

Mafic granulites and clinopyroxenite xenoliths from the Transdanubian Volcanic Region (Hungary): implications for the deep structure of the Pannonian Basin

A. EMBEY-ISZTIN

Department of Mineralogy and Petrology, Hungarian Natural History Museum, 14–16 Muzeum korut, H-1088 Budapest, H-1370 Pf: 330, Hungary

H. G. SCHARBERT, H. DIETRICH

Institute of Petrology, University of Vienna, Dr. Karl Lueger Ring 1, A-1010 Vienna

AND

H. POULTIDIS

Steyrerstrasse 14, A-4531, Kematzen, Austria

Abstract

The Transdanubian Volcanic Region (TVR) is composed mainly of Pliocene alkali basalts, basanites, olivine basalts and olivine tholeiites, as well as rare nephelinites. The partial melting and genesis of alkali basaltic liquids is a consequence of an upwelling of the upper mantle which also caused thinning of the lithosphere and recent sinking of the Pannonian Basin.

Four different types of lower crustal and upper-mantle xenoliths are found within the TVR: garnet-free and garnet-bearing granulites, clinopyroxenites and spinel lherzolites. We present mineralogical and geochemical data on granulite facies and clinopyroxenite xenoliths from three localities in the Hungarian part of the TVR (Bondoróhegy, Szentbékálla and Szigliget). It is concluded that, whilst the protoliths of the granulite facies xenoliths were tholeiitic igneous rocks and could be part of an ancient crust, the clinopyroxenite xenoliths represent recent underplating and may have formed from an alkali basaltic liquid similar to the host lava. Planar contact relations between clinopyroxenites and spinel lherzolites as observed in composite xenoliths, as well as high Al-contents in clinopyroxenes, point to a high-pressure genesis in the upper mantle for these rocks. In contrast, geobarometrical estimates yielded only a moderate pressure range characteristic of lower crustal levels for the garnet-free granulite xenoliths (7–9 kbar). Nevertheless, two-pyroxene geothermometry yielded high temperatures of equilibration (>900°C) for these xenoliths, probably caused by advective heat transfer connected with underplating and in agreement with the high present-day geothermal gradient of this region. In the Central Range localities only garnet-free granulite xenoliths occur, whereas at the border of the TVR both garnet-free and garnet-bearing granulite facies nodules are found. It is possible that the incoming of garnet is retarded by higher temperatures in the lower crust below the Central Range.

It is also suggested that the difference in seismically measured crustal thickness between the Central Range and adjacent basin areas may be connected with different thermal conditions below these regions and that the seismically defined Moho and the petrological Moho do not necessarily coincide.

KEYWORDS: granulite, clinopyroxenite, xenolith, Transdanubian Volcanic Region, Hungary.

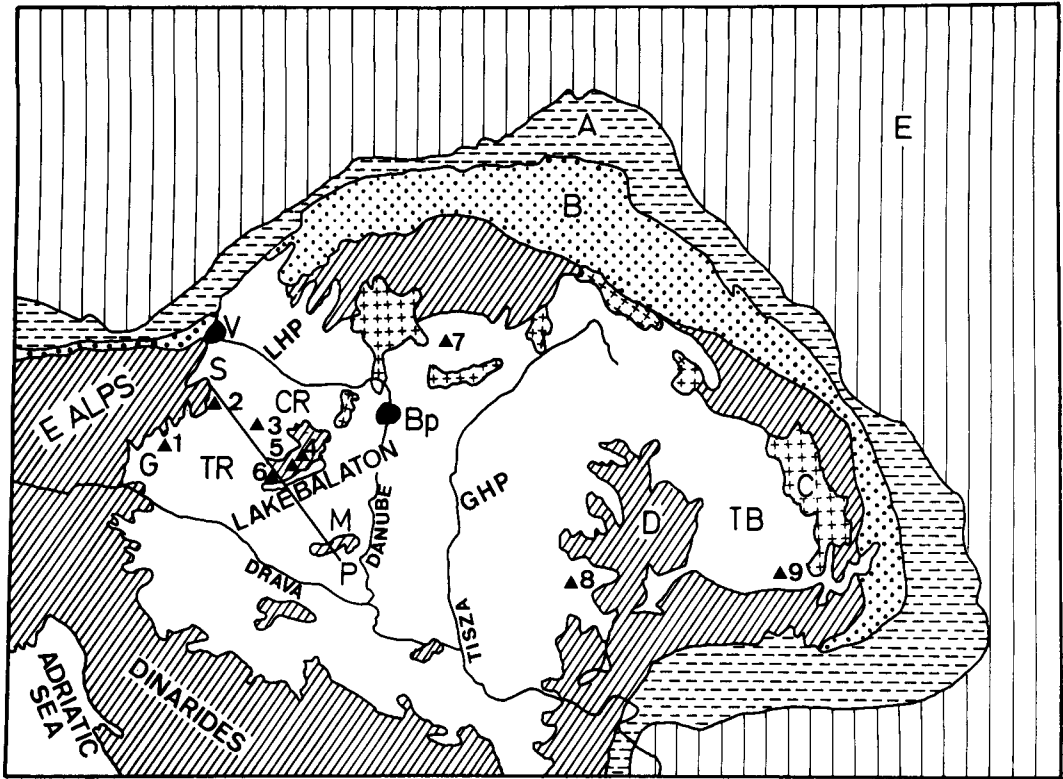


FIG. 1. Simplified geological map of the Pannonian area with the surrounding mountain belts. 1–6: The most important volcanic areas of the TVR: 1 = Graz Basin, 2 = Burgenland, 3 = Gërce, 4 = Bondoróhegy, Kabhegy, Agartető, 5 = Szentbékállá, 6 = Szigliget, Badacsony. 7–9: other young alkali basalt occurrences: 7 = Nógrád district, 8 = Bánát region, 9 = Rákos. A = Molasse Foredeep, B = Alpine-Carpathian Flysch Belt, C = Cenozoic calcalkaline rocks of the Inner Carpathian Volcanic Chain, D = Mesozoic and older inner units of the Alpine, Carpathian and Dinaric mountains, E = Stable Europe, Blank area = Tertiary basin. V = Vienna, G = Graz, Bp = Budapest, TR = Transdanubia, CR = Central Range, LHP = Little Hungarian Plain, GHP = Great Hungarian Plain, M = Mecsek Mts., TB = Transylvanian Basin, Line S–P is the location of seismic profile between Sopron and Pécs shown in Fig. 2.

Structure and evolution of the Pannonian region

THE Pliocene volcanic activity in Transdanubia extends from the Graz Basin and Burgenland (E Austria) to Lake Balaton (W Hungary) (Fig. 1). This arcuate feature has been termed the Styrian Volcanic Arc by several authors (e.g. Hauser, 1954; Heritsch, 1967; Poultidis, 1981). However, since the Styrian Volcanic Arc has no similarity whatsoever with the modern concept of volcanic arcs, i.e. directly related to subduction zones, we propose to replace this expression by Transdanubian Volcanic Region (TVR). The lavas related to this feature changed their character in space and time. A Miocene phase produced trachytic, trachyandesitic and quartz-trachytic lavas restricted to the vicinity of Graz, whereas a Pliocene

phase saw eruption of nephelinitic magmas in the Graz Basin, basanites and olivine tholeiites in Burgenland, basanites, alkali basalts and olivine tholeiites north of Lake Balaton in the Central Range and in the Little Hungarian Plain (Fig. 1). Basalts of Burgenland seem to be older than the much more voluminous basalts in the Hungarian part of the TVR, whereas those of the Graz Basin are younger. The basanites of the Nógrád district, N Hungary (Fig. 1) are also younger (2–2.5 m.y.; Balogh *et al.*, 1986). The alkalic affinity of TVR basalts was confirmed by recent petrological and geochemical studies (Embey-Isztin and Scharbert, 1981; Poultidis and Scharbert, 1986).

It seems probable that the genesis of this

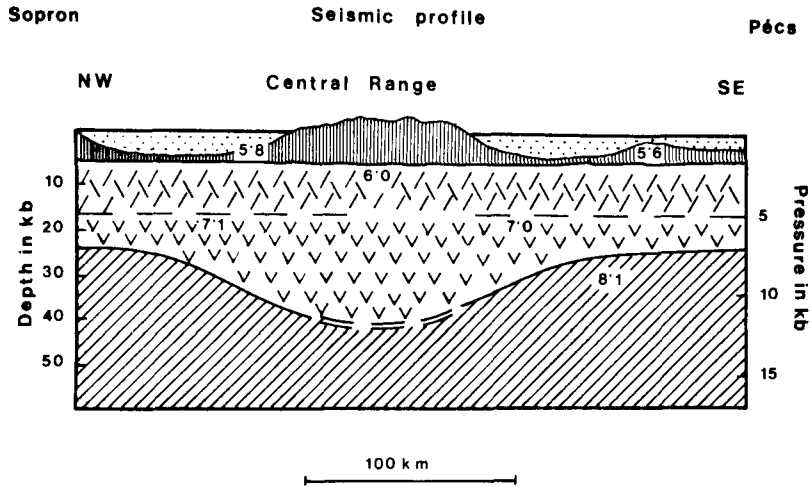


Fig. 2. NW-SE seismic profile through the Central Range and adjacent parts of the Pannonian Basin. Numbers indicate P-wave velocities (Mituch and Posgay, 1972; Posgay, 1975). Note the apparent thickness of the crust beneath the Central Range and the inferred crust-mantle transition.

unevenly distributed, broadly contemporaneous alkali basaltic volcanism in the whole of the Pannonian Basin is linked to the evolution of this basin, the ultimate cause of which is the upwelling of mantle material (Embey-Istzin, 1976a, 1981). As far as the number of eruption centres and volume of volcanic material is concerned, the most important basaltic area is that of the Central Range (Fig. 1).

The Pannonian Basin is a young interarc basin surrounded by the Alpine, Carpathian, and Dinaride orogenic belts (Fig. 1). The evolution of interarc basins is generally considered to be different from that of continental rift zones or grabens. However, a number of present-day geophysical and geotectonic features characteristic of continental rift zones show striking similarities to the Pannonian Basin. Many of these features can probably be extrapolated back over a period of at least 5 m.y. These are:

1. Mainly extensional regime with fault blocks and major sinking of the basement up to 6 km in Kisalföld (Little Hungarian Plain) and Nagyalföld (Great Hungarian Plain). Uplifted block mountains exist in the NE-SW striking Central Range and in the Mecsek. These mountainous regions are built of Palaeozoic and Mesozoic sediments with associated volcanics and Hercynian granites, and are capped by Miocene andesites and Pliocene alkali basalts. Individual members of the Central Range are separated by smaller Paleogene basins, whereas the large basins (Kisalföld and Nagyalföld) were formed by Pliocene

and Quaternary subsidence. The basement is composed of Precambrian, Palaeozoic and Mesozoic rock series (Trunkó, 1969).

2. High but variable heat flow values (70-110 mW/m²) characterize most of the regions of Hungary (Cermák and Rybach, 1979; Horváth *et al.*, 1979; Bodri and Bodri, 1977; Bodri, 1981a; Dövényi and Horváth, 1988).

3. The crustal thickness averages only 27 km (23-35 km) (Posgay, 1982; Meskó, 1982).

4. Below the Bakony Mountains (the largest member of the Central Range) a sharply defined Moho cannot be detected (Fig. 2), but instead there is a transitional crust-mantle boundary zone of about 5 km thickness (Mituch and Posgay, 1972; Posgay, 1982). The disappearance of the Moho seems to be a salient feature of the axis of rift zones (Meissner *et al.*, 1983; Barton *et al.*, 1984).

5. The lithosphere as a whole is also thin. The depth of the low velocity zone (LVZ) is estimated to 53 km deduced from seismic reflections (Posgay, 1982) and between 44 and 66 km based on magnetotelluric measurements (Adám, 1982). However, investigations of surface waves indicated that the LVZ might lie deeper (70-80 km) (Adam *et al.*, 1979; Bodri, 1981a).

6. Eruption of mantle xenoliths in alkali basalts (Embey-Istzin, 1976b; Embey-Istzin and Scharbert, 1989).

The following complexities and apparent con-

traditions arise from consideration of geophysical data:

1. The disappearance of the Moho beneath the Central Range is connected with the region of maximum crustal thickness, instead of the adjacent basin areas where the crust is supposed to be thin (Mituch and Posgay, 1972). Observations in rift and graben areas also show a poorly defined Moho in axial zones; however, these zones are characterized by a minimum crustal thickness (Edel *et al.*, 1975).

2. In the Central Range there is a marked positive Bouguer anomaly, whereas in the basin areas (Kisalföld and Nagyalföld) negative Bouguer anomalies are evident (Meskó, 1982). This is the opposite of what we would expect, since in basin areas the less dense crust is supposed to be considerably thinner, thus the more dense mantle rocks should be closer to the surface. In the Central Range the opposite should be expected i.e. a negative Bouguer anomaly would be predicted instead of the observed positive one since in the Central Range the crust is thought to have a maximum thickness.

3. There is no clear evidence for higher heat flow in the basin areas (Horváth *et al.*, 1979; Bodri, 1981*b*; Dövényi and Horváth, 1988). The Little Hungarian Plain and the Vienna basin are cooler, and low heat flow ($<50 \text{ mW/m}^2$) values characterize the Transylvanian basin; however several regions of the Great Hungarian Plain are hot. The low heat flow in the Central Range is misleading, being caused by the infiltration of meteoritic water into the karstic rocks (Dövényi and Horváth, 1988). Moreover, calculations of Bodri and Bodri (1977) show that at the Moho level the thermal high is strongly concentrated along the NE–SW rift-like axis just below the Central Range. The fact that the greatest number of basaltic eruption centres in the TVR and the only volcanoes producing important volumes of lavas (e.g. Kabhegy, Agártető, Bondoróhegy) are situated in the Central Range also supports the view that the thermal high should be located below this region.

Considering the relatively large time span of the alkali basaltic activity as well as the young ages of many eruptions (Balogh *et al.*, 1986) and the suggestion that thermal anomalies decay in a period of 10 m.y. (Sass and Lachenbruch, 1979), it seems probable that the genesis of alkali basaltic magmas and the present-day geophysical state of the lithosphere in this area are interdependent (Embey-Isztin, 1981).

Upper mantle and lower crustal xenoliths in the TVR

Mantle- and lower crustal-derived xenoliths have been described from the following localities of the TVR:

1. Austria: the tuff of Kapfenstein near Graz (Schadler, 1913; Ross *et al.*, 1954; Kurat, 1971; Kurat *et al.*, 1976, 1980) is famous for its lherzolites, harzburgites, as well as for amphibole- and phlogopite-bearing peridotites and rare garnet websterite cumulates. Dietrich and Poultidis (1985) described small spinel lherzolite and spinel–clinopyroxene–olivine cumulate xenoliths in the nepheline basanite of Klöch, and dunite in the nephelinite of Stradner Kogel, Graz Basin, Styria. The tuff of Tobaj, S. Burgenland, contains amphibole megacrysts, spinel lherzolites and rare garnet websterite nodules (Richter, 1971). Lower crustal xenoliths have not been identified from Austrian localities.

2. Hungary: xenolith-bearing basaltic rocks have been described from: (a) Szentbékálla (Szt), (b) Bondoróhegy (Bo), Szigliget (Szg) and Gérce (G), (Fig. 1) (Embey-Isztin, 1976*b*, 1978, 1984; Embey-Isztin *et al.*, 1989). At Szigliget, Szentbékálla and Gérce xenoliths occur in basaltic tuffs, whilst at Bondoróhegy the nodules are situated in basaltic flows. All of these contain spinel lherzolites of supposed upper-mantle origin and clinopyroxenite nodules. Amphibole-bearing lherzolites are known at Szigliget (Embey-Isztin, 1976*a*). Granulite facies xenoliths have been recorded from Szigliget, Bondoróhegy and Szentbékálla (rare) and they are apparently lacking at Gérce (Embey-Isztin, 1984). With the exception of Szigliget, granulite nodules are garnet-free, whereas at Szigliget both garnet-free and garnet-bearing granulite facies xenoliths occur.

Granulite facies xenoliths

The lower continental crust is probably typically composed of plagioclase- and garnet-bearing granulite material, but generally these lithologies are only observed when brought up as xenoliths in alkali basalts (e.g. Lovering and White, 1964; Hutchison and Gass, 1971; Wilkinson, 1975; Okrusch *et al.*, 1979; Kay and Kay, 1981; Wass and Hollis, 1983; Sutherland *et al.*, 1984; Upton *et al.*, 1984; Downes and Leyreloup, 1986; Griffin and O'Reilly, 1987; Stosch, 1987).

The texture, mineralogy and whole-rock chemistry of the TVR granulite nodules are similar in many respects to xenoliths of other localities, which are widely interpreted as lower crustal

granulite facies rocks. Even details, such as accessory minerals like titanite or scapolite are identical. Scapolite was described in high-grade granulite facies xenoliths from Australia (Lovering and White, 1964) and Germany (Okrusch *et al.*, 1979). Granulite nodules of the TVR must have been brought to the surface by volcanic activity, since no rocks of such high metamorphic grade occur either in outcrop or in wells drilled into the basement in Hungary. The nodules vary in size, but are usually larger at Bondoróhegy (up to 15 cm in diameter) than at Szigliget and Szentbékálla (<10 cm).

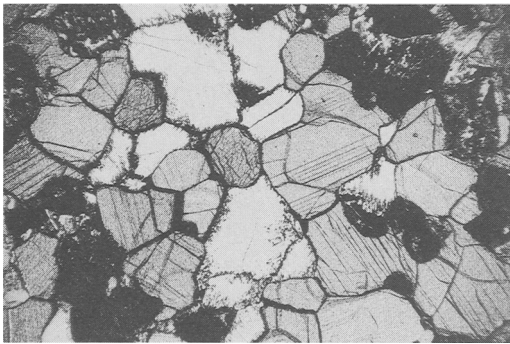


FIG. 3. Typical equilibrium texture with triple points (Bo-3006), crossed nicols.

The dominant texture type of the granulites is polygonal, granoblastic and equigranular (Fig. 3). The microstructure is equilibrated and annealed. These textures are typical for rocks from the lower crust, where recrystallization occurred under static conditions over long periods of time (Nixon and Boyd, 1973). Larger nodules may show compositional layering in mm to cm scale. Preferred orientation of grains is rare as is undulose extinction or other evidence of strain. The different minerals are optically homogeneous. The texture can perhaps be best interpreted as the result of annealing in a static environment reflecting the effect of heating. Rarely there are indications of earlier relict, probably magmatic microtextures (e.g. large plagioclase crystals poikilitically enclosing clinopyroxene crystals) suggesting an igneous protolith.

Mineralogy and texture of the garnet-free granulite facies xenoliths from Bondoróhegy, Szigliget and Szentbékálla are similar. The mineralogy is plagioclase + clinopyroxene + orthopyroxene + amphibole + aluminium spinel + ilmenite + titanomagnetite + titanite + scapolite. Titanite and scapolite are very rare, amphibole is scarce,

orthopyroxene, if present, generally appears in smaller amounts than clinopyroxene. The garnet-bearing xenoliths from Szigliget have similar mineralogies except that they contain highly variable amounts of garnet and plagioclase may become subordinate. Partial melting and subsequent quenching phenomena, probably due to falling pressures and elevated temperatures during eruption, exist in most of the nodules. These processes gave rise to a secondary paragenesis consisting of glass, plagioclase, clinopyroxene, orthopyroxene and olivine. Microscopic descriptions of the investigated xenoliths are given in Appendix A.

Bulk-rock chemistry of the granulitic xenoliths

The compositions (Table 1) of the analysed granulite-facies nodules were partly determined by means of XRF at the Institute of Petrology, University of Vienna. The ratio of Fe^{3+}/Fe^{2+} were obtained by wet chemical methods. Other samples were analysed in the Department of Mineralogy of the Natural History Museum of Budapest using entirely wet chemical methods. A few duplicate analyses have also been performed in the two laboratories. The agreement is generally good except for the ratio of Fe^{3+}/Fe^{2+} .

Most of the granulite facies xenoliths are basic in composition; hy- and ol-normative rocks prevail. Some are even strongly hy-normative (e.g. Bo-3001 and Szig-3001). A few nodules are however slightly ne-normative (e.g. Bo-3007, Szig-3004). The calculation of norms was carried out supposing a Fe_2O_3/FeO ratio of 0.15 (in mol.%) to cancel the effect of oxidation and analytical uncertainties. Plagioclase-rich specimens like Bo-3001, are c-normative. In other words, olivine tholeiitic and olivine basaltic compositions prevail. (The recalculation of the Fe^{3+}/Fe^{2+} ratio increases the degree of undersaturation, so the natural rock composition may be somewhat more shifted towards tholeiitic compositions). Also the garnet-bearing samples of Szigliget (Szig-3015, Szig-3017 and Szig-3018, Table 1) have olivine tholeiitic compositions. Again the plagioclase-rich sample of Szig-3017 show normative c. In the AFM diagram (Fig. 4D) the composition of the nodules indicates a differentiation trend with enrichment of iron similar to that of the tholeiitic rock series. The fact that most of the samples seem to have reasonable normative compositions should be emphasized, since it suggests that the protoliths represented liquid rather than cumulate compositions.

Trace element concentrations in the granulite

Table 1. Chemical composition of granulite xenoliths from Bondoróhegy, Bo ; Sziliget, Szg; and Szentbékálla, Szt; TVR, W-Hungary. x = XRF, w = wet chemical analyses.

	Bo-3001	Bo-3002	Bo-3006	Bo-3006	Bo-3007	Bo-3007	Bo-3009	Szg-3001	Szg-3003	Szg-3004	Szt-3001	Szg-3015	Szg-3017	Szg-3018
	w	w	x	w	w	x	w	w	x	x	x	w	w	w
SiO ₂	46.21	46.21	47.85	47.07	44.68	44.94	42.50	47.91	47.57	49.63	47.14	46.67	47.06	43.61
TiO ₂	1.78	1.90	0.74	0.76	1.12	1.17	2.26	1.72	1.74	1.36	0.60	0.82	2.70	2.44
Al ₂ O ₃	24.32	18.97	14.89	15.41	15.93	15.74	14.78	15.33	15.49	16.67	15.76	12.96	22.98	15.10
Fe ₂ O ₃	3.11	3.31	2.72	6.48	2.77	6.35	4.15	2.55	2.57	3.18	3.36	3.03	2.38	3.73
FeO	7.23	8.49	8.20	4.65	8.29	5.14	10.89	9.38	9.46	6.67	5.90	5.01	7.35	9.76
MnO	0.27	0.26	0.22	0.24	0.17	0.19	0.26	0.19	0.21	0.19	0.19	0.18	0.24	0.44
MgO	4.96	5.15	8.95	9.03	9.53	9.80	10.44	9.48	9.36	5.73	5.31	19.38	5.30	11.92
CaO	6.42	12.85	12.70	12.26	14.95	14.05	12.21	10.25	9.95	14.73	12.40	9.46	6.41	9.23
Na ₂ O	2.90	2.37	2.32	2.24	1.37	1.41	1.32	2.01	2.04	1.84	1.83	1.29	3.42	0.94
K ₂ O	0.35	0.18	0.18	0.29	0.04	0.24	0.24	0.27	0.45	0.39	0.03	0.00	0.44	0.04
P ₂ O ₅	0.08	0.03	0.21	0.10	0.12	0.05	0.07	0.08	0.03	0.19	0.02	0.02	0.04	0.07
H ₂ O	1.22	0.80	0.67	1.64	0.50	1.35	0.91	0.07	1.39	2.09	1.29	1.25	1.91	2.29
Total	98.85	100.42	99.65	100.12	99.47	100.48	100.03	99.24	100.42	99.19	98.85	100.13	100.23	99.57
m%	47	45	60	61	61	62	56	59	59	50	67	82	50	62
CIPW-Norms														
q	0.3	-	-	-	-	-	-	-	-	-	-	-	-	-
c	7.9	-	-	-	-	-	-	-	-	-	-	-	5.4	-
or	2.1	1.1	1.1	1.8	0.2	1.5	1.5	1.6	2.7	2.4	0.2	-	2.7	-
ab	25.3	19.4	20.0	19.3	8.0	9.4	9.4	17.3	17.6	11.7	16.0	11.1	29.6	38.4
ne	-	0.5	-	0.1	2.1	1.5	1.6	-	-	4.6	-	-	-	-
an	32.4	41.1	30.2	31.2	37.9	36.7	34.4	32.5	32.3	35.3	35.8	30.1	32.4	38.4
di	-	19.4	26.7	24.4	30.1	27.9	22.2	15.3	14.5	34.5	22.4	14.0	-	7.2
hy	26.2	-	1.2	-	-	-	-	18.3	16.2	-	7.9	14.0	21.3	25.7
ol	-	13.6	17.7	19.6	18.1	19.5	25.8	10.2	12.0	7.4	15.5	28.3	2.3	14.1
il	3.5	3.7	1.4	1.5	2.2	2.3	4.4	3.3	3.4	2.7	1.2	1.6	5.3	4.8
mt	1.1	1.2	1.1	1.1	1.2	1.2	1.6	1.2	1.3	1.0	1.0	0.8	1.0	1.4
ap	0.2	0.1	0.5	0.2	0.3	0.1	0.2	0.2	0.1	0.4	0.1	0.1	0.1	0.2

Table 2. Trace element contents of granulite facies xenoliths from the TVR, W-Hungary; contents in ppm.

	Bo-3006	Bo-3007	Szg-3003	Szg-3004	Szt-3001	Szg-3015	Szg-3017	Szg-3018	Bon-683	Bad-560
Ni	125	109	106	91	131	401	66	113	455	128
Cr	875	390	254	228	351	1100	204	166	459	136
Co	42	47	35	26	35	51	44	80	-	-
Sc	21	27	28	22	26	38	28	47	10	18
Cu	17	41	1	27	1	109	56	59	32	36
Zn	44	91	111	95	54	50	59	106	80	88
Zr	76	41	127	120	27	39	314	63	260	306
Nb	-	-	-	-	-	-	-	-	72	75
Y	24	21	34	26	10	22	32	-	25	29
V	257	360	429	318	225	-	-	-	144	187
Ba	131	131	208	157	50	75	412	212	772	886
Sr	165	131	230	310	96	56	785	18	790	902
Rb	4	3	5	4	1	-	-	-	56	64
U	-	-	-	-	-	4	7	16	-	-
Th	-	-	-	-	-	-	-	-	2	2
Pb	-	-	-	-	-	-	-	-	3	4
La	-	-	-	-	-	2	32	32	47	53
Ce	-	-	-	-	-	2	36	25	90	107
Nd	-	-	-	-	-	3	8	32	37	44

Bon = Bondoróhegy, Bad = Badacsony host alkali basalts, unpublished XRF analyses
University of Edinburgh

xenoliths (Table 2) further strengthen the evidence, gained from the major element compositions, that the xenoliths are very different from the host basalts. Granulite xenoliths are strongly depleted in incompatible elements such as Rb, Ba, K, Sr, P and Zr, whereas alkali basalts are enriched in these elements. This fact renders any genetical link between the host basalt and the granulite xenoliths very improbable. Furthermore, it is also improbable that the granulite xenoliths could have been contaminated by the host basalt owing to the very low abundances of incompatible elements in the xenoliths relative to their hosts. There are however, differences in the trace element contents of the nodules from the three localities. The Szigliget granulite-facies rocks have higher contents of K, Ba, Sr, Zr, Zn and lower contents of Ni, Cr, Co and Cu than those of Bondoróhegy. The Szentbékálla samples show the lowest concentrations of incompatible elements. Even this feature is probably primary and cannot originate from intergranular contamination by the host alkali basalt because the plagioclase in the Szigliget samples also shows significantly higher contents of potassium (Table 4).

The trace element abundances indicate that most of the samples may reasonably represent basic, tholeiitic liquid compositions. This is also suggested by the immobile element contents in the Ti-Zr-Y diagram (Fig. 4a) where most of the

xenoliths plot in the fields of low-potassium tholeiites and ocean floor basalts. In the Ti-Zr-Sr diagram (Fig. 4b) they tend to concentrate in the field of low-potassium tholeiites and in the Ti-Zr diagram (Fig. 4c) again in the fields of low potassium tholeiites and ocean floor basalts. One garnet granulite xenolith (Szg-3018) however, plots outside all of the fields in Figs. 4a,b,c due to its very low concentration in Y and high concentration of Ti relative to Zr, probably caused by the presence of a cumulus Ti-bearing phase. High Cr abundances in samples Bo-3006 and especially in Szg-3015 (1100 ppm) point to the presence of cumulus pyroxene. The high MgO and low Al₂O₃ in this latter garnetiferous sample also show that the composition is close to that of a pyroxenite. Naturally, subsequent modifications of the original trace element concentrations by metamorphism and metasomatism cannot be excluded. Perhaps, the extremely low concentrations of incompatible elements in the sample Szt-3001 may be explained by depletion during metamorphism.

Mineral chemistry of granulitic xenoliths

The minerals forming the granulite-facies nodules have been analysed by means of an electron microprobe ARL-SEMQ using standard operating conditions. Data corrections were made by the method of Bence and Albee (1968). Rep-

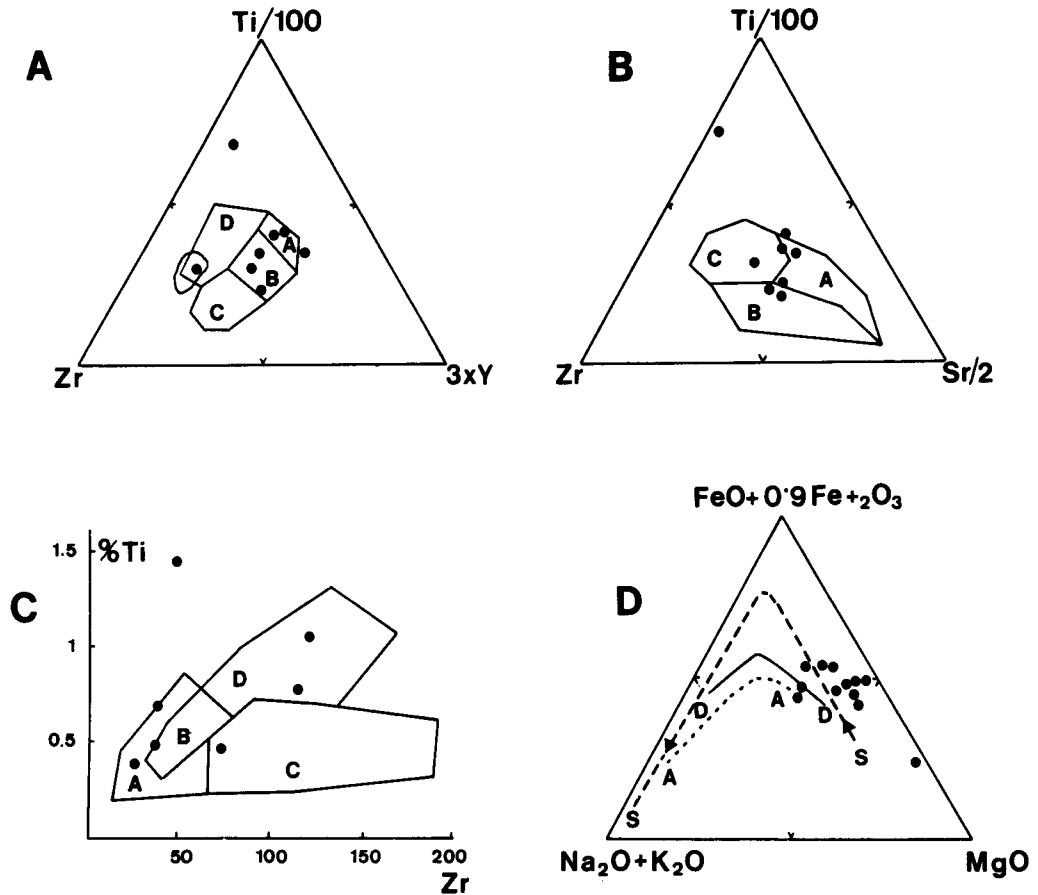


FIG. 4. (A) Ti-Zr-Y discrimination diagram (Pearce and Cann, 1973) for metaigneous mafic granulite xenoliths. Fields: A, low-K tholeiites; B, low-K tholeiites, ocean floor basalts and calc-alkaline basalts; C, calc-alkaline basalts; D, within-plate basalts. Encircled area: composition field of the TVR and Nográd District alkali basalts (unpublished XRF analyses, University of Vienna and Edinburgh). (B) ZR-Ti-Sr discrimination diagram for the same granulite xenoliths. Fields: A low-K tholeiites B calc-alkaline basalts, C ocean floor basalts. (C) Ti-Zr discrimination diagram for the granulite nodules. Fields: A low-K tholeiites, B low-K tholeiites, ocean floor basalts and calc-alkaline basalts, C calc-alkaline basalts, D ocean floor basalts. (D) AFM diagram for granulite xenoliths. Trends S-S Skaergaard, A-A alkaline basalt-trachyte series, Hawaii, D-D diabase-granophyre series, Dillsburg, Pennsylvania (Turner and Verhoogen, 1960).

representative analyses given in Tables 3–6 are averages of several measurements on the same grain and on different grains of the same phase in the sample. In general, the minerals are homogeneous with low standard deviations in individual samples. The compositions of minerals from different nodules are broadly similar.

The clinopyroxenes are salites (Fig. 5) variably enriched in FeO and always poor in Cr. The mg-value $100 \text{ Mg}/(\text{Mg}+\text{Fe})$ shows a wide range (82.6–51.2). The clinopyroxenes of the garnet-free samples have Al_2O_3 contents ranging from

3.8 to 7.4 wt.% (Table 3), one of the garnet-rich granulites (Szg-3018) falls into this range, but Szg-3017 shows however a considerably higher content of Al_2O_3 (13.61%). The $\text{Al}^{\text{IV}}/\text{Al}^{\text{VI}}$ ratio in this Al-rich clinopyroxene is 1.32, whereas this ratio is below 1 in clinopyroxenes of all the remaining samples (Fig. 6). Na_2O is below 1 wt.% in most of the pyroxenes but it is as high as 1.88% in the garnet-bearing sample Szg-3015. The amount of jadeite + tschermakite components are high in clinopyroxenes from garnet granulite sample Szg-3015 and in another sample rich in garnet

Table 3. Representative compositions of clinopyroxenes from granulite facies nodules, Transdanubian volcanic region (W-Hungary). FeO as total iron.

	Be-3006	Be-3007	Szg-3001	Szg-3003	Szg-3004	Szt-3001	Szg-3015	Szg-3017	Szg-3018
SiO ₂	49.12	48.65	50.44	48.26	48.89	49.86	50.01	45.28	49.41
TiO ₂	0.52	1.01	0.60	1.07	0.98	0.74	0.90	0.03	1.02
Al ₂ O ₃	5.74	7.43	3.79	5.82	5.89	5.13	8.75	13.61	5.79
Cr ₂ O ₃	0.12	0.04	0.08	0.03	0.06	0.04	0.09	0.06	0.04
FeO	9.87	8.31	10.02	12.71	12.16	7.90	5.44	10.59	10.48
MnO	0.19	0.11	0.16	0.14	0.18	0.14	0.13	0.19	0.22
MgO	12.88	12.36	13.68	10.14	10.62	13.88	14.50	6.23	11.51
CaO	20.50	20.90	20.64	20.45	20.24	20.58	18.39	20.86	21.21
Na ₂ O	0.91	0.86	0.60	0.79	0.80	0.97	1.88	1.19	0.77
Total	99.85	99.67	100.01	99.41	99.82	99.24	100.09	98.04	100.45

Cations based on 6 O

Si	1.824	1.806	1.675	1.834	1.845	1.849	1.807	1.735	1.841
Al ^{IV}	0.176	0.194	0.125	0.166	0.155	0.151	0.193	0.265	0.159
Al ^{VI}	0.075	0.132	0.041	0.095	0.107	0.074	0.160	0.350	0.095
Ti	0.015	0.028	0.017	0.031	0.028	0.021	0.024	0.001	0.029
Cr	0.004	0.001	0.002	0.001	0.002	0.001	0.003	0.002	0.001
Fe ³⁺	0.134	0.066	0.091	0.067	0.050	0.104	0.093	0.000	0.061
Fe ²⁺	0.173	0.192	0.220	0.337	0.334	0.141	0.071	0.339	0.265
Mn	0.006	0.003	0.005	0.005	0.006	0.004	0.004	0.006	0.007
Mg	0.713	0.684	0.758	0.574	0.597	0.767	0.781	0.356	0.639
Ca	0.816	0.831	0.822	0.833	0.818	0.818	0.712	0.857	0.847
Na	0.066	0.062	0.043	0.058	0.059	0.070	0.132	0.088	0.056
Mg/Mg+Fe	70.1	72.7	70.9	58.7	60.9	75.8	82.6	51.2	66.2
Mg/Mg+Fe ²⁺	80.5	78.1	77.5	63.0	64.1	84.5	91.7	51.2	70.7
Al ^{VI} /Al ^{IV}	0.43	0.68	0.33	0.57	0.69	0.49	0.93	1.32	0.60

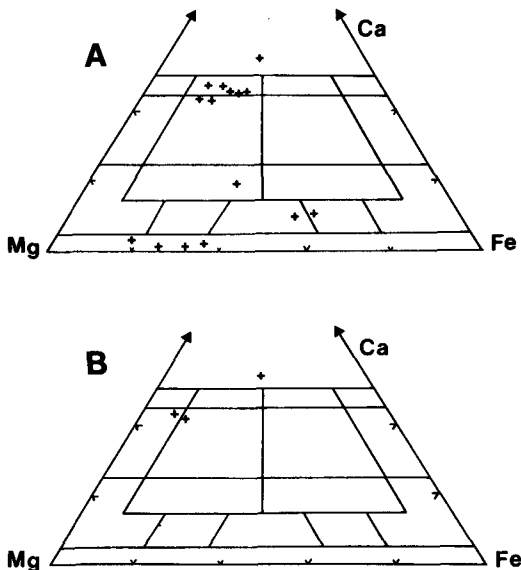


Fig. 5. (A) Composition of the Ca-rich and Ca-poor pyroxenes of granulite xenoliths in the pyroxene quadrilateral. (B) Composition of the clinopyroxenes from clinopyroxenite nodules.

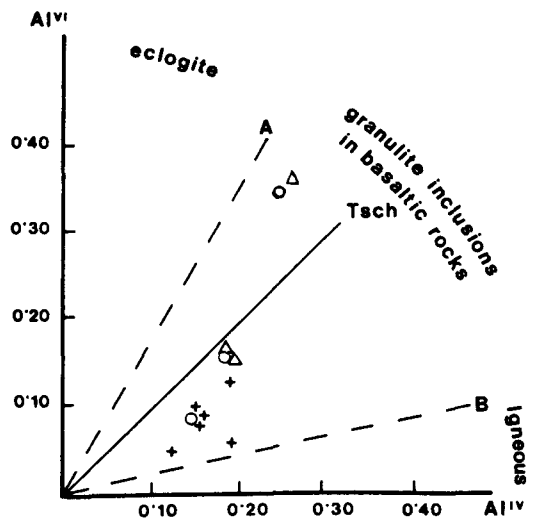


Fig. 6. Al^{VI} vs. Al^{IV} diagram for clinopyroxenes after Aoki and Kushiro (1968). Crosses, garnet-free or garnet-poor granulites; circles, garnet-rich granulites; triangles, clinopyroxenite xenoliths.

and plagioclase (Szg-3017). However the similarly garnet-rich Szg-3018, as well as other Szliget samples containing only minor amounts of garnet, are indistinguishable in this respect from the gar-

net-free nodules. These figures indicate that the rock Szg-3015 and Szg-3017 may have equilibrated at higher pressures than most of the samples and at least some of the garnet-bearing

Table 4 Representative analyses of orthopyroxenes and plagioclases from granulite xenoliths, TVR. Total Fe given as FeO.

	†								
	Bo-3006	Szg-3001	Szg-3030	Szg-3015	Szg-3017	Szt-3001	Bo-3006	Bo-3007	Szg-3001
SiO ₂	50.19	51.12	47.60	49.56	52.09	51.29	55.40	48.68	55.95
TiO ₂	.12	.22	.17	.52	.20	.18	.03	.07	.02
Al ₂ O ₃	4.11	2.32	3.80	7.80	3.79	3.36	27.48	31.82	27.16
Cr ₂ O ₃	.08	.01	.02	.06	.06	.00	-	-	-
FeO	20.24	22.62	30.00	11.52	19.85	16.18	.25	.60	.17
MnO	.33	.29	.99	.48	.25	.22	-	-	-
MgO	24.11	22.29	11.90	27.73	23.66	26.09	.03	.10	.03
CaO	.82	1.07	4.60	1.83	.61	.71	10.26	15.56	9.91
Na ₂ O	.06	.03	.03	.12	.04	.03	5.55	2.48	5.52
K ₂ O	.00	.00	.00	.09	.00	.00	.09	.05	.45
Sum	100.05	99.97	99.11	99.62	100.56	98.08	99.09	99.36	99.21
mg	68.0	63.8	41.4	81.1	68.0	74.2	An 50.3 Ab 49.2 Or .5	77.4 22.3 .3	51.8 48.9 2.6

	x				x				
	Szg-3003	Szg-3003	Szg-3004	Szg-3015	Szg-3017	Szg-3018	Szg-3018	Szg-3030	Szt-3001
SiO ₂	55.74	44.58	56.08	51.26	57.34	48.22	45.24	56.20	58.81
TiO ₂	.05	.02	.03	.11	.02	.04	.00	.02	.01
Al ₂ O ₃	27.24	34.66	27.63	30.40	26.80	32.54	34.24	28.30	26.95
FeO	.24	.87	.21	.56	.09	.45	.92	.14	.17
MgO	.01	.09	.03	.20	.01	.13	.17	.02	.03
CaO	10.46	18.68	9.43	13.46	9.07	16.34	18.90	10.50	9.53
Na ₂ O	5.04	.71	5.56	3.81	6.22	2.06	.62	4.50	6.12
K ₂ O	.51	.05	.92	.09	.34	.18	.05	.20	.05
Sum	99.29	99.64	99.87	99.87	99.87	99.96	100.13	99.90	99.67
An	51.8	93.3	45.8	65.8	43.8	80.6	94.1	55.6	46.1
Ab	45.2	6.4	48.9	33.7	54.3	18.4	5.6	43.1	53.6
Or	3.0	.3	5.3	.5	1.9	1.1	.3	1.3	.3

† = secondary, quench pigeonitic pyroxene, x = secondary, quench plagioclase

nodules as well as the garnet-free ones may have equilibrated at similar pressures.

Orthopyroxene (Table 4) has been found in five samples. They are bronzites (Szt-3001, Szg-3015) and hypersthene (Bo-3006, Szg-3001 and Szg-3017). CaO is ≤ 1 wt. % with the single exception of Szg-3015 (1.83%). This orthopyroxene coexists with a Ca-poor and Na-rich clinopyroxene (Table 3) reflecting probably an equilibration at higher temperatures.

In a few samples (Szg-3003, Szg-3018, Szg-3030) another Ca-poor pyroxene was identified which is apparently secondary in origin because it was found as small euhedral quench crystals in contact with glass and other quench phases. These pyroxenes are enriched in Fe and Mn and

their composition corresponds to that of an intermediate or ferriiferous pigeonite, however the quench pyroxene in Szg-3018 is a subcalcic augite (Fig. 5).

The compositions of plagioclases are given in Table 4. Except samples Bo-3007 and Szg-3018 which contain bytownite, all the other xenoliths carry plagioclases with intermediate compositions (andesine-labradorite). The K-content is high in the Szigliget samples (or-molecule up to 5.3 mol. %), whereas it is negligible in the Bondoró-hegy granulites. The small, secondary, quench plagioclase crystals are extremely calcic (anorthite).

The garnets (Table 5) are homogeneous. From the garnet-rich samples, the garnet (e.g. Szg-

Table 5. Representative compositions of garnets, amphiboles and olivine from granulite facies nodules, Transdanubian volcanic region (W-Hungary). FeO as total iron.

	Garnet					Amphibole		Olivine
	Szg-3001	Szg-3003	Szg-3015	Szg-3017	Szg-3018	Bo-3006	Szg-3001	Szg-3004
SiO ₂	38.76	38.64	40.77	39.79	39.00	40.73	42.05	40.75
TiO ₂	0.15	0.16	0.17	0.21	0.21	2.40	3.42	0.00
Al ₂ O ₃	22.30	21.12	22.19	22.08	22.03	13.72	12.82	0.04
Cr ₂ O ₃	0.11	0.06	0.09	0.09	0.03	0.19	0.15	0.03
FeO	20.41	24.14	13.00	21.16	22.54	13.78	12.66	9.57
MnO	0.52	0.70	0.52	0.62	1.13	0.13	0.11	0.12
NiO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.38
MgO	9.64	6.24	17.58	10.72	7.16	13.05	12.67	49.04
CaO	6.71	8.82	5.41	5.53	8.65	10.66	11.13	0.10
Na ₂ O	n.d.	n.d.	n.d.	n.d.	n.d.	2.80	2.11	n.d.
K ₂ O	n.d.	n.d.	n.d.	n.d.	n.d.	0.07	1.09	n.d.
Total	98.60	99.88	99.73	100.20	100.75	97.53	96.21	100.03
Cations based on 12 O						23 O	4 O	
Si	2.975	2.994	2.958	2.997	2.976	6.051	6.187	0.999
Al IV	0.025	0.006	0.042	0.003	0.024	1.949	1.813	0.001
Al VI	1.991	1.923	1.855	1.957	1.957	0.453	0.410	0.000
Ti	0.009	0.009	0.009	0.012	0.012	0.268	0.379	0.000
Cr	0.007	0.004	0.005	0.005	0.002	0.022	0.017	0.001
Fe 3+	0.000	0.064	0.138	0.024	0.032	n.d.	n.d.	n.d.
Fe 2+	1.310	1.501	0.651	1.309	1.407	1.712	1.557	0.196
Mn	0.034	0.046	0.032	0.040	0.073	0.016	0.014	0.002
Ni	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.007
Mg	1.103	0.721	1.901	1.204	0.815	2.890	2.779	1.732
Ca	0.552	0.732	0.421	0.446	0.707	1.697	1.775	0.003
Na	n.d.	n.d.	n.d.	n.d.	n.d.	0.807	0.602	n.d.
K	n.d.	n.d.	n.d.	n.d.	n.d.	0.013	0.205	n.d.
Alm	43.7	50.0	21.7	43.7	46.9			Fo 90.1
Sps	1.1	1.5	1.1	1.3	2.4			
Pyr	36.8	24.0	63.3	40.1	27.1			
Grs	17.6	20.6	6.4	12.6	21.3			
And+Uva	0.7	3.8	7.6	2.1	2.3			

3015) is dominated by the pyrope molecule, whereas that of Szg-3018 is almandine-rich similar to the garnets of Szg-3001 and Szg-3003.

Amphiboles (Table 5) have only been detected in two samples (Bo-3006, Szg-3001) with restricted modal contents. They form texturally equilibrated grains. The amphiboles are magnesiohastingsites (Leake, 1978) and have broadly similar compositions. However, just like the coexisting plagioclase, the amphibole in the Szigliget sample is considerably richer in potassium.

Olivine of primary origin has not been found in the granulite samples; however secondary, quench olivine crystals (Table 5) are occasionally present. Just like the quench pyroxenes, they are enriched in Fe and Mn. Ca is also high, reflecting a high *T*, low *P* event, perhaps related to eruption.

The composition of spinels has been measured in five rock samples (Table 6). All are aluminous forming solid solutions of hercynite (41–57%) and spinel (27–49%) with smaller amounts of magnetite (8–10%) and ulvöspinel (2–5%). Additionally, other members of the spinel-group can coexist with the Al-spinel e.g. titanomagnetite with 66 and 72 mol.% ulvöspinel occurring in Szg-3004 and Bo-3007 respectively, Ni-magnetite with 14% trevorite in the sample Bo-3006. Ilmenite (Table 6) was found only in the Szigliget samples.

They contain 9–23 mol.% geikielite and 1–1.5% pyrophanite component. The composition of sphene in Szg-3003 was also determined (Table 6).

Comparison with other granulite samples

Petrographically the TVR granulite xenoliths are quite different from exposed granulite facies rocks of the Variscan collision belt, e.g. Moldanubia (Scharbert, 1971; Scharbert and Kurat, 1974). For example, the ubiquitous equilibrium texture in the xenoliths are absent from Variscan granulites and the latter rocks contain a high proportion of felsic varieties, whereas, the granulite nodules are invariably mafic. However, xenoliths with similar textures, mineralogy and chemistry described in this work are known and interpreted as lower crustal rocks from Western Europe, e.g. Massif Central (Dupuy *et al.*, 1977; Dostal *et al.*, 1980; Downes and Leyreloup, 1986); Eifel (Okrusch *et al.*, 1979; Stosch *et al.*, 1986; Stosch, 1987); also from other continents, e.g. South East Australia (Lovering and White, 1964; Wilkinson, 1975; Wass and Hollis, 1983; Griffin and O'Reilly, 1986); Kilbourne Hole, New Mexico (Padovani and Carter, 1977). Though mafic granulites seem to be the major constituent of the lower crust,

petrographical details may be different in samples originating from different regions of the world. Thus, the microstructural types of the lower crustal samples show a broad correlation with tectonic setting (Griffin and O'Reilly, 1987); however, even samples of the same tectonic regime (e.g. continental rifts) exhibit varying petrographical

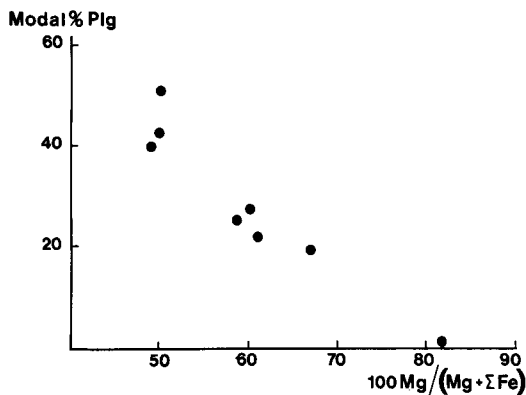


Fig. 7. Mg-value [$100 \text{ mg}/(\text{Mg} + \Sigma \text{Fe})$] vs. modal plagioclase content in granulite nodules of the TVR.

features from one place to another. For example, 40% of the granulite nodules from Massif Central are metasedimentary and the meta-igneous rocks have a calc-alkaline affinity (Downes and Leyrelop, 1986), whereas no metasedimentary granulite xenolith was found in the TVR and the metaigneous samples have low potassium tholeiitic affinities. While in our samples the mg-value is negatively correlated with the modal abundance of plagioclase (or positively with that of pyroxene) (Fig. 7) suggesting a common origin for the rocks, Stosch (1987) reported exactly the opposite, i.e. in the Eifel the plagioclase-rich xenoliths have the highest and the mafic nodules the lowest mg-numbers.

Pressure-temperature estimates

The chemical homogeneity of the constituent phases and the microstructure suggest that equilibrium has been reached among the coexisting minerals. Even so, the application of published geothermometers and barometers should be considered with caution. The pyroxene solvus thermometers are insensitive to composition below 900°C and the unknown ratio of $\text{Fe}^{3+}/\text{Fe}^{2+}$ in pyroxene may cause serious errors in the estimate of temperature with the garnet-pyroxene ther-

mometers. In addition, application of different methods commonly yield widely divergent results.

For the two-pyroxene plagioclase granulites the use of the thermometers of Wood and Banno (1973) and that of Wells (1977) gave, however, comparable results (in the range $932\text{--}947$ and $946\text{--}968^\circ\text{C}$ respectively). The respective values are higher for the two-pyroxene garnet granulite (Szg-1015) (1051 and 1035°C). The application of the method of Ellis and Green (1979) based on the Fe-Mg exchange between garnet and clinopyroxene resulted in an unreasonably high T (1245°C , assuming $P = 11$ kbar) for this sample. Another garnet-rich sample (Szg-1018) yielded a T of 1012°C at $P = 10$ kbar with the same formulation.

Phase constraints allow the bracketing of the pressure of equilibration. The granulite xenoliths with the orthopyroxene-clinopyroxene-spinel assemblage must have been formed or subsequently equilibrated on the higher pressure or lower temperature side of the reaction olivine + calcic plagioclase \rightleftharpoons orthopyroxene + clinopyroxene + spinel. The reaction takes place at about 7 kbar at 900°C and at 7.5–8 kbar at 1000°C (Herzberg, 1978). The absence of garnet gives a further constraint on the pressure; two-pyroxene granulite transforms to garnet granulite at about 9 kbar at 900°C and at about 10 kbar at 1000°C (Kornprobst, 1970; Irving, 1974). In addition, the application of the garnet-orthopyroxene barometer of Harley and Green (1982) for the sample Szg-3015 yielded a pressure of 11 kbar at 1050°C in good agreement with the phase constraint. The clinopyroxene of Szg-3015 shows a much higher proportion of the jadeite molecule than the other samples. However, there seems to be no systematic correlation between the jadeite component in the clinopyroxenes and the $\text{An}/(\text{An} + \text{Ab})$ ratio in the coexisting plagioclases as well as the pyrope content of garnets. This suggests that the xenoliths represent rather more modally and chemically different rock samples than chemically more or less homogeneous granulites with different metamorphic grade.

It is an important question whether the inferred P , T values (7–8 kbar and about 950°C for the two-pyroxene plagioclase granulites and about 11 kbar and 1050°C for the garnet granulite sample) represent peak values of the granulite facies metamorphism (fossil geotherm) or whether they characterize ambient conditions of the lithosphere at the time of entrainment of the xenoliths in the host magma. Unfortunately, no isotopic age determination is available but by analogy to some granulites of the Variscan fold-belt a Hercynian age of the metamorphism is probably likely. The

Table 6. Chemical composition of oxide minerals (spinel, ilmenite) and sphene from granulite facies nodules, Transdanubian volcanic region (W-Hungary). Fe2O3 calculated after the formula R 2+ R2 3+ O4 (spinel) and R R O3 (ilmenite).

	Bo-3006 (Sp)	Bo-3007 (Sp)	Bo-3007 (Sp)	Szg-3003 (Sp)	Szg-3004 (Sp)	Szg-3004 (Sp)	Szg-3004 (Sp)	Szg-3018 (Sp)
TiO2	0.10	0.98	26.88	2.50	1.37	0.37	26.81	1.82
Al2O3	1.16	51.22	3.20	48.56	53.29	52.78	7.31	55.81
Cr2O3	0.00	0.23	0.93	0.64	0.71	13.06	0.34	0.32
Fe2O3	66.18	12.06	15.19	9.68	8.74	3.59	11.58	5.34
FeO	26.02	20.12	49.51	31.70	26.85	11.04	49.75	27.52
MnO	0.07	0.33	0.76	0.32	0.40	0.15	0.73	0.51
ZnO	0.00	0.13	0.13	n.d.	0.10	0.08	0.00	n.d.
NiO	4.40	0.09	0.07	n.d.	0.03	0.38	0.05	n.d.
MgO	0.18	12.25	4.00	6.29	9.44	19.06	4.49	9.59
CaO	0.06	0.68	0.21	0.11	0.05	0.01	0.16	0.02
Total	98.17	98.09	100.88	99.80	100.98	100.52	101.22	100.93
Cations based on 4 O								
Ti	0.003	0.021	0.717	0.055	0.029	0.007	0.696	0.038
Al	0.053	1.698	0.134	1.664	1.745	1.642	0.297	1.807
Cr	0.000	0.005	0.026	0.015	0.016	0.273	0.009	0.007
Fe 3+	1.942	0.255	0.406	0.212	0.183	0.071	0.301	0.110
Fe 2+	0.848	0.473	1.469	0.771	0.624	0.244	1.436	0.632
Mn	0.002	0.008	0.023	0.008	0.009	0.003	0.021	0.012
Zn	0.000	0.003	0.003	n.d.	0.002	0.002	0.000	n.d.
Ni	0.138	0.002	0.002	n.d.	0.001	0.008	0.001	n.d.
Mg	0.010	0.514	0.212	0.273	0.391	0.750	0.231	0.393
Ca	0.003	0.020	0.008	0.003	0.001	0.000	0.006	0.001
Szg-3001 (Ilm) Szg-3003 (Ilm) Szg-3004 (Ilm) Szg-3017 (Ilm) Szg-3018 (Ilm) Szg-3003 (Sphene)								
SiO2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	29.69	
TiO2	52.07	53.26	51.88	53.53	53.25	53.25	38.55	
Al2O3	0.45	0.21	0.60	0.74	0.33	0.33	1.62	
Cr2O3	0.13	0.00	0.07	0.16	0.00	0.00	0.00	
Fe2O3	6.01	0.17	4.77	1.01	0.66	0.66	n.d.	
FeO	35.62	42.57	38.39	39.71	41.24	41.24	1.26	
MnO	0.60	0.77	0.71	0.43	0.22	0.22	0.05	
MgO	5.93	2.38	4.18	4.48	2.97	2.97	0.05	
CaO	0.10	0.24	0.04	0.01	0.10	0.10	28.26	
Total	100.91	99.60	100.69	100.07	99.77	99.77	99.48	
Cations based on 3 O								
Si	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.979	
Ti	0.938	0.995	0.947	0.979	0.989	0.989	0.956	
Al	0.019	0.006	0.017	0.021	0.010	0.010	0.063	
Cr	0.002	0.000	0.001	0.003	0.000	0.000	0.000	
Fe 3+	0.108	0.003	0.087	0.018	0.012	0.012	n.d.	
Fe 2+	0.713	0.885	0.779	0.807	0.852	0.852	0.035	
Mn	0.012	0.016	0.015	0.009	0.026	0.026	0.001	
Mg	0.212	0.088	0.151	0.162	0.109	0.109	0.002	
Ca	0.003	0.006	0.001	0.000	0.003	0.003	0.998	

crystallization age of the igneous protolith could be considerably older; e.g. this was estimated to be 1500 m.y. (maximum age) for the Eifel metaigneous mafic granulites (Stosch and Lugmair, 1984) and 800–2100 m.y. (maximum age) for similar xenoliths from the Massif Central (Downes and Leyreloup, 1986). Thus, it is probably more likely that our *P*, *T* values reflect recent re-equilibration. These temperatures are unusually high for granulites and may be connected with the heat anomaly beneath the Central Range to be discussed later. The temperature values cannot be considered as frozen in at a given *T* of the continuously cooling lithosphere because of the lack of zoning and in addition, calculations suggest that

diffusion rates at *T* > 700°C are much faster than the cooling of the lithosphere after a heating episode (Wilson and Smith, 1984; Smith and Ehrenberg, 1984). The measured high heat flow values in the Pannonian area are not inconsistent with the assumption that the actual geotherm and the one deduced from the *P*, *T* values of the granulite as well as spinel peridotite xenoliths (Embey-Isztin *et al.*, 1989) cannot be very different from each other.

Clinopyroxenite xenoliths

In this category of xenoliths there are clinopyroxenites (*sensu stricto*) as well as olivine clinopyr-

Table 7. Representative bulk-rock and mineral analyses of clinopyroxenite nodules, TVR. cpx = clinopyroxene, ol = olivine, sp = spinel, mt = magnetite. Total Fe given as FeO. Trace elements are given in ppm.

	Bo-3018	Bo-3020	Szt-2001	Szt-2001	Bo-3018	Szt-2001	Bo-3018	Bo-3020	Szt-2001	Bo-3018	Bo-3018
			cpx-rich	ol-rich	ol	ol	cpx	cpx	cpx	sp	mt
SiO ₂	44.54	42.66	48.44	42.77	40.12	39.83	50.06	45.94	49.79	nd	nd
TiO ₂	0.24	1.88	1.04	0.39	0.03	0.03	1.11	0.27	1.13	0.33	0.08
Al ₂ O ₃	4.80	15.53	7.86	3.45	0.08	0.05	8.28	13.58	7.96	60.27	1.43
Cr ₂ O ₃	nd	nd	nd	nd	0.00	0.01	0.24	0.00	0.15	3.72	0.09
FeO	7.16	6.19	3.18	8.92	12.39	14.09	4.40	10.58	5.26	13.33	89.16
MnO	0.15	0.23	0.12	0.16	0.16	0.18	0.12	0.19	0.12	0.10	0.14
NiO	nd	nd	nd	nd	nd	0.12	nd	nd	nd	nd	nd
MgO	35.03	8.21	14.31	33.80	47.27	45.74	14.94	6.44	14.97	21.44	0.96
CaO	3.60	18.29	18.56	6.45	0.13	0.14	19.11	20.72	18.86	0.01	0.00
Na ₂ O	0.48	1.15	1.16	0.37	nd	nd	1.43	1.22	1.27	nd	nd
K ₂ O	0.10	0.00	0.02	0.02	nd	nd	nd	nd	nd	nd	nd
P ₂ O ₅	0.04	0.04	0.02	0.04	nd	nd	nd	nd	nd	nd	nd
H ₂ O	0.00	1.59	1.63	0.00	nd	nd	nd	nd	nd	nd	nd
Total	98.51	100.48	99.19	98.93	100.18	100.19	99.69	98.94	99.51	99.21	91.92
mg	85.9	58.9	81.3	86.4	87.2	85.3	85.8	52.1	83.6	-	-
Ni	1496	112	205	826							
Cr	2770	347	832	407							
Co	97	43	25	105							
Cu	1	4	9	1							
Sc	7	42	23	12							
Zn	56	122	14	56							
V	88	207	295	122							
Y	3	30	13	7							
Ba	21	152	44	30							
Rb	1	0	6	3							
Sr	40	50	103	45							
Zr	23	103	50	29							
U	nd	4	nd	nd							
La	nd	23	nd	nd							
Ce	nd	19	nd	nd							
Nd	nd	25	nd	nd							

oxenites and wehrlites. They are less abundant at Szigliget and Bondoróhegy than the granulites and much less abundant than the Type I Cr-diopside lherzolites. The mineralogy is simple (clinopyroxene + olivine + spinel + magnetite + apatite). The texture is unequilibrated, coarse (from 1–2 mm up to 1–2 cm), often xenomorphic granular; however poikilitic fabrics with olivine + spinel sub- and euhedra enclosed in large clinopyroxene grains also occur. The textures can be interpreted as igneous, probably cumulate in origin, although some samples show partial grain boundary adjustment suggesting an incipient stage of metamorphism. The description of the analysed samples can be found in the Appendix B.

It is evident that such clinopyroxene-rich rocks cannot possibly represent liquid compositions and

therefore they may be of cumulate nature. The composition of the wehrlite (Bo-3018) (Table 7) is compatible with an origin by crystal accumulation from a relatively undifferentiated basaltic liquid, due to its high abundance of compatible elements such as Ni and Cr and the low concentration of the highly incompatible elements e.g. Rb or Ba. Since a meaningful bulk-rock analysis could not be performed on the sample Szt-2001, clinopyroxene- and olivine-rich portions were analysed separately. The smaller, but still considerable Ni and Cr abundances and higher concentrations of incompatible elements as well as the lower mg-value of constituent minerals (Table 7) suggest that this rock could possibly be a cumulate segregated from a differentiated basaltic liquid. However, the Bo-3020 clinopyroxenite is highly enriched in Fe and has low Cr and relatively

high Sr, Ba, Zr and light *REE* abundances; consequently it could be in chemical equilibrium only with an extremely evolved liquid.

Textural and chemical evidence unanimously suggests that the clinopyroxenite and wehrlite nodules of the TVR correspond to those described from worldwide localities such as Type II, Al-augite group, black pyroxene suite xenoliths (e.g. Wilshire and Shervais, 1975; Frey and Prinz, 1978). Such xenoliths are generally considered to be high-pressure, high-temperature cumulates (characteristic of upper-mantle conditions) segregated either from the host alkali basalt or from a similar liquid of an earlier magma batch (e.g. Irving, 1980). Type II xenoliths from the TVR do not have appropriate mineralogies for thermobarometry; however several lines of evidence point to upper-mantle conditions. The most important are: (A) the presence of composite xenoliths in which spinel lherzolite is in planar contact with Type II pyroxene and/or amphibole-rich material (Embey-Isztin, 1976a; 1984; Embey-Isztin *et al.*, 1989); (B) the composition of the clinopyroxenes shows a higher proportion of the high-*P* end-molecules, such as jadeite and Ca-Tschermak molecule, than pyroxenes of the pyroxene-plagioclase granulites; (C) clinopyroxenes with similar compositions to ours have been shown to crystallize from alkali basalts as high-*P, T* liquidus or near liquidus phases (e.g. Bultitude and Green, 1971).

Discussion

Both texturally and chemically, the two kinds of xenoliths (granulites and clinopyroxenites) are so different from each other that they must reflect contrasting genetical histories. In spite of the fact that the pyroxene-plagioclase + minor garnet granulite nodules may represent basaltic liquid compositions, they cannot be cogenetic even 'in sensu lato' with the alkali basalt host rocks because of the important differences in major- and minor-element whole-rock chemistry. These granulites are metagabbros of olivine tholeiitic affinity with very low LIL-element contents. The metamorphic microstructure, the complex crystallization history of these mafic granulites deduced from the textural evidence in itself suggests that the xenoliths are accidental and of much greater age than the host basalts. It is also clear that the granulites are unrelated to the older (Miocene) andesitic volcanism of the Pannonian area as well; this latter shows typical calc-alkaline major and trace element characteristics (Pantó *et al.*, 1988) and conversely, no corresponding tholeiitic magmatism is known in this area. We think

it probable that these granulites are pieces of an ancient crust which by analogy with the Eifel and Massif Central samples, may have metamorphosed in Hercynian and formed in even older times. As to their actual position, three lines of evidence indicate that they may represent a major constituent of the present-day lower crust. These are: (A) measured compressional seismic velocities in the lower crust beneath the TVR ($V_p = 7.1$ to 7.3 km/sec) (Posgay, 1975) correspond well with mafic granulites poor in garnet (Ringwood, 1975); (B) no mafic granulite was found in planar contact with mantle-derived spinel lherzolite; (C) in Hungary, no basic granulites occur either in outcrop or in well drilled into the basement.

The type II clinopyroxenite xenoliths have compositions compatible with a crystallization from the host alkali basalt or a similar magma at high pressure. The undeformed microstructures also suggest that the pyroxenites cannot be significantly older than the host basalt. However, the variable microstructure and the presence of incipient metamorphism in some of the samples attest to a complex evolution. Rocks showing grain boundary adjustment could have been formed at an earlier phase of the Pliocene alkali basaltic volcanic cycle. The existence of Type II/Type I clinopyroxenite/lherzolite and amphibolite/lherzolite nodules (Embey-Isztin, 1976a; Embey-Isztin *et al.*, 1989) is important evidence that the upper mantle is veined beneath the Central Range. These composite xenoliths suggest that the lava flows are not the only products of the volcanic activity; volumetrically important bodies (veins, dykes and sills) of frozen liquids and cumulates may be especially abundant in the uppermost mantle and in the lowermost crust. This underplating process is probably controlled by the contrast in density between the crust and the mafic magmas and by rheological contrasts between the crust and mantle. We suggest that the disappearance of the Moho below the Central Range (Figs. 2, 8) and the 5 km thick seismic crust-mantle transitional zone (Mituch and Posgay, 1972) are related to a complex interlayering of mafic (Type II) and ultramafic (Type I) rocks (Fig. 9) as postulated in other continental areas characterized by high heat flow, e.g. Eastern Australia (O'Reilly and Griffin, 1987) and Spitzbergen (Amundsen *et al.*, 1987).

It has been suggested that a thermal high and perturbation develops in the lithosphere beneath such rift zones (Sass and Lachenbruch, 1979) and the underplating process (dyking) is largely responsible for the heat input (convective heat transport) (e.g. O'Reilly and Griffin, 1985, 1987). It is thus probable that the high temperatures of

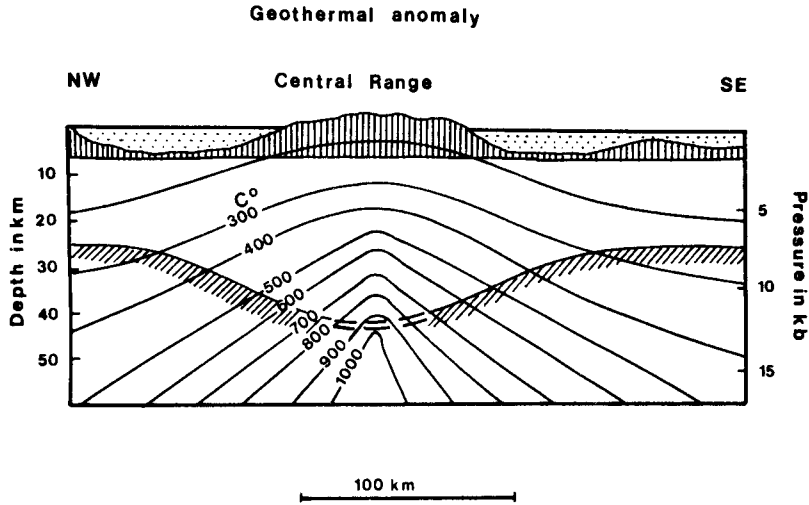


Fig. 8. A non-quantitative thermal model of the Central Range and the adjacent basin areas based on the assumption of a thermal high caused by dyking in the crust-mantle transition zone (advective heat transfer).

the TVR granulite samples obtained by the two-pyroxene methods (930–960°C at 7–8 kbar and 1050°C at 11 kbar) reflect ambient conditions at the time of entrainment of xenoliths by the host magma. These P , T values would fit a steeper than oceanic geotherm like that of Kilbourne Hole, New Mexico (Padovani and Carter, 1977) or Spitzbergen (Amundsen *et al.*, 1987), both of which were interpreted as reflecting ambient P , T conditions of the lithosphere. However, this model of convective heat transfer related to dyking implies that the thermal high in the lower crust beneath the volcanic region of the Central Range should rapidly attenuate with increasing distance in lateral direction since geologically, the total volume of the volcanic products is insignificant (Fig. 8). This suggestion accords with calculations by Bodri and Bodri (1977), who demonstrated a very pronounced heat flow and temperature anomaly at the Moho level within the NE–SW striking rift axis of the Central Range. They also showed that this anomaly diminishes rapidly away from the axis.

In the light of the above thermal model (Fig. 8) we can probably explain the absence of garnet in the Bondoróhegy granulite samples and their presence in some of the Szigliget xenoliths. Bondoróhegy is situated where the volcanic activity was particularly intensive. By far the largest volumes of lavas were produced around this locality compared with the whole of the TVR. In contrast, Szigliget is located rather more in the border zone

of the TVR, in a region characterized by volcanic vents producing tuffs and only insignificant quantities of lavas. It thus seems probable that higher ambient temperatures in the lower crust beneath Bondoróhegy may have suppressed the formation of garnet, whereas below Szigliget, the lower crust being somewhat cooler, the formation of garnet became possible as indicated by the experimental work of Green and Ringwood (1967). If this thermal model is correct, we can also suppose that garnet granulite or even eclogite may become progressively important components in the lower crust with increasing distance away from the thermal high.

Geologists and geophysicists widely believe that the Moho is a stratigraphic boundary with ultramafic rocks below and mafic rocks above it. In their study on the xenoliths of the Bullenmerri and Gnotuk maars (SE Australia) Griffin *et al.* (1984) have shown however, that this is not always true. They demonstrated that advective heat transfer related to dyking can effectively move the seismic Moho downwards due to the decrease of V_p with increasing T and in some regions depending on the thermal conditions, the seismic Moho is not related to lithological changes but to phase transformation. In other words, the petrological and the seismic Moho can decouple. We think that the application of this model to the TVR and to the Pannonian area in general can successfully explain relationships between lithological (xenolithic) and geophysical obser-

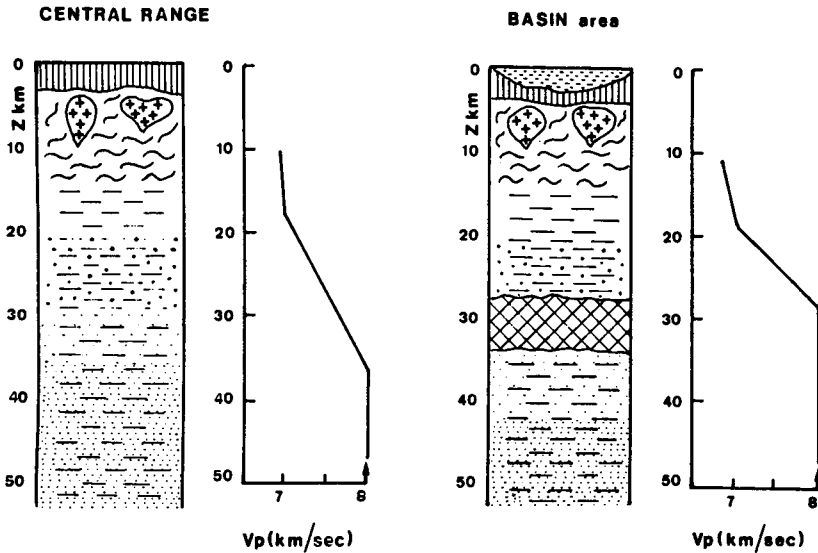


FIG. 9. Simplified xenolith based models suggested for the composition of the lower crust and upper mantle beneath the Central Range (left) and adjacent basin area (right). Dark lines: mafic dykes, cross-hatched area: garnet pyroxenite heavy stippling: ultramafic rocks. The seismic profile is adopted from Mituch and Posgay (1972); Posgay (1975).

variations. Fig. 9 shows such an attempt to demonstrate the differences in the lithosphere beneath the Transdanubian Central Range and adjacent basin areas. Below the Central Range the seismic Moho lies 8–10 km deeper than beneath the basins and it is not sharp, but transitional (Mituch and Posgay, 1972). This can be explained by the formation of mafic dykes (evidenced by Type II xenoliths) at the crust–mantle interface. In the transitional zone mafic and ultramafic lithologies occur together and the $V_p = 8.0$ km/sec is reached where the ultramafic rocks become overwhelming. However, the thermal high can make the seismic Moho move downwards so that the exact location of the seismic Moho is the combined result of these two effects. In addition, the thermal high suppresses the formation of garnet in the granulites of the lower crust. In contrast, beneath the adjacent basin areas lower crustal temperatures are supposed to be lower (Fig. 8) and as indicated by the presence of garnet in some of the Szigliget granulite nodules, the lower crust may be composed of garnet granulites (Fig. 9). This can dramatically raise the position of the seismic Moho. Consequently, it seems probable that the 8 to 10 km difference in the position of the seismic Moho does not mean that the thickness of the mafic lower crust would be considerably different below these regions.

If we abandon the long-lived idea of a thinner

mafic lower crust below the basins we can better explain some of the other geophysical controversies mentioned earlier. In particular the controversy of the negative Bouguer anomaly in the basins and the positive one in the Central Range emphasized by Meskó (1982) becomes more understandable. The larger Bouguer anomaly in the Central Range may perhaps be related to the upwelling of a smaller hot mantle diapir densely interlayered by mafic rocks. Finally, it is also possible that the active volcanism and the underplating of the lower crust by basic magmas is responsible for the uplift of the Central Range since this has been considered to be a main force in epi-orogenic uplift in general, whereas the surrounding areas underwent subsidence.

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Appendix A:

Petrographic descriptions of analysed granulitic samples

Bo-3006: Granoblastic texture with straight line grain boundaries and 120 triple-point junctions (Fig. 3). Average grain size is 0.7 mm. Coarser parts with curvilinear grain boundaries may perhaps indicate relics of magmatic crystallization. Modal composition: clinopyroxene 66%, plagioclase 25%, orthopyroxene 7%, amphibole 2%, magnetite 0.5%.

Bo-3007: Similar texture to Bo-3006 but somewhat coarser-grained. Mode: clinopyroxene 70%, plagioclase 22%, Al-spinel 7%. Melted parts (1%) are composed of dark brown glass and quench crystallites of anorthite, pigeonite and Fe-rich olivine.

Szg-3001: Equilibrium texture. Average grain size is 0.6 mm. Contains a few grains of pleochroic orthopyroxene, brown amphibole and also some garnet and ilmenite. Mode: clinopyroxene 73%, plagioclase 25%, accessories 2%.

Szg-3003: Similar texture to Szg-3001; however a few large (3–4 mm) relict clinopyroxenes can also be seen. Modal composition: clinopyroxene 53%, plagioclase 39%, spinel 4%, ilmenite 2%, garnet 2% and a few crystals of titanite. Anorthite and pigeonite crystallites are quench phases.

Szg-3015: Medium-grained (1–2 mm), equilibrium and slightly foliated texture. Modal composition: clinopyroxene 65%, orthopyroxene 15%, garnet 18%, plagioclase 2%.

Szg-3017: Coarse granoblastic with irregular grain outlines. Rich in plagioclase (49%) and garnet (40%) and very poor in clinopyroxene (1%). Orthopyroxene and ilmenite comprise 5% each.

Szg-3018: Medium-grained granoblastic texture partly with straight-line, partly with curvilinear grain boundaries. Strongly affected by secondary melting (15%) and subsequent quenching. Mode is dominated by clinopyroxene (55%) and garnet (20%) with subordinate amount of plagioclase (6%) and spinel (4%). Quench phases are: subcalcic augite and anorthite.

Szg-3030: Typical fine-grained equilibrium texture, but affected by melting. Clinopyroxene dominates the mode (70%), plagioclase (25%) with some Al-spinel. The melted areas (5%) are composed of glass and quench plagioclase, pigeonite and Fe-rich olivine.

Szt-3001: Fine to medium grained equilibrium texture with modal layering. Clinopyroxene is the dominant phase (80%), plagioclase (19%), plus a few grains of orthopyroxene (1%).

Appendix B:

Petrographic descriptions of analysed pyroxenitic samples

Bo-3018: Wehrlite composed of olivine (58%), clinopyroxene (39%), Al-rich spinel (2%) and traces of plagioclase (labradorite) (1%). Coarse-grained (up to 5 mm), curvilinear grain boundaries prevail. Addition-

nally, there appear small rounded and sometimes subautomorphic crystals. Olivine is generally small and xenomorphic, and spinel appears interstitially.

Szt-2001: A reasonable modal estimate cannot be proposed since this nodule has two contrasting parts: one is rich in olivine with subordinate clinopyroxene; the other is clinopyroxene-dominated and poor in olivine. Phase chemistry is however identical in both parts. It is probable that this sample comes from a layered cumulate igneous body. The texture is coarse-grained (up to 1 cm), curvilinear grain boundaries prevail with incipient polygonization.

Bo-3020: Black clinopyroxenite consisting almost entirely of clinopyroxene plus subordinate oxides and apatite. Coarse-grained (up to 1–1.5 cm) with rounded clinopyroxene crystals.

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