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FELDSPAR THERMOMETRY: A VALUABLE TOOL FOR DECIPHERING THE THERMAL HISTORY OF GRANULITE-FACIES ROCKS, AS ILLUSTRATED WITH METAPELITES FROM SRI LANKA

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Abstract

Two-feldspar geothermometry is applied to granulite-facies metapelites from the Highland Complex of Sri Lanka, for which very high peak temperatures of metamorphism have been inferred. Two-feldspar thermometry can be applied, even where only one feldspar has preserved its original bulk-composition in terms of Al–Si content, in spite of or because of unmixing to perthite or antiperthite. Different methods of integration of the unmixed feldspars to get the original bulk-composition are evaluated. Several types of feldspar assemblages are distinguished: (1) disequilibrium and near-equilibrium perthite – plagioclase pairs (several generations), (2) disequilibrium antiperthite – perthite pairs, (3) single mesoperthite, (4) near-equilibrium antiperthite–mesoperthite pairs. The highest temperatures, above 900°C, are derived from antiperthite porphyroclasts with evenly spaced exsolution lamellae and rods, from antiperthite – mesoperthite pairs, and from perthite inclusions in garnet porphyroclasts. The causes for the preservation of ternary feldspar compositions are discussed. Feldspar recrystallization in response to strong deformation is related to thrusting of the Highland Complex onto the Vijayan Complex. It occurred at about 830–900°C. In the main part of the Highland Complex, the feldspars recrystallized at *ca.* 680–760°C. Mylonitization along the Digana shear zone took place at about 710°C. Late retrograde recrystallization or growth of feldspar rims occurred in the range 460–590°C. Two-feldspar geothermometry combined with published geochronological data reveal new insights in the metamorphic evolution of the Highland Complex of Sri Lanka.

Keywords: feldspar thermometry, antiperthite, perthite, mesoperthite, metapelites, granulite facies, metamorphic evolution, Sri Lanka.

Sommaire

La géothermométrie fondée sur la coexistence de deux feldspaths a été appliquée aux métapélites recristallisées à une température apparemment très élevée, au faciès granulite, dans le socle dit Highland, au Sri Lanka. Cette approche se veut utile même là où un seul des deux feldspaths présents à l'origine subsiste avec sa composition globale Al–Si originelle, malgré les effets d'une exsolution pour donner soit une perthite, soit une antiperthite. On peut utiliser de différentes méthodes d'intégration des feldspaths exsolvés pour en obtenir la composition globale originelle. Plusieurs types d'assemblages sont présents: (1) paires perthite – plagioclase (en plusieurs générations), en déséquilibre ou presqu'en état d'équilibre, (2) paires antiperthite – perthite en déséquilibre, (3) grains isolés de mésoperthite, (4) paires antiperthite – mésoperthite près de l'état d'équilibre. Les températures les plus élevées, dépassant 900°C, sont dérivées de porphyroclastes d'antiperthite démontrant un espacement constant des lamelles et de bâtonnets d'exsolution, provenant de paires d'antiperthite et de mésoperthite, et d'inclusions de perthite dans des porphyroclastes de grenat. Les causes de préservation des compositions ternaires sont évaluées. La recristallisation des feldspaths accompagnant une forte déformation serait liée au chevauchement du socle Highland par dessus le complexe de Vijayan, à une température entre 830 et 900°C. Dans la partie principale du socle de Highland, les feldspaths ont recristallisé à environ 680–760°C. La mylonitisation le long de la zone de cisaillement de Digana a eu lieu à environ 710°C. Une recristallisation tardive rétrograde ou une surcroissance de feldspath s'est formée à environ 460–590°C. La géothermométrie fondée sur la coexistence de deux feldspaths, évalué à la lumière des données géochronologiques déjà dans la littérature, révèle des aspects nouveaux de l'évolution métamorphique du socle Highland du Sri Lanka.

Mots-clés: thermométrie des feldspaths, antiperthite, perthite, mésoperthite, métapélites, faciès granulite, évolution métamorphique, Sri Lanka.

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INTRODUCTION

Two-feldspar geothermometry improved by thermodynamic modeling of ternary feldspars (Fuhrman & Lindsley 1988), based on the experimental data by Seck (1971a, b), and further modified for the application to slowly cooled rocks by Kroll *et al.* (1993), is shown to be a valuable tool for deciphering the temperature development from high-grade to retrograde stages of metamorphism of granulite-facies rocks. Assuming that the Al–Si exchange between plagioclase and alkali feldspar ceases at high temperatures, whereas alkali exchange continues during cooling, a minimum temperature of feldspar equilibration or recrystallization can be derived. The composition of the feldspars that coexisted at the time of closure of intercrystalline Al–Si exchange may be calculated by reversing the K–Na exchange through shifts of the Ab,Or contents of both feldspars at constant An content until the equilibrium tie-line and the common isotherm on the ternary feldspar solvus are found (Kroll *et al.* 1993).

During cooling below the closure temperature for intercrystalline Al–Si exchange, the alkali feldspar exsolves and, by intracrystalline Al–Si and coupled Ca–K,Na exchange, the Ca partitions into the exsolved strings of plagioclase. The plagioclase grains, while exchanging K–Na with the exsolving alkali feldspar, may remain homogeneous, but may also exsolve and transform into antiperthite. In high-grade metamorphic rocks, antiperthite is rather common (*e.g.*, Heinrich 1956, Sen 1959); however, probably because of the commonly coarse and irregular exsolution-related bodies, few attempts have been made to determine the original composition of the ternary feldspar and the temperature of its formation (*cf.* Sen 1959, Lamb 1993, Snoeyenbos *et al.* 1995, Braun *et al.* 1996).

In the pelitic granulites of the Highland Complex of Sri Lanka (Fig. 1), antiperthite with evenly spaced rods and lamellae of K-feldspar is widespread. Bulk compositions of the antiperthite indicate very high peak temperatures of metamorphism. Evidence for very high temperatures (about 900°C) in different parts of the Highland Complex is provided by the occurrence of inverted pigeonite and other exsolved high-temperature pyroxenes (Schenk *et al.* 1988, 1991), as well as the paragenesis sapphirine + quartz (Osanai 1989). High estimated temperatures also have been obtained from alkali feldspar that recrystallized after near-peak granulite-facies deformation in the eastern part of the Highland Complex (about 850°C, Voll *et al.* 1994) and from garnet–orthopyroxene geothermometry (700–900°C, Schumacher *et al.* 1990, Faulhaber & Raith 1991, Schumacher & Faulhaber 1994).

This study applies two-feldspar geothermometry to different types of disequilibrium and near-equilibrium feldspar pairs in metasedimentary rocks from the Highland Complex of Sri Lanka. The highest temperatures are derived from the bulk composition of antiperthite with evenly spaced exsolution lamellae and rods. Further, pairs of ternary feldspars that unmixed into antiperthite and mesoperthite have been found to lie on a common isotherm near 900°C. The causes for the preservation of high-temperature ternary feldspar compositions are discussed.

GEOLOGICAL SETTING, METAMORPHISM, AND P–T PATH

The Highland Complex of Sri Lanka constitutes a broad high-grade metamorphic belt through the central part of the island, bordered by two granite–gneiss terranes (Fig. 1). It consists of a layered sequence of metasedimentary and concordant metamorphosed igneous rocks. In most parts of the Highland Complex, metapelites, metapsammites, quartzites, marbles, and calc-silicate rocks are interlayered with charnockites and metabasites. The present paper is focussed on the metapelites of the Highland Complex. In the western and southwestern part, migmatitic garnet- and cordierite-bearing metapelites and charnockitic gneisses predominate. In the eastern part, especially above a near-horizontal thrust boundary against the Vijayan Complex (Fig. 1), metasedimentary as well as magmatic rocks were strongly deformed, with development of a strong flattening-induced foliation, stretching lineation, and isoclinal folds (Kröner *et al.* 1994b, Voll & Kleinschrodt 1991, Kleinschrodt 1994). The whole layered sequence was folded into km-scale upright open folds coaxial with the older stretching lineation. Flattening, stretching and folding essentially occurred during granulite-facies conditions. In response to high-grade deformation, the feldspar matrix of the rocks recrystallized thoroughly, and large plates of quartz developed. Single crystals of platy quartz are 0.2–1 mm thick and 5–25 mm in length, indicating a flattening of previously isometric grains to about 1/25th of their original thickness (Voll & Kleinschrodt 1991). In localized shear zones, ductile deformation continued to lower temperatures.

In spite of the strong deformation and recrystallization of the rocks, relics of older cm-size unmixed feldspar and orthopyroxene grains are found and give evidence for very high temperatures of peak metamorphism (about 900°C: Schenk *et al.* 1988, 1991, Raase & Schenk 1994). The preliminary results given by Raase & Schenk (1994) on relict antiperthite and mesoperthite from metapelites are elaborated in the present study. The retrograde P–T path in the Highland Complex is characterized by isobaric cooling from very high temperatures prior to uplift (Schenk *et al.* 1988, 1991, Schumacher *et al.* 1990). This is indicated by garnet + clinopyroxene coronas mantling large grains of orthopyroxene and plagioclase, and later orthopyroxene – plagioclase symplectites developed between garnet and clinopyroxene in metabasites from the eastern Highland Complex. On the other hand,

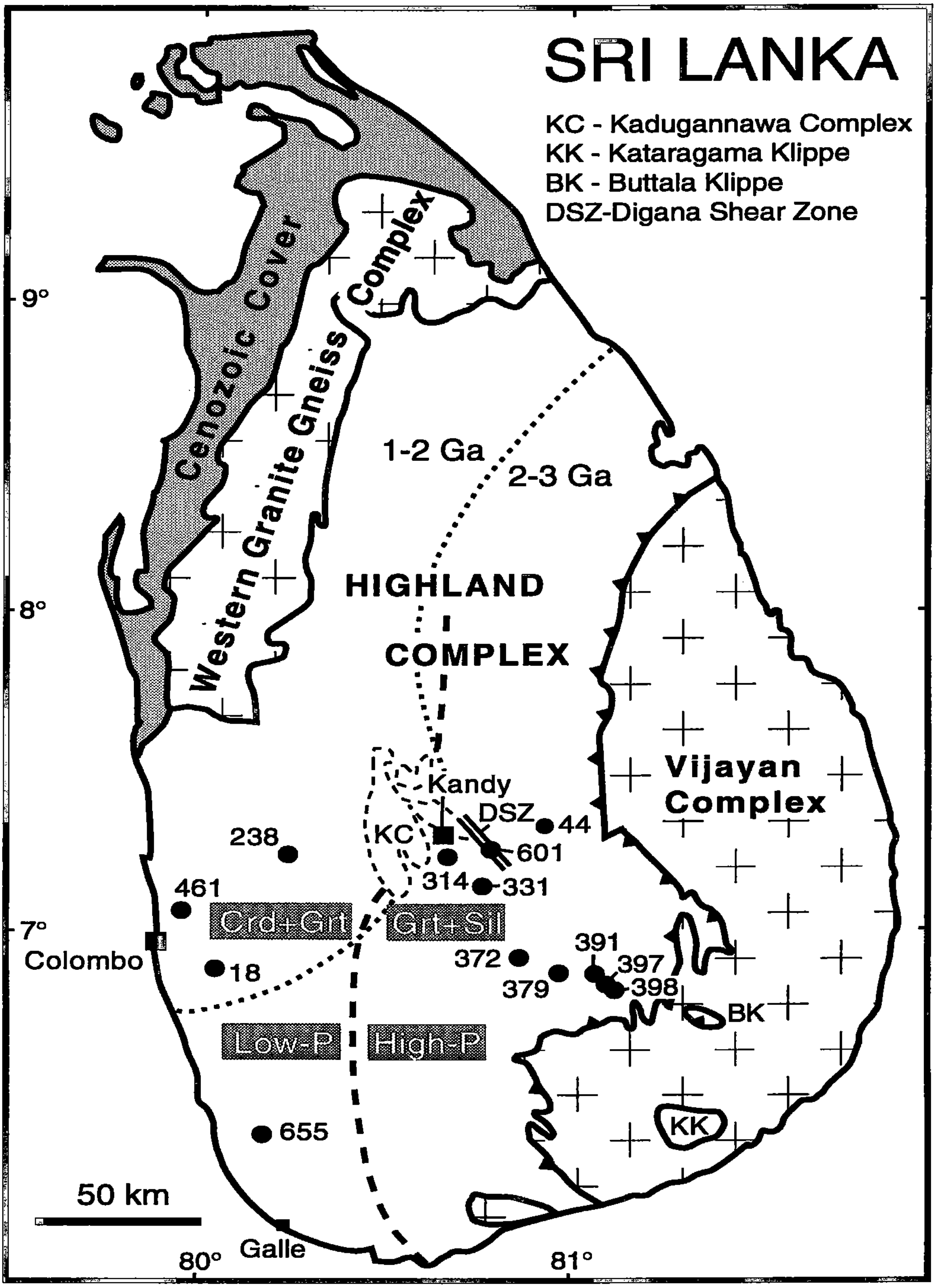

FIG. 1. Simplified geological map of Sri Lanka showing the major lithotectonic units and the sample localities. The Highland Complex is divided into an eastern high-pressure part (>7 kbar) with garnet–sillimanite-bearing khondalitic metapelites, and a western low-pressure part (<7 kbar) with garnet–cordierite gneisses (heavy dashed line: Raase & Schenk 1994). The eastern border of <2 Ga Nd model ages according to Milisenda *et al.* (1994) is given by a dotted line. Other boundaries of lithotectonic units according to the Geological Map of Sri Lanka (1982).

patterns of inclusions in garnet porphyroblasts in metapelites were interpreted by Hiroi *et al.* (1994) to have formed during prograde decompression, whereas Raase & Schenk (1994) inferred prograde increase in pressure followed by temperature increase. Further, postkinematic rims of garnet around sillimanite, formed through the reaction biotite + sillimanite + quartz $\rightarrow$ garnet + K-feldspar, indicate a temperature increase following pervasive deformation (Raase & Schenk 1994). Aware of these conflicting lines of evidence, Kriegsman (1993, 1996) interpreted the garnet – clinopyroxene coronas in metabasites to have formed during prograde increase in pressure. However, this proposal would imply that the strong flattening-induced deformation of the rocks occurred early during the prograde stage, which is in contrast to the results of Voll & Kleinschrodt (1991); furthermore, it would deny the existence of a very-high-temperature stage prior to the growth of the garnet – clinopyroxene coronas.

In the western part of the Highland Complex, garnet – quartz coronas around orthopyroxene in charnockitic rocks (Schumacher & Faulhaber 1994) and garnet – sillimanite rims separating spinel and quartz (Raase & Schenk 1994) indicate near-isobaric cooling. A later corona of cordierite around garnet indicates subsequent uplift (Raase & Schenk 1994). Geothermobarometry on metabasic and metapelitic assemblages suggests significantly lower pressures but similar temperatures in the west (4–7 kbar, 600–850°C) as in the east (7–10 kbar, 700–900°C: Schumacher & Faulhaber 1994, Raase & Schenk 1994). The spread in the P–T estimates can be attributed in part to variations reflecting different stages of the cooling path where cation exchange was frozen in, and in part to problems inherent in the geothermobarometric models (*i.e.*, inaccurate activity models for the coexisting minerals and local retrograde cation-exchange, which affected different minerals to a different extent).

On the basis of Nd model ages, the Highland Complex has been divided by Milisenda *et al.* (1988) into an older (2–3 Ga) eastern–southwestern part, and a younger (1–2 Ga) northwestern part ("Wanni Complex" according to Kröner *et al.* 1991), which includes the Western Granite Gneiss and the Kadugannawa complexes (Fig. 1). Most of magmatic activity occurred at 1.8–1.9 Ga in the eastern and at 0.8–1.1 Ga in the western part (Hölzl *et al.* 1994). Sedimentary and igneous rocks of the two different Nd age provinces were metamorphosed at about 610 Ma (Hölzl *et al.* 1994) and cooled to the closure temperature for Rb–Sr exchange in biotite at about 460 Ma (Hölzl *et al.* 1991).

Characterization of the Feldspars

In spite of processes of deformation, recrystallization, cation exchange and exsolution, which modified the original feldspars in the metapelites and semipelites from the Highland Complex, several types of feldspar assemblages can be distinguished: (1) recrystallized perthite and plagioclase (several generations), (2) antiperthite porphyroclasts located beside recrystallized perthite and plagioclase, (3) single grains of mesoperthite, (4) antiperthite and mesoperthite (a) in various layers of banded metapelite, and (b) as porphyroclasts set in a recrystallized matrix of perthite and plagioclase. Each type is described below.

Recrystallized perthite and plagioclase

This first type of feldspar assemblages is the most common, and several generations of perthite and plagioclase may be distinguished within one specimen of metapelite.

In strongly flattened garnet – sillimanite gneisses (khondalites) near the thrust boundary toward the Vijayan Complex, feldspars recrystallized in response to the main high-temperature deformation. This thrust-related deformation most probably occurred prior to the peak of granulite metamorphism, since thin prograde reaction rims of garnet and alkali feldspar, which developed between sillimanite and quartz, have not been sheared off (Raase & Schenk 1994). Finely laminated samples of metapelite closest to the thrust boundary display this reaction texture and are almost unaffected by retrograde alteration, except for exsolution of the alkali feldspar and late recrystallization on grain boundaries. In these rocks, perthite and plagioclase display a granoblastic–polygonal texture with high-angle grain boundaries (Fig. 2a). Perthite and plagioclase occur in separate lenses or layers, suggesting recrystallization of larger primary grains. Grains of alkali feldspar consist of string perthite, whereas plagioclase is apparently not unmixed. Exsolution of coarse strings of plagioclase in the alkali feldspar occurred during cooling, probably enhanced by late-stage deformation. Further, grain boundaries of alkali feldspar locally became indented, and a late generation of very fine-grained feldspar crystallized (Fig. 2b). This fine-grained alkali feldspar lacks coarse strings, but later exsolved very fine strings and films oriented parallel to the commonly observed Murchison plane (near $\bar{6}01$). Film perthite is also found in between the primary coarse strings and at the rim of the coarser grains (Figs. 2b, 3b, 4b). Fine-scale strings and films are perfectly coherent and exsolved at 300–350°C in a metastable disordered structural state (Evangelakakis *et al.* 1993).

In most parts of the eastern and western Highland Complex, the laminated texture of the metapelites, formed during early granulite-facies metamorphism, is more or less obscured by grain coarsening due to static annealing (Voll & Kleinschrodt 1991), later deformation, and recrystallization of the feldspar matrix. Coarse string perthite is rare in such samples

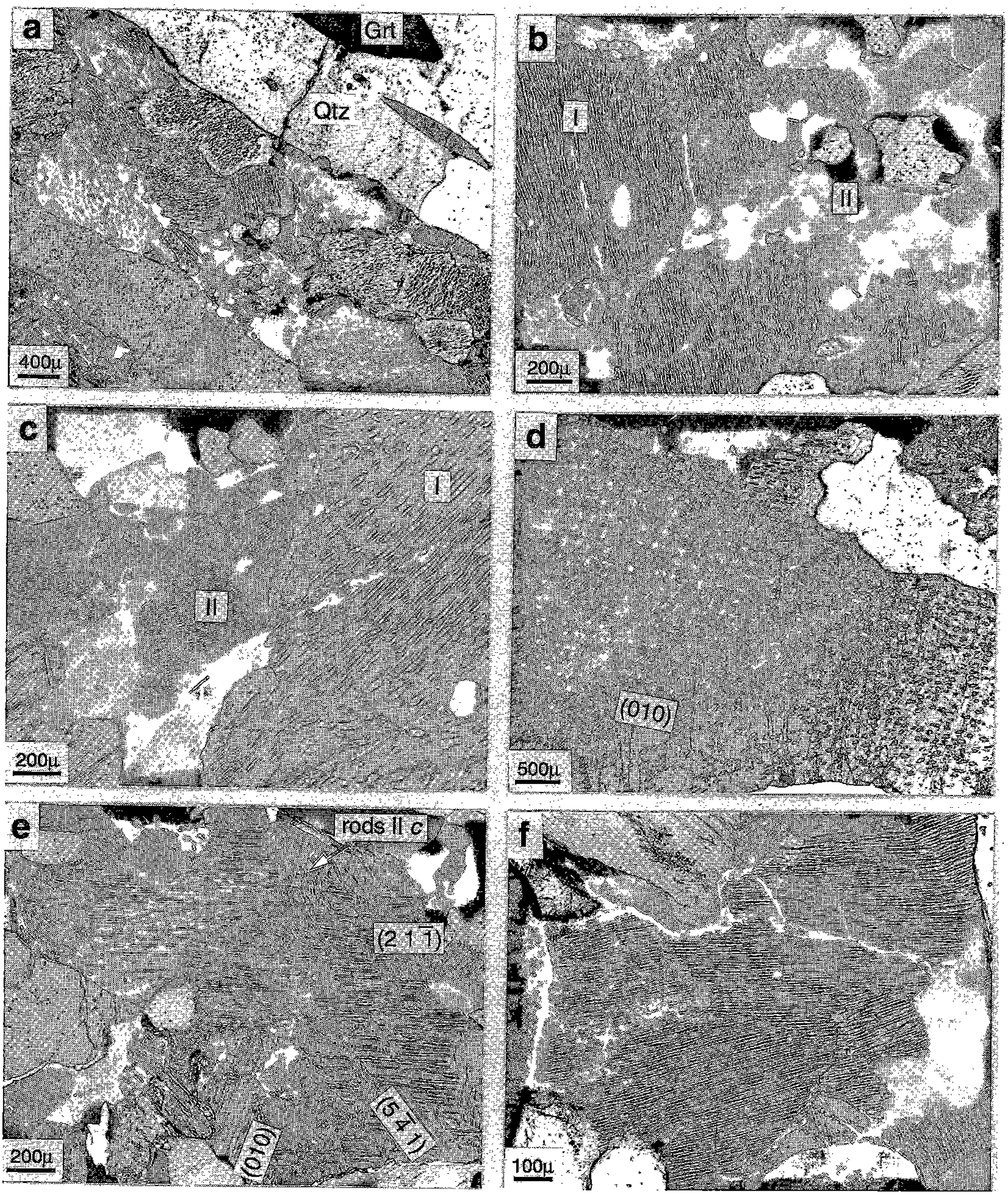

FIG. 2. Photomicrographs showing textures of exsolved feldspars in metapelites from the Highland Complex. (a) Polygonal aggregate of string perthite recrystallized in response to strong deformation at granulite-facies conditions. Laminated garnet – sillimanite gneiss from the basal part of the Highland Complex. Sample 398, oblique polars. (b) Coarse string perthite (I) and low-temperature products of recrystallization (II). Metapelite from the southeastern Highland Complex. Sample 397, crossed polars. (c) Coarse-grained leucosome perthite (I) with locally coarsened strings and recrystallized perthite (II) with very fine strings. Migmatitic garnet – cordierite gneiss from the south. Sample 655, crossed polars. (d) Exsolved K-feldspar rods cutting perpendicular to the *c* axes of an antiperthite porphyroclast are rather evenly distributed though obviously nucleated at albite twin lamellae. Garnet – cordierite gneiss, western Highland Complex. Sample 18, crossed polars. (e) Antiperthite showing two sets of K-feldspar exsolution lamellae, approximately parallel to $(21\bar{1})$ and $(54\bar{1})$, local rods parallel to *c*, and albite twin lamellae. Garnet – cordierite gneiss, western Highland Complex. Sample 461, crossed polars. (f) Polygonized mesoperthite exsolved after deformation and recrystallization. Strongly flattened, laminated garnet – sillimanite gneiss, eastern Highland Complex. Sample 331, oblique polars.

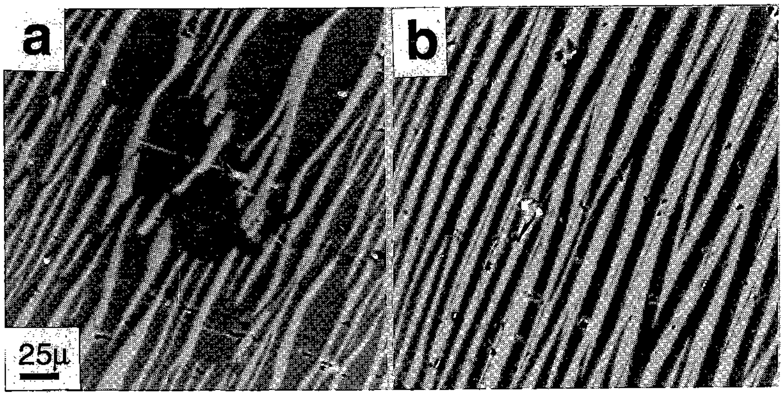

FIG. 3. Back-scattered electron images of ternary feldspars from two adjacent layers in a sillimanite – biotite – garnet gneiss (sample 314): (a) antiperthite with 33 vol.% of K-feldspar lamellae (light) and locally coarsened plagioclase blebs; (b) mesoperthite with 44 vol.% of exsolved plagioclase lamellae (dark) and K-feldspar lamellae (light) containing fine films of secondary albite.

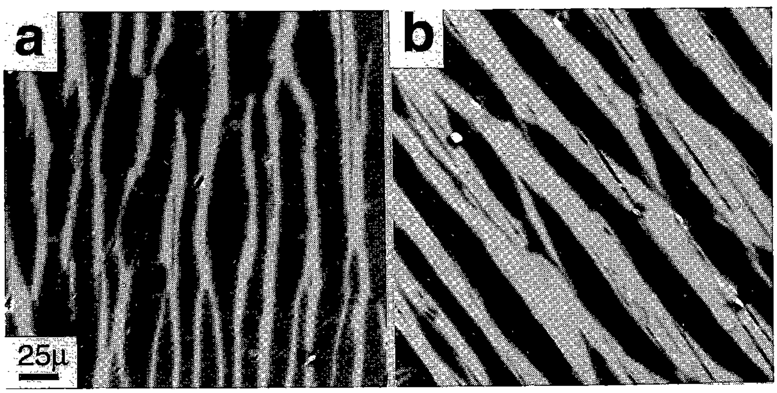

FIG. 4. Back-scattered electron images of relict grains of perthite and antiperthite from a garnet – sillimanite gneiss (sample 391a): (a) antiperthite with bifurcating K-feldspar lamellae (34 vol.%); (b) perthite with coarse plagioclase strings (41 vol.%), and fine films of secondary albite (dark).

and largely replaced by alkali feldspar containing very fine strings and films. Coarse string perthite is only preserved in coarse feldspar in leucosome horizons (Fig. 2c).

Antiperthite

Antiperthite porphyroclasts, though observed as relics in all parts of the Highland Complex, are most common in the migmatitic plagioclase-rich cordierite – garnet gneisses from the western part. These rocks are deformed and partly recrystallized to perthite and plagioclase, but large grains of antiperthite with a rather homogeneous exsolution texture are still preserved. Exsolution textures in antiperthite are quite variable, with orthoclase rods and lamellae of different orientations (Figs. 2d, e). Rods and lamellae of the K-rich phase are generally <20 µm in thickness, but locally coarsened exsolution-induced bodies occur as well. The exsolved rods and lamellae nucleated at twin boundaries or other crystal imperfections but, in large grains with evenly spaced albite twin lamellae, the exsolution bodies are quite homogeneously distributed (Figs. 2d, e). Optical studies show that all the exsolved rods and lamellae of K-feldspar within a plagioclase host have the same lattice orientation. However, three distinctly different sets of orientation have been found from X-ray precession photographs of antiperthite from Sri Lanka (Evangelakakis *et al.* 1991). Unmixing must have occurred after deformation of the rocks, since the narrowly spaced and commonly wedge-shaped and bent twin lamellae most probably formed in response to this deformation.

Single mesoperthite

Mesoperthite with fine exsolution-induced lamellae (0.5–5 μm wide) is found as the only feldspar in a strongly flattened sillimanite – garnet gneiss in the central part of the Highland Complex (sample 331, Fig. 1). In response to deformation, the fine-grained alkali feldspar in this sample of khondalite is polygonized, with subgrain boundaries (Fig. 2f), and partly recrystallized, with high-angle boundaries. Unmixing occurred after deformation and recrystallization at relatively low temperatures and inferred dry conditions, in view of the homogeneous exsolution-induced texture and the narrow spacing of the lamellae. Coarsening of the exsolution lamellae to about 5 μm is observed near the high-angle boundaries as well as near subgrain boundaries, whereas in the core of the subgrains, the width of the lamellae grades to a submicroscopic size. Clear rims devoid of exsolution lamellae locally developed at high-angle boundaries, but rarely at subgrain boundaries.

Antiperthite and mesoperthite

Similar mesoperthite to that in sample 331 was found in one layer of a banded metapelite (sample 314). In an adjacent layer, antiperthite occurs next to recrystallized plagioclase without an exsolution texture. The antiperthite, in some grains or parts of grains, displays a similar lamellar exsolution-induced texture to that in the mesoperthite, but with a higher proportion or rather greater width of the plagioclase lamellae compared to the lamellae of the K-rich phase (Figs. 3a, b). In other grains, coarsening obviously occurred, yielding lensoid bodies of the plagioclase phase and thickened lamellae of the K-rich phase adjacent to the lensoid bodies (Fig. 3a).

In spite of strong deformation recorded in the metapelites, rare porphyroclasts of antiperthite are preserved near the eastern thrust-boundary (sample 391a). The porphyroclasts display lamellar patterns of exsolution similar to those in mesoperthite. Although the proportions of plagioclase and of the exsolved K-feldspar phase are roughly 2:1, the exsolution lamellae of plagioclase tend to have a convex outline as in perthite, whereas the K-rich phase forms thin bifurcating lamellae (Fig. 4a). The feldspar matrix of the garnet – sillimanite gneiss consists of fine-grained perthite with only a few porphyroclasts of antiperthite and a few larger grains of perthite. The latter, which contain a high percentage of lensoid exsolution bodies of plagioclase (Fig. 4b), may have coexisted with the antiperthite during the peak of metamorphism, before unmixing and before deformation.

Feldspar Geothermometry

Re-integration of exsolved feldspars

Re-integration of the original composition of perthite, mesoperthite, and antiperthite was done by two different methods. Firstly, in the case of fine-scale exsolution textures (lamella spacing <5 μm) in perthite and mesoperthite, re-integration was done by doing a sufficiently large number of electron-microprobe analyses with a slightly defocussed beam, which automatically scanned an area of about 20 × 20 μm. Traverses across exsolved grains were done manually, avoiding imperfections in the polish and inhomogeneously exsolved grain parts, especially at margins. Averaging the results of 10 to 20 analyses was generally sufficient to give reproducible results. However, this microprobe re-integration technique involves a systematic matrix-effect error. During the scan, the microbeam excites areas of different composition in the grain of exsolved alkali feldspar. Although different compositions require different absorption, fluorescence and atomic number corrections, the computer program calculates corrections based on a non-existent homogeneous phase. The resulting error in the averaged composition may be small, as suggested by Bohlen & Essene (1977). However, a comparison of the results obtained by this technique with those obtained by the second technique applied, image analysis, shows that the bulk compositions of the alkali feldspar obtained by the first method are too low in Ca. In order to assess this error, the area for the microprobe scanning analysis was chosen to cover two portions of lamellae of exactly the same size in a coarsely unmixed grain of mesoperthite. A comparison of the result of this analysis (48.3% Ab, 45.1% Or, 6.6% An) with the bulk composition (47.7% Ab, 44.3% Or, 7.3% An) calculated from the separately analyzed compositions of the two feldspar phases, suggests a relative error due to matrix effect of about 10% for the An content and 2% for the Or content. This error in An content of the alkali feldspar yields two-feldspar temperatures that are too low by about 10–20°C. On the other hand, a relative error of 2% in Or content of a grain of antiperthite will give higher temperatures by about 5–10°C. The errors in proportion of Ab–An components of the plagioclase and of Ab–Or components of alkali feldspar are insensitive to the calculated two-feldspar temperature, since the solvus isotherms are nearly parallel to Ab–An and Ab–Or joins of the ternary feldspar system.

Secondly, the composition of the unmixed phases of sufficiently coarse grains of perthite and antiperthite was analyzed by electron microprobe using a slightly defocussed beam in order to prevent Na evaporation, and in order to re-integrate very fine second-stage strings or lamellae. Then, the volume percentage of the exsolution bodies and host was estimated by computer image analysis of back-scattered electron (BSE) images or images of K, Ca, or Na distribution in selected sections of grains cut approximately perpendicular to the exsolved lamellae, rods or spindles. Very fine lamellae of secondary albite were generally included in the volume percentage of the K-rich phase, except for one sample of perthite (398), in which strings of primary and secondary plagioclase and host K-feldspar were analyzed separately. In order to take into account any change of lamella thickness through image processing, the thickness of some lamellae or strings in the original BSE image was compared with the corresponding thickness in the final analyzed image obtained after contrast enhancing. The volume percentages of lamellae and host were converted to weight percentages using densities from Tröger (1971) for plagioclase, and from Smith & Brown (1988) for alkali feldspar. The bulk molar composition was obtained from the weight percentages and the composition of the exsolved phases.

The second method is favored, since it largely avoids the matrix-effect error described earlier. Furthermore, although chemical heterogeneities in the K-rich phase due to late secondary exsolution may result in an error in the K–Na ratio of the bulk alkali feldspar, this does not yield a significant error in the two-feldspar temperature, since the isotherms are nearly parallel to the Ab–Or join (see below). The error in An content, which could have a significant effect on the temperature calculated, must be small, because the integrated secondary exsolution lamellae are Ab-rich and An-poor, so that the An content of the bulk alkali feldspar is mainly determined by the An content (and volume percentage) of the strings of primary exsolved plagioclase. A variation in the An content of the plagioclase strings has not been detected, although it was not possible to make point analyses very close to the rim of the strings, since the electron beam must be slightly defocussed in order to prevent Na evaporation.

In case of antiperthite, some inhomogeneities in An content of the plagioclase host were detected, with an increase of up to 5% An near the K-feldspar exsolution-induced bodies (samples 18 and 238). The error in bulk An content of the integrated antiperthite should, however, be less than 2%, resulting in only insignificant errors in inferred temperature.

The largest error in composition and derived temperature is clearly due to heterogeneities of the exsolution texture, although much care was taken to avoid the rim portion of the grains and domains with a coarsened exsolution texture.

Disequilibrium perthite–plagioclase pairs

Table 1 lists the integrated bulk-composition of the alkali feldspar perthite in terms of mol% Ab, Or, An and Cn, the composition of coexisting unexsolved plagioclase, the recalculated compositions obtained by reversing the intercrystalline K–Na exchange according to Kroll *et al.* (1993), and the two-feldspar temperature values calculated with Margules parameters from Fuhrman & Lindsley (1988), Lindsley & Nekvasil (1989), and Elkins & Grove (1990). Recalculation was done by aid of the computer programs F–THERM, L–THERM, and E–THERM (Kroll *et al.* 1993), which are modified versions of M–THERM3 by Fuhrman & Lindsley (1988). The temperature data represent the closure temperatures for intercrystalline Al–Si exchange and may thus be regarded as minimum temperatures. However, the calculated temperatures should be close to temperatures of peak metamorphism, since Al–Si exchange is very slow, even at high temperatures (Grove *et al.* 1984, Yund 1986, Liu & Yund 1992), where dry conditions typical of granulite-facies rocks prevail.

The celsian component (Cn) is very low in the plagioclase, but amounts up to 1.2 mol% in the alkali feldspar. Two-feldspar temperatures computed with Cn according to Fuhrman & Lindsley (1988) are higher by up to 27°C than those ignoring Cn. Temperatures calculated with Margules parameters (*W*) of Lindsley & Nekvasil (1989) are generally lower by about 10–30°C, and in some cases, by more than 50°C, than those calculated with *W* values from Fuhrman & Lindsley (1988). In contrast, two-feldspar temperatures calculated with *W* values from Elkins & Grove (1990) are higher by about 40–50°C (Table 1). Since the Fuhrman & Lindsley (1988) temperatures corrected for K–Na exchange (excluding Cn) compare quite well with temperatures estimated from garnet – orthopyroxene thermometry (Schumacher & Faulhaber 1994), these two-feldspar temperatures are considered in the text that follows.

The analyzed and calculated compositions of *recrystallized matrix feldspars* in the metapelites are plotted in An–Ab–Or diagrams (Figs. 5a, b). In the strongly flattened garnet – sillimanite gneisses from the lowermost tectonostratigraphic part of the Highland Complex in the southeast (samples 391a,b, 397, 398, Fig. 1), two-feldspar temperatures for core compositions of perthite and plagioclase are in the range 835–900°C (Table 1). As shown in Figure 5a, the analyzed plagioclase and alkali feldspar compositions do not plot on the same isotherm, suggesting that the plagioclase lost most of its Or content, whereas the perthite retained relatively high An contents. By reversing the K–Na exchange, the plagioclase becomes much more Or-rich, whereas the Ab component of the alkali feldspar changes only little. This is due to the low ratio of plagioclase to alkali feldspar in the metapelites. The

TABLE 1. COMPOSITIONS (ANALYZED AND CALCULATED) AND TEMPERATURES FOR FELDSPAR PAIRS FROM PELITIC GRANULITES OF THE HIGHLAND COMPLEX, SRI LANKA

Sample	Method	Feldspar	Composition				C (FL+Cn)		C (FL)		C (LN)		T(FL+Cn)	T (FL)	T(LN)	T(EG)	P
			Ab	Or	An	Cn	Ab	Or	Ab	Or	Ab	Or	(°C)	(°C)	(°C)	(°C)	(kbar)
		Feldspars from strongly flattened garnet-sillimanite gneisses															
391a	A	Pl core	83.3	1.3	15.4	-	67.6	17.0	62.2	18.4	72.2	12.4					
	B	Af core	32.8	62.5	4.4	0.3	39.6	55.7	38.8	56.8	41.2	53.4	890	900	841	935	10
	A	Pl rim	83.9	0.9	15.2	-	83.2	1.6	83.1	1.7	83.8	1.0					
	A	Af rim	11.0	88.4	0.1	0.5	11.7	87.7	11.8	88.1	8.6	91.3	449	456	380	478	4
391b	A	Pl core	75.5	1.8	22.6	0.1	64.9	12.4	64.4	13.0	67.6	9.8					
	B	Af core	22.7	73.0	3.8	0.5	29.3	66.4	29.2	67.0	31.6	64.6	853	863	824	911	10
397	A	Pl core	78.1	1.4	20.4	0.1	65.6	13.9	64.5	15.1	68.7	10.9					
	B	Af core	31.3	63.8	4.3	0.6	33.0	62.1	32.6	63.1	35.3	60.4	872	886	841	932	10
	A	Pl rim	80.4	1.2	18.4	-	75.6	6.0	75.3	6.3	76.9	4.7					
	A	Af rim1	24.8	73.8	1.0	0.4	25.0	73.6	25.1	73.9	23.7	75.3	677	684	625	712	5
	A	Pl rim	80.4	1.2	18.4	-	78.3	3.3	78.1	3.5	79.1	2.5					
	A	Af rim2	18.5	80.6	0.4	0.5	18.4	80.7	18.5	81.1	16.2	83.4	568	576	512	601	4
398	A	Pl core	77.7	1.8	20.5	-	67.9	11.6	67.5	12.0	70.8	8.7					
	B	Af core	26.7	70.0	3.0	0.3	29.2	67.5	29.0	68.0	30.2	66.8	831	835	786	877	10
	A	Pl rim	78.9	1.2	19.9	-	78.2	1.9	78.1	2.0	78.6	1.5					
	A	Af rim	10.9	88.6	0.2	0.3	12.8	86.7	12.9	86.9	11.1	88.7	491	496	438	522	4
601	A	Pl incl.	69.9	5.8	24.2	0.1	59.1	16.6	58.6	17.1	63.0	12.7					
	B	Af incl.	25.5	68.1	6.1	0.3	32.1	61.5	31.8	62.1	36.5	57.4	925	931	905	991	12
	A	Pl matrix	70.7	1.2	28.1	-	66.5	5.4	66.1	5.8	67.1	4.8					
	A	Af matrix	18.7	79.0	1.6	0.7	20.9	76.8	21.0	77.4	21.1	77.3	703	716	683	760	6
		Feldspars from metapelite leucosomes															
379	A	Pl	68.2	2.3	29.5	-	61.9	8.6	60.9	9.6	62.4	8.1					
	B	Af leuc.	19.8	76.0	3.4	0.8	24.6	71.2	24.5	72.2	27.8	68.8	808	826	809	883	8
	A	Pl	68.2	2.3	29.5	-	63.1	7.4	62.4	8.1	63.6	6.9					
	B	Af recryst.	24.0	72.4	2.7	0.9	22.9	73.5	22.8	74.5	25.6	71.7	775	791	770	845	8
655	A	Pl leuc.	74.5	1.5	24.0	-	64.1	11.9	63.2	12.8	65.6	10.4					
	B	Af leuc.	32.1	63.1	4.3	0.5	31.9	63.3	31.6	64.1	35.0	60.7	859	871	845	922	7
	A	Pl recryst.	74.5	1.2	24.3	-	68.5	7.2	67.9	7.8	69.1	6.6					
	A	Af recryst.	27.1	70.2	2.1	0.6	26.6	70.7	25.9	72.0	27.5	70.4	749	763	729	803	5
		Feldspars from recrystallized metapelites															
44	A	Pl	74.1	2.0	23.8	0.1	70.8	5.4	70.0	6.2	71.3	4.9					
	A	Af	19.3	78.5	1.3	0.9	21.3	76.5	21.9	76.7	22.5	76.1	684	711	666	749	7
372	A	Pl rim	62.3	1.3	36.3	0.1	61.6	2.0	61.5	2.2	61.6	2.1					
	A	Af rim	12.7	85.9	0.6	0.8	12.5	86.1	12.7	86.7	14.7	84.7	556	569	552	608	4
18	A	Pl core	65.2	1.9	32.9	n.d.			62.6	4.5	63.0	4.1					
	B	Af core	14.3	84.3	1.4	n.d.			18.3	80.3	20.9	77.7		685	667	732	5
	A	Pl rim	59.8	1.1	39.1	n.d.			58.9	2.0	58.9	2.0					
	A	Af rim	12.0	87.4	0.6	n.d.			12.7	86.7	15.0	84.4		564	555	605	3
461	A	Pl core	66.2	0.7	33.1	-	60.9	6.0	60.4	6.5	61.0	5.9					
	B	Af core	16.3	80.7	2.4	0.6	22.1	74.9	22.1	75.5	25.5	72.1	750	764	753	817	5
	A	Pl rim	66.2	0.7	33.1	-	64.4	2.5	64.2	2.7	64.4	2.5					
	A	Af rim	13.6	85.2	0.7	0.5	14.4	84.4	14.5	84.8	16.4	82.9	588	593	571	633	3

Calculated compositions (C) and temperatures (T) obtained by reversing the K-Na exchange according to Kroll et al. (1993):

FL - with Margules parameters (W's) from Fuhrman & Lindsley (1988), FL+Cn - same including the celsian component, LN - with W's from Lindsley & Nekvasil (1989), EG - with W's from Elkins & Grove (1990). Method A: microprobe reintegration; method B: image analysis.

Pressures are estimated applying the Grt-Pl-Sil/Ky-Qtz geobarometer (Koziol & Newton 1988), or obtained from Schumacher & Faulhaber (1994) applying the Grt-Opx-Pl-Qtz geobarometer (Newton & Perkins1982).

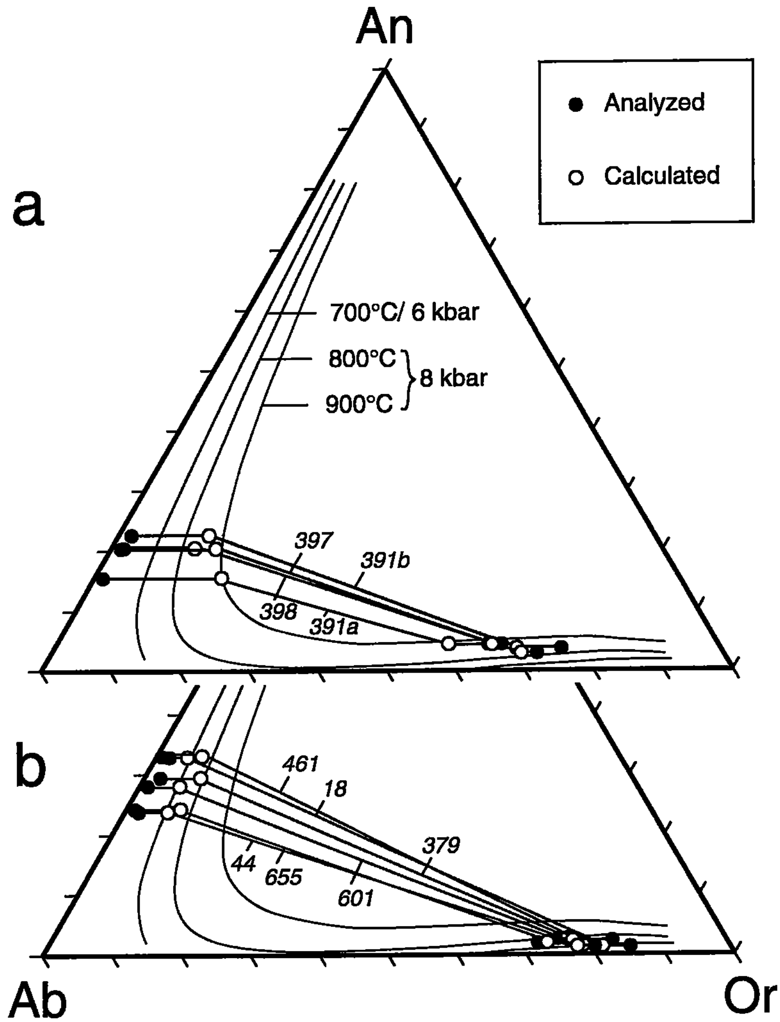

FIG. 5. Analyzed and calculated compositions of recrystallized perthite – plagioclase pairs in (a) strongly flattened garnet – sillimanite gneisses from the basal part of the Highland Complex, (b) metapelites from the central and western part of the Highland Complex. Isotherms according to the model of Fuhrman & Lindsley (1988).

two-feldspar temperatures obtained for the flattened metapelites mark the stage of recrystallization after strong deformation of the rocks. In the central and western part of the Highland Complex, where the metapelites were less intensely deformed, the feldspars recrystallized at lower temperatures in the range 685–790°C (Table 1, Fig. 5b). The highest temperature, approximately 790°C, is calculated for feldspar in recrystallized leucosome (sample 379) from the southeastern part, close to the strongly sheared base of the Highland Complex (Fig. 1). Since the metapelites from the western Highland Complex (samples 18 and 461) show higher modal proportions than in the eastern part, the recalculated compositions of the alkali feldspar for these samples are significantly more Ab-rich than the analyzed compositions, demonstrating clearly the effect of retrograde intercrystalline K–Na exchange between plagioclase and alkali feldspar (Fig. 5b). In the Digana shear zone east of Kandy (locality 601, Fig. 1), a temperature of 716°C is

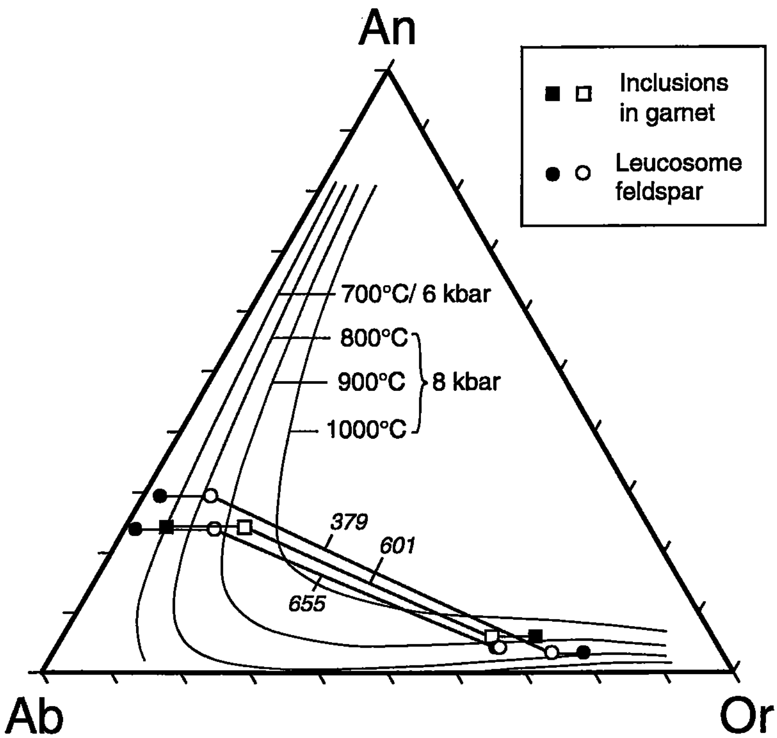

FIG. 6. Analyzed (closed symbols) and calculated compositions (open symbols) of perthite – plagioclase pairs: Sample 601: feldspar inclusions in garnet; sample 379, 655: coarse leucosome feldspars. Isotherms according to Fuhrman & Lindsley (1988).

obtained for the recrystallized plagioclase – perthite pair in the fine-grained mylonitic matrix of a metapelite (sample 601, Table 1, Fig. 5b).

Indications of a high temperature of metamorphism are given by alkali feldspar in *leucosomes*. A sample of migmatitic sillimanite – biotite – garnet – cordierite gneiss from the southern part of the Highland Complex contains An-rich coarse string perthite in the leucosome (Fig. 2c), and plagioclase at the margins of the leucosome (sample 655). A temperature of about 870°C is obtained for this feldspar pair (Table 1, Fig. 6). A slightly lower temperature (826°C) is calculated for leucosome from metapelite from the eastern part of the Highland Complex (sample 379). Both samples contain more alkali feldspar than plagioclase, and no antiperthite was found.

In the *garnet porphyroclasts* of the mylonitic metapelite from the Digana shear zone (sample 601), inclusions of plagioclase and alkali feldspar perthite are preserved. Their compositions give a high two-feldspar temperature (930°C, Fig. 6), probably indicating near-peak conditions of metamorphism. The plagioclase inclusions contain the highest observed amount of Or component (5.8 mol%) retained in the Sri Lankan plagioclase without visible exsolution lamellae.

Disequilibrium antiperthite–perthite pairs

For the ternary feldspar compositions obtained by antiperthite re-integration, two-feldspar temperatures cannot be calculated by the method of Kroll *et al.* (1993), since the re-integrated compositions of the associated alkali feldspar perthite plot on an isotherm for a much lower temperature than the antiperthite (Fig. 7). Obviously, the alkali feldspar largely recrystallized at lower temperatures, and exchanged Al–Si as well as K–Na with the plagioclase after exsolution of the latter. Although relics of primary alkali feldspar with coarse exsolution-induced bodies were sought for analysis, these turned out not to be in equilibrium with the antiperthite bulk feldspar. However, the temperature of equilibration of the original ternary feldspar can be estimated from the position of the isotherms in the ternary feldspar plot (Fig. 7), or can be calculated by varying the alkali feldspar composition at fixed

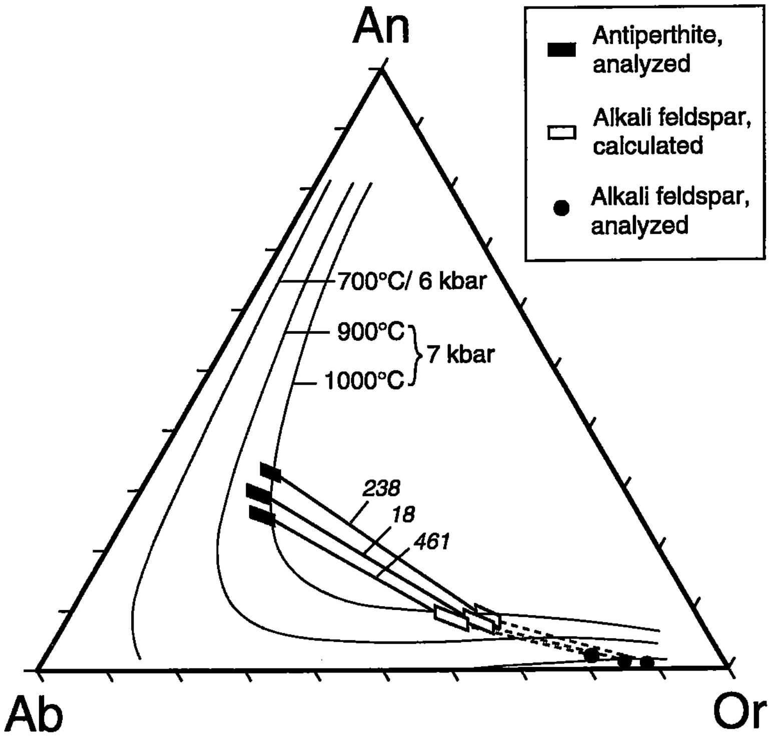

FIG. 7. Re-integrated compositions of antiperthite and perthite, and calculated compositions of ternary alkali feldspar which should have coexisted with the original ternary plagioclase compositions. Samples 18, 238, 461: migmatitic garnet – cordierite gneisses from the western Highland Complex. Isotherms according to Fuhrman & Lindsley (1988).

plagioclase composition until the three temperatures calculated on the basis of equal activities of the Ab, Or, and An components in coexisting plagioclase and alkali feldspar are the same.

Relics of antiperthite are found in all parts of the Highland Complex of Sri Lanka. However, grains with exsolution bodies sufficiently evenly spaced to be reliably integrated are found mostly in the less-deformed garnet – cordierite gneisses from the western part of the Highland Complex. Three samples of antiperthite have been analyzed, one of which mainly contains K-feldspar rods (sample 18), and the other two, both rods and lamellae (samples 238 and 461). The integrated bulk-compositions given in Table 2 are plotted in the feldspar diagram, together with the calculated composition of the alkali feldspar that should have coexisted with the ternary feldspar (Fig. 7). Very high temperatures, in the range 970–990°C, result for these antiperthite bulk-compositions if the Fuhrman & Lindsley (1988) model is applied (Table 2). Even higher temperatures, above 1000°C, are calculated with Margules parameters from Lindsley & Nekvasil (1989) and Elkins & Grove (1990).

Mesoperthite

In a sample of flattened garnet – sillimanite gneiss from the central part of the Highland Complex (sample 331, Fig. 1), the bulk composition of the fine-grained mesoperthite, which recrystallized after strong deformation, indicates a temperature of 916°C by reference to the isotherms calculated after Fuhrman & Lindsley (1988) (Table 2, Fig. 8a). This estimate must be taken as a minimum, since the mesoperthite does not coexist with any plagioclase. Further, this temperature must be too low by about 10–20°C, as the composition of the finely exsolved mesoperthite was obtained by integrating results of electron-microprobe analyses.

TABLE 2. REINTEGRATED ANTIPERTHITE AND PERTHITE/MESOPERTHITE COMPOSITIONS AND ESTIMATED TEMPERATURES

Sample	Method	Feldspar	Composition			Calculated composition			T(FL)	P
			Ab	Or	An	Ab	Or	An	(°C)	(kbar)
			Antiperthite-perthite disequilibrium pairs							
18	B	Antiperthite	53.3	17.3	29.4				974[a])	7
	B	Perthite	14.3	84.3	1.4	32.1	59.7	8.2		
238	B	Antiperthite	49.7	17.2	33.1				992[a])	7
	A	Perthite	11.2	87.7	1.1	30.0	61.0	9.0		
461	B	Antiperthite	54.5	19.6	25.9				983[a])	7
	B	Perthite	16.3	81.3	2.4	35.3	56.0	8.7		
			Single mesoperthite							
331	A	Mesoperthite	55.9	37.4	6.7				916[c])	9
			Antiperthite-mesoperthite near-equilibrium pairs							
314	B	Antiperthite	61.9	28.9	9.2	61.9	28.9	9.2	928[a])	
						67.8	23.0	9.2	890[b])	9
	B	Mesoperthite	42.6	53.6	3.8	51.8	42.3	5.9		
						47.8	48.4	3.8		
391a	B	Antiperthite	61.2	28.4	10.4	61.2	28.4	10.4	942[a])	
						61.4	28.2	10.4	940[b])	10
	B	Mesoperthite	41.6	52.3	6.1	49.8	43.9	6.3		
						49.1	44.8	6.1		

Method A: microprobe re-integration; method B: image analysis; T(FL): temperatures calculated with the Margules parameters of Fuhrman & Lindsley (1988).
[a]) Calculated with the two-feldspar thermometer programs modified by Kroll *et al.* (1993) with fixed plagioclase composition; [b]) calculated according to Kroll *et al.* (1993) with fixed An-content of both feldspars; [c]) minimum one-feldspar temperature.

Near-equilibrium antiperthite–mesoperthite pairs

Antiperthite and mesoperthite found in adjacent layers of a metapelite (sample 314) may have coexisted before exsolution of the feldspars. During cooling, Al–Si exchange and probably also the K–Na exchange did not take place between the layers except in a narrow contact-zone. The bulk compositions of antiperthite and mesoperthite in the center of the respective layers could thus indicate the peak metamorphic conditions. Figure 8a shows that the compositions of antiperthite and mesoperthite do not fall on the same isotherm. Further, the error in alkali feldspar composition based on image analyses of several grains is rather high because of heterogeneous exsolution textures and because of compositional variations within the mesoperthite layer. Two-feldspar thermometry based on fixed An contents (Kroll *et al.* 1993) yields a temperature of 890°C and a shift of both feldspar compositions to lower Or contents (Table 2). A lower Or content of the primary ternary plagioclase is not likely, whereas loss of An content from the alkali feldspar appears to be a probable process; the temperature estimate of about 930°C calculated with fixed bulk-composition of antiperthite may thus be more realistic (Table 2).

A strongly flattened but largely recrystallized feldspar-rich sillimanite – garnet gneiss (sample 391a) from near the southeastern thrust-zone (Fig. 1) contains alkali feldspar and Ab-rich plagioclase (An_{15}) recrystallized from antiperthite found as relics. The bulk composition of the alkali feldspar varies within the thin section from $Ab_{30}Or_{68}An_2$ to $Ab_{42}Or_{52}An_6$, the latter composition obtained from the core zone of larger grains, which might be relics of the primary alkali feldspar. A two-feldspar temperature of 940°C is computed with the presumed bulk-compositions of the primary alkali feldspar and antiperthite when calculated with fixed An contents of both feldspars according to Kroll *et al.* (1993), using the Margules parameters of Fuhrman & Lindsley (1988) (Table 2, Fig. 8b). The calculated composition of the alkali feldspar is richer in Ab than the analyzed composition, whereas the mean of the ternary plagioclase composition is unchanged. This may mean that some retrograde K–Na exchange between the coarse primary feldspars took place

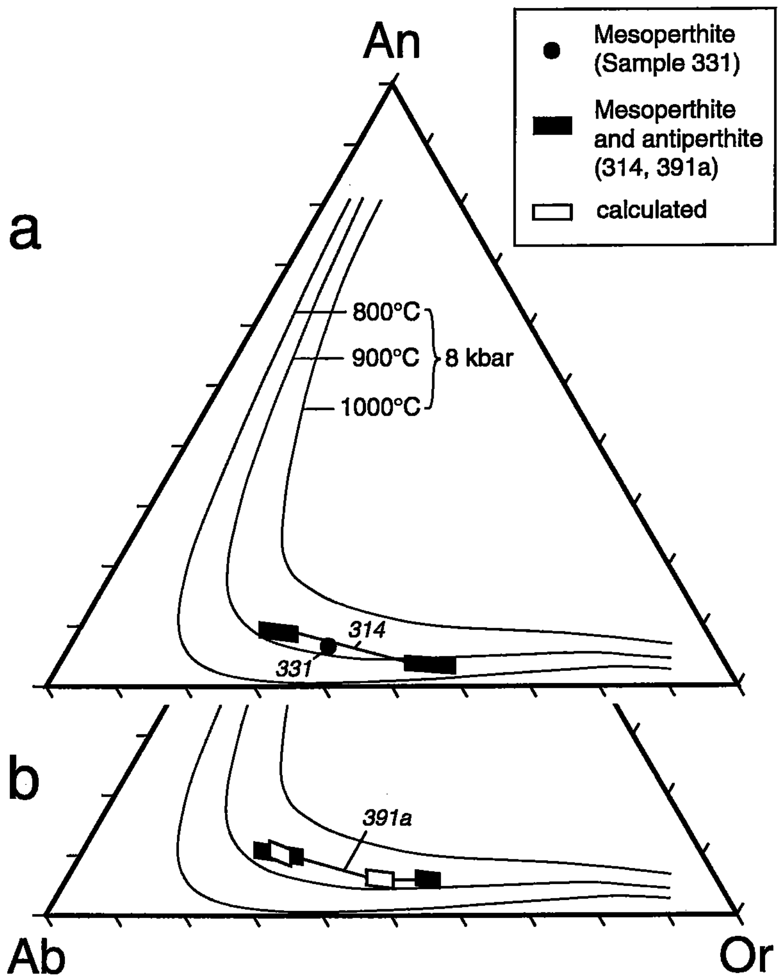

FIG. 8. Re-integrated feldspar compositions from Ca-poor metapelites from the eastern Highland Complex. The size of the symbols indicates the range of analytical uncertainty. (a) Sample 331: recrystallized mesoperthite in a laminated sillimanite – garnet gneiss; sample 314: mesoperthite and antiperthite from adjacent layers of a biotite – garnet gneiss. (b) Sample 391a: relict antiperthite and mesoperthite porphyroclasts in a sillimanite – garnet gneiss. Isotherms according to Fuhrman & Lindsley (1988).

before deformation, recrystallization, and unmixing. The essentially unchanged mean composition of the antiperthite is explained by the predominance of modal plagioclase in this metapelite, whereas the narrowed error-bar results from the recalculation procedure to accommodate K–Na exchange.

Near-equilibrium, retrograde plagioclase–perthite pairs

Rims of recrystallized fine-grained feldspar around coarser grains of feldspar were formed in response to late deformation of the rocks. The composition of these rims yields temperatures of about 560–590°C in the western and central part of the Highland Complex and 460–580°C in the strongly flattened rocks from the southeast (Table 1). There are only slight compositional shifts in the Or component of plagioclase resulting from the recalculation procedure (Kroll *et al.* 1993), meaning that the K–Na exchange stopped shortly after this late recrystallization (Fig. 9). The grains of coarse string perthite in the strongly flattened metapelites locally exhibit two generations of rims, which could be

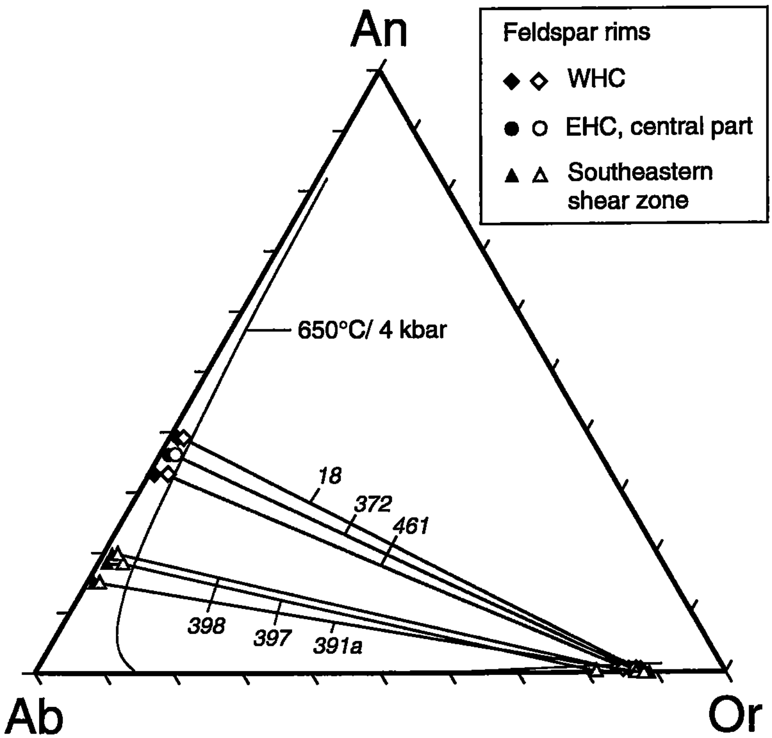

FIG. 9. Analyzed (closed symbols) and recalculated compositions (open symbols) for late products of feldspar recrystallization (partly rim compositions) from migmatitic garnet – cordierite gneisses (samples 18, 461: western Highland Complex), sillimanite – garnet gneisses (sample 372: eastern Highland Complex), and strongly flattened metapelites (samples 391a, 397, 398: southeastern Highland Complex). Isotherm according to Fuhrman & Lindsley (1988).

analyzed separately in sample 397 (Table 1): an older rim (1) grew at about 680°C and a younger rim (2) formed at about 580°C. In most cases, the composition of rim (1), which obviously formed as an overgrowth on the coarse string perthite after or during exsolution of the latter, changed by retrograde diffusion, trending toward the composition of rim (2).

Discussion and Conclusions

Different types of feldspar pairs, near-equilibrium and disequilibrium pairs, can be used to estimate the closure temperature for intercrystalline Al–Si exchange. In case of perthite–plagioclase disequilibrium pairs, the bulk alkali feldspar composition can be re-integrated, and the ternary plagioclase composition recalculated by reversing the intercrystalline K–Na exchange according to the method of Kroll *et al.* (1993). In case of antiperthite–perthite disequilibrium pairs, temperatures of metamorphism can be estimated from the position of the ternary plagioclase on the solvus. Both methods have been applied to the antiperthite–mesoperthite near-equilibrium pairs found in the eastern Highland Complex.

The highest temperatures, derived from antiperthite porphyroclasts and from perthite inclusions in garnet, are in the range 925–990°C, and indicate near-peak conditions of metamorphism. These temperatures appear unrealistically high, and may result from a nonlinear dependence of Margules parameters on temperature and pressure. However, further evidence for very high maximum temperatures near 900°C is given by inverted pigeonite and other exsolved high-temperature pyroxenes occurring in charnockitic and metabasic layers, and by the peak-metamorphic assemblage spinel + quartz recognized in some metapelites (Schenk *et al.* 1988, 1991, Raase & Schenk 1994). Whereas the exsolved pyroxenes could be interpreted as magmatic relics, spinel + quartz as well as the ternary feldspars in metapelites unequivocally indicate an event of high-temperature metamorphism. Evidence for very high temperatures of metamorphism has been reported from adjacent terranes of Gondwana [East Antarctica: Harley & Hensen (1990); Eastern Ghats, India: Sengupta *et al.* (1990); Palni Hills, southern India: Raith *et al.* (1997)].

High temperatures of peak metamorphism, above 900°C, imply that the metapelites must have been partially molten. Leucosomes, however, though widespread, are not volumetrically abundant, indicating strongly reduced activities of H_2O or fluid-absent conditions. Extensive partial melting under fluid-absent conditions takes place at about 850°C in metapelites and at 950°C (at 10 kbar) in biotite–plagioclase gneisses, with the melt production largely dependent on the amount of biotite in the rock (Patiño Douce & Johnston 1991, Vielzeuf & Montel 1994, Patiño Douce & Beard 1995). The plagioclase-dominated metapelites or, more precisely, semipelites in the western Highland Complex, which yield very high antiperthite-derived temperatures, should behave more like the biotite–plagioclase gneiss studied by Vielzeuf & Montel (1994). The ternary plagioclase that unmixed into antiperthite occurs in layers rich in garnet – cordierite – biotite (restite) as well as in the leucosomes. Coarse alkali feldspar perthite in leucosomes in the eastern and southwestern part of the Highland Complex yield significantly lower temperatures of crystallization, in the range 830–870°C, in conformity with the higher ratio of alkali feldspar to plagioclase in these metapelites.

Several causes for the preservation of high-temperature ternary feldspar compositions are suggested from the present study:

(1) Fluid-absent conditions or low activities of H_2O during cooling after the metamorphic climax. H_2O-rich fluids would largely enhance coarsening of exsolved feldspars and migration of the exsolved feldspar component to the grain margins (*e.g.*, Parsons & Brown 1984).

(2) Early incipient exsolution, possibly induced by deformation, will inhibit later intercrystalline exchange of cations, thus preserving the primary bulk-composition.

(3) Coarse grain-size largely prevents intercrystalline Al–Si exchange, even at high temperatures. In the core of the feldspar grains, primary bulk-compositions can be preserved.

(4) Predominance of one feldspar reduces the extent of intercrystalline exchange of the cations Al–Si as well as K–Na. The composition of the predominant feldspar does not change significantly, whereas the composition of the subordinate feldspar phase changes significantly. In case of a single feldspar present (*e.g.*, mesoperthite), no intercrystalline exchange of cations can take place. In this case, however, only a minimum temperature is obtained, since the feldspar composition does not need to lie on the solvus at the conditions of metamorphism.

(5) Feldspars of different compositions in adjacent domains of a rock may have coexisted at peak conditions of metamorphism. Retrograde exchange of cations may be confined to the contact zone between these two domains. Outside the contact zone, primary bulk-feldspar compositions can be preserved.

(6) Feldspar inclusions in garnet porphyroblasts can preserve their primary bulk-composition since they are protected against later intercrystalline exchange of cations. In cases where they were enclosed near the peak of metamorphism, feldspar inclusions should give near-peak temperatures.

Several of these causes are responsible for the preservation of primary compositions in the feldspars studied, *i.e.*, fluid-absent conditions, coarse grain-size, and the predominance of one feldspar. Together, these factors led to the preservation of the primary bulk-composition of the antiperthite porphyroclasts. Concerning the finer-grained generations of feldspars formed through recrystallization at lower temperatures, the effect of smaller grain-size counteracts the effect of lower diffusion-rate. Whereas the recrystallized plagioclase does not exsolve alkali feldspar further below about 800°C but exchanges K–Na, the alkali feldspar unmixes without significant change of bulk composition. Below about 600°C, intercrystalline K–Na exchange ceases, and both feldspars become closed systems. However, this is true only under dry conditions, as prevailed in the metapelites of Sri Lanka, whereas deuteric reactions below about 500°C may lead to coarse exsolution and mutual replacement of alkali feldspar by albite and microcline (Parsons & Brown 1984).

Several stages of the retrograde evolution of the metapelites are documented by generations of recrystallized feldspar: (1) Feldspar recrystallization at about 830–900°C after peak metamorphism in response to strong flattening of the rocks near the base of the Highland Complex in the southeast, (2) feldspar recrystallization in the range 680–760°C in the central and western part of the Highland Complex, (3) deformation and recrystallization of feldspars in the Digana shear zone at about 710°C, (4) late retrograde recrystallization or growth of feldspar rims in the temperature range of 460–590°C in the western and in the eastern Highland Complex.

The temperatures inferred for feldspar recrystallization may be compared with orthopyroxene – garnet Fe–Mg exchange temperatures in metabasites and charnockitic rocks (Schumacher & Faulhaber 1994). For the strongly flattened rocks in the southeastern part of the Highland Complex, feldspar temperatures (830–900°C) are in accordance with orthopyroxene – garnet temperatures (800–910°C) calculated for core compositions with the thermometer of Bhattacharya *et al.* (1991), whereas those calculated with the thermometer of Harley (1984) are significantly lower (710–830°C). Further, the temperatures needed for the development of garnet – clinopyroxene reaction rims in the metabasites (700–820°C: Schumacher *et al.* 1990) and of garnet – quartz reaction rims in charnockitic rocks (650–700°C: Faulhaber & Raith 1991) are comparable with temperatures of feldspar recrystallization, in the range 680–760°C. In metapelites from the western part, garnet – cordierite Fe–Mg exchange temperatures are in the same range (680–770°C: Raase & Schenk 1994). Where the garnet – biotite thermometer is applied, a large range of mostly unrealistic temperatures is obtained because of partial re-equilibration (Raase & Schenk 1994).

Considering the prerequisites discussed for the preservation of high-temperature feldspar compositions, feldspar thermometry is believed to be an effective tool for identifying and quantifying an early, ultra-high-temperature metamorphic event. As demonstrated, furthermore, feldspar thermometry based on the approach of Kroll *et al.* (1993) is a valuable tool for deciphering the retrograde history of granulite-facies rocks.

Geological implications

Relict antiperthite porphyroclasts and feldspars recrystallized after intense deformation in the metapelites from the eastern Highland Complex, and record two high-temperature metamorphic events that are separated by a strong flattening deformation. The causes and age of the early ultra-high-temperature event are disputable. Heating by magmatic intrusions into the lower to middle crust may be responsible for the high temperatures of metamorphism. Charnockitic, enderbitic and metabasic rocks constitute significantly more than half of the rocks in the eastern Highland Complex (Kröner *et al.* 1991). Upper-intercept U–Pb zircon ages of 1.8–1.9 Ga indicate the time of intrusion (Hölzl *et al.* 1994). Rb–Sr whole-rock ages near 2 Ga are considered to date the time of granulite-facies metamorphism (Crawford & Oliver 1969), but Hölzl *et al.* (1994) interpreted these dates as indicative of timing of magmatic intrusion. Provided that the widespread magmatic intrusions are related to the very high-grade early metamorphism documented by the ternary feldspars, the youngest upper-intercept zircon ages of about 1.9 Ga for metasedimentary rocks in the eastern Highland Complex (Hölzl *et al.* 1994) may indicate the time of metamorphism and migmatization. This is at variance with Hölzl *et al.* (1994), who interpreted this date as the age of the detrital zircon.

The episode of recrystallization subsequent to the strong flattening deformation is presumably dated by the lower-intercept U–Pb zircon age at 610 Ma obtained by Hölzl *et al.* (1994), corresponding to the Pan-African event. The temperatures of feldspar recrystallization of about 850°C, obtained for the flattened metapelites near the thrust contact with the Vijayan Complex, imply that thrusting must have happened during granulite-facies conditions. Prograde postkinematic garnet reaction-rims around sillimanite further indicate heating after strong deformation. Thus, a second prograde granulite-facies event is documented; it appears to be significantly younger than the early ultra-high-temperature event. However, since there is no evidence for retrograde alterations in the time span between these two events, the rocks probably resided in the lower crust during Proterozoic times. These may have cooled down slowly and then were reheated, presumably by frictional heat during thrusting of the Highland Complex toward the southeast onto the Vijayan Complex. Prolonged isobaric cooling in the lower crust is inferred from reaction textures in metabasites of the Highland Complex (Schenk *et al.* 1988, 1991) and is further suggested for adjacent terranes of Gondwana (East Antarctica: Harley & Hensen 1990; Eastern Ghats, India: Sengupta *et al.* 1990).

Strong deformation lasted or revived in the Digana shear zone (Voll & Kleinschrodt 1991) until about 550 Ma, as inferred from the lower-intercept zircon age obtained for strongly deformed granitic gneisses from this shear zone (Baur *et al.* 1991). Feldspar recrystallization under amphibolite-facies conditions at about 710°C in the Digana shear zone is indicated by feldspar thermometry.

In the western part of the Highland Complex, which is younger according to the Nd model ages (1–2 Ga: Milisenda *et al.* 1988, 1994), feldspar thermometry suggests an ultra-high-temperature event similar to that in the east. Heat supply for high-grade metamorphism could be related to the intrusion of I-type charnockitic, enderbitic and mangeritic rocks, which have zircon ages of 1.0–1.1 Ga (Kröner *et al.* 1994a). Furthermore, Rb–Sr whole-rock isochron ages of ortho- and paragneisses are near 1 Ga (Cordani & Cooray 1989, Milisenda *et al.* 1994). However, one sample of sillimanite-bearing migmatitic metasedimentary gneiss containing zircon with apparently "igneous" morphology, from near locality 461 (Fig. 1), yielded a well-defined upper-intercept zircon age of 793 Ma (Hölzl *et al.* 1994), which may alternatively be interpreted as the age of ultra-high-temperature metamorphism and migmatization in the western Highland Complex.

The last metamorphic event in the western Highland Complex, which is manifested by feldspar recrystallization at about 680–760°C and growth of cordierite at the expense of garnet in the metapelites, took place at about 535–560 Ma. This date is based on dm-scale Rb–Sr and Nd–Sm isotopic homogenization isochrons (535 Ma: Burton & O'Nions 1990), a zircon U–Pb lower-intercept age (560 Ma: Baur *et al.* 1991), zircon single-grain data (550–560 Ma: Kröner *et al.* 1994a) and monazite ages (540–550 Ma: Hölzl *et al.* 1994), and corresponds to late Pan-African times. Further, post-tectonic alkali granites and pegmatites were emplaced at about 550 Ma (Tilton & Aldrich

1955, Gottfried *et al.* 1956, Hölzl *et al.* 1994). The metamorphic overprint in the western Highland Complex, which must be related to the beginning of uplift, corresponds, in age as well as temperature, to the stage of deformation and recrystallization in the Digana shear zone (550 Ma: Baur *et al.* 1991) in the eastern Highland Complex.

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References

Baur, N., Kröner, A., Liew, T.C., Todt, W., Williams, I.S. & Hofmann, A.W. (1991): U–Pb isotopic systematics of zircons from prograde and retrograde transition zones in high-grade orthogneisses, Sri Lanka. *J. Geol.* **99**, 527-545.

Bhattacharya, A., Krishnakumar, K., Raith, M. & Sen, S.K. (1991): An improved set of *a-X* parameters in Fe–Mg–Ca garnets and refinement of the orthopyroxene – garnet thermometer and the garnet – orthopyroxene – plagioclase – quartz barometer. *J. Petrol.* **32**, 629-656.

Bohlen, S.R. & Essene, E.J. (1977): Feldspar and oxide thermometry of granulites in the Adirondack Highlands. *Contrib. Mineral. Petrol.* **62**, 153-169.

Braun, I., Raith, M. & Ravindra Kumar, G.R. (1996): Dehydration-melting phenomena in leptynitic gneisses and the generation of leucogranites: a case study from the Kerala Khondalite Belt, southern India. *J. Petrol.* **37**, 1285-1305.

Burton, K.W. & O'Nions, R.K. (1990): The timescale and mechanism of granulite formation at Kurunegala, Sri Lanka. *Contrib. Mineral. Petrol.* **106**, 66-89.

Cordani, U. & Cooray, P.G. (1989): Rb–Sr ages of granites and gneisses from the Precambrian of Sri Lanka. *J. Geol. Soc. Sri Lanka* **2**, 35-43.

Crawford, A.R. & Oliver, R.L. (1969): The Precambrian geochronology of Ceylon. *Geol. Soc. Aust., Spec. Publ.* **2**, 283-316.

Elkins, L.T. & Grove, T.L. (1990): Ternary feldspar experiments and thermodynamic models. *Am. Mineral.* **75**, 544-559.

Evangelakakis, C., Kroll, H. & Voll, G. (1991): Exsolution and ordering structures in feldspars from high-grade metamorphic rocks in Sri Lanka. *In* The Crystalline Crust of Sri Lanka, Part I, Summary of Research of the German-Sri Lankan Consortium (A. Kröner, ed.). *Geol. Surv. Dep. Sri Lanka, Prof. Pap.* **5**, 268-271.

________, ________, ________, Wenk, H.-R., Meisheng, H. & Köpcke, J. (1993): Low-temperature coherent exsolution in alkali feldspars from high-grade metamorphic rocks of Sri Lanka. *Contrib. Mineral. Petrol.* **114**, 519-532.

Faulhaber, S. & Raith, M. (1991): Geothermometry and geobarometry of high-grade rocks: a case study on garnet-pyroxene granulites in southern Sri Lanka. *Mineral. Mag.* **55**, 33-56.

Fuhrman, M.L. & Lindsley, D.H. (1988): Ternary-feldspar modeling and thermometry. *Am. Mineral.* **73**, 201-215.

Geological Map of Sri Lanka (1982): Geol. Surv. Dep., Colombo, Sri Lanka (scale: 8 miles to one inch).

Gottfried, D., Senftle, F.E. & Waring, C.L. (1956): Age determinations of zircon crystals from Ceylon. *Am. Mineral.* **41**, 157-161.

Grove, T.L., Baker, M.B. & Kinzler, R.J. (1984): Coupled CaAl–NaSi diffusion in plagioclase feldspar: experiments and applications to cooling rate speedometry. *Geochim. Cosmochim. Acta* **48**, 2113-2121.

Harley, S.L. (1984): An experimental study of the partitioning of Fe and Mg between garnet and orthopyroxene. *Contrib. Mineral. Petrol.* **86**, 359-373.

________ & Hensen, B.J. (1990): Archaean and Proterozoic high-grade terranes of East Antarctica (40–80°E): a case study of diversitiy in granulite facies metamorphism. *In* High-Grade Metamorphism and Crustal Anatexis (M. Brown & J.R. Ashworth, eds.). Allen & Unwin, London, U.K. (320-370).

Heinrich, E.W. (1956): *Microscopic Petrography.* McGraw Hill, New York, N.Y.

Hiroi, Y., Ogo, Y. & Namba, K. (1994): Evidence for prograde metamorphic evolution of Sri Lankan pelitic granulites, and implications for the development of continental crust. *Precambrian Res.* **66**, 245-263.

Hölzl, S., Hofmann, A.W., Todt, W. & Köhler, H. (1994): U–Pb geochronology of the Sri Lankan basement. *Precambrian Res.* **66**, 123-149.

________, Köhler, H., Kröner, A., Jaeckel, P. & Liew, T.C. (1991): Geochronology of the Sri Lankan basement. *In* The Crystalline Crust of Sri Lanka, Part I, Summary of Research of the German-Sri Lankan Consortium (A. Kröner, ed.). *Geol. Surv. Dep. Sri Lanka, Prof. Pap.* **5**, 237-257.

Kleinschrodt, R. (1994): Large-scale thrusting in the lower crustal basement of Sri Lanka. *Precambrian Res.* **66**, 39-57.

KOZIOL, A.M. & NEWTON, R.C. (1988): Redetermination of the anorthite breakdown reaction and improvement of the plagioclase – garnet – Al_2SiO_5 – quartz geobarometer. *Am. Mineral.* **73**, 216-223.

KRIEGSMAN, L.M. (1993): *Geodynamic Evolution of the Pan-African Lower Crust in Sri Lanka.* Ph.D. thesis, Univ. Utrecht. *Geol. Ultraiectina* **114**, 1-207.

________ (1996): Divariant and trivariant reaction line slopes in FMAS and CFMAS: theory and applications. *Contrib. Mineral. Petrol.* **126**, 38-50.

KROLL, H., EVANGELAKAKIS, C. & VOLL, G. (1993): Two-feldspar geothermometry: a review and revision for slowly cooled rocks. *Contrib. Mineral. Petrol.* **114**, 510-518.

KRÖNER, A., COORAY, P.G. & VITANAGE, P.W. (1991): Lithotectonic subdivision of the Precambrian basement in Sri Lanka. *In* The Crystalline Crust of Sri Lanka, Part I, Summary of Research of the German-Sri Lankan Consortium (A. Kröner, ed.). *Geol. Surv. Dep. Sri Lanka, Prof. Pap.* **5**, 5-21.

________, JAECKEL, P. & WILLIAMS, I.S. (1994a): Pb-loss patterns in zircons from a high-grade metamorphic terrain as revealed by different dating methods: U–Pb and Pb–Pb ages for igneous and metamorphic zircons from northern Sri Lanka. *Precambrian Res.* **66**, 151-181.

________, KEHELPANNALA, K.V.W. & KRIEGSMAN, L.M. (1994b): Origin of compositional layering and mechanism of crustal thickening in the high-grade terrain of Sri Lanka. *Precambrian Res.* **66**, 21-37.

LAMB, W.M. (1993): Retrograde deformation within the Carthage–Colton Zone as recorded by fluid inclusions and feldspar compositions: tectonic implications for the southern Grenville Province. *Contrib. Mineral. Petrol.* **114**, 379-394.

LINDSLEY, D.H. & NEKVASIL, H. (1989): A ternary feldspar model for all reasons. *Eos, Trans. Am. Geophys. Union* **70**, 506 (abstr.).

LIU, M. & YUND, R.A. (1992): NaSi–CaAl interdiffusion in plagioclase. *Am. Mineral.* **77**, 275-283.

MILISENDA, C.C., LIEW, T.C., HOFMANN, A.W. & KÖHLER, H. (1994): Nd isotopic mapping of the Sri Lanka basement: update, and additional constraints from Sr isotopes. *Precambrian Res.* **66**, 95-110.

________, ________, ________ & KRÖNER, A. (1988): Isotopic mapping of age provinces in Precambrian high-grade terrains: Sri Lanka. *J. Geol.* **96**, 608-615.

NEWTON, R.C. & PERKINS, D., III (1982): Thermodynamic calibration of geobarometers based on the assemblages garnet – plagioclase – orthopyroxene (clinopyroxene) – quartz. *Am. Mineral.* **67**, 203-222.

OSANAI, Y. (1989): A preliminary report on sapphirine/kornerupine granulite from the Highland Series, Sri Lanka. *Seminar on Recent Advances in Precambrian Geology of Sri Lanka, IFS, Kandy, Extended abstr.*

PARSONS, I. & BROWN, W.L. (1984): Feldspars and the thermal history of igneous rocks. *In* Feldspars and Feldspathoids (W.L. Brown, ed.). *NATO Adv. Sci. Inst., Ser.* **C137**. Reidel, Dordrecht, The Netherlands (317-371).

PATIÑO DOUCE, A.E. & BEARD, J.S. (1995): Dedydration melting of biotite gneiss and quartz amphibolite from 3 to 15 kbar. *J. Petrol.* **36**, 707-738.

________ & JOHNSTON, A.D. (1991): Phase equilibria and melt productivity in the pelitic system: implications for the origin of peraluminous granitoids and aluminous granulites. *Contrib. Mineral. Petrol.* **107**, 202-218.

RAASE, P. & SCHENK, V. (1994): Petrology of granulite-facies metapelites of the Highland Complex, Sri Lanka: implications for the metamorphic zonation and the P–T path. *Precambrian Res.* **66**, 265-294.

RAITH, M, KARMAKAR, S. & BROWN, M. (1997): Ultra-high temperature metamorphism and multistage decompressional evolution of sapphirine granulites from the Palni Hill Ranges, southern India. *J. Metamorphic Geol.* **15**, 379-400.

SCHENK, V., RAASE, P. & SCHUMACHER, R. (1988): Very high temperatures and isobaric cooling before tectonic uplift in the Highland Series of Sri Lanka. *Terra Cognita* **8**, 265.

________, ________ & ________ (1991): Metamorphic zonation and P–T history of the Highland Complex in Sri Lanka. *In* The Crystalline Crust of Sri Lanka, Part I, Summary of Research of the German-Sri Lankan Consortium (A. Kröner, ed.). *Geol. Surv. Dep. Sri Lanka, Prof. Pap.* **5**, 150-163.

SCHUMACHER, R. & FAULHABER, S. (1994): Evaluation of P–T estimates on garnet – pyroxene – plagioclase – quartz-bearing granulite-facies rocks from Sri Lanka. *Precambrian Res.* **66**, 295-308.

________, SCHENK, V., RAASE, P. & VITANAGE, P.W. (1990): Granulite facies metamorphism of metabasic and intermediate rocks in the Highland Series of Sri Lanka. *In* High-Grade Metamorphism and Crustal Anatexis (M. Brown & J.R. Ashworth, eds.). Allen & Unwin, London, U.K. (235-271).

SECK, H.A. (1971a): Koexistierende Alkalifeldspäte und Plagioklase im System $NaAlSi_3O_8 – KAlSi_3O_8 – H_2O$ bei Temperaturen von 650° bis 900°C. *Neues Jahrb. Mineral. Abh.* **115**, 315-342.

________ (1971b): Der Einfluß des Drucks auf die Zusammensetzung koexistierender Alkalifeldspäte und Plagioklase im System $NaAlSi_3O_8 – KAlSi_3O_8 – CaAl_2Si_2O_8 – H_2O$. *Contrib. Mineral. Petrol.* **31**, 67-86.

SEN, S.K. (1959): Potassium content of natural plagioclases and the origin of antiperthites. *J. Geol.* **67**, 479-495.

SENGUPTA, P., DASGUPTA, S., BHATTACHARYA, P.K., FUKUOKA, M., CHAKRABORTI, S. & BOWMICK, S. (1990): Petrotectonic imprints in the sapphirine granulites from Anantagiri, Eastern Ghats mobile belt, India. *J. Petrol.* **31**, 971-996.

SMITH, J.V. & BROWN, W.L. (1988): *Feldspar Minerals.* Springer-Verlag, Berlin.

SNOEYENBOS, D.R., WILLIAMS, M.L. & HANMER, S. (1995): Archean high-pressure metamorphism in the western Canadian Shield. *Eur. J. Mineral.* **7**, 1251-1272.

TILTON, G.R. & ALDRICH, L.T. (1955): The reliability of zircons as age indicators. *Trans. Am. Geophys. Union* **36**, 531-536.

TRÖGER, W.E. (1971): *Optische Bestimmung der gesteinsbildenden Minerale. Teil 1 Bestimmungstabellen.* Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, Germany.

VIELZEUF, D. & MONTEL, J.M. (1994): Partial melting of metagreywackes. I. Fluid-absent experiments and phase relationships. *Contrib. Mineral. Petrol.* **117**, 375-393.

VOLL, G., EVANGELAKAKIS, C. & KROLL, H. (1994): Revised two-feldspar geothermometry applied to Sri Lankan feldspars. *Precambrian Res.* **66**, 351-377.

________ & KLEINSCHRODT, R. (1991): Structural, magmatic and metamorphic development of a Gondwana fragment. *In* The Crystalline Crust of Sri Lanka, Part I, Summary of Research of the German-Sri Lankan Consortium (A. Kröner, ed.). *Geol. Surv. Dep. Sri Lanka, Prof. Pap.* **5**, 22-51.

YUND, R.A. (1986): Interdiffusion of NaSi–CaAl in peristerite. *Phys. Chem. Minerals* **13**, 11-16.

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