg} = 0.65$, $Al_2O_3 = 4.27$ wt%). We note that Al content tends to decrease with increasing X_{Mg} in the orthopyroxene. Compared to orthopyroxene from other UHT areas worldwide, those of the MCB are richer in Fe³⁺, as calculated from charge balance. Fe³⁺/Fe^{tot} ranges from 0.15 to 0.25 and in a sample from the Mollendo area (sample 62) almost 50% is Fe³⁺ (Table 4), attesting to the extent of the paired substitution: Fe²⁺ + Si = Fe³⁺ + Al. Finally, MnO reaches 2.7% in pyroxene coexisting with

Table 4. Chemical characteristics of principal minerals.

spessartine-rich garnet of the Cocachacra area, whereas Na and Ca contents are very low in all types of orthopyroxene.

Sapphirine

Two main types of sapphirine are found in the MCB. The first type comprises relatively large sapphirine (up to a few tenths of a mm) associated with Fe-Ti oxides and separated from quartz by a thin plagioclase aureole. This sapphirine is rare and occurs mainly in aluminous gneisses of the Camana area and its condition of formation is still unclear. The second type are smaller sapphirine occurring in symplectites together with cordierite and orthopyroxene, in samples showing evidence for garnet or orthopyroxene breakdown (essentially in mesosomes from the Cocachacra area). It also occurs as dactylitic intergrowths stemming from corroded sillimanite. All the analyzed sapphirine (Table 3) is relatively Fe-rich (0.79 < $X_{\rm Mg}$ < 0.87; calculated with Fe²⁺ only; Fig. 10). Based on a 2-2-1 structural formula (7 cations normalized to 10 oxygen and charge balance determined Fe^{3+} up to 0.29), Si ranges from 0.71 to 0.85 and Al from 4.19 to 4.54. This sapphirine thus tends towards the 7.9.3 pole of the solid solution (Fig. 14), the most aluminous ones being from the Camana area, those from Cocachacra being closer to the 2-2-1 pole. The Fe^{3+}/Fe^{tot} is highly variable, but always very high (from 0.16 to 0.56; Table 4), resulting in noticeable differences in $X_{\rm Mg}$ whether calculated with Fe²⁺ or Fe^{tot}.

Cordierite

Cordierite occurs in several habits and assemblages: large cordierite grains surrounding corroded garnet in sillimanite-bearing migmatites of the Camana, Quilca and Pampa Blanca areas; small grains associated with orthopyroxene in the Arantas area; small (few tenths of a

		Opx						G	rt		Crd		Bt		Spr		
		X _{Mg}		Al ₂ O ₃		Fe ³⁺	X_{Mg}	X _{Ca}	X _{Mn}	X _{Mg}	X _{Mg}	TiO ₂	F	X _{Mg}	Fe ³⁺ /Fe ^{tot}		
Sample	min	max	\bar{x}	min	max	\bar{x}	x										
								CAMANA									
2	0.65	0.68	0.66	4.27	6.19	5.63	0.03	0.43	0.05	0.04							
15	0.75	0.79	0.77	6.84	9.48	8.29	0.05										
20								0.54	0.03	0.08		0.80	3.31	1.43			
28	0.74	0.80	0.77	6.61	10.33	9.28	0.06	0.53	0.03	0.10		0.81	3.66	1.18	0.83	0.38	
								QUILCA									
43	0.73	0.79	0.75	7.50	8.89	8.18	0.05	0.48	0.03	0.07							
132	0.74	0.80	0.78	7.58	9.58	8.82	0.06	0.53	0.03	0.09		0.84	3.45	1.15			
								ARANTAS									
138	0.75	0.79	0.77	6.47	8.45	7.57	0.06	0.48	0.03	0.03							
139	0.73	0.76	0.74	7.68	8.76	8.19	0.05	0.52	0.03	0.03							
140	0.74	0.76	0.75	6.96	8.01	7.44	0.04	0.52	0.03	0.07	0.90	0.80	2.88	0.82			
							PAM	MPA BLAN	NCA								
102								0.46	0.04	0.08	0.89	0.74	3.99	0.31			
103								0.42	0.03	0.14	0.88	0.75	3.22	0.97			
							N	IOLLEND	0								
53	0.80	0.83	0.82	7.04	9.06	8.26	0.06	0.58	0.02	0.04							
62	0.83	0.89	0.85	7.00	8.05	7.77	0.10	0.62	0.04	0.02							
74	0.78	0.80	0.79	6.73	8.14	7.65	0.04	0.61	0.04	0.04							
78	0.82	0.82	0.82	5.67	6.95	6.22	0.03	0.58	0.04	0.04		0.85	2.92	1.22			
80	0.78	0.84	0.80	7.78	9.33	8.44	0.05	0.59	0.03	0.03							
85	0.79	0.83	0.81	7.35	10.06	8.72	0.06	0.59	0.05	0.02							
112	0.80	0.87	0.82	6.95	8.49	7.68	0.05	0.60	0.05	0.03							
113	0.78	0.85	0.80	5.48	8.52	7.46	0.04	0.61	0.04	0.03		0.89	2.57	1.66			
120	0.80	0.88	0.83	5.62	8.89	7.66	0.06	0.60	0.04	0.05							
122	0.79	0.82	0.81	6.11	8.28	7.14	0.05	0.58	0.05	0.03		0.83	3.12	0.90			
123	0.81	0.84	0.82	6.19	8.06	7.29	0.04	0.64	0.04	0.04							
124	0.80	0.85	0.82	6.72	8.65	7.62	0.06	0.60	0.04	0.05							
127	0.79	0.83	0.81	6.29	7.54	6.96	0.04	0.60	0.05	0.03		0.86	3.22	1.71			
							CC	OCACHAC	RA								
93	0.77	0.80	0.78	8.34	9.48	8.91	0.06	0.51	0.03	0.20	0.91				0.83	0.38	
95	0.76	0.80	0.78	6.74	10.13	8.16	0.05	0.54	0.04	0.06	0.91				0.85	0.51	
95A								0.53	0.04	0.09							



Fig. 14. Composition of sapphirine (structural formulae normalized to 10 oxygen) plotted in terms of total Al + Fe^{3+} vs. Si + Mg + Fe^{2+} full line: ideal solution between 7.9.3 and 2.2.1 sapphirine.

mm) and even μ m-scale cordierite in microscopic and submicroscopic symplectites together with orthopyroxene and eventually sapphirine (Cocachacra area); and myrmekite-like interstitial pockets made of vermicular intergrowths of cordierite, K-feldspar, plagioclase and quartz. No cordierite has yet been found in the Mollendo area. All analyzed cordierite (Table 3) is highly magnesian, compositionally homogeneous at the thin section scale with minor regional variations of X_{Mg} (from 0.86 to 0.92; Fig. 10; Table 3). The oxide total is usually between 93.9 and 97.5%, which suggests the presence of substantial amounts of fluids in the structure. CO₂ and H₂O were analyzed by Secondary Ion Mass Spectrometry (SIMS) at the University of Edinburgh according to the procedures described in Thompson *et al.* (2001) and Harley & Carrington (2001). In the four analyzed cordierite (samples 102 & 103: Pampa Blanca area; sample



Fig. 15. CO_2 and H_2O content of cordierite (analyses, University of Edinburgh, S. Harley).

93 & 206: Cocachacra area), H_2O is close to 1 wt% whereas CO_2 varies from 0.6 to 1.3wt%, corresponding to a total oxide wt% between 98 and 98.4 (Fig. 15).

Feldspar

K-feldspar (mesoperthite, perthitic orthoclase or microcline) occurs together with quartz (and occasionally plagioclase) in the leucosomes whereas it is less abundant and randomly distributed in the matrix of nonmigmatitic gneissic rocks. The presence, in the same area, of mesoperthite, perthitic orthoclase, microcline and plagioclase indicates a long evolution from hyper-solvus to subsolvus crystallization and subsolidus recrystallization. Plagioclase most commonly occurs as coarse andesine grains or as rims around more or less resorbed garnet.

Biotite

Biotite is abundant in Camana, Quilca and Arantas areas, in the mesosomes and the aluminous gneisses. It is rare or absent in migmatites from the Mollendo and Cocachacra areas. Biotite is phlogopite-rich ($X_{Mg} = 0.74-0.89$; Fig. 10; Table 3) with relatively high Al₂O₃ contents (from 14.74 to 17.85 wt%). Fluorine contents vary from low (< 0.4% wt%: Pampa Blanca area) to high (1.8 wt%: Arantas and Mollendo areas). No extreme values of fluorine content have been found but according to experimental data on vapourabsent melting of biotite (e.g. Patiño Douce & Johnston, 1991) values above 1% F are compatible with partial de-hydroxylation of primary biotite at temperatures of c. 1000 °C. Fluorine-poor biotite, on the other hand, may represent that precipitated from melts. The TiO₂ content of biotite shows a negative correlation with X_{Mg} (Fig. 16), from > 3.5% in the Camana, Quilca and Pampa Blanca areas to <3.5% for the Mollendo and Arantas areas. Although increased Ti stabilizes biotite towards higher temperatures, this effect may be tempered with increasing X_{Mg} . Increasing fluorine contents also has a stabilizing effect that may compensate for Ti decrease in Mg-rich biotite (Table 3).

Aluminium silicates

Sillimanite is concentrated in the mesosomes all over the MCB, but isolated grains also occur in the leucosomes. Several samples (e.g.



Fig. 16. TiO₂ vs. X_{Mg} plot for biotite of the MCB (X_{Mg} calculated with Fe^{tot}).

sample 79, 206) of the Mollendo and Cocachacra areas contain large, subrectangular grains with an oblique extinction, identified as kyanite on the basis of Raman spectra. Kyanite has been identified occasionally from UHT terranes in India and Scotland (Raith *et al.*, 1997; Baba, 1999). Analyses of a few sillimanite grains have given about 0.1 Fe^{3+} (charge balance calculation on a 20 O basis).

Spinel and Fe-Ti oxides

Spinel is rare, preferentially found in the mesosomes where it is usually included within or juxtaposed to magnetite or hemo-ilmenite. According to textural criteria, spinel most likely results from granular exsolution of a more complex Al-Fe-Mg-Zn-Ti-Mn-Cr solid solution. In some cases, it contains exsolutions of magnetite (sample 95) or corundum. Occasionally, it occurs as symplectite within biotite. Where it is not entirely entrapped within magnetite, spinel may be in contact with sillimanite, plagioclase or garnet, but not with quartz. Analyzed spinel (Table 3) are hercynite-spinel-gahnite solid solutions with low concentrations of galaxite (MnO ranging from 0.02 to 1.13 wt%) and very low concentration of picotite (Cr2O3 ranging from 0.16 to 0.87 wt%). The hercynite-spinel ratio ranges from 29 to 64 with variable but relatively high amounts of gahnite (from 5 to 42%) and a tendency for gahnite to be incorporated in hercynite-rich phases. These gahnite values are well above average for both granulite facies and UHT rocks. Reintegration of a pre-exsolution spinel composition is hampered by the lack of readily interpretable textures. However, according to the modal proportions of magnetite and Al-spinel, the original phase must have been oxidized and iron-rich ($X_{Mag} > 0.5$) with minor amounts of Mg, Zn and Ti.

The general pattern from the above analytical data is, $X_{Mg}Crd \ge X_{Mg}Bt \ge X_{Mg}Opx \ge X_{Mg}Spr > X_{Mg}Grt$ (Fig. 10). Unfortunately, spinel composition being dependent upon the amount and nature of exsolution, its actual X_{Mg} may not be representative of X_{Mg} at peak conditions. It should be noted that the close partitioning of Fe and Mg between several phases is likely to favour degenerate reactions thus, giving a false impression of high variance of the system.

Fe-Mg PARTITIONING BETWEEN GARNET AND ORTHOPYROXENE AND AI-CONTENT OF ORTHOPYROXENE

Garnet and orthopyroxene are ubiquitous Fe-Mg minerals in the aluminous (Sil-bearing) migmatites of the MCB. Owing to the possibility of Fe-Mg exchange between these two minerals, $K_{\text{DGrt-Opx}}$ (calculated as (Fe/Mg)Grt/(Fe/Mg)Opx from core compositions) is sensitive to temperature and may be used as a semiquantitative approach of temperature variations in the area. Average K_{D} values for rocks of the MCB range from 2.01 to 2.82 with the lowest values (highest *T*) recorded in the dry rocks of the Mollendo area, highest values (lowest *T*) being found in samples from the Arantas and Cocachacra areas (Table 5). Likewise, the high alumina content of orthopyroxene from the MCB is probably the result of a high temperature of crystallization, and as Al₂O₃ relationship is expected. However, in the orthopyroxene analyzed from MCB rocks, Al₂O₃

Table 5. K_D Opx-Grt and Al-content of orthopyroxene (all but sample 2 are sillimanite-bearing).

Sample	Fe/Mg grt	Fe/Mg opx	Al ₂ O ₃ -max	Al ₂ O ₃ -opx	
			CAMANA		
2	1.361	0.566	2.40	6.19	5.63
28	0.93	0.401	2.32	9.80	8,38
			QUILCA		
132	0.911	0.384	2.37	9.58	8.64
43	1.093	0.404	2.71	8.89	8.18
			ARANTAS		
138	1.089	0.386	2.82	8.45	7.57
139	1.017	0.420	2.42	8.76	8.19
140	0.973	0.391	2.49	8.01	7.44
			MOLLENDO		
53	0.74	0.301	2.46	9.06	8.26
62	0.638	0.317	2.01	8.05	7.77
74	0.665	0.327	2.03	8.14	7.65
80	0.706	0.320	2.21	9.33	8.44
85	0.713	0.309	2.31	10.06	8.72
112	0.685	0.308	2.22	8.49	7.84
113	0.684	0.307	2.23	8.52	7.46
120	0.696	0.294	2.37	8.89	7.66
122	0.771	0.303	2.54	8.28	7.14
123	0.598	0.276	2.17	8.06	7.27
124	0.665	0.299	2.22	8.65	7.62
127	0.691	0.289	2.39	7.49	6.96
			COCACHACRA		
93	0.985	0.365	2.70	9.48	8.91
95	0.903	0.357	2.53	10.13	8.16
95 A	0.919	0.357	2.57		8.16

varies from about 5–10 wt% independently of $K_{\text{DGrt-Opx}}$, evidence for the decoupling of $K_{\rm D}$ and alumina contents already noted in other UHT rocks (e.g. Harley, 1998b). At the thin section scale, the X_{Mg} of orthopyroxene is constant, whereas the Al₂O₃ varies by as much as 3 wt%. Moreover, the lack of correlation between $K_{\rm D}$ and Al content of orthopyroxene may be related to the type of pyroxene. For instance, samples 62 and 74 (Mollendo area) have the lowest $K_{\rm D}$ values and thus potentially the highest temperature of equilibration. However, these samples are characterized by small granoblastic pyroxene (Opx II) notably less aluminous than the porphyroblastic aluminous pyroxene found in the leucosomes from the same area. Rather than diffusion, these variations in Al2O3 may reflect crystallization of orthopyroxene during changing temperature conditions. On the contrary, the quasi-constant X_{Mg} (Fig. 17) may reflect some kind of re-equilibration at a single temperature, the lowest $K_{\rm D}$ not necessarily representing peak temperature.

CHEMICAL SYSTEMS AND PETROGENETIC GRIDS

Rocks of the MCB are usually quartz-bearing at the sample scale but a distinction is made between leucosomes that are always quartz-bearing, and mesosomes that are usually quartz-free. At the sample scale and particularly in the presence of a melt fraction, the



Fig. 17. A/AFM vs. X_{Mg} plot of orthopyroxene and garnet from the five areas of the MCB.

availability of quartz may have been limited during the peak of metamorphism inasmuch as the activity of SiO_2 in the melt was lower than unity. Several mineral assemblages can thus be legitimately treated as quartz-free.

Except for cordierite, the only hydrated mineral in the migmatites and the aluminous gneisses of the MCB is biotite. Biotite-bearing assemblages (up to 20% Bt) are preferentially found in the Camana, Quilca, Arantas and Pampa Blanca areas, where they may be juxtaposed with anhydrous (orthopyroxene-bearing) assemblages at the outcrop, sample or thin section scale. In the Mollendo area, biotite is rare and occurs only in the mesosomes whereas this mineral is virtually absent from rocks of the Cocachacra area. Generally speaking rocks of the MCB can be described as dominantly anhydrous with layers and pods containing small amounts of hydrous minerals.

The FKMASH and FMAS grids

Mineral assemblages of the migmatites from the MCB can be represented by the KFMASH system for biotite (and/or K-feldspar)-bearing rocks that contain evidence of partial melting (perthite-rich veins and pods) and the FMAS system for biotite-, K-feldspar-, H₂O-free assemblages (essentially melt-absent, anhydrous assemblages that constitute the mesosomes). The P-T grid for the KFMASH system (applicable to the leucosomes) has nine potential invariant points (invariant points, univariant, bivariant and trivariant assemblages are designated according to the absent phase conven-

tion), among which (Os) and (Bt) are relevant to the P-T domain of UHT metamorphism under consideration (Fig. 18a). These points are well constrained by experimental work (Carrington & Harley, 1995) and thermobarometric calculations (Harley, 1998b). The *P*-*T* grid for the FMAS system (applicable to the mesosomes and to some aluminous gneisses) only allows solid-solid equilibria and permissible minerals are orthopyroxene, garnet, sillimanite, cordierite, spinel and sapphirine (\pm quartz). However, two types of petrogenetic grids are available for the FMAS system at high temperature. The first one (Hensen, 1971; revised by Harley, 1998b) is valid for low fO_2 and is supported both by experimental and field evidence. The second one (Vielzeuf, 1983; Hensen, 1986) is valid for high fO_2 conditions but it is less well constrained experimentally, although it has been successfully applied to some field occurrences. Pending more elaborate experiments relative to the influence of fO_2 on phase equilibria in FMAS, the low fO_2 grid (Fig. 18b) will be applied here because it fits more adequately the observations. Univariant reactions emanating from invariant points (Spl) (Opx) and (Qtz) in the case of FMAS and those that emanate from (Os) and (Bt) invariant points in the case of KFMASH are the most relevant here.

REACTION TEXTURES

Relatively few textures from the prograde path are expected to have resisted the high temperatures attained. However, mineral assemblages that involve porphyroblastic phases with extreme chemical



Fig. 18. (a) P-T grid for the KFMASH system (from Carrington & Harley, 1995) with shaded fields relevant to the migmatites of the MCB area (numbers in circles correspond to univariant reactions; others correspond to bivariant reactions); (b) P-T grid for the FMAS system (from Hensen, 1971; modified by Harley, 1998b) with shaded areas relevant to the mineral assemblages of the mesosomes from the MCB.

Table 6.	Summary	of FMAS	reactions a	and mineral	assemblages	for the MCB.
					6	

		Area						
Reactions FMAS System	Figure	CA	QU	PB	AR	МО	СО	Textures
Univariant Reactions related to (Spl) invariant point garnet + cordierite = orthopyroxene + sillimanite + quartz (Spl, Spr)								
Related bivariant assemblages observed:								
garnet - sillimanite - quartz - cordierite (Spl, Spr, Spx)	19c	+	+	+				Crd moats around Grt
orthopyroxene - sillimanite - quartz - cordierite (Spl, Spr, Grt)					+			Crd rims around Opx in Qtz Sil-bearing rocks
garnet - quartz - orthopyroxene - sillimanite (Spl, Spr, Crd)	19d, e					+		checkerboard Opx-Sil around Grt
garnet - quartz - orthopyroxene - cordierite (Spl, Spr, Sil)	20e						+	Grt relict in an Opx-Crd symplectite
orthopyroxene + sillimanite = garnet + sapphirine + quartz (Spl, Crd)								
Related trivariant assemblage observed								
garnet-sapphirine-quartz	20a						+	Corroded Grt in contact with Spr in Qtz matrix
orthopyroxene + cordierite + sillimanite = sapphirine + quartz (Spl, Grt)								not observed
orthopyroxene + sillimanite = garnet + sapphirine + cordierite (Spl,Qtz)	20d						+	Spr in a Crd aureole around Sil in contact with Opx
Related bivariant assemblages observed								*
orthopyroxene - sillimanite - sapphirine - cordierite (Spl, Qtz, Grt)	20b, c						+	Spr-Crd symplectite in contact with corroded Sil
garnet - orthopyroxene - sapphirine - sillimanite (Spl, Qtz, Crd)	20c, e						+	Relict Grt in an Opx-Crd symplectite

compositions could be considered to have crystallized near peak conditions. Numerous textures record the breakdown of these porphyroblasts and represent arrested reactions. Although these reactions can usually be treated as melt-absent, it is likely that some phases precipitated directly from melts whereas others are produced during back-reaction with melts. These will be best portrayed in KFMASH system. The most spectacular reactions take place in the mesosomes and involve FMAS phases (orthopyroxene, garnet, sillimanite) back-reacting either with solids or with melts during high-T retrograde events. The phases that occur as coronas, symplectites and pseudomorphs (products) include cordierite, sapphirine, biotite, orthopyroxene, plagioclase and less commonly garnet. The majority of reaction textures described below involve the breakdown of garnet and the growth of cordierite, a situation that indicates a pressure decrease. Tri-, bi- and univariant assemblages (Table 6) and the corresponding textures will be discussed starting from peak conditions. Most of these textures have already been described in great detail by previous authors from Eastern Ghats, Antarctica, Sahara, Scotland and the Alps (for references see Harley, 1998b).

PEAK ASSEMBLAGES

The orthopyroxene-sillimanite-K-feldspar-quartz-(melt) assemblage

Northern part of the MCB

In the Camana, Quilca and Arantas areas, quartzperthite leucosomes from migmatites (samples 15, 25, 28, 39, 239, 241) contain coarse, pleochroic, Al₂O₃-rich orthopyroxene ($X_{Mg} = 0.75-0.79$; Al₂O₃ up to 10.30 wt%) occasionally laden with sillimanite inclusions (Fig. 19b). The absence of garnet-biotite-quartz mutual contacts suggests that the conditions were reached for incongruent melting of biotite through the discontinuous (KFMASH) reaction,

biotite + garnet + K-feldspar + quartz =
orthopyroxene + sillimanite + (melt),
$$(1)$$

a vapour-absent melting reaction with a subvertical slope (Fig. 18a) that takes place at temperatures slightly in excess of 910 °C for pressures of 1.0 ± 0.1 GPa (Carrington & Harley, 1995). The resulting assemblage is orthopyroxene-sillimanite-K-feldspar-quartz- $(\pm \text{ garnet})$ -(+ melt). Additional components like Ti and F, preferentially incorporated in biotite (Dooley & Patiño Douce, 1996), may increase the variance of this reaction and displace the stability field of biotite toward higher temperature. Some garnet-absent assemblages contain potentially residual primary biotite (Bt I: sample 20: $X_{Mg} = 0.80$; TiO₂ = 3.3 wt%; F = 1.4 wt%) suggesting that reaction (1) was arrested after exhaustion of garnet although incomplete melting may have occurred according to a bivariant garnet-absent reaction such as:

> biotite + sillimanite + quartz = orthopyroxene + K-feldspar + (melt), (2)

which takes place a few tens of degrees above reaction (1). Melting might even proceed at higher temperature after exhaustion of quartz.

Also, in garnet-poor or garnet-absent assemblages (sample 15; Fig. 5), the presence of mm-scale oxide grains with sillimanite inclusions suggests that Ti released by biotite melting, together with excess Fe released by garnet resorption, precipitated as ilmenite and magnetite. Experiments on fluid-absent melting of biotite (Patiño Douce & Johnston, 1991) have demonstrated a significant increase in FeO and TiO₂ in the melt fraction with increasing temperature. In the present case, melting of 20% (modal) biotite containing 3% TiO₂ releases 1.2% TiO₂ into the melt, if meltrock proportions are 1:1. As Ti and Fe do not easily dissolve in silica-rich melts, they may have precipitated in the form of Fe-Ti oxides or may have formed pockets of immiscible liquids, both processes being favoured by high fO_2 (Naslund, 1983). Upon cooling,



Fig. 19. (a) Secondary biotite (Bt II) resulting from melt precipitation (sample 2); (b) sillimanite inclusions in primary orthopyroxene (Opx I; sample 239); (c) cordierite moat around garnet (sample 103); (d) orthopyroxene (Opx II)-sillimanite at the contact between garnet and quartz (sample 112); (e) back-scattered image of orthopyroxene (Opx II) and sillimanite symplectite including a resorbed garnet initially in contact with quartz (sample 122); (f) low-alumina orthopyroxene (Opx II) around secondary biotite, itself resulting from melt precipitation (sample 78).



Fig. 20. (a) sapphirine-quartz-garnet assemblage (back-scattered images; sample 93); (b) sapphirine-cordierite symplectite formed at the expense of sillimanite and orthopyroxene (sample 93); (c) spessartine-rich garnet in a cordierite-orthopyroxene (Opx II) symplectite (sample 93); (d) cordierite (\pm sapphirine) reaction rims between an orthopyroxene (Opx I) porphyroblast and matrix sillimanite, and with included sillimanite (sample 95); (e) garnet relict in the core of an orthopyroxene (Opx II)-cordierite symplectite symplectite symplectite of a primary garnet in a sillimanite-rich mesosome (sample 95); (f) cordierite–K-feldspar symplectite, probably representing melt precipitate, surrounding corroded garnet (white line underlining the initial shape of garnet) (sample 93).

after the crystallization of sillimanite, TiO₂ combined with FeO to produce ilmenite and magnetite forming an interprecipitate around sillimanite prisms. The resulting oxide-sillimanite-rich pockets rather than belonging to mesosomes or melanosomes, may thus be part of the leucosome. Otherwise, the absence of garnet from magnetite-rich layers (Fig. 3) could result from the instability of the almandine component under high fO_2 but the equivalent spinel-magnetite-quartz assemblage has not been observed. Whether due to melting reactions or to oxidation there is a tendency for garnet to be consumed in high-grade assemblages of this area and the presence of quartz, plagioclase and sillimanite inclusions in garnet shows that it does not result from the prograde destabilization of orthopyroxene in the presence of sillimanite. Consequently, biotite-garnet-sillimanite-K-feldspar-quartz-(melt) is the likely precursor of many orthopyroxene-sillimanite-K-feldspar-quartz-(\pm garnet)-(melt) assemblages.

Southern part of the MCB

In the Mollendo and Cocachacra areas, quasiexhaustion of biotite through reactions (1) and (2) leads to the ubiquitous bivariant (KFMASH) assemblage orthopyroxene-sillimanite-K-feldspar-quartz-(\pm garnet)-(+ melt). Kyanite, found locally and considered as a relict phase, is inherited from a premelting prograde stage or from initial melting in the biotite-kyanite field whereas sillimanite inclusions in porphyroblastic orthopyroxene are compatible with protracted melting in the sillimanite field culminating with reactions (1) and (2). Porphyroblastic orthopyroxene (Opx I) has Al_2O_3 contents ranging from 7.49 to 10.6 wt%, and X_{Mg} ranging from 0.78 to 0.83, whereas garnet has an X_{Mg} ranging from 0.58 to 0.64. Since $K_{DGrt-Opx}$ has the lowest regional values (2.01-2.54; Table 5), it is likely that the highest temperatures were attained in the Mollendo area. Moreover, the high X_{Mg} in both garnet and orthopyroxene is an indication that peak pressures also were probably attained in these areas (see Harley, 1998b). The upper limit of pressure is provided by the intersection of reaction (1) with the kyanite-sillimanite inversion curve, namely about 1.2 GPa (Fig. 18a; see also Fig. 21a).

In most samples of the Mollendo area, orthopyroxene does not show any sign of instability even where in direct contact with sillimanite. Sample 74 contains small (recrystallized?) orthopyroxene, with low to average Al-content (Al₂O₃ = 6.13–8.14 wt%) and $X_{Mg} = 0.78$ –80. The corresponding K_{DGrt-Opx} (2.03) has the lowest value for the whole area. If this orthopyroxene is secondary, low K_D may represent a value corresponding to a lower than peak temperature. In quartz-rich aluminous granulites from the Cocachacra area (sample 218), sillimanite prisms included in large quartz grains, suggest that associated orthopyroxene (< 0.3 mm) crystallized via melt-absent, quartz-forming, univariant (FMAS; Spl, Spr) reaction,



Fig. 21. (a) *P*–*T* paths for migmatites of the Mollendo–Camana block, interpreted from the KFMASH grid of Carrington & Harley (1995); (b) *P*–*T* paths for melanosomes of the Mollendo–Camana block, interpreted from the FMAS grid of Hensen (1971) modified by Harley (1998b); same legend as for Fig. 18.

cordierite + garnet = orthopyroxene + sillimanite

$$+$$
 quartz, (3)

Crossing this univariant boundary (Fig. 18b) implies a negative ΔV and thus a pressure increase. This boundary is the upper limit of prograde cordierite-bearing assemblages and the resulting orthopyroxene-sillimanite-quartz assemblage is interpreted as a relict of the melt-absent part of the prograde path in this sample.

Finally, a unique occurrence of the assemblage sapphirine-quartz is found in a biotite-free rock, 10 km west of Cocachacra (sample 93). The coexistence of sapphirine and quartz is an index of extreme conditions of metamorphism, unless fO_2 conditions are very high. Whether prograde or retrograde, the restricted occurrence of this assemblage means that either the P-T (and fO_2) conditions for its stability were attained only locally or that the assemblage was not preserved upon retrogression in the rest of the MCB. According to textural evidence (Fig. 20a), the assemblage is garnet-sapphirine-quartz, with evidence for garnet resorption. This trivariant assemblage indicates that a univariant boundary involving sapphirine and quartz has been crossed to enter the Spr-Qtz stability field. In view of the absence of primary cordierite in the area, we favour univariant reaction (Spl, Crd),

orthopyroxene + sillimanite
= garnet + sapphirine + quartz,
$$(4)$$

for this boundary (Fig. 18b) rather than univariant reaction (Spl, Opx) to enter the Spr-Grt field. Bivariant reaction (Spl, Crd, Opx),

$$garnet + sillimanite = sapphirine + quartz,$$
 (5)

related to reaction (Spl, Crd) would account for the textural relationships and the preservation of corroded garnet after exhaustion of sillimanite. It should be noted that this reaction, cannot be precisely determined due to the imprecision of sapphirine thermodynamic data, but, because of its positive ΔV it may be related to a pressure decrease.

The cordierite-garnet-sillimanite-quartz assemblage

Some 15 km inland and 1200 m above sea level, in the Pampa Blanca area, orthopyroxene is absent and none of the observed textures suggest that it was once present. The univariant boundary corresponding to reaction (3) was probably not crossed during the prograde path and the peak assemblage in aluminous gneisses and biotite-rich migmatites of this area is garnet-sillimanite-quartz-(\pm cordierite).

RETROGRADE ASSEMBLAGES

Cordierite growth after garnet

In the mesosomes of quartz-rich, biotite-poor migmatites and aluminous gneisses from the Camana, Quilca and Pampa Blanca areas (Fig. 19c; samples 102, 103), cordierite porphyroblasts (X_{Mg} Crd = 0.88–0.89) surrounding lobate garnet ($X_{Mg} = 0.42$ –0.46) probably originated through the (FMAS) bivariant reaction,

$$garnet + sillimanite + quartz = cordierite.$$
 (6)

Cordierite moats around garnet are not accompanied by K-feldspar nor biotite and thus probably formed through (FMAS) solid-solid reaction (6) rather than from a (KFMASH) garnet + (melt) reaction. They are thus the likely product of pressure decrease. The H_2O content of cordierite is close to 1% (Fig. 15) and, as free water is not expected to be present in partially molten rocks, H_2O of the cordierite must, at some stage, have been in equilibrium with that from the melt. According to experimentally determined partitioning of water between cordierite and melt (Carrington & Harley, 1996), the H_2O content of the melt was probably around 4%, provided equilibrium was attained and the fluid content of cordierite has not been altered by subsequent events. Melting thus occurred under strongly water-undersaturated conditions.

Secondary cordierite also formed in orthopyroxene bearing rocks. In an orthopyroxene-sillimanite migmatite ($X_{Mg-Opx} = 0.75-0.79$; $Al_2O_3 = 6.47-8.45$ wt%) from the Arantas area (sample 140), along the boundary with a perthite-rich leucosome, and adjacent to cordierite-plagioclase-quartz symplectites, cordierite rims ($X_{Mg} = 0.90$) around orthopyroxene appears to result from the bivariant (FMAS) reaction,

orthopyroxene + sillimanite + quartz
= cordierite,
$$(7)$$

which probably represents the destabilization of the assemblage orthopyroxene-sillimanitetrivariant quartz under pressure decrease. In the same sample, garnet ($X_{Mg} = 0.47-0.49$) is rare and occurs in the trivariant (cordierite-free) subassemblage garnet-sillimanite-quartz, which suggests that garnet composition is very close to bulk X_{Mg} for the corresponding layer. The above assemblages are related to univariant (FMAS; Spl, Spr) reaction (3) and are compatible with a pressure decrease at constant temperature (Fig. 18b). We note that Al_2O_3 is high in orthopyroxene grains that have probably preserved their high-T alumina content, as their involvement in cordierite-forming reactions does not involve temperature decrease.

Plagioclase rims around garnet

In cordierite-free rocks (Camana, Quilca and Arentas areas) that contain the assemblage garnet-biotite-sillimanite-quartz, grossularite-poor garnet ($X_{grs} = 0.03$) is rimmed by oligoclase-andesine feldspar (sample 16, 20, 23, 27, 28, 29, 41). As garnet does not show evidence of resorption in these rocks, such rims are interpreted as due to the breakdown of the sole grossularite component. The corresponding reaction:

could be due to either an isothermal decompression or an isobaric heating. The first path is more plausible in view of the decompression-associated breakdown of garnet into cordierite observed in the same areas.

Orthopyroxene recrystallization

In rocks that contain the assemblage sillimaniteorthopyroxene-K-feldspar-quartz-(\pm garnet), primary Al-rich orthopyroxene may have recrystallized into Al-poor orthopyroxene according to the reaction,

Al-orthopyroxene
$$I + garnet + quartz$$

$$=$$
 plagioclase + orthopyroxene II. (9)

This reaction may in part account for the variation of both the Al content of orthopyroxene and modal proportion of garnet in similar assemblages from nearby areas.

Orthopyroxene-sillimanite incompatibility

In biotite-free rocks of the Cocachacra area (samples 93 & 95), the breakdown of stubby orthopyroxene (Opx I) gave rise to cordierite-sapphirine symplectites (Fig. 20b), whereas dactylitic sapphirine (Spr II) preferentially grew at the expense of sillimanite (Fig. 20b,c). These textures suggest the breakdown of peak orthopyroxene reacting with sillimanite to produce sapphirine and cordierite through the bivariant (FMAS) reaction:

which is related to the (Spl, Qtz) univariant (FMAS) reaction (Fig. 18b):

The absence of garnet as a breakdown product of orthopyroxene suggests that the univariant (Spl, Qtz) boundary was not crossed, although newly formed garnet may have reacted with sillimanite and SiO₂-rich melts to form cordierite and sapphirine. This may account for the occurrence of minute sapphirine crystals in large cordierite (Fig. 20d; samples 95 & 140). Even if univariant boundary (11) is intersected, garnet may be absent due to the instability of the almandine component at relatively high fO_2 , which, in turn, accounts for the presence of fine magnetite inclusions in the cordierite.

Garnet breakdown in SiO₂-saturated assemblages

In quartz-rich samples from the Mollendo area, relict garnet (sample 112: $X_{Mg} = 0.60$, Fig. 19d; sample 122: $X_{Mg} = 0.58$; Fig. 19e) is rimmed by a first moat of plagioclase, itself surrounded by Al-poor orthopyroxene (sample 112: $X_{Mg} = 0.82$; $Al_2O_3 = 6.95$ wt%; sample 122: $X_{Mg} = 0.82$, $Al_2O_3 = 6.11-8.28$ wt%). Garnet breakdown can thus be summarized as,

$$garnet + quartz (or SiO_2 from melt)$$

= orthopyroxene + plagioclase, (12)

a reaction that involves a volume increase consistent with a pressure decrease. The grossularite component of reacting garnet provides anorthite to the product plagioclase, whereas the albite component more likely comes from the melt. It is worth noting that in the same samples, the boundaries of quartz inclusions within porphyroblastic garnet do not show any sign of reaction. Whereas the volume increase consecutive to garnet breakdown is readily accommodated if the reaction products crystallized in contact with a melt, the same does not happen in trapped inclusions.

Secondary, relatively Al-poor orthopyroxene (Opx II; Sample 112: $Al_2O_3 = 6.95-8.49 \text{ wt}\%$; $X_{Mg} = 0.80-0.87$) is also formed in checkerboard-like orthopyroxenesillimanite symplectites (Kriegsman & Schumacher, 1999) and preferentially grew at the expense of garnet ($X_{Mg} = 0.60$), in quartz-present domains (Fig. 19d,e). The bivariant (FMAS) reaction,

garnet + quartz = orthopyroxene + sillimanite, (13)

responsible for these textures is related to the univariant (Spl, Spr) reaction (Fig. 18b) and is also favoured by a pressure decrease.

In biotite-free, quartz-saturated rocks from the Cocachacra area, mm-scale garnet from an early garnet-orthopyroxene-sillimanite-quartz assemblage is almost completely replaced by either of the following two types of pseudomorphs. In sample 95, an orthopyroxene (Opx II)-cordierite (± plagioclase-magnetite) symplectite, itself surrounded by a partial rim of orthopyroxene (Opx II) (\pm plagioclase-magnetite) is developed only where initial garnet was in contact with quartz or with the leucosome (Fig. 20e). In the core of the symplectites, spessartine-rich garnet relicts (Sp_{10}) are present whereas the Mn content of orthopyroxene II is about 1.17 wt%. In the groundmass, sillimanite is separated from primary, Mn-poor (0.4 wt%) orthopyroxene (Opx I) by a fine rim of cordierite with magnetite droplets and vermicular sapphirine (see reaction (10) above). These complex textures can be explained by the following succession of bivariant (FMAS) reactions,

orthopyroxene + sillimanite + quartz

garnet + quartz = orthopyroxene + cordierite, (15)

which are related to the univariant (FMAS; Spl, Spr) reaction (3).

Crossing this boundary implies a major volume increase and is thus favoured by a pressure decrease (Fig. 18b). The sapphirine vermicules found in the cordierite may be explained if, in an orthopyroxenesillimanite aggregate from which quartz has been exhausted due to reaction (14), further reaction proceeds under decreasing pressure according to reaction (10), which implies crossing the univariant boundary (11).

In sample 93, a cordierite-sapphirine-plagioclase- $(\pm \text{ magnetite } \pm \text{ orthopyroxene})$ symplectite is observed. In this case, the pseudomorphs of primary garnet

consist of cordierite with sapphirine vermicules (\pm plagioclase \pm magnetite). Spessartine-rich garnet relicts (Sp₂₀; Fig. 20c) occur in the centre of the symplectites, whereas an Mn-rich orthopyroxene (Opx II; MnO = 2.70%) forms a partial rim around the symplectite. In this case, garnet breakdown probably took place under silica-undersaturated conditions according to the bivariant (FMAS) reactions,

garnet + sillimanite = sapphirine + cordierite, (16)

$$+$$
 sillimanite, (17)

both being related to univariant reaction (Spl, Qtz). Crossing this univariant boundary in the sense of sapphirine is also compatible with a pressure decrease (Fig. 18b; see also Fig. 21b).

It should be noted that although these reactions are valid for silica-undersaturated systems, quartz is present in the sample, but it is likely that at the onset of decompression much if not all of it was incorporated in the melt within which silica activity was less than unity. This illustrates the need for considering mm-scale subassemblages, since reactions that should only occur in silica-undersaturated solid rocks can be observed in melt-present silica-saturated rocks. Finally, the presence of submicroscopic spinel grains within newly formed cordierite (samples 95, 93) suggests crossing of the (Opx, Qtz) boundary under still declining pressure. As spinel does not occur together with quartz, the univariant boundary (Opx, Spr) was not intersected which implies an inflection of the decompression path (Fig. 18b; see also Fig. 21b).

BACK-REACTIONS INVOLVING MELT AND MELT PRECIPITATION

Several textures are strongly indicative of solid-melt reactions or melt precipitation. In the Camana area, the modally limited but ubiquitous subassemblage biotite-quartz(\pm plagioclase) suggests coprecipitation from silicate melts, and as resorbed garnet is often associated with this texture (Fig. 19a; Sample 2), it may be involved in (KFMASH) bivariant reaction,

Alternatively, garnet may be a product of melt precipitation. For instance, the orthopyroxene-poor, sillimanite-free quartzofeldspathic sample 2 contains two generations of garnet (Fig. 17): a pyrope-rich garnet $(X_{Mg} = 0.53)$ is found isolated in the matrix as is the orthopyroxene (Al₂O₃ = 4.3–6.2 wt%; $X_{Mg} = 0.65$ – 0.68), whereas a pyrope-poorer garnet ($X_{Mg} = 0.39$) is associated with quartz-biotite-plagioclase symplectites in microdomains that may represent melt segregations. The pyrope-poor garnet and the biotite could be considered as products of melt precipitation according to reaction (1) operating backwards until exhaustion of sillimanite (isobaric cooling). In the case of sample 2 however, considering the significant difference in the pyrope content of garnet, it is more plausible to invoke a lower T- lower P discontinuous (KFMASH) reaction such as,

cordierite + orthopyroxene + (melt) = biotite II

+ garnet II + K-feldspar + quartz, (19)

that proceeded until exhaustion of cordierite (Figs. 18a & 21a). This evolution presumes that either cordierite was present in peak assemblages in the case where pressure was not higher than about 1.0 GPa (see Harley, 1998b) or more likely that primary garnet was previously converted into cordierite during the first stages of decompression.

In a quartz-saturated sillimanite-orthopyroxene migmatite of the Arantas area, interstitial, mm-scale, myrmekite-looking symplectites of cordierite-plagioclase-quartz can be confidently interpreted as due to melt precipitation. As neither K-feldspar nor biotite are involved in the assemblage, the reaction probably took place between K-poor melt, sillimanite and orthopyroxene.

In the Cocachacra area (sample 93a), mm-scale myrmekite-looking, vermicular intergrowths of cordierite and K-feldspar surrounding corroded garnet in quartz-bearing rocks (Fig. 20f) that could be misinterpreted as retrograded osumilite, result from the continuous solid-melt reaction,

$$garnet + sillimanite + (melt)$$
$$= cordierite + K-feldspar.$$
(20)

This reaction is sensitive to the X_{Mg} of garnet and, due the relatively flat Mg/Mg + Fe isopleths, the instability of garnet in the presence of melt will be favoured by a pressure decrease (Spear *et al.*, 1999).

SECONDARY MELTING

Renewed melting related to a second stage of decompression is found in several samples from the Mollendo area (e.g. sample 78), where large biotite (Bt II; $X_{Mg} = 0.85$; TiO₂ = 2.92 wt%; F = 1.2 wt%) may be overgrown by a relatively Al-poor orthopyroxene (Al₂O₃ = 6.2 wt%) forming a partial corona between biotite and either quartz or the leucosome (Fig. 19f). This texture could be explained by a fluid-absent (KFMASH) multivariant melting reaction related to the (Os, Sil) univariant reaction (19). This late melting event took place after major decompression and cooling from ultra-high temperatures.

PHYSICAL CONDITIONS OF METAMORPHISM

The identification of mineral assemblages and textural relationships together with the compositions of Fe-Mg

phases allow a semiquantitative evaluation of the physical conditions and their evolution during the metamorphism and migmatization of the MCB. Pending the determination of reliable P-T estimates, peak and postpeak P-T evolution can be monitored using reaction textures and appropriate P-T grids (Fig. 21).

Water content of assemblages

The proportion of hydrous minerals (essentially biotite and cordierite) within rocks of the MCB varies from 20 to almost 0%. Several samples from the Mollendo and Cocachacra areas are devoid of hydrous minerals whereas migmatites from Quilca may contain up to 20% biotite. It is likely that among the biotite, the F-rich ones may have survived melting while F-poor ones precipitated from melts. The distribution of water within the assemblages was thus variable throughout the area, but it is likely that, at least in the south-eastern part of the MCB, water was only present as a component of the melt at the peak of metamorphism. Cordierite is able to trap water from the melt with a melt-solid partition coefficient of about 4, as experimentally determined by Carrington & Harley (1996). In Pampa Blanca and Cocachacra areas, the H₂O content of cordierite is about 1%, which means that cordierite may have been in equilibrium with a melt containing about 4% H₂O.

Oxygen fugacity

Oxygen fugacity is difficult to estimate in rocks from the MCB because of extensive exsolution-re-equilibration of Fe-Ti oxides during cooling. Textural evidence, however, attests to the presence, during peak metamorphism, of an ilmenite_{ss}-magnetite_{ss} buffer in most assemblages. In the Pampa Blanca area, magnetite ($X_{mag} = 0.69, X_{uSpl} = 0.30$) coexists with ilmenite ($X_{ilm} = 0.80, X_{hem} = 0.20$) which corresponds to a log $fO_2 = -11$ at 950 °C (Ni-NiO buffer; Spencer & Lindsley, 1981). However, in many samples, pure magnetite $(X_{mag} = 0.99)$ is found coexisting with Fe-Mg spinel, in keeping with granular exsolution and re-equilibration at lower temperatures. The Fe^{3+} content of several silicates as obtained from chargebalance calculation also attests to relatively high fO_2 . Such is the case for sapphirine and orthopyroxene that have Fe³⁺/Fe^{tot} ratios of about 0.4 and 0.2, respectively (see Table 4). Sillimanite has about 0.1 Fe^{3+} (on a 20 O basis) substituted for Al. Finally, the absence of rutile and graphite is in keeping with a relatively high fO_2

Partial Melting

In Camana, Quilca, Arantas, Mollendo and Cocachacra areas, the assemblage orthopyroxene-sillimanite-K-feldspar-quartz- $(\pm \text{ Grt})$ shows that peak temperature conditions were above those of the (Os, Crd) major boundary (Fig. 21a), and probably well above 900 °C if maximum F and Ti enter biotite (Patiño Douce & Jonhston, 1991). Minimum pressures are provided by the absence of osumilite and primary cordierite, and those of the (Bt) invariant point where these minerals would coexist with melt. Local evidence of vapour-absent melting of biotite- $(\pm \text{ garnet})$ according to a quartz-free melting reaction such as,

biotite + sillimanite +
$$(\pm \text{ garnet}) = \text{orthopyroxene}$$

+ $(\pm \text{ sillimanite}) + \text{K-feldspar} + (\text{melt}),$ (21)

suggests that either biotite was previously exhausted from quartz-rich assemblages or that quartz was exhausted via reaction (1) in quartz-poor assemblages. In fact, during prograde melting, temperature was probably buffered, in the sillimanite field, at the (Os, Crd) boundary until exhaustion of biotite in guartzrich rocks, and subsequent temperature increase permitted continuous biotite melting in quartz-free rocks. This situation may have prevailed in Mollendo and Cocachacra areas where melt production has dramatically decreased the modal proportion of garnet and virtually removed biotite from the assemblages. Considering the very low water content of generated melts (from 0 to 4%), any pressure decrease at high temperature would prevent crystallization, thus favouring a continuous process of melt segregation. On the other hand, the high viscosity of quasi-anhydrous melts is probably responsible for the limited amount of melt migration and for the preservation of the stromatic habit of migmatites. In that respect it is worth noting that evidence of melt migration in the form of aplitic dykes and granitic stocks (still undated but probably related to the UHT event) are more abundant in the northern part of the MCB where water content of the melts was higher and viscosity lower.

Peak temperature and pressure

Besides extensive melting in the orthopyroxenesillimanite field, evidence for ultra-high temperature is provided by the high Al content of orthopyroxene. From Camana to Mollendo, Al₂O₃ values approaching 10% (corresponding to X_{A1} of c. 0.10 on a 10 O basis) are not exceptional and are compatible with temperatures of about 1000 °C (e.g. Harley, 1998b). Also, a unique occurrence of sapphirine in contact with quartz in the Cocachacra area may indicate temperatures above 1000 °C (Fig. 21b). The absence of primary (prograde) magnesian cordierite suggests pressures in excess of 1.0 GPa (e.g. Harley, 1998b). Moreover, retrograde cordierite in Pampa Blanca area is close to pure Mg end-member, and probably grew under already declining pressures around 1.0 GPa. Finally, the presence of kyanite porphyroblasts in the Mollendo area suggests an early stage of

higher pressure or lower temperature metamorphism. This early kyanite must be inherited from a pre- or an early melting stage and it is unlikely that melting occurred in the kyanite-orthopyroxene field, because sillimanite is the only Al-silicate so far found as inclusions in the orthopyroxene. The vapour-absent discontinuous melting curve of Mg-rich biotite (Carrington & Harley, 1995) intersects the kyanitesillimanite inversion curve at about 920 °C and 1.22 GPa and at slightly higher conditions if biotite is stabilized by significant incorporation of Ti and F. If melting occurred between 950 and 1050 °C in the sillimanite field, pressure thus cannot have exceeded 1.3 Gpa (Fig. 21a). Based on the above elements, a reasonable estimate for the ubiquitous orthopyroxene-sillimanite-quartz-(\pm melt) assemblage in the MCB should therefore be in the range 900-1000 °C and 1.0-1.3 GPa. Lower conditions for the peak of metamorphism may apply to the Pampa Blanca area where no trace of the orthopyroxene-sillimanite assemblage was found (Fig. 21b). A regional pressure gradient from Camana to Mollendo is independently suggested by the increasing X_{Mg} of garnet (from about 0.30 to over 0.60) and orthopyroxene (from 0.61 to 0.84) in rocks of otherwise similar bulk X_{Mg} . Finally, it should be remembered that this UHT metamorphism has a regional character as it occurs in an area of several thousands km², and not as boudins or localized outcrops as is the case for several UHT area so far described.

The retrograde path

Within the P-T range defined above for peak conditions, the low fO₂ FMAS grid (Hensen, 1971; Harley, 1998b) is dominated by the assemblages orthopyroxene-sillimanite-quartz and garnet-cordierite-sillimanite-quartz, whereas the spinel-quartz assemblage is restricted to lower pressures and higher temperatures, namely away from the stable invariant point (Spl). For the same P-T range, the high fO_2 grid requires the presence of stable invariant points (Spr) and (Grt) and hence the stability of the assemblage spinel-quartz. Among the observations suggesting that the low fO_2 grid is more appropriate for UHT rocks of the MCB we note that: (a) garnet was stable during peak conditions even if it has been largely consumed by redox and melting reactions; (b) neither spinel-quartz assemblages, nor garnet and/or sillimanite coronas around spinel have been observed in any of the above areas, suggesting that spinel never coexisted with quartz; (c) retrograde textures illustrated by the formation of orthopyroxene-sillimanite and orthopyroxene-cordierite symplectites can be accounted for by bivariant (Spl. Spr, Crd) and (Spl, Spr, Qtz) reactions attesting to the stability of the (Spl) point of the low fO_2 grid; they cannot be accommodated with any of the stable invariant points of the high fO2 grid. Thus, it seems that although fO_2 was relatively high during UHT metamorphism of the MCB, it was not high enough to support an inversion in the topology of Hensen's original FMAS grid.

A preliminary interpretation of the retrograde textures could stress the breakdown of garnet and the formation of cordierite, whether or not associated with sapphirine or orthopyroxene. There is unequivocal evidence for a major initial stage of decompression at high-T in Mollendo and Cocachacra areas. In the orthopyroxene-sillimanite-quartz domain, orthopyroxene-sillimanite symplectites grew after garnetquartz according to reaction (13). Then the entry into the cordierite field is marked by orthopyroxenecordierite symplectites after garnet-quartz according to reaction (15). Evidence for further decompression in low-SiO₂ assemblages is provided by garnet breakdown into fine aggregates of orthopyroxenesapphirine-sillimanite according to reaction (17) and sillimanite-orthopyroxene giving cordierite-sapphirine according to reaction (11). Finally, the formation of very fine spinel-cordierite assemblages suggests that the decompression path went on as to intersect the (Opx, Qtz) reaction (Fig. 21), before switching from decompression to cooling. Following the initial stage of decompression, isobaric cooling took place. This stage probably lead to P-T conditions of the (Os) invariant point as suggested by the tendency of X_{Mg} of garnet and orthopyroxene and X_{A1} of orthopyroxene of the MCB to converge toward the values obtained by Harley (1998b) for calculated isopleths at this invariant point.

CONCLUSIONS

orthopyroxene-sillimanite-quartz-(\pm garnet-K-The feldspar) assemblage, usually considered as diagnostic of UHT metamorphism (e.g. Harley, 1998b), is ubiquitous in the Camana, Quilca, Arantas, Mollendo and Cocachacra areas although by no means this implies uniform temperature throughout the area. The Pampa Blanca area, which is devoid of this assemblage, may represent lower metamorphic conditions, and as this area is the most inland, it is likely that an 'orthopyroxene-sillimanite isograd' runs more or less parallel to the Pacific coast. The highest temperatures attained in the Mollendo and Cocachacra areas may have been locally in excess of 1000 °C at pressures around 1.2 GPa. Mineral assemblages and textural relationships suggest a three-stage retrograde P-Tpath, starting with an initial decompression at high-Tfollowed by an isobaric cooling stage and a second decompression. The first stage is best recorded in the Cocachacra area by solid-solid FMAS reactions and is characterized by a quasi-isothermal decompression of about 0.3 GPa. The isobaric cooling stage does not produce any typical reaction texture and is rather interpreted from the absence of osumilite, the absence of spinel-quartz assemblages and the tendency for garnet and orthopyroxene to stabilize their compositions close to those theoretically in equilibrium in the vicinity of the (Os) invariant point. The melting of secondary biotite with formation of a late low-Al orthopyroxene provides evidence for a second decompression stage, so far observed in the Mollendo area only, a reaction favoured by isothermal decompression. Retrograde *P*-*T* paths, well established for Mollendo and Cocachacra areas, are not well established for the north-western part of the MCB.

A remarkable feature of these UHT rocks is the relative scarcity of readjustment to lower temperature that may be due to the combination of rock dryness and rapidity of cooling. In fact, it is in the high-alumina low-K mesosomes, that retrograde reactions are best developed. The leucosomes, being partly molten during high-T conditions, porphyroblastic minerals continuously readjusted their compositions to varying P and T and registered the conditions of crystallizing melts only. In the mesosomes, back-reactions are rather limited to the initial decompression stage.

Among the dozen or so P-T paths that have been determined for UHT rocks, those of Forefinger Point (Harley et al., 1990) and Palni Hill (Raith et al., 1997) are the most akin to the MCB path. However, isotopic data are still too sketchy to demonstrate whether the various sections of the path belong to the same Grenvillian tectono-thermal event or whether they represent a polycyclic evolution, the UHT one being Palaeoproterozoic, the second, lower-T one, being Grenvillian. In the case of a monocyclic UHT evolution, the fact that all the crustal levels appear to end up near the (Os) after the first stage of decompression and cooling suggests that the deepest parts of the section underwent greater and probably faster exhumation that the less buried ones. The exhumation being essentially isothermal, it can hardly be driven by erosional processes and is more likely controlled by tectonic processes. Two scenarios are plausible: diapiric-like rising of the hottest parts (Mollendo area) of the migmatite complex, and unloading via northdipping extensional faults or detachment. These mechanisms are not exclusive, as an ascending hot zone, reminiscent of core complex, may trigger detachment. A second, less well-defined, exhumation stage is suggested by renewed anhydrous melting of secondary biotite in the Mollendo area.

Two types of heat perturbations have been advocated to explain the high temperatures attained during UHT metamorphism: advection of heat due to emplacement of mafic or super-heated silicic magmas and the upwelling of asthenospheric mantle at the base of continental crust. The MCB is almost devoid of mafic rocks, while UHT rocks outcrop along the Pacific coast for more than 100 km, implying a thermal influx that can hardly be attributed to local emplacement of magmas. An upwelling of asthenospheric mantle into an already overthickened crust is thus the more likely source of heat.

ACKNOWLEDGEMENTS

This work was initiated during a sabbatical stay of JM in Montpellier in 1995 and we thank A. Leyreloup for introducing us to the orthopyroxene-sillimanite rocks of the Mollendo area. The local assistance of the French IRD staff in Lima was instrumental in the planning of fieldwork and special thanks are addressed to G. Carlier, R. Morroco and T. Sempere for their collaboration. G. Poirier is thanked for his assistance with the McGill electron microprobe and P. Pouget for devotion to the maintenance of the THEBA software. Raman spectroscopic determinations on Al-silicate were conducted at the Département de Chimie de l'Université de Montréal with the aid of S. Elouatik. Fieldwork in Peru and laboratory expenses was possible thanks to NSERC grants to JM. The manuscript has benefited from constructive comments by S. Baba, S. Harley and A. Indares.

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Received 10 December 2001; revision accepted 28 August 2002.