Garnet-sillimanite bearing gneisses from Darjeeling, eastern Himalaya: Textural relationship and P–T conditions

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The area around Darjeeling consists of medium grade metamorphic rocks and provides a classic example of inverted Himalayan metamorphism. The area under investigation shows upper amphibolite facies metamorphism (sillimanite-muscovite subfacies), rocks are intimately associated with the migmatites and granites. The presence of quartzite, calc-silicate rocks, graphitic schist and abundance of aluminous minerals like kyanite or sillimanite in these rocks indicate their metasedimentary character. Garnet-sillimanite bearing gneisses occupy most of the area of Darjeeling but not persistent throughout. Textural relationship suggests sequential growth of progressively higher-grade metamorphic minerals during D₁ and D₂ deformation. The relative X_{Mg} in the minerals varies in the order: biotite>staurolite>garnet, and the X_{Mn} decreases in the order: garnet>staurolite>biotite. The P–T evolution of these garnet-sillimanite gneiss has been constrained through the use of conventional geothermobarometry, internally consistent TWEEQU programme and Perple_X software in the KFMASH model system, the combination of these three approaches demonstrates that the Darjeeling gneisses experienced peak pressure and temperature at 7.0 ± 0.3 kbar and $700 \pm 30^\circ\text{C}$. The observation in this study has important bearing on the inverted metamorphism in the Himalayan metamorphic belt.

1. Introduction

The term ‘Darjeeling Gneiss’ was first used by Mallet (1875) for crystalline gneiss of Darjeeling hills of the Darjeeling Himalaya. Later this term has been used extensively by Auden (1935), Acharya (1978), Lal *et al.* (2005), Prakash and Tewari (2013), etc. This region constitutes a part of the lesser Himalaya and consists of metamorphic rocks belonging to Daling–Darjeeling Group (Sinha Roy 1974) of probably Precambrian to Palaeozoic age. These rocks are thrust over younger metasedimentary rocks of Gondwana group of Permocarboniferous age towards the south, while the north these are separated from the medium to high grade rocks of the central crystalline thrust. The deformation

accompanying or post-dating the thrusting movements has folded the rocks of the group in major synform (Gansser 1964). The area under investigation consists of a part of lesser Himalaya, north-eastern part of India and falls between $27^\circ 02' 45''$ – $27^\circ 03' 53''\text{N}$ and $88^\circ 17' 27''$ – $88^\circ 21' 08''\text{E}$ in the Survey of India toposheet no.78A/8 (figure 1).

Preliminary geological mapping of the Lower Himalaya of Darjeeling were carried out by Mallet (1875), Bose (1891), Auden (1935), Heim and Gansser (1939) and others which form the base for further detailed investigations (Mukhopadhyay and Gangopadhyay 1971; Jangpangi 1972; Acharya 1978). These studies have revealed that the region has undergone multiple deformation and at least three major phases of deformation have been

Keywords. Garnet-sillimanite gneiss; textural relationship; pseudosection; P–T conditions; Darjeeling; Eastern Himalaya.

Figure 1. Generalized geological map of the Himalayas. The location of Darjeeling is shown in the box.

recognized which presumably occurred during the Tertiary Himalayan orogeny.

Isochemical phase-diagram, calculated with the thermodynamic software *Perple_X* (Connolly 2005) coupled with its internally consistent thermodynamic database (Holland and Powell 1998) is valid for a unique chemical composition in a model system. If it can be assumed that the analysed bulk rock composition was the effective chemical composition of the equilibrated system at the time of crystallization of garnet, the isochemical phase-diagram drawn for this composition and showing different isopleths for garnet can then be used as thermobarometric tool allowing the determination of P–T condition. In Darjeeling–Sikkim area, number of earlier workers have provided a detailed geological and structural account (e.g., Mukul 2000; Gupta *et al.* 2010; Prakash and Tewari 2013), however, the metamorphic conditions of the rocks are not constrained through conventional geothermobarometers and isochemical phase-diagram approach. In this contribution, textural relationship, mineral chemistry and P–T conditions of garnet-sillimanite

bearing gneisses are presented. Furthermore, isochemical phase-diagram modelling of mineral assemblage is used to investigate the metamorphic evolution of garnet-sillimanite bearing gneisses.

2. Geological setting and structure

Mallet (1875), first distinguished the lithological difference in between Daling group and Darjeeling group. Wager (1939) considered Daling to be equivalent of the pelitic series of Mt. Everest, and Darjeeling gneiss to be injected at higher structural level. Gansser (1964) considered the Daling and Darjeeling to be portions of a single stratigraphic unit and divided them on the basis of their metamorphic characters and placed garnetiferous mica-schists in between them. Gansser (1964) clearly states that there is no evidence of thrusting between Daling and Darjeeling rocks particularly in the Tista valley–Darjeeling hill section. He maintains that the level of disharmonic disturbances may coincide with contact of gneiss and underlying argillaceous

rocks in view of the difference in their respective competence. The inverted sequence of lesser Himalaya is classic itself. Ray (1947) first mapped the metamorphic zones of Barrovian type on the basis of the first appearance of index minerals, viz., chlorite, biotite, garnet, staurolite, kyanite and sillimanite. He also identified that the rocks of the sillimanite-bearing gneisses occur at highest structural levels while chlorite-rich phyllite and schist occur at the bottom of the metamorphites. The Darjeeling formation comprises of migmatitic gneisses, Kyanite-staurolite-garnet-mica gneisses and sillimanite-garnet-mica ($\pm$ staurolite) gneisses.

The garnet-sillimanite bearing gneisses occupy higher elevations of the area and their occurrence is not persistent throughout. It is hard, compact and shows alternating bands of quartzo-feldspathic and micaceous material. It is composed of quartz, biotite, muscovite, garnet, feldspar and sillimanite. Needle-like sillimanite and pink blebs of garnet can be seen by naked eyes (figure 2). The sillimanite needles are maximum up to 1 cm in length. It is 60-m thick at Darjeeling town and extends further northwards for about 1.5 km along Kursyong road. It shows gradational contacts with migmatite. The garnet-sillimanite bearing gneiss dips 30° – 55° , NE. The sillimanite lineation plunges 12° – 16° , NNE. The gneiss is puckered, folded and boundinaged at Mangpu. Three generations of folds are recognized in Sikkim–Darjeeling Himalaya by Mukhopadhyay and Gangopadhyay (1971). First fold F_1 is very tight, near isoclinals and is recognizable on minor scales. Trend of the lineation (L_1) is NE/ENE–SW/WSW. F_2 is moderately tight to open and

the trend of F_2 is NW–SE. The youngest fold F_3 are open folds and the axis usually trend NE–SW. The second deformation (D_2) associated with the Himalayan orogeny is responsible for the generation of appressed, isoclinal, recumbent/reclined flattened flexure slip fold (F_2) on all scales with occasional axial planar crenulation cleavages and pervasive schistosity (figure 3a). These folds apparently represent a stage of superimposed fields of strain during progressive deformation. These folds indicate that the strain exhibited is the result of the operation of extension which has been superimposed on previous overall compression. The area under study also exhibits intense degree of mobilization marked by ptygmatic veining (figure 3b). Extreme cases of mobilization of gneisses result in the formation of locally granitized zones, leading to biotite granite gneisses.

3. Textural relations and chronology of the phases of deformation and metamorphic crystallization

The pelitic rocks of Darjeeling area are represented by the coarse-grained garnet-sillimanite bearing gneisses. The high-grade gneisses with fibrolite to phyllitic composition of rocks occurring from Darjeeling to Tista valley represent a classical Barrovian metamorphism. Five metamorphic zones have been established on the basis of development of typical index minerals, viz., chlorite, biotite, almandine, kyanite and sillimanite. The area around Darjeeling is unique in the sense that it

Figure 2. Close-up view of hand-specimen of Darjeeling gneiss showing garnet, sillimanite and biotite with other minerals.

Figure 3. (a) Isoclinal folding in Darjeeling gneiss and (b) an irregular, lobate layer in Darjeeling gneiss showing ptigmatic folding.

is situated at higher structural levels and belongs to sillimanite zone. Prograde chlorite is absent in this zone. Garnet, commonly occurs as porphyroblasts are xenoblastic and sieve textured with inclusions of quartz, biotite, muscovite, ilmenite and rarely staurolite, plagioclase and kyanite (figure 4a). Various macro and micro structures are present in this area and the rocks of the area have suffered at least three phases of deformation. The original bedding plane (S_0) changes into the schistosity plane (S_1) due to the D_1 deformation and F_1 folding followed by the D_2 deformation and F_2 folding of the S_1 plane forming S_2 crenulation cleavage. Lastly, the D_3 deformation developed by the refolding of F_2 into F_3 and formation of the S_3 axial-plane crenulation cleavage. The textural features suggest syn-tectonic to post-tectonic crystallization of muscovite with respect to D_1 , followed by post-crystalline deformation during the subsequent folding movements. Sillimanite commonly occurs as fine-grained needle-like grains (fibrolite). Decrease

in the modal content of kyanite, and the absence of K-feldspar association with kyanite indicate that sillimanite could have been derived by the polyphase transformation of kyanite (Mohan *et al.* 1989; Lal *et al.* 2005).

The reaction (1) may also be a result of the breakdown of garnet and staurolite-bearing assemblages, instead of the breakdown of muscovite in the presence of quartz (Banerjee and Bhattacharya 1981; Mohan *et al.* 1989; Lal *et al.* 2005). Garnet porphyroblasts showing replacement by sillimanite, which is intergrown with biotite, indicate the resorption of garnet through the reaction (figure 4b):

Mostly, the sillimanite as fibrolite mats is intergrown with micas (figure 4c). Kyanite is present as relict in the sample no. T4 (figure 4d). The

Figure 4. Photomicrographs illustrating textural relations in the Darjeeling gneiss. (a) Poikiloblastic garnet containing inclusions of quartz, biotite and muscovite (in XPL), (b) sillimanite occurring along with biotite at the rim of garnet porphyroblast (in XPL), (c) sillimanite (fibrolite) intergrown with muscovite (in XPL), and (d) relict kyanite enclosed within an aggregate of biotite (in PPL).

crystallization of garnet began prior to that of staurolite and kyanite; it continued to grow even after the formation of these minerals, and its crystallization ceased before the development of fibrolite (cf. Prakash and Tewari 2013).

4. Mineral chemistry

The mineral chemistry were carried out by SX100 CAMECA Electron Probe Micro Analyzer (EPMA) at Wadia Institute of Himalayan Geology, Dehradun, India under comparable operating conditions of 15 keV accelerating voltage, 20 nA beam current and beam size of 1 μm . The representative microprobe analyses of various minerals are listed in table 1. The stoichiometry of the garnets corresponds closely to the ideal formula: $(\text{Fe}^{2+}, \text{Mg}, \text{Mn}, \text{Ca})_3(\text{Al}, \text{Ti}, \text{Cr}, \text{Fe}^{3+})_2\text{Si}_3\text{O}_{12}$. Garnet is essentially almandine (~ 75 mol%)–pyrope (~ 15 mol%) rich. Grossular and spessartine components are rather low.

The structural formulae of the biotites are similar to other biotites from metamorphic rocks (Deer *et al.* 1962, vol. 3) and deviate from the general formulae of ideal trioctahedral mica as the total number of cations at the Y-site falls short of the possible six per formula unit. This is apparently related to heterovalent substitution in octahedral layer as discussed by Foster (1960). The alkali deficiency, i.e., $X < 2.0$, present in the biotites is not abnormal for these metamorphic trioctahedral micas (cf. Guidotti 1974). Biotite has a homogeneous composition and shows extensive Fe–Mg solid solution. Biotite in immediate contact with garnet is Fe-rich compared with biotite in the matrix. Biotite with ~ 4 wt% TiO_2 is characteristic of this mineral when associated with pelitic gneisses. The X_{Mg} of biotite ranges between 0.36 and 0.38.

The staurolite has slightly higher Al-content and lower octahedral sum compared with the ideal formula of $(\text{Fe}, \text{Mg})_2\text{Al}_5\text{Si}_4\text{O}_{23}(\text{OH})$ proposed by Thompson (1976). Similar feature has been

Table 1. Representative microprobe analyses of the coexisting minerals (sample no. T4).

	Grt(c)	Grt(r)	Bt(m)	Bt(r)	St(c)	St(r)	Mus(m)	Mus(r)	Pl(c)	Pl(r)	Ilm
SiO ₂	37.18	36.07	34.66	34.04	27.94	27.75	46.27	47.51	61.57	60.72	0.02
TiO ₂	0.01	0.00	4.06	3.49	0.87	0.82	0.85	0.87	0.00	0.00	53.78
Al ₂ O ₃	21.26	21.55	19.00	18.76	52.97	52.62	35.09	35.46	24.29	24.82	0.02
Cr ₂ O ₃	0.04	0.05	0.00	0.00	0.00	0.000	0.00	0.00	0.00	0.00	0.04
FeO	34.33	33.97	21.76	22.67	13.67	13.94	1.23	1.35	0.03	0.05	46.92
MnO	2.65	3.88	0.13	0.21	0.11	0.18	0.00	0.00	0.00	0.00	1.03
CaO	1.33	1.45	0.00	0.02	0.00	0.00	0.01	0.02	5.37	5.86	0.02
MgO	3.79	3.06	7.29	6.98	1.69	1.59	1.15	1.19	0.00	0.00	0.20
K ₂ O	0.00	0.00	9.24	9.20	0.00	0.00	9.94	10.39	0.32	0.21	0.00
Na ₂ O	0.00	0.00	0.17	0.16	0.00	0.00	0.29	0.39	8.42	8.31	0.00
Total	100.60	100.03	96.30	95.52	97.75	97.46	94.84	95.91	100.01	99.98	102.03
Oxygen basis	12	12	22	22	23	23	22	22	8	8	4
Si	2.974	2.923	5.290	5.270	3.889	3.884	6.160	6.128	2.733	2.701	0.001
Al ^{IV}	0.026	0.077	2.710	2.730	0.111	0.116	1.840	1.872	1.271	1.301	0.000
Al ^{VI}	1.978	1.981	0.700	0.690	8.578	8.562	3.666	3.634	0.000	0.000	0.000
Ti	0.001	0.000	0.460	0.410	0.091	0.086	0.085	0.086	0.000	0.000	3.896
Fe ³⁺	0.019	0.016	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.754
Cr	0.003	0.003	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001
Fe ²⁺	2.277	2.286	2.770	2.930	1.591	1.631	0.137	0.149	0.001	0.002	3.462
Mn	0.180	0.266	0.020	0.030	0.013	0.021	0.000	0.000	0.000	0.000	0.084
Mg	0.451	0.369	1.660	1.610	0.351	0.332	0.228	0.234	0.000	0.000	0.029
Ca	0.114	0.126	0.000	0.000	0.000	0.000	0.001	0.003	0.255	0.279	0.003
K	0.000	0.000	1.800	1.820	0.000	0.000	1.688	1.746	0.018	0.012	0.000
Na	0.000	0.000	0.050	0.050	0.000	0.000	0.075	0.100	0.725	0.717	0.000
X _{Mg} or X _{Ca}	0.165	0.139	0.375	0.355	0.181	0.169	0.622	0.605	0.256	0.277	–

X_{Mg} = Mg/(Fe²⁺ + Mg); X_{Ca} = Ca/(Ca + Na + K); m=matrix; c=core; r=rim.

observed by microprobe analyses of staurolite from Sini area in Singhbhum district of Bihar (Lal and Ackermann 1979; Mohan *et al.* 1989). Staurolite is Fe-rich as X_{Mg} ranges between 0.17 and 0.18.

Muscovite shows X_{Mg} in the range of 0.60–0.62 and it is a solid solution between muscovite and phengite. A celadonite-poor white mica crystallized in higher grades is related to the muscovite-phengite by the Tschermak substitution, i.e., Si^{IV}(Mg, Fe^{VI}) = Al^{IV}Al^{VI}. X_{Cel} component is low.

The formula of plagioclase shows persistent deficiency of $\sum \text{Ca} + \text{Mg}$ cations. A similar feature has been observed by Schreyer *et al.* (1975). They have proposed that the observed deficiency may be explained through deviations from ideal feldspar stoichiometry towards higher siliceous solid solution, for example, in pure albite by virtues of substitution Si=Na, Al; and in anorthite through Si=Al, Ca 0.5. Plagioclase is sodium rich and anorthite content is ~25 mole%.

5. Geothermobarometry and phase-equilibria modelling

Metamorphic conditions of garnet-sillimanite bearing gneisses have been constrained through the

application of (a) various geothermobarometric models of mineral equilibria, (b) internally consistent thermodynamic datasets of Berman (1988, 1990) and the updated TWEEQU computer programme, and (c) *Perple_X* 6.66 (Connolly 2005). The P–T estimates for the core and the rim compositions obtained from the selected thermometers and barometers are given in table 2. Temperature estimates were made using garnet-biotite, garnet-muscovite, garnet-staurolite and plagioclase-muscovite thermometers and pressure estimates with a garnet-plagioclase-biotite-muscovite-quartz and garnet-plagioclase-aluminosilicate-quartz barometer. The thermometric models of Goldman and Albee (1977), Dasgupta *et al.* (1991) and Bhattacharya *et al.* (1992) provide average temperatures of 684±17° and 650±15°C for core and rim, respectively. However, the temperature values are raised by a minimum of 40°–50°C using the models of Perchuk and Lavrent'eva (1983), Perchuk *et al.* (1985) and Indares and Martignole (1985). The P–T conditions deduced from these geothermometers give an average temperature estimate of 706±31 (core) and 678±40 (rim) °C. The garnet-plagioclase-biotite-muscovite-quartz and garnet-plagioclase-aluminosilicate-quartz assemblages permit the use of calibrations of Hoisch (1990), Newton

Table 2. *P–T* results using preferred models of geothermobarometry.

Models	Temp. °C calculated at 6 kbar	
	Core	Rim
Geothermometry		
Garnet-biotite thermometry		
Goldman and Albee (1977)	664	632
Perchuk and Lavrent'eva (1983)	720	691
Perchuk <i>et al.</i> (1985)	741	714
Indares and Martignole (1985)	725	709
Dasgupta <i>et al.</i> (1991)	693	658
Bhattacharya <i>et al.</i> (1992)	695	659
Garnet-muscovite thermometry		
Hoisch (1990)	756	747
Garnet-staurolite thermometry		
Perchuk (1990)	672	631
Plagioclase-muscovite thermometry		
Green and Uzdansky (1986)	621	606
Average	706 ± 31	678 ± 40
Models	Pressure (kbar) calculated at 600 °C	
	Core	Rim
Geobarometry		
Garnet-plagioclase-biotite-muscovite-quartz barometry		
Hoisch (1990)	6.5	6.3
Garnet-plagioclase-aluminosilicate-quartz barometry		
Newton and Haselton (1981)-Sill	6.3	6.2
Newton and Haselton (1981)-Ky	6.6	6.5
Hodges and Crowley (1985)-Sill	6.3	6.2
Hodges and Crowley (1985)-Ky	6.7	6.6
Koziol (1989)-Sill	7.4	7.1
Koziol and Newton (1988)-Sill	6.7	6.5
Average	6.6 ± .37	6.4 ± .31

and Haselton (1981), Koziol (1989), Koziol and Newton (1988) and Hodges and Crowley (1985) to estimate pressure conditions of chemical equilibration. Pressure conditions from different pertinent calibrations are estimated to be around 6.6 ± 0.3 (core) and 6.4 ± 0.3 (rim) kbar. Variations in pressure estimates are a consequence of the difference in the activity composition relations and the thermodynamic data utilized in these models.

P–T estimates were also obtained using the internally consistent thermodynamic datasets of Berman (1988, 1990). The end-member phases used in the TWEEQU program calculations include almandine, pyrope, grossular, muscovite, annite, eastonite, phlogopite, anorthite, sillimanite, rutile, ilmenite and alpha-quartz. There are four possible equilibria that can be written for the selected end-member phases. Using these equilibria, core and rim compositions for sample T4 give precise intersection (figure 5 table 3) at 7.1 kbar, 690 °C (core) and 6.8 kbar, 640 °C (rim).

P–T data have been obtained from equilibrium assemblage diagram (EAD, often referred as

pseudosections), which were constructed with the software Perple_X 6.66 (Connolly 2005) (figure 6) and the internally consistent thermodynamic dataset in the KFMASH system. The detailed formula, notation and sources of the solution models are given in table 4. P–T pseudosections for the pressure and temperature range (2–10 kbar and 200–800 °C) was computed for garnet-sillimanite gneiss (sample no. T4) from the study area in the model system KFMASH ($\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}$) (figure 6). In the construction of the pseudosection for the sample no. T4, following phases and the corresponding phase components (in the parentheses) are chosen from those listed by the Perplex software, e.g., Grt (alm, prp and grs), Mica (phen, pa and ma), Chl (Mn chl), Bt (phl and east), Pl (ab and an) and St (fst and mst) and silicate melt phase (h2oL). Sillimanite and quartz are taken as pure phases. This diagram was contoured by X_{Mg} , X_{Ca} and X_{Mn} garnet isopleths, e.g., to the content of molar fractions of garnet components. The peak assemblage biotite-mica-garnet-sillimanite-

albite is stable above 600°C. The calculated isopleths for X_{Mg} and X_{Ca} garnet intersect at 7 kbar and 690°C for the core composition whereas rim

composition intersect at 6.5 kbar and 680°C. The calculated isopleths for garnet are well corroborated with the garnet composition obtained by EPMA. The X_{Mg} and X_{Ca} isopleths of garnet suggest peak P–T condition of 7 kbar at 700°C for the formation biotite-mica-garnet-sillimanite-albite-bearing assemblage of the study area.

6. Discussion and conclusion

Petrological studies on the Lesser- and Higher Himalayan crystallines have led to characterization of metamorphic reactions, including the melting reactions in the higher grades from fluid-induced melting to muscovite dehydration melting and biotite dehydration melting (Spear *et al.* 1999; White *et al.* 2003; Dasgupta *et al.* 2009). These studies have established that varying bulk compositions affected the distribution of metamorphic isograds. The productivity of the melt depends on which melting reactions are reached during the P–T–time evolution, the availability of fluids and the bulk rock composition. The latter can be reasonably estimated from the observed composition. The dynamics of crustal melting in collisional settings has major implications. The collisional environment and melting of high grade gneisses have been envisaged for the origin of leucogranite formation in Himalayas (LeFort *et al.* 1987; Crawford and Windley 1990). In this model, the melting has been induced in sillimanite gneisses of Tibetan crystalline slab by ‘hot over cold’ thrusting and subsequent fluxing of fluids derived from dehydration of metasediments overthrust by the gneisses. Subsequent fluid flux caused ‘wet melting’ in the gneisses to generate minimum melt which moved a short distance from the source and

Figure 5. Simultaneous calculations of P–T condition by TWEEQUE program for core and rim composition.

Table 3. Simultaneous calculation of P–T by TWEEQUE program (version Win 2.32).

	dH	dS(JK ⁻¹)	dV(Jbar ⁻¹)
Equilibria plotted in figure 5			
Sample no. T4 (core)			
1. Gr+ α Qtz+2Sil = 3An	-12.3	-4.67	-0.23
2. Ms+Gr+Alm = Ann+3An	-12.2	-3.55	-0.21
3. 4 α Qtz+3Eas+3Alm = 3Ann+2Py+4Sil	-13.4	-0.79	-0.01
4. 2Ms+Phl+Py = 6 α Qtz+3Eas	-0.5	-0.24	0.07
5. 3Ms+2Py = 2Sil+7 α Qtz+3Eas	-10.6	-0.26	0.09
Sample no. T4 (rim)			
1. 3Rt+Alm = 3Ilm+2 α Qtz+Sil	3.4	-0.65	-0.1
2. 6 α Qtz+Py+2Gr+3Eas = 6An+3Phl	-0.3	-2.71	-0.13
3. 2Sil+ α Qtz+Gr = 3An	-12.3	-4.67	-0.23
4. Ms+Gr+Alm = Ann+3An	-12.2	-3.55	-0.21
5. Phl+Alm = Py+Ann	-22.7	-0.83	-0.01

dS and dV are calculated at 1 kbar and 298 K.

Figure 6. Calculated P–T pseudosection for Darjeeling gneiss (sample no. T4) in the KFMASH system. Bulk composition in wt% (Na₂O 1.19, MgO 3.34, Al₂O₃ 19.07, SiO₂ 62.66, P₂O₅ 0.11, K₂O 4.25, CaO 0.24, TiO₂ 0.55, Mn 0.04, Fe₂O₃ 6.24). The pseudosection has been contoured for the isopleth X_{Mg} , X_{Ca} and X_{Mn} garnet. Calculation of P–T condition using core and rim composition of garnet by the intersection of X_{Mg} (garnet) and X_{Ca} (garnet).

generated leucogranites. In the high Himalaya, the central crystallines or the Tibetan slab gneisses are thrust over Lesser Himalaya along the MCT. The above model has been reviewed in detail by Dasgupta *et al.* (2009), Basu (2013) and Rubatto *et al.* (2013).

A comparison of the estimated temperature range of 700°C for the sillimanite zone with the minimum anatexis curve valid for plagioclase An₂₆ (Winkler 1978) suggests that widespread melting

in the gneiss should occur in this zone through the melting reaction

The melting reaction lies at higher temperatures (~650°C/7 kbar) than the minimum anatexis of granite through the reaction (Winkler 1978)

Table 4. *Solution notation, formula and model sources for phase diagram calculation.*

Symbols in Perple_X solution model file	Symbols used indrawn pseudosections	Solutions name	Formula	Source
Bio(HP)	Bt	Biotite	$K[Mg_xFe_yMn_{1-x-y}]_{3-w}Al_{1+2w}Si_{3-w}O_{10}(OH)_2, x + y \leq 1$	Powell and Holland (1999)
TiBio(HP)	Bt	Biotite	$K[Mg_xFe_yMn_{1-x-y}]_{3-w-z/2}Ti_zAl_{1+2w}Si_{3-w}O_{10}(OH)_2, x + y \leq 1$	Powell and Holland (1999); White <i>et al.</i> (2000)
Gt(HP)	Grt	Garnet	$Fe_{3x}Ca_{3y}Mg_{3z}Mn_{3(1-x-y-z)}Al_2Si_3O_{12}, x + y + z \leq 1$	Holland and Powell (1998)
Chl(HP)	Chl	Chlorite	$[Mg_xFe_wMn_{1-x-w}]_{5-y+z}Al_{2(1+y-z)}Si_{3-y+z}O_{10}(OH)_8, x + w \leq 1$	Holland <i>et al.</i> (1998)
Mica(CHA)	Mica	Muscovite	$K_yCa_xNa_{1-x-y}(Mg_{1-v}Fe_v)_zMg_wTi_wAl_{3+x-w-z}Si_{3-x+z}O_{10}(OH)_2,$ $x + y \leq 1, w + z \leq y$	Coggon and Holland (2002); Auzanneau <i>et al.</i> (2010)
Pheng(HP)	Mica	Muscovite	$K_xNa_{1-x}Mg_yFe_zAl_{3-2(y+z)}Si_{3+y+z}O_{10}(OH)_2$	Holland and Powell (1998)
St(HP)	St	Staurolite	$Mg_{4x}Fe_{4y}Mn_{4(1-x-y)}Al_{18}Si_{7.5}O_{48}H_4, x + y \leq 1$	Holland and Powell (1998)
Ctd(HP)	Ctd	Chloritoid	$Mg_xFe_yMn_{1-x-y}Al_2SiO_5(OH)_2, x + y \leq 1$	White <i>et al.</i> (2000)
Melt(HP)	Melt	Melt	Na-Mg-Al-Si-K-Ca-Fe hydrous silicate melt	Holland and Powell (2001)
Pl(I1, HP)	Ab, An	Plagioclase Feldspar	$K_yNa_xCa_{1-x-y}Al_{2-x-y}Si_{2+x+y}O_8, x + y \leq 1$	Holland and Powell (2003)

Although in less aluminous pelitic gneisses, K-feldspar+muscovite do occur in this zone, sillimanite and K-feldspar are not found together in the aluminous gneisses. The absence of co-existing sillimanite+K-feldspar and the stable occurrence of muscovite in the estimated P–T conditions of this zone could be estimated as follows:

- Plagioclase in the gneisses is oligoclase and thus the melting reaction would shift to higher temperatures. Tracy (1978) calculated temperatures around 700°C for the reaction (4) in which plagioclase is anorthite instead of albite.
- It is also likely that the gneisses may not be saturated with water and thus the condition may have been between those of reaction (3) and the reaction (5)

The latter reaction occurs at 750°C/7 kbar (Thompson 1974) and would shift to still higher temperatures with the incorporation of anorthite molecule in plagioclase (cf. Tracy 1978).

Several other reactions have been proposed for the formation of sillimanite other than the

above reactions. These may be summarized as follows:

Reaction (6) has been given by Turner and Verhoogen (1960) and Winkler (1978).

Above reaction (7) has been given by number of researchers, viz., Chakraborty and Sen (1967); Thompson and Norton (1968); Hollister (1969); Carmichael (1970); Froese and Gasparrini (1975); Lal and Singh (1978); Spear *et al.* (1999).

Above reaction has been given by Guidotti (1974).

The reactions (6) and (8) are not applicable in the area because of the instability of chlorite in the presence of muscovite + quartz and absence of K-feldspar in the sillimanite bearing gneisses

respectively. Textural relations (see above) are also not in favour of the reaction (7).

According to Carmichael (1970), although sillimanite is rarely texturally associated with kyanite, the inversion of kyanite to sillimanite takes place by a series of reactions involving other phases which act as catalyst that is consumed at one step of the sequence of reaction at a later step. All the intermediate products cancel out of the reaction that expresses the net change taking place in the system, i.e., kyanite = sillimanite (reaction 1). Thus even though sillimanite is remote from kyanite in these rocks, the simple net reaction (1) may have been in progress. The common association of sillimanite with biotite according to Chinner (1961) appears to be nucleation phenomenon. The trigonally arranged chains of oxygen tetrahedra and octahedra in the alternate mica sheets acting as nuclei for the growth of the tetrahedral and octahedral chains constitutes the sillimanite structure. Nucleation is probably dominantly epitaxial, and alumina and silica for sillimanite growth are mainly derived from unstable kyanite and transferred to the sillimanite nuclei through the medium of fluid phase.

It is generally believed that melting is associated with process of anatexis in regions of high heat flow (Turner 1968; Winkler 1978). It is apparent from the disposition of the tectonic units (see figure 1) and the metamorphic zones that the central crystalline belts may have served as the locus of metamorphism producing widespread melting and higher grade metamorphic rocks in the regions of high heat flow and successive lower grade rocks away from it. Later thrusting during D₂ deformation transported these rocks from regions near to central crystalline to lower Himalayas, thus producing inverted sequence of metamorphic zones. Therefore, in order to explain the pattern of regional metamorphism, it may be necessary to assume an abnormally high rate of heat flow in the region of Main Central Thrust from the underlying basement extending to the mantle to provide heat for the metamorphism. Due to random heat flow in some of the regions in lesser Himalaya, the muscovite dehydration melting reaction was not encountered and rock did not witness sillimanite-K-feldspar isograd.

Considering these effects of anorthite-content of plagioclase and the activity of water, and possible error in models of geothermometry and isochemical phase-diagram applied for estimated temperatures which certainly do not give results better than $\pm 50^\circ\text{C}$ (cf. Tracy *et al.* 1976; Thompson 1976), the observed absence of K-feldspar in the sillimanite-bearing gneisses does not appear to be anomalous. Therefore, the above study has an important bearing on the inverted metamorphic sequence in the Sikkim-Darjeeling Himalaya.

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