Contrib Mineral Petrol (1983) 84:1–5

Contributions to Mineralogy and Petrology

Density Constraints on the Formation of the Continental Moho and Crust

C.T. Herzberg[1], W.S. Fyfe[2], and M.J. Carr[1]

[1] Department of Geological Sciences, Rutgers University, New Brunswick, New Jersey, USA 08903
[2] Department of Geology, University of Western Ontario, London, Ontario, Canada N6A 5B7

Abstract. The densities of mantle magmas such as MORB-like tholeiites, picrites, and komatiites at 10 kilobars are greater than densities for diorites, quartz diorites, granodiorites, and granites which dominate the continental crust. Because of these density relations primary magmas from the mantle will tend to underplate the base of the continental crust. Magmas ranging in composition from tholeiites which are more evolved than MORB to andesite can have densities which are less than rocks of the continental crust at 10 kilobars, particularly if they have high water contents. The continental crust can thus be a density filter through which only evolved magmas containing H_2O may pass. This explains why primary magmas from the mantle such as the picrites are so rare. Both the over-accretion (i.e., Moho penetration) and the under-accretion (i.e., Moho underplating) of magmas can readily explain complexities in the lithological characteristics of the continental Moho and lower crust. Underplating of the continental crust by dense magmas may perturb the geotherm to values which are characteristic of those in granulite to greenschist facies metamorphic sequences in orogenic belts. An Archean continental crust floating on top of a magma flood or ocean of tholeiite to komatiite could have undergone a major cleansing process; dense blocks of peridotite, greenstone, and high density sediments such as iron formation could have been returned to the mantle, granites sweated to high crustal levels, and a high grade felsic basement residue established.

Introduction

Despite our inability to directly observe the inaccessible regions of the lower continental crust, there is good evidence that it differs compositionally from crust near the surface. This has arisen from heat flow constraints which require the upper crust to be enriched in the LIL elements such as U, Th, and K relative to deeper levels (Heier 1973). A number of simple two-layer geometries have thus emerged, granodiorite being an average composition of the uppermost part (Taylor 1979) and granulite facies rocks of uncertain composition below. The formation of granulites by dehydration and partial melting processes is likely to be important for transporting the LIL elements from the lower to the upper crust (e.g., Fyfe 1973; Heier 1973), and could be compatible with a lower crust ranging from an intermediate composition favoured by Tarney and Windley (1979) to more basic compositions than the basaltic-andesite preferred by Taylor (1979). Uncertainties in providing a more exact composition stem from obvious sampling problems and from the model-dependent nature of mass balance calculations which proportion a granodiorite upper layer to various assumed bulk crust compositions.

Exposures of obducted portions of the upper and lower continental crust are now thought to exist in cross-section at a number of sites on the Earth's surface; these have been reviewed in detail by Fountain and Salisbury (1981). An important outcome of the raw geological data is that it illustrates how overly simple is the above two-layer model for continental structure. Superimposed on prominent metamorphic facies changes is an extensive vertical and lateral compositional variability in addition to complexities associated with deformation. Although mafic granulites are important constituents of the lower crust in some areas, in others the prominent rock type is intermediate to acid granulite.

New high resolution seismic data are now showing that the original Moho concept for the crust-mantle boundary beneath continents is also much too simple. Obviously, a petrologically complex lower continental crust and upper mantle must be separated by an equally complex "interface". To quote Oliver (1982) "Many recent observations indicate that the current view of the Moho, the crust-mantle boundary, needs revision. In particular, the widely held concept of the Moho as a sharp laterally uniform boundary appears misleading in its simplicity, and may even be a major barrier to a better understanding of the nature and evolution of continents". Recent COCORP (Brown et al. 1981) and BIRPS (Matthews 1982) seismic data for deep crustal regions clearly show the complexity of deep crustal structure. Data like Fig. 3 from Brown et al. (1981) is rather typical and shows for a region of Texas that there is no single well defined continental Moho. Similar data led Missner (1973) to conclude that the Moho in some areas is a transition zone characterized by velocities greater than mean deep crustal values and yet less than mantle P-wave velocities; the interfingering of ultramafic intrusions and granulites is an interpretation of this transition zone which is receiving support from field and petrological studies of the peridotites and associated rocks in the Invrea Zone (e.g., Fountain and Salisbury 1981; Rivalenti et al. 1981). Brown et al. (1981) show that structures may change radically over very short lateral distances and could be strongly controlled

Offprint requests to: W.S. Fyfe

by local rheology, density or composition. The new seismic data also show the existence of rather low angle thrust faults in continental crust, thrusts which may go far below conventional Moho depths (Smythe et al. 1982; Matthews 1982). Certainly there are places where a good 30 km Moho appears to be present, but one wonders if this is not often simply a reflection of the resolution of seismic techniques (e.g., see Bolt 1982; p 74–78).

It is clear that any hypothesis which attempts to explain the origin of the structure of the continental crust and Moho must provide a mechanism for understanding the lithological diversity involved. This mechanism must address the control of mass transfer during igneous and metamorphic processes throughout geological time. As a step in this direction, we wish to point out some rather simple density relationships amongst rocks and magmas which place constraints on the distribution of mass in the continental crust and upper mantle.

A number of recent observations tend to suggest that all models have underestimated the importance of a basaltic component in both the lower crust in general and in the formation of present-day subduction zone magmatism in particular. Basaltic and kimberlitic pipes commonly contain pyroxene and garnet granulite nodules which are basaltic in composition and which originated at pressures corresponding to those of the lower crust (e.g., Rogers 1977; Arculus and Smith 1978; Padovani and Carter 1977; McCulloch et al. 1982). Subduction zone magmas of Central America are typically basaltic rather than andesitic, and include primitive MORB-like tholeiites (Carr 1983). Similar primitive magmas have been modelled as parental to the high alumina basalts in the Aleutians (Conrad and Kay 1982). Indeed, basaltic inclusions in crystals of olivine found in andesites, dacites as well as basalts have lead Anderson (1982) to the general conclusion that basaltic liquid is parental to the formation of new continental crust at convergent plate boundaries.

In Central America many of the magma chambers which fed the volcanoes appear to be located at the base of the crust (Carr 1983), in indicating that the Moho is a boundary layer under which magmas from the mantle are temporarily ponded. A magma ponding or underplating process has previously been considered in an attempt to explain the formation of continental crust during the Archean (Fyfe 1973, 1974, 1978), and is now being encorporated into models of the formation of continental flood basalts (Cox 1980) and all magmatic processes at convergent plate margins (Hildreth 1981). The idea is that basaltic magma will from a stable fluid layer beneath 'granitic' material of lower density as long as partial melting or plastic deformation of this buoyant layer can temporarily seal fractures and prevent the formation of a hydraulic head via extrusion. In the model of Cox (1980) the magma is a picrite which underplates and fractionates until the residual liquid becomes basaltic; the Moho becomes the petrological interface between olivine cumulates added to the mantle and gabbros added to the lower crust.

Magma Densities

With the densities of magmas to high pressures now better understood (Stolper et al. 1981; Herzberg 1983; Kushiro 1982) it is possible to examine the underplating model with new constraints. Figure 1 shows the densities of a range

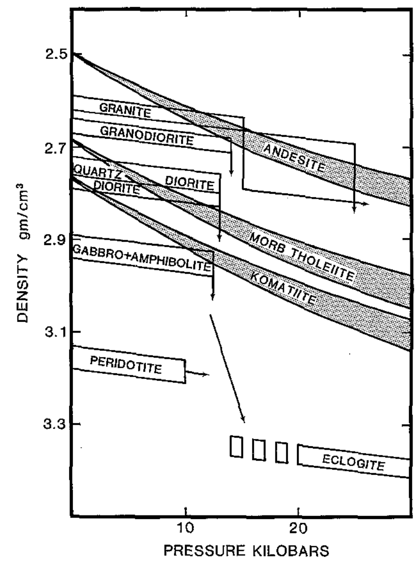

Fig. 1. Densities of rocks and magmas. Rock densities are averages listed in Clark (1966) calculated for lower crustal temperatures of 600–1,000° C using $1/V(dV/dT) = 3 \times 10^{-5}$ ° C^{-1} and $-1/V(dV/dP) = 1.1 \times 10^{-3}$ $kbar^{-1}$. Magmas are Mt. Hood andesite given in Bottinga and Weill (1970), average FAMOUS MORB glass in Elthon (1979), and P9–118 komatiite in Arndt et al. (1977) at 1,100, 1,250, and 1,600° C respectively. Densities of magmas at 1 atmosphere are from Nelson and Carmichael (1979). The range of magma densities at high pressures arise from uncertainties in the isothermal bulk modulus K^0 and its pressure derivative K' in an equation of state (Stolper et al. 1981; Herzberg 1983; Kushiro 1982); the values chosen are $K^0 = 150$ to 200 kbar at $K' = 7$. Densities of the high pressure mineralogies increase continuously and discontinuously due to solid solution effects and the stabilization of jadeite, garnet, kyanite, and coesite at pressures higher than 10–15 kbar (Stern et al. 1975; Stern and Wyllie 1981; Herzberg 1978)

of rocks which make up the crust and upper mantle, in addition to those of magmas which range from komatiite to andesite. The most important observation is that primitive MORB-like tholeiitic magmas are more dense than diorite, quartz diorite, granodiorite, and granite at 10 kilobars; they are also denser than most common sedimentary rocks. A maximum H_2O content of such magmas which can be parental to high alumina basalts and andesites is likely to be less than 1.6 wt.% (D.M. Harris, pers. comm.), thus reducing their density by about 0.06 gm cm^{-3} at these pressures (Bottinga and Weill 1970). A water content of 3% contained in some basaltic glasses considered by Anderson (1982) would reduce the magma density even more, but it would still be less than granodiorite at 10 kilobars. Although the dissolution of H_2O can clearly reduce the densities of magma at the base of the crust by a significant amount, this can be partly offset by the effect of suspended crystals in a vigorously convecting parcel of magma. For example, a dry tholeiitic liquid laden with 30% olivine crystals in suspension can have a bulk density of about 2.9 gm cm^{-3} at 10 kilobars, far greater than diorite and quartz diorite.

For relatively dry tholeiitic and picritic magmas, the continental Moho can be an effective boundary layer across which only heat could be transferred to the crust above. Indeed, these density relations explain why primary magmas from the mantle (i.e., picrites and komatiites) are so rare; quite simply, their densities require that they remain in the mantle or pond below a low density boundary layer. A breakdown of this boundary layer may only occur if the major element chemistry of these magmas evolve and if they gain volatiles such as H_2O. The continental crust could thus become a kind of density filter through which evolved tholeiites and magmas of intermediate composition could pass, particularly if they gained H_2O. Magmas which underplate the crust or penetrate the Moho and reside at various crustal levels could have their trace element and isotopic signatures extensively modified by contamination (e.g., Carter et al. 1978; Thompson et al. 1982; Carlson et al. 1981).

An objection which is periodically raised to a basaltic lower continental crust is that it would crystallize to eclogite with densities and P-wave velocities which are higher than the known geophysical properties of the crust at 10 kilobars (e.g., Heier 1973). This objection is based on the use of phase diagrams of the gabbro to eclogite transformation which were constructed from high temperature experiments extrapolated linearly to the temperatures of interest at the base of the crust (Green and Ringwood 1972; Kennedy and Ito 1972). Because these extrapolations require eclogite to be the stable basaltic mineralogy at all points along a normal geotherm, hypothetical reaction kinetics associated with this transition has been invoked to explain why we rarely see eclogite in crustal rocks (Kennedy and Ito 1972). However, it has been pointed out that these phase diagrams based on linear extrapolations of high temperature experimental data may be in error (Herzberg 1978). Indeed, recent experimental and thermochemical work has shown that garnet granulite rather than eclogite is the stable assemblage for many anhydrous metabasic compositions at 9 kilobars and 700–900° C (Newton and Perkins 1982). Additionally, any H_2O introduced could transform both garnet granulite and eclogite into amphibolite assemblages (e.g., Wyllie 1977). Finally, large volumes of magmas ponded below the crust would cool very slowly and differentiate much more efficiently than similar magma batches at higher crustal levels. Strong flotation of plagioclase and sinking of olivine and pyroxene cumulates could produce a layered ultramafic-anorthosite complex, not eclogite. As pointed out above, both seismic and field observations support the concept of the Moho as a transition zone of interlayered intrusions and granulites at the base of the crust in some areas.

It is worth considering, however, some consequences of an eclogite mineralogy in the lower portions of a very thick Andean-type continental crust where pressures in the 10–20 kilobar range may be reached. Layers or pods of eclogite with densities higher than that of mantle peridotite below would certainly be unstable; decoupling of eclogite from the lower crust and subsequent foundering into the mantle wedge above the subduction zone could occur. The fate and identify of the lower crust may then be very complex and depend on the total thickness and tectonic history of the continental crust being ponded. A basaltic component at the base of a very thick crust could begin to autodestruct on formation, leaving behind a stable lower crust with a buoyant mineralogy appropriate to compositions which are less basic.

The nature of the lithological diversity bounded on each side of a Moho can be even more complex when the origin of large volumes of Precambrian amphibolite facies orthogneisses are considered. Those from West Greenland and Zimbabwe have been interpreted to be juvenile material produced from an irreversibly differentiating Earth (Moorbath and Taylor 1981). However, the possibility that there has been mixing of continental crust and mantle on a large scale will render this interpretation ambiguous (Fyfe 1981). Regardless of whether the mass of the continental crust has grown or has been conserved with time, the isotopic data indicate that large volumes of calcalkaline magmas could have originated from the mantle during certain episodes. For a detailed discussion of the relationship between isotopic "age peaks" and "quantum" jumps in convection styles in the mantle, the reader is referred to Herzberg and Forsythe (1983). Of immediate concern here, however, is the fate of large quantities of these low density magmas. Figure 1 shows that such magmas of intermediate composition are less dense than granodiorite at 10 kilobars. These could have easily disrupted any pre-existing ancient Moho during their ascent to high crustal levels. This corresponds to the "over-accretion" model of Wells (1980) rather than the "under-accretion" or under-plating model which we have discussed above. Intuitively, the consequences of each model on the spatial distribution of igneous rocks, regional metamorphism, and partial melting in the crust would be profoundly different. Whereas the underplating model would tend to distribute the heat of crystallization of basaltic magmas over a larger lateral extent and produce a more sharply-defined Moho (i.e., but see below), the overaccretion model could result in larger lateral modifications to pre-existing lithologies and a severely disrupted Moho. Since the geological record points to a nonsteady state cooling Earth with periods of enhanced magmatic activity and metamorphism followed by relatively quiescent periods (e.g., Herzberg and Forsythe 1983), we should expect that there have been various ways in which mass has been transferred from the mantle to the crust. The production of a wide range of lower crustal lithologies and Moho geometries by these different accretion mechanisms is totally consistent with the geological and seismic observations outlined above.

With these alternative accretionary mechanisms in mind, we think that the obducted portions of the upper and lower crustal columns reviewed by Fountain and Salisbury (1981) point to the overall importance of a magma underplating model. Common to the cross-sections at the Invrea Zone, Fraser and Musgrave Range, Pikwitonei Belt, and the Kasila Series is a change in metamorphic grade from the granulite facies at the base to greenschist facies at the top. To quote Fountain and Salisbury (1981) "the most prominent layering in the crust is not compositional but metamorphic". Although detailed geothermometry and geobarometry remains to be done for these areas, most granulite facies rocks from all terrains studied in detail record 700–1,000° C and 5–10 kilobars (Bohlen and Boettcher 1981; Johnson and Essene 1982; Newton and Perkins 1982; Bohlen et al. 1983). If these equally apply to the granulite terrains cited above, the metamorphic geotherm would range from about 20–55° C km^{-1}, and could be representative of the actual geothermal gradient at their time of formation. Although high metamorphic geotherms have been interpreted as the product of rapid uplift and erosion for the Dalradian (e.g., England and Richardson 1977), we think that they are also the consequence of a vast "contact aureole" effect due to magma underplating. Indeed, both processes could have competed to boost the thermal state of the crust particularly if significant crustal updoming occurs above sites of rising mantle convection cells or diapirs.

A continental crust floating on an underplate magma flood would be subject to partial melting or degassing, and become mechanically weak. Complex styles of deformation near the Moho with rapid lateral changes (Brown et al. 1981) would hardly be surprising. Further, underplating provides an elegant mechanism for "cleaning" continental crust by a density filtration process. For example, if granite-greenstone terrains of the Archean were underplated, high density lithologies such as peridotites, greenstone and iron formations could have been returned to the mantle, low melting granites sweated out to higher crustal levels, and a high grade felsic basement residue established with the constraint that P basement $<$ P underplate magma. Sinking of high density H_2O-CO_2-S rich materials could flood the lower crust with such volatiles. We suggest that this could have been particularly important in the Archean characterized by the flotation of continental crust on a magma flood or ocean of dense tholeiite to komatiite.

Acknowledgements. This research was partly supported by NASA Under Grant NAGW-177. This paper was motivated by the organizers and participants of the Workshop on the Lower Continental Crust held in Princeton, 1982.

References

Anderson AT, Jr (1982) Parental magmas in subduction zones: implications for continental evolution. J Geophys Res 87:7047–7060

Arculus RJ, Smith D (1979) Eclogite, pyroxenite and amphibolite inclusions in the Sullivan Buttes latite, Chino Valley, Yavapai County, Arizona. In: Boyd FE, Meyer HOA (eds) The mantle sample: Inclusions in kimberlites and other volcanics. Am Geophys Union 309–317

Arndt NT, Naldrett AJ, Pyke DR (1977) Komatiitic and iron-rich tholeiitic lavas of Munro Township, Northeastern Ontario. J Petrol 18:319–369

Bohlen SR, Boettcher AL (1981) Experimental investigations and geological applications of orthopyroxene geobarometry. Am Mineral 66:951–964

Bohlen SR, Wall VJ, Boettcher AL (1983) Experimental investigation and application of garnet granulite equilibria. Contrib Mineral Petrol in press

Bolt BA (1982) *Inside the Earth.* WH Freeman and Co, San Francisco, 191 pp

Bottinga Y, Weill DF (1970) Densities of liquid silicate systems calculated from partial molar volumes of oxide components. Am J Sci 268:169–182

Brown LD, Oliver JE, Kaufman S, Brewer JA, Cook FA, Schilt FS, Albaugh DS, Long GH (1981) Deep crustal strucutre: Implications for continental evolution. In: O'Connell RJ, Fyfe WS (eds) Evolution of the Earth. Geodynamics series vol. 5, Am Geophys Union 38–52

Carlson RW, Lugmair GW, MacDougall JD (1981) Columbia River volcanism: the question of mantle heterogeity or crustal contamination. Geochim Cosmochim Acta 45:2483–2499

Carr MJ (1983) Crustal influence on magmatic diversity along the Central American volcanic front. J Volcanol Geotherm Res, submitted

Carter SR, Evensen NM, Hamilton PJ, O'Nions RK (1978) Nd- and Sr-isotopic evidence for crustal contamination of continental volcanics. Science 202:743–747

Clark SP, Jr (1966) Handbook of Physical Constants. Geol Soc Amer Mem 97

Conrad WK, Kay RW (1982) Ultramafic and mafic inclusions from Adak Island: Crystallization history, and implications for the nature of primary magmas and crustal evolution in the Aleutian Arc. J Volcanol Geotherm Res (in press)

Cox KG (1980) A model for flood basalt volcanism. J Petrol 21:629–650

Elthon D (1979) High magnesia liquids as the parental magma for ocean floor basalts. Nature 278:514–518

England PC, Richardson SW (1977) The influence of erosion upon the mineral facies of rocks from different metamorphic environments. J Geol Soc London 134:201–213

Fountain DM, Salisbury MH (1981) Exposed cross-sections through the continental crust: implications for crustal structure, petrology, and evolution. Earth Planet Sci Lett 56:263–277

Fyfe WS (1973) The granulite facies, partial melting and the Archean crust. Phil Trans R Soc London A273:457–461

Fyfe WS (1974) Archean tectonics. Nature 249:338

Fyfe WS (1978) The evolution of the Earth's crust: modern plate tectonics to ancient hot spot tectonics? Chemical Geology 23:89–114

Fyfe WS (1981) How do we recognize plate tectonics in very old rocks? In: Kroner A (ed) Precambrian Plate Tectonics. Elsevier 549–560

Green DH, Ringwood AE (1972) A comparison of recent experimental data on the gabbro-garnet granulite-eclogite transition. J Geol 80:277–288

Heier KS (1973) Geochemistry of granulite facies rocks and problems of their origin. Phil Trans R Soc London A273:429–442

Herzberg CT (1978) Pyroxene geothermometry and geobarometry: Experimental and thermodynamic evaluation of some subsolidus phase relations in the system $CaO-MgO-Al_2O_3-SiO_2$. Geochim Cosmochim Acta 42:945–957

Herzberg CT (1983) Solidus and liquidus temperatures and mineralogies for anhydrous garnet-lherzolite to 15 GPa. Phys Earth Planet Inter (in press)

Herzberg CT, Forsythe RD (1983) Destabilization of a 650 kilometer chemical boundary layer and its bearing on the evolution of the continental crust. Phys Earth Planet Inter (in press)

Hildreth W (1981) Gradients in silicic magma chambers: implications for lithospheric magmatism. J Geophys Res 86:10153–10192

Johnson CA, Essene EJ (1982) The formation of garnet in olivine-bearing metagabbros from the Adirondacks. Contrib Mineral Petrol 81:240–251

Kennedy GC, Ito I (1972) Comments on: "A comparison of recent experimental data on the gabbro-garnet granulite-eclogite transition". J Geol 80:289–292

Kushiro I (1982) Density of tholeiite and alkali basalt magmas at high pressures. Carnegie Inst Wash Yearbook 81:305–309

Matthews DH (1982) BIRPS: deep seismic reflection profiling around the British Isles. Nature 298:709–710

McCulloch MT, Arculus RJ, Chappel BW, Ferguson J (1982) Isotopic and geochemical studies of nodules in kimberlite have implications for the lower continental crust. Nature 300:166–169

Meissner R (1973) The "Moho" as the transition zone. Geophys Surveys 1:195–216

Moorbath S, Taylor PN (1981) Isotopic evidence for continental growth in the Precambrian, In: Kroner A (ed) Precambrian Plate Tectonics, Elsevier 491–526

Nelson SA, Carmichael ISE (1979) Partial molar volumes of oxide components in silicate liquids. Contrib Mineral Petrol 71:117–124

Newton RC, Perkin D (1982) Thermodynamic calibration of geobarometers based on the assemblages garnet-plagioclase-orthopyroxene (clinopyroxene)-quartz. Am Mineral 67:203–222

Oliver J (1982) Changes at the crust-mantle boundary. Nature 299:398–399

Padovani ER, Carter JL (1977) Aspects of the deep crustal evolution beneath south central New Mexico. In: Heacock JG) The Earth's crust: its nature and physical properties. Am Geophys Monogr Ser 20:19–55

Rivalenti G, Garuti G, Rossi A, Siena F, Sinigoi S (1981) Existence

of different peridotite types and of a layered igneous complex in the Invrea Zone of the Western Alps. J Petrol 22:127–153
Rogers NW (1977) Granulite xenoliths from Lesotho kimberlites and the lower continental crust. Nature 270:681–684
Smythe DK, Dobinson A, McQuillan R, Brewer JA, Matthews DH, Blundell DJ, Kelk B (1982) Deep structure of the Scottish Caledonides revealed by the MOIST reflection profile. Nature 299:338–340
Stern CR, Huang W-L, Wyllie PJ (1975) Basalt-andesite-rhyolite-H_2O: crystallization intervals with excess H_2O and H_2O-undersaturated liquidus surfaces to 35 kilobars, with implications for magma genesis. Earth Planet Sci Lett 28:189–196
Stern CR, Wyllie PJ (1981) Phase relationships of I-type granite with H_2O to 35 kilobars: The Kinkey Lakes biotite-granite from the Sierra Nevada batholith. J Geophys Res 86:10412–10422
Stolper E, Walker D, Hagar BH, Hays JF (1981) Melt segregation from partially molten source regions: the importance of melt density and source region size. J Geophys Res 86:6261–6271
Tarney J, Windley BF (1979) Continental growth, island arc accretion and the nature of the lower crust – a reply to SR Taylor and SM McLennan. J Geol Soc London 136:501–504
Taylor SR (1979) Chemical composition and evolution o the continental crust: the rare earth element evidence. In: McElhinney MW (ed) The Earth: Is origin, structure, and evolution. Academic 353:376
Thompson RN, Dicken AP, Gibson IL, Morrison MA (1982) Elemental fingerprints of isotopic contamination of Hebridean Palaeocene mantle-derived magmas by Archean sial. Contrib Mineral Petrol 79:159–168
Wells PRA (1980) Thermal models for the magmatic accretion and subsequent metamorphism of continental crust. Earth Planet Sci Lett 46:253–265
Wyllie PJ (1977) Crustal anatexis: an experimental review. Tectonophys 43:41–71

Accepted May 3, 1983