

**Evolution of the Main Central Thrust Zone  
in the Higher Garhwal-Kumaun Himalayas:  
An integrated structural and textural study**

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**Dedication**

***This thesis is dedicated with deep respect to the loving memory of my grandfather, the late Balaknath Ghosh, who was a poor rural peasant and, unfortunately like many of my countrymen still today, did not know even how to hold a pen, but who knew how to inspire his son to become an enlightened teacher.***

### **DECLARATION**

I declare that this thesis has been composed by myself and contains results of my own research. Work of others, wherever consulted, has been duly acknowledged.

(Subrata Ghosh)

## ABSTRACT

The thesis provides as much of basic geological information as possible on the Main Central Thrust Zone (MCT-Zone) of the Himalayas taking two sections into focus: the Alaknanda section between Joshimath and Gulabkoti was studied in greater detail than the Darmaganga section between Sobala and Khet, the former in Garhwal and the latter in Kumaun. Thrusting along the MCT-Zone has brought high grade crystalline rocks of the Higher Himalayas on top of the low grade Lesser Himalayan rocks. A classic inverted metamorphic sequence is noticed across the MCT-Zone into the crystallines above. The major deformation and metamorphism episodes that affected the MCT-Zone and its neighbourhood, the broad physical conditions operating during the MCT-emplacement, and the relationship of MCT-emplacement with the inverted metamorphism in the area have been worked out.

The MCT-Zone ('Munsiari Formation' after Valdiya, 1980) is essentially a complex shear zone with lithology-controlled strain-partitioning, but shows a general downward increasing gradient in strain. The upper and lower boundaries of the zone are given by the Vaikrita Thrust and the Munsiari Thrust respectively. These two boundaries have been clearly located in the field unlike before. Two generations of stretching lineations are present in the MCT-Zone; the dominant set corresponds to the time of main emplacement of the MCT-sheet, while the latter is of post-retrogressive reactivational origin.

Four episodes (F1, F2, F3 & F4) of folding could be recognised taking the MCT-Zone and the hangingwall into consideration. F2 is the most dominant fold-episode. Though MCT-stretching or shearing started broadly synchronously with the F2-episode, the peak of the main stretching event (i.e. the main MCT-emplacement time) came about postdating the F2-episode and broadly synchronously with the F3 episode. The fold episodes (FI, FII & FIII) in the footwall (Berinag-Mandhali formations) could not be correlated with those above.

Four episodes of metamorphism (M1, M2, M3 & M4) affected the central crystallines as well as the MCT-Zone rocks. There was no one-to-one correspondence between the folding and metamorphic episodes except the F3 and M3 both of which took place broadly synchronously with the peak of MCT-shearing. M4 corresponds to the retrogressive metamorphism consequent upon and subsequent to the main MCT-emplacement. The *lower amphibolite* facies M3 metamorphism indicates the physical conditions prevailing during the MCT-emplacement. M2 was the highest grade of metamorphism attained in the area; the progressive Barrovian metamorphic zones that now occur in an inverted order originally developed at the M2 time. M2 was slightly earlier than F2-folding, while the M1 metamorphism was pre- or syn-F1. Events from M2 onward were in direct or indirect response to continuing migration of India following the collision with the Eurasian plate at 50±5 Ma, while M1 and/or F1 was pre-collisional (probably pre-Tethyan, i.e. Precambrian). Imprints of four episodes of metamorphism (MI to MIV) are recognised in the Berinag-Mandhali formations; their correlation with the hanging wall events is not conspicuous.

Inverted metamorphism in the area is shown by the M2 isograds disposed in an inverted order on the overturned limb of a largescale multiorder overturned antiformal F2-fold. Thrusting along the MCT played a passive thickness-modifying role upon the already inverted metamorphic zones and, therefore, the MCT-emplacement was not genetically related to the inverted metamorphic sequence developed at the base of the Higher Himalayas.

## PREFACE

(with acknowledgements)

It is said that the present status of our knowledge about Himalayan geology is comparable to that of the Alps 100 years ago. After having worked for this thesis I realise the full significance of this saying.

The thesis deals with the tectonometamorphic evolution of the Main Central Thrust Zone of the Himalayas as revealed through an integrated structural and textural study of the rocks across the zone from two sections: Alaknanda section (Joshimath area) in Garhwal, and Darmaganga section (Sobala area) in Kumaun. The main focus is on the Joshimath area (Alaknanda section), because this area exposes a very complete succession of the MCT-Zone rocks. The Sobala area (Darmaganga section) has been studied for the purpose of comparison. The reader should take this thesis as a basic groundwork that paves the way for future detailed study in these areas. For obvious reasons I had to put much emphasis on noting the field attributes of the rocks and collecting important specimens, mostly oriented. Most of the specimens are kept in the Grant Institute of Edinburgh University, the rest are kept in the Geology Department of Presidency College, Calcutta.

Strictly speaking, the thesis gives a much generalised picture i.e. a broad framework of the tectonometamorphic evolution of the Main Central Thrust Zone and its immediate environs. Further advanced study is necessary in order to bring out the details, thereby tracing out a more complete picture of the overall evolution. The observations embodied in the thesis are only part of a whole wealth of information to be accrued from this classic area where shear-related structural and chemical features are so well preserved and in such a wide range of scale. To this end, the thesis could be taken as a guide and preliminary source-book of data.

It is a great pleasure to record with heartfelt gratitude my sincere thanks to all those individuals and organisations but for whose help and cooperation the present work could not have been completed. Even though it would be futile to attempt naming them all, I cannot but take this opportunity to mention the following:

The funds for the project came mainly from two sources-- first, from the Government of West Bengal (India) in the form of a State Scholarship, and secondly, from the Committee of Vice-Chancellors and Principals (CVCP), U. K., in the form of an ORS- Award. In the final stage

thanked for their kind help. Dr J. Behrmann of Giessen, Germany very kindly provided the results of X-ray texture analyses of 6 quartzite specimens.

On my way to or from the field, I benefitted from discussions held with Prof K. S. Valdiya at Nainital, Drs V. C. Thakur, A. K. Sinha and A. K. Dubey at Dehra Dun, Prof D. Mukhopadhyay at Dhanbad (now at University of Calcutta), Dr Subhasis Sengupta, Profs P. K. Gangopadhyay and A. K. Saha, and Messrs Ananda Chakrabarti and Pradyot Banerjee in Calcutta. Special help was rendered by the Geology Departments in the Presidency College, Calcutta and in the Indian School of Mines, Dhanbad for trimming, packing and mailing of the rock specimens collected in the field. Mr S. Mukherji of the Indian High Commission in London always completed the formalities regarding field-trips very sincerely.

In the field, accommodation was kindly provided by the Sankaracharya Asrama at Joshimath (Garhwal) and the National Hydroelectric Power Corporation (NHPC) of India at their project-site at Sobala/Nyu (Kumaun). Without such base camps, doing fieldwork in the mighty Himalayan mountains would have been much more difficult. However, it must be mentioned that being an Indian myself, I was in a much better position logistically to carry out fieldwork in those strategically sensitive areas.

Gregory, Angelika & Steve, my flatmates and close friends in Edinburgh, provided great moral support over the last few months.

At the home front, there was no dearth of encouragements from my loving parents, brother, brother-in-law and sister. A special salutary mention is made of Susmita, my beloved wife and an invaluable source of inspiration. She gave me strong moral as well as material support when I needed them most. Without her by my side probably I could not have been able to submit my thesis. Much is also due to our little 'Gora', who missed me as much as I missed him.

I sincerely thank them all again.

Finally I would like to mention that the responsibility for any omissions or commissions in the contents of the thesis solely rests upon me.





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## Chapter-1

# INTRODUCTION

*Everything around us has its own story to tell. We are constantly striving to read as many of such stories as possible, and in the process gradually discovering that these are, in fact, interconnected parts of a still more charming story.....*

The Main Central Thrust Zone (MCT-Zone) is one of the most important geotectonic elements of the Himalayas. It lies at the junction between the Higher Himalayas with older crystalline rocks to the north and the Lesser Himalayas with younger metasediments to the south. Representing perhaps the largest shear zone of the Himalayas, the MCT-Zone has accommodated a significant proportion of the shortening near the northern margin of the Indian plate in response to its northward migration after collision with the Asian plate. Although the tectonic implications are fairly well-known, there are serious gaps in our knowledge about the basic geological character of the MCT-Zone.

### 1.1 OBJECT AND SCOPE OF THE PRESENT WORK

The primary objective of the present work is to remedy the lack of detailed geological information on the MCT-Zone and its immediate neighbourhood, and thereby to provide a framework based on rigorous field and petrographic observations for the study of the tectonometamorphic evolution of the MCT-Zone. Basically the thesis is intended to find answers to the following questions:

(i) Is it possible to locate precisely the Vaikrita Thrust and the Munsiri Thrust that bound the MCT-Zone? What is the nature of these thrusts, and of the MCT-Zone (i.e. the Munsiri Formation) as a whole? What is the most likely reason for the lateral variation in the thickness of the MCT-Zone? Could we specify the magnitude and direction of transport along the MCT-Zone?

(ii) Was the thrust-emplacement along the MCT-Zone a single event or was it episodic? Can we relate this with the post-collisional migration (-cum-rotation) of

the Indian plate? What is the age of the MCT-emplacment? What are the age relations between the Vaikrita Thrust and the Munsiri Thrust?

(iii) What are the styles of folding and distribution of folding across the MCT-Zone? How many episodes of folding could be recognised in the MCT-Zone and its immediate neighbourhood?

(iv) What is the time relation of MCT-emplacment with the fold episodes established? Is there any conspicuous effect of the stretching/shearing event/s upon the folds that developed prior to the stretching event?

(v) How many episodes of metamorphism could be recognised in the three units in Joshimath area and what is their time relation with the folding and thrusting events? What were the broad P-T conditions during the different metamorphic episodes?

(vi) What is the nature and distribution of the metamorphic minerals, particularly the members of the Barrovian Series, across the MCT-Zone, and how is the well-known 'inverted metamorphic sequence' defined in the area? Which model of inverted metamorphism is favoured by the data collected through the present study? Could it be possible to solve the long-sustaining controversy on the origin of the Higher Himalayan inverted metamorphic sequence?

The study concentrated mainly on the Joshimath area in Garhwal, where a very representative, well-preserved sequence across the MCT-Zone exists along the Alaknanda valley section. For the purpose of comparison, the Sobala area in Kumaun, about 100 km ESE of Joshimath, was looked into following a road-section along the Darmaganga valley. Another purpose for extending the study into the latter area was to investigate the reason for the marked variation in the thickness of the MCT-Zone from about 2.5 km in Joshimath area to only about 400 m in Sobala area.

## 1.2 THE STUDY AREAS

Fig. 1.1 shows the location of the two study areas. The Vaikrita Thrust and the Munsiri Thrust are respectively the upper and lower bounding thrusts of the MCT-Zone. The Joshimath township and the Sobala village are both situated just above the MCT-Zone that runs for c.2000 km along the Himalayas. The two study areas span across the junction between the Higher Himalayas and the Lesser

Himalayas along the Alaknanda and Darmaganga sections respectively. The Alaknanda section from Gulabkoti in the south to Joshimath in the north preserves a very representative succession across the MCT-Zone as well as encompassing parts from both of its hanging-wall and foot-wall blocks. The Vaikrita Crystallines (Joshimath Gneiss) form the immediate hanging wall, while the Berinag-Mandhali formations constitute the immediate footwall to the MCT-Zone.

### 1.2.1 Location

Both Joshimath and Sobala are included in the U.P. Himalaya which is a part of the Central Himalayas and is commonly known as Kumaun Himalaya. [U.P. stands for Uttar Pradesh, i.e. 'Northern Province' in English translation; Uttar Pradesh is one of the largest states in North India]. The U.P. Himalaya stretches from the Kali river in the east defining the India-Nepal border to the Tons-Pabar valleys in the west demarcating the eastern border of Himachal Pradesh (Himachal Himalaya). Precisely speaking, the U.P. Himalaya has two divisions -- the western division is called Garhwal and the eastern called Kumaun. Of the eight administrative districts falling in the domain of U.P. Himalaya, five viz. Dehra Dun, Uttarkashi, Tehri, Pauri & Chamoli are included in the Garhwal region and the other three viz. Nainital, Almora and Pithoragarh are included in the Kumaun. Joshimath is one of the subdivisional ('Tehsil') headquarters in the Chamoli district in Garhwal, while Sobala is a small village in the Dharchula subdivision ('Tehsil') of Pithoragarh district in Kumaun.

### 1.2.2 Accessibility

Access to the study areas is difficult. The Joshimath field-area which ranges in altitude from about 1500 metres to 3300 metres above MSL is connected only by a narrow metalled road to Rishikesh, some 200 km away towards south-west, which in turn, however, is well connected (by rail and road) with Hardwar and Dehra Dun in the Himalayan foothills. Joshimath is situated approximately 45 km WNW of Mt. Nanda Devi (7817 mtr) (see Fig. 1.2). The Sobala area, which is closed for political reasons to all but Indian nationals, is about 100 km ESE of Joshimath and is still more difficult to reach. Only a fair weather road connects the Nyu village near Sobala with Tawaghat which, in turn, is connected by a bus-route to Tanakpur via Dharchula and Pithoragarh. One can also go by bus from Kathgodam via Nainital and Almora to Pithoragarh on way to Dharchula/Tawaghat. Dehra Dun, Rishikesh, Kathgodam and Tanakpur are among the important railway-heads serving the Garhwal-Kumaun region. Fig. 1.3 shows the most convenient routes to the study

areas. For further details on communication network in the region see Fig. 1.3 in Valdiya (1980, p. 3).

### **1.2.3 Exposure level & logistics**

The amount of rock exposure is quite good in both Joshimath and Sobala areas. However, being the mighty mountainous regions as they are, one must not think that one can reach and study all the rock-exposures in those areas. There are good road sections that expose almost the complete sequence of rocks across the MCT-Zone (Munsiari Formation) and a fairly representative fraction of the two other units, the underlying Berinag-Mandhali and the overlying Vaikrita or Joshimath Gneiss. Barring a few dangerous ones, the mountain slopes are quite negotiable and most of them are below the snow line and free from dense vegetation. The logistics are assisted by the existence of a few small villages scattered throughout the areas. The local people are very cooperative and peace loving.

### **1.2.4 Physiography & Climate**

These areas are characterised by rugged, high, mountainous topography. Mainly sectional view of the rocks is available, not plan view; hence in most cases it is difficult to measure the lengths of different structures, such as late shear zones, boudin axes etc. Plate-I shows the essential physiographic features of the core of the Joshimath field-area. One could clearly visualise from here the location in the field of the two bounding thrusts of the MCT-Zone (viz. the Vaikrita Thrust above and the Munsiari Thrust below) in relation to the major valley/s and approach roads. The lithology and structure of the rock horizons have an obvious control on the topography of the study areas. Usually the gentler northerly mountain slopes are dip slopes, those opposing the dip of the rock horizons are generally steep and scarp-like. The formation of lower order valleys transverse to the main drainage channel/s has been, to a large extent, influenced by lithology, the less resistant rock horizons guiding the locations of such valleys. The topography of the areas is at a youthful stage; rapid erosion is going on by the action of glacial meltwater and rainwater. Landslides are a major manifestation of erosion. The eroded materials are being drained from the Joshimath area by the SW-flowing river Alaknanda, the longer of the two major tributaries of the Ganga at its higher reaches. River Darmaganga is the main drainage channel in the Sobala area and falls into the Kali river near Tawaghat. More active erosion is going on in the Joshimath area than in the Sobala area. Generally speaking, the Garhwal region shows somewhat less mature topography/physiography than the Kumaun. In Kumaun region, particularly in its

Lesser Himalayan parts, the valleys are wider and better vegetated, river gradients are gentler, landslides are less common and ridge crests are less sharp or irregular than in Garhwal. Most probably an over-all higher annual rainfall and a higher proportion of carbonate rocks in the Kumaun region are two of the major factors leading to this difference.

The mountain ranges or ridges in and around the study areas are broadly NW-SE trending, mutually more or less parallel, but longitudinally they do not continue for long distances and they decrease in height, more or less steadily from NNE to SSW across the Lesser Himalayas. This last feature along the Alaknanda section from Joshimath southward can be seen clearly in Fig. 1.4(a); while Fig. 1.4(b) shows the snow-clad Higher Himalayan ranges looking northward from nearly the same spot. Fig. 1.5 shows a topographic section across the whole of Lesser and Sub-Himalayas and part of the Higher Himalayas through Vishnuprayag (Joshimath). The slope of a statistical (hypothetical) enveloping surface touching the tops of the peaks (i.e. ridge-tops) in this section would be about  $3^{\circ}$  to  $4^{\circ}$  SSW. It is presumed that such a uniformity in lowering of height of the ridges may not only coincidentally represent an erosional feature, but could be a reflection of the effects of tectonic wedging as well (Davis et al., 1983).

Climatically, the areas belong to a monsoon region. So fieldwork is not advisable from July until October. Moreover in December and January normally there is heavy snowfall, so communication is impaired.

Clear and sunny weather from February till mid-July, fresh i.e. less weathered rocks, good road-cut and natural cliff exposures, simple and friendly local inhabitants are some of the advantageous points for carrying out field studies in the areas. However, there are some disadvantageous points as well. For instance, lack of proper transport and communication makes field mapping a hazardous task; much of the time in fieldwork sessions is wasted on simply walking/trekking while trying to reach suitable exposures. The areas are near strategically sensitive international borders, so movements may be restricted at many places. In fact, Sobala area is closed for all but approved Indian workers. Good quality topographic base maps are very difficult to obtain. Due to stringent custom regulations there are often problems even to bring rock specimens for laboratory work.

In the title of the thesis the whole 'Garhwal-Kumaun Himalayas' is mentioned, even though the main work was concentrated on the Joshimath area. This is because:

(a) The Joshimath area is situated roughly at the middle of the length for which the MCT-Zone runs through the Garhwal and Kumaun Himalayas and the area exposes a very representative section across the MCT-Zone along with its contiguous parts from both the hanging wall and the footwall.

(b) Though Joshimath is located in Garhwal, the Sobala area in Kumaun was also looked into for the purpose of comparison. Sobala is c.100 km (straight line distance) ESE of Joshimath along the strike trend. As per the geological map of Valdiya (1980) the Central Crystallines (Vaikrita Gneisses) apparently come directly on top of the Berinag-Mandhalis (i.e. the Lesser Himalayan rocks) telescoping the MCT-Zone (Munsiari Formation) in the Sobala area. In reality, however, a thin MCT-Zone (~400 mtr in thickness) is found to exist just below the Central Crystallines in Sobala (see Chapter-2, Plate-V).

(c) Due to the inherent nature of the terrain, geological studies spreading quite homogenously throughout the Himalayas are not possible. Detailed study along selected traverses and reasonable extrapolation and interpolation of results are inevitable. Scientifically it is perhaps correct to assume that structural style does not change abruptly from area to area in a thrust zone of tectonic scale, particularly when it is widely known that thrusting along the MCT-Zone has brought high grade Higher Himalayan crystalline rocks on top of the low grade Lesser Himalayan rocks almost all along the length of the Himalayas.

### 1.3 METHODOLOGY USED

For the present study the emphasis was on systematic field studies and detailed petrographic (textural) work using the optical microscope in order to work out the deformation-metamorphism relationships. Altogether I carried out three sessions of fieldwork, each session lasting for three months and a half on average. Magnetic declination in the study areas is negligible; so, for the purpose of mapping, the magnetic north was taken as geographical north. The main objects in the fieldwork sessions were to carry out detailed mapping and structural analysis, making field metamorphic observations and systematic collection of rock specimens.

In the laboratory, detailed petrographic work with the help of optical microscopes were coupled with XRD and electron microprobe analyses for mineral identification, XRF analyses for the determination of bulk rock chemistry of selected rock specimens and U-stage work for some petrofabric studies. Mineral separation using magnetic mineral separator and hand-picking was done for some biotites,

muscovites and hornblendes before sending them to SURRC (Scottish Universities Research and Reactor Centre) at East Kilbride for dating by K-Ar, Rb-Sr and Ar-Ar laser probe methods. A selected number of quartzite specimens were analysed for quartz petrofabrics using X-ray texture goniometer (with kind help from Dr. J. Behrmann at Giessen, Germany). However, it is the field structural and metamorphic observations and the textural (optical petrographic) observations which will be mainly reported in this thesis (see title of the thesis and also section 1.1).

A critical appreciation of the present trend of tectonometamorphic studies being carried out in the Himalayas in general suggests that a significant number of these studies is devoted towards testing and/or proposing model/s for the Himalayan inverted metamorphism. Modelling the inverted metamorphism of the Himalayas is now-a-days being based more on indirect evidence like isotopic and geothermobarometric data than on direct microtectonic evidences and mesoscopic field evidences. Ideally, in order to carry out any good tectonometamorphic study with regional implications there must be a particular sequence of work to be followed in which the essential first step is to make a detailed geological map based on field observations. In the next step detailed petrographic (textural) study of systematically collected rock specimens has to be done with an aim of gleaning as much information as possible on tectonometamorphic relationships; this will then pave the way for the next step i.e. advanced isotopic and geothermobarometric studies. Otherwise our data cannot be put into the right perspective and so our interpretation of those data cannot be able to claim proper scientific legitimacy. Compared to other well-studied regions on the globe like the Scottish Highlands or the Alps, or the Appalachians, detailed mapping in the Himalayas is still lacking.

Field-studies and petrographic (textural) work are the best tools for the study of deformation-metamorphism relationships. Without demonstrating such relationships, it is rather difficult to convince ourselves about the relative timing of deformation and mineral growth (metamorphism) from the isotopic and geothermobarometric results only. The geothermobarometric and isotopic analyses are to be taken only to provide additional support, but never the other way round. This realisation has dictated the strong emphasis put in the present work on field geological observations and correlative textural/petrographic studies.

## 1.4 THE HIMALAYAS AND THE MCT-ZONE: AN OVERVIEW OF LITERATURE

Himalaya means 'an abode of snow' [Bengali (originally derived from Sanskrit): Him  $\equiv$  Snow, Alaya/s  $\equiv$  Abode/s]. Home for more than two-thirds of the world's highest peaks the entire Himalayan mountain belt extends for some 3000 km from Afghanistan in the west to Burma in the east. However, the typical Himalayas are represented by the c. 2500 km long central segment of this belt, arcuate, convex toward south, from the north-western syntaxis at Nanga Parbat (8125 m) where the antecedent Indus river cuts through the belt to the eastern syntaxis at Namche Barwa (7755 mtr) where the Brahmaputra river crosses the belt via its famous re-entrant.

The Himalayas are regarded as the most impressive example of a collision-type mountain belt which has resulted from the closure of the Tethys ocean (specifically, the Neo-Tethys) and the subsequent and continuing collision of India with Asia (Dewey & Bird, 1980; Sengor, 1979, 1984, 1985, 1986a & b, 1989; Shackleton et al., 1988; Windley, 1988). Introducing the special issue of the *Journal of Metamorphic Geology* (vol.7, No. 1, 1989) dedicated to Himalayan metamorphism, Barnicoat & Treloar (1989) comments, "The Himalayan region is one of great interest to petrologists because it is an area with recently uplifted metamorphic rocks clearly associated with continental collision. It provides a modern, evolving field laboratory in which to study the intimately associated processes of deformation and metamorphism, as well as to test thermal models of continental crust undergoing thrust stacking and uplift".

### 1.4.1 Sub-divisions of the Himalayas

The Himalayas are divisible into five distinct longitudinal physiotectonic zones viz. from N to S, the Trans-Himalaya, Tethyan Himalaya, Higher Himalaya, Lesser Himalaya and Sub-Himalaya or Foothills Himalaya (Gansser, 1964; Le fort, 1975; Valdiya, 1980; Windley, 1983, 1988). Each of these zones is delimited by important tectonic boundaries (see Fig. 1.6).

For ease of reference there has been transverse subdivision of the Himalayas as well, based on political or administrative zones viz. NW Himalaya (or, Kashmir/Punjab Himalaya), Himachal Himalaya, U.P. Himalaya, Nepal Himalaya, Darjeeling-Sikkim Himalaya, Bhutan Himalaya and NE Himalaya (or, Assam and Arunachal Himalaya) (see Fig. 1.7). Together the Himachal and the U.P. Himalayas

are normally regarded as the Central Himalayas. Many workers prefer to call the U.P. Himalaya as the Kumaun Himalaya, but in fact the U.P. Himalaya has two divisions - the western one is called Garhwal Himalaya and the eastern the Kumaun Himalaya.

Of the five longitudinal divisions, the Trans-Himalaya represents the Andean-type magmatic belt developed above the subducting Tethyan oceanic crust. Along with its extension in the NW Himalaya, i.e. the Kohistan-Ladakh island arc, the Trans-Himalaya is thought to be the last exotic terrane to be accreted to Asia along the Banggong/Shyok Suture prior to the main India-Eurasia collision (Pettersen & Windley, 1985; Debon et al., 1986; Coward et al., 1986).

The Indus-Tsangpo Suture (ITS) is the main collisional suture between the Indian plate and the Eurasian plate. The ITS clearly separates the Trans-Himalaya from the Tethyan Himalaya (see Fig. 1.6). Occurrence of deep-water Mesozoic sediments and melange material as well as ophiolitic rocks marks the suture zone. The largest ophiolitic bodies occur as klippe e.g. at Jungbwa in southern Tibet between Kailas and Mt Nanda Devi, and at Spontang in Ladakh (Gansser, 1964, 1979; Reibel & Reuber, 1982). Within the Indus-Tsangpo suture zone, evidence of high-pressure metamorphism is preserved in scattered outcrops of blue schists (Jan, 1985; Le Fort et al., 1989). The high-pressure assemblages are found largely in meta-igneous blocks contained within the suture zone melange.

South of the Indus-Tsangpo Suture is the Tethyan Himalaya consisting of an almost continuous succession of sediments from the Cambrian to the Eocene, which were deposited on the passive continental margin of the Indian plate (Gansser, 1964; Le Fort, 1975; Gupta & Kumar, 1975; Fuchs, 1979; Thakur, 1981; Sinha, 1981; Baud et al., 1984; Gaetani et al., 1986). The Tethyan sediments overlie the Higher Himalayan crystalline rocks that form the Precambrian metamorphic basement (Barnicoat & Treloar, 1989). The contact between the Tethyan sediments and the Higher Himalayan crystallines is marked by the Tethyan Thrust (TT; locally known as 'Malari Thrust' in Garhwal-Kumaun region) which runs through a zone of stratigraphical unconformity. During late-Tertiary gravitational spreading of the Himalayan crust, the TT was reactivated inversely; there has been movement of several tens of kilometres along this reactivated normal fault (Burg et al., 1984; Burchfiel & Royden, 1985; Windley, 1988).

South of the Tethyan Himalaya and separated from it by the TT, is the Higher Himalaya made up of mainly Precambrian, possibly Proterozoic, crystalline rocks

('Central Crystallines' of the Himalayas). The Central Crystallines constitute the most extensive tract of high-grade metamorphic rocks in the Himalayas. Le Fort (1981), Searle & Fryer (1986) have noticed that parts of the highest grades of metamorphism in the Higher Himalayas are associated with migmatites and anatectic granites. Much of the Higher Himalayas has been metamorphosed during Himalayan orogenic events. A series of Barrovian metamorphic zones disposed in an inverted order is seen in the Central Crystallines. The studies reporting such inverted disposition of Barrovian metamorphic zones at the base of the Higher Himalayas range along the whole length of the Himalayan chain, from Darjeeling in the east (Lal et al., 1981) through Nepal (Pecher, 1978) to Simla (Naha & Ray, 1970) to Kashmir (Searle & Fryer, 1986) in the west. Papers by Hubbard (1989), Mohan et al. (1989), Pecher (1989), Staubli (1989), Searle & Rex (1989) and Treloar et al. (1989) also document this inversion. A separate review of the Himalayan inverted metamorphism will be given shortly.

The rocks of the Higher Himalayas are separated from those of the Lesser Himalayas by the Main Central Thrust (MCT). See later in this section for more detailed review on the MCT. The Lesser Himalayan domain comprises dominantly of low-grade late-Proterozoic to Paleozoic sediments, parts of which have been overridden by the klippen of the high-grade Higher Himalayan gneisses (Stocklin, 1980; Valdiya, 1981; Sinha, 1981; Windley, 1988; Barnicoat & Treloar, 1989). Part of the metamorphic imprint seen in the Lesser Himalayan rocks is pre-Himalayan and possibly early Paleozoic or older (Pognante & Lombardo, 1989; Barnicoat & Treloar, 1989). Johnson & Oliver (1990) argued vigorously in favour of the presence of pre-Himalayan (pre-Eocene) metamorphism in the Lesser Himalayas. Clear indication of pre-Himalayan (pre-MCT) metamorphism has been found in the Lesser Himalayan rocks that were included in the present study (see later in Chapter-5). Recent studies by Morrison & Oliver (1992) in and around the Kathmandu Klippe in Nepal Himalaya indicated a pre-Himalayan (possibly, lower Paleozoic) metamorphism of the Lesser Himalayan rocks.

The Sub-Himalaya, separated from the Lesser Himalaya by the Main Boundary Thrust (MBT), is made up of mostly Neogene Siwalik molasse sediments that were deposited in foreland basins and have been affected by the more recent of Himalayan deformation. The Siwalik sediments are gradually being thrust upon the alluvium of the Indo-gangetic plains of North India along the Main Frontal Thrust (MFT) of the Himalayas (see Fig. 1.6).

### 1.4.2 General tectonics of the Himalayas

The present-day physiographic and plate tectonic setting of the Himalayas and surrounding regions is indicated in Figs. 1.8a-c. Representing in many ways a classic type of collisional orogenic belt the Himalayas contain every significant stage in the Wilson cycle from Permian intracontinental rifting to the Quaternary neotectonics and rifting of Tibet. Precollisional ophiolite obduction and thrusting of shelf sediments, syncollisional suturing and thrusting, and postcollisional indentation of Eurasia by India, giving rise to late thrusts, inverted isograds and Miocene leucogranites, are all easily distinguishable (Shackleton et al., 1988).

Continental collision between the Indian and the Asian plates along the Indus-Tsangpo Suture took place in Eocene time at about  $50 \pm 5$  Ma, based on —

(a) the age of the earliest molasse sediments (Late Eocene - Oligocene) (Searle, 1983), (b) change in the rate of motion of the Indian Plate (from 10 cm/yr to 5 cm/yr at ca.50 Ma) (Dewey et al., 1989), (c) zig-zag pattern of movement of the Indian Plate at around 55-50 Ma (Patriat & Achache, 1984; Besse & Courtillot, 1988), (d) intersection with the projected edge of the Indian Plate, (e) from peak metamorphism and cooling dates (about 55 Ma+) (Coward, 1992, oral presentation, Edinburgh). Coward (ibid.) suggested that the total syn- to post-collisional shortening involving Tibet and the Himalayas together is of the order of 2500 km of which 1750 km is of the Greater India i.e. the Himalayas and 750 km is of Tibet. Clearly a significant proportion of the post-collisional shortening across the Himalayas was accommodated by the shearing along the MCT-Zone. However, taking both sides of the collisional suture zone into account, the post-collisional convergence since Eocene has been taken up by a combination of intracontinental thrusting, homogeneous crustal thickening and the lateral expulsion of material along strike-slip fault systems developed in the Asian plate. A lively debate is continuing on the relative importance of these mechanisms (e.g. Mattauer, 1984; Tapponier et al., 1986; England, 1987; Dewey et al., 1989).

### 1.4.3 Inverted metamorphism in the Himalayas

'One of the most problematic facts of the whole Himalayan range' (Gansser, 1964, p.99) is the inverted metamorphism. Since Medlicott's study in 1864 it has been known that almost throughout the entire length of the Himalayas the highgrade rocks occur structurally and topographically above the lowgrade rocks. In fact, there are two facets of Himalayan inverted metamorphism - firstly, there is inverted

metamorphic stratigraphy the straightforward cause of which is in most cases the occurrence of thrusts bringing higher grade and older rocks on top of the lower grade and younger rocks; and the other, which is more important, is the occurrence particularly at the base of the Higher Himalayas of a sequence of metamorphic isograds in reverse order. While in normal settings we would expect the Barrovian 'chlorite' zone to be underlain by 'biotite' zone, that in turn underlain by 'garnet' zone and then successively by 'staurolite', 'kyanite' and 'sillimanite' zones, one below the other, the Himalayan example shows the position of these zones in opposite order - 'chlorite' overlain by 'biotite' that, in turn, by 'garnet' and so on up to the 'sillimanite' so that the 'chlorite'-zone is encountered in the field at a lower level and the 'sillimanite'-zone at a higher level. It is this second aspect of Himalayan inverted metamorphism that concerns us most. Ray (1947) for the first time clearly mapped such inverted disposition of isograds in the Darjeeling Himalaya (see also Mukhopadhyay & Gangopadhyay, 1971). In many instances it has been observed that the metamorphic isograds are discordant to the lithological or stratigraphical boundaries (see Auden, 1935; Ray, 1947; Powell & Conaghan, 1973a etc.). However, in Joshimath area such discordance is not very apparent.

There has been no unanimity in opinion as to the specific cause of the development of the inverted metamorphic sequence in the Himalayas. Explanations for the metamorphic inversion are wide ranging:

- 1) The reverse or inverted metamorphism is entirely a thermal metamorphic effect due to the occurrence of igneous intrusives in the Higher Himalayas and the Himalayan rocks do not necessarily involve inverted stratigraphic succession due to thrusting or overturned folding (Mallet, 1875; Auden, 1935 etc.)
- 2) Huge recumbent or overturned folding following the main isograd forming metamorphism has resulted in the inverted disposition of isograds in the overturned limb (Heron, 1934; Heim & Gansser, 1939; Ray, 1947; Frank et al., 1977; Searle & Rex, 1989)
- 3) The inverted metamorphic sequence is a result of thrusting on a multiplicity of planes affecting an already existing metamorphic terrain (Pilgrim & West, 1928; Bordet, 1961; Nagdir et al., 1967; Naha & Ray, 1971; Hashimoto et al., 1973; Saxena, 1973; Ghose et al., 1974; Treloar et al., 1989).

- 4) The inverted metamorphism is a consequence of the transient inversion of isotherms during movement along the MCT ---- the so-called 'Le Fort' or 'hot iron' model. The idea is that a nappe of hot material (the Central Crystallines) was thrust over cold material of the Lesser Himalaya with isotherms that were folded by the thrusting (Le Fort, 1975; Pecher, 1978). This model has subsequently been refined and it has been suggested that although the upward increase in grade seen below the MCT is a result of the 'hot iron' effect (Caby et al., 1983; Brunel & Kienast, 1986; Pecher, 1989; Staubli, 1989), the downward decrease in grade seen in the Central Crystallines (the hanging wall of the MCT) is the result of a retrogression of earlier assemblages.
- 5) Though shear heating has been discounted by some workers (e.g. Hubbard, 1989) as a significant effect, a modified version of the 'Le Fort' model with an additional component of shear heating added to it has been suggested by Molnar & England (1990) and England et al. (1992).
- 6) Searle et al. (1988) envisaged as a possibility, but eventually disfavoured, a model involving two-stage development of the inverted metamorphic sequence. In this model an earlier emplacement of a 'hot' High Himalayan hanging-wall on to a 'cold' Lesser Himalayan footwall at intermediate pressures (kyanite grade) was followed by a higher T - lower P (sillimanite grade) metamorphism related to granitic magmatism. They (Searle et al., *ibid.*) discounted this model on the basis that any thermal imprint from partial melting would not be so spatially extensive (up to 50 km away from the nearest granite source region) and that there would be no effect of any shear heating outside the immediate vicinity of the high-strain shear zones, mainly because of the extensive fluid and volatile fluxing within the High Himalaya.

The relative merits<sup>and</sup> demerits of the different models of inverted metamorphism will be discussed in Chapter-6 mainly in the light of the results obtained from the present study. However, it is worth noting here that there is a general trend now-a-days to overemphasize the variations in tectonometamorphic relationships in the Himalayas in different sections across the MCT-Zone. While suggesting or testing a model for the inverted metamorphism, it is always important to remember that the Himalayas form a distinct chain of mountains on the earth's

surface; the MCT is an important geometric element of the Himalayas; thrusting along the MCT took place at a particular zone at a particular geological time; occurrence of the inverted sequence of isograds at the base of the Higher Himalayas is also a major feature of the Himalayas. So it is quite unreasonable in scientific terms, which is to say that it would be an exception rather than the rule, to expect different reasons for the development of this inverted metamorphism in different sections through the Higher Himalayas, as much as it may be unreasonable to expect different causes for the thrusting along the MCT in different sections.

Generally speaking, the discussions on inverted metamorphism models in the Himalayas do not specify the mutual geometric relations among isobars, isotherms and metamorphic isograds. Bhattacharya (1981) pointed out the importance of analysing these relations. In recent years, considerable emphasis is being placed on the results of geothermometry and geobarometry together, in a few cases, with P-T path analysis (see, for example, Hodges et al., 1988; also the special issue of the *Journal of Metamorphic Geology*, vol. 7, No. 1, 1989).

As appropriately highlighted by Gansser (1991), one good thing is that now the Himalayan geologists do not suffer from a dearth of working models. Indeed probably all the plausible models for the Himalayan inverted metamorphism are at hand now. The task is essentially to choose the right one which will fit the observations and have general applicability, rather than local and restricted, throughout the entire length of the Himalayas. From amongst the existing models the one that is applicable, statistically speaking, to a greater number of sections across the Higher Himalayas has a greater chance of being the correct one. Detailed discussion on modelling the Higher Himalayan inverted metamorphism will be given in Chapters 5 and 6.

#### **1.4.4 The Main Central Thrust Zone (MCT-Zone)**

Separating the Higher Himalayan domain from the Lesser Himalayas the Main Central Thrust is not a discrete thrust plane, but a considerably thick and complex movement zone (i.e. a shear zone). The MCT-Zone (equivalent to the root zone of the 'Munsiari Formation' as recognised by Valdiya, 1980) is bounded at the top and bottom by what have been designated by Valdiya (ibid.) as the Vaikrita Thrust and the Munsiari Thrust respectively. These two boundaries of the MCT-Zone are referred to by some workers as MCT1 and MCT2 respectively (Hodges et al., 1988; Morrison & Oliver, 1992). However, the only slight difficulty with this

system of nomenclature is its obvious age connotation. The age relationship between the two boundaries is yet to be established accurately. Thickness of the MCT-Zone varies laterally e.g. in Joshimath area it is 2.5 km, while at Nyu (Sobala) it is only about 400 m in thickness. The 'Main Central Thrust' as a term was first introduced in literature by Heim & Gansser (1939) to refer to the present day lower boundary of the MCT-Zone. Valdiya's Munsiri Formation (i.e. the MCT-Zone) extends across the Lesser Himalayas as a thrust-nappe which is now preserved in the form of different detached klippen, such as at Askot, Baijnath, Almora, Nandprayag, Lansdowne etc. (see the geological map in Valdiya, 1980). Details of regional geological setting of the Lesser Himalayan region of Garhwal-Kumaun Himalayas in relation to the MCT-Zone (Munsiri Formation) will be given in Chapter 2, while the features of the two boundaries of the MCT-Zone are more fully dealt with in Chapter 4.

The generally accepted but poorly constrained age-bracket for the emplacement of the MCT-sheet is 30-10 Ma B.P. (Windley, 1983, 1988; England et al., 1992). Hubbard (1988) demonstrated from petrographic study of an amphibolite specimen collected from the MCT-Zone near the Everest region of Nepal that amphiboles grew during the main shearing event. From radiometric dating of these amphiboles (hornblende) and using hornblende blocking temperature (550°C), Hubbard & Harrison (1990) measured an  $^{40}\text{Ar}/^{39}\text{Ar}$  age of 20-21 Ma for the MCT-emplacement. This is so far the most acceptable date of MCT emplacement. Precisely speaking, however, this gives the age when the MCT was active. Radiometric or any reliable evidence is lacking on the mutual age relations of the Vaikrita Thrust and the Munsiri Thrust which are regarded as the upper and lower boundaries of the MCT-Zone respectively. Applying the 'piggy-back' model of thrust development, Johnson (1986) argued that the thrusts in the (Kumaun) Himalayas developed in a 'hinterland-to-foreland' sequence. For modelling the sequence of thrust development in the Kumaun Himalaya, Johnson used the geological observations of Valdiya (1980) and found a similarity in its general geological set-up with that of the Canadian Rockies, and pointed out that the arguments Dahlstrom (1970) used for the thrust sequence development in the Canadian Rockies could apply for the Kumaun Himalaya as well. The big puzzling point is that in none of the Lesser Himalayan klippen, do we find typical Central Crystallines; the klippen are occupied largely by the Munsiri Formation (see the geological map in colour in Valdiya, 1980). So it is very difficult to know how far to the south the Vaikrita Thrust extends. It is not yet clear whether translation along the Vaikrita Thrust was more than or less than or equal to that along the Munsiri Thrust.

The precise magnitude of transport along the MCT is unknown (Windley, 1983; Johnson, 1986; Brunel, 1986). A lower limit of at least 100 km is given on the basis of the distance between the northernmost outcrop of the Munsiri Formation and its southernmost outcrop in a klippe in the Lesser Himalaya (say, the distance between Helang & the southern margin of the Lansdowne klippe near Lansdowne) (see also Andrieux et al., 1981).

Brunel (1986) has shown that there is a correlation between the arcuate shape of the Himalayan chain and mean stretching lineation orientation at the base of the Higher Himalayas implying that there is variation in the direction of transport of the MCT sheet from area to area (see Fig. 1.9, reproduced from his Fig. 18, p. 262, *ibid*). There is one criticism of Brunel's data. In his compilation of stretching lineation data from numerous sources, he did not specify whether or not all these data refer to a single generation of stretching lineation. The present author has found two generations of stretching lineations, particularly in Joshimath area: one dominant set that plunges NNE is earlier, and the other weaker set that plunges North and is found mainly in the lower part of the section has developed postdating a late retrogressive metamorphism. From the average NNE-ward ( $33^\circ \rightarrow N35^\circ E$ ) plunge of the dominant stretching lineation and a SSW-ward sense of overthrust shear indicated by shear indicators found in both the Alaknanda & Dhauliganga traverses, the main direction of emplacement of the MCT-sheet is deduced to be towards SSW. This SSW-ward direction of transport could be taken to be representative of the Central sector (Garhwal-Kumaun region) of the Higher Himalayas. Further discussion on lineations will be given later in Chapters - 3, 4 & 6.

The time relation and contribution of the MCT-emplacement to the development of the inverted metamorphic sequence in the Higher Himalayas is still a point of major controversy (Le Fort, 1975; Brunel, 1986; Searle et al., 1988). Considerable attention will be given in the present study to resolve this controversy (see Chapters - 5, 6 & 7).

#### **1.4.5 Higher Himalayan Uplift**

There is a diversity of cause of uplift along the Himalayas. Except in some parts of the NW Himalayas, uplift of the metamorphic pile in the Higher Himalayas is presumed to have been accomplished by crustal thinning, due to normal faulting (Barnicoat & Treloar, 1989). Large extensional faults that separate the Tibetan Tethys zone from rocks of the Central Crystallines occur in the Everest region (Burg

et al., 1984; Burchfiel & Royden, 1985) and Zanskar (Herren, 1987). These structures are Miocene to Recent in age, and hence may have been active while the Higher Himalayan and Lesser Himalayan rocks were still quite deeply buried. In Zanskar, they are of greenschist and lower grade, and are hence late- to post-metamorphic (Herren, 1987).

On the other hand, uplift of the Nanga Parbat massif in the north-western Himalayas has been shown to have occurred by westward thrusting, accompanied by rapid erosion. No extensional structures have been found in this region, and the area has inherited a pre-Himalayan drainage pattern, which indicates that erosion kept pace with the uplift (Butler & Prior, 1988). This region has very young cooling ages and analysis of these has revealed that the area is still being uplifted (and hence eroded) at rates of up to 7 mm/yr (Zeitler et al., 1982; Zeitler, 1985; Chamberlain et al., 1989).

#### **1.4.6 Some early ideas on the evolution of the Himalayas**

This review would remain incomplete without reference to some of the comparatively early ideas on the orogenic evolution of the Himalayas. In the fifties and sixties, some geoscientists used to draw strong analogy between the Himalayas and the Alps (Hagen, 1959; Krishnaswamy & Swami Nath, 1965; Valdiya, 1969). The evolution of the Himalayas used to be attributed by many workers in the sixties or earlier to the processes of mountain-building caused by the northward movement of the Indian Shield (Wadia, 1931; Gansser, 1964; Pande, 1967; Mithal, 1968). A generally held view was that the Himalayas, before the advent of the Himalayan orogeny, involved two major geosynclines - the southern Lesser Himalayan miogeosyncline and the northern Tethyan eugeosyncline aligned parallel to and separated by the Central Crystalline Axial Zone in the form of a geanticline (Wadia, 1957).

Both Powell & Conaghan (1973) and Le Fort (1975) advocated two distinct stages in the orogenic evolution of the Himalayas, instead of a single continuous and progressive one. The first stage, according to them, refers to the time around the initial continent-continent collision (Late Cretaceous - Early Tertiary), whereas the second stage refers to deep crustal fracturing and subsequent intracontinental subduction or underthrusting (Mid-Tertiary to Miocene and onward).

Le Fort (1975) gave some remarks on paleogeography of the Himalayan region. Distinguishing between the Lesser Himalayan sedimentary realm and the Tethyan sedimentary realm with the divide in between of crystalline Higher Himalayas it has been noted that the Tethyan area represents 'a relatively homogeneous marine paleogeographic domain' dominated by 'detrital and platform carbonate lithofacies', probably deposited on "a thinned margin along the edge of a continent widely open on the Tethys", while the Lesser Himalayan area is made up of rocks that are mainly deposited in shallow water, tidal, lagoonal or even continental conditions along with some turbiditic flysch in several basins more or less parallel to the present Himalayan trend (see also Gansser, 1964; Kummel & Teichert, 1970; Colchen, 1974; Valdiya, 1970; Frank & Fuchs, 1970; Jain et al., 1971; Bhargava, 1972).

#### **1.4.7 Works closely related to the areas of the present study**

There have been no detailed modern structural and metamorphic studies on the areas of the present study. There is, for that matter, a general lack of detailed work on the MCT-Zone as a whole, as already emphasized.

Geological information of general interest on Joshimath area are found in Valdiya (1980, 1981, 1988), Viridi (1986), Pati & Rao (1983) etc. Valdiya's memoir (1980) is a valuable work and gives a good portrayal of the regional geological setting of the Lesser Garhwal-Kumaun Himalayas. Roy & Valdiya (1988) attempted some modern studies on tectonometamorphic relations in the Higher Garhwal Himalayas. Paul Metcalfe completed his doctoral thesis (Metcalfe, 1990) mainly working on metamorphism and cooling-uplift history of rocks from the MCT-Zone and above in the Bhagirathi valley section in western Garhwal.

As far as the Sobala area or, broadly, the eastern Kumaun region, is concerned, basic reviews on general geology are given in Gansser (1964), Powar (1972), Thakur & Chowdhury (1983), Valdiya & Gupta (1972), Fuchs & Sinha (1978) and Sinha (1981).

### **1.5 LAY-OUT OF THE THESIS**

The main textual part of the thesis consists of seven chapters. This is followed by bibliography and the appendices respectively. The investigation for this

thesis concentrated largely on the Joshimath area because it exposes a very representative section across the MCT-Zone, while the Sobala area was looked into mainly for the purpose of comparison.

This first chapter (Introduction) in the thesis is followed, in turn, by the under-noted ones:

Chapter - 2 (Tectonostratigraphy and Lithology) which gives a tectonostratigraphical and lithological summary of the study areas in the light of the regional geological setting of the Central Himalayas.

Chapter - 3 (Structures and Deformation) highlights the aspects of mainly the mesoscopic and microscopic secondary deformational structures and establishes the sequence of deformation episodes.

Chapter - 4 (The Major Thrusts) deals with the position, nature etc. of the two boundaries of the MCT-Zone i.e. Munsiri Formation and lists the characteristic indicators found for the determination of shear sense during the MCT emplacement.

Chapter - 5 (Textures and Metamorphic History) discusses at length the textural and microstructural aspects of the metamorphic minerals that have direct bearing on the deformation-metamorphism relationships, and establishes the sequence of metamorphic episodes in terms of the deformational episodes (i.e. fold episodes).

Chapter - 6 (Summary of results and Discussion) summarises the main observations from the present study and discusses their implications. Particular attention is given to modelling the Higher Himalayan inverted metamorphic sequence.

Chapter - 7 (Conclusions) forms the final chapter of the thesis. In addition to giving the main conclusions arrived at, this last chapter also provides a list of suggested follow-up work.

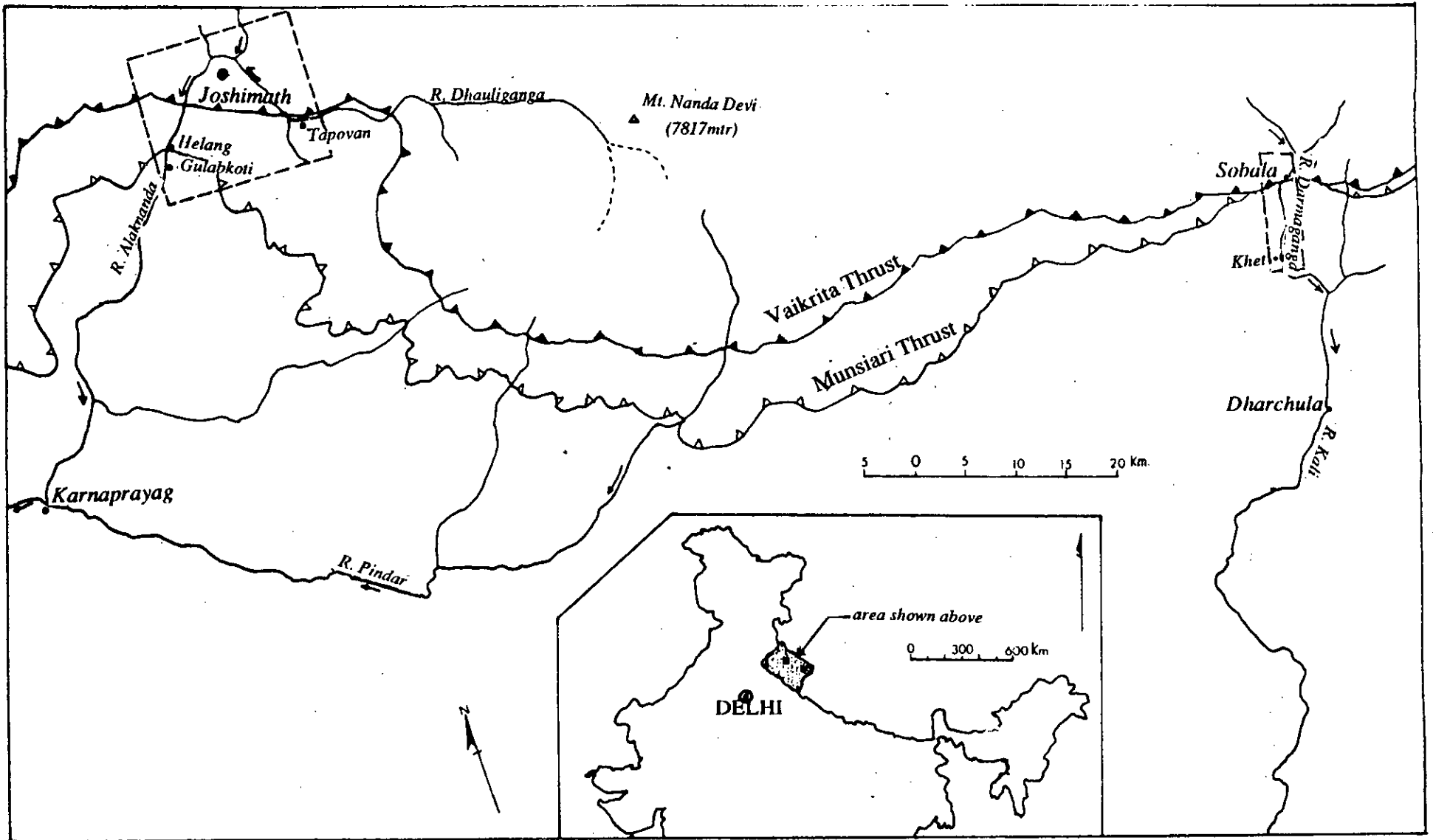
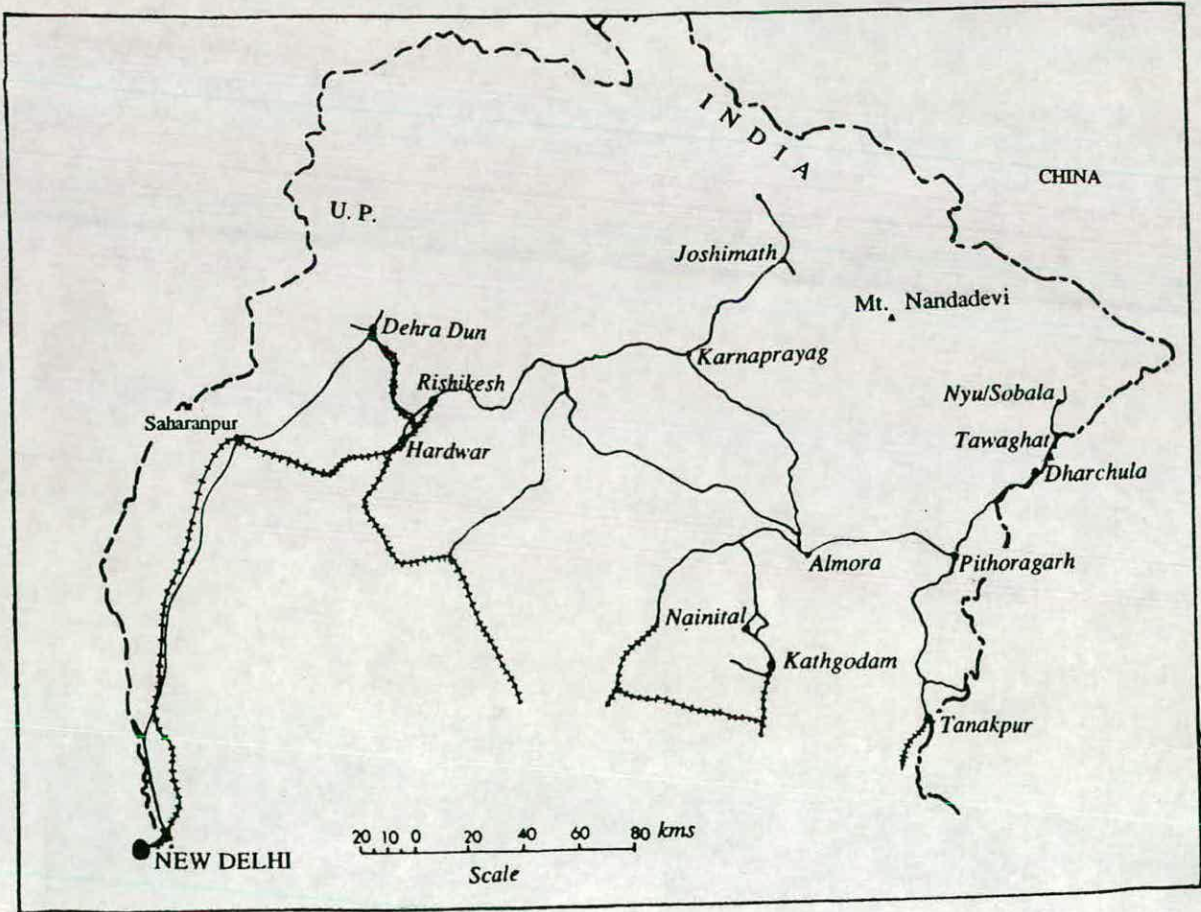


Fig. 1.1: Location of the two study areas. Inset shows approximate position of the Garhwal-Kumaun Himalayas in a regional context.



**Fig. 1.2:** Mt. Nanda Devi (7817 mtr) as seen from Ghotsu, south of Auli. The peak, about 40km ESE of Joshimath, is at far centre of the photograph. River Dhauliganga is seen in the left foreground. Munsiri Formation forms the lower part of the valley extending upto the road above the river. Above this level, the mountain ranges are formed of Vaikrita Crystallines; whereas the top of the Nanda Devi peak is made up of Tethyan sediments.



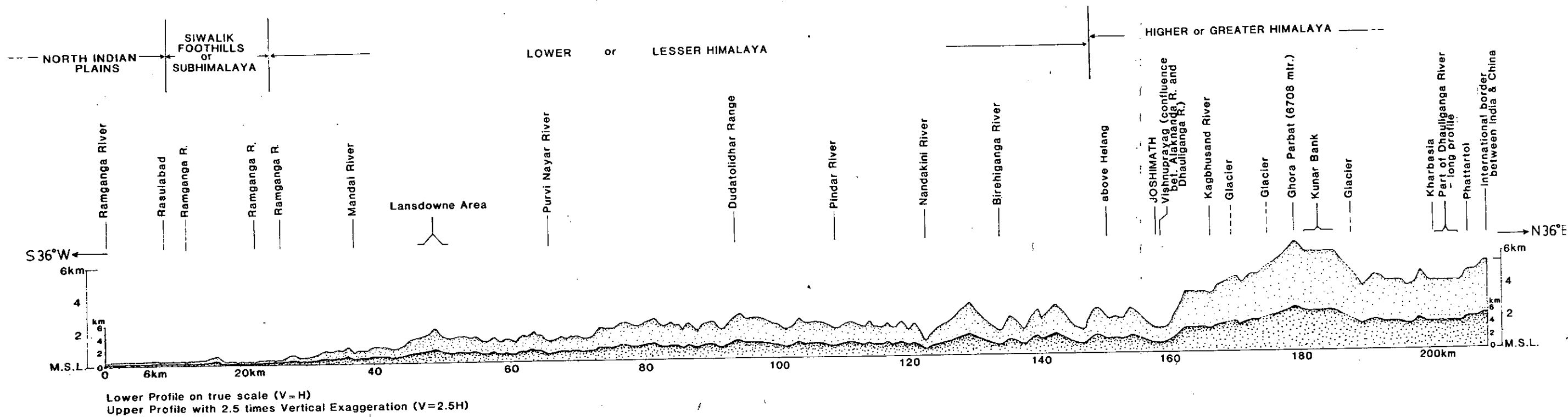
**Fig. 1.3:** A route-map for the study areas.



**Fig. 1.4(a):** Progressive southward lowering of height of the mountain ranges across the Lesser Himalaya. Photo taken looking south from slightly above the surface trace of the Vaikrita thrust (i.e. just above the MCT-zone), NE of Shelang village.



**(b):** Looking N from roughly the same spot as above, the snow-clad Higher Himalayan ranges. The MCT-zone runs approximately along the physiographic break between the Higher Himalayas and the Lesser Himalayas.



**Fig. 1.5:** A topographic profile across the Indian part of the Himalayas. The lower profile is in true scale, while the upper profile has exaggerated vertical scale. Based on contour data collected from Survey of India topographic maps of 1 : 250 000 scale.

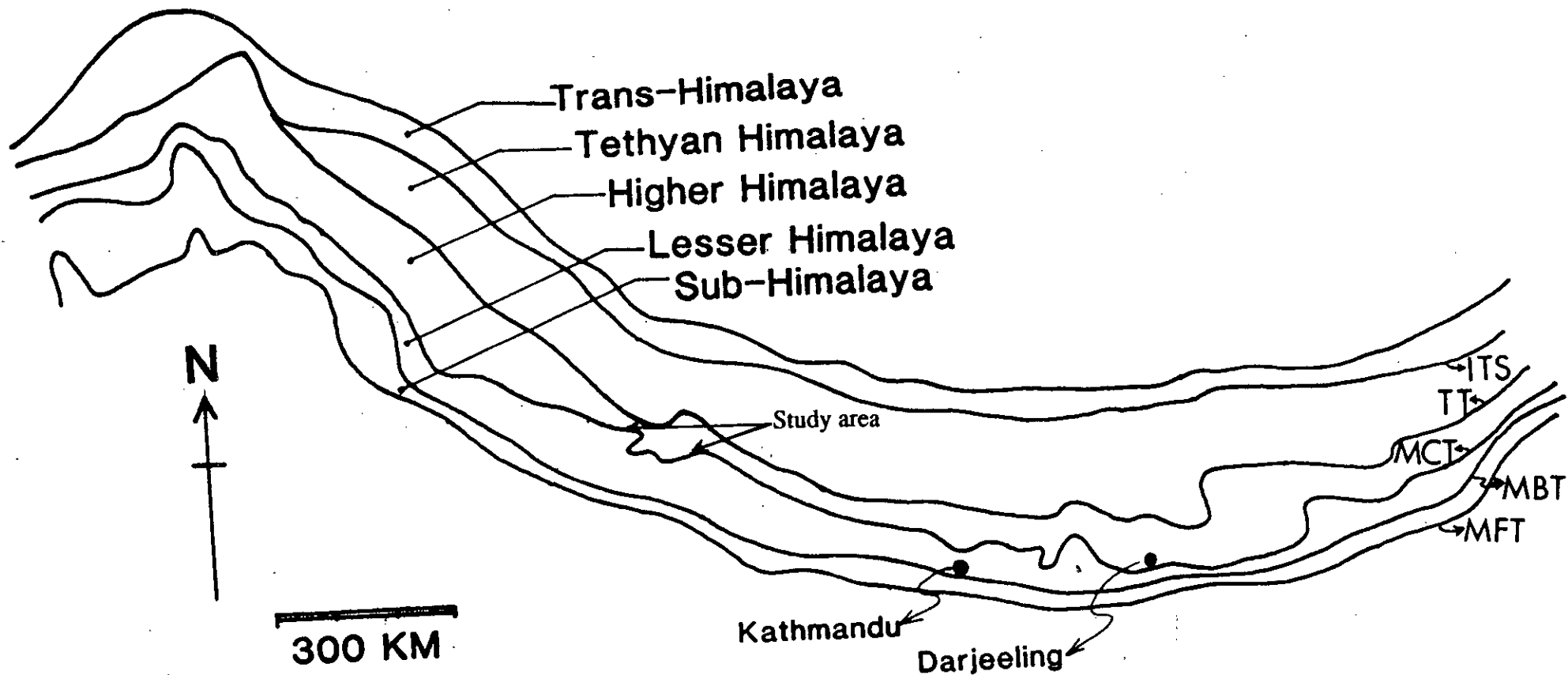


Fig. 1.6: Longitudinal zonation of the Himalayas. Transhimalaya is bounded to the north by the Banggong/Sbyok Suture. ITS = Indus-Tsangpo Suture, TT = Tethyan Thrust, MCT = Main Central Thrust, MBT = Main Boundary Thrust, MFT = Main Frontal Thrust. After, Barnicoat & Treloar, 1989.

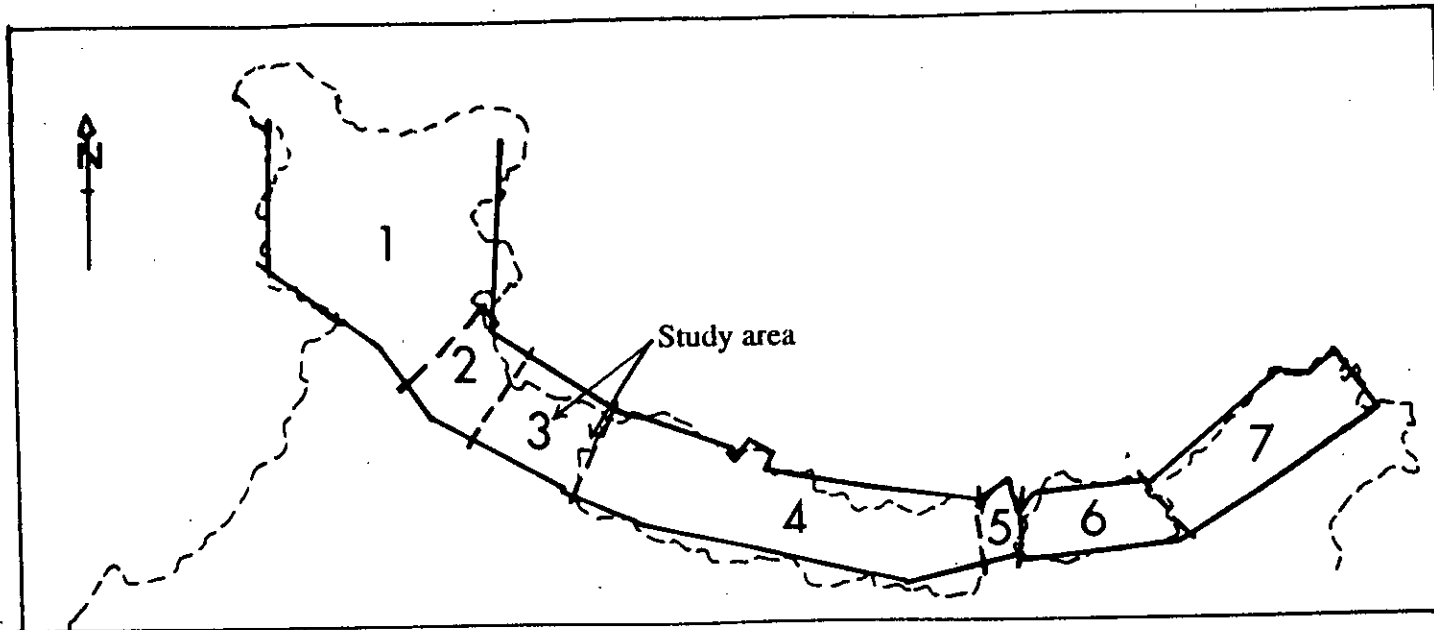
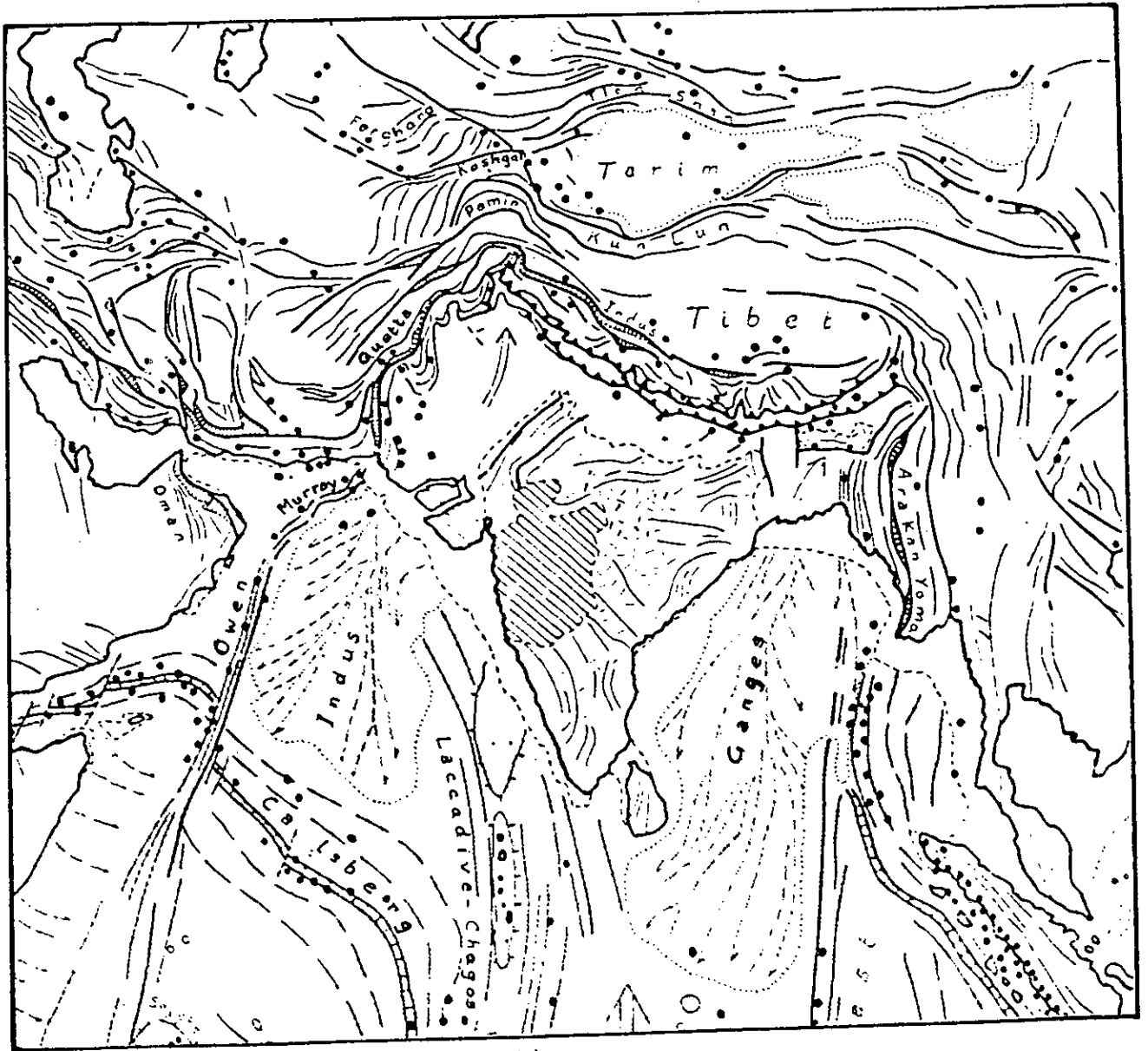
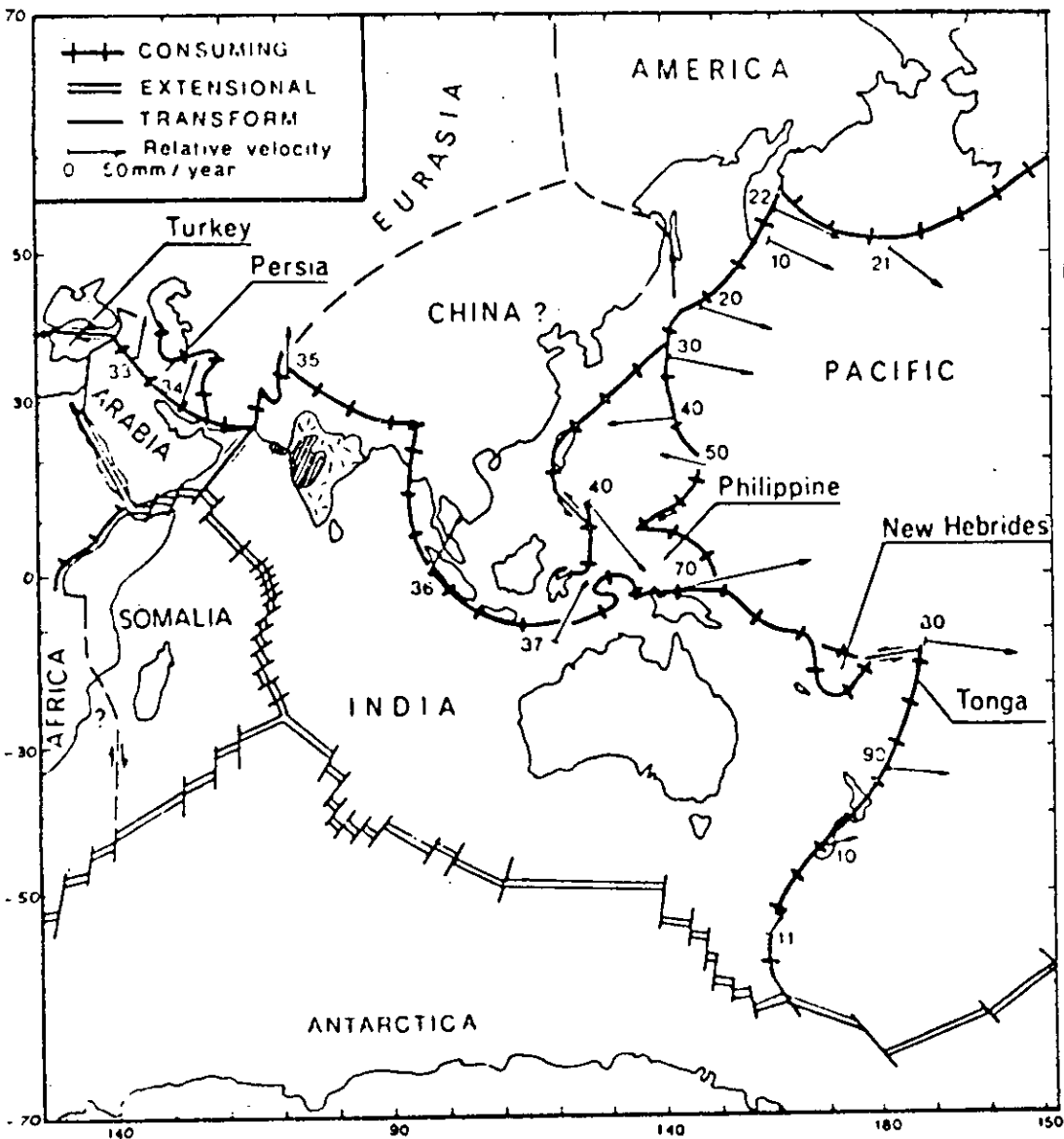


Fig. 1.7: Transverse sub-divisions of the Himalayas.

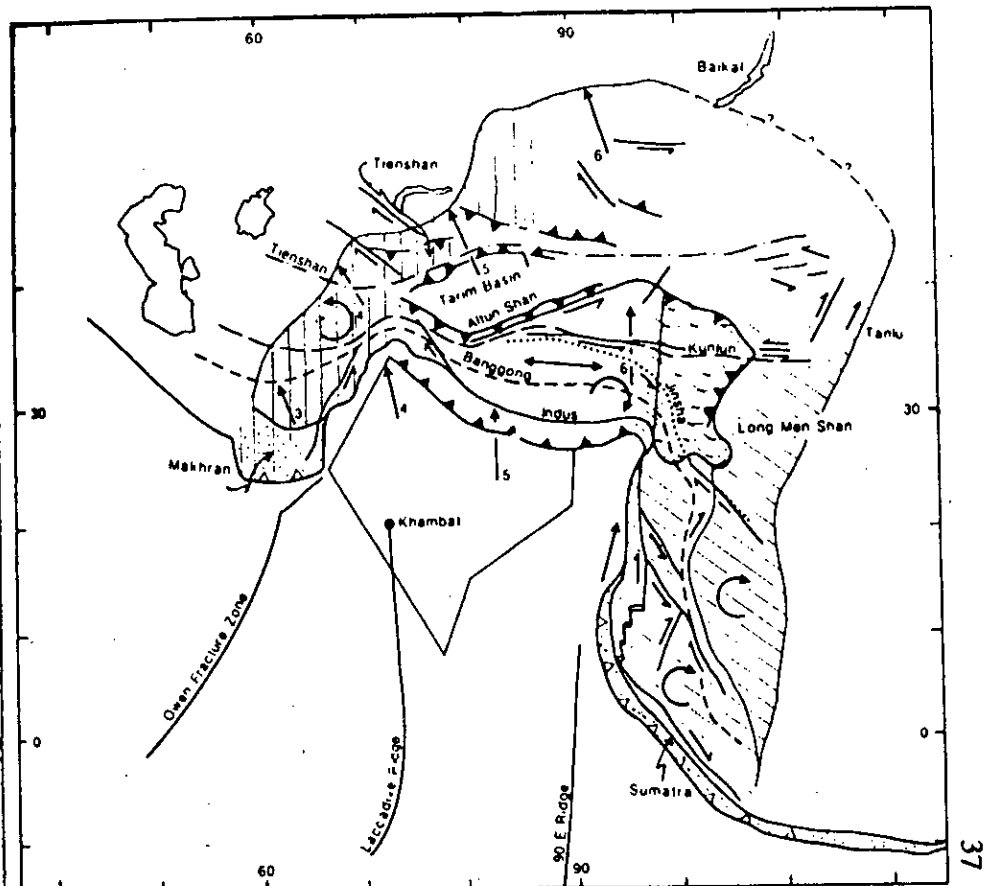


(a)

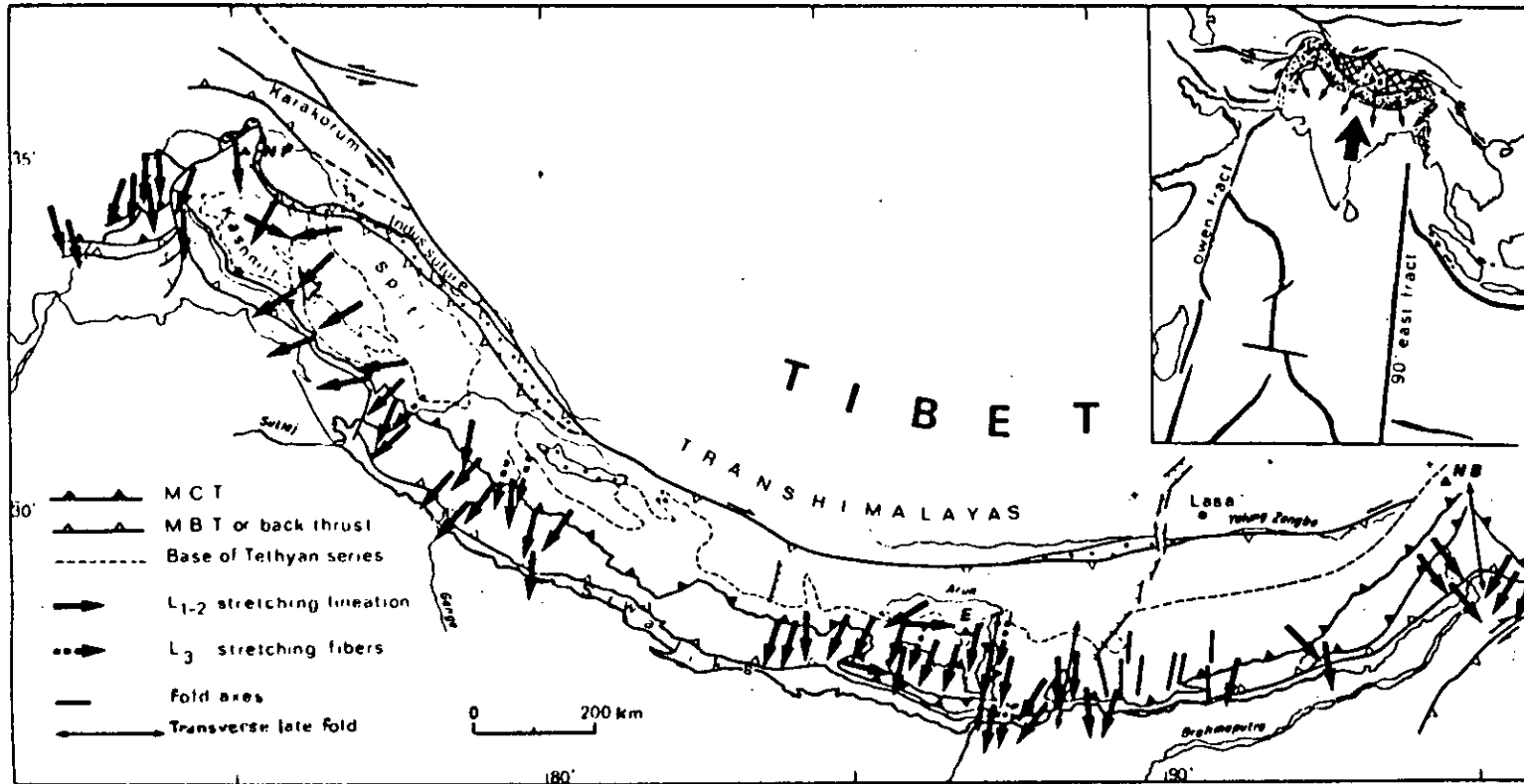
**Fig. 1.8:** (a) Physiographic milieu around the Himalayas and India (after Gansser, 1966); (b) Regional plate tectonic setting around India (after Le Fort, 1975); (c) Major tectonic zones in the India-Eurasia convergent plate boundary complex (after Dewey et al, 1989). See next page for (b) and (c).



(b)



(c)



**Fig. 1.9:** Varying orientation of stretching lineation along the Himalayan Belt. Thrust transport direction varies due to the arcuate shape of the Himalayas (i.e. the collisional plate boundary zone). After, Brunel (1986).

## TECTONOSTRATIGRAPHY AND LITHOLOGY

### 2.1 INTRODUCTION

The most crucial problem in any geological study of regional nature in the Himalayas is posed by the lack of our accurate understanding about the stratigraphy and correlation of the Himalayan rocks. This problem stems mainly from the fact that barring a few, most of the Himalayan formations are devoid of fossils. Despite this, through his sustained efforts for over three decades Prof. K.S. Valdiya of Kumaun University, Nainital (India) was able to systematise to a significant extent the stratigraphic correlation in the central sector of the Himalayas (i.e. Garhwal-Kumaun region which is commonly known as Kumaun Himalaya) in particular and for the Lesser Himalayan domain of the entire mountain chain in general. However, pending thorough confirmation from later studies by other workers, the scheme established by Prof. Valdiya should be taken, strictly speaking, as a reflection of his personal understanding of the geology of the Kumaun Himalaya. There is no scope in the present thesis to go into the details of stratigraphical arguments, and the tectonic and stratigraphic subdivision suggested by Valdiya (1980, 1981, 1988) will be followed for the purpose of this work.

The areas of the present study lie at the junction between the Higher and Lesser Himalayas. In order to appreciate better the position of the study areas with reference to the regional geological setting it would be useful to have a brief summary on the tectonostratigraphy and lithology of various geological units of the Central Himalayas.

This chapter will discuss the tectonostratigraphic subdivision of the study area and the distribution of lithology in those tectonostratigraphic units in the light of the regional geological setting of the Central Himalayas (Kumaun Himalaya). In addition, the possible paleogeographic pattern across the MCT-Zone will be briefly discussed. A discussion on a representative selection of radiometric age data available in the literature from across the Central Himalayas will also be given, followed by a note of guidance at the end on the usage of some lithological and tectonostratigraphical terms in the thesis.

## 2.2 GEOLOGICAL SETTING OF THE KUMAUN HIMALAYA

The regional geological setting of the Kumaun Himalaya is discussed below giving the lithotectonostratigraphic succession along with a brief description of most of the groups and formations within them. The description will be given following the order of structural packages i.e. the sequence of nappes, from top to bottom. The objective is to set the scene for putting the findings from the present study in a clear perspective.

The Kumaun Himalaya has seven major lithotectonic units -- (i) Tethyan unit, (ii) Vaikrita unit, (iii) Munsiri/Almora unit, (iv) Ramgarh unit, (v) Krol-Berinag unit, (vi) 'Neo-autochthonous'/para-autochthonous' unit and (vii) Siwalik unit (Valdiya, 1980). Except the last two, each of the other five units derives its name from the immediate underlying thrust that carries it. The Tethyan unit and part of the Vaikrita unit fall within the domain of the Higher Himalayas, and the Siwalik unit in the Sub-Himalaya, whereas the Kumaun Lesser Himalayan domain encompasses the Vaikrita unit (part), Munsiri/Almora unit, Ramgarh unit, Krol-Berinag unit and the 'Neo-autochthonous'/para-autochthonous' unit. Fig. 2.1 shows the location of these units in the Lesser Himalaya of Garhwal-Kumaun region (note that the Krol-Berinag unit is subdivided into two in the map, and to emphasize correlation the unit-names that are in vogue in the adjacent Himachal Himalaya are also shown as prefix in the legend). For the mutual disposition in a generalised transverse cross section of all the seven lithotectonic units in the Kumaun Himalaya, see Fig. 2.2. Note that this generalised cross section (Fig. 2.2) is roughly along a line passing through Mt Nanda Devi which is some 40 km ESE of Joshimath; in the figure the Malari Thrust (a local name) stands for the Tethyan Thrust, and each major thrust marks the lower boundary of the lithotectonic unit named after it.

Table - 2.1 below gives further details of the lithotectonostratigraphic succession of the Higher and Lesser Kumaun Himalayas with emphasis on the Lesser Himalayan section. The Sub-Himalaya is excluded from discussion, so no mention of the Main Boundary Thrust (MBT), Main Frontal Thrust (MFT) or the Siwaliks is included in Table 2.1.

**Table - 2.1**

Lithotectonostratigraphic succession in the Kumaun Himalaya (after Valdiya, 1980)

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Indus-Tsangpo Suture Zone

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	TEJAM GROUP		
[Forming the 'Neo-autochthonous/ Para-autochthonous zone']		Deoban Formation	Up-Mid Riphean
		Rautgara Formation	Lr-Mid Riphean
	DAMTHA GROUP	Chakrata Formation	Lower Riphean

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The Vaikrita/Central Crystalline Nappe and the Almora-Munsiari-(Jutogh) Nappe together could be considered to form a combined 'Himalayan Crystalline Nappe', so that the former represents the 'Upper Crystalline Nappe' and the latter the 'Lower Crystalline Nappe' (Fuchs & Frank, 1970; Fuchs, 1975). Description of the Tethyan rocks is not included below as the Tethyan Group has rather little relevance to any discussion on the geological setting of the Lesser Himalayan Nappe sequence. For concise details on this group, see Sinha (1981; particularly his Table-II, Fig.-8 and related text). Another point worth emphasizing is that because in Joshimath area the Chandpur Formation is absent as the footwall to the MCT-Zone (Munsiari Formation) and the Berinag Formation is in tectonic juxtaposition with the Mandhali Formation or even with the Deoban Formation of the underlying Tejam Group, and the juxtaposed Berinag and Mandhali formations occur near the base of the Higher Himalayas, I shall refer to this immediate footwall to the MCT-Zone as 'Berinag-Mandhali formations'. They belong to the so-called "Garhwal Group" (Jain, 1971; Gaur et al., 1977a & b; Agarwal & Kumar, 1980; <sup>equivalent</sup> to 'Garhwal Series' of Auden, 1949).

Bounded by the Main Central Thrust Zone (MCT-Zone) to the north and the Main Boundary Thrust (MBT) to the south the Kumaun Lesser Himalaya is built up of a number of thrust sheets (nappes) resting on the foundation of a 'neo-autochthonous'/para-autochthonous' unit exposed in the tectonic windows. Each of these thrust sheets has fairly distinctive lithology, structural pattern and magmatic history (Valdiya, 1980).

The outer Lesser Himalaya (Nag Tibba Range) adjacent to the Siwalik subprovince comprises a succession of three nappes piled one on top of the other. The lowermost nappe is made up of a large thickness of possibly Late Precambrian - Early Paleozoic sediments (Mandhali Formation, and Jaunsar and Mussoorie groups). The intermediate nappe consists of mildly metamorphosed flyschoid sediments with large volumes of early Precambrian granitic porphyroid intrusives belonging to the Ramgarh Group. The uppermost sheet is made up of possibly late Precambrian

moderately metamorphosed sediments and granitic - granodioritic augen gneisses belonging to the Almora Group.

The inner Lesser Himalaya is largely embraced by the neo-autochthonous Precambrian sedimentaries (Damtha and Tejam groups), covered by two thrust sheets -- the lower one is constituted of quartzites and basic volcanics (Berinag Formation), while the upper sheet is represented by the detached pieces or klippen of the same crystalline nappe that caps the Dudhatoli-Ranikhet-Champawat Range of the outer Lesser Himalaya. This crystalline nappe is made up of the Almora Group (Munsiari Formation) of rocks and is rooted at the base of the Higher Himalaya, the root-zone being delimited at the bottom by the Munsiari Thrust and at the top by the Vaikrita Thrust. Fig. 2.3 schematically shows the geometry of the Almora thrust sheet over the Lesser Himalaya, as envisaged by Valdiya (1980) and now commonly accepted by others. Different klippen belonging to the Almora Nappe are clearly depicted in this figure. The "root-zone" of the Almora Nappe, specifically designated by Valdiya as the 'Munsiari Formation' represents the MCT shear zone.

Now a summary is given mainly following the account of Valdiya (1980) on the salient features of the different geological units of the Kumaun Lesser Himalayan Nappe sequence. For the location of different rivers and localities mentioned below, please refer to the geological map (plate) in colour in Valdiya (1980).

### **2.2.1 Vaikrita Group**

The Vaikrita Group consists of a high-grade assemblage of metamorphic rocks commonly intruded by young Tertiary granitoid intrusives, concordant as well as discordant. These rocks were originally referred to as 'Vaikrita system' by Greisbach (1891) after the Sanskrit word meaning "transformed vicareously". Heim & Gansser (1939) included these rocks as an integral part of their 'Central Crystalline Zone'. Attempting lithological subdivision of these "Central Crystallines", Valdiya (1973) separated the lower unit of low-grade metamorphics that constitute the Munsiari Formation from the group of high-grade metamorphics. Four lithological units have been recognised in the Vaikrita Group viz. (from top to bottom) Budhi Schist, Pindari Formation, Pandukeshwar Formation and Joshimath Gneiss Formation (see Table-2.1 above). Valdiya (1980), Valdiya & Goel (1983) and Roy & Valdiya (1988) gave the following lithological descriptions of the four formations in the Vaikrita Group:

### **Budhi Schist**

This is mainly a porphyroblastic biotite-bearing calc-schist interbedded with micaceous schist, with phyllites toward the top. Locally, carbonaceous, pyritic, or staurolite-bearing horizons are also present.

### **Pindari Formation**

Predominantly banded calc-silicate gneiss and marble interbedded with subordinate biotite-psammitic gneisses and schists with or without kyanite and sillimanite. This part of the Vaikrita Group is extensively penetrated by a network of dykes and veins of aplite, pegmatite and granite. Pervasive penetration and invasion of granite plutons (e.g. Badrinath Granite) has brought about widespread migmatisation and development of porphyroblastic augen gneisses, characterised by sillimanite and garnet. In the Malari-Badrinath-Kedamath-Gangotri belt the tourmaline-bearing leucocratic granites have assumed huge batholithic dimension. In the Pindari Formation nebulitic types of pegmatitic horizons are common.

### **Pandukeshwar Formation**

Biotite- and/or muscovite-rich quartzite is intercalated with kyanite-garnet-bearing mica schists and psammitic gneiss. Locally, lenses and subordinate layers of calc-silicate gneisses and garnet-bearing amphibolite are present. Where garnet has developed in abundance the quartzite resembles a light-coloured very high-grade metamorphic rock.

### **Joshimath Gneiss Formation**

In this study the attention is focussed on the lowermost unit of the Vaikrita Group i.e. the Joshimath Gneiss Formation. In the thesis, it will be referred to simply as Joshimath Gneiss, which consists essentially of streaky and banded gneiss and garnet-kyanite-rich biotite schists with phyllonites at the base. Very subordinate and local lenses of calc-silicate gneiss are also present. A cursory reconnaissance field-traverse from Vishnuprayag to Badrinath following the road-section suggests that the Pindari Formation has lithological affinity with the Joshimath Formation and therefore could be a repetition due to large-scale southvergent overturned folding (the dominant dip of the main foliation in all these formations is towards NE/NNE). Valdiya's lithological subdivision of the Vaikrita Group did not take into consideration the aspect of repetition due to large-scale folding and, therefore, the compartmentalisation of the Vaikrita Group into four formations is more of working (i.e. practical) value rather than strictly geological.

Even though there is no clear-cut structural or geomorphic evidence, Valdiya (1980) placed a thrust-contact below the Joshimath Gneiss Formation (Vaikrita Group) mainly on the ground of 'abrupt change in lithology and grade of metamorphism' and from a regional tectonic consideration. For example, in the Nepal Himalaya, the French and Japanese workers recognised the tectonic contact at the corresponding level as the M.C.T. For further discussion on this boundary above the Munsiri Formation and the one underlying it, see later (Chapter 4, The Major Thrusts).

The 'Tibetan Slab' (Dalle du Tibet) defined by Lombard (1958) is most probably the extension of the Vaikrita Group into Nepal. Lithologically and tectonically the two are alike. Hagen's (1958) 'Khumbu Nappes' are possibly comparable with the Vaikrita. In Eastern Himalaya the sillimanite-kyanite-bearing gneisses of the 'Darjeeling Series' (Ray, 1947) are analogous to the Joshimath Gneiss of the Vaikrita Group. In Bhutan the extension of the Darjeeling Gneiss is known as 'Thimpu Formation' (Nautiyal et al., 1964) or as 'Chasilakha-Takhtsang Gneisses' (Gansser, 1964). In the Kameng District of Arunachal Pradesh the unit has been described as the 'Sela Group' (Das et al., 1975). Valdiya's investigation in the Satluj valley in Himachal Himalaya has demonstrated northwestward extension towards Spiti-Kinnaur-Lahul of all his four units of the Vaikrita Group.

As they stratigraphically underlie the well-dated 'Cambrian-to-Eocene' sedimentary succession of the Tethys Himalaya, the Vaikritas are almost certainly Precambrian; but it is difficult to ascertain to which level of Precambrian the Vaikritas belong (Gansser, 1964; Le Fort, 1975; Valdiya, 1980; Windley, 1983).

### 2.2.2 Almora Group

The Almora Group was referred to as the 'Crystalline Zone of Almora' by Heim & Gansser (1939). This thick overthrust sequence of a variety of schists, micaceous quartzites, amphibolites, quartzofeldspathic protomylonites-mylonites and phyllonites is preserved in longitudinally trending synformal structures. It belongs to the lower to middle amphibolite facies of regional metamorphism and contains concordantly emplaced plutonic bodies of granodiorites and granites. It is named after the town of Almora in Kumaun. The three commonly used terms - 'Almora Group', 'Almora Nappe' & 'Almora klippe' need to be unambiguously defined. "Almora Group" is a lithostratigraphic term that refers to the group of rocks as mentioned

above making up the Almora Nappe, while "Almora Nappe" is a structural/tectonic term implying the long-travelled thrust sheet made up of Almora Group rocks. The Almora Nappe is not exposed as a single continuous thrust sheet, but in detached or separate outcrops i.e. in various klippen across the Lesser Himalayas. The "Almora Klippe" is the largest one of such klippen cropping out in and around the Almora township/district (see geological map in colour in Valdiya, 1980).

Petrographic study of the Almora metamorphics started with Middlemiss (1887, 1888, 1890). Later studies by Valdiya (1962, 1963, 1980), Misra & Sharma (1967), Vashi & Merh (1965), Vashi & Laghate (1972), Das (1971, 1974), Desai (1973), Misra & Sharma (1973), Das (1969, 1973) etc. have provided fairly detailed accounts of the petrographic characters and grade and history of metamorphism of the rocks of the Almora Group. Essentially a tectonostratigraphic rather than purely stratigraphic unit, it is divisible into three units: (1) the 'Saryu Formation' forms the lower part in the northern flank in the type area, (2) the batholithic sill of 'Champawat Granodiorite' emplaced in the southern flank, and its extension as a sill in the Nill-Chhira-Panar belt in the northern flank, and (3) the 'Gumalikhhet Formation' forming the upper part. The 'Munsiari Formation' is the part proximal to the "root" of the Almora Nappe and is bound by the Munsiari and Vaikrita thrusts at the base and top respectively. Between the overthrust Almora Group and its "root" towards the north there is a chain of klippen, representing the detached pieces of presumably a once continuous thrust sheet. These klippen are known as the crystalline zones of Nandprayag, Baijnath, Dharamghar, Askot and Chhiplakot (see Valdiya, 1980). To the southwest of the Almora Nappe there is yet another large klippe known as the Amri, forming the Lansdowne Hills, and its detached pieces at Satengal and Banali in the Mussoorie Hills (Auden, 1937). These klippen are made predominantly of the lithological elements of the Saryu Formation and the metamorphosed equivalents of the Champawat Granodiorite.

The lower limit of the Almora Group, that is the base of the Saryu Formation, is defined by the Almora Thrust, the North Almora Thrust in the northern flank and South Almora Thrust in the southern. The North Almora Thrust is a very well-defined tectonic line of division, separating the "autochthonous" sedimentaries below from the metamorphic and granitic rocks above. Mylonitisation of granitic rocks and augen gneisses, and the development of a persistent band of chlorite-sericite-phyllonite along the thrust plane marks out the plane of separation. The extension of the Almora Thrust defines the base of the crystalline formation of the various klippe mentioned above. In the "root-zone" at the base of the Great Himalaya, the Munsiari Thrust,

whose extension is the Almora Thrust, separates the sedimentaries (Berinag/Mandhali/Deoban) from the crystallines of the Munsiri Formation at the base of the Great Himalaya.

Because the Almora Nappe represents the uppermost of the existing thrust sheets in the Lesser Himalaya, the Saryu Formation demarcates the upper limit of the Almora Group and its extensions in the various klippen. But the situation is quite different in the "root-zone", where the Munsiri rocks are succeeded by another group of higher-grade metamorphic rocks (Vaikritas), the plane of separation being recognised as the Vaikrita Thrust.

### Saryu Formation

This lithological unit is named after the river in eastern Kumaun along which its typical sections are exposed. It comprises chlorite-sericite schist, often phyllonitic at the base, followed higher up by garnetiferous muscovite-schists alternating with micaceous quartzites. In the basal part in the northern flank of the Almora synformal klippe there are bands of strongly mylonitised quartz-porphyry and ultramylonite within chloritic phyllonite, as seen between Ghat and Saryu-Panar confluence. Towards the upper part of the Saryu Formation there is a chain of lenticular bodies or sills of porphyritic granite grading marginally into augen gneiss. In the southern flank the Saryu Formation has been much overshadowed by the emplacement of the huge sill of the Champawat Granodiorite so that between the rivers Kali and Ladhiya, by Mornaula, the Saryu Formation is represented by augen gneisses interbedded with biotite-rich mica schists and micaceous flaggy quartzites. However, NW of Mornaula, where the Champawat pluton is in the form of thin lenticular sills of granite or augen gneiss, the schists and flaggy quartzites assume dominance (Valdiya, 1980).

### Champawat Granodiorite

Champawat Granodiorite is a composite batholithic body concordantly emplaced towards the upper part of the Saryu Formation. It is constituted predominantly of granodiorite, composition often varying from tonalite (quartz-diorite) to adamellite (quartz-monzonite). An early synkinematic suite of trondhjemitic rocks, but conspicuously rich in biotite & quartz, is intruded by a younger set of post-kinematic biotite-poor leucocratic adamellite and tourmaline-bearing aplitic or pegmatitic granite. The main body is massive non-foliated coarse-grained, equigranular to locally porphyritic, in the central part and becoming progressively foliated or gneissose towards the margins. Although the northern border

is sharp, being overlain rather abruptly by chloritic phyllite of the Gumalikheth Formation, the southern border is transitional, the gneissose granite giving way imperceptibly to augen gneiss intercalated with biotite schists and micaceous quartzites (Valdiya, 1980).

### **Gumalikheth Formation**

This upper unit of the Almora Group comprises black carbonaceous phyllites alternating with black, fine-grained biotite-rich greywacke. Northwestwards, the carbonaceous phyllite becomes graphitic schist and the greywacke is converted into biotite-rich schist. In the same direction (beyond Kosi River; for location, see geological map in colour in Valdiya, 1980) garnetiferous mica schists and micaceous flaggy quartzites predominate over the black rocks and the Gumalikheth becomes indistinguishable from the Saryu.

### **Munsiari Formation**

Of the huge pile of metamorphics and gneisses of the Higher Himalaya, the lower part exhibiting relatively low-grade metamorphism in greenschist to almandine-amphibolite facies has been distinguished by Valdiya (1973a) as the Munsiari Formation. The Munsiari succession is characterised by a high proportion of mylonitic rocks and related brittle-ductile deformation. Presence of a strong and pervasive stretching lineation and of well-defined shear structures associated particularly with the mylonites and granitoid augen gneisses indicates a high degree of shearing deformation. Effects of consequent retrograde metamorphism are also highly prevalent.

Petrologically the Munsiari is similar to the Almora and its remnant in the klippen of Askot-Bajjnath-Nandprayag and Chhiplakot. This fact has led Valdiya to consider the Munsiari Formation as representing the "root-zone" of the Almora Nappe. The lower limit of the Munsiari Formation is defined by the Munsiari Thrust, and the upper limit is marked by the Vaikrita Thrust across which there are abrupt changes of the lithology and grade of metamorphism. However, in practical terms, these tectonic breaks are quite difficult to recognise in the field and no straightforward geomorphological criteria are applicable in these rejuvenated youthful mountains of the Great Himalayas. The Munsiari Formation comprises a relatively greater volume of granitic rocks (augen gneisses) than the Almora Group in its type area. Commonly the augen-sizes reach up to 15 cm across, for example, in the Darmaganga valley along Khet-Pangu belt, or in the Tons valley north of Mori (see the geological map in colour in Valdiya). The typical mylonitic-protomylonitic rocks

are normally quartzofeldspathic; the porphyroclastic grains in them are mostly made up of plagioclase, alkali feldspar and/or quartz. In addition to being reflected by increased frequency of ribboning of feldspar grains and/or quartzofeldspathic aggregates, progressive mylonitisation (i.e. grainsize reduction) down the section across the Munsiri Formation is occasionally reflected also through advanced myrmekitisation whereby original bigger alkali feldspar grains in the quartzofeldspathic protomylonitic-mylonitic horizons are transformed into tiny recrystallised quartz and plagioclase grains. Valdiya (1980) considers that there are two suites of gneisses in the Munsiri Formation; the synkinematically emplaced biotite-quartz rich granodioritic gneisses are intruded by post-tectonic strikingly leucocratic tourmaline-bearing granite, adamellite and aplite. The other metamorphics in the Munsiri Formation include phyllonite, chlorite-sericite schists, garnetiferous mica schists alternating with micaceous quartzite, graphitic/carbonaceous schists locally interbedded with blue-black marble and amphibolite. Traverses along the valleys of Goriganga, Alaknanda, Dhauliganga, Mandakini, Bhagirathi, Yamuna and Tons give a fairly good idea of the lithological succession of the Munsiri Formation (for location of these rivers, see Fig. 1.8 in Valdiya, 1980). Two sections as described by Valdiya (1980, p.75) are reproduced below:

I. Girgaon-Lilam section (Type area), Goriganga valley.

Base : Berinag quartzites and Mandhali slates.

----- Munsiri Thrust -----

- (a) Mylonitised biotite-quartzite, paragneiss, sheared amphibolite and biotite-sericite-chlorite schist with lenses of white quartzite. The schist is locally garnetiferous. (Girgaon to Kalamuni, and in Ramganga valley near Satgarh)
- (b) Biotite-rich augen gneiss exhibiting flaser-structure, and streakiness resulting from shearing and fragmentation of phenocrysts. Bands of biotite-quartz paragneiss and sheared amphibolite. Intruded by veins, sills and schlierens of medium-grained leucocratic aplogranite, showing pygmatic folding such as seen at Dummar. (Kalamuni through Munsiri to 1 km S of Lilam; in Ramganga valley from N by Satgarh through Kethi to Siuni and beyond).

----- Vaikrita Thrust -----

Top : Kyanite-gneiss of Joshimath Formation

II. Pana-Kunwarikhal-Tapovan-Reni section (Dhauliganga valley).

Base : Berinag quartzites.

----- Munsiri Thrust -----

- (a) Chlorite-sericite schist, which in the middle of the succession is garnetiferous. Interbedded with a horizon of gneissose quartz porphyry and several bands of amphibolite grading into chlorite schist and two horizons of blue flaggy marble. (N by Pana to Kunwarikhal).
- (b) A large body of gneissose granite porphyry grading into augen gneiss overlain by pyritous chlorite-sericite and biotite-chlorite schist and black biotite-rich fine-grained quartzite. Higher up near Dhak Gad another band of augen gneiss. Intruded by leucocratic adamellite and aplite. (Kunwarikhal to Tapovan).
- (c) Sheared and shattered (to powdery state) finer-grained quartzite (non-sericitic) of grey, brown and white or light green colours. Interbedded with very subordinate intercalations of sericite schist, locally garnetiferous. (Tapovan to 3 km N of Tapovan bridge).
- (d) Pyritous black schist interbedded with marble giving way upwards to alternations of fine-grained black biotite-rich quartzite and pyritic biotite schist. In the Rishiganga valley brown and grey garnetiferous sericite schist characterised by pyrite and limonite. Interbedded with subordinate grey quartzite. (From 3 km N of Tapovan to Reni).

----- Vaikrita Thrust -----

Top : Joshimath Formation.

In terms of regional correlation with equivalent formations in other parts of the Himalaya it is widely considered that the Munsiri Formation represents the southeastern extension of the "Jutogh Formation" of the Himachal Himalaya (Pilgrim & West, 1928; Valdiya, 1964a & b, 1980). Thus the Almora Group forms the huge nappe of the Jutogh-Munsiri-Almora unit, and the Champawat Granodiorite can be compared, with regard to composition, petrogenesis and lithotectonic setting, with the Chor Granite which intruded the Jutogh Formation. Towards the east, the Almora-Munsiri rocks are comparable with the "Lower Crystallines" of Fuchs & Frank (1970), the "Upper Midland Formation" between Thulo Bheri and Gandaki valleys of Bordet et al. (1972), Le Fort (1975), Pecher (1976) and Hashimoto et al. (1973), the "Kathmandu unit" in Central Nepal of Hagen (1958) and the "Irkhua Crystallines" including "Barun Gneiss" in Central Nepal of Hashimoto et al. (1975). The lower part of the succession of the Darjeeling Hills and adjoining Sikkim comprising what Ray (1947) had recognised as golden and silvery mica schist with augen gneiss and graphitic schist with garnetiferous mica schist can be equated with the Almora-

Munsiari Group. This unit is now known as the "Paro Formation" (Acharyya, 1975). In southern Bhutan the Paro is missing. Further east in the Kameng district of Arunachal Pradesh the "Bomidila Formation" comprising mica schist, garnetiferous mica schist, marble, quartzite and porphyroblastic gneiss (Das et al., 1976) represent the extension of the Almora-Munsiari.

Valdiya (1980) recognised that of the four main types of schistose rocks occurring in the Almora Group viz. chlorite schist, sericite-chlorite schist, biotite-sericite schist and graphitic or carbonaceous schist, the last one forms a prominent but subordinate component in the upper part of the succession while the last three are common in the lower part. Porphyroblastic garnets (2.5 mm across) are locally very common in some schistose rock horizons; these garnet grains show an inner zone of concentration of inclusions of quartz, mica, and black opaques. The inclusions in some places display sigmoidal disposition in the central part probably indicating rotation during growth. Further discussion on garnet textures will be given in Chapter - 5 (Metamorphism), section - 5.4. Some garnetiferous carbonaceous schist horizons show staurolite occurring as small porphyroblasts bearing inclusions of quartz, mica and magnetite. Blades of kyanite have been reported from some horizons of pelitic schists in the Dudhatoli region in south-central Kumaun (SE of Rudraprayag); these kyanites also contain inclusions of quartz, muscovite, biotite and opaque minerals and are bordered by mica flakes. Near the Almora Thrust tectonic deformation and syn-thrust metamorphism have led to development of phyllonites from schists. The porphyroblasts of garnet and feldspars are granulated, stretched, streaked out and reduced in size, and some of the sericite and biotite grains have become extremely finegrained and chloritised; this implies that these minerals formed prior to the thrusting or dislocation. The phyllonites are interbedded with augen mylonite and mylonitic granite (Valdiya, 1980).

Two main metamorphic mineral assemblages have been recognised by Valdiya (1980), particularly in the lower part of the Almora Group:

- a) quartz - garnet - sericite - biotite - albite - epidote, and
- b) quartz - perthite (microcline) - albite - biotite - muscovite - almandine  $\pm$  staurolite.

He suggests that the former is indicative of greenschist facies of metamorphism, while the latter represents lower amphibolite-almandine facies.

The widespread occurrence of garnet in the pelitic schists and even in psammites and the presence of perthite and oligoclase coexisting with epidote in the



associated gneisses indicate that the grade of metamorphism once rose to almandine-amphibolite facies. In the northeastern part of the Dudhatoli mountain, the grade rose still higher as indicated by kyanite. Valdiya considers that in the lower basal part of the succession the dislocation metamorphism with attendant retrograde metamorphism has brought down the grade of metamorphism to greenschist facies as borne out by chloritisation of biotite and garnet, biotitisation of staurolite and garnet and fragmentation of felsic minerals. Relics of garnet in the chlorite-sericite schists in some pockets bear evidence of their former higher-grade metamorphism. *It will be argued below that retrogression followed the main thrusting event in time. Stretching or fragmentation of retrogressed (chloritised) minerals is not found; on the contrary, generally many of the stretched high-grade minerals have suffered retrogression.*

Thus, based on the evidence of included minerals and their highgrade hosts as reported by Valdiya (1980), there is a clear indication that the Almora Group rocks suffered at least two episodes of progressive metamorphism. The retrogressive metamorphism followed the prograde episodes in time. Detailed discussion on these aspects based on my own observations will be given in Chapters 5<sup>and</sup> 6.

Biotite from Almora granite and gneiss has been reported by Sarkar et al. (1965) to be of lower Oligocene age (K-Ar age of 40-30 Ma) that probably marks the end of the last major thermal event registered by the metamorphics of the Almora Nappe. Jager et al. (1971) dated one biotite sample from the Chaur granite in Himachal by the Rb-Sr method at  $50 \pm 10$  Ma. These dates suggest that the dominant progressive metamorphism in the Almora Group took place well before the Mid-Miocene time which is the widely accepted age for the MCT-thrusting (Windley, 1988; Hodges et al., 1988; Molnar & England, 1990; England et al., 1992).

### 2.2.3 Ramgarh Group

This thrust-bound tectonic unit is analogous to the Chail of Himachal Pradesh and comprises a quartz-porphyry and porphyritic granite suite occurring within a succession of phyllites, finegrained quartzwackes and metasilstone, and carbonaceous pyritous slates alternating with banded white-blue marble. Lithologically, the Ramgarh Group is divided into two units: (i) Nathuakhan Formation, (ii) Deb guru Porphyroid (Valdiya, 1980).

#### Nathuakhan Formation

This formation largely consists of brown, grey and olive green phyllite which are locally schistose, alternating or intercalated with fine-grained sericitic flaggy quartzite and metasilstone. Facies variation is very common in this lithological unit.

## Debguru Porphyroid

The granitic bodies in the Ramgarh Group are collectively recognised as 'Debguru Porphyroid'. The country rocks in the granitic belt are metamorphosed to albite-epidote-amphibolite facies. In the proximity of the bounding thrust planes, the rocks are retrogressed.

The Ramgarh succession is characterised by quite extensive <sup>acid</sup> magmatism. Its lower boundary is defined by the Ramgarh Thrust which in the SE was called the Ladhiya Thrust by Valdiya (1962a, 1963). The upper limit is given by the S. Almora Thrust which has brought the huge pile of metasediments and granitic gneisses of the Almora Group over the Ramgarh. A persistent zone of phyllonite follows the plane of thrust throughout the belt.

### 2.2.4 Sirmur Group

The Sirmur Group consists of two formations: (i) the Subathu Formation of Lower Eocene age, and (ii) the underlying Bansi or Singtali Formation of Upper Cretaceous or Paleocene age (Valdiya, 1980).

#### Subathu Formation

The Subathu Formation is the youngest lithostratigraphic unit of the Lesser Himalaya and occurs unconformably upon the Tal as well as on the Chakrata. It comprises alternating maroon or red and grey-green greywackes, siltstone and mudstone, with intercalations of characteristically nummulitic and shelly limestone of Eocene age. Medlicott (1864) first studied this unit. Raiverman & Raman (1971) noted that the green-grey and the red facies of the Subathu are repeated in the sequence due to stratigraphic intertonguing. The Subathus are partially preserved in several tectonic windows and under the Krol-Tons and Bijni/Saklana thrusts. Because it contains identifiable fossils of limited stratigraphic range, the Subathu plays a very important role in stratigraphic correlation in a generally unfossiliferous Lesser Himalayan domain.

#### Bansi or Singtali Formation

The Bansi or Singtali Formation underlies the Subathu Formation below an unconformity in the Lansdowne and Mussoorie Hills. It comprises sandy, characteristically oolitic and shelly dense limestone grading into calcareous sandstone, showing cross-bedding. The limestone may be locally crystalline, and blue-black in colour, but commonly weathering to brown. On the basis of Tewari &

Kumar's (1967) discovery of cyclotomatous bryozoan fossil *Laterocavea* and calcareous algae *Neomeris* from the limestones of the Bansi Formation in the Nilkanth area, the Bansi Formation is assigned a Cretaceous to Paleocene age.

### 2.2.5 Mussoorie Group

The Mussoorie Group consists of three formations, viz. Tal, Krol and Blaini.

#### **Tal Formation**

This youngest formation of the Mussoorie Group shows variation in lithology from area to area. In the Mussoorie Hills there is a dominance of carbonaceous mudstones, phosphatic limestone and chert; whereas in the Lansdowne Hills there are conglomeratic mudstone and greywacke. In addition, in both the areas there are some quartzite and black as well as purple slate bands. The Tal Formation in the Lansdowne Hills has yielded datable fossils, such as fenestellids, bryozoans and brachiopods assigned to the Permian by Valdiya (1980). However, this age is not unanimously accepted (Azmi, 1981).

#### **Krol Formation**

The Krol Formation is a sequence of limestones, slates and siltstones, that follows the Blaini without perceptible break. This dominantly carbonate formation constitutes the most prominent upper part of the synformal mountain range that extends for over 200 km from the Yamuna valley to Nainital. The lower limit of the Krol is not sharply defined, and the lithology of the lower member (Lower Krol) has many things in common with the upper member of the Blaini (Infra-Krol). But the upper limit is very sharply defined against the overlying flyschoid sequence of the Tal Formation which begins with very distinctive phosphatic rocks in the Mussoorie Hills and with conglomeratic black shale-greywacke assemblage in the Lansdowne Hills. The phosphatic horizon of the Tal Formation has been taken as marking an unconformity. Based on conodont fossil finds Azmi & coworkers (Azmi, 1981; Azmi & Pancholi, 1981; Azmi & Joshi, 1981) considered that the Upper Krol is of Cambrian age. However, the age controversy still remains. Recently, Brasier et al. (1992) suggested an age correlation of the Krol Belt rocks with Neoproterozoic to Lower Cambrian strata from Yangtze Platform (S. China), Alborz Mountains (Iran) and Gobi-Altay Mountains ('outer' Mongolia) based on carbon isotope stratigraphic evidences. Further suggestion for an uppermost Proterozoic age for the Krol carbonate sequence has come from Frank et al. (1992).

## **Blaini Formation**

The Blaini Formation consists of conglomerate with siltstone, greywacke and slates of grey, olive green and black colours, and impersistent lenticular beds of purple or red limestone, associated with purple slate and sandstone. It rests upon the Nagthat quartzites. The Blaini Formation forms an unbroken girdle all around the Mussoorie-Lansdowne and Nainital Hills. The fine banding of siltstone, slate and rhythmite occurring in the upper part of the Blaini Formation has been interpreted by some workers (e.g. Fuchs & Sinha, 1975) as varvite. Throughout its extent the Blaini is separated from the underlying Nagthat by the sharpest unconformity of the Lesser Himalaya, marked by a well-defined and laterally persistent conglomerate horizon bearing pebbles of typical Nagthat rocks. But the upper boundary of the Blaini Formation is rather blurred. The slates with intercalated limestones imperceptibly grade into the limestone-predominant calcareous succession of the Lower Krol. This upper line of division is, more often than not, placed arbitrarily (Valdiya, 1980).

### **2.2.6 Jaunsar Group**

The Jaunsar Group consists of Nagthat Formation and Chandpur Formation. According to Valdiya the equivalent of the Nagthat Formation in the inner parts of Lesser Himalayas is known as the Berinag Formation. A sizable component of the Nagthat Formation is given by a suite of basic volcanic rocks collectively known as 'Bhimtal Volcanics'. The description below is given in the following order: Nagthat Formation, Bhimtal Volcanics, Berinag Formation and Chandpur Formation.

### **Nagthat Formation**

The Nagthat Formation is an assemblage of purple, fawn, white and green quartz-arenites which are locally pebbly or conglomeratic, and interbedded with green and purple slates. It is in transitional contact with the underlying Chandpur Formation. Green tuffaceous phyllite and quartzite and penecontemporaneous basic volcanics are essential components of the Nagthat lithology. In places the basic volcanics i.e. the Bhimtal Volcanics assume dominance over the sediments within the Formation.

### **Bhimtal Volcanics**

Penecontemporaneous lava-flows of spilitic composition and tuffites are interbedded with the Nagthat quartzites. In the Bhimtal area near Nainital the across-strike width of the outcrops of basic rocks reaches about 5 km. The volcanic suite comprises amygdaloidal, vesicular and massive basalts now converted partially or

wholly into amphibolite or epidiorite, and the tuffite altered to chlorite schist in the vicinity of thrust zones. The older volcanics are penetrated by dykes and sills of relatively fresh and little-metamorphosed dolerite and minor gabbro which obviously are younger in age. These younger intrusives are injected also into the overlying Blaini and Krol formations. Thus, while the older suite of volcanics is penecontemporaneous with the Nagthat sediments, the younger set could be as young as the Krol, or may be still younger.

### **Berinag Formation**

The Berinag Formation is lithologically similar and possibly correlatable to the Nagthat Formation of the outer Himalayan belt. But its structural position is quite different in that it forms a thrust sheet bounded by thrust planes of regional dimensions. The Berinag Formation is thrust over the Deoban or the Mandhali Formation of the Tejam Group and is itself thrust over by the Munsiri crystallines. The Berinag Formation covers a vast region in the inner Lesser Himalaya and consists of massive, coarse-grained to pebbly or even bouldery and usually very sericitic quartz-arenite of white, pale purple and green colour with metamorphosed amygdaloidal vesicular basalts and tuffites. The Berinag quartzites are in all likelihood the northerly prolongation of the Nagthat (Valdiya, 1980). The only difference is in the relatively higher grade of metamorphism exhibited by the Berinag rocks which comprise sericitic quartzite, sericite-quartz schist, schistose amphibolite and chlorite schist, all of which bear evidence of mylonitisation. Considering these differences and their tectonic severance, its separate identity under the name 'Berinag' is retained. The lower as well as the upper limits of the Berinag are demarcated by thrust planes of regional dimension. The Berinag Thrust below has presumably eliminated the Chandpur and in some areas the Mandhali also and thus brought the quartzite-volcanics assemblage over the Mandhali or the Deoban Formation of the Tejam Group.

### **Chandpur Formation**

The Chandpur Formation consists of olive green and grey phyllites interbedded and finely interbanded with metasilstones and very fine-grained wackes, with local metavolcanics. Building the base of the Krol Nappe the Chandpur Formation covers the bulk of the Nag Tibba range in Garhwal from the Eastern Nayar to the valley of Tons and beyond. The lithology of the Chandpur Formation represents mildly metamorphosed greywackes with load casts and lustrous phyllites, and is not a turbidite as some workers tend to believe. While in most sections the

lower boundary of the Chandpur Formation is represented by a thrust (Krol/Berinag Thrust), its upper boundary with the Nagthat Formation is gradational or transitional.

### 2.2.7 Tejam Group

#### **Mandhali Formation**

The Mandhali Formation consists of greyish green and black carbonaceous pyritic phyllites or slates interbedded and interbanded with marmorized and plastically folded blue-banded limestones and a variety of lentiform paraconglomerates.

#### **Deoban Formation**

The Deoban Formation is an extensive succession of the stromatolite-bearing cherty dolomite, and dolomitic limestone with bands and intercalations of blue limestone and grey slate that overlies the Rautgara Formation of the Damtha Group. The Deoban rocks were designated by Heim & Gansser (1939) as "Calc Zones of Badolisera and Tejam" and they were erroneously equated in age with the Krol Formation. Stromatolites from the Deoban Formation indicate an age of Middle-Upper Riphean (c. 1100 - 900 Ma). According to Valdiya (1969a), bioherms of branching-columnar stromatolites belonging predominantly to the *Baicalica* group occur in the Deoban dolomites. Sporadic sulphide mineralisation is discernible throughout the "Gangolihat Dolomite" which designates the mineralised facies of the Deoban developed in eastern Kumaun. The contact between the Deoban and the Rautgara Formation of the underlying Damtha Group is usually perfectly conformable, without any suggestion of tectonic or sedimentational break. But the transition is quite abrupt everywhere and the very shallow-water carbonates suddenly succeed the slates and quartzites of the flyschoid facies of relatively deep water environment.

### 2.2.8 Damtha Group

The Damtha group is made up of two formations: the Rautgara Formation and the Chakrata Formation.

### Rautgara Formation

The Rautgara Formation occupies the deepest level in the southeastern part of the Kumaun Himalaya, but it transitionally succeeds the Chakrata Formation in the northwestern region. The Rautgara Formation is an extensive lithostratigraphic unit of fine- to medium-grained muddy quartzite (subgreywacke to sublitharenite) of cream, white, pink, purple, grey and brown colours, sparse lentils of conglomerates and olive green and purple slates, often superficially oxidised to deep red soils, and characterised by extensive occurrences of basic sills, dykes and lava-flows. The Rautgara Formation is a flyschoid formation with characteristic sedimentary structures and represents the upper part of the deeper-water turbidite flysch of the Chakrata Formation. The Rautgara Formation is older than the roughly 1000 Ma old Deoban Formation.

### Chakrata Formation

The Chakrata Formation is a thick succession of dominantly greyish green greywackes and siltstones rhythmically alternating with slates, and exhibits many turbidite features. Cyclicity of red and green facies of the turbidites is a notable feature of this oldest sedimentary formation of the Kumaun Lesser Himalaya. In the north it underlies the Deoban Formation of upper Middle Riphean age with an impersistent intervening horizon of the Rautgara Formation, while in the southern part the Mandhali Formation of Upper Riphean to Vendian age has been thrust over it.

The base of the Chakrata Formation has not been discovered yet. So, it is not possible to surmise what lies beneath it. The upper limit is variable. To the south of the Chakrata township in SW Garhwal the formation is separated by a regional thrust, i.e. the Tons Thrust, from the overlying Mandhali Formation, with patches of Subathu Formation partially sandwiched in between. Almost similar is the situation in the Bidhalna and Pharat windows in southern Tehri where the Subathu-capped Chakrata is thrust over by the Chandpur of the Krol Nappe. In the Chakrata area generally the northern and southern limits of the Chakrata Formation are tectonic; but at places the greywacke-slate assemblage of the Chakrata gradually passes into the quartzite-slate succession of the Rautgara. On the basis of the upper Middle Riphean stromatolites (900-1000 Ma) in the Deoban, Valdiya (1970a) considered the Chakrata to belong to the Precambrian, most probably to Lower Riphean times.

### 2.3 A DISCUSSION ON SELECTED RADIOMETRIC AGE DATA

Table-2.2 gives a selection from the available radiometric age data in the literature in order to highlight the fact that there was early metamorphism (most possibly, Precambrian) of the crystalline rocks of the Himalayas in addition to their Himalayan age metamorphism/s. In this selection of age data the whole rock ages have been given preference over the mineral ages.

**Table - 2.2**

A list of selected radiometric age data from Himalayan crystalline rocks

<u>Sl. No.</u>	<u>Method</u>	<u>Age (in Ma)</u>	<u>Rock-unit/Location</u>	<u>Source/Reference</u>
1.	Rb/Sr (whole rock)	675 ± 70	Biotite granite intruding Vaikrita gneisses in Sutlej & Baspa valleys, H.P.	[Sharma (1983)]
2.	Rb/Sr (whole rock)	467 ± 46	Similar setting as above near Manikaran, H.P.	[Bhanot et al. (1979)]
3.	-do-	494 ± 50	Vaikrita crystalline gneiss near Karcham, H.P.	[Sharma (1983)]
4.	-do-	495 ± 16	Jaspa Granite & nearby gneisses from Spiti-Lahul area	[Frank et al. (1977)]
5.	-do-	1900 ± 100	Granitic gneisses (Munsiari Fm.)	[Singh et al. (1986) quoted in Valdiya (1988)]
6.	-do-	c. 1950	Crystallines of Munsiari Formation	[Gopalan, pers. comm. to Valdiya (1988)]
7.	-do-	1830 ± 200	Higher Himalayan crystalline gneiss near Kalamuni Pass	[Bhanot et al. (1977)]
8.	-do-	1170 ± 120	Ramgarh (Debguru) porphyritic granite near Koidal	[Bhanot et al. (1976)]
9.	-do-	1765 ± 60 to 1875 ± 90	Granitic porphyroid of Ramgarh unit	[Trivedi et al. (1984)]
10.	-do-	1110 ± 131 to 1584 ± 194	Amritpur Granite (Ramgarh Group) from SE Kumaun	[Singh et al. (1986)]
11.	K/Ar (K-felds)	315 ± 5	Almora Group augen gneiss	[Krummenacher (1971)]
12.	-do-	363 ± 5	Almora "granite"	-do-

13. Rb/Sr (whole rock)	500 ± 100	Mandi "granite", H.P.	[Jager et al. (1971)]
14. K/Ar (whole rock)	265	Dandeldhura "granite"	[Talalov (1972)]
15. Rb/Sr (whole rock) (max. apparent age)	930 to 980	Selected crystalline rocks from Himachal & Kumaun	[Crawford (1981, p. 196)]
16. -do-	650	Ramgarh (Chail) crystallines, H.P.	-do-
17. -do-	512 ± 16	Granites in gneisses & schists, S of Chandra Valley, Punjab Him.	[Frank et al. (1973) & Frank (1977)]
18. K/Ar (whole rock)	277 ± 8 & 315 ± 9	Central crystalline gneisses from Kulu-Manali-Rohtang Pass area, Himachal Himalaya	[Mehta & Rex (1977)]
19. Rb/Sr (whole rock)	580 ± 9	Migmatitic central crystalline gneiss from Manali-Rohtang Pass area	[Mehta (1977)]
20. -do-	517 ± 8	Similar rocks as above from Kulu area	-do-
21. -do-	1620 ± 90	Higher Himalayan crystalline gneiss from Almora-Askot area	[Powell et al. (1979)]
22. -do-	345 & 390	Himalayan Central Gneiss of Lahul	-do-
23. -do-	c. 1000	Daling schists (eqv. to Ramgarh) from Darjeeling Himalaya	[Crawford (1981)]
24. U/Pb (zircon)	970 to 1700	Metaclastic protoliths of the Higher Himalayan crystallines, Langtang valley, NNW of Kathmandu, Nepal	[Parrish et al. (1992)]
25. U/Pb (monazite & xenotime)	20.4 to 20.7	Semipelitic gneisses and leucogranites from the same area as above	-do-
26. K/Ar (muscovite)	728 ± 12	Quartzite from Higher Himalayan Nawakot Nappe, Nepal	[Krummenacher (1961)]
27. K/Ar (sericite)	1280	Carbonaceous schist near Tensing, Nepal	[Krummenacher (1966)]
28. K/Ar (muscovite)	872	Base of Kunchla Sr. (eqv. to Rautgara) in Nepal	-do-
29. K/Ar (uralite)	819 ± 80	From a metabasic extrusive from near the above location	-do-
30. K/Ar (biotite)	53	Ulleri augen gneiss, central Nepal	[Krummenacher (1971)]
31. K/Ar (muscovite)	51 ± 3	Palung "granite", central Nepal	[Khan & Tater (1970)]
32. Rb/Sr (biotite)	50 ± 10	Chor "granite", Kumaun	[Jager et al. (1971)]
33. Fission-track (muscovite)	48 ± 24	-do-	[Nagpal & Nagpal (1973)]

34. F-T (apatite)	15 ± 0.2	-do-	[Nagpal et al. (1973)]
35. F-T (apatite)	4.7 ± 1.1	Dharamsala granite, Himachal	-do-
36. -do-	36 ± 1.2	Mandi granite, Himachal	-do-
37. Rb/Sr (biotite)	24 ± 2.4	-do-	[Jager et al. (1971)]
38. -do-	31.4 ± 2.9	-do-	-do-
39. Ar/Ar (hornblende)	38	From an amphibolite band in Vaikrita Gneiss near Surathota, Dhauliganga valley, Kumaun	[Silverberg, D.S., pers. comm. to Valdiya (1987), quoted in Valdiya (1988)]
40. K/Ar (biotites)	8.5 to 3.7	From rocks around the equivalent of Munsiri Thrust in E. Nepal	[Quoted in Roy & Valdiya (1988)]

Krummenacher (1961, 1966, 1971) strongly inferred from his age data analyses a Precambrian metamorphism in the Himalayan crystalline metamorphics. It must also be noted that Kumar et al. (1978) gave evidence of an angular unconformity within the Precambrian in central Nepal and suggested a Precambrian deformation.

Giving a compilation of radiometric dates available for granitoid rocks from different sections of the entire Himalayan belt Sharma (1983) suggested that the Precambrian basement (Vaikrita Group) underlying the sediments of the Tethyan basin was once metamorphosed in late Precambrian time. He cited three pieces of evidence for this - (a) 580 ± 9 Ma age (Rb-Sr whole rock isochron) of the Rohtang migmatitic gneisses from Manali-Rohtang area (Himachal Himalaya) as determined by Mehta (1977); (b) presence of xenoliths of kyanite-sillimanite psammitic gneiss in the Kinnar Kailas granite (675 ± 70 Ma old; Sharma, 1983) and tongues of the latter in outcrops of the former, and (c) observed presence of an angular unconformity between the fossiliferous early-Middle Paleozoic and Precambrian rocks around Kathmandu, Central Nepal, and recognition of additional foliation and lineation in the older rocks (Kumar et al., 1978). Sharma (1983) did not rule out additional high-grade metamorphism of the Higher Himalayan rocks during the Himalayan orogeny. He says that the metamorphism of the Precambrian Vaikrita rocks appears to have gradually decreased in the upper level; as a result the base of the Vaikrita thrust sheet is represented by the high-grade crystallines, while in the upper levels it grades through less metamorphosed to unmetamorphosed Precambrian sediments.

Miller & Frank (1992) have attempted temporal grouping of the granitoid bodies in the NW Himalayas based on isotope age data (mainly whole-rock Rb/Sr). They suggest that the granitic rocks in the Himalayas can be classified mainly into three different age-groups corresponding respectively to:

- (i) A Proterozoic plutonism-volcanism dated 1900-1800 Ma;
- (ii) A very extensive lower Paleozoic magmatic event around 500 Ma;
- (iii) The late Tertiary leucogranite formation

with another additional group of rather sporadic and geologically less well-defined granitoids of 1100-1000 Ma age.

Many of the windows in the Lesser Himalaya contain evidence for Proterozoic magmatism only. Proterozoic granitoids also occur within distinct zones as mylonites and augen gneisses at the base of the Higher Himalayan crystallines (HHC). Only in western Nepal, these 1800 Ma old granitoids are an integral part of the HHC. These granitoids are predominantly peraluminous with  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios lying in the range of 0.708 - 0.713, thereby implying a substantial contribution from old crustal material.

Based on a survey of literature on radiometric ages of the rocks from Himachal and Punjab Himalayas (both whole rock and mineral ages), Mehta (1977, 1978) argued that the data reflect at least three major events in the evolutionary history of the Himalaya:

I. Large-scale recumbent folding, high-grade metamorphism and synkinematic granite intrusion took place between 600 and 500 Ma ago. This event produced a 'protoform' of the Central Crystalline axis, later rejuvenated.

II. Then there was a 'Hercynian Magmatic-Epeirogenic Cycle', indicated by ages of 360 - 290 Ma.

III. The 75 - 10 Ma mineral ages are those of the 'Himalayan Orogeny' which is associated with open folding, low-grade metamorphism, uplift, thrusting, nappe development and 'regional retrogression'.

However, according to Powell & Conaghan (1978) the pre-Cenozoic (or pre-late Mesozoic) ages do not relate to the major folding and Himalayan regional metamorphism, but possibly to granites incorporated in thrust sheets where regional metamorphism was weak and affected only the biotites. The Paleozoic mineral ages (notably, muscovite) probably reflect the age of crystallisation. Powell & Conaghan

(ibid.) insisted that the early major folding in the Chandra Valley is post-Calloviaian, as it deforms fossiliferous rocks near Tandi (Pickett et al., 1975); also Frank et al. (1977) traced this folding to the limit of the metamorphic terrain. Thus, while not explicitly denying the existence of old granites and related local metamorphism, Powell & Conaghan (1978) considered that the early metamorphism was not of regional nature and not related to the main Himalayan orogeny. They thought that the Himalayan crystalline rocks suffered their first major metamorphism only in Tertiary time during the Himalayan orogeny.

The old age (whole-rock) of most of the granites and gneisses amply suggests that the Higher Himalayan crystalline rocks suffered at least another, probably the first, fabric-forming metamorphic event prior to the high-grade Barrovian event. Clear intrusive relation of many of these older granitic bodies with the crystalline country rocks certainly means that the country rocks had acquired their first metamorphic fabric (foliation) before the intrusions took place (see also Guru Rajan & Viridi, 1984, p. 526). Thus the older dates place an upper bound for the time of an early metamorphism in the Himalayan crystalline rocks. Most possibly, this early metamorphism took place in Precambrian time. The strong effects of repeated deformation and metamorphic recrystallisation that took place during the Himalayan orogeny pose serious hindrance to any thorough mineralogical and textural investigation of the early metamorphic event, hence the difficulty of understanding the physical conditions operating at that time.

Crawford (1981, pp. 198-199) in his synthesis commented, "*Though the age data are still very few, and of variable quality.....reliable Precambrian ages have been found. ....It seems probable that along the entire length of the Lesser Himalaya there are Upper Precambrian rocks. The oldest reliable age is for the Almora-Askot gneisses of Kumaun,  $1620 \pm 90$  Ma (Powell et al., 1979 i.e. entry 21 in Table-2.2). These Rb-Sr ages accord with the 1000 - 2000 Ma age for the granitic porphyroid intrusions from the same area. It is also relevant that for the 'Central Crystalline Gneiss' of the Pithoragarh region of the Kumaun Higher Himalaya Bhanot et al. (1977) obtained an Rb-Sr age of  $1830 \pm 200$  Ma (see entry 7 in Table-2.2).....*" (italics are mine).

Crawford (1981, p. 199) further commented, "If these Precambrian ages are those of metamorphism, then their interpretation in terms of orogeny becomes a question of whether they are associated with Himalayan deformation. This is a matter to be solved by fieldwork" (underlined by me).

After reviewing the published radiometric age data, Roy & Valdiya (1988) distinguished five tectonothermal episodes in the evolution of the Higher and inner Lesser parts of the Central Himalayas: (i) Peak thermal conditions during the immediate post-collisional compressional episode (55 - 20 Ma); (ii) Cooling of the early main Barrovian Himalayan metamorphism (38 Ma); (iii) Cooling at the time of uplift and unroofing of the cover during later deformation (20 - 9 Ma); (iv) Latest phase of movement (main emplacement) along the MCT (14 Ma), and (v) Final phase of uplift, so far, of the Himalayan orogeny (8.5 - 3.7 Ma).

The first of the above five episodes indicating thermal plateau of broad duration from 55 Ma to 20 Ma in response to post-collisional compression is suggested by a concentration of whole rock Rb - Sr isochron ages of anatectic granitoids (see, for example, Hamet & Allegre, 1978; Krummenacher et al., 1978; Dietrich & Gansser, 1981; Ferrera et al., 1983).

The second i.e. the cooling age at 38 Ma (Late Eocene) of the early Barrovian metamorphism is indicated by  $^{40}\text{Ar}/^{39}\text{Ar}$  mineral age of hornblendes derived from an amphibolite horizon in Vaikrita Gneiss outcrops near Surraithota, ENE of Tapovan in the Dhauliganga valley (D.S. Silverberg, pers. comm. to Valdiya, 1987). This cooling age implies that the main progressive Barrovian metamorphism in the Higher Himalayas took place during all or part of the time between 50 Ma (India-Eurasia collision age) and 38 Ma. Given the generally accepted (Hubbard, 1988, 1989; Hubbard & Harrison, 1990; Windley, 1988; Molnar & England, 1990; England et al., 1992) age for peak MCT-thrusting at mid-Miocene time (c. 20-21 Ma), this deduction strongly corroborates my results obtained through detailed petrographic (textural) and field studies in that peak MCT-movement postdated a high-grade Barrovian (isograd-forming) metamorphic event. However, I find some discrepancy when Roy & Valdiya (1988) suggest that the peak thermal condition (i.e the first episode) continued over the period from 55 Ma to 20 Ma B.P. and yet the cooling age of the Barrovian metamorphism is placed at 38 Ma. It leaves us wondering about the thermal condition during 38 Ma to 20 Ma B.P.

The third episode (20 - 9 Ma) i.e. at middle to upper Miocene probably giving cooling ages during uplift and unroofing of the cover at the time of later deformation is indicated by K/Ar mineral (mica) ages (see Mehta, 1980).

The fourth episode at 14 Ma possibly implying the latest pulse of the main phase of movement along the MCT is indicated by fission-track annealing ages of garnets, apatites etc. The data also possibly suggest a vertical uplift rate of 1.1 mm/yr (Saini, 1982).

The latest episode at 8.5 to 3.7 Ma (uppermost Miocene to middle Pleistocene) possibly represents the final phase (Siwalik phase) of uplift of the Himalayan orogeny and the age data comes from K/Ar biotite ages from rocks around a thrust in Eastern Nepal which probably corresponds to the Munsiri Thrust of Kumaun (Roy & Valdiya, 1988).

## 2.4 TECTONOSTRATIGRAPHY AND LITHOLOGY OF THE STUDY AREAS

For the present study, the main focus of attention was on the Joshimath area which exposes along the Alaknanda valley a very characteristic succession across the MCT-Zone (Munsiri Formation) within an outcrop-distance of 4.25 km. True thickness of the MCT-Zone in this section is 2.5 km. About 100 km ESE of Joshimath is the Sobala area i.e. the Darmaganga section which was looked at for the purpose of comparison. Interestingly the Sobala section shows a thickness of only about 400 m of the MCT-Zone (contrast with the 2.5 km thick MCT-Zone in Joshimath section).

Plate - II gives the lithological map of Joshimath area and Plate - IV shows the details of the lithotectonic sequence along the Alaknanda section from Gulabkoti to slightly NW of Joshimath. For the structural map of Joshimath area and related discussion see Plate - III and relevant text in Chapter - 3. On the lithological map of Joshimath area (Plate-II) the most conspicuous bends or 'V's in the outcrops are due to topographic reasons. However the main foliation data presented on the litho-map show gradual lowering of dip of the structurally higher litho-horizons in their southerly exposures at higher topographic levels; for instance, see the central part of the map, SW of Auli. Here the trend of the main foliation shows a local change clearly implying the existence of a broad, gentle to open, E-plunging, late antiformal fold. The existence of this antiform, coupled with the fact that the structurally higher litho-horizons occurring at higher topographic levels indicate a tendency to show gradual lowering of dip southward, probably indicate that the hinge of a still large-scale antiformal fold lies farther south, and the fold discernible in the map could be a parasitic fold developed on the northern limb of that vast antiform (or, 'anticlinorium'). From Gulabkoti Bus Stop looking across the Alaknanda river at the near-vertical rock-cliff, one could recognise the tendency in the litho-horizons (here, Berinag-Mandhali formations) to show the progressive, gentle antiformal warping

southward. It is quite possible that these folds relate to the "Chamoli Antiform" occurring some 15 km farther to the SSW.

Plate - V gives the lithotectonic sequence along the Darmaganga valley (Sobala area) from Khet to Sobala. Note the thin (~400 m) Munsuari Formation (MCT-Zone) immediately below the Vaikrita Gneiss. Compared to the Joshimath-Helang section, the Sobala-Khet section shows less diverse lithology in the Munsuari Formation; in the root-zone Munsuari Formation there is a dominance of amphibolitic lithology, whereas farther south in the klippe-zone the quartzofeldspathic augen gneisses are predominant.

The study areas include three tectonostratigraphic units viz. the Joshimath Gneiss Formation (belonging to the Vaikrita Group), the Munsuari Formation (Almora Group) and the Berinag-Mandhali formations (i.e. a juxtaposition of parts from the Jaunsar and Tejam groups). The general setting of the three units in Joshimath area is indicated in Fig. 2.4. The Munsuari Formation essentially represents the MCT-Zone bounded to the north (i.e. at the top) by the Vaikrita Thrust and to the south (i.e. at the bottom) by the Munsuari Thrust. Earlier Gansser (1964) recognised the presence of only the Munsuari Thrust and designated it as the Himalayan Main Central Thrust. Later studies (e.g. Valdiya, 1980) established the presence of the Vaikrita Thrust as well. In fact, Valdiya (1980) favoured the Vaikrita Thrust as the real M.C.T. of the Himalayas.

Broad lithological attributes of the three units involved in the study area have been mentioned in subsections - 2.2.1, 2.2.2, 2.2.6 & 2.2.7. The details of the lithological succession in the three units are given in Plates IV and V. I tried to make these plates as self-explanatory as possible; so in order to avoid the possibility of repetition I do not consider it necessary to give much further discussion on the details of lithological variations etc in the text here. However, it must be mentioned that for the purpose of mapping in Joshimath area on 1 : 25,000 scale (maps now reduced in size as Plates II & III for ease of presentation), no lithological subdivision was possible of the Joshimath Gneiss, but distinct mappable litho-horizons could be recognised in the Munsuari Formation and Berinag-Mandhali formations.

Even though for the purpose of mapping on 1:25000 scale lithological subdivision of the Joshimath Gneiss did not seem to be practicable, on a slightly finer scale the lithology shows quite significant variation. The most dominant of the gneissose country-rocks in the Joshimath Gneiss outcrops has a semipelitic

composition with plagioclase (mostly oligoclase-albite) dominating over k-feldspar, and clinozoisite occurring as a frequent constituent. In many cases, plagioclase is the only feldspar present in these rocks. However, there are intermittent highly pelitic, psammitic, calc-silicate, carbonaceous and amphibolitic horizons as well. Proportion, type and the size of the mica minerals present, colour and overall grain-size of the rock, degree and scale of development of the main planar fabric form some of the very basic, but important tools in field for the initial recognition of lithological variations/alternations in the Joshimath Gneiss. Discussion on the high-grade mineral constituents of the Joshimath Gneiss will be taken up in Chapter 5 (Metamorphism). Features observable in the field i.e. the alternating sequence of semipelitic, pelitic, psammitic, calc-silicate etc varieties of lithology in the Joshimath Gneiss, the very clearly defined fine banding (probably reflecting a control of primary lithological banding) in a number of psammopelitic horizons, and features observable under the petrographic microscope, notably the absence of any distinguishable remnant or indication of primary igneous textures, strongly suggest that the Joshimath Gneiss is essentially a paragneiss formation. It is very likely that originally it was a sedimentary succession consisting of calcareous shale, sandstone, sandy shale, marly limestone, carbonaceous shale etc. There are many intrusive sills and veins of granitoid materials (i.e. aplitic, pegmatitic and common silica/quartz etc.) and also some veins of carbonate present within the Joshimath Gneiss Formation.

A fact of vital importance is that the Munsiri Formation in the Joshimath area consists of a distinct succession of 12 (twelve) mappable (on 1:25000 scale) lithological horizons (see Plate II). These lithological subdivisions in the Munsiri Formation (i.e. within the MCT-Zone) have not been clearly recognised before, and they have proved to be of importance in understanding the geometrical, rheological and other related attributes of the MCT-Zone bearing on the history of its tectonometamorphic evolution. For example, 'asymmetric' repetition within the succession in the order of appearance of different lithological horizons could strongly indicate the existence of fault/s within the unit and if such repetitions are numerous we could deduce a 'schuppen' geometry for the unit. The Munsiri succession does not show any such clear-cut repetition on the map-scale. Nor is any mappable 'symmetric' i.e. 'mirror-image' repetition indicating the presence of map-scale folds clearly discernible. This, however, does not preclude the possible existence of internal movement zones or fault-zones within the Munsiri Formation. The absence of map-scale repetition due to folding may imply either (i) there is no large-scale folding in the area, or, (ii) a large part of the Munsiri Formation probably lies on a

single limb of a large-scale fold or fold-system. The second of these two possibilities is shown to be more likely (see Chapter-6). Some features observed in the field strongly suggest that many of the lithological contacts within the MCT-Zone acted as movement surfaces and retrogressive fluid pathways. For example, in many instances, the contact/s among different mylonite-protomylonite horizons show the effect of phyllonitisation (retrogression). I have seen in a number of cases that, where thin amphibolite bands are sandwiched between comparatively thicker quartzofeldspathic augen gneiss and/or mylonite bands, there has been strong retrogression of amphiboles to biotite, so much so, that many of these former bands have now been converted into thin zones of 'biotite phyllonite'. Interestingly, many of these thin (sandwiched) and dark phyllonite (i.e. biotitised amphibolite) horizons have deep greyish black colour and a very lustrous (shiny) appearance which resembles the body-lustre of a black snake. As will be discussed in greater detail later (Chapters 4 & 6), the Munsiri Formation shows lithology-controlled strain-partitioning. Typical mylonitic rocks are mostly quartzofeldspathic in composition. Obviously, therefore, the shear displacement has been accommodated through microstructural changes more in some lithological bands than in others. All these features point to the fact that the entire Munsiri Formation (MCT-Zone) at the base of the Higher Himalayas is largely a schuppen zone. Except in one or two rather doubtful cases in quartzite bands, no clear-cut "way-up" structures i.e. stratigraphical markers such as cross-bedding, graded bedding, sole marks etc could be identified in the Munsiri Formation. The observations from the present study suggest that the Munsiri unit has a less unique stratigraphic status than the overlying Joshimath Gneiss Formation and the underlying Berinag-Mandhali formations. As it stands today, the Munsiri Formation represents more of a structural or tectonostratigraphic succession than a pure stratigraphic one. It is a tectonostratigraphic unit having a 'mixed' or 'hybrid' character.

The Munsiri Formation probably incorporates parts from the overlying and the underlying units so that one could now recognise altogether at least three contributory lithological elements in it -

- (a) one element refers to the parts probably derived from the Joshimath Gneiss and that lie in the northern part of the Munsiri outcrops just underlying the Vaikrita Thrust. This is probably the tectonised equivalent of the Joshimath Gneiss.
- (b) another element noticeable in the lower and lower-middle parts of the Munsiri Formation has probably some equivalence with the Berinag-

Mandhali formations (i.e. Lesser Himalayan formations). This element has a distinct Lesser Himalayan affinity.

- (c) the most important and characteristic element of the Munsiri Formation is recognised at its core and lower parts. This typical element of the Munsiri Formation is given by the quartzofeldspathic mylonite, protomylonite and granodioritic and granitic augen orthogneissic horizons and the phyllonite horizons. Some of the mylonite horizons show polyphase characteristics. In the lower and lower-middle parts of the Munsiri Formation there is some intermingling of this typical Munsiri element and the element of the Lesser Himalayan affinity.

Most of the fine-grained quartzofeldspathic protomylonite-mylonite and some of the phyllonite horizons in the area were originally coarse-grained augen gneisses as evidenced by the occurrence of remnant feldspar and/or quartz "augens" (porphyroclasts). A very enigmatic feature observed in the field is that horizons of fresh (non-retrogressed) augen mylonite, retrograde phyllonite and fresh augen gneiss occur side by side in many exposures. The obvious questions are: What does this feature imply? Was mylonitisation and retrogression episodic, and therefore, did they take place at different stages? Or, did fluids invade selective rock-horizons and, therefore, even if mylonitisation was at one single phase, do we find phyllonites (affected by fluids) side by side with unreactive (for some reasons) mylonite horizons? - If so, why? Or, was the shearing/mylonitisation quite a prolonged, fairly continuous and progressive phenomenon, and did the different augen orthogneiss horizons intrude at different times with respect to the shearing event so that the earlier ones among them suffered more stretching (or grain-size reduction) and/or more retrogression than the later ones?

That many of the phyllonites in the Munsiri Formation are transformed from original augen gneiss or mylonites is clearly indicated by the occurrence in these phyllonites of flattened clots of scaly muscovite or sericite that represent original feldspar augen/porphyroclasts. If we could find out the crystallisation or formation age of this muscovite/sericite (in phyllonite) which is the product of retrogressive muscovitisation of feldspars from the original augen gneisses, then that would probably give a crude upper i.e. younger bound on the age of MCT-shearing, because such retrogression must have taken place at the late stage or immediately following the MCT-emplacement. Dating of the porphyroclastic muscovites (e.g. 'muscovite fishes') recovered from a highly deformed, but fresh/unaltered augen gneiss or mylonite horizon may give a crude lower i.e. older bound of age (maximum) for the

start of movement along the MCT-Zone. These dates coupled with the age determined from the tiny muscovite flakes formed during the MCT-shearing in many mylonite bands could provide a well-constrained data-set for the age of MCT-emplacment. In this connection it is worth mentioning that the Ar/Ar hornblende date of 21 Ma determined by Harrison & Hubbard (1989) most probably gives the time when the MCT was active. The exact span of time during which the main MCT-emplacment took place is not properly known yet.

Contrary to the common expectation, the more highly strained mylonites, notably in the lower part of the Munsiri Formation in Joshimath area, are characterised more by the overall grainsize reduction than by the presence of ribboned quartz or streaked out feldspar porphyroclasts or quartzofeldspathic aggregates. In this case mylonitic recrystallisation appears to be assisted by progressive symplectitisation (?myrmekitisation). The sequence of events can be crudely shown in the following way:

---

Medium-to-coarse-grained  
quartzofeldspathic augen ----->  
gneiss with mainly  
feldspar augens  
(Protolith)

High degree of  
SYMPLECTITISATION ----->  
(or, ?myrmekitisation)

Profuse development of fine-  
grained recrystallised quartz &  
feldspar grains in the product  
rock  
(Mylonite)

=====> Progressive Mylonitisation ==>====

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The most important implication of this feature is that mylonitisation did not take place in retrograde conditions, but at elevated temperatures probably in the middle to upper greenschist facies environment.

The Berinag-Mandhali formations show typical low-grade metasedimentary rocks belonging to the Lesser Himalayan domain; the main varieties are quartzites, pure and impure carbonates/marbles, phyllites/schists and amphibolites (see Plates II & IV). Interestingly, roughly midway between Gulabkoti and Helang, kyanite is found to occur in low-grade setting in an exposure of quartzite on the road section (vide Plates II & IV to see the exact location of this kyanite quartzite exposure). A local thrust is recognised just above the amphibolite horizon, N of Gulabkoti (see Plate-IV). Valdiya (1980) probably recognised this discontinuity as the 'Berinag Thrust'. The equivalents of the Berinag-Mandhali formations in Sobala section is commonly referred to as 'Sirdang sedimentary zone' (Powar, 1972; Valdiya & Gupta,

1972; Sinha, 1981) which covers the ground between the MCT-Zone and the North Chhiplakot Thrust (i.e. a southern prolongation or reappearance of the Munsiri Thrust due to large-scale folding).

For a comparison between the Joshimath and Sobala sections, about 100 km apart from each other, across the MCT-Zone, please inspect thoroughly the two composite road traverses (Plates IV & V). The most readily identifiable difference in terms of the features of the MCT-Zone is its thickness variation. The 2.5 km thick MCT-Zone of Alaknanda section becomes as thin as only about 400 m in the Darmaganga section. No evidence was found which could suggest that the thin MCT-Zone of Sobala area shows generally higher strain than the comparatively thicker MCT-Zone of Joshimath area. Discussion on the possible cause of this thickness variation will be given in Chapter 6. The other most notable difference is the generally less variable lithology in the MCT-Zone or Munsiri Formation in Sobala area than in the Joshimath area. There is a predominance of amphibolites in the root-zone and of augen orthogneisses elsewhere in the Munsiri Formation in Sobala area. For more details on the general geology of NE Kumaun please refer to Powar (1972), Fuchs & Sinha (1978), Valdiya & Gupta (1972), Valdiya (1980, 1981).

In the study areas the topmost Joshimath Gneiss Formation (Vaikrita Group) is typically a Higher Himalayan sequence, and the Berinag-Mandhali formations are Lesser Himalayan, while the Munsiri Formation (MCT-Zone) spans the junction between the Higher and Lesser Himalayas.

Many workers (e.g. Le Fort, 1975; Bordet, 1973; Pecher, 1989) dealing with the inverted metamorphism at the base of the Higher Himalayas in the Nepalese sector tend to ignore the existence of two thrusts bounding a variably thick Munsiri Formation. Instead they normally tend to consider the existence of only one thrust i.e. the upper one (equivalent to the Vaikrita Thrust), designated by them as the Main Central Thrust. The Munsiri Formation is not given clear identity or recognition in many of their accounts. In my opinion, it would be fallacious if we try to model the inverted metamorphism taking only one of the two thrusts into account. However, recent studies by Morrison & Oliver (1992) in and around Kathmandu Klippe suggest the presence of two mappable tectonic boundaries (i.e. thrusts) on either side of what they locally recognised as "highly sheared Benighat Slates", which is most likely the counterpart of the Munsiri Formation in Nepal.

Some features tend to defy assigning a coherent tectonostratigraphic status of the Munsiri Formation. The Munsiri Formation (Almora Group) appears at different structural levels in the Lesser Himalaya in the form of klippen possibly due to bend-over-ramp or post-thrust folding (Valdiya, 1980, 1981, 1988; Johnson, 1986). The northernmost outcrops of the Munsiri Formation just underlying the Vaikrita Crystallines are considered to represent the root-zone, i.e. the now-traditional MCT-Zone, of the thrust-bound Munsiri unit. Now, it is quite a perplexing fact that compared to those in its so-called root-zone, the augen gneiss bodies in the Munsiri Formation occurring at different structural levels across the Lesser Himalaya are much thicker. In the so-called root-zone these bodies are quite thin, or even non-existent, in some instances (e.g. in Sobala). Presumably this has led some workers to think that the huge Lesser Himalayan augen gneiss masses could represent basement slices moved up along deep-reaching faults (Dubey & Bhat, 1986). It is quite natural that in the first instance these huge bodies may appear as 'para-autochthonous' or 'semi-in situ' intrusive masses, rather than far-travelled thrust masses. Often it is difficult to visualise how the huge 'Chhiplakot orthogneissic mass' got squeezed through such a narrow MCT-Zone in between Sobala and Nyu villages and still without leaving any recognisable trail or tail of augen gneiss in that zone. Comparatively thicker Munsiri Formation towards the south is expected to result in the Munsiri Thrust in Sobala area to be gentler in dip than the Vaikrita Thrust. In reality, however, we find a reverse situation, because the Vaikrita Thrust seems to be gentler than the Munsiri Thrust in the Sobala area. To consider the different Lesser Himalayan crystalline bodies as part of the same Munsiri Formation whose root-zone lies below the Vaikrita Group, is still largely a conjectural supposition, not based on any foolproof evidence such as from drill-core information or bore-hole geophysical data or other forms of geophysical information e.g. shallow seismic sounding data etc. However, for the sake of arguments at different points in this thesis the generally accepted 'Nappe-Klippen' hypothesis will be followed, as already mentioned in the introduction to this chapter.

## **2.5 PALEOGEOGRAPHIC STATUS OF THE CENTRAL CRYSTALLINES AND THE MUNSIARI FORMATION**

For the study of the tectonometamorphic evolution of the MCT-Zone (Munsiri Formation), it is important to understand the paleogeographic and early tectonic status of the Central Crystallines as well as Munsiri Formation. Effectively this will help to clarify the position of the Munsiri Formation (MCT-Zone) in the

regional tectonometamorphic setting. The discussion here will focus on the immediate pre-collision stage.

The important domains or units to be considered in this discussion are - (a) the Tethyan realm above the Central Crystallines, (b) the Central Crystallines themselves, (c) the Munsiri Formation just underlying the Central Crystallines, and (d) the Berinag-Mandhali formations or, more generally, the Lesser Himalayan metasedimentary formations structurally below the Munsiri Formation.

The Tethyan sediments consist of a more or less continuous succession of largely unmetamorphosed or poorly metamorphosed sediments ranging in age from Cambrian to Early Eocene (Valdiya, 1980; Sinha, 1981; Windley, 1988; Le Fort, 1975; England et al., 1991 ms. etc.). The Inner Lesser Himalayan formations are low-grade metasedimentary rocks most of which are considered to have been deposited in late Precambrian to early Paleozoic time (Valdiya, 1980, 1983; Sinha, 1981, 1987 etc.). It might be easy for us to visualise that immediately prior to the India-Eurasia collision the zone occupied by the Vaikrita Gneiss and Munsiri Formation was flanked on either side (both north and south) by a sedimentary basin - the southern basin was older and already filled up with Lesser Himalayan sediments, while in the northern Tethyan basin active sedimentation was still going on concomitant with the northward subduction of the Indian plate and consumption of the Tethyan oceanic crust (see Fig. 2.5).

While it is easy to visualise the southern boundary of the Tethyan realm, the northern margin of the Lesser Himalayan basin cannot be easily demarcated. This clearly highlights the difficulty and at the same time points out the importance of recognising the early tectonometamorphic and paleogeographic status of the Vaikrita Crystallines and the Munsiri Formation. As mentioned already, where we get a fairly complete exposure of the northernmost 'root-zone' of the Munsiri Formation, as in the Joshimath-Helang section, we can recognise three distinct lithotectonostratigraphic characteristics in it. First, in the upper i.e. northern part it bears some resemblance with the Vaikrita Crystallines. Second, in its lower and lower-middle parts some intermittent lithohorizons bear similarities with the Lesser Himalayan metasediments. The third but the most important characteristic refers to the most typical attribute of the Munsiri Formation which is found in its middle to lower parts. This is the occurrence of phyllonites, quartzofeldspathic mylonites-protomylonites and granodioritic and granitic augen orthogneissic horizons. This characteristic of the Munsiri Formation is clearly attributable to later tectono-

magmatic(/anatectic) events and, therefore, may not be directly relevant to our discussion on paleogeographic relations. But if we try to understand the reason why the zone occupied by the Munsiri Formation has been subjected to such important and intense events (e.g. granitoid intrusions, thrusting, mylonitisation etc.), we may get some clues as to its original paleogeographic characteristics. Presumably the present Munsiri Formation originally encompassed a potentially weak contact zone separating two substantially different lithostratigraphic realms which is why it became the main accommodating zone for tectonic strain and intrusions. It will be erroneous to consider the Munsiri Formation as a discrete part either of the Vaikrita Crystallines or of the Lesser Himalayan formations. Fig. 2.6 shows a series of cartoons depicting the suggested evolution of the Munsiri Formation.

In this connection, a paragraph from Valdiya's 1981 paper will be worth quoting where he expressed his view on the geodynamics of the Munsiri root-zone (Valdiya, 1981, p.101) -

"Judging from the scale and intensity of deformation suffered by the rocks in the proximity of the Munsiri Thrust, and appreciating the fact that its squeezed out and far-travelled part - the Almora Nappe and the chains of klippe - register a width of thrusting of the order of 120 km or so, it is quite obvious that the root zone had experienced severe compression and tectonic dynamism. The age of the granite gneiss of the Munsiri Formation near Kalamuni has been dated at  $1895 \pm 100$  m.y. and of the Askot crystallines near Didihat at  $1960 \pm 100$  m.y. (Bhanot et al., 1977) and the leucocratic granite of the Almora unit near Almora, approximately at 700 m.y. (pers. comm. V.B. Bhanot) (underlined by me). The granitic rocks thus point not only to the Almora-Munsiri rocks being of early Precambrian age (considerably older than the oldest of the sedimentary formations the Lower Riphaean Chakrata and Rautgara of the autochthon) but also to the lithotectonic unity of the Munsiri and Askot-Baijnath crystallines. Possibly these crystallines constitute the basement on which the sedimentary rocks of the Lesser Himalaya were deposited. Strong deformation in the northern extremity of the basin and consequent thrusting along the Munsiri and Vaikrita thrusts may have uplifted and pushed southward the superficial part of the postulated basement. It must be admitted, however, that there is no evidence in support of this speculation, and hence it remains in the realm of possibility."

Shortly afterwards he wrote (p.102) - "...In my opinion, the Vaikrita Group represents the basement of the Tethyan sedimentary pile, just as the Munsiri rocks possibly formed the infrastructure of the Lesser Himalayan sediments".

The radiometric dates of the crystallines from the Munsiri Formation show rather wide variation from place to place. The c.2000 m.y. age (Rb-Sr whole-rock) of Askot crystallines indirectly implies that the country-rock (sedimentary/metasedimentary) of the Munsiri Formation was even older. <sup>Valdiya (1981) noted</sup> the problem of recognising the basement to the Lesser Himalayan formations, <sup>and expressed</sup> doubts about the status of the Munsiri Formation. His speculation on considering the Vaikrita Group as basement to the Tethyan sediments and the Munsiri Formation as basement to the Lesser Himalayan rocks leaves us contemplating the relation between the Vaikrita Group and the Munsiri Formation.

There are at least four possibilities for the mutual paleotectonic and/or paleogeographic relationships, when we take Vaikrita Crystallines, Munsiri Formation and the (Inner) Lesser Himalayan formations together into consideration:

I. Vaikrita Group, Munsiri Formation and the Inner Lesser Himalayan formations belong to a single, more or less continuous succession, progressively younger in age.

II. Vaikrita Group and Munsiri Formation together were part of the basement terrane upon which the Lesser Himalayan rocks were deposited. This was Valdiya's tentative view (1981, p.101).

III. Vaikrita Group formed the basement terrane upon which originally the Munsiri Formation and the Lesser Himalayan formations were deposited successively.

IV. Vaikrita Group formed the basement terrane upon which the Lesser Himalayan formations were deposited in sedimentary basin/s; in the early part of its history the "Munsiri Formation" probably developed as a kind of basin margin/floor facies underlying the Lesser Himalayan formations, but through later tectonism it incorporated part of the basement and parts from the overlying sedimentary succession, thereby ultimately establishing itself as a spatial connection between the 'pure' basement to the north and 'pure' basinal deposits to the south. The most typical

attributes of the present Munsiri Formation developed through tectonism and metamorphism when quartzofeldspathic granitoid bodies intruded and mylonitisation affected the vulnerable lithohorizons.

The fourth of the above possibilities is most favoured by my observations. It is worth mentioning that if we follow the strict implication of its definition after Valdiya (1980, p. 71), the 'Munsiri Formation' is to be regarded as a tectonostratigraphic unit which acquired its formational status consequently to movements of the Vaikrita and Munsiri thrusts. The literature does not provide much help when we want to look back to explore the specifics about its original limits and course of evolution before the two bounding thrusts formed.

Results from recent geochronological work <sup>in Nepal</sup> by Parrish et al. (1992) are very interesting. Though apparently these results tend to suggest an altogether new possibility, in fact they provide indirect support for the fourth possibility mentioned above, highlighting that the Munsiri Formation could indeed be a unit of hybrid character. U-Pb radiometric analysis of detrital (protolith) zircon single crystals collected from both sides of the Vaikrita Thrust (named by them as MCT) shows that generally the hanging wall zircons are younger in age (0.97 - 1.7 Ga) than the footwall zircons (1.86 - 2.66 Ga). Parrish et al. (ibid.) considered this age difference between zircons as indicating a difference in age of the protolith provenance. This, they thought, would imply two different provenances; the provenance for the MCT-footwall metasediments was older and different from the one for the hanging wall rocks. I think, Parrish et al's (ibid.) interpretation of the data is correct and provides a concrete demonstration that the Lesser Himalayan metasediments were originally derived from older basement rocks than the Higher Himalayan and/or Tethyan sediments. For <sup>a</sup>long time it has been speculated that while the Lesser Himalayan basin/s had connections and contributions from old peninsular Indian landmass, the Higher Himalayas or Tethyan Himalaya had not. Continuation of some Indian peninsular orogenic trends into the Sub-Himalaya and Lesser Himalaya, and their correlation with some Lesser Himalayan transverse structures, are already known (Valdiya, 1980; see also Valdiya, 1992). I think, an older age of zircon in the MCT (or, Vaikrita)-footwall rocks than in the hanging wall in Langtang valley of Nepal might simply mean that the footwall samples came from rocks with comparatively pure Lesser Himalayan affinity. Therefore, Parrish et al's results indirectly give support for the Munsiri Formation being a multi-component hybrid unit. It is quite possible that in another section where the Higher Himalayan component in the

Munsiari Formation is more well preserved, the age difference between zircons may not be as conspicuous.

## 2.6 A NOTE OF GUIDANCE ON THE USAGE OF SOME LITHOLOGICAL AND TECTONOSTRATIGRAPHICAL TERMS

For the purpose of mapping for the present work I used a working definition of the rocks without bothering too much about strict formalisation of the rock-names. Unlike igneous rocks, the metamorphic rocks are still termed largely in a loose manner (Mason, 1978). Standardisation of rock-nomenclature for the present areas of study based on field, petrographical and chemical attributes would, by itself, make a major project. Except the 'fault-related' rocks, the naming of other metamorphic rocks in the area has been done broadly in the light of the definitions suggested by Spry (1963), Mason (1978) etc.

The ambiguity or incompleteness in the definitions of 'fault-related' rocks, i.e. mylonite, cataclasite etc., is nicely pointed out by the humorous, allegorical starting comment of a useful article by Wise et al. (1984) - "Many traditional terms for fault-related rocks have undergone recent dynamic metamorphism under high-pressure discussions by various groups of specialists." Rapid advances in understanding the processes operating in both brittle and ductile fault zones have left terminology of this field in uncertain condition (Tullis et al., 1982). For <sup>a</sup>long time since Lapworth (1885), mylonitic rocks used to be regarded as products of pulverisation until Bell & Etheridge (1973) in their pioneering work demonstrated clearly that mylonites are products of ductile flow and crystal-plastic grain-size reduction processes and not of brittle comminution of grains. But still '....some terms and definitions inappropriate to modern concepts have survived as relics from older literature' (Wise et al., 1984, p. 391). Notable among the individual efforts for standardising the nomenclatorial system for mylonitic rocks are Higgins (1971), Sibson (1977), Hatcher (1978) and, in particular, White (1982). Taking into account these workers' views, Wise et al. (1984) suggested a conceptual framework of definition into which the most commonly used terms and mechanisms could fit. Their proposed scheme of nomenclature is shown in Fig. 2.7 (reproduced from their Fig. 1, p. 392). The scheme does provide quite a practical 'framework within which both the field geologist and the rock deformation specialist can operate.' However, a major criticism is that the system is too much process-oriented bringing gneisses, schists, mylonites, cataclasites etc. on to the same platform.

Quite recently, the International Committee on the nomenclature of fault-related rocks attempted to resolve the problem of definition of fault-rocks (Ben Harte, pers. comm., 1989). Following the questionnaire that was circulated by the committee to structural geologists asking for their opinion, I personally favour the following definitions of fault-rocks:

**Mylonite:** A rock produced by tectonic reduction of grain size in localised zones (shear zones), resulting in the development of a penetrative fine scale foliation and often layering, and often with an associated mineral and stretching lineation. Porphyroclasts and lithic fragments are commonly present of similar composition to minerals in the fine-grained matrix. Brittle deformation of some minerals is possible, but deformation is mainly by crystal plasticity. Crystallographic preferred orientation is usually present. Many mylonites exhibit asymmetrical fabrics that provide evidence for shear sense of the fault zone.

**Cataclasite:** A cohesive rock containing generally angular fragments in a fine-grained matrix of similar material produced by fracturing, rotation and frictional sliding of particles (cataclasis). Generally no preferred orientation of grains or individual fragments will be present as a result of the deformation, but fractures may have a preferred orientation.

Thin section examination generally allows distinction between mylonites formed due to dominantly crystal plastic processes and often producing a crystallographic preferred orientation from foliated cataclasites which are produced as a result of cataclastic flow. In many cases the distinction of mylonites can also be made in hand specimen if features such as a mineral stretching lineation, or clearly elongate single crystals (e.g. quartz ribbons), are present, both of which often indicate crystal plastic deformation. Fig. 2.8 shows a typical-looking quartzofeldspathic mylonite from Joshimath area. Note the presence of feldspar porphyroclasts and ribboned quartzofeldspathic aggregates on the XZ-face of the hand specimen.

**Protomylonite:** A rock showing incipient mylonitisation. Normally if in a mylonitic rock less than about 50% of the rock has undergone grain-size reduction it could be called a protomylonite.

Augen mylonite: A mylonite containing distinctive large porphyroclasts or lithic fragments around which the fine-grained banding is wrapped. The term 'blastomylonite' is not favoured as a replacement for 'augen mylonite'.

Ultramylonite: A mylonite in which most of the porphyroclasts or lithic fragments have been eliminated (>90% finegrained matrix). An ultramylonite need not be "ultra" fine-grained because the grain-size reduction is relative to the initial grain-size.

Phyllonite: A hydrated, phyllosilicate-rich recrystallised mylonite is called phyllonite. Sometimes it may be difficult to distinguish in hand specimen between a first-order typical phyllite and a phyllonite derived from ultramylonite; but in the field or under the microscope generally from the rock associations and/or overprinted metamorphic fabric, phyllonites are quite easily recognised.

Fault Gouge: It is an incohesive fine-grained fault-rock which may display a foliation and is usually rich in clay minerals due to chemical transformation of country rock. Lithic clasts of various sizes are usually dispersed throughout the gouge body.

Evidently, the International Committee proposes a set of qualitative, rather than strictly quantitative or process-oriented definitions. I found it quite useful in the field.

It must be made clear that I did not lay strong emphasis on seeking the strict definitions of the lithologies for the purpose of the present work. In fact, the decision on the choice of lithological names for the purpose of preparing the maps and sections was taken on the basis of field identification of the lithologies even prior to thorough petrographic examination of the rocks under the microscope. So the lithological names are rather loosely defined and the reader is requested to take them with considerable latitude. Ideally, the system of naming a metamorphic rock should be tri-partite, taking into account (i) the composition of the rock, (ii) index minerals present, and (iii) texture of the rock.

The implications of a few tectonic/tectonostratigraphic terms used at different points in this thesis are explained below. They mostly refer to the Joshimath area, because this area shows a more complete sequence than the Sobala area.

**MCT-Zone:** This abbreviation from the 'Main Central Thrust Zone' refers to the Munsiri Formation in the present study area/s, bounded on the top by the Vaikrita Thrust and at the bottom by the Munsiri Thrust (thrust-nomenclature after Valdiya, 1980). Some workers prefer to refer to these two boundaries as the MCT-I and MCT-II respectively (e.g. Hodges et al., 1988; Morrison & Oliver, 1992). But, as mentioned earlier, there is a slight problem with this nomenclature, because it has an unqualified apparent age connotation and also gives the apparent impression that the zone between is tectonically less important than the two boundaries. In any discussion on Himalayan Tectonics, this whole zone is to be regarded as representing a single unit or element which is a shear zone of complex nature lying at the base of the Higher Himalayan crystalline stack. In order to highlight this point this zone will be called in the thesis either as Munsiri Formation or, more commonly, as the MCT-Zone. Details about the two boundaries of this zone are given in Chapter - 4.

**MCT-sheet:** Taking the whole Munsiri Formation as representing the MCT-Zone, the term 'MCT-sheet' is used to refer to the entire rock sequence of the Higher Himalayas up to the Indus Suture Zone including the Munsiri Formation.

**MCT hanging wall block:** For the purpose of the present study this term is taken to refer to the Munsiri Formation along with that part of the Joshimath Gneiss Formation which is included in the study area/s.

**MCT foot-wall block:** This term is taken to refer to that part of the Berinag-Mandhali formations which lies within the area/s of the present study. It is re-emphasised that this foot-wall block i.e. the Berinag-Mandhali formations does not form an autochthonous block.

In the next chapter, we shall investigate dominantly the mesoscale and microscale structural and deformational aspects of the rocks of the study areas, with particular attention to the Joshimath area for reasons already explained.

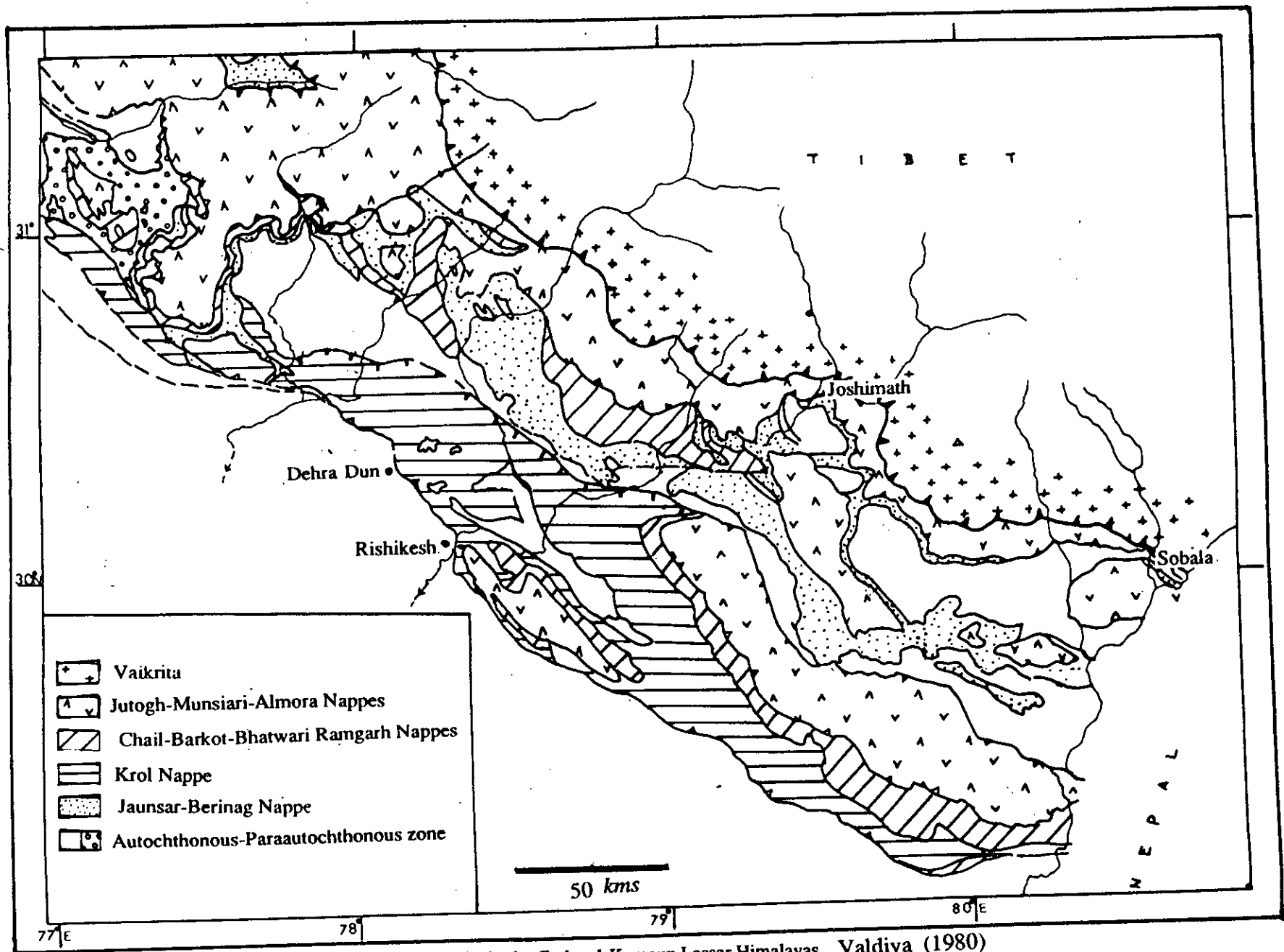


Fig. 2.1 The main lithotectonic units in the Garhwal-Kumaun Lesser Himalayas. Valdiya (1980)

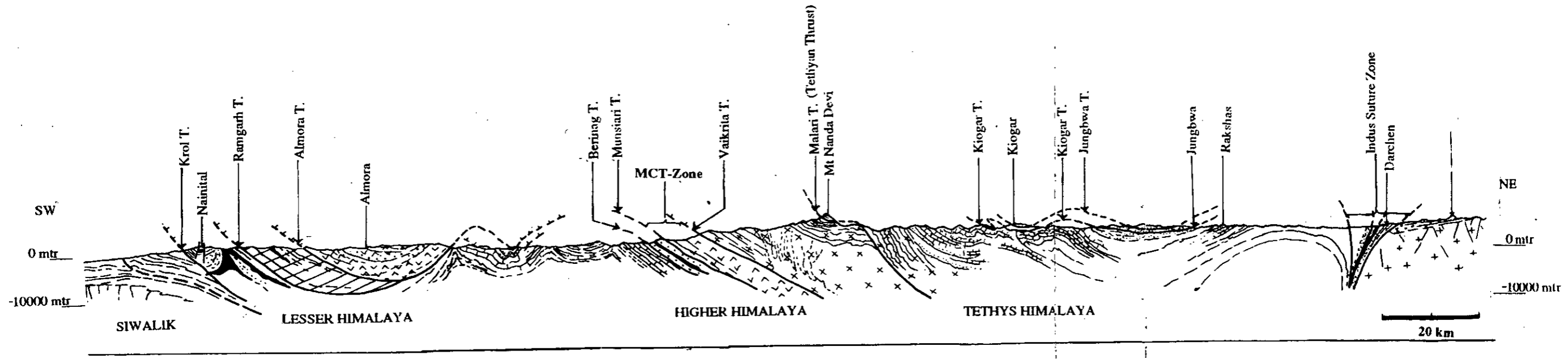


Fig. 2.2: A generalised transverse geological cross section across the Garhwal-Kumaun Himalayas (modified after Valdiya, 1979)

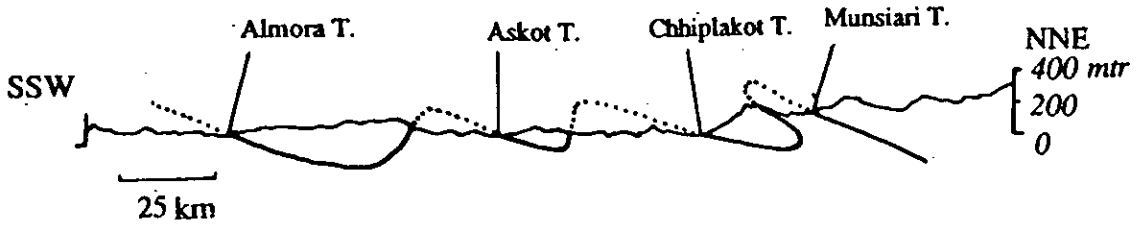


Fig. 2.3 The Almora-Munsiari Thrust and its different klippen across the Lesser Himalayas. (after Valdiya, 1980)

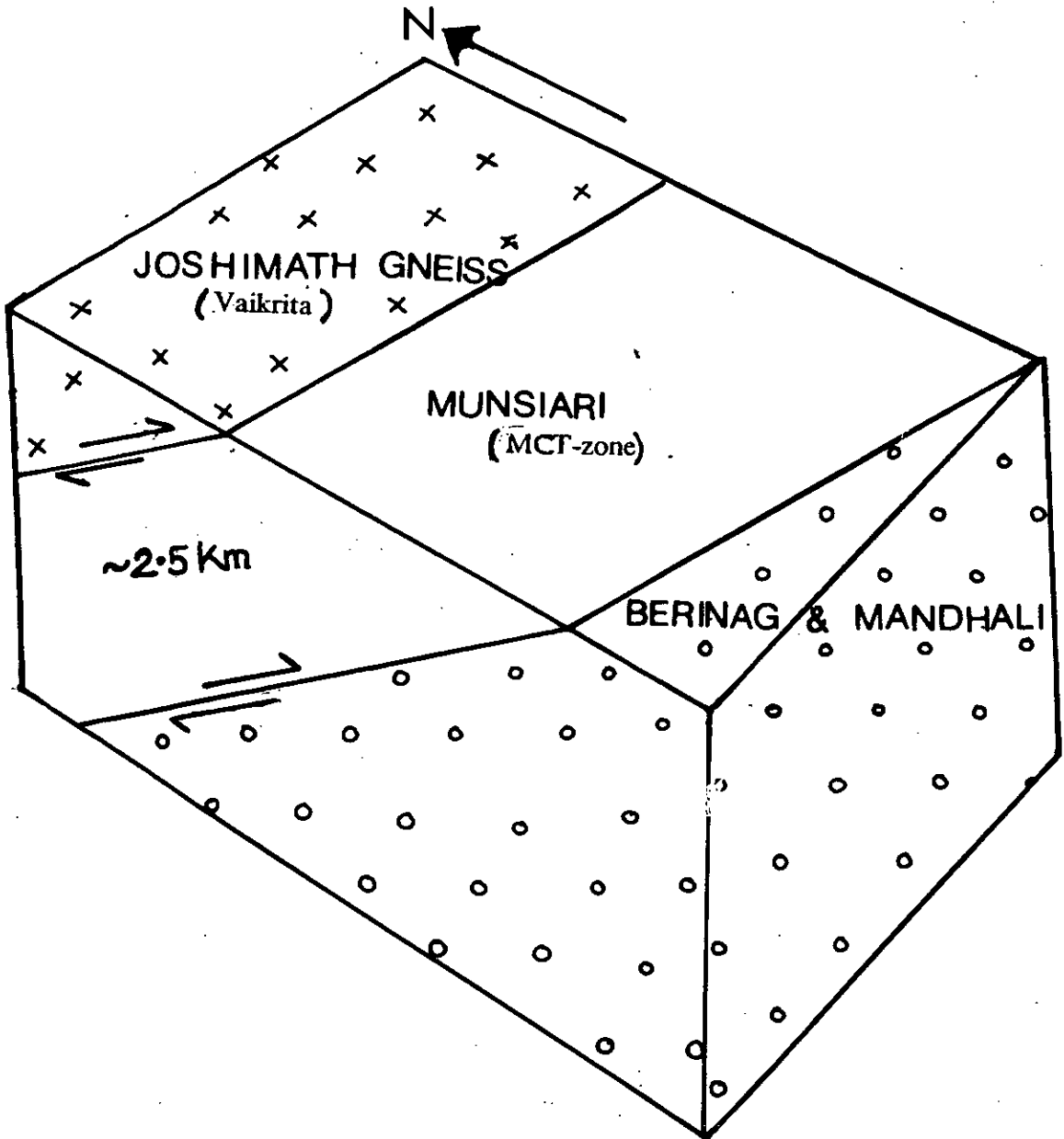
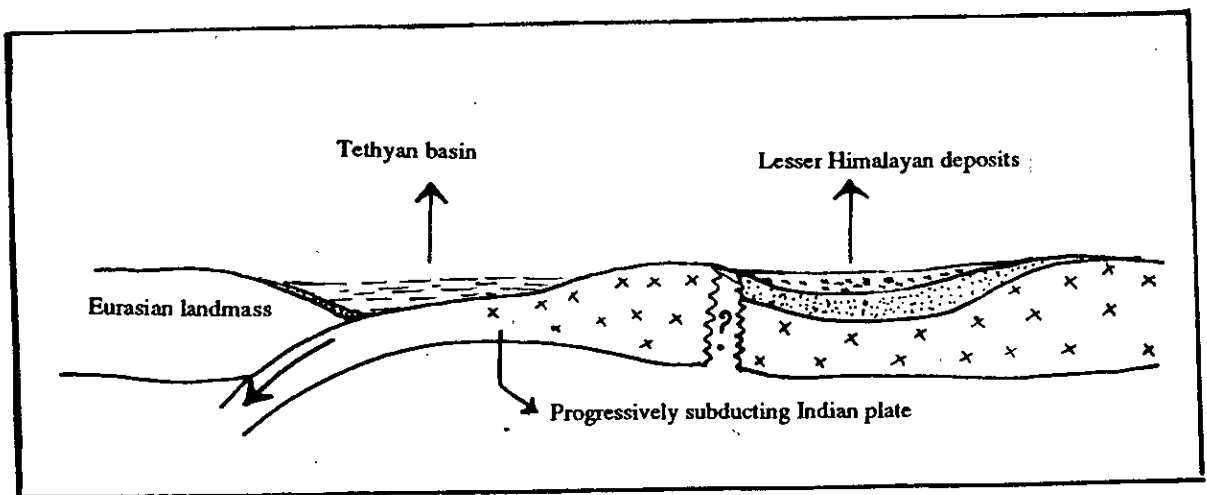
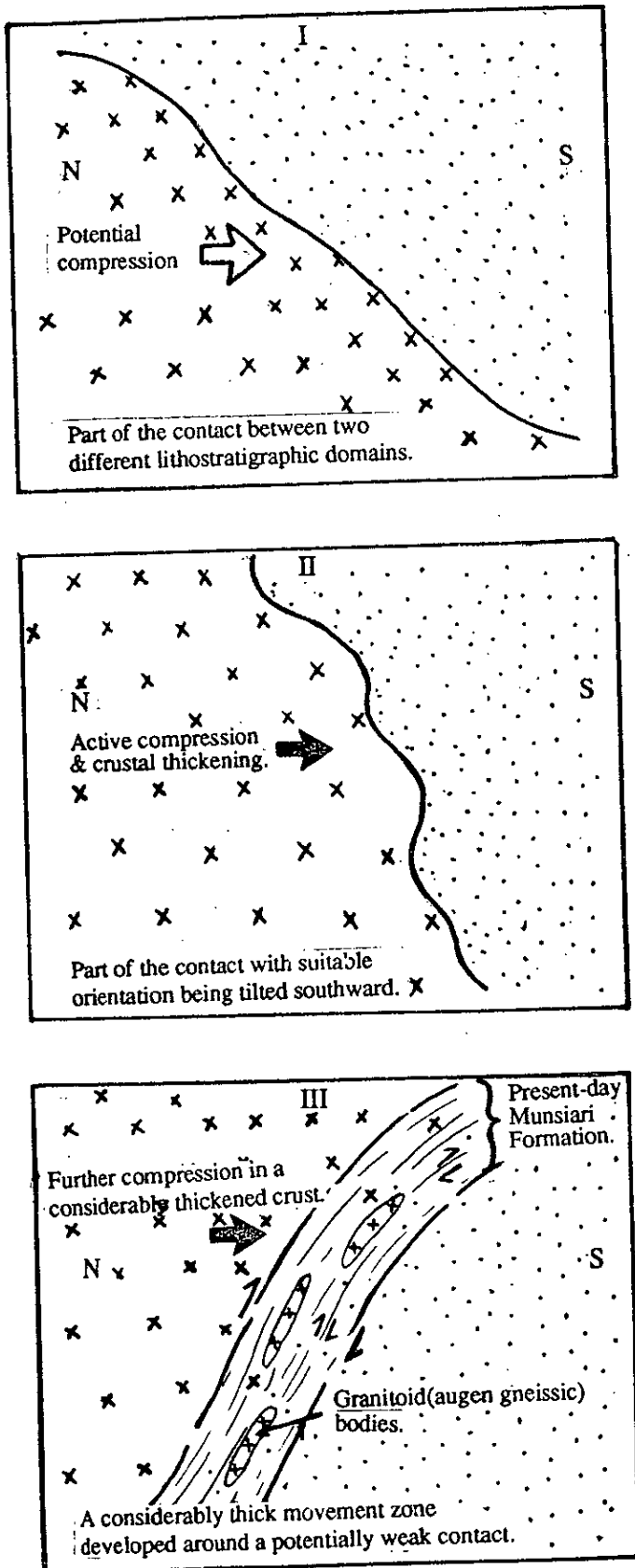


Fig. 2.4 Block diagram showing the generalised tectonic setting of the three units studied in Joshimath area.



**Fig. 2.5** Cartoon depicting possible plate tectonic and palaeogeographic set-up of the Himalayan region shortly before India-Eurasia continental collision.



**Fig. 2.6** Cartoon depicting a possible course of progressive tectonic development of the Munsiri Formation. For the sake simplicity only a part about the present level of exposure is considered.

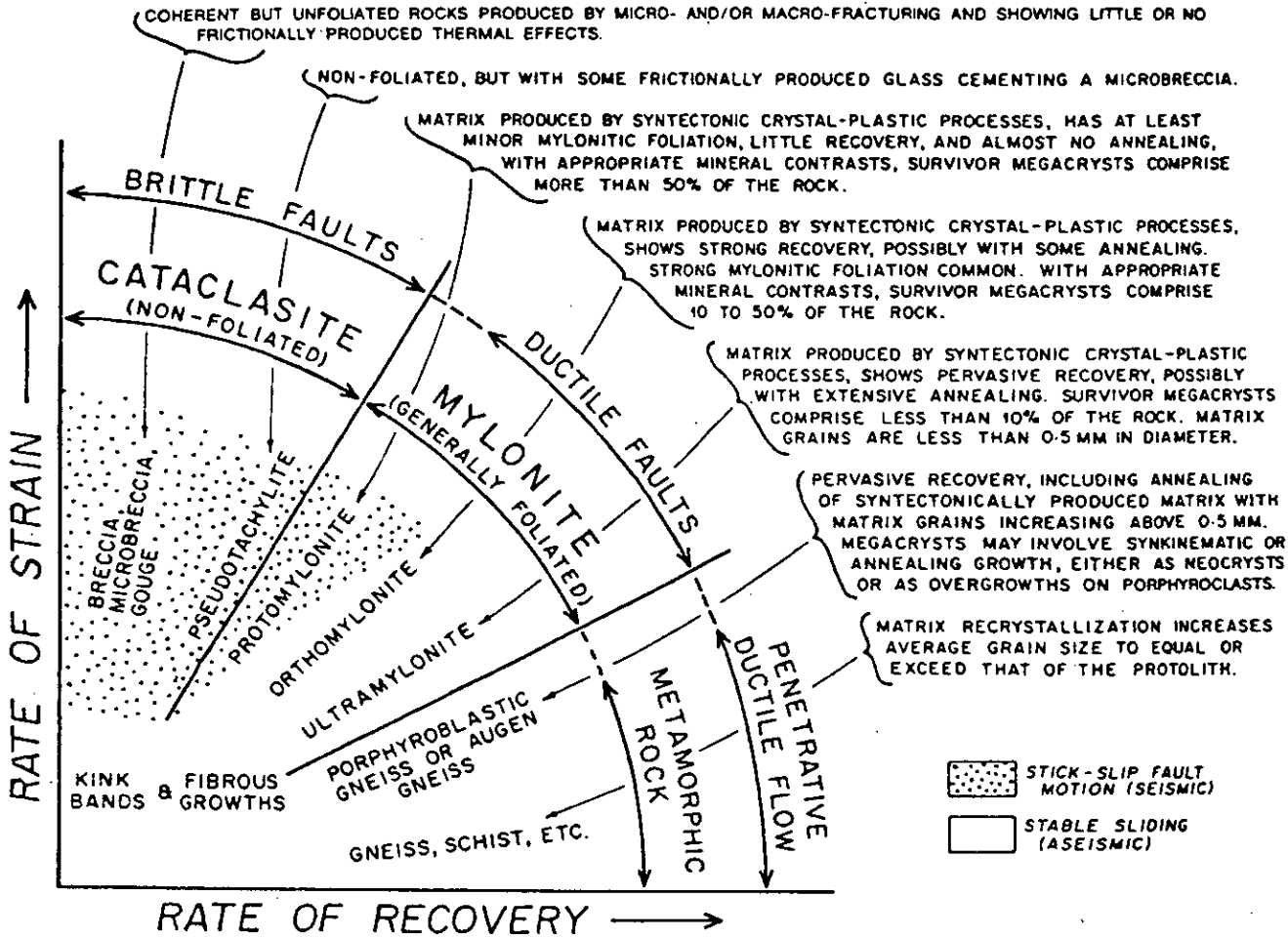


Fig. 2.7 Terminology of fault-related rocks. Horizontal and vertical scales are variable depending on composition, grain size, and fluids. From, Wise et al. (1984).



**Fig. 2.8** Photograph showing a typical quartzofeldspathic mylonite specimen from Joshimath area. Note the presence of clear stretching lineation on top face (XY) and ribboned quartzofeldspathic aggregates together with feldspar porphyroclasts on the XZ face. The effect of stretching is not at all conspicuous on the YZ face which shows somewhat irregular pattern of foliation.

## Chapter-3

# STRUCTURES AND DEFORMATION

### 3.1 INTRODUCTION

There is a wide variety of structures displayed by the rocks of the area. Most of these structures are secondary (deformational) in origin. Rarity of primary structures in the rocks is attributed to their obliteration by intense deformation and metamorphism. After a brief commentary on the primary ones, this chapter deals mainly with the secondary structures of mesoscopic and macroscopic scale, and attempts to establish their sequence of formation.

#### 3.1.1 Primary Structures

Among the rare primary structures are possible relics of original sedimentary layering/bedding as found in the psammitic exposures of Joshimath Gneiss, for instance near the Jogidhara Falls (for the location of Jogidhara Falls see Fig. 4.1). This is discussed in more detail in the subsection on foliations (subsection 3.2.1). Probable bedding and cross-beds are recognised in a few quartzite horizons in the Munsiri and Berinag-Mandhali formations. In two examples within the Munsiri Formation, the angle between foresets and regional bedding in the quartzite is  $18^{\circ}$ - $20^{\circ}$  giving an indirect measure of strain. The younging direction in both the examples is towards north. One of these cross-beds is asymmetrically folded (Fig. 3.1) and the fold asymmetry indicates a 'top-to-south' sense of shear. However, in the absence of sufficient data on younging directions, the present two data indicating a northward younging are considered to have only local significance. It must also be pointed out that some features apparently looking like cross-beds are, in fact, wellpreserved long limbs of disrupted tight overturned asymmetric folds (Fig. 3.2). Roy and Valdiya (1988) also noted their occurrence from this area.

#### 3.1.2 Secondary Structures

The secondary or deformational structures present in the area include almost all the structural elements-

Planar- Foliations, fold axial planes, faults, fractures/joints etc.

Linear- Different types of lineations, fold hinges, boudin axes, mullion or rodding etc.

An important point about the penetrative planar structures is that most of them have a variable but broadly N/NE dip (see Fig. 3.49). This may have several explanations--

- (1) repetition due to large scale folding/faulting
- (2) the stress set-up for their development remained similar throughout the structural history. This, in turn, may have important tectonic implications.
- (3) inherent asymmetry in how the stress worked relative to geographical coordinates.

The general N/NE dip of the rock horizons has some important consequences as is evident in a number of instances from the following discussions. For convenience in mapping, a single term 'main foliation' has been used to denote the most prominent and easily recognisable foliation in the rocks. Thus it stands for the typical gneissosity in Joshimath Gneiss, the mylonitic foliation of the mylonitic-protomylonitic rocks in the Munsiri or the dominant schistosity in phyllonites and other low grade rocks. It is emphasised that the term 'main foliation' as used here does not necessarily bear any particular age connotation.

The study areas show many mutually exclusive structures, which do not often show clear-cut and direct overprinting relationships, even though from the circumstantial evidence it may be understood that they originated at different times. For example, in some instances due to the absence of direct overprinting relationships among different fold episodes, clues about the relative age determination of folds were provided by their relationships with the stretching lineation. Ignoring the possible limitation imposed by the inaccessibility to suitable exposures, this must be attributed to:

- (a) in many cases, the lack of considerable time gaps between successive generations of structures;
- (b) the difference in the relative scale and place of maximum development of such structures.

Detailed analysis of the entire structural sequence may not, therefore, lead to distinct compartmentalisation of the different generations of structures. More often than not, fold episodes and foliation development, especially in thrust zones, occur in a relayed and overlapping fashion, and deformation as a whole follows a kind of continuum (see Platt, 1987; Tobisch & Paterson, 1988; Paterson et al, 1989). With

some generalisations, however, it is possible to indicate in a time sequence the peaks of different deformation events.

In view of this, first it is intended to summarise the fold episodes recognised in the area following a brief summary of the criteria used for such recognition. This will give a reference-frame on which to tie the time of development of other structures. Next will follow a description of other structural elements, especially those that are in some way related to the folds, together with a detailed account of the geometric features of the folds. Initially the structures in the tectonostratigraphic units viz. Joshimath Gneiss, Munsiri and Berinag-Mandhali will be described separately. Later, discussion will concentrate on how relationships with other structural elements (particularly with stretching lineation) helped in identifying the fold episodes and in correlation of folds and folding from one unit to the other. Also given are detailed descriptions of those structural elements that are apparently not related to folding viz. boudinage and faults, joints, late shear zones, etc. respectively preceding and after this. The chapter closes with a summary of structural history and a short commentary on the trend in structural evolution of the area.

### 3.1.2.1 Criteria used for recognition of fold episodes

The main criteria based on which the different episodes of folding were recognised are:

- (a) Nature of fold-defining foliations.
- (b) Consistency or systematic variation in the attitude of fold axes. Orientation of axial planes did not provide much help because most of the folds in the area have similarly oriented axial planes.
- (c) Morphology of the folds with special attention to the character of the axial planar structures, particularly in fold-cores.
- (d) Style or geometry of the folds including interlimb angle, hinge-curvature, class etc.

and finally, in most cases providing the confirmatory evidence--

- (e) Direct and indirect overprinting relationships. Indirect overprinting relations involving other structural elements viz. stretching lineation etc. were found to be nearly as useful as direct overprinting relationships.

The sequence of fold episodes was established essentially through studies in the field further corroborated by microscopic study. Detailed description of the textural and microstructural features will be given in later chapters.

### 3.1.2.2 Fold episodes

Two distinct domains can be recognised in the present area in respect of styles, morphology and orientation of folds. One domain consists of the Joshimath Gneiss and the Munsiri formations, the other includes only the Berinag-Mandhali formations. Most of the folds in the area are inclined (overtuned), with broadly N'ly-NE'ly dipping axial planes reflecting a uniformity of the kinematics during folding. The folds vary in size from small scale crenulations (a few mm. across) to the scale of an exposure. The discussion on folds (and other related structures) will be made in subsections - 3.2.3, 3.3.3, 3.4 etc generally taking each of the tectonostratigraphic units separately; mutual correlation among the folds and folding in these units will be discussed after that. Now, the fold sequence is given below in terms of the two above-mentioned domains. Dominant features of each episode are mentioned alongside. It is worth noting that direct overprinting or refolding relations among the different fold episodes is not always conspicuous; however, the  $F_1 - F_2$  interference patterns are not rare. Important indication about relative age of the fold episodes comes from the indirect overprinting relationships with the stretching lineation. It is proposed to discuss how the fold sequence was established based on these relationships only after the discussion about the stretching lineations in the area is given. Here the sequence of folding which has been established through the present study is introduced.

#### Fold-sequence: a summary

##### A. [Folds in the Joshimath Gneiss and the Munsiri formations]

$F_1$  folds are rare and difficult to recognise. Mainly confined in the Joshimath Gneiss, they are high amplitude tight folds plunging broadly to NE/ENE.

$F_2$  folds are the most intensely developed ones, especially in the Joshimath Gneiss. They are overturned, asymmetric, mostly southerly verging close to tight folds, often multiover, developed from microscopic to outcrop scale, with well developed axial planar schistosity/foliation (average dip  $\sim 48^\circ$ ) i.e. subparallel to the main foliation. A set of crenulations is associated with these folds. The folds are approximately coaxial with  $F_1$ , but southward the axes rotate to parallelism with a dominant set of NNE-trending stretching lineation (Fig. 3.3). Detailed

discussion on this feature will be given later (subsection 3.7). There is also an appreciable tightening of these folds southward.

**F<sub>3</sub>** folds are mainly crenulations developed commonly near the Joshimath Gneiss-Munsiari transition zone in the Munsiari Formation. They are dominantly small scale folds, maximum size up to 40 cm. and are tight asymmetric semi-reclined, with axes broadly N/NW-plunging and axial planes subparallel to the main foliation. A generation of puckers also belongs to this episode.

**F<sub>4</sub>** folds are developed mainly within the Munsiari Formation. Their axial planes dip ENE at a slightly steeper angle than main foliation. The folds range in size from tens of centimetres to a few metres. F<sub>4</sub> folding was not as intense as the F<sub>2</sub> folding.

**F<sub>4a</sub>** folds include mainly a set of crenulations with steep (>50°) N-dipping axial planar cleavage and eastward plunging axes. They are more common in the Munsiari Formation than in the lower part of the Joshimath Gneiss. In terms of regional tectonic significance, the F<sub>4a</sub> folds are not as important as the other four preceding episodes.

#### B. [Folds in the Berinag-Mandhali formations]

Folds observed in the Berinag-Mandhali formations are different in their axial orientation and morphology from the ones seen in Joshimath Gneiss and Munsiari. Three episodes of mesoscale folding are recognised in the Berinag-Mandhali--

**F<sub>I</sub>** - Tight to isoclinal folds developed on compositional layering which is almost completely transposed by the well developed axial planar foliation (i.e. the main foliation in the exposures). Axes of these folds plunge to the NW. Fold size is normally measurable in centimetres. Parallel to the compositional layering, there is a micaceous foliation which is also crenulated by this phase of folding (see subsection 3.4).

**F<sub>II</sub>** - Close to tight semi-reclined folds, sometimes giving rise to distinct mullions in the quartzite horizons, plunge towards slight west of north. Crenulations are associated with these folds. Axial planes are subparallel to the main foliation.

**F<sub>III</sub>** - Open, asymmetric folds with near-horizontal axes trending 127°-307° and steep NE-dipping axial planar spaced crenulation/fracture cleavages. This fold axial direction is almost perpendicular to the regional stretching lineation.

In the Berinag quartzite there are some highly asymmetric southerly verging folds with uniformly spaced, parallel, planar quartzose veins in their very short inverted limbs. The folds are most likely to have resulted from a late phase of shearing (Fig. 3.4) and could be termed **F<sub>IV</sub>**. However, these folds are much more local in nature than the other three episodes in the Berinag-Mandhali formations.

Finally, large scale gentle and upright warping with an approximate E-W trend has affected the whole area. Whether this warping has variable axial plunge could not be determined, but in the present area an eastward plunge is indicated by the way in which the trend of the rock horizons has been modified as seen in the higher parts of the mountain SSW of Auli (see Plates II & III). Also one can distinguish the effect of this warping by looking from the road at Gulabkoti to the near-vertical high cliffs on the opposite side of the Alaknanda river. This warping episode could correspond to that of the 'Chamoli Antiform' located farther south. An inspection of the main foliation data in Plate II indicates that, generally speaking, the main foliation becomes slightly gentler as one goes from N towards S. This feature corresponds well with the possibility of approaching the hinge zone of a large scale upright antiformal fold located towards south.

A notable feature of most of the folds in the area is that they are confined within the limits of exposures or outcrops, that is to say, along the profile sections the folds cannot be traced across successive layers for long distances and are therefore enclosed above and below by mutually parallel foliations. Thus most of the folds are intrafolial in a broad sense and have well defined 'confining area' or 'profile area' (Fig. 3.5a & b). The effects of these fold episodes are not clearly borne out by the outcrop pattern of the different lithological horizons on the map (scale 1:25000 or larger) of the area (cf. Plate II).

## 3.2 STRUCTURES IN THE JOSHIMATH GNEISS

### 3.2.1 Foliations

In the Joshimath Gneiss, the most common foliation is a gneissosity in which the quartzofeldspathic component is well segregated from the micaceous components.

Thickness of individual bands varies from 0.5 cm. to 10-12 cm. The quartzofeldspathic bands are generally thicker ; however, at places both types of bands are equally thick. This reflects variation in original lithological composition from place to place. In the gneiss the quartzofeldspathic domains are not totally devoid of micas and these micas are commonly of smaller size and are in more haphazard orientation than those making up the micaceous domains where they are in more pronounced preferred parallel alignment. The gneissic foliation has a generally uniform NNE-ward dip throughout the Joshimath area ( see Plates II & III, Fig. 3.6). It is possible sometimes to distinguish different lithologic units even in single exposures (Fig. 3.7a & b). Near the Jogidhara Falls, finescale gneissic foliations group together to form thicker bands (Fig. 3.8a & b); this can be termed as 'multi-order gneissic foliation'. The rock here breaks more easily along the boundaries of the composite thicker bands. Local people use these rock-slabs as roofing tiles. It is very likely that these features viz. divisibility into distinct small scale litho-units, multi-order foliations etc. collectively indicate an influence of primary layering on the development of gneissosity. Another convincing evidence is shown in Fig 3.9. Here the rock is of moderately high metamorphic grade (see Chapter 5) and there are examples that the relict primary structures are sometimes better preserved in high grade rocks than in lowgrade ones (Miyashiro, 1973, p. 75). However the rate of metamorphism and the degree to which deformation is associated with metamorphism are also to be considered as prime factors for this. All these seem to suggest a possible paragneissic character of Joshimath Gneiss. However, further geochemical, mineralogical and textural data would be able to confirm this. Within the Joshimath Gneiss outcrops, there are intermittent quartzite horizons in which the main foliation is sharply defined so as to impart a flagginess in the rocks. Individual micaceous bands are extremely thin. These sharper foliations give a fissile character to the rocks. It is found that some massive landslides of the area have their heads (originating points) within these quartzite exposures at the higher parts of hillslopes; the main foliation served as the landslide surface. In addition there are many fracture-controlled landslides in the area.

The other foliations in Joshimath Gneiss include cleavages or schistositys evidently of later generations than the main gneissosity. Especially in the more northerly outcrops of Joshimath Gneiss, a schistosity is found axial planar to a generation of outcrop scale asymmetric, overturned, close to tight folds (F2). Except for the fold hinge-zones the angular discordance between this schistosity and the main foliation (i.e. gneissosity, here) decreases towards south ; this is a consequence of progressive tightening of the folds in that direction. Cleavage sets corresponding to

still later generations of crenulations or minor folds (F3, F4a) are also recognised. Some of the cleavage sets are of very restricted/local occurrence. The foliations whose status could not be properly established have been collectively termed as 'undifferentiated foliations'. Fig. 3.10a shows the poles of the undifferentiated foliations plotted compositely on the equal area projection.

### 3.2.2 Lineations

Lineations are very common in the Joshimath Gneiss as in the other two formations. Of the different types of lineation, three deserve special mention- (1) mineral lineation, (2) stretching lineation, and (3) intersection lineation. Small scale fold-hinge or crenulation lineations are discussed in a separate section (subsection 3.5). More than one set of stretching lineation and crenulation lineation is present.

In the more northerly outcrops of Joshimath Gneiss, there is a NE/NNE-trending mineral lineation defined mainly by the preferred linear alignment of micas (muscovite and biotite) on the main gneissosity.

A typical stretching lineation of similar trend comes in place of the mineral lineation in the southern parts of Joshimath Gneiss. Stretched garnet grains with their pressure shadows are found defining this lineation on the main foliation (Fig. 3.11). Fig. 3.11a gives an equal area plot of the stretching lineations from the Joshimath Gneiss. The mineral lineation ~~in the~~ north thus grades into a typical stretching lineation towards south. This gradual change in the character of lineation without any corresponding variation in trend/plunge possibly indicates on one hand a broad synchronicity in their development and, on the other, a gradual passage into a thrust or shear zone.

Among the intersection lineations, only those developed on recognisable foliation surfaces have been considered. The ones found on joint/fracture surfaces have not generally been measured. In the Joshimath Gneiss, due to the existence of coarse grained gneissose fabric it is sometimes difficult to readily identify the intersection lineation on gneissosity surfaces. The intersection lineations found in the Joshimath Gneiss normally belong to one of the following groups:

- (1) gneissosity-axial planar schistosity/cleavage intersections;
- (2) gneissosity-undifferentiated cleavage intersections;
- (3) intersections between shear bands.

Most common intersection lineations belong to groups (1) or (2) and normally for ease of operation measurements were made more on the fold-axes rather than on intersection lineations when both were common in the exposures. Lineations whose status could not be unequivocally established were collectively grouped as 'undifferentiated lineations'. Fig. 3.10b shows the equal area projection of all the undifferentiated lineations for the whole area covering all the three formations; many of these are intersection lineations of some sort.

Now a short discussion will be given on the relationship between the stretching lineation and folding. The areal distribution of fold episodes and the general sequence of deformation in the area will be given in tabular form near the end of this chapter. The NNE/NE-plunging main stretching lineation and mineral lineation in Joshimath Gneiss are not folded by the F1 folding, but their geometric relationship with the F2 folding is inconclusive. On the F2 hinge zones in the northerly part of the area neither of the lineation types can be recognised clearly, whereas farther south in the lower Joshimath Gneiss or upper Munsiri Formation the stretching lineation pervasively runs through the limb and hinges of the F2 folds. This means that the lineation cannot be pre-F2, it could be either syn-F2 or post-F2. However, no evidence was found which could suggest that the lineation is either strictly syn-F2 or strictly post-F2. Lineation probably started to develop synchronously with F2 folding and then continued its development even later than the F2 fold episode. The peak development of the stretching lineation probably came about after F2 folding.

### 3.2.3 Folds

Although not equally developed in the outcrops three episodes of folding are clearly recognisable in the Joshimath Gneiss viz. F1, F2 and F3. F1 folds are quite rare. Their occurrence is relatively more common in the northerly parts of Joshimath Gneiss. The most dominant set of mesoscale or larger scale folds (F2) found in Joshimath Gneiss clearly fold the typical gneissosity. Interestingly, the F1 folds also appear to be defined, but rather crudely, by the same gneissic foliation in Joshimath Gneiss (see Fig. 3.12); in other words, fold definition by the gneissosity is not as distinct in case of F1 as for F2. This indicates that possibly there was a phase of fabric-enhancement post-dating F1 but prior to F2 and at the same time points to the difficulty in pinpointing the age of metamorphic differentiation/s in Joshimath Gneiss. The infrequent occurrence of F1 folds compared to F2 implies a strong transposition associated with F1 folding. Effects of strong metamorphism further

complicated the situation. The first metamorphic differentiation obviously took place before F1 folding, but it is not unlikely that the presentday main gneissosity in Joshimath Gneiss developed originally as an axial planar fabric to the F1 folding and subsequent recrystallisation enhanced its development as a well differentiated metamorphic foliation. Fig. 3.12 shows a good example of an F1 fold being refolded by F2.

The F2 folds constitute the most common and dominant group of folds in Joshimath Gneiss. A well developed axial planar schistosity is found to be steeper than gneissosity (main foliation in Joshimath Gneiss) on normal limbs, but gentler on the inverted limbs (Fig. 3.13) of the mesoscale or exposure-scale F2 folds which are southerly verging overturned asymmetric close to tight in nature. The F2 folds are usually multi-order. The congruent geometric relation with the axial planar schistosity holds true for the pumpellyan minor folds as well. Very often the long limbs of the larger scale F2 folds are thicker than the short limbs. Deformation mechanisms operating on the grain scale and the pattern of strain partitioning have been shown to vary from one limb to the other in overturned folds (Mitra, 1978; Ramsay & Huber, 1983, pp.120-124). The short (inverted) limbs are usually much attenuated, often giving rise to small scale thrusting, whereby the antiforms override the complementary synforms (Fig. 3.14a & b). Whereas in the Joshimath Gneiss these small scale slip zones along the short limbs are not occupied by veins, in the Munsiri these limbs are almost always traversed by veins. Hudleston (1989) cited a situation where veins occupied fractures that developed before the folding, or in other words, the folds developed as a consequence of the heterogeneity imparted by fracturing in a simple shear regime. Apparently there is a crude similarity between the kinematic set up suggested by Hudleston for folds with veins and that by Platt and Vissers (1980) for foliation boudinage. When arranged in a series or train the disrupted small scale F2 folds affecting the quartzfeldspathic bands in the Joshimath Gneiss give the appearance of imbricate (or rotated) boudins (Fig. 3.15a & b). The parasitic folds (higher order) are in some instances found to be equally well developed on both the limbs (Fig. 3.16a) or are confined mainly to the long (normal) limbs, the short (inverted) limbs being comparatively straight and thinned in the multiorder F2 folds within the Joshimath Gneiss (Fig. 3.16b).

Small scale tight to isoclinal rootless intrafolial folds are common in Joshimath Gneiss, as also in the Munsiri (to be discussed later). Most of these small scale intrafolial folds in Joshimath Gneiss belong to the F2-episode (Fig. 3.17a & b), though representatives from other episodes, notably F1, are also present. This

suggestion is supported by the spread of the axes of the intrafolial folds on a composite equal area projection (Fig. 3.18c).

The largest directly observable folds in the whole area are found in Joshimath Gneiss and they belong to the F2 episode. These larger profiled folds have their mesoscale equivalents mostly in the form of congruent pumpellyan parasitic folds. It is mainly these mesoscale or hand specimen scale F2 folds which have been used for geometrical classification. Hand-tracings made from enlarged photographs of thin sections or from polished surfaces of hand specimens giving profile-view of the folds were used for the purpose. Class 2 (similar fold) geometry is found to be quite common among the F2 folds (see Fig.3.20c). Such a geometry is particularly common where the competence-contrast (as determined from the mineral composition) between adjacent folded layers is low or negligible. Where this contrast is high or there were irregularities in the original thickness of the layers, fold geometry is quite irregular (see Figs. 3.22 & 3.23). Further discussion on the geometric patterns and some peculiarities in them will be given later. At places (eg. Loc. 4'87), where more or less homogeneous siliceous lithology is exposed within the Joshimath Gneiss, some mesoscale F2 folds in the exposure appear like the following (Fig. 3.19).

These folds, which might escape notice in the field, are defined by the zones of closespaced, mica-rich, fissility surfaces which are mutually parallel as well as parallel to the axial plane of the fold itself, and physically they resemble slip/shear folds (card-pack model). It seems to be logical to presume that a shearing mechanism was responsible for the development and/or modification of the F2 folds. Clear-cut evidence of the F2 folds folding the stretching lineation is lacking. This implies that for the most part the stretching event is not pre-F2. For further discussion on this, see subsection 3.2.2.

In the Joshimath Gneiss, the F3 episode was dominantly an episode of small scale folding. They are common in its southern parts nearing the contact with the Munsiri. A distinct set of crenulation belongs to this episode.

Now the results of a geometrical classification of the F2 folds following Ramsay's method (1967) are discussed. Both the isogon method and thickness-ratio method have been applied. The discussion is based on three examples, all taken from the Joshimath Gneiss: Specimen Number 59'89 (fig. 3.20a), Sp. No. 58'87 (Fig. 3.20b) and Sp. No. 22/4/88B (Fig. 3.20c).

Sp. No. 59'87 (Fig 20a) The geometrical analysis of this fold gives many insights into the relative merits of different fold classification schemes. Even the most modern scheme for fold classification suggested by Ramsay (1967) cannot account for certain geometrical aspects of these folds. Here it is found that the results of classifying a folded layer using the dip isogon method do not properly match with the results obtained from thickness ratio plots. Band 1 is of class 1C geometry as indicated by both the isogon pattern and thickness ratio distribution; but in band 2, even though the isogon pattern suggests class 3 geometry, the thickness data indicate that on the right limb the folded band shows class 1C geometry whereas the left limb shows class 3 geometry. The same is true of band 4. This feature is not properly highlighted in literature and is probably indicative of initial thickness variation in the bands resulting from an earlier deformation event (F1 folding). Bands 3 and 5 were parts of a single straight and continuous band initially. Despite the F1 folding, the isogon pattern and thickness variation due to F2 folding are quite similar in both bands which is attributed to the relatively greater competence and hence more active response to folding. Broadly similar are the patterns for bands 7 and 8, but in bands 2 and 6 which were also originally continuous the fold-styles are not similar to each other. While band 2 shows class 3-1C geometry band 6 shows class 1A geometry. Most probably this is due to the fact that these bands are made of comparatively incompetent materials and responded rather passively during folding.

Another important feature to note is the selective/restrictive discontinuity of dip isogons. See, for example, the behaviour of the 45°-dip isogon on the left limb. This 45°-dip isogon is corresponding approximately to the steepest dip of the left limb in bands 1, 2 and 3, and also for band 8 and in its immediate surroundings, whereas bands 4, 5, 6 and 7 have no 45°-dip component and so no corresponding dip isogon. The 45°-dip isogon therefore runs through bands 1,2 and 3, but disappears across bands 4, 5, 6 and 7 only to appear again from below band 7. On the right limb a slightly different, but more usual situation is developed. Here the successive bands become gentler as we go towards the core region of the fold; consequently the isogons corresponding to gradually steeper dips become shorter and therefore pass across fewer bands. For instance, 0°- and 15°-dip isogons pass through all the bands even beyond band 8, whereas 30°-isogon stops just below band 8, 40°-isogon continues only upto the top of band 5, and 60°-isogon stops at the top of band 4. This latter situation is discussed in literature (Hudleston, 1973; Ramsay & Huber, 1987, p.356), but the former peculiarity is neither documented nor explained. Probably this is an extreme case where initial thickness variation has greatly influenced the overall

fold geometry. Also a likely corollary may be that the F2-strain is not large enough to wipe out the influence of F1 or earlier deformation.

The thickness distribution pattern for all the bands (the whole  $t'\alpha$ - $\alpha$  plot) looks like 'a long-legged spider in motion' reflecting the complex shapes of these folded bands. Ramsay (1967, pp.369-371) suggested that first-order and second-order derivatives of  $t'\alpha$  with respect to  $\alpha$  be used in order to bring out the relative influences of different classes in the overall geometry of such folded bands. However, for detailed classification one might need to go for repeated differentiation well beyond second-order and so the method is still not very precise; successive differentiation in this example would bring in more noise and therefore the graphs using the derivatives would be less well defined and less significant.

Sp. No. 58'87 (fig. 3.20b)- Here except for band 1, the bands are considerably thinner on the right limb than on the left limb. This discrepancy is more marked in the pelitic bands than in the quartzofeldspathic bands. One likely reason for this sort of thickness variation is the involvement of at least one earlier episode of folding the hinges of which are not present in this specimen. Compare the results and sketch from Sp. No. 59'87 (Fig. 3.20a) where the early fold hinges are clearly preserved. These two specimens were collected within a distance of about 200 metres from each other. In the present scheme for the classification of folds into different classes according to Ramsay, scope for handling this aspect is limited. A cursory look at the specimen (sketch) would indicate a fairly regular and systematic fold pattern but in reality the  $t'\alpha$  plots show wide scatter in the thickness data. Note how markedly the folds (graphs) cross the class boundaries. Obviously there must have been a strong variation in band thickness prior to the F2 folding. This shows that even when dealing with similar lithology (similar layer-competence), caution must be exercised in the attempts to characterise a particular fold episode in terms of fold classes. This is especially so in areas of refolded folding.

Sp. No. 22/4/88B (Fig. 3.20c)- As evident from the fold-sketch showing the dip isogons in the three different bands, there is obvious difference in the isogon-pattern between the quartzofeldspathic bands and the pelitic band. Even in a single band (e.g. Band 1), the isogon pattern in the antiform (class 1) is different from that in the synform (class 3). The isogons in the quartzofeldspathic band are either convergent or divergent thus showing either class 1 or class 3 pattern respectively, whereas those in the pelitic band are mutually parallel suggesting class 2 (similar fold) geometry. Such difference in the isogon patterns in the bands of different composition obviously

indicates the difference in their competence, mica-rich layers being less competent than the siliceous layers.

Thickness distribution (t'a - a plot) in the folded layers corroborates the above findings. The graph for band 2 closely follows the curve for class 2 (similar fold), and those for the two limbs of the synform (band 1) lie mostly within the class 3 field, although very near to the class 2 curve (except for the steepest part of the left limb). However, closer observation shows that the graph for one limb sometimes falls in a field which is different from the field for the other limb even for the same band (eg. bands 1 and 3, antiform) and also that the graph occasionally crosses the fold fields even within a single limb (viz. left limbs for band 1 for both the antiform and the synform). Such deviations, however, are not unexpected in natural folds (see Ramsay, 1967; Ramsay & Huber, 1987; Hudleston, 1973).

All the graphs lie within a narrow zone around the curve for class 2 folds. This probably points to the fact that a shearing mechanism (card-pack model) has influenced the development of these folds and the mechanism was broadly the same for all the layers.

Note the value of interlimb angle ( $40^{\circ}$ - $50^{\circ}$ )- close fold (Fleuty, 1964).

The above discussion on fold classification indicates two things:

- (a) though in many cases fold geometry may apparently seem to be uniform, in detail they may be quite irregular;
- (b) the methods of classifying natural folds still need much improvement.

### 3.3 STRUCTURES IN THE MUNSIARI

#### 3.3.1 Foliations

The gneissose fabric of Joshimath Gneiss is found to be gradually replaced by a schistose fabric as we enter into the top part of the Munsiri Formation. Farther down into the Munsiri Formation, there appears, depending on the lithology involved, a mylonitic-protomylonitic foliation or a prominent phyllonitic schistosity representing the main foliation in the rocks. However, in the augen gneiss horizons the main foliation is a gneissosity. The change-over in the character of the main planar fabric between Joshimath Gneiss and Munsiri is more of a function of

tectonism involving retrograde metamorphism, rather than being due to gradually northward increasing grade of a single phase of progressive metamorphism (this point will be discussed in detail in Chapter-5). The average orientation of the main foliation in the Munsiri is not appreciably different from that in the Joshimath Gneiss or Berinag-Mandhali (compare Fig. 3.21a with Figs. 3.6 & 3.21b). As mentioned above the dominant planar fabric in most of the lithologies in the Munsiri is a tectonic (mylonitic) foliation whereas in the phyllonitic horizons its equivalent is a schistosity. There is a slight difference in the amount of dip between the <sup>fraction</sup>schistosity of phyllonitic horizons and the mylonitic foliation in other horizons. This difference is interpreted as due to refraction (Fig. 3.22). Depending on the variation in lithological composition the foliation-defining minerals may vary. The main foliation in augen mylonite-protomylonite is characterised by dark, thin (1-2 mm.) phyllosilicate-rich bands alternating with thicker bands of streaked-out feldspars, platy quartz, elongate feldspar augens and lenticular quartzofeldspathic aggregates (Fig. 3.23). In the amphibolite horizons, the amphibole grains are normally elongate and arranged, more or less, parallel to one another thus defining the main foliation and a crude lineation on it (L-S fabric). The amphibole-rich bands alternate with bands rich in quartz and feldspar grains which are normally platy in nature. In the feldspathic horizons of semipelites as also in some quartzites a distinct flagginess is observed. The phyllonitised horizons show a very prominent *foliation* which is usually sigmoidally wavy in nature. Even though the mylonitic/protomylonitic foliation is almost strictly parallel to the lithological boundaries, the main *foliation* in phyllonites is slightly steeper, probably indicating a kind of refraction (see Fig. 3.22). Some lithological bands, especially some siliceous semipelite horizons, apparently do not show any colour or compositional banding and appear fine-grained, massive, bluish grey in color, but on closer observation in the field a distinct layer-parallel planar tectonic fabric is identified. This planar fabric is an expression of the preferred shape fabric of the constituent mineral grains mainly in quartz and feldspars. In the strongly schistose phyllonite bands, interference by other foliations is easily recognisable with clear evidence of repeated transposition (Fig. 3.24).

The Munsiri rocks, in general, show a greater variety of foliation than do the overlying Joshimath Gneiss and the Berinag-Mandhali below. This is particularly true if we consider the phyllonitic horizons in the Munsiri where the main foliation (schistosity) is transected often by several foliations. Distinct shear bands are observed to lie tangential to the extremities of 'S'-shaped segments of sigmoidal/wavy schistosity (Fig. 3.25). Wherever present, these shear bands are found to have a uniformly low NE'ly dip (see fig. 3.26a and Plate III).

### 3.3.2 Lineations

The stretching lineation referred to in subsection 3.2.2 is present south of the Vaikrita boundary; indeed it gains prominence in the Munsiri which shows the acme of development of the stretching lineation in the area. In the quartzite horizons of the Munsiri Formation the stretching lineation is very pervasive. A stripiness on the main foliation defines the stretching lineation within the augen gneiss horizons of the Munsiri. The stripiness is due to the concentration of dark coloured micas (mainly biotite) in broadly rectilinear zones which swing around intermittent bulges on the foliation surface enclosing augens of feldspars or quartzofeldspathic aggregates. The bulges are aligned in linear fashion defining the alternate light coloured (whitish) zones. Three-dimensional exposures in the field show that the augens (elliptical porphyroclasts) have their longest dimension parallel to the stretching lineation. More than two-thirds of the porphyroclasts in the augen gneiss are elliptical in outline, <sup>while</sup> the rest are circular or rectangular when observed on rock-surfaces nearly at right angles to the main foliation but subparallel to the stretching lineation (Fig. 3.27). The rectangular porphyroclasts are oriented so that a diagonal is often parallel to the stretching lineation and in rectangular rhombus-shaped porphyroclasts it is normally the longer diagonals that are parallel to the stretching direction. Yet the long dimensions of some of these regular geometric shaped porphyroclasts show the maximum deviation from parallelism with the stretching lineation, whereas this is never the case with the elliptical or sigmoidal porphyroclasts which are invariably conformable to the stretching direction (Fig. 3.28). It therefore seems likely that deformation did not modify the outlines of the regular geometric shaped porphyroclasts; in their case the effect of strain is probably expressed through body rotation, whereas the elliptical or sigmoidal outlines of the porphyroclasts must indicate modification through deformation.

The stretching lineation is more clearly visible where the rock is relatively less phyllonitised. Indeed in some phyllonites it is often difficult to identify the stretching lineation conclusively and for practical purposes during mapping the lineation (normally due to mica-alignment) that corresponds in orientation to the stretching lineation observed in nearby exposures has been taken to stand for stretching lineation. As mentioned in the earlier chapter, <sup>the</sup> proportion of phyllonitic rock increases towards the bottom of the Munsiri Formation, hence the typical NE/NNE-trending stretching lineation is observed rather infrequently towards the lower part of the Munsiri Formation; nevertheless in the intermittent competent

horizons of mylonite and amphibolite here the NE/NNE-trend of stretching lineation is well preserved.

The stretching lineation is defined by features such as parallel alignment of stretched grains of garnets, feldspars etc. or quartzofeldspathic aggregates, preferred alignment of pressure-shadow trails around porphyroblasts/porphyroclasts, also quartz roddings, and alternate grooves and ridges on the main foliation surface (Fig. 3.30b). In the field, the stretching lineation is most commonly identified on the surface of the main foliation; elongate flakes of biotite, muscovite, sericite, etc are aligned parallel to it. At places, a high degree of stretching in the rocks is indicated in sections perpendicular to foliation but parallel or subparallel to stretching lineation by the presence of fairly homogeneously distributed thin lenticular quartzofeldspathic fragments.

Around locations NR51, 52 etc. (see Plate III for location) within the Munsiri outcrops, a distinct N-trending stretching lineation is observed on the main foliation in the semipelitic schistose phyllonites. It is defined by long pressure shadow trails lying only on one side (up-plunge side) of unstretched garnet grains. When the main foliation is viewed orthogonally up the plunge of the lineation, these garnet grains with their pressure-shadow trails look like 'falling fireballs' (see Fig. 3.29d). In similar situations with the NE/NNE-trending lineation farther north, there are grooved trails lying on the down plunge side of the garnet grains seen on the main foliation surface. Together with these trails the garnet grains look like 'a swarm of upward moving tadpoles in water' (Fig. 3.30a). These two different features indirectly point to the difference in the flow behaviour of the rocks during the two stretching events; this must have been due to a change in the ambient P-T conditions. At the time of development of the N trending stretching lineation, the P-T condition of the rocks was not high enough to allow purely ductile stretching/shearing; the condition was lowgrade (probably nearer the surface of the earth) and the stretching event took place in a quasiductile-semibrittle environment. This N-trending stretching lineation appears like slickenside grooving on the retrograde chlorite-rich surfaces of the main foliation. It is not as pervasively developed as the NE/NNE trending stretching lineation which is considered to represent the regional stretching lineation. Where the N trending lineation is strongly developed, we cannot readily identify the NE/NNE trending stretching lineation; but one can fairly easily see the presence of the N trending lineation even where the NE/NNE trending stretching lineation is quite strongly developed, particularly in the lower part of the Munsiri. Metamorphic retrogression has altered the minerals defining the NE/NNE lineation which is almost

obliterated where the N lineation is strongly developed. This implies that the N trending stretching lineation is later than the NE/NNE trending one. Figs. 3.29a & b indicate on equal area projection diagrams the orientation distributions respectively of the main stretching lineations in the Munsiri Formation and later stretching lineations dominantly from the same formation (compare also the composite plot of the main stretching lineation from across the whole of Joshimath area in Fig. 3.29c)

#### Lower boundary of the Munsiri Formation vs. Stretching lineation:

Unfortunately there is a lack of good and accessible in-situ exposures around the lower boundary of the Munsiri Formation, i.e. the contact between the Munsiri and the Berinag-Mandhali formations. However, it is evident that near the boundary zone the NE/NNE lineation is infrequent, and more northerly trending lineation becomes prominent farther south. The available field evidence suggests that-

- (1) the presence or absence of the NE/NNE trending stretching lineation should not be the only criterion to be used for demarcating the lower boundary of the Munsiri Formation. The definition of the boundary must also be based on the differences in microstructural character of the lithologies involved and also in the character of other deformational features (mesoscopic as well as microscopic).
- (2) the boundary might be a diffused one or transitional in character unlike the classic discrete thrusts such as the Moine thrust or the different Alpine or Rocky Mountain thrusts.

As found in the foot track section between Salur village and Gulabkoti, at the sharpest bend in the track on the spur axis south of Helang at 6930'-level (~2112 m), a fine elongate mica alignment lineation, NE/NNE trending (plunge 28° towards 31°), is present on the main foliation in an approximately 3.5 m. thick greenish white coloured quartzite horizon. This quartzite lies below a low-grade greenish phyllonite horizon which is over 15 m in thickness and contains biotites that are almost completely chloritised. No conspicuous lineation is found in the phyllonite horizon and this, in turn, lies below an augen mylonite horizon (thickness not precisely known) which shows distinct stretching lineation plunging 36° towards 18°. Obviously this trend is somewhat different to that of the characteristic NE/NNE lineation. The contact between this augen mylonite and the phyllonite is not exposed. The above mentioned quartzite band is also underlain by a thick low grade greenish coloured phyllonitic rock horizon below which appear still thicker horizons of carbonate and quartzite; in these lower quartzite horizons, a quite distinct stretching

lineation is observed on the main foliation plunging towards  $15^{\circ}$ - $18^{\circ}$  (cf. the section between Helang and Gulabkoti in Plates II, III & IV).

Based on field-data alone, two possibilities can be suggested:

- (a) the abovenoted augen mylonite should be regarded as the lowermost horizon of the Munsiri Formation; or
- (b) the Munsiri Formation ends with the phyllonite horizon just overlying the 3.5 m thick band of quartzite.

Which of these two possibilities is to be favoured and on what further grounds will be discussed in next Chapter.

In the Munsiri, the types of intersection lineation are the same as in the Joshimath Gneiss. But the main foliation on which these are developed is not a typical gneissosity everywhere as in the Joshimath Gneiss.

The hinge of shear bands is much more common in occurrence in the Munsiri than in the Joshimath Gneiss or Berinag-Mandhali. It is normally the phyllonitic rock-horizons that show the shear band hinges most prominently.

### 3.3.3 Folds

Folds belonging to the F1 episode recognised in the Joshimath Gneiss are difficult to find in the Munsiri. Representatives of F2 folds are present, but are less common than in Joshimath Gneiss. Generally, the F2 folds found in the Munsiri Formation are of diminished size, with their axes reoriented to parallelism with the main NE/NNE trending stretching lineation (see Fig. 3.3); these folds also show appreciable tightening gradually towards south. There is some doubt as to whether the isoclinal intrafolial small scale (measured in cms) folds found just N of the newly constructed Helang bridge on the Karmnasa Nala (river) can be properly correlated with the F2 folds. If the correlation is correct, then these isoclinal folds would represent the extremely tightened version of the F2 folds. It must be noted here that this fold-exposure is within only ~300 m above Gansser's MCT. In a similar set-up at the Almora klippe which is considered as a part of the MCT sheet, Bhattacharya & Siawal (1985) have shown that the folds get tightened as the thrust is approached, though it is not clear from their paper whether all those folds belong to the same episode.

F3 folds are more common in the Munsiri Formation than in the Joshimath Gneiss. However, it is the F4 folds that are most easily recognisable and most

dominant folds within the Munsiri. The type-exposure of the F<sub>4</sub> folds can be found at Locs. 17, 18. In some close, angular F<sub>4</sub> folds, 'S'-shaped parasitic folds (observed looking eastward) are developed only on the overturned limb and thin quartzose veins are found to run through its main axial trace (Fig. 3.31). Within the mylonite horizons in the Munsiri Formation, there are F<sub>4</sub> folds in which a progressive increase in tightness and an increasingly well developed axial planar fabric towards the fold-core are observed. This axial planar fabric is locally intense enough to completely transpose the mylonitic foliation. In some instances this axial planar cleavage shows fanning. In another instance, mylonitic bands of different thicknesses define folds of different wavelengths (Fig. 3.34). These two features i.e. progressive tightening towards the core (reflecting class 1A-1B fold geometry of Ramsay) and thickness-dependent wavelength variation of folds defined by different bands strongly suggest that these folds have developed by a buckling mechanism. Another point suggesting buckling is the presence of curved stretching lineation on the folded surface maintaining a more or less constant angular relation with the fold axis-- the measured angle was 36° in one case. No evidence was found which could clearly indicate whether flexural slip or tangential longitudinal strain was the specific mode of internal deformation in these folds. Later modification by flattening strain was probably not very significant. The curving of stretching lineation around the hinges necessitates that the F<sub>4</sub> folds developed later than the stretching lineation. F<sub>4a</sub> (from orientational attributes and spatial closeness these folds seem somewhat akin to F<sub>4</sub> folds, hence a separate term is not used) crenulations and the corresponding axial planar cleavage are rather heterogeneously developed in the area; a distinct lithological control on their development is very apparent. Mainly the mica-rich semipelite and phyllonite horizons show their presence. In a way, the inhomogeneous development across different lithologies is quite a characteristic of all the different fold-episodes in the area.

The following two paragraphs deal with the results of geometrical analysis for classifying an unassigned (probably F<sub>4</sub>) fold developed in an amphibolite horizon in the Munsiri Formation (Fig. 3.35).

Slide-Loc. 18(2)'87 ( ?F<sub>4</sub> fold in Munsiri amphibolite; Fig. 3.35)

Fold trace was sketched from the negative print of thin section. Note how the general fold class defined by different bands changes towards the core (from class 3 to class 1); however, excepting the fold defined by band C, no other clearly falls strictly into a single class; often fold class indicated by one limb differs from that of the other. Difficulty arises while trying to draw dip isogons to the digitations in band

C, except for the one exactly at the hinge. These minor folds die out within the band C and thus make the measurement of their limb-thickness impossible. To establish the general fold-pattern a surface is drawn freehand enveloping the crests of minor folds in band C. For practical purposes, this enveloping surface (dotted line in Fig. 3.35c) is considered as the upper boundary of band C. Band D is also treated in a similar way. The minor folds in band D, however, are analysed separately as indicated; a common tendency in their geometry is to lie around class 2; isogon-pattern reveals this well.

An important point is that though, in a strict sense, classes cannot characterise the different fold episodes (Hobbs et al, 1976; Ramsay, 1967; Ramsay & Huber, 1987), it is found that each episode is dominated either by one particular class or, more commonly, by a combination of two contiguous classes of folds.

### 3.4 STRUCTURES IN THE BERINAG-MANDHALI

Only a small part of these formations are included in the present study. In the following paragraphs the structures observed in them are described in the same order as for the previous two formations but without designating them as separate sections.

The main foliation (which is a planar schistose fabric) seen in the Berinag-Mandhali formations is more closely spaced and more readily recognisable in the lowgrade phyllites than in the alternating greenish-grey quartzite horizons where it is rather obscure. However, in the carbonate horizons and in the milky white sugary-textured (granular) quartzites farther south a distinct flagginess characterises the main foliation.

The Berinag-Mandhali rocks show a dominant stretching lineation which is somewhat intermediate in trend between the NE/NNE lineation and the N lineation of the Munsiri (see Fig. 3.51 and compare with Figs 3.29a & b). This poses a great difficulty in correlation of stretching lineations across the Munsiri/Berinag-Mandhali contact. The lineation in the Berinag-Mandhali is not developed on a retrograde foliation surface. On grounds of great similarity and persistence in orientation, one could argue in favour of correlating this lineation with the N-trending later stretching lineation of the lower Munsiri. If this correlation holds true, then two very important inferences can be drawn

- (i) There are evidences of superimposed shearing events in the area;

(ii) Focus of deformation (shearing) migrated southward, and with time the condition of deformation changed from more ductile to less ductile for progressively younger events.

On the other hand, if this lineation represents a slightly rotated version of the dominant NNE trending lineation of farther N, then a wider tectonic implication involving rotatory migration of <sup>the</sup> Indian plate may be invoked. More discussion on these aspects is given in Chapters 4 & 6.

The Berinag-Mandhali rocks show a very spectacular set of intersection lineations developed as a striping on the main foliation. The lineation is due to the intersection of compositional layering with an axial planar schistosity. Owing to a high degree of transposition the axial planar schistosity is the main foliation in the rocks. Crudely developed shear band hinges are recognised, especially in the semipelitic (schistose) limestone horizon about 750 m south-west of Helang Bus Stop, at the northern part of a high road-cut cliff exposure where the road takes a very distinct bend. In the southern part of this exposure, the abovenoted striping lineation is most readily recognised.

The three main sets of folds recognised in the Berinag-Mandhali Formation can be clearly observed from the same road-cut exposure. The earliest set of recognisable mesoscale folds are tight to isoclinal with axes plunging to the NW and strong axial planar foliation transposing the earlier compositional banding. The main foliation in the exposure is represented by this axial planar schistosity. This folding, however, does not represent the earliest deformation/recrystallisation suffered by these rocks. Microscopic study reveals that parallel to the compositional layering there is a micaceous metamorphic foliation which has been crenulated by the above folds. The second set of folds are close to tight semi-reclined in nature plunging to slightly W of N. Associated with the second folds are well developed mullions in the quartzite horizons and crenulations in phyllosilicate-rich horizons. The third set of folds are open asymmetric, nearly non-plunging with steep NE dipping axial planar spaced crenulation cleavage.

A group of asymmetric southerly verging folds with uniformly spaced, mutually parallel planar quartzose veins occupying the short inverted limbs could not be properly placed in the fold sequence; these folds found in the quartzite are most likely to have resulted from a late phase of shearing. See Fig. 3.5 showing a series of systematically oriented dilatational shear veins intimately associated with the folds.

### 3.5 SMALLSCALE FOLD HINGE OR CRENULATIONS AND OTHER UNASSIGNED STRUCTURE

#### Smallscale fold hinge or crenulations

With differing intensities these structures are present in all the three formations. The fact that all the rock horizons in this area dip towards N-NE and that the larger scale folds are all overturned towards south and are mostly close to tight in nature makes it easier to group the crenulations into different sets according to their orientation, coupled with the character (roundness) of the hinge zone, the interlimb angle and the character of the axial planar structures. However, due to a general lack of direct interference relations among these different sets, it is difficult to place them into a definite time-sequence. For this reason, <sup>an</sup> attempt is made to establish whether a particular set of crenulations or minor folds is related to or equivalent in terms of orientation to any particular phase of major folding (mesoscale or outcrop-scale) and by this way the different sets of crenulations have been fitted into a time-sequence as indicated below. Most are found to correspond to the major fold episodes. Sets developed on a very restricted (local) scale are not described here.

C2 (= F2): Tight or close semi-reclined crenulations with moderately steep, broadly NE-dipping axial planar cleavage and axial plunges to NE or E.

C3 (= F3): Tight to close inclined-type crenulations N plunging, with moderately steep NE/ENE-dipping axial planes.

C4a (= F4a): Crenulations asymmetric, close to open type, E plunging and with N dipping axial planar cleavage at an angle  $>50^\circ$ .

The above sets of crenulations are not represented in the Berinag-Mandhali where, as already mentioned, a distinct set of crenulation has horizontal axes (trend  $127^\circ$ ) and are open asymmetric with steep NE-dipping axial planar spaced crenulation cleavage. There is a marked scarcity of upright crenulations in the area; most of them are with inclined axial planes.

### 3.6 BOUDINAGE

This phenomenon is exceptionally well exhibited in the rocks of the area and on a wide range of scales. Boudinage is more common in the Munsiri and Berinag-

Mandhali formations than in the Joshimath Gneiss. Most of the boudinaged rock-bands are made up of amphibolites and the enclosing rocks are semipelites, mostly phyllonites. However, quartzites and intrusive aplite or pegmatite bands also often show boudinage. Though not strictly looking like classical boudinage, quartzofeldspathic segregation bands as well as veins are highly stretched and torn apart giving rise to lenticular fragments occurring fairly pervasively in the pelitic horizons in the lowest part of Joshimath Gneiss and in the Munsiri in particular. Veins, particularly the stretched and fragmented ones, are quite rare in the Berinag-Mandhali rocks.

Boudins are long, slightly flattened sausage-shaped or cylindrical objects lying side by side and made up of materials more competent than the matrix surrounding them. They were first described from Bastogne, Belgium by Lohest, Stanier et al (1909). Since then different workers dealt with such structures from various parts of the world, both theoretically and in experiments (Quirke, 1923; Holmquist, 1931; Wegmann, 1932; Read, 1934; Cloos, 1947; Ramberg, 1955; Rast, 1956; Coe, 1959; Sanderson, 1974; Hambrey & Milnes, 1975; Platt & Vissers, 1980; Lloyd & Ferguson, 1981; Sengupta, 1983; Mandal & Karmakar, 1989; etc.). Boudinage, as a tectonic process, is defined as involving layer-parallel extension of a more competent layer enclosed in a more incompetent medium. The extension could result in tensile failure producing fractures separating individual boudins (discontinuous boudinage), or, in differential thinning producing 'pinch-and-swallow' structures (continuous boudinage). The tension giving rise to boudinage may be direct, or, more commonly, indirect due to, in fact, a compression acting in a perpendicular direction or at an oblique (high) angle to the layer to be boudinaged, as indicated in the experimental and theoretical studies by Ramberg (1955), Griggs & Handin (1960), Paterson & Weiss (1968), Stromgard (1973), Gay & Jaeger (1975), Selkman (1978) etc. When tension can operate freely in all directions parallel to the competent layer during the same boudinage-event, 'chocolate-tablet boudinage' results (Wegmann, 1932; Ramsay, 1967, pp.112-113); similar features might probably also develop due to superposed extensional deformation.

Following Wegmann (1932) and Wilson (1982), a boudin typically has the following dimensional properties: length or long axis, width and thickness; another attribute i.e. 'separation' between adjacent boudins is also taken into account. The length of a boudin is also called boudin axis. The orientation of this axis coupled with the magnitude of 'separation' and the shapes of boudins especially around the separation-zones have important tectonic significance (as discussed in the following).

When 'pinch-and-swell' structure develops due to either low competence-contrast between the boudinaged layer and the enclosing matrix, or to a weak tension, neck-zones or pinch-in zones develop in place of boudin separation zones. In progressive development of boudinage, initial necking leads to rupture and then separation of the boudins takes place along rupture surfaces. The boudin separation zones widen with progressive straining.

As boudinage is normally developed due to layer parallel extension in a direction perpendicular to the (potential) boudin axis, so the orientation of boudin axis is measured to determine the local extension direction. The orientation of boudin neck-zones or boudin separation zones generally corresponds to the orientation of boudin axis. When, as often happens in field exposures, it is difficult to measure the boudin axis directly, the orientation of boudin separation zones or neck-zones are normally measured and taken to represent the boudin axis orientation. It should be mentioned, however, that Burg & Harris (1982) claimed to have found boudinage that developed at oblique angles (mostly at  $65^{\circ}$ - $70^{\circ}$ ) rather than at  $90^{\circ}$  to the maximum extension direction. Hobbs et al (1976, p.283) comment "The presence of elongate boudins indicates that the direction parallel to the boudinaged layer and perpendicular to the length of the boudins was a direction of extension during at least part of the deformation"; they point out the possible role of deformation history in the apparent reorientation of boudin lines with respect to the direction of maximum finite bulk elongation.

Mostly we see only the cross-sections of boudins in exposures. From characteristic profile-shapes of boudins in the boudin profile plane and the pattern of arrangement of matrix-material around the boudin separation zones, we can make qualitative estimates of the relative competence of the rock-layers involved and in some cases also of the relative ratios between the principal extensions of the strain ellipse in two dimensions (Ramsay, 1967, pp.104-106).

Fig. 3.36(reproduced from Ramsay, 1967, p.106, fig.3-44) clearly indicates that boudinage as seen in stage III can be broadly classified into three types--

Brittle boudinage (exemplified by band 1)

Semi-brittle boudinage (exemplified by band 2), and

Ductile boudinage (exemplified by band 3).

Even though this is a subjective scheme of classification, it has obvious practical value.

Thus barrel-shaped or rectangular boudins with slightly curved or nearly flat edges indicate brittle boudinage; a lenticular boudin-shape is indicative of semi-brittle/quasi-ductile boudinage, and attenuated lenticular shape of boudins with highly pointed edges indicate ductile boudinage. Obviously, the mutual contrast in ductility or competence between the boudinaged bands and their enclosing matrix is the main reason for the development of different types of boudinage. Boudinage cannot develop when the competence of the band-material is equal/comparable to that of the matrix. Homogeneous thinning occurs in such a case (e.g. band 4 in Ramsay, 1967). Brittle boudinage takes place when the competence-contrast is high, ductile boudinage at low contrast and semi-brittle boudinage at an intermediate contrast. Also with change in physical conditions of deformation there is a change in the type of boudinage. It has been shown that even with a particular band-matrix combination the rock-band showing brittle boudinage at one place might show ductile boudinage at another place due to a change in ambient physical conditions (Ramsay & Huber, 1987, pp.399-401). Sengupta (1983) showed good examples from Vaddo area, Sweden and Chhotanagpur area, India of brittle boudins that have behaved in ductile manner at a later reactivated stage (superposed deformation). All these reported changes in behaviour of a particular rock-band at different places or in different times are due to metamorphic transformation of the minerals constituting the rock in response to changes in tectonothermal conditions.

In the Joshimath area, the boudin thicknesses differ in profiles in different bands due to the difference in the original band-thickness; however, within bands the size variation of boudins is comparatively small. As far as the dimensions of individual boudins are concerned, it is difficult to find out their lengths from the rock exposures; but in cross-sections, the longer dimensions are always more than or equal to 1.5 times the shorter dimension. The scale of profile-section of individual boudins is from centimetres up to that of an exposure. Boudins of all the different types viz. brittle, semi-brittle and ductile are found in the area (Figs. 3.37a, b, c, 3.38a, b, and 3.39a, b). There is little systematic variation or sequential arrangement of different boudin types across the area, but in general, semibrittle-ductile boudinage appears to be more common and boudin separation zones widen as we go southwards across the Munsiri Formation. It is also important to note that post-boudinage deformation and metamorphism have a more serious obscuring effect on boudin-matrix relationships towards the lower part of the Munsiri Formation than in its upper part or in the Joshimath Gneiss. There is wide variation in modal composition, grain-size and texture among the boudinaged rock-bands. No boudinaged layer has been found without distinct metamorphic foliation and this foliation is most spectacularly

affected due to boudinage. Thus boudinage cannot be earlier than this foliation and in all likelihood there was no appreciable time-gap between one type of boudinage and the other. Varying viscosity-contrast was the main reason for the development of the various types of boudinage viz. brittle, semibrittle and ductile, and a southward increasing magnitude of boudinage-strain played a modifying role. (At places, only 'pinch-and-swell' structures have developed instead of complete boudinage). In the rock-bands showing brittle or semi-brittle boudinage, the boudin separation zones are frequently occupied by quartzose vein-materials. Orientations of boudin axes have been measured often indirectly. In most cases, the orientations of boudin neck-zones or of boudin separation lines/zones have been measured and considered to give the orientations of boudin axes. Boudin axes have generally a low plunge (equal or less than  $30^\circ$ ) and their modal trend is spectacularly perpendicular to the mean trend of the main stretching lineation (see Fig. 3.26b; cf. Fig. 3.29c). This means the main boudinage event took place at the time of the main stretching lineation.

An interesting kind of structure which may be termed 'inverse boudinage' has been found (Fig. 3.40); they resemble mica-fish in a large scale and presumably develop when relatively thin incompetent layers sandwiched between thicker competent layers undergo extensional deformation. In literature due attention has not yet been given to this type of structures which may well develop side by side with conventional boudinage, given proper setting of lithology.

Interestingly, in a number of instances of semibrittle boudinage occurrence of 'small scale boudins within larger boudins' has been observed (Fig. 3.41a & b). The small scale boudinage appears as nodding in well foliated rocks; nodal fractures and nodal veins are also commonly associated. It is found that in addition to the well developed tectonic (mylonitic-phyllonitic) foliation and distinct stretching lineation on it, the other very significant class of small scale structure that characterises most of the rocks of the Munsiri Formation includes such nodes/ nodal fractures which occur in almost all the relatively mica-deficient fine grained rock-horizons. This is a special kind of boudinage where within a single lithological horizon a particular bunch of main foliations is necked or noded, and these necks/nodes are not arranged in a series involving the same bunch of foliation everywhere. Mostly, along such nodes there are quartzose veins and local bending of the mylonite/protomylonite foliation. Where no vein-material occupies the node, there is a more or less clean cut nodal fracture and a little void with some discoloration (due to localised mineral transformation induced by P-T change) of the rock immediately around it. On closer observation in the field, it is recognised that such nodding/boudinage is usually

involved with ductile-brittle microthrusting and the foliation around it is folded to some extent. Almost always it is the lower (northern) boudin/segment that shows an antiformal bend, whereas the upper (southern) segment/boudin shows complementary synformal bend near the thrust/neck. Often the rupture is along a very smooth surface and there is a gap which is occupied by the veins. The apparent fold-hinges (antiformal ones) developed around the nodes have subhorizontal plunge ( $8^\circ \rightarrow 284^\circ$ ;  $9^\circ \rightarrow 300^\circ$ ); this is consistent with a SSW-directed overthrust shear affecting the rocks. Each of such individual nodal zones associated with folding seems to represent a 'break thrust zone' in miniature.

Features similar to the nodes mentioned above have been described in literature as 'foliation-boudinage'. Hambrey & Milnes (1975) first described them from glacial ice. Platt & Vissers (1980) found similar structures in strongly foliated quartzofeldspathic schists lying within a shear zone in Archean metasedimentary complex near Agnew, Western Australia, (see their figs. 1&5 pp. 398-399). All these authors recognised two kinds of foliation-boudinage: symmetrical and asymmetrical. Commenting on the possible mode of their origin Platt & Vissers (1980) noted that unlike the classical types of boudins which result when extension fracturing is preceded by necking of the competent unit surrounded by incompetent mass, foliation-boudinage takes place by initial fracturing followed by further inhomogeneous straining. Foliation-boudinage is nothing but perturbations in the ductile deformation field around fractures. Platt & Vissers (1980) emphasise that a well developed foliation means strong planar anisotropy which can limit the rate of ductile extension parallel to foliation. Therefore, formation of extensional fractures normal to foliation and their widening /opening take place easily due to layer-perpendicular compression. Further compression leads to pinching-in of the fractures and adjacent foliations giving rise to finite pinch-and-swell geometry i.e. symmetric foliation-boudinage (see Platt & Vissers, 1980, fig. 3, p. 399). The asymmetric foliation-boudins are closely associated with the symmetric ones and the two types are probably formed under similar conditions. But the asymmetric boudins are caused by shear fractures, rather than extension fractures and the sense of displacement along both sets of shear fractures is such as to cause extension along foliation. The apparent pinching in of the foliation (reverse drag) results from rigid body rotation of both the fracture and the foliation adjacent to it, in compensation for slip on the fracture (see their Fig. 4, p. 399).

Recent experiments and theoretical considerations of Mandal & Karmakar (1989) have confirmed the idea of Platt & Vissers (ibid) regarding the origin of

foliation-boudinage. Mandal & Karmakar show that symmetric foliation-boudins can develop only in coaxial deformation history with the foliation and the extension fracture remaining mutually perpendicular throughout the entire course of deformation; whereas asymmetric foliation-boudinage may develop either in coaxial or in noncoaxial bulk deformation in the neighbourhood of short segments of fractures, the characteristic asymmetry developing due to the oblique positions of the foliation or the fracture or both in relation to the principal axes of stress.

The nodes observed in the Munsiri resemble foliation-boudinage as described in literature, but there are some additional features associated with the Munsiri nodes which are yet to be explained properly. These are:

- (1) the quite well developed fold-bends associated with many nodal zones/fractures. Probably the reverse drag-effect associated with asymmetric foliation-boudinage would be too weak to result in such well developed folds as are found associated with the nodal zones in the Munsiri. So post-folding thrusting must have taken place in many cases associated with nodal fractures/zones;
- (2) the occurrence of larger boudins encompassing smaller nodal boudins (cf. Fig. 3.41a). It is very likely that the formation of small scale boudins preceded large scale boudinage even though the time gap may not be a long one. Probably in the limited number of studies published on foliation-boudinage so far, adequate attention could not be given to all the associated structures, and so this sort of 'small scale boudinage within large scale boudins' has escaped attention. Thus sufficient room has to be made now in the theory of foliation-boudinage to account for these features. Alternatively, these features cannot be properly grouped into classical type of foliation-boudinage and now we probably have to look for an altogether new and unifying theory to explain the origin of these and other related structures.

Another very important aspect of the nodal structures is their nearly constant association with veins which may suggest interrelation of fluid-flow, boudinage and thrusting/shearing in the rocks.

In the Munsiri within some of the amphibolite bands showing large scale boudinage, folded mylonitic foliation can also be recognised. This implies that the large scale boudinage event post-dated both the grain-size reduction event and a folding event. At Stn. NR33 the asymmetric shape of amphibolite boudins and the characteristic fabric in the country rock surrounding them point to the role of shear in the development of boudinage.

The above discussion on the boudinage of all scales in Joshimath area indicates that possibly there were two distinct stages of boudinage: the earlier one on a small scale gave rise to nodal structures broadly contemporaneously (early synchronously) with the main phase of stretching or grain size reduction, whereas at a later stage a large scale boudinage event took place probably representing the peak of the main stretching or shearing event. It is important to mention that large scale boudinage with similar orientation as in the Munsiri has been observed also in the upper part of the Berinag-Mandhali formations.

### 3.7 FOLD-EPIISODES: THE SEQUENCE DEFINITION AND CHARACTERISTICS

The relationship between the main (NE/NNE) stretching lineation and the folds is crucial for establishing the fold sequence in the area.

The F1 folds do not have any control over the orientation of the stretching lineation. They neither deform the lineation nor have any systematic orientational relation with it. This means that the lineation must have developed later than F1 folding. The geometric (morphological) relationship of the stretching lineation with the F2 folding is rather inconclusive. In the northern parts of the outcrops of Joshimath Gneiss, where the stretching lineation is at a high angle to the F2 axes, it is difficult to recognise stretching lineation in the F2 hinge zones, but southward in the lower part of the Joshimath Gneiss or upper part of the Munsiri the difference in attitude between the F2 axes and stretching lineation becomes very small and the stretching lineation runs pervasively along the F2 hinges as well as the limbs of the F2 folds. Here the stretching lineation is very distinct on the F2 axial planar schistosity as well. This means that the stretching lineation cannot be pre-F2; it could be either syn-F2 or post-F2. No evidence was found which could unequivocally suggest whether the lineation is syn-F2 or post-F2. The stretching lineation presumably started to develop synchronously with F2 folding and continued its development even beyond the F2-episode. The acme of development of stretching lineation probably occurred following the F2 folding.

As regards the geometric relation between the stretching lineation and the F3 folds no clear evidence has been found that could indicate folding of the stretching

lineation by the F3 folds. Instead it is observed that the stretching lineation is undeflected by these folds (see Fig. 3.42).

This indicates at least two things:

- (a) stretching was taking place during the F3 folding and possibly continued until slightly after the F3 episode. Thus shearing/stretching was a prolonged phenomenon in the area.
- (b) the time-gap was not long between the F2 & F3 episodes of folding.

The geometric relation of the F4 folds with the stretching lineation is much more clear. In a few selected exposures the F4-hinges show curving of the stretching lineation around them. Thus the F4 fold-episode postdated the stretching lineation.

Thus the relationships with the main stretching lineation have proved to be crucially important for the fold sequence definition.

As already discussed (subsections 3.2.3 and 3.3.3), the F2 and F4 episodes were more intense than other episodes. F2 was the strongest episode producing folds of a wide range of scale, including those of the largest scale in the area. These show a concentration mainly in the Joshimath Gneiss. Most probably, the F2 folds developed largely through a shearing mechanism of fold-formation. The F4 folds found almost exclusively in the Mungsiari Formation are simple buckle folds.

It is interesting to note that in this area the reorientation of early folds by later folds is insignificant on a regional or mappable scale. This is brought out very clearly on the equal area projection diagrams (see Figs. 3.43, 3.44, 3.45 & 3.46). The dominant orientation of each episode of folds (except F1) is well-preserved. As was indicated nearly at the beginning of this chapter, there may be two reasons for this--

- (a) there is an obvious influence of the difference in the scale, intensity and site of maximum development of the different episodes of folds;
- (b) the time-gaps are comparatively short between the successive episodes of folding in comparison to that observed in the older so-called 'polyphase tectonite terrains'.

In addition to these factors, there is probably the limitation imposed by the inaccessibility to, or nonavailability of, suitable exposures preserving the clearcut refolding relationships if any among different fold-episodes.

As mentioned at the end of section 3.1, most of the folds in the area are found to have well defined 'confining areas' or 'profile areas'. It is not very clear what the main factors are in determining the size of such 'confining areas' or simply 'profile areas' of the folds. When all the folds in the area are considered together, there is an appreciable change in the area of the fold-profile from N to S across the region of study. Folds having larger profile-areas are much more common in the northerly parts, especially within the Joshimath Gneiss, than in the Munsiri or Berinag-Mandhali to the south. This means either the presence of a gradient in folding strain across the area at least over a particular span of time, or that there was probably a kind of strain partitioning in a larger scale. Three possible alternative reasons could give rise to this--

- (1) Larger-sized folds of the north are not represented in the south, because we are dealing with different thrust sheets: Vaikrita (Joshimath Gneiss) sheet in the north, Munsiri sheet in the middle and the Berinag-Mandhali footwall block in the south.
- (2) Originally the larger folds were also present in the south, especially within the Munsiri. But due to increasing shear and repeated strong transposition *most* of these folds have been obliterated from the southern parts.
- (3) Fold-formation in the area was diachronous, taking place in a time-transgressive manner. While folding was accommodating the strain in the northern parts, in the southern parts the strain was probably being accommodated largely by a different mechanism. Fold-forming pulses invaded the southern parts later on, but with diminished strength. In other words, folding gradually moved towards south, later episodes producing generally smaller scale folds than the earlier episodes.

Strictly speaking, these alternative possibilities, particularly the second and the third, are not mutually exclusive. Through further discussions in Chapters 4 & 5, an attempt to assert which of the above possibilities is most likely will be made.

### 3.8 LATE BRITTLE AND DUCTILE STRUCTURES

Among the late structures, faults and fractures/joints are the most important brittle structures. In addition there are discrete ductile shear zones in the area (see Figs. 3.47). The following discussion deals first with faults and fracture/joints along with slickenside lineations developed on faults, and then with the ductile shear zones.

#### 3.8.1 Faults and fracture/joints

Most of the faults and fracture/joints developed late in the history of deformation of the area and they affect almost all other structures. Faults of diverse orientations are present here. These faults did not develop at the same time. Most of the foliation-parallel faults are earlier than others. In the area, particularly within the Mungsiari, there are some thin zones (less than 12 cm in thickness) made up of phyllonites which contain mostly partly chloritised biotite, and discrete chlorite, sericite and muscovite, that are sandwiched (concordantly) between thicker horizons of mylonite-protomylonite, etc. These zones acted as movement zones and as major pathways for fluid flow as evidenced by the prolific retrogression of the minerals along them. Sometimes extreme retrogression of thin amphibolite bands has given rise to thin phyllonitic zones, dark brownish, shiny black in colour, the rock lustre comparable to that of the body of a black snake. These movement zones are probably the earliest among the fault zones and can be regarded as small scale tectonic slides (Hutton, 1981). In many cases lithological contacts have acted as movement surfaces.

Reverse faults and normal faults are more common than the wrench faults. Fig. 3.48 clearly shows variable orientations of different faults, but there is a distinct group subparallel to the main foliation (Fig. 3.49; cf 3.6, 3.21a & b). Displacement magnitudes along the faults vary for different faults; many show displacements measurable in centimetres only. Most of these low displacement faults show an anastomosing (branching and joining) pattern bounded within zones of certain width. Thus they show distributed displacement. Along some of the late faults there has been development of fault gouges, while many others are without them. Although it has been attempted to estimate the magnitude of fault displacement from gouge thickness elsewhere (Robertson, 1982, 1983), the principle cannot be directly applied to this area because most of the faults show gouge development in a complex manner. At many places there are heterogeneous gouge development across the fault zones.

Also gouge development is found in the anastomosing faults with distributed displacement. From Chain village, looking across the Alaknanda River, the new road-cut exposure below the Marwari locality shows a well developed system of faults/fractures that are steeper than the main foliation in the Joshimath Gneiss. Locally these faults/fractures control the hillslopes; displacement along these faults does not appear to be large.

It is important to note that because the road sections provide the best of the accessible rock exposures in the area, one has to be careful not to confuse the man-made (blasting-induced) faults and fractures with the natural ones. To glean profitable

information regarding the behaviour of the rocks even from these blasting-induced fractures would require further study.

Fractures/joints are more common than faults. Restricted fracture zones are more frequent and within them there are often several sets of close spaced joints. Most of these joints are shear fractures/joints.

Shear fracture zones showing normal and reverse displacement sense are present. These shear fracture zones are more common in the Munsiri Formation. As Fig. 3.49 suggests, there are three main directions for shear fractures in the area: ESE-WNW, NNW-SSE and ENE-WSW. Not all these coexist in equal proportion everywhere. Fig. 3.50 indicates the presence of two conjugate sets of high angle joints/fractures in the area. Joint development has been largely guided by the host lithology, thus the relatively competent quartzite and quartzofeldspathic mylonite-protomylonite horizons of the Munsiri Formation are preferred hosts for joints. In the Joshimath Gneiss a ramp-flat geometry is shown by some fractures within an alternating sequence of psammitic and pelitic layers. The fractures here run parallel to the main foliation through the less competent mica rich horizons (flats) whereas they cut across the more competent siliceous horizons at high angles to the main foliation (ramps).

### 3.8.2 Slickenside lineations on faults

These are recognised mainly on faults developed late in the sequence of deformation. Their intensity varies. It is sometimes difficult to identify them, for example, on faults with well developed gouge or where thin films of carbonate (homogeneous solution precipitate) on the fault surface mask the slickensides. They are totally obscured in the foliation-parallel faults along which post-faulting recrystallisation has taken place. It is very likely that these latter faults are older than other faults.

There is less unanimity in opinion regarding the definition, classification, the possible modes of origin or even the significance of the different types of slickenside lineations (see Fleuty, 1975; Means, 1987; Will & Wilson, 1989 etc.). Slickenside lineations do not always give unequivocal sense of slip along faults (Paterson, 1958; see also Tzia, 1964; Norris & Barron, 1969; Gay, 1970; Durney & Ramsay, 1973). The main reason for this is the confusing presence almost always of what are called 'step structures' on the slickensides (see Hobbs et al, 1976, pp.304-5). Hills (1972, pp.178-180) thought that the step structures are essentially broken up tip-zones of

pinnate shears associated with the faults and suggested that they were ambiguous indicators of the sense of fault displacement. In the present area in some instances the slip sense along faults have been recognised from the existence of unidirectionally pointing quartz-fibres defining the slickenside lineation. Sometimes, following the slickenside lineation there has been systematic frictional bevelling of protruding quartz crystals on the fault surface; as a result now the edges of the crystals are very sharp resembling the head of a carpenter's chisel. The sharpness is felt most strongly when one rubs the palm along the slickenside lineation following the direction of movement of the wall holding the crystals.

As far as the evolutionary history of a fault zone is concerned, this feature has an additional significance because the protruding crystals had to grow in the fracture before the fault-movement took place. Though it is not possible to recognise whether there were any earlier stages of movement along the fault prior to the growth of the crystals, the fault evolution encompasses at least three stages: (i) initial fracture opening, (ii) formation of thin vein-like body in the fracture and development of the protruding crystals, and (iii) fault-displacement. Altogether these might have taken a considerable time unlike short duration, seismic faulting. It has been demonstrated that sometimes even veins can evolve through millions of years (cf. Ramsay & Huber, 1983, p.262; Durney & Ramsay, 1973).

Pitch of slickenside lineations varies on different faults. In general, the pitch is higher (more than or equal to  $60^\circ$ ) on low to moderate dipping faults (dip less than or equal to  $50^\circ$ ), whereas on the high angle faults, the pitch is lower. Thus along the high angle faults in the area there is a strong component of strike-slip displacement, whereas the low-to-moderate dipping faults have stronger dip-slip components.

### 3.8.3 Late ductile shear zones

There are many examples of late discrete ductile shear zones developed in the rocks of all the three rock formations. Most of these shear zones are cutting across the main foliation at high angles. Wrench shear is rare compared to the normal, reverse or oblique shears along these shear zones. There has been spectacular retrogression and grain size reduction along these shear zones e.g. amphibolites locally turned into biotitised phyllonites along them or high grade gneissic rocks changed into low grade schists show distinct biotitisation (see Fig. 3.46). Most late shear zones have widths between 5 and 15 cms. Lack of suitable exposures makes it impossible to determine the length/width ratios of the zones. However, their dip length/width ratios

are found to be invariably  $\gg 30:1$ . In some instances, the widths vary so as to give pinch-and-swells in the shear zones. This feature is very well marked where pegmatitic veins have been intruded along the shear zones. These veins also show pinch-and-swell and occupy only the middle part of the shear zone dividing it more or less symmetrically into two halves. No extra displacement is shown by the affected foliations along the surface of these veins. The pinch-and-swells are asymmetric. The swollen parts are asymmetric and spindle-shaped in outline; that is, a swell on one side is not directly opposing a swell on the other side of the zone (the same is true for the pinched parts).

These features are good and direct indicators of shear sense. For any asymmetric spindle, the position of a bulge on one side in relation to that on the other side gives the sense of shear. When the bulge on the hanging wall side goes down relative to the bulge on the footwall side, a normal shear is indicated. The reverse situation implies a thrust shear. The validity of this criterion is checked through a direct comparison with the sense of displacement of the main foliations in the shear zone. Before applying such a criterion, however, one has to make sure that the vein-intrusion and the pinch-and-swell structure are definitely syn-shearing. In the present case, it has been possible to measure the amount of shear displacement of successive foliations. Within the limit of the exposure it is found to be constant for all the successive foliations. This implies that the development of pinch-and-swell, the intrusion of the vein and shearing all took place simultaneously. This feature is, therefore, also an example of shear-induced boudinage.

### 3.9 SUMMARY OF STRUCTURAL HISTORY

#### 3.9.1 Areal distribution of folds

Table 3.1

Folds in Joshimath

Gneiss & Munsiriari :

F1

F2

F3

F4

F4a

Berinag-Mandhali

folds:

F<sub>I</sub>

F<sub>II</sub>

F<sub>III</sub>

Final gentle  
warping  $\gamma$

	Joshimath Gneiss	Munsiari	Berinag-Mandhali
F1		---	
F2		-----	
F3	-----	-----	
F4		-----	
F4a	-----	-----	
F <sub>I</sub>			-----
F <sub>II</sub>			-----
F <sub>III</sub>			-----
Final gentle warping $\gamma$			

### 3.9.2 Simplified sequence of deformation

In the following table (Table 3.2) showing the generalised sequence of deformation, the positions of the two major metamorphic episodes are also indicated.

Table 3.2

	Stretching & Mylonitisation	Boudinage	Late Shears	Very late brittle structures	Metamorphism	
					Initial highgrade	Late retrograde
F1					Initial highgrade	
F2	Initial highgrade	Initial highgrade				
F3	Initial highgrade	Initial highgrade				
F4	Initial highgrade					Late retrograde
	Late retrograde		Late retrograde	Late retrograde		

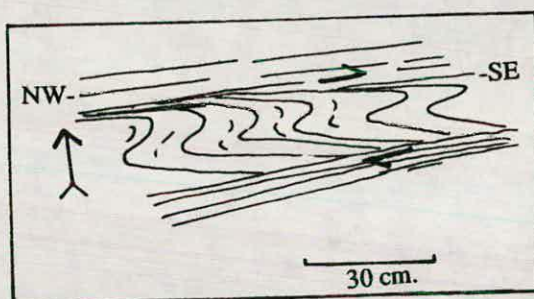
### 3.9.3 Concluding remarks

It is to be emphasised that in the present study of an area lying within a younger orogenic belt, we are interested in getting a magnified view of the deformational phenomena which may have occurred within a relatively thin slice of geological time involving the emplacement of the MCT. Therefore, the conventional terminology-- F1, F2, F3 etc. that is used here to designate different episodes of folding is to be taken with a slightly different significance from that in case of studies on older polyphase tectonite terrains. As already mentioned in the introductory part of the chapter, fold episodes and foliation development in the thrust or shear zones, occur in a relayed and overlapping fashion, and deformations as a whole follow a continuum. Perhaps in the older orogenic belts this is not so apparent because

repeated major deformation events obliterate much of the finer details of earlier deformations, and our observational tools are still not sophisticated enough so as to distinguish all the successive phases of transposition. Holdsworth (1990) indicated the changes necessary in our approach to structural analysis particularly in thrust or shear zones. However, the conventional approach to structural analysis can still indicate in a time sequence the relative positions of the peaks of different deformation episodes, even though such episodes may be considerably overlapping in time. This chapter attempted to fit the deformation episodes into such a sequence.



Fig. 3.1 (a) Field photograph showing probable cross-bedding, asymmetrically folded and right-side-up (below the hammer-head), in a quartzite horizon, Munsiri Formation. Location on the Bypass Rd, NE of the confluence between Poini Nala and Alaknanda river (Original Location NR88). Photo taken looking N60° E.



(b) Handsketch highlighting the folded cross-beds from the middle portion of the above photograph.

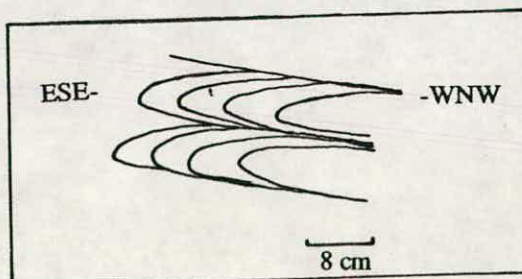


Fig. 3.2 Field-sketch of well-preserved long limbs of disrupted tight overturned asymmetric folds appearing superficially as "cross-beds" in a quartzite horizon, Joshimath Gneiss Formation. Location 4km N of Vishnuprayag on way to Badrinath.

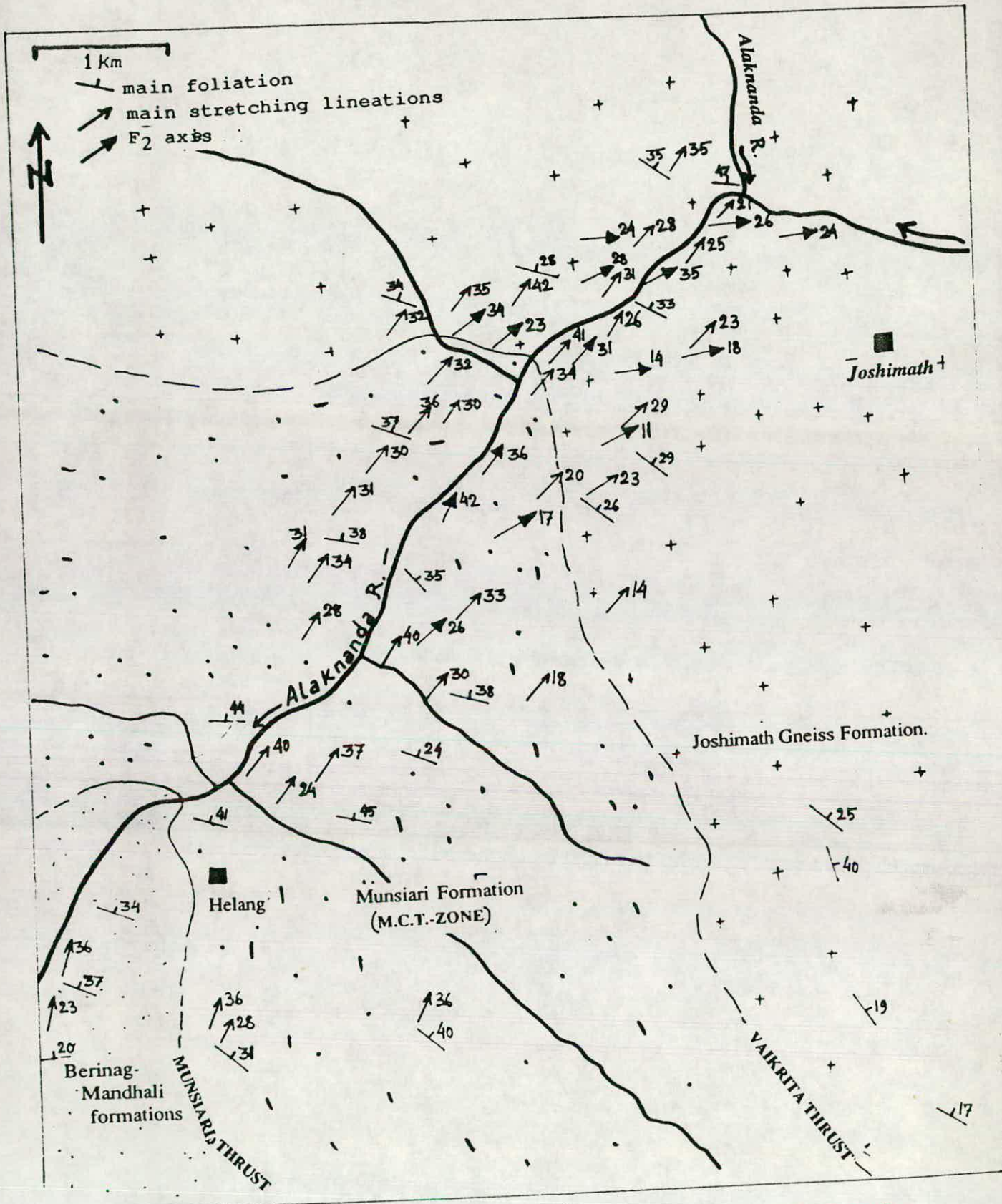
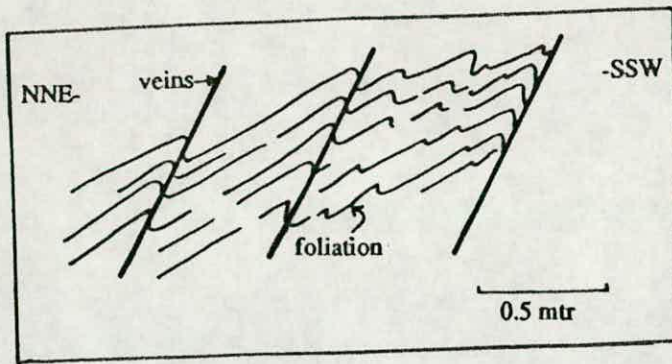
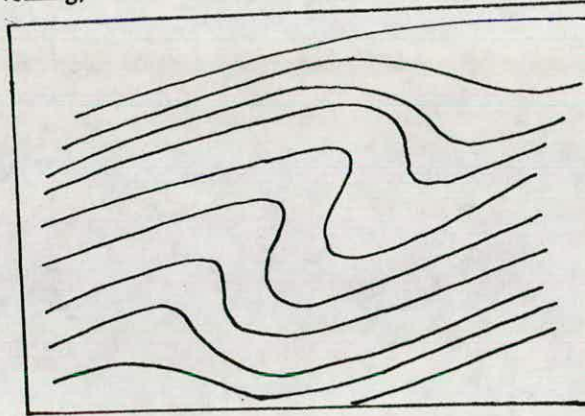


Fig. 3.3 Progressive parallelism between  $F_2$  axes and main stretching lineation towards south from Joshimath Gneiss into the MCT-zone.



**Fig. 3.4** Highly asymmetric southerly verging folds in a quartzite horizon of the Berinag-Mandhali formations, near Helang Bus Stop. Note the occurrence along the short overturned limbs of straight quartzose veins following the axial planar orientation. Probably there was no appreciable time-gap between folding and veining, and the folds resulted from a late phase of shearing.



(a)



(b)

**Fig. 3.5 a & b.** Respectively a schematic sketch (not to scale) and a field photograph from a mylonite horizon in Munsiri Formation (SW of Shelang village) showing well-defined 'confining area' or 'profile area' of folds enclosed within mutually parallel foliations. This feature is probably a common characteristic of shear zone folds.

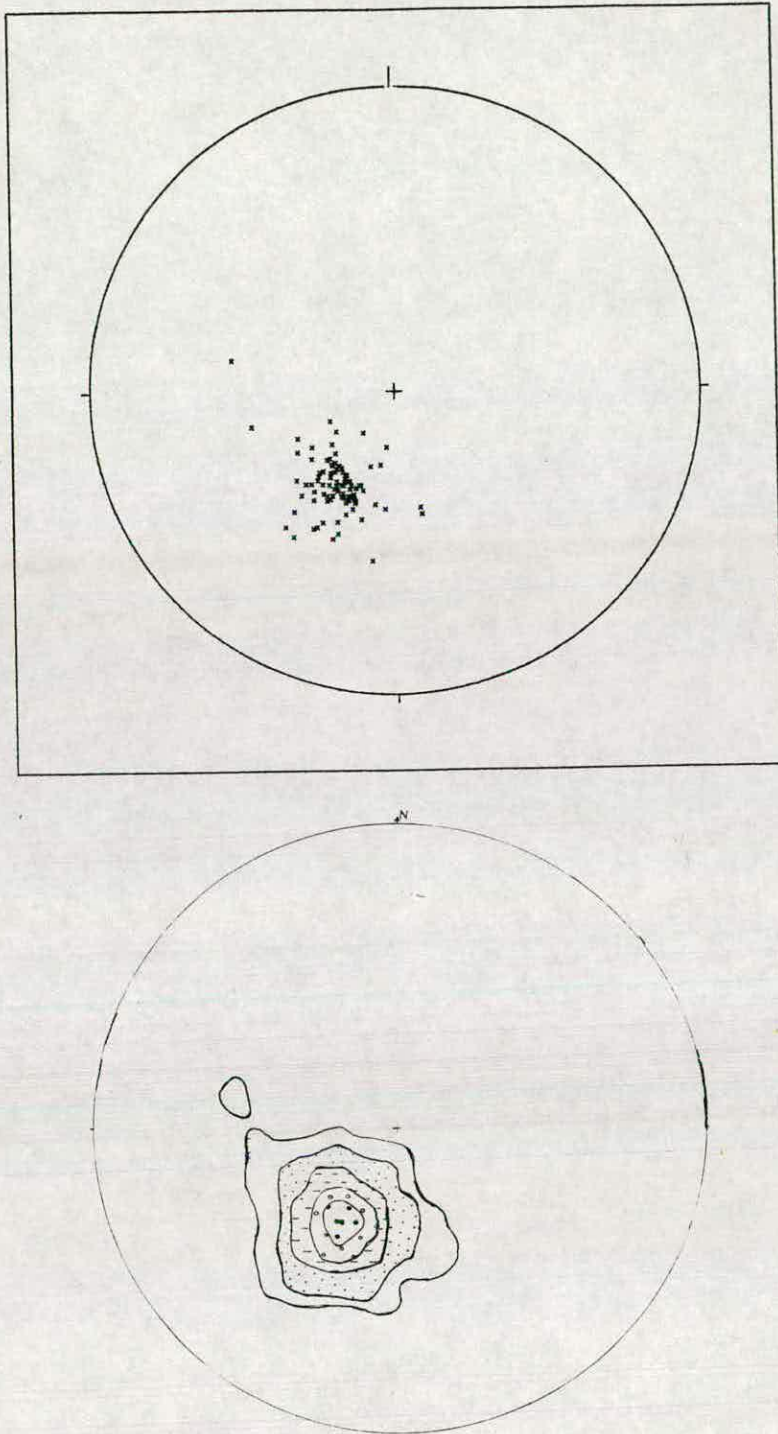


Fig. 3.6: 89 poles to main foliation in Joshimath Gneiss. Contour percentage per 1% area; 1.124-3.37-14.6-28.089-43.82 (corresponding number of data points: 1-3-13-25-39). maximum concentration 65.17% (58 data points).



(a)

**Fig. 3.7 a & b.** Distinguishable lithological variations even within the limit of individual exposures. Field photos from the 'intraformational slide zone' in the Joshimath Gneiss Formation (Orig. Loc. between NR25 & 26). For details of location see Plate-IV placed in the back pocket. Views looking E.



(b)

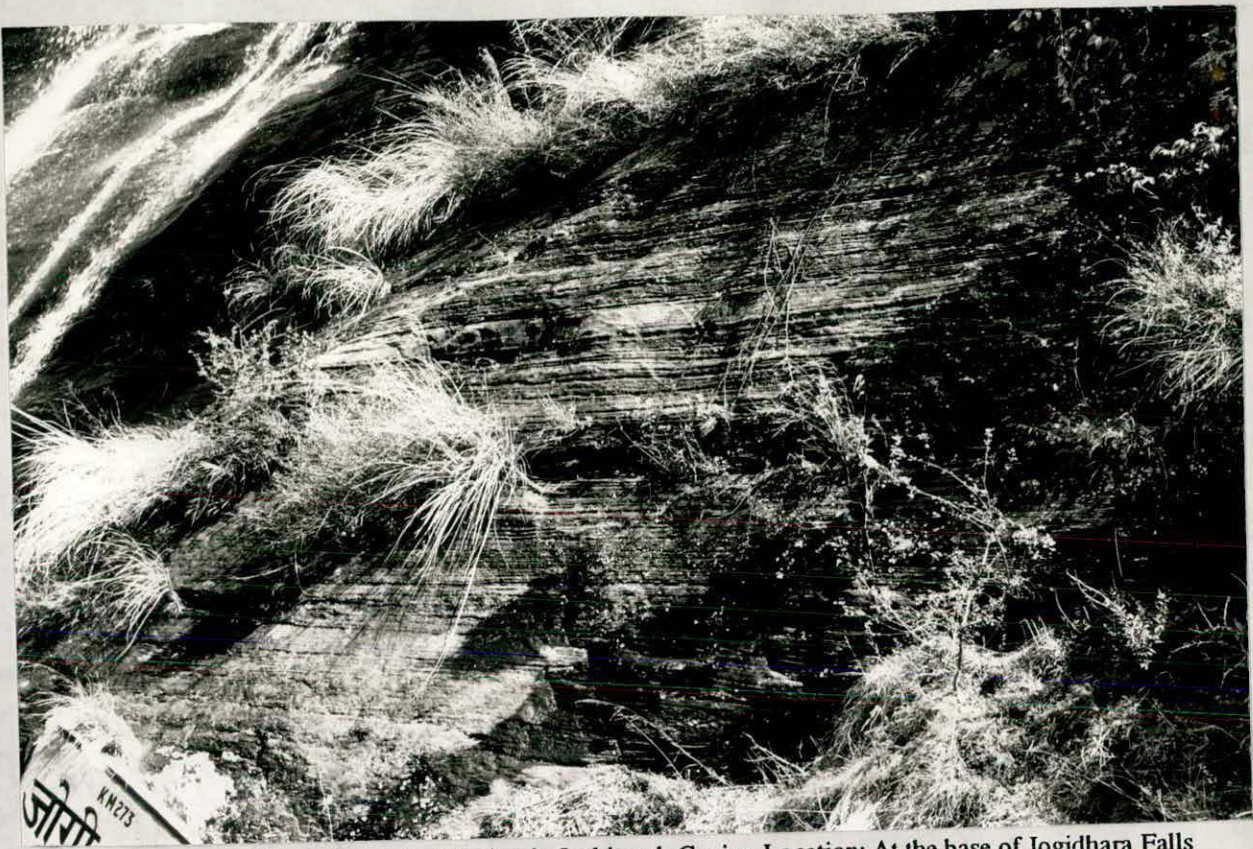


Fig. 3.8 (a) 'Multiorder' gneissic foliation in Joshimath Gneiss. Location: At the base of Jogidhara Falls (Note the water fall at upper left). Photo taken looking E-ward).

(b) Sketch showing the details of 'multiorder foliation'.

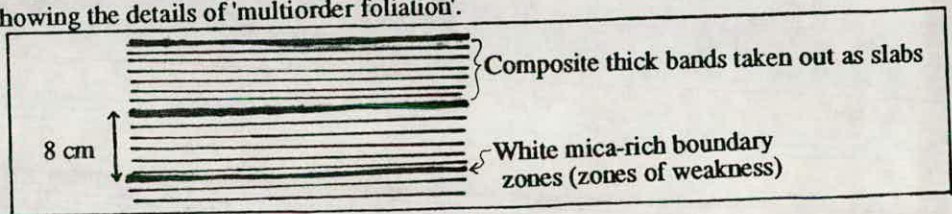
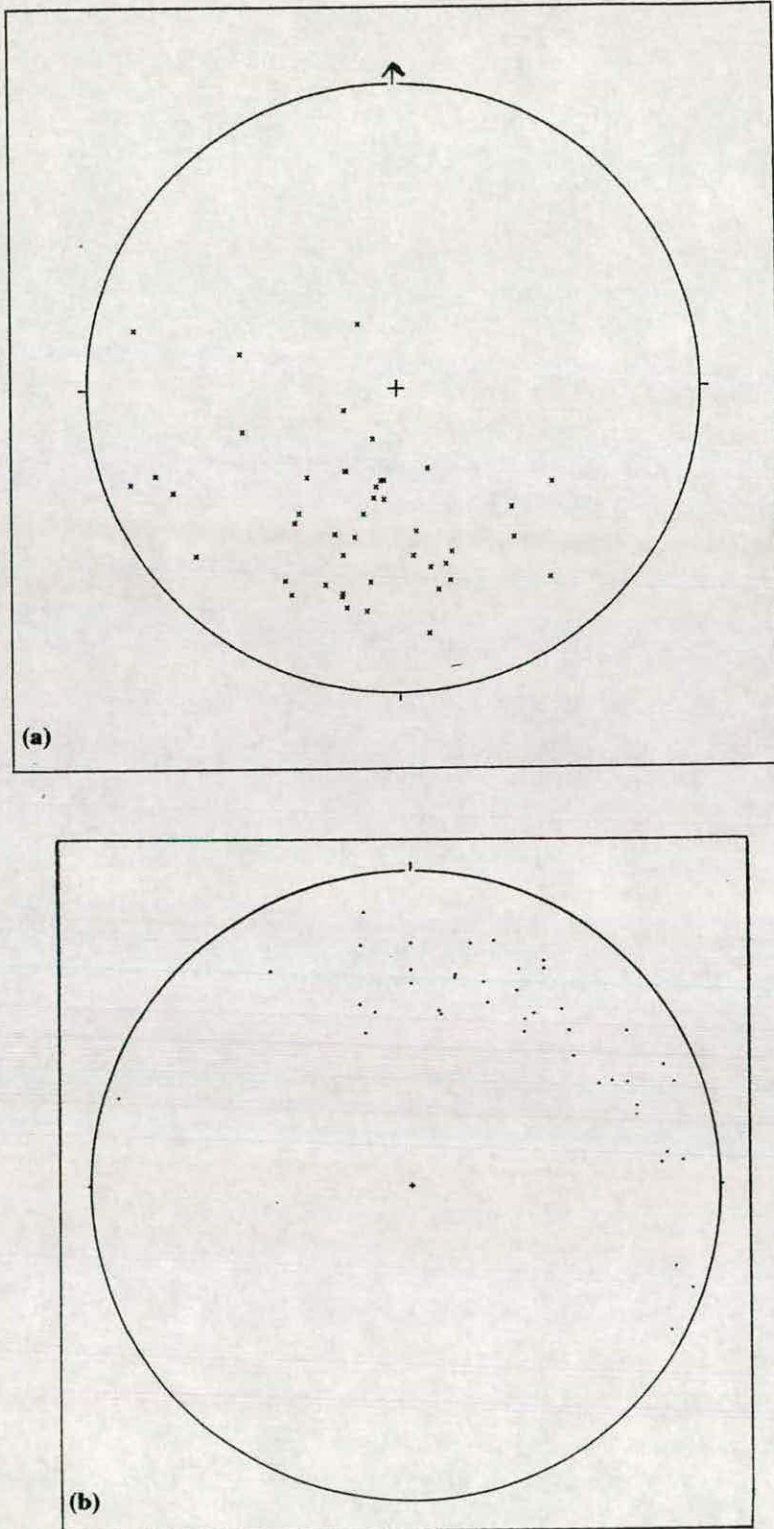
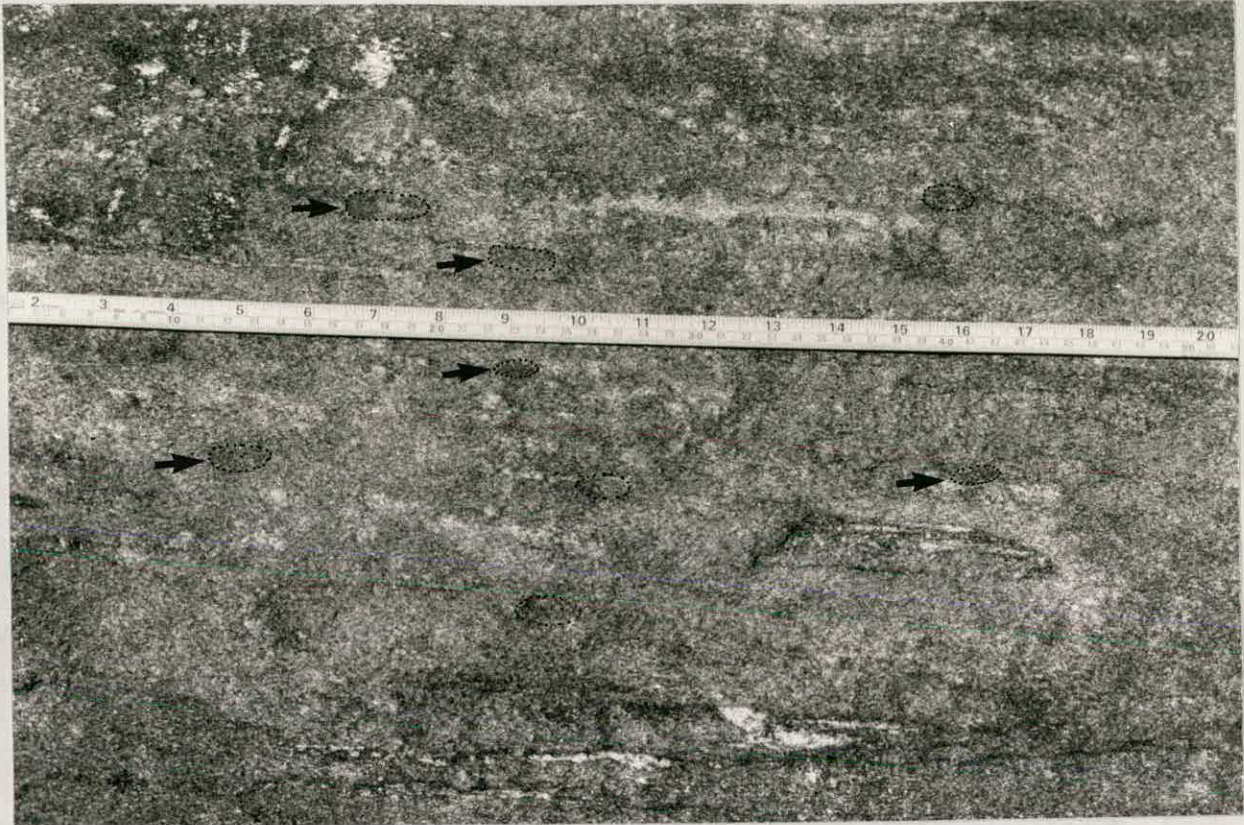


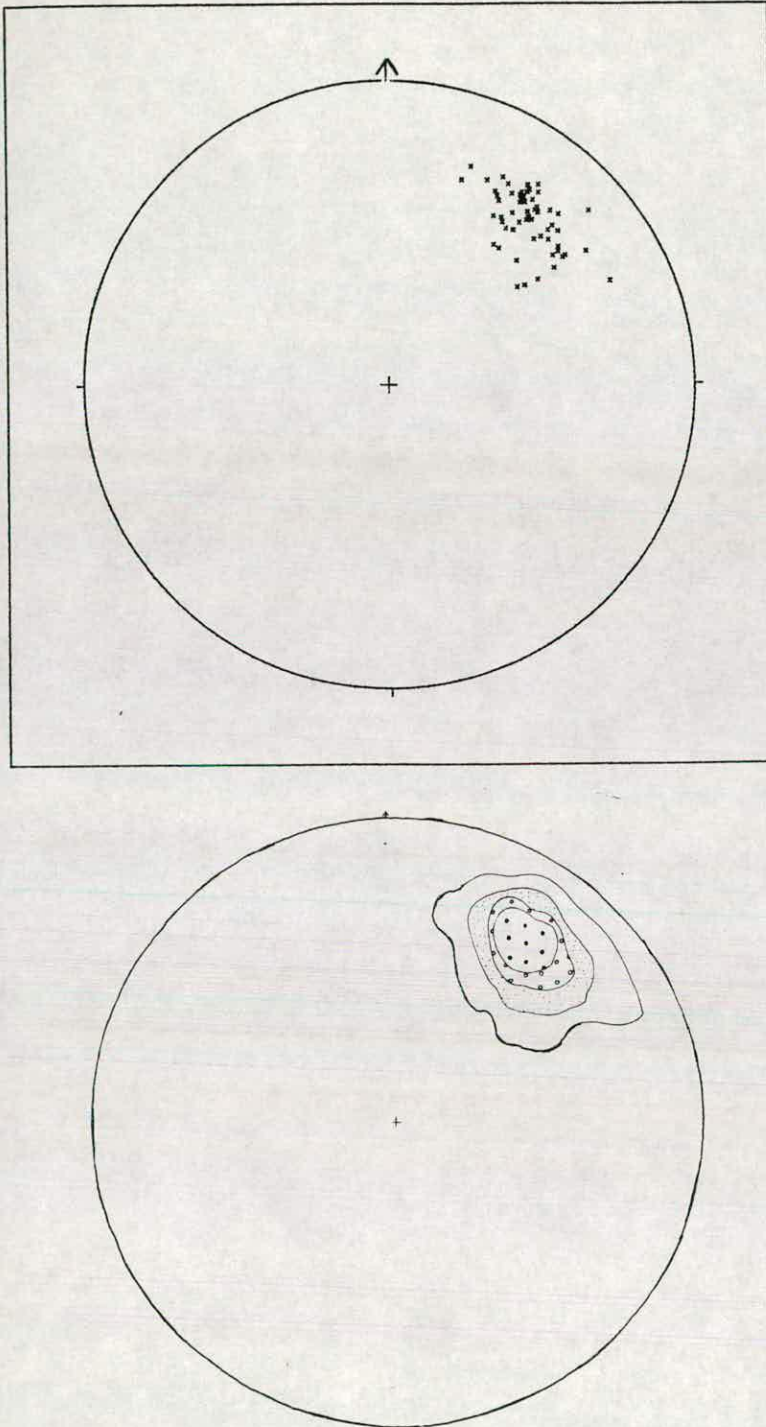
Fig. 3.9 'Multiorder foliation' involved in folding. Location: A rolled block WSW of Tapovan. The influence of primary lithological banding is quite apparent here.



**Fig. 3.10:** (a) 46 poles to undifferentiated cleavage; (b) 43 undifferentiated lineations.



**Fig. 3.11** Stretched garnet porphyroblasts (indicated by arrows) in Joshimath Gneiss lie parallel to the main stretching lineation (orientation given by the alignment of the ruler). Location near Jogidhara Falls. Photo taken looking downward, facing W.



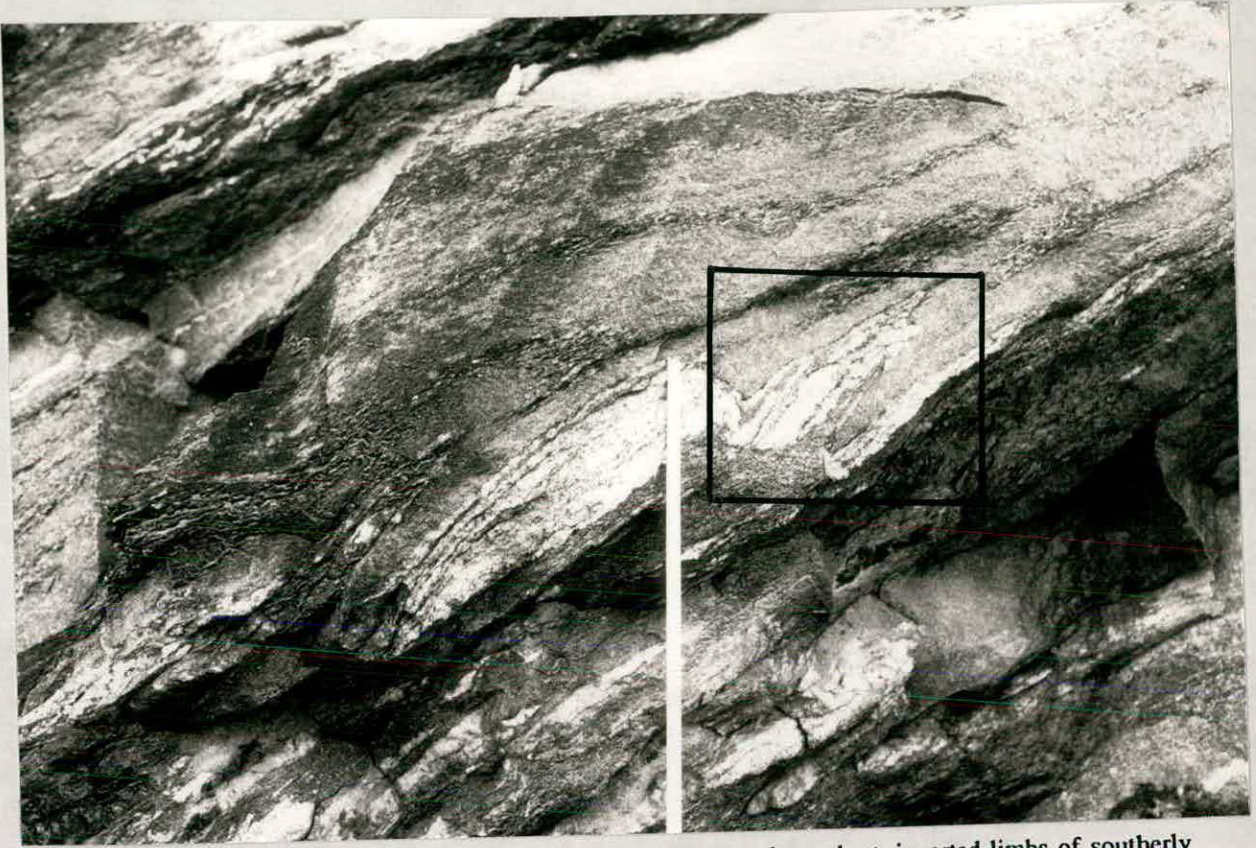
**Fig. 3.11a:** 62 main stretching lineations from Joshimath Gneiss. Contours at 1.613-9.68-22.58-37.09% per 1% area (corresponding number of data points: 1-6-14-23). Maximum concentration 53.23% (33 points).



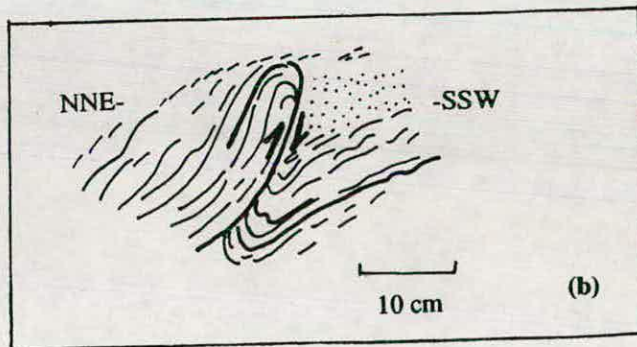
Fig. 3.12 Hook-shaped i.e. Type-3 fold interference pattern between  $F_1$  &  $F_2$  in Joshimath Gneiss. Photo taken from a rolled block after crossing the Singdhar Bridge on Alaknanda river, on way to Badrinath.



Fig. 3.13  $F_2$  axial planar schistosity (S) slightly gentler than the gneissosity (G) or main foliation in Joshimath Gneiss on the overturned limb of a outcrop-scale  $F_2$  fold. The fold-hinge and the normal limb are not included in the photograph. Photo taken looking ESE (Orig. Loc. NR3).



**Fig. 3.14 (a)** Extreme attenuation leading to microthrusting along short, inverted limbs of southerly verging asymmetric  $F_2$  folds in Joshimath Gneiss. Note the boxed part in particular. View looking E. (orig Loc. 2).



**(b)** Field sketch of a more conspicuous example of the above kind of feature. Location: A few metres south of the above photographed spot. Note in both (a) & (b) the absence of any vein in the dislocation zone/s.



Fig. 3.15 a & b. Attenuation and disjunction of inverted limbs and overriding of antiformal hinges above the complementary southern synformal hinges in a series of  $F_2$  folds developed in quartzofeldspathic bands give rise to what apparently look like 'imbricate (rotated) boudins'. Sense of direction is the same in both (a) & (b). Orig. Loc. 2.

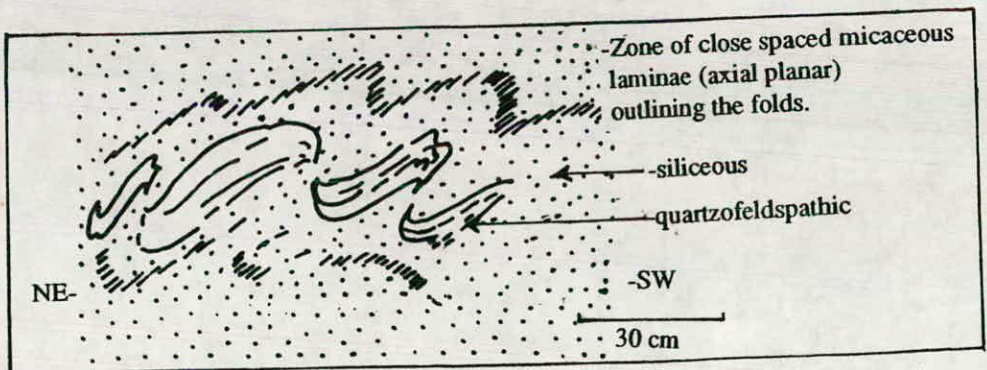
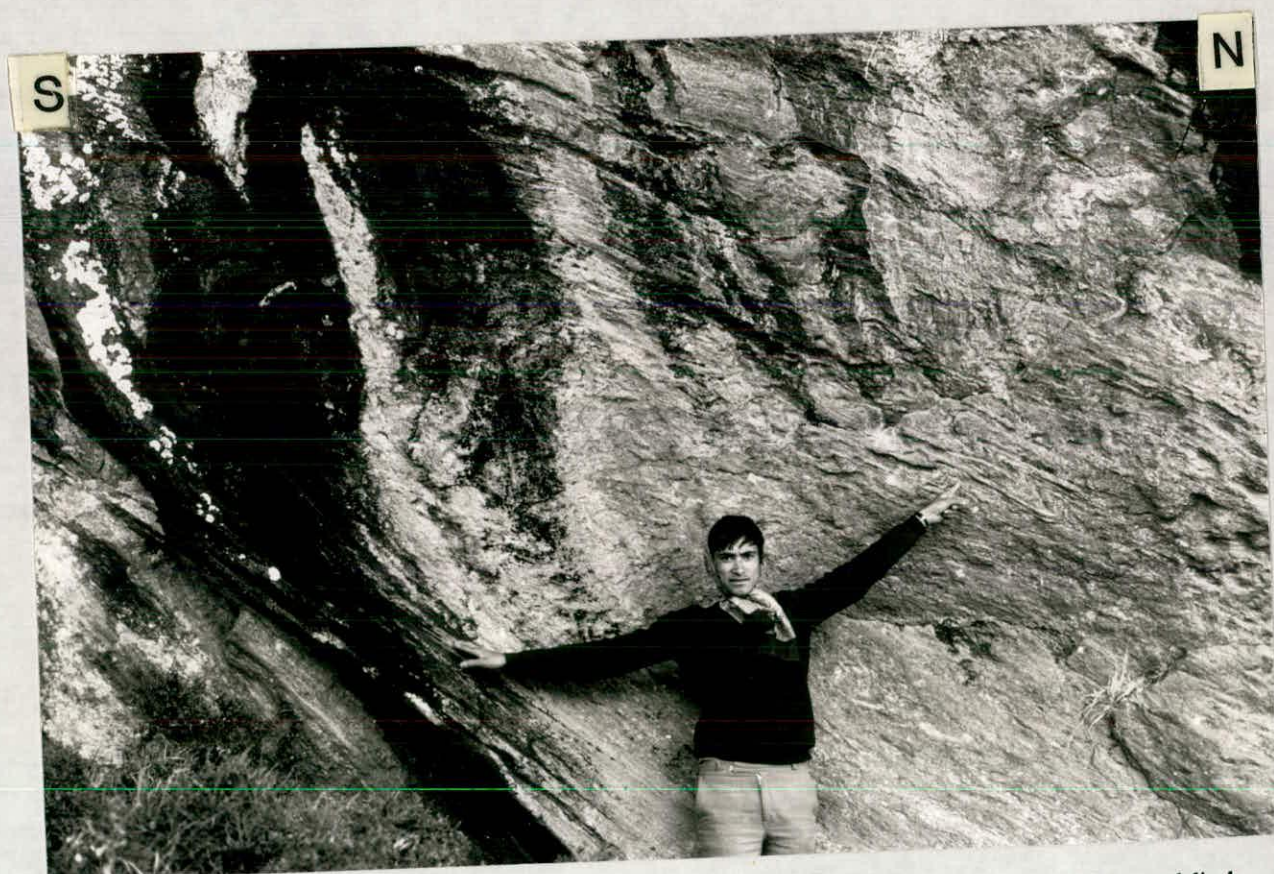
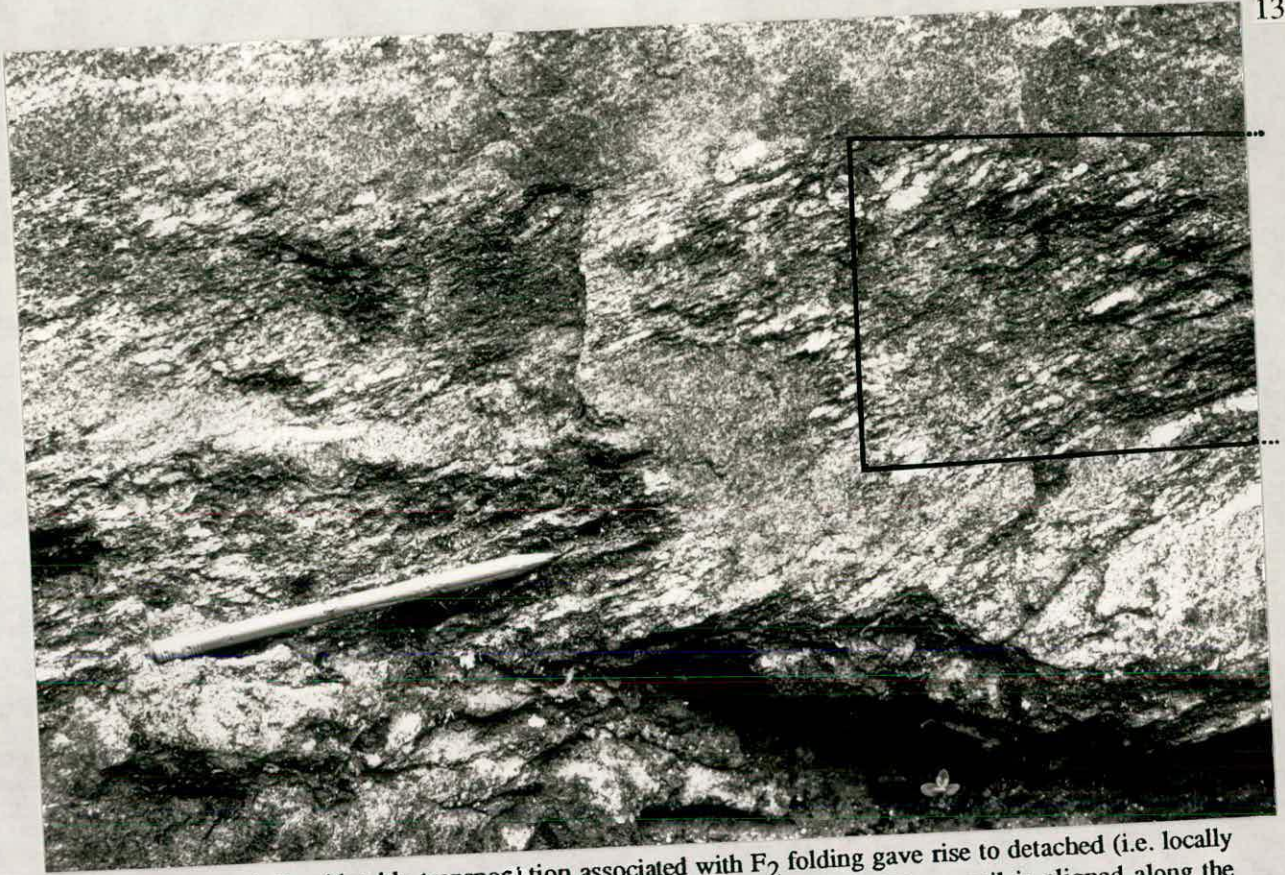




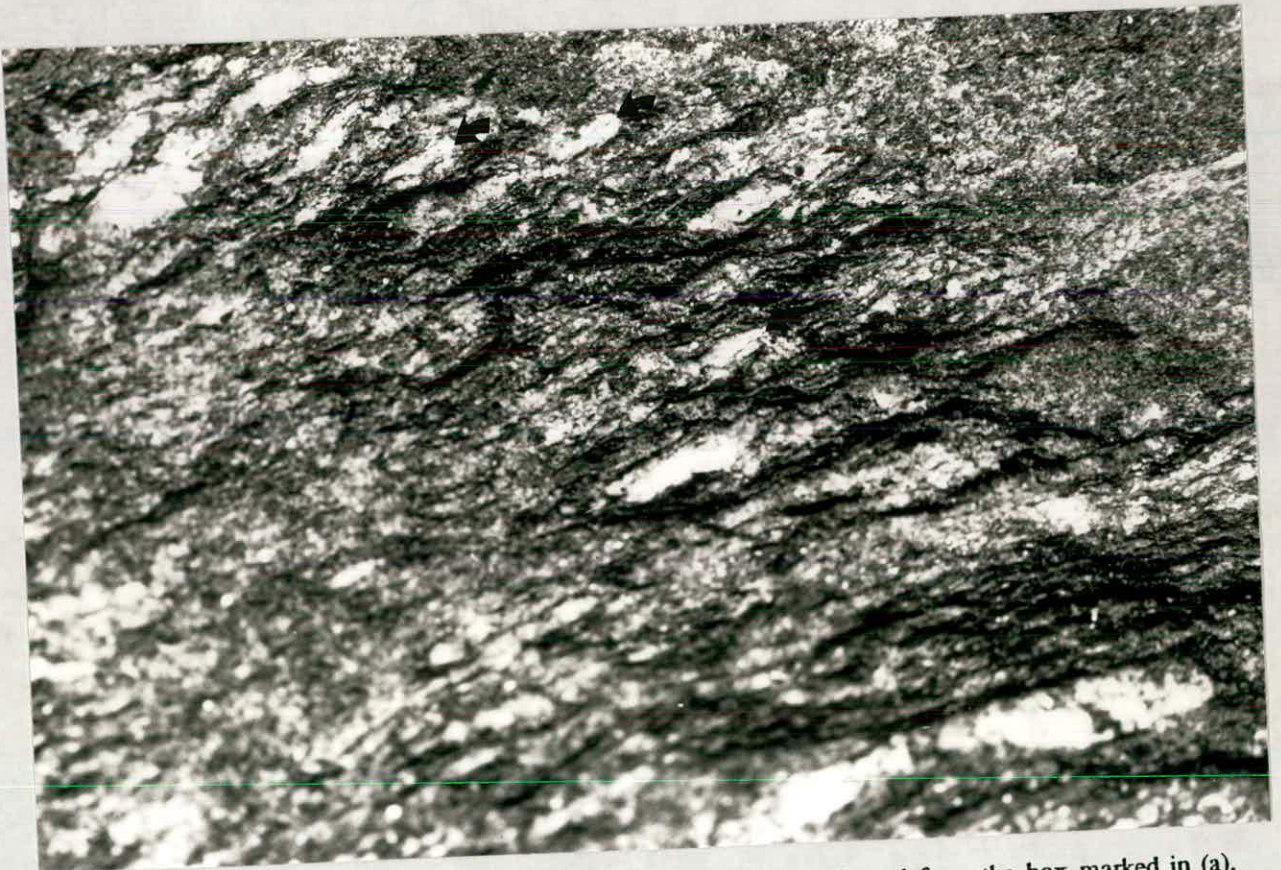
Fig. 3.16 (a) Clear 3D-exposure of  $F_2$  fold with minor folds developed on both the limbs of a lower order fold at Loc. NR31 (i.e. the Jogidhara causeway on the Lower Bypass Road.). Note that the  $F_2$ -axis is already nearly parallel to the stretching lineation. Photo taken looking downward facing ESE.



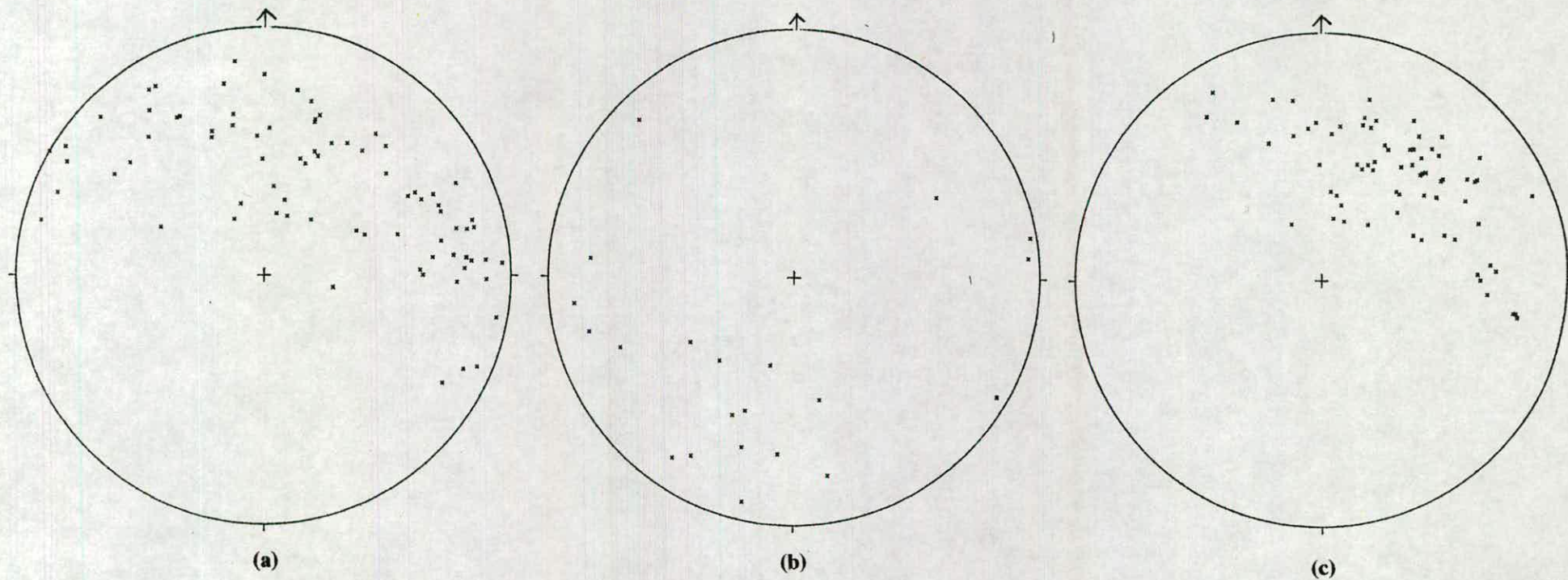
(b) An excellent example of an exposure-scale  $F_2$  fold with long, attenuated inverted limb free from parasitic folds, but comparatively short and thick normal limb with profusion of parasitic folds. Photo from Loc. 21 (at Auli), looking WNW.



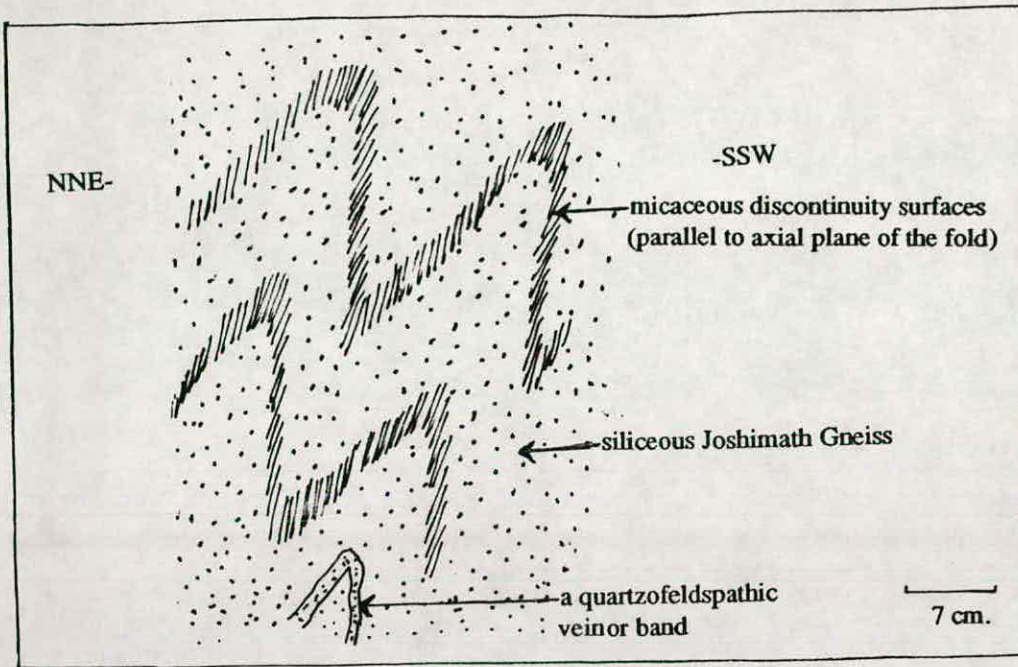
**Fig. 3.17 (a)** Considerable transposition associated with  $F_2$  folding gave rise to detached (i.e. locally 'intrafolial') tight smallscale fold hinges in quartzofeldspathic bands. The pencil is aligned along the axial trace of the dominant fold here.



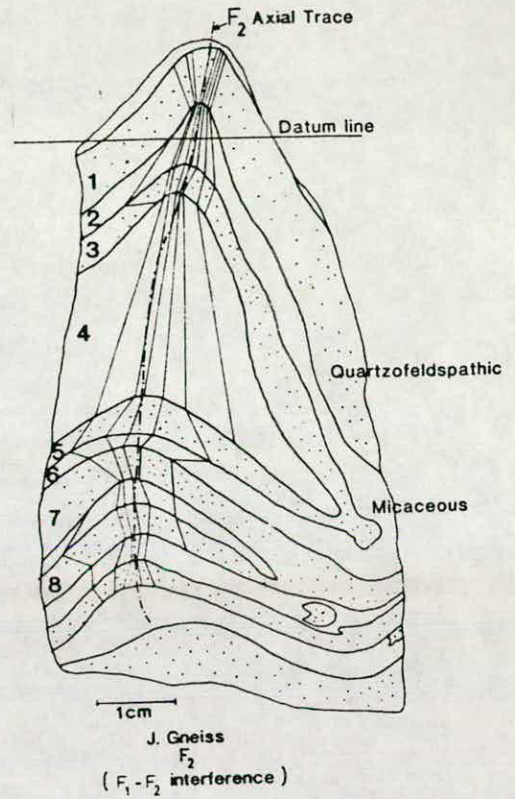
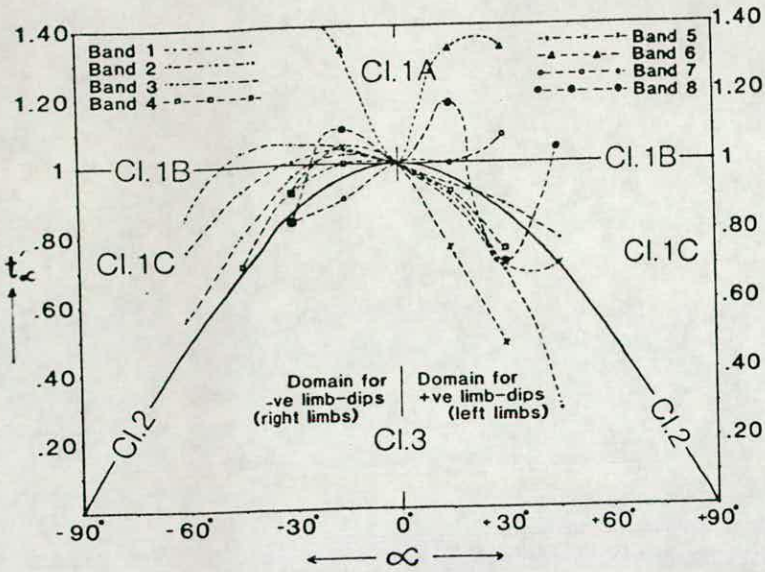
**(b)** Details of some detached fold hinges (arrowed) enlarged from the box marked in (a). Orig. Loc. NR8. Photos taken looking broadly ESE.



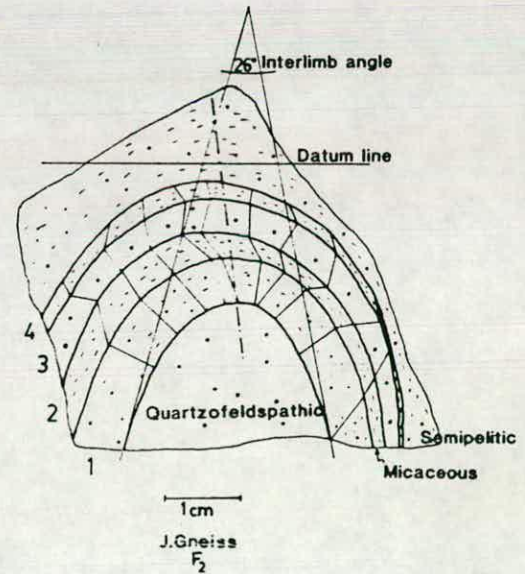
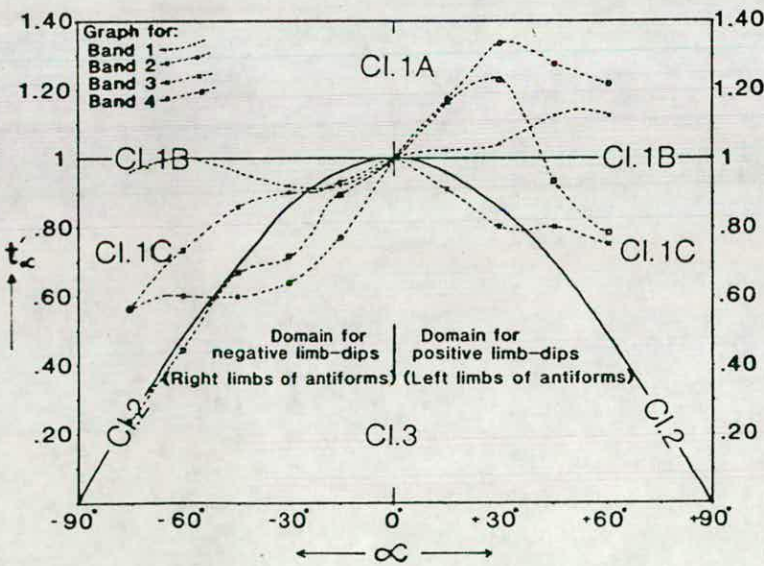
**Fig. 3.18:** (a) 80 undifferentiated fold axes, (b) 22 poles to axial planes of undifferentiated folds, (c) 71 unclassified intrafolial fold axes.



**Fig. 3.19** The outline of  $F_2$ -folds defined mainly by localised mica-rich zones in a siliceous horizon of Joshimath Gneiss. Orig. Loc. 4.

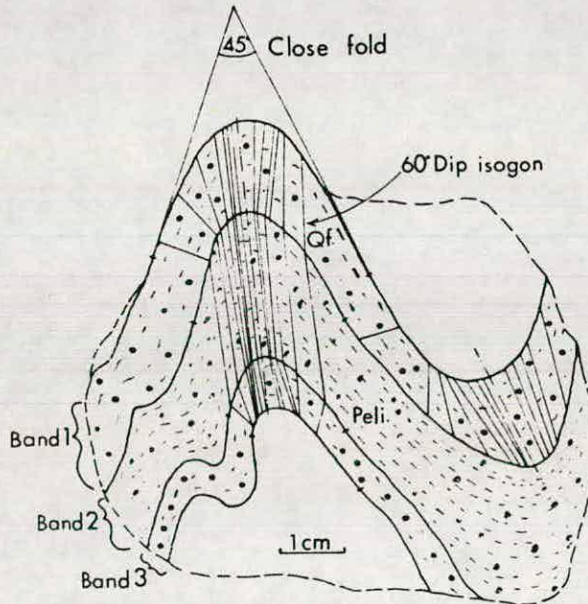
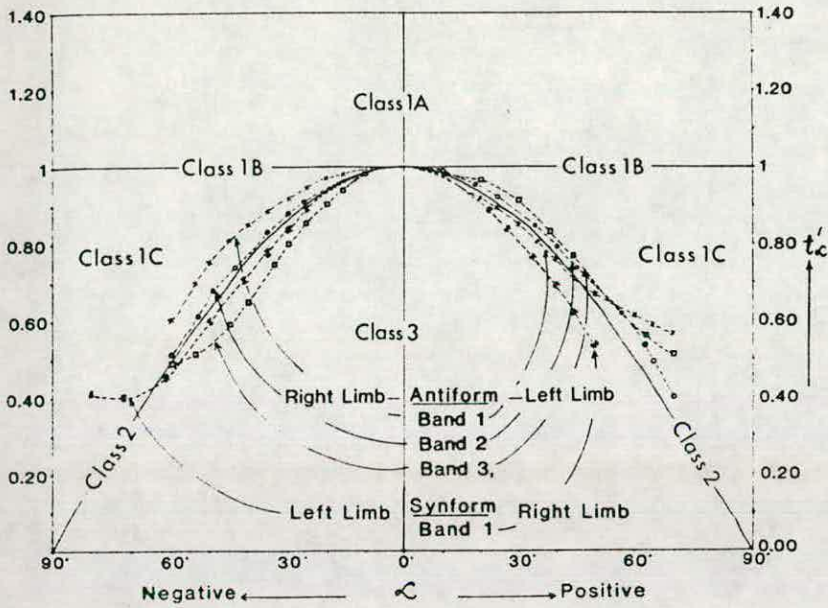


(a) from Specimen No. 59'87.



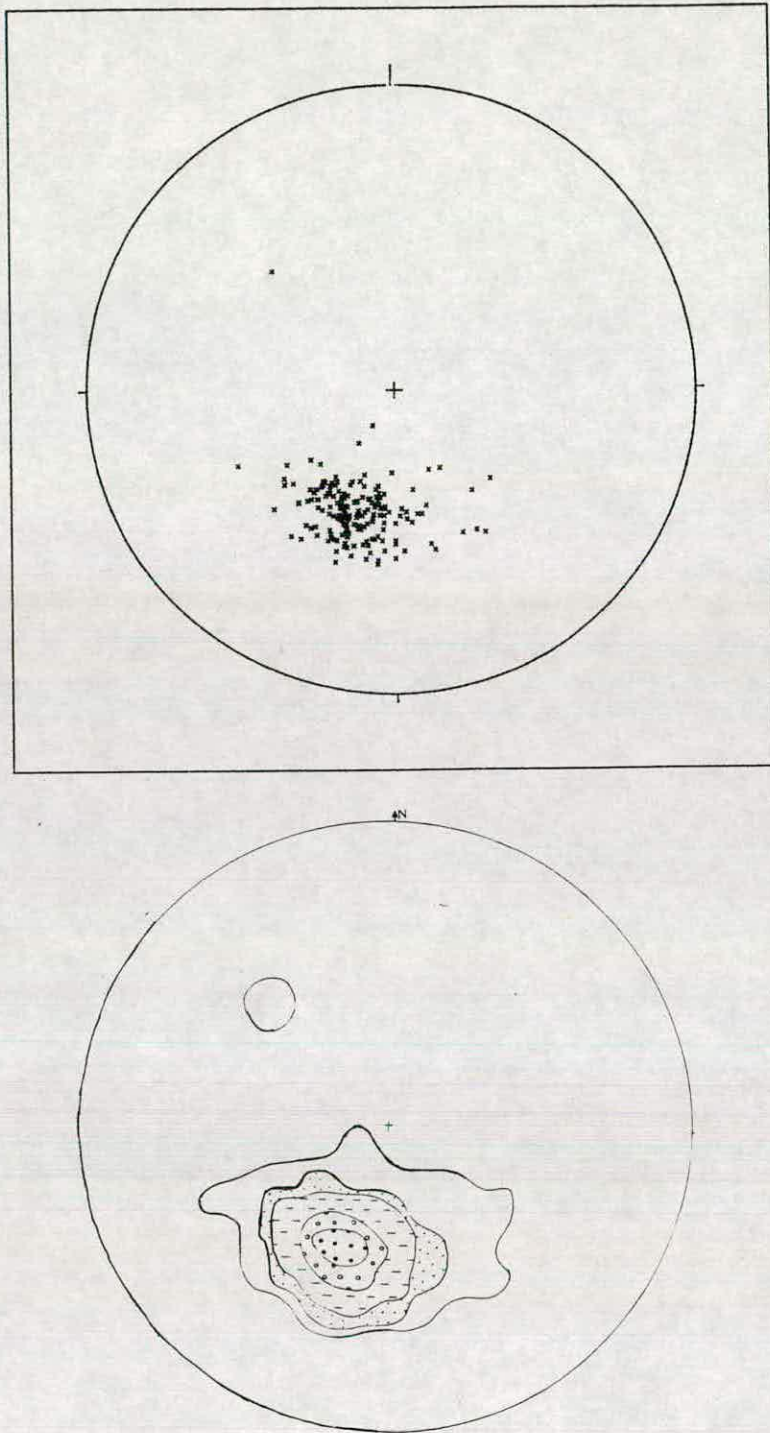
(b) from Specimen 58'87;

Fig. 3.20

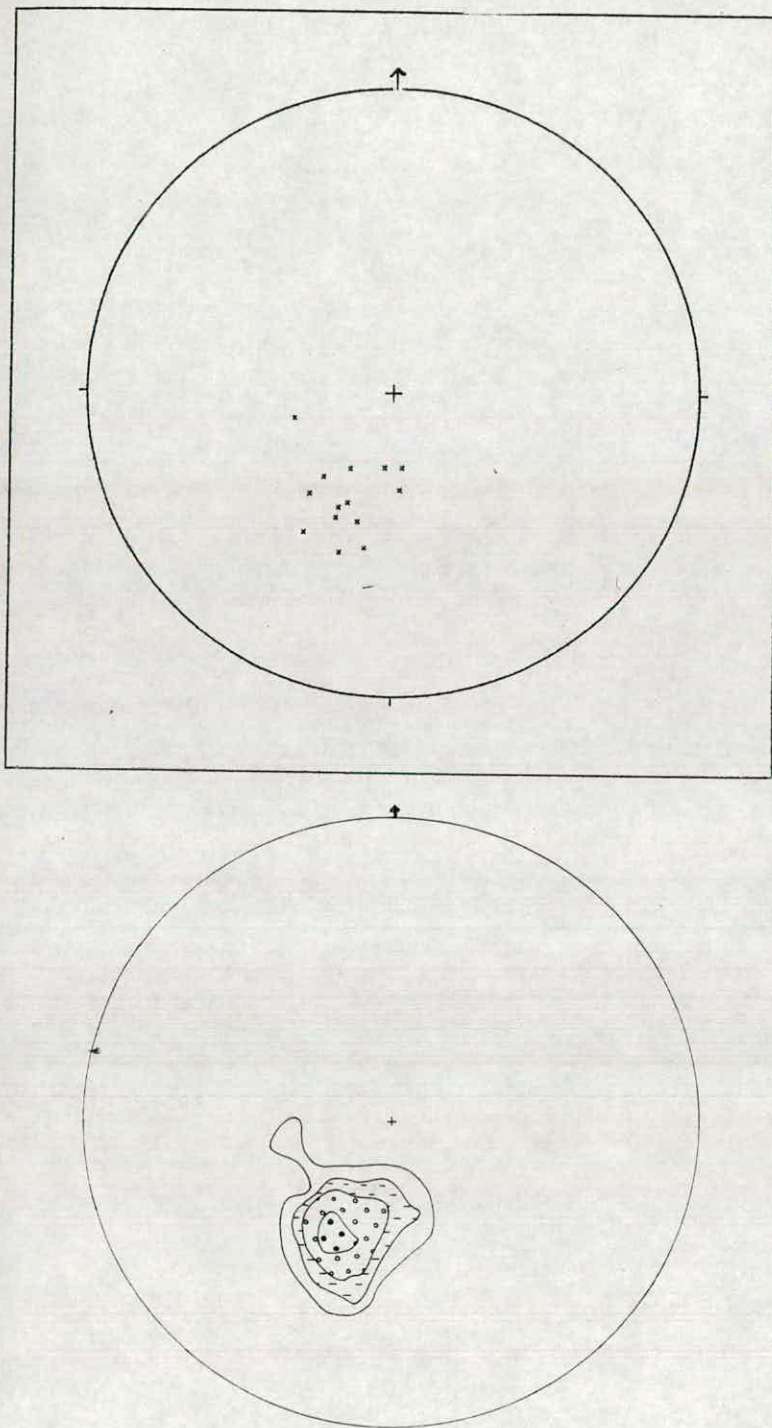


(c) from Specimen No. 22/4/88B.

**Fig. 3.20** Geometrical analyses of 3  $F_2$  folds in Joshimath Gneiss following dip isogon and thickness ratio methods (after Ramsay, 1967). (a) from Specimen No. 59'87. Note the clear type-3 interference between  $F_1$  &  $F_2$ ; (b) from Specimen 58'87; (c) from Specimen No. 22/4/88B. Further details in the text.



**Fig. 3.21a:** 179 Poles To Main Foliation In The Munsiri Formation. Contour percentage per 1% area: 6.15-22.34-33.52 (number of data points per 1% area:1-5-11-40-60). Maximum concentration 45.25% (81 data points).



**Fig. 3.21b:** 14 poles to main foliation in the Berinag-Mandhali formation. Contours at 7.14-14.25-21.43-35.7% per 1% area (corresponding number of data points 1-2-3-5).

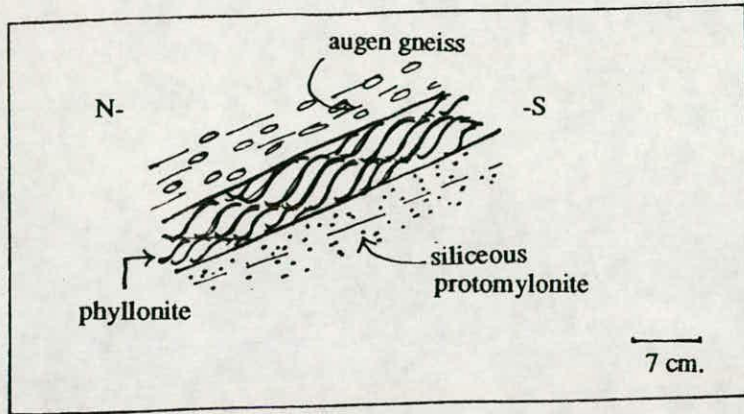


Fig. 3.22 Field sketch from Loc. 15 (SW of Jharkula village) showing difference in the orientation of main foliation in different lithologies. The dominant foliation in phyllonite is slightly steeper than main foliation in the surrounding proto-mylonite and augen gneiss.



Fig. 3.23 Typical appearance of augen mylonite in an exposure near Loc. NR82. View looking E.

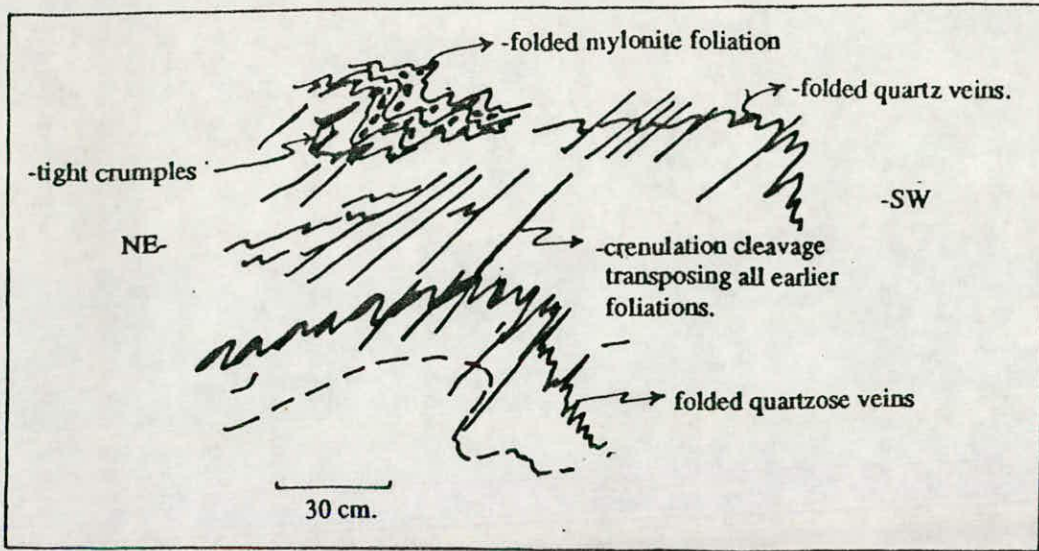
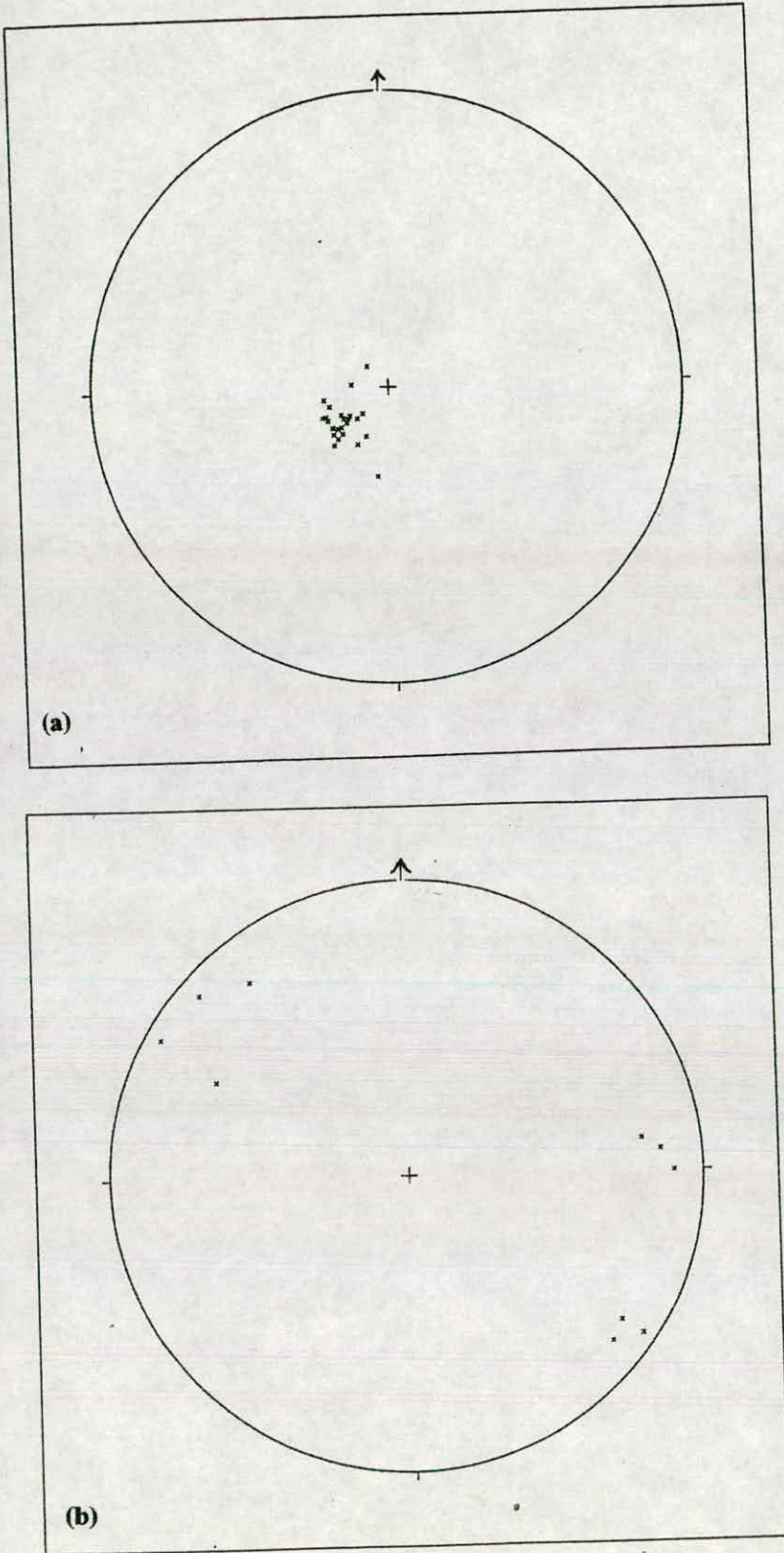


Fig. 3.24 Field sketch showing transposition of an early mylonitic (phyltonitised) fabric by later development of crenulation foliation associated probably with  $F_4$  folding. Loc. MR5 (SW of Animath village where the Lower Bypass Road meets the Main Road).



Fig. 3.25 Mesoscale S-C banding in phyllonite horizon. Pencil follows the trace of a C-band. The waviness of schistosity (S-banding) is conspicuous. View looking eastward. Loc. NR81.



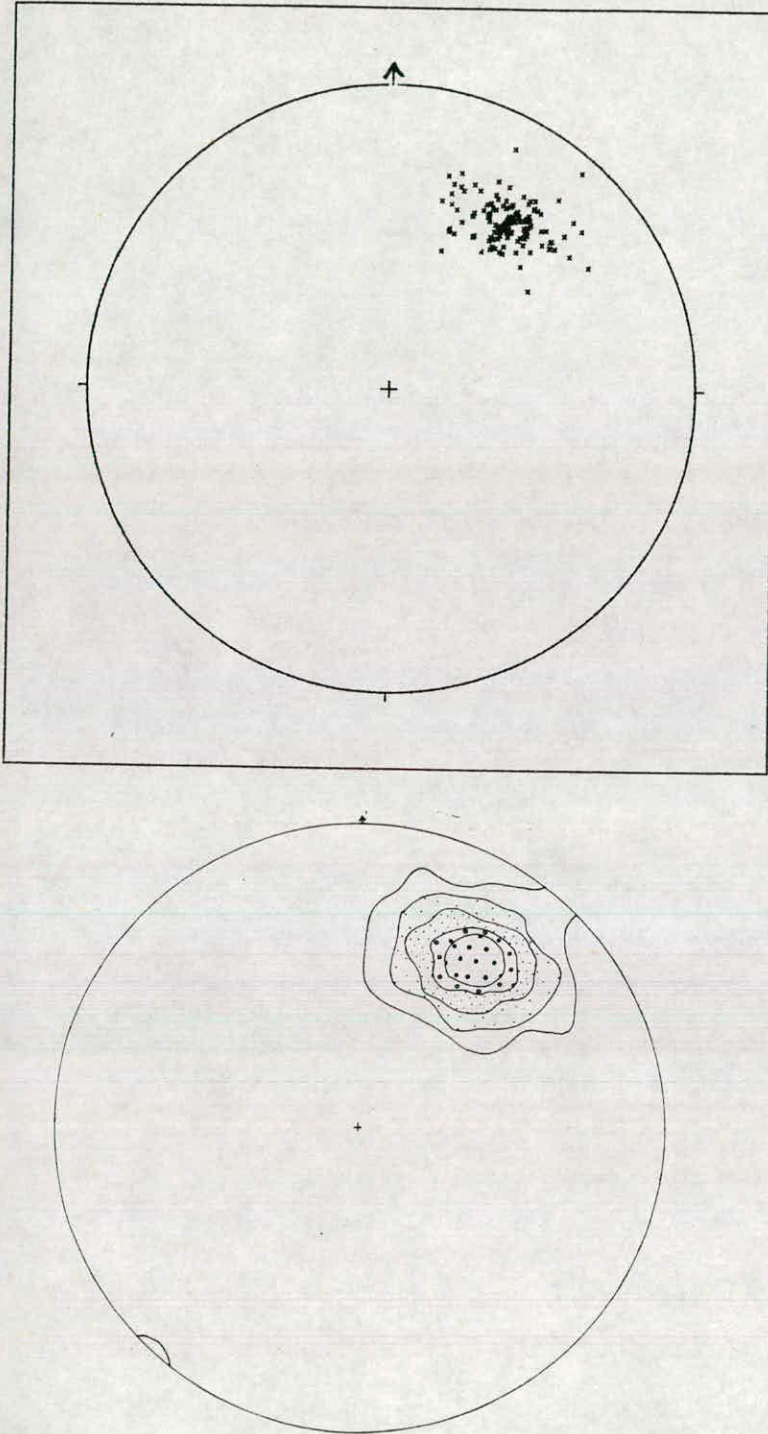
**Fig. 3.26:** (a) 26 poles to shear bands (C-bands), (b) 10 Boudin lines; note the remarkable orthogonality between average boudin line orientation and main stretching lineation (cf. Fig. 3.29c).



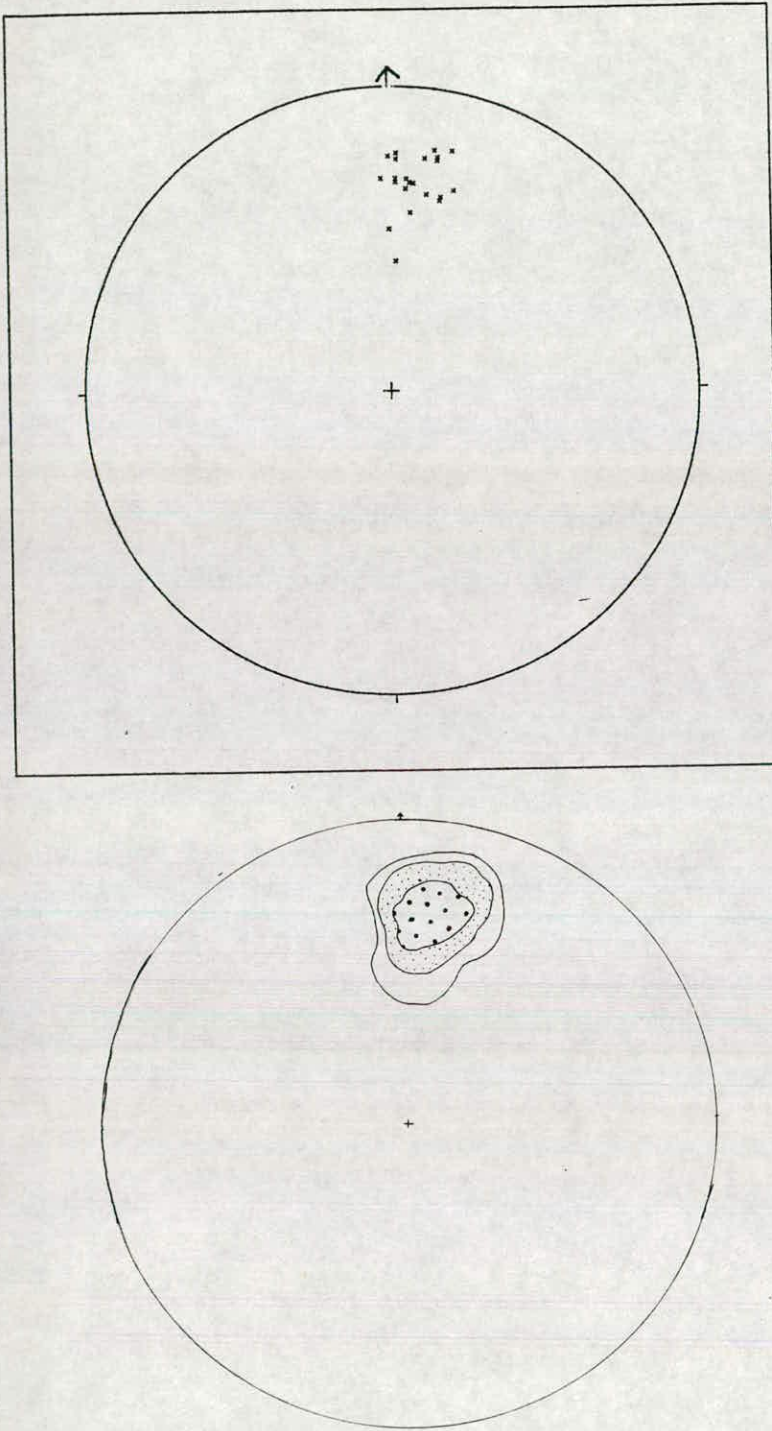
**Fig. 3.27** On an exposure-surface running across foliation, but subparallel to main stretching lineation at Loc. 15 (SW of Jharkula village) feldspar augens of different shapes in augen gneiss. View looking ESE.



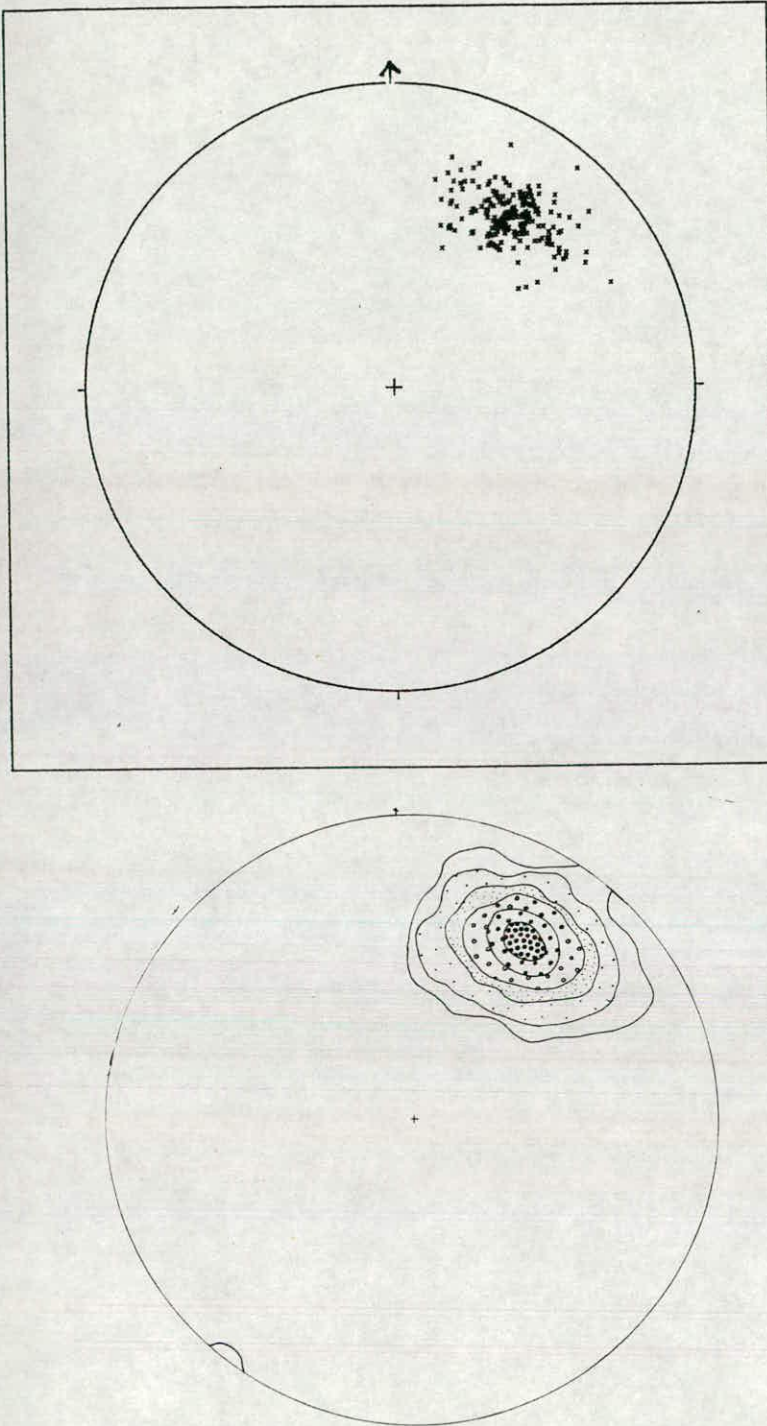
**Fig. 3.28** Conformable orientation of elliptical and/or sigmoidal feldspar and quartz porphyroclasts with the stretching direction. The exposure surface at Loc. 15 (SW of Jharkula village) is across the main foliation and subparallel to main stretching lineation in augen gneiss. View looking E-ward.



**Fig. 3.29a:** 128 stretching lineations(main) from the Munsiri formation. Contours at 0.78-3.906-10.16-28.125-38.28% per 1% area (corresponding number of data points 1-5-13-36-49). Maximum concentration 54.7% (70 points) per 1% area.



**Fig. 3.29b:** 22 later stretching lineations from Joshimath area. Contours at 4.55-13.6-36.4% per 1% area (corresponding number of data points 1-3-8). Maximum concentration 68.2% (15 points) per 1% area.



**Fig. 3.29c:** 200 stretching lineations(main) from across the whole of the Joshimath area. Contours: 0.5-1.5--7.5-12-28-45% per 1% area (corresponding number of data points 1-3-15-24-56-90). Maximum concentration 49% (98 points) per 1% area.



**Fig. 3.29d** Garnet porphyroblasts along with their pressure shadow trails define 'falling fire-balls' feature associated with the N-trending later stretching lineation. Orig. Loc. NR52. View looking downward at an angle, facing S.



(a)



(b)

**Fig. 3.30 (a)** Garnet porphyroblasts partially sliding during the main stretching event got stuck at some stage leaving behind them well-marked trails (grooves). These trails, together with the garnet porphyroblasts, define what may be called as 'upward moving tadpoles' feature. Orig. Loc. NR41. View looking downward facing  $S30^{\circ}E$ .

**(b)** A typical exposure of the main stretching lineation. Loc. 38. (SW of Main Road & Bypass Road junction near Animath). View facing south looking downward at an angle.



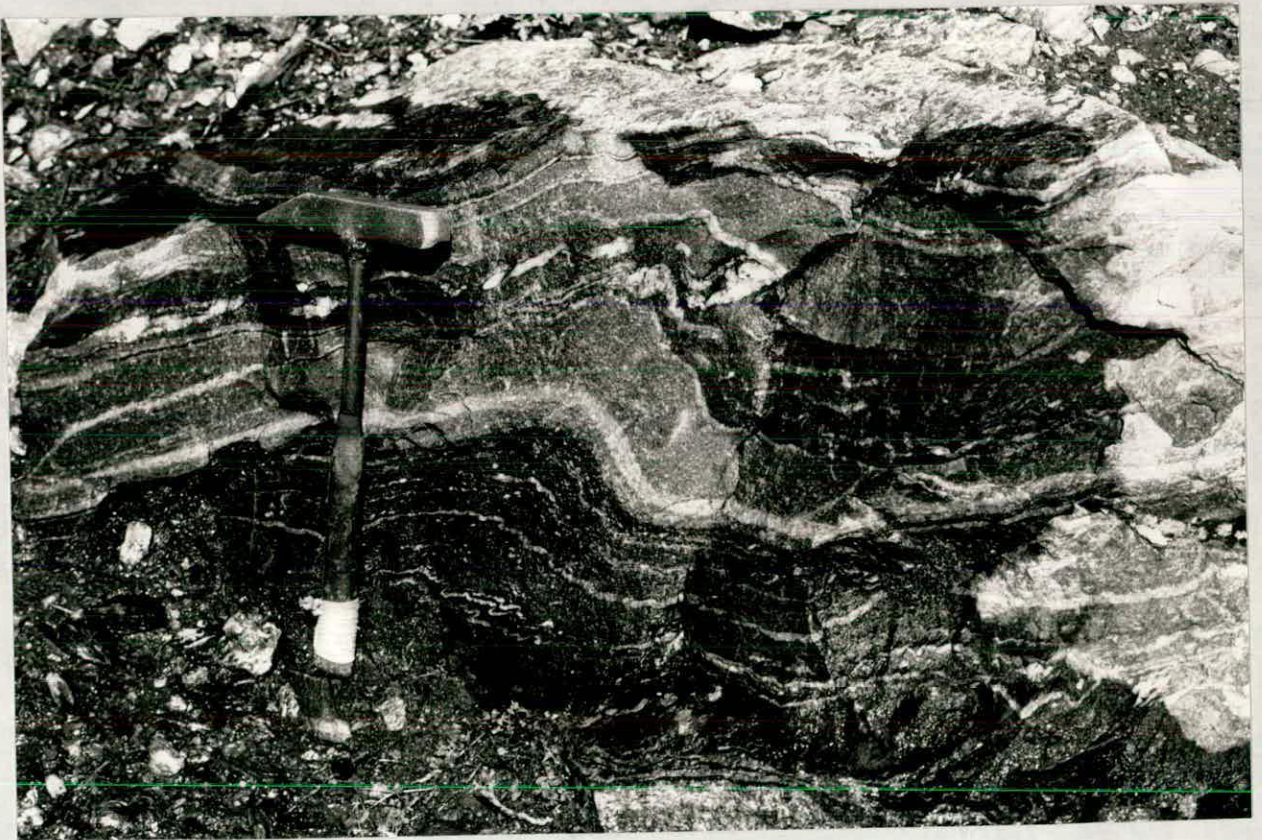
Fig.3.31 A close, angular south-vergent  $F_4$  fold. Note the thin quartzose vein running along the axial trace and the S-shaped minor fold developed on overturned limb. Orig. Loc. 28 (SW of Shelang village). View looking  $N75^\circ E$ .



Fig. 3.32 Progressive increase in tightness towards the core in an example of  $F_4$ -fold. View looking  $N70^\circ E$ , Loc. 18 (WSW of Shelang village). The features in exact core of the fold lying below the lower left corner of this figure is shown in Fig. 3.33.



**Fig. 3.33** The core region of the fold shown above has well developed isoclinal folds giving rise locally to a new set of transposition foliation (axial planar) (cf. Fig. 3.24). View looking N65° E.



**Fig. 3.34** A part of an  $F_4$  fold showing open pattern of folding. Note mylonitic bands of different thickness gave rise to folds of different wavelengths. Loc. 17. View facing E-ward looking down at an angle.

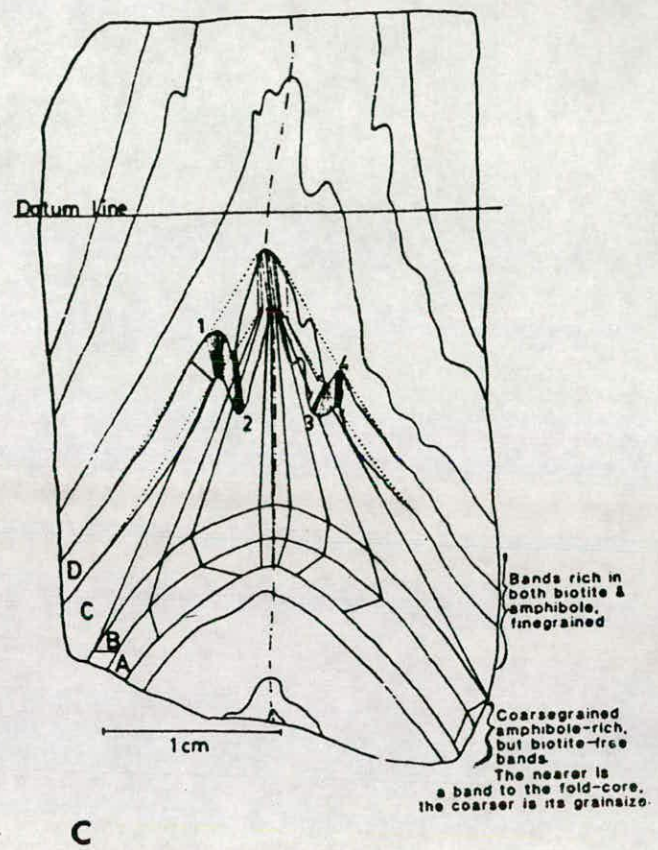
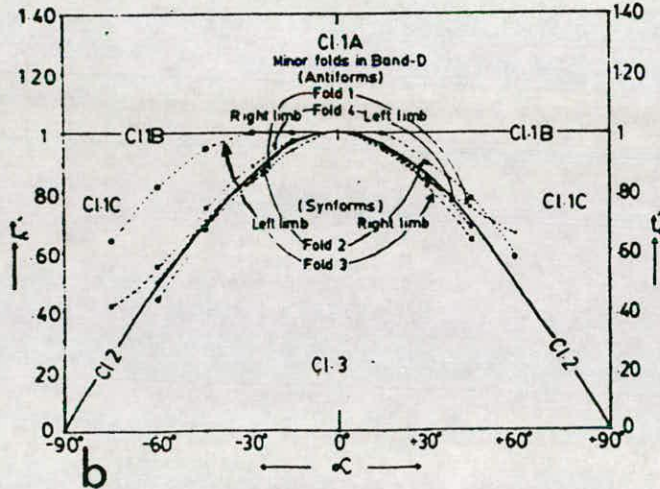
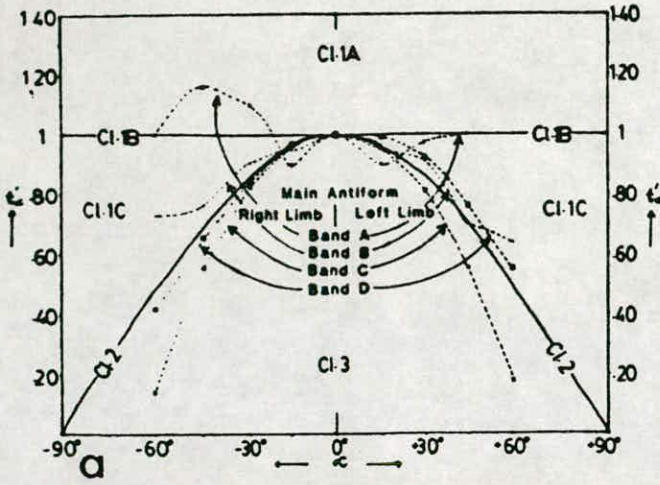


Fig. 3.35 Geometrical analysis of a thin section scale fold ( $F_4$ ) from an amphibolite horizon of Munsiri Formation. See text for details. Slide No. -Loc. 18(2).

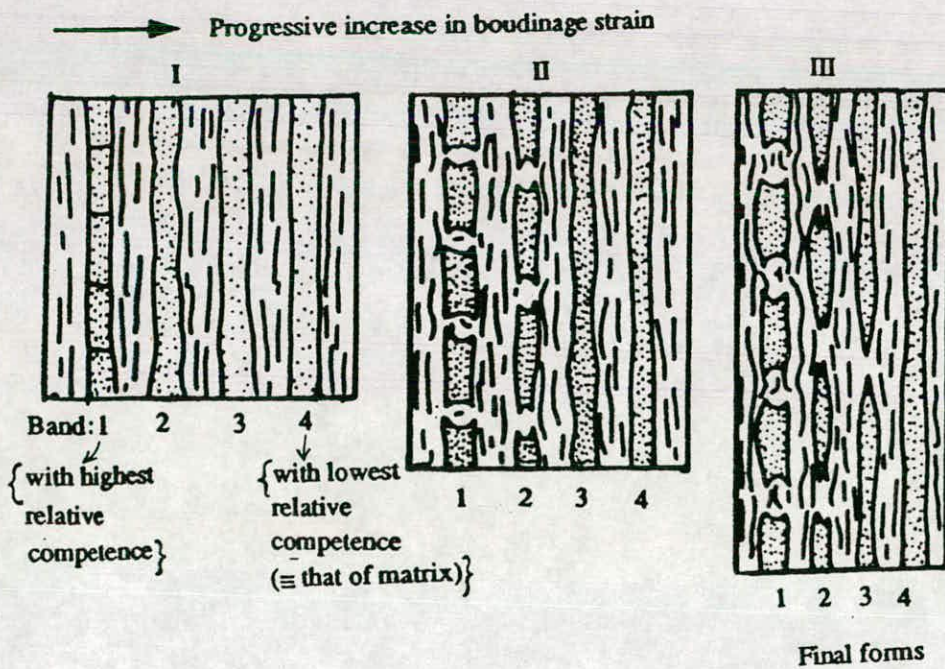
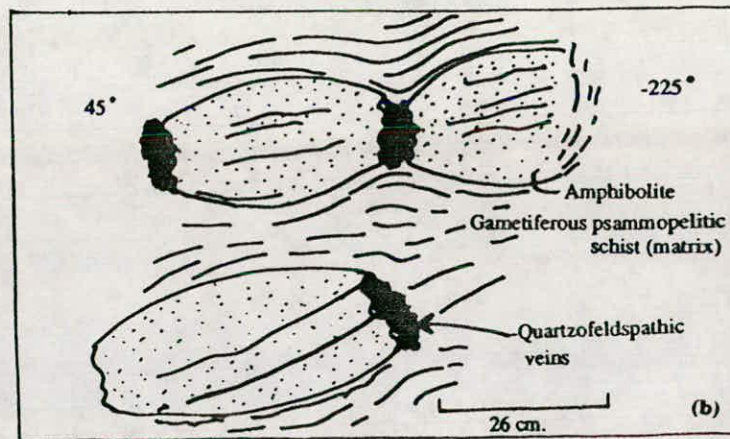
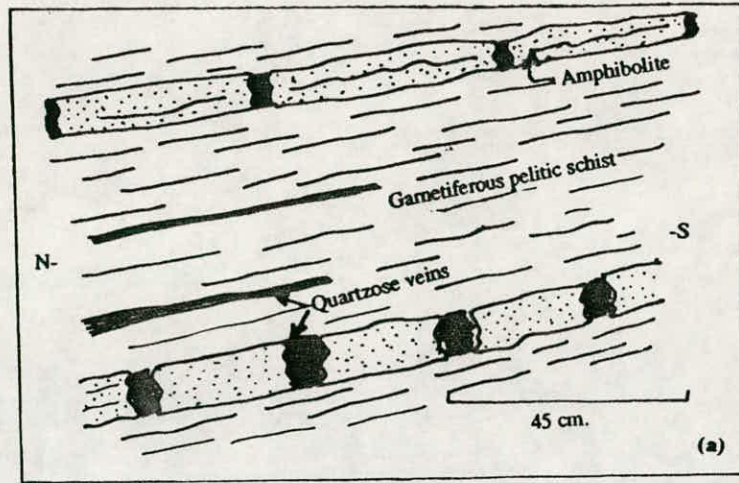


Fig. 3.36 Different types of boudinage (see stage III) depending on the relative contrast in competence between matrix and the boudin角度/s. After Ramsay(1967).



(c)

Fig. 3.37 Different types of boudinage in Joshimath area. (a) Brittle boudinage (field sketch from ~8mtr above road level at Loc. 8, N of Jharkula village); (b) Semibrittle boudinage (field sketch from 1 mtr above road level at Loc. 8); (c) Field photograph showing semi-brittle to quasi-ductile boudinage of a thickly banded quartzite band enclosed within schistose pelite (Orig. Loc. NR61; view looking ESE).



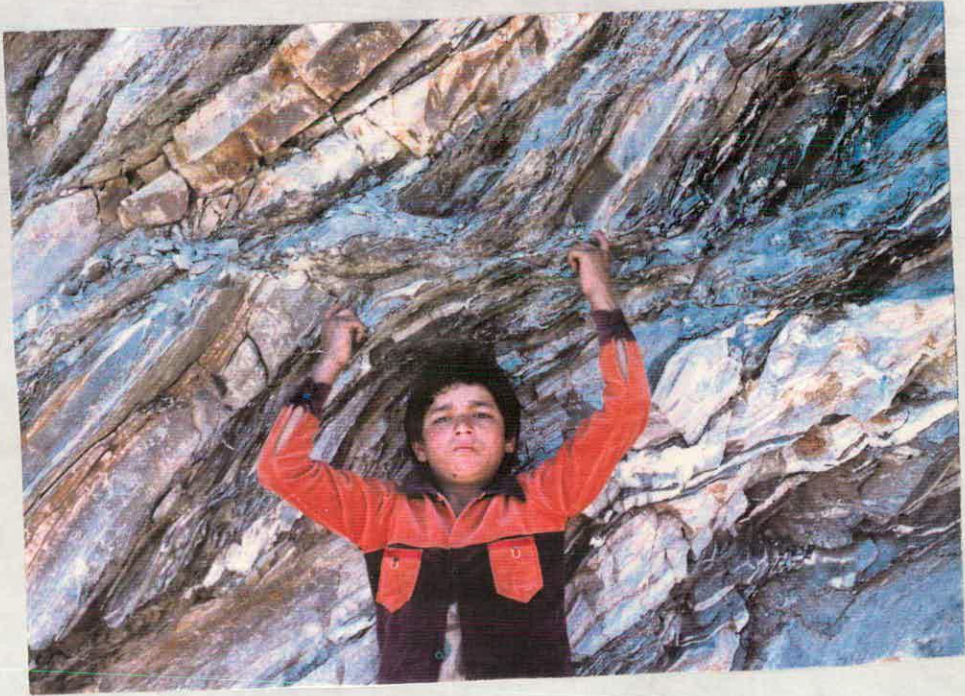
**Fig. 3.38 (a)** Ductile boudinage of amphibolites in a matrix of schistose garnetiferous pelites. Note the dark eye-shaped boudin towards the lower left (under the Brunton compass). View looking E.



**(b)** An enlarged view of eye-shaped boudin. The Brunton compass (as scale) is kept at the same position as above. View looking ENE. Orig. Loc. NR52.



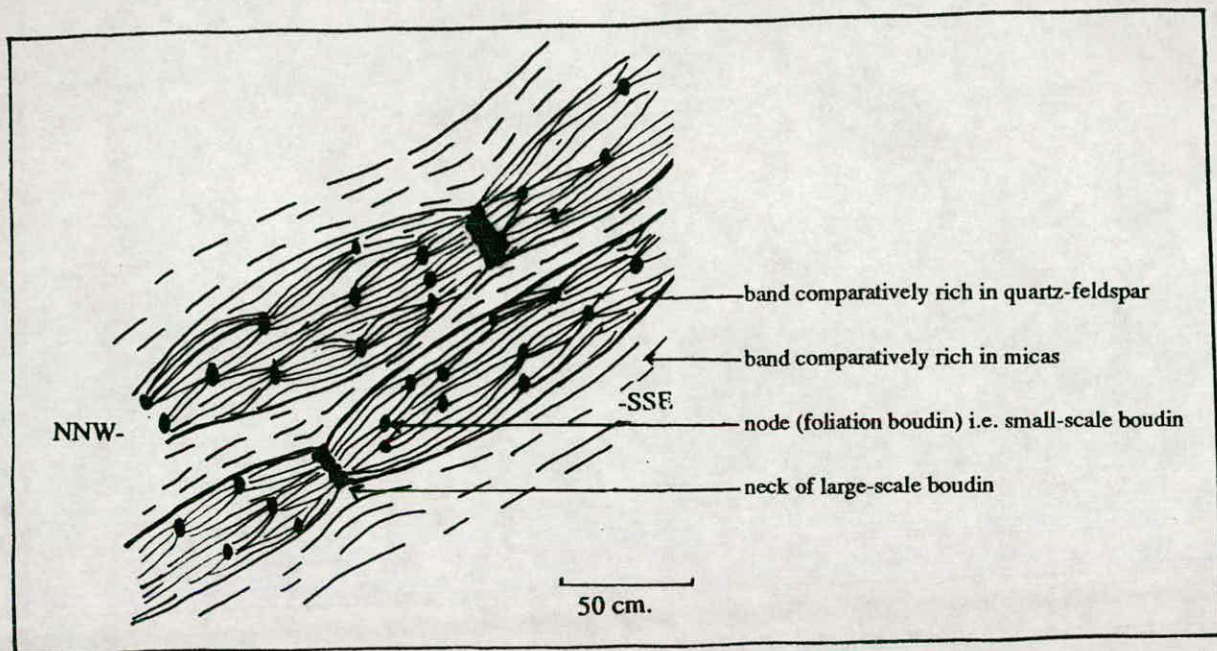
**Fig. 3.39 (a)** Ductile boudinage in a series of more than four amphibolite bands enclosed within garnetiferous psammopelite. Note also a low angle, somewhat listric shear zone cutting across the foliation and boudinaged bands and showing a top-to-right sense of displacement. Probably such shear zones are large scale equivalent of C-bands. Orig. Loc. NR60. View looking ESE.



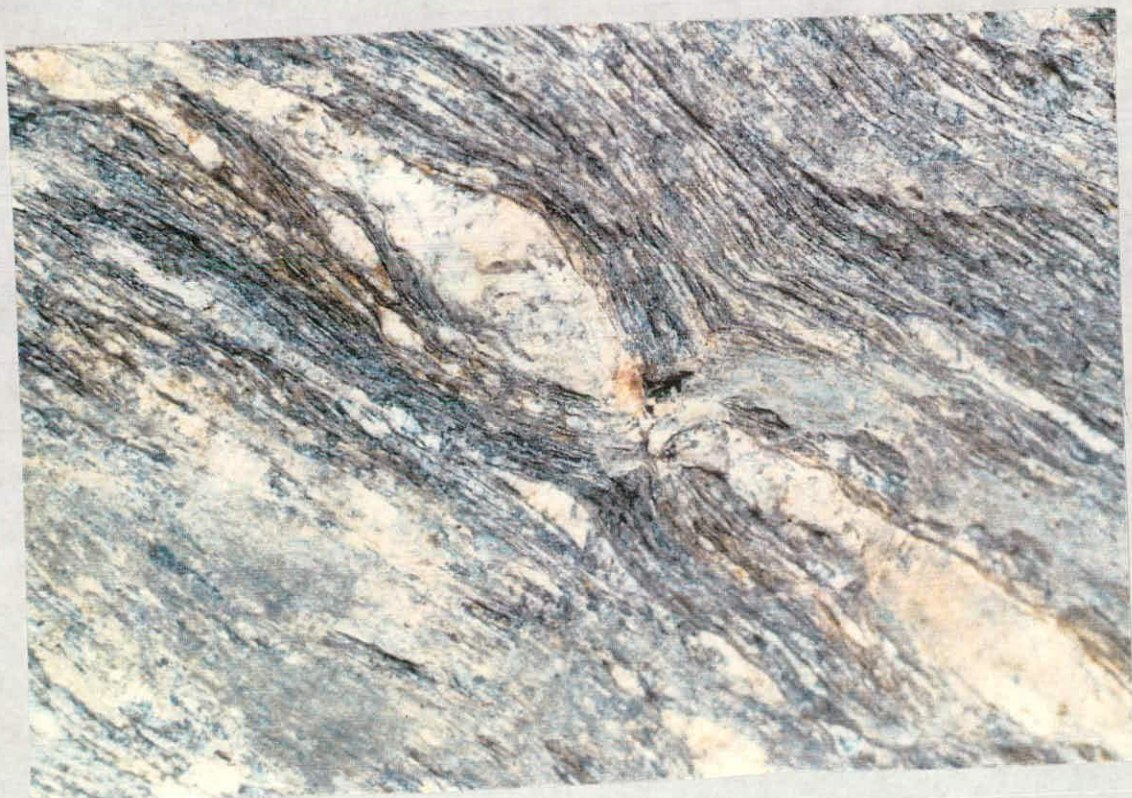
**(b)** Another example of ductile boudinage of amphibolite band/s affected by low angle shear zone having the same sense of displacement as above. View looking ESE. Loc. 17 (S of Jharkula village).



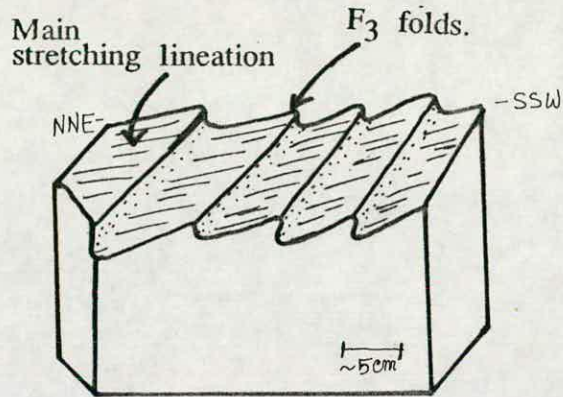
**Fig. 3.40** Example of 'inverse boudinage'. Field photograph looking ESE. Orig. Loc. NR46.



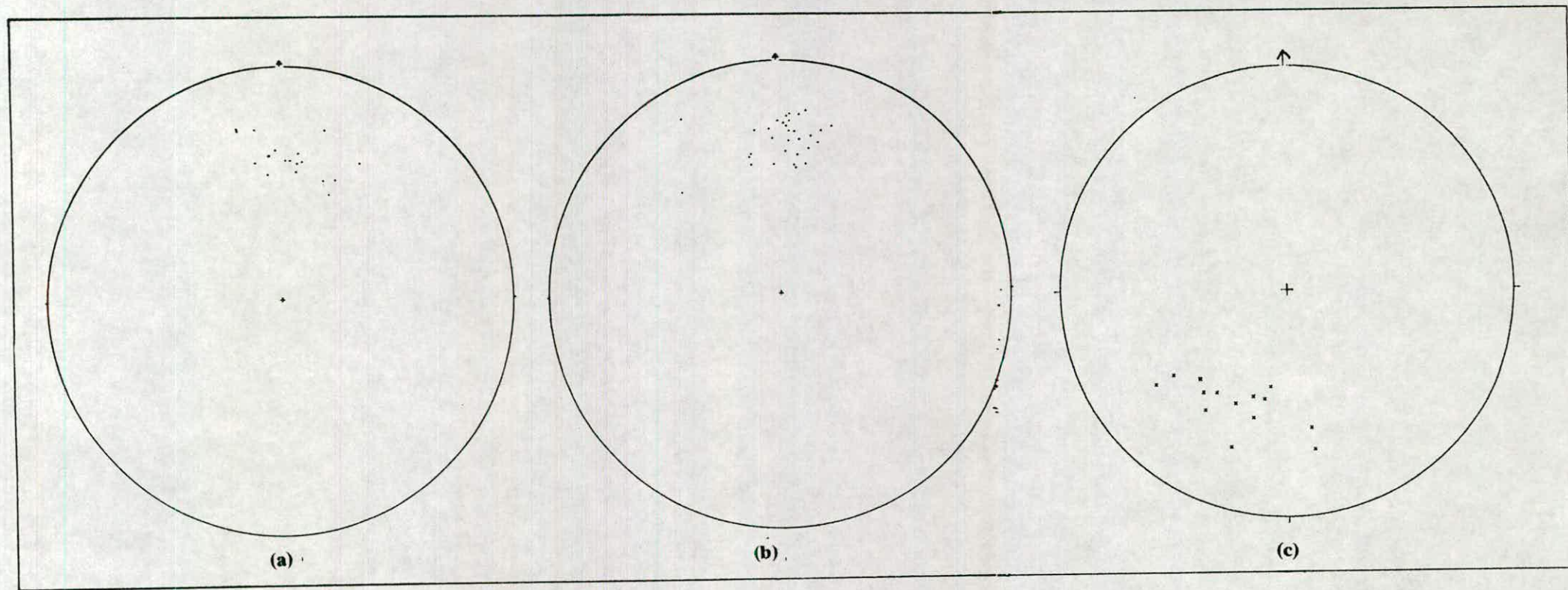
**Fig. 3.41 (a)** Schematic handsketch showing small-scale boudinage (nodding or foliation boudinage) within larger boudins.



**(b)** Field photograph showing close-up of a slightly bigger node (i.e. boudin) in augen protomylonite. View looking WNW. Location: 100 mtr N of Loc. 141 (N of Karchhigaon village). Photograph length 10 cm.



**Fig. 3.42** Schematic sketch showing relation between  $F_3$ -minor folds and stretching lineation. The lineation is distinct on the long, normal limbs of the folds, but not visible on the short limbs; it looks, as if there is a shadow zone for stretching lineation neighbouring the short limbs and synformal hinges. Evidently the stretching lineation cannot be earlier than the folds here ( $F_3$ ). The sketch depicts part of an exposure at Loc. NR31.



**Fig. 3.43:** (a) 19  $F_3$  minor fold axes, (b) 31  $F_3$  puckers, (c) 15 poles to  $F_3$  axial planes.

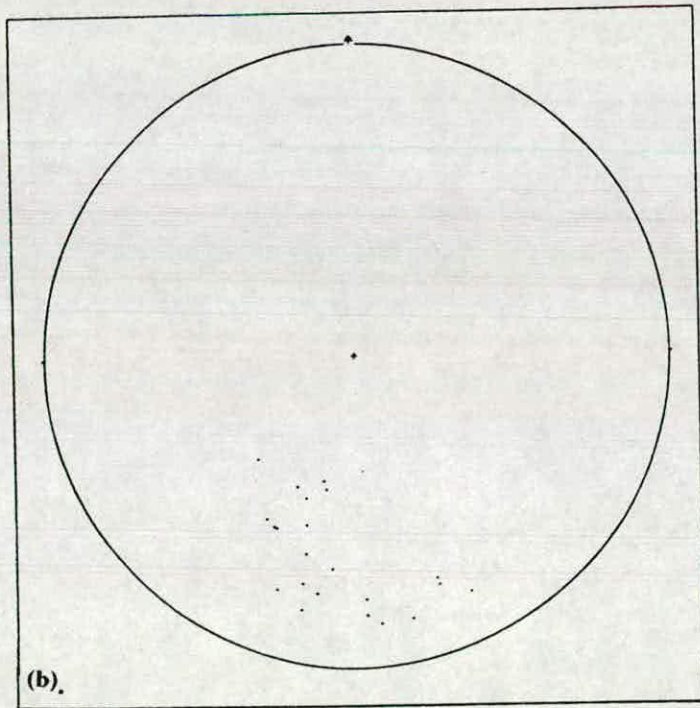
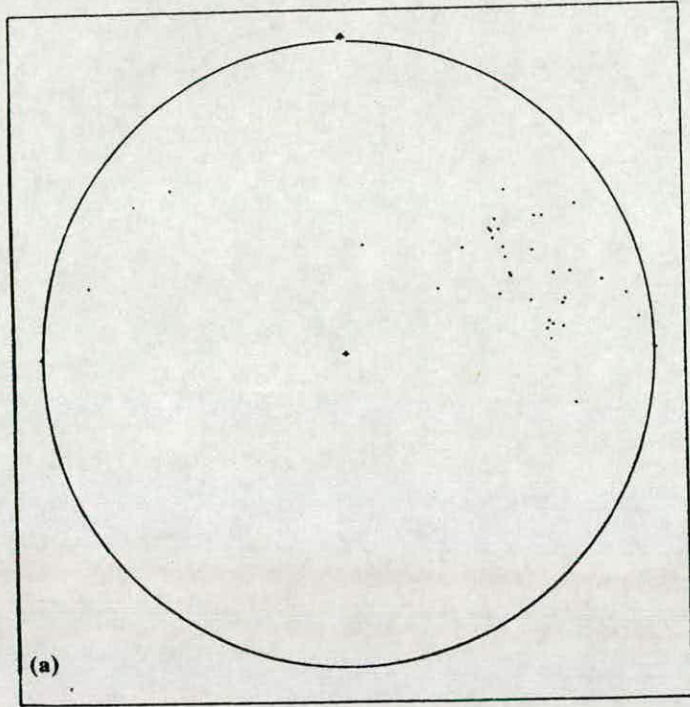
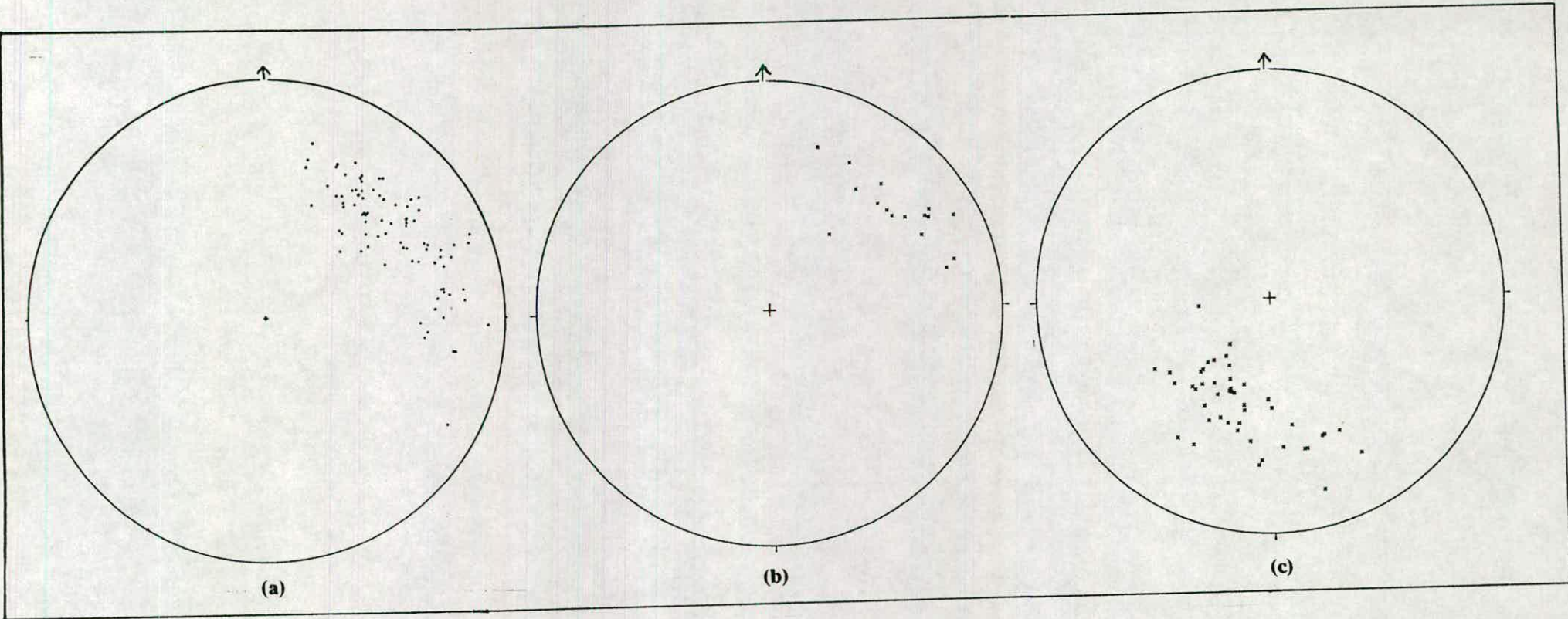
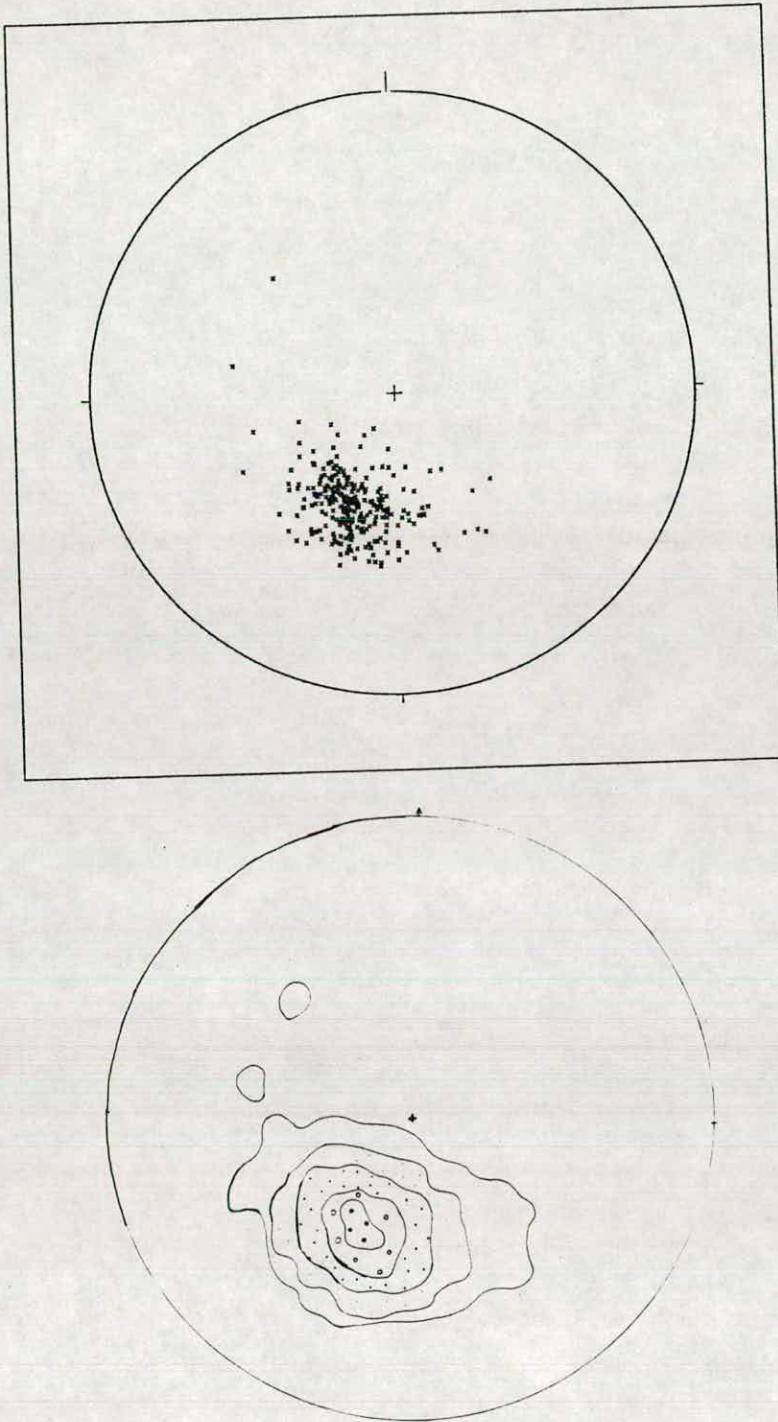


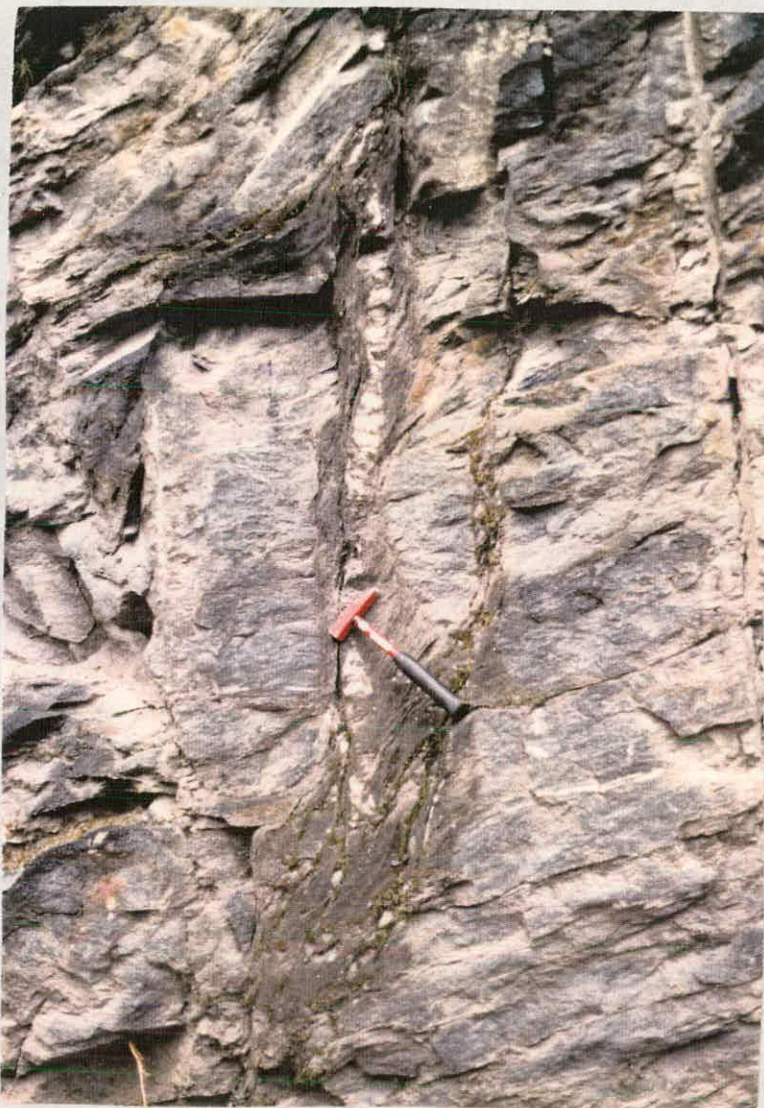
Fig. 3.44: (a) 36  $F_4$  fold axes, (b) 23 poles to  $F_4$  axial planes.



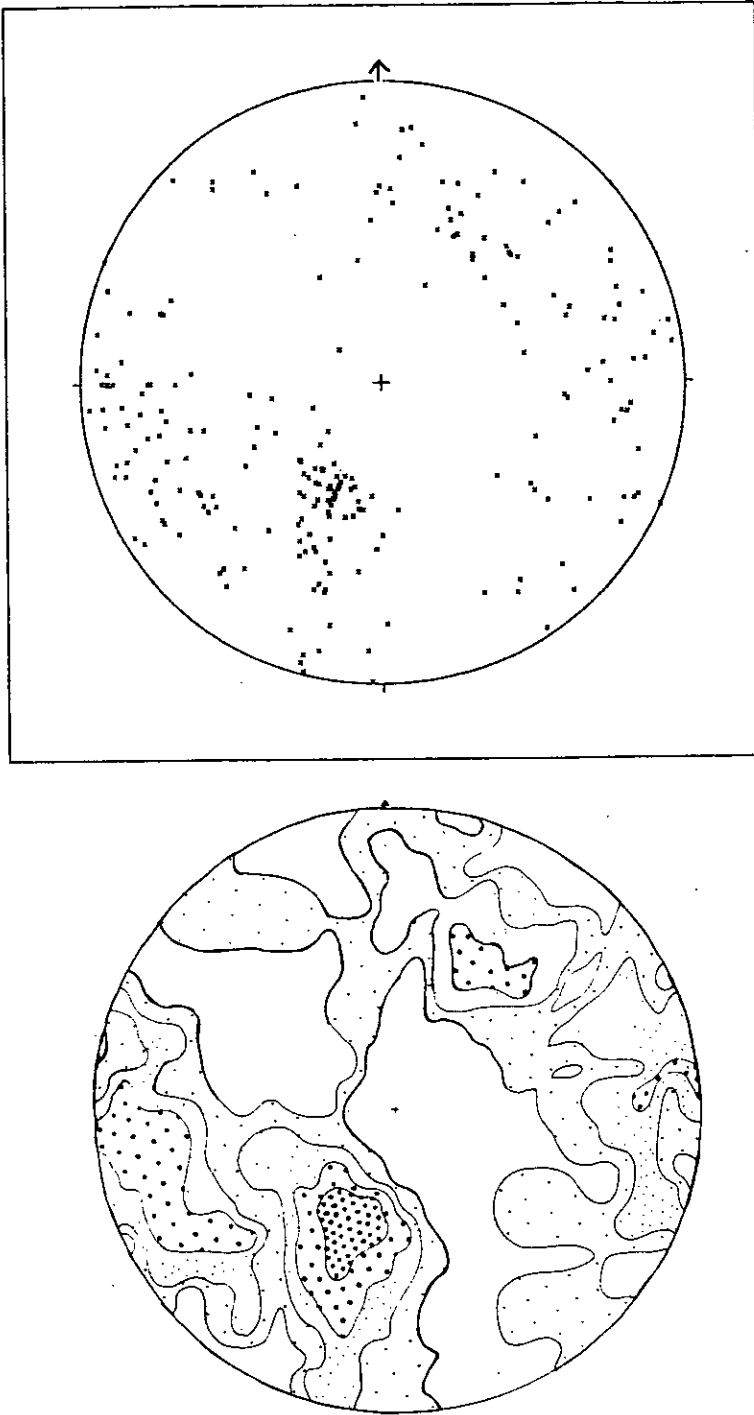
**Fig. 3.45:** (a) 87 axes of  $F_2$  minor fold, (b) 16  $F_2$  pucker axes, (c) 47 poles to  $F_2$  axial planes.



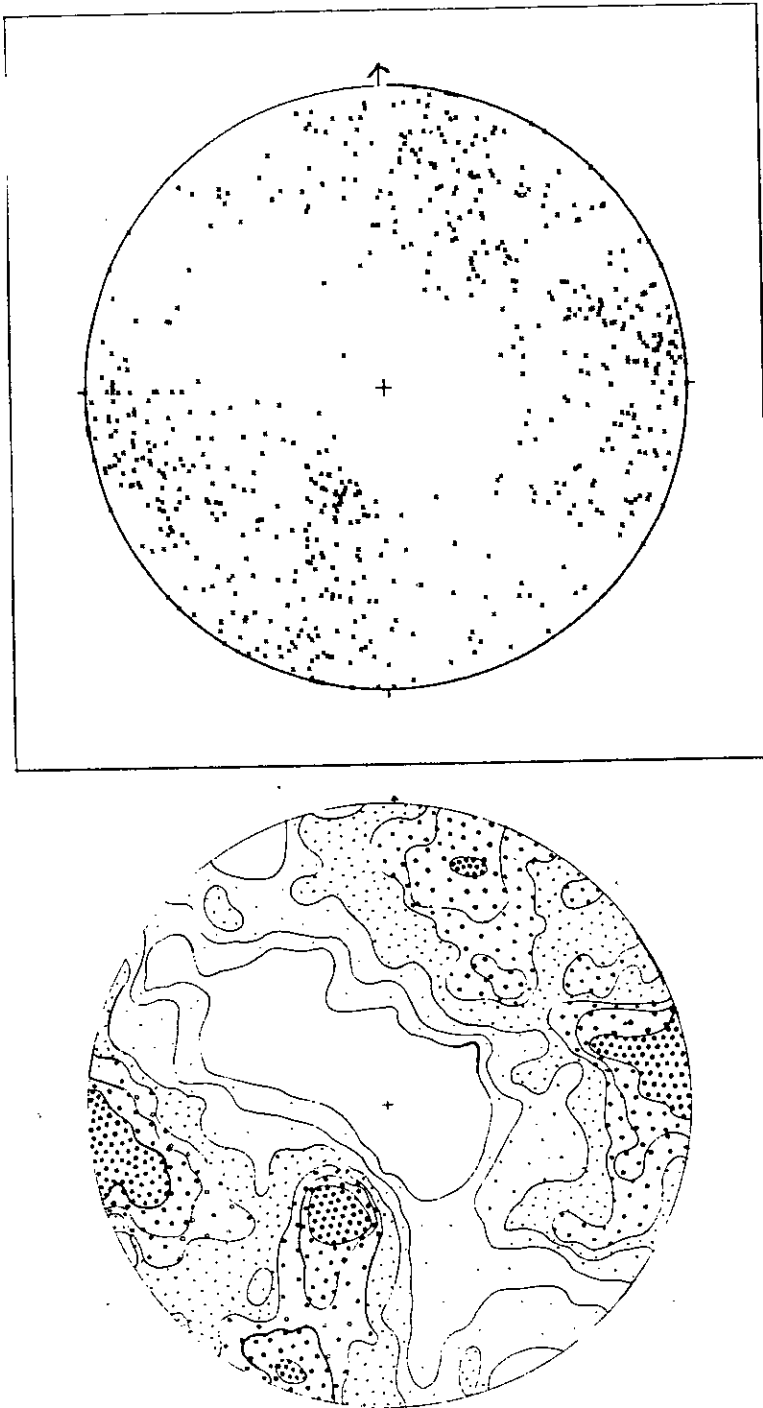
**Fig. 3.46:** 282 poles to main foliation. Composite data from three units in Joshimath area. Contours: 0.35-1.8-6-18-36 percent per 1% area (corresponding number of data points: 1-5-17-51-101). Maximum concentration 39.75 (112 points).



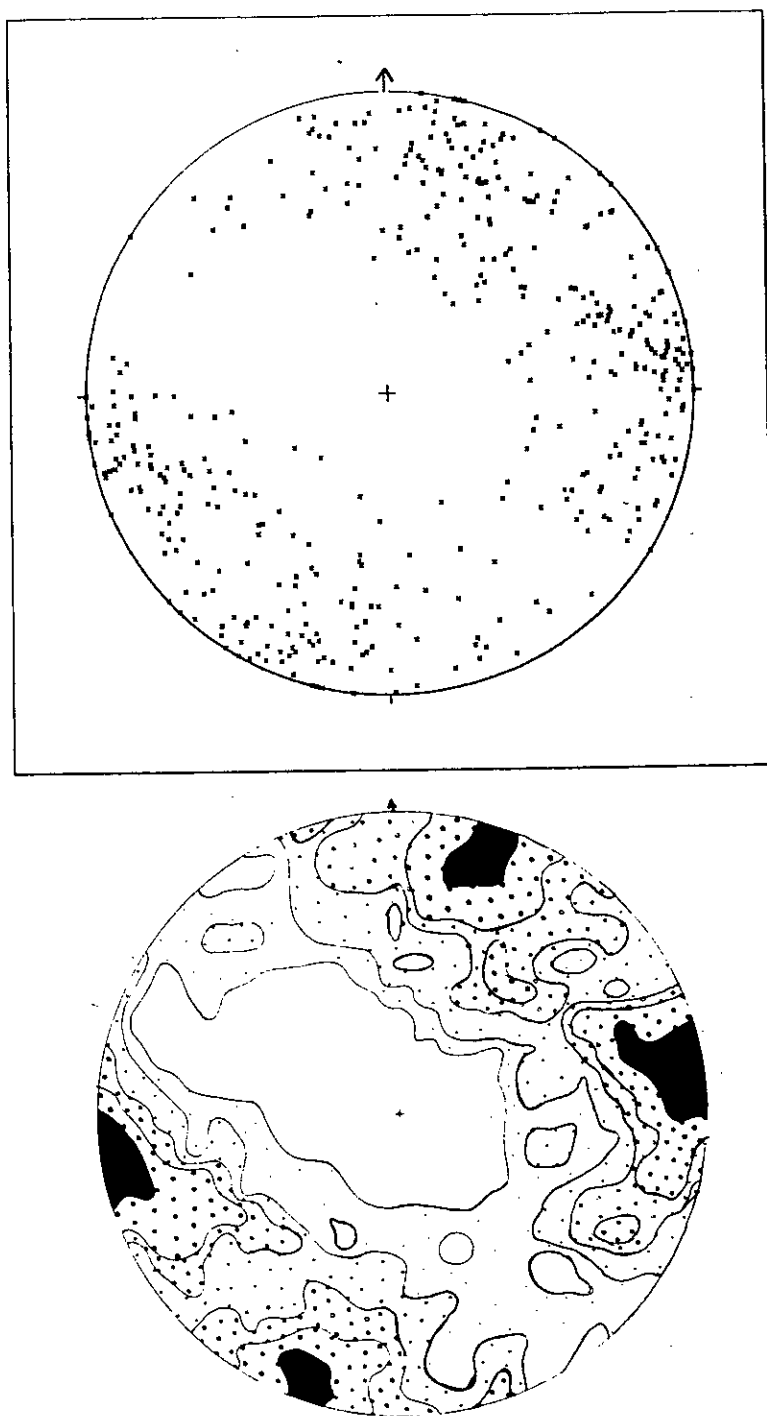
**Fig. 3.47:** Late ductile shear zones occurring in a fine-grained amphibolite horizon in the lower part of the Munsiri formation at a road-cut exposure, mid-way between Helang Bus Stop and the New Helang Bridge. Note the occurrence of differentially 'pinched and swollen' quartzofeldspathic vein along the shear zones. Amphibole is transformed into biotite and chlorite in the shear zones, hence their distinctive dark grey colour. View looking E.



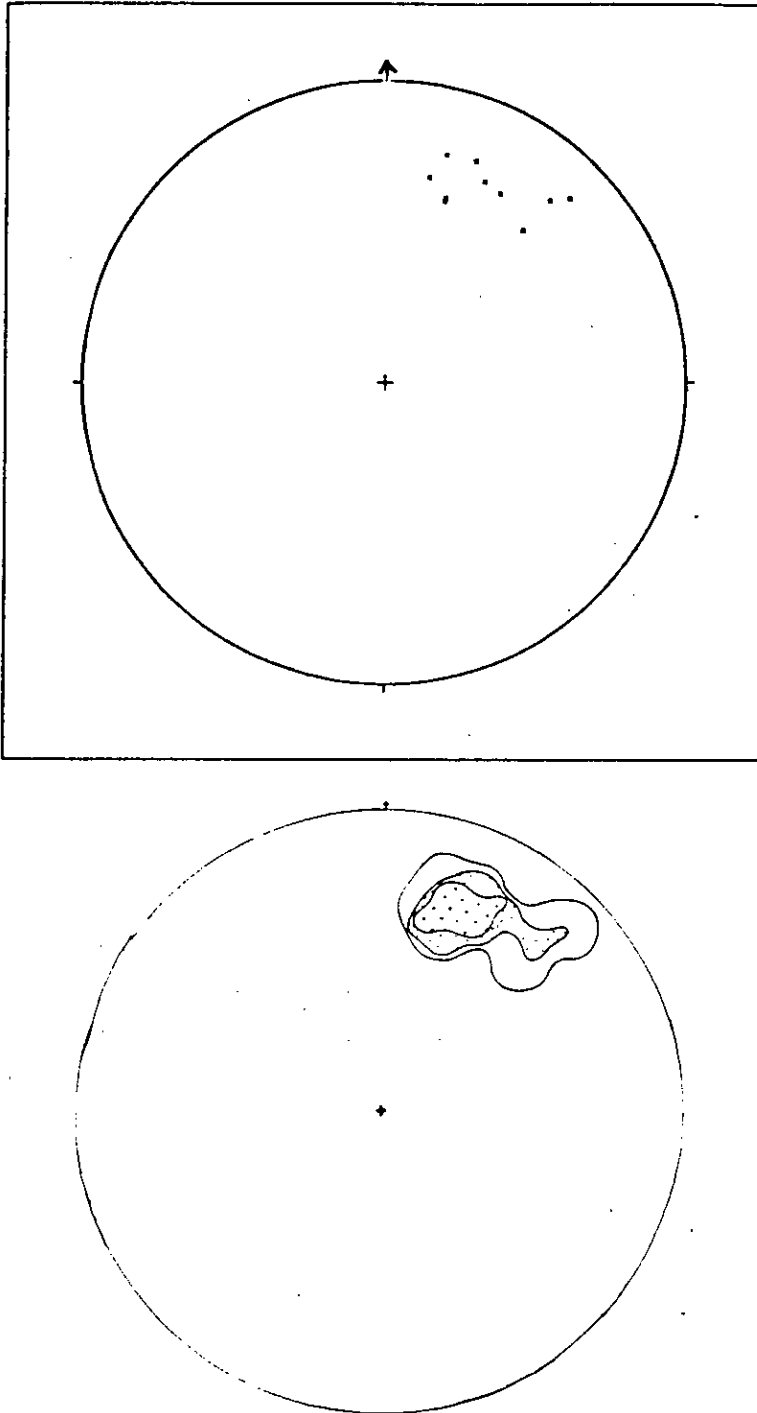
**Fig. 3.48:** 225 poles to faults. Data collected from across the whole of Joshimath area. A considerable number of faults (i.e. movement zones) is parallel to main foliation. Contours at 0.44-1.33-2.66-6.22% per 1% area (corresponding number of data points 1-3-6-14). Maximum concentration 12% (27 points) per 1% area.



**Fig. 3.49:** Composite equal area plot of 643 poles to fractures and faults from across the whole of Joshimath area. Note that a significant number of faults or movement zones is parallel to main foliation (see also next figure for comparison). Contours at 0.155-0.47-0.93-1.55-2.33-3.11% per 1% area (corresponding number of data points 1-3-6-10-15-20). Maximum concentration 4.82% (31 points) per 1% area.



**Fig. 3.50:** 418 poles to late fractures (composite data scattered across whole of the Joshimath area). Presence of a sub-vertical conjugate set is apparent. Note that the maximum compression stress direction is broadly NNW\_SSE. Contours: 0.24-0.72-1.44-2.15-3.83% per 1% area (corresponding number of data points 1-3-6-9-16). Maximum concentration 6.7% (28 points) per 1% area.



**Fig. 3.51: 10 probable main stretching lineation from the Berinag Mandhali formations. Contours: 10-20-30% per 1% area (corresponding to data points: 1-2-3). Maximum concentration 50%(5pts). The maximum density contour (i.e. the 1 point contour) has been drawn by Mellis(Circle) method and in absence of sufficient data this is the most significant contour.**

## Chapter-4

# THE MAJOR THRUSTS

### 4.1. INTRODUCTION

In this chapter an attempt is made to establish the location and the nature of the Vaikrita Thrust and the Munsiri Thrust, mainly using field evidence. The major gaps in our knowledge related to the thrusts are concerned with:

- (1) the positioning or delineation of the thrusts
- (2) the nature of the thrusts viz. whether they are ductile or brittle and the precise direction and magnitude of transport along them etc.
- (3) the age relationship between the two thrusts.

The object of the present chapter is to attempt to fill these gaps. To accomplish this the discussion will be made in the following sequence:

Firstly, a traverse across the Vaikrita Thrust will be described in order to document the nature of the Joshimath Gneiss and Munsiri Formations in its vicinity. Secondly, the Munsiri Thrust (Gansser's Main Central Thrust) will be described to show the natures of the Munsiri formation and Berinag-Mandhali formations near to its supposed trace. Thirdly, in the light of these accounts I shall discuss briefly the nature of ductile shear zones and slides, the aim being to see which corresponds well with the observed characteristics.

Before going into the discussion as proposed above, it must be noted that there are many controversies in the literature regarding the thrusts in the area. Such controversies are related mainly to three aspects- (a) Number of thrusts, (b) Location of the thrusts, and (c) Nomenclature.

At the root of all these <sup>controversies</sup> are the basic stratigraphical uncertainties surrounding the rock-units. Because most of the rocks have suffered intense deformation and/or high degree of metamorphism, we cannot simply apply general criteria for recognition of the thrusts. Occurrence of high-grade metamorphics overlying low-grade ones does not necessarily prove thrusting. Radiometric ages of the associated rocks are poorly constrained, though the general opinion is that the successively overlying units are

comparatively older (see Gupta et al., 1982, p. 214; Windley, 1983, p. 851). The kinematic indicators which are particularly common within the Munsiri Formation support a SSW-ward thrusting sense. Le Fort (1975) clearly points out that workers on Himalayan Geology, particularly those prior to mid-seventies tended to use innumerable names for Himalayan thrusts without suggesting or attempting any useful correlation. And yet barring only a few publications (e.g. Naha & Ray, 1971; Ray & Naha, 1971), it is extremely difficult to know the criteria used by different workers for the recognition of the thrusts. As pointed out by Le Fort (1975), presumably one or more of the following points were used by different workers for recognition of thrusts in the Himalayas:

- a) Abnormal superposition suggested on the basis of assumed stratigraphic sequence
- b) Sharp modification of the lithology
- c) Cataclasis and diaphoresis on a large scale
- d) Injection of basic sills
- e) Apparent jump in the intensity of the metamorphism
- f) Reverse metamorphic zonation
- g) Retrogression effects
- h) Change in the tectonic pattern
- i) (Previous recognition)

Given (i) the unfossiliferous nature and unclear stratigraphy of most of the Higher Himalayan and Lesser Himalayan rocks involved in thrusting, (ii) the deepseated origin and metamorphic environment of the thrusts, (iii) absence of conspicuous angular discordance between the thrusts and foliation in the thrust formations and (iv) control of metamorphic mineralogy, in most cases, by varied primary rock compositions, the abovenoted criteria are not very useful for recognising the 'peculiar' Himalayan thrusts.

Heim & Gansser (1939) and Gansser (1964) recognised one thrust i.e. the Main Central Thrust (MCT), separating the Higher Himalayan crystalline rocks from the low-grade Lesser Himalayan rocks. Gansser (1964, pp.103-104) considered it to be the major tectonic element of the Higher Himalayas which is well defined both geologically and geomorphologically. According to him, 'One could enter the High Himalayan Ranges only after crossing this major structural element and from here the rise of the high mountains is sudden and abrupt. Along most contacts, at or near this Main Central Thrust, occur the widespread amphibolitic sills of gabbro-dioritic composition.' Whereas Valdiya (1980) recognised two major thrusts-- the Munsiri Thrust (i.e. the lower one and

equivalent to Gansser's MCT) and the Vaikrita Thrust (i.e. the higher one). Valdiya also points out that the MCT as identified by Bordet (1973), Hashimoto et al. (1973) in western Nepal is more in line with his 'Vaikrita Thrust' in Garhwal-Kumaun. In terms of the tectonostratigraphic subdivision suggested by Valdiya, the Main Central Thrust of Gansser (Munsiari Thrust of Valdiya) forms the boundary between the Munsiari Formation and the underlying Berinag-Mandhali formations. The second thrust --the Vaikrita Thrust -- is placed between the Munsiari and the overlying Joshimath Gneiss. Thus, according to Valdiya, the Munsiari Formation is bounded by two major thrusts-- Vaikrita Thrust above and Munsiari Thrust (Heim & Gansser's MCT) below. For descriptive purposes in the present chapter, the lower thrust will be referred to as 'Munsiari Thrust' and the upper one as the 'Vaikrita Thrust'. It is commonly believed that the whole Munsiari Formation is a schuppen zone bounded between these two thrusts (Roy & Valdiya, 1988; Valdiya, 1988).

Unfortunately, nowhere is it mentioned where exactly one can find these thrusts in the outcrops, nor is the nature of these thrusts clearly discussed in the literature. Prof. K.V. Hodges (1988) comments-- ".....We must recognise that 'Main Central Thrust' has become an unfortunate generic term for any fault-zone that separates the Greater and the Lesser Himalaya. It is quite likely that the MCT as mapped in Garhwal is totally unrelated to the MCT as mapped in Darjeeling ..." (Hodges et al., 1988, p.279). Brunel (1986) states, 'Large thrusting along the MCT of the highly metamorphic rocks over less metamorphic, is now an indisputable fact, but the precise location of this thrust is not well-defined.' In the Nepal Himalaya apparently one thrust is present which is recognised as the MCT, instead of the two as recognised by Valdiya in Garhwal. In Nepal the MCT is chosen by many workers as a lithological change below the lowermost gneiss layers of the Higher Himalayas, while others consider it to be represented by the line marking the first appearance of kyanite (LeFort, 1975; Stocklin, 1980).

Indiscriminate and nonstandardised nomenclature is a crippling problem in Himalayan Geology as a whole, which is why any sort of correlation in the Himalayas is so difficult (see Rupke, 1974; Le Fort, 1975). Confusion arises, for instance, when Valdiya (1980, p.107) proposes to rename Gansser's 'Main Central Thrust' as the 'Munsiari Thrust' and also when he suggests renaming his 'Vaikrita Thrust' as the 'Main Central Thrust'. As mentioned before, for the sake of consistency in terminology for descriptive purposes, the upper thrust will be referred to as 'Vaikrita Thrust' and the lower one as the 'Munsiari Thrust' in this chapter.

#### 4.2. TRAVERSE ACROSS THE VAIKRITA THRUST

The boundary between the Joshimath Gneiss Formation (Vaikrita Group) and the Munsiri Formation represents the Vaikrita Thrust which Valdiya (1980) regarded as the real MCT that divides the higher grade crystallines of Higher Himalayas from the low-grade Lesser Himalayan rocks. The main lines of evidence put forward by Valdiya (1980) favouring the status of this boundary as a thrust are as follows--

- (1) An abrupt change in the grade of metamorphism from the "granulite facies" of the overlying Vaikrita (Joshimath) Gneiss to greenschist-- lower amphibolite facies of the underlying Munsiri Formation. He says, "The appearance of kyanite (and sillimanite) bearing schists and gneisses with euhedral garnet as large as a half to one centimetre across just above the schists made up of sericite-chlorite and very fine-grained biotite, with very locally developed tiny garnet 1 to 2 mm. across, cannot be interpreted as signifying gradual increase in metamorphic grade".
  - (2) The "pervasive retrogression" and profound mylonitisation in the rocks of the Munsiri Formation are products of enormous shearing which he attributes to the overriding of the thick Vaikrita sheet.
  - (3) The style and pattern of folding in the Munsiri Formation and the Vaikrita formation are quite different. While the Vaikrita formation shows one set of disharmonic sigmoidal isoclinal recumbent plastic folds, there are multiple generations of folding in the Munsiri Formation.
  - (4) Some of the hot springs of the Great Himalaya are sited on the Vaikrita Thrust, or on its related faults, for they spout hot water in its close proximity.
- and,
- (5) The seismic plane with a dip of 40° towards NE recognised by Kaila & Narain (1976) corresponds more closely to the 30°-45° dipping Vaikrita Thrust than to the low-dipping (5°-20°) Munsiri Thrust (Gansser's MCT). Thus the high seismicity of the belt is attributed to active movement along the Vaikrita Thrust.

Now a critique is given of these lines of evidence:

##### (1) Is there a metamorphic break?

Observation along sections across the Vaikrita/Munsiri boundary does not always clearly indicate an abrupt break in the grade of metamorphism. For instance, along the

Joshimath-Helang main road section the thick horizons of highly garnetiferous schistose psammopelitic rocks N of Jharkula chatti (village) belonging to the upper part of the Munsiri Formation are probably a medium grained equivalent of Joshimath (Vaikrita) Gneiss. Also the Joshimath Gneiss is not of 'granulite' facies (see Chapters 5 & 6). Some horizons in the Upper Munsiri schists contain small fragments of staurolite crystals (vide subsection- 5.3.5 in Ch.5). Such occurrence of staurolite in the upper part of the Munsiri Formation some distance below the kyanite-bearing Joshimath Gneiss would naturally indicate a progressive downward decrease in the grade of metamorphism.

However, in the lower road section (along the Bypass Road now under construction following the Alaknanda River) which shows a much more complete and continuous series of exposures, we find a rather sudden appearance of fine-grained (average grain-size less than or equal to 0.2 mm) retrograde-looking (shale-grey in colour) schistose horizons below a kyanite and garnet bearing, fragmentary textured horizon of Joshimath Gneiss (avg. matrix grain-size more than or equal to 0.5 mm, garnet-size 1.0 mm, kyanite-lengths 3 mm). In the exposures the garnets in the retrograde looking schistose rocks appear to be in disequilibrium with the surrounding matrix, biotites are retrograded to chlorite. Clearer evidence of retrogression is found in the nearby amphibolite horizons where amphibole is being replaced by biotite. If there is any thrust that separates the Joshimath Gneiss from the Munsiri Formation, it must pass through the zone separating the above mentioned retrograde schistose rocks from the overlying kyanite-bearing Joshimath Gneiss horizon. See Figs. 4.1 & 4.2 for a complete log across this boundary and Fig. 4.3 for the details of the exposure showing this boundary. The section (Fig. 4.3) shows five different lithological horizons, termed A to E for convenience. Among these, E can be regarded as a more schistose equivalent of D (avg. grain-size 0.1 - 0.2 mm). In between D and E, there is a disturbed (movement) zone. This zone is about 20 cm wide including a profusion of quartzofeldspathic veins occupying what were originally an anastomosing array of fractures; the upper boundary of this zone is a clean cut (discrete) discontinuity surface. Horizon B is the most fine-grained of all with average grain-size  $\ll 0.1$  mm. and a distinct flagginess; it is psammopelitic in composition. The uppermost horizon A containing kyanite, garnet etc. is typically a part of Joshimath Gneiss (average quartzofeldspathic grain-size more than or equal to 0.5 mm). However, the contact between this gneissic horizon A and the flaggy horizon B does not appear to be a clear cut faulted contact.

Now there are three possibilities as to the specific location of the Vaikrita/Munsiari boundary (the Vaikrita Thrust of Valdiya).

The boundary is,

- (a) In the disturbed zone in between D and E;
- (b) In the flaggy horizon B;
- (c) Represented by the contact between A and B horizons.

Even though it doesn't appear like a clearcut tectonic contact, the A/B boundary evidently indicates a sudden break in average grain-size and metamorphic grade at least locally. It is mainly on these grounds that this contact is considered as the boundary between the Joshimath Gneiss and the Munsiari (Valdiya's Vaikrita Thrust). This contact is structurally much above the staurolite-occurrence mentioned earlier. On the foot track from Vishnu Gad foot-bridge to Kalpeswar, on the west of Alaknanda River, there are exposures north-west of Loani village that show clear differences in the grain-size and metamorphic characters of the rocks. Sulphur and iron oxide-stained, kyanite and garnet-bearing coarse-grained Joshimath Gneiss horizons are found to lie above the comparatively fine-grained, fresh-looking muscovite-rich schistose rocks of the Munsiari Formation which make up the local hillslope. Apparently much of the perfectly planar hillslope (dip slope) below Loani village follows the orientation of the Vaikrita/Munsiari contact.

Roy and Valdiya (1988, p.115) were so impressed by the disparities in metamorphic grade between the Munsiaris and the Joshimath Gneiss that they went so far as to claim that metamorphism in the Munsiari was pre-Himalayan, while that in Joshimath Gneiss is Himalayan. They specifically suggested that the index minerals (e.g. staurolite) in the Munsiari Formation are of pre-Himalayan origin whereas those (e.g. kyanite, sillimanite, etc.) in the Joshimath Gneiss are synkinematic i.e. syn-Himalayan or syn-thrusting in origin. In my view this conclusion cannot be supported and such a strong disparity is unacceptable (see Chapter-5 for details). Whatever disparity is apparent could probably be explained if we think in terms of a limited amount of transport along the Vaikrita Thrust. This is not unlikely when we consider the close chemical (mineralogical) and textural resemblance between the upper Munsiari psammopelites and pelites with the Joshimath Gneiss (see Chapters 2, 5 & 6 and Appendix -II). The difficulty in clear cut recognition of the thrust may also be partly attributed to this reason, as suggested by the section (Fig. 4.3). Another important point to note is that the Vaikrita rocks (Joshimath Gneiss) or their equivalents are not reported from any klippe existing in the Lesser Himalaya, whereas Munsiari rocks are reported from most of the Lesser Himalayan

klippen. Probably, erosion cannot be the sole factor for this; limited movement along the Vaikrita Thrust in comparison to that along the Munsiri Thrust may be another reason.

Valdiya's evidence of decreasing garnet size is not valid because there are some horizons, particularly in the upper part of the Munsiri Formation where the garnets are in profusion and their size is not smaller than garnets in most of the Joshimath Gneiss. The occurrence and size of metamorphic mineral grains depend to a large extent on the original composition of the host lithology (Barker, 1990). Even within the Joshimath Gneiss, there are many examples where the garnet grains are comparatively smaller in the siliceous (quartzofeldspathic) bands than in the alternating micaceous (pelitic) bands. Such a control of parent lithology is also reflected by the way kyanites occur in the Joshimath Gneiss. Strictly speaking, the occurrence of kyanite is not confined to a particular horizon/band at a specific structural level (see Plate IV). It occurs intermittently such that one might tend to think of the existence of either a number of kyanite zones or a very wide kyanite zone with a few intervening kyanite-free horizons. If the former is true, then this could suggest three possibilities--

- (a) the multiplicity of such kyanite-bearing horizons is a primary metamorphic feature controlled by the original rock composition;
- (b) the multiplicity is due to tectonic repetition caused by folding or faulting, or both, following metamorphism;
- (c) it is a combined effect of deformation and metamorphism.

If (c) is true it becomes rather difficult to explain the occurrence of the wide intermediate horizons which are devoid of kyanite. Further discussion on these aspects will be made in Chapter 5. However, it is quite obvious that the occurrence and intensity of development of kyanite are controlled, to a great extent, by the original composition of the host lithology. Although the Vaikrita/Munsiri boundary coincides with the southernmost limit of these high-grade occurrences of kyanite, kyanite has also been found in low-grade association within the Berinag-Mandhali formations well below the Munsiri Thrust (see Chapter - 5). In Nepal, the first appearance of kyanite in amphibolite facies rocks while approaching from the Lesser Himalaya has been regarded as one of the most important criteria for defining the location of the MCT (Bordet et al., 1972; Le Fort, 1975; Pecher, 1977; Hashimoto et al., 1973; Brunel, 1986). Subscribing to this view Valdiya regards his Vaikrita Thrust as the real Main Central Thrust (Roy and Valdiya, 1988, p.107). But there is an inherent difficulty in using the changing metamorphic grade as an evidence of thrusting.

Thus, if we confine our observations only to the Vaikrita Thrust and its immediate surroundings (as in Fig. 4.3), we find a rather abrupt metamorphic break across the thrust, but when the whole area is taken into consideration the change in metamorphism from Joshimath Gneiss to the Munsiri appears gradational.

### (2) Retrogression-Mylonitisation as evidence of a thrust:

It is difficult to appreciate this argument in relation to the Vaikrita Thrust. We would expect to find the greatest shearing and mylonitisation in the basal part of the Vaikrita sheet just above or within the thrust-zone itself. Instead, such features are seen within the footwall block (here the Munsiri Formation). A similar argument applies for retrogression as well. Retrogression is expected to affect the lower part of the hanging wall block more than the footwall block which may, in fact, rise in metamorphic grade as a result of being overridden by an immensely thick sheet. Contrary to this, in the present case, in fact the footwall (Munsiri) rocks are mylonitised and retrograded to a great extent.

Thus the whole of the Munsiri Formation could be regarded as belonging to a thick shear zone or it may represent the sheared and retrograded basal part of a very thick thrust sheet overlying the Gansser's MCT. The point of retrogression and mylonitisation as presented by Valdiya(1980) is not particularly pertinent as regards the existence of the Vaikrita Thrust.

### (3) Fold style and orientation:

The postulated difference between the Vaikrita Group (Joshimath Gneiss Formation) and the Munsiri Formation in terms of the style and pattern of folding is not clearly borne out by the results of the present study, as discussed in subsections 3.5 & 3.7. Both the formations show multiple generations of folding. As already mentioned in Chapter 3, the early fold sets of Joshimath Gneiss are not represented in the Munsiri. This is most likely to be due to their obliteration through transposition and metamorphism and the gradually increasing intensity of shearing strain from N to S moving from the Joshimath Gneiss into the Munsiri (see Chapters 3 & 6). One plausible explanation of whatever differences we find in fold patterns between the two formations, could be in terms of progressive southward increase in deformation. At one point, Valdiya recognises this fact and says, "... there is no structural discordance indicating the existence of a thrust plane (i.e. the Vaikrita Thrust), nor are there geomorphic expressions of the weak zone associated with the thrust." However, he

subsequently emphasises that "there is an abrupt, dramatic change in the grade of metamorphism and the constitution of lithologies." (Valdiya, 1980, p.107).

#### (4) Hot springs:

The argument based on the location of hot springs is not very convincing because across the whole of the Higher Himalayas and the inner Lesser Himalayas there is no preferential distribution of hot springs around the Vaikrita/Munsiari boundary. Hot springs are found as far up into the Higher Himalayas as Badrinath in the Alaknanda valley, Dar-village in the Darmaganga valley or as far below the Vaikrita/Munsiari boundary as to the south of Nyu village (near the Nyu Primary School) or Jhamrigaon in the Darmaganga valley or SE of Loani village in the Alaknanda valley. My impression is that the hot springs could be related to some late leucogranitic or pegmatitic intrusions in the Himalayas.

#### (5) Dip of thrusts:

In the Joshimath area there is no appreciable difference in dip between the Vaikrita/Munsiari boundary (Vaikrita Thrust) and the Munsiari/Berinag-Mandhali boundary (Munsiari Thrust) (cf. Figs. 3.6, 3.21a & 3.21b in Chapt. 3 ). In the Sobala area, in fact, the Vaikrita Thrust is gentler than the Munsiari Thrust in its northernmost exposure. So a unique linkage between the seismic plane of Kaila and Narain (1976) and the Vaikrita Thrust is questionable. Whether the high seismicity of the belt is due to temporally repetitive slip or reactivation along one of these boundaries or due to recent faulting confined within or above the weak zone (i.e. the Munsiari Formation) enveloped by these two boundaries is still not very clear.

Concerning the location of the MCT (i.e. equivalent to Valdiya's Vaikrita Thrust) south of the Annapurna Range in the Central Nepal, Pecher & Le Fort (quoted in Le Fort, 1975, p. 19) recognised a distinct change in litho-association across the movement-zone, retrogression features within the movement zone, location of kyanite isograd just above it, presence of conspicuous shear foliation and a distinct NNE-SSW trending stretching lineation in and around the movement zone.

The main observations from my study on the Vaikrita/Munsiari contact are summarised below. See also Table - 4.1 at the end of this chapter for a comparison between the features across the Vaikrita Thrust and the Munsiari Thrust.

(a) The lithological contrast across the Vaikrita/Munsiari contact is largely one of dominantly gneissic rocks coming against schistose rocks dominated by psammopelites, quartzofeldspathic protomylonite, phyllonite, augen mylonite etc. If the average grain-size of the two formations are considered side by side, then there is obvious difference between the two. The average grain-size of the Munsiari is considerably smaller than that of the Vaikrita (Joshimath Gneiss). But there are some restricted zones of fine-grained, flaggy horizons even within the Joshimath Gneiss (see Plate IV). Immediately above the contact, the rocks are of Barrovian kyanite grade, whereas those below are of garnet-biotite grade now under the influence of retrogression. Above the contact, within the Joshimath Gneiss only the fracture/fault surfaces show conspicuous signs of retrogression (mainly hydraulic). But the Munsiari rocks immediately below the Joshimath Gneiss show frequent alteration of biotite and garnet to chlorite, amphibole to biotite etc.

(b) In terms of folding and other strain features, there is no apparent difference immediately across the Vaikrita/Munsiari contact. Deformation intensifies more or less progressively southward. The rocks, in general, become more strongly foliated as the contact towards the south is crossed. The main stretching lineation passes undeflected in orientation across this contact and gradually becomes more intensely developed within the Munsiari Formation. Later stretching lineation is inconspicuous around this contact zone. The large  $F_2$  folds which are so common in the Joshimath Gneiss outcrops distant to the contact become very rare or absent in and around this contact zone. However, relatively small scale  $F_2$  folds are noticeable even within the Munsiari Formation.  $F_1$  folds are very difficult to find in the Munsiari Formation. Generally speaking, perfect isoclinal folds are not very common around this boundary. Most folds are asymmetric, inclined with the major component of their vergence towards south.

(c) Boudinage (particularly of amphibolite bands) is very much more common in the Munsiari Formation below this contact than in the Joshimath Gneiss (above the contact). Foliation boudinage is almost totally absent in the Joshimath Gneiss, whereas this feature is nearly a characteristic of the Munsiari rocks below (see section - 3.6 in Chapt. 3). However, differential thinning and thickening of white quartzofeldspathic segregation bands within Joshimath Gneiss is very frequent.

(d) Veins are more common in the Munsiari Formation than in Joshimath Gneiss. Concordant veins intrusive into the Joshimath Gneiss may be confused with

quartzofeldspathic segregation bands of metamorphic origin. There is a predominance of concordant/peneconcordant veins over discordant ones above the Vaikrita/Munsiari contact, whereas within the Munsiari Formation concordant and discordant veins are nearly in equal proportions. Veins are always more common in mica-rich (pelitic/phyllonitic) horizons than in siliceous ones.

(e) Shear sense indicators found on either side of the contact all point to a SSW-directed overthrust shear. Such indicators are more common within the Munsiari Formation than in the Joshimath Gneiss and are defined by asymmetric microfolds, S-C bands, sigmoidal pressure shadow trails around porphyroblasts/ porphyroclasts, asymmetric S/Z-shaped augens or porphyroclasts, 'mica-fish'-structures and their mesoscopic equivalents, characteristic boudin imbrications etc., all giving a consistent SSW-ward sense of overthrust shear.

#### 4.3. TRAVERSE ACROSS THE MUNSIARI THRUST

The Munsiari Thrust is primarily the boundary/contact between the Munsiari Formation above and the Berinag-Mandhali formations below. This was originally considered by Heim & Gansser (1939) and Gansser (1964) as the Main Central Thrust of the Himalayas. As mentioned earlier, locating this thrust in the field is a difficult problem. Attributing this difficulty generally to the presence of an imbricate stack of thrusts in the footwall underlying the it, Valdiya (1980, p.105) notes that there are, however, some sections where imbrication is less pronounced and in those sections this thrust is rather sharply defined. The present area of study around Joshimath is supposed to provide one such favourable section. According to Valdiya, the following features are the proofs or consequences of thrusting along the Munsiari Thrust --

(1) The selective thinning or, at places, even complete elimination of the dominantly quartzitic Berinag Formation which lies just below the Munsiari Thrust, indicates that it has been sliced off or truncated by thrusting.

(2) Where the Berinag is missing, the crystallines above the thrust come directly on top of the Mandhali rocks which show folds indicative of movement along the thrust. A high frequency of landslides is also noted in the thrust zone.

(3) The occurrence of medium grade metamorphics with granitic intrusives (i.e. the Munsiari Formation) over sediments showing low-grade metamorphism (i.e. Berinag-Mandhali formations) indicates that the former is thrust onto the latter.

(4) The tremendous shearing and mylonitisation of the rocks of the Munsiri Formation speak volumes for its thrusting. The pervasive and profound post-crystallisation cataclastic deformation of the rocks, especially augen gneiss and granitoids, testifies to the enormity of movement along the thrust plane, as do flat cleavages parallel to the bedding planes and the pronounced penetrative NNE lineation.

(5) The abrupt and precipitately steep rise of the mountain ranges with dizzy scarps facing the Lesser Himalaya is the geomorphic expression of the youthful movement along the Munsiri Thrust. The lessening of the gradient of the rivers, and the transition of the vertical-walled valleys of the Great Himalayas to the broad V-shaped ones of the Lesser Himalaya south of the thrust are further pointers to recent or sub-recent movement along it.

The abovementioned points invite some criticisms.

(1) The alternation and/or elimination of Berinag quartzite may not be solely due to truncation by a thrust (the Munsiri Thrust). The possibility of having some stratigraphic pinch-outs cannot be totally ruled out, especially in view of the occurrence of these rocks near the northern border of the so called 'Lesser Himalayan sedimentary basin' (see Ahmad & Alam, 1978; Kumar, 1980; Saxena, 1980).

(2) I have noticed that even where the intervening Berinag rocks are present, the underlying Mandhali rocks show folds that indicate the influence of an overriding thrust, but not necessarily the Munsiri Thrust. Also correlation of recent landslides with thrust is difficult to establish, especially when the thrusts are not brittle thrusts. Landslides may occur due to so many different reasons. Neither in Joshimath area nor in Sobla area have I observed any preferred occurrence of landslides along the line of thrusts. In fact, it has been observed that landslides in and around the area of my study are largely controlled either by lithology or by late fractures; no evidence are found of more landslides taking place along the boundaries of the MCT-Zone. Within the Munsiri Formation, landslides are developed mainly in the fractured quartzite horizons. Some distance south of the present study area around Joshimath, there are many massive landslides around Ganai village which are mainly due to the existence of a particular kind of lithology, whitish in colour, which is either Deoban-Mandhali dolomites (containing ?magnesite) and/or fine-grained tuffaceous rocks (Ganai volcanics). The observation that on the whole landslides are more frequent in Garhwal region than in the Kumaun region would point to the range of factors that may be involved in the formation of landslides.

(3) The third point is a highly valid one, but the problem is that in most of the sections the metamorphic discordance is not well marked. Mostly, the situation appears to be one of classic metamorphic convergence (Read, 1934). Fig. 3.12 in Valdiya (1980, p.106) shows a good example of this. Retrogression must have played a major role in such convergence. Valdiya says, 'The low-grade metamorphism at the base of the otherwise medium-grade metamorphics is but a consequence of retrogressive changes attending mylonitisation.'

(4) The extensive shearing and mylonitisation of the rocks belonging to the Munsiri Formation are features more of the entire formation rather than of its contact with the Berinag-Mandhali. Also the 'cataclastic' deformation mentioned by Valdiya should probably refer to the crystal-plastic deformation suffered by the original intrusive granitoid bodies in order to give rise to the augen gneiss, mylonites etc. This again is not a clear and direct testimony of extensive movement along the Munsiri Thrust which is described in passing as a discrete planar thrust. Unfortunately, therefore, still there remains a distinct blank about the real nature of the Munsiri/Berinag-Mandhali contact, i.e. the Munsiri Thrust.

(5) Physiographic expressions could be equally due to reactivation of the thrusts. However, it is indeed very difficult to assess the real magnitude or scale of movement along thrusts from such physiographic features. The approach of Seeber & Gornitz (1983) is useful for analysing recent uplifts. Seeber & Gornitz (ibid.) found that the long profiles of most of the transverse Himalayan rivers are characterised by a zone of high gradient which correlates with (a) the topographic front between Lesser and Higher Himalayas, (b) a narrow belt of intermediate magnitude thrust earthquakes, and (c) the occurrence of the Main Central Thrust Zone. This correlation led them to suggest that differential erosion is cause for this break in gradient but tectonic uplift of the Higher Himalaya relative to the Lesser Himalaya.

In a regional perspective, the Munsiri Thrust (Heim & Gansser's MCT) is generally considered as the root of the Almora Thrust and its equivalent thrust sheets represented by various klippen exposed in the Lesser Himalaya, such as the Amri, Almora, Baijnath-Dharamghar-Askot and Chhiplakot klippen (Valdiya, 1980, p.104; and also his geological map; Johnson, 1986). It is postulated that these klippen represent the remnants of a once continuous thrust sheet with its root in the northern outcrop of the

Munsiari Formation. The klippen occur in asymmetric synforms developed on the thrust, such folds generally having their northern limbs steeper than the southern limbs. The asymmetry in these folds, increases progressively northwards towards the root, so much so that the northernmost klippe (Chhiplakot) has an overturned northern flank dipping NE-wards. A comparatively higher structural level of the Almora Nappe which lies in the inner Lesser Himalaya has led Valdiya (1980, p.104) to think that the emplacement of the Almora Nappe represents the latest thrust movement, whereas the Amri Thrust in the outer Lesser Himalaya which has brought the crystallines over the Lower Eocene Subathu rocks, is considered to have been emplaced earlier (post-Lower Eocene). This point is important because it implies that in the Lesser Himalaya thrusting propagated in a 'foreland to hinterland' sequence which is just opposite to what Johnson (1986) has proposed. Assuming a piggy-back sequence of thrusting (Dahlstrom, 1970; Elliot and Johnson, 1980; Boyer and Elliot, 1982), Johnson (1986) demonstrated that in the Lesser Himalaya the thrusts have developed in a 'hinterland to the foreland' sequence-- the Vaikrita Thrust developed first, followed in time by the Krol, Berinag and the Main Boundary Thrusts. Definite proof is lacking but this order of development of the thrusts is now widely accepted (Valdiya, 1988, p.166; Barnicoat & Treloar, 1989, p.6).

The main observations from the present study on the Munsiari/Berinag-Mandhali boundary are enumerated below. See also Table - 4.1 at the end of this chapter for a comparison among the features across the Munsiari Thrust and the Vaikrita Thrust.

(a) As we go from the Munsiari Formation to the Berinag-Mandhali formations crossing the Munsiari Thrust lithology changes from one with quartzofeldspathic rylonite, phyllonite, amphibolite etc. to another consisting dominantly of low-grade quartzite, phyllite, calcareous semipelite (see Fig. 4.4). Overall grain-size variation in the rocks across the thrust is not very appreciable. Generally speaking, however, the Munsiari shows smaller grain-size (except for micas in the mica-rich phyllonites) than do the Berinag-Mandhali. But the effects of grain-size reduction i.e. mylonitisation are much more pronounced in the Munsiari Formation than in the Berinag-Mandhali formations. Metamorphism in the rocks across the boundary shows an interesting feature. There is distinctly a case of metamorphic convergence across the Munsiari Thrust. Considered separately, now the grades of metamorphism on either side of the thrust show little difference: biotite-chlorite grade above the thrust and chlorite-sericite grade below it.

(b) In terms of the degree of deformation, there is a wide difference between two sides of the Munsiri Thrust. The Munsiri rocks are much more intensely deformed and tectonised e.g. augen mylonite, phyllonite etc., whereas the Berinag-Mandhali rocks are comparatively much less deformed and tectonised. The main stretching lineation is rare below the Munsiri Thrust; wherever it is present near the Munsiri Thrust or below, it is found to be rotated slightly towards north from its usual NNE trend above (the shift in trend is from  $N35^{\circ}E$  to  $N23^{\circ}E$ ). It is yet to be established what this shift means. One possibility is that it reflects the simultaneous anticlockwise rotation of the Indian plate at one stage during its continued northward drift after collision with the Eurasian plate. The later stretching lineation (i.e. the N-trending one) is well developed particularly in the mica-rich phyllonite horizons in the middle to lower part of the Munsiri Formation. This later lineation appears prominently from about the middle of the Munsiri Formation and continues down-section into the Berinag-Mandhali formations below the Munsiri Thrust. Dominant fold-sets on either side of the thrust are not readily correlatable. Outcrop-scale folds are not found. Mesoscale isoclinal, often intrafolial, folds are common, particularly some distance above the thrust. In a section in Pithoragarh district in Eastern Kumaun, Bhattacharya & Siawal (1985) noticed progressive increase in flattening of mesoscale folds when they approached the MCT (Munsiri Thrust) from either side. The total width of the flattened zone was noticed to be 5 to 6 km. However, they did not make it clear to which generation those folds belong. Fold-vergence, in most cases, in the present area of study is broadly towards south.

(c) Boudinage is found in the formations above and below the Munsiri Thrust, but it becomes less common from about 1.5 km. farther down. Within the Munsiri Formation, foliation boudinage is found to be less pronounced in the intensely mylonitic and mica-rich phyllonitic rocks.

(d) The Berinag-Mandhali rocks show fewer veins than the overlying Munsiri Formation. In and around the Munsiri Thrust, discordant veins are slightly more common than the concordant/peneconcordant ones. Here also veins have a preference for the mica rich (pelitic) horizons.

(e) Shear sense indicators above and below the Munsiri Thrust all point to a broadly SSW-directed overthrust shear. They are more common within the Munsiri Formation, but some calcareous (semipelitic) horizons in Berinag-Mandhali formations

also show well developed shear bands (?C/S-bands) all with consistent SSW-ward shear sense.

#### 4.4 SHEAR INDICATORS IN THE MCT-ZONE

There are many characteristic kinematic indicators in the zone bounded by the two major thrusts and also in its surroundings.

Deformational features preserved in rocks that indicate the kinematics of the associated deformation are called kinematic indicators. In shear zones such indicators are used for recognising the sense of shear.

Since late seventies there has been great upsurge among structural geologists to establish the accurate sense of shear in shear zones using various such criteria (Berthe, et al., 1979; Bouchez et al., 1983; Paschier & Simpson, 1983; Platt, 1984; Etchecopar & Malavielle, 1987; Bell & Johnson, 1992).

There could be two aspects of shear sense determination (Ramsay & Huber, 1987, p. 632): one where the associated ductile shear zones are comparatively of smaller scale and have well defined boundaries; and the other where the shear zones are very wide and/or their boundaries are obscure. In the first case, one or more of the following criteria are commonly used:

- a) change in finite strain state,
- b) sigmoidal form of schistosity induced during the deformation,
- c) the orientation of folded and boudinaged competent layers, dykes etc.,
- d) the geometry of en-echelon vein arrays,
- e) fracture openings, together with fibrous shear veins.

In the second case, while the above criteria may be used in suitable situations, normally other small scale structures are used to define the shear sense. These structures can be classified into the following groups:

- a) sense of fold overturning and vein asymmetry;
- b) asymmetrical pressure shadows;
- c) spiralled inclusion fabrics and rolling structures;
- d) S-C fabrics, shear bands and mica fish;

e) grain-shape fabrics;

f) crystallographic preferred orientation pattern.

For details on these features see Barker (1990, pp.89-103), Ramsay & Huber (1987, pp. 632-634).

In the present study emphasis has been given on recognising unambiguously the regional sense of shear related with the MCT-emplacement; therefore, the second of the above aspects is more relevant here.

Some of the unambiguous mesoscale shear indicators recognised in Joshimath area are shown in Fig. 4.5. For approximate location from where these shear indicators have been noted in the field, see Plate IV. In the Sobala area the shear indicators are commonly found in quartzofeldspathic augen gneiss and protomylonite-mylonite horizons. The above indicators are associated with the main stage of MCT emplacement and give SSW-ward sense of overthrust shear. Some microscopic evidence of fractured and partially rotated fragments of garnet porphyroblasts give an unequivocal sense of shear particularly at the final stage of main MCT emplacement. (see Figs. 4.6 & 4.7).

The Fig. 4.7 is important because it shows a good example of pulled apart garnet indicating the effect of stretching. The two garnet fragments are tilted in such a way that the disjointing fracture surface on either of them slope towards each other, taking the main foliation horizontal. This coupled with the converging arrangement of micas pointing southward in the separation zone is a very good shear indicator-- 'top-to-south'. The most likely course of evolution of this feature is also depicted in fig. 4.7(a & b).

Shear indicators associated with the later stretching event are relatively scarce. The 'falling fire-ball' garnet, associated with later stretching lineation indicate presence of strain shadow zones on the up-plunge side.

The process involved in the creation of such features is quite similar to glacial 'crag and tail'. The shear was 'top-up-the-plunge' giving an overthrust sense.

Thus, the kinematic indicators confirm the compressional deformation associated with the MCT zone bounded by the Vaikrita and Munsiri Thrusts.

#### 4.5 CONCLUSIONS

As shown by Ramsay & Graham (1970), 'Ductile shear zones'(DSZ) show<sup>a</sup> steep strain gradient across narrow well defined boundaries. DSZ normally results from noncoaxial simple shear deformation. The deformation may be continuous without any loss of cohesion. However, Berthe et al. (1979) showed that a combination of continuous and discontinuous deformation leads to the development of mylonitic shear zones. The Vaikrita/Munsiari boundary and the Munsiari/Berinag-Mandhali boundary do not represent straightforward ductile shear zones of classic type. Evidently these are not discrete brittle thrust surfaces either. The main foliations in the rocks within and outside these boundaries are all mutually parallel and lithological boundaries also follow this foliation. Though locally there is severe and repeated transposition, its effect on the large scale in the form of lithological redistribution is not apparent at all. So we find a marble-band or a quartzite band side by side with semipelite bands or siliceous protomylonite band lying beside augen gneiss. Could the two boundaries represent tectonic slides?

Originally used by Bailey (1922) and later formalised and popularised by Fleuty (1964) and Hutton (1981), the term 'tectonic slide' is defined as a fault that forms during metamorphism in orogenic belts. According to Hutton 'A tectonic slide is a (ductile) fault which forms in metamorphic rocks prior to, or during, a metamorphic event. It occurs within a zone of coeval penetrative deformation that represents an intensification of a more widespread, often regionally developed deformation phase. Within this zone of high strain, slides may lie along and be subparallel to the boundaries of lithological, tectonic and tectonometamorphic units'. A good example of a tectonic slide is the Sgurr Beag slide in Scottish Highlands (Kelley & Powell, 1985). Slides normally form in metamorphic rocks exhibiting lower greenschist to upper amphibolite facies metamorphism. These definitions fit well with what we observe in and around the two boundaries in question.

Thus it is concluded that these boundaries in themselves can be considered neither as brittle thrusts nor as narrow discrete shear zones. They appear more like tectonic slides. The whole Munsiari Formation can be regarded as a complex and thick shear zone with lithology-controlled spatial strain partitioning within it. The two 'thrusts' represent the two boundaries of this zone. The upper boundary represented by the Vaikrita Thrust

is, tectonically speaking, more diffuse than the lower boundary i.e. the Munsiri Thrust. The relative age of these two boundaries has not yet been unequivocally established. Generally it is believed that probably the Vaikrita Thrust developed earlier than the Munsiri Thrust (Johnson, 1986; Searle, 1986). Finally, Table - 4.1 summarises how the different features vary across the two thrusts/boundaries.

**Table - 4.1**

(A comparison of different features across the two thrusts)

Features	Across the VAIKRITA THRUST	Across the MUNSIARI THRUST
1. Lithological contrast	Gneissic rock against schistose rocks (dominated by protomylonite, phyllonite, augen mylonite etc.)	Mylonitic and phyllonitic rocks against low-grade quartzites and phyllites
2. Grain-size variation	Overall variation appreciable	Not appreciable
3. Barrovian grade of metamorphism	Kyanite grade above, garnet-biotite grade below	Biotite-grade above, chlorite-sericite below
4. Retrogressive metamorphism	Fairly abrupt increase in retrogression downward	Retrograded Munsiri juxtaposed with low-grade Berinag-Mandhali (apparently a case of metamorphic convergence)
5. Grade of tectonism	Apparently less difference	More tectonised above (augen mylonite-phyllonite lithology dominant), less tectonised below
6. Main stretching lineation	Passes undeflected; more intensely developed within the Munsiri below	Less conspicuous below; it takes a slight N-ward swing from the usual NE/NNE-trend
7. Later stretching lineation	Inconspicuous or absent	Almost pervasively present, particularly in the mica-rich horizons
8. Fold features	Large $F_2$ -folds less common near and below the thrust. Isoclinal folds not very common. Most folds asymmetric, broadly S-verging.	Outcrop-scale folds not found. Some distance above, mesoscale, isoclinal, often intrafolial folds common. Fold-vergence broadly S-ward. Dominant fold-sets not properly correlatable across the thrust.
9. Boudinage	More common in the Munsiri	Common on either side of Munsiri

(below the thrust). Foliation  
boudinage not seen above thrust.

Thrust. Foliation boudinage common in  
Munsiari, not so in Berinag-Mandhali

10. Veins

More common in Munsiari Formation  
than in Joshimath Gneiss.

Conspicuously less common  
in the Berinag-Mandhali  
than in the Munsiari.

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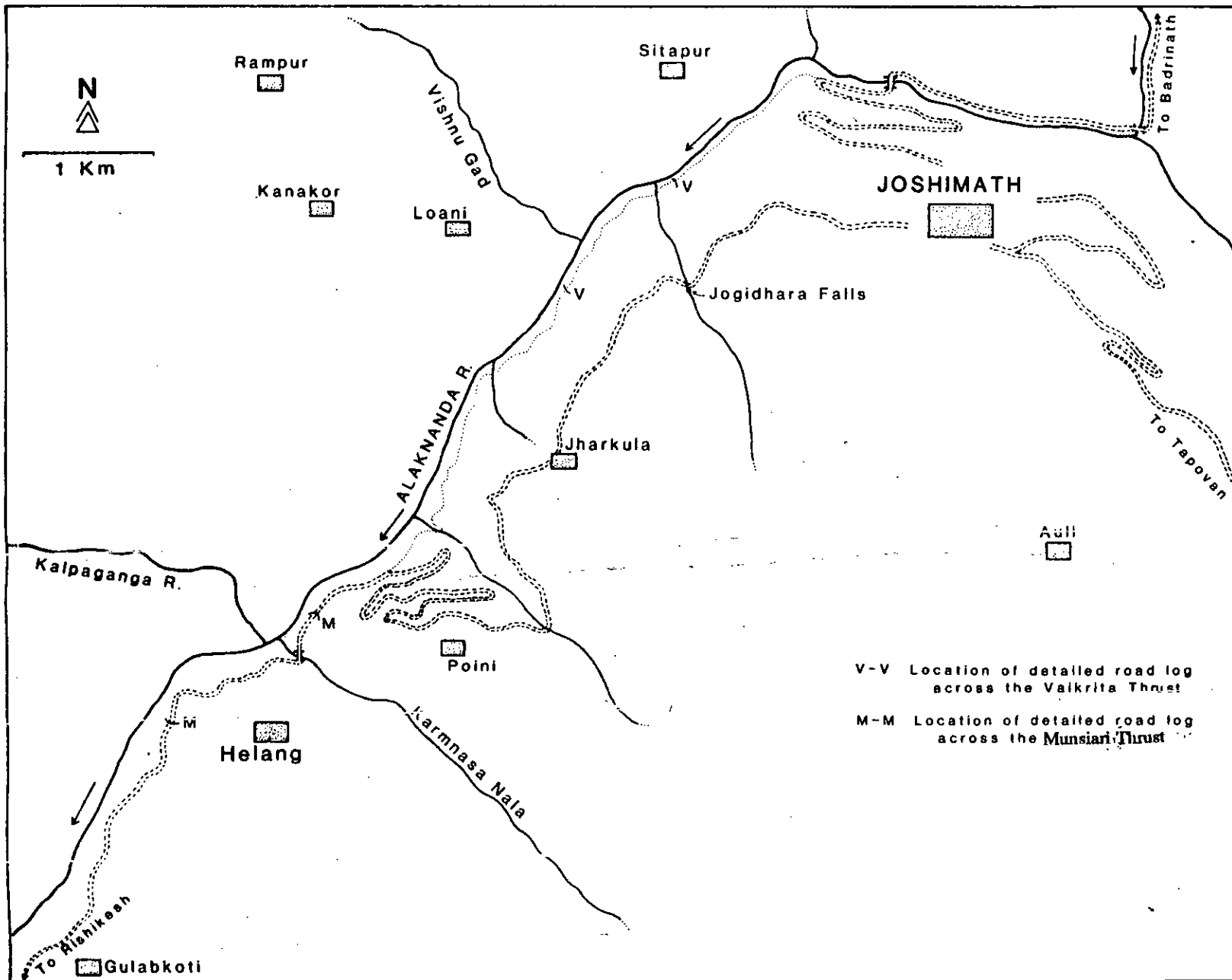


Fig. 4.1 Location of the road logs across the two major thrusts.

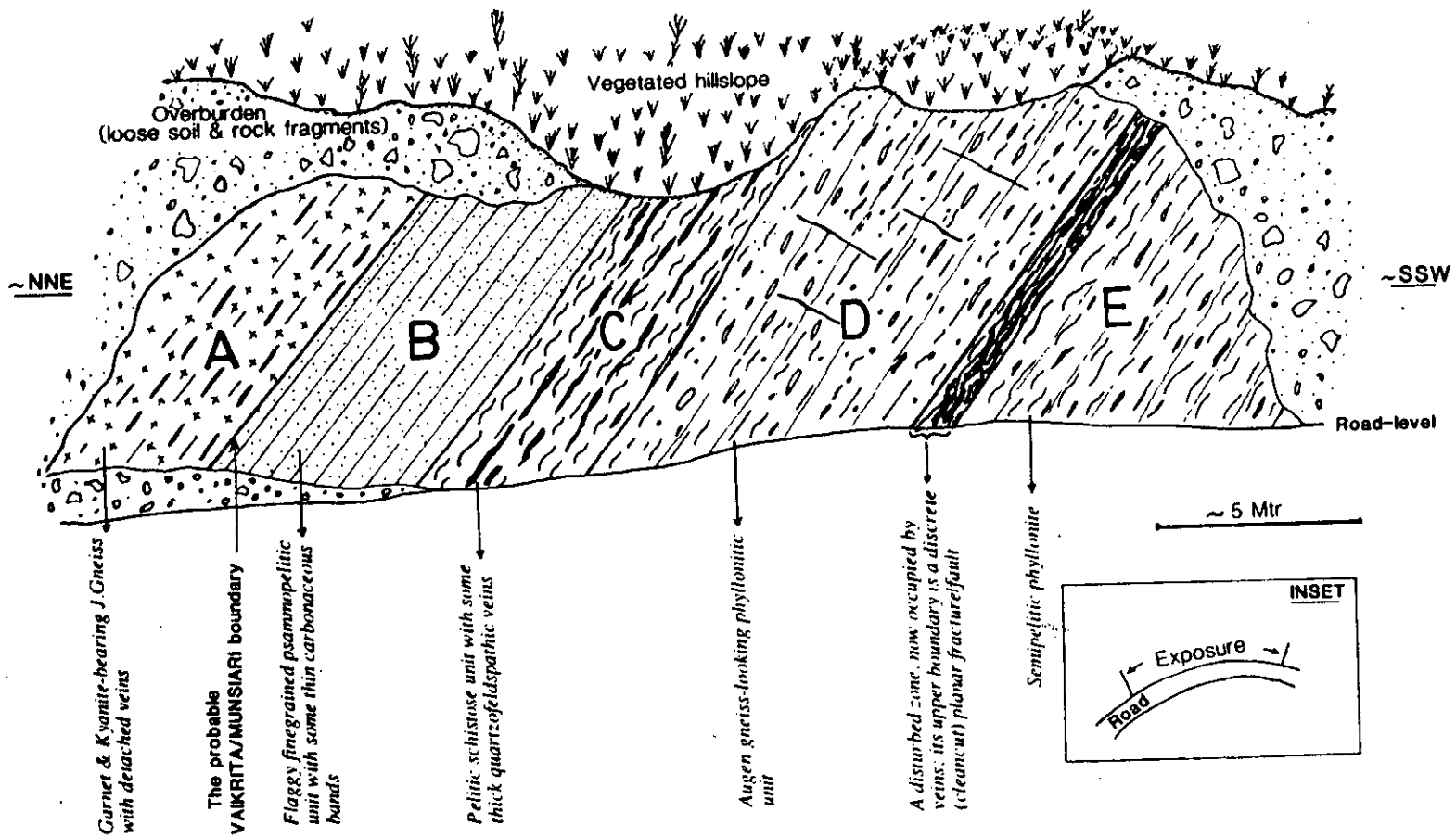
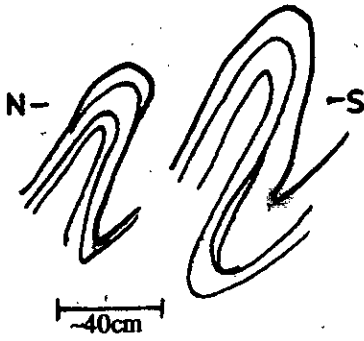
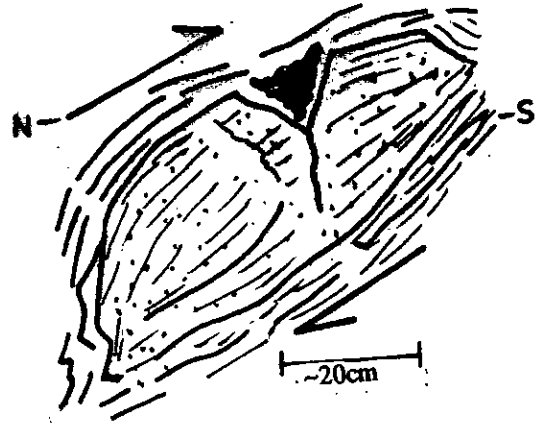


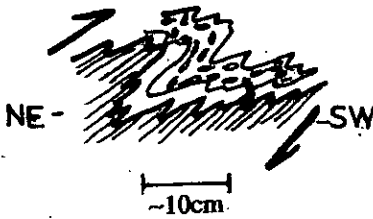
Fig. 4.3 Details of the road-section exposure showing the VAIKRITA / MUNSIARI boundary  
See text for details.



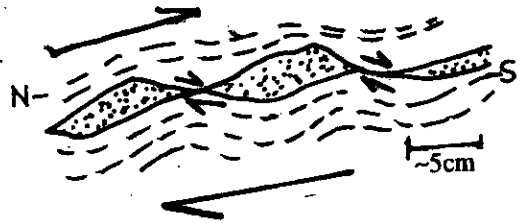
Asymmetric overturned folding in Joshimath gneiss.



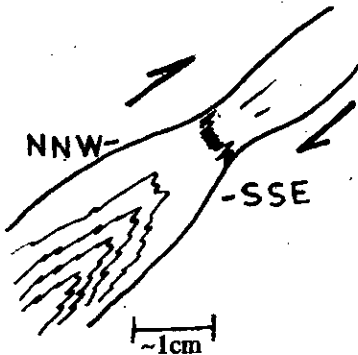
Gaping fracture in amphibolite boudin— an excellent shear sense indicator.



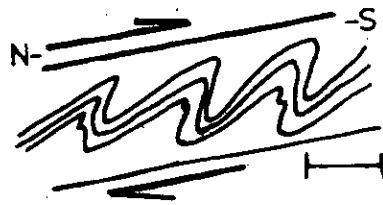
Asymmetric folding in augen phylonite horizon.



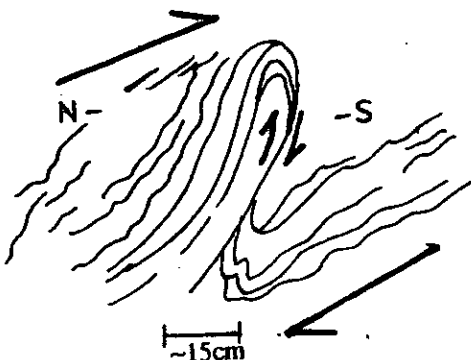
Quartzofeldspathic band suffered complementary extension as a result of shearing. Enclosing medium is of psammopelite.



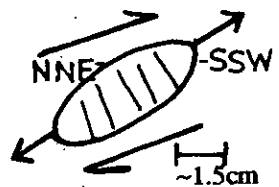
Boudinage of calc-silicate band with intrafolial folds in it.



Asymmetric overturned folds in Joshimath Gneiss.



Asymmetric south vergent folding with attenuated and thrust overturned limb.

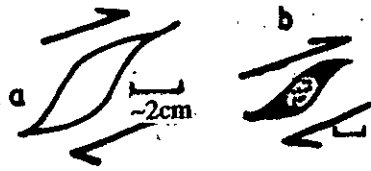


Elliptical feldspar porphyroblast in augen gneiss developed tensile fractures as a result of complementary tension during shearing.

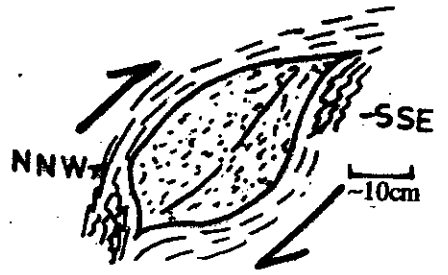
Fig. 4.5 Some mesoscale shear indicators from Joshimath area.

(For approximate level of occurrence in the section see Plate IV.)

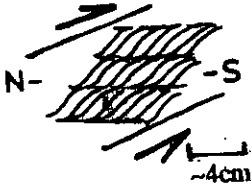
Cont'd.



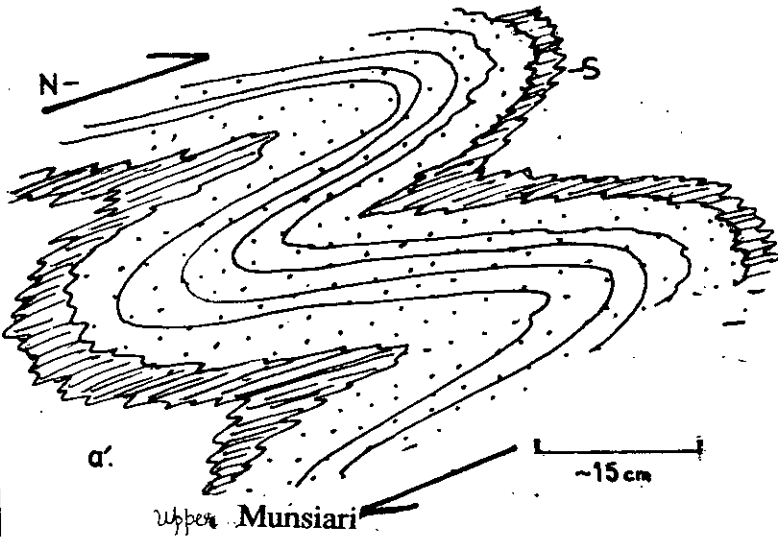
a Sigmoidal augen-shape of feldspar porphyroblast  
 b Sigmoidal pressure shadow trails around garnet porphyroblasts



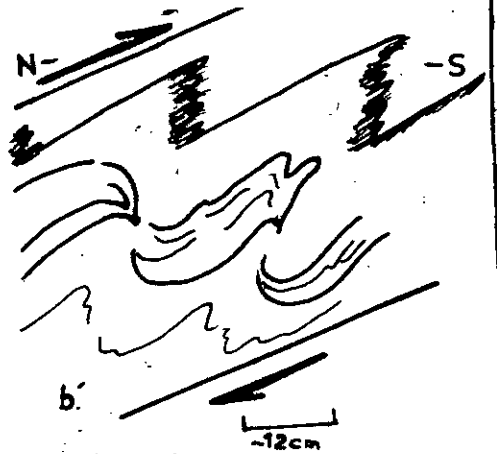
Asymmetric amphibolite boudin with characteristic matrix microstructure.



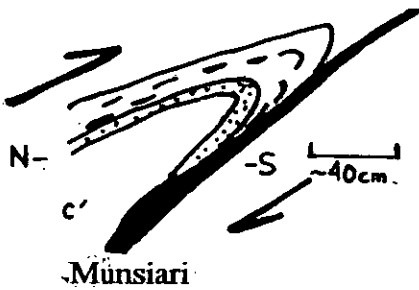
C-S bands in quartzofeldspathic protomylonite.



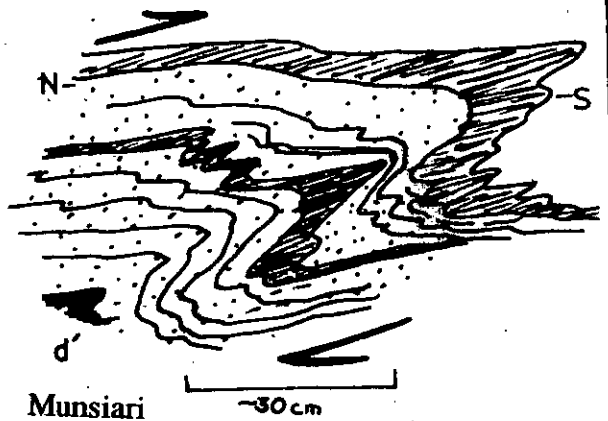
Upper Munsiri



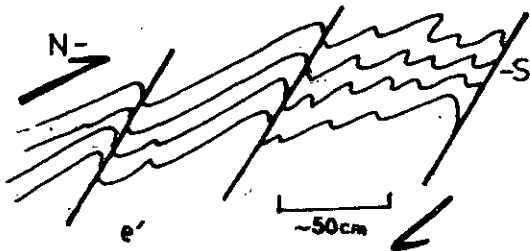
Joshimath Gneiss



Munsiri



Munsiri

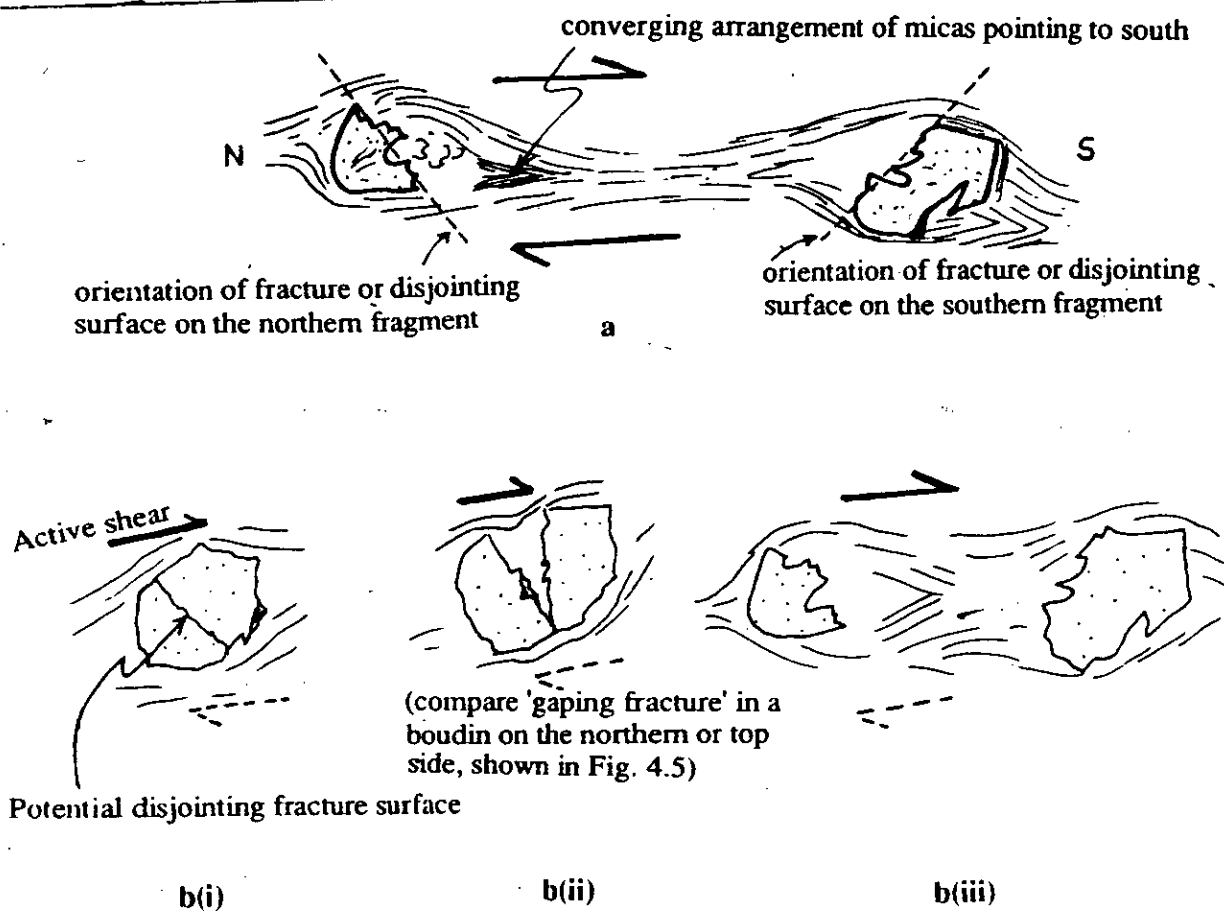


Berinag-Mandhali

a'-e': Asymmetric folds.



Fig. 4.6 A tilted triangular feldspar porphyroclast in an augen phyllonite. Its tilted position, the pattern of its wrapping by matrix foliation and the difference in the length of pressure shadow on either side give unequivocal shear sense as indicated.



4.7 (a) An excellent example of stretched and pulled apart garnet. See (b) for possible course of pulling apart.

(b) The successive stages of stretching/shearing affecting the garnet porphyroblast.

## Chapter-5

# TEXTURES AND METAMORPHIC HISTORY

### 5.1 INTRODUCTION

This chapter deals with the establishment of the metamorphic history from the petrographic study of the textural relations among the different metamorphic minerals in the rocks. The objective is to work out the deformation-metamorphism relationships. The chronology for metamorphic evolution is relative to the time of the deformation episodes as discussed in Chapter 3. For identification of minerals and for the study of textural relations, petrographical microscopes were mainly used. In some instances, use of X-ray diffractometer, electron microprobe or of a U-stage, fitted to a petrographic microscope, helped in the confirmation of mineral identity. Following a brief review of some established textural criteria bearing on deformation-metamorphism relationships the chapter gives sequentially discussions on the distribution and textures of the high-grade minerals viz. sillimanite, kyanite & staurolite, then garnet followed by amphiboles and other minerals before giving a brief summary of retrograde metamorphism. Finally, on composite fish diagrams, the paragenetic relations of the minerals are depicted as a pre-requisite to a discussion of the metamorphic history and deformation-metamorphism relationships and on aspects of inverted metamorphic sequence developed in the area/s.

### 5.2 RECOGNITION OF DEFORMATION - METAMORPHISM RELATIONS: A HISTORICAL PERSPECTIVE

H.J. Zwart's pioneering works (1960a, b; 1962, 1963) in the Central Pyrenees initiated the modern approach to the study of textural relations in metamorphic rocks with a view to unravelling the deformation-metamorphism relationships. Zwart demonstrated clearly that following detailed field analysis, careful microscopic studies of textural relations among different minerals in metamorphic rocks can resolve their time of formation in relation to deformation episodes. Around the same time Johnson (1963) carried out studies on establishing the time relations of movement and metamorphism in parts of the Scottish Highlands. Later studies by Naha (1965), Harte & Johnson (1969), Misch (1971), Ferguson & Harte (1975), Platt & coworkers (Platt, 1982; Platt et al., 1983, 1984; Platt & Behrmann, 1986), Bakker

et al. (1989) and Bell & coworkers (see Bell et al., 1992 and references therein) are also relevant.

Traditionally the deformation-metamorphism relationships centred around the use of porphyroblast-matrix microstructural relations. Zwart (1962, see his Fig.1) distinguished diagnostic forms of pre-, syn- or post-kinematic porphyroblasts and suggested the possible pattern of Si-Se (Schistosity internal - Schistosity external) relations corresponding to dominantly three basic types of deformation: one, when the schistosity is a plane of slip; the second, where the schistosity is a plane of flattening; and the third, when the schistosity is micro-folded as a result of foliation-parallel compression. Thus, in total, nine diagnostic forms have been given. Although in nature, the relations between Se and Si may show a large variation, they can all be deduced from these basic types.

The relations between the fabric of the rocks and the internal structure of the porphyroblasts (minerals) help us in distinguishing whether a particular mineral (porphyroblast) is pre-, syn- or post-kinematic with respect to the fabric-forming deformation. There has been extensive search for criteria distinguishing pre-tectonic crystals, syntectonic crystals and post-tectonic crystals (for an appreciation see reviews in Spry, 1963; Vernon, 1968; Barker, 1990 etc). However, of these three broad types the establishment of the syntectonic porphyroblasts drew maximum attention (further details in Section 5.4 on garnet textures).

Vernon (1968) emphasized the use of common sense for identifying 'which grew when' among the metamorphic minerals using textural criteria. Apart from the scale I find a crude analogy between the determination of relative age of minerals from petrographic (textural) study of thin sections and the establishment of the relative chronostratigraphy (which is older, which is younger) among different rock horizons using field relations or map relations.

### **5.3 DISTRIBUTION AND TEXTURES OF HIGH-GRADE MINERALS (SILLIMANITE, KYANITE & STAUROLITE)**

As pointed out in Chapter-4 (The Major Thrusts), even though locally the metamorphic break between the Joshimath Gneiss and Munsiri in the MCT hanging wall block may appear sharp, broadly it is gradational. Probably it is just a coincidence that the first appearance of kyanite is correlatable with the location of the

Vaikrita Thrust. It is noted that to some extent there is alternation between kyanite and staurolite zones and, therefore, a repetitive occurrence of kyanite-bearing horizons in the Joshimath Gneiss (as studied along the Alaknanda section). This repetition is presumably due to folding. Also noted is a thick kyanite-bearing horizon below Sitapur village on the western slope of the Alaknanda valley, whereas at the corresponding level on the eastern slope, the kyanite-bearing horizon is much thinner. Whether this is due to the inherent metamorphic pattern or due to the presence of a fault locally following the course of the Alaknanda river could not be settled. Broadly speaking, however, in the MCT-hanging wall block consisting of the Munsiri Formation and the Joshimath Gneiss Formation, the highgrade minerals viz. staurolite, kyanite and sillimanite occur in successive zones starting from the upper Munsiri northward into the Joshimath Gneiss. Garnet occurs all through the Joshimath Gneiss and southward well beyond the staurolite zone down to the lower-middle part of the Munsiri, though sporadic occurrences of garnet are found even farther south down-section in the very top part of the Berinag-Mandhali formations. In addition kyanite is found to occur in Berinag quartzites in one exposure about 1.25 km below the Munsiri Thrust. Thus garnet and kyanite are the only highgrade minerals recorded from the Berinag-Mandhali footwall block and they occur at two isolated exposures --- garnet just below the MCT, kyanite further 1.25 km below it. In the hanging wall block, sillimanite and kyanite occur only in the Joshimath Gneiss, whereas staurolite occurs in both the Munsiri Formation and Joshimath Gneiss. Garnet is present in both the units alongside these high grade minerals. Garnet textures will be dealt with later in section - 5.4; here sillimanite, kyanite and staurolite textures are described. The discussion on kyanite and staurolite will get prominence, because sillimanite is not as prevalent and its occurrence not as clear as the kyanites and staurolites.

It is stressed that the high grade minerals do not occur uniformly in all the three units viz. Joshimath Gneiss, Munsiri and Berinag-Mandhali. There is a gradation in their distribution so that the highest grade minerals occur to the north in the Joshimath Gneiss and as we go southwards we come across the occurrences of relatively lower grade minerals. In any attempt to establish the effects of successive deformational events on these highgrade minerals as reflected in their textures, or more precisely, while attempting an interrelation of different deformational (folding) events and the formation of the high grade minerals, it is to be remembered that the peak development of the successive fold events is noticed successively towards south. The effects of later fold events are not clearly recognisable towards the north.

For a brief recapitulation of the deformation (fold) episodes the reader is referred to Subsection 3.1.2.2 (Fold episodes etc.). Essentially, of the three main fold episodes (viz.  $F_1$ ,  $F_2$  and  $F_3$ ) recognised in the Joshimath Gneiss,  $F_2$  is the most prevalent and most well developed one and thus provides a very important time marker. Stretching/shearing was a prolonged phenomenon whose peak intensity came about following the  $F_2$ -folding and broadly spanning around the  $F_3$ -folding. The  $F_2$ ,  $F_3$  and  $F_4$  episodes could be considered as taking place successively under the broad influence of a progressive simple shear regime. The Munsiaris show clear evidence of  $F_2$ ,  $F_3$  and  $F_4$  folding, while the  $F_1$  folds are presumed to have been obliterated due to high strain. The folds in the footwall block could not be properly correlated with the hanging wall folds. Mainly three episodes ( $F_I$ ,  $F_{II}$  &  $F_{III}$ ) of folding have been recognised in the Berinag-Mandhali of which the third episode was evidently a postmetamorphic one.

The essential aim in this section is to identify if there are more than one generation of the high grade minerals in the rocks and their time relation in terms of the deformation episodes. The discussion is given in the following order - (1) Sillimanite, (2) Kyanite & (3) Staurolite. Here the arguments are based mainly on collective textural information. Detailed petrographic description of relevant thin sections are given in Appendix-II. A representative list of minerals occurring in the rocks that contain the high-grade Barrovian index minerals is given below (Table - 5.1).

**Table - 5.1**  
List of mineral assemblage in rocks bearing the high-grade Barrovian minerals

Unit	Name of the high-grade mineral	A thin section reference	Mineral assemblage
Joshimath Gneiss	Sillimanite(?)	1F87	Quartz + plagioclase + orthoclase (much less than plag.) + biotite + muscovite + garnet + sillimanite(?) + epidote + apatite + opaques
-do-	Kyanite	22/4/88B	Quartz + plagioclase + orthoclase (much less than plag.) + biotite + muscovite + kyanite (large & small) + garnet + apatite
-do-	Staurolite	5B <sub>a</sub> '87	Quartz + plagioclase + orthoclase (in very small amount) + biotite + muscovite + garnet + staurolite + clinozoisite + epidote + apatite + zircon + sphene + black & red opaques

Munsiari Formation	Staurolite	7H'87	Quartz (predominant over feldspars) + plagioclase + orthoclase + muscovite + biotite + chlorite (secondary) + garnet + staurolite + black opaques + tourmaline + apatite + clinozoisite
Munsiari Formation	-do-	NR58/59	Quartz + plagioclase + orthoclase (in very small amount) + biotite + chlorite (secondary) + muscovite + garnet + staurolite (included within garnet) + black opaques + tourmaline + apatite + clinozoisite

### 5.3.1 Sillimanite in the Joshimath Gneiss

Very distinct examples of sillimanite have not been found in the thin sections studied so far. However, in a few thin sections, extremely small quantities of fibrous masses are recognised (see Fig. 5.1) which are most probably aluminosilicates, and possibly sillimanite (fibrolite).

It is rather surprising that even though sillimanite is so often referred to in the literature (see, for example, Valdiya 1980, 1988; Roy & Valdiya, 1988; Gupta, 1978; Viridi, 1986; Gairola & Ackermann, 1988), no clear example of sillimanite has been found in the rocks that I have studied? Two possible reasons for this could be-

- (i) Incomplete sampling or inspection in the present study;
- (ii) Reports in the literature may largely represent misidentification of sillimanite or an over-exaggeration about the occurrence of sillimanite.

Given that good care has been taken to make thin sections of rocks from different levels in Joshimath Gneiss starting from Vishnuprayag southward and that potential sillimanite-bearing thin sections were carefully examined without much success, it seems that the second possibility is the more likely one. On two occasions in course of the present study, a different mineral was misidentified as sillimanite (cf. Figs. 5.2 & 5.3).

As far as the time of formation of the possible sillimanites is concerned, fool-proof textural evidence is lacking in terms of the relations of sillimanite growth and fold hinges. However, the preferred alignment of the sillimanite fibres following the foliation that is folded by  $F_2$  is taken to indicate that sillimanite grew before the  $F_2$  folding. Though by itself this evidence is not an unambiguous proof of the pre- $F_2$  origin of sillimanite, other points favour such an origin e.g. the absence of any

immediate heat source, sillimanite as a part of the progressive regional Barrovian sequence to which kyanite, staurolite etc also belong, and kyanite being certainly established as a major pre-F<sub>2</sub> phase (see the next subsection for the time of major kyanite formation). It is not unreasonable, particularly in view of the compositional similarities (see Table-5.1) and spatial closeness of the index mineral-bearing rock horizons, to consider that sillimanite forms an integral part of a major episode of progressive regional Barrovian metamorphism of which kyanite, staurolite etc are also a part, and that sillimanite grew broadly at the same time as the kyanite, staurolite etc. As discussed later, major kyanite formation in the Joshimath Gneiss took place before the F<sub>2</sub> folding, but during or, more possibly, after the F<sub>1</sub> folding; so most probably sillimanite also grew at the same time as a part of the same progressive Barrovian sequence.

However, this does not mean that there was no later episode of sillimanite development, particularly in the still higher parts of the Higher Himalayas. In fact, farther to the N, near Badrinath, development of sillimanite as a contact metamorphic effect of late leucogranitic intrusions has been reported (e.g. Sinha, 1977c, 1987 Gupta, 1978 b & c; Viridi, 1986). Reference to two generations of sillimanite can be found also from a comparison of the works of Hodges et al. (1988) and England et al. (1992). Hodges et al. (1988) refer to an early generation of sillimanite and their sillimanite-bearing assemblage (i.e. qtz + biot + musc + plag + gar ± k-felds ± sillim) matches closely with the corresponding assemblage recognised in the present study (cf. Table-5.1). England et al. (1992) mention a later generation of sillimanite in <sup>the</sup> Annapurna-Manaslu region of Central Nepal corresponding to anatectic granitoid intrusions. As will be discussed later (Chapter 6), such anatectic granitoid intrusions took place postdating the high-grade regional Barrovian metamorphic event in the Higher Himalayas and broadly synchronously with the MCT-shearing (although emplacement of some of the major granitoid bodies could have started prior to the MCT-thrusting). So, temporally speaking, the later generation of sillimanite in the Higher Himalayas developed synchronously or late synchronously with the MCT emplacement.

For a diagrammatic representation of the time of sillimanite growth with respect to fold episodes in <sup>the</sup> Joshimath area, see Fig. 5.4 (note that the approximate position of the second generation of sillimanite reported by other workers from the higher parts of the Higher Himalayas is also indicated).

### 5.3.2 Kyanite in the Joshimath Gneiss

In Joshimath Gneiss, two distinct size groups of kyanite have been found: the most conspicuous group of kyanite is the larger ones; the other group consists of tiny prismatic grains in which cleavage cannot be recognised. This second group of kyanites occurs mostly as trails. Given below are the criteria and arguments concerning the relative age of the two groups of kyanites.

#### A. The larger kyanites

Among all the high-grade minerals studied petrographically, kyanite shows the clearest evidence about its time of formation. As stated earlier, the  $F_2$  folding is the most prominent time-marker, especially in the Joshimath Gneiss. The major phase of kyanite formation (i.e. corresponding to the larger kyanite grains) in the Joshimath Gneiss predated the  $F_2$  episode of folding. As indicated in Fig. 5.5 which is a profile-view of an  $F_2$  hinge, a kyanite grain has been distinctly folded by the  $F_2$  fold. So, kyanite must have grown before the  $F_2$  fold episode. As we already know, the  $F_2$  folding was the strongest fold episode and developed in a wide range of scales especially in the Joshimath Gneiss. Thus if kyanite (the larger ones) has grown before  $F_2$  folding, we would now expect that the kyanite grains, in general, would appear strained. This proved to be so, because the kyanite grains, no matter whether or not they are near or at a conspicuous fold-hinge, generally have a very strained appearance showing undulose extinction, many with broadly wavy, bent or irregular outlines (see, for example, Figs 5.6 & 5.7 and cf. Fig. 5.14). Now having been definitely established as pre- $F_2$ , when did kyanite grow in terms of the  $F_1$  episode of folding? Unfortunately, for microstructural study I couldn't find any suitable example of an  $F_1$  fold-hinge which contains kyanite as well. So, direct evidence is lacking as to the kyanite vs.  $F_1$  time relation. However, none of the folded kyanite grains studied show any refolding pattern; they are deformed by only one episode of folding, that is  $F_2$ . The later episodes, such as  $F_3$  etc, have not left any recognisable imprints either on these kyanite grains. So, it is almost certain that kyanite formation was not pre- $F_1$ . Now three possibilities are left open:

- (1) Kyanite was syn- $F_1$
- (2) Kyanite was syn- to post- $F_1$
- (3) Kyanite was post- $F_1$  and developed at the interkinematic stage between  $F_1$  and  $F_2$ .

Nowhere is kyanite aligned exactly parallel to the axial planes of the  $F_1$  folds. Had it been strictly syn- $F_1$  we would expect a strong orderliness in the present distribution of the kyanite grains, but this is lacking. So, most probably, kyanite did not grow strictly at the syn- $F_1$  time. In Chapter 3 reference was made to the possibility of a

stage of fabric enhancement or fabric coarsening during the interval between  $F_1$  and  $F_2$ . Thus the third of the above possibilities regarding kyanite formation demands special consideration. But the possibility of kyanite development starting from a late-syn- $F_1$  time cannot be ruled out, even though no foolproof evidence for this could be found. It is concluded that the major episode of kyanite development in the Joshimath Gneiss took place at a pre- $F_2$ , but late-syn- $F_1$  or, more possibly, post- $F_1$  time.

This timing of major kyanite growth assumes great importance in view of the fact that in many cases for various reasons it is difficult to find suitable 'fold hinge - mineral growth' relations. Such a problem becomes very apparent when we try to establish the time relations of other high grade minerals (as already mentioned for sillimanites; see later for staurolites). That the different Barrovian zones of high grade minerals in Joshimath Gneiss (Vaikrita Group) are essentially members of the same progressive metamorphic sequence is a well established fact of literature [see, for example, Valdiya & Goel (1983) for Garhwal-Kumaun Himalaya, Brunel & Kienast (1986) for Nepal Himalaya, Ray (1947) and Mohan et al. (1989) for Darjeeling Himalaya, Searle et al. (1988) for NW or Zaskar Himalaya]. I did not find any evidence contrary to this. Because the above-noted kyanites (bigger grains) are an essential or integral member of the prograde Barrovian metamorphic sequence which is largely preserved in the area, it is likely that other minerals of the sequence would also be broadly synchronous with them. This is to say that probably one would not expect any significant time-gap between the formation of the different high-grade minerals as members of a single and the same progressive regional metamorphic sequence.

Therefore, if the members other than kyanite of the Barrovian sequence failed to reveal distinct evidence of 'fold hinge--mineral growth' relations, then the approach has been to look for indirect evidences that could tell us something about their time of formation with respect to folding. Basically it was attempted to explore if the textures of such minerals could be compared with those of the kyanite grains situated away from any conspicuous fold hinge. If the features are comparable, we could then possibly safely extrapolate the timing of kyanite-growth to the other high-grade minerals concerned, remembering that both of them belong to the same progressive metamorphic sequence.

### B. The smaller kyanites

Rock horizons at different structural levels within the Joshimath Gneiss have been found to contain tiny prisms of kyanite that often lie in trails partly bordering large grains of quartz and feldspar (Fig. 5.8.). In some horizons the tiny kyanites occur in addition to the larger kyanite grains which are normally platy or bladed in habit; these latter grains do not have any specific regular orientational relation with the adjacent quartz or feldspar grains (see Fig. 5.9). Where both types of grains occur in the same rock, there are instances where the tiny kyanite prisms are adjacent to large kyanite grains (Fig. 5.10). In several thin sections, no representative of the bigger kyanite grains is found, but only the bundles or trails of tiny kyanite prisms are present. A slight increase is noticed in the population of the smaller kyanite grains from north to south within the Joshimath Gneiss.

The possibility that the tiny prisms are broken up fragments of the large kyanite grains is not totally ruled out for at least a small proportion of the tiny grains which are intimately associated with the bigger, partly broken or fractured grains of kyanite. However, the majority of the smaller grains are well outside the vicinity of any larger grain of kyanite and have grown round quartz and feldspar grains that are often augen-shaped (Fig. 5.11). In several cases even where the smaller and the larger grains are mutually adjacent, the smaller grains are oriented with apparent disregard to the orientation of the larger grains (Fig. 5.12). The occurrence of tiny grains is noted from different structural levels sometimes even from lithological horizons where no larger grains of kyanite have been found. This evidence indicates that most of the tiny grains belong to a generation different from that of the larger kyanites. Their distinctly smaller grain size and fairly unstrained character are in contrast to the strained and, in many cases, bent larger kyanite grains. This suggests that the smaller grains grew later than the larger grains. Had it not been so, the smaller grains would possibly have acted as nuclei upon which the bigger grains were to grow. Evidently this has not happened. So, why and specifically when did the smaller kyanites grow?

The small kyanite grains are found in several instances to be aligned in the main foliation which, in turn, is folded by the  $F_2$  folding (cf. Fig. 5.5). By itself, this cannot tell whether the small kyanites are pre-, syn- or post- $F_2$ . However, on closer observation, these small kyanite grains are found not to be strictly aligned along the foliation, instead they are rather haphazard in orientation, broadly confined in linear zones that define the trace of the main foliation. Also, <sup>the</sup> majority of them do not show any undulose extinctions. This probably means that these kyanites were not pre- $F_2$ ; they were either syn- or post-  $F_2$ . Had they been pre- $F_2$  and pre-gneissosity

formation, then we would expect them to show considerably stronger preferred orientation along the main foliation. The grain reorientating effect due to the formation of gneissosity followed by a strong episode of folding would have been quite significant. Moreover, where these small kyanite grains are found around some augen-shaped grains of feldspar or quartz, they are oriented at high angle or sub-perpendicular to the augen boundary; many project well into the augens (Figs. 5.13a & b). Such a feature implies that these kyanite grains developed at a stage either synchronous with or postdating the development of the augen shapes of the feldspar and quartz grains. Now, if we assign this augen development to the peak of the stretching or shearing event, which is not an unreasonable proposition, the smaller kyanites assume a post- $F_2$ , syn-stretching age. In any case, therefore, these small kyanite grains are not pre- $F_2$ . This finding has an obvious importance insofar as the determination of physical conditions during the main phase of stretching/shearing is concerned. It indicates that MCT thrusting did not take place in cold, retrograde conditions.

That (i) the occurrence of the tiny kyanite crystals is concentrated mainly along zones following the main planar fabric, (ii) many of the crystals lie surrounding the shear-modified porphyroclastic grains of quartz or feldspar, and (iii) there is a slight general increase in the population of these tiny grains towards the MCT-Zone, makes a case for shear heating due to MCT emplacement to be an important contributor for the development of the smaller kyanites in Joshimath Gneiss. The heat generated could easily affect the bottom part of the Joshimath Gneiss lying immediately above the MCT-Zone. That the tiny kyanites occur only in the Joshimath Gneiss in selected rock horizons and not in the Munsiri Formation itself (i.e. the MCT-Zone) is probably due to a primary compositional control.

In addition, mainly towards the southern part of Joshimath Gneiss near the Vaikrita Thrust, some gently bent larger kyanite grains show minor recrystallisation along their transverse boundaries (Fig. 5.14). The bending or folding is probably related to  $F_2$  folding and the minor recrystallisation on the boundaries of these high energy grains is accordingly late-syn- $F_2$ . The reason why these kyanite grains are considered to have been affected by  $F_2$  folding is a broad similarity in the hinge-curvature as defined by the cleavage in the grains and strain-shadow pattern with those in the kyanite grains lying at the hinge of well established  $F_2$  folds found in other horizons. And because the recrystallisation is in geometric conformity with the general folded shape of the grain, a late-syn- $F_2$  time is assigned for the recrystallisation. Now the obvious question is, why is such recrystallisation found

more commonly towards the south than to the north? Probably this is due to the nearness of the MCT-Zone, reflecting the intense shearing or stretching characteristic of that zone. Additional straining imposed on grains that already had surplus strain energy probably triggered the recrystallisation.

The times of kyanite growth in the Joshimath Gneiss in terms of the fold episodes are shown diagrammatically in Fig. 5.15 (sillimanite is also shown for comparison). Clearly the two groups of kyanite have grown at two different times -- the larger and major ones earlier, the tiny ones later.

### 5.3.3 Kyanite in the Berinag-Mandhali (footwall block)

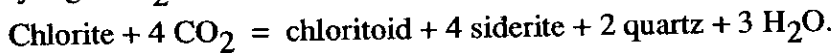
About 1.25 km (stratigraphically or, more precisely, structurally) below the Munsiri Thrust, kyanite occurs in a low-grade setting in a quartzite horizon in the Berinag-Mandhali footwall block, but no sillimanite or staurolite has been found in the specimens studied from this unit (see Plate-IV for precise location of the kyanite quartzite exposure on the Helang-Gulabkoti road section). Occurrence of kyanite in a low-grade setting is not very rare, though quite unusual. This occurrence of kyanite-quartzite south of Helang was noticed earlier by Brunel (1983) and Brunel & Kienast (1986), but they did not give any detailed study of the textural relations or phase relations in the rock.

Rawat (1982) reported occurrences of kyanite from the Lesser Himalayas from farther south near Ghimtoli and gave an explanation of their origin in terms of their metamorphic phase relations. The stable assemblage he recognised in the kyanite-bearing Ghimtoli Quartzite is: quartz + muscovite + kyanite + albite + tourmaline. The rock is rich in muscovite and tourmaline and contains traces of chlorite as well as carbonaceous material. Rawat (ibid.) found that the tourmaline content of the rock increases in the mylonitised parts in the vicinity of faults and shear zones and, therefore, suggested that the tourmaline was introduced later into the rock due to flow of boron-rich fluids along the tectonically disturbed zones. Referring to the coexistence of stable minerals like albite and muscovite with kyanite, he argued that the kyanite-bearing quartzite formed due to greenschist facies regional metamorphism of aluminous sediments. Presence of carbonaceous material in the sediments might have played a helpful role during the metamorphism by providing CO<sub>2</sub>-rich fluids.

In the SW Scottish Highlands, kyanite has been reported from low grade associations (Burgess et al., 1981). Here regional metamorphic kyanite has been found near Port Ellen (Islay) in highly aluminous phyllitic metasediments together with chloritoid-bearing assemblages within the *greenschist facies*, although chlorite itself is absent in the mineral assemblages at these localities. The mineral assemblages are:

- (i) Quartz + muscovite + kyanite + siderite + rutile + hematite  $\pm$  paragonite  
 $\pm$  ankerite;
- (ii) Quartz + muscovite + chloritoid + rutile  $\pm$  hematite  $\pm$  paragonite;
- (iii) Quartz + muscovite + siderite and/or ankerite + rutile + hematite  
 $\pm$  paragonite  $\pm$  albite.

Burgess et al. (ibid.) emphasized the role of fluids in the metamorphism with a relatively broad range of CO<sub>2</sub>-H<sub>2</sub>O compositions and argued that the occurrence of chloritoid-phyllite instead of normal chlorite-phyllite by itself would imply a relatively high CO<sub>2</sub> content in the fluid phase, as indicated by the following reaction:



It has been suggested on the basis of textural and carbon isotope evidence that probably the oxidation of organic carbon (e.g. graphite or hydrocarbons) has produced the CO<sub>2</sub>.

The kyanite-bearing quartzite horizon of the Berinag-Mandhali formations has the following mineral assemblage-

Quartz + kyanite + muscovite + tourmaline + black opaques + zircon + ?rutile  
+ traces of chlorite (probably secondary).

Interestingly, there are important similarities between this kyanite-bearing Berinag Quartzite and the Ghimtoli Quartzite reported by Rawat (1982). Except the albite, the stable assemblage looks almost the same and tourmaline is indeed a significant component of the rock. Recrystallised quartz veins are also present (see Appendix-II). So, obviously there was access to fluids during metamorphism. Collectively, these features suggest that the kyanite-bearing Berinag Quartzite did not undergo any metamorphism beyond the greenschist facies.

Emphasis will now be given on recognising the time of formation of the kyanites in the Berinag-Mandhali formations with respect to the footwall deformation episodes. Relative age between the kyanite-forming event in the footwall and the major kyanite-forming event in the hanging wall will also be established.

There is a strong linear fabric in the rock which is defined by fine trails of elongate tourmaline needles or threads, alignment of fine micas and a fairly strong

preferred orientation of kyanite grains with their lengths parallel to the lineation. When observed in sections sub-parallel to the foliation, quartz grains also show a crude elongation parallel to the lineation. In such sections, however, a number of kyanite grains are noticed which are aligned oblique or at a high angle to the lineation; many of these kyanite grains are folded (see Figs. 5.16a & b). The fact that some of the linearly oriented kyanite grains are pulled apart following the direction of lineation suggests that it is a stretching lineation (see Fig. 5.17).

Kyanite is evidently porphyroblastic in origin and forms most of the largest grains in the rock. Their dimensions are generally >2mm in length and over 1mm across. The kyanite grains are not homogeneously distributed throughout the rock, but have preferential concentration along micaceous foliation surfaces. In contrast to quartz the kyanite grains appear much strained; almost all of them show strong undulose extinction and strain shadow patterns. Some of them are conspicuously folded or kinked (Fig. 5.18; see also Figs. 5.16a & b). That the stretching has affected kyanites, as reflected by tensile fracturing (pulling apart) in the direction of stretching lineation of several kyanite grains, clearly indicates that kyanite developed prior to this stretching event. The conspicuous folding of a number of kyanite grains, on the other hand, implies that kyanite grew before a folding event as well. So, an obvious question is, what is the time-relation between the stretching event and the kyanite-folding?

Both the stretching event and the folding event postdated the formation of kyanite. The fold episode here refers to that responsible for the folding or kinking of some kyanite porphyroblasts. There could be three possible hypotheses as to the time relation between the folding and the stretching events:

- (i) Kyanite-folding was later than the stretching lineation
- (ii) Kyanite-folding was synchronous with the stretching lineation development
- (iii) Kyanite-folding was earlier than the stretching lineation and, therefore, the lineation development may have played a modifying role upon the orientation of the kyanite grains after folding.

We need to test which of these hypotheses is correct.

Thorough study of the rock (petrographic, hand specimen and in field) reveals different features from which different inferences can be drawn as to the time relation between kyanite-folding and the lineation. These observations and the corresponding inferences are given below:

**Observation I** - The kyanite grains with their elongation parallel to lineation never show folding (see Fig. 5.16a)

**Inferences:** (a) Folding could be either synchronous to or later than lineation development and is due to a shortening that acted perpendicular to the lineation.

(b) Folding of kyanites was earlier than and hence modified by the lineation. Lineation-parallel kyanite grains presumably belonged to the limb areas of folds whereas the folded grains at high angles to the lineation could represent remnant parts of the hinge areas:

**Observation II** - The sizes and broad shapes of kyanite are not widely varying (see Fig. 5.16a)

**Inference:** All the kyanites belong to the same generation.

**Observation III** - Fine streaks or trails with development of small-sized tourmaline needles/threads and white micas along them (especially within the kyanite grains) mainly define the lineation. Such streaks are present even within the folded kyanite grains and are, more or less, following the fold axial traces (Fig. 5.19).

**Inference:** Folding could be synchronous or earlier than lineation.

**Observation IV** - Kyanite grains at high angle to lineation do not always show folding (Fig. 5.20).

**Inference:** Folding is less likely to be later than or synchronous with the lineation; it is probably earlier than lineation.

**Observation V** - Preferred alignment of many kyanite grains parallel to the lineation is quite distinct. Some of the kyanite grains (relatively larger ones) appear to have been fractured and torn apart (pulled apart) following the lineation direction (see Fig. 5.17)

**Inference:** Kyanite formed earlier than the lineation and the lineation is to be interpreted as a stretching lineation. Possibly, folding was earlier than stretching.

**Observation VI** - There is one folded kyanite grain a part of which is pulled apart parallel to the stretching lineation (Fig. 5.21).

**Inference:** It could not be properly understood whether stretching took place after folding or folding took place after stretching. This recalls the general problem of deciding whether a particular structure is a 'folded boudinage', or a 'boudinaged fold'.

Most probably folding did not take place after stretching, because in the groundmass surrounding the kyanite grain, there is no evidence of folding.

Thus, though some of the observations lead to different inferences, most favour the third hypothesis that kyanite-folding was earlier than and modified by stretching. Now let us see if this can be confirmed by using a special geometric test and following inductive logic.

Let us assume that stretching lineation and folds were mutually related i.e. either they developed simultaneously, or folding took place later due to a compression that acted preferentially in a direction perpendicular to the lineation. In that case we would expect that depending on their present orientation relative to the lineation, the folded grains of kyanite would show systematic variation in the amount of their shortening due to folding. Data measured from selected grains, most of them showing conspicuous folding (see Figs. 5.16a & b), are presented in Figs. 5.22a & b (the graph and circle plots). It is clear from these figures that there is no systematic variation in the amount of shortening due to folding of the kyanite grains relative to their orientation with respect to lineation. If our assumptions were correct, the data in Fig. 5.22a would have defined a U-shaped curve implying that there were more shortening in the grains aligned at high angle to lineation and less shortening in those that are subparallel to lineation; in other words, the data would have defined a strain ellipse in two dimensions in the simple case when an initial uniformity in kyanite grainsize and homogeneous diversity in their orientation are assumed. Fig. 5.22a clearly shows that the distribution of data is unsystematic and irregular; Fig. 5.22b indicates that the distribution of shortening data is far from defining a strain ellipse.

*[Fig. 5.22b relates to the kyanite grains depicted in Fig. 5.16b and has been constructed such that corresponding to the orientation of each folded kyanite grain (precisely speaking, that of its enveloping surface) with respect to the mean direction of lineation (m.d.l.) there is a line following a diameter of the original circle, and the length of this diametrical line is proportional to the shortening value of the said kyanite grain. Thus, the more shortened a grain, the shorter is the length of the diametrical line corresponding to it. The tips of the adjacent diametrical lines are finally joined free-hand in order to find out whether or not the resulting outline defines an ellipse (strain ellipse). Clearly the outline is irregular (amoeboid) and does not define an ellipse.]*

Therefore, neither of the assumptions that folding and stretching took place simultaneously or folding was later than stretching was correct; folding was unrelated to lineation i.e. folding and lineation development were not parts of the same

deformation episode. Our hypothesis testing, therefore, confirms that the third hypothesis is correct. Folding of kyanite was earlier than the development of the stretching lineation.

As mentioned in Chapter-3, three episodes of mesoscale folding ( $F_I$ ,  $F_{II}$ ,  $F_{III}$ ) have been recognised in the Berinag-Mandhali rocks. The folding of the kyanite grains is probably correlatable with the  $F_{II}$  episode of the footwall fold-sequence. This correlation is on the basis that -


- (a) the  $F_{II}$  folds are comparatively more common than the other two episodes (viz.  $F_I$  and  $F_{III}$ ) in the vicinity of the kyanite-quartzite exposure.
- (b) the  $F_I$  folds are conspicuously isoclinal, while almost none of the kyanite folds have strictly isoclinal geometry. The kyanite fold geometry matches better with the  $F_{II}$  fold geometry.
- (c) the scale of  $F_{III}$  folds is considerably bigger than the size of individual kyanite grains, and these folds are not so pervasively and penetratively developed as to affect the kyanite grains on such a small scale. Also, as mentioned above, the distinct  $F_{III}$  folds are not common in and around the kyanite quartzite exposure.

I presume kyanite developed at a time intermediate between  $F_I$  and  $F_{II}$ . Concentration of shear stress in a particular horizon in the footwall block consequent upon the emplacement of the MCT sheet above might be called for as a major reason for the development of these kyanites. The  $F_I$  folds in the footwall block were most likely to be pre- to early-syn-MCT. The direction of plunge of these folds is perpendicular to the direction of emplacement of the MCT-sheet and the fold axial planes parallel the main planar fabric in the MCT-Zone; these facts probably suggest a causal connection between the development of these folds and the MCT-emplacement. The  $F_{III}$  folding is a considerably later, post-metamorphic event. For further comments on correlation and details of observation see Section-5.8 and Appendix-II respectively.

The position of the formation of foot-wall kyanites in a time-sequence is shown in Table-5.2 (Details of the tectonometamorphic sequence in the foot-wall block are discussed in Section-5.8).

**Table - 5.2**  
**Time of kyanite development in the Berinag-Mandhali formations**

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I. First metamorphic imprint [Initial low-grade (?burial) metamorphism]	
II. Second metamorphism -accompanying- <i>F<sub>I</sub> fold episode</i>	 Main MCT-emplacment time
III. Third metamorphism --- "KYANITE PORPHYROBLASTESIS"	
IV. - - - - - <i>F<sub>II</sub> fold episode</i>	
V. Fourth metamorphic imprint accompanying later stretching/shearing	
VI. - - - - - <i>F<sub>III</sub> fold episode</i>	

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It must be stressed that correlation could not be established between the fold episodes of the footwall block (Berinag-Mandhali) and the hanging wall block (Munsiari and Joshimath Gneiss together). However, as mentioned above, it is highly probable that the episode-I folding in the Berinag-Mandhali formations is pre- to early-syn-MCT emplacement. Therefore, the main kyanites in Joshimath Gneiss cannot be time-equated with the Berinag-Mandhali kyanites. Because, if the footwall kyanites have developed postdating the  $F_I$  episode of folding and thus broadly synchronously or late synchronously with the main stretching event in the MCT-Zone, then there was evidently a time-gap between their formation and the formation of the Joshimath Gneiss kyanites (bigger or dominant ones). It has already been established that the main kyanite-forming event in Joshimath Gneiss took place before  $F_2$  folding which, in turn, is earlier than the main stretching event. So, the footwall kyanites are relatively younger than the Joshimath Gneiss kyanites.

#### **5.3.4 Staurolite in the Joshimath Gneiss**

Staurolite in Joshimath Gneiss is found to occur in discrete, prismatic grains, generally 1mm or less in length and 0.2mm or less in width. Their occurrence was inferred in the field and confirmed through subsequent microscopic studies. Although no clear example has been found where staurolite is directly involved in conspicuous fold hinges, sufficient circumstantial evidence suggests their broad time of formation

with respect to the fold episodes. Staurolite prisms are oriented parallel to the main foliation (gneissosity) which defines F<sub>2</sub> hinges, in a few instances, only a short distance away from the staurolite crystals. But unfortunately nowhere have I found staurolite prism/s exactly at a fold-hinge. Most staurolite prisms show undulose extinction, but the undulose character is not as prominent as in kyanites in similar settings in the kyanite-bearing rocks. Differences in physical and rheological properties between the two minerals may be the reason for this. Some staurolite prisms show transverse or oblique cracking, and even partial breaking (Fig. 5.23); this is in agreement with the features of some bigger kyanite grains (cf. Fig. 5.5). Mineral grains (mostly of quartz and micas) surrounding the staurolite prisms show strain (as evidenced by their undulose extinction). These features collectively indicate that staurolite did not grow as a post-tectonic phase after the major fold episodes. That staurolite represents one of the early high grade minerals is strongly supported by the

ensuing discussion on Munsiri staurolites which are included within garnet (see Fig. 5.25). The fact that (i) staurolite grains follow the main foliation in nearby F<sub>2</sub> hinges, (ii) there is a broad similarity in the strain features of the staurolites and confirmed pre-F<sub>2</sub> kyanites, and (iii) staurolite occurrences in the field are located well within the domain under the strong influence of F<sub>2</sub> folding suggests, on one hand, a pre-F<sub>2</sub> time of development of staurolite and, on the other, its existence as an integral part of the same Barrovian metamorphic sequence to which kyanite and sillimanite also belong. By inference, a pre-F<sub>2</sub>, but syn- to post-F<sub>1</sub> time is obtained for the staurolites in the Joshimath Gneiss.

### 5.3.5 Staurolite in the Munsiri Formation (MCT-Zone)

No difference in the attributes of the staurolite crystals is found between the Munsiri and the Joshimath Gneiss formations, except that the Munsiri staurolites are more stretched and fragmented, particularly when they are not included within another phase such as garnet. Within the Munsiri Formation staurolite occurs in two modes: one, in the matrix (Fig. 5.24; thin section reference - 7H'87), and the other, as inclusions within garnet (Fig. 5.25; thin section reference - NR58/59). However, they are not present together in the same lithological horizon.

The matrix staurolite crystals which are stretched and fragmented (pulled apart or torn apart) follow the trace of the main foliation in the rock (Fig. 5.26). The staurolites included within garnet appear intact, in distinct prismatic or thick needle-like crystals and quite diversely oriented (precisely speaking, however, there are two mean directions of their orientation, mutually at about 85°-90° to each other and at

about 45° to the present external foliation; see Fig. 5.27). Broadly speaking, there is no major difference in the sizes of the included and the matrix staurolite crystals. Probably, the reason why the included staurolites escaped stretching is that before the stretching took place garnet grew enclosing the staurolites and thereby protecting them. The garnet shows elliptical outline implying that it suffered stretching. Now, texturally speaking, can it be possible to find out whether or not the matrix (stretched) and the included (unstretched) staurolites were of the same generation, and also their time/s of formation?

The following points are noteworthy in this context:

(a) If these are two different generations of staurolite, then it wouldn't be unreasonable to expect that one of them is equivalent to the later generation of kyanite in Joshimath Gneiss. On the other hand it could be argued that both types of staurolite represent the same generation and that neither is synchronous with the later kyanites, but could be contemporaneous with the early kyanites (larger ones).

(b) Of the two types of staurolites in the Munsiri, it is quite obvious that the included (unstretched) ones cannot be later than the matrix (stretched) ones. They could either be earlier than the matrix ones or are synchronous with them. Thus for the purpose of comparison of the time of formation between the later kyanites and staurolite, we should take up only the matrix staurolites.

(c) In the Joshimath Gneiss, the second generation kyanites are in the form of tiny grains and they are rarer than the first generation kyanites (larger ones). There is no recognisable effect of stretching on the later kyanites and these grains are mostly strain-free. Even though collectively the later kyanites appear to be confined within linear zones following the main foliation (gneissosity), on a finer scale the grains individually do not show any strong linear preferred orientation. As mentioned before, their formation is presumed to be broadly synchronous with the main stretching event, postdating the F<sub>2</sub>-folding. But clearly the matrix staurolite grains of the Munsiri Formation are much larger than the later kyanite grains of the Joshimath Gneiss; the former are evidently stretched and fragmented (pulled apart), show internal strain features (exemplified by undulose extinction), and are aligned along the present main foliation (schistosity). In several instances, they are found to maintain a conformable relation with some smallscale folds (microfolds) of the main foliation. Collectively all these features indicate that the Munsiri matrix staurolites grew before the main stretching event, and possibly before an important fold event (F<sub>2</sub>). In terms of body-strain these matrix staurolites closely resemble some members

of the earlier group of larger kyanites of the Joshimath Gneiss. In all likelihood the matrix staurolites grew well before the later kyanites of Joshimath Gneiss (cf. point (b) above) and most probably at the same time as the staurolite, kyanite and sillimanite of the Joshimath Gneiss, thereby defining a single progressive Barrovian sequence.

(d) Having established that the matrix staurolites grew well before the later kyanites, it is now necessary to find out when the included staurolites grew. As mentioned before, the included staurolites must have grown either earlier than or synchronously with the matrix staurolites. The obvious difficulty posed by the first possibility is how to explain the localised development of staurolite before even the main Barrovian metamorphism took place. That the matrix staurolites formed as a part of the main Barrovian sequence has been indicated above. No other related evidence of attainment of P-T conditions high enough for the formation of staurolites before the main Barrovian metamorphic event have been found in the area particularly on a local scale. As we have already seen (Sub-section 5.3.4), the higher grade Joshimath Gneiss occurring above the Munsiri contains only one generation of staurolite. Thus it is unlikely that the included staurolites were earlier than the matrix staurolites.

(e) Garnet occurs in all the rocks containing staurolite and it developed in different generations in the area (see discussion below and also Section 5.4). Generally speaking, from the wide range of its occurrence garnet can be considered as one of the most readily crystallisable metamorphic minerals. Multiple growth stages and site-preferred growth depending on the local availability of the necessary reactants are common phenomena for garnets in comparison to other members of the index mineral series. Thus if there were multiple generation of garnets in the staurolite-bearing rocks, it is possible that at one stage garnet grew selectively in one horizon enclosing the staurolite crystals, while in the other horizon staurolites remained unenclosed. Having multiple generation of garnets could, therefore, be an additional indirect support in favour of a single episode of staurolite formation.

(f) Foliation transposition occurs very noticeably in the Munsiri Formation. If garnet-growth was followed by foliation transposition, then the possibility of having two different generations of staurolite would be less likely.

(g) As demonstrated below, the existence of multiple generation of garnets in the same rock as the included staurolites and also the distinct evidence of foliation transposition tend to support the view that the included staurolites and the matrix

staurolites belong to the same generation. Note that no compositional difference between the two staurolites could be recognised through optical studies (electron microprobe work on the staurolites is yet to be undertaken).

A detailed petrographic study of the thin section - NR58/59 which contains the example of staurolites included within garnet, furnishes some crucial evidence of the existence of more than one datable generation of garnet, foliation transposition etc. from which to glean the essential information for establishing the time of development of the included staurolites. The observations and their implications are discussed below:

- (i) There are spongy garnet porphyroblasts as well as homogeneous garnet porphyroblasts in the rock. The inclusions of staurolite crystals occur in one of the spongy garnet porphyroblasts (see Fig. 5.25). Whilst it is difficult to recognise any opaque inclusion trails in the spongy garnets, such trails are well defined within the homogeneous garnets.
- (ii) Opaque minerals in the matrix of the rock occur in definite trails (wavy in pattern) which define the local orientation of the main foliation. These trails pass undeflected through the homogeneous garnet porphyroblasts (Fig. 5.28) which implies that the homogeneous garnets overgrew matrix foliation.
- (iii) The opaque trails defining the matrix foliation swerve around the projected edges of spongy garnet porphyroblasts (Fig. 5.29). This implies that the spongy garnet was already in the rock when the present external foliation developed through transposition. Thus, clearly there were two different generations or, in other words, two different stages of garnet growth in the rock -- the spongy garnets grew earlier than the homogeneous ones.
- (iv) There is an example where a homogeneous garnet rim appears to have partially overgrown a spongy garnet core (Fig. 5.30); the opaque trails are less conspicuous in the spongy inner part, but continue undeflected from the matrix through the homogeneous outer part. This suggests that the present main foliation developed through a phase of strong foliation transposition postdating the formation of the spongy garnets, whereas the homogeneous garnet growth took place more or less concomitantly with the foliation transposition.
- (v) The thin section does not preserve any evidence of folding (say, F1) that might have taken place prior to the formation of the included staurolite grains. The staurolite crystals are quite irregularly oriented within the garnet which indicates that in between the staurolite formation and the growth of the enclosing garnet there was no episode of folding. If there were one, that would have imparted some sort of

regular arcuate distribution of the staurolite crystals. This also gives an indirect indication that if at all there was a fold episode it must have been either concomitant with or postdating the spongy garnet growth.

(vi) In addition to the elliptical grains, some of the spongy garnets show folded outlines (Fig. 5.31) which means that indeed a fold episode took place either synchronously with or following the development of the spongy garnets, and most possibly before the stretching event. Given the isotropic nature of garnet, it is very difficult to know whether a garnet grain is internally strained or not. A mineral grain of pre-folding origin would obviously show greater internal strain than a syn-folding one. Some clues are provided in the present case by the anisotropic included phases within the garnet/s. There is a lack of strong undulose extinction in those included grains. This along with the observation that the garnet shows only mild (gentle) folding possibly favours an early-syn-folding origin of the garnet, rather than a distinct pre-folding origin.

(vii) Pieced together, these evidences build up a sequence of events which closely resemble that found in the Joshimath Gneiss, that is, formation of high-grade metamorphic mineral e.g. kyanite, followed by an episode of folding i.e. F<sub>2</sub>, followed in turn by the main stretching event. The only difference in the present case is an additional episode of garnet growth (the spongy ones) broadly synchronously with the folding (F<sub>2</sub>). This is hardly surprising when we are dealing with a thick, complex and repeatedly active shear zone such as the MCT (i.e. the Munsiri Formation). Also as already highlighted, garnet is one of the few metamorphic minerals that show exceptional readiness to grow in multiple stages in preferred sites depending on the local availability of necessary reactants. The broadly syn-F<sub>2</sub> time of development of the spongy garnets implies that the included staurolites were pre-F<sub>2</sub>. As indicated above at point (v), the included staurolites were either syn-F<sub>1</sub> or, more likely, post-F<sub>1</sub>, because there was no folding after the staurolite formation, before they got enclosed within the spongy garnet. This pre-F<sub>2</sub>, but syn- to post-F<sub>1</sub> time of development of the included staurolites, established independently, suggests that the included staurolites are synchronous with the matrix staurolites.

The thin section (NR 58/59) preserves clear evidence of at least two generations of garnet growth and provides information which leads to the following reconstruction of events --

Following the staurolite formation the growth of the spongy garnets took place broadly synchronously with an episode of folding (F<sub>2</sub>) which was followed in time by a strong event of foliation transposition and stretching that gave rise to the present main foliation along which the opaque trails are

aligned. The spongy garnets enclosed some staurolites from selected horizons, thereby protecting them from the effects of later realignment and fragmentation due to foliation transposition and stretching/shearing. The remaining matrix staurolites being 'unprotected' became aligned and stretched along the main foliation. The strong event of foliation transposition is almost certainly associated with the main stretching or shearing event related to the MCT-emplacment. The homogeneous garnets grew either synchronously or late synchronously with this foliation transposition thereby overgrowing the opaque trails ( i.e. foliation).

It is concluded that there was essentially one generation of staurolite in the Munsiri Formation, but in two different modes of occurrence. The Munsiri staurolites were also an integral part of the pre-F2, but syn- to post-F1 progressive regional Barrovian metamorphic sequence described already from the Joshimath Gneiss. In other words, the Munsiri-staurolites and the Joshimath-Gneiss-staurolites belong to the same metamorphic episode. On the mineral paragenetic diagram, therefore, the position of the Munsiri staurolites corresponds to that of the Joshimath Gneiss staurolites (Fig. 5.32).

## 5.4 GARNET TEXTURES

Garnets are all porphyroblastic in origin and show a wide variety of textures. This section first gives a review of the literature on garnet textures, particularly on porphyroblast inclusion fabrics and then a description of the major varieties of garnet textures recognised in the Joshimath area, along with an attempt at classification. Wherever possible, attempts are also made to find out if such grouping has any age connotation. Textures of garnet refer to the patterns of distribution of inclusions within the garnet, the overall shape of the garnet grains, the geometric relation of the inclusion fabric (Si) with the external foliation/fabric (Se), and the mode of occurrence of the garnets in relation to the surrounding mineral phases in the matrix.

### 5.4.1 A review of literature on garnet textures

The literature is replete with studies on the textures of porphyroblastic garnets, particularly on the origin and significance of inclusion trails commonly observed in them. The inclusion trails in garnet porphyroblasts are often curved, S-shaped or spiral in nature and are widely designated as the internal schistosity (Si).

The geometric relations of these internal inclusion trails (Si) with the external foliation (Se) in the matrix surrounding the garnet porphyroblasts are useful in working out the tectonometamorphic history of the rocks. Though the patterns of inclusions within porphyroblasts and their geometric relations with the external foliation have long been used to distinguish between pre-, syn-, and post-tectonic mineral growth (see, for instance, Read, 1949; Rast, 1958; Zwart, 1960a & b, 1962, 1963; Johnson, 1963; Spry, 1963; Harte & Johnson, 1969; Ferguson & Harte, 1975), the literature on syntectonic porphyroblastesis presents conflicting interpretations on various aspects, viz. on the mechanism of formation of the S-shaped trails (Spry, 1963; Schoneveld, 1977; Carmichael, 1969; Bell et al., 1992), on the nature of matrix deformation (Spry, 1963; Wilson, 1971; Bell, 1985; Bell & Johnson, 1989), on the tectonic significance of porphyroblast rotation (Spry, 1963; Ghosh, 1975; Schoneveld, 1977; Lister & Williams, 1983; Bell & Johnson, 1992; Bell et al., 1992), on the three-dimensional shape of the inclusion trails (Spry, 1963, 1969; Treagus, 1964; Powell & Treagus, 1967, 1970; Rosenfeld, 1970; Bell et al., 1992) and also on the problem of whether it is the porphyroblasts or their matrix which rotate (Ramsay, 1962; Spry, 1963; Schoneveld, 1977; Wilson, 1971; Bell, 1985; Bell & Johnson, 1989; Bell et al., 1992).

The geometry and origin of inclusion patterns in garnet porphyroblasts are explained in various ways by various authors (for an appreciation, see Schmidt, 1918; Spry, 1963, 1969; Powell & Treagus, 1967, 1970; Carmichael, 1969; Wilson, 1971, 1972; Rosenfeld, 1970; Harvey & Ferguson, 1973; Cox, 1969; Bell, 1985; Bell & Johnson, 1989; Bell et al., 1992). However, there are broadly three groups of hypotheses as to the origin of the porphyroblast inclusion trails (Si) respectively involving:

(1) rotation of the porphyroblasts (see, for example, Peacey, 1961; Spry, 1963; Cox, 1969; Powell & Treagus, 1970; Dixon, 1976; Schoneveld, 1977; Williams & Schoneveld, 1981; Lister & Williams, 1983)

(2) rotation of the matrix (see, for example, Ramsay, 1962; Wilson, 1971; Bell & Rubenach, 1983; Bell, 1985; Bell & Johnson, 1989; Bell et al., 1992), and,

(3) differential rotation of the porphyroblast and the matrix (see, for example, Ramsay, 1962; Wilson, 1971; Bell & Rubenach, 1983; Olesen, 1978).

*Spry (1963) favoured the rotational origin of inclusion trails; in fact, he gave a firm footing to the 'Snow-ball' hypothesis of porphyroblast rotation. The hypothesis*

of rotational origin of porphyroblast inclusion trails was supported by the work of Schoneveld (1977). Meanwhile Wilson (1971) mentioned the possible role in the development of the sigmoidal inclusion trails of overgrowth of a progressively deforming matrix by a growing porphyroblast. Finally, Bell (1985), Bell & Johnson (1989) and Bell et al. (1992) came up with the suggestion that porphyroblast rotation is extremely rare in nature; according to them, deformation partitioning controls how and when the inclusion trails can develop in porphyroblasts.

Given below is a more detailed review of the works of Spry (1963), Schoneveld (1977), Wilson (1971), Bell (1985) and Bell & Johnson (1989); these works made significant impact on our understanding about porphyroblast growth and inclusion trail development.

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### Spry's work

Spry (1963) emphasized that the development of sigmoidal syntectonic inclusion trails in garnet is due to simultaneous rotation and growth of the garnet porphyroblasts. Drawing an analogy between the development of such a porphyroblast and a snowball rolling down a snow-covered slope, he says "The snowball rotates and picks up extra snow thus increasing in radius as it rolls. The points of contact between the snowball and the slope become enclosed when the next layer of snow is attached and the locus of the points of contact within the ball is a spiral. The nature of this spiral is controlled by the rate of growth and the rate of rotation' (Spry, 1963, p.217). Snowball structures in porphyroblastic minerals have attracted the attention of various workers since early part of this century (see, for example, Flett, 1912; Schmidt, 1918; Mugge, 1930; Fairbairn, 1949; Read, 1949). Spry's hypothesis suggests that the garnet porphyroblast grows preferentially along the foliation and that the series of elongate quartz inclusions are not parts of the once-continuous external foliation (Se). As the porphyroblast rotates, preferential growth always occurs along Se even though the included Si have been rotated away from the Se-direction (Fig. 5.33).

As to the behaviour of the matrix, Spry (ibid.) mentioned that the medium (matrix) probably flows in a streamline pattern around the porphyroblast. The crystal deflects the flow layers sideways as it grows but the close presence of a number of rigid crystals may prevent flow within some layers and increase the viscosity. The couple acting across the crystal is the cause of rotation and this may change during growth resulting in a change in the rate of rotation. The rate of growth may be

controlled by the availability of the material, and by its viscosity and temperature which may also vary with time.

### Schoneveld's work

Schoneveld's (1977) work can be regarded as the culminating influence so far by the advocates of rotational origin of syntectonic porphyroblast inclusion trails. Early supporters of porphyroblast rotation dealt with inclusion patterns only in two dimensions, perpendicular to the supposed rotation axis. Powell & Treagus (1967 & 1970) made the first attempts to discuss in more detail the three dimensional shape of inclusion surfaces, but they considered only those garnets which suffered 90°-rotation. Rosenfeld's (1970) comprehensive study dealing with highly rotated garnets came up with a ring model for the three dimensional shape of the central inclusion surface. Rosenfeld identified two types of included surfaces in garnets and considered them broadly as schistosity and compositional banding, but didn't clarify exactly how they come to be preserved as well defined alternate spirals, sometimes with varying widths. Moreover, the occurrence of double quartz spirals and/or double opaque-graphite-mica spirals were not explained. Schoneveld's work aimed to further our understanding of the two dimensional inclusion patterns in central sections perpendicular to the rotation axis in garnets that suffered rotation of 180° or more. It highlighted the role of porphyroblast growth rate vs. rotation rate ( $g/r$ ) on the progressive thickening or thinning of inclusion spirals from centre outward of the porphyroblasts.

The importance of Schoneveld's model stems from the fact that it can explain many enigmatic attributes of the inclusion trails observed in nature. Schoneveld (1977) demonstrated that the factor ( $g/r$ ) given by the ratio of porphyroblast growth rate ( $g$ ) to porphyroblast rotation rate ( $r$ ) plays an important role to induce different characters to the inclusion trails. The variation in inclusion trail thickness may depend on a change in the  $g/r$  ratio, where  $g$  is given in terms of the increase in radius of the porphyroblast. If  $g/r$ -ratio is kept constant, the width of inclusion spiral remains unchanged. If the  $g/r$ -ratio decreases (implying decreasing growth rate, or increasing rotation rate or both), a continuous thinning of the quartz spiral towards the porphyroblast rim is observed; also the Si hinges become sharper and less frequent. Broadening of quartz spiral towards the rim can take place when there is a relative increase in growth rate (i.e. when  $g/r$ -ratio increases). The angle between the incremental growth surfaces of porphyroblast and the quartz and/or opaque inclusion spirals emerging out of the porphyroblast is directly related to the  $g/r$ -ratio; the lower

this ratio, the lower the angle, meaning thereby that in a relatively slowly growing porphyroblast the spirals tend to become parallel to the porphyroblast faces. Narrow quartz spirals lacking distinct Si hinges (suggesting low  $g/r$  ratios) are therefore expected to run almost parallel to the crystal outline of the porphyroblast; thus the double quartz spiral may mimic the euhedral form of the crystal (porphyroblast).

The occurrence of garnet porphyroblasts without any quartz inclusions, but showing only opaque spirals is common. Referring to Rast (1965) and Spry (1969, p. 170), Schoneveld (*ibid.*) mentioned that the ability of a crystal to include foreign material depends on several factors. For example, depending on its rate of growth and the interfacial energy variation between the host and the impurity, the host crystal may become strongly poikiloblastic or may show no inclusions at all. Such factors may apply to rotated garnets as well, which may sometimes preserve the opaque minerals and eliminate the quartz. Also due to varying diffusional properties between garnet and quartz, and the obvious scarcity in the quartz-rich pressure shadow areas of suitable reactants for garnet formation, garnet-growth may selectively avoid the quartz-rich zones, so quartz does not become enclosed within garnet. In a slightly different, but broadly similar situation, a rotating garnet can have near-concentric, circular inclusion trails. However, circular inclusion trails (spherical in three dimensions) have been explained in terms of replacement of matrix coupled with displacement by 'crystallisation force' by Harvey & Ferguson (1973), though this suggestion has been later questioned (see Spry, 1974).

The main contribution of Schoneveld (*ibid.*) came from the results of his ingeniously performed 'String Experiment' to simulate simultaneous growth and rotation of porphyroblast in a quartz-mica rich foliated rock (see pp. 455-457, particularly Figs. 1 & 2 in his paper). Fig. 5.34 (modified after his Fig. 2) shows four stages (a to d) in the progressive rotation and growth of an idealised circular garnet porphyroblast with accompanying Si-Se fabric. The diagrams were originally sketched (traced) from photographs taken at four stages of the 'String Experiment'. The lines (strings) outside and inside the circular garnet model indicate Se and Si respectively. Where the string/line segments are more closely spaced they correspond to the mica-rich (or opaque/graphite-rich) domains and areas where they are more widely spaced represent quartz-rich domains. The amount of rotation is given in degrees under each of the four sketches; the progressive increase in porphyroblast size (from a to d) is self-evident. More clearly distinguishable in (d), the white areas inside and outside the garnet model represent respectively the quartz-rich relict (i.e. enclosed) pressure shadows and the present ones. Thus in garnet porphyroblasts with

large rotations (say,  $360^\circ$  or more), we get a pair of double spirals - a double quartz spiral (white bands) and a double opaque spiral (dark bands). Note that the lines (Si trails) get their maximum curvature when they individually pass through the quartz-rich pressure shadow trails (white). Schoneveld referred to these strongly curved segments as "hinges of Si trails", though they do not represent real fold hinges. An important point to note is that the opaque-mica-graphite rich spiral domains within the porphyroblasts do not themselves represent the early foliation of the rock, although many workers consider them to be so; except in the very core region of the porphyroblast these simply represent, according to Schoneveld, a new internal fabric within the porphyroblast developed as a result of bunching together of fine scale early foliation (Si) consequent upon the porphyroblast rotation.

The few points that are not clearly dealt with in Schoneveld's paper are -

(a) The criteria based on which the orientation of porphyroblast rotation axes have been identified. A porphyroblast may have different rotation axes at different points of time in its history.

(b) Concerning the time-relation between the growth of pressure shadow quartz and the garnet porphyroblast, when he (Schoneveld, 1977, p. 454) said that 'they are essentially produced synkinematically', he probably had compared the later growth stages of garnet with the pressure shadow grains. But there might be appreciable gap between their initial nucleation times, because pressure shadows can form only when an already existing relatively more competent (rigid) object behaves as a barrier against the pressure to act homogeneously on the rock.

(c) How far the suggested model of inclusion trail development would be applicable to porphyroblasts of minerals other than garnet.

(d) The influence of foliation transposition during or immediately after the growth of the porphyroblast.

### Wilson's work

Wilson (1971) pointed out that in nature the deformation of the matrix surrounding the porphyroblasts may be much more complicated than normally considered by workers advocating a rotational origin of sigmoidal inclusion trails in porphyroblasts. He emphasized the incremental aspect of deformation and mentioned that the supposed rotation axis may vary in orientation with time. 'The situation is analogous to the change in direction of fold axes generated during progressive three-dimensional deformation (Flinn, 1962; Ramsay, 1967, p.177)'. However, in some

cases the attitude of the porphyroblast rotation axis for each incremental deformation may remain constant throughout the whole duration of deformation.

Referring to the origin of S-shaped inclusion trails, Wilson (*ibid.*) made a case whereby the gradually growing porphyroblast encloses at successive stages parts of a progressively deforming or rotating matrix fabric (Fig. 5.35). The final internal fabric (Si) would thus become a record of the successive orientations of matrix fabric adjacent to the garnet. If, therefore, somehow the growth increments of the porphyroblast could be distinguished (*cf.* Harte & Henley, 1966), then each growth increment would preserve the Si which will give the matrix orientation at that time and place. Wilson's work essentially highlights that the development of sigmoidal inclusion trails in porphyroblasts in some cases can be attributed to the progressive rotation of the foliation surrounding the porphyroblasts due to increasing deformation and/or transposition.

The limitation of Wilson's model is that it is inadequate to explain inclusion spirals which imply relative rotation of more than  $90^\circ$  (see Barker, 1990, p.84). Note that the sense of curvature of the inclusion trails in Wilson's model is opposite to what would be expected if the porphyroblast itself were to rotate.

#### **Bell and coworkers' work**

Bell and his coworkers (for example, Bell, 1981, 1985, 1986; Bell & Rubenach, 1983; Bell et al., 1986; Bell et al., 1989; Bell & Johnson, 1989; Bell et al., 1992) are the strongest proponents of the idea that most inclusion trails originate without involving active rotation of the host porphyroblasts. Bell (1985) emphasized the role of deformation partitioning for the syntectonic development of porphyroblasts without their active rotation. Ductile deformation affecting a rock-mass would be partitioned broadly into four different types of zones - (i) that takes up no strain (competent or rigid zone), (ii) dominated by progressive shortening strain (*i.e.* zone of progressive coaxial deformation), (iii) dominated by progressive shortening plus shearing strain (*i.e.* zone of progressive non-coaxial deformation), and (iv) dominated by progressive shearing strain (*i.e.* another zone of progressive non-coaxial deformation); see Fig. 5.36 where the first three types are clearly indicated.

Taking the most likely deformation regime that may induce porphyroblast rotation, that is non-coaxial bulk inhomogeneous shortening, Bell (1985) suggested that the shearing component of deformation is partitioned about the porphyroblast

which normally acts as a rigid undeformable object and thus protects an ellipsoidal island of matrix from the effects of progressive shearing. Thus the porphyroblasts escape rotation, while the surrounding foliation cannot but be reactivated due to partitioning of the deformation around the porphyroblasts. Consequently, progressively growing porphyroblasts commonly preserve in their strain shadows or within their bodies in the form of inclusion trails the orientation of early foliations and stretching lineations, even if these structures have been rotated or obliterated in the matrix due to subsequent deformation. This reinforces the fact that porphyroblast inclusion trails are very useful tools for the study of foliation development through repeated transpositions and therefore, for deciphering deformation history. Fig. 5.37 schematically shows porphyroblasts with differing internal fabric but all with broadly contemporaneous development overgrowing different stages and places of formation of a crenulation cleavage belonging to a second deformation episode. Bell mentioned that porphyroblasts can rotate only when the deformation cannot partition and involves progressive shearing with no combined bulk shortening component. In this context, the model put forward by Ghosh (1975) of rotation of a rigid sphere in a homogeneous isotropic matrix or Lister & Williams' (1983) model of rotation of a porphyroblast due to vorticity during 'fluid' flow becomes quite relevant.

Taking representative rocks carrying inclusion trail-bearing garnet porphyroblasts (so-called 'snow-ball' garnets) from three areas of the world viz. Chester Dome area of Vermont in U.S.A., Modi Khola Valley of Nepal and Hunza Valley of Pakistan, Bell & Johnson (1989) carried out a detailed microstructural analysis in order to find out how the inclusion trails developed in the garnet porphyroblasts. They suggest that the spiral-shaped inclusion trails do not form by rotation of the growing porphyroblasts relative to the geographical coordinates; instead the trails form by progressive growth by porphyroblasts over several successive sets (often, 6) of near-orthogonal foliations with alternate sub-vertical and sub-horizontal orientations attained through repeated transposition (see their Figs. 1 to 14 and also Fig. 20 reproduced here as Fig. 5.38). Thus this idea of inclusion trail development and porphyroblast growth is essentially an extension of that in Bell (1985) and in a somewhat similar vein to that of Wilson (1971).

Bell & Johnson (1989) also suggested that the above mentioned process of inclusion pattern development in growing porphyroblasts reflects the basic pattern of orogenesis that involves, in their opinion, a multiply repeated two-stage cycle of - (1) crustal shortening, thickening and uplift giving rise to near-vertical foliation, followed by (2) gravitational instability and collapse of the uplifted mass with the

development of near-horizontal foliation (see their Figs. 24 to 26). They say that thrusting in orogenic belts mostly occurs in the second stages as a result of gravitational spreading and near-coaxial vertical shortening.

Bell and coworkers' model calls for several criticisms.

Firstly, it is implied in the model that the suggested mode of growth of porphyroblasts preserving the imprints of so many (often, 6) successive phases of foliation transposition is not special for garnets only; all porphyroblastic minerals seem to be equal. Taking into account a noncoaxial bulk inhomogeneous shortening regime (Bell, 1985), garnets and other porphyroblastic minerals would probably be comparable in terms of strain partitioning capability. Yet we hardly ever find porphyroblasts of any other mineral which show such a prolonged and complex growth history as has been depicted in the model taking garnets as example. This suggests that definitely there is some inconsistency or lacuna between the prediction (the suggested model of porphyroblast growth) and what actually happens in nature.

Secondly, the differences in the crystal structures, external shapes and physical properties among porphyroblastic minerals must be important factors which are neglected in the model proposed by Bell & his coworkers. Highlighting the importance of the external shape of the porphyroblastic grains, Prior (1987, p.27) commented, 'Unless the deformation is so strongly partitioned that no deformation of the porphyroblasts and their immediate surrounds occurs, inequidimensional porphyroblasts will rotate'.

In contrast the porphyroblast rotation model has the obvious underlying assumption that the ease of rotation would depend much on the shape and properties of the porphyroblastic crystal and therefore it can readily explain why garnets, having isotropic-dodecahedral character, tend to suffer more rotation and, therefore, show much more complex inclusion trail fabric than other minerals.

Thirdly, it is quite an intriguing coincidence how almost each and every garnet porphyroblast studied by Bell & Johnson (1989) from the thin sections of rocks collected from three widely separated areas on the surface of the globe invariably preserves five or even six generations of foliations! This would mean that garnet-growth was taking place for long enough time to witness almost the entire history of tectonometamorphism affecting the rocks since the nucleation of the porphyroblasts. This is in contrast to the findings of Cashman & Ferry (1988) which indicated that garnet porphyroblasts may grow to their maximum size well within 100,000 years.

Fourthly, the suggested process of orogenesis involving successive stages of horizontal compression and horizontal extension (i.e. spreading as a result of vertical compression due to gravity) seems to be too simplistic to be true and the suggested symmetrical geometry of orogenic belts seems to be untenable particularly in relation to the collisional orogenies.

Fifthly, in scientific terms it is rather an unacceptable approach to infer about the overall deformational pattern of megascopic mountain belts just from the observations on some microscopic porphyroblasts. There must be sufficient coherent and correlative findings in the wide intermediate range of scale.

Finally, their claim that porphyroblast rotation is rather unlikely in nature seems to be somewhat dogmatic. The examples shown in Chapter 4 of this thesis of pulled apart and/or tilted garnet or other porphyroblasts or their fragments within mylonitic schists formed in a shear regime with complementary extension suggest that porphyroblast rotation is not so rare in nature.

However, the strongest influence of Bell and coworkers' work is that it now makes us observe the porphyroblast-related microstructures very carefully and not to take it for granted that any sigmoidal inclusion trail-bearing syntectonic porphyroblast is of rotational origin. What Bell and his coworkers suggested about the development of porphyroblast inclusion trails may not be entirely implausible. The suggested model may well be true at least for some garnet porphyroblasts in nature.

An important by-product of Bell and coworkers' analyses is the strong support for the recognition of the fact that texturally distinct concentric or sub-concentric zones in the porphyroblasts are representative of discrete growth zones no matter whether there is any compositional variation among those zones or not. The detailed review given above indicates that in this particular aspect there is no diversity of opinion among the different workers. Compositional variation by itself is not a sufficient criterion to indicate multiple growth stages, but its presence in addition to textural variation provides a further confirmation. However, this does not mean that the absence of textural and/or compositional variation in a porphyroblast automatically implies that it has developed through a single stage of growth, though this is a very likely possibility. Also simply from a recognition of the presence of multiple generation of porphyroblast growth, no firm indication could be given as to

the rate of growth of the porphyroblasts in different stages. In order to estimate such rates, we need to know the ages of different growth zones of the porphyroblasts. For the present thesis the aim is to identify whether there are multiple generations of porphyroblast growth in the area and if so, to establish when such growth took place in terms of the deformation episodes. Precise dating (absolute) of the porphyroblasts and estimation of porphyroblast growth rates would form an important subject-matter for future study.

We, therefore, find that most of the early workers (e.g. Flett, 1912; Schmidt, 1918; Mugge, 1930; Fairbairn, 1949; Read, 1949; Peacey, 1961; Spry, 1963) did not believe in any other process than the rotational 'snowball' origin of the porphyroblast inclusion trails. Ideas gradually evolved and now we recognise that for the development of the inclusion trails rotation of the porphyroblasts is not always necessary. In many cases the porphyroblasts can grow *statically*, but it is due to the ~~progressive rotation of the foliated matrix surrounding the growing porphyroblasts~~ that the curved inclusion trails develop in them (Wilson, 1971; Bell, 1985; Bell & Johnson, 1989, Bell et al., 1992).

The fairly detailed review given above on the different models of porphyroblast growth and inclusion trail development makes it clear that no one model is universally acceptable at the exclusion of others. Because the mode of growth of a porphyroblast does not depend only on the intrinsic properties of the mineral concerned, but the tectonometamorphic environment also plays an influential role, it is obvious that depending on the variation in the tectonometamorphic regime, local as well as regional, the porphyroblasts will show different modes of growth. This point will be evident from the following discussion on the wide textural variety of garnet porphyroblasts from the Joshimath area.

#### 5.4.2 Garnets of the Joshimath area

Garnet is common in the Joshimath Gneiss and the Munsiri Formation, whereas in the Berinag-Mandhali formations it is noticed only in a thin phyllitic band sandwiched between two quartzite horizons only a few metres below the Munsiri thrust. In the upper part of the Munsiri Formation, garnet is very common in the semipelitic horizons, but it becomes less common down-section. The quartzofeldspathic mylonite-protomylonite and/or augen gneissic horizons do not contain garnets. Generally speaking, garnet is noticed quite frequently (except in the quartzofeldspathic horizons) down to the lower-middle part of the Munsiri

Formation below which its occurrences are sporadic. Many amphibolite horizons both in the Munsiri and in the Joshimath Gneiss are garnet-bearing. As mentioned before, all garnets in the area are porphyroblastic. The characteristics of the garnets are broadly the same in the Joshimath Gneiss as in the Munsiri Formation except that the Munsiri garnets generally show a more complex internal inclusion trail pattern and a stronger effect of stretching. The results from the present study do not support the view of Valdiya (1980, p. 87) that the variation in the size of garnet porphyroblasts can be considered as a criterion distinguishing the Munsiri rocks from the Joshimath Gneiss. Valdiya stated that the Munsiri garnets are 'commonly very small-sized' and are 'confined to a few horizons which are few and far between'. But in fact some garnet-amphibolite horizons of the Munsiri Formation show the largest among the garnet porphyroblasts from the present area of study. This subsection gives a description of the different textural varieties of the garnet porphyroblasts present in the area. The object is to classify them into distinct groups with the ultimate aim of recognising their time/s of growth with respect to the deformation episodes.

Texture of garnet porphyroblasts is taken to refer to the overall shape and size of the garnet grains, the patterns of distribution of inclusions within the garnet, the geometric relation of the inclusion fabric (Si) with the external foliation (Se) and the mode of occurrence of the garnets in relation to the neighbouring mineral phases in the matrix. As far as the rotational or non-rotational origin of the inclusion trail-bearing porphyroblasts is concerned, Bell and his coworkers (see review above and references therein) suggested that mainly millepede microstructures and abrupt termination or truncation of one inclusion trail against another imply a non-rotational origin of the porphyroblasts. Absence of any such structures and presence of smooth inclusion spirals or sigmoidal trails within porphyroblasts are generally taken to be indicative of a rotational origin.

The inclusions in the garnet porphyroblasts are mostly of quartz, opaque minerals including graphite, and/or micas (mainly biotite). But in a number of instances inclusions of tourmaline, carbonate, clinozoisite, chlorite etc. are also noticed.

Theoretically possible time slots for the development of garnet porphyroblasts in the Joshimath area, particularly in the Joshimath Gneiss and the Munsiri Formation, are as follows:

- Post-stretching (thus, broadly post-F3) [Here stretching refers to the main stretching event]
- Syn-stretching (broadly syn-F3)
- Post-F2, pre-stretching
- Syn-F2
- Pre-F2, post-F1
- Syn-F1
- Pre-F1

As mentioned earlier, clear-cut evidence of mineral growth during F1 folding is relatively scarce due to the lack of suitable specimens of F1 fold hinges. Thus it could not be properly determined whether garnets grew specifically at pre-F1 or syn-F1 time. If any mineral is found to have grown before the pre-F2, post-F1 stage, it will be assigned a pre- to syn-F1 age. Also because it is quite difficult to determine whether a garnet porphyroblast developed at syn-stretching time or post-stretching time, the two different time slots (syn-stretching and post-stretching) are combined into one (syn- to post-stretching). Therefore, practically we are left with five time slots:

- Syn- to post-stretching (broadly syn- to post-F3 time)
- Post-F2, pre-stretching
- Syn-F2
- Pre-F2, syn- to post-F1
- Pre- to syn-F1

#### Criteria used for relative age determination of garnets

Wherever suitable, the following criteria have been used for the determination of relative age of the garnet porphyroblasts:

1. The mode of occurrence of garnet grains relative to age-defining microstructures. For instance, if in a rock (thin section) we find some garnet grains elongate in habit and are either folded or lying along a foliation that defines a fold, whereas other garnet grains, spheroidal in habit, occur along the axial planar foliation, then we can deduce that the former group of garnets are pre-folding, whereas the latter group are most probably syn- or post-folding (syn- or post-axial planar foliation) (see Fig. 5.39a).
2. Texturally distinct groups of garnets in a rock with homogeneous structure and composition are likely to be of different ages. For instance, if in the same rock there

are garnets with homogeneous bodies and well defined euhedral outlines, while other garnets show skeletal or filamental growth features, then it is possible that the skeletal garnets are later than the homogeneous, euhedral garnets. Probably the skeletal group would represent a stage of garnet growth when the rock was already depleted in garnet-forming reactants (Fig. 5.39b). Note that the situation would be reverse if either the skeletal pattern in the garnet is a replacement or alteration feature (Fig. 5.39c) or, the overall outline of the skeletal porphyroblast suggests deformation (Fig. 5.39d). In these two latter cases, the skeletal or spongy garnet would be earlier than the euhedral, homogeneous-bodied garnet.

3. Si-Se relations. If a foliation passes undeflected through some garnets in a rock, but wraps around others then obviously the latter garnets antedate the former (Fig. 5.39e). Single porphyroblasts may show multiple growth zones where Si and Se are continuous across the outer zone but not in the inner zone, implying that there was a distinct gap in time between the formation of the inner and outer growth zones. The criteria discussed by Zwart (1962) could also be relevant at this point.

4. Degree of alteration or chloritisation. By itself this is comparatively a weak criterion. However, if we can assume that access to fluid per unit time is more or less the same for all garnets in a rock, then the more altered grains are likely to be older than the less altered. This is akin to the aging process. In suitable circumstances, the assumption about fluid access can be checked under the microscope, for example, by noting whether there is any microfracture, vein or chlorite-bearing micaceous foliation in close vicinity of the garnet grains.

5. Growth zoning features. Textural zoning features are very good indicators of multiple growth stages of garnets. Though by themselves these features cannot always clarify the relations between garnet growth and deformation episode, they provide very useful circumstantial evidence and stimulate a search for more obvious examples of different generations of garnet.

### Garnets in the Joshimath Gneiss

Broadly five texturally distinct classes of garnet porphyroblasts have been recognised in the Joshimath Gneiss. They are:

[Note that, if not otherwise mentioned, the illustrations referred to below are from thin sections cut across the main foliation, but parallel or sub-parallel to the main stretching lineation.]

I. Spongy/Skeletal/Filamental garnets (see Fig. 5.40a; thin section reference - 'Chain-rock'). Almost invariably the skeletal character of these porphyroblasts is a growth feature rather than secondary fragmentation feature, because each porphyroblast occurs as a single grain and has discrete outline; moreover, the branching parts within it neither show any zig-zag fracture features nor any patterned distribution. Fig. 5.40a shows just a small part (about one-fifth in size) of a large spongy porphyroblast. The overall outline of the porphyroblasts is elliptical with their elongation parallel to the direction of stretching lineation, a fact which suggests that these porphyroblasts developed prior to the main stretching event. The rocks in which these skeletal garnets have been found do not show any other kind of garnet porphyroblasts, and the composition of the rock is not highly favourable for profuse garnet development. This is why the second criterion of relative age determination mentioned above cannot be applied properly in this case. However, the deformed outline of the porphyroblasts possibly indicates an early age.

II. Elongate porphyroblasts with length-parallel elongate inclusions. No clear crystal outline is found on these porphyroblasts. In one case, folded outline of such a porphyroblast is noticed at a fold-hinge (related to a local fold possibly developed shortly after F2) (Fig. 5.40b; thin section reference - NR25/26F(2)).

III. Almost entirely homogeneous porphyroblasts with euhedral outline, occasionally including only a small central spotty core of inclusions (Fig. 5.40c; thin section reference - 1F'87).

IV. Circular or rounded porphyroblasts with homogeneous outer rim surrounding a comparatively large inner zone with inclusions. Two different types are recognised within this class - (a) those with the inclusions in the inner zone arranged in concentric trails (Fig. 5.40d; thin section reference - NR30F), and (b) those with a large inner inclusion-bearing zone surrounded by thin homogeneous inclusion-free outer rim. Crude spiral arrangement of the inclusions in the core region is noticed (Fig. 5.40e; thin section reference - NR33B).

V. Porphyroblasts with well developed inclusion spirals. In this class also two slightly different types are recognised - (a) porphyroblasts with subcircular outline showing no apparent effects of stretching (Fig. 5.40f; thin section reference - 12ab), and (b) inclusion spiral-bearing porphyroblast with somewhat elliptical outline - the ellipse long axis follows the stretching direction (Fig. 5.40g; thin section reference - 4Ab).

Generally, in most classes of garnet porphyroblasts the associated matrix foliation wraps around the garnets. It is noted that distinct S-shaped inclusion trail-bearing garnet porphyroblasts are rather rare in the Joshimath Gneiss in contrast to the Munsiri Formation (see later); probably this is not unexpected, given that the Munsiri Formation represents the main shear zone (MCT-Zone). As far as the mode of origin of the porphyroblasts is concerned, it is very likely that the class V porphyroblasts had a rotational origin (cf. Schoneveld, 1977 reviewed above). The class IV porphyroblasts show an early rotational phase with a subsequent static homogeneous overgrowth. Porphyroblasts of classes IV & V came from relatively near the Vaikrita Thrust. The other three classes (viz. classes I, II, & III) did not probably involve any rotation while growing. Presumably while class III porphyroblasts represent static growth in all directions from a nucleus, classes II and I represent overgrowth of an earlier fabric. The relative rarity of elongate porphyroblasts even on XZ-sections in Joshimath Gneiss implies a generally weak effect of stretching upon the rocks compared to the Munsiri Formation.

From the patterns of inclusion textures coupled with external shapes of the porphyroblasts and their relationships with external foliations and/or microstructures (folds etc.) suggestions could be made as to the relative times of growth of the above classes of garnet porphyroblasts in the Joshimath Gneiss. Porphyroblast classes V & IV most probably had their growth broadly synchronously with the main shearing or stretching event in the area - their rotational origin might have been aided by shearing; however, the homogeneous outer rim-overgrowth in class IV porphyroblasts may signify a syn- to post-stretching garnet growth phase. The folded outline of the class II porphyroblast implies that the porphyroblast grew before the folding. In the field it could not be properly ascertained whether this particular fold-hinge represents an F2 fold or not. From the orientational attributes and the relatively isolated development of the fold-hinge it was surmised that it represents a local fold probably developed at the waning stage and thus postdating the main phase of the F2 folding. Hence I am inclined to consider this class II folded garnet to be broadly syn-F2 in origin. Class III porphyroblasts without any internal schistosity (Si), but with euhedral outline and class I porphyroblasts with skeletal form and elongate-elliptical overall outline most probably represent the earliest among the above five classes of porphyroblasts. However, these two classes of porphyroblasts have not been found to occur together in any rock-horizon. Development of the skeletal garnets is attributed to an insufficiency of garnet-forming reactants due to the unfavourable composition of the host lithology. I conclude that the earliest classes (III & I) of garnet

porphyroblasts developed during the early episodes of metamorphism. Garnet porphyroblastesis in the Joshimath Gneiss took place probably in four stages viz. (a) at pre- to syn-F1 time, (b) at pre-F2, but syn-, or more possibly, post-F1 time, (c) at syn-F2 time and (d) at syn- to post-stretching time (broadly syn- to post-F3). See, for a comparison the discussion on garnet porphyroblasts from the Munsiri Formation given below.

### Garnets in the Munsiri Formation

Garnets in the Munsiri Formation, in general, show more complex internal textures than those in the Joshimath Gneiss. The other difference is the much stronger effect of stretching particularly on garnets occurring in the lower middle or lower part of the Munsiri Formation. Most of the largest garnet porphyroblasts in the area are found in some garnet amphibolite horizons of the Munsiri Formation.

What I propose to do here is: first, to suggest a broad classification of the garnet porphyroblasts from the Munsiri Formation based on their textural features. Second, to describe in the light of this classification the different textural varieties of garnet porphyroblasts from a representative set of thin sections. Wherever possible it will be attempted to deduce the time of formation of the porphyroblasts from their textural relations. Third, to enumerate the features of garnets from two thin sections (viz. 7M'87 & NR58/59) that will provide confirmatory evidence about the times of garnet porphyroblastesis in the Munsiri Formation.

Based on their textural patterns, the garnet porphyroblasts in the Munsiri Formation can be grouped into the following classification -

#### I. Porphyroblasts recording single growth episode:

- (a) Skeletal or meshy porphyroblasts
- (b) Circular or ovoid or euhedral outlined porphyroblasts with straight inclusion trails
- (c) Porphyroblasts with well developed inclusion spirals or sigmoidal inclusion trails

#### II. Porphyroblasts recording multiple growth episodes:

- (a) Ovoid or irregularly outlined porphyroblasts with inner core bearing spotty inclusions, or sigmoidal or spiral inclusion trails, surrounded by homogeneous rims
- (b) Euhedral porphyroblasts with inner core bearing spotty inclusions, or sigmoidal or spiral inclusion trails, surrounded by homogeneous rims

III. Porphyroblasts in which growth episodes cannot be recognised e.g. partly or wholly retrograded porphyroblasts with irregular outlines.

#### Class I(a) porphyroblasts

[Example from slide - NR 81]: The thin section is approximately an XZ section and shows a number of bigger, stretched and fragmented garnets that now appear as skeletal porphyroblasts (one is shown in Fig. 5.41). These porphyroblasts are definitely of pre-stretching origin, but it is difficult to say exactly when they developed in terms of fold episodes. A notable feature of the larger (skeletal) garnets is that their component segments do not appear fresh and homogeneous; including lots of minute non-opaque inclusions they show diffused extinction and have altered appearance.

[Example from slide - 39G'87]: Part of a skeletal garnet from this garnetiferous amphibolite is shown in Fig. 5.42. Here most presumably the skeletal nature is a primary growth feature, rather than a secondary fragmentation feature. The entire outline (not shown in Fig. 5.42) of the porphyroblast is elliptical with its elongation parallel to the stretching direction. This means the garnet must be pre- or early-syn-stretching in age.

[Example from slide - NR 53]: A large meshy garnet from amphibolitic rock (Fig. 5.43). The small component parts of the porphyroblast are of various shapes and sizes - most are without any regular outlines, but some have euhedral or subhedral crystal shapes. The overall shape of the porphyroblast is elongate in the direction of stretching implying that the stretching event postdated the development of the porphyroblast.

[Example from slide - NR39/40A]: The rock is a garnet amphibolite and the section is approximately of XZ orientation. Fig. 5.44 shows an example of broadly a class I(a) garnet which seems to have been pulled apart as a result of stretching. The elongate elliptical shape of the quartz-rich zone enclosing the garnet porphyroblast clearly brings out the effect of stretching. The porphyroblast has an irregular outline. Inclusions are aligned parallel to the grain elongation and are thus parallel to the external foliation (Se) in this case. Generally speaking, the inclusion grain size is smaller than the matrix grain size. The garnet is obviously pre-stretching in origin.

[Example from specimen - 6E'87]: Class I(a) type of porphyroblasts (i.e. skeletal garnets) are found in both XZ and YZ sections cut from the specimen. Fig. 5.45 shows an example from the XZ-section (6Ea'87). Note that the overall outline of the skeletal grain is elliptical with its long axis coinciding with the stretching direction. Generally speaking, the aspect ratios (long axis:short axis) of such ellipses are larger in the XZ section than in the YZ section. This evidence clearly shows that these skeletal garnets developed prior to the peak of the stretching event. Less clear is their relation to the inclusion trail-bearing garnets present in the thin section. However, on the basis of the second criterion discussed earlier for porphyroblast age determination it is possible that the skeletal garnets represent a relatively later growth.

#### Class I(b) porphyroblasts

[Examples from slide - 46A'87]: All the garnet porphyroblasts in the thin section belong to the class I(b). They have elliptical outlines, elongate parallel to the stretching lineation. Each porphyroblast shows straight inclusion trails at high angles to the porphyroblast elongation direction (Fig. 5.46). There is a remarkable parallelism in inclusion trail orientation from one porphyroblast to another except only in one garnet grain. The different Si-orientation in only one garnet suggests that the porphyroblast encloses part of a hinge of an earlier tightly or isoclinally folded fabric. No evidence is found which can indicate a rotational origin of these garnet porphyroblasts. The size of the inclusions is consistently smaller than the matrix grain size. This suggests that foliation transposition due to shearing or flattening took place after garnet growth. The elliptical outline of the porphyroblasts probably developed due to the same flattening or shearing that caused the transposition of the foliation. Matrix coarsening took place after the foliation transposition. Thus, the sequence of events is:

1. Development of an early metamorphic fabric
2. Tight to isoclinal folding of the early fabric
3. Garnet porphyroblastesis that overgrew the early fabric
4. Foliation transposition and garnet flattening
5. Matrix grain coarsening

Now *we examine* the time relation of these events with the fold episodes? As we already know, there was a metamorphic fabric (corresponding probably to the earliest metamorphism) before the high grade minerals developed in the area at pre-F2, syn- to post-F1 time. The consistency in the orientation of the early fabric in the

form of Si in the garnet porphyroblasts save only one grain in the present thin section leads us to presume that this early fabric is a representative of the earliest recognisable metamorphic foliation which was subjected to tight to isoclinal folding. Clearly a very plausible candidate for this folding is the F1 fold episode. Thus, presumably the garnet porphyroblastesis took place after this F1 folding. The elliptical outlines of the porphyroblasts parallel to the stretching lineation indicate that these garnets developed before the stretching event (i.e. before the F3 episode). Thus we are left with three time slots for this garnet porphyroblastesis viz. (a) at pre-F2, syn- to post-F1 time, (b) at syn-F2 time, or (c) at post-F2, pre-F3 time, of which the first two demand strong consideration. The third time-slot is probably too late for these garnet porphyroblasts, because we do not find any effect of the strong F2 folding upon the orientation of the Si in garnet porphyroblasts. If the garnets were of post-F2, pre-F3 time, then it would have been very unlikely for the early fabric (now, the inclusion trails) to escape the influence of the major F2 fold episode. On the other hand, if the garnets had developed during the pre-F2, syn- to post-F1 time, then they would have survived a very strong fold episode i.e. F2 and perhaps would not have looked as fresh and well defined as they appear now. So the first time slot also could possibly be discounted. I think, most probably these garnets grew during the syn-F2 time. We shall see later that the syn-F2 time is also an important time for garnet porphyroblastesis in the Munsiri Formation, as in the Joshimath Gneiss.

[Examples from slide - NR 40/41]: There is considerable variation in the textures of the garnet porphyroblasts in this thin section. The most notable ones are described here.

Fig. 5.47 shows a garnet porphyroblast which is so full of inclusions (quartz) that it almost appears like a skeletal garnet. The inclusions are broadly arranged in straight trails orthogonal to the matrix foliation (Se) which wraps around the porphyroblast. The porphyroblast has a circular outline and records only a single growth episode. The inclusion quartz grain size is comparable to matrix quartz grain size. There is no conspicuous sign of rotation of the porphyroblast. Collectively all these points suggest that (a) the Si represents an early foliation and the Se a later transposed foliation, (b) the porphyroblast grew before the foliation transposition, (c) the comparable inclusion and matrix quartz grain size suggests that no matrix coarsening took place after the foliation transposition, or in other words, the rock records evidence possibly suggesting that garnet growth was followed by a relatively short time interval by the foliation transposition. This argument is favoured by the

textural features (described below) of some garnets in the same thin section that possibly record porphyroblast-growth at different stages of foliation transposition (see review of Wilson's work given earlier and also Fig. 5.37).

Fig. 5.48 shows a porphyroblast which has an irregular trapezoid outline and even though the angular discordance between the  $S_i$  and  $S_e$  is small, the distinctly smaller inclusion grain size compared to matrix grain size indicates that the  $S_i$  represents an early fabric. The porphyroblast developed overgrowing this early fabric. This is also supported by a slightly different combination of minerals present as inclusions than in the matrix. The inclusions are of quartz, muscovite, chlorite, calcite, opaque and biotite. The matrix minerals include quartz, biotite, plagioclase, muscovite, tourmaline, clinozoisite, apatite, opaques and sphene. Matrix coarsening took place after the growth of the porphyroblast. There is no indication of any rotation suffered by the porphyroblast.

Fig. 5.49 gives an example where the inclusion grain size is comparable to matrix grain size and the  $S_i$  is in angular conformity with the  $S_e$ . Though the micas (biotites) in the matrix are found to wrap around the porphyroblast, some quartz grains project from the matrix into the porphyroblast. These features probably mean that the formation of the main foliation, i.e. foliation transposition, followed closely in time the growth of the garnet porphyroblast.

Fig. 5.50 shows three garnets which obviously grew before the formation of the main foliation. It is, however, difficult to resolve whether there was any time difference among the formation of these three garnet porphyroblasts.

On the whole the textural investigation of the garnets in the present thin section (NR 40/41) suggests that probably there were subtle differences in the times of formation of the garnet porphyroblasts. Had all the porphyroblasts developed at the same time, we would expect to see broadly the same size of inclusions in all of them, and also a similar relationship with the matrix grain size. Clearly, this is not the case. However, all are broadly pre-tectonic in terms of the main (matrix) foliation. Some porphyroblasts developed much earlier than the foliation transposition which led to the formation of the main foliation and matrix coarsening subsequent to it, while some others were broadly concomitant with the foliation transposition, various porphyroblasts among this latter group recording various stages of the transposition. Any distinct evidence of porphyroblast rotation is lacking; no clear inclusion spiral or sigmoidal inclusion trail is noticed. The differences in the  $S_i$ - $S_e$  angular relations

from one porphyroblast to another could be explained most readily as due to porphyroblast growth at different stages and places of a progressively transposing foliation (see review of Wilson's work and Fig. 5.37 given earlier).

#### Class I(c) porphyroblasts

[Example from slide - 7Pb'87]: The thin section is cut perpendicular to the stretching lineation, which means it is a YZ section. The porphyroblast preserves typical example of spiral inclusion trails (Fig. 5.51). These spiral trails probably suggest a rotational origin of the garnet porphyroblast (see review of Schoneveld's work given earlier). It is not known whether the same porphyroblast would show spiral inclusion trails on an XZ section. But the present case clearly indicates the inherent difficulty in using rotated garnets as shear sense indicators particularly in areas like the present one where there is such a wide textural variety of garnets. Unless the complete rotation histories since their nucleation of a statistically representative number of porphyroblasts in an area are worked out properly, the shear sense determined from rotated porphyroblasts cannot be dependable. For instance, in the present case, the implied sense of shear would be layer-parallel and perpendicular to the stretching lineation indicating almost a strike-slip shearing, a possibility which can be ruled out totally from all the other available evidence. It is very difficult to be sure of spiral inclusion trails indicating the sense of shear in the present area, given the repeated foliation transposition and the different generations of garnet porphyroblasts. Indeed only the tilted fragments of pulled apart garnets may be considered as recording a reliable sense of shear, but this refers to a very late stage of the main shearing or stretching process in the MCT-Zone (see later).

[Examples from slide - NR 52]: Two porphyroblasts are shown with very well defined spiral inclusion trails (see Figs. 5.52a & b). To some extent the matrix foliation wraps around the porphyroblasts, but there are also examples of matrix micas abutting against the garnets which cast some doubt as to whether the porphyroblasts were entirely earlier than matrix foliation. The spiral inclusion trails presumably developed through rotation. So, it is quite possible that these porphyroblasts developed broadly synchronously with the shearing that led to the formation of the present matrix foliation through transposition. An intriguing fact is that the sense of rotation given by the inclusion spires is different in the two porphyroblasts. This again highlights the difficulty in using rotated porphyroblasts for shear sense determination. The irregularities in the outline of the porphyroblasts are probably due to one of the following reasons -

- (a) Primary growth feature
- (b) Secondary fragmentation feature
- (c) A result of differential dissolution by later chemical reaction
- (d) A combination of two or all of the three above-noted reasons.

The tiny patch of garnet on the righthand side of Fig. 5.52b probably represents a separate grain by itself. Also there is no very obvious evidence of differential chemical dissolution or secondary mechanical fragmentation. So, I presume the irregularities in outlines are essentially a growth feature.

[Example from slide - 6A'87]: Here a spectacular S-shaped inclusion trail-bearing garnet porphyroblast is shown (Figs. 5.53a & b). The inclusions are of quartz, opaques and some micas. The nearly circular outline of the porphyroblast, the sigmoidal Si and its relation with the Se suggest that the garnet grew during a non-coaxial deformation and involved rotation (see review of Schoneveld's work given earlier). The relation of the Se with the Si is somewhat ambiguous; in part there is a continuity between Si & Se, yet in general the matrix foliation (Se) wraps around the porphyroblast. Clearcut continuity between Si & Se would indicate that the porphyroblast grew broadly synchronously with the folding of the matrix foliation, but the wrapping or swerving relationship suggests a possible pre-tectonic origin of the porphyroblast in terms of the matrix foliation. On balance I presume the porphyroblast here developed early synchronously with the transposition that led to the formation of the present matrix foliation.

#### Class II(a) porphyroblasts

[Example from slide - NR39/40A]: A distinct example of a class II(a) garnet porphyroblast in an approximate XZ-section is shown in Fig. 5.54. The porphyroblast clearly records two different stages of growth - the inner linear inclusion trail-bearing zone is surrounded by a homogeneous inclusion-free outer rim. The inclusion trail geometry in the inner core region resembles millipede microstructures of Bell (see review given earlier). Note the presence of a mica-free (quartz-dominated) band looking like a 'corona' enveloping the garnet porphyroblast (cf. Fig.- 5.61 from slide - 64A'87 given later that shows a partially retrograded i.e. partially chloritised garnet porphyroblast, surrounded by a distinct 'corona' of white mica).

[An example from slide - ~NR39"]: The thin section is approximately an XZ section in terms of the stretching lineation. The porphyroblast (see Fig. 5.55) has a distinctly

elliptical outline with its elongation parallel to the stretching direction. Broadly two growth zones are present - spotty inclusions are concentrated in the inner/central zone, while the outer rim is almost completely free from inclusions; however, no distinct growth ring/line is found distinguishing between the two zones. Inclusion grainsize is very small compared to the matrix grainsize. No distinct spiral/linear/circular arrangement of the inclusions is recognised.

[Example from specimen - 6E'87]: Figs. 5.56a & b show two class II(a) porphyroblasts from a YZ section (slide - 6Eb'87). The sigmoidal nature of the inclusion trails is very distinct. The Se wraps around the porphyroblasts and the outline of the porphyroblasts appears rather irregular. More than one growth stage is recorded in these porphyroblasts. The latter figure (Fig.- 5.56b) clearly shows an inner core region of S-shaped inclusion trails surrounded by a fairly homogeneous outer rim that probably replaced matrix foliation. Although, in general, the present overall shape of the porphyroblasts is equidimensional in this YZ section, the inclusion trail-bearing zone has a slightly elongate elliptical shape. This enforces the conclusion that possibly the later overgrowths on these porphyroblasts are syn- to post-stretching, while the inclusion trail-bearing zones are pre- to early syn-stretching. If the inclusion-bearing zone can be treated as crude strain ellipsoids, then a comparison of the shape of the porphyroblasts in the XZ & YZ sections suggests that the strain ellipsoid for stretching is of triaxial prolate nature with  $x > y > z$ . This is also corroborated by the aspect ratios of the class I(a) skeletal porphyroblasts discussed earlier from the same specimen.

[Example from slide - 7C'87]: The section was cut at high angle to the stretching lineation. The class II(a) porphyroblast distinctly shows at least two or, more possibly, three stages of growth (see Fig. 5.57). The earliest growth stage, represented by the central ovoidal zone, shows clear S-shaped inclusion trails which indicate rotation during this stage. The inclusion trails evidently do not continue into the later growth stage/s. Later overgrowth did not take place homogeneously on all sides of the inner ovoidal core. The porphyroblast shows partial retrogression (note the presence of chlorite partially surrounding the porphyroblast). The Si-Se or other relations are not clear enough to suggest anything about the possible time difference between the successive stages of growth.

### Class II(b) porphyroblasts

[Example from slide - 23/88]: The rock is a garnet amphibolite. The remarkably euhedral garnet porphyroblast has a somewhat meshy body (Fig. 5.58). It is surrounded by altered/semi-altered plagioclase grains (not shown in the sketch). The rock is possibly an orthoamphibolite. The extreme outer rim of the porphyroblast is inclusion-free, but inside there are sigmoidal inclusion trails. The main body of the porphyroblast shows a rather meshy structure. No effect of stretching or flattening is evident, because the euhedral hexagonal shape of the porphyroblast is virtually intact. This suggests that possibly the growth of the porphyroblast continued up to syn- to post-stretching time. Effect of retrogression (mainly chloritisation) is quite common within the garnet.

[Example from slide - 8G'87]: A number of porphyroblasts with euhedral outline and multistage growth are shown in Fig. 5.59. The inner zones are inclusion-bearing, the outer zones are homogeneous and free from inclusions. The growth rings/lines separating one growth zone from the other are very distinct in some of the porphyroblasts. Lack of suitable time-marking microstructures in the matrix makes it difficult to establish the times of development of these different growth zones. Most possibly the outermost homogeneous overgrowth took place in a relatively quiescent period that was followed by the main stretching event. Note that there are some distinct examples of pulling apart of these euhedral porphyroblasts; this pulling apart is probably related to the main stretching event.

### Description of porphyroblast classes II(a) & II(b) occurring together in slide - 7G'87:

Note how contrasting the textural patterns are in two mutually adjacent garnet porphyroblasts (Fig. 5.60). One (class II(a)) shows a very irregular outline with lots of visible inclusions arranged in splayed trails, while the other (class II(b)) has well developed crystal outline and a fairly homogeneous body with only a few inclusions. Closer observation, however, suggests that in addition to some visible quartz and opaque inclusions there are innumerable fine dusty inclusions of opaque and/or graphite in the latter homogeneous euhedral porphyroblast. These fine dusty inclusions define a pattern crudely resembling the picture of a 'rotating nebula'. The side by side occurrence of these two porphyroblasts, with widely different textures, is difficult to explain. They could mean either

- (a) simultaneous crystallisation, but developed at different points in a partitioned deformation space, or

- (b) growth (nucleation) at two different points of time. The complex inclusion trail-bearing irregular-outlined garnet grain might have started to grow earlier than the homogeneous one.

It is difficult to resolve clearly which of the above two possibilities is correct. There is clear textural zonation in both the garnets; broadly three zones in the meshy garnet, while in the euhedral garnet two distinct zones are present. The latest growth zone in the meshy garnet is represented by the somewhat isolated curved rectangular patch (just by the fracture running across the thin section) with a few elongate inclusions within it defining an Si more or less parallel and continuous with the Se. The latest growth zone in the homogeneous garnet is completely free of inclusions. Note that the latest growth zone in the euhedral porphyroblast is thicker in the direction of stretching than in the direction perpendicular to it. Keeping in mind that the thin section is roughly an XZ section, we can recognise that the distinct ellipticity in the overall outline of both the porphyroblasts is parallel to the stretching direction. All these features suggest that the latest growth zones in both the porphyroblasts developed simultaneously with each other and were broadly synchronous with the stretching/shearing event. Probably both the porphyroblasts involved rotation at some stage during their early development. These evidences indicate that both of the forementioned possibilities are correct in part and probably the meshy garnet nucleated earlier.

### Class III porphyroblasts

[Examples from slide - 64A'87]: Two examples are illustrated, one partially and the other totally altered. A spectacular example where the garnet porphyroblast has undergone partial retrogression (chloritisation) along the rim and the body-fractures is shown in Figs. 5.61a & b. The degree of chloritisation is more prominently observed under the microscope between cross polars (see Fig. 5.61b above). Note a rim of white mica defining a 'corona' surrounding the altered garnet porphyroblast (cf. Fig. 5.54 from slide - NR 39/40A described earlier). A few irregularly distributed opaque inclusions are present within the porphyroblast. Fig. 5.62 is probably an example of a wholly chloritised garnet porphyroblast (chlorite pseudomorph after garnet). There is radial arrangement of chlorite grains within the pseudomorph.

[Example from slide - 7C'87]: Fig. 5.63 shows a class III porphyroblast. The porphyroblast appears to be partly skeletal and does not show significant inclusion features. A notable point is that the porphyroblast with all its arms and branches

occurs as 'island/s in a lake' or chlorite; this indicates an advanced stage of retrogression of the porphyroblast.

The above description of various types of garnet porphyroblasts from different parts of the Munsiri Formation gives ample indication that certainly all these porphyroblast types did not develop at the same time. Now, some crucial evidence from two thin sections - NR 58/59 and 7M'87 will be discussed which will clarify and confirm the timing of garnet porphyroblastesis in the Munsiri Formation.

#### Evidence from slide - NR 58/59

As already discussed in detail in subsection-5.3.5 when exploring the time of formation of staurolite in the Munsiri Formation, there are mainly two kinds of porphyroblasts in the thin section (NR 58/59) - (i) spongy garnets, and (ii) homogeneous garnets. There are also examples where in the same porphyroblast there are spongy core and homogeneous rim (see Fig. 5.30). One of the spongy garnet porphyroblasts contains spectacular inclusions of staurolite crystals (see Fig. 5.25) that formed at pre-F2, syn- to post-F1 time. Obviously, therefore, the spongy garnets developed after this period of the main high grade Barrovian metamorphism in the area. The matrix foliation in the rock is defined by alignment of micas and quartz along with trails of opaque minerals. It is found that the opaque trails representing the matrix foliation (Se) continue undeflected through the homogeneous garnet porphyroblasts, whereas they swerve round the spongy garnets. This observation coupled with the fact mentioned above that in some porphyroblasts a spongy core is surrounded by a homogeneous rim unequivocally establishes that the spongy garnets are earlier than the homogeneous garnets. Also the evidence of indentation or pushing aside of the matrix foliation by the pointed edge of a spongy garnet porphyroblast (see Fig. 5.29) implies that the spongy garnet was already in existence when the present external foliation developed. The above relations of matrix foliation with spongy garnet and homogeneous garnet suggest that the matrix foliation developed through a phase of strong foliation transposition postdating the formation of the spongy garnets, and that the homogeneous garnets grew more or less concomitantly with the foliation transposition. It is established that the strong foliation transposition that led to the development of the present main i.e. matrix foliation in the rock took place at the time of the main shearing or stretching event. Therefore, the homogeneous garnet porphyroblasts are syn- to post-stretching. Fig. 5.31 (see subsection - 5.3.5) shows a gently folded spongy garnet which implies that a fold episode took place either synchronously or following the development of the

spongy garnets. As argued in subsection - 5.3.5, the spongy garnets (including the one enclosing staurolite crystals) grew broadly synchronously with an episode of folding (most possibly, F2).

Thus the thin section NR 58/59 supplies evidence for two definite generations of garnet:

- (i) Syn-F2, and
- (ii) Syn- to post-stretching.

As a corollary it also hints at a very likely possibility of having at least another generation of garnet in the area that developed synchronously with the staurolites as a part of the early main Barrovian metamorphic sequence at pre-F2, but syn- to post-F1 time.

#### Evidence from slide - 7M'87

The thin section is, in fact, an F2 profile section (Fig. 5.64). In the broad fold hinge zone there are two main subsidiary hinges. The fold hinge is clearly defined by two broad bands of different composition - the outer band (arc) is quartz-rich, while the inner band (arc) is rich in mica and garnet. Garnet porphyroblasts occur in varying modes in the two broad bands. Within the outer quartz-rich band, thinner sub-bands are recognised viz. coarse quartz-rich sub-band, fine quartz-rich sub-band, mica-opaque-garnet-quartz-bearing sub-band and apatite-garnet-quartz-bearing sub-band. Within the inner mica-rich band pronounced axial planar crenulation foliation renders further subdivision into sub-bands difficult. The axial planar fabric is much less pronounced in the outer quartz-rich zone. As already indicated, garnet porphyroblasts are much more common in the inner mica-rich band (see Fig. 5.64).

(Garnets in the inner mica-rich zone)

Fig. 5.65a shows a garnet porphyroblast with at least two or, more probably, three growth stages. First growth stage indicated by the sigmoidal inclusion trail-bearing part is followed by possibly two successive stages of more or less homogeneous garnet growth. Precisely speaking, however, these apparently homogeneous later overgrowths contain boundary-parallel linear trails of dusty inclusions. An adjacent porphyroblast (not shown in figure) preserves only the earlier S-shaped inclusion trail-bearing stage. There is a striking similarity in the appearance of dusty inclusion trails in garnet porphyroblasts with those in adjacent white mica aggregates. This probably implies that garnets overgrew earlier micas. While micas have been chemically transformed into garnet, fine/dusty opaque or graphite

inclusions remained unconsumed. This is supported by the fact that the trails of dusty inclusions are found mainly in the overgrowths of the garnet porphyroblasts and that there is angular conformity between these inclusion trails (Si) and adjoining matrix foliation (Se).

Fig. 5.65b shows an elongate garnet porphyroblast aligned parallel to the axial planar foliation. The porphyroblast has broadly three segments and records at least two growth stages; the middle segment with distinct inclusion trails represents the early growth stage (I), while the two segments on either side of it represent a later growth stage (II). The porphyroblast elongation parallel to axial planar foliation here probably reflects favoured growth in the direction of the cleavage rather than rotation and realignment of an original elongate grain. This possibly indicates that the later growth stage of the garnet was synchronous with the development of the axial planar foliation.

Fig. 5.65c shows a garnet porphyroblast whose relations with the enclosing foliation probably suggest a broadly syntectonic development of the porphyroblast in respect of the enclosing axial planar fabric. The matrix foliation clearly is not indented by the garnet porphyroblast.

An oval porphyroblast of garnet with its elongation across the matrix foliation (axial planar foliation) is shown in Fig. 5.65d. This is an ideal situation where one could expect 'apathetic' indentation or pushing aside of the enclosed foliation, if the porphyroblast were definitely pre-tectonic. But what we see is, in fact, a simple 'sympathetic' enveloping of the porphyroblast by the mica grains of the matrix. This feature probably suggests that the porphyroblast is broadly syntectonic with respect to the matrix foliation. Also note the parallelism between the Si and Se.

None of the garnet porphyroblasts in the mica-rich inner band (arc) shows strong indentation or bending of the enclosing matrix foliation (F2 axial planar foliation) - a fact that suggests a broadly concomitant development of these garnets (particularly their later overgrowths) and the axial planar foliation. Had these porphyroblasts been definitely earlier than the enclosing foliation, then we would have expected to see pronounced indentation of the foliation adjacent to the projected edges of the porphyroblasts. Therefore, the later growth stage/s of the garnet porphyroblasts in question are synchronous with the F2 folding. The multistage growth of the garnet porphyroblasts indicate that even though the final growth stages may be broadly synchronous with the formation of the F2 axial planar foliation, the

early growth stages may be earlier than the foliation. It is not unlikely that the earliest growth stages of these porphyroblasts may correspond to the pre-F2, but syn- to post-F1 main Barrovian metamorphism. See below for discussion on the other garnets in the thin section.

(Other garnets in the thin section)

Additional information is provided by two garnet porphyroblasts occurring in the outer quartz-rich band insofar as the determination of the time of garnet porphyroblastesis is concerned (see Fig. 5.64).

One long and roughly prismatic garnet porphyroblast occurs in the mica-opaque-quartz-rich sub-band mentioned earlier. The occurrence of this porphyroblast makes the sub-band look prominent under the microscope. The sub-band represents the early foliation that is folded by the F2 folding. On closer observation, using higher power objective, the porphyroblast is found to show gentle folding/kinking that corresponds to lower order F2 folds (see Fig. 5.66). Therefore, this garnet porphyroblast is pre-F2 in origin. Now there are two distinct possibilities - either the garnet is a part of the main Barrovian sequence developed at pre-F2, but syn- to post-F1 time, or, it is still earlier possibly corresponding to the metamorphism when the earliest foliation was formed at pre- to syn-F1 time.

The close surrounding of the garnet porphyroblast and the general appearance (rugged look) of the porphyroblast itself have considerable similarity with those associated with some staurolite porphyroblasts of the Munsiri Formation; also a somewhat similar kinking behaviour has been noticed in a staurolite prism of the Joshimath Gneiss (cf. Fig. 5.23). Thus it is quite probable that this garnet porphyroblast developed at the same time as the staurolite porphyroblasts in the area as a part of the early main Barrovian metamorphic sequence at pre-F2, syn- to post-F1 time.

But the alternative possibility seems to be more likely, because the elongate porphyroblast lies along, and thereby helps in defining the early foliation that corresponds to the earliest recognisable metamorphic differentiation in the rock (see Fig. 5.64); so, most probably, the garnet porphyroblast developed at the time of the first metamorphic fabric formation i.e. at pre- to syn-F1 time.

The other porphyroblast in the outer quartz-rich band is a typical skeletal garnet (Fig. 5.67). The skeletal nature is almost certainly due to local paucity of

garnet-forming reactants, given that the porphyroblast is developed in the band rich in coarse-grained quartz. In some respects, there are similarities between this skeletal garnet and the elongate garnet mentioned above. For example, the elongate garnet also shows partly skeletal character at one extremity of the grain. An imaginary line circumscribing this skeletal garnet would approximately define an ellipse whose long axis would parallel the compositional banding which, in turn, defines the F2 fold hinge/s; this implies a pre-F2 time of development of the skeletal garnet. These features may induce one to consider that both the garnets belong to the same generation (i.e. pre- to syn-F1 in age). But closer observation suggests that, in fact, the skeletal garnet, having a wide elliptical overall shape, overgrew the fine-scale earliest foliation present in the outer quartz-rich band (see Fig. 5.64) which means the porphyroblast developed later than the pre- to syn-F1 time and is, therefore, younger than the elongate garnet. So the most obvious time of development of the skeletal garnet is pre-F2, but syn- to post-F1 time, corresponding to the time of the major high-grade Barrovian event in the area. The considerable retrogression (chloritisation) in the skeletal porphyroblast may support an older age, but this criterion loses much of its significance due to the involvement of a fracture running right across the porphyroblast; the fracture must have acted as a fluid pathway which aided in chloritisation.

Thus the outer quartz-rich band in the thin section (7M'87) preserves two garnet porphyroblasts, one of which developed at pre-F2, syn- to post-F1 time as a part of the major Barrovian metamorphic sequence in the area, and the other at pre- to syn-F1 time probably during the first metamorphic differentiation; whereas the inner mica-rich band showing pronounced F2 axial planar foliation preserves evidence of additional growth of garnets at syn-F2 time.

Combining the results obtained from detailed textural observation on garnet porphyroblasts from the Munsiri Formation, we therefore come to the conclusion that there were four times of garnet porphyroblastesis viz.

- i) Pre- to syn-F1
- ii) Pre-F2, syn- to post-F1
- iii) Syn-F2
- iv) Syn- to post-stretching (broadly syn-F3).

As already highlighted, the observations on garnets in the Joshimath Gneiss also indicated the same times of garnet development.

Mainly due to difficulty in sampling and preparation of suitable thin sections, no textural study under the microscope could be undertaken on the garnets of the Berinag-Mandhali formations. On a speculative note, their time of crystallisation will be discussed in section 5.8.

### Conclusions from textural study of garnets in the Joshimath area

1. Principal growth stages of garnet porphyroblasts were pre-stretching.
2. There were garnet growth during: (a) pre- to syn-F1 time; (b) the high-grade Barrovian event at pre-F2, syn- to post-F1 time; (c) syn-F2 time, and (d) syn-F3 (i.e. syn-MCT) time.
3. However, the main phases of garnet porphyroblastesis recognised in the area are,
  - (i) Pre-F2, syn- to post-F1 (during the regional high-grade Barrovian episode)
  - (ii) Syn-F2
  - (iii) Syn- to post-stretching (broadly syn-F3 i.e. syn-MCT time).

Fig. 5.68 (fish diagram) shows the garnet-forming episodes in the MCT-Zone and above more clearly.

## 5.5 AMPHIBOLE TEXTURES

In the Joshimath area, there are amphibolite horizons in all the three units viz. Joshimath Gneiss, Munsiri and Berinag-Mandhali formations, but they are most common in the Munsiri Formation. Some of them are garnet-bearing, while others are not. The garnetiferous amphibolites are not found below the lower middle part of the Munsiri Formation; however, above this level there are numerous non-garnetiferous amphibolite horizons.

For the purpose of textural description, I intend to refer to all the individual species under the same group name 'amphibole'. The amphibole textures are described below considering the three formations together. In some amphibolite horizons particularly in the Munsiri Formation, it is difficult to determine whether the amphiboles are porphyroblastic or porphyroclastic in origin. However, most amphiboles are porphyroclastic in origin. Generally speaking, three broad textural types of amphiboles have been recognised in the area:

- 1) Massive type, mostly occurring with skeletal garnets in the Joshimath Gneiss or in the upper part of the Munsiri Formation. In this type, the individual component grains occur together without any conspicuous spatial separation among them. The overall outline of these massive type amphiboles is irregular.
- 2) Discrete amphiboles. These are in the form of discrete platy or tabular crystals, or in many cases, appear as elongate prismatic grains or as diamond shaped grains. This type of amphibole is rather rare in the Joshimath Gneiss, but is relatively common in the Munsiri Formation and in the Berinag-Mandhali together with other textural types of amphiboles. Usually these discrete amphibole grains have rather irregular outlines and reveal sieve texture, incorporating many inclusions.
- 3) Small sized elliptical, or elongate diamond-shaped, lozenge-shaped or slender prismatic grains of amphibole. These occur close to each other and are restricted to specific bands. They give rise to distinct amphibole-rich bands. Often they are found clustered along <sup>!following the trace of main foliation.</sup> linear trails. This type of amphibole is found in the Munsiri Formation, especially in its middle and lower middle parts.

Systematic observation suggests that the frequency of occurrence of the type-3 amphiboles increases down-section in the Munsiri Formation and this is probably a reflection of increasing stretching/shearing in that direction. In other words, substantial proportion of type-3 amphiboles are recognised to have been derived from class-1 and/or class-2 ones.

Given below is a description of amphibole textures from a few selected thin sections--

- (a) Joshimath Gneiss, Slide-NR25/26L(2): Amphibole grains occur mainly in clusters (Fig. 5.69a); only a few occur as isolated crystals. The grains in the clusters are crudely diamond-shaped and thus show subhedral to euhedral outlines. The

predominant type of grains has a maximum length of about 1.27 mm and breadth or width of about 0.78 mm. The extinction of the amphibole grains is smooth implying that they have very little internal strain. Inclusions are not common and are mostly quartz with rounded circular or elliptical shapes. There are some instances where along some fractures, the amphibole grains have altered to biotite. No folds or strong foliation are observed. So the time of development of these amphiboles is difficult to determine. The other mineral in the thin section comparable in proportion to the amphibole is clinopyroxene (augite) (see Fig. 5.69a). Many of the clinopyroxene grains show incipient alteration into amphibole which is apparently looking like inclusions (Fig. 5.69b).

(b) Munsiri, Slide-15/45G: There are broadly two types of amphibole grains in this rock— one, the relatively larger porphyroclastic grains, and the other, the smaller matrix grains. Along with biotite, clinzoisite, quartz and plagioclase the small amphibole grains define the matrix foliation that swerves around the porphyroclastic amphibole grains or aggregates of them (Fig. 5.70). In many instances the porphyroclastic grains show single dominant cleavage; the cleavage is oblique (at  $\sim 30^\circ$ ) to the main foliation. While most of the porphyroclastic grains are tabular, platy or ovoid in shape, the matrix amphibole grains are rather irregular in shape and size. There are diamond or lozenge shapes, elongate prismatic shapes, elliptical shapes as well as nonspecific shapes. The porphyroclastic grains are normally in the range of 1mm or more in length and 0.5mm or more across. Obviously the porphyroclastic amphiboles have formed earlier than the present main foliation and therefore pre-stretching in origin. While many of the matrix amphibole grains may belong to the same generation as the porphyroclasts, some of them may have formed synchronously with the stretching event. There are evidences of retrogression of amphibole grains into biotite; also many of the amphibole grains show undulose extinction implying internal strain. These features correlate well with the pre-shearing origin of majority of the amphibole grains in this rock. However, it is difficult to determine the time of formation of these amphiboles in terms of the fold episodes. No garnet is seen in this rock.

(c) Munsiri, Slide-8C'87: The thin section shows a somewhat hornfelsic-looking texture particularly in the feldspar-rich portions adjacent to amphibole-rich band. In the field, it was presumed that the amphibolite horizon had an intrusive origin and due to the heat supplied by the intrusive sill, the adjacent parts of the country rock suffered selective melting or homogenisation; such features are more clearly indicated by the feldspathic portions than the quartz-rich portions in the country rock.

Twins in the plagioclase grains have become diffused and irregular, and are, therefore, difficult to recognise. On close observation such homogenisation of plagioclase grains is found to decrease in intensity away from the amphibole-rich band. Note that though tiny biotite flakes are common in the country rock, most of the larger biotite grains occur just in the immediate contact with the amphibole band. There is also evidence of chloritisation of some of these large biotites as well as nearby amphibole grains. A relatively late quartzose vein has intruded along the contact zone for some distance. Also, though garnet occurs both in the country rock and in the amphibole-rich band, there is a strong contrast in the textures of the garnets between the two areas of occurrence. The garnet in the amphibole-rich band is large in size and distinctly skeletal in nature with very irregular and indistinctive outline, while garnets in the country rock are large in number and occur as discrete small grains generally with inclusion-rich inner parts surrounded by more homogeneous outer parts.

The amphiboles here mostly occur as dense mass consisting of many grains in close-knit clusters (Fig. 5.71). The foliation is very poorly defined, if at all, in the amphibole-rich band. Occasionally, several prismatic amphibole crystals showing lamellar twinning occur in diverse orientation defining a star-like pattern (Fig. 5.72). These elongate prismatic grains are normally about 1 mm long and 0.2 mm wide. The grains forming the clusters are of various sizes and shapes, generally irregular in outline; the largest among them are somewhat platy in character with lengths of about 1.65 mm and breadths 1 mm. The grains intermediate in size are broadly diamond-shaped with maximum dimensions 0.84mm/0.36 mm. Inclusions of quartz, opaques, clinozoisite, sphene, biotite and/or calcite are common, particularly in the small to intermediate sized amphibole grains. In a few instances, biotitisation and/or chloritisation of amphibole is noticed (Fig. 5.73). Association with calcite is also present. Extinction of the amphibole grain is considerably undulose; in cases, even the individual cleavage rhombs/diamonds in the same grain vary in extinction position. This implies that there was postcrystalline internal strain upon the amphibole grains. The general distribution of the amphibole grains in the rock probably suggests an intrusive origin of the amphibolite possibly early synchronously with the main shearing event.

(d) Munsiri, Slide-17Rb: The thin section is a YZ section from a highly retrograde amphibolite horizon. Biotitisation has affected most of the amphibole grains along two directions of their cleavage so pervasively that the original amphibole grains now resemble a picture of the grain-bearing part of a stalk of wheat or paddy, the small

individual cleavage rhombs standing for the wheat or rice grain itself (Fig. 5.74a). The rice-shaped individual amphibole grains (cleavage rhombs) are very tiny in size, measuring only 0.06/0.03 mm. But the elliptical (crude diamond shaped) outline of the original (host) amphibole grain is still retained in most cases and their dimensions measure normally in the range of 0.66 /0.33 mm. Occasionally some homogeneous looking, fairly large (0.70/0.36 mm) platy or elongate prismatic amphibole grains are still preserved recording only incipient biotitisation and recrystallisation. A few of these grains show twinning as well. It is clear that the alteration of amphiboles contributed significantly to the present content of biotite in the rock. Also biotites are being chloritised. In fact, the trace of main foliation in the rock is most clearly marked by the biotite-chlorite-rich semi-continuous bands. It is quite certain that originally the amphibole grains were pre-shearing or pre-stretching, but their recrystallisation (development of rice-shaped granules) and present disposition were largely syn-stretching. No garnet is seen in the rock.

(v) Munsiri, Slide-Loc.18(2)'87: This thin section shows a fold profile and has been used for fold class determination using thickness variation and isogon method in Chapter 3. In all probability this is a post-F2 fold, either F3 or, more possibly, F4, because the mylonitic/protomylonitic foliation is distinctly folded. Three main amphibole bearing bands (I, II & III) can be distinguished in the thin section depending on the size and shape of the amphibole grains and the presence or absence of biotite as a coexisting phase with amphibole (see Fig. 5.74b). The innermost and outermost bands have no biotite at all; while the intermediate band has biotite almost in equal proportion with amphibole. Quartz, plagioclase and opaques are present in all the bands, but with differing grain sizes. Generally speaking, the amphibole grain-size decreases from core of the fold outward, so the innermost band (I) has the most stumpy and stout grains measuring upto 0.97/0.60 mm or 1.12/0.18 mm, while the intermediate band (II) has elongate prismatic amphibole grains normally 0.84/0.09 mm dimension or elongate diamond-shaped grain with maximum dimensions normally around 0.33/0.15 mm (even within this intermediate band, the amphibole-size decreases outward), and the outermost band (III) has the tiniest amphibole grains, slender prismatic in shape normally measuring 0.24/0.03 mm. Tiny lozenge-shaped grains with maximum dimensions 0.07/0.04 mm are also common in this outer band. Except for roughly at the middle level of the biotite-bearing intermediate band (2) where whole grains of biotite are found to be chloritised along with the presence of some calcite, the rock shows very fresh-looking amphibole, biotite and quartz-feldspar grains. Though the fold is very clearly defined by the compositionally and texturally different bands, no distinct example has been found where an individual

amphibole grain is folded; in most cases the amphibole grains are simply reoriented in the exact hinge zone. However these features amply suggest that amphiboles formed prior to the fold episode. The mylonitic fabric in the rock is very clearly visible particularly in the outermost band, and this mylonitic foliation is folded. The strong preferred orientation of amphibole grains along the foliation, their homogeneous extinction, (i.e. without any conspicuous internal strain) and the lack of any conspicuous retrogression/alteration of these amphibole grains suggest that possibly these amphiboles developed concomitant with the shearing or stretching and do not represent pre-stretching 'old' grains.

The foregoing description of amphibole textures from a selected number of thin sections clearly indicates that there were mainly two stages of amphibole growth/recrystallisation in the area -- (a) before stretching/shearing, and (b) synchronously with the stretching/shearing event. Clearly the former stage was much more dominant than the latter one. Though, generally speaking, it is difficult to identify the exact time of formation of the pre-stretching amphiboles in terms of deformation episodes, it is likely that there were more than one generation of them.

## 5.6 TEXTURES OF SOME OTHER MINERALS

Here 'other minerals' mainly refer to the matrix phases. Textures of the matrix minerals viz. quartz, feldspars, micas etc. have not been studied in detail. It was mainly in connection with the study of the high grade minerals and garnets that the textures of 'other associated minerals' were also investigated. However, while carrying out some petrofabric analyses with quartz and calcite, textures of these two minerals were studied carefully.

### Quartz

Timing of significant matrix recrystallisation (quartz, in particular) has been indicated in discussions on textures of kyanite, staurolite and garnet in sections 5.3 and 5.4. The most clearly recognisable matrix coarsening event in the hanging-wall block, particularly within the MCT-Zone, took place late synchronously with the main stretching event. This time is very crucial in understanding the tectonic significance of the quartz and calcite petrofabrics.

The textures of quartz in the footwall block, particularly in the kyanite-bearing quartzite horizon deserve special mention. Three thin sections (32A||L, 32A⊥L and F.W.Kya'89) respectively of XZ, YZ and XY orientation from the kyanite quartzite horizon were studied in detail. Of these, the section - 32A||L reveals most interesting quartz textures.

This section (32A||L) shows that there are three groups of quartz grains in the kyanite quartzite of the Berinag-Mandhali formations -- matrix quartz, porphyroclastic quartz and vein/band quartz. The vein/band quartz grains are confined to a band that runs concordant to the main foliation through one side of the thin section. The generally larger grain-size (compared to matrix) and fairly uniform size distribution of quartz in the band and its low mica content indicates that it was originally a concordant quartz vein which intruded the rock at an early stage of stretching. While the porphyroclastic quartz grains are highly deformed in appearance showing heavily undulose extinction with deformation lamellae and embayed outlines, the matrix quartz and the vein/band quartz grains do not show strong undulose extinction. The two latter groups show fairly annealed texture with examples of grain-contact triple junctions (Fig. 5.75). The porphyroclastic quartz grains are the largest in size, while the matrix grains are the smallest. There are clear evidence of recrystallisation of small strain-free quartz grains out of the porphyroclastic quartz grains. It is likely that most of the matrix quartz are products of recrystallisation from such larger porphyroclasts which, in turn, might represent remnants of original clastic grains.

The crystallographic preferred orientation of the three groups of quartz grains are different from one another. Using an accessory sensitive tint plate and taking the main foliation perpendicular to the accessory slot so as to avoid the interference by mica-colours, a dominance of red colour is observed among the matrix quartz grains and that of blue-green among the vein/band quartz grains; while different porphyroclastic grains show different colours viz. green, yellow, blue, red, bluish red, reddish yellow etc. This confirms that the matrix and the vein/band quartz grains are recrystallised and have acquired significant crystallographic preferred orientation through crystal-plastic processes; whereas the haphazard crystallographic orientation of the porphyroclastic quartz grains is a further pointer to their primary clastic inheritance.

An exercise of grain-size analysis from the three mutually perpendicular thin sections (i.e. 32A||L, 32A⊥L and F.W.Kya'89) proved to be important. Table-5.3 gives the analytical data.

**Table - 5.3**

Grain-size data from kyanite-bearing quartzite of the Berinag-Mandhali formations

	- ( XZ ) - <u>Slide-32A  L</u>	- ( YZ ) - <u>Slide-32A⊥L</u>	- ( XY ) - <u>Slide-F.W.Kya'89</u>
Average matrix quartz grainsize	0.134mm/0.09mm	0.115mm/0.10mm	0.4mm/0.2mm
(Aspect ratio)	1.49 : 1	1.15 : 1	1.8 : 1
Average vein/band quartz grainsize	0.25mm/0.16mm	0.28mm/0.21mm	... ..
(Aspect ratio)	1.56 : 1	1.33 : 1	... ..
Av. porphyroclastic quartz grainsize	2.7mm/0.78mm	2.99mm/0.87mm	... ..
(Aspect ratio)	3.46 : 1	3.44 : 1	... ..
Larger muscovite size	1.12mm/0.076mm	0.55mm/0.036mm	
(Aspect ratio)	14.7 : 1	15.3 : 1	... ..
Tiny muscovite size	0.05mm/0.01mm	0.04mm/0.01mm	... ..
(Aspect ratio)	5 : 1	4 : 1	... ..

It is interesting to note that, as we would expect, the aspect ratios are almost as a rule more in the XZ-section than in the YZ-section implying a triaxial prolate shape of the grains with long axes parallel to the stretching lineation; but the porphyroclastic quartz grains are a major exception, and they have similar aspect ratios in both XZ and YZ sections implying oblate ellipsoidal grain shapes. This could mean that the porphyroclastic quartz grains responded to stretching or shearing more by recrystallising themselves than by being bodily stretched. Hence it is quite likely that their present oblate shape is largely inherited. These features also suggest that the porphyroclastic quartz grains are the remnants of early, detrital clastic grains.

The vein/band quartz grains in thin section - 32A||L show an well developed shape fabric. Elliptical in outline, these grains are arranged in such a way as to indicate an imbricate pattern oblique (at  $-40^\circ$ ) to the main foliation. This sort of shape fabric is a good indicator of shear sense -- here 'top-to-south', which is consistent with the regional pattern. But the matrix quartz grains do not show similar shape fabric. Bound within fine micaceous laminae, the matrix quartz grains are mostly elongate parallel to the main foliation (Fig. 5.76). This pattern represents growth under the inhibitory influence of a second phase (here, muscovite) (Hobbs et al., 1976, pp. 113-118).

The still remaining, now heavily strained and embayed bodies of the clastic quartz grains in the kyanite-bearing quartzite point to the possibility of quartz recrystallisation at different stages within the Berinag-Mandhali footwall block.

### Phyllosilicates

Among the phyllosilicates, muscovite is generally more ubiquitous than biotite in the study areas. Texturally speaking, both the minerals have close similarities. Their proportion varies from horizon to horizon; this is most likely an indication of a distinct control of primary lithology upon the development of metamorphic minerals in the area. A variation in phyllosilicate grain shapes and sizes is also noticed. This has been found to be partly due to the presence of multiple generation of the minerals. Strong preferred orientation of the phyllosilicates following the main foliation and, to a considerable extent, the dominant stretching lineation is noticed; the aberrations, where present, are generally shown by the relatively larger-sized grains. Grain shapes of phyllosilicates are quite varied, but four general classes could be recognised -- (i) large, elongate ( $>2\text{cm}$  long), (ii) rectangular (razor-blade shaped),  $\sim 1\text{cm}$  long, (iii) tiny flakes, and (iv) irregularly outlined. As a mineral phase the phyllosilicates existed through  $F_1$ ,  $F_2$ ,  $F_3$  &  $F_4$  fold episodes in the MCT-Zone and its hanging wall as indicated by 'fold-core and foliation development' studies. Their presence is noticed from the earliest to the latest recognisable foliations. But the dominant neomineralisation stages were at pre- or syn- $F_1$ , syn- $F_2$  and syn- $F_3$  (syn-stretching) times.

### Feldspars

In Joshimath area, feldspars in quartzofeldspathic horizons in the upper and middle parts of the Munsiari Formation show deformational textures more commonly in the brittle-semibrittle mode than in the lower parts (Fig. 5.77). Textures implying

crystal-plastic deformation in feldspars are more common in the quartzofeldspathic rocks in the lower-middle and lower part of the Munsiri Formation. As highlighted earlier in Chapter-2, the mylonitic-protomylonitic horizons in Joshimath area are characterised not so much by the presence of streaked out feldspar and/or quartz grains or quartzofeldspathic aggregates as by the crystal-plastic 'discrete' reduction of grain sizes, both porphyroclastic and matrix. A major mechanism of such reduction in grain size is found to be through symplectitisation (or myrmekitisation) of plagioclase feldspars (Fig. 5.78 a & b).

The timing of metamorphic crystallisation or recrystallisation of the dominant matrix phases has been determined either by direct means using similar criteria and procedure as have been applied for the high-grade or other minerals discussed earlier, or by indirect means through recognising stable paragenetic relations with minerals of known age. Further discussion on this aspect will be given in section 5.8.

## 5.7 RETROGRESSION FEATURES

Mineralogical changes characterising the retrograde metamorphism in the study areas include: (i) biotite → chlorite, (ii) garnet → biotite/chlorite, (iii) amphibolite → biotite, (iv) plagioclase → saussurite, (v) alkali feldspar → sericite/muscovite etc. The last feature is most clearly observable in some augen phyllonite horizons in the Munsiri Formations. For general discussion on various aspects of retrogression of alkali feldspar into white micas see McCaig (1988a & b).

The main retrogressive episode of metamorphism was late syn- to post- main stretching event, but before the second/late stretching event. However, retrogression associated with the very late structures such as with the high angle faults, fractures and localised shear zones is a much later phenomenon. Nowhere have I found any chlorite grain that has been pulled apart or stretched by the main stretching event. Evidently chloritisation of the garnet grains or amphibole grains showing pull-apart features occurred after the latter had been stretched. In several instances, clear examples of pseudomorphous alteration of garnet and biotite into chlorite are found (Figs. 5.79 & 5.80; also see Fig. 5.62).

That the later stretching event took place post-dating the retrograde metamorphism is evident from the fact that the later stretching lineation is defined on

chloritised foliation surfaces. Chloritisation is a hall-mark of regional retrogressive metamorphism in the study areas.

Chlorite, sericite etc. are much more common in the lower part of the Munsiri and the upper part of the Berinag-Mandhali formations than elsewhere. However, sporadic occurrences of chlorite either partially or completely replacing discrete grains of biotite, garnet and/or amphibole are found even in the upper part of the Munsiri and the Joshimath Gneiss.

No evidence has been found of chlorite as a discrete mineral phase developed prior to or synchronously with the F<sub>2</sub>-folding. The relatively rare occurrences of chlorite in the form of inclusions in the garnet porphyroblasts are most probably not of primary origin. As we see them in thin section, most of these chlorite inclusions are surrounded on all sides by the garnet. But it is very hard to tell whether or not these inclusions are connected with some body fractures in the garnet porphyroblasts in the third dimension that are connected with any fluid pathways. If they are, then such inclusions would clearly represent secondary chlorites most probably replacing biotites. There are evidences where partial chloritisation of garnet porphyroblasts and/or complete or partial chloritisation of neighbouring biotite grains are found to have been induced by fluids invading through adjoining fractures. Thus it is indeed very likely that the few chlorite inclusions that are found in some garnet porphyroblasts are secondary chlorites, a product of retrogressive alteration of original biotite inclusions.

The main retrogressive metamorphism took over from the peak of stretching or thrusting along the MCT and probably took place, tectonically speaking, in relatively quiet condition without any conspicuous deformation (folding) accompanying it.

## 5.8 METAMORPHIC HISTORY

*relativit...*

In this concluding section of Chapter-5, at first the collective paragenetic sequence of minerals (synoptic fish diagram) is presented taking the hanging wall (Joshimath Gneiss and Munsiri Formation considered together) and the footwall (Berinag-Mandhali formations) separately. This will facilitate the establishment of the metamorphic history of the rocks across the MCT-Zone. Determination of the time and nature of the major metamorphic episodes would effectively help in

clarifying the tectonometamorphic relations within and among the three tectonostratigraphic units and give an approximate idea of the range of physical conditions operating during the deformation events of which MCT-thrusting was the most important one. The section will close with a note on the inverted metamorphism.

### 5.8.1 Mineral paragenetic sequence in the hanging wall and the footwall

Fig. 5.81a gives in a combined list the mineral paragenetic sequence in the Joshimath Gneiss and the Munsiri Formation; the corresponding list for the Berinag-Mandhali formations is given in Fig. 5.81b. These two lists embody the results of the detailed textural studies that have been carried out. However, because the main objective in carrying out the textural work has been to gather information so as to be able to recognise how many metamorphic events have occurred in the area, the two above-noted paragenetic diagrams by no means include all the minerals from each type of lithology present in the area. Emphasis has been given for working out the paragenetic relations of the Barrovian index minerals along with a selected number of other minerals mainly from pelitic-semipelitic, quartzofeldspathic and basic lithologies. Generally speaking, the choice was confined to those minerals which, on account of their inherent genetic significance and/or critical location in thin sections with time-marking fabrics, have been considered to be potential source of crucial tectonometamorphic information. Details of some other minerals are relegated to Appendix II and that for the few remaining have not been included in discussions.

The reason why the mineral paragenetic relations in the Joshimath Gneiss and the Munsiri Formation are shown in a combined list (Fig. 5.81a) is to emphasize the fact that there is a remarkable correspondence between the two units in terms of the sequence of metamorphic mineral formation. Except the sillimanite, kyanite, scapolite and clinopyroxene which have been found to occur in the Joshimath Gneiss only, all the other minerals shown on the list occur in both the units and have developed in equal number of generations in both. However, there was difference in the relative importance of the different generations of minerals between the two units.

Fig. 5.81a clearly shows that most of the high-grade Barrovian minerals in the hanging wall block grew in one generation, more or less, intermediate in time between the  $F_1$  and  $F_2$  fold episodes, with the exception of kyanite which had an additional weaker generation at syn- $F_3$  (syn-MCT) time; the second generation contact metamorphic sillimanite reported from the Higher Himalayas has not been

encountered in the present study, so its position is only approximately shown for reference after assessing the information provided in literature. Garnet grew in multiple generations; in addition to a generation corresponding to the Barrovian sequence there were three other times of garnet development. The earliest garnets grew at pre- to syn-F<sub>1</sub> time, the latest ones at the period of main shearing (syn-F<sub>3</sub>) and there was another generation at the time of F<sub>2</sub> folding. Taking into account both progressive and retrogressive origins, micas (biotite and muscovite) have been found to be the most readily crystallisable of all the enlisted minerals; they formed during each of the identifiable metamorphic mineral-forming events. The micas of retrogressive origin developed mainly at a late stage postdating the F<sub>4</sub> folding, broadly contemporaneously with one generation each of chlorite, sericite, calcite and epidote-clinzoisite. Some chlorite and sericite developed during the syn-F<sub>3</sub> (i.e. main stretching) time as well, when epidote-clinzoisite also formed concomitant with the recrystallisation of calcite, quartz, feldspars and amphibole. There was another generation of epidote-clinzoisite and calcite at the time of the main Barrovian event (pre-F<sub>2</sub>, but syn- to post-F<sub>1</sub> time). During this Barrovian event, some quartz, feldspars (both k-feldspar and plagioclase), clinopyroxene and amphibole (hornblende) also developed, each of them in addition to having a still earlier generation at pre- to syn-F<sub>1</sub> time. Scapolite grew in Joshimath Gneiss synchronously with the F<sub>2</sub> folding. Fig. 5.81a shows that in the hanging wall block, there were four (4) main metamorphic events, in chronological order, respectively at: (i) pre- to syn-F<sub>1</sub> time, (ii) intermediate between F<sub>1</sub> and F<sub>2</sub>, (iii) syn-F<sub>3</sub> (syn-MCT) time, and (iv) post-F<sub>4</sub> time. Metamorphic crystallisation during the F<sub>2</sub>-folding was weak and not comparable in intensity to any of the above four episodes and, therefore, it is kept out of consideration. For convenience, the four main metamorphic episodes recognised from the hanging wall block are termed - M<sub>1</sub>, M<sub>2</sub>, M<sub>3</sub> & M<sub>4</sub> respectively.

The mineral paragenetic relations in the Berinag-Mandhali footwall block have been determined following similar procedure and criteria as in the hanging wall block; the only exception was garnet for which, in the absence of suitable thin sections, the possible time of growth has been worked out using observations made in the field and employing logic based upon a comparison with the footwall kyanite occurrence. In the thesis, the textual discussion of the textures of footwall minerals has been kept to a minimum in order to give prominence to the discussion of textural features of the hanging wall minerals.

Fig. 5.81b indicates that before any recognisable fold episode took place, there was crystallisation of calcite, quartz, feldspars, chlorite, sericite, some biotite and muscovite in the Berinag-Mandhali formations. Collectively speaking, there were three other times of biotite formation of which the latest was mainly through retrogressive alteration from amphiboles during the later stretching event, intermediate in time between  $F_{II}$  and  $F_{III}$ ; the other two times were respectively during  $F_I$  episode and at the interkinematic stage between  $F_I$  and  $F_{II}$ . Muscovite also grew during these two latter times. Although no distinct example of chlorite growth during  $F_I$  episode has been found, some sericite developed at this time and both sericite and chlorite developed as retrogression products at the post- $F_{II}$ , pre- $F_{III}$  stage. Recrystallisation of quartz took place during both  $F_I$  and the later stretching event (post- $F_{II}$ , pre- $F_{III}$ ). Some feldspars and calcite also grew during the  $F_I$  episode. In addition to a generation at  $F_I$  time, epidote-clinozoisite formed as alteration products at the post- $F_{II}$ , pre- $F_{III}$  time. Amphiboles grew mainly during the  $F_I$  episode; whereas tourmaline developed at two stages: one, late synchronously with  $F_I$  (corresponding broadly in time with the main stretching event in the MCT-Zone), and the other, at the time of later stretching (kyanite-stretching in the footwall) i.e. at the interkinematic stage between  $F_{II}$  and  $F_{III}$ . Kyanite developed in an isolated peraluminous quartzitic horizon in the Berinag-Mandhali formations at the interkinematic stage between  $F_I$  and  $F_{II}$ . The garnets in the thin phyllite band sandwiched between two quartzite horizons near Helang Bus Stop, only a few metres below the Munsiri Thrust, possibly grew at the same time as kyanite. Fig. 5.81b clearly shows that there were imprints of four metamorphic events in the Berinag-Mandhali footwall block. Termed  $M_I$  to  $M_{IV}$  in chronological order, these metamorphic events took place at pre- $F_I$  time ( $M_I$ ), syn- $F_I$  time ( $M_{II}$ ), intermediate in time between  $F_I$  and  $F_{II}$  ( $M_{III}$ ) and intermediate in time between  $F_{II}$  and  $F_{III}$  ( $M_{IV}$ ) respectively.

### 5.8.2 Tectonometamorphic summary

As indicated above, four episodes of metamorphism -  $M_1$ ,  $M_2$ ,  $M_3$  and  $M_4$  have been distinguished in the Joshimath Gneiss and the Munsiri Formation. However, the observable influence of these metamorphic episodes are not present with equal intensity in both the units. For instance, the effects of  $M_1$  and  $M_2$  are much more easily distinguishable in the Joshimath Gneiss, but that of  $M_3$  are more prominent than  $M_2$  or  $M_1$  in the Munsiri Formation; also the  $M_4$  episode affected the Munsiri Formation, particularly its lower part, more intensely than the Joshimath Gneiss. This is partly due to the obscuring or interfering effect of later episodes upon

the earlier ones, and partly due to the difference in the original pattern, grade and degree (or intensity) among the metamorphic episodes themselves. Two things are implied by the distribution of these metamorphic episodes: (a) Before the MCT-emplacement, whatever metamorphic episodes affected the Joshimath Gneiss had also affected much of what is now regarded as the Munsiri Formation, especially its root-zone; this is consistent with the possibility that much of the upper part of the Munsiri Formation is tectonised equivalent of Joshimath Gneiss (see Chapter 2). (b) Transport along the Vaikrita Thrust was most probably less than the generally agreed lower limit of total displacement of the MCT-sheet of 'over 100 km' (see Chapters 1 & 4) and also possibly less than that along the Munsiri Thrust. Pre-MCT metamorphic minerals ( $M_2$  ones, in particular) are found to occur on either side of the Vaikrita Thrust.

An inspection of Fig. 5.81a suggests that the grade of metamorphism varied significantly from one episode to the other in the hanging wall block. Most possibly, the  $M_1$  episode corresponds to the time when the first metamorphic foliation (gneissosity) formed or was initiated in the Joshimath Gneiss and/or Munsiri Formation.  $M_2$  corresponds to the highest grade metamorphism in the area when all the high-grade Barrovian minerals (viz. garnet, staurolite, kyanite and sillimanite) formed and defined a regional thermal structure.  $M_3$  episode corresponds to the syn-MCT-emplacement metamorphism, whereas  $M_4$  was an episode of retrogressive metamorphism.

Although the primary concern throughout the textural investigation has been to recognise how many metamorphic events have taken place and at what times relative to the deformation episodes, it offers some scope for broadly establishing the facies of the different episodes of metamorphism.

The development of garnet, muscovite, biotite, amphibole (hornblende), clinopyroxene etc. in suitable lithologies and the high degree of metamorphic differentiation necessitated by the gneissosity or first foliation formation together imply that the  $M_1$  metamorphism probably took place in the upper greenschist to lower amphibolite facies condition. The  $M_2$  episode must have required still higher temperature and pressure, because the Barrovian zones in the area, notably the four high-grade zones viz. garnet, staurolite, kyanite and sillimanite, developed at this time. There has been accompanying recrystallisation or neomineralisation of quartz, biotite, muscovite, feldspars (both k-feldspar and plagioclase), clinozoisite etc. during this episode. Most presumably, the conditions rose up to upper amphibolite facies

during the  $M_2$  metamorphism. Mylonitisation in the area (i.e. within the MCT-Zone) took place during the  $M_3$  episode. My observations suggest that the  $M_3$  event is characterised more by tectonic (mylonitic) fabric formation than by porphyroblastesis (metamorphic mineral formation). However, during this syn-MCT event, there was crystallisation and/or recrystallisation, mainly in the Munsiri Formation, of quartz, micas (both white micas and biotite), feldspars (particularly, symplectitisation of plagioclase), garnet, amphibole (some actinolitic) etc. and, interestingly, of one generation of tiny kyanite grains confined within the Joshimath Gneiss. Mylonitisation of quartzofeldspathic rocks through crystal-plastic processes normally requires a temperature not less than  $350^\circ - 400^\circ\text{C}$  (see White, 1977, 1979a & b; Tullis, 1979; Schmid, 1982). Clearly the  $M_3$  metamorphism was not a retrogressive episode, but took place probably under *lower amphibolite* facies condition. MCT-emplacement, therefore, did not take place in low-temperature retrograde conditions, but in moderately high-temperature *lower amphibolite* facies condition. Finally, the  $M_4$  episode was characterised by retrogressive transformation into minerals, such as chlorite, sericite, biotite, muscovite, epidote-clinozoisite/saussurite etc.. Therefore,  $M_4$  was the episode of retrogressive metamorphism that took place in all probability at normal pressure and temperature.

The deformation-metamorphism time relationships in the hanging wall block are clearly indicated in Fig. 5.81a. Obviously, one must not assume equal time-gap in absolute sense between the successive episodes of deformation and metamorphism. Also there was no one-to-one correlation between the metamorphic episodes and the fold episodes, except at the syn-MCT time. Being synchronous with the main stretching time (the MCT-emplacement time), the  $M_3$  and  $F_3$  episodes were mutually simultaneous. The  $M_1$  episode was either pre- or syn- $F_1$ . The  $M_2$  episode took place clearly preceding the  $F_2$ -folding, but late synchronously with or, more possibly, postdating the  $F_1$ -folding.  $M_2$  was the time of major progressive Barrovian metamorphism when the regional index mineral zones, now defining the well known "inverted metamorphic sequence", originally developed. The  $M_3$  episode was syn-MCT (syn- $F_3$ ). Whereas the retrogressive  $M_4$  metamorphism took place broadly following in time the MCT-stretching and  $F_4$  folding. Movement along the MCT-Zone was indirectly the cause of the  $M_4$  retrogressive metamorphism, which is to say that if there were no MCT movement, there would not have been any  $M_4$  metamorphism in the rocks of the Munsiri and Joshimath Gneiss formations. In other words,  $M_4$  retrogressive metamorphism was a follow-up of the MCT-emplacement at  $M_3/F_3$  time.

In the Berinag-Mandhali footwall block, the variation in metamorphic grade from one episode to the other is not as conspicuous as in the hanging wall block (cf. Figs. 5.81b & a). For obvious reasons, the temporal classification of metamorphism into different events is comparatively more subtle in the low-grade Lesser Himalayan Berinag-Mandhali formations.

In the Berinag-Mandhali formations, the first metamorphism ( $M_I$ ) refers to the event when the earliest metamorphic foliation was introduced in the rocks. It has been observed from the petrographic study of the schistose impure carbonate horizon, located just S of Helang, that the earliest micaceous foliation is not a transposition foliation and it follows a distinct compositional layering which is most probably primary in origin. No conspicuous folding associated with this early metamorphism could be recognised. These features coupled with the fact that only the low-grade minerals developed at this time, suggest, on one hand, that the first metamorphic foliation ( $S_I$ ) mimicked the original sedimentary layering ( $S_0$ ) and the metamorphism was probably in response to burial without much of folding associated with it, and on the other, that the metamorphism most probably took place in lower greenschist facies. The second metamorphism ( $M_{II}$ ) was the most intense one among the footwall events and it took place simultaneously with the  $F_I$ -folding. There was strong transposition of foliation during this period; in fact, the main foliation of the rocks seen in the outcrops originated at this time as an axial planar foliation to the  $F_I$  folds. It is possible that  $F_I$ -folding and the accompanying  $M_{II}$  metamorphism marked the onset of loading of the footwall block due to the beginning of the emplacement of the MCT-sheet above. The group of minerals that crystallised during this event (see Fig. 5.81b) suggests that the metamorphic conditions *rose well up to the middle or upper greenschist facies during the  $M_{II}$  metamorphism.* The nature of the  $M_{III}$  metamorphism in the Berinag-Mandhali formations was somewhat unusual in that, (i) its effects are not pervasively present, instead they are seen only in restricted zones or horizons, and (ii) two so-called high-grade minerals, such as kyanite and garnet, along with some biotites and muscovites have been found to have grown during this event which took place at the interkinematic stage between the  $F_I$  and  $F_{II}$  episodes, either late synchronously with or immediately following the MCT-emplacement. Most presumably, consequent upon the MCT-emplacement the metamorphic conditions locally rose up to the upper <sup>or even lower amphibolite</sup> greenschist facies and led to the development of the so-called high-grade minerals only where conducive environment (e.g. suitable lithology and/or access to fluids) prevailed. The  $M_{IV}$  metamorphism was largely retrogressive in nature and took place at the interkinematic stage between the  $F_{II}$  and

F<sub>III</sub> episodes, either synchronously or early synchronously with the later stretching event.

For a depiction of deformation - metamorphism time relationships in the footwall block, please see Table 5.2 given earlier in subsection 5.3.3. No definite correlation of tectonometamorphic events between the hanging wall and the footwall could be established. Probably there was only a crude correspondence in time between the M<sub>IV</sub> and M<sub>III</sub> events of the footwall with the M<sub>4</sub> and M<sub>3</sub> episodes of the hanging wall respectively.

### 5.8.3 Some remarks on the inverted metamorphism

The well known inverted metamorphism developed at the base of the Higher Himalayas can be inspected clearly in the Helang-Joshimath section. Whereas because the Sobala area does not show a full succession across the Munsiri Formation, we cannot get a full picture of the inverted metamorphic sequence from the Darmaganga section (see Plate V). It must be pointed out, however, that in the Sobala section, the Vaikrita Thrust does not coincide with the line of the first appearance of kyanite in the Joshimath Gneiss (i.e. Vaikrita Gneiss); the kyanite zone starts from some distance (~300 m ) above the Vaikrita Thrust.

Going up-section from the Munsiri Thrust in Helang, apparently the chlorite zone is met first and then biotite zone, followed by more distinct garnet zone and staurolite zone before the Vaikrita Thrust is encountered. The kyanite zone starts right from the Vaikrita Thrust and gives place to sillimanite zone farther up-section. Thus, generally speaking, the progressively higher grade minerals occur at successively higher structural and topographical levels in Joshimath area, thereby defining a typical example of the 'Himalayan inverted metamorphism' (for an appreciation, see Fig. 5.82).

However, on closer observation I found that, not all the metamorphic zonal boundaries are typical isograds; the chlorite and biotite zones are ill-defined and diffused; the so-called 'inverted metamorphic sequence' did not develop through a single episode of metamorphism, but is a collective result of more than one episode of metamorphism, *showing that* the inverted sequence is polymetamorphic in nature. It is the appearance of garnet, staurolite, kyanite and, to some extent, sillimanite zones at successively higher structural/topographical levels which forms the most distinctly observable cogenetic part of the inverted metamorphic sequence in the Higher

Himalayas in <sup>the</sup> Joshimath area. Whereas the low-grade part of the sequence, particularly the chlorite and biotite zones occurring in the lower part of the Munsiri Formation have been heavily affected by late retrogressive alterations during the  $M_4$  metamorphism, so much so that it is difficult to ascertain in the first instance whether these so-called 'chlorite' and 'biotite' zones are of primary or secondary origin. The following facts are to be noted in this context:

(i) No conspicuous example of primary chlorite has been found in the Munsiri Formation or in the Joshimath Gneiss. The most typical chlorites are found in some highly retrogressed phyllonite horizons in the Munsiri Formation.

(ii) Ideally the fresh-looking biotites of the intrusive granitoid bodies should be excluded from the discussion on inverted metamorphic sequence. It is the biotites in the country-rocks which should be taken into consideration; but many of these latter biotites that we see now are not primary, but of secondary retrogressive origin.

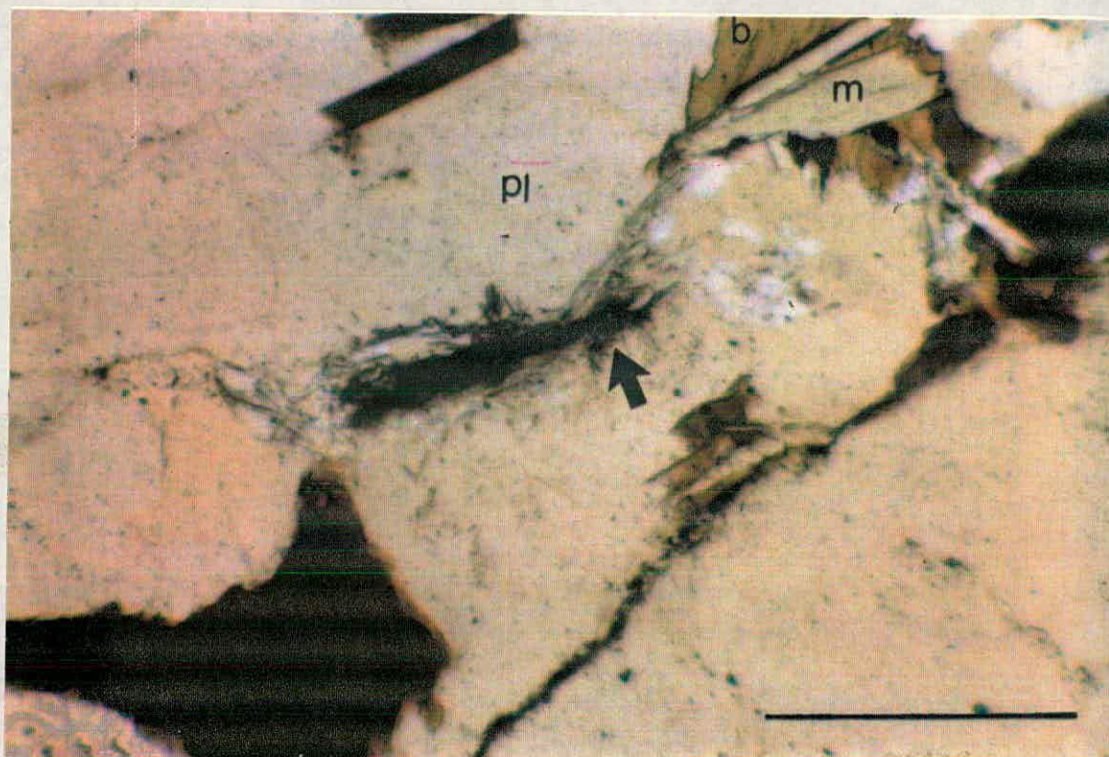
(iii) Relics of stretched garnets occur sporadically in several pelitic as well as calc-silicate bands even near the base of the Munsiri Formation, which is why one could argue that originally the lower part of the Munsiri Formation, just N of Helang, was possibly of garnet grade.

(iv) There were contributions from more than one episode of metamorphism for giving rise to the inverted metamorphic sequence. This is exemplified even by the preservation of early coarse-grained fabrics. Some amphibolite horizons near Helang in the lowermost part of the Munsiri Formation are still coarse-grained, probably implying that they acted as resistant horizons while other weaker horizons, notably the pelitic ones, could not escape retrogressive recrystallisation.

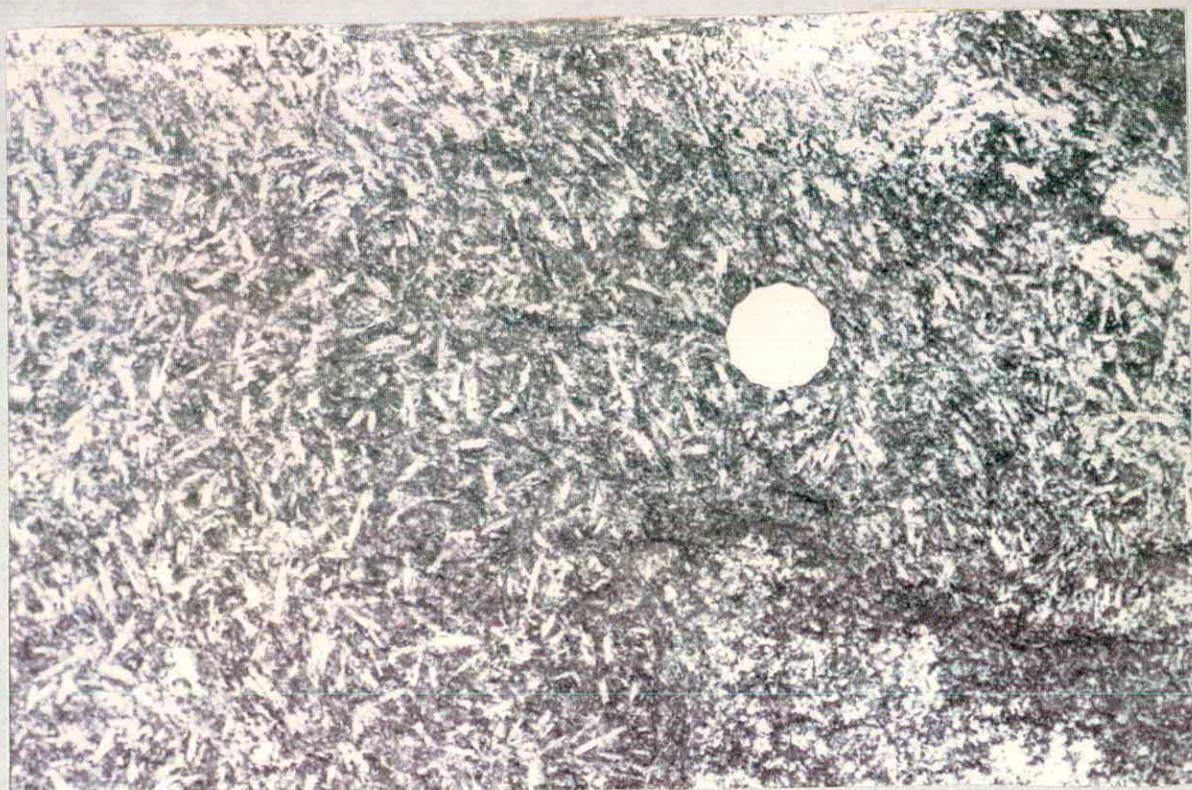
The above-mentioned facts suggest that the part currently appearing as the 'chlorite zone' and possibly also the 'biotite zone' at the basal portion of the Munsiri Formation are most likely of secondary (retrogressive) origin and they were originally included in the garnet zone. In other words, these two lowermost zones ('chlorite' and 'biotite') in the Joshimath - Helang section are not real Barrovian zones, but apparent.

The main cogenetic Barrovian part of the inverted metamorphic sequence is given by the four high-grade zones viz. garnet, staurolite, kyanite and sillimanite which developed during the  $M_2$  episode of metamorphism prior to both the  $F_2$ -folding and the MCT-emplacment. No indication has been found for the regional  $M_2$  Barrovian event to have taken place under the influence of a reverse geothermal gradient; so the Barrovian metamorphic zones must have originally developed in a normal i.e. right-way-up order. The inversion of this original right-way-up sequence

was caused by some later tectonic reasons postdating the  $M_2$  metamorphism. The most likely candidate which caused this inversion is either the  $F_2$ -folding or the MCT-thrusting or a combination of both. The detailed discussion on modelling the Higher Himalayan inverted metamorphism will follow in respective section in the next chapter.



**Fig. 5.1:** Fibrous masses of possible sillimanite (fibrolite) in Joshimath Gneiss. b = biotite, m = muscovite, pl = plagioclase. Photomicrograph, PPL. (Thin section 1F87). Scale bar 0.5mm.



**Fig. 5.2:** Field photograph showing clinozoisite crystals in semipelitic Joshimath Gneiss. From the appearance and mode of occurrence (needle-like whitish crystals oriented randomly on the main gneissosity surface) these were initially misidentified as sillimanites. Near Original Location NR20. Coin diameter 2.5cm.

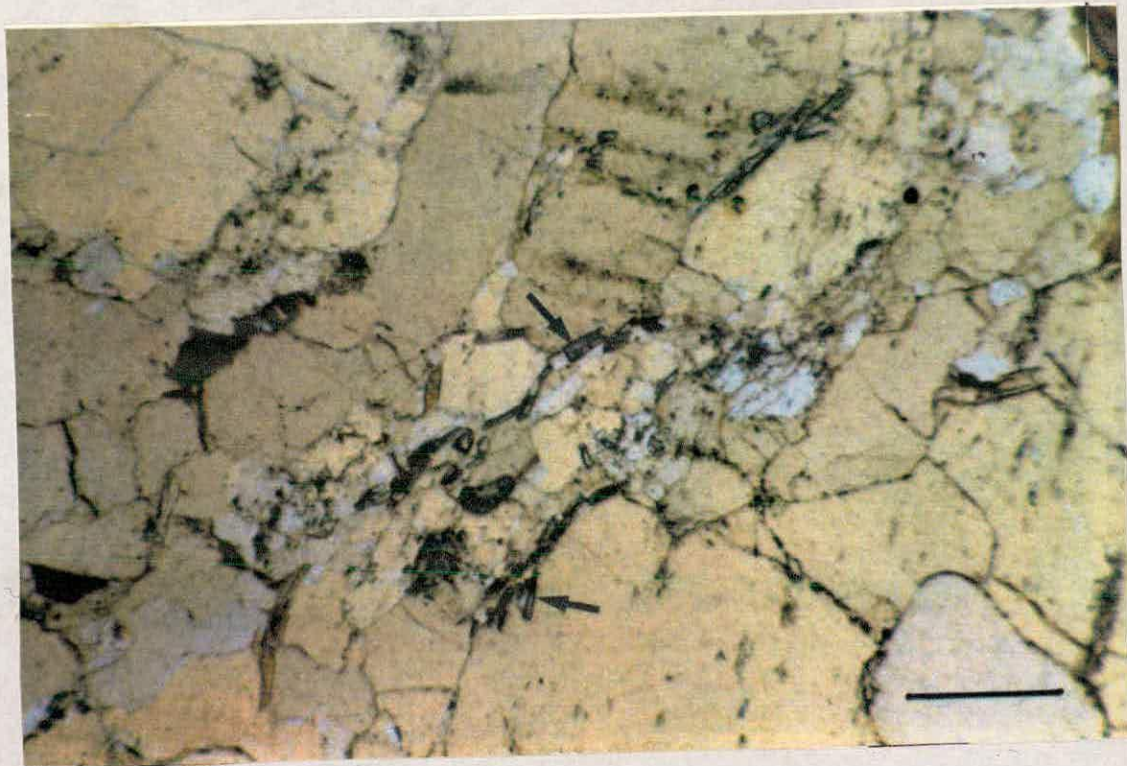
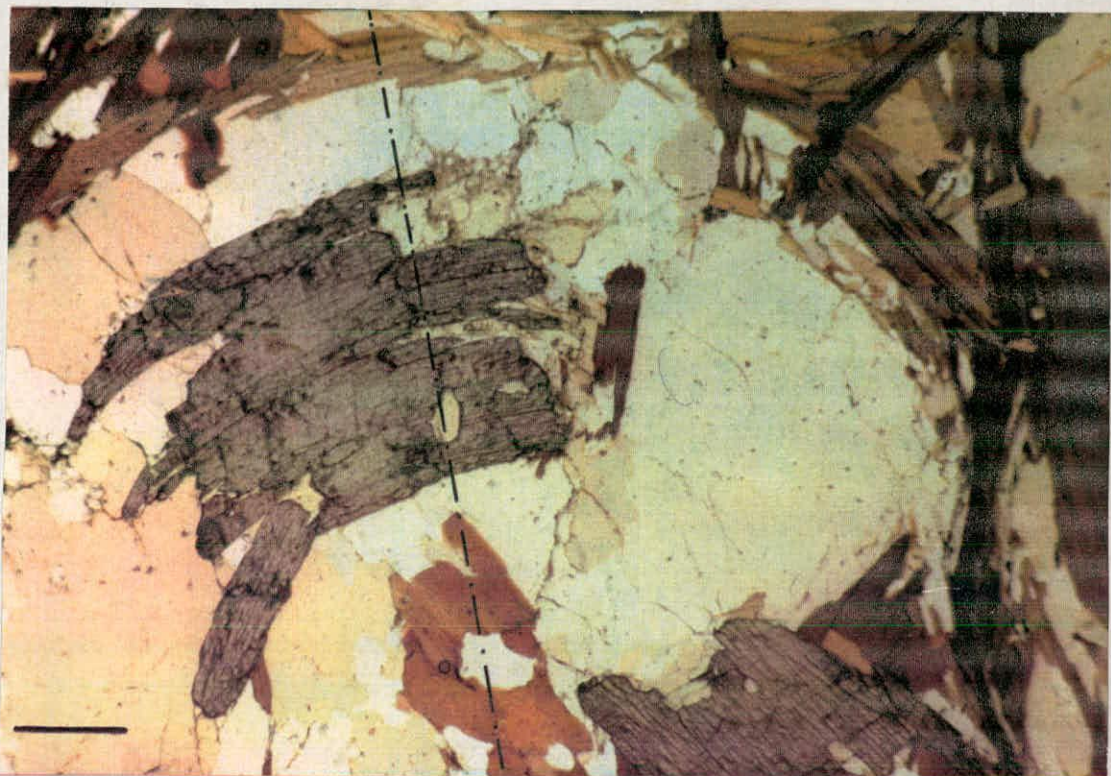


Fig. 5.3: Discrete tiny needles (arrowed) of kyanite, apparently looking like sillimanite. Photomicrograph, PPL. Joshimath Gneiss (thin section 22/4/88B). Scale bar 0.5mm.

	F <sub>1</sub>	F <sub>2</sub>	F <sub>3</sub>	F <sub>4</sub>
Sillimanite	●			○

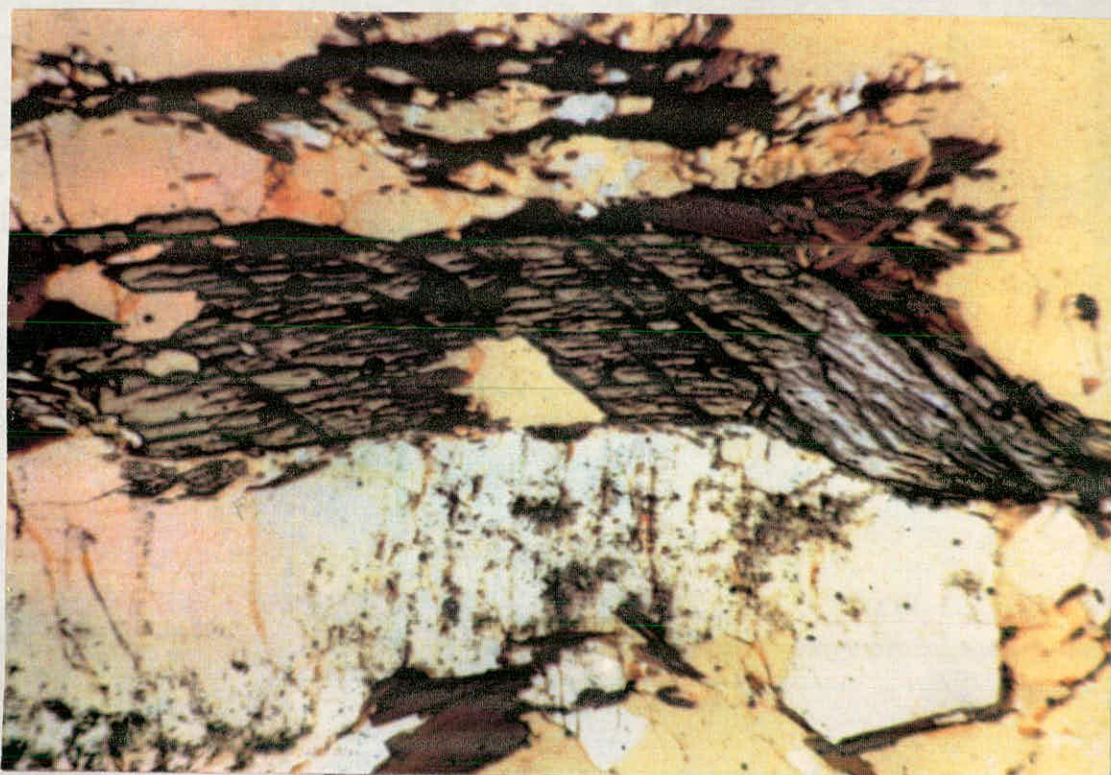
Fig. 5.4: Times of sillimanite growth in Joshimath Gneiss (Vaikrita Crystallines) with respect to fold episodes.



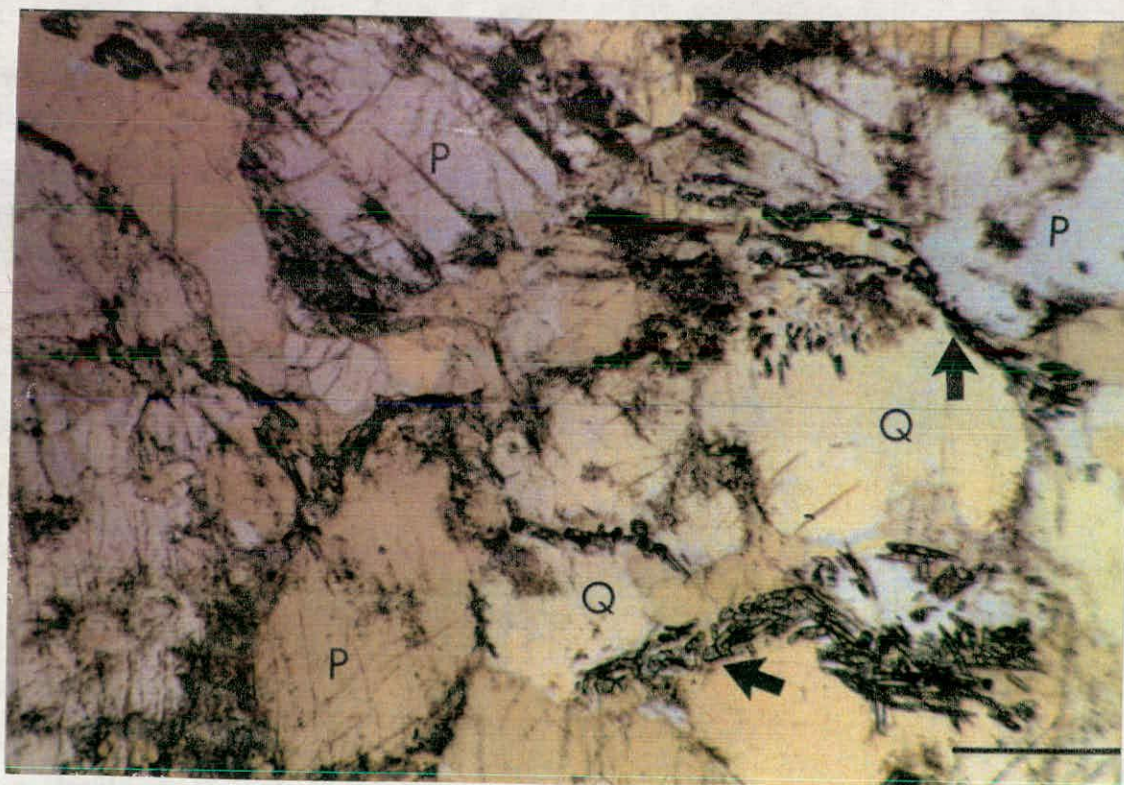
**Fig. 5.5:** In an  $F_2$ -hinge a kyanite grain is distinctly folded, implying that the kyanite is pre- $F_2$  in origin. The dash-dot line gives the  $F_2$  axial trace. Note how the lower left part of the kyanite grain is kinked and broken. Photomicrograph, PPL. Joshimath Gneiss (thin section 22/4/88B). Scale bar 0.5mm.



**Fig. 5.6:** A folded porphyroblast of kyanite in Joshimath Gneiss. A freeze-shot of the migratory pattern of extinction in the grain is clearly visible. Photomicrograph, PPL. (Thin section NR37/38(2)). Photo-base 5.4mm.



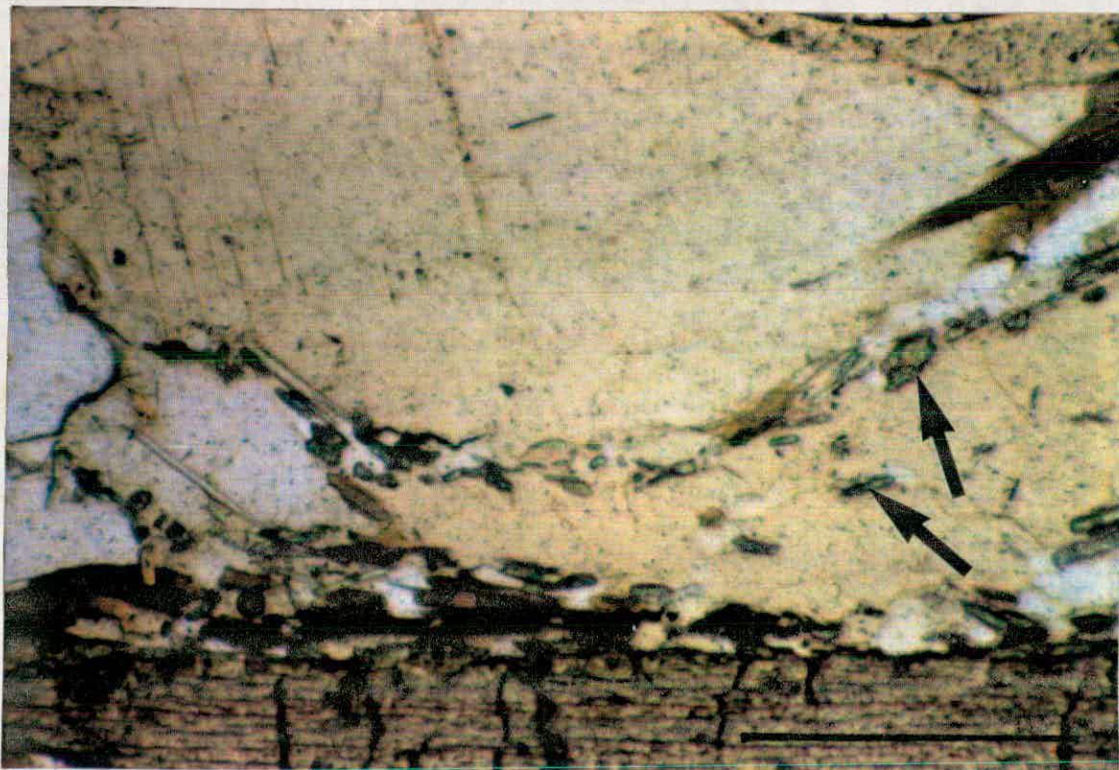
**Fig. 5.7:** Bent outline of an elongate kyanite grain. Photomicrograph, PPL. Joshimath Gneiss (thin section ~NR38/39). Photo-base 5.4mm.



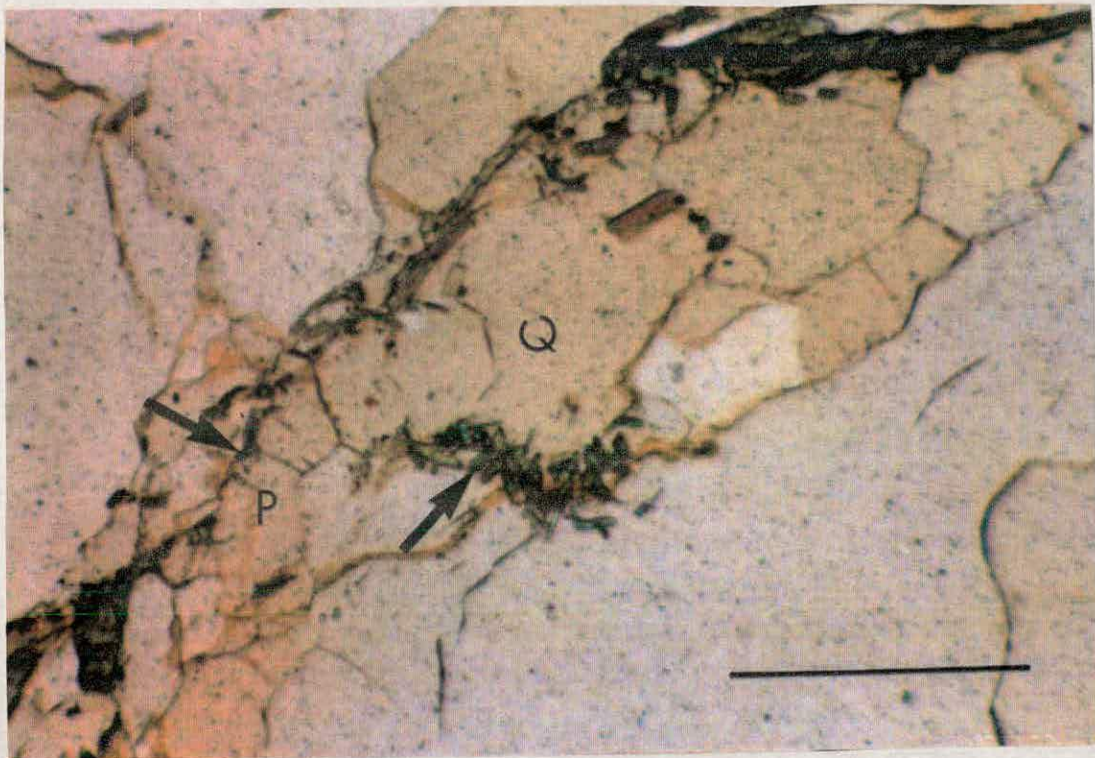
**Fig. 5.8:** Trails (arrowed) of tiny kyanite prisms occur partly bordering large grains of quartz and feldspar. p = plagioclase, q = quartz. Photomicro., PPL. Joshimath Gneiss (thin section 'Kya from foot-track'). Scale bar 0.5mm.



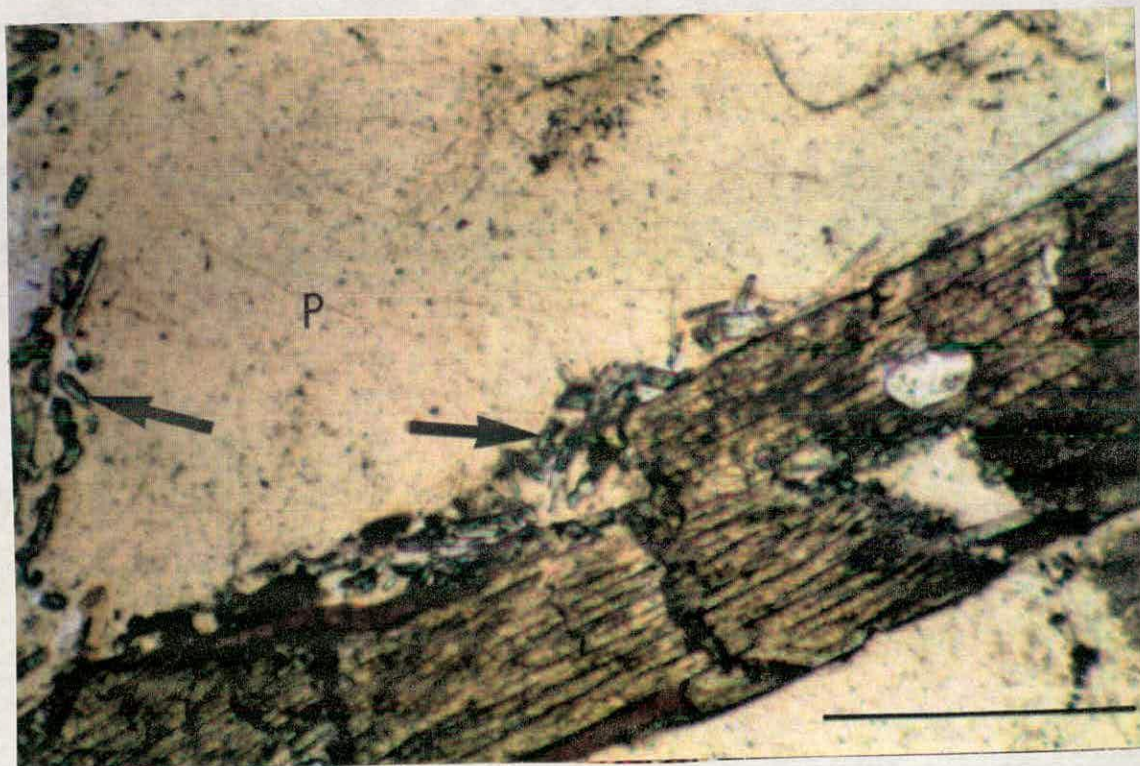
**Fig. 5.9:** Occurrence of larger kyanite grains amidst biotite, feldspar, quartz and muscovite grains, but without any regularity in mutual orientational relations. No conspicuous tiny kyanite grains could be caught in this view. Photomicro., PPL. Joshimath Gneiss (thin section 22/4/88B). Scale bar 0.5mm.



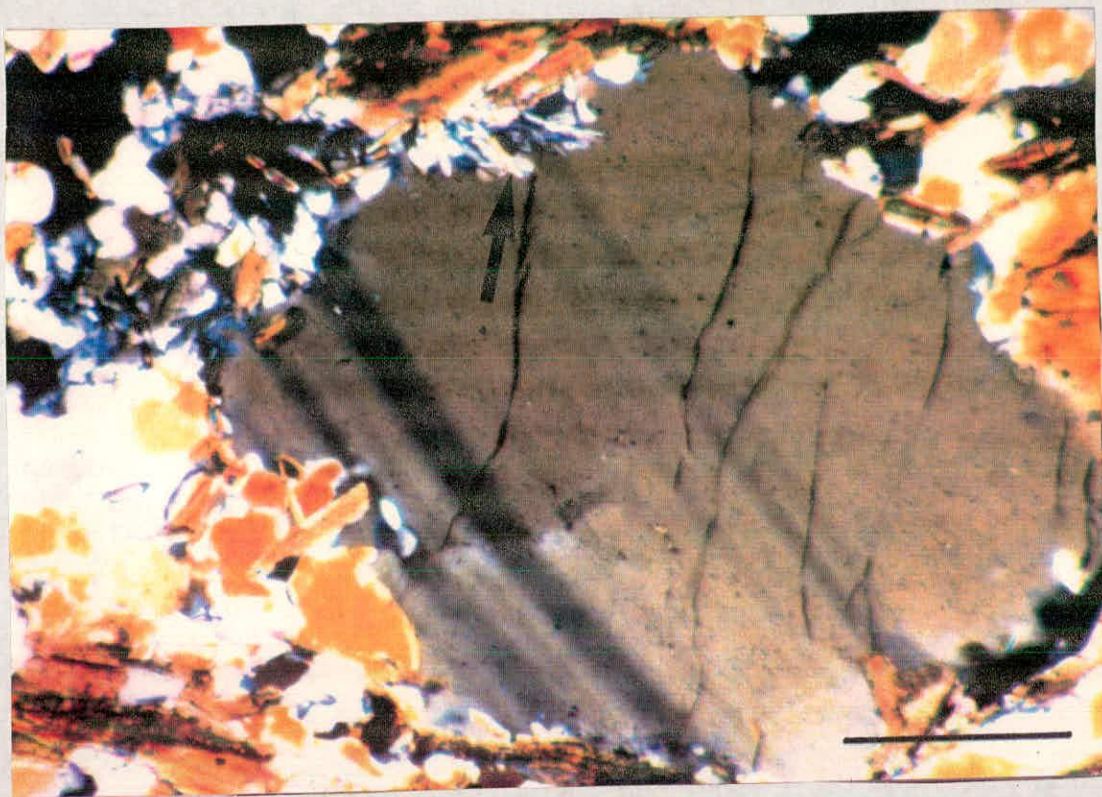
**Fig. 5.10:** Occurrence of tiny kyanite prisms (arrowed) adjacent to a large kyanite grain lying at the bottom. Note that generally the orientation of the tiny crystals does not follow that of the latter. Photomicro., PPL. Joshimath Gneiss (thin section 22/4/88B). Scale bar 0.5mm.



**Fig. 5.11:** Tiny kyanite crystals grown around broadly augen-shaped quartz and feldspar grains outside the vicinity of any larger kyanite grain. p = plagioclase, q = quartz. Photomicro., PPL. Joshimath Gneiss (thin section NR37/38(2)). Scale bar 0.5mm.



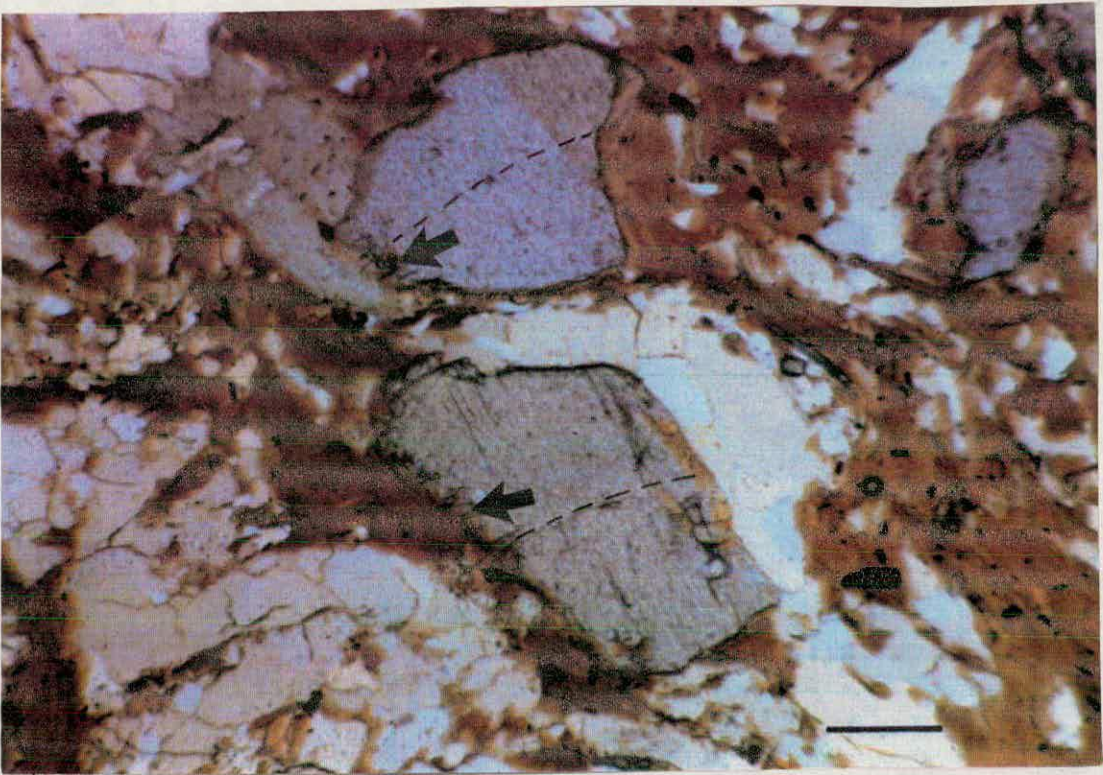
**Fig. 5.12:** Some tiny kyanite crystals occur spatially close to a larger kyanite porphyroblast; but note that the former are distributed with clear disregard to the orientation of the latter. In fact, they are partly bordering an elliptical plagioclase grain (p). Photomicro., PPL. Joshimath Gneiss (thin section 22/4/88B). Scale bar 0.5mm.



**Fig. 5.13: (a)** An example where most of the tiny kyanite crystals (arrowed) are found to be aligned at high angle to the boundary of an augen-shaped plagioclase grain. Photomicro., XPL. Joshimath Gneiss (thin section NR37/38(2)). Scale bar 0.5mm.



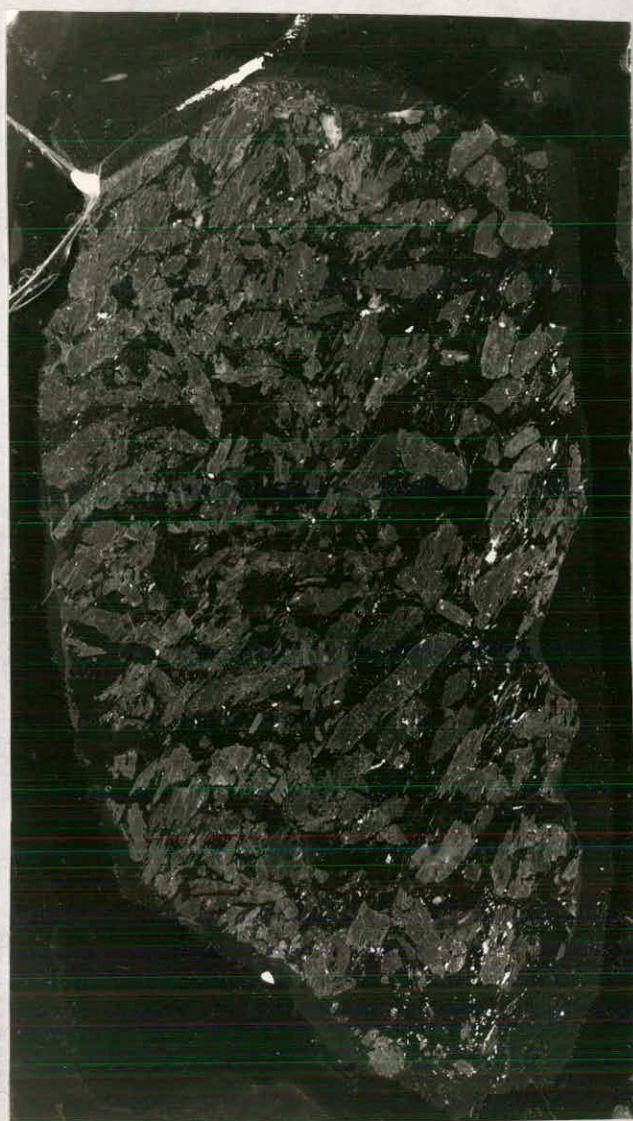
**(b)** Some tiny kyanite crystals located at the boundary of an augen-shaped plagioclase project well into the main body of the augen. Further details are as in (a) above.



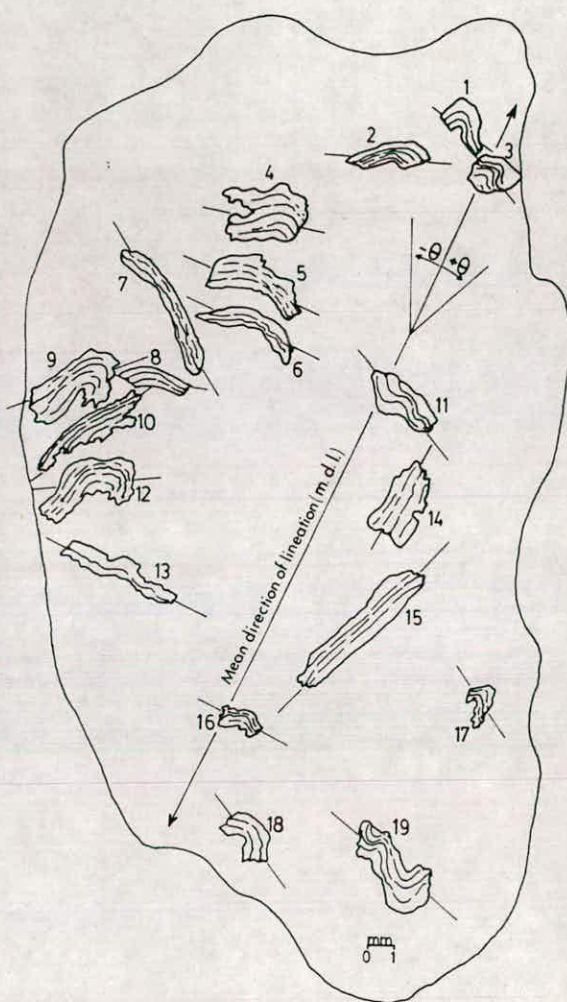
**Fig. 5.14:** Minor recrystallisation along parts of the transverse boundaries of some strained (i.e. gently folded and/or having undulose extinction) bigger kyanite grains. Photomicro., PPL. Joshimath Gneiss (thin section NR37/38(2)). Scale bar 0.5mm.

	F <sub>1</sub>	F <sub>2</sub>	F <sub>3</sub>	F <sub>4</sub>
Sillimanite				
Kyanite				

**Fig. 5.15:** Times of kyanite growth in Joshimath Gneiss with respect to fold episodes. Sillimanite growth times are also indicated for comparison.



(a)



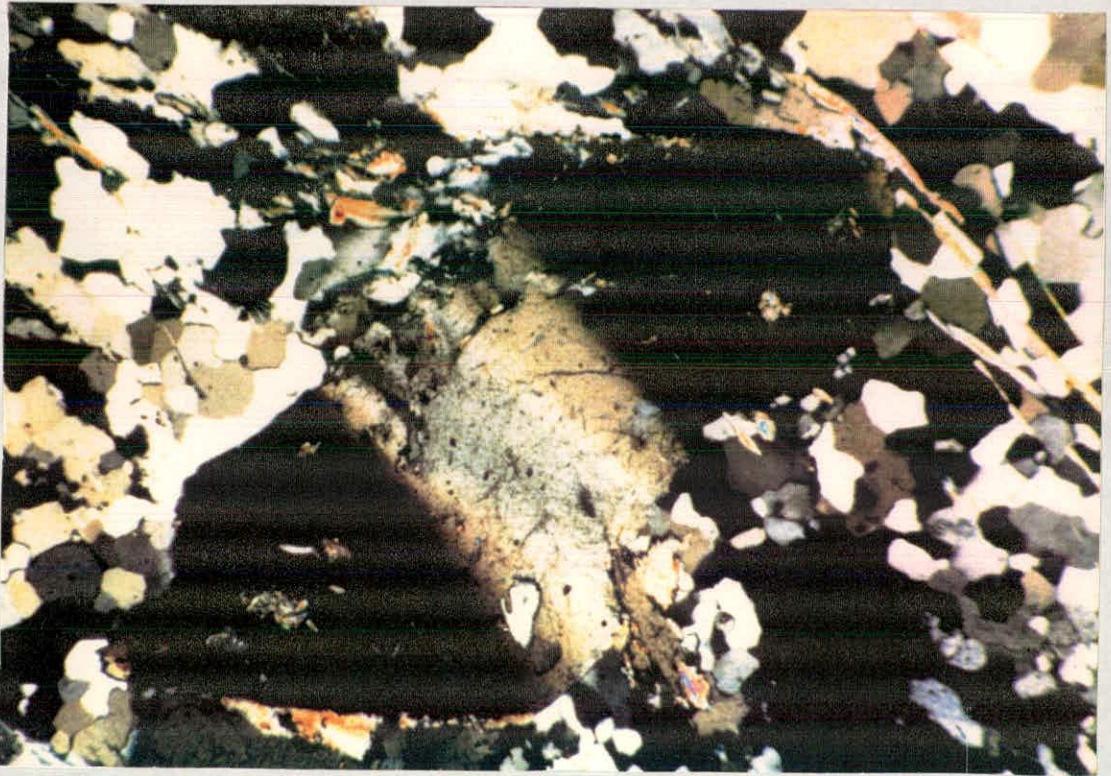
(b)

**Fig. 5.16:** (a) A negative print of the whole thin section - F.W.Kya'89 (this is an XY-section, cut parallel to the main foliation). Note the profusion of kyanite grains in the thin section. A stretching lineation runs from lower left to upper right. There is a fairly strong preferred orientation of the kyanite porphyroblasts, many of them having their lengths parallel or sub-parallel to the lineation. In details the kyanite grains show varied textures. Kyanite quartzite of the Berinag-Mandhali formations.

(b) Tracing of 19 kyanite grains from the same thin section as above. Most of these grains show conspicuous folded outline and are aligned at varying angle to the lineation. The geometrical analyses shown in Fig. 5.22 refer to these grains



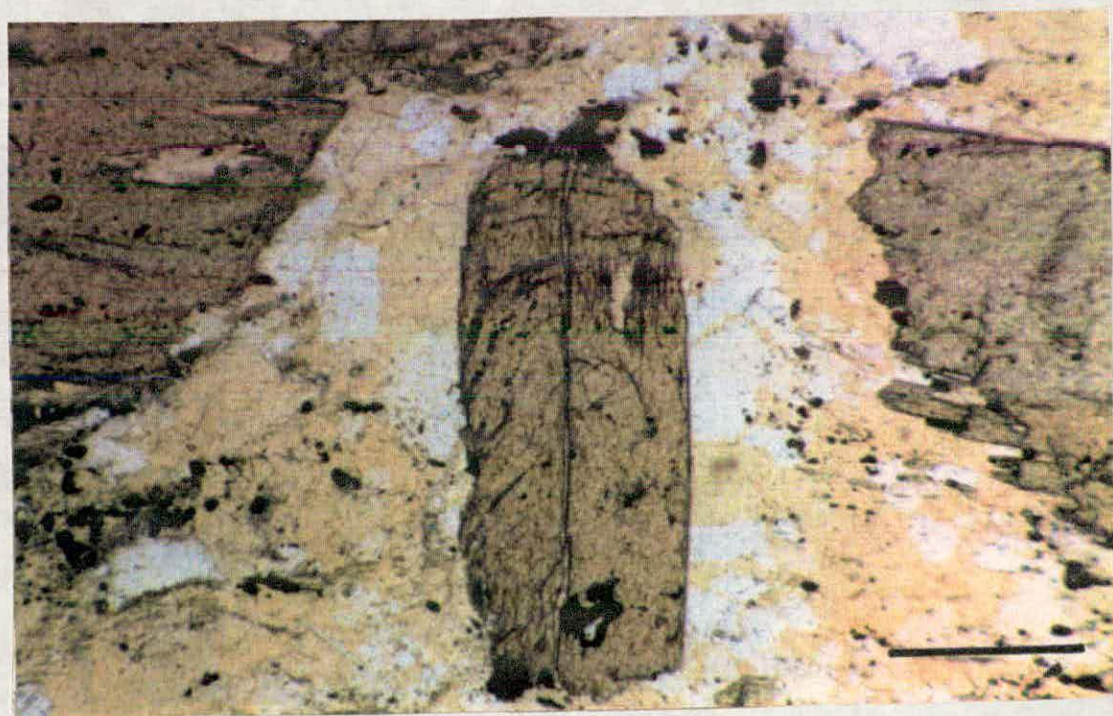
**Fig. 5.17:** Some elongate kyanite porphyroblasts are pulled apart (stretched) along the direction of stretching lineation. The direction of lineation is shown by the dashed line. Pull-apart gaps in three kyanite porphyroblasts are shown by arrows. Photomicro., PPL. Kyanite-bearing quartzite of Berinag-Mandhali formations (thin section - F.W.Kya'89). Scale bar 0.5mm.



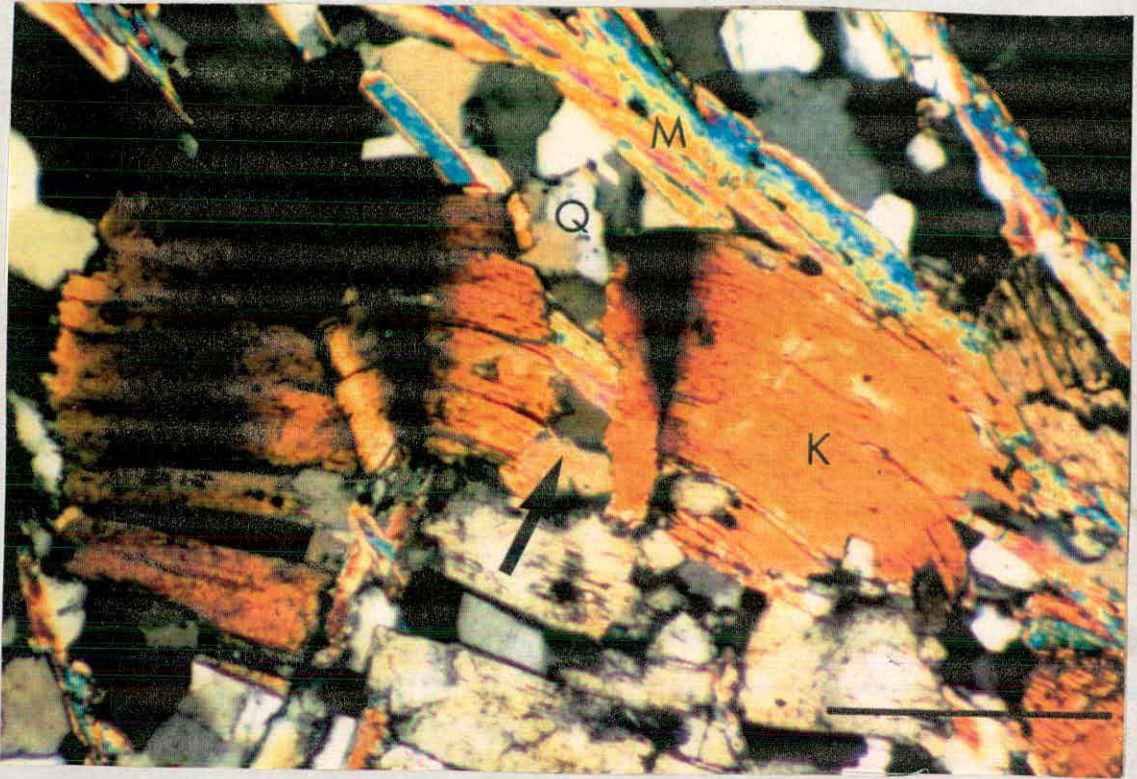
**Fig. 5.18:** A conspicuously kinked or folded porphyroblast of kyanite. The strain shadow pattern in the grain is remarkable; in the present position it looks as though it is deformation twinning parallel to the fold axial trace. The fold axial trace gives the direction of stretching in this case. Photomicro., XPL. Kyanite-bearing quartzite of Berinag-Mandhali formations (thin section - F.W.Kya'89). Photo-base 3.25mm.



**Fig. 5.19:** Fine trails of tourmaline threads (dark greyish blue in colour; arrowed) aligned parallel to the stretching lineation run across a folded kyanite porphyroblast following the fold axial trace. The stretching lineation is vertical in this photograph. Photomicro., PPL. Kyanite-bearing quartzite of Berinag-Mandhali formations (thin section - F.W.Kya'89). Scale bar 0.5mm.



**Fig. 5.20:** Here the stretching direction is kept almost horizontal i.e. sub-parallel to the base of the photograph. Note the kyanite porphyroblast at centre is aligned nearly at right angle to the stretching lineation, yet its outline shows no sign of folding. However, one can see folded tourmaline trails (included) within the porphyroblast; the median fracture in the porphyroblast runs along the axial trace of this fold. Note that the folded tourmaline trails stop at the margin of the host porphyroblast and do not continue into the matrix. Photomicro., PPL. (Thin section - F.W.Kya'89). Scale bar 0.5mm.



**Fig. 5.21:** A folded as well as pulled-apart kyanite porphyroblast. Note the occurrence of muscovite, quartz and fine tourmaline threads in the pull-apart zone (arrowed). Stretching direction is parallel to the base of the photograph. K = kyanite, M = muscovite, Q = quartz. Photomicro., PPL. (Thin section - F.W.Kya'89). Scale bar 0.5mm.

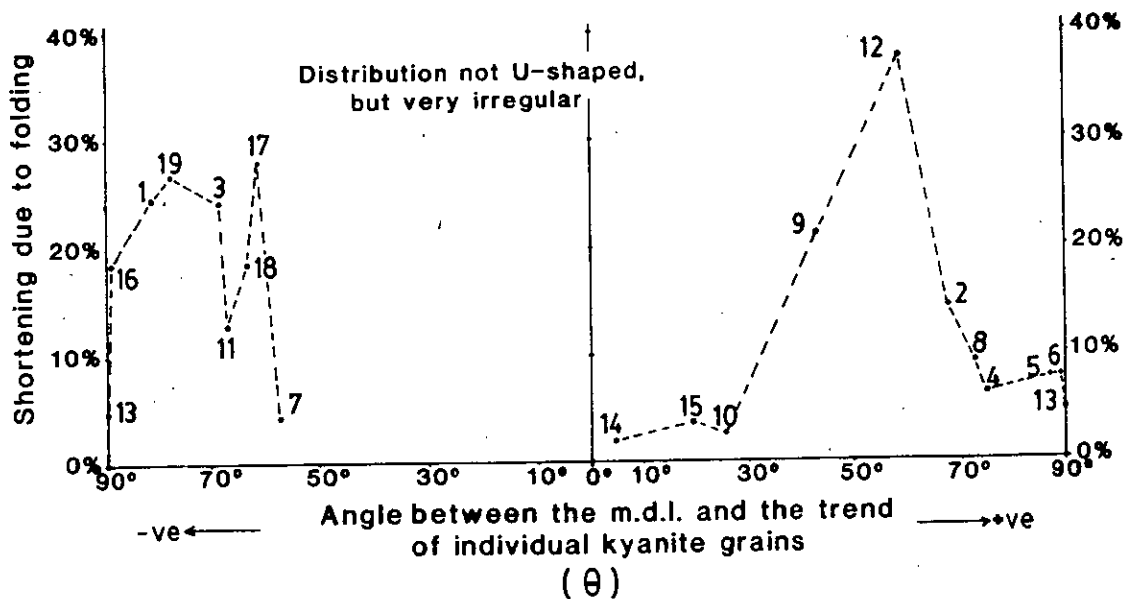
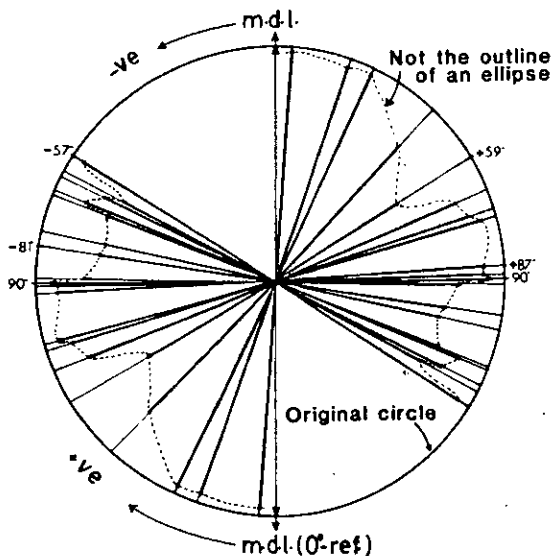
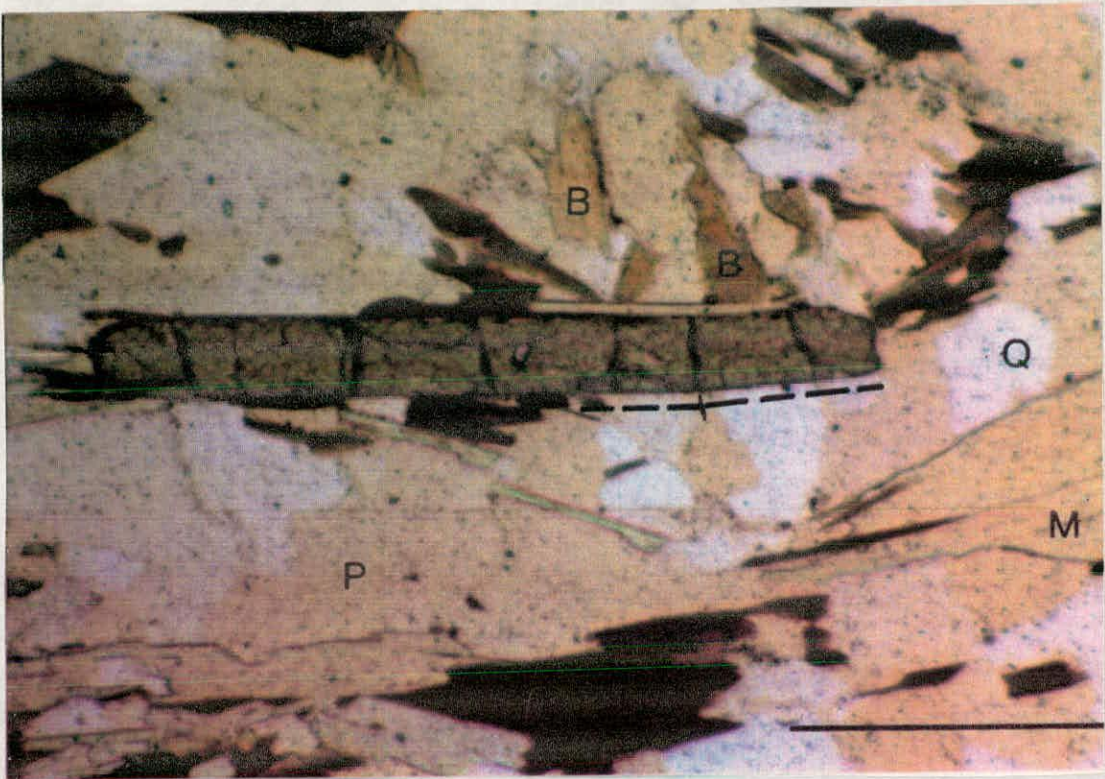


Fig. 5.22: (a) Angular distance from lineation direction vs. shortening (%) shown by kyanite grains in Fig. 5.16b (note the grain serial nos). The pattern is highly irregular and suggests no systematic variation in the amount of shortening due to folding of the kyanite grains relative to their orientation with respect to the lineation. Folding seems to be unrelated to lineation. (m.d.l. = mean direction of lineation)



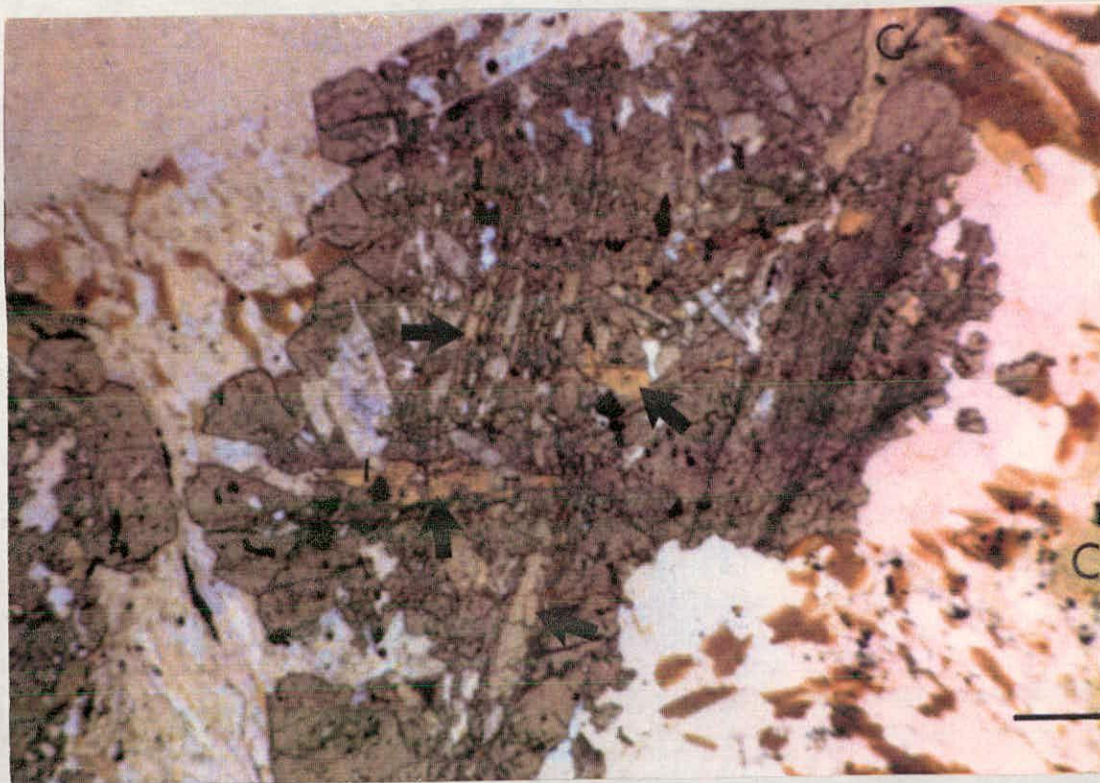
(b) Further demonstration (visual) that the shortening distribution is unsystematic. Different diameter of the original circle representing differently oriented grains are shortened accordingly, but the final outline is far from defining an ellipse (strain ellipse).



**Fig. 5.23:** A staurolite prism showing transverse cracks. On close observation using higher magnification, a partial knee-bend breaking is noticed at the second crack from right. The difference in orientation of the two segments on either side of the crack is shown by the dashed lines. B = biotite, M = muscovite, Q = quartz, P = plagioclase. Photomicro., PPL. Joshimath Gneiss (thin section 5B<sub>~a</sub>'87). Scale bar 0.5mm.



**Fig. 5.24:** Occurrence of staurolite (arrowed), not included in a second phase such as garnet. In the text, these staurolites are referred to as 'matrix staurolites' for convenience. Contrast the much bigger size of the garnet porphyroblasts present in the same rock. Photomicro., PPL. Upper Munsiri semipelite (thin section - 7H'87). Scale bar 0.5mm.



**Fig. 5.25:** Distinct examples of staurolite crystals (arrowed) included within garnet porphyroblast. Note the inhomogeneous-textured (i.e. spongy) body of the host garnet. C = chlorite. Photomicro., PPL. Munsiri Formation (thin section - NR58/59). Scale bar 0.5mm.

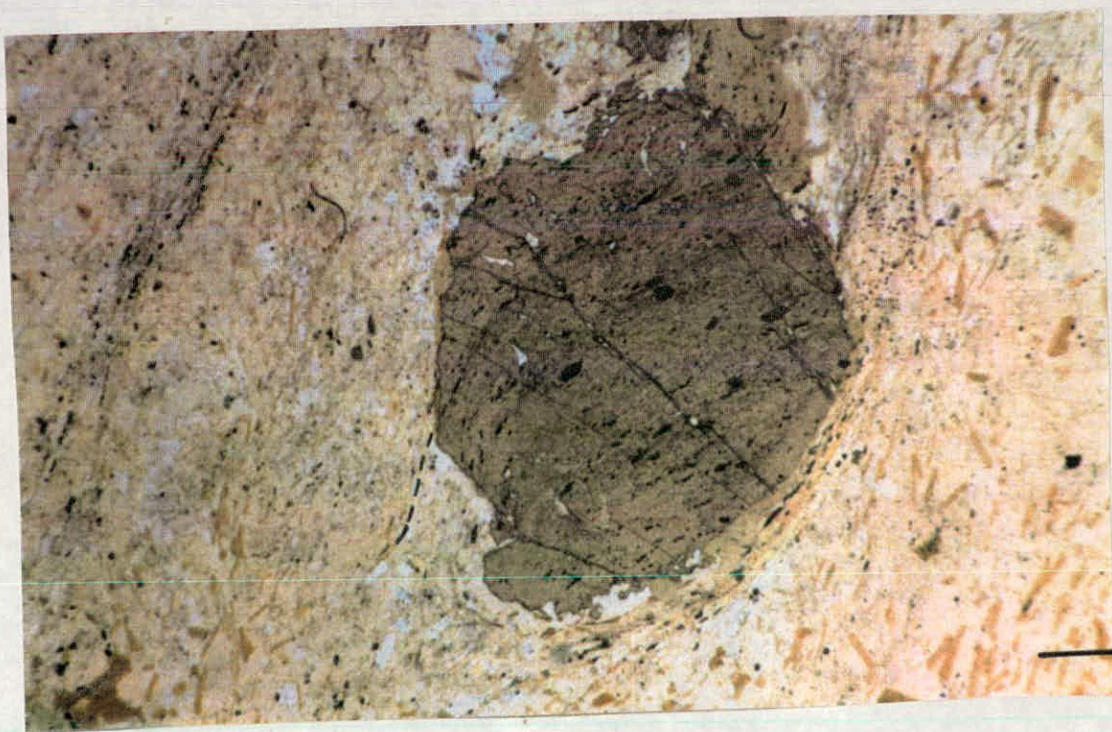


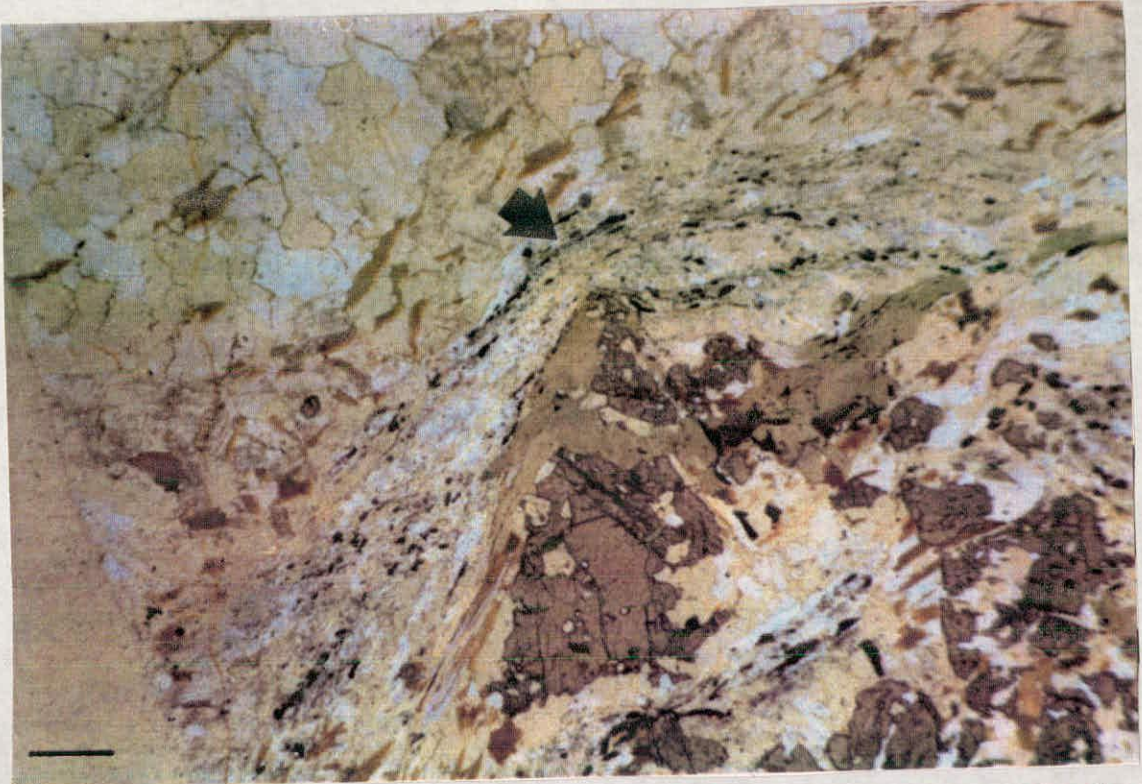
**Fig. 5.26:** A string of pulled apart staurolite fragments. Note none of the fragments is aligned transverse or oblique to the foliation. This is a close-up view of most of the staurolite grains shown in Fig. 5.24. The adjacent green bands below the staurolites show chloritised biotites on either side of a quartz-rich vein. Part of a garnet porphyroblast is at lower right. Photomicro., PPL. Munsiri Formation (thin section - 7H'87). Scale bar 0.5mm.



**Fig. 5.27:** (Left) The staurolite crystals included within garnet porphyroblast in the upper half of the photograph are mostly prismatic or needle-shaped and intact. Note the mutual orientation of the crystals and also with respect to the external foliation whose average direction is given by the dashed line. For other details see Fig. 5.25.

**Fig. 5.28:** (Below) Opaque trails defining the matrix foliation (Se) continue undeflected through a homogeneous garnet porphyroblast. Note particularly the top right and middle left of the porphyroblast. C = chlorite; other matrix phases are quartz, muscovite, biotite, plagioclase & black opaques. Photomicro., PPL. Munsiri Formation (thin section - NR58/59). Scale bar 0.5mm.





**Fig. 5.29:** Matrix foliation swerves round the projected edge of a spongy garnet porphyroblast, implying that the porphyroblast is older than the foliation. Note chloritisation of biotite and garnet. Other details are as in caption of Fig. 5.28.



**Fig. 5.30:** In a garnet porphyroblast the spongy part (sg) has an overgrowth of homogeneous garnet (hg). Opaque trails are common in the homogeneous outer part, but not in the spongy inner part. Photomicro., PPL. Munsiri Formation (thin section - NR58/59). Scale bar 0.5mm.



**Fig. 5.31:** Slightly folded outline of an elongate spongy garnet porphyroblast. Photomicro., PPL. Munsiri Formation (thin section - NR58/59). Scale bar 0.5mm.

	F <sub>1</sub>	F <sub>2</sub>	F <sub>3</sub>	F <sub>4</sub>
Sillimanite				
Kyanite				
Staurolite				

**Fig. 5.32:** Time of staurolite growth with respect to fold episodes in Munsiri Formation as well as in Joshimath Gneiss.

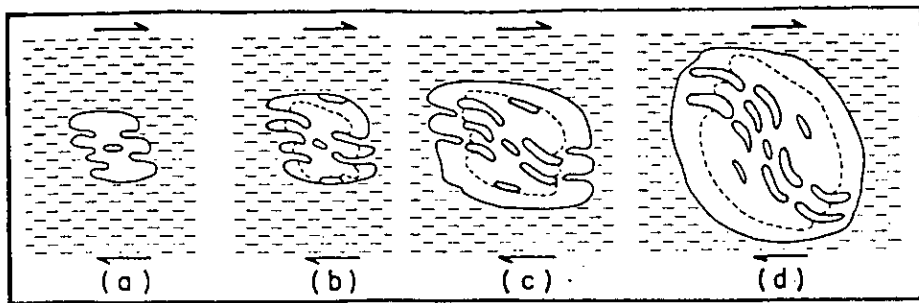


Fig. 5.33: Progressive growth (a to d) of skeletal variety of snowball garnet. The garnet extends digitations and incorporates quartz grains that are elongate along the foliation. The dotted lines in (b), (c) and (d) are given to distinguish the incremental growth in each stage; so they show the porphyroblast outline, albeit subsequently rotated, acquired at the end of immediate previous stage (Spry, 1963).

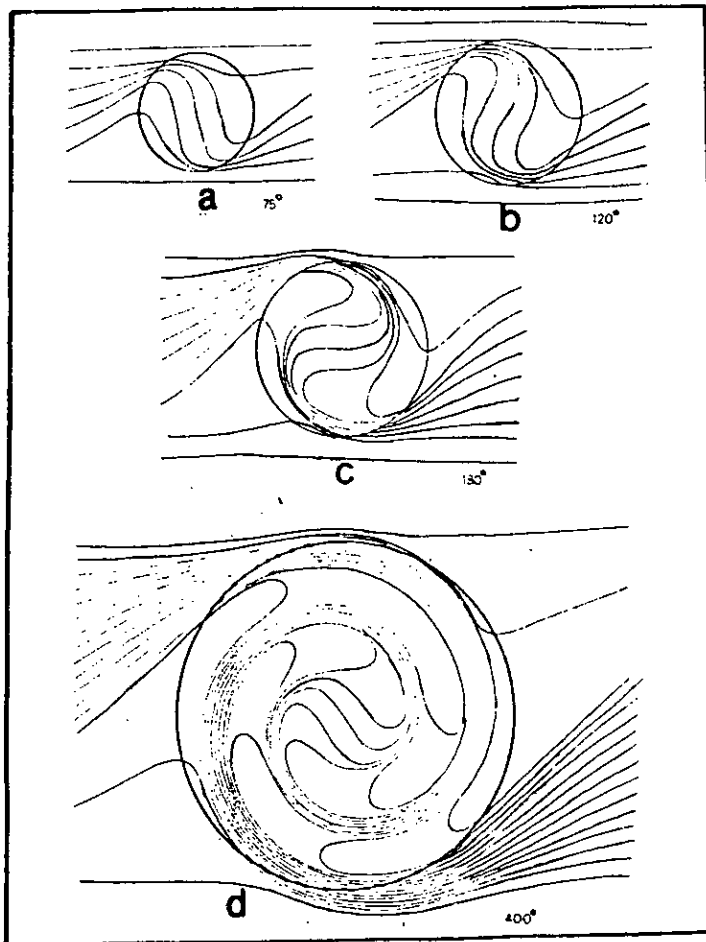


Fig. 5.34: Idealised representation of different stages (a to d) in progressive rotation-cum-growth of garnet porphyroblast with associated Si-Se fabric, as indicated by Schoneveld's String Experiment. The number (in degrees) under each figure gives the amount of rotation suffered by the porphyroblast (after Schoneveld, 1977).

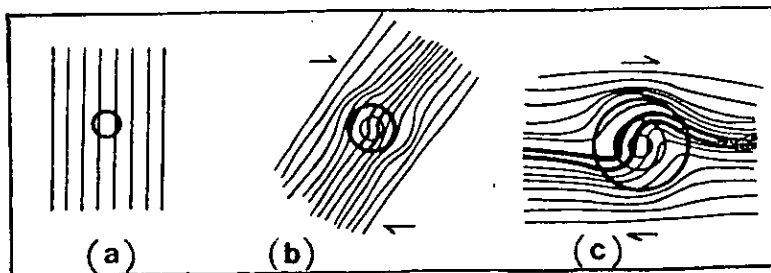
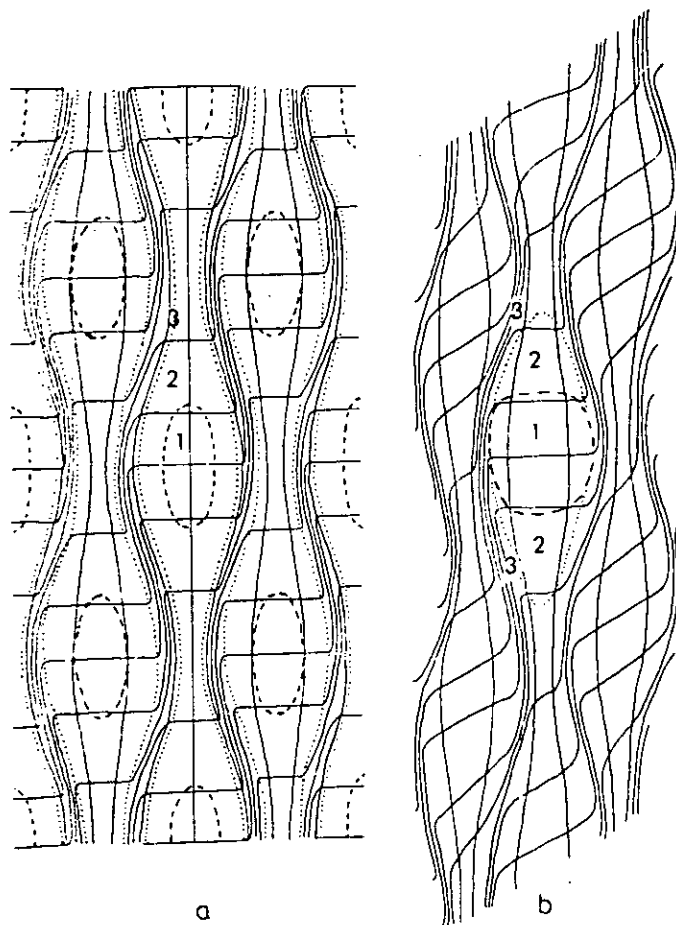
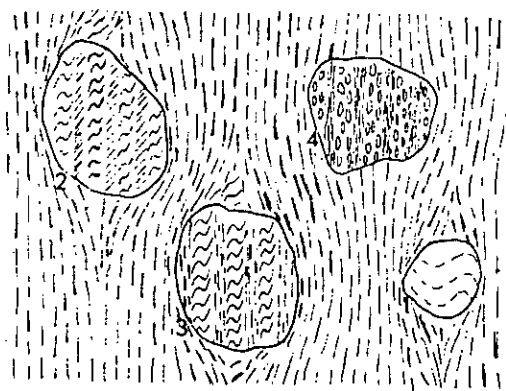


Fig. 5.35: A model showing the development of sigmoidal inclusion fabric in a stationary growing porphyroblast; the external foliation is being progressively transposed (rotated) through  $90^\circ$  (after Barker, 1990 following Wilson, 1971).

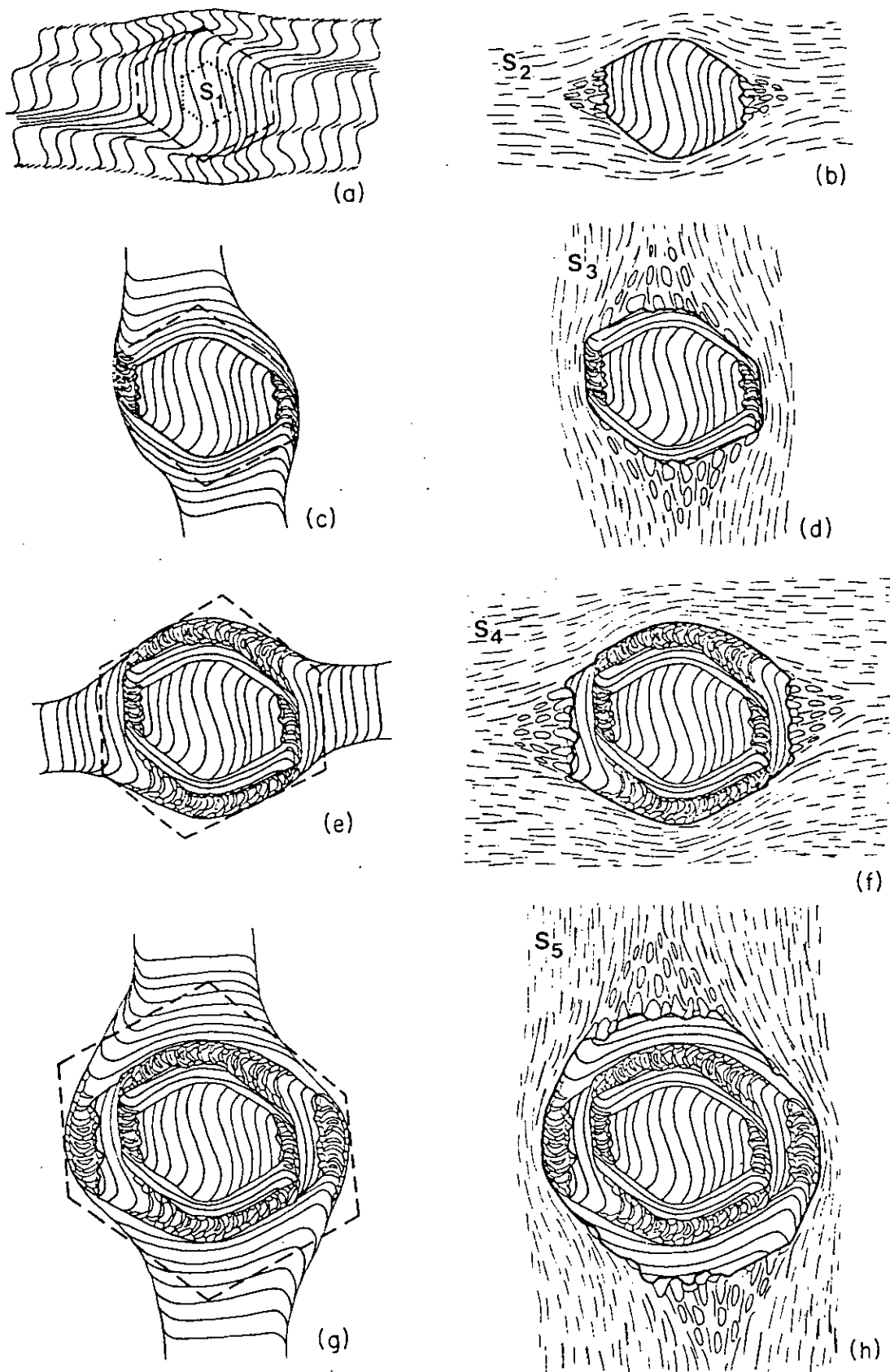


**Fig. 5.36:** (a) Schematically shown distribution of deformation partitioning on a strain-field diagram (XZ-plane) for noncoaxial progressive bulk inhomogeneous shortening. 1 (areas within dashed ellipses) - zones of no strain, 2 (areas between dashed and dotted lines) - zones of progressive shortening strain, 3 (areas lying only between dotted lines) - zones of progressive shortening and shearing strain (after Bell, 1985).

(b) Sketch showing partitioned deformation pattern around a porphyroblast (with dashed outline) as a result of non-coaxial progressive bulk inhomogeneous shortening. Numbering keys are the same as in (a) (after Bell, 1985).



**Fig. 5.37:** Broadly contemporaneous porphyroblasts that have overgrown different stages and places (1 to 4) of development of a second generation crenulation cleavage. Note that the porphyroblast-matrix relations and the distribution of phyllosilicates within the porphyroblasts imply that, relatively and precisely speaking, the formation of porphyroblast-1 preceded that of porphyroblast-4.



**Fig. 5.38:** Model showing the progressive development of spiral-shaped inclusion trails in a garnet porphyroblast. The minerals (notably, quartz) in the pressure fringe/strain shadow zone develop crenulation due to subsequent stage of orthogonal foliation formation (after Bell & Johnson, 1989).

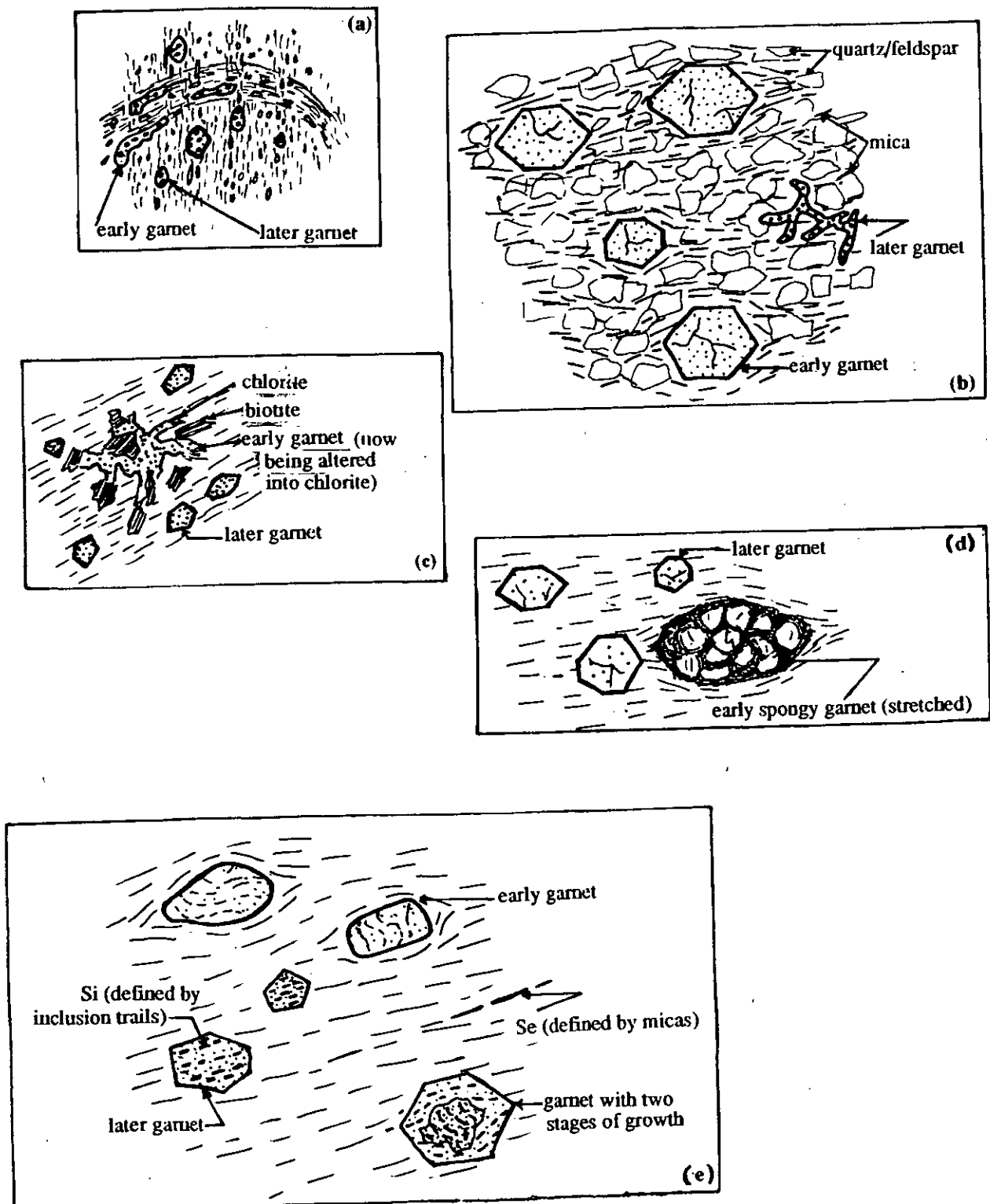
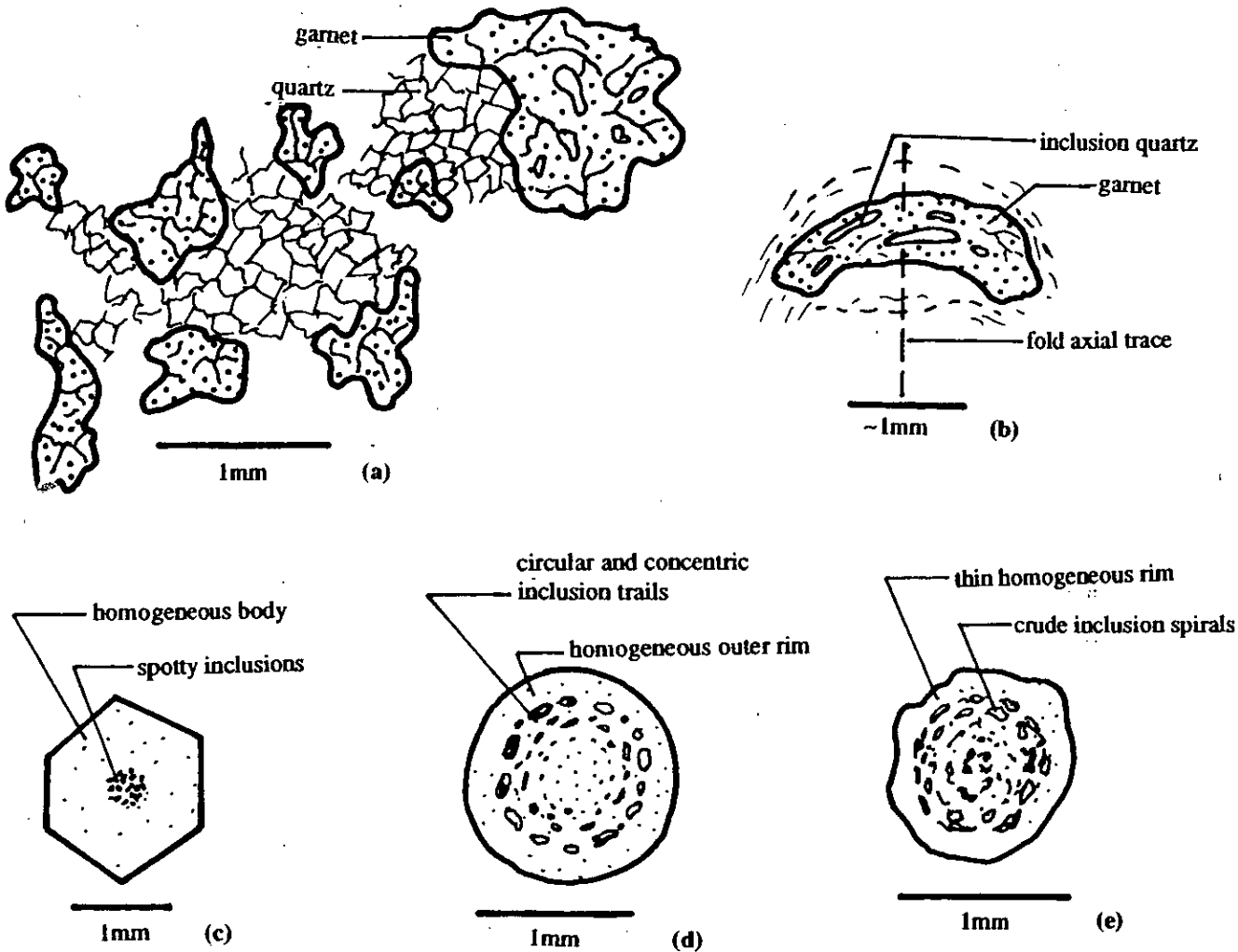


Fig. 5.39: Some textural criteria for relative age determination among multiple generations of garnet porphyroblasts. See text for further details. Sketches are not to scale.



**Fig. 5.40:** (a) Sketch showing part of a large skeletal garnet porphyroblast (class I in Joshimath Gneiss) from a semipelitic horizon. See text for details.

(b) Sketch showing an elongate garnet porphyroblast which is folded (late synchronously with or shortly after  $F_2$  folding) and does not have crystal outline. Note the presence of length-parallel quartz inclusions. (Class II garnet in Joshimath Gneiss). Further details in text.

(c) Sketch of a euhedral outlined garnet porphyroblast with almost entirely homogeneous body except a small central zone of spotty inclusions. (Class III garnet in Joshimath Gneiss).

(d) Sketch of a garnet porphyroblast, circular in outline and having concentric trails of inclusions inside a homogeneous outer rim. (Class IVa type in Joshimath Gneiss). Details in text.

(e) Garnet porphyroblast with rounded outline and a comparatively thin homogeneous rim surrounding a broad inner zone of crudely arranged inclusion spirals. (Class IVb type garnet in Joshimath Gneiss).

(f) B/W photomicrograph of a subcircular garnet porphyroblast with fairly well-developed inclusion spirals. Note the absence of any rim overgrowth or grain elongation parallel to foliation trace (here, stretching lineation). The stretching direction is parallel to the base of the photograph. For details see text. (Class Va type garnet in Joshimath Gneiss). PPL; Scale bar 1mm.

(g) B/W photomicrograph of a garnet porphyroblast with broad elliptical outline; the ellipse is more clearly defined if the pressure shadow areas are also considered. Ellipse long axis parallels the stretching direction. The porphyroblast shows a relatively broad inclusion spiral, but no rim overgrowths. Note the inclusion grain-size increases from centre outward. See text for details. (Class Vb type garnet in Joshimath Gneiss). XPL; long diameter of garnet is 5.8mm.

(f) & (g) figures (B/W photomicrographs) are on next page

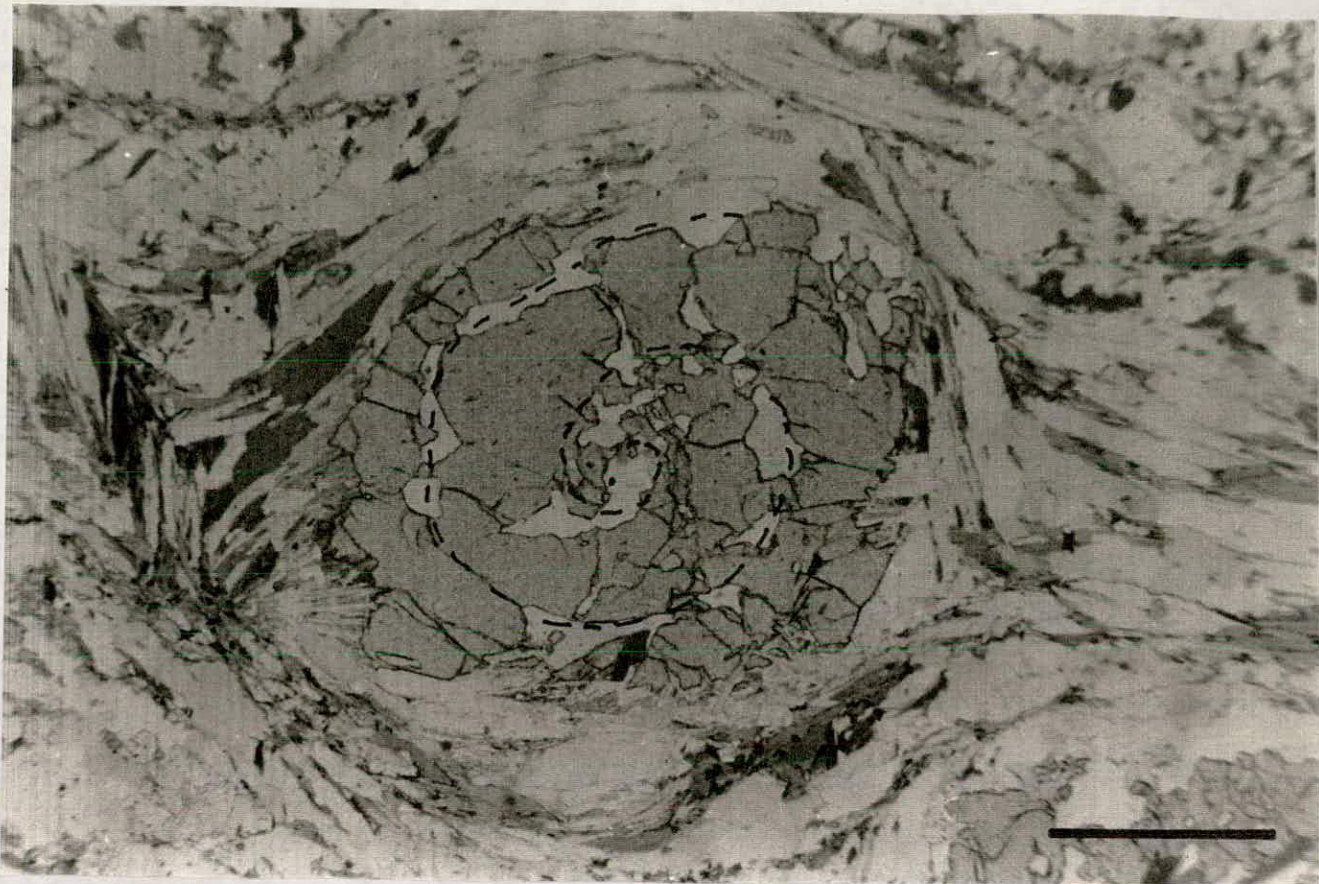


Fig. 5.40: (f)

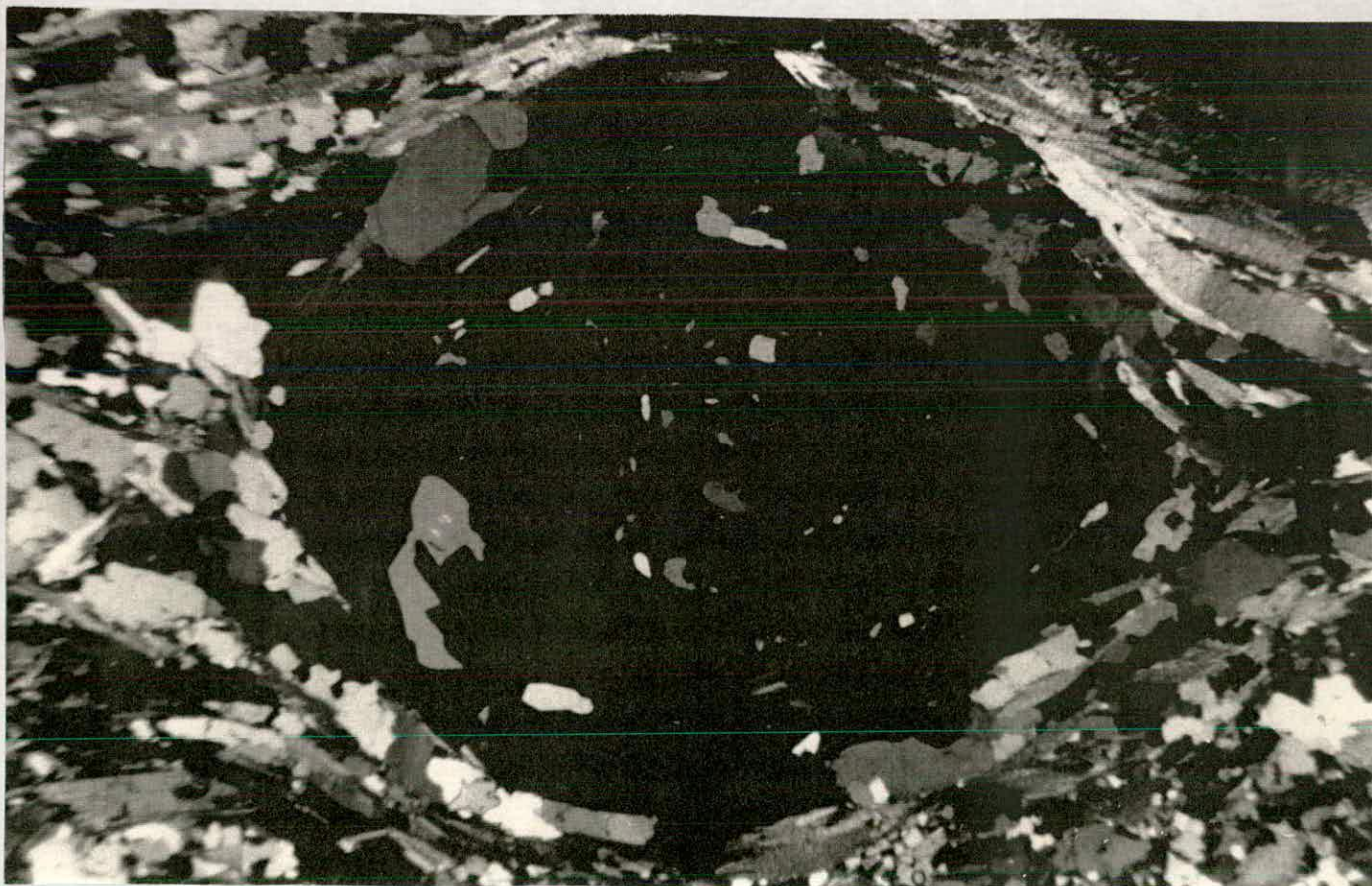
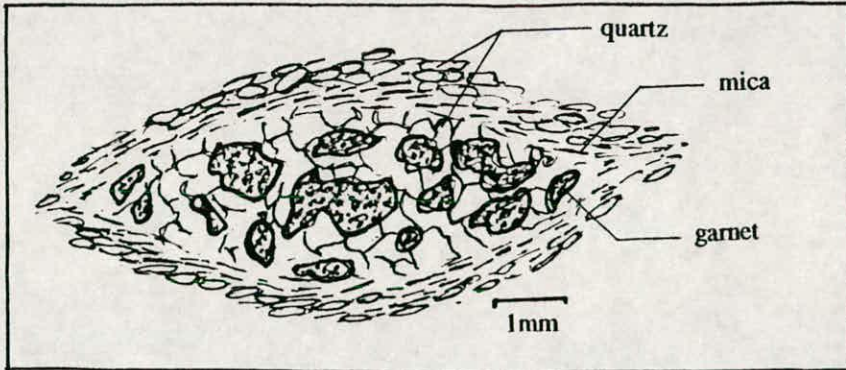
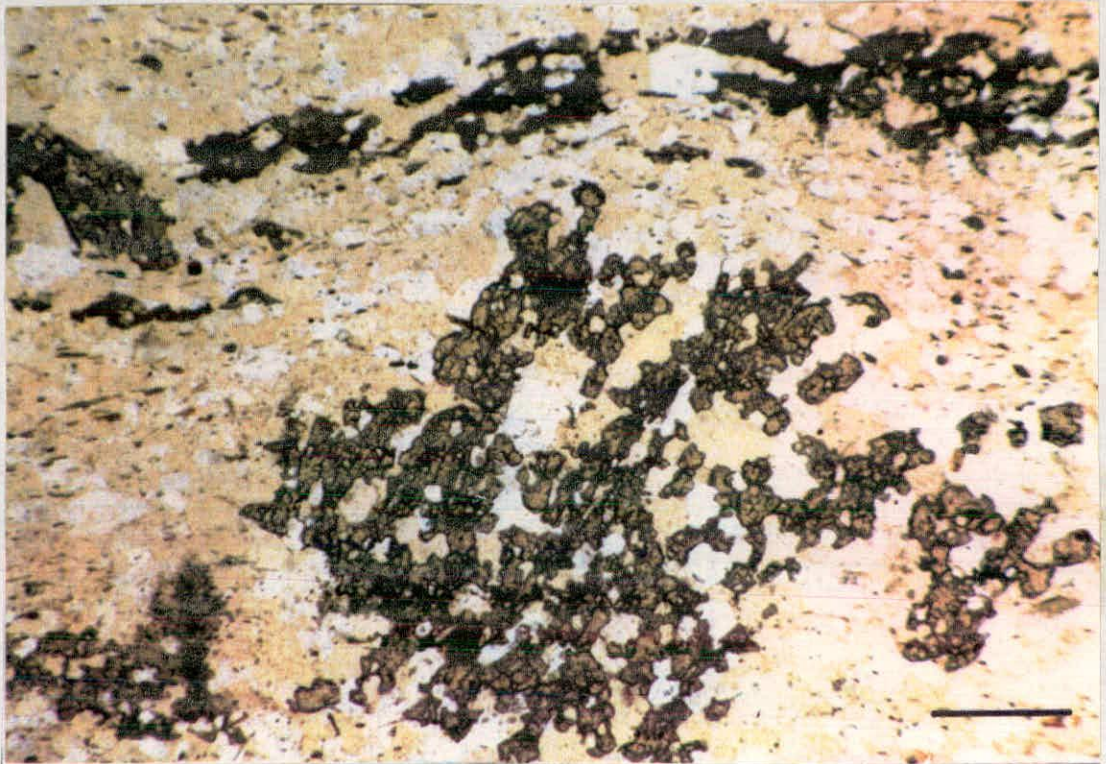


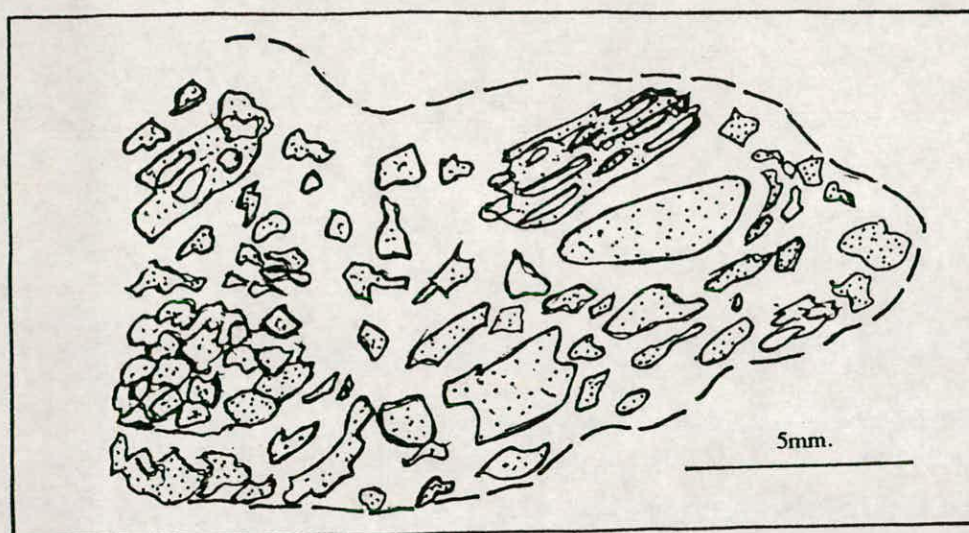
Fig. 5.40: (g)



**Fig. 5.41:** Sketch showing a skeletal garnet porphyroblast whose overall outline is distinctly elliptical (i.e. stretched). See text for further details. Class Ia garnet of Munsiri Formation (from thin section - NR81).



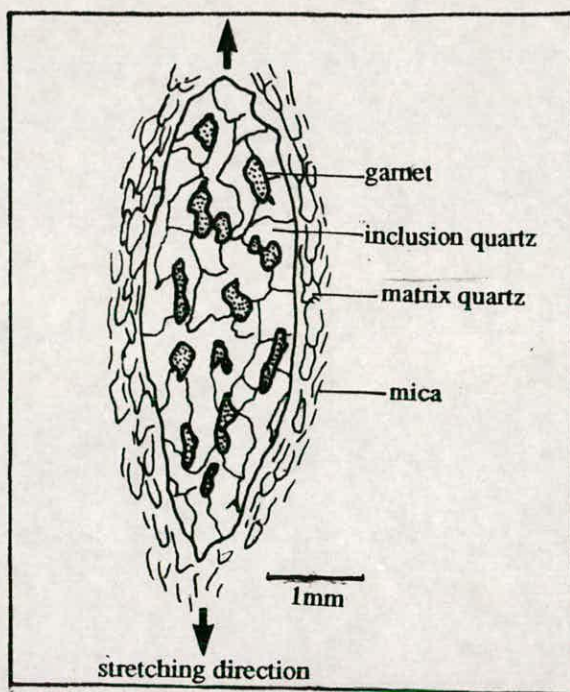
**Fig. 5.42:** Photomicrograph showing large part of a skeletal garnet from an amphibolitic rock. The skeletal nature of the porphyroblast is due to the occurrence of a large number of almost matrix-sized inclusions. The inclusions crudely lie in straight trails oblique to the foliation. But the overall outline of the porphyroblast is elliptical with its elongation parallel to the stretching direction. Dark greenish grains above and at lower left are of amphibole. Munsiri Fmn (thin section - 39G'87). PPL; scale bar 0.5mm.



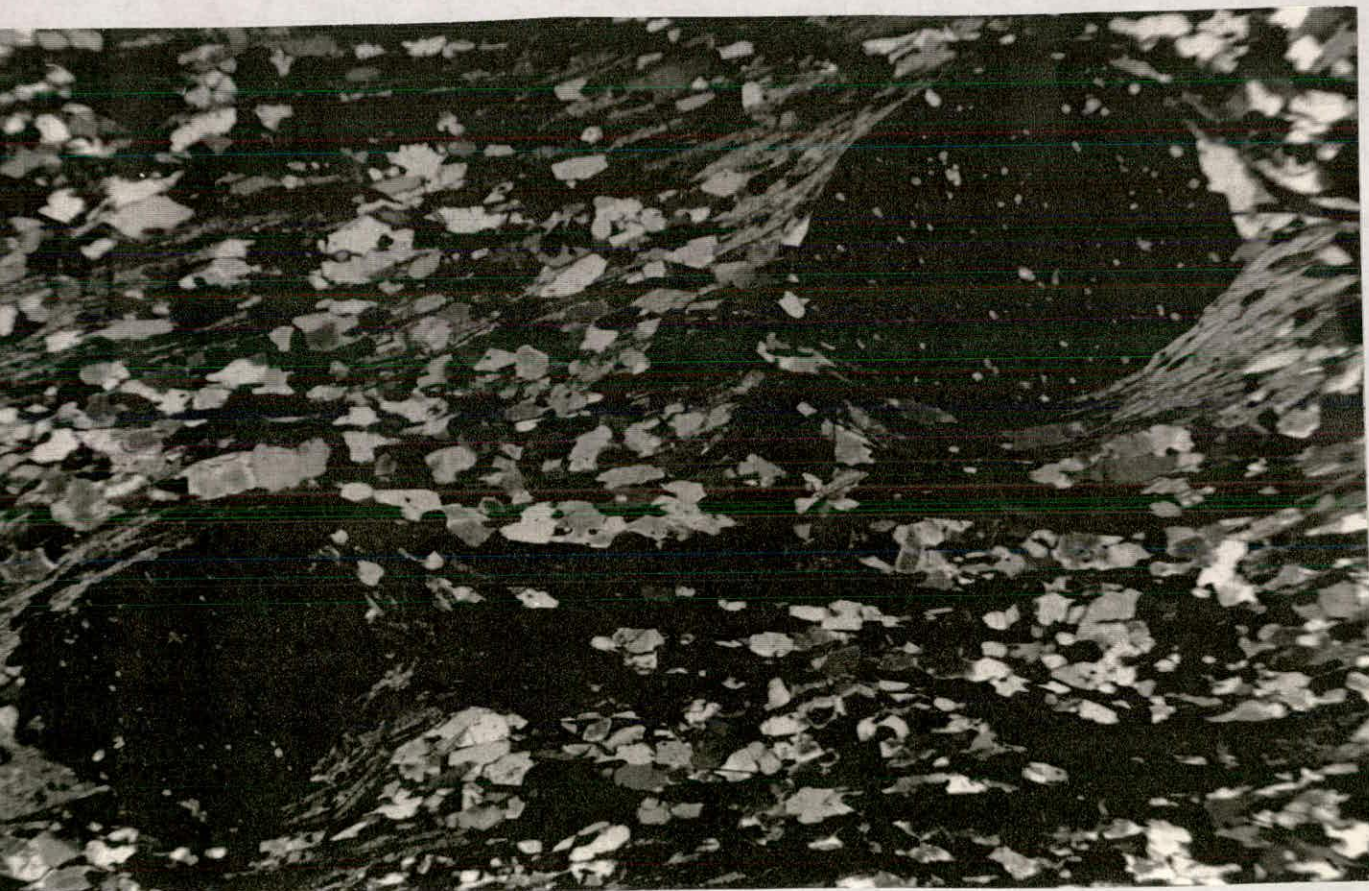
**Fig. 5.43:** Sketch showing part of a large meshy garnet from an amphibolitic rock. See text for further details.



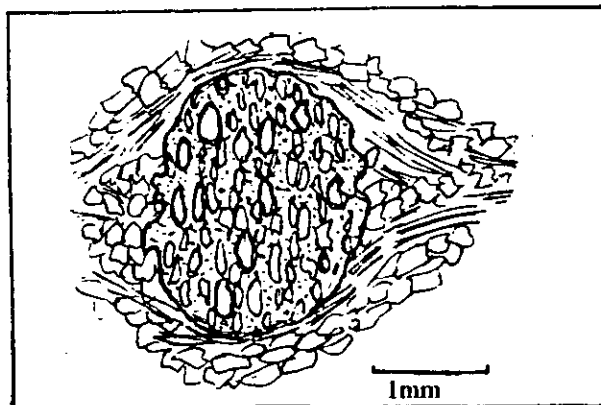
**Fig. 5.44:** A spongy elliptical garnet (g) has been pulled apart due to stretching. Note the slight inward turn in the enveloping matrix foliation at the pull-apart zone (which looks very similar to a 'boudin neck zone'). So the garnet is definitely pre-stretching in origin. For further details see text. (a = amphibole, q = quartz). Photomicro., PPL. Scale bar 1mm.



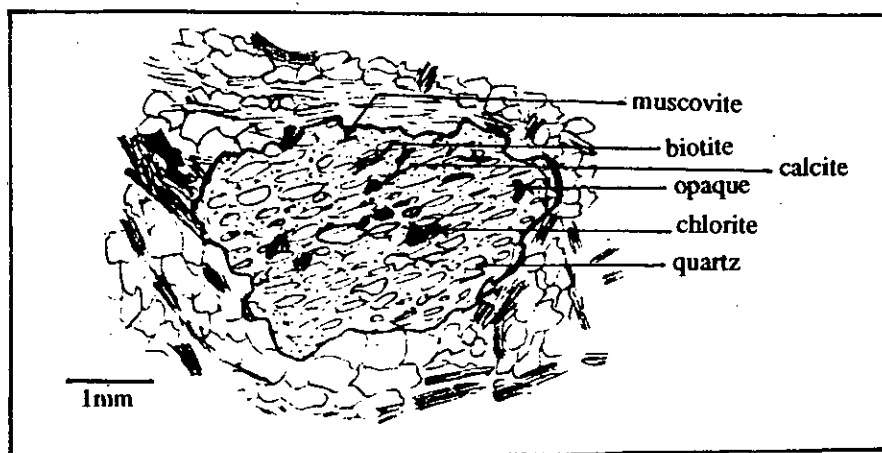
**Fig. 5.45:** Sketch of a skeletal garnet porphyroblast with elongate elliptical outline parallel to the stretching direction. Class I(a) porphyroblast from Munsiri Formation (thin section - 6E'87), See text for details.



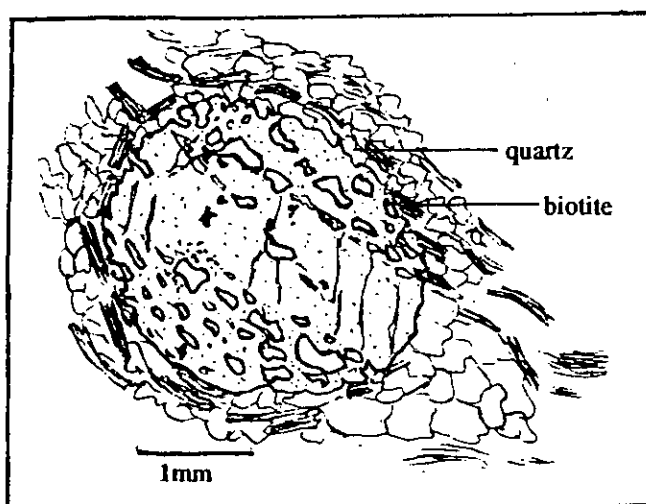
**Fig. 5.46:** Two elliptical garnet porphyroblasts with straight inclusion trails at high angle to the porphyroblast elongation which is parallel to the trace of main foliation (here, giving the stretching direction). Note the mutual parallelism of the inclusion trails in two porphyroblasts, and the considerably smaller inclusion grain size compared to matrix grain size. Matrix is made up mainly of quartz, feldspar, muscovite, biotite and opaques. See text for detailed discussion. Class I(b) porphyroblast from Munsiri Formation (thin section). Photomicro., XPL. Photo-base 5.5mm.



**Fig. 5.47:** Sketch of a class I(b) porphyroblast of garnet from Munsiri Formation. Straight inclusion trails at right angle to matrix foliation. Inclusion grainsize almost comparable to matrix grainsize. Note that Figs. 5.48, 5.49 & 5.50 also come from the same thin section (NR 40/41). See text for details.



**Fig. 5.48:** A class I(b) garnet porphyroblast with an irregular trapezoid outline. There are inclusions of quartz, muscovite, biotite, calcite, chlorite and opaques in the garnet. Inclusions are clearly smaller than the matrix grains. See text for details. Compare Figs. 5.47, 5.49 & 5.50 which also come from the same thin section - NR40/41, Munsiri Formation.



**Fig. 5.49:** In this ovoid garnet porphyroblast the inclusion grainsize is comparable to matrix grainsize. Here Si-Se relation is important. Though the biotites are found to wrap around the porphyroblast, some quartz grains project from the matrix into the porphyroblast. See discussion in text. Munsiri Formation (thin section - NR40/41).

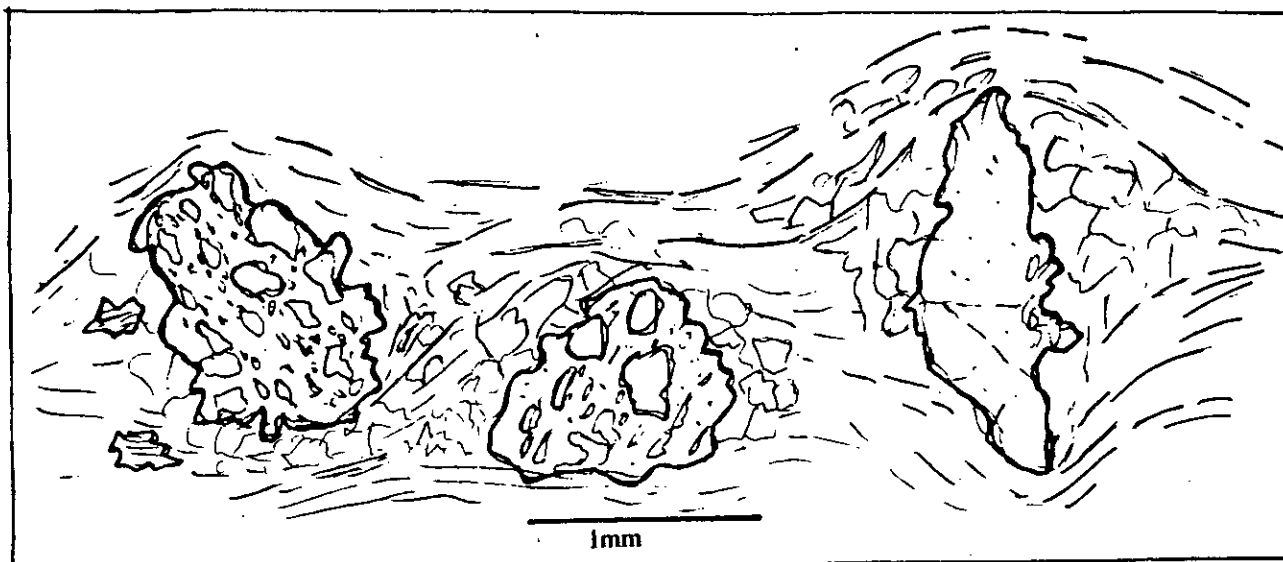


Fig. 5.50: A sketch showing three garnet grains occurring side by side from thin section - NR40/41. All three of them appear to have grown prior to the formation of the enveloping matrix foliation.

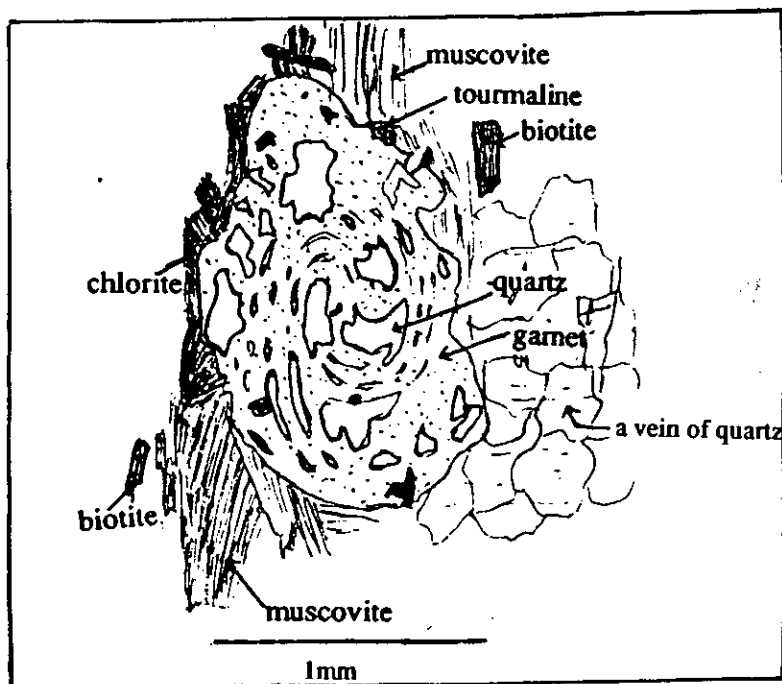
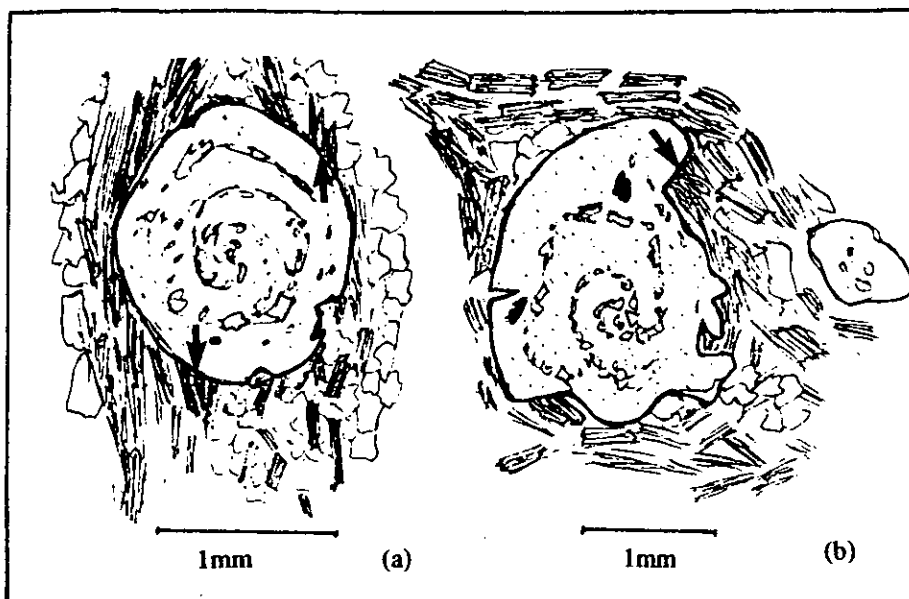


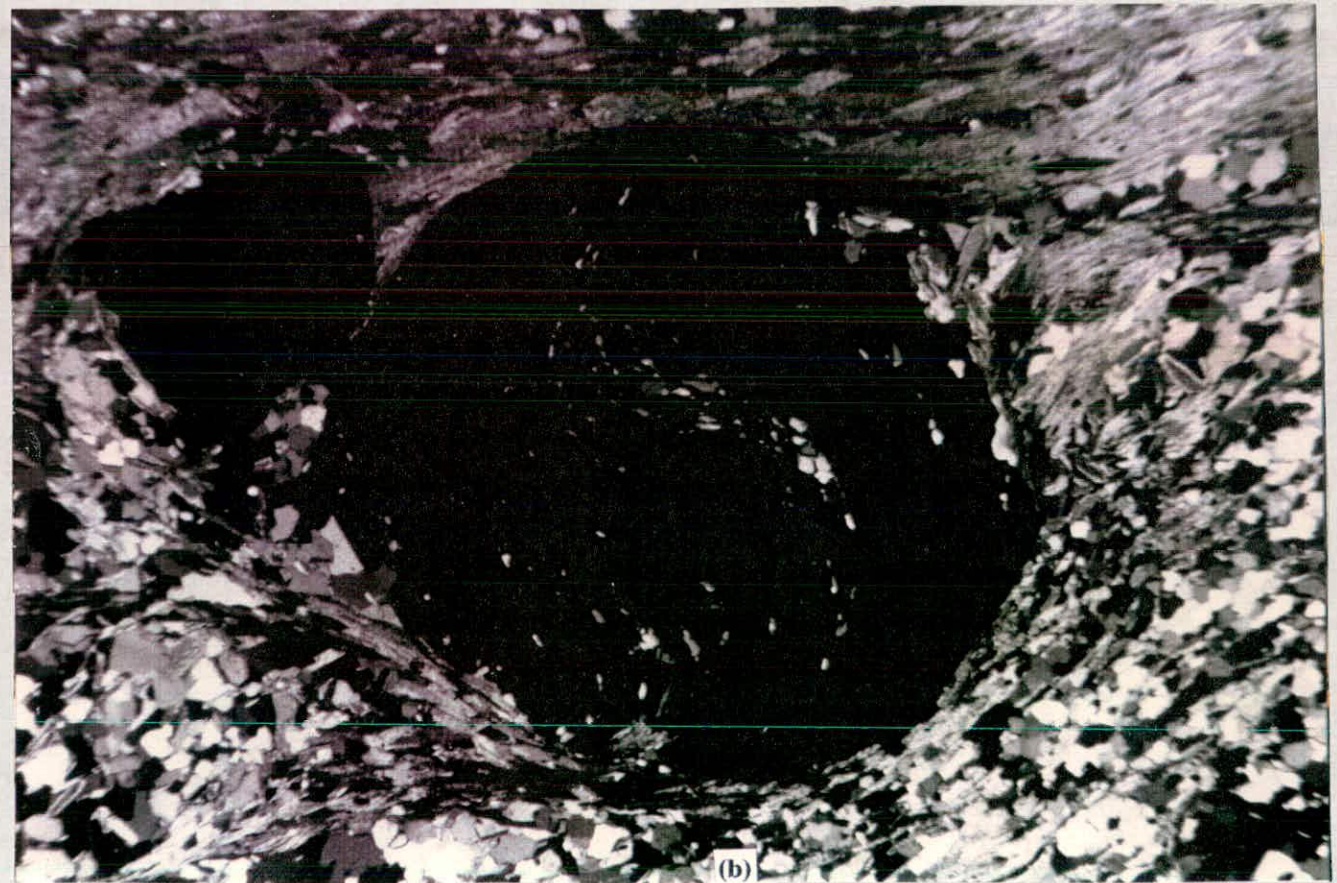
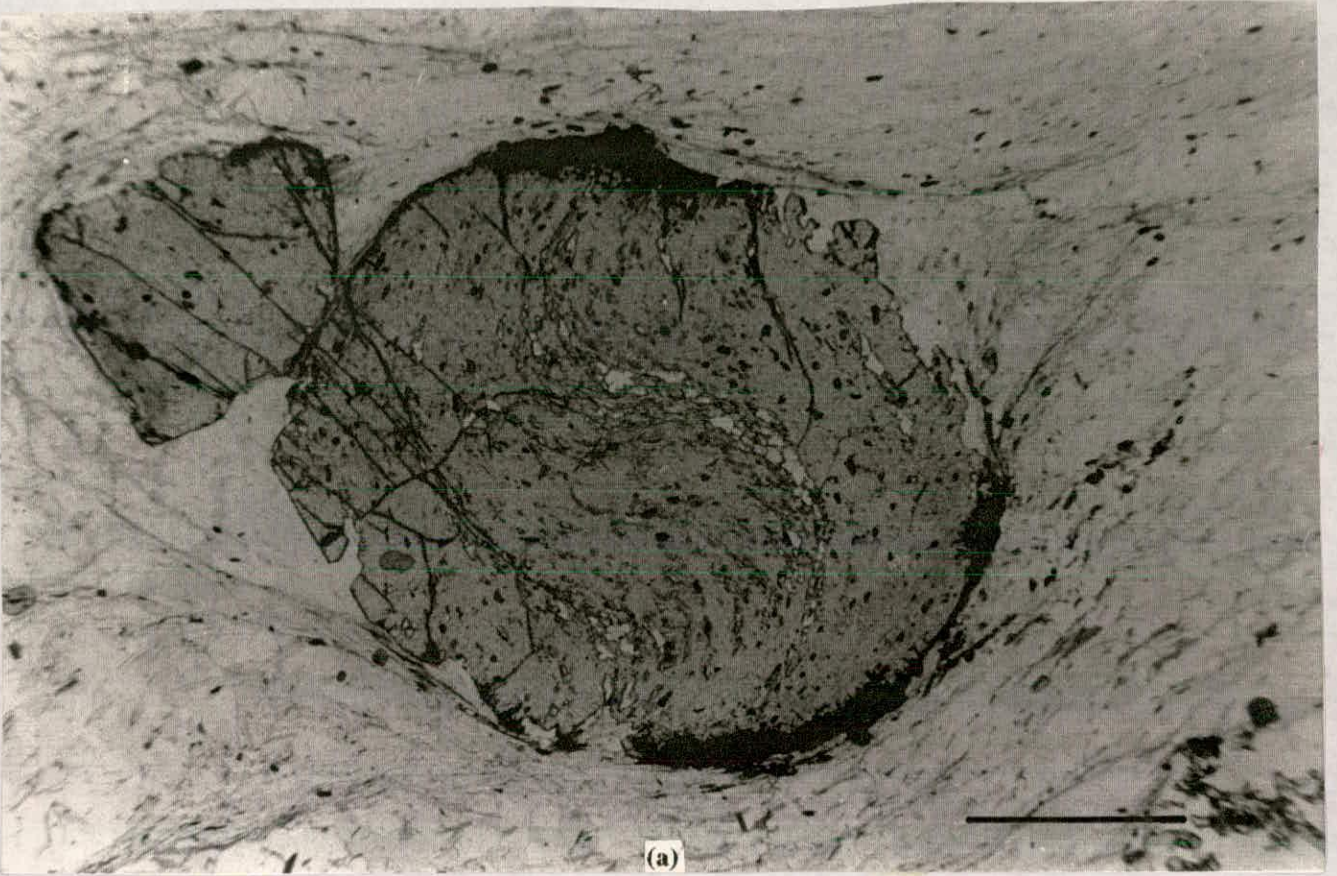
Fig. 5.51: Sketch showing a class I(c) garnet from a thin section of YZ-orientation. Yet it shows typical spiral trails of inclusions. See text for discussion. Munsiri Formation (thin section - 7Pb'87).

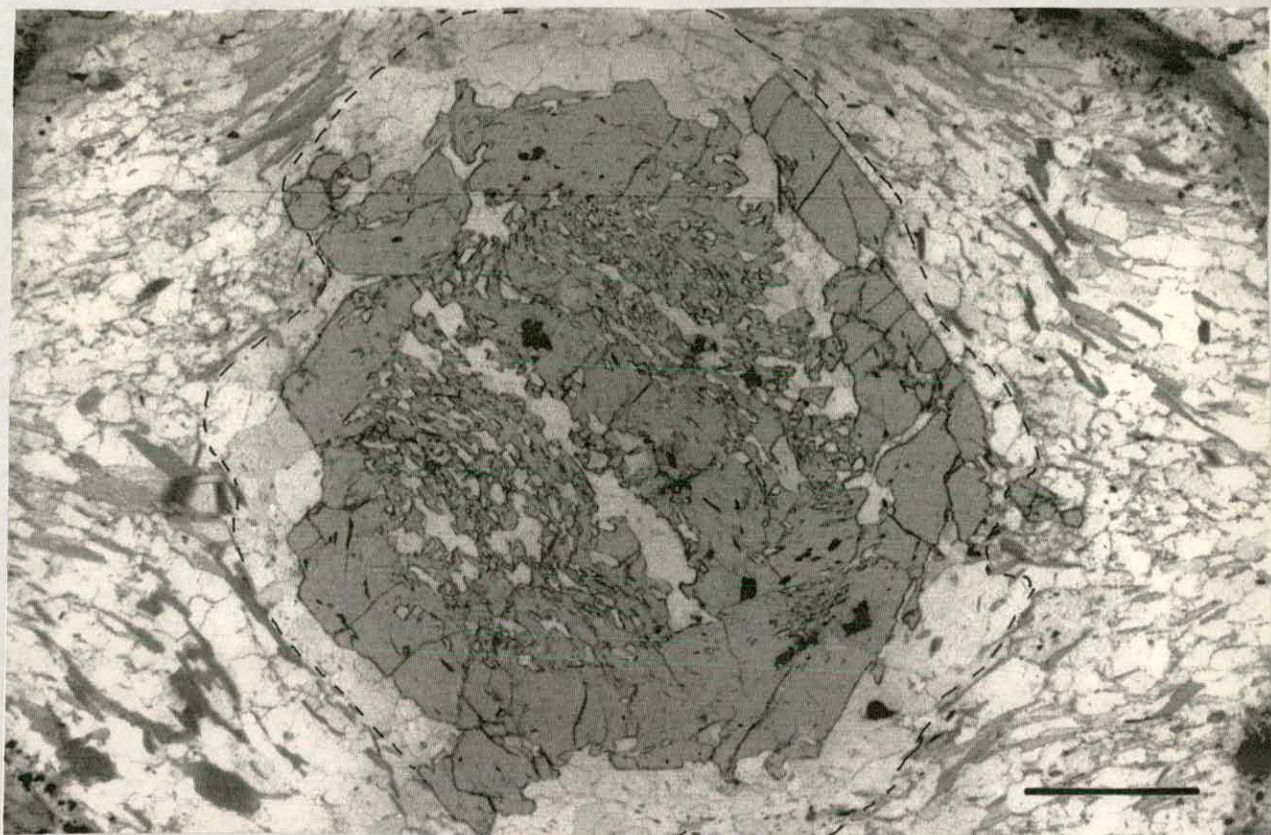


**Fig. 5.52:** (a) & (b). Two class I(c) porphyroblasts of garnet from thin section - NR52 are put side by side. Both show well defined spiral inclusion trails implying possibly a rotational origin of the porphyroblasts. But, intriguingly, the sense of rotation of the inclusion spires is different in the two porphyroblasts. Note the wrapping by matrix foliation is not very conspicuous; some mica grains abut against (arrowed) the porphyroblasts. Further details in text.

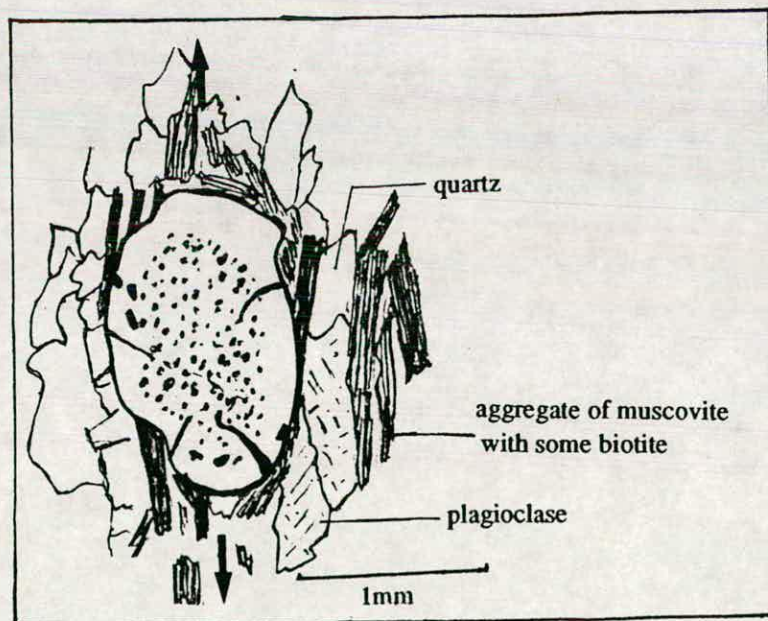
**Fig. 5.53:** (a) & (b). Respectively PPL & XPL photomicrographs (B/W) showing a garnet porphyroblast with spectacular S-shaped inclusion trails implying rotation (anticlockwise, giving locally a 'top-to-left' sense of shear). The small adjoining garnet at upper left is possibly a separate grain; however, the possibility of its existence as a restricted overgrowth to the bigger grain cannot be totally ruled out (note, in PPL-photo (a), the continuation of fractures between the two grains). Upper part of Munsiri Formation (thin section - 6A'87). Scale bar in (a) is 1mm.

( figures are on next page )

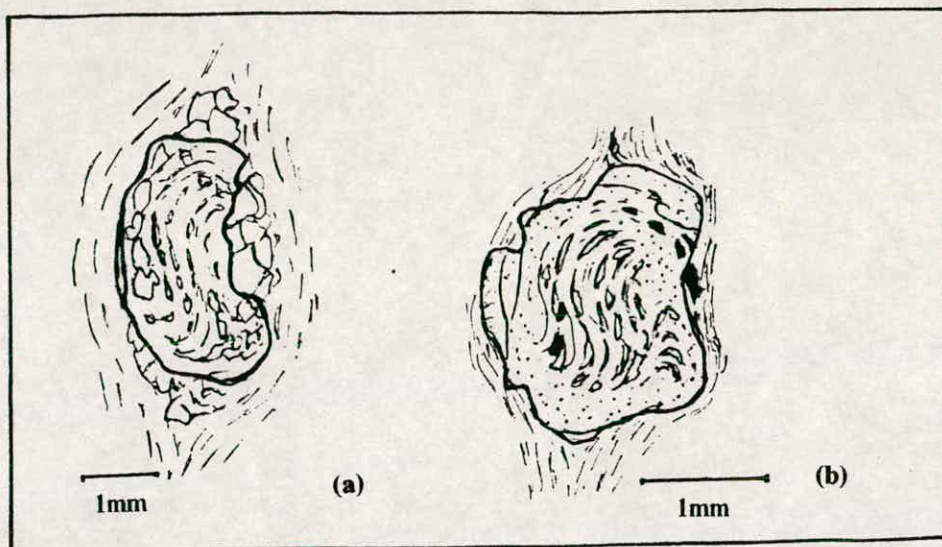




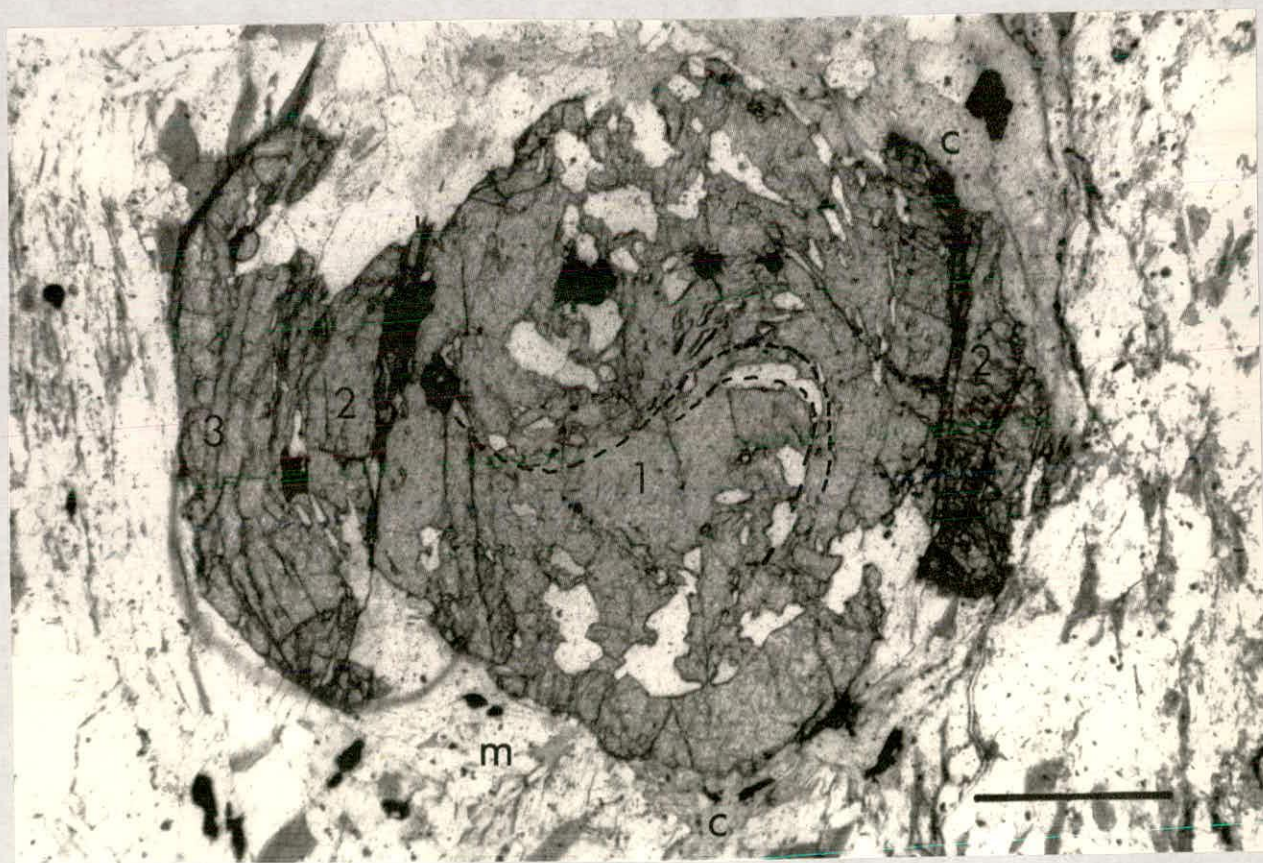
**Fig. 5.54:** A typical example of class II(a) garnet with at least two growth stages. A fairly homogeneous inclusion-free outer rim surrounds an inner zone bearing curvilinear inclusion trails. The inclusion trails define 'millipede' geometry (Bell, 1985). Note the presence of a whitish, quartz-rich and mica-free 'corona' (indicated by dashed line) immediately surrounding the garnet porphyroblast (cf. also Fig. 5.61). Main foliation trace parallel to length of photo. Munsiri Formation (thin section-NR39/40A). B/W photomicro., PPL. Scale bar 1mm.



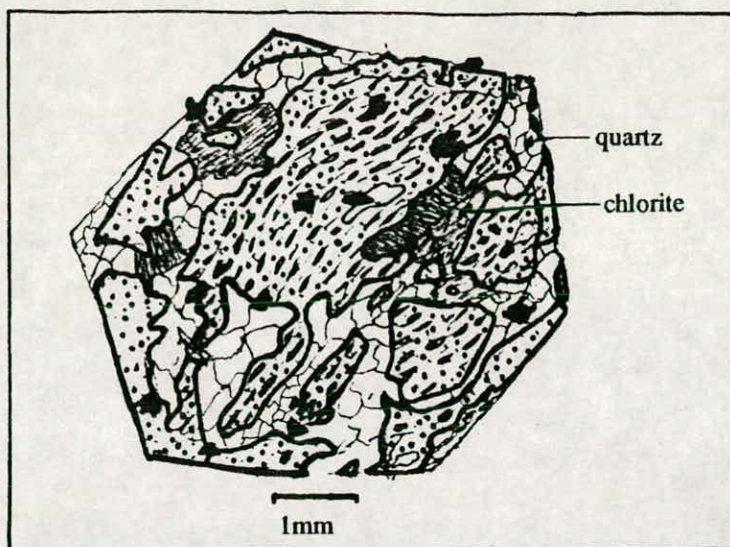
**Fig. 5.55:** Sketch showing an elliptical garnet porphyroblast with elongation parallel to stretching direction (arrows). A smooth i.e. inclusion-free outer zone and an inclusion-studded inner zone are present in the porphyroblast. Note the absence of any linear or spiral arrangement of inclusions. Possibly there were two growth stages of garnet (class II(a)). Munsiri Formation (thin section NR39'').



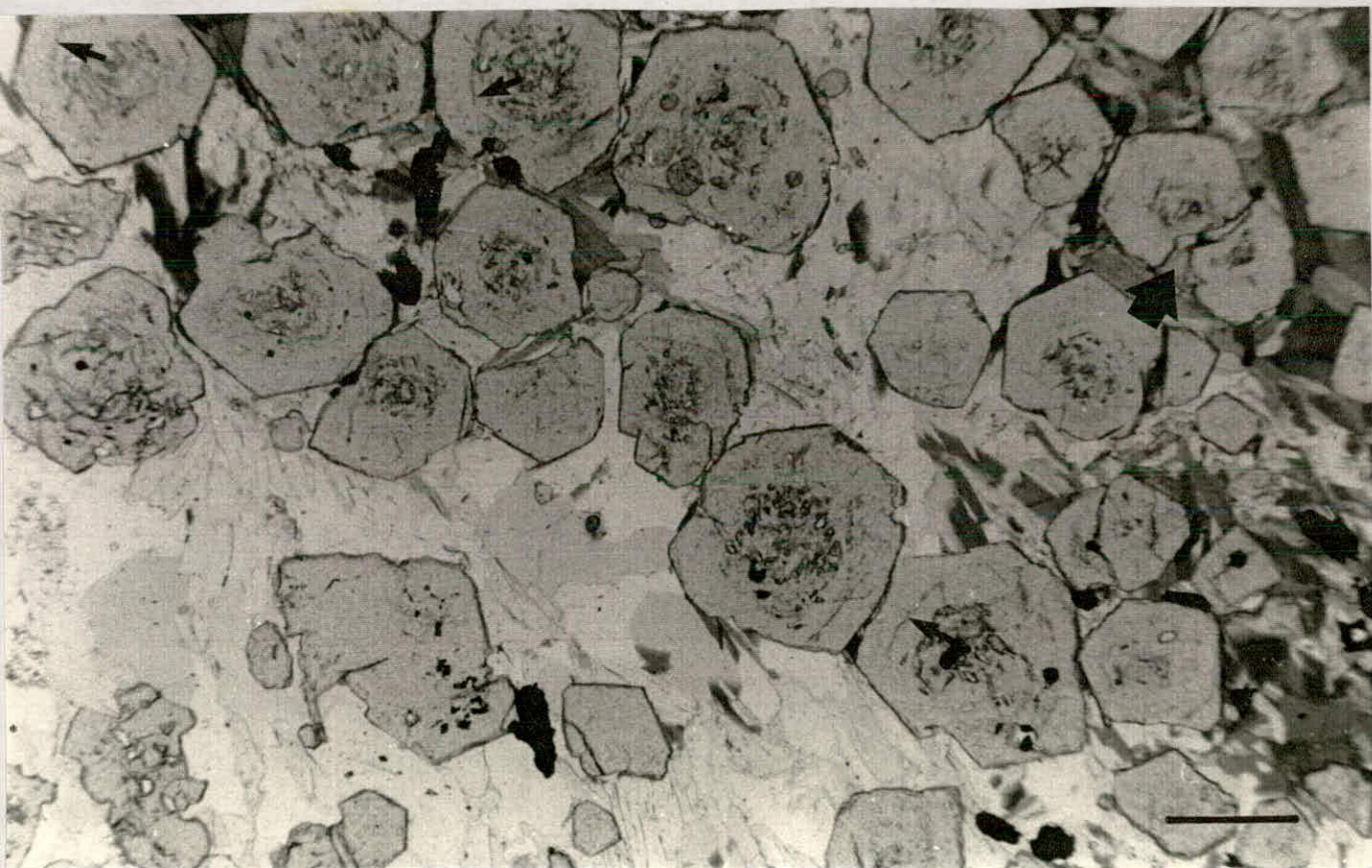
**Fig. 5.56:** (a) & (b). Sketches of two class II(a) garnet porphyroblasts from thin section - 6Eb'87 (an YZ-section). Two growth stages of the porphyroblasts are indicated; distinct sigmoidal inclusion trail-bearing wide inner zones are surrounded by relatively thin, partially preserved homogeneous outer rims. See text for details.



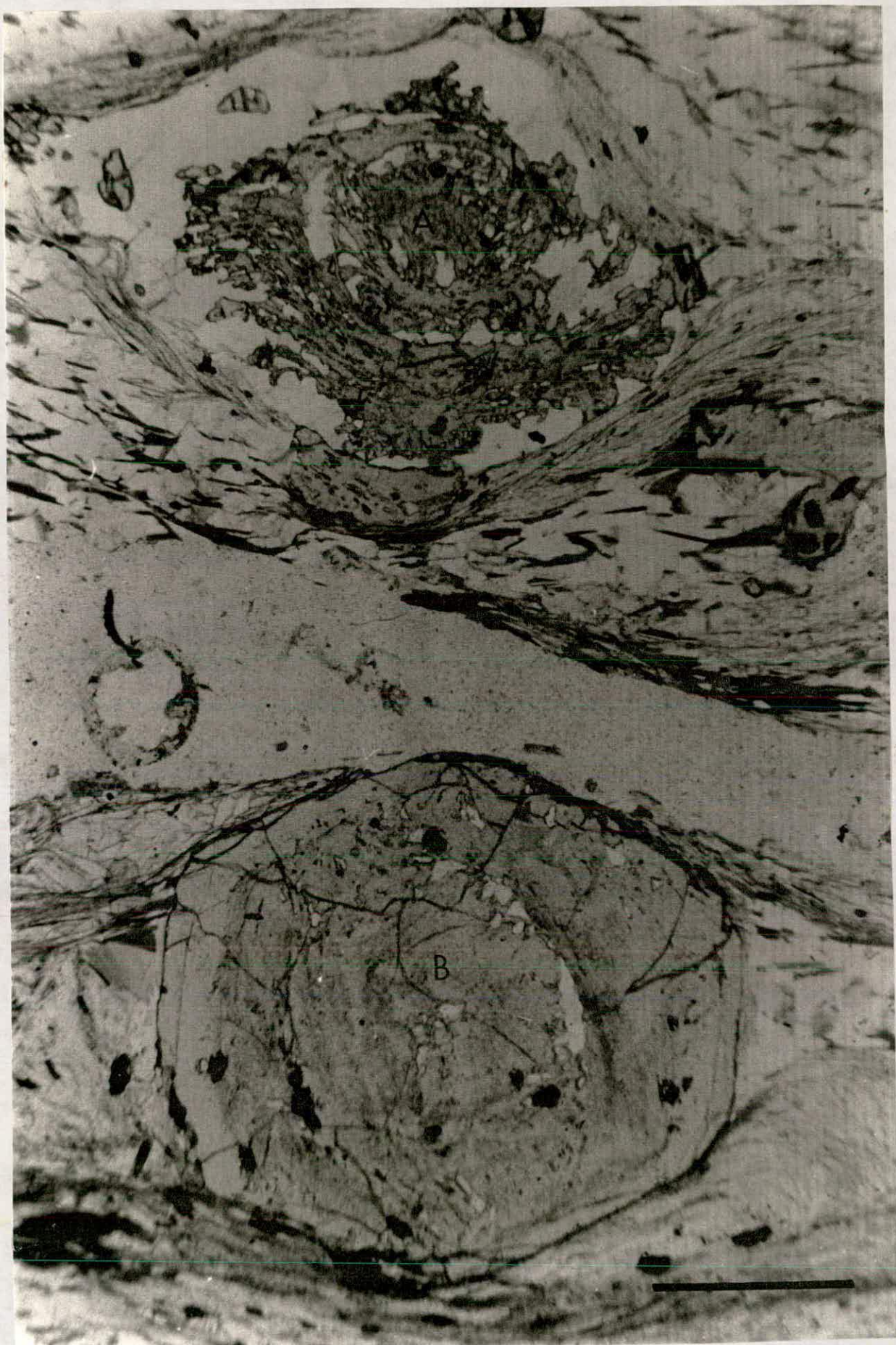
**Fig. 5.57:** Photomicrograph of a class II(a) garnet showing at least two, or possibly, three stages of growth (1, 2, 3). Sigmoidal inclusion trails are present only in the earliest growth stage (1). Retrogression into chlorite (c) is seen at upper right and bottom of porphyroblast. Trace of main foliation parallel to breadth of photo. Further details in text. m = muscovite. Upper part of Munsiri Formation (thin section-7C'87). PPL; scale bar 0.5mm.



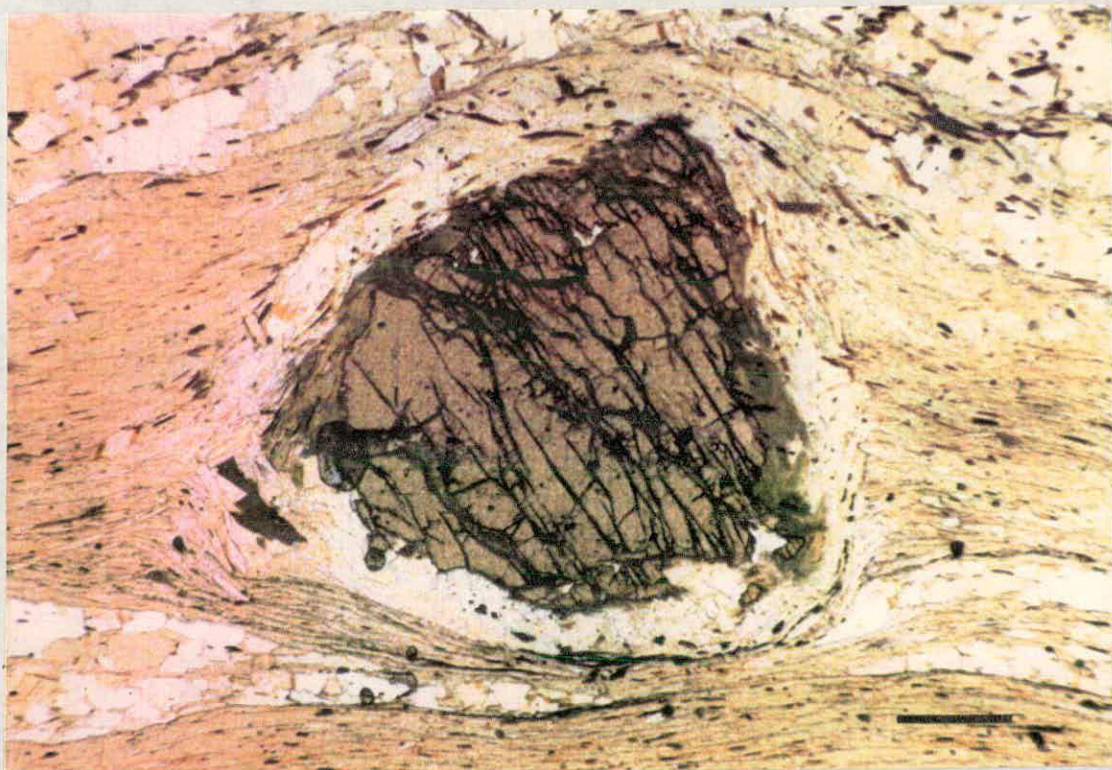
**Fig. 5.58:** Sketch of a remarkably euhedral, yet meshy garnet (class II(b)) from an amphibolite. Body texture of the garnet is quite intriguing. However, two growth zones are clearly distinguishable. The outer rim (present as detached zones) is inclusion-free; the comparatively large and embayed inner zone has sigmoidal inclusion trails. Parts of garnet are altering into chlorite. Munsiri Formation (thin section - 23/88). Scale bar 1mm.



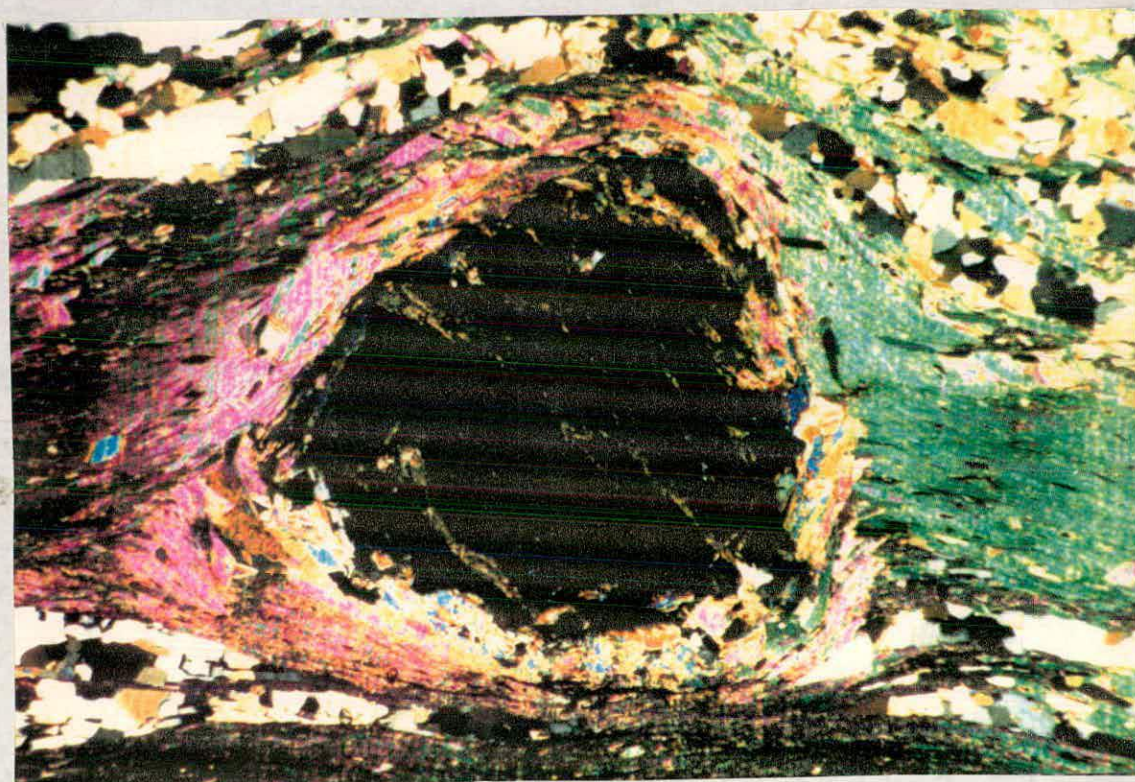
**Fig. 5.59:** Photomicrograph (PPL) showing a group of euhedral to subhedral garnet porphyroblasts showing multi-stage growth. Inclusion-bearing inner zones (cores) are surrounded by inclusion-free homogeneous outer zones. Growth rings/lines (i.e. zonal boundaries) are very clear in some porphyroblasts (arrowed). Though here most garnet grains are equidimensional i.e. not elliptical, pull-apart effect is present in some (an example indicated by bigger arrow). Munsiri Formation (thin section 8G'87). Scale bar 0.5mm.



**Fig. 5.60:** Photomicrograph (PPL) showing two adjacent garnet porphyroblasts with very different textures; one (A) is class II(a) and has very irregular outline and complexly arranged ('splayed') inclusion trails, the other (B) has well developed crystal outline and a fairly homogeneous body with only a few visible inclusions. A fracture runs across the middle of the thin section. See text for detailed discussion on the two porphyroblasts. Upper part of Munsiri Formation (thin section 7G'87). Scale bar Imm.

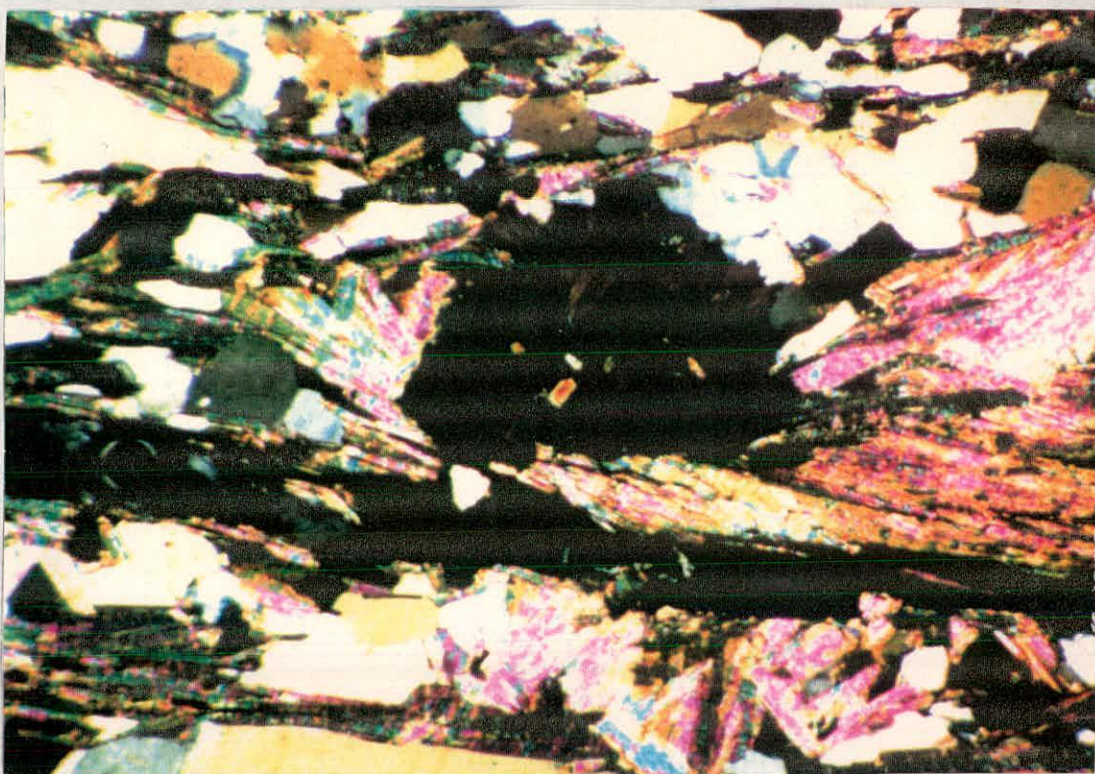


(a)

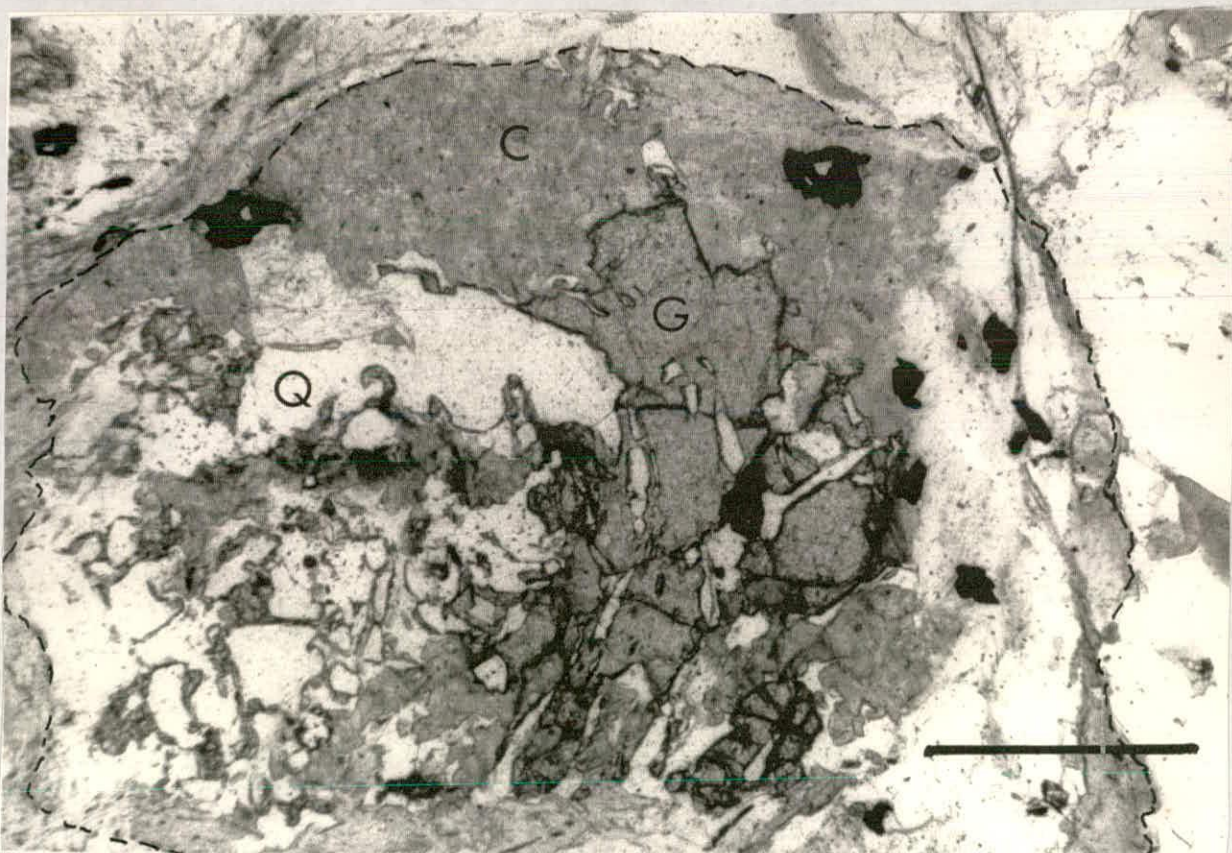


(b)

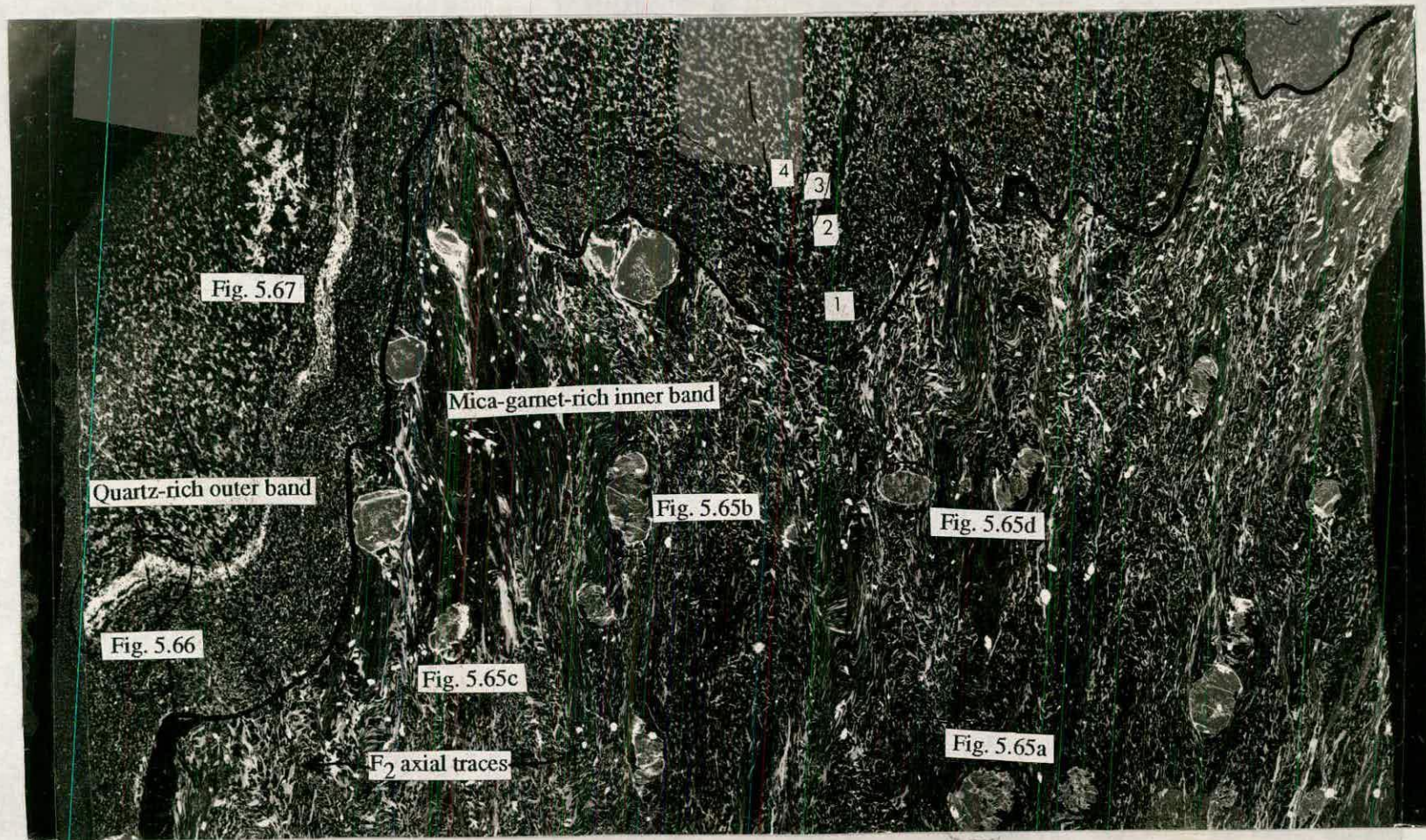
**Fig. 5.61:** (a) & (b). Respectively PPL & XPL photomicrographs showing a class III garnet porphyroblast which gives a spectacular example of active retrogression. Chloritisation is affecting the porphyroblast from the rim inward and along the fractures. Note the presence of a zone ('corona') of white mica bordering the altered garnet (cf. Fig. 5.54 where a quartzose border surrounds a garnet porphyroblast). Muniari Formation (thin section 64A'87). Scale bar in (a) is 0.5mm.



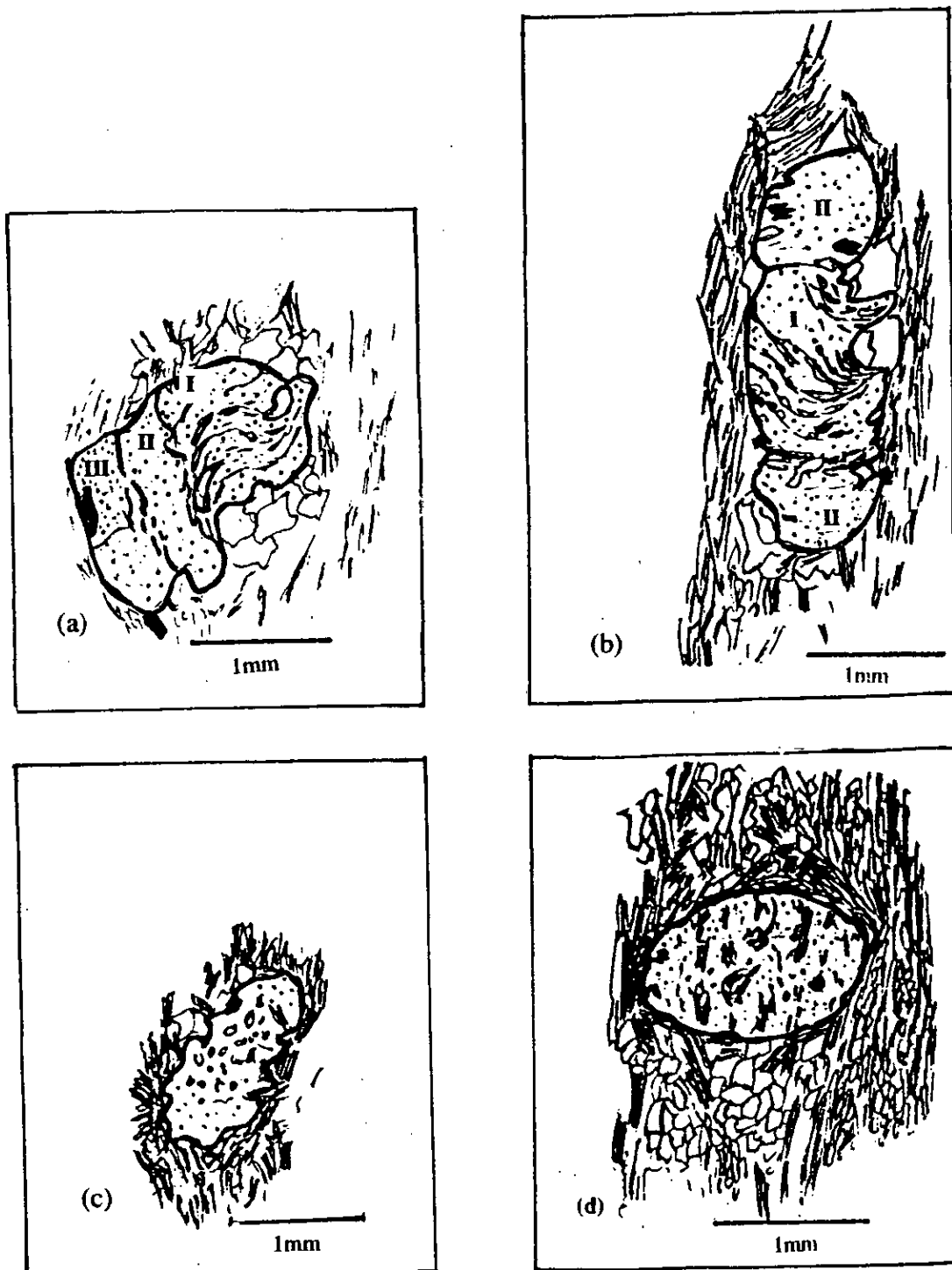
**Fig. 5.62:** Photomicrograph (XPL) shows at the centre a possible example of a wholly chloritised garnet. The chlorite pseudomorph is surrounded mostly by white mica (muscovite). Note the existence of a peneconcordant vein of semi-opaque (?graphitic) material below the pseudomorph. Same thin section as above. Photo-base 2.25mm.



**Fig. 5.63:** Photomicrograph (PPL) of a class III garnet porphyroblast whose area (size) has diminished considerably due to alteration (chloritisation). The remnant garnet, partly skeletal in nature, is preserved as "islands in a 'lake' of chlorite". The outline (dashed) of the 'lake' probably gives the original outline of the garnet porphyroblast and thus suggests a high degree of resorption of the grain. G = garnet, C = chlorite, Q = quartz. Munsiri Formation (thin section 7C'87). Scale bar 0.5mm.



**Fig. 5.64:** A negative print of whole thin section - 7M'87 (upper Munsiri psammopelitic schist). The thin section preserves a broad  $F_2$  fold-hinge with several subsidiary (lower order) hinges, and shows interesting garnet textures. Location of Figs. 5.65 to 5.67 as well as some other relevant features are indicated on the overlay. Further details in text. (1) sub-band rich in fine quartz, (2) mica-opaque-garnet-quartz-bearing sub-band, (3) apatite-garnet-quartz-bearing sub-band, (4) sub-band rich in coarse quartz. Length of view (photo) is 3.78cm.

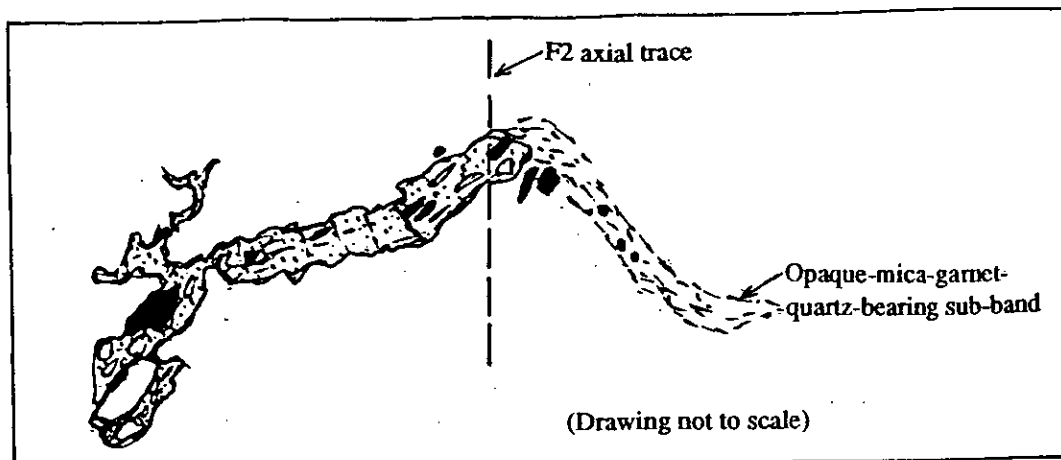


**Fig. 5.65:** (a) A garnet porphyroblast showing two or, more possibly, three growth stages (I, II, III). Please refer to text and Fig. 5.64 for further details.

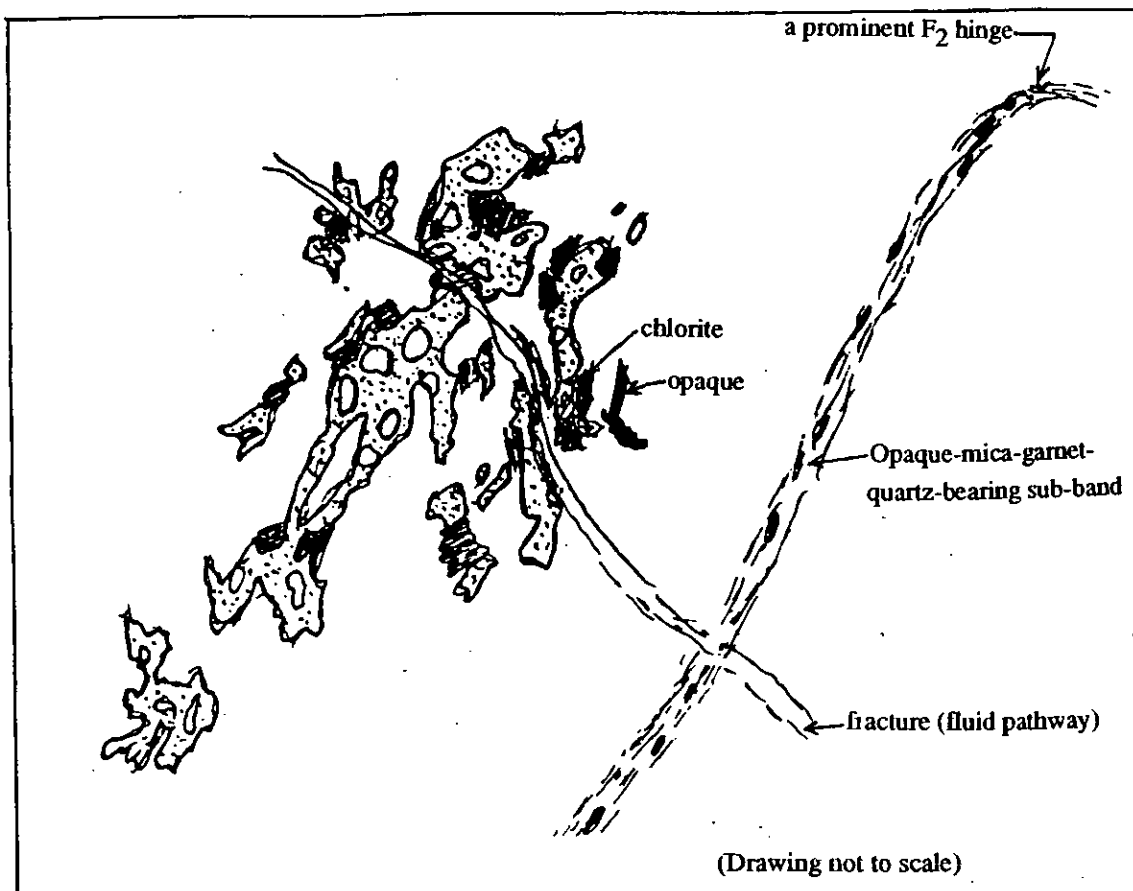
(b) An elongate garnet porphyroblast aligned parallel to the  $F_2$  axial planar foliation shows two growth stages (I & II). The mutual disposition of the growth zones clearly implies favoured growth of the garnet along the foliation.

(c) Micas in the matrix around the garnet porphyroblast are arranged quite baphazardly implying no indentation of the  $F_2$  axial planar foliation by the porphyroblast and hence a broadly syntectonic time of growth of the latter. The foliation trace is vertical in the sketch.

(d) An oval-shaped garnet porphyroblast aligned perpendicular to the  $F_2$  axial planar foliation. Interestingly, the matrix micas generally have a "sympathetic" enveloping relation with the garnet. Note the  $S_i$  in garnet is aligned parallel to the  $S_e$  and some matrix micas abut against the garnet.



**Fig. 5.66:** A long, roughly prismatic grain of garnet aligned along the pre- $F_2$  compositional layering (this layering would be  $S_0$ , if the  $F_2$  axial planar foliation referred to in Figs. 5.64 & 5.65 is considered as  $S_1$ ). Note the slight bending of the porphyroblast across an  $F_2$ -hinge, implying a pre- $F_2$  time of origin of the garnet. Further details in text. For scale see Fig. 5.64.



**Fig. 5.67:** A typical skeletal garnet has a crude overall elliptical outline (see also Fig. 5.64) with long axis nearly parallel to pre- $F_2$  compositional layering. A close inspection of Fig. 5.64 would suggest that the garnet overgrew the relatively fine-scale pre- $F_2$  foliation in the coarse quartz-rich sub-band. For interpretations see text. Note the fracture running across the porphyroblast and chloritisation. For scale see Fig. 5.64.

## Garnet paragenesis

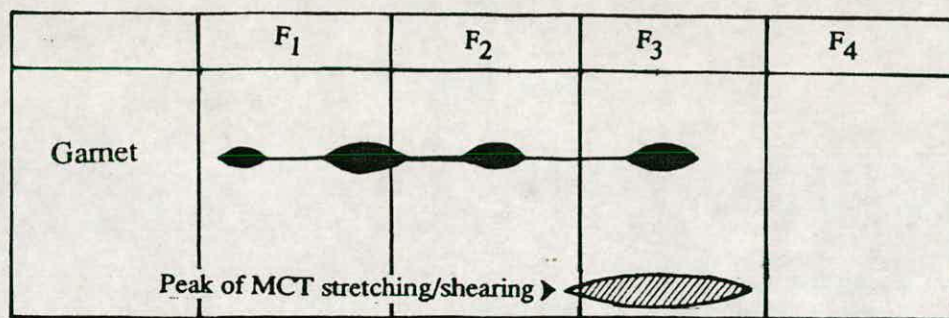


Fig. 5.68: Fish diagram showing garnet crystallisation times in the Munsiri Formation and Joshimath Gneiss with respect to fold episodes.

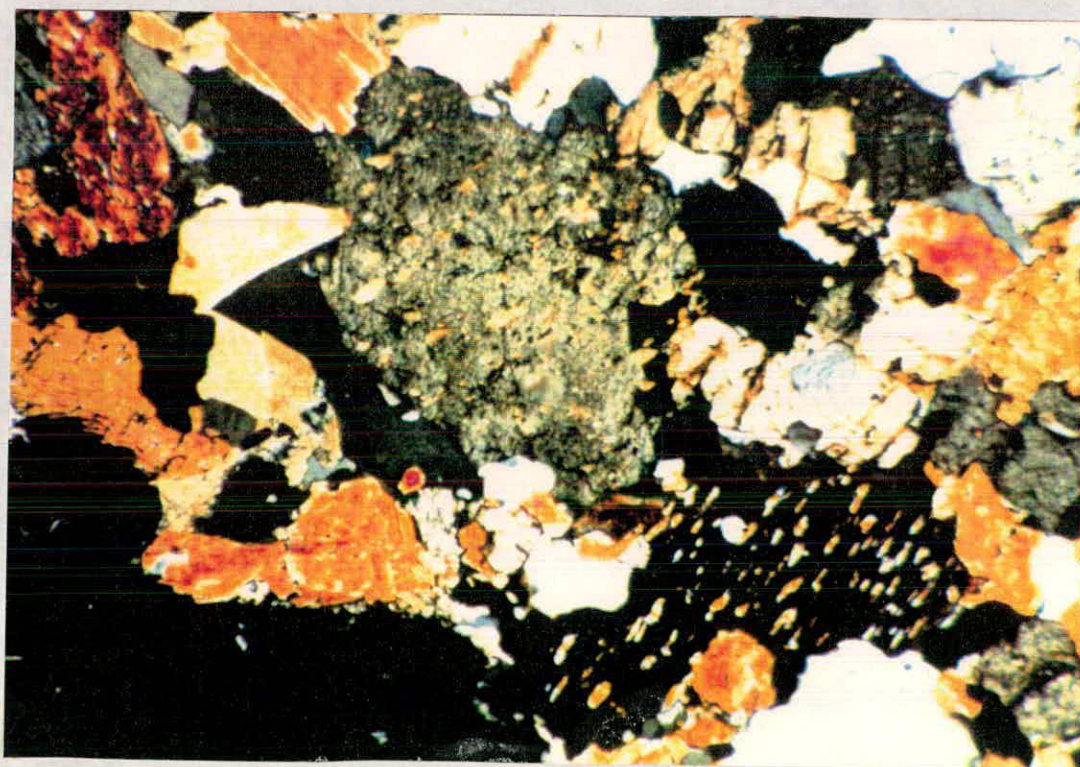


Fig. 5.69: (b) Inclusions of amphibole in clinopyroxene grains (see the large greenish grey grain at centre and the grain under extinction to its lower right). Note the matching colour (yellow) of the inclusions with some discrete amphibole grains outside. Interestingly, all these inclusions in any particular pyroxene grain together come to maximum illumination position between cross polars when the host grain is extinct or nearly so. This suggests that these are oriented 'apparent' inclusions probably indicating incipient alteration of pyroxene into amphibole along cleavage fractures/intersections. Photomicro., XPL. A part of the same thin section as in Fig. 5.69(a). Photo-base 2.8mm.

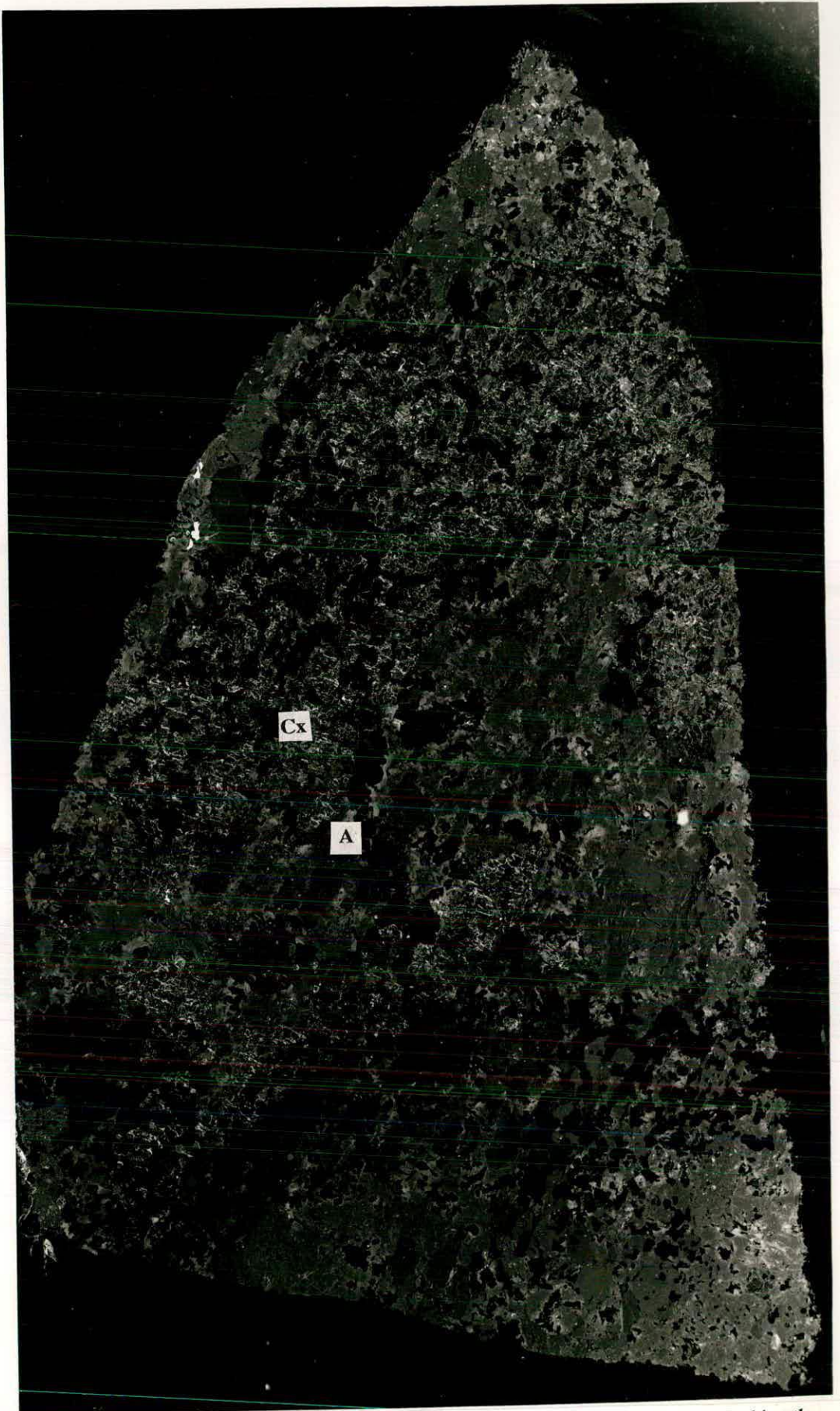
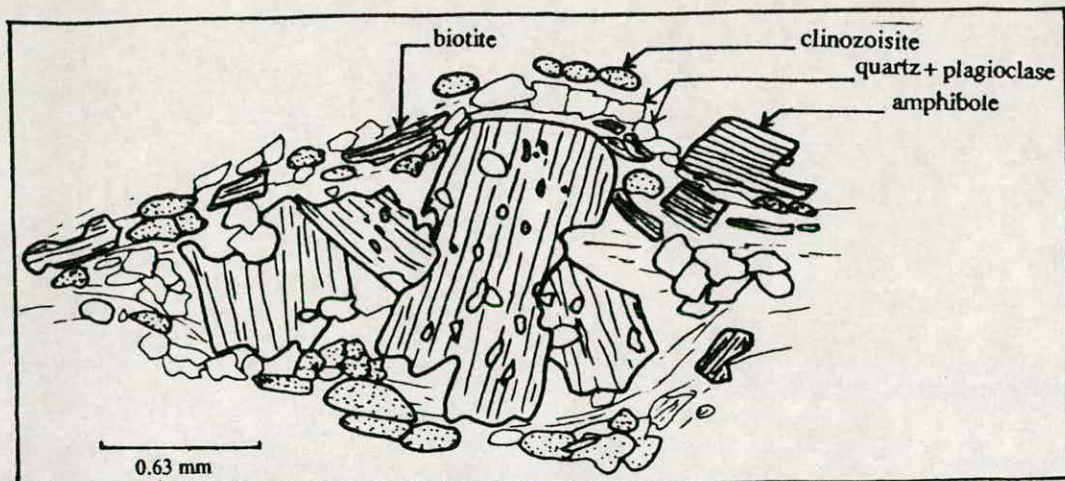
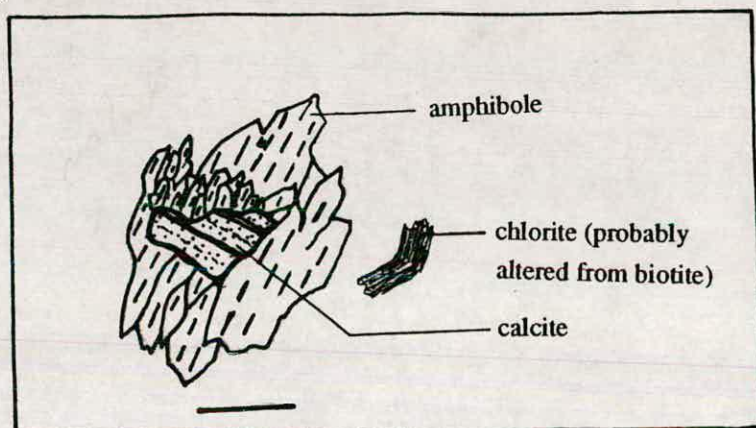


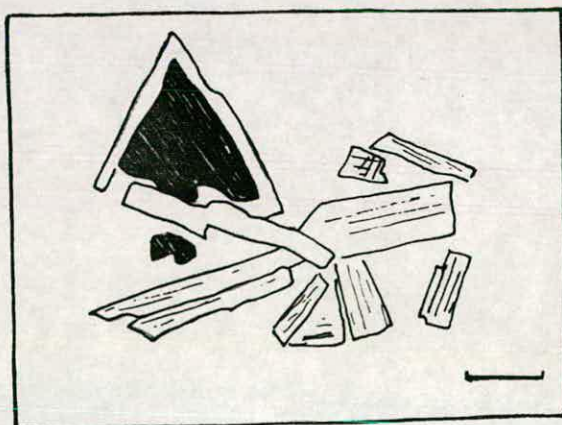
Fig. 5.69: (a) Negative print from thin section (NR 25/26L(2)) of a metabasic horizon in Joshimath Gneiss. The height of the triangular rock section is 3.4 cm. A = amphibole, Cx = clinopyroxene (augite).



**Fig. 5.70:** Sketch showing two size classes of amphiboles. Matrix foliation bearing small amphibole grains and other minerals swerves around the aggregate or cluster of larger porphyroclastic amphibole grains. The porphyroclastic grains are aligned at high angle to the matrix foliation; probably these are older than the matrix amphibole grains. Munsiri Formation (thin section - 15/45G).



**Fig. 5.71:** Sketch showing close-knit cluster of amphibole grains; associated minerals are calcite and chlorite (shown) and clinzoisite, quartz, garnet and opaques (not shown). Munsiri Formation (thin section 8C'87). Scale bar 0.5mm.



**Fig. 5.72:** Clusters of amphibole grains with diversely oriented members defining a star-like pattern. The dark mineral is carbonate. Crude hornfelsic-looking texture is seen in nearby feldspathic patches or bands (not shown in sketch). Munsiri Formation (thin section - 8C'87). Scale bar 0.3mm.

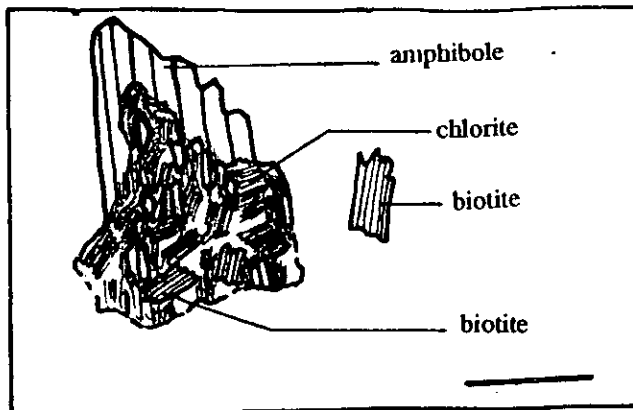


Fig. 5.73: An amphibole grain is being progressively altered into biotite and chlorite. Discrete biotite grains are also present alongside. Munsiri Formation (thin section - 8C'87). Scale bar 0.3mm.

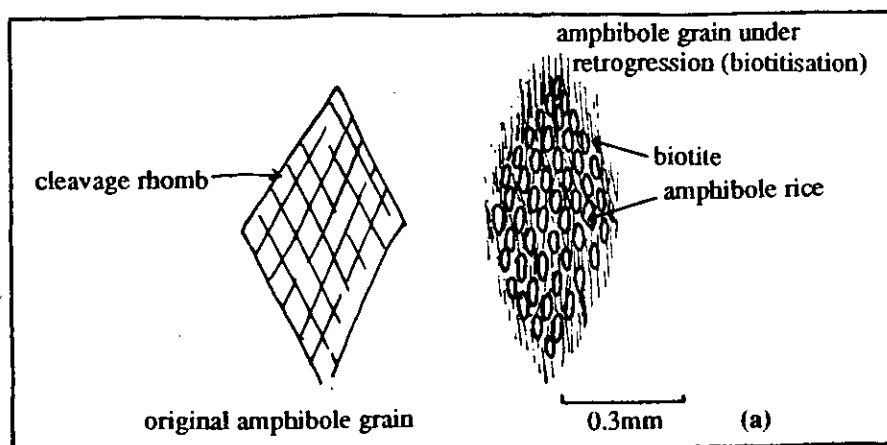


Fig. 5.74 (a): Selective retrogression (biotitisation) following the two cleavage sets in amphibole grain. With progressive retrogression the tiny cleavage rhombs are separated from each other by the intervening biotites and collectively look like a stalk of rice or wheat; however, the original diamond-shaped outline (now elliptical) of the whole grain is still recognisable. Munsiri Formation (thin section - 17Rb'87).

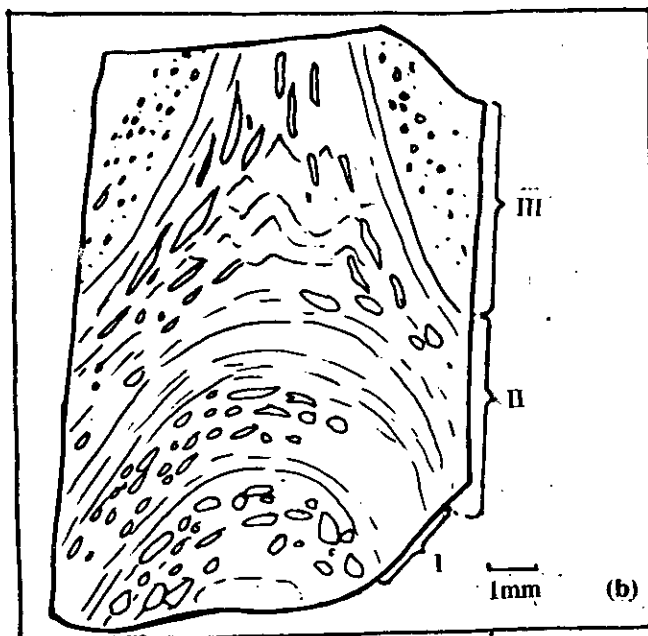
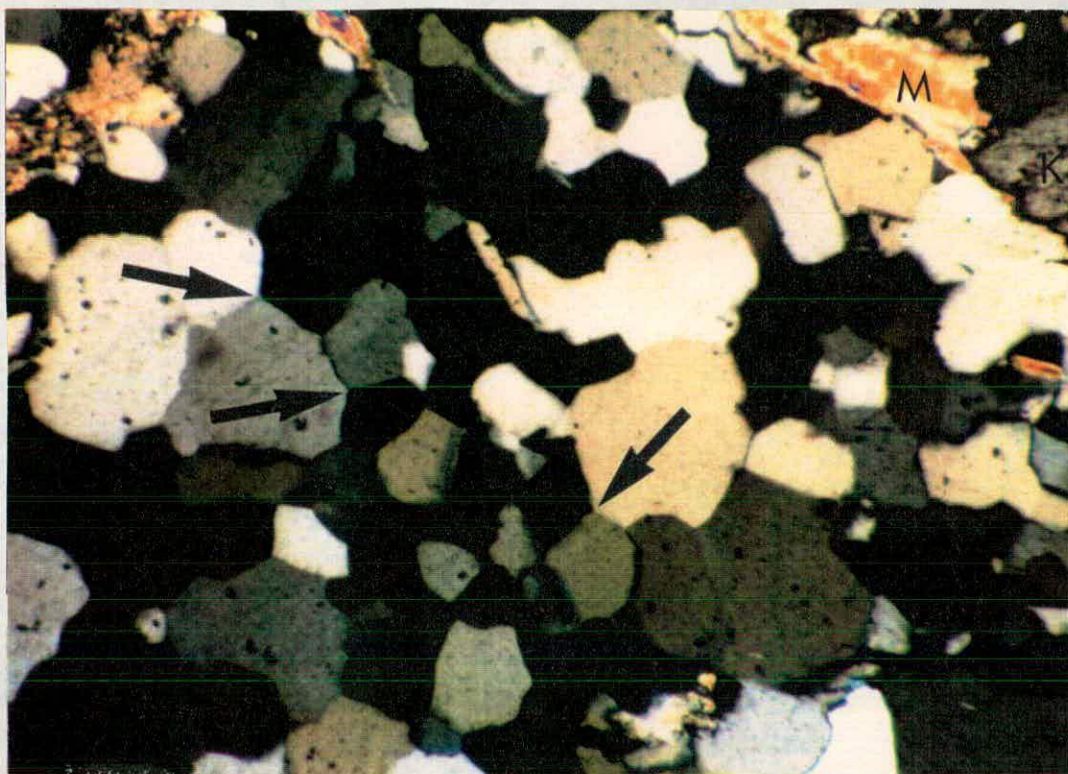
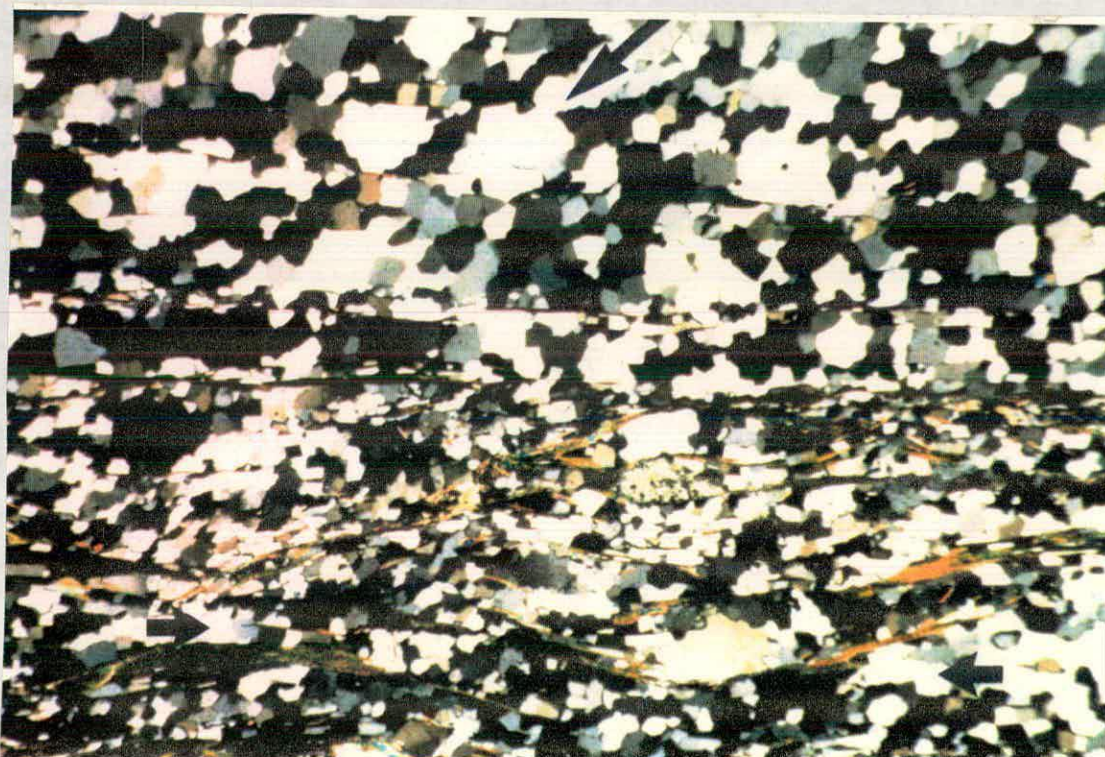


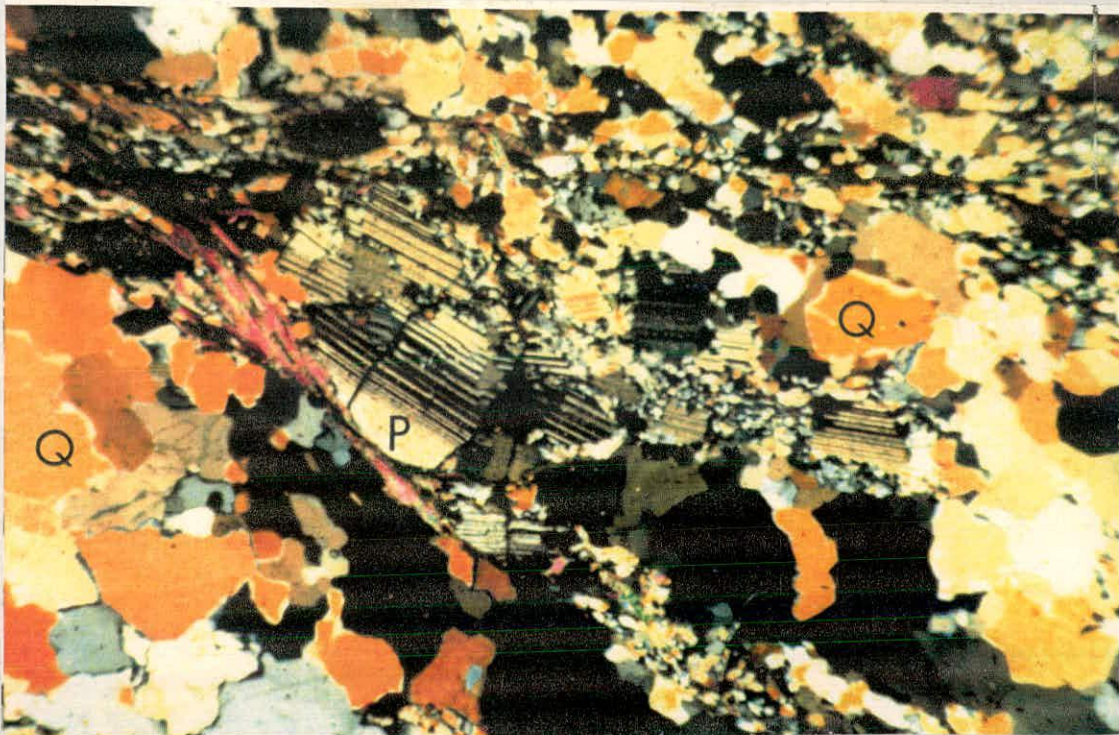
Fig. 5.74 (b): Textural banding in an amphibolite depending on the varying proportion and size of the amphibole grains and associated biotites in the profile section of a fold (possibly,  $F_4$ ). See text for details. Munsiri Formation (thin section - Loc. 18(2)'87).



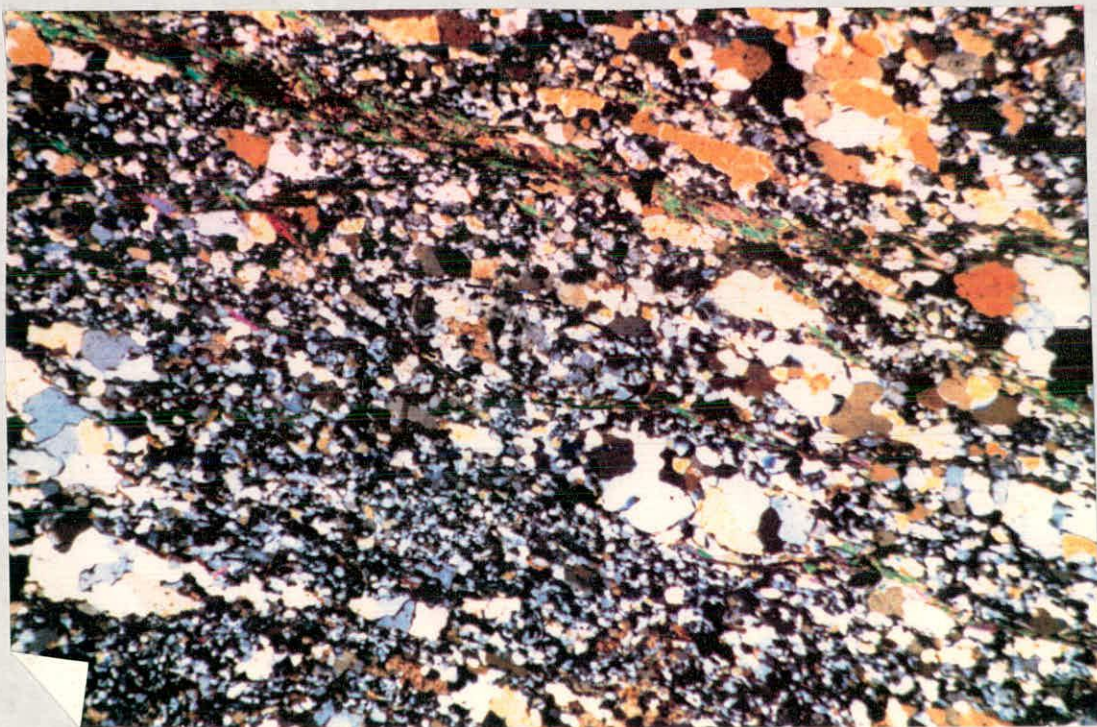
**Fig. 5.75:** Photomicrograph (XPL) showing recrystallised matrix quartz fabric on way to attain equilibrium (annealing). Note the common occurrence of grain contact triple junctions (arrowed). Stretching direction is from left to right parallel to the base of photo. In the upper right corner, M = muscovite, K = kyanite. Kyanite-bearing quartzite of the Berinag-Mandhali formations (thin section 32A||L). Length of view (photo) 2mm.



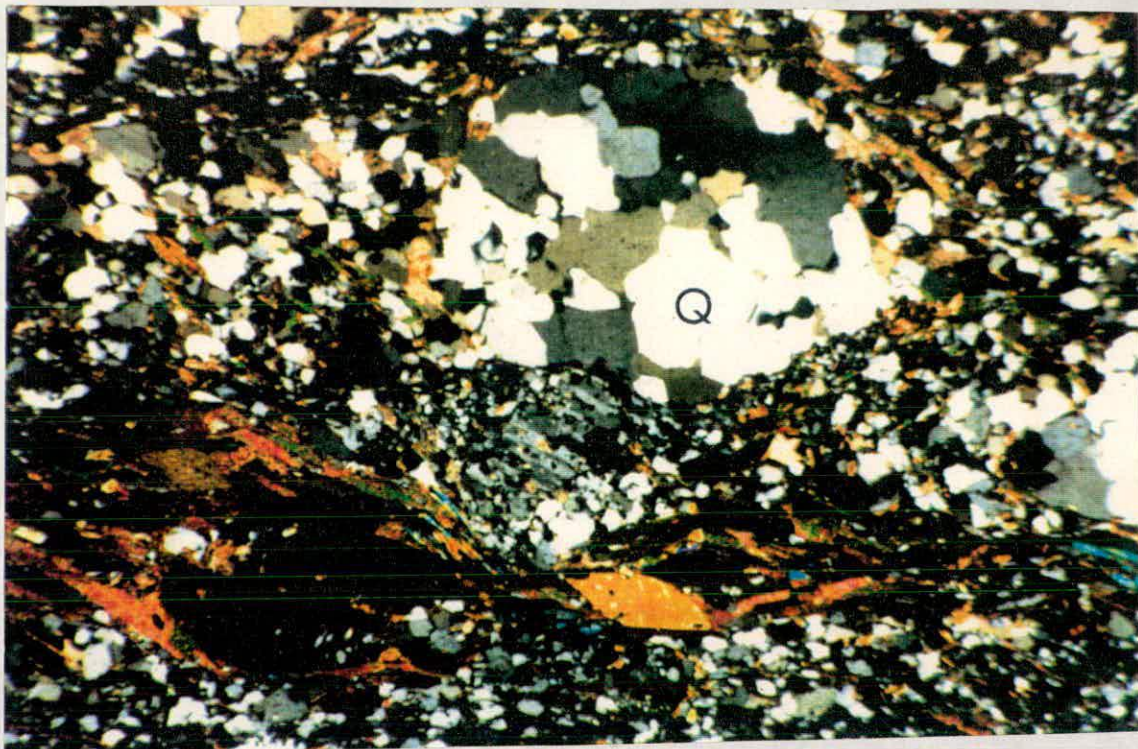
**Fig. 5.76:** Photomicrograph (XPL) showing the shape fabric in matrix quartz (lower half of photo) and vein/band quartz (upper half). Main foliation trace is horizontal in the photo; S is on right, N on left. Note that the vein/band quartz grains show an elongate shape fabric (indicated by longer arrow at top middle) oblique to the main foliation, whereas the matrix quartz grains are generally elongate parallel to the main foliation (smaller arrows at bottom left and right). Main foliation in the matrix is clearly defined by the fine micaceous laminae; but mica is very rare in vein/band. Kyanite quartzite of Berinag-Mandhali formations (thin section 32A||L). Photo-base 7.5mm.



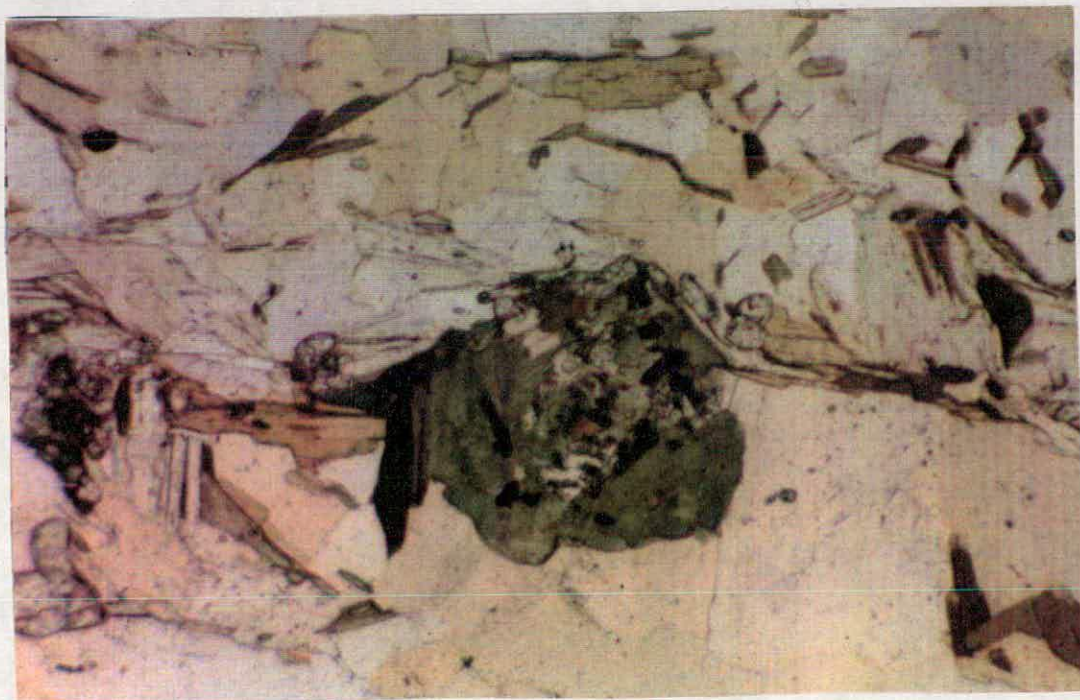
**Fig. 5.77:** Photomicrograph (XPL) shows plagioclase (P) is deformed in brittle mode, while quartz (Q) is not. Quartzofeldspathic protomylonite from Munsiri Formation (thin section 15/22Ab). Photo-base is about 5.5mm.



**Fig. 5.78:** (a) Photomicrograph (XPL) showing break-down of plagioclase through symplectitisation. Most of the fine-grained quartzofeldspathic matrix in this rock is presumed to have recrystallised from plagioclase in this way. Note particularly the part within dashed ellipse at centre; probably this was originally a single plagioclase porphyroclast (with difficulty the shades of lamellar twinning can be seen; compare with the next figure (b)). The fairly intact, grey/white, elliptical porphyroclasts are of quartz; yellow grains are amphibole. Lower part of Munsiri Formation (thin section - Bet. MR18&19). Photo-base 6.5mm.



**Fig. 5.78:** (b) Photomicrograph (XPL) from the same thin section as above shows the existence of porphyroclasts of three minerals side by side; a large ovoid quartz porphyroclast (Q), a plagioclase porphyroclast at centre and an amphibole porphyroclast (now in extinction position surrounded by micas) at lower left. Of the three, the plagioclase is breaking down most rapidly (through symplectitisation) and contributing to the development of fine-grained recrystallised quartz and feldspar in the matrix. However, still one can recognise the remains of the characteristic lamellar twinning in the plagioclase porphyroclast. Cf. Fig. 5.78(a) above. Photo-base 4mm.

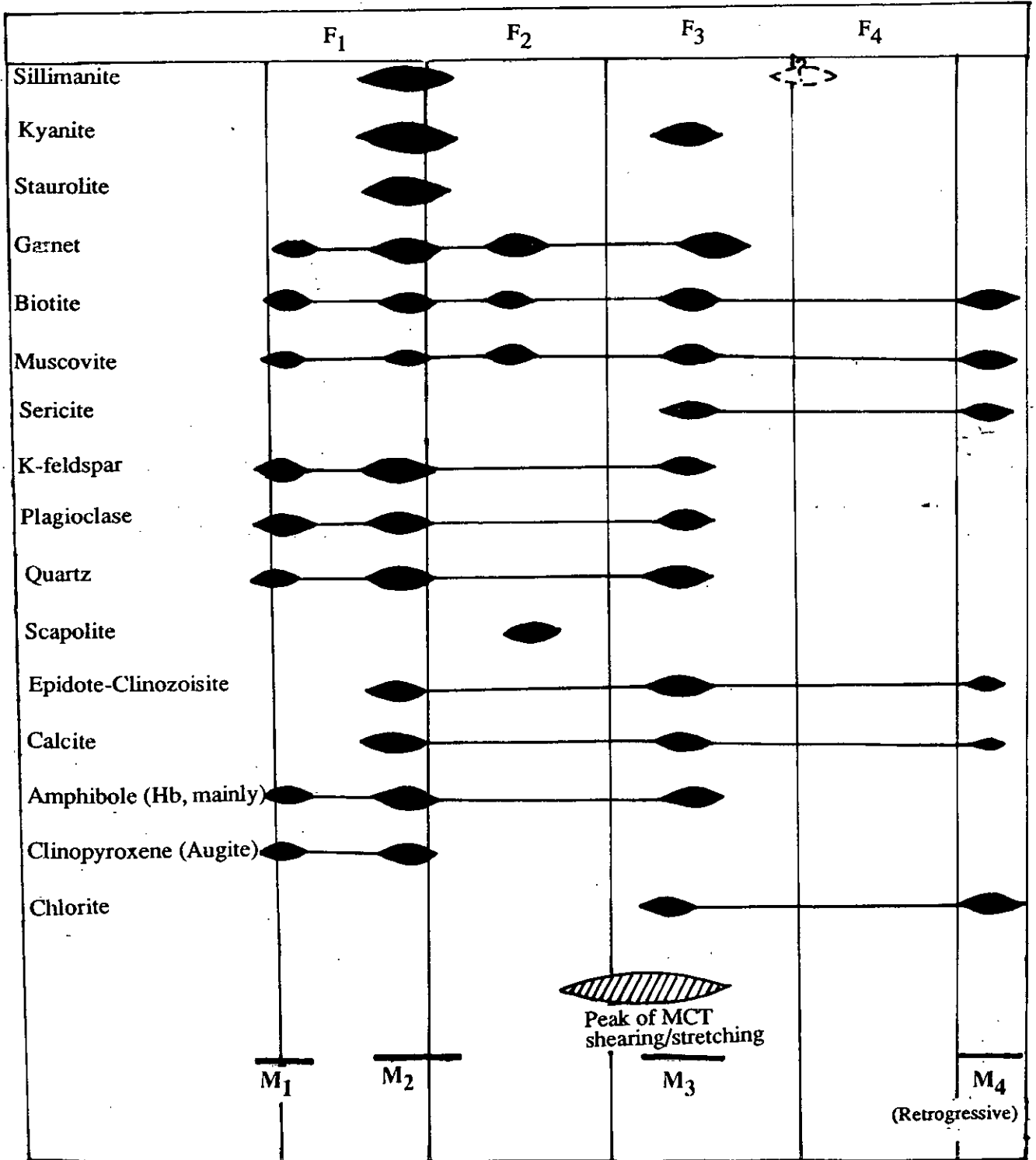


**Fig. 5.79:** Pseudomorphous alteration of a garnet porphyroblast into biotite and chlorite. The outline of the original garnet grain is still preserved. Compare with Fig. 5.62. Photomicrograph, PPL. Munsiri Formation (thin section - 64A'87). Photo-base about 2.5mm.



**Fig. 5.80:** Photomicrograph (PPL) showing example of pseudomorphous alteration of biotite (B) into chlorite (C). In fact none of the chlorite seen this thin section is primary. Note how nicely the form, shape and size of the original biotite have been mimicked by chlorite (shown by arrow at lower left). The arrow on right shows a small remaining portion of an otherwise totally chloritised body of biotite. G = garnet, Q = quartz, M = muscovite. Munsiri Formation (thin section - 7Pb'87). Scale bar 0.5mm.

**Metamorphic mineral paragenesis**  
**Combined list for the Joshimath Gneiss & Munsiri Formation.**



**Fig. 5.81(a):** Composite fish diagram (mineral paragenetic sequence) for Joshimath Gneiss and Munsiri Formation.

**Metamorphic mineral paragenesis**  
**For Berinag-Mandhali formations.**

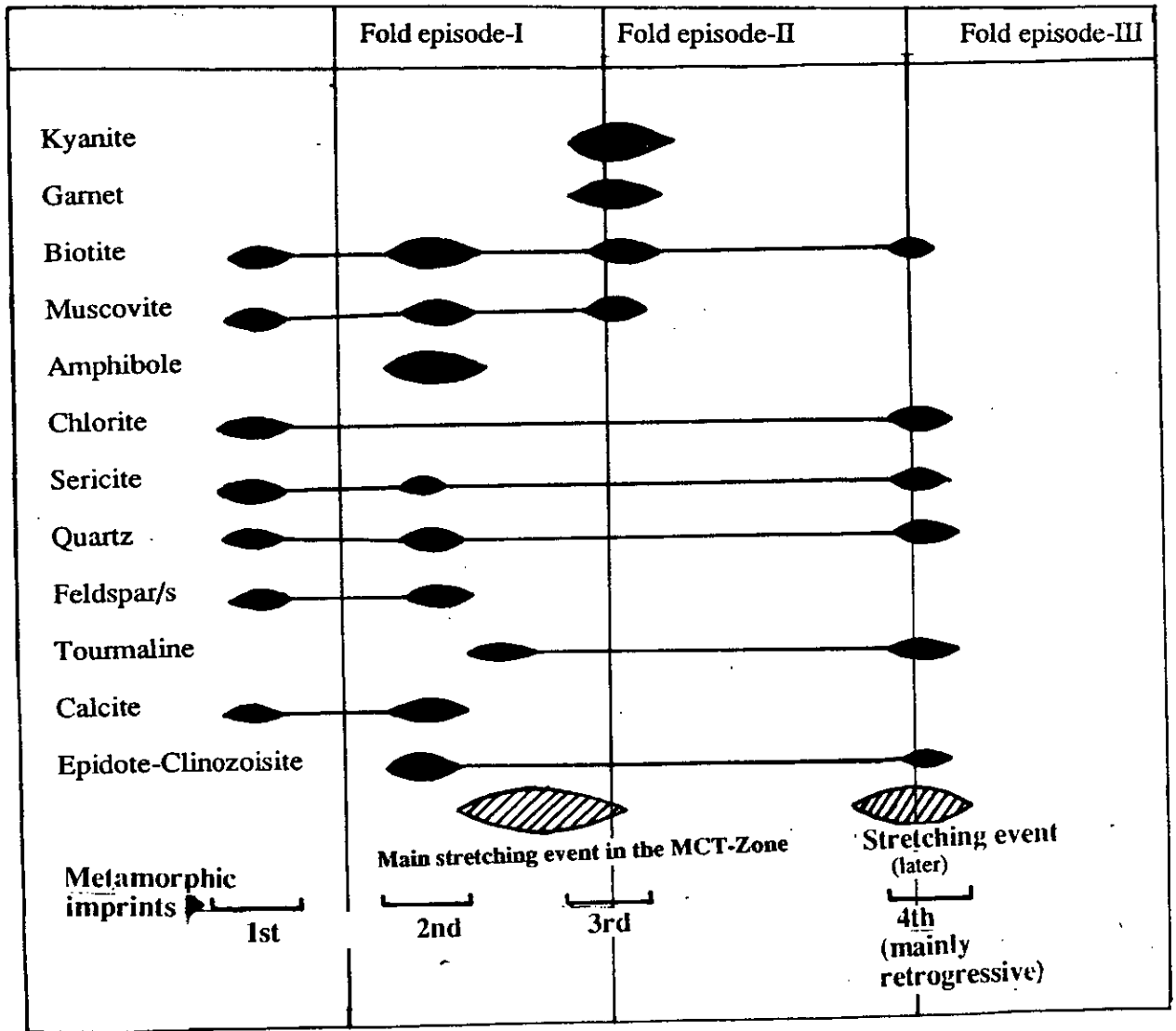


Fig. 5.81(b): Composite fish diagram (mineral paragenetic sequence) for the Berinag-Mandhali formations.

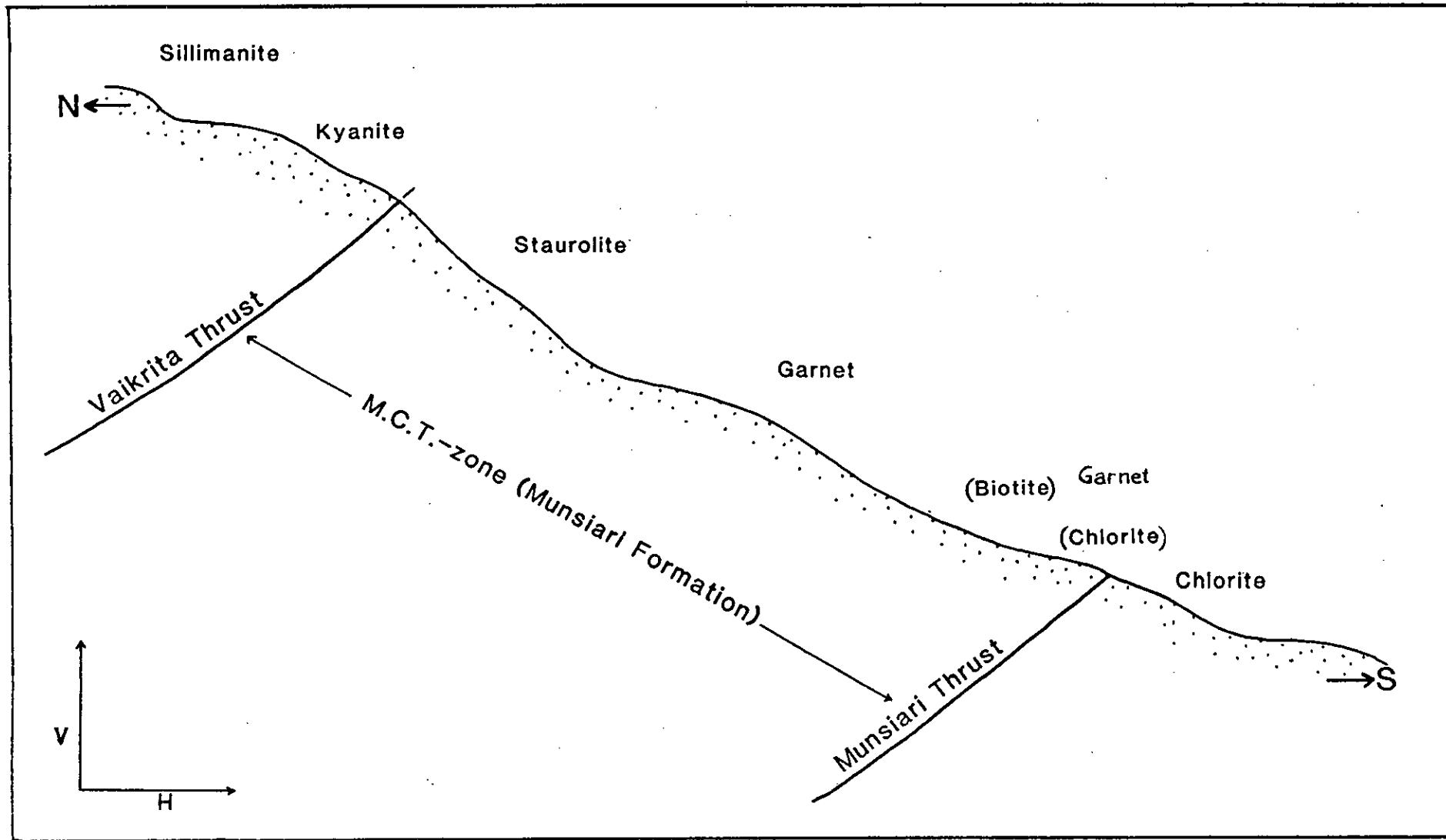


Fig. 5.82: Metamorphic mineral distribution defining the inverted metamorphic sequence across the base of the Higher Himalayas in Joshimath area.

## Chapter-6

# SUMMARY OF RESULTS AND DISCUSSION

In this chapter, the results are summarised and their implications discussed. After giving an overview of the major geological features of the MCT-Zone, the chapter will provide: first, the essence of the structural-tectonic findings; second, the main metamorphic observations which will set the scene for delving into the establishment of a suitable model for the Higher Himalayan inverted metamorphism; third, a critique of the models suggested for inverted metamorphism in general (an introductory discussion on the models that have been applied to the Himalayas has been given in sub-section - 1.4.3 in Chapter 1); fourth, an appreciation in the light of the observations made in the present study, of the relative merits and demerits of the potentially most applicable models for the Higher Himalayan inverted metamorphism. Finally, the model which is most favoured by the data from this study will be established; particular attention will be given to find out the relationship between the inverted metamorphism and the MCT emplacement.

At this point it might be useful for the reader to refer back to subsection-1.1 in Chapter-1 for a recapitulation of the important questions posed at the outset. The following sections are designed to cover the answers to those questions. However, it must be noted that the results obtained mostly refer to the Joshimath area (Alaknanda - Dhauliganga section) unless otherwise mentioned, because this area has been studied in much greater detail than the Sobala area (Darmaganga section).

Fig. 6.1 portrays some of the most important geological features across the MCT-Zone (Munsiari Formation). For instance, it shows how:

(i) the pre-MCT "regional"  $F_2$  folds are modified as a result of syn-MCT stretching/shearing. The  $F_2$  axial orientation comes into parallelism with the main stretching lineation and the folds show appreciable tightening as we enter into the MCT-Zone from above. Such changes are clearly noticeable in the field from about 2 km (outcrop-distance) above the Vaikrita Thrust downward well into the MCT-Zone.  $F_3$  and  $F_4$  are probably "local" shear zone folds.

(ii) the later N-trending stretching becomes important low in section.

(iii) the  $M_4$  i.e. post- $M_3$  retrogression increases down section. This may be due largely to the introduction of retrograding hydrous fluids derived from the Lesser

Himalayan lowgrade metasedimentary footwall into the hanging wall as a consequence of the MCT emplacement.

(iv) in general, the strain increases down section to the Munsiri Thrust. This increase in strain is recognised on the basis of qualitative observations, such as, down the section the matrix and porphyroblast grain size decreases in the quartzofeldspathic augen gneiss and/or mylonite-protomylonite horizons; the number of quartz *and/or* feldspar porphyroclasts decreases in them; there is an increase in the frequency of occurrence of tight to isoclinal folds; there is progressive parallelism of the  $F_2$  axes with the main stretching lineation direction; boudinage structures are more frequent, and the competent/semicompetent pre- or early-syn- $M_3$  porphyroblasts are generally more strongly pulled apart or stretched.

(v) while there is a stronger correlation or continuation of structures across the Vaikrita Thrust particularly in terms of the pre- or syn- $M_3$  ones, the correlation across the Munsiri Thrust could not be established.

All these features probably suggest that the Vaikrita Thrust is, strictly speaking, not as clear a shear zone boundary as the Munsiri Thrust. Considering the MCT-Zone as a shear zone, its upper boundary is more diffused than the lower one. The basal or lower part of the Munsiri Formation suffered maximum strain. Post-MCT folding ( $F_4$ ), retrogression ( $M_4$ ) and the later stretching event suggest repeated movement in the lower part of the MCT-Zone, on or near the Munsiri Thrust.

## 6.1 STRUCTURAL-TECTONIC OBSERVATIONS

### 6.1.1 Location of the Vaikrita and Munsiri thrusts and nature of the MCT-Zone

As already mentioned in Chapters 1 and 3, the literature provides little help as regards the location of the Vaikrita Thrust and the Munsiri Thrust in the field. Detailed studies in the field in connection with the present work has enabled the author to establish the exact field location of the two thrusts in two important sections (viz. Alaknanda section and the Darmaganga section in Garhwal and Kumaun Himalayas respectively). (see Fig. 4.2, 4.3 & 4.6 for the location of these two bounding thrusts of the MCT-Zone or Munsiri Formation and Plates-II, III, IV & V). The Vaikrita and Munsiri thrusts appear more like slide zones i.e. deep crustal faults (for further discussion see section 4.5 in Chapter 4).

Compared to the Joshimath Gneiss Formation above and the Berinag-Mandhali formations below, the Munsiri Formation has a less clear stratigraphic status. Certainly it is a tectonostratigraphic unit and has a less precise lithostratigraphic or chronostratigraphic character than the units overlying and underlying it; it has a more 'mixed' or 'hybrid' character (for further details see Chapter-2, Tectonostratigraphy and Lithology). I find great difficulty in visualising the character of the Munsiri Formation before the Vaikrita and Munsiri Thrusts in the Himalayas came into existence. In Chapter 2, some indications of the paleogeographic and tectonometamorphic status of the Munsiri Formation in a regional context has been given. The whole MCT-Zone or the Munsiri Formation appears to be a complex shear zone whose exact analogue I have not found in literature. Some workers consider the Munsiri Formation as a schuppen zone (Gansser, 1964, 1979, 1980); Saklani & coworkers postulate the existence of large-scale duplexes within the Munsiri Formation in Bhagirathi and Bhilangana valley sections in western Garhwal (see Saklani et al., 1991; Saklani et al., 1989).

Embodying a realm of lithology-controlled strain partitioning the MCT-Zone shows a general increase in strain down the section in Joshimath area. The lithological succession in the Munsiri Formation is quite varied and does not show any mappable repetition (for an appreciation see Plates II & IV, lithomap and composite road-log). Although the Munsiri Formation shows a general increase in strain down the section, the conspicuous lateral variation in thickness of the MCT zone from 2.5 km in Alaknanda section to only 400 m in Darmaganga section is probably not due to strain-induced attenuation. Most likely this thickness variation is a large-scale deep-crustal equivalent of horses accreted to the base of thrusts seen in upper crustal thrust zones or anastomosing fault-zones (see Fig. 6.2a). For example, in the Moine Thrust zone of NW Scotland, Elliott & Johnson (1980, their Fig. 6 and related text) describe horses on various scales characterised by a lenticular shape in three dimensions. They give a genetic explanation to this feature by appealing to a possible lateral variation in the nature of slip (smooth slip vs. rough slip) along the thrust. But in case of the MCT-Zone, it is difficult to tell whether this has been the case and certainly thrusting along the MCT zone took place in a more ductile regime than in the Moine Thrust. While the explanation given by Treloar & Coward (1991) for the mutual closeness of thrusts in the NW Himalaya (Zaskar area) as due to their nearness to the pinning points closely behind the indentation-head of the Indian plate is indeed elegant, such an explanation cannot hold true for the lateral variation in thickness of the MCT zone found elsewhere almost all along the length of the Himalayan belt. That the MCT-Zone may have down dip thickness variation as well is clearly indicated in Valdiya's

(1980) model of tectonic architecture in Sobala region (i.e. NE Kumaun) (see Fig. 6.2b).

I have found two generations of stretching lineations, particularly in Joshimath area: one dominant set that plunges NNE'ly is earlier, and the other weaker set plunging due north and found mainly in the lower part of the section has developed postdating the last retrogressive  $M_4$  metamorphism. Further discussion on these two generations of stretching lineation will follow in Subsection 6.1.4.

### 6.1.2 Styles and distribution of folding across the MCT zone

At least four episodes of folding are recognised in the Joshimath Gneiss and the Munsiri Formation. The  $F_4$  folds are rarely seen in the Joshimath Gneiss and the  $F_1$  folds are difficult to recognise in the Munsiri Formation. The main domain of occurrence of successively younger generations of folds appears successively towards south i.e. down the section (see Fig 6.1). In the Berinag-Mandhali formations, three episodes of folding (I, II & III) could be distinguished. Correlation of these folds with those in the hanging wall block is uncertain. For the styles of the different generations of folds in the area, see Subsections 3.2 and 3.3 in Chapter 3.

If one takes an overall statistical count (irrespective of fold-generations), then the frequency of tight to isoclinal folds increases from the top to the bottom of the MCT zone. This is probably a qualitative indicator of increasing strain down the section across the MCT zone. Bhattacharya & Siwal (1985) in their study across the North Almora Thrust demonstrated quantitatively that the shortening of the tight to isoclinal folds increases progressively towards the thrust. Similarly, the quartzofeldspathic horizons in the Joshimath-Helang section show a general decrease in matrix grainsize and a decrease in the frequency of occurrence and in the size of the feldspar &/or quartz porphyroclasts down the section within the MCT-Zone.

Of the four fold episodes recognised in the hanging wall block  $F_2$  is the most dominant episode, particularly common in the Joshimath Gneiss and developed in a wide range of scale (multi-order folds). The largest recognisable  $F_2$  folds in the exposures are all south-verging and, interestingly, most of them have long attenuated overturned limbs compared to relatively short, thick or wavy normal limb. This probably suggests that these folds belong to the overturned limb of a still higher order regional  $F_2$  fold. Further discussion on this aspect will follow in Subsection 6.4.3.

### **6.1.3 Mutual time relationships between folding and stretching events**

The main stretching/shearing event is equated with the time of main MCT-emplacement. Although the shearing/stretching event was probably initiated during the  $F_2$  fold episode, its peak postdated the  $F_2$ -episode and was broadly synchronous with the  $F_3$ -episode. Thus the main MCT-emplacement phase is considered to have started broadly synchronously with the  $F_2$ -episode, and culminated around the  $F_3$  time.

The later stretching event postdated the  $F_4$  fold episode.

In the Berinag-Mandhali footwall block, the stretching lineation is not uniformly strongly developed. Where present, a northward plunging lineation is usually common. But in some quartzite horizons, particularly in the kyanite-quartzite horizon, a strong  $N18^\circ E$ -ward plunging lineation is observed which is presumed to be later than the dominant stretching lineation in the MCT-Zone and probably correlatable with the later (i.e. post-retrogression) stretching lineation of the MCT-Zone. As argued in Chapter 5, probably the footwall fold episode-I developed in response to the initiation of the MCT emplacement, while episodes II and III took place respectively preceding and following the later stretching lineation (i.e. kyanite quartzite lineation). The later stretching event, even though it postdated most of the fold episodes, did not have any recognisable effect on the folds. But the main stretching event affected the orientation and, to some extent, the style (interlimb angle) of the  $F_2$  folds. When the MCT zone is approached from the north, the  $F_2$  fold axes are reorientated from ENE to NE or NNE so that they gradually come into parallelism with the stretching lineation (see Fig. 3.3). Similarly an appreciable tightening of the  $F_2$  folds is noticed. These features imply a general increase in strain down the section in the MCT-Zone.

### **6.1.4 Evolution of the MCT-Zone**

Rather than being a single event, thrusting along the MCT appears to have been episodic. Even though there was one dominant stage of thrust emplacement, evidence of a later reactivational stage is also found. These two episodes are recognised on the basis of the observation of two distinctly different generations of

stretching lineations. The earlier set is the dominant one: it plunges NNE and has been probably recognised by other workers as the regional stretching lineation which corresponds to the main stage of MCT-emplacement. The later, weaker set plunges N and is found on retrograde foliation surfaces at the middle to lower part of the Joshimath-Helang section. The later set probably indicates a post-retrogressive i.e. post-M<sub>4</sub> reactivational stage of MCT thrusting. These two generations of stretching lineations may have some largescale tectonic implications. In collisional orogenic belts like the Himalayas, it is very likely that different thrust motion trajectories at different times, particularly along the MCT-Zone, could reflect the different post-collisional convergence directions of the Indian plate. In other words, we could presume that the NNE'ly plunging early stretching lineation is in response to a broadly NNE-ward convergence of the Indian plate, whereas the later northward plunging stretching lineation indirectly reflects a later change in the plate convergence direction towards north (Fig. 6.3). As we know from the geophysical investigations carried out in and around the Indian plate, the Indian plate suffered a counter-clockwise rotation while continuing its broad northward flight (see Achache et al., 1983; Besse et al., 1984; Patriat & Achache, 1984; Besse & Courtillot, 1988; Dewey et al., 1989 etc.). To establish the presumed correlation, it would have been very useful to know the dates (absolute) for the two stretching events and thus the time-gap between them. Although some headway could be made on determining the date (discussed below) of the main MCT-emplacement, the age of the later minor reactivational stage is not known. Fig. 3 (p.721) of Dewey et al. (1989), reproduced here as Fig. 6.4 displays very clearly the progressive anticlockwise rotation of the Indian plate concomitant with its general northward migration. The motion path established in Dewey et al.'s study (1989) bears a broad resemblance to those reconstructed by Patriat & Achache (1984) and Besse & Courtillot (1988), but unlike the two latter it does not show the large zig-zag deviations in the early Tertiary time. Although there is some discrepancy between the actual values of the dates and the magnitude of angular rotation with the available MCT-emplacement date/s and the variation in the direction of plunge of the two generations of stretching lineations respectively, I think this is not very serious, particularly in view of the inherent uncertainties/errors involved in reconstructions of early movements of plates. The important thing is the general matching pattern over the time. In this connection, reference could be made to a recent study by Treloar et al. (1991) in the Northwest Himalaya, where they have shown how lineation patterns reflect the rotation of thrust sheets. Using a combination of lineation data and paleomagnetic data, they estimated regional scale rotations of up to 40°, anticlockwise in Pakistan Himalaya and clockwise in Kashmir Himalaya, which, they believe, led to converging thrust

interference now expressed in the form of crustal scale folds (i.e. the NW Himalayan syntaxial bends).

## 6.2 METAMORPHIC OBSERVATIONS

### 6.2.1 Metamorphic observations in a regional perspective

The present study has definitely established that the highgrade metamorphic event (sillimanite - kyanite grade) in the Higher Himalayas predates the MCT-emplacement. So, if we accept a mid-Miocene age for thrusting along the MCT (see Windley, 1983, 1988; Searle et al., 1988; Hodges et al., 1988; England et al., 1992), then the highgrade metamorphism in the Central Crystallines i.e. the Vaikrita Crystallines would certainly assume a pre-Miocene age. This leaves two important possibilities which are not mutually exclusive:

- (a) The Central Crystallines underwent a highgrade metamorphism of Tertiary (Himalayan) age.
- (b) The Central Crystallines underwent a highgrade metamorphism which is a much older, certainly pre-Tertiary, possibly Precambrian event.

According to (a), the highgrade event is related to post-collisional thickening of the northern margin of the Indian plate. It is argued below that the event took place due to P-T increase as an immediate consequence of thickening and burial following the India-Eurasia collision, but it is proposed that there was also another older, possibly Precambrian, highgrade metamorphic event in the Central Crystallines.

It is emphasized that the minerals (viz. staurolite, kyanite, sillimanite etc.) defining the present day inverted metamorphic sequence observed within the Central Crystallines developed during the high-grade event of Tertiary age. Also as already discussed in Chapter - 5 (Metamorphism), this highest grade metamorphic episode ( $M_2$ ) was preceded by at least one other episode of metamorphism ( $M_1$ ) during which the formation of the gneissosity (or metamorphic foliation) was initiated in the Joshimath Gneiss (Vaikrita Group). Reaching at least up to Barrovian garnet grade the  $M_1$ -episode possibly took place in lower amphibolite facies conditions consistent with the associated high degree of metamorphic differentiation.

The important stages in large-scale regional (plate) tectonics in the Himalayan region that are expected to have identifiable metamorphic and structural signatures are:

1. Pre-collisional stage
2. Collision and immediate post-collisional thickening in the northern marginal part of the Indian plate
3. Continuing post-collisional migration of the Indian plate and consequent progressive tectonic thickening (by folding and/or thrusting) of its northern margin south of the collisional suture.

As far as the metamorphic events are concerned, if the late retrogressive  $M_4$  episode is kept aside, then we are left with three other distinct metamorphic episodes viz.  $M_3$ ,  $M_2$  and  $M_1$ . Of these three episodes,  $M_3$  is contemporaneous with the MCT-emplacement (see Chapter 5). There is no doubt that thrusting along the MCT would correspond to the third stage mentioned above (stage 3 of regional tectonics). Therefore the  $M_3$ -metamorphism corresponds to this third stage. Now the most simple and logical step would be to equate the  $M_2$ -episode with stage 2 and  $M_1$  with stage 1. Through the course of discussion that now follows it will be clear that this correlation is correct.

Consultation of literature on Himalayan Geology suggests the following:

(1) Most workers thought (see Naha & Ray, 1970; Ray & Naha, 1971; Powell & Conaghan, 1973a, b; Le Fort, 1975) the highest grade metamorphism records the first metamorphism in the Central Crystallines and this metamorphism is Himalayan (i.e. Tertiary) in age. So far it has not been clearly emphasized in the literature that the Central Crystallines were well-defined metamorphic rocks even before the Himalayan orogeny.

(2) Many workers (e.g. Le Fort, 1975; Bordet et al., 1981; Sinha Roy, 1981; Burg et al., 1987; Windley, 1983, 1988 etc.) considered this highgrade metamorphism is correlatable with the MCT-emplacement. Only in recent years has it been postulated that this highgrade event preceded the thrusting along the MCT (see Brunel, 1983, 1986; Hodges et al., 1988).

(3) Very recently Hodges et al. (1988), England et al. (1992) suggested that the highgrade metamorphism took place during the time

of immediate post-collisional thickening, but prior to the MCT-emplacement. England et al. (1992) considered the highest grade episode as the first major metamorphic episode affecting the Central Crystalline rocks. This episode corresponds to the  $M_2$ -episode recognised in my study.

No distinct metamorphic imprint has been recognised in the Central Crystallines corresponding exactly to the time of India-Eurasia collision. Thus the  $M_1$  episode recognised through the present study must have taken place before the India-Eurasia collision. By itself this is a very important finding, because this means that the Central Crystallines record clear-cut imprint of pre-Himalayan (i.e. pre-collisional) metamorphism as well. When the India-Eurasia collision took place, the Vaikrita or Central Crystallines were already highgrade metamorphic rocks possibly belonging at least to the lower amphibolite facies. In other words, the Central Crystallines were affected not only by the staurolite-kyanite-sillimanite grade upper amphibolite facies highgrade  $M_2$ -event during the early part of the Himalayan orogeny (in Tertiary time), but there was also another older i.e. pre-Tertiary, fairly highgrade, probably lower amphibolite facies metamorphic event in them.

In summary, therefore, the  $M_1$ -metamorphism established in the present study took place during the pre-collisional stage (Stage 1 of regional tectonics mentioned earlier); the  $M_2$ -metamorphism took place during thickening of the northern marginal part of the Indian continental plate closely following the India-Eurasia collision (i.e. during stage 2 of regional tectonics); while the  $M_3$ -metamorphism was a syn-MCT episode corresponding to a relatively later stage of tectonic burial and thickening through thrusting and folding in response to continuing post-collisional northward migration of India (Stage 3).

### **6.2.2 Timing of the earliest metamorphism ( $M_1$ ) in the Central Crystallines**

An important question is : What was the absolute age of the  $M_1$  metamorphism in the Central Crystallines? Unfortunately systematic and detailed radiometric work could not be undertaken in the present study; also there is no mention of this metamorphism and its age in literature. So, any suggestion made here regarding the age of the  $M_1$  metamorphism would be of qualitative nature.

The  $M_1$  metamorphic episode was pre-collision, so certainly it was pre-Eocene in age. Tethyan sediments that overlie the Central Crystallines range in age from Cambrian to Early Paleogene (Le Fort, 1975; Valdiya, 1988; Windley, 1988; Hodges et al., 1988a; Searle et al., 1988; Molnar & England, 1990; England et al., 1992 etc.). No discrete evidence has been found in the Himalayan region of any tectonothermal event on a large scale that took place intermediate in time between the start of deposition of the Tethyan sediments and the onset of the India-Eurasia collision. Effects of the Gondwanaland break-up are not noticeable in metamorphic terms within the Central Crystallines. So, most possibly, for nearly the whole duration of the Paleozoic and the Mesozoic and part of the Cenozoic there was a 'metamorphic quiescence' in the Central Crystallines, and the  $M_1$ -episode, which is the earliest metamorphism recognised in them so far, most probably took place during the Precambrian time, before the start of the Tethyan sedimentation.

### 6.2.3 Metamorphism across the MCT-Zone

Considering the Joshimath Gneiss and Munsiri Formation together, four distinct episodes of metamorphism,  $M_1$ - $M_4$ , could be recognised. All the four episodes are clearly recognisable in the Joshimath Gneiss and in the upper part of the Munsiri Formation, while in the middle and lower part of the Munsiri Formation it is difficult to recognise the effects of  $M_1$  metamorphism probably because of intense deformation and late retrogressive effects ( $M_4$ ). For the time relations of metamorphic episodes ( $M_1$  to  $M_4$ ) with the fold episodes ( $F_1$  to  $F_4$ ) and stretching events, see Fig. 5.81a in Chapter-5.

Imprints of four episodes of metamorphism have been noticed in the Berinag-Mandhali formations. Of these, the earliest one was a lowgrade metamorphism, apparently not involving any deformation. The second metamorphism was synchronous with fold episode-I. The third metamorphism included kyanite porphyroblastesis and was followed in time by fold episode-II which, in turn, was followed by the stretching event (kyanite stretching) accompanied by the fourth metamorphic imprint. Finally came the fold episode-III. There was no one to one correlation between the metamorphic episodes in the MCT-sheet and the Berinag-Mandhali footwall block, except possibly during the later stretching event when appreciable recrystallisation took place, at least in the kyanite quartzite horizon of the Berinag-Mandhali formations.

In the MCT-sheet (Joshimath Gneiss & Munsiri Formation together), during the  $M_1$  the P-T conditions probably corresponded to lower amphibolite facies. The facies determination of the  $M_1$  episode is somewhat speculative and is based more on circumstantial evidence than factual evidence. For example, I noticed through textural (petrographic) investigation that the highest grade mineral that could be recognised to belong to this metamorphic episode ( $M_1$ ) is garnet (in pelitic lithohorizons); other minerals include micas (both muscovite & biotite), quartz, alkali & plagioclase feldspars etc and amphiboles in basic lithologies. But the better indication for the highgrade nature of  $M_1$  episode comes from an indirect feature, that is, a high degree of metamorphic differentiation (gneissosity or first foliation formation/initiation) involved with this episode.

The  $M_2$  event gave rise to the Barrovian isograds in the area which included at least garnet, staurolite, kyanite and sillimanite. There has been accompanying recrystallisation of quartz, biotite, muscovite, clinozoisite, tourmaline, apatite etc. in this metamorphic episode. The  $M_2$  metamorphism, therefore, probably has taken place <sup>for the major part</sup> in upper amphibolite facies condition. Some geothermobarometric studies of the  $M_2$  assemblage (eqv. to their  $M_1$ ) have been done by Hodges & coworkers (Hodges & Silverberg, 1988; Hodges et al., 1988; Hubbard, 1988, 1989). Their P-T data were mainly derived using three well-calibrated pelitic geothermobarometers: the garnet-biotite geothermometer (Ferry & Spear, 1978), the garnet-plagioclase-aluminosilicate-quartz geobarometer (Newton & Haselton, 1981) and garnet-muscovite-biotite-plagioclase geobarometer (Ghent & Stout, 1981; Hodges & Crowley, 1985). Some additional help came from the Gibb's method modelling of garnet zoning (Spear & Selverstone, 1983) and garnet inclusion thermobarometry (St-Onge, 1987). The assemblage they recognised for their  $M_1$  event is quartz + biotite + muscovite + plagioclase + garnet  $\pm$  microcline  $\pm$  sillimanite and their P-T estimate for this episode was  $T = >900$  K (i.e.  $627^\circ\text{C}$ ), and  $P = >960$  Mpa (i.e.  $>9.6$  Kb). They deduced a burial depth for this metamorphism of about 36km. It is interesting to note that the present structural thickness above the Vaikrita Thrust in Garhwal closely matches with this result. Hodges & coworkers also recognised in the higher part of Higher Himalayas (i.e. farther away from the Vaikrita Thrust) a later Buchan-type event,  $M_2$  (eqv. broadly in time to my  $M_3$ ) related to anatectic granitoid intrusions. The  $M_2$ -assemblage recognised by them is quartz + muscovite + biotite + plagioclase + garnet  $\pm$  kyanite and their P-T estimate for this  $M_2$  is:  $P = 317$  to  $523$  Mpa (i.e.  $3.17$  to  $5.23$  Kb) which is much lower than that of the  $M_1$  event, and  $T = >900$  K (i.e.  $627^\circ\text{C}$ ) which is the same as in their  $M_1$ . Note that Hodges & coworkers found sillimanite to belong to their  $M_1$  ( $M_2$ , established in my study) and, probably due to

inadequacy in sampling, did not find any sillimanite in their  $M_2$  assemblage; while England et al. (1992) recognised sillimanite in the Annapurna-Manaslu region in Central Nepal as belonging to this later episode of anatectic granitoid intrusion. Formation of sillimanite in the Higher Himalayas as part of a Buchan-type metamorphism (i.e. low-P) related to Tertiary granitoid intrusions is widely reported in <sup>the</sup> literature (e.g. Gupta, 1978a & b; Pecher, 1989; Le Fort, 1981, 1975, 1988etc.). Therefore, we see that in the Higher Himalayas there are two generations of sillimanite (cf. suggestions made earlier in Chapter 5).

The  $M_3$  i.e. syn-MCT event gave rise to the formation of tiny kyanite grains, some garnets, amphiboles in addition to white micas, quartz etc and myrmekitic recrystallisation from plagioclase. Also mylonitisation in the area took place during this event. The  $M_3$  metamorphism, therefore, probably took place under *lower amphibolite* facies conditions. My observations suggest that the  $M_3$  event is characterised more by tectonic (mylonitic) fabric formation than by porphyroblastesis (metamorphic mineral formation). Hubbard (1989) found syn-MCT (20 Ma old) hornblende whose closure temperature for Ar is 550°C. Also mylonitisation of quartzofeldspathic rocks through crystal-plastic processes normally requires a temperature not less than 350°-400°C (see White, 1976, 1979a & b; Schmid, 1982; Tullis, 1979etc).

Finally the  $M_4$  episode took place in chlorite-grade, retrogressive, lowermost greenschist facies conditions, in all probability at normal temperature and pressure.

#### 6.2.4 Inverted metamorphic sequence in the study area

Going upsection chlorite-biotite grade is found in the lower part of the MCT-Zone (Munsiari Formation), garnet grade in the middle part and staurolite grade in the upper part. At Joshimath the Vaikrita Thrust that separates the Munsiari Formation from the Joshimath Gneiss coincides with the kyanite isograd. However, in the Sobala area the kyanite-isograd appears slightly above the Vaikrita Thrust. The kyanite-bearing Joshimath Gneiss is overlain by a sillimanite zone. Garnet and biotite occur almost throughout the Joshimath Gneiss. In the Munsiari Formation, garnet occurs in profusion in the uppermost part, decreases in abundance southward and is only intermittently found in the lower-middle and lower part.

Detailed observation suggests that the high-grade part of the Barrovian sequence is more clearly defined in the Joshimath area, and out of the four well-defined Barrovian zones two (viz. garnet and staurolite zones) lie below the Vaikrita Thrust and two above (viz. kyanite and sillimanite zones).

### 6.3 A RECAPITULATION OF MODELS OF INVERTED METAMORPHISM

Intuitively speaking, generally two categories of models are possible for inverted metamorphism, or a combination involving one or more models from both the categories. These categories are:

Category I. Syngenetic or synmetamorphic inverted metamorphism involving presence of reverse geothermal gradient, either locally or regionally.

Category II. Post-metamorphic i.e. when normal progressive metamorphic zones are inverted due to some tectonic or other reasons after the metamorphism has occurred.

Category I (i.e. synmetamorphic type) generally covers the following models -

(i) Shear heating, whereby heat is produced as a result of shearing friction when a hanging wall block moves past the footwall block along a fault/shear zone and a local or perched geothermal gradient is introduced to each block. This new geothermal gradient is normal for the hanging wall block, but reverse for the footwall block. So when there is appreciable shear heating, we may get reverse metamorphic zoning in the top part of the footwall block, but normal progressive sequence in the lower part of the hanging wall block above the shear zone. For further discussion on this aspect, see Graham & England (1976), Brewer (1981), Scholz (1980), Molnar & England (1990) etc.

For the Higher Himalayan inverted metamorphic sequence it is unlikely that the shear heating model would be applicable, because shearing along the MCT-Zone postdated the development of most of the isograd-defining minerals. Also to make any case for shear heating both the Vaikrita and Munsiri thrusts have to be taken into account. The relative timing of the two thrusts is not known; but mainly on a speculative basis it has been suggested that the Vaikrita Thrust either predated or was synchronous with the Munsiri Thrust. However, this is not to rule out any shear heating during MCT-emplacement, only that any contribution of shear heating was

not strong enough to subdue the already existing isograd structure inherited from the pre-MCT,  $M_2$  metamorphic episode.

(ii) Hot-over-cold thrusting, whereby the hot hanging wall block gradually heats up the nearby part of the footwall block and thus introduces a local inverted geothermal gradient across the thrust that could cause inverted metamorphism. The leading workers who introduced this model are Oxburgh & Turcotte (1974), Oxburgh & England (1980), Le Fort (1975) etc., In addition England et al. (1992) recently proposed a model for the Higher Himalayan inverted metamorphism incorporating both the 'hot-over-cold' and shear heating models. A review of England et al.'s work will be given later.

In discussions on this model often no clear indications are given of the detailed metamorphic character viz. early isograd distribution etc., in the hanging and footwall blocks before thrusting. For example, in the Himalayan case I found the inverted isograd structure developed during the  $M_2$  event predating the MCT - emplacement. Large-scale overturned  $F_2$  folding prior to or early synchronously with MCT-thrusting was responsible for the inversion of isograds. Le Fort's (1975) model did not recognise this point. For reasons discussed later, I doubt whether the Central Crystallines were hotter than the temperature necessary for myrmekitisation and/or the formation of garnet, amphibole and small kyanites that took place during the syn-MCT  $M_3$  metamorphism. If the Higher Himalayan crystallines (Joshimath Gneiss and Munsiri Formation) were hot enough and if there were no significant time difference between thrusting along the Vaikrita Thrust and the Munsiri Thrust, then we would expect to see recognisable inverted metamorphic zonation in the Berinag-Mandhali rocks underlying the Munsiri Thrust. Evidently we do not see any such zonation in the Berinag-Mandhali rocks.

(iii) Conductive heating by a high level intrusive body (essentially a Buchan-type metamorphism). To explain the Himalayan case, Mallet (1875), Auden (1935) etc. suggested a cause-and-effect relation between the inverted metamorphism and the Higher Himalayan intrusives. While some Buchan-effect is not ruled out, for instance, formation of second generation of sillimanites in the Higher Himalayan crystallines nearby the leucogranitic intrusives underlying or cutting across the Tethyan rocks (Pecher, 1989; England et al., 1992 etc.), generally speaking, these intrusive bodies occur so far to the north that contact metamorphic effect from them cannot have given rise to the inverted metamorphism at the base of the Higher Himalayas. Also generally the rocks showing the inverted isograds are equivalents of the country-

rocks to the leucogranitic intrusive bodies (see Le Fort, 1975, 1981, 1986, 1988; Brunel & Kienast, 1986 etc.) and so the inverted isograd-structure is older than the leucogranites. The origin of these intrusives is in itself a highly debatable point. Several workers (e.g. Le Fort, 1981, 1986, 1988; Le Fort et al., 1987; Deniel et al., 1987; Vidal et al., 1982, 1984; France-Lanord et al., 1988 etc.) have suggested that the Manaslu pluton in north-central Nepal was anatectic melt-product from a part of the Central Crystallines that displays the inverted sequence of isograds in the hanging wall of the MCT. With an attempt to establish an in situ melt origin of many of these granitoid bodies, Jaupart & Provost (1985), Pinet & Jaupart (1987) proposed a 'heat focussing' hypothesis. All the above points suggest that conductive heating by a high level intrusive body cannot account for the formation of the Higher Himalayan inverted isograds.

Category II i.e. Postmetamorphic models include the following-

(i) Postmetamorphic overturned folding of isograds, giving rise to an inverted sequence of isograds on the inverted limb of such a fold. This model is a very attractive one, particularly because a normal sequence of metamorphic isograds has been discovered in the northern part of the Higher Himalayas underlying the Tethyan Thrust. Rather than one single largescale long overturned antiformal fold, probably a largescale fold-system/train runs along the length of the Himalayas following the outcrops of the Central Crystalline rocks (Vaikrita Gneisses & Munsiri Formation). On the inverted limb of the fold-system at the base of the Higher Himalayas we find the inverted metamorphic sequence, whereas on its normal limb at the sub-Tethyan part a normal disposition of isograds is found. Also the existence of the two crustal scale bounding thrusts viz. Tethyan Thrust to the north and MCT to the south of the Central Crystalline domain could readily explain why we cannot see the other folds complementary to the huge antiformal fold-system/train. Further discussion on this model will be given later in this chapter.

(ii) Thrust-stacking resulting in inversion of a progressive sequence of metamorphic zones. There is a serious problem with this model to generate regional scale inverted isograd sequence, as will be demonstrated in detail later.

(iii) Folding of isograds, followed by further geometric *and/or* metamorphic modification during thrusting. This is essentially similar to the first kind of post-metamorphic model, but with some later influence by thrusting mainly in the form of thickness modification of the folded metamorphic zones on the limbs.

(iv) Invasion of lower part of the highgrade hangingwall block and the thrust/shear zone by hydrous retrograding fluids concomitant with or immediately after thrusting. In the MCT- zone this is indicated by the chlorite and biotite zones which appear to be largely of secondary retrogressive origin. But this model is unlikely to produce a spectacular inverted metamorphic sequence of regional scale. The classic (inverted) isograd sequence across the MCT-Zone at the base of the Higher Himalayas originally developed during the major highgrade  $M_2$  Barrovian metamorphism that took place predating the MCT-emplacement. Johnson & Oliver's (1990) results from systematic illite crystallinity studies across the Lesser Himalaya from Hardwar to Joshimath and across some specific Lesser Himalayan klippen suggest that the Lesser Himalayan rocks do not register any imprint of the Higher Himalayan highgrade ( $M_2$ ) metamorphism; so I think there was no role of the fluids derived from the Lesser Himalayan rocks for the development of the Higher Himalayan isograds.

#### 6.4 POTENTIALLY MOST APPLICABLE MODELS -- A DISCUSSION

In my opinion, mainly three models deserve special discussion insofar as the development of the 'traditionally famous' inverted metamorphic sequence at the base of the Higher Himalayas is concerned -- (i) Hot-on-cold ( $\pm$  shear heating) model, (ii) Thrust-stacking model, and (iii) Folded isograd model. Given below is a discussion of relative merits-demerits of each set of these models.

##### 6.4.1 'Hot-on-cold ( $\pm$ shear heating)' model

Originally, Le Fort (1975) proposed the 'hot-on-cold' model which involves transient inverted geotherm for the development of inverted metamorphic sequence. The idea was that dissipative or conductive heating of a cold footwall block underthrust along the MCT below the hot Central Crystalline hanging wall block would develop inverted metamorphic isograds about the thrust zone. Recently Molnar & England (1990) and England et al. (1992) developed this model further and added a shear heating component to it. Here I propose to discuss the merits-demerits of the 'hot-on-cold ( $\pm$  shear heating)' model in the form of a detailed review of Molnar & England (1990) and of England et al. (1992).

The two papers treat at length the many different aspects of heat generation *and* transfer, temperature distribution etc. related to thrusting along the MCT. The authors present a very clear exposition of many of the important problems related to Himalayan metamorphism and thermal evolution. Based on data available from the literature and using rigorous mathematical calculations they have explored conceptually the heat sources involved in the different phases of Himalayan metamorphism. The focus in England et al.'s study was on the Higher Himalayas and the inner Lesser Himalayas in the Annapurna-Manaslu region of the Central Nepal.

In their study Molnar & England (1990) primarily address the question of assessing from heat-flow data the magnitude of shear stress involved in major faults/thrusts, such as subduction zones, with a view to resolving an existing controversy: one group of workers considers that deformation manifested by earthquakes and mountain building involves more than 100 MPa (1Kb) of shear stress (see for example, Brace & Kohlstedt, 1980; Christie & Ord, 1980; Hanks, 1977; Oxburgh & Turcotte, 1970; Scholz, 1980; Scholz et al., 1979; Toksoz et al., 1971; Turcotte & Schubert, 1973), while another group suggests that mean stresses on major faults are below 30 MPa (see Brune et al., 1969; Dalmayrac & Molnar, 1981; Lachenbruch & Sass, 1973, 1980; McKenzie & Jarvis, 1980; Mount & Suppe, 1987; Tapponier & Molnar, 1976; Zoback et al., 1987 etc). Molnar & England's (1990) analysis shows that the higher value of shear stress is quite plausible. They also point out that a controversy on similar vein exists on the role of dissipative heating during rapid slip along major faults for providing the heat necessary for metamorphism and anatexis in orogenic belts; while Barton & England (1979), Graham & England (1976), Scholz et al. (1979), Sibson (1980) etc considered such dissipative heating as an important source of heat, Bird (1978), Molnar et al. (1983) did not. It is in connection with this second controversy that they took into consideration the anatexis and metamorphism near the MCT in the Himalayas. They say that 'the melting and the inverted metamorphism have been interpreted in a variety of ways, but none of the interpretations rests on a quantitative treatment of the thermal development of the major thrust zones'. Therefore, essentially theirs is a theoretical analysis.

According to Molnar & England (1990), two simple geologic facts suggest a major influence of thrust faulting on the thermal evolution of the region surrounding the MCT in the Himalaya, and of such regions in general -- (i) the metamorphic grade near the MCT is commonly inverted, increasing up section before decreasing again; (ii) Late Cenozoic anatectic granitoid bodies are common in the hanging wall of the

MCT and they have been considered by Le Fort (1981, 1986, 1988) to surely represent melting of the crust whilst or shortly after slip along the MCT was occurring. Mainly three points have been cited as evidence for synchronicity and hence active interconnection between MCT-thrusting and anatectic melting in the Higher Himalayas -- (a) oxygen isotope ratios in one formation of the Higher Himalayas (i.e. MCT hanging wall block) have been shown to be similar to those of the neighbouring Manaslu granitoid body (France-Lanord et al., 1988; Vidal et al., 1982) and by implication it has been assumed that the granitic melt was derived from that formation; (b) Le Fort (1981) observed syn-MCT deformational features in the Chhokang arm of the Manaslu granitoid just above the MCT and this observation is taken to imply melting was roughly synchronous with slip on the thrust; (c) Deniel et al. (1987) dated the Manaslu granitoid at 18 Ma using Rb/Sr whole rock isochron age and Hubbard & Harrison (1989) determined the MCT-emplacement date at  $20.9 \pm 0.1$  Ma from syn-MCT hornblende mineral age using  $^{40}\text{Ar}/^{39}\text{Ar}$  method. The closeness of the two dates has been considered to imply that the leucogranites probably formed while the MCT was active. Molnar & England (1990) argued that in the absence of shear heating, movement on the fault should have cooled the region where melting occurred.

With the strong implicit assumption that the MCT hanging wall block was considerably hotter than the footwall block at the time of thrusting, it has been suggested by Molnar & England (1990) that slip on a thrust fault can affect the temperature distribution in the overlying hanging wall in three main ways. First, displacement of the relatively cool footwall, downward along the fault, should draw heat from the overlying hanging wall and cool it. Second, the gradual emplacement of the initially shallow, and therefore radiogenic, crust beneath the hanging wall can eventually cause an increase in temperature. Third, shear heating along the fault will cause an increase in temperature near it. Whether the temperature in the hanging wall rises or falls, and the precise evolution of such changes depend upon several parameters: the magnitudes and distributions of the three sources of heat (conduction from below, radioactivity and shear heating), the rate at which displacement of the lower block of the fault advectively removes heat, and the physical properties of the medium (particularly, its coefficients of thermal conductivity and thermal diffusivity).

Because of the large number of parameters that affect the thermal evolution of such a region, and the difficulties in identifying the influences of each, most studies have considered simplifications for the advective transport of heat. The principal simplifications have been to consider diffusion of heat only in the vertical direction

and to simulate thrust faulting by an instantaneous juxtaposition of the hanging wall, with its initial steady state geotherm, over the footwall whose initial surface temperature is zero (e.g. Bickle et al., 1975; Brewer, 1981; England, 1978; England & Thompson, 1984; Molnar et al., 1983; Oxburgh & Turcotte, 1974). Shear heating is accommodated, when required, as a point source in such a one-dimensional configuration (Turcotte & Schubert, 1973).

In conclusion, assuming a direct link between MCT-thrusting and anatectic granitoid formation, Molnar & England (1990) say, '.....to generate granitoid magmas near the MCT, even at relatively low temperatures of 600°C to 650°C, in rocks initially not hotter than 700°C, while underthrusting at a rate of  $15 \pm 5$  mm/yr occurs, requires a large heat source. Heating from below the lithosphere and radiogenic heating within the crust are insufficient to prevent the temperature near the MCT from dropping hundreds of degrees as advection carries heat away rapidly. Dissipative heating with shear stresses of roughly 100 MPa, or more, seem to be the only plausible source of additional heat'.

But, I think, by itself the similarity in oxygen isotope ratios between Manaslu granitoid body and one particular formation in the Higher Himalayas cannot prove that the granitoid melt was derived from that particular formation. It could be simply a coincidence that there is a match between the oxygen isotope values. I presume, we can find similar isotopic signatures in rocks even from far-off continents; that does not help us in any way. However, it is quite plausible that the melt was not derived from beyond the Higher Himalayas. Brunel (1989) pointed out an important possibility i.e. that of 'in situ' melting for the major Higher Himalayan anatectic granitoid bodies.

The deformation of the Chhokang arm of Manaslu granitoid body by MCT is probably a better indication that the intrusion took place before MCT-emplacment rather than synchronously with it. However this is not to suggest that there was no intrusions in the Higher Himalayas during or after the MCT-emplacment. Syn- or even post-MCT minor leucogranitic or aplitic intrusions/veins also abound in the Higher Himalayas:

The third point highlighting the closeness of radiometric ages determined by two different methods may not be very important either. The Rb-Sr whole rock age may have a completely different significance from that of the Ar-Ar mineral date from hornblende. Moreover, Deniel et al. (1987) also <sup>obtained</sup> a Sm-Nd date of c.25 Ma for

Manaslu granitoid and they suggested melting took place probably throughout the 7 Ma period (Sm-Nd age of 25 Ma minus Rb-Sr age of 18 Ma) which, in fact, implies that bulk of the Manaslu granitoid body was emplaced before thrusting along the MCT took place.

Implicit in Molnar & England's (1990) work is an unqualified assumption that the Central Crystalline hanging wall was substantially hotter than the footwall at the time of MCT-movement. I do not find any reason why the relative difference in mean temperatures between hanging wall and footwall has had to be substantially different from that observed today. Relying on all the available dates, the Higher Himalayan crystallines acquired <sup>high</sup> highgrade character during the M<sub>2</sub> metamorphism soon after collision at 50-55 Ma; so nearly 30 Ma had elapsed since M<sub>2</sub> event before these crystallines came on top of the footwall block due to thrusting along the MCT at c.20 Ma B.P. That means heat could radiate away from the Central Crystallines for c.30 Ma before it came in contact with the footwall block. The Joshimath Gneiss (or Central Crystallines, in general) suffered 'kyanite-sillimanite' grade upper amphibolite facies M<sub>2</sub> metamorphism predating the MCT-movement. During the M<sub>3</sub> event (syn-MCT metamorphism) there was garnet-kyanite-hornblende grade condition of lower amphibolite facies. So, what was happening to the Central Crystallines during the time-gap between M<sub>2</sub> and M<sub>3</sub>? The obvious possibility is that the Central Crystallines were losing heat. Prior to the MCT-emplacement did the temperature of the hot crystallines drop below that necessary for garnet-kyanite-amphibole formation, or did the garnet-kyanite-amphibole form at a particular stage during the progressive cooling of the hot crystallines? Intuitively speaking, metamorphic mineral formation (except the retrogressive alterations) in a rock is more likely when the rock is subjected to an **increase** in temperature and probably it is not so likely when the same temperature-pressure is approached in course of progressive cooling or normalisation of P-T conditions (for some hints on this point, see Harte & Johnson, 1969; England & Richardson, 1977). This is why I suspect that prior to the M<sub>3</sub> episode, but after M<sub>2</sub>, the temperature of the Central Crystallines dropped below that favourable for the formation of garnet, kyanite, amphibole etc.. This temperature-drop was not necessarily down to 0°C; probably it went just below the M<sub>3</sub>-temperature before MCT-thrusting. The rise in temperature and pressure during the MCT-emplacement may be due to shearing. The pressure increase can be readily explained as due to this. Many of the small kyanites and amphiboles occur along the latest transposed (main) foliation, and also along some crude shear bands. In addition many syn-MCT garnets show favoured growth along the latest transposed (main) foliation. This preferential growth along high strain areas suggests that shear-heating was

probably the main cause for this temperature-increase during the M<sub>3</sub> metamorphism (contrast Hubbard, 1989).

Also it has not been duly emphasized in Molnar & England's work (1990) that in most cases the zone of anatectic granitoids is away from the MCT (Vaikrita Thrust) by a considerable distance and the heat produced due to shearing will have strongest effect closest to the shear/fault zone and this effect will be gradually less and less away from it. The available metamorphic evidence suggests that the MCT is overlain by kyanite zone (or part of it) that, in turn, is overlain by sillimanite zone which is just opposite to what one would expect if there were considerable shear heating. Geologically speaking, it would be implausible that heating would be so selective as to bypass the zone closest to the fault and affect the area that lies quite a long distance away from the fault. Lithologically there is considerable homogeneity within the Central Crystallines lying above the MCT (V.T.). Here reference could also be made of the 'heat focussing' hypothesis advanced by Jaupart & Provost (1985) and Pinet & Jaupart (1987).

Molnar & England's (1990) and England et al.'s (1992)

MCT refers to what Valdiya (1980) called as the Vaikrita Thrust. In the present study, I <sup>have</sup> followed Valdiya's nomenclature. It is quite hard to imagine that thermal effects of thrusting along this one plane would give rise to such a wide metamorphic zonal structure and still without showing strict correlation between the location of the thrust and spatial position & thickness of the metamorphic zones. We know that the metamorphic zones vary in their width from section to section in the Higher Himalayas. It is also well-known that the MCT cuts across the kyanite isograd, although in most sections the thrust coincides with it. These are more than conclusive evidence that the thrusting along the MCT postdated the 'traditionally famous' inverted Barrovian isograd sequence at the base of the Higher Himalayas.

The six Barrovian metamorphic zones form a progressive metamorphic series, not a retrogressive series. So we cannot expect the following kind of alterations due to cooling: sillimanite → kyanite, kyanite → staurolite, staurolite → garnet; instead we may get sillimanite/kyanite/staurolite/garnet → biotite, biotite → chlorite, garnet → chlorite etc. Thus unless there has been an exceptional kind of "metamorphic convergence", we cannot expect to find such a complete *and* thick sequence of inverted Barrovian zones about the MCT as a result of heating/cooling due to thrusting. Taking the Munsiri Thrust as the lower boundary of the MCT zone, we do not find any metamorphic zonation in the Berinag-Mandhali footwall block due to heating effect.

Thus the possibilities of having a pre-MCT sequence of inverted Barrovian isograds and a not so hot hanging wall block to the MCT (compared to the footwall block) are unavoidable. In the Higher Himalayas only one generation of staurolite has been found which is certainly pre-stretching (i.e. pre-MCT) in origin. Clear examples of stretched staurolite grains have been recognised particularly within the Munsiri Formation. These staurolites are a significant integral part of a "cogenetic" Barrovian sequence; the model envisaged by Molnar & England (1990) and England et al. (1992; to be reviewed below) cannot give satisfactory explanation of how the staurolite zone formed. All these numerous problems that arise when we look at the features of the rocks in detail keeping the "Le Fort" model or its modified version proposed by Molnar, England & coworkers in mind, are mainly due to a misconception about the relative age of MCT-thrusting and inverted Barrovian sequence. When we understand that the inverted metamorphic sequence in the Higher Himalayas is pre-MCT in origin, it becomes easy to visualise that the Central Crystallines were not much hotter than the footwall block during the time of MCT-thrusting which is why there has not been advective cooling effect to a significant degree; there was some shear heating along and immediately around the MCT-Zone giving rise to the syn-MCT  $M_3$  metamorphism, but the Higher Himalayan anatectic granitoid formation did not owe for its heat-sources to the MCT-shearing. Of course, this does not preclude the possibility of some coincidence in time between thrusting and the anatectic melting. I think, post-collisional thickening (mainly by folding and homogeneous thickening) and burial immediately prior to MCT-thrusting might have been responsible to cause melting at deeper part of the crust and the low density molten mass rose above following the shortest available route to give rise to the present granitoid bodies of the Higher Himalayas. It may be possible that some portion of this molten crust got routed through the MCT-Zone (a weak zone itself) simultaneously with thrusting and gave rise to some of the augen gneissic bodies in the present Munsiri Formation.

England et al.'s work (1992) is a follow-up to Molnar & England (1990). Here the authors deal more or less exclusively with the heat sources and physical processes responsible for the regional metamorphism, anatectic granitoid formation and the inverted metamorphism found near the MCT in the Annapurna-Manaslu region of central Nepal. Not much new geological data or observations have been added into this work, except a more detailed theoretical analysis. The authors suggest, 'Following the collision of India with southern Tibet, the crustal rocks of the leading edge of India (1) underwent regional metamorphism to upper amphibolite grade, (2) were locally melted to produce anatectic granitoids, and (3) were sheared and thrust

onto lower grade rock along the MCT, yielding an inverted metamorphic sequence'. With respect to the major phase of deformation associated with slip along the MCT, they recognised broadly three phases of metamorphism:

- I. The upper amphibolite facies (high T - high P) regional metamorphism of the Higher Himalayan Crystallines (but not in the Lesser Himalayan rocks) that preceded movement on the MCT. Termed 'eo-Himalayan metamorphism' by them, this episode correlates well with the M<sub>2</sub> episode recognised in my study. The general range of P-T conditions of this metamorphic episode was T = 650°C - 750°C, P = 750 MPa - 1000 MPa (Pecher, 1989) at a depth of about 25 to 40km below the then earth's surface.
- II. It has been suggested that the second episode of metamorphism accompanied slip within the MCT-Zone (syn-MCT metamorphism) and involved small change in temperature accompanied by a decrease in pressure. Sillimanite has been suggested to have grown replacing kyanite in the top part of this Central Crystallines during this second phase of metamorphism (Pecher, 1989). This syn-MCT metamorphism is correlatable with the M<sub>3</sub> metamorphism recognised in my study; but while I recognised *lower amphibolite* facies condition at this time in Joshimath area, England et al. (ibid) suggest a nearly comparable P-T condition (slightly higher temperature and lower pressure) to that of the eo-Himalayan phase (their first phase) in the Annapurna-Manaslu area. According to England et al. (1992), 'The most striking feature of the thermal evolution of the Higher Himalayan crystalline series during this phase (*syn-MCT*) of metamorphism is the generation of granitoids at its base during slip on the MCT. ... Unlike the first phase, the second phase.....is also recorded in the Lesser Himalayan formations below the MCT. A significant aspect of this metamorphism is that its grade appears to be higher at the MCT than to either side of it. This observation led Le Fort (1975) to suggest dissipative heating in the MCT zone as a possible contributor to this metamorphism'.
- III. the final, retrogressive, phase of metamorphism (Caby et al., 1983) may have involved reactivation of the MCT in a regime of lower temperature and pressure. This third or last phase (so far) of metamorphism is equivalent to the M<sub>4</sub>-metamorphism recognised in Joshimath area.

England et al.'s (1992) work is concerned with the development of only the eo-Himalayan and syn-MCT metamorphism. The major questions they have posed and sought answers for are: (i) What processes and sources of heat are responsible for the eo-Himalayan metamorphism in Central Crystallines? (ii) Given that the thrusting

of cold rock beneath hotter rock should cool the upper block, what heat source could have enabled melting in the Higher Himalayan Crystalline Series during movement on the MCT? (iii) What processes created the inverted metamorphic isograds at the MCT?

The commonly given explanation for the eo-Himalayan metamorphism is that temperatures within the Higher Himalayan Crystalline Series rose after it was buried beneath the rest of the northern margin of India and southernmost Tibet (e.g. Caby et al., 1983; Hodges et al., 1988; Jaupart & Provost, 1985; Pinet & Jaupart, 1987; Pecher, 1989 etc). The parameters that normally control the metamorphic grade at a particular level within a thickened pile of crust are the depth of that particular level below the land surface, the thermal conductivity of the rocks in the pile, the rate of heat input from below the crust, the amount and distribution of heat generated within the crust, and the length of time that is available for thermal relaxation of the pile (e.g. Bickle et al., 1975; England & Thompson, 1984; Oxburgh & Turcotte, 1974). The depth of burial required for the eo-Himalayan metamorphism has been estimated to be 25-40 km. For  $M_2$  metamorphism in Joshimath area, the above information is quite readily applicable. What is lacking in England et al.'s work is a clear indication of the mineral assemblage of the eo-Himalayan metamorphism. In Joshimath area, the most important 'cogenetic' minerals defining the inverted sequence of isograds have been recognised to belong to the  $M_2$  episode (eo-Himalayan metamorphism). A progressive increase in metamorphic grade towards the Vaikrita Thrust (MCT) from both north and south has not been found in Joshimath area; here the thrust causes an abrupt termination of the kyanite zone lying above.

As suggested by England et al. (ibid), the contribution of heat for the syn-MCT metamorphism and development of inverted metamorphic sequence came from two sources: (a) dissipative or conductive heating due to underthrusting of 'cold' footwall block below 'hot' hanging wall, and (b) Frictional heating or shear heating during movement along the MCT.

In my opinion, the two basic premises (which are, in fact, assumptions and not proven facts as indicated by the authors) that led to the theoretical analyses of Molnar & England (1990) and of England et al. (1990) are essentially that,

- (i) there is an inverted metamorphic sequence and there is a thrust (MCT) at the base of the Higher Himalayas. Due care has not been taken to establish the relative age of these two features. It has been just assumed that these two features are genetically related. Detailed fieldwork and careful microscopic

study of suitable rock specimens should have been undertaken first in order to establish properly their relative ages. My detailed observations indicate that the 'traditionally famous' inverted isograd sequence of the Higher Himalayas is **definitely** a pre-MCT feature and, therefore, there is no genetic connection between this inverted metamorphic sequence and the MCT-emplacement.

(ii) the formation of leucogranitic intrusives in the Higher Himalayas has been suggested on the basis of relatively weak evidence to be synchronous with the MCT-emplacement. And, more importantly, then it has been automatically assumed that there is a close connection i.e a cause-and-effect relation between the two events. They suggested that thermal effects of thrusting along the MCT had contributed towards the heat sources necessary for the anatexis melting. It seems to me that MCT-thrusting has been considered by the authors to be too important an event to lead to the development of an well-defined inverted isograd sequence at the base of the Higher Himalayas as well as to cause anatexis melting in the higher parts of the Higher Himalayas quite a long distance away from this inverted metamorphic sequence.

The other main criticisms are:

Firstly, their MCT equates with the Vaikrita Thrust which is the upper boundary of the MCT-Zone but no reference is made in their studies to any thrust at the base of the MCT-Zone equivalent to the Munsiri Thrust.

Secondly, it has been implicitly stressed that the 'well-known' inverted metamorphism of the Higher Himalayas is noticeable mainly below the Vaikrita Thrust (their MCT) and is, therefore, most likely to be directly related to the MCT-emplacement. In my opinion this is clearly not the case. The 'cogenetic' minerals viz. garnet, staurolite, kyanite and sillimanite that define the inverted metamorphic zonation were clearly pre-MCT emplacement and therefore, the simple attribution of the Himalayan inverted metamorphism to the MCT-emplacement is highly misleading. Probably they missed to identify the real Barrovian character of their Eo-Himalayan metamorphic episode.

Thirdly, they consider the MCT as a discrete planar thrust surface, or at most a narrow zone of very limited thickness. But in reality their MCT (Vaikrita Thrust) is just the upper boundary of a complex movement (shear)-zone i.e. the MCT-Zone whose lower boundary is given by the Munsiri Thrust. I think, in any study involving the MCT-Zone and having regional metamorphic and/or tectonic implications it would be erroneous to take up only one boundary instead of the whole movement-zone.

Fourthly, referring to Pecher (1989), England et al. (1992) suggest that sillimanite in the Higher Himalayas developed only during the time of syn-MCT metamorphism. As explained earlier in this chapter, two generations of sillimanite occur in the Higher Himalayas - the earlier generation is, in fact, an integral part of the inverted Barrovian isograd sequence and belongs to my  $M_2$  or Hodges & Silverberg's (1988)  $M_1$  episode (eqv. to England et al.'s Eo-Himalayan metamorphic episode); but their syn-MCT sillimanites are a secondary overprint as a result of Buchan-type metamorphism due to anatectic granitoid intrusions (broadly at the time of my  $M_3$ , or Hodges & Silverberg's  $M_2$  episode). An important criticism of England et al.'s work is that the true Barrovian character of their Eo-Himalayan metamorphism has not been recognised and, therefore, the inverted Barrovian isograd sequence has been erroneously attributed to their syn-MCT metamorphism.

Fifthly, while they emphasized only three episodes of metamorphism in the Central Crystallines, I could recognise at least four distinct episodes of metamorphism in the Central Crystallines and Munsiri Formation in Joshimath area. There was a significant highgrade metamorphic event (most possibly belonging to lower amphibolite facies) even prior to their Eo-Himalayan metamorphic episode. This means the Higher Himalayan Central Crystallines were well-defined metamorphic rocks even before the India-Eurasia collision. Note that their Eo-Himalayan metamorphism has been regarded to have been due to thickening & burial as an immediate consequence of the collision.

Sixthly, as far as investigation into the cause of the Higher Himalayan inverted metamorphism is concerned, England et al.'s model is simply a combined version of Le Fort's (1975) conductive or dissipative heating model and Graham & England's (1976) shear heating model. Proposing shear heating as a viable mechanism for the development of inverted metamorphic gradient in the Pelona Schist area of southern California, Graham & England (1976) thought the Himalayan inverted metamorphic sequence would also be a possible example. They said (p. 143, *ibid*) that in the Himalayas the inverted metamorphic isograds are all developed underlying the MCT (i.e. the Vaikrita Thrust). Implicit in England et al. (1992) is a similar notion where the authors highlight that the MCT overlies the most important part of the inverted sequence of isograds. The isograd distribution within the Central Crystallines above their MCT (the Vaikrita Thrust) is not explicitly dealt with. But it is a well-known fact that in many sections as in Joshimath area the Vaikrita Thrust coincides with the kyanite isograd (see Brunel, 1986; Valdiya, 1980; Brunel & Kienast, 1986; also the present study); kyanite is only rarely recorded below the Vaikrita Thrust (excepting of course the lowgrade Lesser Himalayan occurrences of kyanite). This means, generally speaking, out of the six Barrovian zones, the lower

four viz. chlorite, biotite, garnet and staurolite-zones are found in the rocks lying below the Vaikrita Thrust (their MCT) and the higher two viz. the kyanite & sillimanite zones are mostly developed in the Central Crystallines lying above the Vaikrita Thrust. Clear recognition of the 'cogenetic' development of the inverted metamorphic zones in the Higher Himalayas poses a great problem for applying the 'modified Le Fort model' proposed by England & coworkers.

And, finally, the conclusions arrived at by England et al. (1992) on the thermal causes of Higher Himalayan inverted metamorphism are not supported by my observations in Joshimath area. For its major part, the Higher Himalayan inverted metamorphic sequence is found to be defined by minerals that grew during the  $M_2$  episode predating both the  $F_2$  fold episode and the MCT-emplacement. So, there was no direct genetic link between the MCT emplacement and the Higher Himalayan 'classic' inverted metamorphism. The MCT-emplacement and resulting shear heating, if any, have had no role to play in developing the classic 'cogenetic' Higher Himalayan inverted isograds. The effect of shear heating, if any, due to MCT-thrusting, is expressed by the addition of some later minerals into the already inverted metamorphic sequence. The shear heating was not strong enough to contribute to the formation of the anatectic granitoids at the higher parts of the Higher Himalayas. The syn-MCT metamorphism gave rise to the formation of some garnets, microscopic kyanite crystals, amphiboles, myrmekitisation of plagioclase etc either within or immediately above the MCT-Zone. Molnar & England (1990) and England et al. (1992) modified the original "Le Fort model" a little and added a component of shear heating into it. Still the reason why this "modified Le Fort model" is so much at odds with the observed details of the geological features is mainly because of the misidentification of the relative age of the MCT-thrusting and the traditionally well-known inverted metamorphic sequence.

#### 6.4.2 'Thrust-stacking' model

The model which involves close-spaced thrust-stacking was very favourite for many workers in the sixties and early seventies of this century. Le Fort (1975) mentions "This hypothesis is so widely favoured now-a-days that some authors take it as a fact and deduce the existence of thrusting phases from the limit of the inverted isograds." Even though thrusting along the MCT postdated the highgrade  $M_2$ -metamorphism, it is difficult to visualise stacking of the different 'Barrovian' metamorphic zones in a reverse order simply due to thrusting. Thrusting or faulting

on a series of planes would give rise to 'asymmetric' repetition of the affected strata or order, instead of their 'symmetric' repetition as in case of folding (see Fig. 6.5). On the left-hand-side in the figure the original i.e. pre-faulting sequence of strata is shown, while on the right is shown the sequence after being affected by two reverse faults. Now, following a traverse from point 'C' to point 'A', one would notice the following kind of repetitive occurrence of the sequence of strata:

$$a/b/c/d / a/b/c/d / a/b/c/d/e/f$$

Obviously the repetition here is not symmetric about the faults; this could be called 'asymmetric repetition'. Whereas in case of folding (see Fig. 6.6) we get 'symmetric' or 'mirror-image'-type of repetition about the fold axial plane or axial trace. Here a traverse from C to A would give us the following kind of repetition -  $a/b/c/d / d/c/b/a$  which is evidently a 'symmetric' type or 'mirror-image' type of repetition.

Fig.6.7 shows similar effects of faulting in a metamorphic isograd sequence. Here two thrusts divide the original uniform right-way-up sequence of isograds into three units, A, B & C (from bottom to top), that show along any particular level of observation (say, that given by the rectangle) groups of progressively higher grade metamorphic zones at successively higher units. But when we look at any individual unit carefully, we find that each of them preserves within its limit the normal right-way-up isograd sequence. Translated to plan-view, this would mean that in the field after crossing one thrust if we move about within the same unit (say, B, here) without crossing the other thrust, we would find different metamorphic zones occurring in a right-way-up sequence. Evidently in this sort of thrust-induced 'repetition' or 'inverted stacking' of isograds, many of the metamorphic zonal boundaries encountered along the traverse are not the original Barrovian isograd boundaries, but are "apparent isograds" (tectonic surfaces). This is exemplified more clearly in Fig. 6.8 taking an extreme case.

Here a normal sequence of Barrovian isograds is shown to have been affected by a stack of five reverse faults (thrusts). The result is that the highgrade metamorphic zones occur at higher levels in successively higher thrust-sheets. An important case would arise if one traverses from point 'A' to point 'C'. One will seem to encounter progressively higher grade Barrovian zones at successively higher structural &/or topographic levels which would appear as an ideal example of "inverted metamorphic sequence". But, in fact, none of these zonal boundaries are real isograds, they are simply fault or thrust boundaries (tectonic planes). Metamorphic zonal boundaries along such a traverse are virtually given by the post-metamorphic tectonic boundaries

only and not by the original metamorphic isograds. In the drawing of this figure several assumptions are involved which are very unlikely to be met collectively in nature -

- (a) all Barrovian zones excepting the chlorite and sillimanite are of comparable thickness;
- (b) displacements along the thrusts are of comparable magnitude;
- (c) the movement surfaces (faults/thrusts) are not subparallel to the true isograds, but at an angle to them;
- (d) the 'AC'-line is taken to represent the topographic level or the line of observation. The observer is assumed not to look around to explore the features in the third dimension.
- (e) the topographic level is deliberately made to pass through the thrust-stack in such a way that the chlorite-zone is encountered at the lowermost level and sillimanite-zone at the highermost level, with the intermediate metamorphic zones met at intermediate levels.

This sort of coincidence could, in extreme cases, be possible only in very localised areas. But in case of the Himalayas the inverted sequence of isograds are reported from various sections at different topographic levels and certainly most of the 'Barrovian' zones are bound by original metamorphic isograds and not by 'apparent isograds' i.e. tectonic planes. In fact, for the latter type we should not use the term 'isograd' at all. Sometimes, however, it is possible that a fault or a thrust coincides with a true metamorphic isograd at least within the limit of the observational area. For instance, the Vaikrita Thrust coincides with the kyanite isograd in many sections including Joshimath area.

So, rephrasing one of Le Fort's criticisms (Le Fort, 1975, p.28) of the thrust-stack model, inside every thrust sheet we would expect to see a right-way-up sequence of isograds and clearly this once 'widely held explanation is **inconsistent** with our knowledge of tectonics and metamorphism relationships'.

### 6.4.3 'Folded isograd' model

This model does not suffer from the limitations of other models, and therefore, as demonstrated below, holds the promise to prove to be the correct one for the development of the Higher Himalayan inverted metamorphic sequence.

The zone of Higher Himalayan inverted metamorphic sequence evolved as a polymetamorphic feature. It has recognisable contributions from three episodes of metamorphism ( $M_2$ ,  $M_3$  &  $M_4$ ) and two episodes of deformation ( $F_2$  & MCT-thrusting). However, the most spectacular 'cogenetic' part that typifies the Higher Himalayan inverted metamorphic sequence owes its origin to the  $M_2$  metamorphic episode. During  $M_2$  event the isograds formed in a right-way-up sequence. The largescale overturned  $F_2$  folding that followed the  $M_2$  event, but preceded the peak of MCT-thrusting caused inversion of these isograds on the overturned limb. In the Joshimath area, the easily distinguishable 'cogenetic' part of the inverted metamorphic sequence is given by the garnet-staurolite-kyanite-sillimanite zones. The two low-grade metamorphic zones, biotite *and* chlorite, are largely of secondary retrogressive origin (influenced by  $M_4$  metamorphic episode).

Strictly speaking, proving the 'folded isograd' model, requires confirmatory study spreading across both the limbs of the large-scale overturned fold. But the present study covers only the inverted limb of this fold; nevertheless I found important circumstantial evidence that supports it. As has been emphasized already, the generation of minerals (particularly the highgrade ones) that define the isograds in the Joshimath Gneiss and Munsiri Formation had developed during the  $M_2$  metamorphic episode that preceded the very intense  $F_2$  fold episode which, in turn, predated the peak of the stretching event (main phase of MCT-emplacement). Also established is the fact that the highgrade  $M_2$  event, strictly speaking, did not accompany any intense fold episode. So the metamorphic isograds almost certainly formed in a right-way-up sequence during the  $M_2$  metamorphism. The  $F_2$  event which was the first major fold episode to affect the metamorphic scenario established through the  $M_2$  metamorphism could, therefore, have folded the metamorphic isograds.  $F_2$  folds are clearly south-vergent overturned folds (see subsection 3.1 & 3.2) and developed in a wide range of scale from microscopic to macroscopic/megascopic (see also Viridi, 1986). Thus, in all likelihood, the inverted metamorphic sequence in the Joshimath area is essentially what is preserved in the inverted limb of a largescale  $F_2$  fold. MCT-emplacement was certainly post-metamorphic in terms of the original development of the metamorphic isograds

during the  $M_2$  event. Thrusting along the MCT-Zone probably played a modifying role on shaping the final disposition of the already inverted isograds.

The overturned fold model does not involve the idea of 'apparent' isograds, unlike the thrust-stacking model. As shown in Fig.6.9, the original isograds are folded in such a way that when one approached the core of the largescale fold across its inverted limb following the line 'AC', one would encounter the true progressively highgrade Barrovian zones at successively higher levels.

From the detailed study of deformation-*and*-metamorphism relationships in a very limited area around Joshimath, I suggest that the metamorphic isograd distribution in Joshimath area is consistent with the inverted limb of a largescale post-metamorphic overturned fold. Any comment from my data on the possible location of the normal limb of this fold would be too speculative and, therefore, unwarranted. However, it is worth emphasizing that (i) Viridi (1986) postulated the occurrence of big south-vergent folds in the Central Crystalline outcrops in the Higher Garhwal Himalaya between Joshimath & Pandukeswar (see his Figs.- 1a & 1b for map & section, pp. 157-159, *ibid*); the axial traces of two of such folds (a complementary antiform - synform pair) are shown to be located close to Joshimath; interestingly, these south-vergent macroscopic folds have been recognised by Viridi (*ibid*) to belong to the second episode of deformation and, therefore, correlate with the  $F_2$ -episode recognised in the present study; Viridi recorded presence of crenulation cleavage as axial planar foliation to these folds just as I also recognised in the  $F_2$  folds. My slight disagreements are in regard of the general plunge directions of these folds and the presence or not of a synform running through just north of Joshimath; while shapes of minor folds on the inverted limb of Viridi's 'Pandukeshwar antiform' are clearly indicated in areas north of Vishnuprayag, those on the normal limb of 'Joshimath/Vishnuprayag synform' are not (see his section i.e. Fig. 1a, p.159); I found the same shapes continuing in areas to the south of Vishnuprayag as to its immediate north, which means probably there exists a single huge fold conforming to his 'Pandukeshwar antiform'; (ii) Maruo (1979) reported a progressive upward decrease in metamorphic grade in the Central Crystallines below the Tethyan sediments in Mt. Nandadevi area only ~35 km E of Joshimath; it may be possible that Maruo's study-area falls in the domain of the normal i.e. upper limb of the huge overturned south-verging antiform; and (iii) Searle et al. (1988) convincingly mapped the normal limb and hinge of such a fold in the Ladakh - Zaskar region, which is some 400 km NW of Joshimath. So, in all probability, the inverted sequence of isograds seen at the base

of the Higher Himalayas is essentially that found at the inverted limb of a large-scale post-metamorphic overturned fold (or fold system).

As mentioned in Chapter 3 (Structure), the  $F_2$  folds are multi-order folds developed in a wide range of scale. In the field some of the largest outcrop-scale  $F_2$  folds measure approximately over 200m in half-wave-length and over 500 m in amplitude e.g. the ones found below the Sitapur village and on the cliffs to the east of the Alaknanda river seen from Sitapur village. There is no doubt folds of still higher order must be present in the area. Interestingly, the largest observable folds in the outcrops of Vaikrita Gneiss in Joshimath area are all southerly verging and most of them have long attenuated inverted limbs and comparatively short & thick normal limbs (see Fig. 3.16). Therefore, when viewed downplunge, these folds appear as 'S'-shaped congruent 'parasitic' folds developed on the inverted limb of a major south-verging antiform (Fig. 6.14). This gives concrete indication about the presence of macroscopic  $F_2$  fold in the area/region.

Crucial evidence that kyanite grains are folded by microscale  $F_2$ -folds has been found. So, these kyanites are obviously pre- $F_2$  in origin. These kyanites form an integral part of the inverted Barrovian isograd sequence developed during the  $M_2$  metamorphism in the Higher Himalayas. I have found no reasons to consider this Barrovian isograd sequence to be polyphasal or diachronous in origin in terms of the fold episodes recognised. Based on this direct evidence and other indirect evidence discussed in Chapter 5 (Metamorphism), I have considered that the progressive Barrovian  $M_2$  metamorphism is pre- $F_2$ . So there could be no doubt that the  $M_2$  isograds were folded by the macroscale  $F_2$  fold.

Now an obvious question is: What about the continuation of the isograd-deforming largescale  $F_2$ -folds to the Lesser Himalaya i.e. MCT footwall block?

As argued earlier, the peak of the main stretching event postdated the  $F_2$  folding. So, effectively the MCT-emplacement postdated the  $F_2$  folding. The existence of two later bounding thrusts, viz. MCT to the south and Tethyan Thrust to the north of the Higher Himalayan Central Crystallines, poses practical problems for exploring the continuation of the  $F_2$ -folding beyond the Central Crystallines area. Moreover, when the  $F_2$  folding took place, there was no MCT and the Central Crystallines and the Lesser Himalayan rocks were at least 100 kms away from each other. I think, the  $F_2$  fold-waves did not affect the part of the Lesser Himalayan formations that is now found to underlie the Central Crystallines below the MCT-

Zone. And by the same count, when the cause of the  $M_2$  metamorphism is regarded to be the thickening *and* burial at the leading edge of the Indian continental plate consequent upon the collision with the Eurasian plate, there is little reason to expect any significant effect of  $M_2$  metamorphism (which was clearly pre-MCT and pre- $F_2$ ) in the Lesser Himalayan metasediments that originally lay in far southerly areas. Johnson & Oliver (1990) did not find significant imprint of highgrade metamorphism in their studies on Lesser Himalayan rocks.

Until thrust-emplacement along the MCT took place, the Lesser Himalayan formations were nowhere near to the Central Crystallines that formed the northern edge of the Indian continental mass facing the collisional suture. The fold-forming pulses and the thickening progressed from N to S initially within the Central Crystallines only and at later stages, concomitant with and following the thrusting along the MCT, the Lesser Himalayan Formations also came under the influence of tectonism and metamorphism resulting from continuing northward migration of India. Over the time, the focus of deformation and metamorphism progressed towards the south from the leading edge of the Indian plate involved in collision. The  $M_2$  and  $F_2$  in the Central Crystallines were probably the successive results of initial shortening & burial as immediate consequence of the beginning of post-collisional northward migration of the Indian plate.

That the MCT-emplacement postdated and is, therefore, totally unrelated to the highgrade metamorphism and isograd-development in the Higher Himalayas (Central Crystallines *and* Munsiri Formation) is also indicated by the fact that although in many sections the Vaikrita Thrust coincides with the kyanite isograd, there are some sections (e.g. Sobala, present study) where the kyanite isograd lies above this thrust and in some it lies below (as in Dudh Kosi and Hongu sections near Mt. Everest, see Fig. 9, p.267 in Hodges et al., 1988).

The effect of MCT-emplacement upon the already inverted metamorphic sequence has been mainly in the form of differential thinning *and* thickening of the metamorphic zones that lie within the MCT-Zone. Fig. 6.10 gives the synoptic evolutionary model for the inverted metamorphic sequence and the MCT-Zone in cartoon form.

Any argument that cites the variable presentday thickness of the metamorphic zones from one section to other across the Higher Himalayas (for instance, the Barrovian sequence in Bhagirathi section in western Garhwal is considerably thicker

than in Alaknanda section in eastern Garhwal) in order to counter the 'folded isograd' model cannot be acceptable because such thickness variation can be due to one or more of at least the four following factors: (i) syn-M<sub>2</sub> original metamorphic feature which means, original metamorphic zones were not equally thick everywhere (we cannot expect the Barrovian zones to develop with profound regularity in thickness in a region); (ii) an influence of F<sub>2</sub> folding, as a result of differential attenuation on the overturned limb/s (longitudinal variation in limb thickness is quite common in natural folds, particularly in the largescale ones); (iii) differential layer thinning &/or due to horse-effect during MCT-thrusting (cf. Fig. 6.2, in Subsection 6.1.1) and, (iv) variable influence of overprinting by M<sub>4</sub> retrogressive metamorphism, particularly in the lower part of the Munsiri Formation.

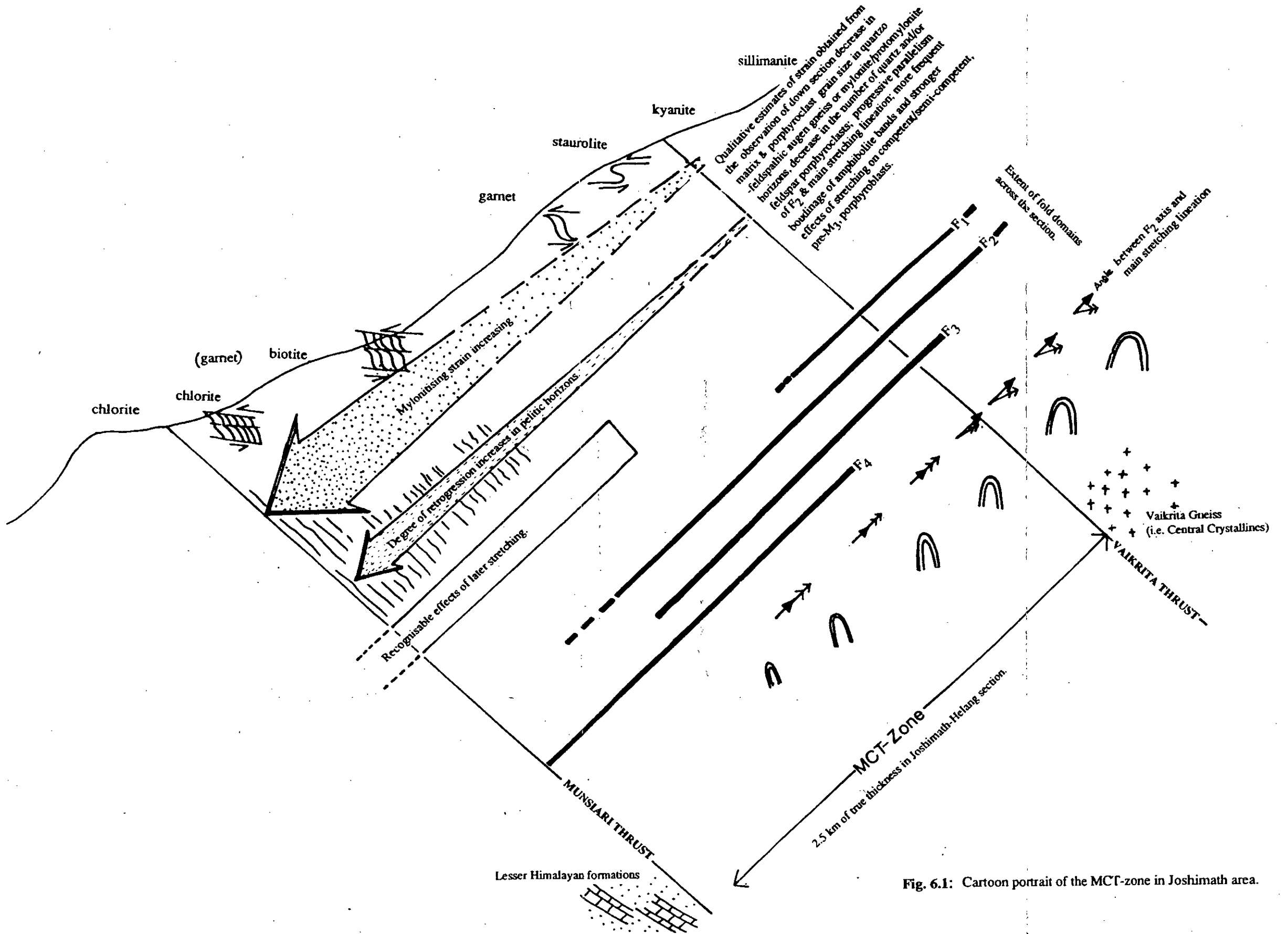


Fig. 6.1: Cartoon portrait of the MCT-zone in Joshimath area.

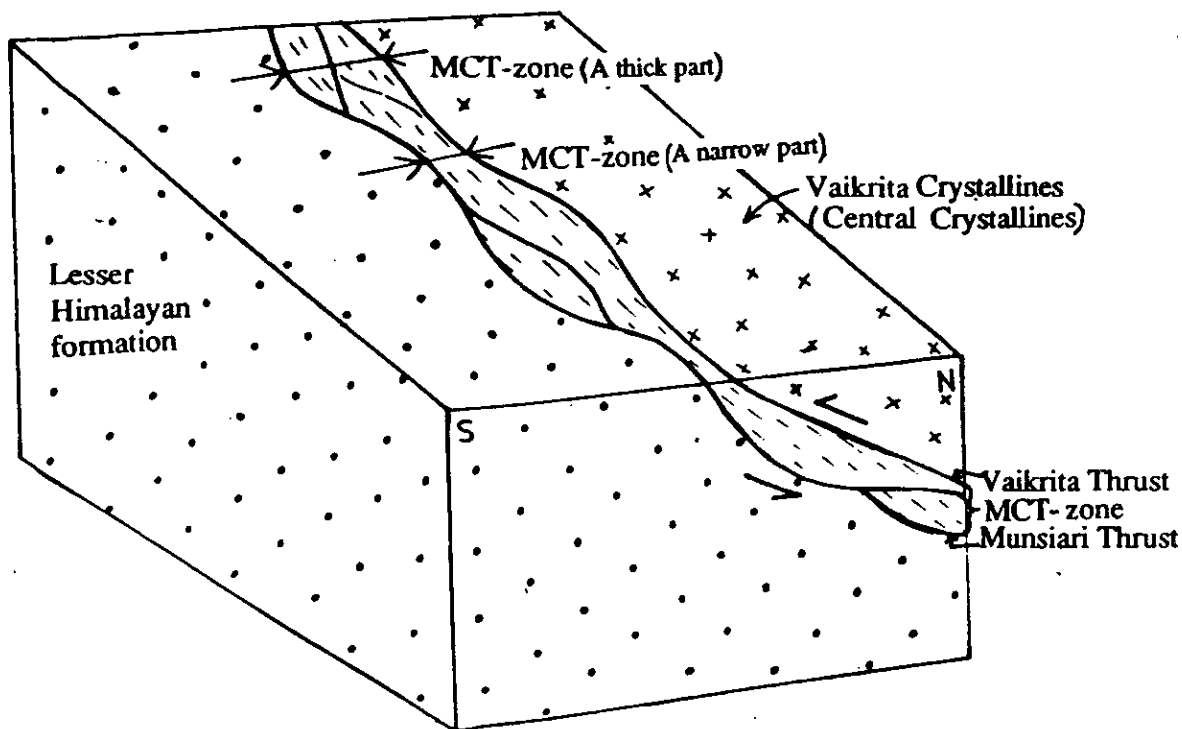


Fig. 6.2(a): Cartoon (not to scale) portraying anastomosing horse that induced along-strike and across strike variation in thickness of the MCT-zone.

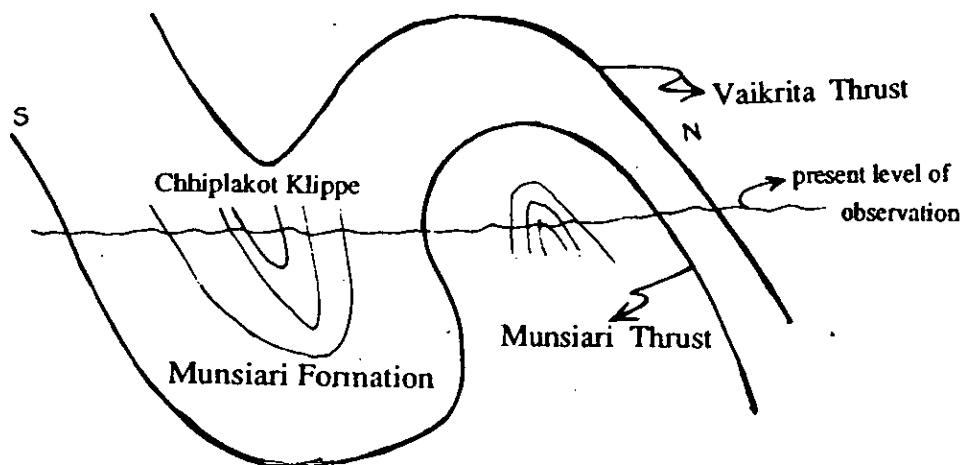
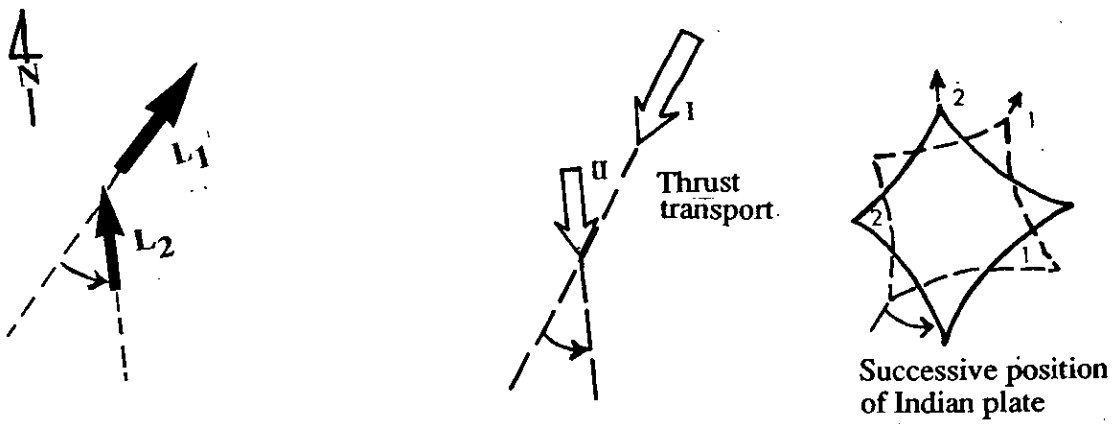
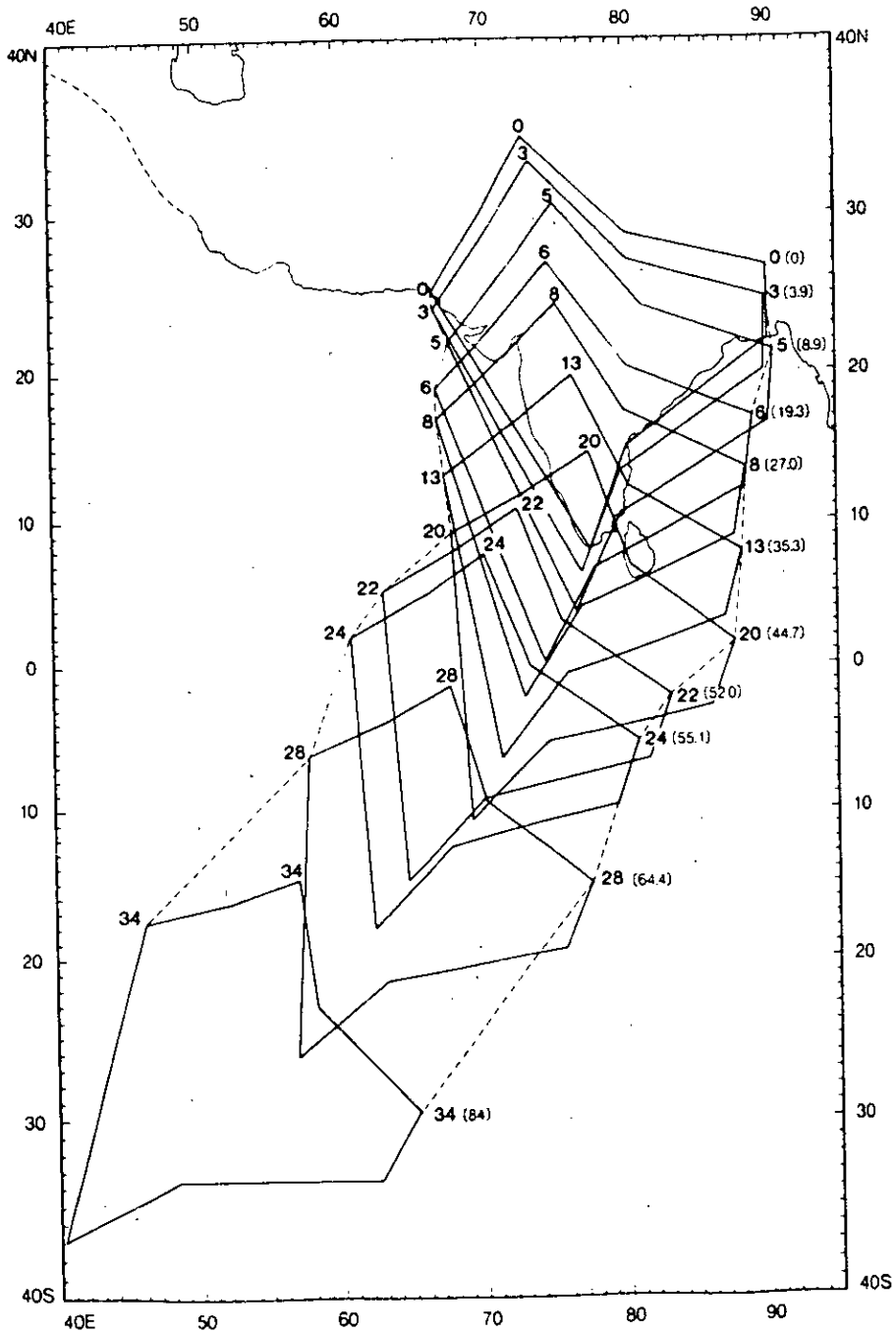


Fig. 6.2:(b): Regional structural interpretation of Sobala section (Darmaganga section) (after Valdiya, 1980).



**Fig. 6.3** Schematic demonstration of a possible correlation between two generations of stretching lineations ( $L_1$  = dominant;  $L_2$  = later), change in MCT movement trajectory over time and counter-clockwise rotation of the northward drifting Indian plate.



**Fig. 6.4:** Successive positions of India in the Eurasian reference frame. Numbers refer to magnetic anomalies; numbers in brackets refer to ages in million years. (from Dewey et al, 1989)

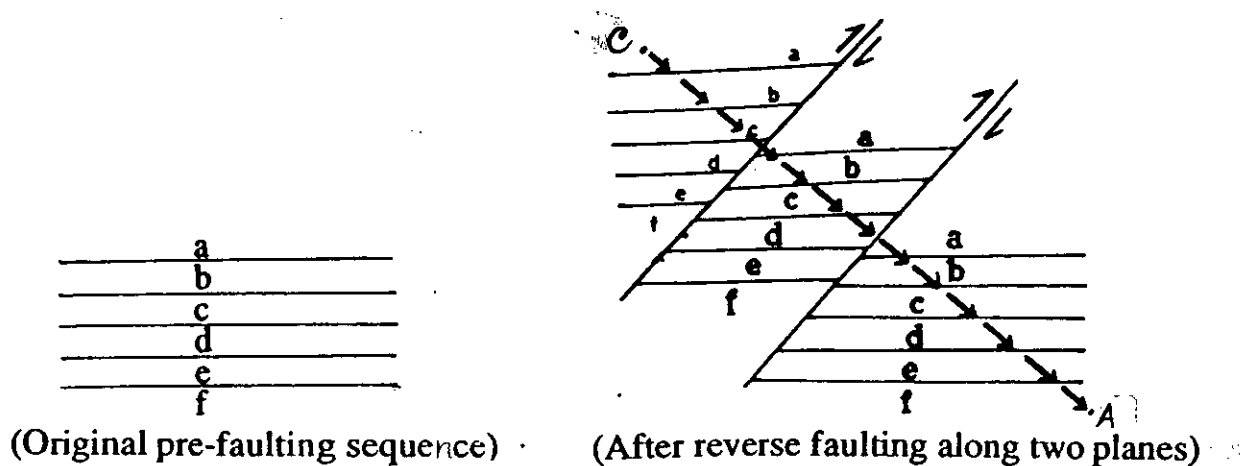


Fig. 6.5: 'Asymmetric' pattern of repetition of a horizontal set of strata after reverse (thrust) faulting along two planes. See text for further discussion.

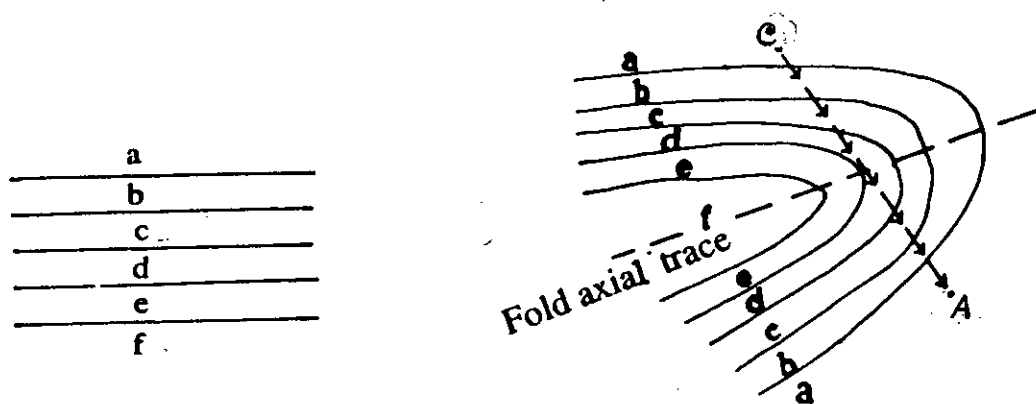
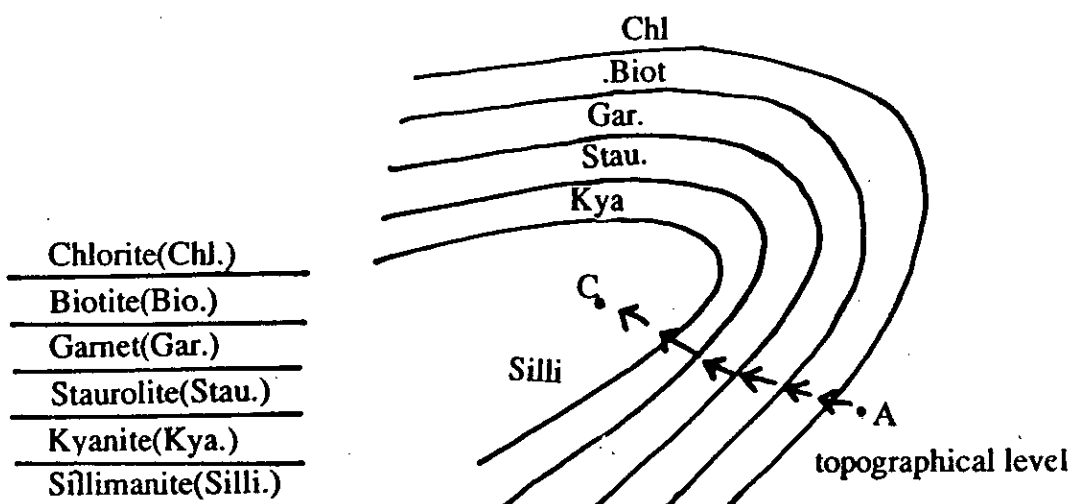


Fig. 6.6: 'Symmetric or Mirror-image' pattern of repetition due to folding. See text for details.





**Fig. 6.9:** Inverted disposition of metamorphic isograds caused by overturned folding. The inverted isograd sequence found at the overturned limb does not involve 'apparent', but 'true' isograds. Further discussion is in the text.

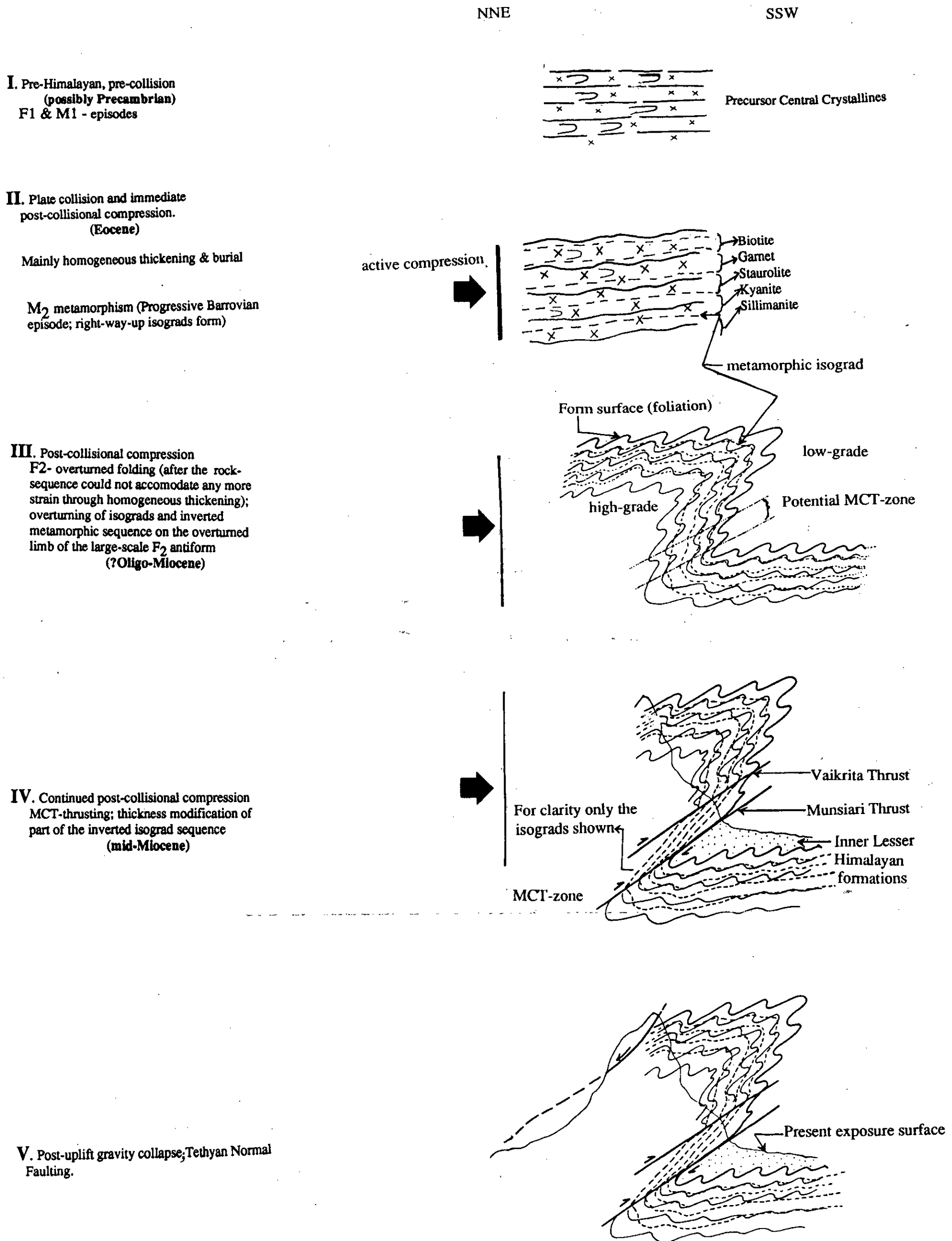


Fig. 6.10: Synoptic evolutionary model showing the development of the Higher Himalayan inverted metamorphic sequence and its relationship with the MCT thrusting. Drawing not to scale.

## Chapter-7

# CONCLUSIONS

### 7.1 THE MAIN CONCLUSIONS

1. (a) The Main Central Thrust Zone (MCT-Zone) is a complex shear zone with a tectonically more clear and abrupt lower boundary than the upper. In metamorphic terms, however, there is an apparent "convergence" due to late retrogression across the lower boundary, whereas a somewhat abrupt break across the upper boundary.

(b) Developed at the base of the Higher Himalayas, the MCT-Zone (or, 'Munsiari Formation' as recognised by Valdiya, 1980) is made up of variably sheared semipelites, pelites, quartzites, augen gneisses, quartzofeldspathic protomylonites and mylonites, phyllonites, amphibolites and occasional marble bands. The MCT-Zone is overlain by highgrade gneissic rock ('Vaikrita Gneiss') and underlain by lowgrade (chlorite-grade) Lesser Himalayan sedimentary rocks (locally called 'Berinag-Mandhali formations').

(c) Sense of shear revealed by indicators of different types and scale reflects a consistent SSW-ward overthrusting that has brought highgrade rocks on top of the lowgrade rocks.

(d) Embodying a realm of lithology-controlled strain partitioning, the MCT-Zone shows a general downward increasing gradient in strain.

(e) Variation in grain-size and mylonitising strain are better criteria for the recognition of the MCT-Zone in the field; the criterion based on the first appearance of kyanite for marking the upper boundary of the zone is not valid in every section.

2. (a) The dominant mesoscopic fabric in the MCT-Zone includes a strong NE-dipping foliation with a well-defined NNE-trending stretching lineation on it. The main foliation is uniform in orientation throughout the MCT-Zone and beyond. A later generation of N-trending stretching lineation is also recognised which gains in prominence in the middle to lower part of the MCT-Zone.

(b) The presence of two generations of stretching lineations implies two stages of emplacement along the MCT-Zone -- the dominant NNE-trending set is earlier and corresponds to the main emplacement, while the later N-trending set represents a reactivational stage postdating regional retrogression (i.e. the  $M_4$  episode of metamorphism). Kinematic indicators associated with both the stretching lineations give overthrust sense of shear.

(c) The two differently trending stretching lineations belonging to two distinctly different times reflect a temporal change in the direction of thrusting along the MCT-Zone. Probably this is an evidence indicating partially the continued change in the motion direction of the Indian plate even after the India-Eurasia collision.

3. (a) Taking the MCT-Zone and Joshimath Gneiss together, dominantly four episodes ( $F_1$ ,  $F_2$ ,  $F_3$  &  $F_4$ ) of folding have been recognised. The frequency of occurrence of folds of successively younger episodes increases from N towards S. However, the second fold episode is by far the most prominent one and is manifested in a wide range of scale. These  $F_2$  folds are multi-order, south vergent, overturned folds.

(b) The shearing corresponding to the main stage of MCT emplacement was a considerably prolonged phenomenon starting broadly synchronously with the  $F_2$  fold episode, but its culmination (i.e. peak of the main stretching event) came about during the  $F_3$  episode.

(c) Of the three episodes of folding ( $F_I$ ,  $F_{II}$  &  $F_{III}$ ) recognised in the Berinag-Mandhali footwall block, the  $F_I$ -episode was more dominant than the other two. The most easily identifiable foliation in the exposures of the Berinag-Mandhali rocks is a typical transposition foliation originally developed as axial planar foliation to the  $F_I$  folding. Most presumably, the  $F_I$  folding in the footwall took place in response to the onset of the MCT emplacement above. No correlation was found between the folds in the hanging wall and footwall blocks.

4. (a) Imprints of four distinct metamorphic episodes ( $M_1$ ,  $M_2$ ,  $M_3$  &  $M_4$ ), of which the last is retrogressive, have been recognised in the MCT-Zone and Joshimath Gneiss. Of these, the  $M_2$  episode records the highest grade attained; all the high-grade minerals viz. garnet, staurolite, kyanite and sillimanite, and the corresponding isograds which are now involved in the inverted metamorphic sequence, originally developed during this high-grade  $M_2$  Barrovian event. No indication has been found

that could suggest the  $M_2$  metamorphism took place under the influence of an inverted geothermal gradient; most probably the  $M_2$  Barrovian zones originally developed in a right-way-up order. While  $M_2$  was almost certainly immediate post-collisional in origin, the  $M_1$  metamorphism was pre-collisional, most probably Pre-Tethyan (i.e. Precambrian) in age.  $M_1$  metamorphism probably took place in lower amphibolite facies condition.  $M_3$  was the syn-MCT emplacement metamorphic episode and took place in *lower amphibolite* facies condition.

(b) There was no one-to-one correlation among the folding (deformation) and metamorphism events except at the  $M_3$  and  $F_3$  time. The  $M_1$  metamorphism took place at pre- or syn- $F_1$  time;  $M_2$  metamorphism at pre- $F_2$ , but syn- or, more possibly, post- $F_1$  time; the  $M_3$  (syn-MCT) metamorphism occurred broadly synchronously with the  $F_3$  folding, whereas the retrogressive  $M_4$  metamorphism took place possibly shortly after the  $F_4$  folding.

(c) Thus, the highest grade metamorphism (sillimanite grade, upper amphibolite facies) recorded in the MCT-Zone rocks and above is a pre-MCT feature. The syn-MCT metamorphism ( $M_3$ ) was not the first metamorphism affecting these rocks.  $F_2$  folding post-dated the highest grade ( $M_2$ ) metamorphism, but was earlier than the peak of the MCT-thrusting.

(d) Thrusting along the MCT was hot enough to induce pervasive recrystallisation of matrix minerals, particularly quartz, and transposition of early foliations. Study of the nature, shape and size of quartz grains in the matrix of different rocks from the MCT-Zone indicates that matrix coarsening took place at the late-syn main stretching stage. This dominant matrix recrystallisation time is vital to the interpretation of any quartz or calcite petrofabric data collected from across the MCT-zone.

(e) In the Berinag-Mandhali foot-wall block, imprints of four metamorphic events ( $M_I$ ,  $M_{II}$ ,  $M_{III}$  &  $M_{IV}$ ) have been recognised. None of these events went beyond the upper greenschist facies of metamorphism. The  $M_I$  event did not accompany any recognisable fold episode and most possibly took place well before the MCT-emplacement. The  $M_{II}$  event accompanied the  $F_I$  folding; the  $M_{III}$  event took place at the interkinematic stage between  $F_I$  and  $F_{II}$  folding, whereas the dominantly retrogressive  $M_{IV}$  event occurred in between the  $F_{II}$  and  $F_{III}$  fold episodes.

(f) No definite correlation of tectonometamorphic events between the hanging wall and the footwall could be established.

5. (a) The inverted metamorphic sequence in the Higher Himalayas is a polymetamorphic feature. But the most classic part of the inverted isograd sequence is defined by minerals that developed cogenetically during the  $M_2$  episode predating both the  $F_2$  folding and MCT thrusting.

(b) The relative chronology of deformation and metamorphic events established through the present study suggests the following:

i) The multi-order, south-vergent  $F_2$  overturned folding which took place following the  $M_2$  high-grade Barrovian metamorphism, most probably folded the originally right-way-up  $M_2$  isograds, so that these isograds are now preserved in an inverted order in the overturned limb of a large-scale antiformal  $F_2$  fold at the base of the Higher Himalayas.

ii) MCT-thrusting post-dated both  $F_2$  folding and  $M_2$  metamorphism, and hence there was no genetic connection between the MCT-emplacement and the Higher Himalayan inverted metamorphic sequence. MCT-thrusting played a passive thickness-modifying role on an already inverted isograd sequence.

iii) The results from the present study favour the "post-metamorphic modification" model for the development of the Higher Himalayan inverted metamorphic sequence.

## 7.2 A LIST OF SUGGESTED FOLLOW-UP WORK

I. An accurate specification of P-T conditions during MCT-emplacement and exploration of the thermal behaviour of the Higher Himalayan crystalline rocks during the time-gap between  $M_2$  and  $M_3$ .

II. Further refinements of the sequence of deformation and metamorphism, and systematic and detailed geothermobarometric and radiometric study along the Alaknanda-Dhauliganga and Darmaganga traverses in order to establish a very accurate P-T-t evolution. An important aspect of this study would be to explore the normal limb of the large-scale  $F_2$  'overfold' that folded the  $M_2$  isograds.

III. Detailed and systematic petrofabric (optical &/or X-ray goniometric) study of quartz, calcite, feldspars &/or amphiboles supplemented by SEM/TEM studies in order to establish the relative roles of different deformation micromechanisms through space and time.

IV. Selected petrochemical study (i) for the establishment of the ortho-/para-origin of the metabasic rocks occurring within and around the MCT-Zone, (ii) for identifying the palaeotectonic setting of the ortho-amphibolites, (iii) for exploring the nature of mylonite protoliths, (iv) for estimating the volume loss during mylonitisation using the mylonite trace element chemistry (following the line established by K. O'Hara, 1988, 1989).

V. Systematic mineral age and closure temperature studies in order to establish the pattern/s of uplift in the Higher Himalayas.

VI. Mineral radiometric studies and fluid inclusion studies of vein & matrix minerals combined with necessary stable isotope analyses in order to establish the role of fluids in the MCT-emplacment and regional tectonometamorphic evolution.

VII. Isotopic and electron and ion-microprobe studies to investigate the role of fluids and microchemical processes involved in the transformation of amphibolite into shiny black lustrous biotite phyllonite.

VIII. Exploring the pattern of recent uplift in the Higher Himalayas from systematic tectonic analysis of the very late brittle/ductile structures.

IX. To work out the tectonometamorphic implications from exhaustive analyses of zoning profiles in garnets.

X. Investigation into physical and chemical processes involved in the development of shear-related boudinage and other structures in the area.

*Taking a holistic view, no two things are mutually exactly alike or equal in Nature; they can't be, because in the time-space continuum everything has its own special place. Indeed there is always a limit to generalisation.*

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## APPENDIX - I

### Computer program used for making equal area projections from field structural data

The following fortran program that was used for preparing equal area projection diagrams from field structural data was written by Susmita Ghosh in the Edinburgh University Geography Department.

The program takes orientation data in the form of lines only. Thus the planar data were fed in polar form, whereas the linear data were readily accepted. Data input was in the form of inclination (A) and azimuth (B). For each of the data that represented horizontal lines, there were two inputs corresponding to two azimuths 180° apart. An Hewlett-Packard Plotter was used for making the outputs. The plotter routine was available from the Department of Geography, University of Edinburgh.

```

C      PPROG --Program to calculate the coordinates of orientational
C      ----- data on Equal-Area Polar Net
      IMPLICIT NONE
      INTEGER COUNT, I, J, PENUP, INTX, INTY, INTL, INTM, INTP, INTQ, INTS
      INTEGER INTE, INTF, INTG, INTH, INTR, INTSYM
      REAL A(1000), B(1000), D, X, Y, R, L, M, VL, VL2, VL3, POD, POD2, POD3, SYM
      REAL E, F, Z, S, Q, P, Q, H, PI, D1
      CHARACTER *40 FILENAME
      PARAMETER(PI=3.14159)

C      -----
      OPEN(UNIT=19, FILE='GEOPROG.PLT', STATUS='NEW')
      SYM=0.04
      WRITE(*,*) 'ENTER DATA FILENAME: (eg.RAT1.DAT)'
      READ'(A)', FILENAME
C      Open the Plot file
      CALL STPLOT
C      Request to input Radius(R) of the Hemisphere which is read in cms
      WRITE(*,*) 'PLEASE TYPE RADIUS(R) OF HEMISPHERE WHICH WILL BE'
      WRITE(*,*) 'READ IN CENTIMETRES'
      READ(*,*) R
      PENUP=1
      SYM=SYM/2
      SYM=SYM*400
      INTSYM=SYM
      WRITE(*,*) 'PLEASE WAIT, CALCULATION IN PROGRESS...'
C      Calculation of x,y coordinates for Hemisphere Circle with radius(R):
      DO 50, I=1, 360, 2
      VL=(I*PI)/180
      VL2=SIN(VL)
      VL3=COS(VL)
      X=R*VL2
      Y=R*VL3

C      To convert values relevant to H.P.Plotter X and Y are multiplied
C      by 400 and to shift initial X,Y to the centre 6000, 4000 are
C      added respectively.
      INTX=(X*400)+6000
      INTY=(Y*400)+4000
      CALL DRPLOT(INTX, INTY, PENUP)
      PENUP=2
50    CONTINUE
C      -----
C      DRAW BOX AROUND THE CIRCLE [P=6000-(R*400)]
      Z=R+1
      Z=Z*400
      P=6000-Z
      Q=4000-Z
C      Assigning value to S which makes the box 1cm outside the circle
C      S = (R*2*400) + 800
      S=R*800
      S=S+800
      PENUP=1
      INTP=P
      INTQ=Q
      INTS=S
  
```

```

CALL DRPLOT (INTP,INTQ,1)
CALL DRPLOT (INTP,INTQ+INTS,2)
CALL DRPLOT (INTP+INTS,INTQ+INTS,2)
CALL DRPLOT (INTP+INTS,INTQ,2)
CALL DRPLOT (INTP,INTQ,2)
-----
C
C Draw cross at the centre of the circle
E=6000
F=4000
R=R*400
G=R+40
H=R+100
INTE=E
INTF=F
INTG=G
INTH=H
INTR=R
-----
CALL DRPLOT (INTE,INTF,1)
CALL DRPLOT (INTE,INTF+40,1)
CALL DRPLOT (INTE,INTF+40,2)
CALL DRPLOT (INTE+40,INTF,1)
CALL DRPLOT (INTE+40,INTF,2)
C Draw lines protruding from the circumference at E,W,N,S direction
CALL DRPLOT (INTE,INTF-INTR,1)
CALL DRPLOT (INTE,INTF+INTG,2)
CALL DRPLOT (INTE-INTR,INTF,1)
CALL DRPLOT (INTE+INTG,INTF,2)
CALL DRPLOT (INTE,INTF+INTR,1)
CALL DRPLOT (INTE,INTF+INTH,2)
CALL DRPLOT (INTE+INTR,INTF,1)
CALL DRPLOT (INTE+INTR,INTF,2)

C Read data from PROJ.DAT file consisting of inclination(A) and
C Azimuth(B) data
OPEN (UNIT=10,FILE=FILENAME,STATUS='OLD')
C Subprogram --Calculation of coordinates(L,M) of orientation data
C where D is the distance from the centre of the circle.
C Calculation of D :
R=R/400
DO 150,I=1,1000
C Read Data
READ(10,*) A(I),B(I)
C Check for end of data where A >90 and B>360
IF (A(I).GT.90.AND.B(I).GT.360) GOTO 200
C Equation for division along meridians:D=R*SQRT2*(SIN(45-(A/2)))
D=(A(I)/2)
D=45-D
D=(D*PI)/180
D=SIN(D)
D1=SQRT(2.0)
D=D1*D
D=R*D
D=D
C Calculation of coordinates of orientation data :
C Finding the coordinates(L,M) as L=D*SIN(B), M=D*COS(B)
POD=(B(I)*PI)/180
POD2=SIN(POD)
POD3=COS(POD)
L=D*POD2
M=D*POD3
INTL=(L*400)+6000
INTM=(M*400)+4000
C Plot a cross at each L,M to represent orientation data :
CALL DRPLOT(INTL-INTSYM,INTM+INTSYM,1)
CALL DRPLOT(INTL+INTSYM,INTM-INTSYM,2)
CALL DRPLOT(INTL-INTSYM,INTM-INTSYM,1)
CALL DRPLOT(INTL+INTSYM,INTM+INTSYM,2)
150 CONTINUE
CALL FIPLOT
CLOSE(19)
200 STOP
END
-----
C

```

```

CA----- Hewlett-Packard Plotter routines -----
C----
C---- This consists of 4 routines to control the Hewlett-Packard 7550 plotter
C---- They routines write to a file on channel 19 which the user must have
C---- previously OPENED. It is also the responsibility of the user to CLOSE t
his
C---- file after the finish plotter routine has been called.
C----
C----
C---- After exiting the program, the file can be sent to the plotter with the
C---- HPLOT filename
C---- command. BEFORE SENDING A PLOT FILE, however, check that the last line
C---- contains the PG; command as otherwise, the next person's plot may be dr
awn
C---- over yours, such to their, and your, annoyance.
C----
C----
C---- The routines are  STPLOT : open the plot file
C----                   FIPLOT : finish the plot file
C----                   DRPLOT : draw a line, or move the pen
C----                   PEPLLOT : change the pen
-----
C---- The parameters are described at the top of the subroutines.
C----
C----
SUBROUTINE STPLOT
IMPLICIT NONE
CC--- OPEN PLOTTER ROUTINE
C----
CP--- no parameters
C----
C----
WRITE(19,*) 'PG;'
WRITE(19,*) 'IN;IP 0,0;SP 1;'
WRITE(19,*) 'VS10;'
C---- the above line sets the pen speed to its slowest, and best quality.
C---- you may increase the speed from 10 to 65 for draft copies.
RETURN
END
SUBROUTINE FIPLOT
IMPLICIT NONE
CC--- CLOSE PLOTTER ROUTINE
C----
CP--- No parameters
WRITE(19,*) 'PG;'
RETURN
END
SUBROUTINE DRPLOT(X,Y,IP)
IMPLICIT NONE
CC--- Draws a line, or moves the pen, to the point X,Y in HP plotter units
C----
CP--- X,Y - INTEGER - coords, in HP units, of point to draw or move to
CP--- IP - INTEGER - pen code =1 pen up and pen is moved without drawing
CP---                   =2 pen down and line is drawn
C----
C----
INTEGER X,Y,IP
C----
IF(IP.LT.2) GOTO 100
WRITE(19,*) 'PD',X,Y,';'
RETURN
100 CONTINUE
WRITE(19,*) 'PU',X,Y,';'
RETURN
END
SUBROUTINE PEPLLOT(IPEN)
IMPLICIT NONE
CC--- changes pen colour
C----
CP--- IPEN - INTEGER - the pen colour requested
CP---                   if outside range 1-8 then pen 1 selected
C----
C----
INTEGER PEN,IPEN
PEN=IPEN
IF(PEN.LT.1.OR.PEN.GT.8) PEN=1
WRITE(19,*) 'SP',PEN,';'
RETURN
END

```

## APPENDIX - II

**Petrographic description and discussion of some important thin sections**

Here description of the petrographic (textural) features from some of the important thin sections will be given as a supplement to the discussions in Sections - 5.3 & 5.4 in Chapter-5.

**1. Slide No. 1F'87 (Possible sillimanite-bearing Joshimath Gneiss)**

The main minerals are - quartz, plagioclase, orthoclase ( <<plagioclase ), biotite, muscovite, garnet, epidote, ?sillimanite, apatite and opaques. At one place in the upper right of the slide, tufts of aluminosilicate occur in between two plagioclase grains. These are most probably fibrolitic variety of sillimanite. The fibrous aggregate is intimately associated with fine white-micas (probably alteration product). The general rock texture is high grade gneissic, so it is quite possible that some sillimanite tufts are preserved here. The general alignment of the fibres follows the orientation of the gneissic foliation. Even though the fibrous mass itself is not involved in smallscale fold hinge, the thin section clearly shows an F2 hinge defined by the gneissic foliation. In fact, on close scrutiny, a well defined hook-shaped, type-3 interference pattern between F2 and F1 is recognised, defined by biotite

**2. Slide No. 22/4/88B (Kyanite-bearing Joshimath Gneiss)**

The original hand specimen comes from over 1.5km above the Vaikrita Thrust. The main mineral constituents are: quartz, plagioclase, orthoclase ( <<plagioclase ), biotite, muscovite, kyanite (in two size classes - larger and smaller), garnet and apatite. In general, garnet is less common in kyanite-bearing horizons than in others within the Joshimath Gneiss.

In the lower-left of the thin section there is an excellent example of a large kyanite grain folded by an F2 hinge. The straining effect on the grain due to folding is reflected by the migratory character of its extinction (strain shadow) and also by kinking and fragmentation of small parts from the sides of the main grain which maintains a conformable relation with the fold. Thus the larger kyanite grains must have developed before F2 folding. The folded kyanite grain is 1.73mm in width, contains inclusions of quartz and micas and is surrounded mainly by quartz which show strong undulose extinction. Many of the other kyanite grains show simple deformation twinning. There has been development of micas (muscovite & biotite) in the fractures within the kyanite grains, possibly at a stage following the F2 folding. The close association of the larger kyanites with biotites possibly suggests a reaction relation between biotite and kyanite.

Another important point is the presence of tiny prismatic grains of kyanite in the slide. These smaller kyanite grains are fresh (i.e. unaltered), do not show any conspicuous undulose extinction and lie along the contacts of bigger plagioclase and/or quartz grains. None of them show any folded outline. At one stage I had misidentified these slender prisms as sillimanites. These smaller kyanites do not show any strong preferred orientation, but they are broadly confined in linear zones that follow the orientation of the gneissosity i.e. main foliation. Even though individually these small prismatic grains are not folded they occupy the main foliation which is folded. While the small kyanite prisms occur mostly at the contacts of plagioclase or quartz grains and are not closely associated with micas, the larger kyanite grains occur without any regard to the quartz or feldspar grains. The smaller and the larger kyanite grains are sometimes closely associated with each other. On closer observation, however, it is found that the precise orientation of the smaller grains is not in conformity with the alignment of the larger grain. Also there are many examples where even though they are spatially close by, the smaller grains are distributed with clear disregard to the orientation of the larger grain.

**3. Slide No. ~NR 38-39 (Kyanite-bearing Joshimath Gneiss)**

The hand specimen comes from a rolled boulder lying in between Stns NR 38 and 39 near the exposure of the Vaikrita Thrust on the new bypass road under construction. The rock probably represents a kyanite-bearing horizon in the lower part of the Joshimath Gneiss lying immediately

above the Vaikrita Thrust. The main minerals are: quartz, plagioclase, some orthoclase, biotite (in part chloritised), kyanite (large and small), garnet, apatite and opaques.

Kyanite occurs in the thin section normally as tabular, platy or bladed grains, some of which are elongate in habit; length of these grains is about 0.6mm and breadth or width about 2.1mm. However, tiny needle-like prismatic grains of kyanite, length ~ 0.27mm and breadth ~ 0.07mm, are also found adjacent to large kyanite grains or bordering some quartz and plagioclase grains. Kyanites, particularly the bigger grains, are closely associated with biotite. Sign of alteration of the kyanite grains is not strong except the occurrence of patchy biotite in the cleavage-fractures of some grains; but there is conspicuous evidence of retrogressive effects in the thin section e.g. biotite is chloritised at many places. In the upper part of the thin section, there is an elongate kyanite grain with folded outline. Migration of extinction is clearly seen across the hinge of this folded grain. This feature clearly indicates that there has been folding following the kyanite formation. A similar-looking folded kyanite grain is also noticed in slide NR 37/38(2). No conspicuous F2 or F3 fold-hinges are found in this thin section and hence it is difficult to correlate the fold in the kyanite grain with any particular fold episode. However, the observation is in conformity with the pre-F2 origin of kyanite as indicated clearly in the thin section 22/4/88B.

#### 4. Slide No. NR 37/38(2) (Kyanite-bearing Joshimath Gneiss)

The hand specimen comes from near the Vaikrita Thrust from a rolled boulder lying on the new bypass road under construction and presumably represents the kyanite-bearing Joshimath Gneiss horizon just overlying the Vaikrita Thrust. The thin section is cut subparallel to main foliation and so most of the biotite grains have 001-faces aligned subparallel to it. Also because the rock is considerably retrogressed as discussed below, the biotite grains look rather patchy in thin section giving a somewhat homfelsic appearance to the rock. The main minerals are: quartz, plagioclase, some orthoclase, biotite, chlorite (altered from biotite), kyanite (larger and smaller grains), garnet, negligible muscovite, apatite, opaques and ?rutile aggregates.

Kyanite is very common in the rock. A number of kyanite grains show spectacular simple twinning. Most of the kyanite grains show migratory (undulose) extinction implying that after the growth of kyanite the rock has suffered considerable strain. The extinction in the quartz and plagioclase in the matrix is heavily undulose. The way in which some of the kyanite grains are wrapped around by matrix foliation and/or the foliation is pushed or bent (that is, indented) by kyanite near the hinge of a fold near top-middle part of the slide suggests that kyanite was pre-tectonic at least in terms of this folding. This is further confirmed by the presence of a folded kyanite grain at the top left part of the slide. These observations match well with the pre-F2 kyanites shown by thin section 22/4/88B. even though it has not been clearly established which generation the present folds belong to.

The most important difference in the character of kyanites in this section from those in section 22/4/88B is that here the kyanites do not look fresh. Evidently retrogression has affected them to a considerable extent. The thin section shows good examples of alteration from biotite to chlorite, garnet to chlorite or even kyanite to biotite &/or chlorite. For instance, one mildly folded kyanite grain at the lower-right part of the section probably shows biotitisation along cleavage-fractures; another smallish kyanite grain in between two bigger grains is completely engulfed within retrograde chlorite. Upon rotation of the microscope stage, extinction migrates across chlorite into the kyanite without any apparent discontinuity at the kyanite grain boundary which is diffused. Also there is greater effect of stretching and deformation because almost no kyanite grain has a smooth cross-boundary. The transverse or cross boundaries are, in most cases, jagged. The particular way in which the section is cut cannot explain the jaggedness which is present in almost all the kyanite grains. In none of the kyanite grains is a well developed crystal outline preserved.

As in many other kyanite-bearing thin sections of Joshimath Gneiss, we find tiny needle-like crystals of kyanite lying in clusters along the borders of some plagioclase or quartz grains. Also found are tiny kyanite crystals which project well into plagioclase augens. The above three figures show tiny kyanite crystals developed round the augenshaped plagioclase and/or quartz grains giving. These figures give sufficient indication that these kyanites grew either synchronously with or postdating the augen development.

Another important fact revealed by the thin section is the evidence of minor recrystallisation along the transverse boundaries of some of the larger kyanite grains. These larger kyanite grains are invariably high energy varieties as indicated by their folded shape and/or quite strongly undulose extinction

#### 5. Slide No. 5B~a'87 (Staurolite-bearing Joshimath Gneiss)

The rock comes from south-west of Jogidhara Falls from a road-cut exposure (Main Road), about 700m above the Vaikrita Thrust. Main minerals are: quartz, plagioclase, negligible orthoclase, biotite and muscovite in equal proportion, garnet, staurolite, clinozoisite, epidote, apatite, zircon, sphene, and black and red opaques.

A 1mm long and 0.1mm wide prismatic crystal of staurolite in the upper right of the slide. The crystal shows transverse cracks and partial knee-bend breaking along one of these cracks. The two partially broken segments show a difference in extinction across this crack the quartz and/or plagioclase (a small opaque grain is also found in contact). The quartz and plagioclase grains show undulose extinction. The cracking, partial breaking and related undulose extinction of the staurolite crystal implies that staurolite growth was evidently not post-tectonic. The staurolite grain follows the orientation of main foliation in association with micas many of which are conspicuously bent. Because no distinguishable fold hinge is preserved in the thin section, it is difficult to establish the time of formation of staurolite in terms of folding. However, it can be argued that staurolite also grew prior to the F2 folding because (a) the staurolite was an integral part of the progressive metamorphic sequence to which kyanite and sillimanite also belong, (b) the rock-exposure is located well within the domain under the strong influence of F2 folding, and (c) there is crude similarity in the strain-features of this staurolite and confirmed pre-F2 kyanites in other sections.

#### 6. Slide No. 2C'87 (Staurolite-bearing Joshimath Gneiss)

The rock comes from about 1km farther north (horizontal distance) from the earlier one and contains quartz, plagioclase, biotite, muscovite, garnet, clinozoisite, staurolite, apatite, sphene and black opaques. Some biotite and garnet grains are partly chloritised. However, there is some doubt as to whether or not the supposed staurolite grains in this thin section are, in fact, a thulitic variety of zoisite. Confirmation through electron microprobing could not be done due to the smallness of the grains and very low quantity of the mineral in the rock. Shows two small prisms of the supposed staurolite (identifying characters - high/very high R.I., faint pale pinkish brown to pale yellow pleochroism, straight extinction, low to moderate birefringence, biaxial positive). The relatively more stout of the two is 0.38mm long and 0.09mm wide, and the other grain at the south-western part is 0.26mm long and 0.05mm wide. Both the grains have transverse cracks and are aligned parallel to the main foliation being closely associated with biotite. The former i.e. the stouter grain, shows distinct undulose extinction. There is no direct way of inferring the time-relationship between this staurolite-growth and any fold episode, because the thin section does not show any distinct fold-hinge. But arguing in the same line as for slide-5B~a'87 and because here also the staurolite grains are not strain-free (the other associated minerals like micas, quartz and plagioclase are strained too), it is likely that this staurolite also grew before F2 episode of folding, as a part of the progressive metamorphic sequence that involved kyanite and sillimanite as well.

It is proposed now to describe the petrography of two thin sections (22/4/88A and NR 25/26L(2)) from the Joshimath Gneiss which do not contain highgrade metamorphic minerals. This description is considered relevant because the first section highlights some special aspects of the F2 folding and its relation to the formation of some interesting minerals, and the second one shows that some horizons in the Joshimath Gneiss do not contain any typical high grade metamorphic minerals (except garnet) and do not show any conspicuous sign of straining due to folding.

#### 7. Slide No. 22/4/88A (An F2 profile section from a semipelitic band in Joshimath Gneiss)

The rock shows very interesting and varied mineralogy in different bands. However, broadly speaking, there are thin quartz-

rich bands alternating with thick mica-rich bands. Except for some relatively smaller-grained micas and occasional calcite, the quartz-rich bands contain virtually only quartz; whereas the mica-rich bands contain in varying proportions in different bands all of the minerals mentioned below: quartz, muscovite, biotite, clinozoisite, skeletal scapolite (probably meionitic), tourmaline, calcite, sphene, reddish black opaques (probably hematitic iron oxides) and apatite. No feldspar or highgrade metamorphic aluminosilicate or garnet is found in the thin section.

Scapolite is identified optically by low to moderate R.I., nonpleochroic white colour (quartz-colour), moderately high birefringence (2nd order yellow int. colour - hence calcic variety i.e. meionitic rather than sodic, marialitic), straight extinction with respect to a prominent set of elongation|| cleavage (another set of cleavage noticed in some grains to be at about  $65^\circ$  with the former set), no twinning, -ve elongation and uniaxial -ve character. The scapolite is in skeletal form; simultaneous extinction distinctly marks out the outline of individual scapolite grains. No undulose extinction is noticed. This observation coupled with the fact that the F2 hinges defined by arrangement of micas, quartz etc. pass undeflected right through many of the scapolite grains proves that scapolite developed through static overgrowth at a post-F2 stage. Matrix-fabric continues unaffected through the scapolite grains. Schiller-like materials (possibly weathering stains) are common along the boundaries and cleavage/fractures in the scapolite. The overall outline of these spongy or skeletal scapolite grains is somewhat elliptical, broadly parallel to the F2 axial trace.

Quartz grains in this section also generally have elliptical outline with their long axes parallelling the F2 axial trace. The most common grainsize of quartz is 0.28mm/0.20mm i.e. two semiaxes (aspect ratio in the thin section is 1.4 : 1). However, where developed in between two mica grains, the shape is often rectangular and the length : breadth ratio is slightly greater than 1.4 : 1. Quartz grains show fairly undulose extinction and lack well defined crystal outlines. Typical triple junctions are almost absent. The texture is evidently not an equilibrated or annealed one. Axial traces of F2 folds (crenulations) are clearly defined across the quartz-rich bands by trails of relatively fine-grained micas running along quartz grain contact zones

Muscovite is more common and larger than biotite in the rock. However, a notable point is that neither the muscovites nor the biotites show a widely varying grainsize and that they occur mostly as short, straight and stout grains, generally free from bending, kinking and undulose extinction (muscovite-size 0.33mm/0.12mm, biotite-size 0.20mm/0.07mm). The F2 hinges are defined, in most cases, by mutually oblique arrangement of several grains of mica rather than by bending of individual large mica grains. These fold-defining mica grains are definitely pre-F2. Another less common group of micas are still smaller in size (0.15mm/0.02mm) and occur mainly along the axial traces of the F2 crenulations, particularly in the quartz-rich bands. Apparently the fold axial traces in the mica-rich zones are marked more clearly by reoriented early micas than by growth of new micas. The relatively few larger elongate grains of mica which occur parallel to the F2 axial trace within the mica-rich zones are almost certainly rotated pre-F2 grains (Fig.....photomicro. at 5X, XPL showing large micas along axial traces as well as along limbs).

Next to quartz and micas, clinozoisite is the most common mineral in the section. Commonly clinozoisite occurs in clusters following the fold traces defined by micas and, therefore, it grew before F2 folding. The clinozoisite grains are prismatic or elliptical; modal grainsize is around 0.24mm/0.13mm. Many grains show undulose extinction, but obviously less conspicuously than quartz.

An important point about the occurrence of tourmaline is that majority of the grains show their circular i.e. basal sections in the slide and therefore do not show pleochroic colour change and appear black between cross polars. The thin section is cut perpendicular to the F2 axis which means that the tourmaline shows a preferred orientation parallel to the F2 axis. So, probably tourmaline is also pre-F2; however, in view of such a strong preferred orientation, some recrystallisation at syn-F2 time is not ruled out. No feldspar is found in the thin section, neither is any aluminosilicate nor garnet.

Strictly speaking, it was not possible to decipher the time of growth of scapolite in terms of the F3 or F4 folding, because no such fold hinge is found nearby. However, that the overall outline of

scapolite grains show somewhat elliptical shape parallelling the F2 axial trace indicates the presence of post-F2 flattening or shearing strain. Therefore, in all probability, scapolite had developed prior to the end of main stretching event in the area and hence the likelihood is that its growth was pre- or syn-F3 folding.

#### 8. Slide No. NR 25/26L(2) (From a metabasic band in Joshimath Gneiss)

The hand specimen came from the lower part of an intraformational slide zone in Joshimath Gneiss more than 1km above the Vaikrita Thrust. The major minerals are: amphibole (common hornblende), clinopyroxene (augite), garnet, quartz, apatite, clinozoisite, specks of biotite (probably alteration product from hornblende) and negligible opaques. No feldspar or any highgrade metamorphic aluminosilicate occurs.

The common hornblende is optically identified by virtue of its high R.I., pleochroism in shades of pale yellow, bluish green and greenish brown, two sets of cleavage at  $\sim 57^\circ$ , moderately high birefringence, inclined extinction ( $z \wedge c = 18^\circ$ ). Confirmation is provided by the pleochroic scheme  $b > g > a$ , positive elongation, high 2V and -ve optic sign. Some of the hornblende grains are twinned.

The clinopyroxene is identified to be augite on the ground of high R.I., lack of colour (nonpleochroic), two sets of cleavage almost at right angles, moderately high birefringence (2nd order blue-green colour), inclined extinction ( $z \wedge c = 45^\circ$ , that is, greater than  $38^\circ$  which is the upper limit for diopside), biaxial +ve with moderately high 2V (definitely  $\gg 30^\circ$ , so not a pigeonite which has a  $2V < 30^\circ$ ). The pyroxenes show well developed schiller structures and some of the grains show incipient alteration to amphibole. Also possibly there are some oriented inclusions of amphiboles within the pyroxenes; these inclusions show maximum illumination between cross polars when the host grains are in extinction or near extinction. The large, discrete amphibole grains, on the other hand, are almost completely devoid of inclusions except occasional ones of quartz.

There is evidence of incipient biotitisation in some amphibole grains. In fact, the few conspicuously identifiable biotite grains in the section are mostly intimately associated with amphiboles and are presumed to be an alteration product from the latter.

It is rather intriguing why there is almost no feldspar in the rock. Most of the white, low R.I. minerals in the section are quartz.

The most interesting point, as mentioned before, is that majority of the mineral grains in this thin section do not appear strongly strained and no distinct influence of any fold episode is apparent, yet the minerals do not have well developed crystal outlines. Most of the grain boundaries are quite irregular. There is no conventional highgrade metamorphic mineral in it, except garnet. It could not be properly resolved when the minerals in the rock grew in relation to the fold episodes.

#### 9. Slide No. NR 58/59 (Garnetiferous Munsiri schist with staurolite included in garnet)

As mentioned before, no sillimanite or kyanite has been found in any of the Munsiri rocks that have been studied. But staurolite has been found in the upper part of the Munsiri Formation. The present rock comes from about 500m below the Vaikrita Thrust. The main minerals are: quartz, plagioclase, possibly some orthoclase, biotite, chlorite (retrogression product), muscovite, garnet, staurolite (only as an included phase within garnet), black opaques, tourmaline, apatite and clinozoisite. Almost all these minerals are also present as small included grains in garnet porphyroblasts. The major difference between staurolite and other included mineral phases is that the latter grains are comparatively much smaller than the staurolite inclusions which are even larger than many of the matrix mineral grains. But staurolite is found in one garnet porphyroblast only. As Fig.....(photomicro. at 5X, PPL from the western ring) shows, an area of diversely oriented staurolite crystals is engulfed by garnet. Broadly, however, there are two modal directions, mutually at  $85^\circ$ - $90^\circ$ , in which the included staurolite grains are oriented; these two directions are at about  $45^\circ$  to the present external foliation. Strictly speaking, the staurolite crystals within the garnet are of various sizes ranging from 0.2mm to 1.25mm in length and 0.02mm to 0.25mm in width and mostly prismatic in shape. No staurolite has been found outside this garnet porphyroblast in the thin section. The garnet

porphyroblast in question is elongate, rather elliptical in outline, with semi-axes 1.64mm and 1.1mm long respectively.

The following features in this thin section are very important:

- (i) There are spongy garnet porphyroblasts as well as fresh homogeneous garnet porphyroblasts in the thin section. The staurolite crystals are found in only one of these spongy garnets. Opaque inclusion trails are more well defined in the homogeneous porphyroblasts.
- (ii) Opaque minerals occur in definite trails (wavy) in the thin section broadly defining the main foliation. These trails pass undeflected through the homogeneous garnet porphyroblasts.
- (iii) There is possible evidence of a fractured and pulled apart garnet in the thin section.
- (iv) There are late fine fractures filled with biotite that are clearly recognisable within the garnets. These fractures are also present in the matrix and run nearly at right angles to the main foliation.
- (v) chlorite as retrograde product is very common in the thin section.

Now the obvious questions are: What is the sequence of mineral formation in the thin section? Are there multiple generations of garnets here? When did the staurolite form in terms of fold episodes? Why are the staurolites confined within a single garnet porphyroblast in a particular area in the thin section?

The elliptical shape of the garnet grain enclosing the staurolite crystals implies that the garnet has been affected most probably at the time of the main phase of shearing or stretching (corresponding to the MCT-emplacement). This confirms that there was a highgrade metamorphic event corresponding to the formation of staurolites well before the MCT emplacement (cf. kyanite development before F2 folding which, in turn, is earlier than the main stretching event). From the points mentioned above, it seems very likely that there are possibly two generations of garnets in the thin section - the spongy ones are earlier (including the staurolite-bearing one) and the homogeneous ones bearing the opaque trails are later. The opaque trails defining the present main foliation in the rock is indented or pushed aside by the pointed edge of a spongy garnet. This implies that the spongy garnet was already in the rock when the present external foliation developed. In contrast the opaque trails pass undeflected through the homogeneous garnets. In another instance, homogeneous garnet appears to have overgrown a spongy garnet; the opaque trails are less conspicuous in the spongy inner part, but continue undeflected through the homogeneous outer part. This suggests that the present main foliation developed through a phase of strong foliation-transposition postdating the formation of spongy garnets; the homogeneous garnet growth took place more or less concomitantly with this foliation transposition.

It is not possible to ascertain from this thin section whether staurolite is pre-F2. But the contrary evidence is not present either.

In view of this, the most plausible story that can be read from the present thin section is as follows:

- III. [Later] Main stretching, foliation transposition and homogeneous garnet formation, more or less, simultaneously
- II. Development of spongy garnets broadly synchronously with an important episode of folding (most possibly, F2)
- I. [Earlier] Formation of staurolite as a part of the early Barrovian metamorphic sequence.

One final point emerges as a corollary, that is, about the possibility of having a generation of garnet which is strictly an integral part of the early Barrovian metamorphic sequence referred to above and is, therefore, related in time with the staurolite, kyanite etc. Effectively this suggests that we must not rule out the possibility of having altogether at least three generations of garnet in the MCT-sheet. Further discussion on this point will be given in Subsection 5.4 dealing with garnet-textures.

**10. Slide No. 7H'87 (Garnetiferous Munsiri schist with stretched staurolites)**

The rock comes from within 400m below the Vaikrita Thrust from an exposure N of Jharkula village on the Main Road. The main mineral constituents are: quartz, both plagioclase and orthoclase (but in very small quantity compared to quartz), muscovite, biotite, chlorite (retrogression product), garnet, staurolite, black opaques in considerable quantity, tourmaline, apatite and clinozoisite. Staurolite is present as a matrix phase, not as inclusions within the garnet porphyroblasts (contrast section NR 58/59 described above).

The staurolite grains in the thin section are quite small and infrequent. The original dimensions cannot be determined, because the original outlines of the grains are not preserved. Most grains have been stretched and fragmented (pulled apart). A number of elongate staurolite fragments occur in a series or trail following the main foliation. Almost undoubtedly majority of these fragments were originally parts of the same crystal that has been stretched and pulled apart. Thus staurolite must have formed before the main stretching event. Direct evidence is lacking that could suggest the time-relation of staurolites with F2-folding. However, there are important indirect evidences bearing on this time relation. A geometrically conformable relation between the quartz-mica-opaque inclusion trails (fabric internal) defining a fold hinge within a garnet porphyroblast and the staurolite-quartz-mica-opaque association defining the matrix schistosity (fabric external) outside the garnet porphyroblast. If the fold hinge within the garnet porphyroblast represents an F2-hinge, then we could certainly infer that staurolite forms an important part of the foliation that defines the F2 fold. The garnet porphyroblast grew either synchronously with or later than the folding. The staurolite grains maintain a conformable relation with some smallscale folds (microfolds) defined by micas on the main foliation. These features coupled with the fact that none of the staurolite grains are found to have perfect euhedral shapes or well defined crystal outlines (rather they all have embayed or zigzag outlines and are internally strained as indicated by their strong undulose extinction) probably mean that staurolite grew prior to a major fold episode, possibly F2. The strongly deformed appearance of two staurolite grains. The folds mentioned above clearly do not postdate the main stretching event; the microfold in the matrix defined by micas could at most be broadly synchronous with the stretching. Most of these folds are probably developed at an early stage of a progressive shear the culmination of which is represented by intense stretching. Thus the essential story appears to be the same, that is, a highgrade metamorphic event, represented here by the development of staurolite, is followed by a major fold event which is followed, in turn, by the main stretching event.

**11. Slide No. F.W.Kya '89 (Kyanite-bearing quartzite from the Berinag-Mandhali)**

The rock comes from the Berinag-Mandhali footwall block from about 1.25km below the Gansser's MCT i.e. the lower limit of the MCT-zone. This is the only exposure in the study area where a highgrade mineral like kyanite is found in the footwall block. As mentioned before, garnet is noticed in the footwall block just below the MCT in a thin phyllite horizon sandwiched between quartzites. The thin section is cut subparallel to the main foliation. The main minerals are: quartz, kyanite, white mica (phengite/sericite), tourmaline, black opaques and zircon. Kyanite occurs in great numbers in the thin section.

Corroborating the field observation the thin section shows the presence of a strong linear fabric in the rock which is defined by fine trails of elongate tourmaline needles or threads, alignment of fine micas and a fairly strong preferred orientation of kyanite grains with their lengths  $\parallel$  to the lineation. Quartz grains also show a crude elongation parallel to the lineation. However, there are a number of kyanite grains which are aligned oblique or at a high angle to the lineation and many of these kyanite grains are folded. The fact that some of the linearly oriented kyanite grains are pulled apart following the direction of lineation suggests that it is a stretching lineation.

The quartz grains are mostly strain-free with well developed crystal outline. Generally they do not show any undulose extinction. Grain-contact triple junctions are very common among them. These features clearly indicate that the quartz fabric in the rock is essentially a recrystallised fabric. Most of the quartz grains have slightly elongate hexagonal shape, their long dimensions coinciding

with the direction of stretching lineation. The average grain size of quartz varies from 0.20mm/0.14mm (i.e. two semi-axes of circumscribing ellipses) to 0.29mm/0.22mm.

Kyanite is evidently porphyroblastic in origin forming the largest grains in the thin section, their dimensions generally being >2mm in length and over 1mm across. In contrast to quartz the kyanite grains do not look at all happy; almost all of them show strong undulose extinction and strain shadow patterns. Some of the kyanite grains are conspicuously folded or kinked. As mentioned above, there is a fairly strong preferred orientation of the kyanite grains (particularly among those which are not folded) with their lengths parallel to the stretching lineation in the rock. That the stretching has affected kyanites, as reflected by tensile fracturing (pulling apart) in the direction of stretching lineation of several kyanite grains, clearly indicates that kyanite developed prior to this stretching event. The conspicuous folding of a number of kyanite grains, on the other hand, implies that kyanite grew before a folding event as well. So, the obvious question is, what is the time-relation between the stretching event and the kyanite-folding? We'll seek an answer to this question shortly afterwards.

The mode of occurrence of tourmaline in the thin section is very interesting. Most tourmalines are in the form of fine long needles or threads bundled together and lying in trails which broadly define the strong linear fabric present in the rock. These trails are found to continue through the matrix of the rock right across many of the kyanite porphyroblasts. Precisely speaking, in most cases, particularly in the matrix of the rock, what appears now as continuous trails of tourmaline are essentially stacks of individual fine tourmaline threads/needles arranged in a close-spaced an echelon or imbricate fashion. It is interesting to note that the individual needles/threads maintain a strict parallelism with the stretching direction, while the collective stack or trail occasionally deviates away from it. These tourmaline threads and the trails might be misconceived as of chlorites. On close observation using high power objective, however, their characteristic pleochroism - from colourless/pale yellowish blue/pale yellow to greenish blue/blue/brownish yellow, deeper colour appearing when the grain lengths are perpendicular to the polariser's vibration direction, pleochroic scheme  $w > e$  and negative elongation make it certain that these are not chlorite, but tourmaline. The moderately high R.I. and birefringence, and uniaxial negative character are clearly recognisable from the thicker among these acicular tourmaline grains. There are discrete tabular tourmaline grains as well in the rock. Both tabular and thread-like tourmaline are also present as inclusions within kyanite porphyroblasts. As indicated above, the thread-like variety in most cases maintains their arrangement in trails even within the kyanites.

The association of tourmaline trails with kyanite is indeed very interesting. There is an example where a straight kyanite grain aligned subperpendicular to the stretching lineation shows folded tourmaline trails within it; the inclusion trails are confined within the kyanite, i.e. they do not continue into the matrix from the kyanite porphyroblast. This observation is very significant, because it implies that kyanite overgrew a folded fabric. There was an episode of folding predating the kyanite development. Essentially, therefore, the thin section preserves the evidence of at least two episodes of folding, one prior to and the other following the kyanite development. For comparison, see Subsection 3.1.2.2 where the fold episodes recognised in the Berinag-Mandhali mainly from mesoscopic field observations are tabulated. However, the kyanite porphyroblasts with folded inclusion trails in them are not very common in the section. Most of the tourmaline trails are closely parallel to the stretching direction both inside the kyanite porphyroblasts and outside them in the matrix of the rock. In fact, as already indicated, the lineation in the slide is defined most clearly by these fine streaks or trails of tourmaline and white micas which pass through undeflected across many of the kyanite porphyroblasts. Thus arises the possibility that we are probably dealing with two generations of tourmaline trails -- one, later; evidently related to the stretching event postdating the kyanites; the other, earlier, that was present even prior to the kyanite development and representing a kind of layering. So, the important questions are:

- (a) When and why did the tourmaline trail/s develop as microstructures?
- (b) When and how did the tourmaline develop as mineral and what is its time relation with kyanite? and finally the general, but a very vital question,
- (c) What is the sequence of mineral formation and tectonic events revealed by the thin section?

#### Time relation between kyanite-folding and stretching lineation

As indicated above, both the stretching event and the folding event postdated the formation of kyanite. The fold episode here refers to that responsible for the folding or kinking of some kyanite porphyroblasts mentioned earlier. It is very important to find out the time relation between this folding event and the stretching event. There are three possible hypotheses:

- (i) Kyanite-folding was later than the stretching lineation
  - (ii) Kyanite-folding was synchronous with lineation development
  - (iii) Kyanite-folding was earlier than lineation. So, lineation development may have played a modifying role upon the orientation of the kyanite grains after folding.
- We need to test which of these hypotheses is correct.

As discussed in text (subsection-5.3.3), folding of kyanite was earlier than the development of stretching lineation, and the deformation that led to the development of the lineation might simply have played a modifying role on the folds. Thus, the succession of events is,

- iii) Development of stretching lineation
- ii) Folding or kinking of kyanites
- i) Formation of kyanite porphyroblasts.

#### Timing of tourmaline trails as microstructures

The two most crucial evidences that have direct bearing on the time of development of the tourmaline trails *sensu lato* are:

(a) Most of the tourmaline trails are parallel to the stretching lineation and pass through the matrix cutting across majority of the kyanite porphyroblasts, folded as well as the non-folded ones. Evidently, the kyanite porphyroblasts cannot be said to have overgrown these lineation-parallel tourmaline trails.

(b) There are instances where the lineation-parallel trails are not found to traverse through kyanite porphyroblasts. These porphyroblasts are themselves not folded, but interestingly contain folded tourmaline-trails (inclusions) in them which means that the kyanite porphyroblasts must have overgrown these trails after their folding. Thus almost certainly there are two generations of tourmaline trails in the rock -- the lineation-parallel ones are the later and these are the dominant ones; the folded tourmaline-trails preserved in only a few kyanite porphyroblasts represent the earlier ones. There is hardly any doubt that the rock shows only one generation of kyanite porphyroblasts. Also the fact that the folded trails are only seen in the straight (nonfolded) kyanite porphyroblasts rather than in the folded ones leaves no scope for assuming that the host kyanites could represent an earlier group of porphyroblasts.

Obviously, therefore, the earlier generation of tourmaline trails developed before the formation of kyanite porphyroblasts whereas the later generation of trails developed at the time of the stretching event postdating the kyanites. Furthermore there was an episode of smallscale tight to isoclinal folding (evidenced by the folded inclusion trails) intermediate in time between the formation of early tourmaline trails and the kyanite porphyroblastesis.

So far as the mode of origin of the tourmaline trails is concerned, the lineation-parallel later trails are obviously related to the stretching event. Here again evidences suggest that possibly there are contribution from two different ways:

- (i) Recrystallisation of tourmaline following the direction of lineation as a result of shearing
- (ii) Mechanical breaking down (brittle to semibrittle deformation) of early tourmaline grains and alignment of the fine broken fragments into grooves/furrows/fractures resulting due to shearing/stretching.

The evidence for the first contributory mechanism is provided by the very regular growth of fine tourmaline needles parallel to the stretching lineation within the pull-apart gaps of kyanite porphyroblasts. Clearly recrystallisation of quartz has also taken place along with tourmaline in those gaps.

The second mechanism is supported by the fact that in several instances we see innumerable small fragments, flakes or particles of tourmaline closely gathered around and partially covering larger tabular grains of tourmaline with broken boundaries; these larger tourmaline grains act as originating points of some lineation-parallel tourmaline trails. The fact that kyanite porphyroblasts have been fractured (pulled apart) due to stretching presents strong evidence for brittle/semibrittle deformation associated with the stretching/shearing event. Intuitively, in the lower temperature range tourmaline is expected to deform even in more brittle way than kyanite.

The origin of the early tourmaline trails (i.e. the folded ones) could not be recognised properly. However, it is speculated that they probably represent the traces of the original pre-metamorphic fabric - either fine scale sedimentary layers rich in detrital tourmalines or impregnations of hydrothermal boron-rich fluids following the original layer orientations.

So, the sequence of events including the formation of tourmaline trails is as follows -

- v) Formation of later tourmaline trails during the stretching event
- iv) An episode of folding (represented by bending of kyanites)
- iii) Kyanite porphyroblastesis overgrowing folded tourmaline trails
- ii) Tight to isoclinal folding affecting the early trails
- i) Formation of early tourmaline trails (possibly representing the original pre-metamorphic fabric in the rock).

Relative time of mineral development: tourmaline vis-a-vis kyanite

The significance of the intimate association of kyanites with the tourmaline trails initially seemed extremely difficult to unravel. Did kyanite and tourmaline grow together? Or, was tourmaline earlier than kyanite? Or, later? These simple questions had originally appeared very difficult to answer.

Now, however, in the light of what have been discussed already, it is quite easy to find out the relative time of growth of tourmaline and kyanite. As stressed before, kyanite seems to have developed in the rock in one single episode of porphyroblastesis. But growth of tourmaline in the rock involved most possibly two processes - (a) detrital deposition of tourmaline grains, and (b) introduction (injection/impregnation) of boron-rich hydrothermal fluid into the rock possibly at multiple stages. In this connection, it is worth recollecting that tabular tourmaline grains are also present in the rock both in the matrix as well as in the form of inclusions within some kyanite porphyroblasts which means that there was a stage of tourmaline formation (corresponding to the discrete tabular grains) before kyanite porphyroblastesis. Does this mean that a boron-rich fluid flowed pervasively through the rock before the kyanites formed? It is very likely that formation of kyanite porphyroblasts amidst a lowgrade surrounding, as it is in the kyanite quartzite and its immediate environs within the Berinag-Mandhali formations, can be attributed to concentration of shear stress in a particular horizon in the footwall block consequent upon the emplacement of the MCT sheet above (also see below, and description from slide - 32A11L). So, it may not be at all unlikely that pervasive fluid flow along favourable rock horizons within the footwall led to the formation of tourmaline, either concomitant with or immediately following the MCT-emplacement. Implicit in this discussion is that the stretching event recorded in the kyanite quartzite is supposedly later than the main stage of MCT-emplacement. There are evidences of quartz vein intrusions in the kyanite quartzite related in time with this later stretching event (see discussion on slide-32A11L); so it is possible that some contribution towards tourmaline formation came from this later hydrothermal fluid injection as well.

The tourmaline- and kyanite-forming episodes can be shown thus in a sequence -

- iv) Possible formation of tourmaline associated with early-syn-stretching quartz vein intrusions
- iii) Kyanite porphyroblastesis
- ii) Possible formation of tourmaline associated with fluid flow concomitant or late synchronously with MCT-emplacement
- i) Detrital deposition of tourmaline along with quartz, mica etc. leading to the formation of original sedimentary fabric (and/or boron-rich hydrothermal fluid impregnations into the original premetamorphic fabric leading to tourmaline formation)

It must be mentioned that quartz in the rock suffered thorough recrystallisation broadly synchronously with the stretching event (see discussion on slide - 32AIII); the main early phase of mica neomineralisation that corresponded to a mild early episode of metamorphism took place even before the first episode of folding recognised in the rock. See in Chapter-3 where it is indicated that the earliest folding recognised in the Berinag-Mandhali folds a metamorphic foliation (cf. Slide - 30E'88).

Sequence of tectonometamorphic events, as read from the thin section

By now, it is clear that the following sequence of events can be reconstructed from the evidences preserved in the thin section (F.W.Kya'89):

- VII. Stretching/shearing event (accompanied by recrystallisation of quartz, tourmaline and possibly some white micas; intrusion of concordant quartz veins during early part of the event)
- VI. A strong episode of folding (corresponding to the bent kyanite grains)
- V. Kyanite porphyroblastesis
- IV. Possible pervasive flow of boron-rich fluids along favourable horizons (related to discrete tourmaline grains and inclusions in kyanite)
- III. An episode of tight to isoclinal folding (corresponding to the curved tourmaline trails preserved in some kyanite porphyroblasts)
- II. A low grade metamorphism episode
- I. Development of original pre-metamorphic fabric (represented by the early tourmaline trails)

Placing the local tectonometamorphic events in a largescale context

Given the local sequence of events viz. the three episodes of folding (I, II & III) in the Berinag-Mandhali established from mesoscale observations (see Subsections 3.1.2.) and the two generations of stretching lineations found in the MCT-zone, the task is now to find out how all these fit into a coherent sequence. The thin section preserves indications of two episodes of folding of which the earlier one is of tight to isoclinal geometry and of smaller scale than the second one. The geometric and morphological features, and the temporal relations of these two episodes of folding match very well with the episodes I and II respectively that have already been recognised from mesoscopic observations. So, the two episodes preserved in the thin section are equated with the episodes I & II respectively in the footwall fold sequence. As emphasized before, episode-III largely represents post-metamorphic folding and so probably postdates the entire sequence of events recognised from the thin section. Of the two sets of stretching lineations recognised in the MCT-zone, the early major one (NE/NNE-trending) is related directly to the emplacement of the MCT; the other set (N-trending) represents a later stage of reactivation. Strictly speaking, the orientation of the stretching lineation in the kyanite quartzite is intermediate between that of the two sets in the MCT-zone. So, it is not easy to equate the present lineation straight away with one of the two sets in the MCT-zone, yet it is obvious that it must be correlatable with one of them. There are, however, some indirect ways to judge the correctness of a correlation in this case; for example, the preferred orientation of kyanite along the lineation, folding of kyanite grains even before the development of the lineation and the origin of kyanite must be properly pigeon-holed in the regional sequence of events. The correlation will fail if it leads to any problem in putting these events into a proper sequence.

Now, the obvious first choice would be to correlate the kyanite-quartzite lineation with the NE/NNE-trending major lineation of the MCT-zone. The involvement of kyanite in F2-folding within the Joshimath Gneiss and the same fold-event (F2) predating the main stretching event provide a very enticing coincidence to suggest the possibility of such a correlation. The similarity in the sequence of tectonic events involving kyanite indeed appears very compelling. Kyanite in this exposure, as also in the Joshimath Gneiss, has been folded by a major fold-event which is followed by a major stretching event. So, did the kyanites in two units (Joshimath Gneiss & Berinag-Mandhali) develop at the same time? If they did, there are many important ramifications. For instance, our tendency would then be to equate the folding (episode II) that folds the kyanite in Berinag-Mandhali with the F2 folds of the

Joshimath Gneiss and likewise episode-III and episode-I folds with the F3 and F1 folds respectively. Later we shall see whether such correlation among fold episodes can really hold true or not. Before that, let's focus on what the implications would be on the pre-kyanite history of the rocks. There was an episode of tight to isoclinal folding before kyanite formation in the Berinag-Mandhali and this represents the earliest recognisable fold event in that unit. An episode of mild metamorphism is recognised to predate this folding. In the Joshimath Gneiss the order of events appears to be very much the same: before or during early part of kyanite development there was F1 folding, and most probably this F1 folding postdated the formation of a gneissosity which represented a distinct metamorphic episode. Thus we see that a remarkable correlation in the broad sequence of events between the footwall and the hanging wall block emerges, when we equate the kyanite-quartzite stretching with the main i.e. the earlier stretching event in the MCT-zone.

Let us now try to compare the characteristics of the fold episodes and other events whose orders of development appear so remarkably well correlatable from the footwall block to the hanging wall block when we suppose that the kyanite-quartzite lineation is equivalent to the dominant NE/NNE-trending lineation of the MCT-zone. However, the correlation certainly appears too well-matched to be true and the very fact that for our comparison we are dealing with two blocks (areas) which were at least 100 kms away from each other before the emplacement of the MCT makes us strongly suspect the validity of such a one-to-one correlation, including even microscale events. Development of kyanite in the Joshimath Gneiss is clearly related to a distinct episode of progressive high grade Barrovian metamorphism, but that of the footwall kyanite is not. The origin of kyanite in the Berinag-Mandhali formations is certainly due to some other reason. So, evidently, there is no definite reason to believe that kyanite in the two blocks developed at the same time. Well-constrained fission track dating of both kyanites could possibly indicate the difference in their age. As far as the morphology (including items like the character of the foliation being folded, the nature of the axial planar structures etc.) and the orientation of the folds are concerned, there is absolutely no similarity between the F1, F2 & F3 folds in Joshimath Gneiss and/or Munsiri and the episodes - I, II & III folds of the Berinag-Mandhali respectively (for details see Chapter-3). This suggests that episode - I, II & III folds cannot be correlated with the F1, F2 & F3 folds respectively. In fact, I could not find out any definite evidence or criterion that could indicate a possible correlation among these fold episodes. Therefore, the abovenoted remarkable correlation is apparent, not real and the basic supposition that led to such apparent correlation is not correct. Our supposition was that the kyanite-quartzite lineation in Berinag-Mandhali is equivalent to the 'dominant NE/NNE-trending stretching lineation' of the MCT-zone. With this supposition, another obvious difficulty arises when we try to explain what happens to the 'later N-trending stretching lineation' that gradually gains in prominence towards south within the MCT-zone (see also Subsection 3.2.2 in Chapter -3). Therefore, the only other plausible alternative open to us is that the kyanite-quartzite lineation in the Berinag-Mandhali postdates the dominant stretching lineation of the MCT-zone and equates better with the 'later N-trending lineation.'

Now, what have been the effects in the footwall block due to the emplacement of the MCT-sheet (corresponding to the 'dominant NE/NNE-trending lineation')? In other words, which of the event/s recognised from the thin section would correlate with the MCT emplacement? The difficulty to explain the origin of kyanite in the footwall has been already highlighted; now an important possibility is that concentration of shear stress selectively in the quartzite horizon consequent upon the emplacement of the MCT might have helped in the development of these kyanites. Also the origin of discrete tourmalines without any preferred orientation in the rock predating kyanites, but postdating episode-I folding can be explained as due to pervasive flow of boron-rich fluids along favourable rock horizons within the footwall concomitant with or immediately following the MCT-emplacement. The direction of plunge of the episode-I folds is perpendicular to the direction of emplacement of the MCT-sheet and the axial planes of these folds parallel the main planar fabric in the MCT-zone; these facts probably suggest a causal connection between the development of these folds and the MCT-emplacement.

The following table shows the suggested correlation of the events identified from the thin section (F.W.Kya'89) with those for the Berinag-Mandhali footwall block (MCT emplacement time is also shown for comparison).

Sequence of events recognised in the slide		Sequence of events in the Berinag-Mandhali
		Fold Episode-III
Later stretching event in MCT-zone	Stretching/shearing event with qtz & tourmaline recrystallisation	I Stretching event with 4th metamorphic imprint
	Kyanite folding	I Fold Episode-II
	Kyanite porphyroblastesis	I 3rd metamorphism
MCT-emplacment (Major stretching Possible flow of boron-rich fluids event)		
	Folding of tourmaline trails	I Fold Episode-I accompanied by 2nd metamorphism
	Initial lowgrade metamorphism	I 1st metamorphic imprint (mild)
	Development of original pre-metamorphic fabric	

As indicated above, kyanite developed in the Berinag-Mandhali in between fold episodes I & II and most possibly immediately following the emplacement of the MCT. We have already seen that the MCT-emplacment (eqv. to the dominant stretching event) followed F2 fold-episode of Joshimath Gneiss in time, and kyanite-development in the Joshimath Gneiss predated this F2 folding. Therefore, certainly there was considerable time-gap between formation of kyanite in the Joshimath Gneiss and in Berinag-Mandhali. The footwall kyanites are younger than the hanging wall kyanites. Fission track annealing dates from both these kyanites could possibly confirm their age difference.

#### 12. Slide No. 32A||L (Kyanite quartzite of the Berinag-Mandhali footwall block)

The hand specimen comes from the same kyanite-quartzite exposure as mentioned before (see description of slide F.W.Kya'89). The section is cut across foliation, parallel to lineation which means that it is an XZ section. In comparison to thin section- F.W.Kya'89 which is subparallel to the foliation, the present section contains much less proportion of kyanite. This gives a further confirmation that there is preferred development and preferred orientation of kyanite along the main foliation within the kyanite-quartzite exposure.

The thin section shows very simple mineralogy – mostly quartz, muscovite and kyanite with accessory zircon and ?rutile; traces of chlorite are also noticed (probably alteration product from biotite). The section does not contain any tourmaline, unlike the section - F.W. Kya'89.

The general rock texture is schistose. Muscovite shows strong preferred orientation; their 001-faces are mostly parallel to foliation and their grain-lengths parallel to lineation. Thus lineation (i.e. stretching) had a very important role in dictating the present arrangement of muscovites and kyanites. Kyanites show strong undulose extinction, some of the grains are fragmented (torn apart), some show lamellar twinning; also there are some bent or kinked kyanite grains. The effect of stretching is also highlighted by the tearing apart of the rutile grains (brownish black, high relief, semi-opaque minerals in the section) and by their stringed arrangement intimately associated with muscovite. Muscovite has grown within the fractures of some kyanite grains implying that kyanite grew earlier than muscovite. Also, generally speaking, strain-features shown by muscovites are less conspicuous than those in kyanites. There is bimodal distribution of grain-size of muscovites. There are larger elongate muscovite grains and also tiny flakes. Given the strong influence of stretching, it is very likely that most of the smaller flakes are, in fact, broken-up parts from the larger grains. However later recrystallisation of smaller muscovite is also a possibility.

In addition to kyanite, quartz is showing very interesting textures. There is a band of quartz running concordantly all along the eastern margin of the section. The size of quartz grains in this band is comparatively larger than the average matrix quartz grain-size in the rock and the size-distribution is not non-uniform; also the proportion of mica is much less in this band. Probably this was originally a concordant quartz vein that intruded the rock at an early stage of stretching (see below for further discussion). Within this band, there is a well developed shape fabric of the quartz grains; the grains are mostly somewhat elliptical or elongate in outline and are arranged in such a way as to indicate an imbricate pattern oblique (at  $\sim 40^\circ$ ) to the main foliation. This sort of shape fabric is a good indicator of shear sense -- here 'top-to-south' which is consistent with the shear sense established within the MCT-zone. Thus the effect of shear or stretching is present also within the Berinag-Mandhali, well below the Munsiri Formation (i.e. the MCT-zone). As established already, this stretching lineation in the kyanite-quartzite of Berinag-Mandhali equates with the later stretching event of the MCT-zone. The type of shape fabric referred to above is less conspicuous within the matrix quartz, probably because of the presence of more micas in the matrix. In many instances within the matrix, quartz grains are found to be elongate parallel to the foliation; such a pattern is representative of growth under the inhibitory influence of a second phase (here, muscovite) (Hobbs et al, 1976, pp.113-118). Also, using a sensitive tint plate, it is found that there is a difference in the crystallographic preferred orientation of quartz in the vein/band and in the matrix. Another important feature shown by the thin section is the presence of larger porphyroclastic quartz grains which show heavily undulose extinction and deformation lamellae or deformation bands, in contrast to the matrix quartz grains. Also there are excellent evidence of recrystallisation of small strain-free quartz grains out of these larger quartz porphyroclasts. It is very likely that most of the matrix quartz are products of recrystallisation from such larger porphyroclasts which might represent original clastic grains. These porphyroclastic quartz grains, among themselves, have variable crystallographic orientations. Using a sensitive tint plate in any fixed orientation of the thin section, different porphyroclasts are found to show different colours: green, yellow, blue, red, bluish red, reddish yellow etc. The crystallographic preferred orientation (C.P.O.) of these porphyroclasts differs from that of the matrix quartz grains. Using a sensitive tint plate, a good majority of the matrix quartz grains show red colour when the trace of main foliation is kept E-W. Thus essentially there are three different classes of C.P.O. in the thin section -- there is dominance of red in the matrix quartz, dominance of blue-green in the vein/band quartz and different colours in different porphyroclasts (this is seen by using an accessory sensitive tint plate and taking the main foliation perpendicular to the accessory slot so as to avoid the interference by mica-colours). The quite haphazard crystallographic orientation of the porphyroclastic grains is a further pointer to the possibility of their being original clastic grains. While the porphyroclastic quartz grains are highly deformed in appearance, the matrix quartz and vein/band quartz grains do not show strong undulose extinction; the latter groups show fairly annealed texture with many examples of triple junctions.

Confining our observations only to this thin section, we can certainly erect the following sequence of events -

Shearing/stretching (accompanied by quartz recrystallisation and possibly immediately postdating mica neomineralisation; at an early stage of stretching, quartz vein intrusion took place so that the solidified vein materials also ultimately suffered stretching)

Bending or kinking of kyanites

Formation of kyanites

Initial metamorphism (possibly along with folding effect of which is not conspicuous in the section)

Clastic quartz deposition

Clearly there is a close match between the above sequence of events and that established from the slide - F.W.Kya'89. Though the two sections come from two different hand specimens, both the hand specimens were collected from the same exposure of the kyanite-bearing quartzite horizon in the Berinag-Mandhali. A comparison between the two thin sections strengthens the probability that there were multi-stage vein intrusions through the rock - at least an early boron-rich fluid infiltration immediately prior to or during kyanite formation, and a later silica-rich veining occurring in the early stage of the stretching event.

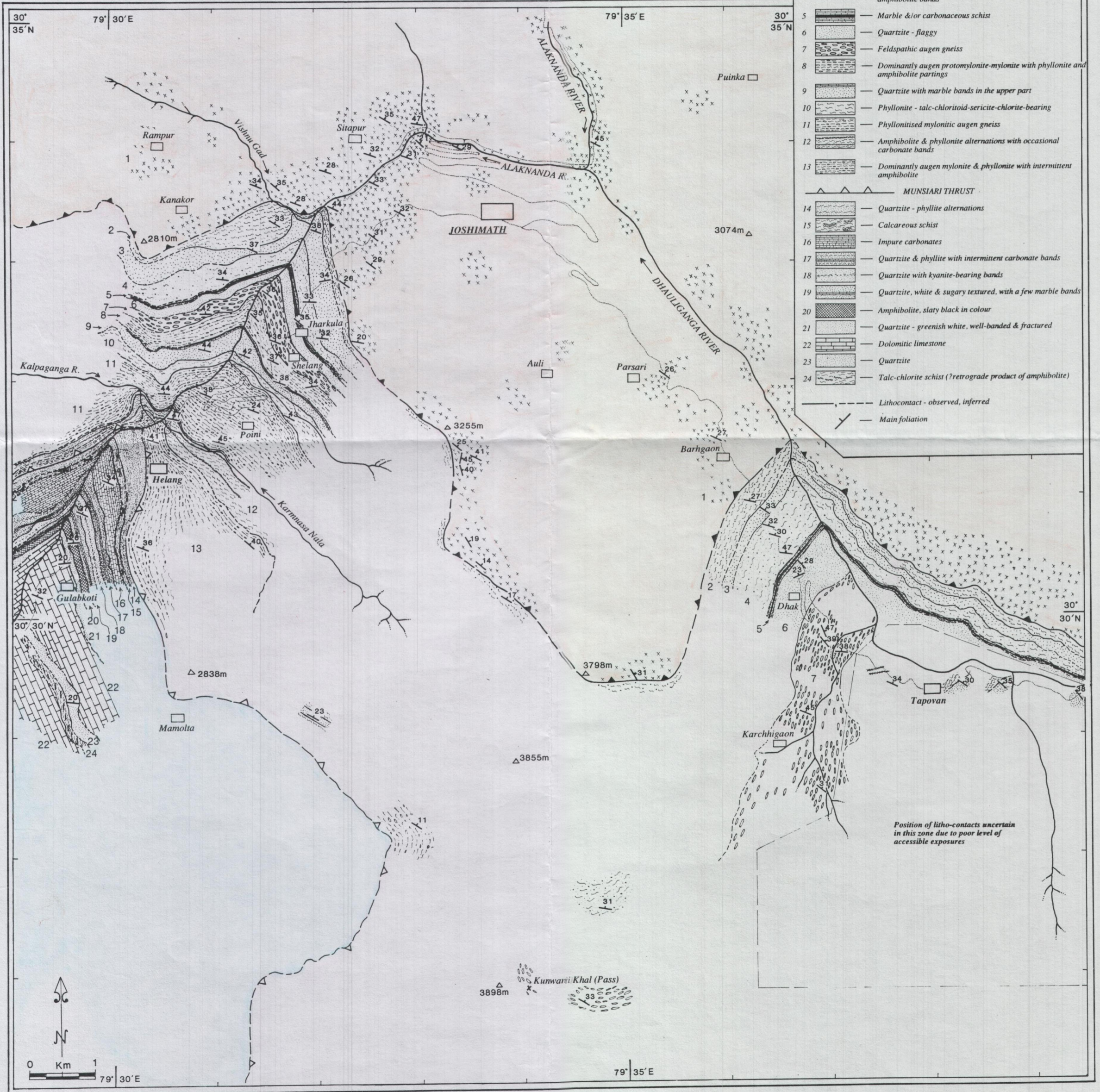
Now let us give a closer look at the grainsize data from the two abovementioned sections and another section (slide - 32A&L) complementary to the section - 32A||L. The thin section - 32A||L is a YZ-section, cut perpendicular to the stretching lineation. As earlier mentioned, slide- 32A||L gives an XZ-section and slide- F.W.Kya'89 gives approximately an XY-section. For measurement of quartz grainsize the two semiaxes of circumscribing ellipses were measured using micrometer oculars; for the porphyroclastic quartz grains the whole elliptical outlines including the recrystallised portions were considered and the measurements were done in plane polarised light; for muscovites the lengths and breadths were measured. Aspect ratios were calculated by dividing the value of longer dimension with that of the shorter dimension. The essential data are given in a table (Table - 5.3) in Chapter - 5.

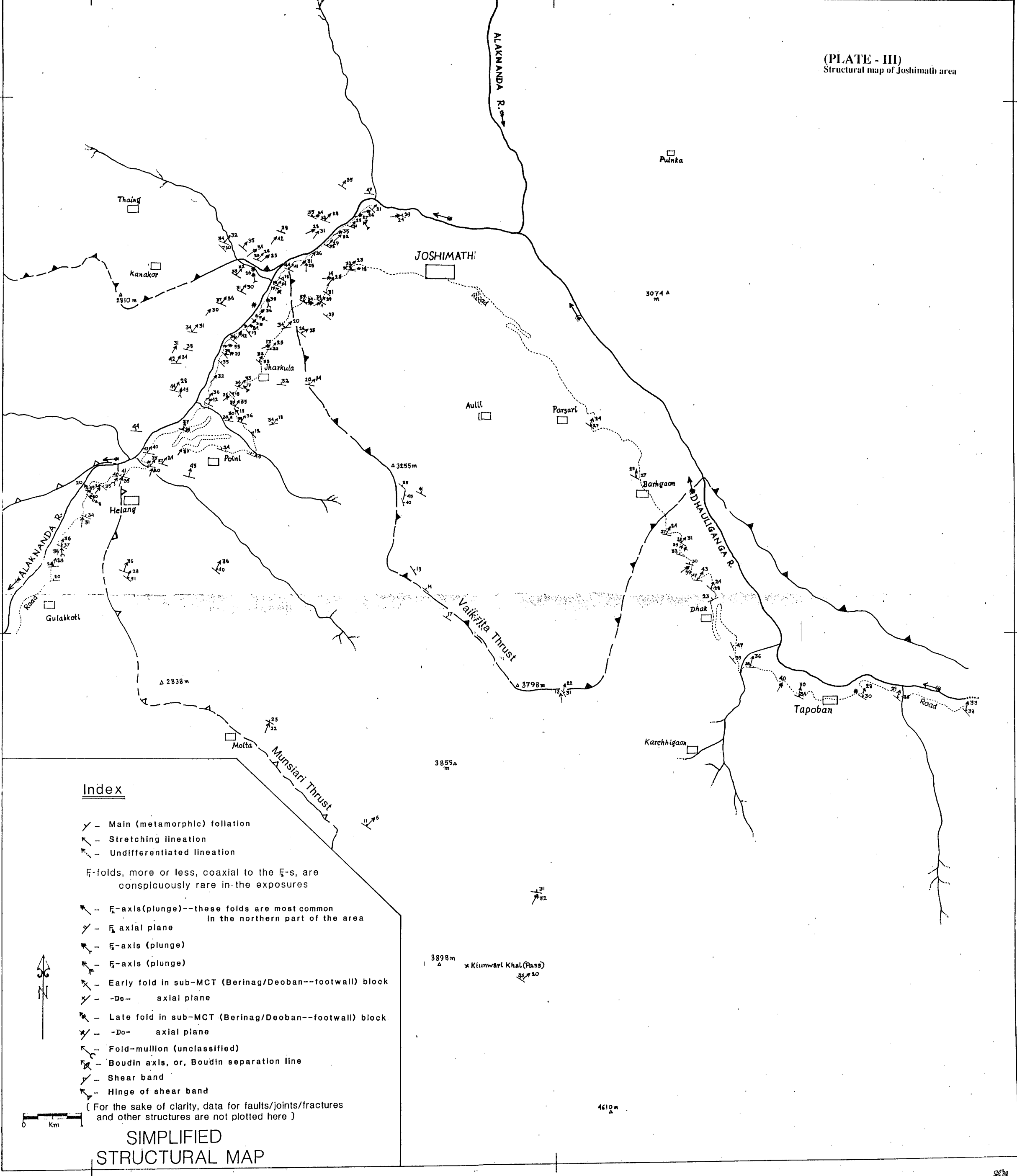
---

Lithological map of Joshimath area

INDEX

- 1 — Joshimath Gneiss Formation
  - 2 — Munsiri Formation
  - 3 — Berinag-Mandhali formations
  - 1 — Joshimath Gneiss, semipelitic
- VAIKRITA THRUST
- 2 — Metapelite - retrograde & garnetiferous with carbonaceous bands in the upper part
  - 3 — Quartzite
  - 4 — Psammopelite, garnetiferous & schistose with occasional amphibolite bands
  - 5 — Marble &/or carbonaceous schist
  - 6 — Quartzite - flaggy
  - 7 — Feldspathic augen gneiss
  - 8 — Dominantly augen protomylonite-mylonite with phyllonite and amphibolite partings
  - 9 — Quartzite with marble bands in the upper part
  - 10 — Phyllonite - talc-chloritoid-sericite-chlorite-bearing
  - 11 — Phyllonitised mylonitic augen gneiss
  - 12 — Amphibolite & phyllonite alternations with occasional carbonate bands
  - 13 — Dominantly augen mylonite & phyllonite with intermittent amphibolite
- MUNSIARI THRUST
- 14 — Quartzite - phyllite alternations
  - 15 — Calcareous schist
  - 16 — Impure carbonates
  - 17 — Quartzite & phyllite with intermittent carbonate bands
  - 18 — Quartzite with kyanite-bearing bands
  - 19 — Quartzite, white & sugary textured, with a few marble bands
  - 20 — Amphibolite, slaty black in colour
  - 21 — Quartzite - greenish white, well-banded & fractured
  - 22 — Dolomitic limestone
  - 23 — Quartzite
  - 24 — Talc-chlorite schist (?retrograde product of amphibolite)
- Lithocontact - observed, inferred
- Main foliation





Index

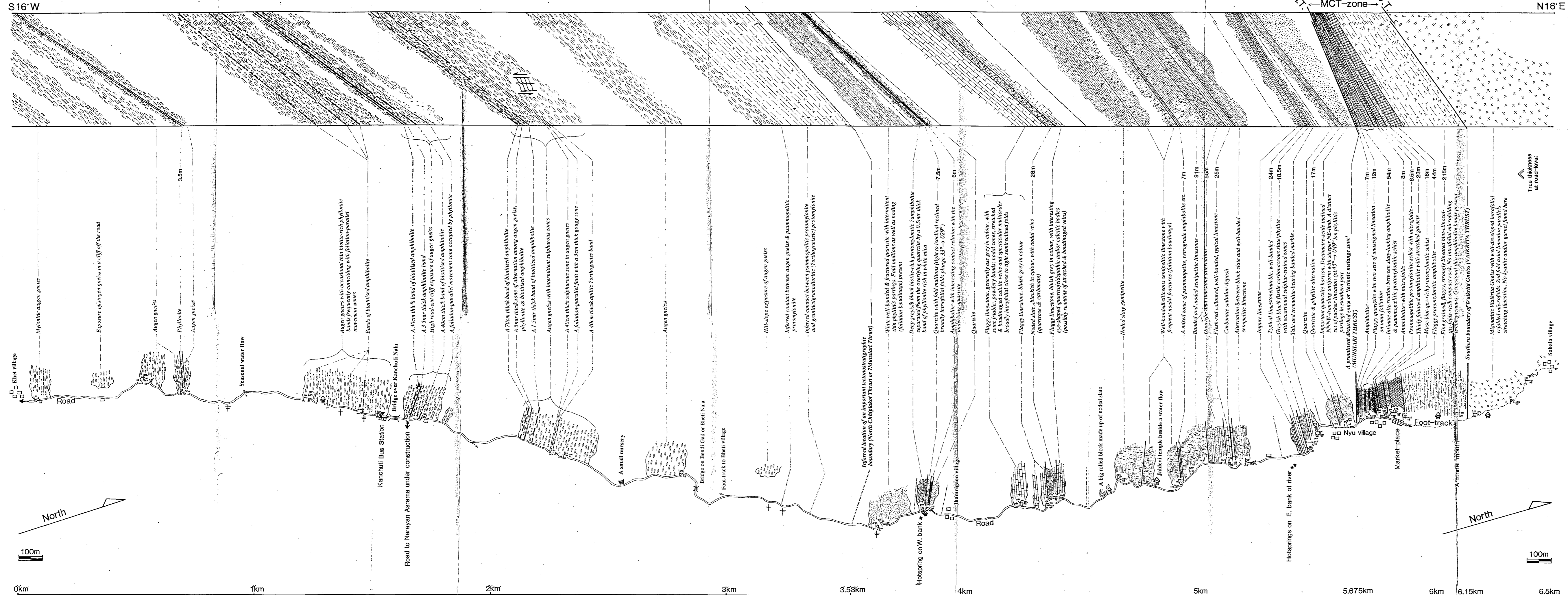
- ↘ - Main (metamorphic) foliation
- ↗ - Stretching lineation
- ↖ - Undifferentiated lineation
- F<sub>1</sub>-folds, more or less, coaxial to the F<sub>1</sub>-s, are conspicuously rare in the exposures
- ↖ - F<sub>1</sub>-axis (plunge) -- these folds are most common in the northern part of the area
- ↘ - F<sub>1</sub> axial plane
- ↖ - F<sub>2</sub>-axis (plunge)
- ↘ - F<sub>2</sub>-axis (plunge)
- ↖ - Early fold in sub-MCT (Berinag/Deoban--footwall) block
- ↘ - -Do- axial plane
- ↖ - Late fold in sub-MCT (Berinag/Deoban--footwall) block
- ↘ - -Do- axial plane
- ↖ - Fold-mullion (unclassified)
- ↘ - Boudin axis, or, Boudin separation line
- ↖ - Shear band
- ↘ - Hinge of shear band

(For the sake of clarity, data for faults/joints/fractures and other structures are not plotted here)



SIMPLIFIED  
STRUCTURAL MAP

Darmaganga Traverse:  
A Composite Road Log/Map  
&  
A Vertical Cross Section



INDEX

- Main stretching lineation
  - Main foliation
  - Main foln. & str. lin. together
  - Unassigned lineation
  - Boudin axis
  - Shear band
  - Important fracture/fault
  - Unassigned fold
    - Axis
    - Ax. plane & axis together
  - Lithocontact - Observed, Inferred
  - Limit of exposure
  - Unornamented areas indicate gaps in exposure.
  - Small temple
  - House/Stall
  - Telegraph Pole
  - Causeway
  - Bridge
- Shear indicator/s and some thin amphibolite &/or phyllonite bands are not to scale.

Alaknanda Traverse :  
A Vertical Cross Section  
and  
A Composite Road Log

(PLATE - IV)

N36° E

Joshimath Gneiss  
(Vaikrita)

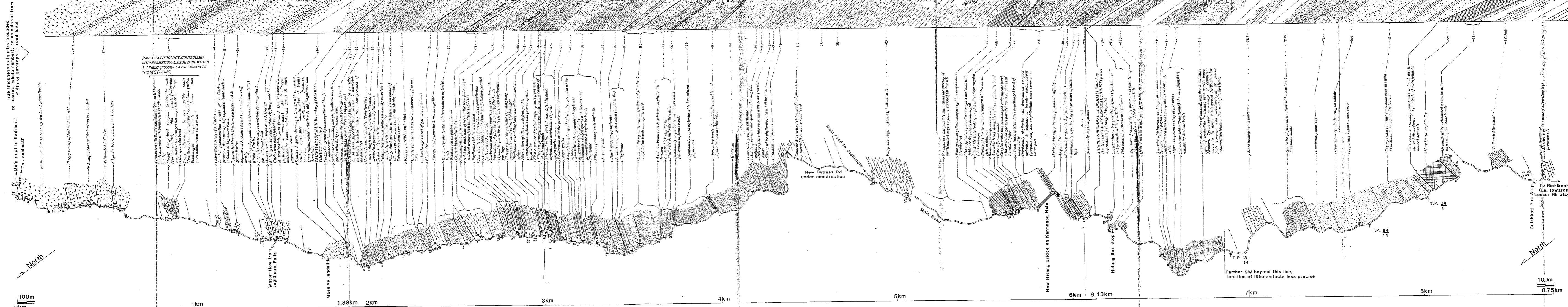
V. T. (MCT I)

MCT-zone  
(Munsiari)

M. T. (MCT II)

Berinag-Mandhali

S36° W



**INDEX**

- Main stretching lineation
- Later stretching lin.
- Main foliation
- Main foln & str. lin. together
- Shear band
- Boudin axis

**Folds**

- (Axes) (Ax. Planes) (Axes & Ax. Pls. together)
- F<sub>1</sub>
- F<sub>2</sub>
- F<sub>3</sub>
- Unassigned

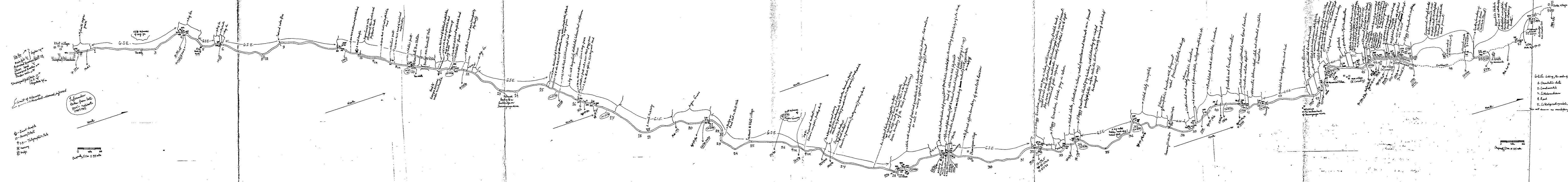
- Lithocontacts— Observed, Inferred
- Limit of exposure
- T.P. Telegraphic pole
- Small temple
- House/Stall

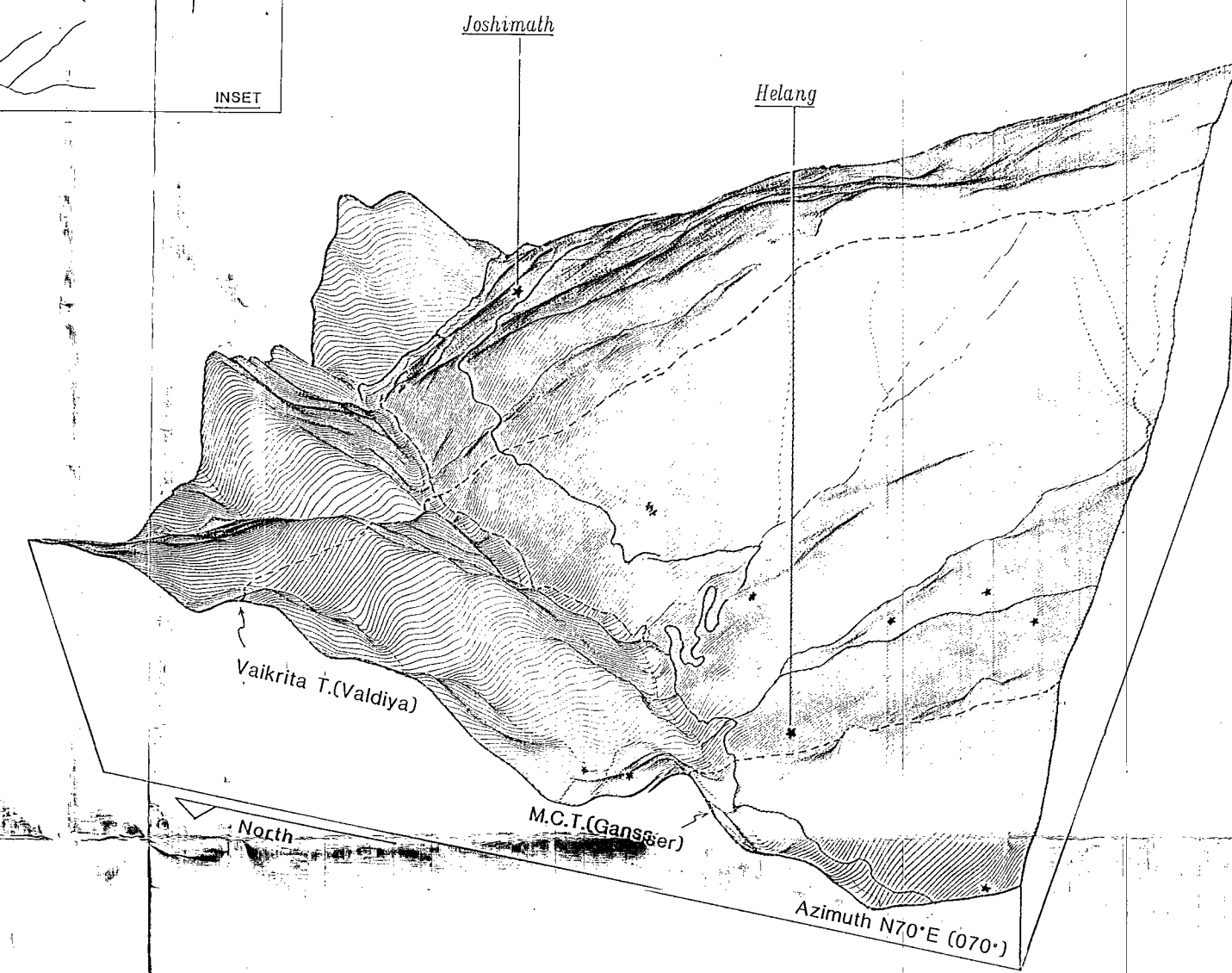
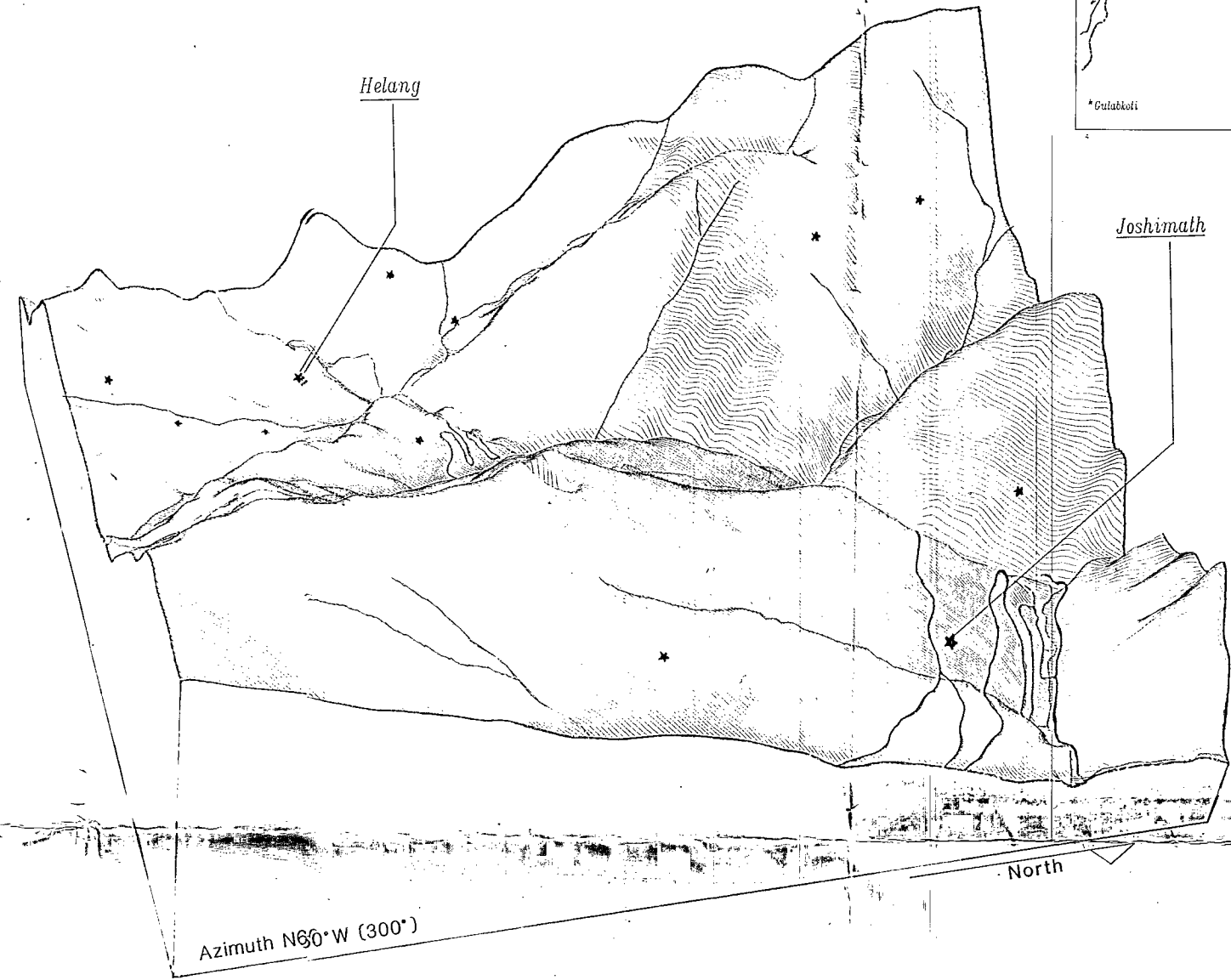
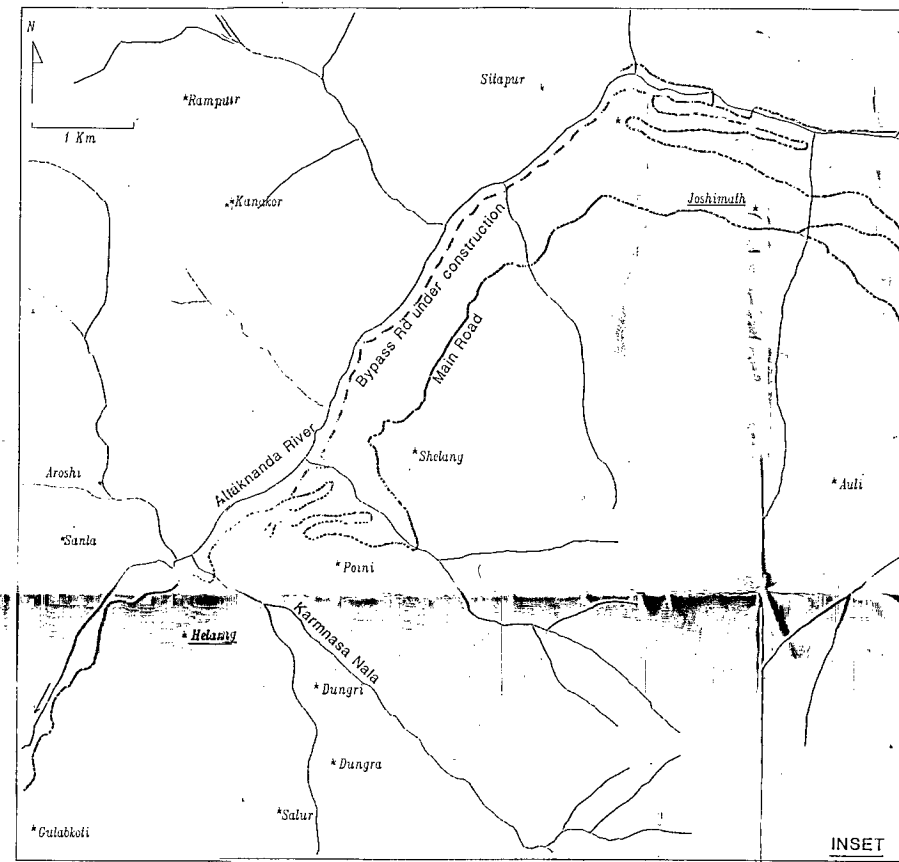
In the section, shear indicators are not to scale.  
Unornamented areas indicate gaps in exposure.

100m  
8.75km



**(PLATE - VII)**  
 The original locations of the specimens and photographs collected from the  
 Darmaganga Traverse. For legends etc see caption of Plate-VI.





Z-scale 0.008  
Altitude 50'

Digital Terrain Models of the core of Jashimath Field Area viewed at two different angles. The two major thrusts are delineated approximately in the DTM on right. Inset map shows names of localities.

Compilation courtesy: Mrs Susmita Ghosh, using ARC/INFO.

# Road log across the Vaikrita Thrust (Valdiya '80)

( Only the new lower Bypass Road is considered )



## INDEX

- Main foliation
- Main stretching lineation
- Later stretching lineation
- Shear band
- Boudin axis
- Mesoscale fold (axis)
- Limit of exposure
- Unornamented parts indicate gaps in exposure along the road section

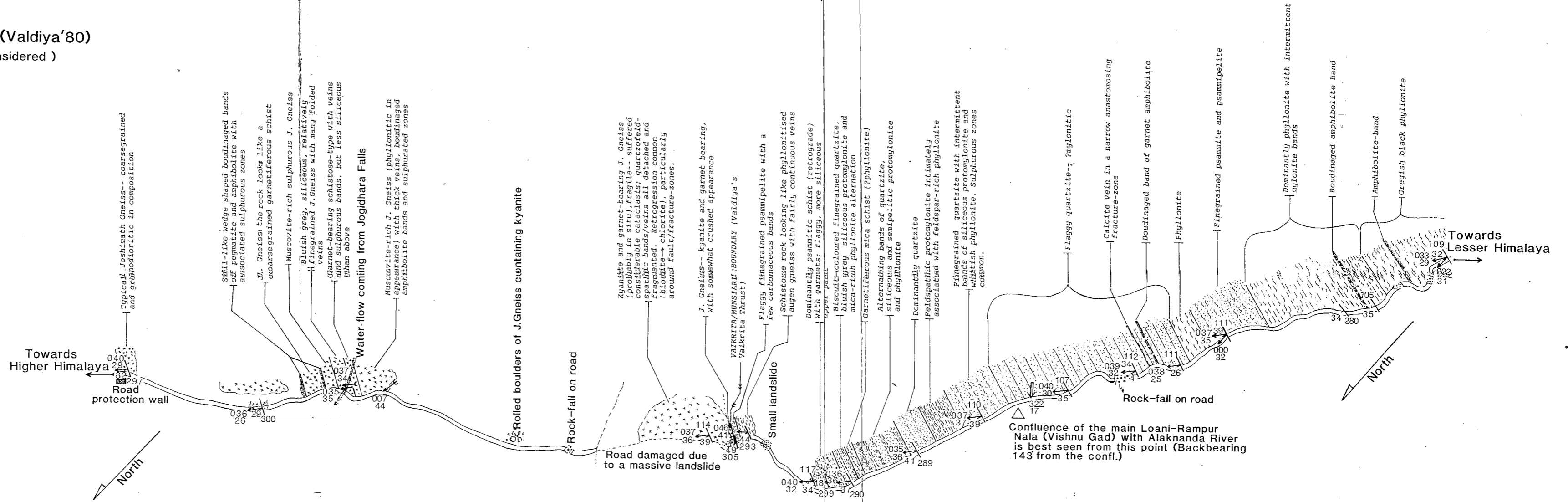


Fig. 4.2

Road log across the Mungsiari Thrust



INDEX

- Main foliation ↘
- Main stretching lineation →
- Later stretching lineation →
- Shear band ↘
- Boudin axis ↗
- Mesoscale folds (unassigned) ↻
- Limit of exposure ———
- T.P.—Telegraphic Pole †
- M.S.— Milestone □
- Unornamented parts indicate gaps in exposure along road-section

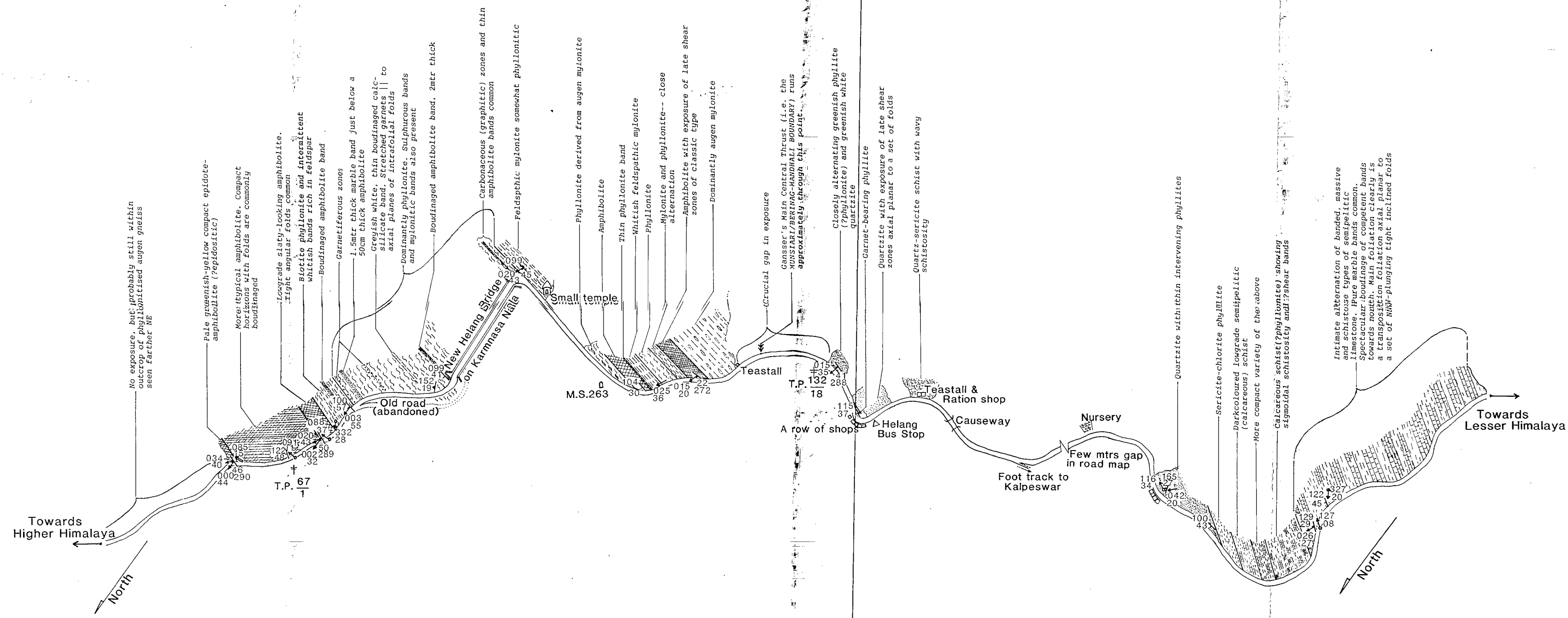


Fig. 4.4

