

## Geothermobarometry and fluid inclusions in a rock from the Doddabetta charnockite complex, Southwest India

JACQUES L.R. TOURET

Institute of Earth Sciences, Free University, de Boelelaan 1085, 1081 HV Amsterdam, The Netherlands

THOR H. HANSTEEN

Mineralogisk-Geologisk Museum, Sars gate 1, N-0562 Oslo 5, Norway

**ABSTRACT.** — A deformed, garnet and orthopyroxene-bearing charnockite from Doddabetta near Ooty, Nilgiri Hills India, contains many CO<sub>2</sub> fluid inclusions in 3 major minerals (garnet, plagioclase and quartz). Detailed study of the mineral phases places PT peak metamorphic conditions at  $8 \pm 0.7$  kb and  $800 \pm 15^\circ\text{C}$ , respectively. The rock is conspicuously deformed and recrystallized (quartz more than plagioclase), whereas garnet has resisted the deformation. The densities of the CO<sub>2</sub> inclusions in each mineral are relatively uniform, and they increase in order of increasing deformation from 0.96 - 1.02 g/cm<sup>3</sup> in garnet to 1.12 g/cm<sup>3</sup> in quartz. Primary inclusions in garnet correspond to a pressure about 1 kbar lower than peak metamorphic conditions. Higher densities in plagioclase and quartz indicate a post metamorphic P-T-trajectory concave towards T axis («isobaric cooling»). Geodynamic implications and fluid regime after metamorphism (important quantity of free CO<sub>2</sub> during mineral growth) are discussed.

### Introduction

The Archean craton of Southern India consist of a southern granulite facies and a northern greenschist-amphibolite facies terrane, connected by a narrow transitional zone from Mangalore in the west, through Mysore and Bangalore to Madras in the east (DRURY and HOLT 1980) (Fig. 1). The granulite facies metamorphism is essentially a postkinematic transformation of a 3400 to 2900 m.y. migmatite-gneiss terrain (BECKINSDALE et al., 1980). CRAWFORD (1969) found a metamorphic age of 2600 m.y. Incipient proterozoic retrogression of the granulite facies rocks occurs mainly along several extensive E-W and NE-SW trending

ductile mega - shear zones (DRURY and HOLT, 1980).

The Doddabetta (or Doddabetta) Charnockite Complex is located in the northwestern part of the Indian «Charnockite Uplands» (Fig. 1). (In the subsequent text, «Doddabetta charnockite» refers to the rock used in this study). Both geographically and structurally, it is marginally connected to the regional shear zones. The Doddabetta charnockite is an example of a rocks well suited for diverse geochemical investigations. It contains a relict mineral assemblage from peak granulite conditions, including several potential geothermometers and-barometers. The rock contains as much as three major minerals with a surprising quantity of fluid inclusions, much more abundant than in most similar rocks in other terranes.

The present study tries to establish the pressure and temperature conditions of high grade metamorphism in the Doddabetta charnockite by a combination of mineralogical and fluid inclusion methods. The main tools are petrographic studies, electron microprobe analyses and microthermometry. Geothermobarometrical data are compared to independent results obtained from a different method (fluid inclusions).

From this, semiquantitative P-T paths are inferred (see e.g. CRAWFORD and HOLLISTER, 1986; TOURET, 1985; NEWTON, 1985, etc.).

This publication is an illustration of the

value of combining independent geochemical methods at the scale of the hand specimen, in order to estimate the conditions of formation for high grade rocks. However, we wish to emphasize that we by no means claim that this study, based on only one sample, is representative for the area in general. All conclusions have to be confirmed by investigations on other specimens.

in the collection of the Free University) classifies as a tonalite in the nomenclature of STRECKEISEN (1976) see (Table 1). A proper rock name would be «enderbitic-granolite» (WINKLER 1974), indicating the strong dominance of plagioclase over alkali feldspar, which occurs only as antiperthitic lamellae in the plagioclase grains. However, the «granolite» nomenclature has never been

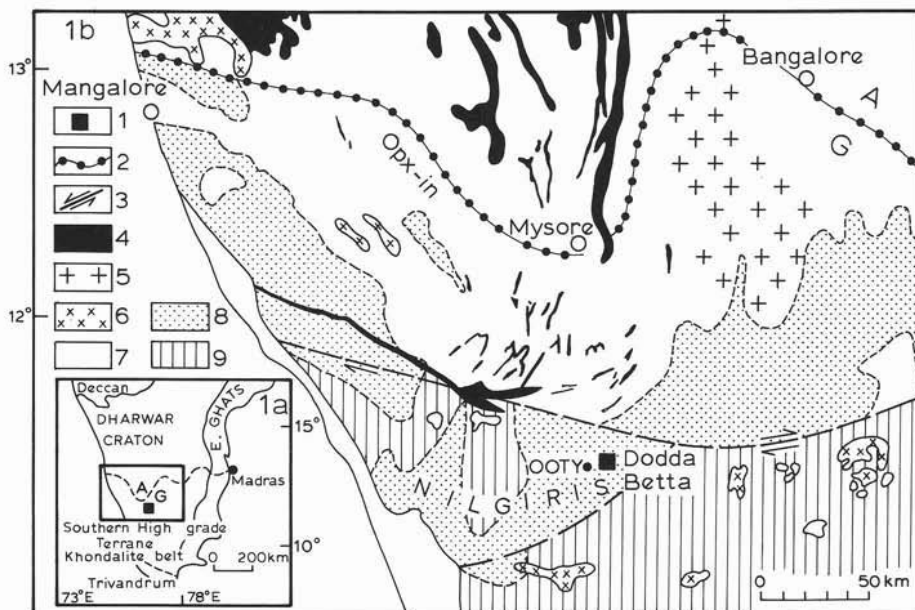


Fig. 1. — Geological setting of Doddabetta

1a. (Insert) Sketch of South India, (after GOPALAKRISHNA et al., 1986). The metamorphic grade increases on a regional scale from the Dharwar craton in the north to the high grade terrane in the south. (Northern part of the «Southern High grade terrane = Charnockite Upland»).

1b. Doddabetta is situated in the Nilgiris hills, near Ooty in the northern part of the Indian «Charnockite Uplands». The high grade terrane is cut by regional, mainly E-W trending ductile mega shear zones, inducing an incipient retrogression of the granulite facies rock (DRURY and HOLT, 1980).

1: Location of the investigated sample (6074, VU collection); 2: Amphibolite/granulite transition (orthopyroxene isograd); 3: Ductile mega shear zones; 4: Supracrustal rocks of the DARWAR and SARGUR group; 5: Proterozoic granites; 6: Closepet plutonic belt; 7: Migmatite gneiss complex; 8: Charnockite/khondalite complex; 9: Charnockite/khondalite complex affected by retrogression. A = Amphibolite, G = Granulite facies.

Earlier results from the charnockites near Doddabetta are given by JANARDHAN et al. (1982), HARRIS et al. (1982) and NEWTON (1985). For a more thorough description of the Indian granulites, the reader is referred e.g. to HOWIE (1955), WEAVER (1980), RAITH et al. (1983), and SANTOSH (1986).

### Petrography

The investigated sample (sample no. 6047

widely accepted and the name «Doddabetta charnockite», taken in a very broad sense, will be sufficient for the purpose of the present paper. Qualitatively, its mineral composition resembles other Doddabetta charnockites, e.g. samples 11.1 and 11.1B, Table 1 in JANARDHAN et al. (1982). However, from comparison with sample 11-1 (obtained by courtesy from R.C. Newton), it is obvious that sample 6047 is more deformed and contains

TABLE 1

*Modal analyses and whole rock chemistry of sample 6047 and other Doddabetta charnockites*  
(JANARDHAN *et al.* 1982)

Modal analyses (i)		Whole rock chemistry (ii)			
(Sample 6047)			(1)	(2)	(3)
Plagioclase (Antiperthite)	45	SiO <sub>2</sub>	67.1	69.9	66.8
Quartz	25	Al <sub>2</sub> O <sub>3</sub>	14.1	13.5	14.3
Garnet	12	TiO <sub>2</sub>	0.50	0.41	0.52
Orthopyroxene	10	FeO (XRF)	4.36	3.43	5.46
Mica (Phlogopite)	5	FeO (wet.an.)	3.6	2.5	4.5
Opauques (Ilmenite)	2	MgO	2.31	1.97	2.58
Calcite	< 1	CaO	2.82	1.98	3.08
Sphene, rutile, zircon	<< 1	Na <sub>2</sub> O	3.91	2.81	3.61
		K <sub>2</sub> O	2.23	3.91	1.31

(1), (2), (3): Doddabetta charnockites, in Janardhan *et al.*, (1982), Contr. Min. Petr., 79, 130-149.

(i) Surface percentage, average of estimates in 5 thin sections

(ii) weight % oxides

## Mineral and fluid inclusion description

### Quartz

Most quartz crystals show strong undulatory extinction. Measured differences in extinction angles between the edges of a single, flattened grain are up to 27 degrees.

Fluid inclusions (in the subsequent text, «fluid inclusions» and «inclusions» are used as synonymous expressions, unless otherwise stated) averaging 15 micrometer in diameter, occur in widely variable amounts along healed microfractures (7, Fig. 2). These paleofractures, outlined by the inclusions, most often attain curved shapes. The inclusion trails show no obvious geometrical relations to other textural or deformational features. Inclusion boundaries range from completely irregular, through subhedral forms (which are most common) to perfect negative crystal shapes (8, Fig. 2). Inclusions occur mainly in the quartz grains which have escaped recrystallization.

### Plagioclase

Plagioclase is somewhat deformed, and is found both as original grains, and as part of the mortar fabric. It shows conspicuous undulatory extinction, but less so than quartz. The plagioclase is unzoned, but features very coarse exsolution textures (antiperthites). Deformation textures commonly include bent albite lamellae.

Occasionally myrmekite intergrowths between quartz and feldspar occur, with quartz appearing as «droplets» inside the feldspar grains. Fluid inclusions in the plagioclase are fracture-bound (6; Fig. 2). Inclusions-filled healed fractures are found both along, and at different angles to, the best cleavage plane (001). Some trails are irregularly curved, due to strain induced transposition. Some trails end abruptly at the grain boundaries, while others extend through the smaller, recrystallized grains. Inclusion outlines show great variations, but often approach negative crystal shapes, which in plagioclase is a distorted, irregular prism (6, Fig. 2). Some of the inclusions contain one

more abundant garnet. As a whole, the upland Dodda Betta charnockite complex is somewhat heterogeneous (metabasites indicated in JANARDHAN *et al.* (1982)), and it contains several charnockite varieties, many of them granitic (some with over 3 wt% K<sub>2</sub>O), and undepleted in Rb and U (JANARDHAN *et al.*, 1982). The most prominent feature in hand specimen (Sample 6047) is a matrix consisting mainly of tightly interlocking quartz and feldspar grains of 1 to 7 mm diameter, containing evenly distributed garnets of 0.5 to 5 mm diameter. On some of the sawn surfaces, a foliation of partly oriented phlogopite and flattened, recrystallized quartz is observed (1, Fig. 2). A visual estimate of the modal mineral composition is given in Table 1. The rock is dense. The colour on a fresh surface is dull, medium grayish green speckled with pale red garnet grains.

Microscopically, the dominant feature is a well developed mortar fabric, indicating extensive recrystallisation of quartz and feldspar. The result is that the original grains are surrounded by a combination of almost discrete cluster and elongate aggregates of newly formed grains, with polygonal to lobate grain boundaries (1, Fig. 2).

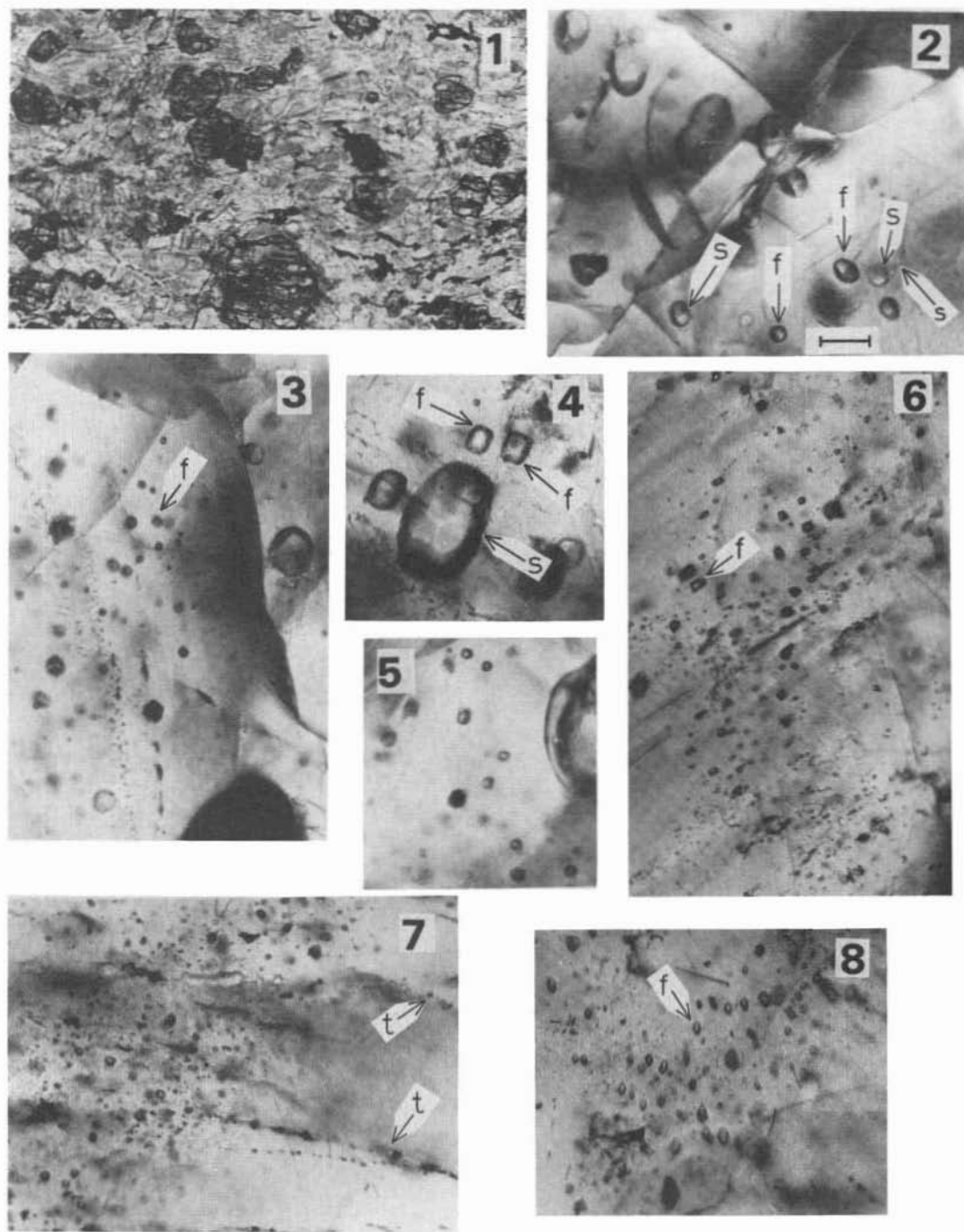


Fig. 2. — Fluid and solid inclusions in the Doddabetta Charnockite. 1: Photograph of a polished slab of the investigated sample. Numerous globular garnets (high relief) occur in a strongly deformed matrix of quartz, antiperthitic plagioclase and orthopyroxene. Note the postkinematic horizontal foliation turning around the garnets. 2 to 5: Fluid and solid inclusions in garnet - (S = Solid inclusion, mainly quartz), f = high density CO<sub>2</sub> inclusion). In 3 and 5, only fluid inclusions. All fluid inclusions are randomly dispersed in the core of the garnet (primary

or two daughter minerals. From the high birefringence, one of the phases can be tentatively identified as a carbonate, probably similar to the ones described in garnet (Fig. 3).

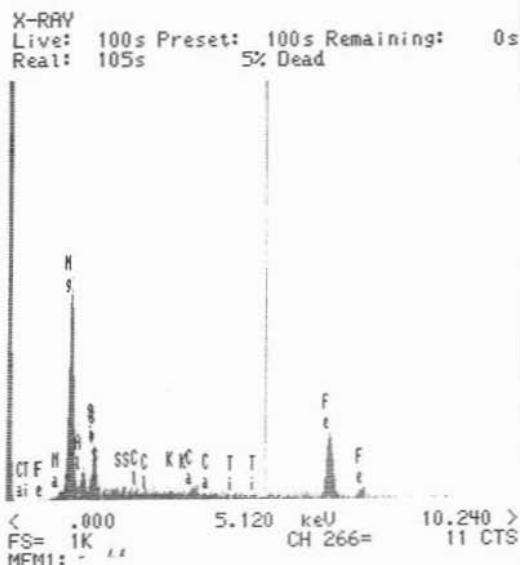
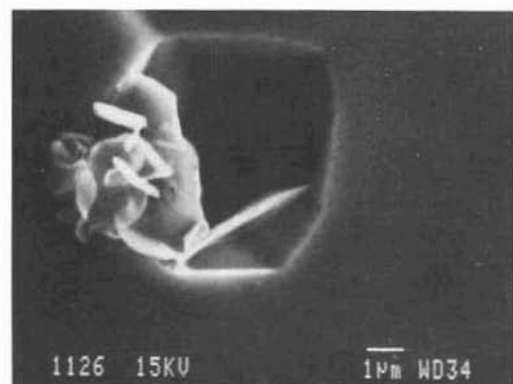


Fig. 3. — Electron microscope imagery of an opened fluid inclusion in garnet.

Left: SEM image (Jeol JSM 840 scanning electron microscope, Institutt for Geologi, Oslo). In the open cavity, which has a clearly visible negative crystal shape, an aggregate of platy (white) daughter minerals remains. Some of them (the smallest) have partly left the cavity; they could correspond to an original bigger crystal which has been broken during the fracturing of the garnet.

Right: EDS-Spectrum of the daughter minerals (lines for Si and Al come from the host mineral garnet).

### Garnet

Garnet occurs largely as subhedral crystals in the quartz-feldspar matrix. Some of the grains are rounded due to deformation, but most appear to be unaffected by strain. The mineral shows a poikilitic texture, with many solid inclusions showing a perfect negative garnet crystal shape (Fig. 2).

Fluid inclusions occur mainly in discrete clusters, well removed from the grain boundaries. The inclusions have regular shapes; some show development of negative garnet crystal faces. Numerous solid inclusions occur in close vicinity to and with the same shapes as the fluid ones. Among them, quartz is by far the dominating mineral, followed by apatite and then carbonates (iron bearing

dolomite). Spatial relations between the different minerals suggest that all phases were present along with the fluid at the time of entrapment (2 and 4, Fig. 2). Daughter minerals in some opened inclusions (Fe, Mg carbonates) are described below (Fig. 3).

### Orthopyroxene

Orthopyroxene occurs as sub- to anhedral grains, often in contact with ilmenite, garnet and/or phlogopite. Some grains are slightly deformed. Reaction rims can be seen in some of the contact areas between garnet and phlogopite grains, indicating the onset of very local retrogressive reactions. However, most opx and gnt grains constitute parts of the

inclusions) and, in three dimensions (much more abundant than the impression given by the photograph). Only a few inclusions strictly in the focusing plane are visible (but changing the focus makes now inclusions appear in the whole mass of the garnet). 6: Fluid inclusions (f) in plagioclase (negative crystal shape; approximately cubic shape). The complex pattern illustrated in the figure corresponds to an early subvertical trail (N-S) later transposed along secondary small trails parallel to the main cleavage (NE-SW). 7 and 8: Trail bound inclusions in quartz (= trace of late, nearly open trails. The sudden disappearance of nearly all inclusions to the right in Fig. 7 corresponds to a post kinematic recrystallization (annealing) of the quartz grain. Length of the bar (photo 2): Photo 1 = 8 mm, all others: 10µm.

stable, primary paragenesis. No fluid inclusions have been found in this mineral, but it contains numerous primary solid inclusions of carbonate.

### Mica

Dark mica (biotite/phlogopite, see below) occurs both as primary and secondary grains, mostly in clusters with other ferromagnesian minerals. The distinction between «primary» versus «secondary» is basically made on the shape (primary = subrectangular sections with well-defined boundaries, homogeneous coloration and strong pleochroism) and the relation with neighbouring minerals (secondary = transional gradation and partial

pseudomorphism of other minerals, e.g. garnet or orthopyroxene). Some of the micas are strongly deformed, and appear to be bent around neighbouring grains.

Some primary phlogopite is found in close spatial connection with garnet. In some circumstances, the mica is completely enclosed in the garnet, without any evidence of reaction rim.

### Fluid inclusion chronology

Crosscutting relations show that the annealing process, which leads to the formation of the mortar fabric, partially postdates the formation of fluid inclusions in quartz and feldspar. In some cases, the

TABLE 2

*Selected mineral analyses (microprobe; Cambridge Microscan-9, Free University, Amsterdam or Cameca Camebax MICROBEAN, Geological Museum, Oslo)*

*(I, II, III = different types of phlogopite as described in Fig. 4 and in the text)*

2a) GARNET :					2b) ORTHOPYROXENE :			2c) FELDSPAR :						
	A2 core	A2 rim	B2 core	C2 rim	A2	B2	C2		A2 plag	A2 kfsp	B2 plag	B2 kfsp	C2 plag	C2 kfsp
SiO2	38.20	38.20	38.70	38.40	51.50	51.70	51.20	SiO2	62.03	63.17	61.50	63.80	61.40	63.90
Al2O3	21.60	21.70	21.90	21.90	1.33	1.86	1.83	Al2O3	23.89	18.34	23.70	18.50	23.80	18.30
FeO	29.10	30.20	28.80	29.20	27.90	25.90	26.40	CaO	5.34	0.02	5.51	0.06	5.69	0.06
MnO	0.81	0.84	0.82	0.79	0.27	0.21	0.21	Na2O	8.93	1.00	8.63	0.99	8.51	0.92
MgO	8.10	7.27	7.96	7.77	18.90	20.10	19.50	K2O	0.27	14.84	0.27	15.60	0.25	15.60
CaO	2.30	2.20	2.60	2.46	0.26	0.29	0.26	Sum :	100.46	97.37	99.61	98.95	99.65	98.78
Sum :	100.11	100.41	100.78	100.52	100.16	100.06	99.40							
Si	2.97	2.98	2.98	2.97	1.96	1.95	1.95	Si	2.74	2.98	2.74	2.97	2.74	2.98
Al	1.98	2.00	1.99	2.00	0.06	0.08	0.08	Al	1.24	1.02	1.25	1.02	1.25	1.01
Fe	1.89	1.97	1.86	1.89	0.89	0.82	0.84	Ca	0.25	0.00	0.26	0.00	0.27	0.00
Mn	0.05	0.06	0.05	0.05	0.01	0.01	0.01	Na	0.77	0.09	0.75	0.09	0.74	0.08
Mg	0.94	0.84	0.92	0.90	1.07	1.13	1.11	K	0.02	0.89	0.02	0.93	0.01	0.93
Ca	0.19	0.18	0.22	0.20	0.01	0.01	0.01	Sum :	5.02	4.98	5.02	5.01	5.01	5.00
Sum :	8.03	8.03	8.02	8.01	4.00	4.00	4.00							
Al IV					0.02	0.03	0.03							
Fe 2					0.89	0.82	0.84							
Mn					0.01	0.01	0.01							
Mg					1.07	1.13	1.11							
Ca					0.01	0.01	0.01							
Sum :					2.00	2.00	2.00							
PHLOGOPITE AND GARNET-PHLOGOPITE PAIRS :														
2d) Primary phlg in qtz (I)				2e) Primary gnt-phlg pairs (II):				2f) Secondary phlg (III):						
	A2d	A2e	A2f	A2a gnt	A2a phlg	A2b gnt	A2b phlg	A2c gnt	A2c phlg	A2	B2	C2		
SiO2	37.72	37.34	37.35	38.53	39.16	38.37	38.45	38.95	39.03	38.30	37.20	37.80		
Al2O3	14.94	15.07	15.00	22.27	16.59	22.39	15.48	22.15	15.51	16.30	14.70	14.70		
TiO2	5.18	5.11	5.02	0.00	2.14	0.02	5.20	0.02	4.82	2.23	5.52	5.08		
Cr2O3	0.13	0.14	0.25	0.06	0.20	0.05	0.06	0.03	0.29	0.10	0.24	0.17		
FeO	14.15	12.53	13.04	29.91	8.05	28.29	9.21	26.94	9.17	8.15	11.40	10.60		
MnO	0.00	0.00	0.00	0.56	0.00	0.63	0.07	0.68	0.00	0.00	0.00	0.00		
MgO	13.77	14.57	14.82	7.64	20.35	7.06	17.46	7.81	17.93	19.30	15.50	16.30		
CaO	0.00	0.03	0.00	2.13	0.00	3.98	0.03	3.51	0.04	0.00	0.00	0.00		
Na2O	0.06	0.07	0.00	0.00	0.04	0.03	0.06	0.00	0.00	0.06	0.07	0.07		
K2O	9.02	9.26	8.87	0.01	9.87	0.06	9.75	0.04	8.85	10.50	10.20	10.40		
Sum :	94.97	94.12	94.35	101.11	96.40	100.88	95.77	100.13	95.64	94.94	94.83	95.12		
Si	2.80	2.79	2.78	2.97	2.79	2.96	2.78	3.00	2.80	2.78	2.75	2.78		
Al	1.31	1.33	1.32	2.02	1.39	2.04	1.32	2.01	1.31	1.39	1.28	1.27		
Ti	0.29	0.29	0.28	0.00	0.12	0.00	0.28	0.00	0.26	0.12	0.31	0.28		
Cr	0.01	0.01	0.02	0.00	0.01	0.00	0.00	0.00	0.02	0.01	0.01	0.01		
Fe	0.88	0.78	0.81	1.93	0.48	1.83	0.56	1.74	0.55	0.50	0.71	0.65		
Mn	0.00	0.00	0.00	0.04	0.00	0.04	0.00	0.04	0.00	0.00	0.00	0.00		
Mg	1.53	1.62	1.65	0.88	2.16	0.81	1.88	0.90	1.92	2.09	1.71	1.65		
Ca	0.00	0.00	0.00	0.18	0.00	0.33	0.00	0.29	0.00	0.00	0.00	0.00		
Na	0.01	0.01	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.09	0.01	0.01		
K	0.86	0.88	0.84	0.00	0.90	0.01	0.89	0.00	0.81	0.98	0.97	0.97		
Sum :	7.69	7.71	7.70	8.02	7.86	8.02	7.72	7.98	7.67	7.96	7.75	7.62		

inclusion trails end abruptly at the grain boundaries; the fluids have evidently escaped during the annealing process. In other cases, the inclusion trails crosscut boundaries between the grains constituting the groundmass of the mortar fabric.

This puts time constraints on the fluid inclusion formation in quartz and feldspar. They are formed after host mineral growth, but at least partially before the annealing process connected to the latest deformational phase.

Fluid inclusions in garnet occur in discrete clusters, well removed from grain boundaries. They are definitely related to the growth of the host mineral and are hence regarded as primary. As quartz, feldspar and garnet are part of the same mineral paragenesis, this puts an additional time constraint on the inclusion chronology: fluid inclusions in garnet are formed previous to fluid inclusions in quartz and plagioclase.

The abundance of inclusions decreases from core to rim of the garnet grains. By optical estimate, the volume proportion of the pure  $\text{CO}_2$  inclusions in some of the garnet cores, are found to be 10 to 15 volume percent. The density of the garnet is roughly  $4 \text{ g/cm}^3$ . With a  $\text{CO}_2$  density of  $1 \text{ g/cm}^3$ , this means that these parts of the garnets contain about 3 to 5 weight percent  $\text{CO}_2$ !

### Mineral analyses

Microprobe analyses were carried out with a Cambridge Microscan-9 microprobe at the Free University of Amsterdam, and with a Cameca Camebax MICROBEAM microprobe at the Geological Museum in Oslo. The chemical composition of coexisting mineral phases were obtained from three different thin sections, denoted A2, B2 and C2. Analytical results are given in Table 2. Garnet (Table 2a), orthopyroxene (Table 2b) and feldspar (Table 2c) analyses from the three different sections show rather small chemical variations, supporting the view that they show a weak retrograde zonation. The composition of the micas is highly variable, but as in general  $\text{MgO}$  content is significantly higher than  $\text{FeO}$ , the name «phlogopite» will be used in the rest of

the study. Strong variations in  $\text{FeO}$ ,  $\text{MgO}$  and especially  $\text{TiO}_2$  indicate the possibility of a continuous evolution and a probable secondary origin for some micas, confirming the petrographical observations. Low  $\text{TiO}_2$ , weakly pleochroic micas are indeed clearly secondary. On the basis of their occurrence in the rocks, 3 groups of phlogopite have been defined and analyzed:

- I) Primary phlogopite entirely isolated in quartz (2d, Table 2).
- II) Primary phlogopite enclosed in solid inclusions in garnet (no reaction rim) (2e, Table 2).
- III) Clearly secondary phlogopite, most of them transitional with the outer limit of the garnet (2f, Table 2).

### Geothermobarometry

Four geothermometers and two geobarometers are applied to the Doddabetta charnockite in order to establish the P-T-conditions during metamorphism. The thermometers used are:

- Orthopyroxene-garnet (SEN and BHATTACHARYA, 1984);
- Alkali feldspar-plagioclase (STORMER, 1975);
- Garnet-biotite (THOMPSON, 1976; FERRY and SPEAR, 1978).

Pressure estimates are obtained by using the two different garnet-orthopyroxene-plagioclase-quartz barometers given by PERKINS and CHIPERA (1985) and NEWTON and PERKINS (1982) (Fig. 4).

### Thermometry

The opx-gnt thermometer (SEN and BHATTACHARYA, 1984) is based on the partitioning of  $\text{Fe}^{2+}$  and  $\text{Mg}$  between coexisting opx and garnet, and the Ca content of the garnet. It is only slightly pressure dependent, but a pressure correction is included in the general formula, given by

$$1) T^{\circ} = \frac{2713 + 0.022 P + 3300 X_{\text{Ca}}^{\text{GNT}} + 195 (X_{\text{Fe}}^{\text{GNT}} - X_{\text{Ca}}^{\text{GNT}})}{-1.9872 \ln K_d + 0.787 + 1.5 X_{\text{Ca}}^{\text{GNT}}} - 273.15$$

where T is the equilibrium temperature of the opx-gnt assemblage, given in degrees C and

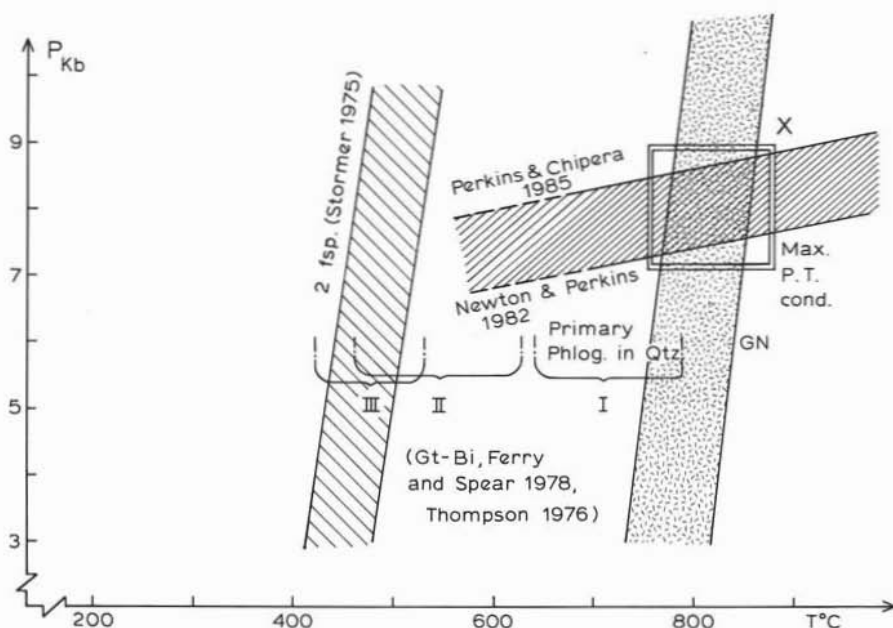


Fig. 4. — Geothermometric and barometric data. A summary. Geothermometry (subvertical curves in the P.T. diagram); Loose hatching: Two feldspar thermometer (STORMER, 1975); I, II and III: garnet-biotite (I: Primary phlogopite entirely isolated in quartz, II: Primary phlogopite in garnet, III: Secondary phlogopite). Note the coincidence between III and the feldspar thermometer. Small streaks: garnet-orthopyroxene equilibria (SEN and BHATTACHARYA, 1985). Geobarometry (dense hatching): garnet-orthopyroxene - plagioclase - quartz equilibria (NEWTON and PERKINS, 1982; PERKINS and CHIPERA, 1985). X = Preferred peak metamorphic P.T. conditions (Gnt.OpX and Gnt.OpX.Plag.Qtz. curves).

P the pressure in bars. The distribution coefficient for  $\text{Fe}^{2+}$  and Mg between orthopyroxene and garnet is given by

$$2) K_D = \frac{\frac{X_{\text{Fe}}^{\text{OPX}}}{X_{\text{Mg}}^{\text{OPX}}}}{\frac{X_{\text{Fe}}^{\text{GNT}}}{X_{\text{Mg}}^{\text{GNT}}}}$$

When the mineral analyses are recalculated to give a sum of 1.00 for the cation content of each of the minerals, and the  $K_D$  is calculated, the general formula can be applied directly. In Table 3, the temperatures are given along with corresponding, selected pressures. (The precisions used in this table, are those stated by the authors in the original publications). Due to the very constant mineral compositions, the temperature show very little spread around an average of about 800°C. The garnet used in these calculations, are the same as those used for gnt-biot thermometry.

The garnet show rather constant composition, while the phlogopites show very variable  $\text{Fe}^{2+}/\text{Mg}$  ratios (see Tables 2d, e and f), giving a broad range of gnt-biot temperatures. Primary gnt-phlog pairs (analyses are given in Table 2e) give temperatures of 425 to 535°C at 6 kb. Due to a weak zonation, the use of core compositions of the garnets give temperatures roughly 30 to 40°C above those obtained from analyses near the grain contacts. The primary phlogopites, occurring as inclusions in quartz, give the much higher gnt-biot temperatures of 639 to 783°C. The secondary, and at least partly synkinematic phlogopites, give temperatures of 461 to 630°C at 6 kb (I, II and III, Fig. 4).

The two feldspar thermometer of STORMER (1975) give temperatures of 480 and 505°C (Table 3). WHITNEY and STORMER's (1977) two feldspar thermometer gives results about 50°C higher. But according to the reasoning given by BROWN and PARSONS (1981), the WHITNEY and STORMER (1977) thermometer should not

TABLE 3

*Thermometry (garnet-orthopyroxene, two feldspar, garnet-biotite)*

(*Sen and Bh* = SEN and BHATTACHARYA, 1984; STORMER, 1985; *F. & S.* = FERRY and SPEAR, 1978; *Thompson* = THOMPSON, 1976).  $Kd = Fe/Mg$

	Gnt-biot				Gnt-biot				Gnt-biot			
	Sample	F & S	Kd	Thompson	Sample	F & S	Kd	Thompson	Sample	F & S	Kd	Thompson
8 Kbar	A2a	431±50	0.101	439±50	A2d	787±50	0.284	696±50	A2	468±50	0.118	468±50
	A2b	496±50	0.131	491±50	A2e	702±50	0.237	639±50	B2	637±50	0.203	594±50
	A2c	532±50	0.149	518±50	A2f	712±50	0.249	645±50	C2	580±50	0.173	554±50
6 Kbar	A2a	425±50	0.101	439±50	A2d	780±50	0.284	696±50	A2	462±50	0.118	468±50
	A2b	490±50	0.131	491±50	A2e	694±50	0.237	639±50	B2	630±50	0.203	594±50
	A2c	526±50	0.149	518±50	A2f	703±50	0.249	645±50	C2	574±50	0.173	554±50

	Sample	Gnt-opx		Two fsp
		Sen & Bh	Kd	
8 Kbar	A2	804±60	0.3554	505±30
	B2	808±60	0.3564	505±30
	C2	815±60	0.3602	500±30
6 Kbar	A2	788±60	0.3554	485±30
	B2	792±60	0.3564	485±30
	C2	799±60	0.3602	480±30

be used on the studied sample, because the model used for Al/Si order/disorder is unsuitable for regional metamorphic rocks. This leaves STORMER's (1975) thermometer as the best choice.

Very coarse exsolution lamellae are found in the feldspar, indicating that subsolvus re-equilibration has occurred. In applying STORMER's (1975) two feldspar thermometer, these exsolution textures pose a danger: Microprobe analyses may give results that are not representative for the average composition in each of the grains. But in our case, all feldspar give very similar results. The temperatures compare favourably with those obtained from the completely re-equilibrated primary garnet-biotite pairs.

In conclusion, three out of the four thermometers, that is the two gnt-biot and the kfsp-plag thermometers, are believed to give less-than-maximum temperature estimates. As gnt and opx are part of the stable granulite assemblage, the temperatures obtained from the gnt-opx thermometry are regarded as peak metamorphic temperatures. Phlogopite in physical contact with garnet show much lower  $Fe^{2+}/Mg$  ratios, resulting in much lower gnt-biot thermometrical values, than those

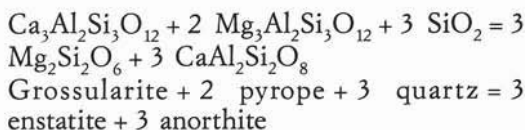
completely enclosed in quartz (group II), whereas all garnets have similar compositions. Due to the low mica content in the rock in comparison with the gnt and opx contents, a phlg-gnt re-equilibration, at a late metamorphic stage, seems to have affected the gnt-biot temperature estimates only to a limited degree. But the gnt-biot temperatures obtained for this rock are probably somewhat doubtful, even though local re-equilibration may have been obtained for coexisting gnt-phlg pairs. Still, the group II phlogopites have  $Fe^{2+}/Mg$  ratios nearly compatible with maximum P-T conditions of the rock.

### Barometry

Two end member reactions are used as a basis for the PERKINS and CHIPERA (1985) geobarometer.

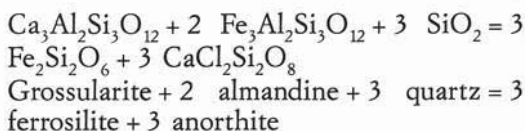
The first is termed the Mg-reaction:

3)



The second is termed the Fe-reaction:

4)



The two equilibrium constants are given by

5)

$$Mg\text{-reaction: } \ln K_1 = \ln \frac{a_{EN}^3 \cdot a_{AN}^3}{a_{GR} \cdot a_{PY}^2}$$

6)

$$\text{Fe-reaction: } \ln K_2 = \ln \frac{a_{\text{FS}}^3 \cdot a_{\text{AN}}^3}{a_{\text{GR}} \cdot a_{\text{ALM}}^2}$$

In order to apply the barometer, the activities of the hypothetical end members are needed. The PERKINS and CHIPERA (1985) paper contains a set of formulas for calculating the garnet end member activities, which are not repeated here, but which calls for a special attention of the reader, as several formulas must be corrected according to the published errata (Contr. Min. Petr., 1985, 90:410). Particularly important is the expression of plagioclase activity (NEWTON, 1983):

7)

$$a_{\text{AN}} = X_{\text{AN}} \cdot \frac{X_{\text{AN}} (1 - X_{\text{AN}})^2}{4}$$

There are three prerequisites for the calculation of Opx end member activities:

A) The first one is the site assignment: All Ca, Mn, Na should be assigned to the M2 site, and all Ti, Fe<sup>3+</sup>, Al (tetrahedral) should be assigned to the M1 site (WOOD and BANNO, 1973).

B) Secondary, the Fe<sup>2+</sup> to Mg ratio should be equal in the two sites. This can be expressed by the expressions:

$$\frac{X_{\text{Fe}^{2+}}^{\text{M1}}}{X_{\text{Mg}}^{\text{M1}}} = \frac{X_{\text{Fe}^{2+}}^{\text{M2}}}{X_{\text{Mg}}^{\text{M2}}}$$

C) The third and probably most essential prerequisite (which is not clearly stated in the paper), is that the cation content in the opx should be recalculated, so that *each* of the M1 and M2 sites sum up to one:

$$M_1 \text{ cations} = M_2 \text{ cations}$$

(Note that recalculated opx analyses are given in Table 2b). Once these requirements are observed, the activities of the opx end members are calculated by simple expressions:

8)

$$a_{\text{Fe}_2\text{Si}_2\text{O}_6} = X_{\text{Fe}}^{\text{M1}} \cdot X_{\text{M}_2}^{\text{Fe}}$$

9)

$$a_{\text{Mg}_2\text{Si}_2\text{O}_6} = X_{\text{M}_1}^{\text{Mg}} \cdot X_{\text{M}_2}^{\text{Mg}}$$

Two general formulas are given for the pressure calculations:

10)

$$P_{\text{Mg}} = 6.1349 - 0.3471 \ln K_1 + 0.02136 T - 0.001140 T \ln K_1$$

11)

$$P_{\text{Fe}} = 0.0630 - 0.3482 \ln K_2 + 0.0143 T - 0.000997 T \ln K_2, \text{ where } P \text{ is given in kilobars and } T \text{ in degrees C.}$$

Results are given in Table 4.

NEWTON's and PERKIN's (1982) gnt-opx-plag-qtz geobarometer is also applied to the rocks. It is based on reaction 3 given above (the Mg-reaction), but a different calibration than the PERKINS and CHIPERA (1985) Mg-reaction barometer.

Calculation procedures for the enstatite end member activities are exactly the same as those mentioned above. The garnet end member activities are based on a different set of solution models given by GANGULY and KENNEDY (1974). Calculations of plagioclase end member activities are based on the «Al-avoidance» model given by NEWTON et al. (1980). The general barometric formula is given by:

12)

$$P = 3,994 + 13.070 T + 3.5039 T \ln K_A$$

whith P in bar and T in K.  $K_A$  is given by:

$$K_A = \frac{(a_{\text{Ca}} \cdot a_{\text{Mg}}^2)^{\text{GNT}}}{(a_{\text{Ca}}^{\text{PLAG}}) (a_{\text{Mg}}^{\text{OPX}})}$$

Results from this geobarometer are presented in two versions in Table 4. Row no. 1 in this table features results obtained following NEWTON and PERKINS (1982) in all calculation steps. Row no. 2 is obtained using the more up to date plagioclase activity model given by NEWTON (1983).

The NEWTON and PERKINS (1982) and NEWTON and PERKINS (1982)/NEWTON (1983) barometers show results in the range 7.6 to 8.7 kbar with an average value of 8.2 Kbar, and

TABLE 4

Barometry: GNT-OPX-PLAG-QTZ T = 800 C

Sample	N & P /		Perkins & chipera 1985	
	N & P	N 1983	P Mg	P Fe
A2	7637±1500	7419±1500	7020±1000	7780±1000
B2	8692±1500	8475±1500	7987±1000	8981±1000
C2	8345±1500	8130±1500	7654±1000	8530±1000
Average:	ca. 8200	ca. 8000	ca. 7600	ca. 8400

7.4 to 8.5 kbar with an average of 8.0 kbar, respectively. This small differences in average pressure is of course due to the different plagioclase activity models used, the one featuring the newest model (NEWTON, 1983) giving the lower pressures. This last one gives results which are comparable to the graphical barometer of BOHLEN *et al.* (1983).

The PERKINS and CHIPERA (1985) barometer pair gives results of about 7.6 kbar for the Mg-reaction and 8.4 kbar for the Fe-reaction. The averaged value of 8.0 kbar is identical to the pressure average of the pressures obtained by the other barometers. A schematic summary of thermobarometric data are given in Figure 4.

NEWTON and PERKINS (1982) found pressures of roughly 7.5 kbar for other rocks in the Doddabetta area. This is consistent with our results obtained with the same barometer. In conclusion, the best pressure estimate for our Doddabetta charnockite lies in the range 7 to 8 kbar for a given temperature of 800 degrees C. Further elaboration concerning these barometric results is beyond the scope of this publication.

Both two feldspar thermometry and gnt-opx-plag-qtz barometry involve the use of plagioclase end member activities. Results from two feldspar-thermometry show that the feldspar have reequilibrated after peak metamorphic conditions, and thereby the end member activities have changed. This, of course, influences the barometrical results. In the case of the two feldspar thermometer, a small change in the Ab content of the alkali feldspar gives a relatively large change in the resulting temperature. This relation is exactly opposite for the plagioclase, because the Ab content of the alkali feldspar is here more important than the Ab content of the plagioclase. But due to the high Ab content in plagioclase, and the high plag/kfsp ratio, the influences of the feldspar reequilibration on the barometrical results are not necessarily large.

Both the Mg and Fe-reactions (equation 3 and 4) involve a calcium exchange between anorthite and grossularite. A change in the plagioclase end member activities would also change the grossularite activity

correspondingly, and again minimize the effect on the barometric results.

### Microthermometry/Fluid inclusions

Many inclusions in garnet contain an aggregate of daughter minerals (Fig. 3). Several of these could remain in an opened inclusion (freshly and gently broken surface) and they have been analyzed using a Jeol JSM 840 scanning electron microscope fitted with a Link AN 10000 energy dispersive analyser (EDS) unit. The EDS spectrum in Figure 3, is interpreted as a Fe, Mg dominated carbonate. As shown in the Figure, the aggregates commonly consist of carbonate platelets. The extremely small Ca content is striking.

At room temperature, each of the inclusions are filled with a monophase liquid. This means, of course, that their final homogenization temperature are less than about 20°C. All inclusions homogenize into the liquid phase (denoted by Th(L)).

Contaminants in the form of other gaseous species (notably N<sub>2</sub>, CH<sub>4</sub>) are commonly found in inclusions from other high grade terranes (see e.g. TOURET, 1985 and SWANENBERG, 1980). This may seriously affect thermobarometric interpretations (BURRUSS, 1981; GUILHAUMOU *et al.*, 1981).

All measured inclusions show melting temperature of -56.6°C, which is the triple point of CO<sub>2</sub>. This observation is of major importance, as it indicates the purity of the CO<sub>2</sub>. The composition of the inclusion contents was also checked with a Laser-Raman microprobe with a multichannel detector system (DILOR MICRოდIL 28). This equipment facilitates fast and reliable in situ analyses of molecular species inside single inclusion cavities. (The resolution lies in the one micron range). The analytical results confirmed the purity of the CO<sub>2</sub> (Fig. 5). (For a description of this system and its advantages, the reader is referred to BURKE and LUSTENHOUWER (1987). Several analyses have been done on inclusions in quartz, plagioclase and garnet, respectively, which all concluded with the absence of any other gas than CO<sub>2</sub> (no CH<sub>4</sub>, N<sub>2</sub>, H<sub>2</sub>, H<sub>2</sub>S). For any of these gases, the detection limit is below 1%.

This, combined with the melting data, will allow us to interpret all present Th data in terms of pure  $\text{CO}_2$ . This does not imply that no other gases will ever been found in the Dodda Betta area, but a more complete investigation is far beyond the scope of the present paper.

(1, Fig. 6), contains two separate peaks, with maxima at  $-5$  and  $-18^\circ\text{C}$ , corresponding to  $\text{CO}_2$  densities of  $0.96$  and  $1.02 \text{ g/cm}^3$ , respectively. (Density data are adapted from ANGUS et al., 1973 and HOLLISTER, 1981). The complete separation of the peaks may be due to variations in  $\text{CO}_2$  pressure during garnet

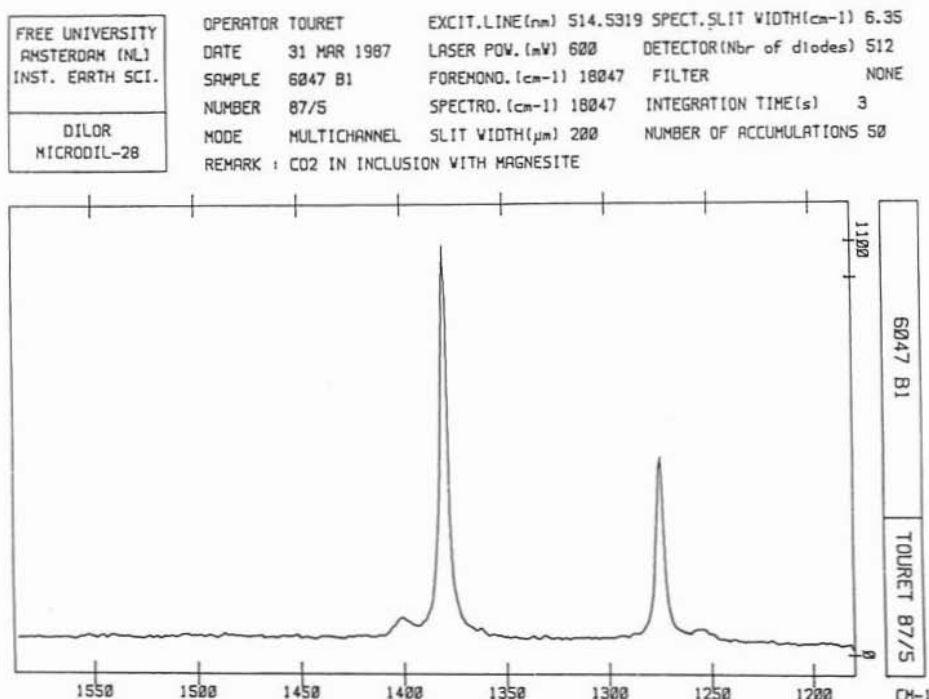


Fig. 5. — Raman spectrum of  $\text{CO}_2$  (Microdil 28, Free University, Amsterdam, Analyst: E. Burke).

Fluid inclusions of sufficient size and abundance for microthermometric studies are found in three minerals; garnet, quartz and plagioclase. The measurements were performed on eight selected grains: 3 garnets, 3 quartzes and 2 plagioclases. The results are presented in Fig. 6. Note that in plagioclase, measurements were done on inclusion trails both parallel to and inclined to the best cleavage plane (001).

Fluid inclusion measurements from each of the mineral species were used to construct frequency histograms, presenting homogenization temperatures relative to the number of measurements (Fig. 6). Some differences between the three histograms are immediately evident. The one with garnet data

growth, possibly reflecting changes in the fluid regime during the early stages of peak metamorphism. However, it has not been possible to observe a clear spatial distribution between inclusions with different densities and both values have been used for the bracketing of the garnet field (Fig. 6).

The plagioclase histogram (2, Fig. 6) seems to consists of two partly superimposed peaks. They occur at  $-27$  and  $-35$  degrees C, corresponding to  $\text{CO}_2$  densities of  $1.06$  and  $1.09 \text{ g/cm}^3$ , respectively. Homogenization temperatures from the quartz inclusions (3, Fig. 6) show a remarkable uniformity. They are concentrated in a single, narrow peak with a maximum at  $-39^\circ\text{C}$ , corresponding to a  $\text{CO}_2$  density of  $1.12 \text{ g/cm}^3$ .

Another histogram, labelled «combi» (4, Fig. 6), combines all Th measurements. The most conspicuous peak is represented by the quartz inclusions. To sum up,  $\text{CO}_2$  densities are very uniform for each separate mineral, compared to many other examples (e.g. in Southern Norway, TOURET, 1985; Th commonly varies between  $-20$  and  $+30^\circ\text{C}$  in a single mineral grain). Moreover, the density tends to increase in order of increasing deformation and recrystallization (increasingly higher densities from garnet, through plagioclase to quartz).

consistent with TOURET and BOTTINGA, 1979). Here they are shown together with P-T-estimates from solid phase thermobarometry. Because these isochores represent frequency histograms, they are not drawn as single lines, but rather as «isochore bands», covering the density ranges indicated in the figure.

Fluid inclusion petrography has shown that the garnet inclusions are primary with respect to peak metamorphic conditions. Plagioclase and quartz inclusions are trailbounded and secondary with respect to host mineral growth. This makes them, at least partially,

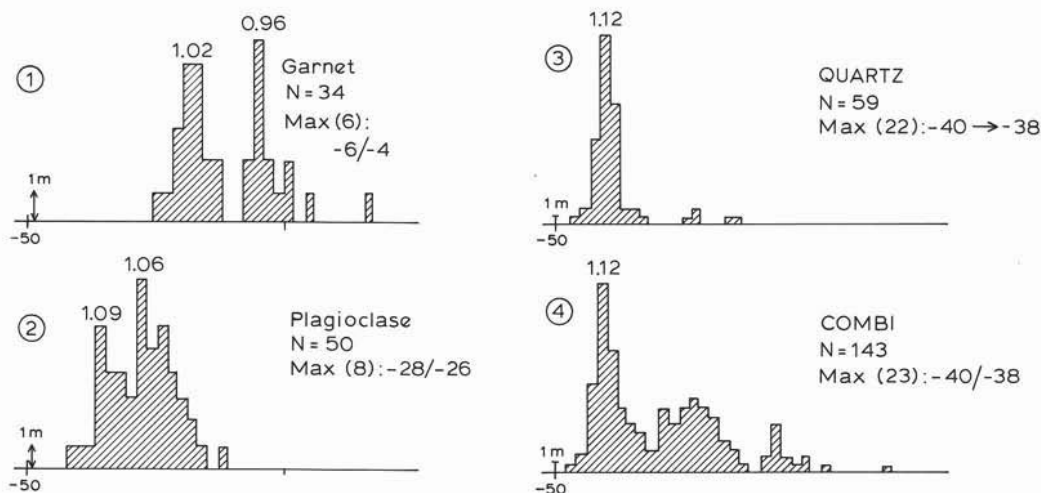


Fig. 6. — Homogenization temperature (Th) histograms for  $\text{CO}_2$  inclusions in garnet (1), plagioclase (2) and quartz (3). (All temperatures denote *liquid* homogenization). The above three histograms are combined in «combi» (4). For each histogram: The first line (N) is the local number of measurements. The second line is the maximum frequency (in number of the total number of measurements) in a given temperature interval. For instance, in histogram No. 4 (combi), 23 of the measurements (maximum frequency) occur between  $-38$  and  $-40^\circ\text{C}$ . Numbers above each peak:  $\text{CO}_2$  density in  $\text{g}/\text{cm}^3$ .

### Evidence for a «isobaric cooling» PT path after the peak of metamorphism

Geothermobarometric calculations show peak metamorphic conditions of  $795$  to  $815^\circ\text{C}$  and  $7.4$  to  $8.7$  kbar. This is illustrated by the «P-T box» in Figure 4. These P-T-estimates are regarded as the starting point for the P-T-story indicated by the present study.

The maximum frequency values from the histograms (Fig. 6) are used to construct isochores, which are drawn in Fig. 7 (Thermodynamic data are taken from HOLLOWAY, 1977; 1981 and they are

later. Knowing this fluid inclusion chronology, the frequency histograms can be used in the further explanations: The average  $\text{CO}_2$  density increases gradually from garnet, through plagioclase to quartz.

Isochores bracketing the «low» density primary inclusions in garnet pass somewhat below the Doddabetta peak metamorphic conditions. Results from other granulite terranes suggest that some  $\text{H}_2\text{O}$  was present during prograde metamorphism (COOLEN, 1982). In our case, this is proven by the presence of primary phlogopite. In cases

similar to this, a water pressure of 1 to 2 kbar in addition to the measured  $\text{CO}_2$  pressure has been indicated (TOURET, 1974). This would make the primary garnet inclusions compatible with the peak metamorphic conditions. The water in the garnet inclusions could have removed through reactions forming daughter minerals, but also at least up to 10 mole%  $\text{H}_2\text{O}$  may remain unnoticed

have lower densities than the quartz ones. (It is also noteworthy that they show much greater internal variations, the ones along the best cleavage plane being less dense). This indicates that most plagioclase inclusions are intermediate in time those from garnet and quartz. All inclusions in quartz have remarkably similar homogenization temperatures, which can only reflect a well

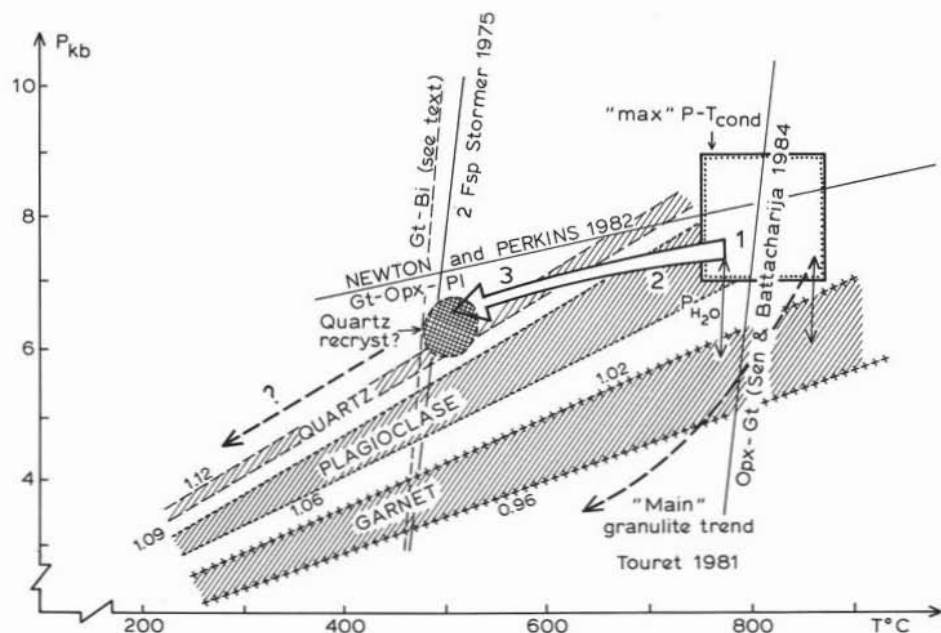


Fig. 7. — Model of post-metamorphic P-T path in the Doddabetta charnockites. Maximum metamorphic P-T conditions and «solid mineral» estimates, see Fig. 4. Hatched:  $\text{CO}_2$  densities in garnet, plagioclase and quartz, respectively. Each band is limited by representative isochores labelled in density ( $\text{g}/\text{cm}^3$ ) and selected from the respective peaks in Th histograms (Fig. 6). White arrow: P-T evolution recorded by the fluid inclusions. The necessity to cross the isochores in the order 1, 2, 3 (garnet, then plagioclase, then quartz) imposes a temperature decrease at almost constant pressure («isobaric cooling»). Note that pressures indicated by the primary fluid inclusions in garnet from peak metamorphic conditions are lower (1-2 kb) than the pressure recorded by the solids. This can tentatively be explained by the formation of retrograde, hydrated minerals (see text).

on the walls of the inclusion cavities.

The  $\text{CO}_2$  densities are increasing through plagioclase and quartz and it is obvious in Fig. 7 that no additional water pressure is required to bring the  $\text{CO}_2$  isochores into the solid mineral P-T field. It can be tentatively estimated that immediately after the peak of metamorphism, all available water has been consumed through the formation of high temperature retrogressive minerals, notably phlogopite.

The majority of the inclusions in plagioclase

defined kinematic event.

Many secondary fluid inclusion trails are destroyed in the formation of quartz-feldspar mortar fabric. Crosscutting relationships show that the accompanying annealing process postdates the quartz inclusion trails. If the  $\text{CO}_2$  inclusions in quartz were formed early during the deformational phase, this would fit well with the observed fluid inclusion chronology. Additional evidence of the relative timing of the deformational phase, is that some of the secondary phlogopite grains

are strongly deformed. Some of the secondary phlogopite is thus pre- to syn-kinematic. The gnt-biot thermometers (FERRY and SPEAR, 1978; THOMPSON, 1976) give a large spread in temperatures for secondary micas. This points toward the development of very local equilibria after peak metamorphic conditions. The temperatures obtained by STORMER's (1975) two feldspar thermometer, are roughly in the same range as the lowest temperature obtained by using the two gnt-biot thermometers on the group 1 primary gnt-phlg pairs, i.e.  $500 \pm 80^\circ\text{C}$ . If the fluid inclusions in quartz are trapped during the deformation phase, and the subsolvus feldspar re-equilibration is contemporaneous with the latest phlogopite re-equilibration, a semiquantitative area on the P-T-path can be defined. This is the area where the quartz isochores intersect the gnt-biot and two feldspar thermometric curves (Fig. 7). Even if the gnt-biot temperatures are erroneous, this area is well defined by the intersection of the quartz isochores and the two feldspar thermometric curve.

When all textural, thermobarometric and microthermometric informations are combined, they reveal the following P-T-story; peak metamorphic conditions of 7.4 to 8.7 kbar and roughly  $800^\circ\text{C}$  were obtained by thermobarometry, and fits relatively well with the densities of primary inclusions found in garnet. Secondary feldspar and quartz inclusions show a gradual increase in  $\text{CO}_2$  densities after peak metamorphism. Formation of secondary phlogopite possibly covers the entire temperature range from peak metamorphism, to less than  $500^\circ\text{C}$ . Deformation and sub solvus feldspar re-equilibration both occur at temperatures of very roughly  $500^\circ\text{C}$  and 5 to 6 kbar. This last recorded event is indicated by the intersection of the two feldspar and gnt-biot thermometric curves with the isochores for quartz (Fig. 7).

#### Comparison with other granulite facies terranes

In many granulites (TOURET in HOLLISTER and CRAWFORD, 1981) both mineral and microthermometric data are compatible with

an adiabatic uplift model (Fig. 8, I). In such cases, the lower continental crust has been uplifted from 6 to 8 kbar (corresponding to a depth of 20-35 km) through a cooling path somewhat convex towards the T axis (see e.g. COOLEN, 1980). This can be regarded as the most «normal» uplift path. The fluid inclusions decrease in density after peak metamorphism. This is exactly the opposite of what is found in Doddabetta.

The case of Southern Norway is more complicated. In the western part (Rogaland), SWANENBERG (1980) found increasing  $\text{CO}_2$  densities in post-metamorphic inclusions and he was then the first to propose a P-T path comparable to the Doddabetta one («isobaric cooling», II, Fig. 8). Some of the Swanenberg's arguments are now discredited (low density  $\text{N}_2\text{-CO}_2$  mixtures mistaken for «high density pure  $\text{CO}_2$ », TOURET and VAN DEN KERKOF, 1986). The concept of an isobaric cooling path, however, found a regional support by the discovery of late, post metamorphic kyanite in the Bamble, eastern part of Southern Norway (see discussion in TOURET and OLSEN, 1985; and TOURET, 1985). In this region, the part of the P-T trajectory convex towards the P axis is explained by the synkinematic emplacement in the lower crust of deep-seated intrusives of basic or intermediate composition, which carry the  $\text{CO}_2$  from the mantle. They are emplaced at high temperatures (at least  $1000^\circ\text{C}$ ) and then the cooling at regional metamorphic temperatures will impose an increase in the  $\text{CO}_2$  densities, resulting in the part of the «isobaric cooling» P-T path illustrated in Fig. 7 and/or 8. The path proposed for the Doddabetta charnockite is in principle similar to the Bamble one, but the cause is obviously different. No deep-seated intrusives can be invoked in this case, and the regional trend (SANTOSH, 1985; HANSEN et al., 1984) as suggested by all earlier published data, is a «normal» one (adiabatic uplift).

Considering the extreme deformation of the rock and the relation with density evolution, it can be speculated that the «isobaric cooling» trend is related to some ductile megashear zone starting at (or close

to) peak metamorphic conditions and developing during post metamorphic regional cooling. This is (freely) illustrated in Fig. 8 III as a thickening of the continental crust due to overthrusting, followed or accompanied by an isostatic readjustment which would bring the lower part of the initially above lying crust to the surface. At any temperature along a «normal» post metamorphic path (e.g.  $T_1$ ,

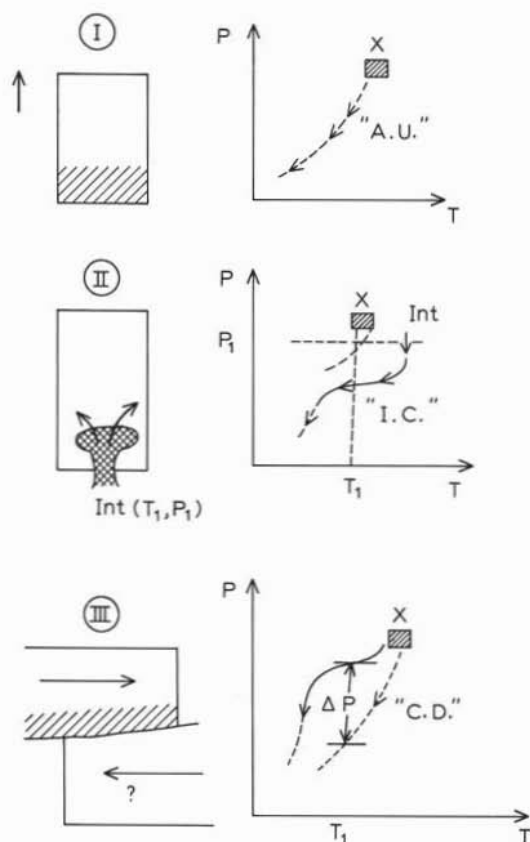


Fig. 8. — Post-metamorphic P-T trajectories in granulites and related rocks. I: «Adiabatic uplift» (A.U.). This trajectory, still regarded as the «normal» granulite trend (TOURET, 1981), corresponds to vertical movements of a continental block (hatched = lower crustal, granulite domain). II: «Isobaric cooling» (I.C.) = Temperature decrease at high pressure, related to the intrusion, at a regional  $P_1$ ,  $T_1$  condition, of a deep-seated intrusion, (Int.). T suddenly increased, and the trajectory is explained by the cooling at depth of the intrusion. III: Isobaric cooling trend explained by continental thrusting (continental doubling = C.D.). Regional trend would be I (isostatic vertical movement with the base of the crust still at high metamorphic temperature), but at any temperature (e.g.  $T_1$ ) an excess pressure  $P$  is caused by a continental overthrusting.

Fig. 8), the overthrusting imposes an additional pressure  $\Delta P$  which explains the observed trajectory (solid line with arrow, III, Fig. 8). Needless to say, this model is entirely speculative. Only one sample has been studied, and drawing conclusions at a megascale is hazardous at best. But at least the reality of a trend concave towards the temperature axis is now firmly established. This makes without any doubt the interpretation of fluid inclusion data much more complicated, but also much more interesting than initially assumed.

### Conclusion

By carefully combining petrography, thermobarometry and microthermometry, severe constraints can be put both on the evolution of the fluid density and on the P-T-paths of a rapidly increasing number of high grade rocks.

This task is never easy, but some specimens are better suited than others. In this respect the Doddabetta is one of the best that we have ever seen, both for the quantity of inclusions and the specific character of the inclusions in a given mineral. One sample is not enough to draw large scale conclusions, and we will restrain to comment here on the much debated problem of fluid regime in the lower crust («vapour absent metamorphism», THOMPSON, 1980) or «carbonic metamorphism», e.g. NEWTON et al., 1980). This problem is adequately covered in a number of recent publications, notably CRAWFORD and HILLOSTER, 1986; and NEWTON, 1986).

It can be stated, however, that the abundance of  $\text{CO}_2$  in the Doddabetta charnockite suggests some local influx of  $\text{CO}_2$  of an external source. Mantle origin has been repeatedly proposed by the senior author for Southern Norway, where  $\text{CO}_2$  inclusions are specially abundant in meta intrusive rocks, associated with carbonate of possibly primary magmatic origin (TOURET, 1985). In that publication, it was stated that this hypothesis was supported by some melting experiments (OTTO, 1984). In fact, OTTO's results were misinterpreted and no experimental proof of

immiscibility between carbonate and silicate melts has ever been observed under lower crustal conditions. But on the other hand, direct evidence of immiscibility between carbonate and silicate melts has recently been observed in mantle xenoliths, in the form of quenched droplets of carbonatic and basaltic glass in lherzolites (AMUNDSEN, 1987). This gives way for interesting aspects concerning the source of  $\text{CO}_2$ . Observing the Doddabetta charnockite, we want to stress the petrographic relation between primary  $\text{CO}_2$  inclusions and a comparable type of carbonate (iron and magnesium rich) as found in the Bamble meta intrusives (TOURET, 1985). We are convinced that some genetic relation does exist between them, and its precise understanding will greatly enhance our knowledge of the source of  $\text{CO}_2$  in the lower crust.

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