



# Dehydration melting and the relationship between granites and granulites



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## ABSTRACT

For more than half a century, thought about granite genesis and crustal evolution has been guided by the concept of partial melting in the lower crust. In this model, granitic magmas produced at depth are lost to shallow levels, leaving behind a more mafic, volatile poor residue that is depleted in incompatible components (H<sub>2</sub>O, alkalis, and heat-producing elements). Although granite extraction must be the dominant process by which crust is modified over time, the preferred model of granite genesis triggered by metamorphic dehydration reactions (dehydration melting) does not adequately explain important aspects of granite formation. The temperatures required for voluminous granite production by dehydration melting need heat and mass input to the crust from mantle-derived mafic magmas. In addition, prediction of the H<sub>2</sub>O contents of granitic liquids by extrapolation from low-pressure experiments to deep-crustal pressures (*P*) and temperatures (*T*) implies that the H<sub>2</sub>O resident in hydrous minerals is insufficient to account for large granite volumes, such as anorogenic granite batholiths in continental interiors. To test this, we conducted new experiments on the H<sub>2</sub>O contents of simple granitic liquids at 10 kbar and 800–950 °C. We confirm previous extrapolations from lower *P* and *T* indicating that a minimum of 3–4 wt% H<sub>2</sub>O is present at the studied *P* and *T* in a granitic liquid in equilibrium with quartz and feldspars. For large-scale melting, this is much more than could have been supplied by the H<sub>2</sub>O resident in biotite and amphibole by dehydration melting at these conditions, unless lower-crustal temperatures were higher than generally inferred. Another problem with the dehydration-melting model is that the crystal chemistry of the large-ion lithophile elements (LILE) does not favor their partitioning into granitic liquids; rather, U, Th, Rb and the rare earth elements (REE) would more likely be concentrated in the postulated mafic residues. Finally, observations of migmatite complexes reveal many features that can not be satisfied by a simple dehydration-melting model.

We suggest that the volatile components CO<sub>2</sub> and Cl are important agents in deep-crustal metamorphism and anatexis. They induce crystallization and outgassing of basalt magmas at lower-crustal levels, where the combination of latent heat and liberated H<sub>2</sub>O may contribute to granite production, leading to larger melt fractions than for simple dehydration-melting models. Since the Cl and CO<sub>2</sub> are very insoluble in granite liquids, granite generation leads naturally to production or separation of a coexisting metamorphic fluid with low H<sub>2</sub>O activity. Such a fluid could coexist with granulite-facies assemblages and yet be capable of dehydration, alkali exchange and LILE extraction to explain many chemical processes of deep-crustal metamorphism not readily explainable by dehydration melting.

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## 1. Introduction

Tuttle and Bowen's (1958) memoir on experimental melting of a simple granite (alkali feldspars + quartz) at elevated H<sub>2</sub>O pressures is one of the most influential works in petrology. They showed that melting temperatures of common quartzofeldspathic rocks are lowered by high H<sub>2</sub>O pressure to the temperature range thought to prevail in high-grade metamorphism. This led to the concept of

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granite formation as the culmination of regional metamorphism (Brown and Fyfe, 1970). To many petrologists, the new experimental work provided a satisfying guide to the interpretation of granite in all of its manifestations. Some aspects of granite described in classical field studies, observations that could be plausibly attributed to the interaction of rocks with migrating pore fluids, or metasomatism (Anderson, 1937; Misch, 1949), were downplayed in the expectation that these features could be explainable in the larger context of magmatism.

In the years following publication of the “Granite Memoir” experimental and theoretical petrology has shown that the temperature scale of crustal metamorphism extends well beyond the melting temperatures of crustal rocks at high H<sub>2</sub>O pressures (~650–700 °C; e.g., Bohlen, 1987; Bohlen and Mezger, 1989; Harley, 1989, 1998; Pattison et al., 2003). An explanation commonly advanced for a refractory lower crust is H<sub>2</sub>O deficiency. Rocks that had experienced granulite-facies metamorphism – an important fraction of the deeper parts of the continental crust – are dominated by anhydrous minerals and possess porosity so low that significant long-term storage of free volatiles is implausible.

Granulites exposed in uplifted deep-crustal terranes commonly contain segregations of feldspars and quartz, the components of the rocks that melt at lowest temperature in the presence of H<sub>2</sub>O; that is, they are migmatized. The H<sub>2</sub>O must have been restricted in amount (or in its activity) to limit the extent of partial melting. The model of “dehydration melting” provides a plausible explanation: the H<sub>2</sub>O available comes entirely from micas and amphiboles initially present in the granodioritic or tonalitic gneisses and, to a lesser extent, metasediments in the deeper parts of the crust.

The concept of partial hydrous melting in the deep crust and extraction of granitic magmas leads naturally to a simple crustal evolution model which could account not only for removal of H<sub>2</sub>O and other hyperfusible components, leaving a refractory granulite-facies lower crust, but might also explain certain trace-element signatures of the exposed high-grade terranes, including depletion, relative to upper crustal rocks of the same general major element compositions, of the large-ion lithophile elements like Cs, Rb, Ba, and, to a lesser extent, K, and the radioactive elements U and Th (Heier, 1973). Such a process might be important in long-term survival of the continents in that it creates a refractory underpinning to continental masses (Collerson and Fryer, 1978).

The dehydration-melting hypothesis satisfies some of the major observational requirements. These include the experimental facts that partial melting of metatonalites and Al-rich metasediments yields, respectively, liquids of the compositions of metaluminous (Rutter and Wyllie, 1988) and peraluminous (Vielzeuf and Holloway, 1988) granites, that the residual minerals are the mafic-aluminous silicates pyroxene, garnet, cordierite, and sillimanite characteristic of the granulites, and that the temperature range of melting, 750–900 °C, is attained in high-grade metamorphism (Bohlen, 1987; Bohlen and Mezger, 1989; Harley, 1989; Pattison et al., 2003). Dehydration melting has been suggested to play a major role in softening of the middle crust, thus expediting detachment faulting and the extraction of granite batholiths (Vigneresse et al., 1996; Vanderhaeghe and Teyssier, 2001). Fluid-absent dehydration melting is commonly invoked to explain crustal differentiation by upward removal of the granitic partial melt, leaving a volatile- and LILE-depleted granulite-facies lower crust (e.g., Clemens, 1990; Clemens et al., 1996; Vielzeuf et al., 1990; Brown, 2001).

Although the dehydration-melting model has gained wide currency, there remain key aspects of granite genesis and its relation to granulite-facies metamorphism that it does not adequately explain. It is difficult to supply sufficient heat to generate observed granite masses solely by dehydration-melting of the lower crust. In addition, the H<sub>2</sub>O available for dehydration melting of

relatively anhydrous lower crustal granulitic rocks yields relatively small melt volumes. The partitioning of LILE between granitic liquids and granulite residues does not adequately explain observations in exposed terranes. The feasibility of efficient extraction of low-degree viscous partial melts from their residua remains uncertain. Finally, segregations or leucosomes with compositions different than those predicted for dehydration melting are commonly observed in granulite grade migmatites.

The purpose of the present paper is to discuss in more detail some of the aspects of the supposed granite-granulite connection that have not been adequately addressed but which may bear on the concept of fluid-deficient dehydration melting. We present new experimental results that support the low liquid fractions produced by typical amphibolite facies rocks during dehydration melting at 10 kbar, as inferred by Johannes and Holtz (1991). We build on this result by discussing additional observations that are not well-explained by the dehydration-melting model. We conclude by proposing that alternative conceptual schemes involving the introduction of volatile components, particularly H<sub>2</sub>O, into the sites of deep-crustal melting can resolve some of these problems.

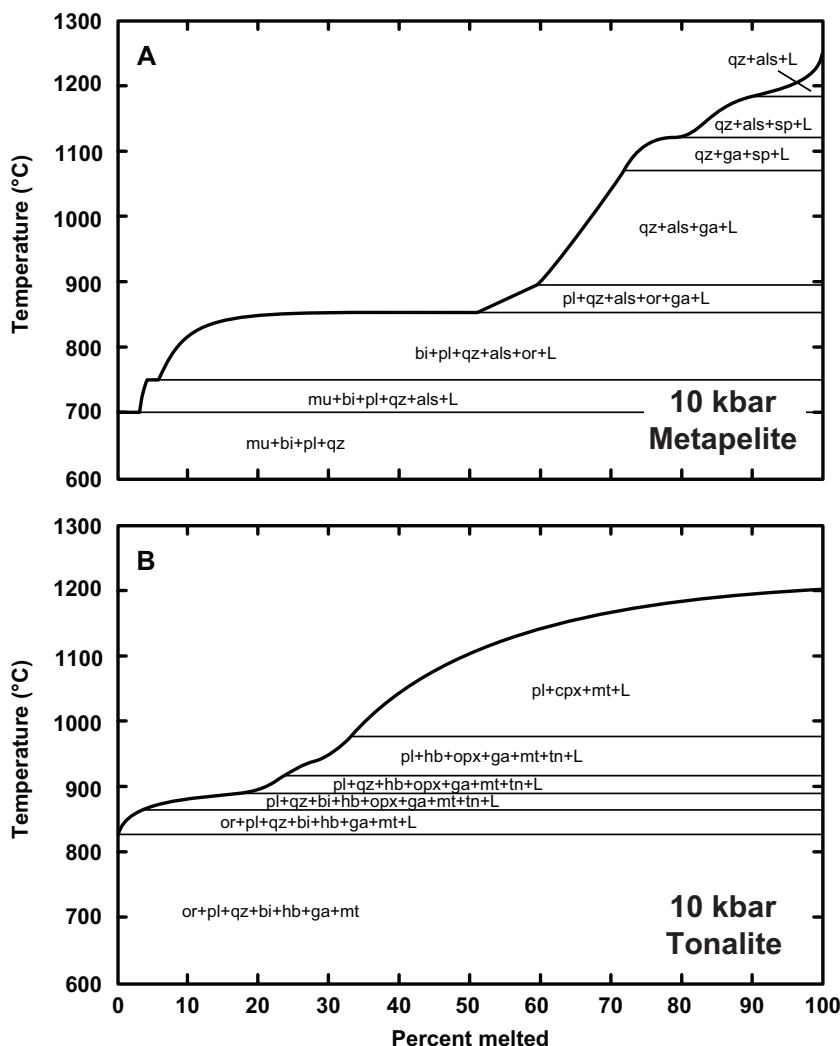
## 2. The problem of heat and mass transfer during crustal melting

Granites are chiefly products of melting of continental crust, with some input of mantle-derived material (Wyllie et al., 1976; Leake, 1990). While some granites may reflect fractional crystallization of a mantle-derived magma (e.g., Fowler et al., 2001), it is difficult to produce granite in the large amounts seen in batholiths by fractional crystallization-differentiation of basalts, or by assimilation of crustal material (Bowen, 1928).

As Bowen (1928) pointed out, simultaneous consideration of the heat and the material budgets places an important constraint on granite genesis. If mantle-derived basalt is the heat source for crustal melting (Harley, 1989; Bohlen, 1987) as well as the source of the extra-crustal material component, the creation of a certain mass of granitic liquid must be attended by crystallization of a nearly equal mass of injected basalt to supply the heat for melting. This would in turn imply that generation of granite in the lower crust also involves large-scale crustal accretion in the form of basalt underplating (Ewart et al., 1980). These considerations have at least two consequences.

First, the postulated mantle-derived basalts would, in some measure, be expected to supply some material to granitic liquids, as well as heat. Yet most granites are demonstrably almost entirely partial melts of pre-existing crust in their major element chemistry; it is in the Sr and Nd isotopes that large amounts of primitive mantle input are implicated (Leake, 1990). It is difficult to see how this incongruent contamination could occur unless some transfer mechanism exists that is not anticipated by models of melting in a closed system such as simple dehydration melting.

Second, the source-rock fertility determines the amount of granitic liquid that can be produced by crustal melting. Variations in fertility arise chiefly from the bulk H<sub>2</sub>O content, as controlled by the abundance and identity of hydrous minerals in the source rock. The larger the H<sub>2</sub>O content, the larger the volume of granitic liquid produced at a given pressure and temperature, all else being equal. The best case for generating substantial quantities of granitic liquid by dehydration melting can be made for Al-rich metasedimentary sources (Vielzeuf and Holloway, 1988; Thompson, 1996; Nabelek and Bartlett, 1998). Fig. 1A shows the percentage melt yield of a metamorphic rock containing 30 modal percent of biotite and muscovite at deep-crustal pressures. As much as 50 vol% of melt



**Fig. 1.** (A) Percent yield of liquid from a metapelite bulk composition with 2.2 wt% H<sub>2</sub>O as a function of temperature at 10 kbar, from Vielzeuf and Holloway (1988). (B) Percent yield of liquid from a tonalite bulk composition with 0.8 wt% H<sub>2</sub>O as a function of temperature at 10 kbar, from Rutter and Wyllie (1988). Als, sillimanite; bi, biotite; cpx, clinopyroxene; ga, garnet; hb, hornblende; L, liquid; mt, magnetite; mu, muscovite; opx, orthopyroxene; or, K feldspar; pl, plagioclase; qz, quartz; sp, spinel; tn, titanite.

can be generated at high-grade metamorphic temperatures of up to 900 °C, leaving residues of garnet and sillimanite. Such a large mass of liquid would be readily extractable, even by conservative estimates (Wickham, 1987).

A metasedimentary source could account for only a fraction of exposed granite terranes, inasmuch as the granitic extract from melting of sedimentary rocks is always peraluminous (e.g., Clemens et al., 1996). For example, modal muscovite is not likely to be a factor in the source region of I- and A-type granites. The melting experiments of Rutter and Wyllie (1988) apply more directly to the tonalitic-trondhjemitic gneisses that are likely the major components of the deeper continental crust. Fig. 1B shows the temperature-dependent yield of granitic liquid from a tonalitic gneiss having 15 modal percent of biotite and amphibole (Rutter and Wyllie, 1988). A reasonable metamorphic temperature for generating granitic liquids would be about 875 °C, where biotite begins to melt. This would still leave some residual refractory biotite in the deep-crustal granulite-facies terrane, as observed in even the highest grade exposed granulite-facies terranes, such as Enderby Land, Antarctica (Ellis et al., 1980). The amount of melting would be about 10 vol% (Fig. 1B). This relatively modest yield would be in accord with Bowen's (1928) heat budget consideration, if injection of basaltic magma is the only feasible heat source (e.g., Rutter

and Wyllie, 1988). The modest amount of granitic liquid generated by the dehydration-melting process would be an upper limit since (1) the deep-crustal sources of Phanerozoic granite terranes could include less-fertile rocks previously metamorphosed in the granulite facies, and (2) the estimate assumes perfect extractability.

These observations highlight the challenge for the dehydration-melting model to produce voluminous granitic terranes. An example is the Transbaikalian granites, which occupy 50% of the surface of an area of subcontinental dimensions and in places comprise half the crustal thickness (Litvinovskii et al., 1994). Voluminous alkali-rich granite magmatism in other localities present similar challenges, including the Caledonide granites of Scotland and Ireland (Harmon et al., 1984) and those of the Nubian-Arabian Shield (Harris, 1982). One potential way to generate large masses of granite by crustal melting would be lower-crustal temperatures that routinely exceeded 900 °C. However, while examples of such "ultrahigh-temperature metamorphism" do exist, they are rare in the geologic record (e.g., Harley, 1998). Even if estimates of granulite peak temperatures are too low (e.g., Pattison et al., 2003), there is currently insufficient evidence to suggest that such high temperatures consistently characterize the crustal sources of granitic liquids, as required by a closed-system melting model.

### 3. The problem of the H<sub>2</sub>O content at the granite liquidus

The water content of source rocks controls fertility because of the strong solubility of H<sub>2</sub>O in granitic liquids. The lever rule dictates that, for a given source rock H<sub>2</sub>O content at fixed *P* and *T*, the volume of melt that can be produced decreases with increasing concentration of H<sub>2</sub>O in the silicate liquid. Surprisingly, the H<sub>2</sub>O content of simple granite liquids has never been directly investigated at pressures above 5 kbar. Most published calculations of the fertility of crustal source rocks therefore rely on liquidus H<sub>2</sub>O contents measured at low pressures and extrapolated to deep-crustal conditions. At 10 kbar and 900 °C, the extrapolation of Johannes and Holtz (1991) predicts that a simple-system granitic magma contains about 2.3 wt% H<sub>2</sub>O at the liquidus. This would allow some possible source rocks – notably, mica-rich metasediments – to be marginally fertile at *P*–*T* conditions that are not extreme for deep-crustal metamorphism. Holtz et al. (2001) subsequently revised the extrapolation upward to about 3.5%. This is sufficiently high that the fertility of all common crustal rocks would be too low to produce significant volumes of granitic liquid by closed-system dehydration melting. However, both of these estimates are uncertain because they involve extrapolations from lower temperatures and pressures.

To address this problem, we carried out new measurements of the liquidus H<sub>2</sub>O content of a simple granite composition. The experiments were conducted at 10 kbar and represent the first determination of the liquidus H<sub>2</sub>O content at conditions appropriate for deep-crustal melting. Three haplogranite glasses were prepared from synthetic glass of albite (NaAlSi<sub>3</sub>O<sub>8</sub>) composition (prepared as described in Hayden and Manning, 2011), Puratronic SiO<sub>2</sub> glass (anhydrous), and sanidine glass prepared by D. B. Stewart (#95GQA1). The weight percentages of the albite:K-feldspar:quartz glass components were 40.0:40.0:20.0, 37.5:37.5:25.0, and 35.0:35.0:30.0. The weight ratios of the alkali feldspars were chosen to be 1:1, based on experimental work at 10 kbar by Ebadi and Johannes (1991). Each glass component was separately fine-ground under acetone in an agate mortar into a smooth paste, then baked at 600 °C. Each of the three starting compositions was prepared by mixing weighed portions of the three glasses with prolonged stirring under acetone. Each mixture was melted to a homogeneous glass in a closed Pt container at 1380 °C for 2 h, quenched, again fine-ground to a smooth paste with acetone, and again baked at 600 °C.

Starting materials were weighed into Pt tube segments of either 2.0 or 3.5 mm diameter, depending on the sample size (20–40 mg). Water was added with a microliter syringe in the amount of 4.1 ± 0.1 wt% of the total charge. Weighing was done on a Mettler UMX2 ultramicrobalance (1σ = 0.2 μg). A small excess of water was added and the weight monitored during slow evaporation until the desired amount was attained, after which the capsule was pinched and sealed by arc-welding. Weight loss on sealing was usually less than 50 μg, all of which is attributable to Pt evaporation, as demonstrated by sealing tests.

All experiments were carried out in a piston-cylinder apparatus of 2.54 cm diameter using NaCl pressure medium and graphite heater sleeves. Pressures were monitored on a Heise bourdon tube gauge. Samples were pressed cold to 7 kbar, then heated. Pressure rose with heating but was held at the desired final pressure by bleeding. For runs above 900 °C, the interior of the furnace surrounding the sample capsule was modified with BN inserts of 0.38 cm height. Calibration of the NaCl-BN furnace assembly was carried out by measuring the solubility of quartz in H<sub>2</sub>O (Newton and Manning, 2000). A pressure correction of 0.6 kbar was required, in agreement with previous calibrations of the NaCl-BN cell at 10 kbar. Accordingly, the nominal pressure was held at 10.6 kbar throughout the runs. Pressure uncertainty is ±0.3 kbar.

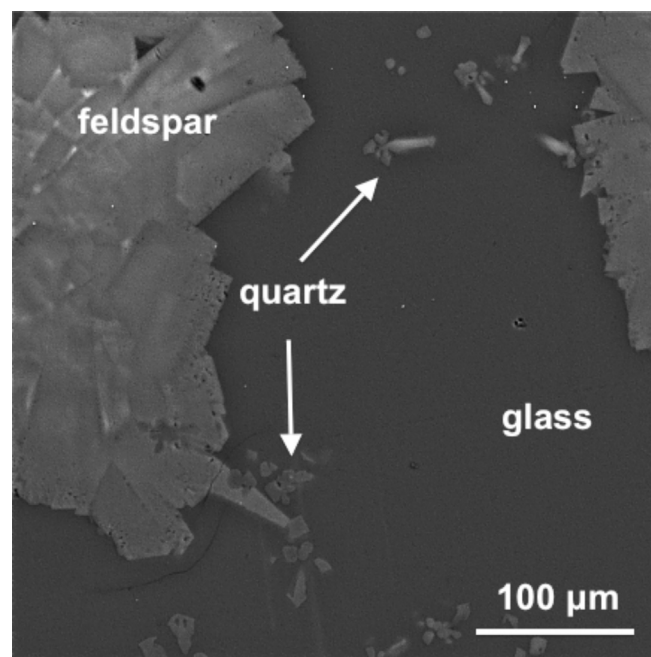


Fig. 2. Cathodoluminescence photomicrograph of charge quenched from high *P*–*T* experiment, showing alkali feldspar, quartz, and glass (quenched liquid). Run #194 (Table 1).

Temperatures were controlled automatically and measured with Pt–Pt 10%Rh (type S) thermocouples with a stated accuracy of ±3 °C. With the exception of run 186, experiments in which crystal growth was anticipated were first held at 950 °C for 2–3 days to homogenize the melt above the liquidus, then the temperature was lowered and held at the final value (Table 1). Durations of the runs varied from 52 to 165 h. Runs were terminated by quenching to below 200 °C in less than 20 s.

Quenched and cleaned sample capsules were weighed to check for any weight loss that could be attributable to leakage during a run. There was never any detectable weight loss, which shows that the initial starting water content was conserved. Capsules were cut open along a margin so as to preserve the quenched glassy slug in an unbroken wafer. The glass was ground flat and polished on both surfaces for petrographic study. The results were readily determined by optical microscopy. Birefringent crystals in a matrix of isotropic glass were either a few large feldspars or many small skeletal β-quartz, or both (Fig. 2).

The results allow depiction of isobaric phase relations as a function of temperature and SiO<sub>2</sub> concentration (Fig. 3). The minimum melt composition lies between 25 and 30 wt% SiO<sub>2</sub> and at a temperature above 800 °C. The determinations may be considered as half-reversals in that they give a minimum liquidus temperature. Higher liquidus temperatures are possible if the experiments are biased by undercooling (i.e., failure to nucleate feldspar and/or quartz until a temperature is encountered below the true minimum melting temperature). Long runs may be needed to assure that, after initial equilibration of a charge at 800 °C, subsequently raising the temperature induces resorption of both quartz and feldspar.

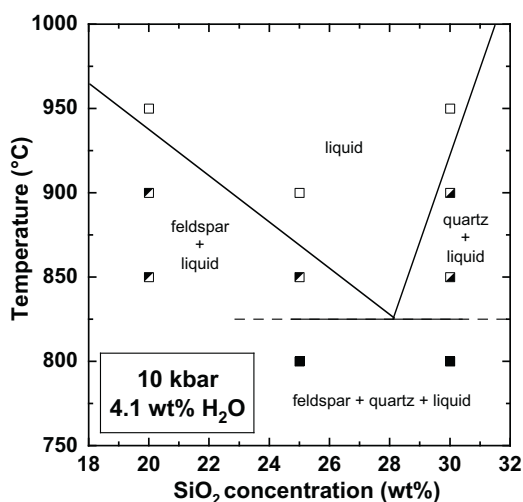
Fig. 4 compares estimates of the H<sub>2</sub>O content along the simple granite liquidus surface at 10 kbar. The curves of Johannes (1985), Johannes and Holtz (1991) and Holtz et al. (2001) are all extrapolations from lower pressure. Our half-reversed bracket is shown for comparison. The data suggest that the liquidus is located at 800–850 °C, 25–30 wt% SiO<sub>2</sub>, but extension to the extreme upper limit of 900 °C should be allowed due to the lack of a full reversal. Considering that the initial 4.1 wt% H<sub>2</sub>O is concentrated in the liquid phase and that the runs at 850 °C contain abundant crystals



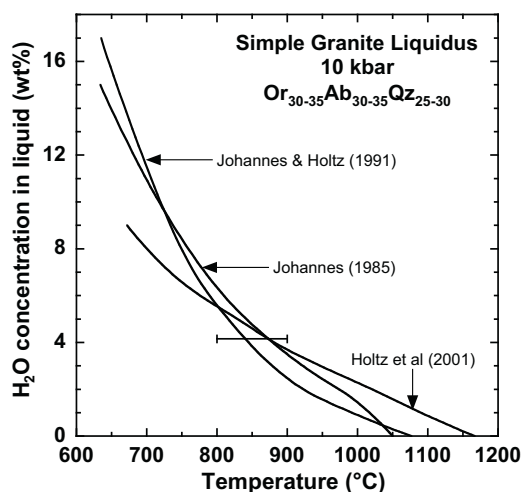
**Table 1**  
Results of experiments on granite melting at  $1.00 \pm 0.03$  GPa and  $4.1 \pm 0.1$  wt%  $H_2O$ .

| Run number | ab, ksp, qz (wt%) | Time (hours) | Temperature ( $^{\circ}C$ ) | Result           |
|------------|-------------------|--------------|-----------------------------|------------------|
| 172        | 35.0, 35.0, 30.0  | 52           | 950                         | glass            |
| 177        | 35.0, 35.0, 30.0  | 70           | 950                         |                  |
|            |                   | 70           | 900                         | qz + glass       |
| 186        | 40.0, 40.0, 20.0  | 66           | 900                         | fls + glass      |
| 187        | 40.0, 40.0, 20.0  | 69           | 950                         | glass            |
| 189        | 40.0, 40.0, 20.0  | 72           | 950                         |                  |
|            |                   | 70           | 850                         | fls + glass      |
| 190        | 35.0, 35.0, 30.0  | 65           | 950                         |                  |
|            |                   | 70           | 850                         | qz + glass       |
| 192        | 37.5, 37.5, 25.0  | 72           | 950                         |                  |
|            |                   | 47           | 850                         | fls + glass      |
| 193        | 37.5, 37.5, 25.0  | 70           | 950                         |                  |
|            |                   | 47           | 900                         | glass            |
| 194        | 35.0, 35.0, 30.0  | 67           | 950                         |                  |
|            |                   | 98           | 800                         | fls + qz + glass |
| 195        | 37.5, 37.5, 25.0  | 67           | 950                         |                  |
|            |                   | 98           | 800                         | fls + qz + glass |

Abbreviations: fls, alkali feldspar; qz, quartz. Two-stage experiments involved an initial stage at temperature greater than the liquidus, then a second step at lower temperature; results reflect observations on quenched run products after final step (see text).



**Fig. 3.** Stability fields of phase assemblages of simple granite compositions at 10 kbar, 4.1 wt%  $H_2O$ , and equal weight fractions of albite and orthoclase (Table 1).



**Fig. 4.** Liquidus  $H_2O$  contents of a simple minimum-melting granite composition (Ab-Or-Qz) according to model extrapolations of Johannes (1985), Johannes and Holtz (1991), and Holtz et al. (2001), compared to our bracket for liquidus temperature at 10 kbar and  $4.1 \pm 0.1$  wt%  $H_2O$ . The lower end of solid bracket ( $800^{\circ}C$ ) is an absolute minimum, and the  $H_2O$  contents at higher temperatures must be higher than 4.1 wt% because of the presence in the charges of non-negligible volumes of crystals at  $850^{\circ}C$ . The upper bound of the bracket is  $900^{\circ}C$ .

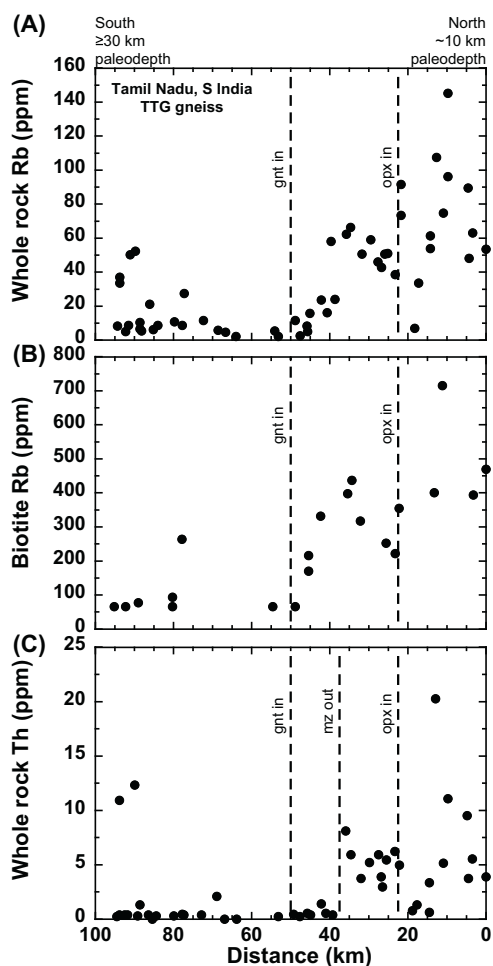
as well as glass, the high- $T$  half of the  $800$ – $900^{\circ}C$  bracket may be more appropriate.

Fig. 4 demonstrates that our results confirm the extrapolations of Johannes (1985), Johannes and Holtz (1991) and Holtz et al. (2001) at 4.1 wt%  $H_2O$ . As pointed out by Johannes and Holtz (1991), dehydration melting of an intermediate gneiss with typical modal amounts of biotite and amphibole ( $\sim 0.6$  wt%  $H_2O$ ) would only yield small amounts of granitic liquid at temperatures that are at the high end of the range considered appropriate for deep-crustal melting. The apparent success of this preliminary determination of liquidus  $H_2O$  content of a simple granite in the deep-crustal pressure range suggests that the methods used here may remove some of the uncertainty apparent in Fig. 4. Thus, the conclusion is the same as that when considering the heat budget of melting: melt fractions generated by dehydration-melting are problematically low unless lower-crustal temperatures higher than  $900^{\circ}C$  prevail over large regions of the lower crust during melting.

#### 4. The problem of trace-element depletion

If dehydration melting leads to extraction of a granitic liquid from the lower crust, then a volatile-poor granulite residue should be generated (e.g., Fyfe, 1973). This potential complementary relationship between granites and granulites can be tested by evaluating whether trace elements such as Rb and Th agree with predictions for the residue from which granitic liquid has been extracted. Lambert and Heier (1968) and Heier (1973) found these elements to be moderately or severely depleted in granulites from Australia and Norway. This is now known to be a feature of only the highest grade metamorphic rocks. Some transitional granulites are actually enriched in Rb, such as the incipient charnockites of southern Karnataka, India (Stähler et al., 1987), and some are greatly enriched in Th, including the charnockitic veins of Namibia (Andreoli et al., 1993). Nevertheless, depletion of these elements, along with U, and to a lesser extent, K, appears to be a characteristic feature of the highest grade of granulite-facies metamorphism.

Fig. 5A shows whole-rock Rb content across what appears to be an unbroken cross-section of Late Archean crust in southern India (Hansen and Harlov, 2007). The metamorphic grade increases from north to south and is a function of paleodepth, from about 15 km at the onset of the granulite facies, to at least 30 km in the highest-grade garnet-bearing orthogneiss (peak  $T \geq 800^{\circ}C$ ). The rocks of this traverse are tonalitic-trondhjemitic-granodioritic (TTG) gneisses representing much of the middle and lower continental crust. They are biotite bearing, with some of the more



**Fig. 5.** Compositional variations in Archean TTG gneisses, Tamil Nadu State, South India (Hansen et al., 2002; Hansen and Harlov, 2007). (A) Whole-rock Rb. Note that some orthopyroxene (opx)-bearing rocks are not severely depleted in Rb and that some non-garnetiferous (gnt) rocks are severely depleted. The four high values between 90 and 100 km of the traverse are from a boundary shear zone and are likely rocks retrogressively enriched in Rb. (B) Rb content of biotite. The refractory high-grade biotite is very low in Rb, which cannot be explained by subtraction of a granitic melt to account for the high-grade depleted granulites. (C) Whole rock Th. The monazite (mz)-out isograd coincides with severe depletion of Th. This effect is unlikely to result from melt extraction (see text). The two high values between 90 and 100 km of the traverse are from the retrogressive bounding shear zone.

mafic lithologies containing hornblende and/or clinopyroxene, and garnet in the highest grade, deepest section. There is a distinct orthopyroxene-in boundary marking the onset of the granulite facies.

Rubidium becomes strongly depleted as metamorphic grade increases from north to south (Fig. 5A). The Rb depletion does not coincide with the first appearance of orthopyroxene, but becomes pronounced well into the granulite facies. Thus, the breakdown of biotite generally described by the reaction of biotite + quartz to orthopyroxene + K-feldspar + H<sub>2</sub>O, does not, by itself, account for severe Rb depletion, though biotite is the principal carrier of Rb. Severe depletion occurs well before the appearance of garnet, though garnet is generated largely by the breakdown of biotite. Enough biotite remains in the rock to accommodate a significant amount of Rb. Fig. 5B shows that biotite itself becomes depleted in Rb with advancing metamorphic grade.

The question may be asked whether extraction of a silicate melt from gneisses undergoing anatexis may be the cause of Rb depletion. If this were the case, the partitioning of Rb between melt and biotite would have to strongly favor the melt, since the whole-rock

chemistry of the granulite-facies gneisses is little different from that of the amphibolite-facies gneisses (Hansen et al., 1995), requiring that amounts of extracted granitic liquid must have been small. However, analyses of coexisting biotite and glass in lavas (Mahood and Hildreth, 1983) show that the opposite is true. Biotite strongly concentrates Rb relative to a coexisting melt phase ( $Rb_{\text{bio}}/Rb_{\text{liq}} \sim 4.0$ ). Clearly, escape of a partial granitic melt could not account for the trend of Fig. 5B.

Thorium depletion begins in the South Indian Archean Craton when the metamorphic grade is quite advanced in the granulite facies (Fig. 5C). Hansen and Harlov (2007) identified the mineralogical control on the Th depletion as the sudden disappearance of monazite in the sequence of modal changes in the progressive metamorphism. As with Rb, a model in which U and Th are depleted in the lower crust by extraction of a granitic liquid may be explored. However, the major carriers of U, Th and the REE – zircon, apatite and monazite – are relatively insoluble in granitic liquids and would be concentrated in mafic restites (Watson, 1979; Watson and Harrison, 1983, 1984; Harrison and Watson, 1984; Rapp and Watson, 1986; Boehnke et al., 2013). Thus, anatexis is not an effective mechanism for causing U and Th depletion of the lower crust.

### 5. The problem of the extractability of granitic partial melt

The literature on the extractability of granitic partial melt from its source rocks is discouragingly inconsistent. Wickham (1987) calculated that efficient escape of the melt fraction of a partially molten rock would be virtually impossible. Thirty to fifty percent melting would result in diapirism of the whole melt-restite unit. This analysis was based on the high viscosity of even H<sub>2</sub>O-rich granitic melts at temperatures in the supposed anatectic range: 700–900 °C, and the small buoyancy of these liquids relative to the host rock. Using a different set of criteria, Connolly and Podladchikov (2007) calculated that large-scale segregation of basalt from its peridotite source might occur after only 20–25 vol% of melting. The temperature of this kind of partial melting would be much higher and the viscosity lower than for granitic liquids in crustal melting.

It has been argued that deformation during melt production will enable melt segregation in much smaller amounts (see Brown, 2001, 2007, 2013). Inger and Harris (1993) inferred that some leucogranites in the Himalayas represent low-fraction dehydration melts (~12%) that were completely removed from their muscovite–biotite source rocks by shear stresses in the Himalayan thrust zones. Sawyer (1994) proposed that, because of the directed stress factor, granitic melt in amounts of only 5% or less can be continually drained from a rock undergoing partial melting and deformation. Sawyer (1999) supported this by postulating effective microscale melt migration along grain boundaries. In this interpretation, the quartz and feldspar grain-boundary microveins noted in many granulites (Hansen et al., 1995; Harlov and Förster, 2002) represent the vestiges of a former intergranular melt, most of which was extracted in response to differential stress.

Two considerations highlight that the extraction of granitic liquids at such low melt fractions may not be the norm. First, there are numerous examples of terranes preserving voluminous amounts of segregated former granitic liquids that apparently did not leave the deep-crustal environment as liquids (Percival, 1991; Jung et al., 2000; Makitie, 2001; Redler et al., 2013). If melt is readily extractable in small volumes, the large amounts of leucosome in some migmatites (metatexites and diatexites) would have to represent either partial melts arrested in a state of incomplete escape or liquids from a foreign source arrested in transit through a terrane undergoing high grade metamorphism, with or without partial melting.

An interesting example of a granulite-facies terrane with a high percentage of leucocratic veins believed to be former granitic melts was described recently by Morfin et al. (2013). This terrane in the Archean Craton of northern Quebec is a complex of orthopyroxene- and garnet-bearing metagraywackes and metapelites with granitic dikes, pygmatic veins and veinlets as thin as a few millimeters that can be traced for meters, all occupying as much as 60% of the surfaces of outcrops. The authors termed the migmatitic terrane an “injection complex”. They postulated the intimate penetration of granitic sheets on all thickness scales into a terrane undergoing granulite-facies metamorphism and anatexis. The injected melts were presumably of low H<sub>2</sub>O content, in order to be compatible with granulite-facies wall rocks, yet capable of the most intimate penetration into, and replacement of, large tracts of the deep crust. In an even more extreme view of the penetrative ability of granite melts, Hasalová et al. (2008) advocated large-scale metasomatism by massive volumes of pervasively flowing intergranular melt.

The second problem with granite extraction at low melt fraction is the interpretation of the origin of the quartz and feldspar microveins. Quartzofeldspathic granulites commonly contain grain-boundary veinlets of quartz and feldspar lining garnet, plagioclase and orthopyroxene. Examples have been described from Finland (Perchuk and Gerya, 1992), SE Alaska (Todd and Evans, 1993; Harlov and Förster, 2002), S. India (Hansen et al., 1995), Sri Lanka (Perchuk et al., 2000), S. Norway (Harlov et al., 1998), the Ivrea Zone, Italy (Harlov and Förster, 2002), and the Furua Complex, Tanzania (Coolen, 1980). Such grain boundary features differ from microtextures of quartz and feldspar demonstrably resulting from melting (e.g., Vernon and Collins, 1988). Sawyer (1999) interpreted apparently similar veins as the residuum of a former interstitial melt phase, most of which was drained. However, Harlov (2012) noted that such veins occur in orthopyroxene-bearing rocks that display no evidence of having melted. Perchuk and Gerya (1992, 1993) described the chemical zoning of plagioclase, orthopyroxene and garnet against the veinlets in some granulites. The plagioclase rims against veins are more Ca-rich than the interiors of the crystals, but freezing of a grain-boundary granitic liquid would have resulted in the opposite zonation. Perchuk and Gerya (1992, 1993) argued that the veins were more likely to have resulted from the action of an aqueous fluid rich in alkalis and of low H<sub>2</sub>O activity, most likely a high *P*–*T* brine (e.g., Harlov and Förster, 2002). These examples highlight that grain-boundary microveins of quartz and feldspar can form by subsolidus fluid-rock interaction and do not necessarily represent the residue of granitic liquids. Even in granulites where plagioclase segregations are also present and for which partial melting has been inferred (Sawyer, 1999; Hartel and Pattison, 1996), infiltration of aqueous fluids should not be ruled out.

In sum, the question of the minimum melt fraction required to extract granitic liquid from its source still awaits definitive resolution. It has not been demonstrated conclusively that anatectic liquids formed by small degrees of partial melting can be extracted in quantities sufficient to form large granite masses.

## 6. The problem of the relation of granites to migmatites

### 6.1. Non-granitic leucosomes

Non-granitic veins and segregations of tonalite, K-feldspar-quartz, plagioclase-quartz, or nearly pure quartz are commonly reported in high-grade rocks. Many of these veins contain fresh orthopyroxene and are synmetamorphic. It is typically difficult to explain these non-granitic leucosomes by partial-melting or melt-segregation processes that are consistent with the local geological context.

There is controversy as to whether migmatitic leucosomes of tonalitic or trondhjemitic composition may be products of partial melting. Amit and Eyal (1976) described tonalitic leucosomes of migmatites in the Sinai Peninsula and considered them non-anatectic. The tonalitic leucosomes of the Skagit area, Washington, USA, were considered non-anatectic by Babcock and Misch (1989) but reinterpreted as at least partly magmatic by Whitney and Irving (1994). Morfin et al. (2013) regarded the granitic and tonalitic veins in their “injection complex” in northern Quebec as both igneous. Newton et al. (1998) found that mobilized and cross-cutting tonalitic leucosomes in the granulite-facies migmatites of southern Karnataka, India, had bulk compositions nearly identical to those of their host rocks. The conditions under which an intermediate quartzofeldspathic rock may yield tonalitic partial melts are not known. Quartz-K-feldspar veins containing orthopyroxene have been reported from the transitional granulite-facies terrane of S. India (Fig. 6).

Quartz veins in orthopyroxene-bearing rocks are common in high-grade terranes (examples: S. India, Newton et al., 1998; Fig. 6; Bamble, Norway, Van den Kerkhoff et al., 1994; and the Southern Marginal Zone of the Limpopo Complex, South Africa: Fig. 6). Although it is conceivable that quartz veins can be formed magmatically by flow differentiation or by closed-system metamorphic segregation, these mechanisms seem unlikely in the cited occurrences. More probably, the quartz segregations were formed by aqueous fluids of low H<sub>2</sub>O activity, perhaps with high salinity. Although the reduction of H<sub>2</sub>O activity generally decreases quartz solubility, the effect is lessened if due to the presence of dissolved halide salts (Newton and Manning, 2000, 2010; Shmulovich et al., 2001, 2006).

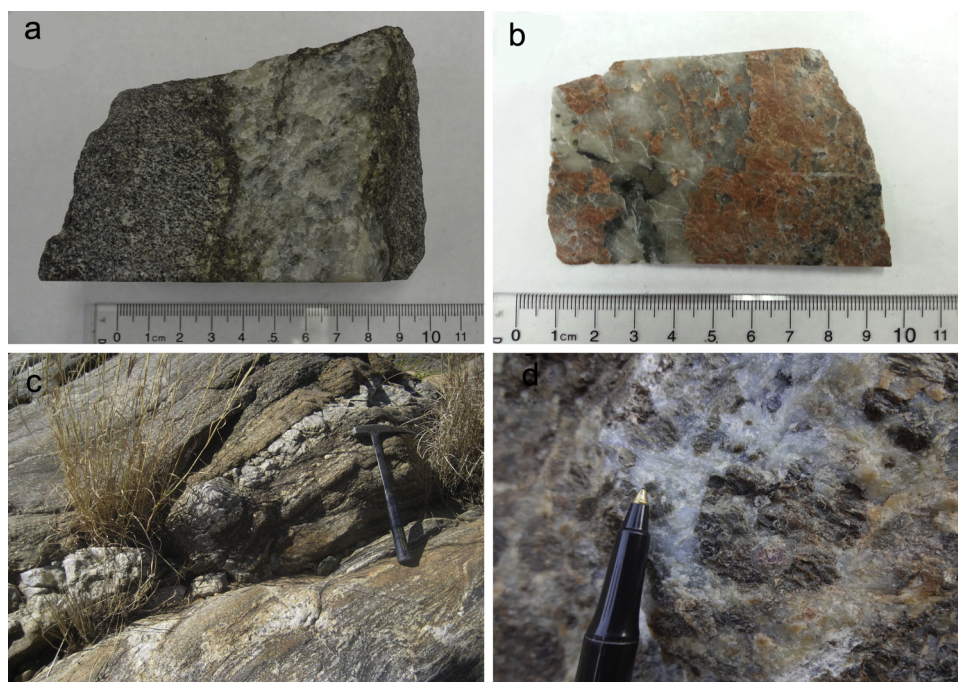
### 6.2. Partial melting of basic protoliths

Another challenge for the dehydration-melting model is the formation of granite leucosomes in mafic migmatites. In situ development of granitic migmatites within granulite-facies mafic gneisses has been documented in a number of Precambrian terranes (e.g., Eastern Transbaikalia: Gavrikova and Zharikov, 1984; Sharyzhgalskiy complex, Eastern Siberia: Kurdyukov and Berdnikov, 1987; Kapuskasing Structural Zone, Ontario: Hartel and Pattison, 1996; Lapland granulite belt, Baltic Shield: Korikovsky and Aranovich, 2010). Typical examples of migmatitic mafic gneisses from the two latter works are shown in Fig. 7. Any model of in situ melting must adequately address two major issues, melting temperature and melt composition at moderate melt fractions.

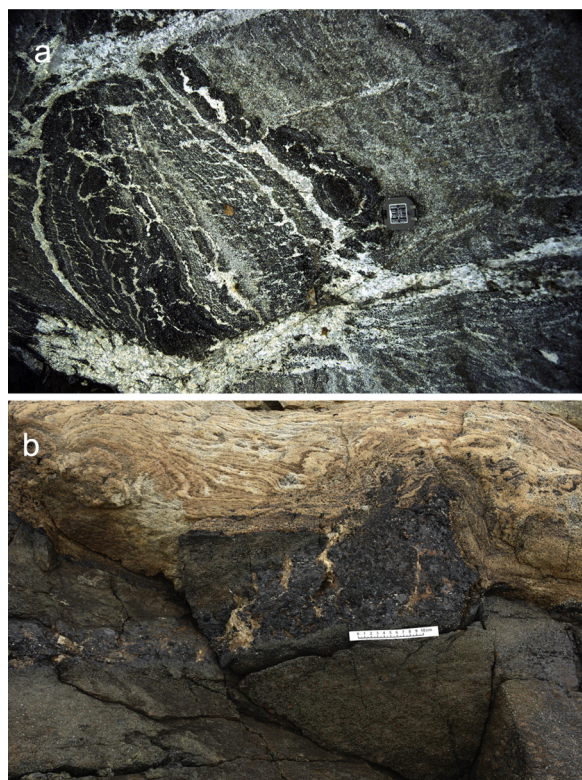
Temperature estimates from Fe–Mg exchange thermometry for the host gneisses almost always yield values in the range 720–780 °C, which is 50–100 °C lower than is required for dehydration melting of the amphibole-bearing basic rocks according to most experimental results at <10 kbar (Rutter and Wyllie, 1988; Beard and Lofgren, 1991; Rushmer, 1991; Winther and Newton, 1991; Wolf and Wyllie, 1994; Patiño Douce and Beard, 1995). This discrepancy is not explained by resetting of geothermometers during cooling. The Al-in-Opx thermometer is much less susceptible to the retrograde resetting (Anovitz, 1991; Aranovich and Berman, 1997) but also gives lower temperatures than required for dehydration melting, and correction of Fe–Mg–Al distribution for possible resetting in mafic granulites does not shift average temperatures above the dehydration-melting solidus (Pattison et al., 2003).

The discrepancy in temperature is exemplified by the calculated phase relations (Fig. 8) for the mafic Hbl + Opx + Cpx + Pl ± Bt gneiss hosting charnockitic leucosomes shown in Fig. 7B. Calculations used the gneiss composition from Korikovsky and Aranovich (2010) (Table 2) with initial H<sub>2</sub>O content 1.5 wt% (equivalent to about 35% amphibole + 7% biotite), and the Perple\_X software (Connolly, 2005) and thermodynamic data set for minerals and melts of White et al.



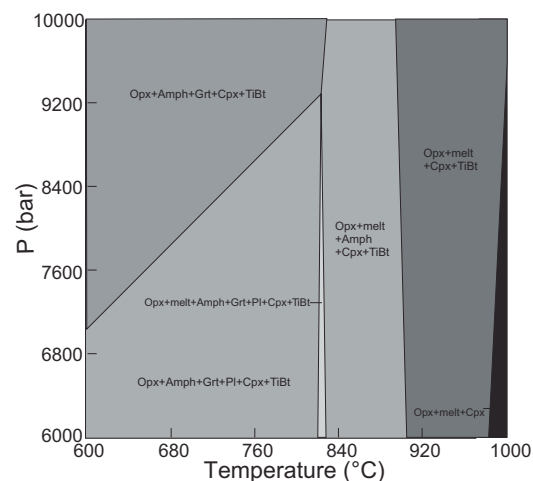


**Fig. 6.** Non-granitic granulite-facies leucosomes with orthopyroxene. (A) Bull quartz vein with selvages of orthopyroxene and K-feldspar, from Archean granulite-facies terrane near Halaguru, Karnataka, S. India (Hansen et al., 1984). (B) Quartz-K-feldspar vein in granite with a large orthopyroxene crystal, from the transitional granulite-facies interval near Satnur, Karnataka, S. India (Hansen et al., 1984). (C) Quartz vein with disseminated orthopyroxene in granulites of the Southern Marginal Zone of the Limpopo Belt, northern South Africa (present authors, July, 2013). (D) Large orthopyroxene crystals in vein of (C).



**Fig. 7.** Granitic migmatites in mafic amphibole-bearing granulites. (A) Kapuskasing Structural Zone, Ontario (Hartel and Pattison, 1996); compass is 7 cm across (photograph courtesy D. Pattison). (B) Lapland granulite belt, Baltic Shield (Korikovsky and Aranovich, 2010).

(2001). The phase compositions and mineral proportions are predicted reasonably well, except for the appearance of a small amount of garnet that was not observed in the sample. The calculated onset of melting at the expense of amphibole occurs at  $\sim 850^\circ\text{C}$ . This is in good agreement with experimental results on dehydration melting of amphibolite (Beard and Lofgren, 1991; Patiño Douce and Beard, 1995), but is  $70^\circ\text{C}$  higher than that estimated from mineral thermometry (Korikovsky and Aranovich, 2010), again highlighting the inconsistency between experimentally predicted melting temperature and that inferred from a natural example.



**Fig. 8.**  $P$ - $T$  pseudosection calculated for the  $\text{Hbl} + \text{Opx} + \text{Cpx} \pm \text{Bt}$  gneiss bulk composition (see Table 2), with 1.5 wt%  $\text{H}_2\text{O}$ , in the  $\text{Na}_2\text{O}-\text{K}_2\text{O}-\text{CaO}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{TiO}_2-\text{H}_2\text{O}$  system. Calculations employed *Perple.X* software (Connolly, 2005) and thermodynamic data set for minerals and melts of White et al. (2001). Opx, orthopyroxene; Amph, amphibole; Grt, garnet; Cpx, clinopyroxene; TiBt, Ti-biotite; Pl, plagioclase.



**Table 2**  
Comparison of observed and calculated composition of migmatites.

|                                | Mafic gneiss <sup>a</sup> | Charnockite <sup>a</sup> | Calculated <sup>b</sup> |
|--------------------------------|---------------------------|--------------------------|-------------------------|
| SiO <sub>2</sub>               | 50.55                     | 70.45                    | 67.28                   |
| TiO <sub>2</sub>               | 0.93                      | 0.48                     | –                       |
| Al <sub>2</sub> O <sub>3</sub> | 4.87                      | 13.47                    | 16.17                   |
| Fe <sub>2</sub> O <sub>3</sub> | 3.39                      | 0.74                     | –                       |
| FeO                            | 12.57                     | 2.76                     | 4.65                    |
| MnO                            | 0.28                      | 0.05                     | –                       |
| MgO                            | 15.19                     | 1.32                     | 1.42                    |
| CaO                            | 10.55                     | 2.65                     | 0.26                    |
| Na <sub>2</sub> O              | 1.06                      | 3.07                     | 8.17                    |
| K <sub>2</sub> O               | 0.60                      | 3.26                     | 2.05                    |

<sup>a</sup> Mafic Hbl + Opx + Cpx ± Bt gneiss host and charnockite leucosome (Fig. 7B), from Korikovsky and Aranovich (2010).

<sup>b</sup> Composition of 5 vol% melt derived from the mafic gneiss (Fig. 8), as calculated with Perple.X software (Connolly, 2005) and the White et al. (2001) dataset. Measured and normative compositions are on an anhydrous basis.

Another problem is the composition of the melt phase that is generated by dehydration melting of mafic amphibolites. The composition of the charnockitic leucosome in Fig. 7B is given in Table 2. This rock lies at the boundary of the granodiorite-granite field in a normative Ab–An–Or ternary (Barker, 1979), chiefly due to its high K<sub>2</sub>O content. The composition differs from those produced by experimental investigations of dehydration melting of mafic lithologies, regardless of bulk composition, temperature, pressure, H<sub>2</sub>O content, or melt fraction (Hacker, 1990; Beard and Lofgren, 1991; Rapp et al., 1991; Rushmer, 1991; Winther and Newton, 1991; Wolf and Wyllie, 1994; Rapp and Watson, 1995; Clemens et al., 2006; Qian and Hermann, 2013). The only experimental study in which similar melt compositions were produced was that of Patiño Douce and Beard (1995); however, this study employed a synthetic quartz-amphibole rock that was not intended to be representative of natural amphibolites. In short, even low-volume, near-solidus liquids in experimentally melted natural amphibolites differ from those illustrated in Fig. 7B.

As with the temperature discrepancy, calculated phase relations of the migmatite in Fig. 7B emphasize this problem. Table 2 gives the calculated melt composition at 5 vol% on an anhydrous basis (870 °C, Fig. 8), corresponding to the minimum amount of melt sufficient to segregate from the rock matrix (e.g., Brown et al., 1995). The calculated normative Ab–An–Or content corresponds to trondjemite (Barker, 1979). Although the melt model utilized for this calculation (White et al., 2001) was not optimized for mafic lithologies (for example, the CaO content is very low), the calculated composition is nevertheless K<sub>2</sub>O poor and more similar to experimentally derived liquids than to the observed K<sub>2</sub>O-rich leucosome. The substantial disagreement between observed leucosome composition and calculated or experimentally derived melts supports consideration of the participation of an externally derived fluid phase and solutes (e.g., K<sub>2</sub>O) in the melting process.

## 7. A role for fluids during deep-crustal melting?

The preceding sections have pointed out the serious problems that arise if the lower crust, as the source of granite by partial melting, and as a desiccated, LILE-depleted crustal foundation, is considered a closed system except for the extraction of granitic liquid. An open-system lower crust that can be metasomatically altered may be more consistent with an uppermost subcontinental mantle, which, according to many studies, has been subjected to metasomatic alteration (Stolz and Davies, 1989; Wooley, 1989; Barker, 1996). Such open-system behavior is integral to models of silicic magmatism in convergent margin, post-convergent, and within-plate settings (e.g., Hildreth, 1981, 2004; Hildreth and Moorbath, 1988; Wooley, 1989; Annan et al., 2006; Martin, 2006).

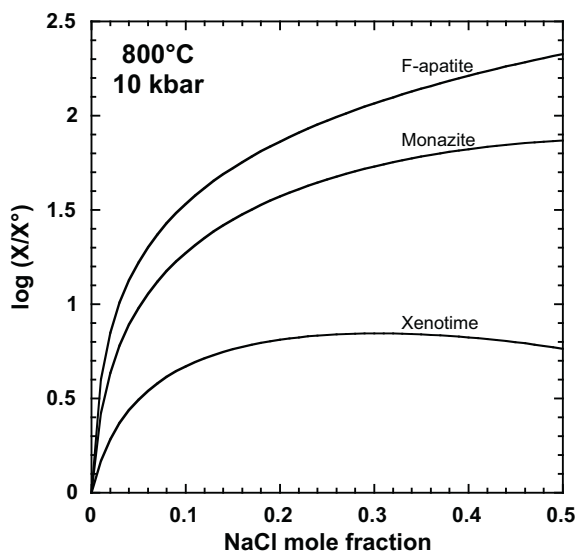
A model in which volatile-bearing, mantle-derived magma intruded into the deep crust is an important crust-modifying agent (e.g., Litvinovsky and Podladchikov, 1993) could explain some of the major problems of the genesis of the alkaline metaluminous granites. In addition to supplying H<sub>2</sub>O, and possibly granitic components, the source of heat for large scale melting of the crust is supplied by the latent heat of freezing of the basalt, which is forced to congeal because of loss of its H<sub>2</sub>O. The problem of decoupling of the major and trace elements is addressed by transfer of these components by magma-derived fluids. Enrichment of mantle-derived Sr and Nd into a crust-derived granite melt could have the result that the isotopes suggest a primitive mantle source of the granite. The upwardly departing granite is replaced by a nearly equal volume of basalt; the process implies substantial crustal accretion, as envisioned by Ewart et al. (1980). By this mechanism, a mafic lower crust is developed during the evolution of the continent interiors.

The roles of the volatile components CO<sub>2</sub> and Cl may be important in the above process. CO<sub>2</sub> is soluble in basaltic liquids at high pressures but less so at lower pressures. This promotes outgassing and crystallization of the basalt at depth (Holloway, 1976). A similar role has been advocated for Cl (Webster et al., 1999). These volatile components are much less soluble in granitic liquids than in basaltic liquids; hence, once liberated, they will be incorporated into metasomatizing fluids of low H<sub>2</sub>O activity. The fluids would inevitably split into carbonic and saline fractions as they cool and decompress (Johnson, 1991). If they are able to advect heat (Ganguly et al., 1995; Bachmann and Bergantz, 2006), alkalis and halogens, they could promote mid-crustal migmatization. Localization of partial melting in the mid-crust may be explainable by the positive *dP/dT* slopes of liquidus curves of hydrous granite in the presence of concentrated brines (Aranovich et al., 2013; Manning and Aranovich, 2014).

The saline fluid component would be capable of much more intimate penetration of grain boundaries of rocks than would a viscous silicate melt. The grain-boundary microtextures observed in many granulites can be explained by migrating saline solutions of low H<sub>2</sub>O activity (Perchuk and Gerya, 1992, 1993; Harlov and Förster, 2002). These brines could perform metasomatism, such as the development of scapolite observed in some high-grade terranes like the Furua Complex of Tanzania (Coolen, 1980) and the Western Gneiss Region of Norway (Austrheim, 2013). The high sulfate content of most granulite-facies scapolites may implicate sulfur as an important component of deep-crustal metamorphic fluids, with possible application to the highly oxidized granulite-facies terranes like Wilson Lake, Labrador (Cameron and Hattori, 1994). However, it has not yet been demonstrated experimentally that high-sulfate scapolite can form in the presence of chloride-rich fluid. In any case, localized metasomatism by these fluids may provide an alternative explanation for features often attributed to non-equilibrium partial melting under the dehydration melting hypothesis.

Antignano and Manning (2008) and Tropper et al. (2011) showed experimentally that apatite, monazite and xenotime, major carriers of radioactive and rare-earth elements, are orders of magnitude more soluble at high temperatures and pressures in concentrated saline solutions than in pure H<sub>2</sub>O (Fig. 9). Dissolution of monazite in hypersaline metamorphic fluids seems to be a feasible means of severe depletion of Th in deep-crustal terranes like the Archean Craton of S. India; whereas removal of granitic melt would have the opposite effect of concentrating Th (Boehnke et al., 2013). Hansen et al. (2002) showed that equilibration of biotite with a high pressure and temperature chloride solution would be effective in depleting Rb in granulites, while preserving the biotite; whereas melt removal would not.

In summary, many petrologic features of granulites, granites and migmatites that present conceptual problems for the dehydration-melting model for the generation of granites in the crust can be explained if the volatile components H<sub>2</sub>O, CO<sub>2</sub>, halogens and alkalis



**Fig. 9.** Experimental solubility of F-apatite (Antignano and Manning, 2008), monazite ( $\text{CePO}_4$ ) and xenotime ( $\text{YPO}_4$ ; Tropper et al., 2011) in high  $P$ - $T$  NaCl solutions, as mole fractions ( $X$ ) relative to those in pure  $\text{H}_2\text{O}$  ( $X^\circ$ ). The diagram shows that these carriers of REE and the radioactive elements can be mobilized by concentrated saline solutions. The phosphate minerals are very insoluble in silicate melts (Harrison and Watson, 1984; Rapp and Watson, 1986).

play significant roles as components of mineralizing fluids accompanying deep-crustal processes. As discussed by Manning and Aranovich (2014), plausible sources of such fluids include metamorphic fluids enriched in salts by fractional  $\text{H}_2\text{O}$  loss, crystallizing and degassing basaltic intrusions in the lower crust, and metasedimentary sources such as marbles and metaevaporites.

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