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# Metamorphic evolution of the northern Himachal Himalaya: phase equilibria constraints and thermobarometry

by Martin Wyss<sup>1</sup>

## Abstract

In the northern Himachal Himalaya, the Himalayan orogen consists of three tectonic domains, the Tethyan Himalaya, the High Himalayan Crystalline and the Lesser Himalayan Sequence, from NE to SW. Along the transect from the Spiti valley through eastern Lahul to the Parvati valley, the metamorphic evolution of these domains was studied in pelitic rocks of the Precambrian to Lower Cambrian Phe Formation and the Proterozoic Berinag Series. In this area four ductile deformational phases occur, including a first phase  $D_1$  of unknown geometry, the NE-verging  $D_2$  stacking of the Shikar Beh Nappe, SW-verging  $D_3$  folding and SW-directed  $D_4$  extrusion of the Crystalline Nappe between the SW-directed Main Central Thrust at the base and NE-directed normal movement in higher crustal levels. Detailed microtextural analyses allow these deformational phases to be related to four stages of metamorphic crystallization  $M_1$  to  $M_4$ . Phase equilibria based on discontinuous reactions allow to place constraints on the metamorphic conditions of each stage. The peak conditions of the main metamorphism  $M_2$  in the High Himalayan Crystalline are estimated by thermobarometry.

Metamorphism  $M_1$  reached sillimanite zone peak conditions for metapelites in the High Himalayan Crystalline with temperatures ranging between 620 and 760 °C. From the Tethyan Himalaya downsection to the underlying High Himalayan Crystalline a gradual succession of chlorite, biotite, garnet and kyanite Barrovian metamorphic mineral zones are observed that formed during  $M_2$  and  $M_3$  respectively, indicating a gradual transition between these two domains. As a consequence, the Tethyan Himalaya and the High Himalayan Crystalline represent metamorphic zones rather than tectonic units in the studied area.  $M_2$  is the main metamorphism in the High Himalayan Crystalline. It is a Barrovian-type metamorphism, reaching kyanite zone peak conditions of  $T \approx 700$  °C and  $P \approx 850$  MPa.

By contrast,  $M_3$  is the main metamorphism in the Tethyan Himalaya. In the High Himalayan Crystalline  $M_3$  peak conditions were close to  $M_2$  conditions within the kyanite zone. During  $D_4$  thrusting of the Crystalline Nappe over the Lesser Himalayan Sequence along the Main Central Thrust, the  $M_2/M_3$  Barrovian kyanite zone mineral assemblage was overprinted by a retrograde  $M_4$  greenschist facies metamorphism at the base of the High Himalayan Crystalline. In the underlying Lesser Himalayan Sequence  $D_4$  shearing was related to a prograde  $M_4$  greenschist facies metamorphism. Concurrently with greenschist facies  $M_4$  overprint along the Main Central Thrust, the  $M_2/M_3$  kyanite zone mineral assemblage was overprinted by a sillimanite-bearing assemblage in NE-dipping  $D_4$  normal shear zones in higher crustal levels. Barometry indicates that  $D_4$  normal shearing was associated with increasing decompression going upsection in the High Himalayan Crystalline. The overgrowth of the  $M_2/M_3$  mineral assemblage by  $M_4$  during extrusion of the Crystalline Nappe results in an apparent inverted metamorphic zonation, from greenschist facies at the base of the High Himalayan Crystalline grading upsection into kyanite- and then into sillimanite zone conditions.  $M_2$  metamorphic peak conditions decrease downsection from  $T \approx 700$  °C and  $P \approx 850$  MPa to  $T \approx 650$  °C and  $P \approx 720$  MPa in the kyanite zone. These data suggest that, in addition to the apparent inverted metamorphic zonation caused by  $M_4$  overprint,  $M_2$  isograds were passively deformed during SW-verging  $D_3$  folding and/or during  $D_4$  Crystalline Nappe extrusion, leading to a real inverted metamorphic zonation.

*Keywords:* polymetamorphism, inverted metamorphic zonation, thermobarometry, P-T path, nappe extrusion, synconvergent extension, Crystalline Nappe, High Himalayan Crystalline, Himachal Pradesh.

## 1. Introduction

The Himalayan orogen is the result of continental collision between the Indian and Asian plates that

began between the latest Paleocene and early Eocene (PATRIAT and ACHACHE 1984; GARZANTI et al. 1987; GARZANTI et al. 1996). The orogen is traditionally subdivided into five laterally contin-

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Tab. 1 Stratigraphy and petrography of the Tethyan Himalaya (TH), the High Himalayan Crystalline (HHC) and the Lesser Himalayan Sequence (LHS) in the the Spiti valley-eastern Lahul-Parvati valley area. Note the gradual transition between the Tethyan Himalaya and the High Himalayan Crystalline reflected by continuous stratigraphy and by gradually increasing metamorphic conditions.

Age	Formation	Petrography	
Upper Permian	Kuling	Siltstones, pelites, sandstones, limestones, marls	TH
Upper Permian	Ganmachidam	Conglomerates, sandstones	
Middle to Upper Carboniferous	Po	Siltstones, pelites, sandstones, limestones, marls	
Lower Carboniferous	Lipak	Limestones, dolomites, marls, sandstones, shales, gypsum	
Middle Devonian	Muth	Quartzsandstones	
Middle Ordovician	Thaple	Sandstones, siltstones, shales, dolomites	
Lower Cambrian	Angular unconf. Karsha	<b>Lower chlorite zone:</b> Siltstones, greywackes, sandstones, dolomites	HHC
Precambrian to Lower Cambrian	Phe	<b>Lower chlorite zone:</b> Siltstones, greywackes, sandstones	
		<b>Upper chlorite zone:</b> Slates	
		<b>Biotite zone:</b> Slates, phyllites	
		<b>Garnet zone:</b> Phyllites, garnet-bearing two-mica schists and biotite schists	
Lower Ordovician Hanuman Tibba intrusion Granites, gneiss		<b>Kyanite zone:</b> Garnet-, kyanite- and/or sillimanite-bearing two-mica gneisses and schists	
Precambrian	Mélange zone	Quartzite, calcschists, phyllites	LHS
	Berinag Group	Quartzite, biotite-chlorite schists, phyllites	

uous major tectonometamorphic units (GANSSE, 1964; Fig. 1, inset A). These are from north to south: (1) the Indus-Tsangpo Suture Zone (ITSZ), (2) the Tethyan Himalaya (TH), (3) the High Himalayan Crystalline (HHC), (4) the Lesser Himalayan Sequence (LHS) and (5) the Subhimalaya (SH). The Tethyan Himalaya consists of generally nonmetamorphic to low-grade Precambrian to Eocene sediments that were deposited on the northern margin of the Indian continent. The High Himalayan Crystalline represents the high-grade metamorphic zone of the orogen and was thrust over the Lesser Himalayan Sequence along the Main Central Thrust (MCT). The Lesser Himalayan Sequence is thrust over the Subhimalaya along the Main Boundary Thrust (MBT) and the Subhimalaya is thrust over the alluvial deposits of the Indo-Gangetic plain along the Himalayan Frontal Thrust (HFT). Locally, wedged between the High Himalayan Crystalline and the Lesser Himalayan Sequence, a sixth unit, the Lesser Himalayan Crystalline (LHC), occurs (VANNAY and GRASEMANN, 1998).

According to paleomagnetic data (BESSE et al., 1984), approximately 1000 km of the Indian plate margin has been underthrust below and/or accreted to the Asian continent after closing of the Neotethys ocean at the Indus Suture Zone. This shortening caused intracontinental thrusting and nappe formation associated with a polyphase folding and metamorphism within the leading edge of the subducted Indian plate (e.g. EPARD et al., 1995; WYSS et al., 1999). Nappes mainly developed within the High Himalayan Crystalline and the Tethyan Himalaya (FRANK et al., 1973; BASOULET et al., 1980; STECK et al. 1993).

One of the most debated features of the Himalayan geology is the inverted metamorphic zonation with metamorphic conditions increasing up-section from the Lesser Himalayan Sequence toward the High Himalayan Crystalline and also within the High Himalayan Crystalline, that has been reported from many sections all along the Himalayan orogen. Already observed by MALLET (1875), it has given rise to numerous models that have been proposed in the last 25 years. Some interpretations suggest a mechanical syn- to post metamorphic deformation of isograds through recumbent folding, thrusting or shearing (e.g. FRANK et al., 1977; THÖNI, 1977; BRUNEL and KIENAST, 1986; SEARLE and REX, 1989; TRELOAR et al., 1991; SWAPP and HOLLISTER, 1991; JAIN and MANICKAVASAGAM, 1993; EPARD et al., 1995; GRUJIC et al., 1996; HUBBARD, 1996 and references therein). Other models invoke heat transfer between the hot hanging wall and the colder foot-wall (LE FORT, 1975) or more complex interac-

tions between the tectonic evolution and thermal processes such as crustal radiogenic heating, heat advection, shear heating and heat focussing (e.g. HUBBARD, 1989; JAMIESON et al., 1996; DAVIDSON et al., 1997 and references therein) or the coupling of pre-deformational thermal disturbances with varying thermal conductivities between different rock types (GRASEMANN, 1993). In the investigated area, a real inversion of isograds occurs, but also an apparent inverted metamorphic zonation is observed, that is the result of a polyphase metamorphic evolution. Accordingly, given the important geological variations observed along the whole Himalayan orogen, it is not surprising, that a great number of models have been proposed until today and that there is little consensus about them.

Another much debated problem is the transition between the Tethyan Himalaya and the High Himalayan Crystalline in eastern Lahul. In the Zaskar area, the Zaskar Shear Zone (Fig. 1) juxtaposed low-grade metamorphic rocks of the Tethyan Himalaya against high-grade metamorphic rocks of the High Himalayan Crystalline by ductile normal shearing, causing a well-defined boundary between these two domains (SEARLE, 1986; HERREN, 1987; SEARLE et al., 1988; SEARLE and REX, 1989; GAPAIS et al., 1992; SEARLE et al., 1992; DÈZES et al., 1999). Consequently, the Tethyan Himalaya and the High Himalayan Crystalline have been interpreted as tectonic units. In order to link the Zaskar Shear Zone with comparable shear zones observed in Garhwal (PÈCHER and SCAILLET, 1989; PÈCHER, 1991; METCALFE, 1993; VANNAY and GRASEMANN, 1998; Fig. 1), SEARLE (1986) and SEARLE et al. (1988) proposed that this structure continues towards SE across eastern Lahul. This contrasts with observations made by HAYDEN (1904), FRANK et al. (1973), FRANK et al. (1977), FUCHS (1987), VANNAY (1993), STECK et al. (1993), VANNAY and STECK (1995) and WYSS et al. (1999), who reported neither a sedimentary nor a tectonometamorphic discontinuity between the Tethyan Himalaya and the High Himalayan Crystalline in Lahul.

Along the Spiti valley–Lahul–Parvati valley section, the polyphase metamorphic evolution of the Tethyan Himalaya and of the High Himalayan Crystalline can be studied in the pelitic Precambrian to Lower Cambrian Phe Formation rocks that occur all along the section. These metapelites recorded a great number of deformational phases and stages of metamorphic crystallization. The evolution of the Lesser Himalayan Sequence is best documented in this area in calcschists and pelitic schists of the Berinag Group.

On the basis of structural, microtextural and petrographic observations on outcrops and in

thin sections, four stages of metamorphic crystallization can be distinguished, that are related to four important phases of deformation. These phases of deformation result from a detailed investigation on the structural evolution of the Spiti valley–Lahul–Parvati valley transect by WYSS et al. (1999). The metamorphic conditions of the four stages of crystallization can be roughly determined on the basis of mineral assemblages. In the High Himalayan Crystalline, a P-T path and peak P-T conditions of the main metamorphism are determined using phase equilibria constraints based on discontinuous reactions and P-T estimates based on mineral chemistry data. The results are discussed with reference to the tectonometamorphic evolution of the Tethyan Himalaya, the High Himalayan Crystalline and the Lesser Himalayan Sequence and with particular emphasis on the inverted metamorphic zonation in the High Himalayan Crystalline and on the transition between the Tethyan Himalaya and the High Himalayan Crystalline.

## 2. Geological setting

The studied traverse consists of two sections. Section I climbs up from the Spiti river along the Taktsi valley, over Kun Zam La (La = pass) to Batal in eastern Lahul and then down the Upper Chandra valley, over Sara Umga La and down the Tos valley to the Parvati River. Section II

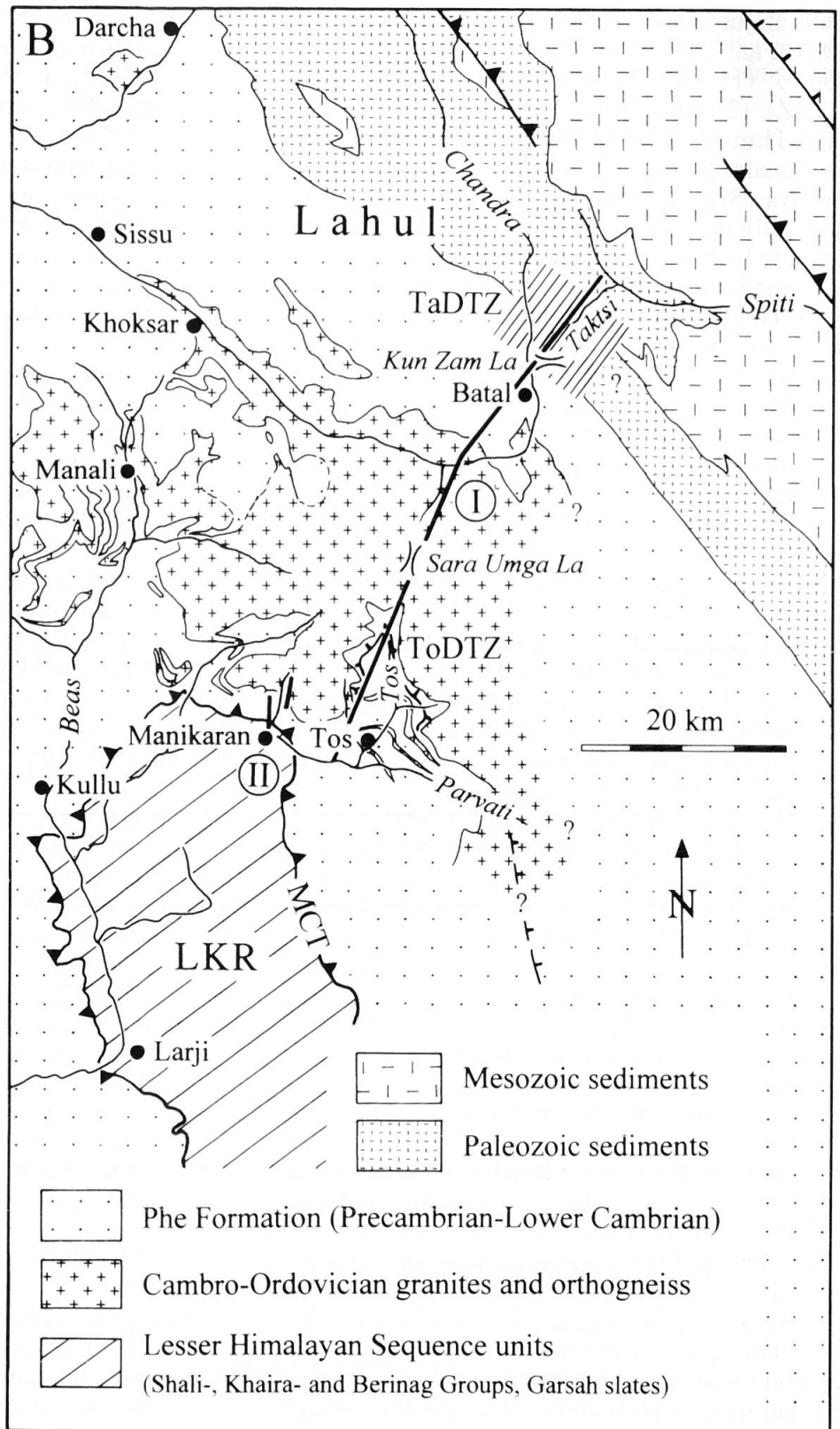


Fig. 1 Generalized map of tectonic zones and mineral zones in the NW Himalaya between Simla and Padum. Mineral zones are compiled after SPRING (1993), STECK et al. (1993), VANNAY (1993), VANNAY and GRASEMANN (1998), DÉZES et al. (1999) and own data. Inset A: Tectonic map of the Himalaya, modified after GUNTLI (1993). Inset B: Geological map of the study area with tracks of the sections I and II indicated (see plate 1). Section II is composite. Compiled after THÖNI (1977), FRANK et al. (1995), VANNAY and STECK (1995) and own data.

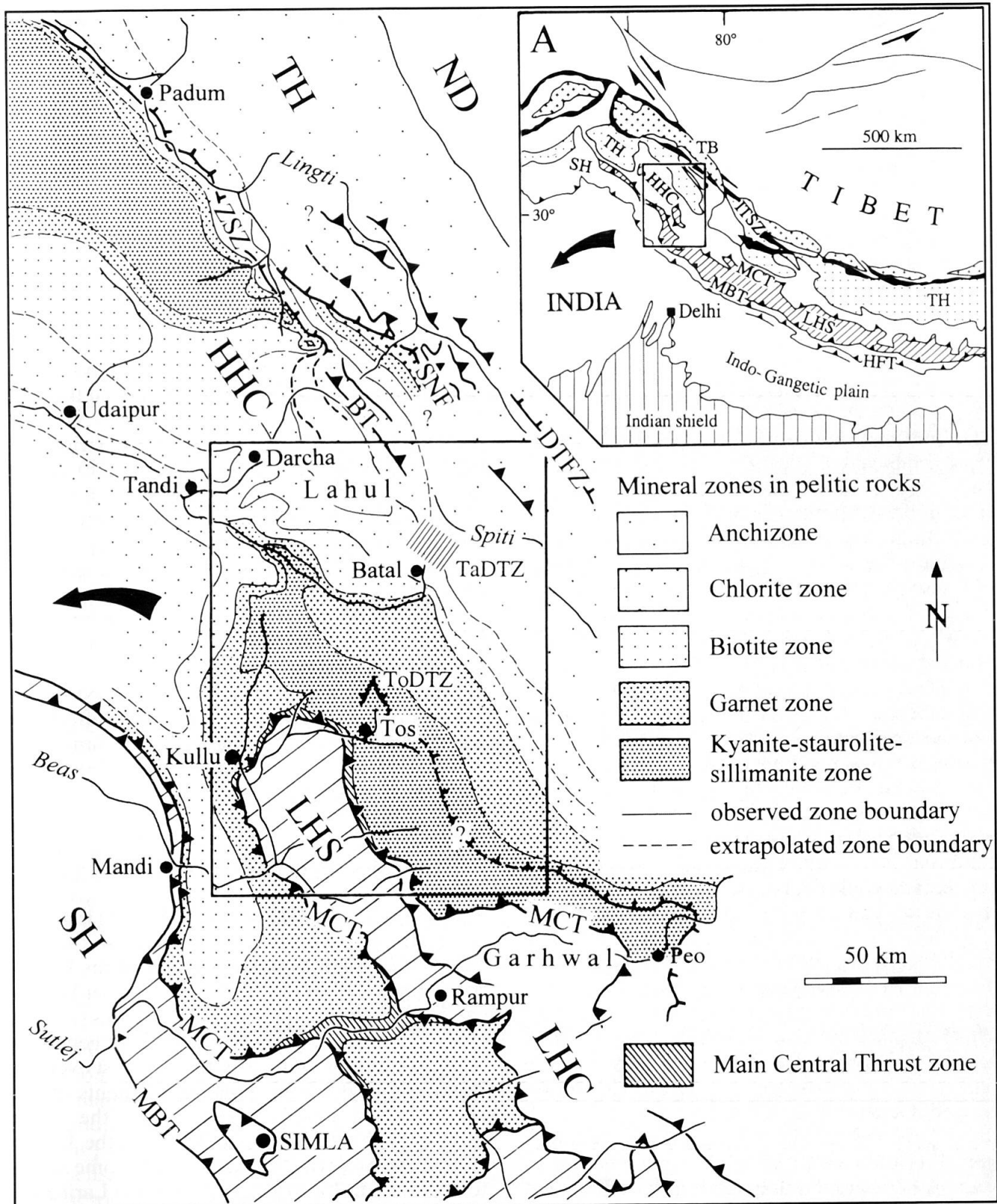


Fig.1 (cont). ND = Nyimaling Dome, TB = Transhimalayan Batholith, TH = Tethyan Himalaya, HHC = High Himalayan Crystalline, LHC = Lesser Himalayan Crystalline, LHS = Lesser Himalayan Sequence, SH = Subhimalaya, LKR = Larji-Kullu-Rampur tectonic window, DTFZ = Dutung-Thaktote Fault Zone, SNF = Sarchu Normal Fault, ZSZ = Zaskar Shear Zone, TaDTZ = Taktsi Dextral Transtension Zone, ToDTZ = Tos Dextral Transtension Zone, BT = Baralacha La Thrust, MCT = Main Central Thrust, MBT = Main Boundary Thrust, HFT = Himalayan Frontal Thrust. Note that the boundary between the Tethyan Himalaya and the High Himalayan Crystalline is well defined along the Zaskar Shear Zone (ZSZ), where it is clearly tectonic. Toward SE, the Zaskar Shear Zone dies out and the transition between these two domains is gradual as indicated by metamorphic mineral zones in metapelitic rocks.

consists of two short sections following the Brahmanganga valley N of Manikaran (Fig 1, inset B; Plate 1). Along this traverse, the Tethyan Himalaya, the High Himalayan Crystalline and the Lesser Himalayan Sequence are exposed from the NE to the SW (Fig. 1, Tab. 1). The entire traverse has first been studied in detail by WYSS et al. (1999). Earlier investigations were restricted to the northeasternmost (HAYDEN, 1904; STECK et al., 1993) and southwesternmost areas (FRANK et al., 1973, 1977; THÖNI, 1977).

## 2.1. LITHOTECTONIC UNITS AND STRATIGRAPHY

### 2.1.1. Tethyan Himalaya

The Tethyan Himalaya (Fig. 1) is the tectono-stratigraphically highest domain of the Himalayan orogen. In the NW Himalaya it represents a stack of north Himalayan nappes that were thrust from the northern edge of the subducted Indian continent towards SW (e.g. the Zaskar Nappes of BASSOULET et al. (1980), the Nyimaling-Tsarap Nappe of STECK et al. (1993) and the Mata Nappe of STECK et al. (1998)). Generally, the Tethyan Himalaya is composed of a nearly complete stratigraphic section of nonmetamorphic or low-grade metamorphic Upper Precambrian to Lower Eocene sediments that were deposited on the northern margin of the Indian plate (GAETANI and GARZANTI, 1991). In the studied area, the Tethyan Himalaya crops out in the Taktsi valley, where it consists of low-grade metamorphic Precambrian to Upper Permian sediments (Tab. 1), the metamorphic degree of which increases from the NE to the SW.

### 2.1.2. High Himalayan Crystalline

The Tethyan Himalaya is underlain by the High Himalayan Crystalline (Fig. 1), that represents the high-grade metamorphic zone of the Himalayan orogen. Generally, the High Himalayan Crystalline consists of middle- to high-grade metamorphic Precambrian to Jurassic sediments and of Cambro-Ordovician and Tertiary plutonic complexes. (FRANK et al., 1973; STECK et al., 1993). In the studied area, the High Himalayan Crystalline crops out in eastern Lahul, in the Tos valley, in the Parvati valley and in the Brahmanganga valley, where it is mainly made up of schists and paragneisses with a metamorphic grade increasing gradually from NE to SW, from greenschist facies to amphibolite facies. These rocks are metamorphic equivalents of the pelitic Precambrian to

Lower Cambrian Phe Formation sediments in the Tethyan Himalaya, indicating that the High Himalayan Crystalline represents the stratigraphic continuation of the Tethyan Himalaya downsection (STECK et al., 1993; VANNAY, 1993; WYSS et al., 1999). The rock types in the Phe Formation are summarized in table 1. According to HAYDEN (1904), FRANK et al. (1973, 1977), FUCHS (1987), VANNAY (1993), STECK et al. (1993), VANNAY and STECK (1995) and WYSS et al. (1999), neither a tectonic nor a metamorphic discontinuity occurs between the Tethyan Himalaya and the High Himalayan Crystalline in Lahul and Spiti. The Tethyan Himalaya is therefore considered to be transitional into the underlying High Himalayan Crystalline. Between the Chandra valley in Lahul and the Parvati valley area, the Phe Formation was intruded by the Hanuman Tibba Intrusion of Ordovician age ( $495 \pm 16$  Ma, Rb/Sr whole rock isochron, FRANK et al., 1977), which has a well-developed metamorphic contact aureole. WYSS et al. (1999) and WYSS and HERMANN (subm.) reported small mafic intrusions in the lowermost Phe Formation paragneisses of the Tos valley. In the SW, the High Himalayan Crystalline was thrust over the Lesser Himalayan Sequence along the Main Central Thrust with a hanging wall-to-the SW sense of shear, forming the Crystalline Nappe (FRANK, 1973; THÖNI, 1977). Thrusting along the Main Central Thrust affected a zone 4–5 km wide including the High Himalayan Crystalline and the underlying Lesser Himalayan Sequence (WYSS et al., 1999).

### 2.1.3. Lesser Himalayan Sequence

The Lesser Himalayan Sequence (Fig. 1) is generally made up of the Proterozoic to Lower Cenozoic Gondwanian sedimentary cover of the Indian continent (FRANK et al., 1977, 1995) and of Proterozoic metavolcanics, diabases and granites (FRANK et al., 1977; BHAT and LE FORT, 1992) of generally low metamorphic grade. This unit is thrust over the Paleogene to Neogene molassic sediments of the Subhimalayan foreland basin along the Main Boundary Thrust. In the studied area, the Lesser Himalayan Sequence is exposed in a dome structure that forms the northern part of the Larji-Kullu-Rampur tectonic window (LKR in figure 1), cropping out in the Parvati valley. In this area, it consists of greenschist facies, massive Precambrian quartzites of the Berinag Group, intercalated with pelitic and calc-pelitic schists and phyllites (Tab. 1) that underwent a polyphase structural evolution. On top of the Berinag Quartzites and below the base of the High Himalayan Crystalline a tectonic mélange zone occurs that is made up of lenses of

quartzite surrounded by calcschists and phyllites (Wyss et al., 1999).

## 2.2. TECTONICS

The studied transect reveals a complex tectonic evolution including an early Himalayan phase  $D_1$ , a NE-verging nappe formation  $D_2$ , SW-verging folding  $D_3$ , SW-directed nappe extrusion  $D_4$  and several late-stage phases. In this chapter an outline of the structural evolution is given based on detailed data presented by Wyss et al. (1999). Fig. 2 is a tectonic model for the phases  $D_2$ ,  $D_3$  and  $D_4$  based on STECK et al., 1998 and WYSS et al., 1999.

*Abbreviations:* S = schistosity, cleavage, F = fold, L = lineation, D = deformational phase, M = stage of metamorphic crystallization. Structural elements and stages of metamorphic crystallization that are related to the same deformational phase bear the same number, e.g.  $S_2$ ,  $F_2$ ,  $L_2$  and  $M_2$  are related to  $D_2$ .

### 2.2.1. Phase $D_1$ : Early event

The oldest deformation observed in the studied area is documented by a relict schistosity  $S_1$  and by rare isoclinal  $F_1$  folds in the High Himalayan Crystalline. The deformational conditions of  $D_1$  are poorly constrained. However,  $D_1$  is considered to be an important event, since  $D_1$  structures are well recognizable and their imprint is locally very strong although they were overprinted by several subsequent deformational phases.

### 2.2.2. Phase $D_2$ : NE-verging Shikar Beh Nappe

In the High Himalayan Crystalline, the main deformation is the isoclinal folding  $F_2$  that is related to the main schistosity  $S_2$ , to a penetrative, SW-NE-trending (or locally WSW-ENE and W-E-trending) mineral stretching lineation  $L_2$ , and to shear sense indicators such as rotated garnets and asymmetrically boudinaged dikes, indicating NE-directed deformation. These structures are interpreted to have formed during NE-directed stacking of the  $D_2$  Shikar Beh Nappe that has first been described by STECK et al. (1993) and VANNAY (1993). It is suggested that NE-directed stacking of the Shikar Beh Nappe occurred contemporaneously with SW-directed stacking of the north Himalayan nappes within the Tethyan Himalaya (STECK et al., 1993, 1998; EPARD et al., 1995). In

the Tethyan Himalaya, the existence of  $D_2$  structures is not unambiguously documented.

### 2.2.3. Phase $D_3$ : Initial stage of Crystalline Nappe formation

The subsequent SW-verging folding  $F_3$  represents a transitional phase of southwestwards propagating deformation between late thrusting at the front of the north Himalayan nappe stack (the Nyimaling-Tsarap Nappe of STECK et al. (1993) and the Mata Nappe of STECK et al. (1998)) and incipient  $D_4$  SW-directed thrusting of the Crystalline Nappe.  $F_3$  folding did not create a nappe structure but large-scale folds that are referred to as the Main Fold Zone.  $D_3$  occurs in both the Tethyan Himalaya and in the High Himalayan Crystalline, documenting the common evolution of these domains during  $D_3$ . In the Tethyan Himalaya  $F_3$  folds and the related main schistosity  $S_3$  represent the main phase of deformation. In the High Himalayan Crystalline by contrast,  $D_3$  is of subordinate importance with respect to  $D_2$  and a related schistosity  $S_3$  is rarely observed.

### 2.2.4. Phase $D_4$ : SW-verging Crystalline Nappe extrusion

During  $D_4$  the Crystalline Nappe was thrust towards SW over the Lesser Himalayan Sequence along the Main Central Thrust. Thrusting on the Main Central Thrust occurred in a zone 4–5 km wide and included both the High Himalayan Crystalline and the underlying Lesser Himalayan Sequence (Fig. 3, Plate 1). It is characterized by a NE-SW-trending mineral stretching lineation and by S-C fabrics, indicating a hanging wall-to-the SW sense of shear. In the Crystalline Nappe stretching lineations and shear bands are concentrated into shear zones some meters to some hundred meters in width, the density of which increases with decreasing distance from the base of the nappe. In the Lesser Himalayan Sequence  $D_4$  shearing is strongest in the schists and phyllites of the mélangé zone directly below the base of the Crystalline Nappe. In the underlying Berinag Group,  $D_4$  shear zones are restricted to thin layers of schists and phyllites within the massive Berinag Quartzite (Wyss et al., 1999).

According to Wyss et al. (1999), SW-directed thrusting at the base of the Crystalline Nappe occurred simultaneously with NE-directed dextral transtension in higher crustal levels of the High Himalayan Crystalline and at the transition between the High Himalayan Crystalline and the Tethyan Himalaya. In the High Himalayan Crys-



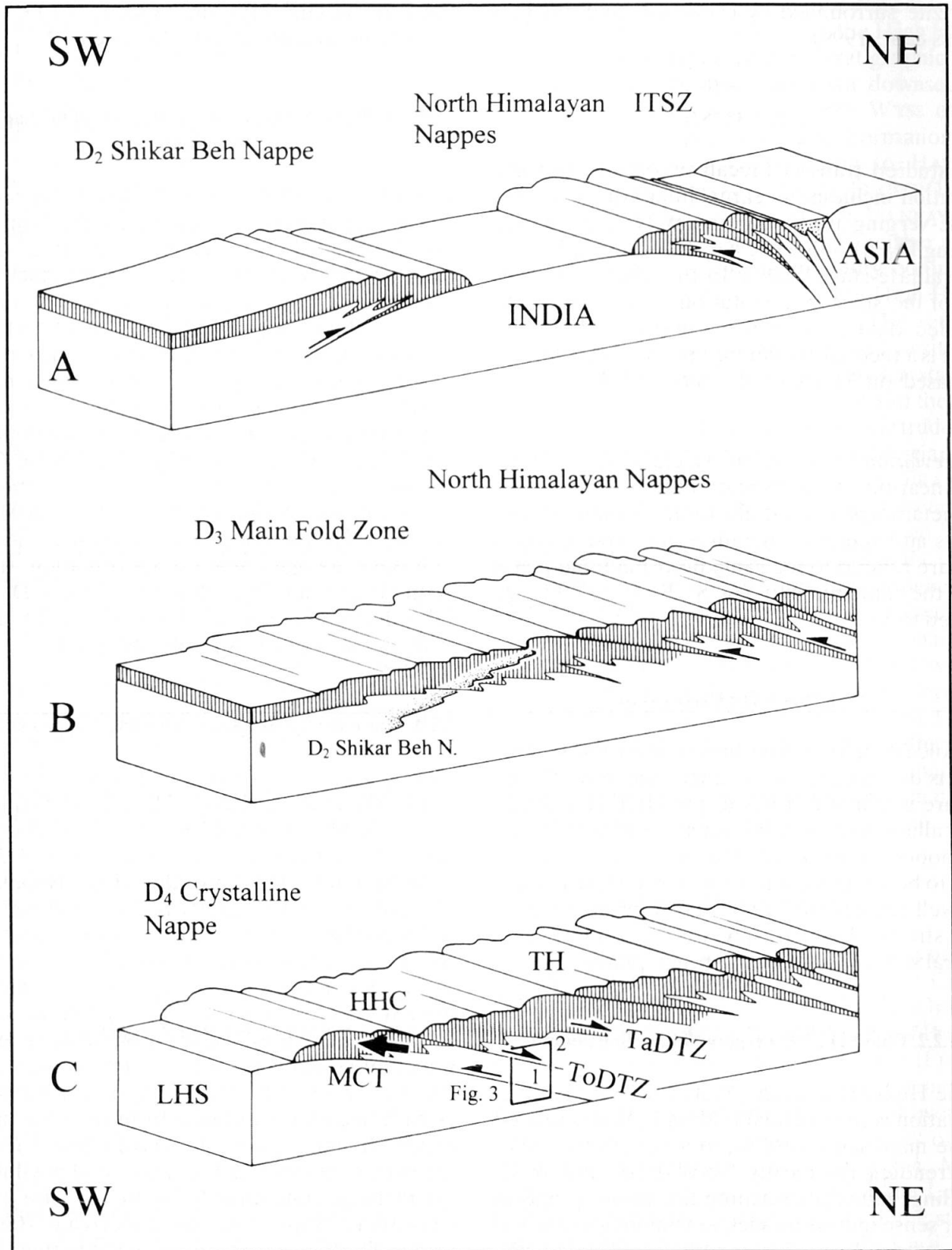


Fig. 2 Model for the tectonic evolution of the NW Himalaya from the Indus Suture Zone (ITSZ) in the NW to the Parvati valley in the SW. Stages B and C focus on the section from the Spiti valley through Lahul to the Parvati valley. The hatched layer is drawn to clarify the geometry of deformation, it has no stratigraphic or tectonic meaning. (A) Initiation of the NE-verging D<sub>2</sub> Shikar Beh Nappe within the Indian crust SW of the subduction zone and contemporaneous initiation of the SW-verging north Himalayan nappes from the northeastern edge of the Indian crust. The Indian continent was subducted at the base of the north Himalayan nappes. (STECK et al., 1993, 1998; EPARD et al., 1995; WYSS et al., 1999). (B) Ongoing subduction of the Indian continent at the base of the north Himalayan nappes. The front of the north Himalayan nappes reached the area NE of the front of the Shikar Beh Nappe. Initiation of the D<sub>3</sub> SW-verging Main

talline ductile  $D_4$  dextral transtension is revealed by a ENE-plunging  $L_4$  mineral stretching lineation on NE-dipping shear bands that crosscut  $S_2$ , with shear sense criteria indicating a hanging wall down-to-the ENE shear sense. These shear bands are rarely longer than some decimeters, but they are often concentrated locally and combined into huge shear zones of some hundred meters in length and several meters in width. Locally,  $D_4$  dextral shearing rotated  $L_2$  stretching lineations from an originally SW-NE-trending orientation to a W-E-trending orientation. This ductile dextral transtension can be well observed in the Tos valley, where it is distributed over a zone about 5 km wide, referred to as the Tos Dextral Transtension Zone (Fig. 3, Plate 1).

At the transition between the Tethyan Himalaya and the High Himalayan Crystalline,  $D_4$  is characterized by clockwise rotation and NE-directed overturning of formerly SW-verging  $F_3$  folds that occurred in a zone about 6 km wide referred to as the Taktsi Dextral Transtension Zone. The lateral limits of the Taktsi Dextral Transtension Zone are not known, therefore, it is indicated as a diffuse zone in figure 1. Concurrent SW-directed thrusting at the base of the Crystalline Nappe and NE-directed extension in two higher tectonic levels indicate that the Crystalline Nappe was extruded in two parts towards SW between the Lesser Himalayan Sequence and the Tethyan Himalaya (1 and 2 in figure 2, plate 1).

$D_4$  transtension is interpreted to be part of a large-scale system of normal and/or dextral strike-slip shear zones that were reported from many sectors of the Himalayan orogen between the Tethyan Himalaya and the High Himalayan Crystalline (e.g. BURG et al., 1984; BRUN et al., 1985; BURCHFIEL and ROYDEN, 1985; SEARLE, 1986; HERREN, 1987; SEARLE et al., 1988; PÉCHER and SCAILLET, 1989; PÉCHER, 1991; PÉCHER et al., 1991; GAPAIS et al., 1992; BURCHFIEL, 1992; METCALFE, 1993), termed the North Himalayan Shear Zone (PÉCHER et al., 1991) or the South Tibetan Detachment System (BURCHFIEL et al., 1992). In these shear zones, normal and/or dextral strike-slip shearing are generally related to a well defined structural detachment that juxtaposes the

nonmetamorphic or low-grade metamorphic rocks of the Tethyan Himalaya against high-grade metamorphic rocks of the High Himalayan Crystalline (e.g. the Zaskar Shear Zone: Fig. 1; SEARLE, 1986; HERREN, 1987; SEARLE et al., 1988; SEARLE and REX, 1989; GAPAIS et al., 1992; SEARLE et al., 1992; DÉZES et al., 1999). The Taktsi Dextral Transtension Zone fits well into the Central Himalayan Shear Zone system according to its sense of movement and its tectonic position. However, fold reorientation did not cause a disruption of previous structures. Consequently, in the study area, a tectonic boundary between the Tethyan Himalaya and the High Himalayan Crystalline is not observed.

### 2.2.5. Late stage phases

$D_4$  is postdated by a NE-verging backfolding and doming phase, by restarting of SW-directed nappe extrusion and finally by compression parallel to the mountain chain. As none of these phases are related to important stages of metamorphic crystallization, they are not considered here when discussing the metamorphic evolution.

## 3. Metamorphic petrography

This study focuses on pelitic rocks as they constitute the major part of the rocks exposed along the Spiti valley-Lahul-Parvati valley transect and as they recorded a great number of metamorphic mineral assemblages and deformational structures. The pelitic rocks of the Karsha and Phe Formations make up most of the Tethyan Himalaya in the studied area and the High Himalayan Crystalline is dominated by Phe Formation metapelites. In the Lesser Himalayan Sequence, pelitic schists and phyllites are intercalated with the Berinag Quartzites (Tab. 1). The Karsha Formation consists of metasandstones, metagreywackes and metasiltstones intercalated with dolomite beds. The Phe Formation consists of monotonous alternations of subarkosic to arkosic metasandstones, metagreywackes and metasiltstones with

### Fig.2 (cont.)

Fold Zone SW of the north Himalayan nappe stack.  $D_3$  is a southwestward progressing transitional phase between late thrusting of the north Himalayan nappes and early formation of the Crystalline Nappe (Wyss et al., 1999).

(C) Formation of the Main Central Thrust (MCT) and SW-directed  $D_4$  extrusion of the Crystalline Nappe between the Lesser Himalayan Sequence (LHS) and the north Himalayan nappe stack. This is the result of contemporaneous SW-directed thrusting along the Main Central Thrust and NE-directed normal movement in the Tos Dextral Transtension Zone (ToDTZ) and in the Taktsi Dextral Transtension Zone (TaDTZ). As a consequence, the Crystalline Nappe is subdivided into a lower (1) and an upper (2) part (Wyss et al., 1999). During  $D_4$ , the zone of subduction has stepped from the base of the north Himalayan nappes to the Main Central Thrust. The TaDTZ has no well defined limits. HHC = High Himalayan Crystalline, corresponding more or less with the Crystalline Nappe; TH = Tethyan Himalaya.

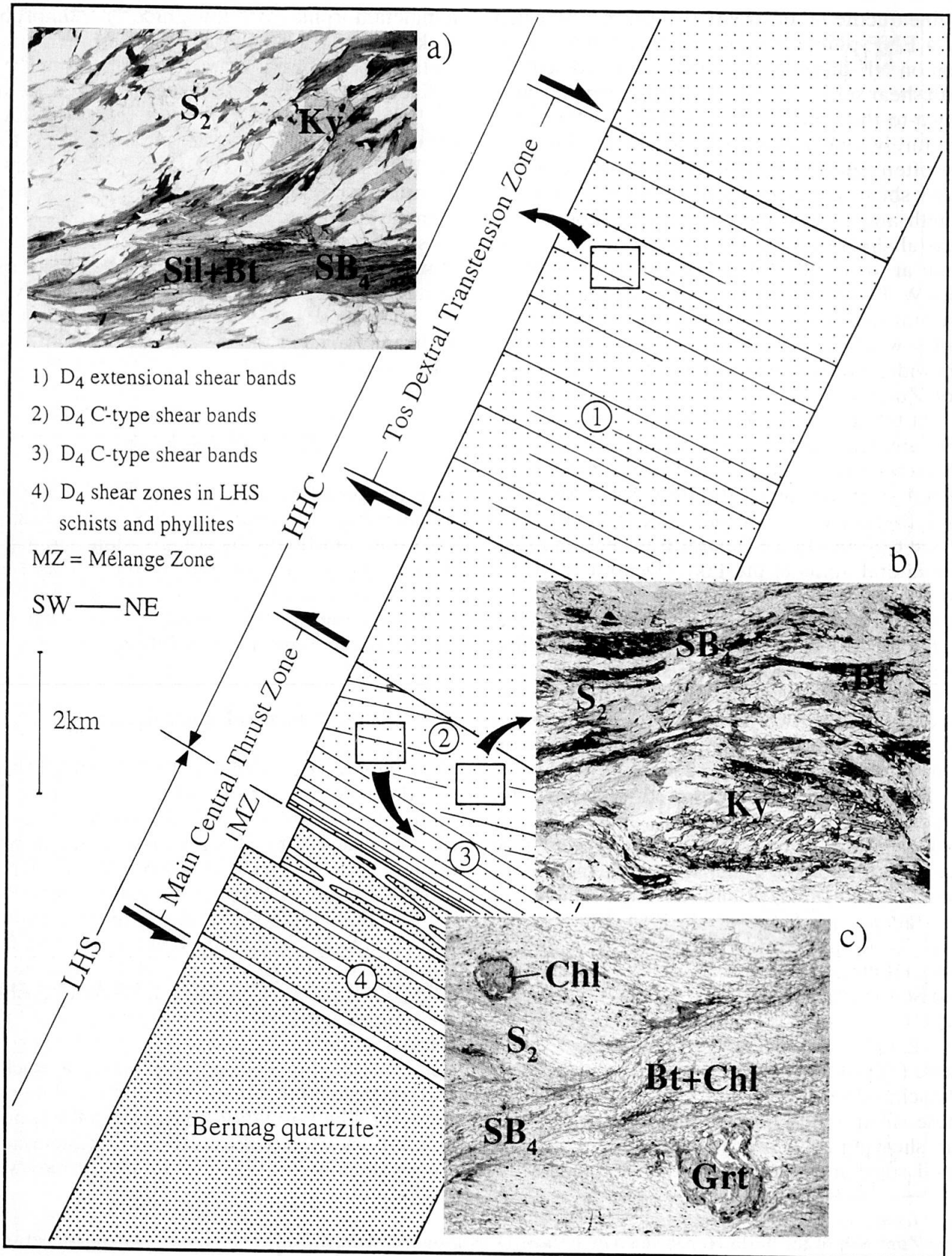


Fig. 3 Simplified section across the Main Central Thrust Zone and the Tos Dextral Transtension Zone (LHS = Lesser Himalayan Sequence, HHC = High Himalayan Crystalline, corresponding to the Crystalline Nappe in the Tos valley-Parvati valley area). Insets: Selected photomicrographs illustrating D<sub>4</sub>/M<sub>4</sub>. The scale given within brackets corresponds to the long dimension of the images, SW is to the left hand side. (a) Sillimanite and biotite on a D<sub>4</sub> extensional shear band, crosscutting kyanite on S<sub>2</sub> [20 mm]. (b) Kyanite on S<sub>2</sub>, partly transformed to muscovite [9.5 mm]. (c) D<sub>4</sub> C-type shear bands SB<sub>4</sub> crosscutting S<sub>2</sub>. Garnet rims are transformed to chlorite [15 mm].

millimeter- to meter-scale bedding that are mostly low-Al pelites (SPEAR, 1993; Fig. 4). Locally, carbonaceous graphitic quartzites and thin marly layers are intercalated. The schists and phyllites intercalated with massive quartzites of the Berinag Group in the Lesser Himalayan Sequence are also low-Al pelites (Fig. 4).

The four tectonic events  $D_1$  to  $D_4$  are documented by structures as well as by the related stages of metamorphic crystallization  $M_1$  to  $M_4$ . With the exception of stage  $M_{1a}$ , all stages of metamorphic crystallization are related to the formation of a schistosity and are therefore considered to represent syn-tectonic crystallization. Metamorphic main mineral assemblages in the metapelites of the Tethyan Himalaya and the High Himalayan Crystalline allow four prograde Barrovian metamorphic mineral zones to be distinguished. These are, from NE to SW, the chlorite zone, corresponding roughly with the Tethyan Himalaya and the biotite-, garnet- and kyanite zones, corresponding with the High Himalayan Crystalline (Plate 1). In the garnet and kyanite zones,  $M_2$  is the main metamorphism (related to the main schistosity  $S_2$ ) and consequently, these zones are defined on the basis of the  $M_2$  mineral assemblages. In the chlorite and biotite zones,  $M_3$  is the main metamorphism (related to the main schistosity  $S_3$ ) and therefore, these zones are defined on the basis of the  $M_3$  mineral assemblages. In the Main Central Thrust Zone, the  $M_2$  kyanite

zone assemblage was retrogressively overprinted by a greenschist facies metamorphism at the base of the High Himalayan Crystalline. In the uppermost Lesser Himalayan Sequence, a coeval prograde greenschist facies metamorphism occurred that reached biotite zone conditions (Plate 1).

The interpretation of the pressure and temperature stability conditions of the characteristic assemblages in the metapelites is based on a composite petrogenetic grid adapted to the observed compositions (Fig. 5). Most of the reaction curves limiting these stability fields have a steep  $dP/dT$  slope, implying a strong temperature control on the characteristic mineral equilibria.

*Mineral abbreviations:* Ab = albite, As = aluminosilicate, An = anorthite, Bt = biotite, Carb = carbonate, Chl = chlorite, Cld = chloritoid, Grt = garnet, Kfs = K-feldspar, Ky = kyanite, Ms = muscovite, Pa = paragonite, Phl = phlogopite, Pl = plagioclase, Qtz = quartz, Sil = sillimanite, St = staurolite, Zo = zoisite.

### 3.1. TETHYAN HIMALAYA AND HIGH HIMALAYAN CRYSTALLINE

#### 3.1.1. Chlorite zone

##### *Lower chlorite zone*

In the lower chlorite zone the pelites of the

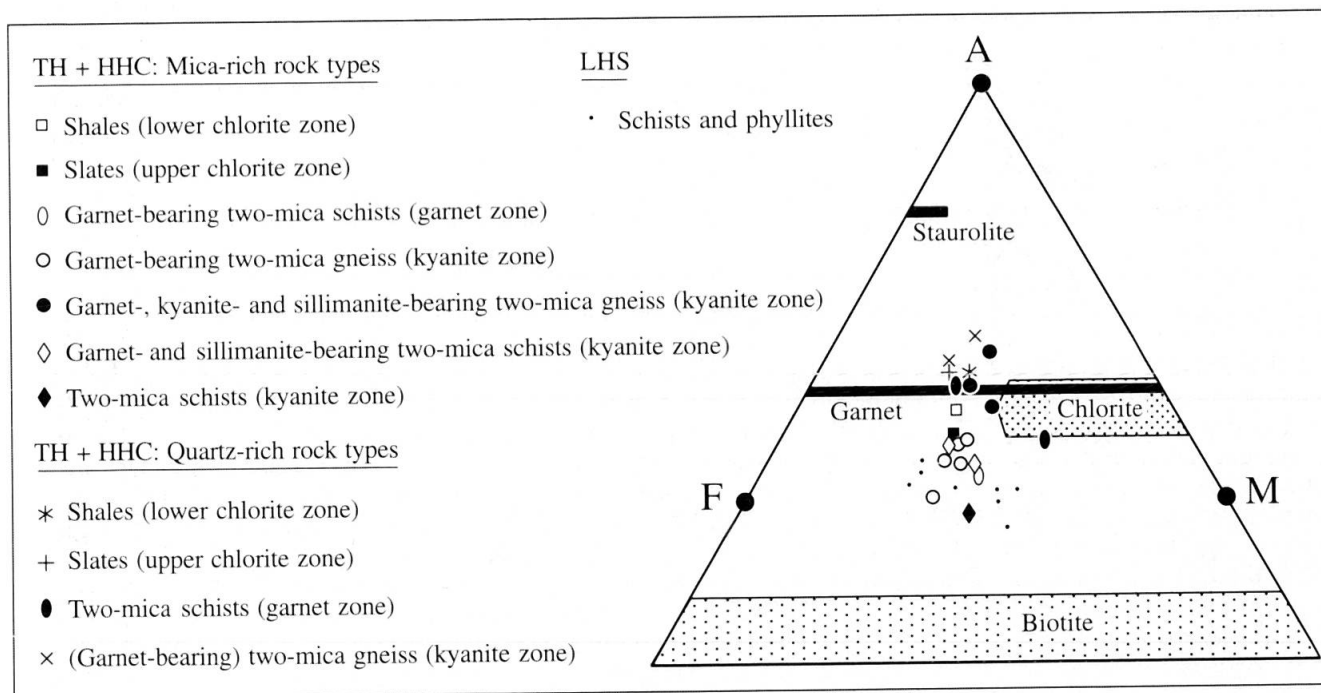


Fig. 4 Bulk compositions of Phe Formation pelites from the Tethyan Himalaya (TH) and the High Himalayan Crystalline (HHC) and of pelitic schists and phyllites from the Berinag Group in the Lesser Himalayan Sequence (LHS), summarized in an AFM diagram. All analyzed samples are low-Al pelites (SPEAR, 1993). The analytical procedure is given in the appendix.

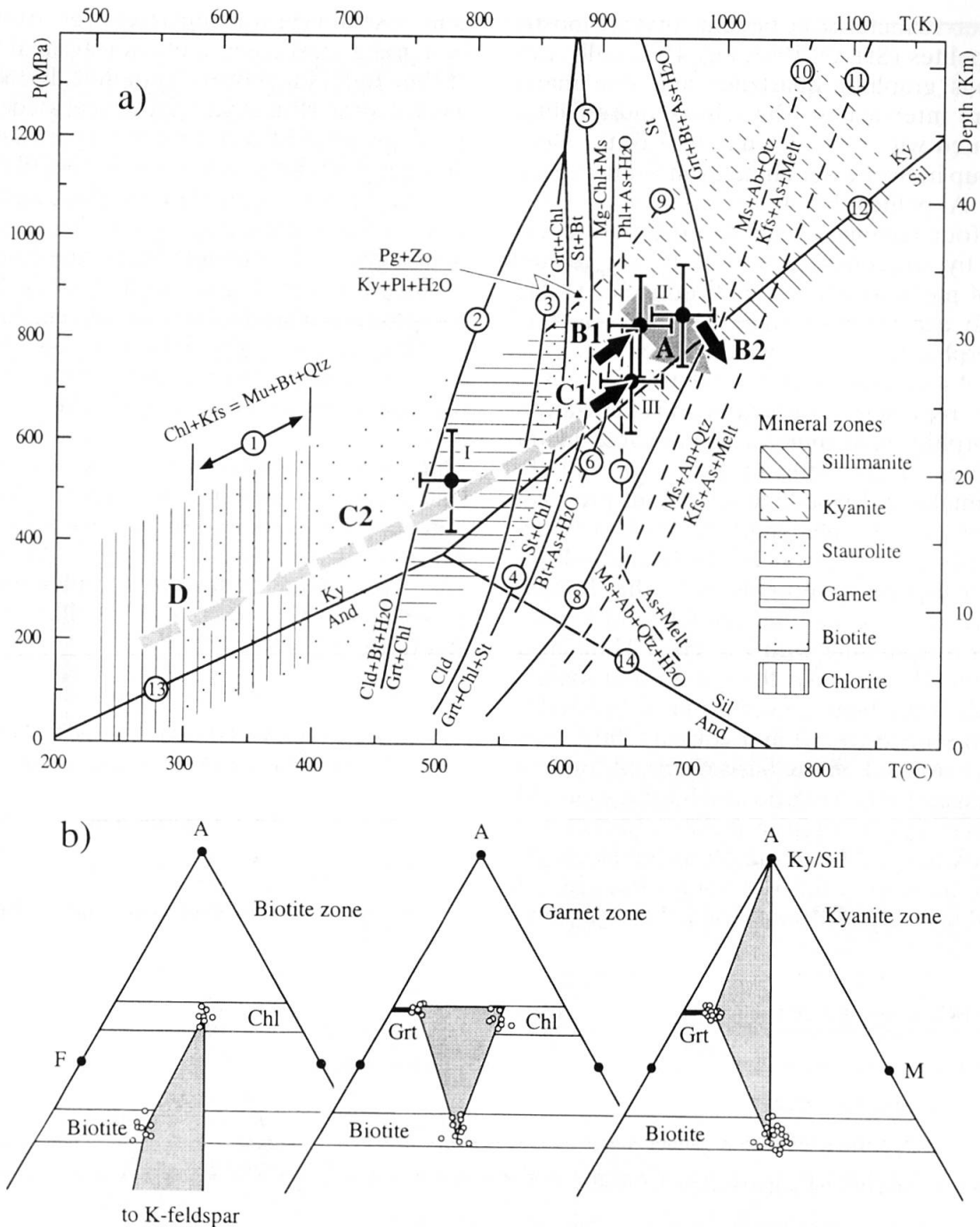


Fig. 5 (a) Composite petrogenetic grid for the metapelites in the Spiti valley-eastern Lahul-Parvati valley area. P-T stability fields for the mineral assemblages characteristic of the observed Barrovian mineral zones are indicated. Solid lines correspond to reactions in the (Mn)KFMASH system and dashed lines correspond to reactions in the CKNASH system. Reactions (2) and (4) that limit the garnet zone field are plotted for  $X_{\text{sps}} = 0.2$ , corresponding to the maximum  $X_{\text{sps}}$  in the analyzed garnets. Reactions (1) to (6), (8), (9) and (12) to (14) are according to SPEAR and CHENEY (1989) and SPEAR (1993). The vapor present melting reaction (7) represents the minimum conditions for anatexis in water-saturated Ms + Pl + Qtz metapelites and the reactions (10) and (11) correspond to vapor-absent melting conditions (THOMPSON and TRACY, 1979; LE BRETON and THOMPSON, 1988). P-T estimate I corresponds to sample EMP 1 from the garnet zone, II corresponds to EMP 2 (right) and EMP 8 (left) from the kyanite zone and III corresponds to EMP 11 from the Main Central Thrust Zone. The arrows A-D indicate P-T path segments for the stages of metamorphic crystallization  $M_1$  to  $M_4$  deduced from textural relations discussed in the text. Arrow A:  $M_{1a}$ - $M_{1b}$ ; arrows B1 and B2:  $M_2$ - $M_4$  in the garnet and kyanite zones and in the Tos Dextral Transtension Zone; arrows C1 and C2:  $M_2$ - $M_4$  in the High Himalayan Crystalline part of the Main Central Thrust Zone; arrow D:  $M_4$  in the Lesser Himalayan Sequence part of the Main Central Thrust Zone. Black arrows are constrained by phase equilibria and P-T estimates, whereas grey arrows indicate qualitative tendencies of P-T paths. (b) Mineral compositions and assemblages characteristic of the biotite-, garnet- and kyanite zones are represented in AFM diagrams. Note the decreasing FeO/(FeO + MgO) ratio in biotite with increasing metamorphic grade.

Phe and Karsha formations display all the characteristics of sedimentary rocks. Graded bedding, rhythmites of alternating sandstones and siltstones as well as cross stratifications are often observed, and on bed surfaces, ripple marks, desiccation cracks, synaeresis cracks and trace fossils are abundant. The metasiltstones are slates that display strong cleavages  $S_2$  and  $S_3$  (Fig. 6a), whereas cleavages in more arenitic layers are weak. The rocks mainly consist of detrital monocrystalline grains of quartz, plagioclase, sericitised K-feldspar, muscovite and accessory minerals (detrital rutile, titanite, tourmaline, zircon, apatite, epidote, clinozoisite and opaque minerals). Muscovite is abundant as buckled detrital grains, generally subparallel to bedding. This preferred orientation  $S_2$  most probably represents the record of the deformation phase  $D_2$ , but it may also represent a "sedimentary cleavage" that formed due to a preferential orientation of mica during sedimentation. The main schistosity  $S_3$  is defined partly by reoriented detrital muscovite and partly by more fine-grained muscovite that is interpreted as the product of recrystallization of the original detrital pelitic fraction during the main metamorphism  $M_3$ . Detrital muscovite is often observed as intergrowths with chlorite. The crystallization of chlorite can not directly be related to the formation of  $S_3$ . Consequently, it is not clear, if chlorite crystallized during  $M_3$  or during an earlier stage of metamorphic crystallization that could have been related to  $D_2$ .

#### *Upper chlorite zone*

Sedimentary bedding still dominates the aspect of the Phe Formation in the upper chlorite zone. Sedimentary structures, however, are rarely preserved. The metapelites consist of slates displaying components of detrital quartz, K-feldspar and plagioclase in a matrix generally made up of muscovite, chlorite and locally also of graphite, that define the main schistosity  $S_3$ . During the main metamorphism  $M_3$  fine-grained muscovite grew coarser from NE to SW and detrital muscovite mostly recrystallized. The grain size of chlorite also increases considerably from NE to SW.  $S_2$  is preserved as rare relics marked by muscovite and chlorite within quartz layers (Fig. 6b). By contrast with the oldest schistosity in the lower chlorite zone, these  $S_2$  relics are considered to definitely document the existence of the deformation  $D_2$  and the metamorphism  $M_2$  in the upper chlorite zone as they are unlikely to represent a sedimentary cleavage.

The stability of the  $M_2/M_3$  assemblage chlorite + K-feldspar in the chlorite zone is limited by re-

action (1) in the KFMASH system. The P-T region corresponding to this assemblage is not certain, maximum temperatures ranging between  $T \approx 300$  and  $400$  °C (Fig. 5).

#### **3.1.2. Biotite zone**

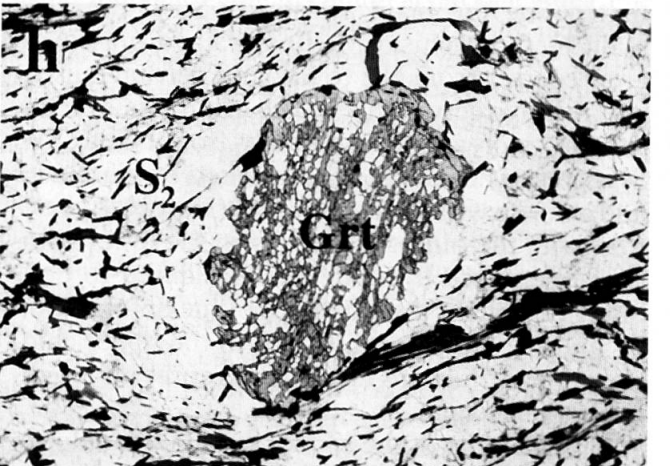
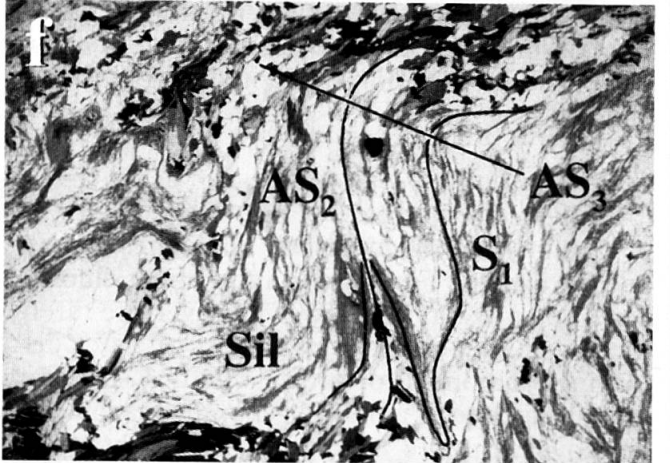
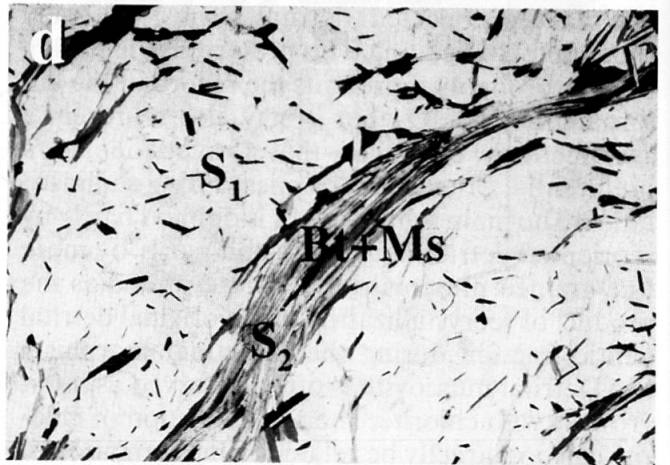
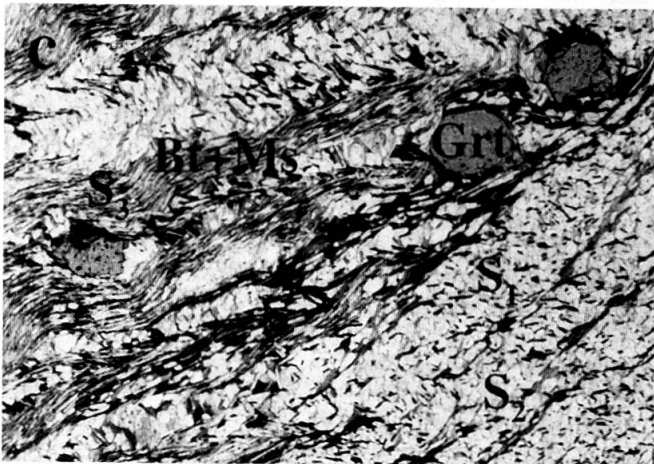
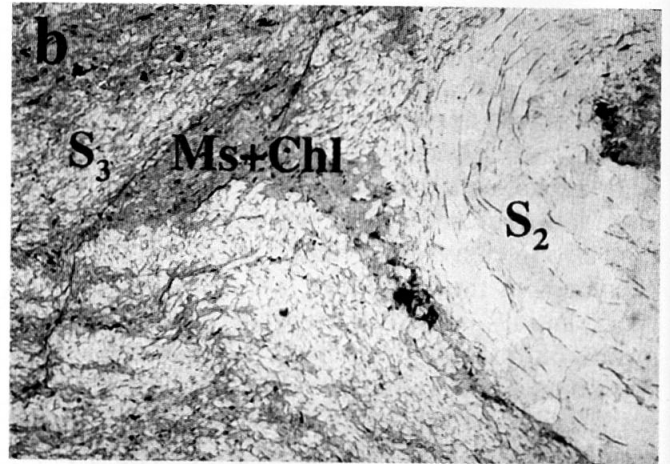
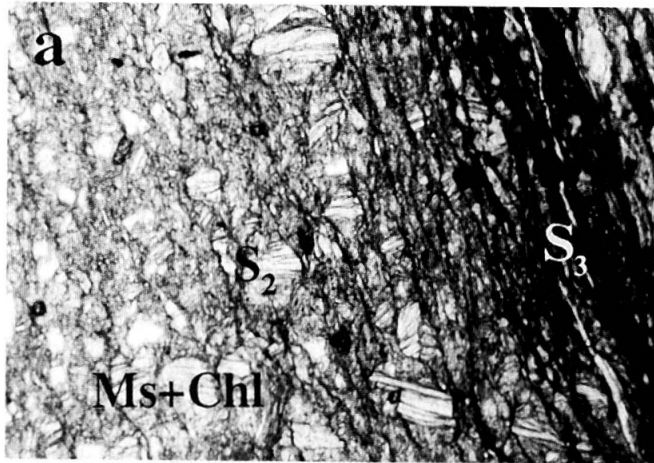
In the biotite zone the Phe Formation metapelites are slates and phyllites that differ from the upper chlorite zone slates by the presence of additional biotite on the main schistosity  $S_3$ , indicating that  $M_3$  reached biotite zone conditions in this area. Relics of  $S_2$  within quartz layers by contrast are free of biotite. Accordingly, the main metamorphism  $M_3$  reached biotite zone conditions, whereas  $M_2$  did not exceed chlorite zone conditions.

The first appearance of biotite in low-Al pelites is defined by reaction (1) in the KFMASH system, the location of which is most likely between  $T \approx 300$  and  $400$  °C (Fig. 5).

#### **3.1.3. Garnet zone**

In the garnet zone the metapelites are mainly phyllites, two-mica schists and biotite schists with additional garnet. The sedimentary bedding is well preserved on all scales, the main schistosity is  $S_2$ . The rocks consist of quartz, plagioclase, biotite, muscovite, chlorite and garnet poikiloblasts. Locally, graphite is abundant. The main schistosity  $S_2$  as well as  $S_3$  and relics of  $S_1$  are marked by biotite, muscovite and chlorite (Fig. 6c). Garnet is on one hand wrapped around by  $S_2$  in an augen-like texture, documenting its existence prior to  $S_2$ , on the other hand it overgrows  $S_1$ , establishing that it crystallized post- $D_1$  and pre-or syn- $D_2$ . As a consequence, garnet growth is considered to be related to both stages of metamorphic crystallization  $M_1$  and  $M_2$ , indicating that both of them reached garnet zone conditions in this area. By contrast, no garnet growth related to  $S_3$  is observed. It is, however, important to note that garnets do not show any signs of resorption. This may indicate that the  $M_3$  metamorphic conditions were also within or very close to the stability of garnet.

The stability of the main  $M_2$  mineral assemblage garnet + biotite + chlorite + muscovite + quartz is limited by the reactions (2) and (4) in the MnKFMASH system (for  $X_{\text{sps}}$  up to 0.2, the maximum content measured in the analyzed garnets), constraining the temperature between  $T \approx 460$  and  $600$  °C (Fig. 5). The discontinuous reaction (2) implies that garnet should crystallize consuming biotite and chloritoid. However, as no chloritoid occurs in the low-Al pelites of the biotite zone,



this reaction is unlikely to account for the formation of garnet. Garnet is more likely to have formed by the continuous reaction chlorite + quartz + muscovite = garnet + annite + H<sub>2</sub>O. The P-T conditions of this reaction depend on the mineral compositions in the KFMASH system and therefore, it does not show up in the grid in figure 5.

### 3.1.4. Kyanite zone

This zone is referred to as the kyanite zone because over 90% of the rocks display the kyanite zone M<sub>2</sub>/M<sub>3</sub> metamorphic mineral assemblage. However, the M<sub>1b</sub> and M<sub>4</sub> sillimanite-bearing mineral assemblages also occur in this zone. In plate 1, the occurrence of M<sub>4</sub> sillimanite in the Tos Dextral Transtension Zone is indicated. The metapelites in the kyanite zone are generally medium- to coarse-grained garnet-bearing two-mica gneisses and schists with additional kyanite and/or sillimanite, consisting of quartz, plagioclase, biotite, white mica, garnet, kyanite, sillimanite, tourmaline and rare K-feldspar. Staurolite, apatite, rutile, ilmenite, zircon and monazite are accessories. These rocks display a strong metamorphic mineral segregation separating quartz- and plagioclase-rich microlithons from biotite- and muscovite-rich cleavage domains that represent the main schistosity S<sub>2</sub> (Fig. 6d). Relics of S<sub>1</sub> occur within the microlithons and in quartz layers, defined by biotite and muscovite. Locally, in mica-rich layers, a S<sub>3</sub> schistosity occurs that is also defined by biotite and muscovite. The primary sedimentary bedding is indicated by variable mineral ratios. Kyanite and/or sillimanite are abundant only in a few beds that constitute about 5% of the Phe Formation paragneisses. In the remaining 95%, aluminosilicates are rare.

The first stage of metamorphic crystallization M<sub>1a</sub> is represented by relics of kyanite in the center of aggregates of fibrolitic sillimanite that are oriented parallel to S<sub>1</sub> (Fig. 6e). M<sub>1a</sub> is therefore interpreted to be related to an early stage of D<sub>1</sub>

that reached kyanite zone conditions. The second stage of metamorphic crystallization M<sub>1b</sub> is characterized by the mineral assemblage biotite, muscovite and locally fibrolitic sillimanite on S<sub>1</sub> (Fig. 6f). Relics of K-feldspar frequently occur in samples in which S<sub>1</sub> is preserved. The M<sub>1b</sub> mineral assemblage biotite + muscovite + sillimanite ± K-feldspar indicates conditions close to the vapor-absent melting reactions (10 and 11 in figure 5) and between the kyanite = sillimanite (12) and sillimanite = andalusite (14) equilibria. This limits the temperature range between T ≈ 620 and 760 °C, and pressure may have attained 900 MPa.

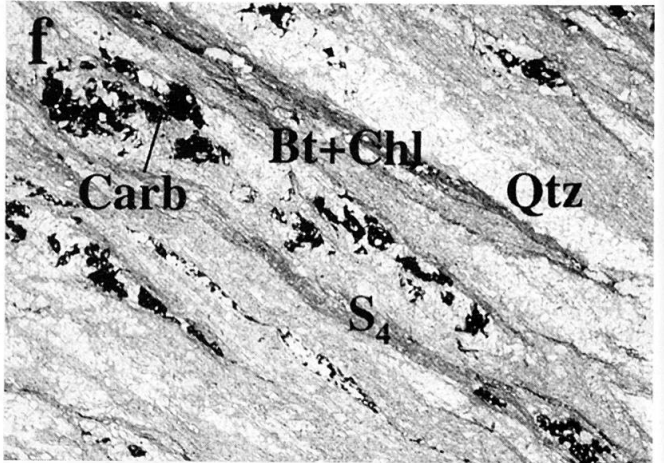
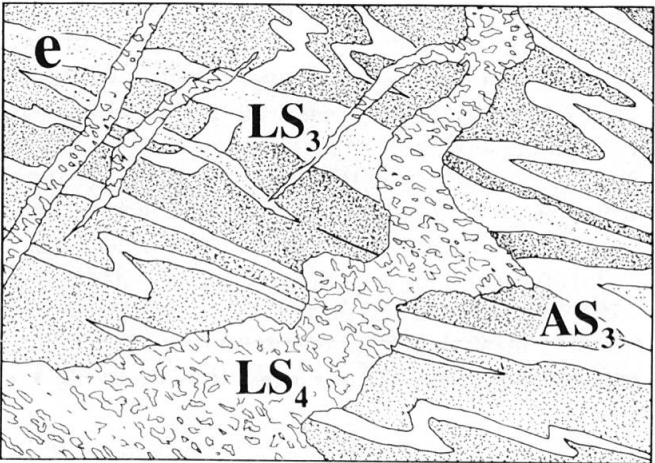
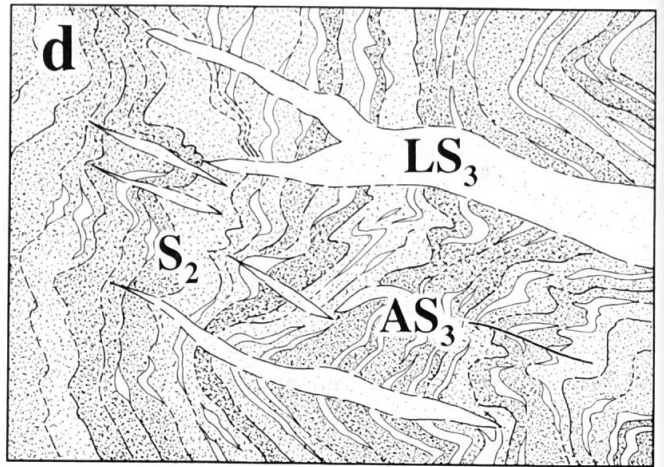
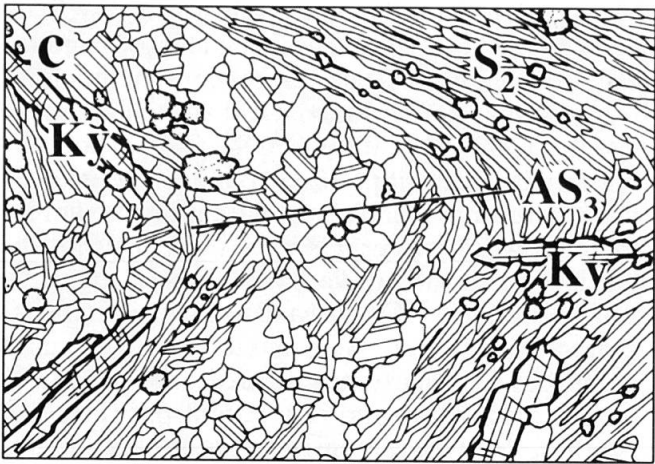
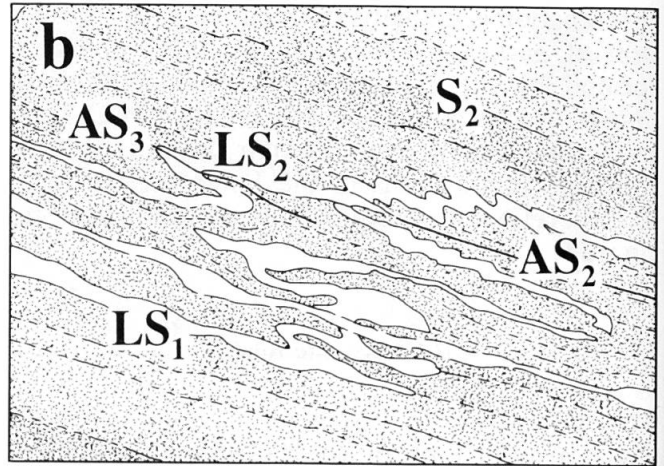
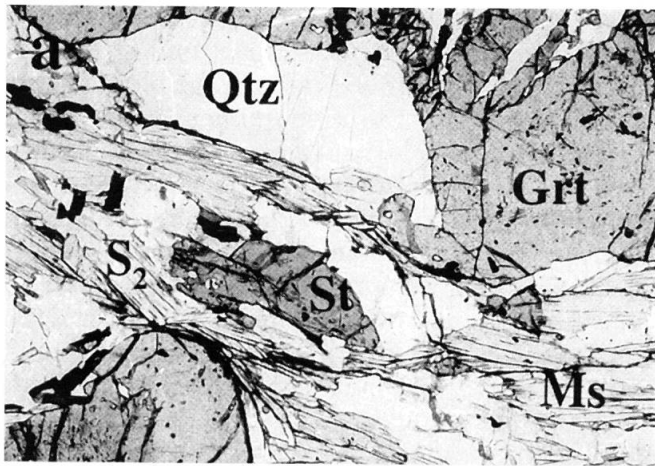
The third stage of metamorphic crystallization M<sub>2</sub> is related to the main schistosity S<sub>2</sub>. S<sub>2</sub> is marked by biotite and muscovite (Fig. 6d) and locally by kyanite crystals several centimeters long (Fig. 6g) that define the mineral stretching lineation L<sub>2</sub>. Chlorite is absent on S<sub>2</sub>. Garnet poikiloblasts with inclusions of S<sub>2</sub> marked by biotite, muscovite, quartz and plagioclase indicate syntectonic growth with respect to S<sub>2</sub> (Fig. 6h). Staurolite is rare in the studied area. It was found as small relics on S<sub>2</sub> (Fig. 7a) and as inclusions in D<sub>2</sub>-syntectonic garnet. Associated with M<sub>2</sub>, small quantities of leucosomes formed, that consist of quartz, plagioclase and rare K-feldspar (Fig. 7b). As these melts rarely migrated more than half a meter, they are considered to represent in-situ melting of the metapelites. The M<sub>2</sub> assemblage plagioclase (X<sub>An</sub> ≈ 20) + quartz + biotite + muscovite + garnet + kyanite + melt ± staurolite indicates metamorphic conditions within the kyanite zone for M<sub>2</sub>, its stability being limited by the reactions (6), (8) and (9) as well as by the kyanite = sillimanite equilibria (12) in figure 5, defining a P-T range of T ≈ 640–720 °C and P ≈ 600–1100 MPa. The widespread presence of small quantities of melt implies a temperature above the vapor-saturated melting reaction (7), that is, above T ≈ 650 °C.

During the fourth stage M<sub>3</sub>, minor mineral growth is documented. Biotite and muscovite on S<sub>2</sub> were crenulated during D<sub>3</sub> and locally they form a schistosity S<sub>3</sub>. Occasionally, small kyanite

← Fig. 6 Textural and structural relationships. The scale given within brackets corresponds to the long dimension of the image, SW is to the left hand side. All photomicrographs are from Phe Formation metapelites.

(a) Slate from the lower chlorite zone. S<sub>2</sub> is defined by muscovite intergrown with chlorite and S<sub>3</sub> is defined by fine-grained muscovite [1.5 mm]. (b) Phyllite from the upper chlorite zone showing a F<sub>2</sub> fold. S<sub>2</sub> and S<sub>3</sub> are defined by muscovite and chlorite [5 mm]. (c) Schist from the garnet zone, showing the schistosities S<sub>1</sub> and S<sub>2</sub> in the competent layer on the right hand side and S<sub>2</sub> and S<sub>3</sub> in the incompetent layer on the left hand side [9.5 mm]. (d) Crenulation cleavage S<sub>2</sub>, marked by biotite and muscovite, overprinting S<sub>1</sub> that is preserved in microlithons [13.5 mm]. (e) Kyanite surrounded by aggregates of fibrolitic sillimanite [1.5 mm]. (f) Fibrolitic sillimanite on S<sub>1</sub>, folded by F<sub>2</sub> and F<sub>3</sub>, that form a convergent-divergent interference pattern [18.5 mm]. (g) Kyanite on S<sub>2</sub>, broken by F<sub>3</sub> folding. Kyanite contains small garnet inclusions that may represent relics of M<sub>1</sub> [18 mm]. (h) D<sub>2</sub>-syntectonic garnet, containing S<sub>2</sub> [16 mm]. (d)–(h) are schists and gneisses from the kyanite zone.





crystals grew parallel to the axial surface of  $F_3$  folds (Fig. 7c). Broken kyanites on  $S_2$  are a common feature in  $F_3$  fold hinges (Fig. 6g). Leucosomes that are oriented parallel to  $F_3$  axial surfaces are interpreted to represent in-situ melting during  $M_3$  (Fig. 7d). Although little mineral growth is documented during  $M_3$ , the presence of kyanite on  $S_3$  and migmatization indicate conditions in the same range as for  $M_2$ .

During both  $M_2$  and  $M_3$ , small quantities of leucosomes formed, that are interpreted to be the result of vapor-saturated melting. This indicates that during  $M_2$ , not all vapor was consumed. This feature is noteworthy and may indicate that  $M_3$  represents the direct continuation of  $M_2$  metamorphic conditions during  $D_3$ .

The fifth stage of metamorphic crystallization  $M_4$  is related to  $D_4$  transtension in the Tos Dextral Transtension Zone (Figs 1, 2 and plate 1). It is characterized by fibrolitic sillimanite intergrown with biotite on  $D_4$  shear bands (Fig. 3, inset a). Kyanite is crosscut by sillimanite and partly resorbed, indicating that sillimanite was the stable aluminosilicate during  $M_4$ .  $D_2$  and  $D_3$  structures are crosscut by leucosomes representing migrated melts (Fig. 7e). This indicates local, but intensive melt formation after  $D_3$  and therefore most probably during  $M_4$ . Consequently, the vapor-absent melting reaction (10) was crossed during  $M_4$ , which restricts the P-T range to above  $T \approx 720^\circ\text{C}$  and below  $P \approx 820\text{ MPa}$ . This implies that the staurolite-consuming reaction (8) was also crossed during  $M_4$ , which is consistent with the rare occurrence of staurolite in the form of strongly resorbed crystals.

### 3.1.5. Zone of retrograde greenschist facies metamorphism

In a zone 2–2.5 km wide at the base of the High Himalayan Crystalline, the kyanite zone mineral assemblage in the Phe Formation pelites was retrogressively overprinted by a greenschist facies metamorphism related to  $D_4$  thrusting with a top-to-the SW shear sense along the Main Central Thrust (Fig. 3).

In general, the rocks are more fine-grained, K-feldspar is lacking, muscovite is more abundant and kyanite was transformed to muscovite during retrogression (Fig. 3, inset b).  $D_4$  thrusting is concentrated in shear zones, the frequency and intensity of which is strongest at the base of the High Himalayan Crystalline. These shear zones are characterized by C- and C'-type shear bands that overprint  $S_2$  and that are marked by biotite and chlorite, defining the schistosity  $S_4$  (Fig. 3, inset c). In  $D_4$  shear zones, chlorite replaced biotite on  $S_2$  and garnet rims were transformed to chlorite (Fig. 3, inset c).

### 3.2. LESSER HIMALAYAN SEQUENCE

In the Lesser Himalayan Sequence,  $D_4$  thrusting is concentrated in a zone about 2 km wide in the calcschists and phyllites of the mélangé zone below the base of the Crystalline Nappe and in thin layers of schists and phyllites within the underlying massive Berinag Quartzite (Fig. 3). The mineral content of these rocks is variable, the calcschists consisting of calcite, dolomite, quartz, muscovite, chlorite, biotite, graphite, opaque minerals, detrital sericitized plagioclase and K-feldspar (Fig. 7f). Detrital tourmaline and zircon are accessory minerals. The schists and phyllites are made up of muscovite, chlorite, biotite, graphite and ribbons of equigranular-polygonal quartz.  $D_4$  thrusting is indicated by C- and C'-type shear bands marked by muscovite, biotite and chlorite, forming  $S_4$  that overprints all earlier structures. Locally,  $S_4$  is very strong and represents the main schistosity (Fig. 7f). The assemblage chlorite + biotite in the schists and phyllites is defined by reaction (1) in the KFMASH system, the location of which is most likely between  $T \approx 300$  and  $400^\circ\text{C}$  (Fig. 5).

### 4. Thermobarometry

Despite the complex tectonometamorphic evolution of the studied section, the major metamorphic minerals observed can either be assigned to the stage of metamorphic crystallization  $M_2$  in the

← Fig. 7 Textural and structural relationships. The scale given within brackets corresponds to the long dimension of the image, SW is to the left hand side. (a)–(e) and (g)–(i) are pelitic Phe Formation schists and gneisses. (a)–(e) and (h), (i) are from the kyanite zone. (a) Relic of staurolite on  $S_2$  [8 mm]. (b)  $LS_2$  leucosomes oriented parallel to  $S_2$  that formed during  $F_2$  folding and  $M_2$  metamorphism.  $S_2$  and  $LS_2$  were subsequently folded by  $F_3$  folds. The leucosomes folded by  $F_2$  folds ( $LS_1$ ) most probably formed during  $M_1$  [40 cm]. (c)  $S_2$  schistosity folded by  $F_3$  folds. Kyanite occurs on  $S_2$  and parallel to the axial surface of  $F_3$  [15 mm]. (d)  $LS_3$  Leucosomes oriented parallel to the axial surface of  $F_3$  folds, related to  $D_3/M_3$  [55 cm]. (e)  $LS_4$  leucosomes crosscutting  $F_3$  folds, related to  $D_4/M_4$  [65 cm]. (f) Calcschist from the mélangé zone on top of the Berinag Quartzites;  $S_4$  is the main schistosity [9 mm]. (g) Euhedral garnet from a garnet zone schist [1 mm]. (h) Euhedral garnets from a kyanite zone gneiss [3 mm]. (i) Resorbed garnets from a kyanite zone gneiss, crosscut by sillimanite on a  $SB_4$  shear band [5.5 mm].

garnet and kyanite zones, to  $M_3$  in the chlorite and biotite zones or to  $M_4$  in the Main Central Thrust Zone. In the kyanite zone this is documented by

$D_2$ -syntectonic garnet growth and by biotite, muscovite and kyanite defining  $S_2$ . In these rocks textural relationships suggest equilibrium between

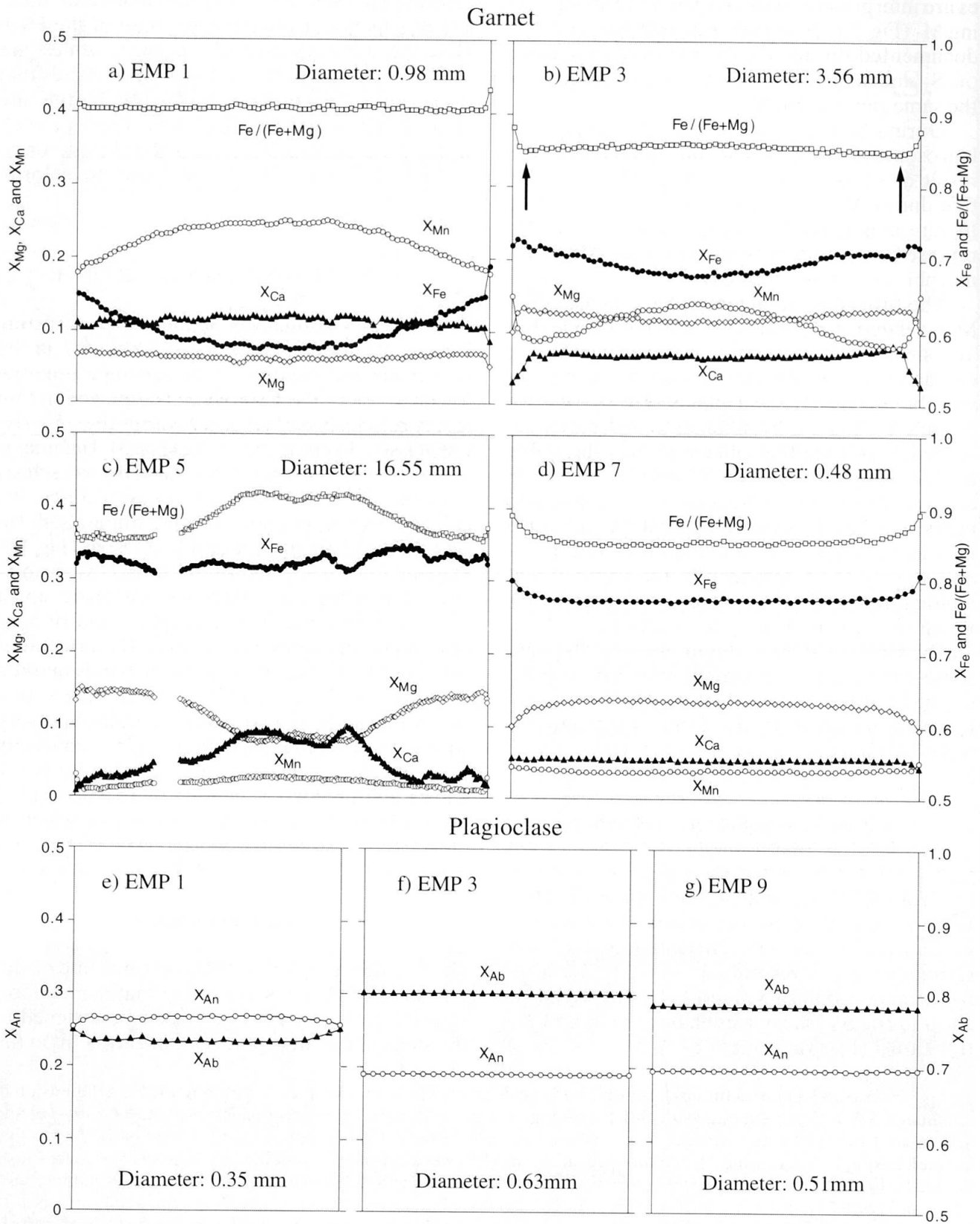


Fig. 8 Representative zoning profiles in garnet and plagioclase from metapelites in the Spiti valley–eastern Lahul-Parvati valley transect. EMP 1 is from the garnet zone, EMP 3, 5, 7 and 9 are from the kyanite zone. Arrows in (b) indicate the  $Fe/(Fe + Mg)$  minimum. See figure 9 for sample locations.

the major phases garnet, biotite, muscovite, plagioclase, quartz and kyanite during  $M_2$ . This allows the application of garnet–biotite (GARB) thermometry as well as garnet–aluminosilicate–quartz–plagioclase (GASP) and garnet–biotite–muscovite–plagioclase (GMAP) barometry to quantify the equilibrium P–T conditions during  $M_2$  throughout the High Himalayan Crystalline. Analytical procedure and mineral compositions are given in the Appendix.

#### 4.1. MINERAL ZONING PROFILES

To characterize the zoning in garnet and plagioclase, the two most typical minerals for fractional crystallization during progressive metamorphism, compositional traverses have been measured in all samples analyzed for thermobarometry.

##### 4.1.1. Garnet zoning

###### *Growth zoning*

Garnets from the garnet zone show slightly bell-shaped compositional profiles characterized by a core to rim  $X_{Fe}$  increase, revealing a typical growth zoning (Fig. 8a). Such a growth zoning is interpreted to represent a sequence of equilibrium compositions adapted to increasing P–T conditions during garnet growth. A similar zoning is shown by garnets of > 1 mm diameter from the kyanite zone (Fig. 8b). With increasing diameter, zoning in these garnets becomes stronger, the most extreme example occurring in an extraordinarily large garnet of 16.5 mm diameter (Fig. 8c). This large garnet shows a strong bell-shaped zoning with a core to rim  $X_{Fe}$  and  $X_{Mg}$  increase and parallel  $X_{Ca}$  and  $Fe/(Fe + Mg)$  ratio decrease. The abrupt change in composition after 1/3 of the radius may reflect the effect of the garnet-consuming NCKFMASH reaction garnet + chlorite + muscovite = staurolite + biotite + plagioclase + quartz +  $H_2O$ . Small size garnets (< 1 mm diameter) from the kyanite zone, by contrast, show rather homogeneous compositional profiles (Fig. 8d). The change in the strength of zoning from small size garnets to large size garnets in the kyanite zone is interpreted to indicate a homogenization of the garnet composition through diffusion at high temperatures that progressed from rim to core. As a consequence, small garnets were entirely homogenized whereas the core composition of large garnets was only slightly modified or even entirely preserved.

###### *Retrograde zoning*

Retrograde zoning in garnet is generally considered as the result of either a Fe–Mg exchange reaction involving garnet and biotite or both a Fe–Mg exchange reaction and a net transfer reaction. Exchange reactions cause only a change in the composition of minerals and do not affect the modal proportions of the phases involved. As a consequence, garnet crystals maintain their euhedral shape. In high-grade metamorphic pelites in the KFMASH system, garnet + K-feldspar +  $H_2O$  = sillimanite + biotite + quartz is the possible net transfer reaction (SPEAR, 1993). This reaction consumes garnet and accordingly, garnet crystals should show resorbed rims.

Throughout the garnet and kyanite zones, the garnets systematically show an increase in  $Fe/(Fe + Mg)$  ratio towards the rims, indicating Fe–Mg exchange between garnet and biotite during cooling. This increase is strongest where garnet is in direct contact with biotite. From the garnet zone to the kyanite zone, the penetration of Mg and Fe diffusion increases as a function of the increasing equilibration temperature (Fig. 8a vs Figs 8b–d). Garnets from the garnet zone always show euhedral grain boundaries (Fig. 7g), whereas in the kyanite zone, both euhedral garnets and garnets with resorbed rims occur (Figs 7h, i), indicating that in the kyanite zone the above net transfer reaction may be of a certain importance during retrogression. In fact, most of the resorbed garnets occur in the Tos Dextral Transtension Zone, where  $D_4$  sillimanite- and biotite-bearing extensional shear bands occur and where the rocks are nearly K-feldspar-free.

##### 4.1.2. Plagioclase zoning

Compositional profiles of plagioclase show no zoning except for garnet zone plagioclases (Figs 8e–g). Cation diffusion in plagioclase is very slow (GROVE et al., 1984) and once plagioclase is produced, it cannot easily change composition during retrograde equilibration. Consequently, these homogeneous profiles indicate that plagioclase grew with a stable composition in the kyanite zone.

#### 4.2. ESTIMATES OF APPARENT PEAK P–T CONDITIONS

Both net transfer and exchange reactions during cooling have a strong influence on the composition of garnet and biotite crystals, specifically on the  $Fe/(Fe + Mg)$  ratio. Consequently, as the GARB thermometer is particularly sensitive to

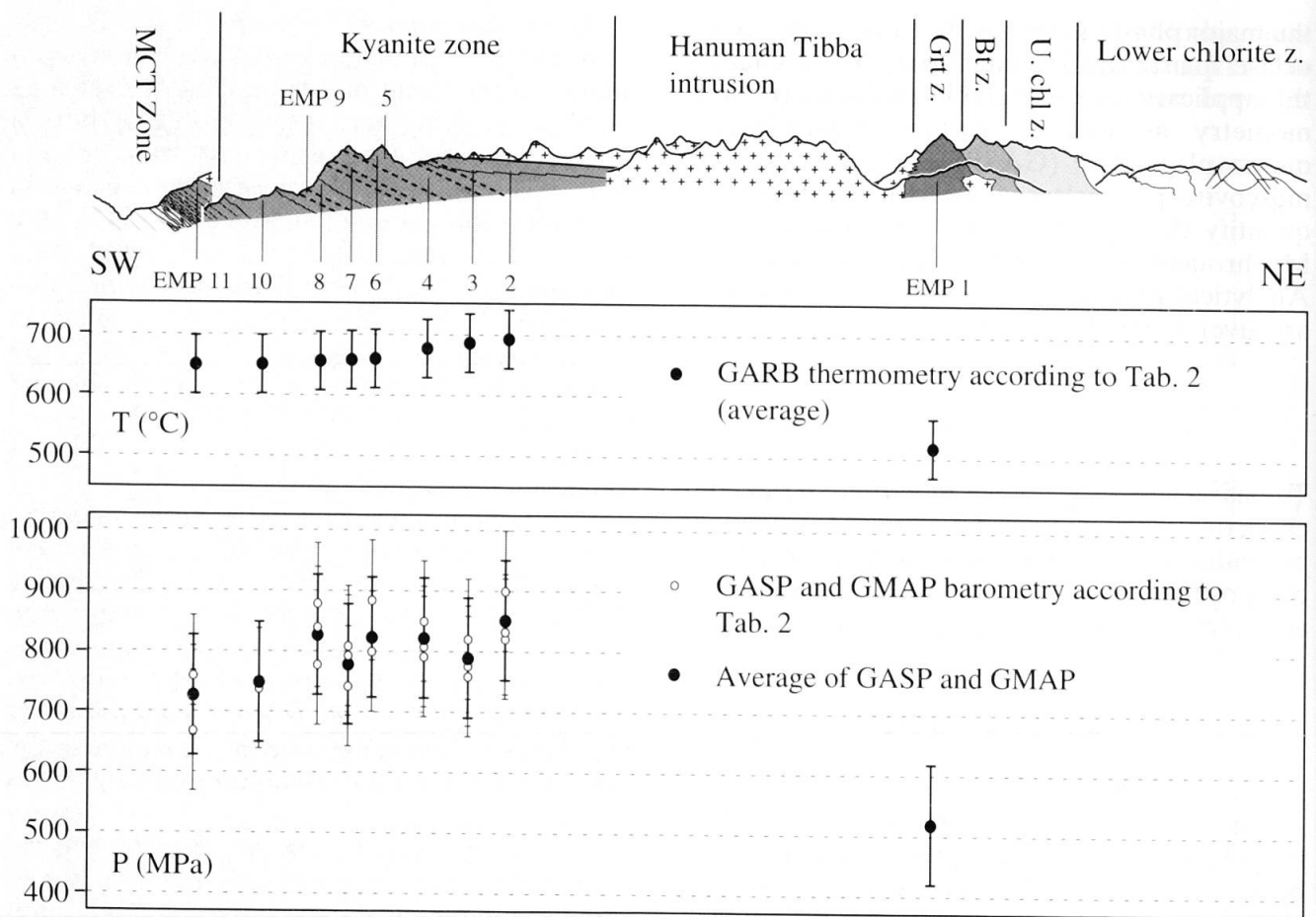


Fig. 9 Pressure and temperature profiles for the Spiti valley–eastern Lahul-Parvati valley transect, based on the results presented in table 2. The section is a reduction in size of plate 1.

the Fe/(Fe + Mg) ratio, temperature estimates based on GARB thermometry do not represent peak temperatures and are therefore referred to as apparent temperatures. In the case of exchange reactions only, the estimated temperature will be lower than the peak temperature and in the case where both exchange and net transfer reactions

can operate, thermometry will yield a temperature that is higher than the peak temperature (SPEAR, 1991; SPEAR and FLORENCE, 1992; SPEAR, 1993).

To minimize the influence of retrograde reactions for the estimation of peak temperatures, garnet compositions with the smallest possible Fe/

Tab. 2 Summary of thermobarometry results. EMP 1 is from the garnet zone, EMP 2 to EMP 10 are from the kyanite zone and EMP 11 is from the zone of retrograde greenschist facies metamorphism (Main Central Thrust Zone). See figure 9 for sample locations.

Sample	Apparent peak P–T estimates						Rim T estimates
	GARB (HODGES and SPEAR, 1982) GASP (HODGES and SPEAR, 1982)		GARB (HODGES and SPEAR, 1982) GASP (HODGES and CROWLEY, 1985)		GARB (HODGES and SPEAR, 1982) GMAP (HODGES and CROWLEY, 1985)		GARB (HODGES and SPEAR, 1982)
	T (°C) ± 50	P (MPa) ± 100	T (°C) ± 50	P (MPa) ± 100	T (°C) ± 50	P (MPa) ± 100	T (°C) ± 50
EMP 1	–	–	–	–	515	520	505
EMP 2	690	830	695	900	690	825	520
EMP 3	685	775	690	820	685	760	520
EMP 4	675	790	680	850	680	810	560
EMP 6	660	800	665	880	660	800	540
EMP 7	660	740	665	810	665	790	550
EMP 8	730	780	635	880	635	840	540
EMP 9	–	–	–	–	–	–	535
EMP 10	650	750	650	750	650	740	520
EMP 11	650	760	650	760	645	670	530

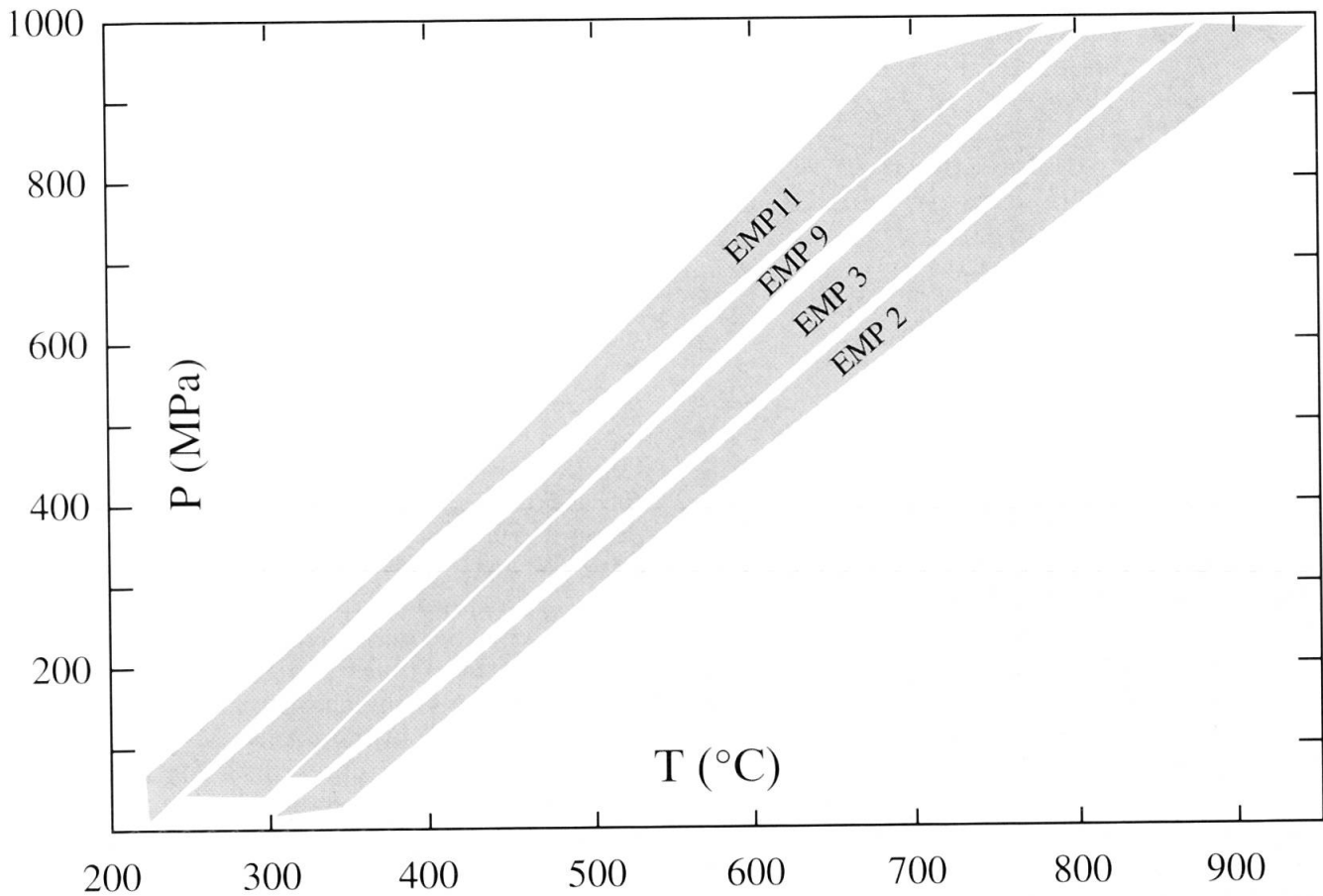


Fig. 10 Summary of GASP and GMAP reaction curves for rim barometry of four selected samples according to the calibrations of HODGES and SPEAR (1982) and HODGES and CROWLEY (1985), representing a semiquantitative approach to present the upsection decrease of pressure in the High Himalayan Crystalline during  $M_4$  from sample EMP 11 to sample EMP 2. See figure 9 for sample locations.

(Fe + Mg) ratio near the rim were used for thermobarometry (indicated by arrows in figure. 8b). These compositions are suggested to be the closest possible to the compositions of the peak of metamorphism, as they were apparently not influenced by retrograde reactions yet.

Retrograde Fe–Mg exchange reduces and net transfer reactions enlarge the Fe/(Fe + Mg) ratio in biotite. However, as diffusion in biotite is rapid relative to cooling rates, the interior of the biotite crystals quickly equilibrate with the rim and biotite crystals remain homogeneous. As a consequence, it is not possible to control the effect of retrograde reactions on biotite. In all samples used, the quantity of garnet is very small with respect to biotite. Garnet therefore represents a small reservoir for cation exchange with biotite and accordingly, the effect of cation exchange between these two phases can be considered as small for biotite. Consequently, it is suggested that the influence of both types of retrograde reactions on biotite is negligible in the analyzed samples, and that the biotite compositions are close to peak compositions.

GASP and GMAP barometers are particularly sensitive to the Ca-content of garnet and plagioclase. The homogeneous compositional profiles of kyanite zone plagioclases indicate stable plagioclase composition during prograde growth. Consequently, in this special case, the plagioclase composition is treated as if it had been in equilibrium with every point in garnet and it is used to estimate the pressure at the Fe/(Fe + Mg) minimum. Ca diffusion in garnet is slow during retrograde equilibration with respect to Fe and Mg diffusion. Accordingly, the Ca-zoning is considered to represent the Ca-content during prograde garnet growth. During prograde growth on a “nappe formation” P–T path,  $X_{Ca}$  remains relatively stable and decreases during decompression only (SPEAR, 1993). In all garnet compositional profiles used for thermobarometry,  $X_{Ca}$  starts to decrease at a point that coincides with the Fe/(Fe + Mg) minimum. Therefore,  $X_{Ca}$  at the Fe/(Fe + Mg) minimum is suggested to represent the Ca-content near the peak of metamorphism, indicating that the estimated pressures may be close to peak pressures.

Apparent peak P-T estimates are summarized in table 2 and presented in figure 9 as a function of their structural position in the section. P-T results for aluminosilicate-bearing rocks obtained from two independent sets of thermobarometer (GAR-B-GASP and GAR-B-GMAP) and with two different calibrations for GAR-B-GASP (HODGES and SPEAR, 1982; HODGES and CROWLEY, 1985) generally overlap within error, suggesting that the analyzed mineral assemblages correspond to well-equilibrated systems.

Apparent peak P-T estimates from the garnet zone indicating a temperature of 515 °C and a pressure of 520 MPa fit the stability conditions deduced from the petrogenetic grid (I in figure 5). In the kyanite and MCT zones peak P-T conditions decrease slightly from the NE to the SW from  $T \approx 690$  °C and  $P \approx 850$  MPa to  $T \approx 650$  °C and  $P \approx 730$  MPa (II and III in figure 5). Samples that most probably underwent retrograde Fe-Mg exchange only and samples that underwent both Fe-Mg exchange and net transfer reaction fit within the same trend of apparent temperatures. This may indicate that the influence of the garnet-consuming net transfer reaction was small in the analyzed samples and, as a consequence, that the presented temperature estimates represent minimum values of true peak temperatures.

#### 4.3. RIM P-T ESTIMATES

Rim P-T conditions have been estimated using garnet compositions from the outermost rims that are in contact with biotite. As all samples analyzed underwent at least retrograde Fe-Mg exchange, the estimated temperatures are expected to be higher than  $T \approx 500$  °C, that is the temperature at which Fe-Mg diffusion in garnet ceases to penetrate measurably (SPEAR and FLORENCE, 1992), and lower than the true metamorphic peak temperature (Tab. 2). As it is not known to what extent the analyzed samples were influenced by retrograde reactions, the presented rim temperatures may represent any mixture between the peak temperature and  $T \approx 500$  °C.

Barometry is less sensitive to the Fe/(Fe + Mg) ratio and therefore the results of rim barometry may represent true pressures attained during cooling. However, as the rim temperature estimates do not yield values that can be used to fix pressures on the GASP and GMAP reaction curves, a semiquantitative approach is selected to present the evolution of pressure. A comparison of the results of rim GASP and GMAP barometry for several High Himalayan Crystalline samples shows a decrease of the rim pressure towards the

Tos Dextral Transtension Zone (Fig. 10). These results fit well with the structural model of the Crystalline Nappe that implies decompression by NE-directed normal movement in the Tos Dextral Transtension Zone (Figs 2, 3).

## 5. P-T Path

### 5.1. HIGH HIMALAYAN CRYSTALLINE

No tectonic discontinuities have been observed between the Barrovian mineral zones within the Tethyan Himalaya and the High Himalayan Crystalline. The mineral assemblages characterizing the various zones are therefore considered to be related to a continuous metamorphic evolution during  $M_2$ ,  $M_3$  and  $M_4$ .

Structural and microtextural observations imply that  $M_2$  is the main metamorphism in the garnet and kyanite zones and that  $M_3$  is the main metamorphism in the chlorite and biotite zones. In the garnet and kyanite zones  $M_3$  played a less important part, as it represents a continuation of  $M_2$  conditions, whereas in the chlorite and biotite zones,  $M_2$  is only of minor importance. The chlorite and biotite zones are therefore mainly defined by  $M_3$  mineral assemblages, whereas the garnet and kyanite zones are defined by  $M_2$  assemblages including  $M_3$  assemblages.

Petrographic features and P-T estimates allow constraints to be placed on the metamorphic evolution for garnet- and kyanite zone samples. On this basis, the P-T path during  $M_1$  can be roughly constrained and a continuous P-T path from  $M_2$  to  $M_4$  can be drawn for the High Himalayan Crystalline.

The oldest observed metamorphic event  $M_1$  is characterized by relics of kyanite ( $M_{1a}$ ) in the center of aggregates of fibrolitic sillimanite that occur together with the mineral assemblage biotite + muscovite  $\pm$  K-feldspar ( $M_{1b}$ ). Most probably, a small quantity of melt was generated during  $M_1$  as indicated by leucosomes deformed by  $F_2$  folds (Fig. 7b). This indicates that during  $M_1$ , the kyanite = sillimanite reaction (12 in figure 5) was crossed close to the vapor-absent melting reactions (10 and 11). Arrow A in figure 5 indicates the transition from stage  $M_{1a}$  to stage  $M_{1b}$ . The exact position and  $dP/dT$  slope of arrow A, however, cannot be determined.

The stability conditions deduced from the petrogenetic grid for the  $M_2$  garnet zone mineral assemblage are limited by the reactions (2) and (4) in figure 5. They coincide with the P-T estimates of garnet zone sample EMS 1, which represents the starting point of the  $M_2$  to  $M_4$  P-T path

(I in figure 5). In the  $M_2$  kyanite zone mineral assemblage chlorite is completely resorbed and small quantities of migmatites occur, representing in-situ melting related to  $M_2$ . Together with a systematic presence of plagioclase in kyanite-bearing assemblages, this implies a prograde path below reaction (9) and crossing reaction (6) and (7) in the kyanite stability field (Fig. 5, arrow B1). This fits well with the maximum P-T estimates for the samples EMP 2 and EMP 8 (II in figure 5). It is suggested that the  $M_2$  peak metamorphic conditions did not cross the vapor-absent melting reaction (10), because only small amounts of melt formed during  $M_2$ . The  $M_3$  mineral assemblage indicates that the metamorphic crystallization during  $M_3$  was a direct continuation of  $M_2$  under similar conditions but with minor mineral growth.

In the Tos Dextral Transtension Zone  $M_2$  kyanite crystals were cut by  $D_4$  shear bands marked by  $M_4$  sillimanite and biotite. Migmatites representing migrated melts crosscut  $D_2$  and  $D_3$  structures, suggesting local, but intensive melt formation posterior to  $D_3$  and therefore most probably during  $M_4$ . Staurolite occurs as rare relics only, and garnet recorded decompression after passing the pressure peak of metamorphism as indicated by Ca zoning and by rim barometry, implying that garnet continued to crystallize, whereas staurolite was consumed. These features suggest a decompressive path crossing the staurolite-consuming reaction (8) and the vapor-absent melting reaction (10) close to the  $Ky = Sil$  equilibria (12) during  $D_4$  extension (Fig. 5, arrow B2). Crossing of reaction (10) implies the formation of K-feldspar. This mineral, however, is rare in the kyanite zone pelites and, if it occurs, it is strongly resorbed. This could indicate that K-feldspar was consumed by the retrograde reaction  $garnet + K\text{-feldspar} + H_2O = sillimanite + biotite + quartz$  that is suggested to have operated during retrogression. Decompression after the peak of metamorphism increases in the Tos Dextral Transtension Zone from the SW to the NE (Fig. 10). The maximum amount of decompression, however, cannot be estimated as no quantitative constraints were obtained from the analyzed samples.

Towards the base of the High Himalayan Crystalline, in the zone of retrograde greenschist facies metamorphism, the  $M_2$  stability conditions are given by the presence of garnet, kyanite and small quantities of  $M_2$  leucosomes, indicating that reaction (7) was crossed above reaction (12) during  $M_2$  (Fig. 5, arrow C1). The  $M_2$  peak P-T estimates are slightly lower than in the center of the nappe (III in figure 5). Sillimanite, K-feldspar and leucosomes representing migrated  $M_4$  melts do not occur, implying that the reactions (10) and

(12) were not crossed after the peak of metamorphism in this part of the section. The widespread presence of biotite and chlorite on shear bands related to  $D_4$  SW-directed thrusting as well as chlorite overgrowing garnet rims and muscovite replacing kyanite in the zone of retrograde greenschist facies metamorphism indicate that the metamorphic conditions decreased to biotite zone conditions during  $D_4$  thrusting of the High Himalayan Crystalline over the Lesser Himalayan Sequence (Fig. 5, arrow C2).

## 5.2. LESSER HIMALAYAN SEQUENCE

During  $D_4$  thrusting the calcschists and phyllites of the mélangé zone and thin layers of schists and phyllites within the massive Berinag Quartzite reached  $M_4$  biotite zone metamorphic conditions similar to  $M_4$  conditions at the base of the overlying Crystalline Nappe. It cannot be excluded that the Lesser Himalayan Sequence rocks were already metamorphic prior to  $M_4$ . The conditions of this metamorphism, however, did not exceed  $M_4$  conditions. Consequently, the Lesser Himalayan Sequence rocks followed a prograde P-T path during  $D_4$  thrusting in the Main Central Thrust Zone (Fig. 5, arrow D). Because the P-T conditions of the greenschist facies metamorphism in the Main Central Thrust Zone cannot be determined precisely, the arrows C2 and D are poorly constrained and should be considered as purely qualitative.

## 6. Discussion

### 6.1. METAMORPHISM $M_1$

Well preserved relics of a strong  $S_1$  schistosity and rare isoclinal  $F_1$  folds within the High Himalayan Crystalline in the garnet- and kyanite zones indicate that  $D_1$  was an important phase of deformation. However, due to the subsequent  $D_2$ ,  $D_3$  and  $D_4$  overprint, no information about the vergence and the orientation of fold axes and axial surfaces of  $F_1$  folding is preserved and therefore,  $D_1$  deformational conditions are poorly known.

The most important information about  $D_1$  comes from the  $M_{1a/b}$  metamorphic mineral assemblages in the Phe Formation metapelites (Plate 1), indicating kyanite zone conditions in the Tos valley area during  $M_{1a}$ . This is in agreement with metamorphic conditions deduced from  $M_1$  corona textures overprinting the magmatic mineral assemblages in rare olivine gabbro bodies that occur in the lowermost kyanite zone Phe For-



mation rocks in the Tos valley (WYSS and HERMANN, *subm.*). For these orthopyroxene + clinopyroxene + garnet + amphibolite + spinel coronas, peak conditions of  $T \approx 700$  °C and  $P \approx 800$ – $1000$  MPa have been determined. During  $M_{1b}$ , sillimanite formed on  $S_1$ , indicating a change of the metamorphic conditions to the sillimanite zone. As the subsequent  $M_2$  metamorphism was prograde, reaching kyanite zone peak conditions in the Tos valley,  $M_{1a}$ – $M_{1b}$  are interpreted to represent an independent metamorphic cycle. Therefore,  $D_1$  was an independent tectonometamorphic event that occurred prior to the NE-verging  $D_2$  stacking of the Shikar Beh Nappe (WYSS *et al.*, 1999). The Shikar Beh Nappe has for a long time been considered to be the first record of tectonic activity in the Himachal Himalaya (STECK *et al.*, 1993, 1999; VANNAY, 1993; VANNAY and STECK, 1995).

Several authors found indications for pre-Himalayan orogeny. POGANANTE and LOMBARDO (1989) reported metabasites with relics of a granulite facies metamorphism from Zaskar and Lahul, and WALKER *et al.* (1999) found relics of an old metamorphism in the Zaskar area that are interpreted to be of pre-Himalayan age. ARGLES *et al.* (1999) discovered garnets with Sm–Nd ages  $>500$  Ma in the Garwhal Himalaya and outside the NW Himalaya, FERRARA (1983) reported a  $449 \pm 56$  Ma Rb/Sr age on paragneisses from the Mt. Everest area. The predominating arguments for a pre-Himalayan event, however, are based on the sedimentary record showing several angular unconformities and sedimentation gaps. A prominent angular unconformity between the Lower Cambrian Karsha Formation and the Middle Ordovician Thaple Formation (Tab. 1) is interpreted as an exposure to erosion and continental clastic resedimentation due to the uplift of a Cambro-Ordovician mountain belt that may be related to Pan-African orogeny (GARZANTI *et al.*, 1986; GAETANI and GARZANTI, 1991; VALDIYA, 1995). Some authors invoke that the concurrence of Cambro-Ordovician magmatic activity and exposure to erosion is the strongest argument for pre-Himalayan orogeny (SAXENA, 1980; GARZANTI *et al.*, 1986). PANDE and SAXENA (1968), SAXENA (1973) and FUCHS (1992) interpreted sedimentary gaps as the evidence for Caledonian and Hercynian orogeny. JAIN *et al.* (1980) remain more general and invoke late Paleozoic epeirogenic movements.

If these unconformities are interpreted to represent erosion of an orogenic belt, it is expected that the structural and metamorphic record below and above the unconformities should indicate pre-erosional deformation and/or metamorphism. In other words, in such a case below the

unconformity, at least one phase of deformation and/or one stage of metamorphic crystallization more should be observed than above the unconformity. Such a feature, however, has not been described until now. It is therefore suggested that these arguments should be treated with caution.

All these observations and their interpretations show that the existence of a pre-Himalayan orogeny is subject to investigation for a long time, but there is little consensus as to what the age of such an event could be. In the Phe Formation metapelites,  $M_1$  overprints the hornfelsic contact aureole of the  $495 \pm 16$  Ma old Hanuman Tibba intrusion (Rb/Sr whole rock isochron, dated by FRANK *et al.*, 1977), indicating that  $M_1$  must be younger than this intrusion (WYSS *et al.*, 1999). As a consequence,  $D_1$  may be considered as a pre-Himalayan or as an early Himalayan event.

## 6.2. METAMORPHISM $M_2$

In the High Himalayan Crystalline, structural and textural observations allow the peak P-T conditions during Tertiary Himalayan nappe stacking to be assigned to the stage of metamorphic crystallization  $M_2$  that is related to formation of the NE-verging Shikar Beh Nappe. This feature has only been reported from the central and northern Himachal Himalaya until now (EPARD *et al.*, 1995, STECK *et al.*, 1999; WYSS *et al.*, 1999). In the other part of the Himalayan orogen, peak P-T conditions are reported to be related to SW-directed nappe stacking (e.g. VANNAY and GRASEMANN, 1998 and references therein).

During the Barrovian-type  $M_2$  metamorphism, the High Himalayan Crystalline rocks followed a prograde P-T path that reached different peak P-T conditions according to the tectonostratigraphic depth (Plate 1, Fig. 9). A prograde P-T path for  $M_2$  is also supported by prograde mineral zoning in  $M_2$  amphiboles from olivine gabbro bodies (WYSS and HERMANN, *subm.*). This indicates that  $M_2$  represents an independent metamorphic stage and cannot be considered as a retrograde stage of  $M_1$ . It is important to note that the peak P-T conditions decrease downsection from the NE to the SW in the kyanite zone (Figs 5, 9), indicating an inverted zonation of  $M_2$  isotherms and isobars. This feature will be discussed in the last section.

The NE-verging  $D_2$  Shikar Beh Nappe is documented by  $S_2$  and  $F_2$  folds in the garnet zone, in the kyanite zone and in the zone of retrograde greenschist facies metamorphism along the Main Central Thrust, but by relics of  $S_2$  solely in the biotite zone. In the chlorite zone  $D_2$  is not clearly

documented. In the garnet- and kyanite zones  $M_2$  is the main metamorphism, whereas in the biotite zone  $M_3$  is the main metamorphism. At the transition between the garnet- and biotite zones, the  $M_2$  metamorphic conditions decrease abruptly from garnet zone conditions to chlorite zone conditions (Plate 1). NE of this boundary the metamorphic zoning is the result of  $M_3$  and SW of this boundary it is the result of  $M_2$ ,  $M_3$  being of subordinate importance. The abrupt decrease in the intensity of  $M_2$  and  $S_2$  and the disappearance of  $F_2$  folds between the garnet- and biotite zones could be interpreted to represent the front of the NE-verging Shikar Beh Nappe. This assumption, however, requires the record of a large-scale thrust zone between the garnet zone and the biotite zone, a feature that is not documented in the field. STECK et al. (1998) interpreted a NE-directed thrust in the Lagudarsi La area NE of Spiti valley as a frontal thrust of the Shikar Beh Nappe. This may explain, why in the chlorite zone a schistosity  $S_2$  occurs prior to the main schistosity  $S_3$ . However, if the nappe front is assumed to be situated more than 20 km to the NE,  $F_2$  folds are not expected to disappear NE of the garnet zone to biotite zone boundary. Alternatively, the sudden decrease of  $M_2$  metamorphic conditions could reflect a constriction of  $M_2$  isogrades due to extensional normal shearing prior to the  $M_3$  metamorphism. Widespread normal shearing, however, is not observed between the phases  $D_2$  and  $D_3$ . Accordingly, based on the present knowledge, this situation is inconsistent and needs further investigation.

### 6.3. METAMORPHISM $M_3$

SW-verging  $F_3$  folding formed at the front of the north Himalayan nappe system and propagated towards SW (WYSS et al., 1999), affecting both the Tethyan Himalaya and the High Himalayan Crystalline. A common evolution of these two domains during  $D_3$  is also reflected by  $M_3$  metamorphic conditions that increase gradually from the NE to the SW from the lower chlorite zone in the Taktsi valley to the kyanite zone in the Tos valley (Plate 1). In the chlorite- and biotite zones,  $M_3$  is a prograde metamorphism that is related to folding and subduction below the front of the North Himalayan Nappe System (WYSS et al., 1999). In the garnet- and kyanite zones, the  $M_2$  mineral assemblage remained stable during  $M_3$ . Locally in these zones, a weakly developed  $M_3$  mineral assemblage crystallized, containing similar minerals as the  $M_2$  assemblage, indicating a continuous evolution from  $M_2$  to  $M_3$ . As a consequence, in the High Himalayan Crystalline,  $F_3$  folding occurred

under conditions that were close to  $M_2$  peak conditions without major cooling in the meantime and therefore, SW-verging  $F_3$  folding affected the structures of the  $D_2$  NE-verging Shikar Beh Nappe before exhumation started. According to EPARD et al. (1995), in the neighbouring Kullu valley area SW-verging folding was related to a slightly retrograde evolution compared with  $D_2$ .

### 6.4. METAMORPHISM $M_4$

During  $D_4$  extrusion of the Crystalline Nappe the base of the High Himalayan Crystalline underwent a different  $M_4$  metamorphic evolution than the central part (Figs 3 and 5, Plate 1). Related to SW-directed thrusting on  $D_4$  shear bands in the Main Central Thrust Zone, the base of the High Himalayan Crystalline retrogressed from  $M_2/M_3$  amphibolite facies conditions to greenschist facies conditions. The central part of the High Himalayan Crystalline by contrast underwent decompression under constant or slightly increasing temperatures within amphibolite facies conditions, coupled with NE-directed normal shearing and dextral strike-slip shearing within the Tos Dextral Transtension Zone. This is supported by rim P-T estimates indicating a pressure decrease from the SW to the NE in the High Himalayan Crystalline during  $M_4$  (Fig. 10) and temperatures above  $T = 500^\circ\text{C}$  (Tab. 2).

### 6.5. THE TETHYAN HIMALAYA – HIGH HIMALAYAN CRYSTALLINE TRANSITION

$D_4$  Normal- and dextral strike-slip shearing is generally reported to be restricted to large-scale shear zones between the Tethyan Himalaya and the High Himalayan Crystalline. In these shear zones, the pre- $D_4$  isogrades were constricted and therefore the low-grade metamorphic Tethyan Himalaya rocks appear to be juxtaposed directly against the high-grade metamorphic rocks of the High Himalayan Crystalline as reported, for example, from the Zaskar Shear Zone by DÉZES et al. (1999). On this basis, the Tethyan Himalaya and the High Himalayan Crystalline were considered as purely tectonic units. In the studied area by contrast,  $D_4$  normal and dextral strike-slip shearing are distributed tectonostratigraphically over about 20 km, including the Tos- and Taktsi Dextral Transtension Zones. The geometrical arrangement of the shear zones as indicated in figure 1 suggests that the Taktsi Dextral Transtension Zone could be the equivalent of the Zaskar Shear Zone in the study area. Normal movement

in the Taktsi Dextral Transtension Zone, however, is characterized by fold reorientation, which accommodated only a relatively small shear displacement compared to normal shearing in the Zaskar Shear Zone. Consequently, the  $M_3$  Barrovian-type metamorphic mineral zones remained undisturbed between the Tethyan Himalaya and the High Himalayan Crystalline, resulting in a gradual transition between these two domains. As a consequence, the Tethyan Himalaya and the High Himalayan Crystalline should be regarded as metamorphic zones rather than as tectonic units. It is proposed that the Tethyan Himalaya is restricted to the lower chlorite zone, whereas the biotite, garnet and kyanite zones make up the High Himalayan Crystalline. The upper chlorite zone represents the transition between these two domains. To compensate for a reduced shear displacement at the transition between the Tethyan Himalaya and the High Himalayan Crystalline,  $D_4$  shearing shifted deep inside the High Himalayan Crystalline, where it formed the ductile Tos Dextral Transtension Zone.

#### 6.6. INVERTED METAMORPHIC ZONATION

Two types of inverted metamorphic zonations occur in the High Himalayan Crystalline. The first type is a real inversion of metamorphic isograds by post-metamorphic deformation and the second one is the result of metamorphic overprint. The real inversion of metamorphic isograds is not recognizable in the field and was detected by thermobarometry. It is characterized by  $M_2$  peak P-T conditions that decrease downsection within the kyanite zone, indicating an inversion of  $M_2$  isotherms and isobars, a feature that cannot be explained by the geometry of the NE-directed  $D_2$  Shikar Beh Nappe. It is suggested that the  $M_2$  isograds were passively inverted during  $F_3$  folding and/or during  $D_4$  extrusion of the Crystalline Nappe, forming an open, SW-verging fold. A similar situation has already been encountered by EPARD et al. (1995) along the Mandi-Khoksar transect 50 km west of the studied section. On the basis of model calculations assuming a pre-Main Central Thrust metamorphism ( $M_2$  in this study) and varying thermal conductivities of the Tethyan Himalaya, the High Himalayan Crystalline and the Lesser Himalayan Sequence during thrusting in the neighbouring Kullu valley area, GRASEMANN (1993) proposed that inversion of isotherms may be possible in the overthrust Crystalline Nappe due to thrusting along the Main Central Thrust. These inverted isotherms are considered to be the result of a pre-thrust metamor-

phism ( $M_2$  in this study) and a syn-thrust metamorphism ( $M_3$  and/or  $M_4$  in this study) during a continuous metamorphic evolution. Applied to the studied area, GRASEMANN'S (1993) model suggests that the downsection decrease in peak metamorphic conditions within the Crystalline Nappe is the result of an inversion of  $M_3$  and/or  $M_4$  isotherms during  $D_3/D_4$  due to previous  $M_2$  thermal disturbances. This contrasts with the results presented in this study that indicate a passive deformation of  $M_2$  isotherms during  $D_3/D_4$ . Consequently, as it lays emphasis on the syn-Main Central Thrust metamorphic crystallization for the formation of inverted isotherms and as it assumes that the pre-Main Central Thrust metamorphism  $M_2$  is of subordinate importance in the High Himalayan Crystalline, GRASEMANN'S (1993) model cannot be adapted to the studied traverse.

The second type of zonation is a combination of  $M_4$  greenschist facies and sillimanite zone metamorphic overprint of the  $M_2/M_3$  kyanite zone mineral assemblage in the  $D_4$  Main Central Thrust Zone and in the Tos Dextral Transtension Zone, respectively.  $D_4$  thrusting along the Main Central Thrust Zone occurred under  $M_4$  greenschist facies conditions that were prograde in the underlying Lesser Himalayan Sequence and retrograde at the base of the Crystalline Nappe, where a mineral assemblage containing mainly chlorite, biotite and muscovite overprinted the  $M_2/M_3$  kyanite zone assemblage. This gives the impression of an increase upsection in metamorphic conditions from  $M_4$  greenschist facies rocks in the Lesser Himalayan Sequence and at the base of the Crystalline Nappe to overlying  $M_2/M_3$  kyanite zone rocks (Plate 1). Retrogression at the base of the Crystalline Nappe may be explained as an effect of cooling by heat flow into the cold overridden Lesser Himalayan Sequence, that was simultaneously heated up by the hot overriding Crystalline Nappe, according to the model proposed by LE FORT (1975). Changing metamorphic conditions due to decompression during  $D_4$  transtension in the Tos Dextral Transtension Zone lead to an overprint of the  $M_2/M_3$  kyanite-bearing mineral assemblage by  $M_4$  sillimanite in  $D_4$  shear zones, giving the impression of metamorphic conditions increasing upsection from the  $M_2/M_3$  kyanite zone to the  $M_4$  sillimanite zone (Plate 1). As the increase in metamorphic conditions upsection, from  $M_4$  greenschist facies conditions in the Main Central Thrust Zone over  $M_2/M_3$  kyanite zone conditions to  $M_4$  sillimanite zone conditions in the Tos Dextral Transtension Zone is the result of a polymetamorphic evolution, it has to be referred to as an apparent inverted metamorphic zonation.

## 7. Conclusions

Along the traverse from the Spiti valley through eastern Lahul to the Parvati valley, four stages of metamorphic crystallization  $M_1$  to  $M_4$  are distinguished, related to the four deformational phases  $D_1$  to  $D_4$ .  $M_1$  represents an independent metamorphic cycle that first reached kyanite and then sillimanite zone peak conditions in the High Himalayan Crystalline, indicating considerable crustal thickening, that is either of pre- or of early Himalayan age.

$M_2$  is the main Tertiary metamorphism in the High Himalayan Crystalline, related to the NE-verging  $D_2$  stacking of the Shikar Beh Nappe.  $M_2$  peak conditions increase from the top of the High Himalayan Crystalline downsection from  $T \approx 520$  °C and  $P \approx 500$  MPa to  $T \approx 700$  °C and  $P \approx 850$  MPa in the center of the High Himalayan Crystalline. From there on,  $M_2$  conditions decrease downsection towards the base of the High Himalayan Crystalline to  $T \approx 650$  °C and  $P \approx 720$  MPa. This inversion of  $M_2$  isotherms and isobars is suggested to result from passive folding of  $M_2$  isogrades during  $D_3$  and/or  $D_4$ , representing a real inverted metamorphic zonation.

$M_3$  is the main metamorphism in the Tethyan Himalaya, related to the SW-verging  $F_3$  folding. In the High Himalayan Crystalline by contrast,  $M_3$  represents the continuation of  $M_2$  conditions and involved only minor mineral growth.  $M_3$  metamorphic conditions increase gradually from the Tethyan Himalaya to the underlying High Himalayan Crystalline, indicating that these domains represent metamorphic zones rather than tectonic units in the studied area.

During  $D_4$  extrusion of the Crystalline Nappe,  $M_4$  conditions were related to different structures and reached different metamorphic conditions at the base and in the center of the High Himalayan Crystalline. At the base of the High Himalayan Crystalline  $M_4$  is characterized by retrograde greenschist facies conditions related to SW-verging thrusting along the Main Central Thrust, that overprinted the  $M_2/M_3$  kyanite zone mineral assemblage. Retrogression to greenschist facies conditions is interpreted as the result of cooling by heat flow into the cooler, overridden Lesser Himalayan Sequence, that simultaneously underwent prograde greenschist facies metamorphism. In a higher crustal level  $M_4$  reached sillimanite zone conditions related to  $D_4$  normal- and dextral strike-slip movement, overprinting the  $M_2/M_3$  kyanite zone mineral assemblage in the Tos Dextral Transtension Zone.  $D_4$  is associated with decompression that increases upsection from the base of the High Himalayan Crystalline to its center.

The  $M_4$  overprint on the  $M_2/M_3$  mineral assemblage leads to an increase in metamorphic conditions upsection from  $M_4$  greenschist facies conditions over  $M_2/M_3$  kyanite zone conditions to  $M_4$  sillimanite zone conditions in the High Himalayan Crystalline. This increase is the result of a polymetamorphic evolution and has therefore to be referred to as an apparent inverted metamorphic zonation.

This study demonstrates that a real and an apparent inverted metamorphic zonation occur within the same section in a relatively small area compared to the whole Himalayan chain. As a consequence, for further studies about this phenomenon, it will be important to distinguish between inverted metamorphic field gradients assigned to one stage of metamorphic crystallisation only and apparent inverted metamorphic zonations resulting from metamorphic overprint due to a polyphase metamorphic evolution. Given the significant geological variations observed all along the Himalayan orogen, it is expected that several models must be allowed to coexist to explain the different features integrated into what is referred to as the inverted metamorphic zonation.

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## Appendix

### THERMOBAROMETRY

In the garnet and kyanite zones of the studied section 36 samples containing relevant mineral phases were chosen for thermobarometry. On these samples a two-step analytical procedure was applied.

*Step 1:* To characterize the compositional zoning in garnet a rough garnet composition traverse was measured in all samples chosen for thermobarometry (Figs 8, A1). On the basis of these traverses 11 samples were chosen that meet to the following requirements: (1) they are representative for the garnet and kyanite zone assemblages and (2) their traverses show a compositional zoning.

*Step 2:* To estimate the apparent peak temperatures the zone of Fe/(Fe + Mg) minimum was determined for each garnet on the basis of the compositional traverse established in step 1 and for each garnet a series of 10 to 20 additional radial compositional traverses across this zone towards the rim was measured (Fig. A1). These traverses comprise the compositions at the Fe/(Fe + Mg) minima and the rim compositions, each of them contains 20 to 40 analyses.

Compositional traverses across plagioclase, biotite and muscovite were also measured. With the exception of plagioclase from the garnet zone, these minerals do not show considerable variations from core to rim. An average of 60 analyses were performed for each of these mineral phases per sample.

Mineral compositions were measured using a Cameca SX50 electron microprobe. The acceleration voltage was 15 kV for all phases and the nominal beam current was 25 nA for garnet, 15 nA for biotite and muscovite, and 10 nA for plagioclase.

Representative garnet compositions at the Fe/(Fe + Mg) minimum, garnet rim compositions and plagioclase, biotite and muscovite compositions are presented in tables A1 to A4. Pressures and temperatures were calculated with a computer program written by Frank S. Spear (GTB 68), using the thermobarometers and calibrations listed in table 1.

### BULK ROCK ANALYSES

The bulk chemical compositions were determined by X-ray fluorescence (XRF) analyses with a sequential spectrometer using natural USGS rock samples for calibration. Rocks were ground in a tungsten carbide mill. Major elements were determined using glass beads which were fused from ignited rocks powders (at 1050 °C) mixed with  $\text{Li}_2\text{B}_4\text{O}_7$  in a 1/5 ratio in gold platinum pans at 1150 °C. The intensities were corrected for instrumental drift, background and matrix effects. The trace elements were determined by XRF analyses of 10g rock powder samples using the synthetic background method for which major elements have to be known (NISBET et al., 1979).

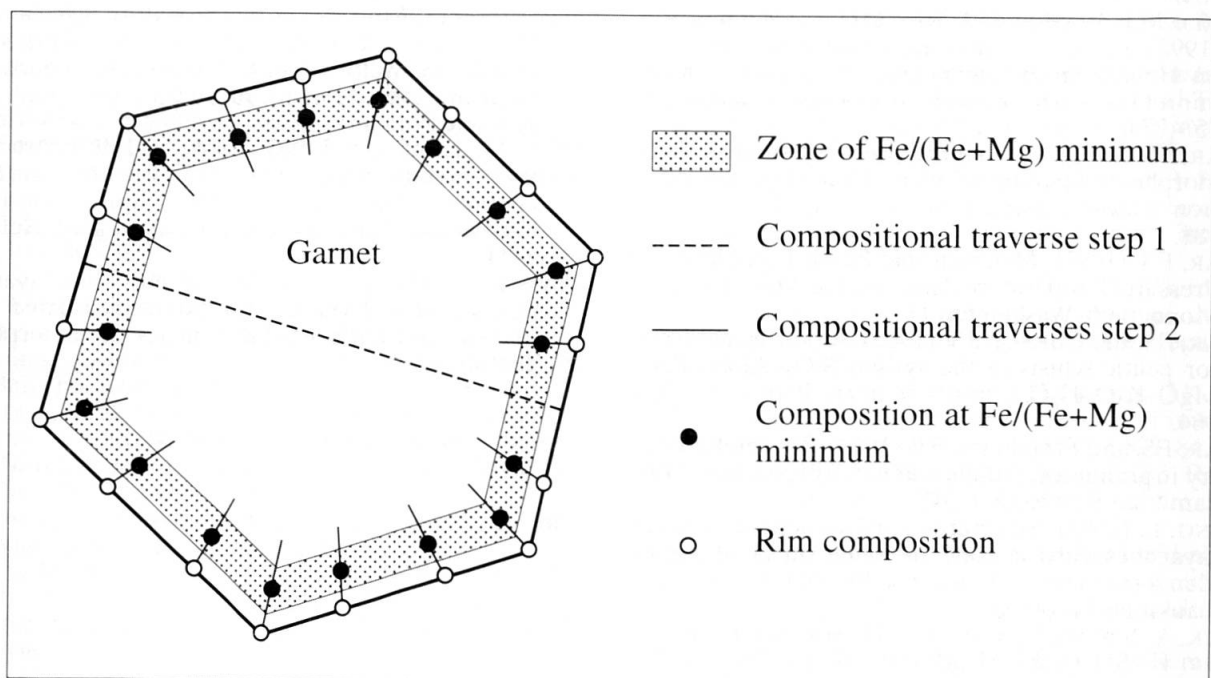


Fig. A1 Two-step analytical procedure for garnet compositional traverses.

Tab. A1 Representative garnet compositions

<b>Garnet compositions at Fe/(Fe + Mg) minima</b>									
Sample	EMP 2	EMP 3	EMP 4	EMP 6	EMP 7	EMP 8	EMP 10	EMP 11	
SiO <sub>2</sub>	36.64	36.60	37.45	37.58	37.05	37.35	37.18	37.38	
TiO <sub>2</sub>	0.02	0.01	0.03	0.03	0.01	0.03	0.00	0.02	
Al <sub>2</sub> O <sub>3</sub>	20.43	20.23	20.96	20.89	20.55	20.78	20.85	20.75	
Cr <sub>2</sub> O <sub>3</sub>	0.05	0.09	0.04	0.04	0.04	0.05	0.04	0.05	
FeO	32.50	33.19	34.13	34.05	35.53	33.81	34.65	34.36	
MnO	4.00	4.23	2.26	1.20	1.84	2.23	0.74	0.53	
MgO	3.48	3.30	4.17	4.40	3.36	4.15	4.59	4.63	
CaO	2.37	1.99	1.97	2.89	1.96	2.28	2.06	2.40	
Total	99.52	99.64	101.02	101.05	100.33	100.67	100.12	100.19	
Cations normalized to 12 oxygens									
Si	2.97	2.98	2.98	2.98	2.99	2.98	2.98	2.98	
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	
Al	1.95	1.94	1.97	1.95	1.95	1.96	1.97	1.96	
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	
Fe <sub>tot</sub>	2.20	2.26	2.27	2.26	2.39	2.25	2.32	2.29	
Mn	0.27	0.29	0.15	0.08	0.12	0.15	0.06	0.03	
Mg	0.42	0.40	0.49	0.52	0.40	0.49	0.55	0.56	
Ca	0.20	0.17	0.17	0.24	0.17	0.19	0.18	0.20	
Sum	8.04	8.05	8.04	8.04	8.04	8.04	8.04	8.03	
X <sub>Fe</sub>	0.70	0.72	0.74	0.72	0.77	0.72	0.75	0.74	
X <sub>Mg</sub>	0.13	0.13	0.16	0.16	0.13	0.16	0.18	0.17	
X <sub>Mn</sub>	0.08	0.09	0.05	0.02	0.04	0.05	0.01	0.02	
X <sub>Ca</sub>	0.06	0.06	0.06	0.08	0.06	0.63	0.06	0.07	
Fe/(Fe+Mg)	0.84	0.85	0.82	0.81	0.86	0.82	0.80	0.81	
<b>Garnet rim compositions</b>									
Sample	EMP 1	EMP 2	EMP 3	EMP 4	EMP 6	EMP 7	EMP 8	EMP 10	EMP 11
SiO <sub>2</sub>	36.46	35.96	36.35	36.86	37.17	37.05	36.97	37.15	36.83
TiO <sub>2</sub>	0.01	0.02	0.03	0.02	0.03	0.00	0.02	0.01	0.04
Al <sub>2</sub> O <sub>3</sub>	20.35	20.23	20.32	20.78	20.56	20.56	20.54	20.58	20.53
Cr <sub>2</sub> O <sub>3</sub>	0.03	0.06	0.03	0.07	0.06	0.04	0.06	0.02	0.04
FeO	29.26	32.06	31.84	23.92	33.61	36.20	34.56	35.63	35.94
MnO	8.15	7.17	7.30	5.33	3.27	1.99	3.56	0.91	0.98
MgO	1.67	2.21	2.10	3.10	3.04	2.54	2.62	3.38	3.29
CaO	3.66	1.35	1.24	1.43	2.88	2.04	1.93	2.52	2.32
Total	99.36	99.09	99.22	100.51	100.65	99.93	100.29	100.23	99.97
Cations normalized to 12 oxygens									
Si	2.98	2.96	2.99	2.97	2.99	2.97	2.99	2.99	2.98
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.99
Al	1.96	1.96	1.97	1.98	1.96	1.97	1.96	1.95	1.96
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe <sub>tot</sub>	2.00	2.21	2.19	2.22	2.26	2.46	2.34	2.40	2.43
Mn	0.56	0.50	0.50	0.37	0.22	0.13	0.24	0.06	0.07
Mg	0.20	0.27	0.26	0.37	0.37	0.30	0.31	0.40	0.34
Ca	0.32	0.12	0.10	0.12	0.25	0.18	0.16	0.22	0.20
Sum	8.03	8.04	8.02	8.03	8.04	8.04	8.02	8.03	8.04
X <sub>Fe</sub>	0.64	0.72	0.71	0.72	0.73	0.81	0.76	0.77	0.79
X <sub>Mg</sub>	0.06	0.08	0.09	0.12	0.12	0.10	0.10	0.13	0.12
X <sub>Mn</sub>	0.18	0.16	0.16	0.11	0.07	0.04	0.08	0.02	0.02
X <sub>Ca</sub>	0.10	0.03	0.04	0.04	0.08	0.06	0.06	0.07	0.06
Fe/(Fe+Mg)	0.90	0.89	0.88	0.85	0.86	0.88	0.88	0.86	0.87



Tab. A2 Representative plagioclase compositions.

Plagioclase rim compositions									
Sample	EMP 1	EMP 2	EMP 3	EMP 4	EMP 6	EMP 7	EMP 8	EMP 10	EMP 11
SiO <sub>2</sub>	61.01	62.97	63.22	63.12	62.61	61.56	60.20	62.90	62.56
Al <sub>2</sub> O <sub>3</sub>	23.82	23.48	22.83	23.66	23.73	24.37	24.45	23.47	23.45
FeO	0.19	0.18	0.07	0.18	0.15	0.04	0.08	0.08	0.05
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	5.48	4.68	4.02	4.94	5.08	5.83	5.84	4.91	4.89
Na <sub>2</sub> O <sub>3</sub>	8.56	9.03	9.37	8.87	8.78	8.26	8.32	8.96	8.94
K <sub>2</sub> O	0.12	0.07	0.17	0.09	0.11	0.14	0.11	0.05	0.06
Total	99.19	100.44	99.69	100.86	100.50	100.20	99.04	100.36	99.96
Cations normalized to 8 oxygens									
Si	2.73	2.77	2.80	2.77	2.76	2.72	2.70	2.78	0.77
Al	1.26	1.22	1.19	1.22	1.23	1.27	1.23	1.22	1.23
Fetot	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca	0.26	0.22	0.19	0.23	0.24	0.28	0.28	0.23	0.23
Na	0.74	0.77	0.80	0.76	0.75	0.71	0.72	0.76	0.77
K	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.00	0.00
Sum	5.01	5.00	5.00	4.99	5.00	4.99	5.01	4.99	5.00
X <sub>ab</sub>	0.73	0.77	0.80	0.76	0.75	0.71	0.71	0.77	0.77
X <sub>an</sub>	0.26	0.22	0.19	0.23	0.24	0.28	0.28	0.23	0.23
X <sub>or</sub>	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.00

Tab. A3 Representative biotite compositions.

Biotite rim compositions									
Sample	EMP 1	EMP 2	EMP 3	EMP 4	EMP 6	EMP 7	EMP 8	EMP 10	EMP 11
SiO <sub>2</sub>	35.71	35.22	35.16	36.32	35.46	35.42	35.36	36.23	35.96
TiO <sub>2</sub>	1.81	1.93	2.70	2.38	2.46	2.72	2.44	1.70	1.72
Al <sub>2</sub> O <sub>3</sub>	18.82	18.77	18.85	19.38	19.11	18.50	18.62	18.66	18.92
FeO	19.87	19.89	20.14	18.13	20.07	20.34	19.39	16.77	17.08
MnO	0.20	0.18	0.19	0.15	0.15	0.04	0.10	0.08	0.04
MgO	9.50	10.03	8.86	10.39	9.12	9.34	10.00	12.12	11.76
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na <sub>2</sub> O	0.10	0.30	0.23	0.32	0.20	0.25	0.20	0.31	0.32
K <sub>2</sub> O	9.54	9.17	9.19	9.11	9.16	9.14	8.78	8.86	8.67
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	95.57	95.49	95.33	96.18	95.75	95.77	94.91	94.98	94.85
Cations normalized to 11 oxygens									
Si	2.71	2.68	2.68	2.71	2.69	2.69	2.69	2.72	2.71
Ti	0.10	0.11	0.15	0.13	0.14	0.16	0.14	0.09	1.00
Al	1.69	1.69	1.69	1.7	1.70	1.66	1.67	1.66	1.68
Fe <sub>tot</sub>	1.27	1.27	1.28	1.13	1.27	1.29	1.24	1.06	1.08
Mn	0.01	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.00
Mg	1.07	1.14	1.00	1.15	1.03	1.06	1.14	1.36	1.32
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.01	0.04	0.04	0.05	0.03	0.04	0.31	0.04	0.05
K	0.92	0.89	0.90	0.86	0.89	0.89	0.85	0.85	0.83
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Sum	7.80	7.83	7.77	7.76	7.77	7.78	7.77	7.80	7.79
X <sub>ann</sub>	0.44	0.43	0.45	0.39	0.44	0.45	0.42	0.36	0.37
X <sub>phl</sub>	0.37	0.39	0.35	0.40	0.36	0.37	0.39	0.45	0.46

Tab. A4 Representative muscovite compositions.

Muscovite rim compositions									
Sample	EMP 1	EMP 2	EMP 3	EMP 4	EMP 6	EMP 7	EMP 8	EMP 10	EMP 11
SiO <sub>2</sub>	46.16	45.86	45.71	45.89	46.01	46.07	45.97	45.98	46.16
TiO <sub>2</sub>	0.48	0.77	0.77	0.91	0.89	0.94	0.86	0.67	0.81
Al <sub>2</sub> O <sub>3</sub>	34.75	34.92	34.48	35.40	35.24	34.86	35.13	35.30	34.97
FeO	1.64	1.94	2.11	1.15	1.27	1.43	1.31	1.16	1.13
MnO	0.05	0.02	0.03	0.02	0.02	0.03	0.02	0.00	0.02
MgO	0.91	0.72	0.74	0.75	0.73	0.81	0.83	0.79	0.93
CaO	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.02
Na <sub>2</sub> O	0.90	1.27	1.09	1.10	0.87	0.98	0.96	1.67	1.58
K <sub>2</sub> O	10.17	9.68	9.86	9.91	10.15	10.09	10.06	9.05	9.09
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	95.08	95.17	94.80	95.15	95.18	95.19	95.15	94.69	94.71
Cations normalized to 11 oxygens									
Si	3.08	3.06	3.07	3.06	3.07	3.07	3.06	3.06	3.07
Ti	0.02	0.04	0.04	0.05	0.04	0.05	0.04	0.03	0.04
Al	2.73	2.75	2.73	2.78	2.77	2.74	2.76	2.77	2.75
Fe <sub>tot</sub>	0.09	0.11	0.12	0.06	0.07	0.08	0.07	0.07	0.06
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	0.09	0.72	0.07	0.74	0.07	0.08	0.08	0.08	0.09
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.12	0.16	0.14	0.14	0.11	0.13	0.13	0.22	0.20
K	0.87	0.83	0.85	0.84	0.86	0.86	0.86	0.77	0.77
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Sum	7.01	7.02	7.02	7.00	6.99	7.00	7.00	7.00	6.99
X <sub>ms</sub>	0.88	0.83	0.86	0.86	0.85	0.87	0.87	0.78	0.79
X <sub>pg</sub>	0.12	0.17	0.14	0.14	0.15	0.13	0.13	0.22	0.21



*Plate 1* Sections for the Spiti valley-eastern Lahul-Parvati valley area showing metamorphic mineral zones for pelitic rocks. Tracks of the sections I and II are indicated in figure 1 (inset B), simplified after WYSS et al. (1999). For each mineral zone, the mineral assemblages corresponding to the stages of metamorphic crystallization  $M_1$  to  $M_4$ , the index mineral, schistosity  $S_x$  and deformational phases  $D_x$  are indicated. In the kyanite zone, the occurrence of sillimanite in the Tos Dextral Transtension Zone is indicated. Note that the transition from the Tethyan Himalaya into the High Himalayan Crystalline is gradual and note the abrupt decrease of  $M_2$  metamorphic conditions between the garnet zone and the biotite zone. For abbreviations, see figure 1. Sed = sedimentary minerals, Met = metamorphic minerals, 1 = mélangé zone, 2 =  $D_4$  shear zones in schists and phyllites within the Berinag Group quartzites.

