THE MUTH FORMATION IN THE PIN VALLEY (SPITI, N-INDIA): DEPOSITIONAL ENVIRONMENT AND ICHNOFAUNA OF A LOWER DEVONIAN BARRIER ISLAND SYSTEM

144 pages, 108 figures, 3 tables, 2 appendices

eingereicht von

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„Wer einmal nicht nur mit den Augen,
sondern mit der Seele in Indien gewesen ist,
dem bleibt es ein Heimwehland.“

Hermann Hesse

Thank U India.

Tibi Bokri Pyramide (6408 m) in the upper reaches of the Parahio Valley from the crestline of the Muth Fm. anticline, SE of Mikkim, towards the West.

Even in hundreds of ages of the gods
I could not tell you enough
about the glory and wonder of the Himalaya.

Puranas

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ABBREVIATIONS

Facies association	FA
Formation	Fm.
Higher Himalaya	HH
Higher Himalaya Crystalline	HHC
Indus Yarlung Suture Zone	IYSZ
Lesser Himalaya	LH
Litho-units	LU
Main Boundary Thrust	MBT
Main Central Thrust	MCT
Main Frontal Thrust	MFT
South Tibetan Detachment Zone	STDZ
Tethyan Himalaya	TH

PHOTO SCALES

50 Paisa coin	23 mm
1 Rupee coin	25 mm
2 Rupee coin	26 mm
Nikon lens cap	53 mm
Swiss army knife	85 mm
Lighter	60 mm
Hammer	570 mm
Estwing Hammer	330 mm

LITHOLOGICAL INDEX

Lithologies

- Shale
- Siltstone
- Black shale
- Sandstone
- Quartzarenite
- Conglomerate
- Intraformational breccia
- Dolomitic sandstone
- Sandy dolomite
- Sandy dolomite with crinoids
- Silicified dolomite
- Dolomite
- Dolomitic limestone
- Calcareous sandstone
- Sandy limestone
- Limestone
- Stromatolitic limestone
- Nodular limestone
- Marl
- Oolithic limestone
- Black calcareous mudstone
- Bioturbation

Fossils

- Bone
- Brachiopoda
- Conodont
- Coral
- Echinoderm (crinoids, echinids)
- Gastropoda
- Ooid
- Plant fragment
- Radiolaria
- Stromatolites, algal mat
- Trilobite
- Undefined shell

Sedimentary structures in the arenites of the Muth Fm.

- Structureless quartzite
- Horizontal lamination
- Tabular cross-bedding
- Tangential cross-bedding
- Trough cross-bedding
- Reactivation surface
- Ripples
- Domal stromatolites
- Mounds on bedding surface
- Bioturbation on bedding surface
- Root traces
- Sandstone diatremes
- Deformation bands
- Arthropod trackway

1. SUMMARY

Uppermost Proterozoic to Lower Paleozoic sedimentary sequences in the Pin Valley (Spiti) have been investigated; emphasizes has been laid on the sedimentology and ichnofauna of the Muth Fm. Lithological sections of the Pin Fm., Muth Fm. and Lipak Fm. have been measured, and useful, clear recognizable boundaries between the Formations have been defined. Typical lithological variations permit the regional correlation with Tethyan sediments from Zanskar to Kumaon. The boundary between the Pin Fm. and the Muth Fm. represents a disconformity that probably stands for a considerable time gap in the sedimentary record.

Sedimentary structures indicate that the Muth Fm. was deposited in a barrier-island system. Based on lithological variations and different sedimentary structures, four facies (FA1-4) have been distinguished, comprising (beginning at the base) shoreface to foreshore, coastal dune, lagoon and again shoreface depositional environments. The occurrence of the arthropod dominated ichnofauna is restricted to FA2 (beach to coastal dune environment).

The ichnoassemblage consists of abundant *Palmichnium antarcticum* and *Diplichnites gouldi* with rarer *Diplopodichnus biformis*, *Taenidium barretti*, *Didymaulichnus* cf. *lyelli*, *Didymaulyponomos* cf. *rowei*, *Metaichnia* isp. and vertical burrows of unclear affinity. The similarity of this ichnofauna to other trace fossil assemblages in marginal marine environments of similar age, in Antarctica and Australia, suggests that a recurrent Early Devonian ichnocoenosis around the margins of eastern Gondwana may be recognized.

Due to the lack of age-indicative fossils in the Muth Fm. the lower age limit of the Formation is constrained by the age of the overlying Lipak Fm. Conondont samples have been taken at the lowermost occurrence of limestone beds in the Lipak Fm. (*c*. 30 m above the base of the Formation), that yielded nicely preserved conodonts of Givetian age.

Deformation bands are very common in the Muth Fm. SE of Mikkim. Restored to their original orientation, they represent near vertical, E-W striking faults, indicating pre-Himalayan deformation with an older age limit represented by the age of sedimentation and an younger age limit represented by the age of complete cementation of the arenite. However, the exact age remains undetermined.

ZUSAMMENFASSUNG

Im Pin Tal (Spiti) wurden die Jungproterozoischen und Altpaläozoischen Sedimentserien bearbeitet, wobei der Schwerpunkt auf die Untersuchung der Sedimentologie und Spurenfossilien der Muth Fm. gelegt wurde. Es wurden lithologische Profile von der Pin Fm., Muth Fm. und Lipak Fm. angefertigt und logische, sowie leicht erkennbare Grenzen zwischen den Formationen definiert. Charakteristische lithologische Variationen ermöglichen eine überregionale Korrelation dieser Serien mit den Tethys Sedimenten von Zanskar bis Kumaon. Die Grenze zwischen der Pin Fm. und der überlagernden Muth Fm. stellt eine Diskonformitätsfläche mit einer vermutlich beträchtlichen Schichtlücke dar.

Die Muth Fm. wird anhand ihrer sedimentären Strukturen als ein Barriereninsel System interpretiert. Basierend auf lithologischen Veränderungen und unterschiedlichen sedimentärer Strukturen, werden 4 Fazies Assoziationen (FA1-FA4) unterschieden. Vom Liegenden ins Hangende finden sich *beach, shoreface* bis *foreshore*, Küstendünen, Lagune und abschließend wieder *shoreface* Ablagerungen. Arthropoden Laufspuren kommen ausschließlich in FA2 (Strand und Küstendünen) vor.

Die Spurenfossilien Vergesellschaftung besteht aus zahlreichen *Palmichnium antarcticum* und *Diplichnites gouldi*. Deutlich seltener finden sich *Diplopodichnus biformis, Taenidium barretti, Didymaulichnus* cf. *lyelli, Didymaulyponomos* cf. *rowei, Metaichnia* isp. und verticale Bohrgänge mit unklarer systematischer Zuordnung. Die Ähnlichkeit der Spurenfossilien Vergesellschaftung mit jenen von anderen, etwa zeitgleichen in Australien und der Antarktis bietet Hinweise auf eine einheitliche, unterdevonische Spurenfossilien Assoziation an der Küste von Ostgondwana.

Da in der Muth Fm. keine Fossilien gefunden wurden, die für eine Altereinstufung verwendbar wären, läßt sich lediglich das Mindestalter der Formation anhand der Datierung der überlagernden Lipak Fm. feststellen. Es wurden deshalb aus den ersten Kalkbänken der Lipak Fm. Conodontenproben genommen (etwa 30 m oberhalb der Basis). Die gut erhaltenen Conodonten Faunen dieser Proben zeigen ein mitteldevonisches Alter (Givet) an.

SE von Mikkim wurden zahlreiche *deformation bands* im gesamten Aufschlußbereich gefunden. Die Rekonstruktion dieser Strukturen auf ihre Orientierung vor der Verfaltung durch die Himalaya Orogenese, zeigt Ost-West streichende, subvertical einfallende Störungen, die eine Deformation anzeigen, die älter ist als die Himalaya Orogenese. Da es sich dabei um Strukturen handelt, die aus porösen Sandsteinen beschrieben sind, liegt das Alter dieser Deformation zwischen dem Sedimentationsalter der Muth Fm. und deren vollständiger Zementation.

RÉSUMÉ

Au Val de pin, des couches de sédiments du Protérozoïque supérieur et du Paléozoïque inférieur ont été étudiées portant principalement sur la sédimentologie et sur les traces de fossiles de la Fm de Muth. Des profils lithologiques de la Fm de Muth et de la Fm de Lipak ont été faits définissant les couches logiques et facilement reconnaissables. Avec les sédiments de Téthys du Zanskar au Kumaon, des variations lithologiques caractéristiques permettent une corrélation de ces strates. Les couches superposées entre la Fm de Pin et la Fm de Muth présentent une discontinuité de surface avec probablement une importante partie manquante.

La Fm de Muth est interprétée à l'aide de ses structures sédimentaires comme un système de barrière d'îlots. Basé sur des changements lithologiques et sur des structures sédimentaires différentes, 4 faciès d'associations (FA1 - FA4) se distinguent. De la salbande inférieure à la salbande supérieure se trouvent des sédiments de rivage, *shoreface* jusqu'à *foreshore*, de dune côtière, de lagune et pour finir, de nouveau, de plage. On trouve des traces de pas articulés exclusivement en FA2 (sédiments de plage et de dune côtière).

L´association de traces de fossiles composées de nombreux *Palmichnium antarcticum* et *Diplichnites gouldi*. Plus rarement on trouve des *Diplodpodichnus biformis, Taenidium barretti,*

Didymaulichnus cf. *Lyelli*, *Didymaulyponosmos* cf. *Rowei*, *Metaichnia* isp. et des forures verticales avec une coordination confuse et systématique. La similitude de cette association de traces de fossiles avec celles que l'on trouve, entre autres, pour à peu près la même période, en Australie et dans l'Antarctique offre une indication d'uniformité avec les traces de fossiles du Dévonien inférieur sur la côte de Gondwana-Est.

N'ayant pas trouvé de fossiles dans la Fm de Muth dont on aurait pu se servir pour une datation on ne peut fixer l'âge minimum de la formation qu'à l'aide de la datation de la Fm de Lipak. C'est pourquoi des échantillons Conodontes furent prélevés des premières couches calcaires de la Fm de Lipak, à 30 m environ au-dessus de sa base. Les échantillons de cette faune bien conservée indique un âge Dévonien moyen (Givet).

Au SE de Mikkim, de nombreuses *deformation bands* ont été trouvées dans l’entière zone. La reconstruction de ces *deformation bands* selon leur orientation avant la dégradation par l'orogenèse de l'Himalaya montre d'est en ouest des failles étendues subverticales plongeantes et qui indiquent probablement une déformation plus ancienne que l’orogenèse de l'Himalaya. Comme il s'agit de structures composées de grès poreux, on situe l'âge de cette déformation entre l'âge sédimentaire de la Fm de Muth et celui de sa cimentation complète

2. INTRODUCTION

2.1. Geographical setting

The Himalaya represents one of the most spectacular mountain belts on Earth with ten of the highest peaks in the world situated there. The Himalaya (Sanskrit: *him* = snow; *alaya* = home) extends from the western syntaxis at the Nanga Parbat to the eastern syntaxis at the Namche Barwa for some 2500 km along strike, its width ranging 230 to 320 km. The orogen steeply ascends from the Indus-Ganges alluvial plains in the South, forming a snow and ice covered rim at the southern margin of the Tibetan Plateau, representing an important climatic, floral and faunal boundary in South Asia (Fig. 2.1).

Politically, Spiti forms part of the Lahaul-Spiti district of the Indian state of Himachal Pradesh covering 3225 km^2 of a rough mountainous area with some peaks more than 6500 m high and valley floors at altitudes of some 3600 m. The Spiti River rises near Kunzam La in the NW of Spiti flowing towards SE and joins the Sutlej River near Khab after collecting the waters of its main tributaries, the Pin and Lingti Rivers (Fig. 2.3). The landscape resembles those of Ladakh and parts of Tibet (Fig. 2.8). Its position north of the main Himalayan range (Fig. 2.7) shields it from most of the benefit of the monsoon reducing the annual precipitation to less than 200 mm, resulting in a semi-arid climate, the major part of the precipitation in form of snow. It is difficult to imagine, but according to travelers' reports, Spiti has been much drier in former times, today's situation just reflecting the every year increasing rainfall of the last decade. Hutton (1839) states: "Rain is here almost unknown, falling only like angel's visits, and even then sparingly as to be of no use except to allay the clouds of dust for a few hours".

The temperatures are extreme both in summer and winter; as a result the mountains are barren and bleak. Vegetation is mainly restricted to river terraces, stripes along mountain creeks and artificially watered surfaces (Fig. 2.6). Today's existence of small forests is a result of intense reforestation efforts of the Indian government that brought trees to an area, which, at least in historical times, hardly saw them before. Fieldwork in Spiti is possible from mid-June to mid-October, with the best weather conditions in July and mid-September to mid-October. River crossings can be problematic, particularly in July and August when rivers are full with melt water (Fig. 2.10), residual snow bridges early in summer might be very useful.

The special climatic and terrain conditions of this region resulted in an unique fauna and flora. The Indian government made allowance for this specific situation and constituted the Pin Valley National Park in 1984, covering a core zone of 675 km^2 and a buffer zone of 1150 km^2. Snow and ice throughout the year cover a considerable part of the park. There are more than 20 species of animals in the park like ibex, golden eagle, bearded vulture, red fox, weasel, *etc.*, but it is especially renown for the protection of the elusive endangered snow leopard.

Historically and culturally Spiti belonged to the Tibetan sphere of influence for most of the time, forming part of the West-Tibetan kingdoms, its faith closely linked to the one of Ladakh. Local rulers, who had the title of Nonos, were either descendants of native Spiti families or emissionaries sent to look after the affairs of Spiti by the rulers of Ladakh. Whenever the central-power of Ladakh was weak, Spiti became autonomous and it became practically independent after

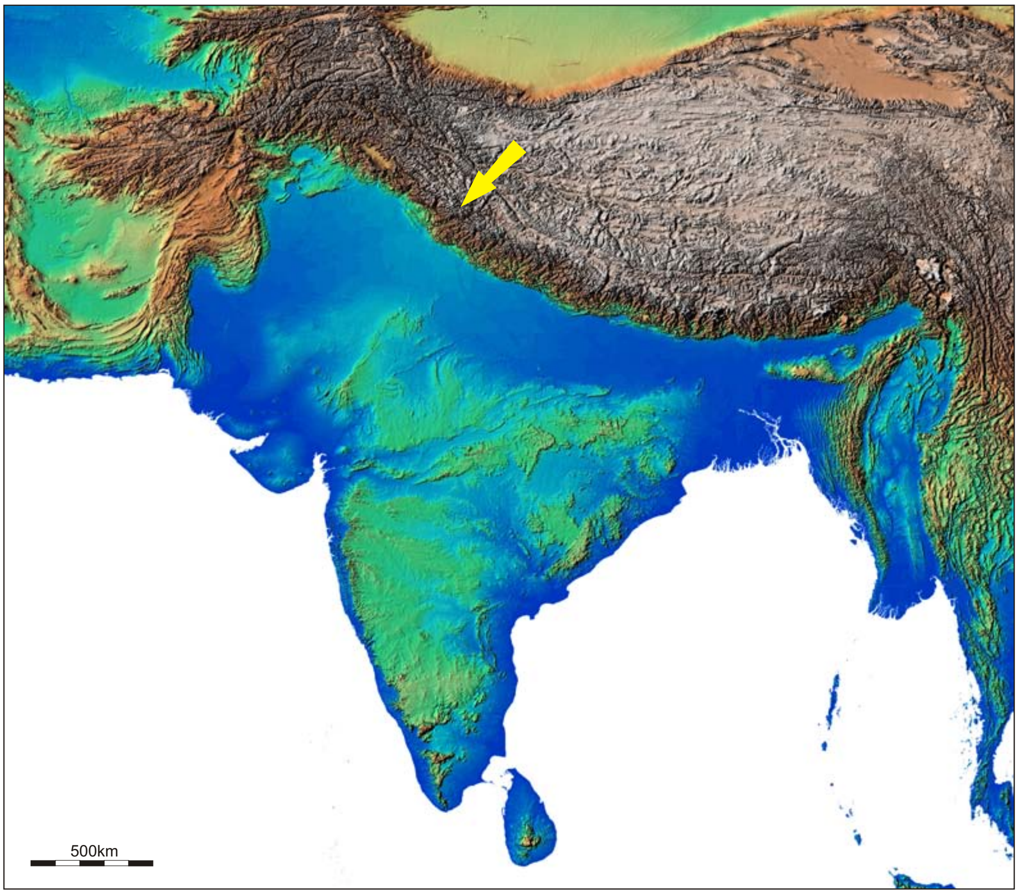


Fig. 2.1: Digital elevation model (1x1 mile USGS dataset) from the Indian subcontinent. Note the steep front of the Himalayan range towards the South and the huge Tibetan plateau in the North. The two syntaxes near the Nanga Parbat and the Namche Barwa are nicely visible. An arrow indicates the location of the Pin Valley.

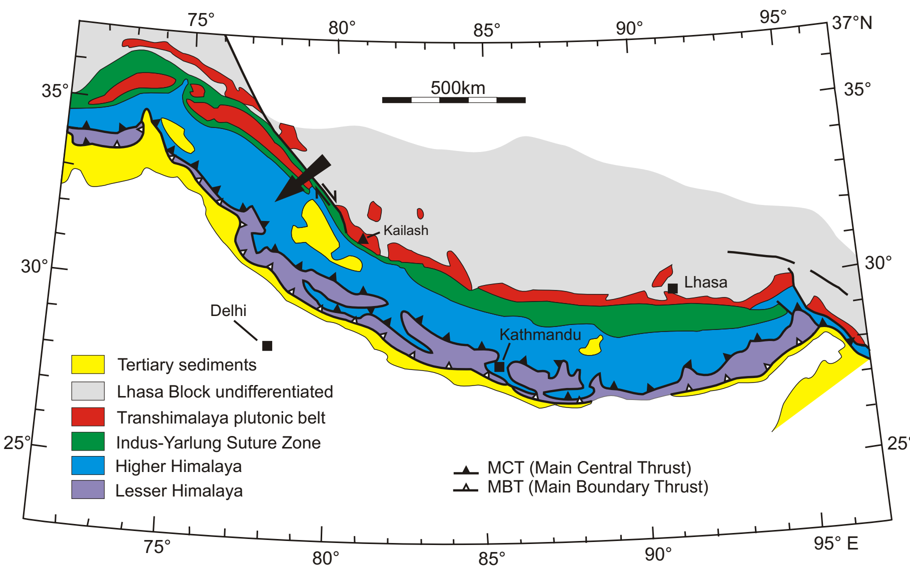


Fig. 2.2: Simplified tectonic map of the Himalayan orogen modified after Schuster (1999). An arrow indicates the location of the Pin Valley.

the Ladakh-Tibet war of 1681-83. Later on the Sikhs established a loose control over the principality, then passed on to the East India Company after the defeat of the Sikhs and finally became part of the independent India.

Due to the strategic important border situation to Chinese occupied Tibet, Spiti has remained restricted area for foreigners for several decades. In 1992 Spiti has been opened to foreign visitors without special permits and has become a popular destination for scientists and tourists, although this territory and particularly the area of interest, the Pin Valley, are difficult to access, being sheltered by unfriendly mountain ranges. In principal there are three possibilities to reach this area (Fig. 2.3). (i) Entering the valley from the NW by crossing Kunzam La (4590 m), which is possible by jeeps and busses from Mid-June to Mid-October. (ii) The entrance to the Spiti Valley from the South, from the Sutlej valley, is a rare example of crossing into a major valley deep in the inner Himalaya without actually crossing a pass. This way is usually open all year, but due to major landslides in the Sutlej Valley during monsoon not advisable in these periods. (iii) The most impressive, but also challenging passage is to cross Pin-Parbati Pass (5319 m), which is open between July and October and connects the Parbati Valley with the upper Pin reaches. Crossing this pass is only advisable with local guides.

2.2. History of geological investigations

Spiti and in particular the Pin Valley represent an “El Dorado” for geologists and paleontologists, renown for the excellent outcrop situation, rich fossil content, lithological variation and the almost continuos stratigraphic range from Neoproterozoic to Cretaceous. Bhargava & Bassi (1998) have summarized the history of geological investigation and the reader is referred to this publication for an exhausted list of references.

Early in the 19th century Gerard (1827) made first reports about fossil occurrences in the sedimentary successions of Spiti. In 1862 Thomas Oldham, the first director of the Geological Survey of India, visited Eduard Suess in Vienna and asked for a paleontologist for the Indian survey (Sengör 1998). Suess recommended Ferdinand Stoliczka from the Austrian Geological Survey who published his pioneering results in 1866. Unfortunately, it was not granted to him to work in this area for a long time and he died early. Griesbach (1891) continued these investigations. Hayden (1904, 1908) wrote brilliant and irreplaceable fundamental publications about the geology of Spiti. He mapped large areas and established the still valid stratigraphy.

Reed (1910, 1912) described Lower Paleozoic fossils of the Pin and Parahio Valleys in detail. The numerous publications of Carl Diener (see Gansser 1964 for references) with emphasis about Triassic successions in the Himalaya are still valuable sources for this period. Auden (1935) introduced the term “Tethys Himalaya” (TH) for the fossiliferous sequences north of the Higher Himalaya Crystalline (HHC). Geological fieldwork in the Pin Valley would be unthinkable without the detailed mapping by Fuchs (1982) in the scale of 1:50.000. Srikantia (1981) and Bagati (1990) carried out the helpful work of straightening up the stratigraphy of the Spiti sedimentary successions.

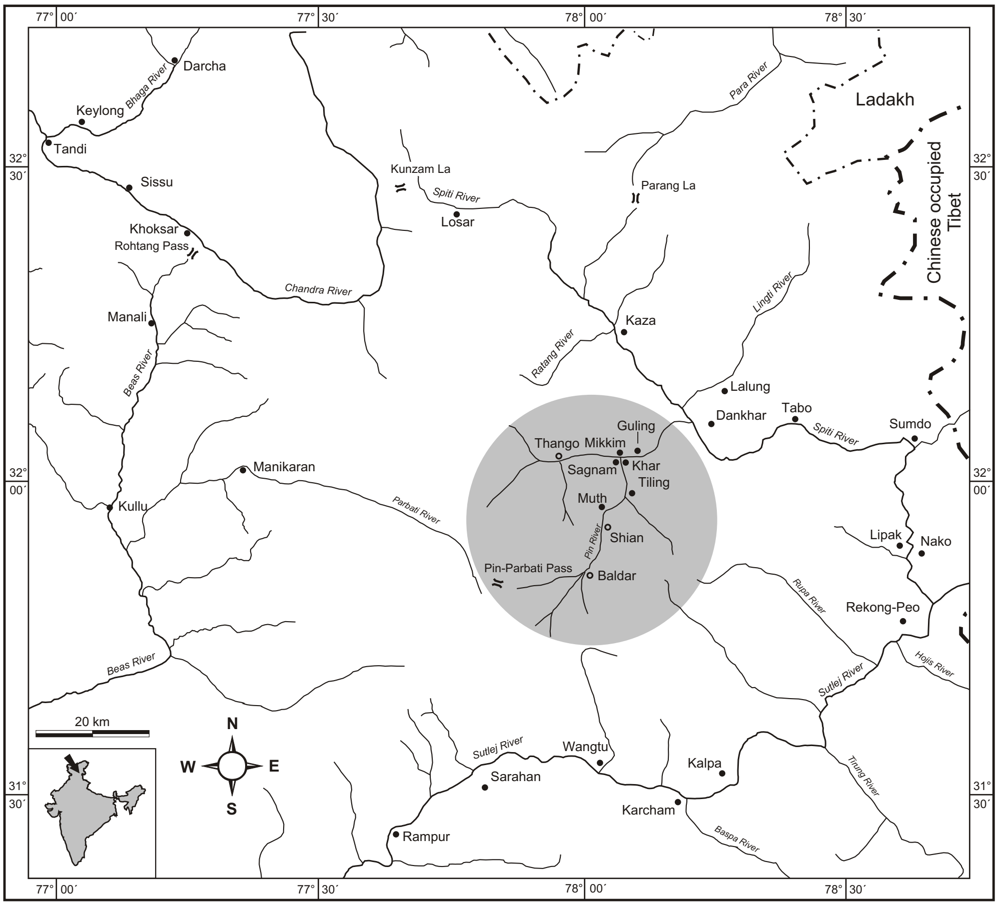


Fig. 2.3: Map of the NE part of Himachal Pradesh, the big gray spot indicates the position of the Pin Valley.

Fig. 2.4: View from the western termination of the anticline towards the North to Mikkim. Muth Fm. in the foreground.

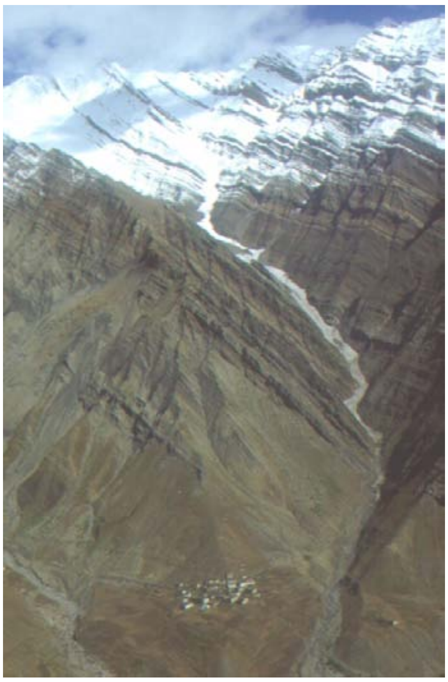


Fig. 2.5: View from the right side of the Pin River towards the NW to village Muth. Ravine left of Muth leads to section G-G´.

In the 80´s and first half of the 90´s a lot of valuable sedimentological and stratigraphic research has been carried out by earth scientist from Milano (*e.g.* Gaetani & Garzanti 1991, Garzanti *et al.* 1996a,b and references cited therein). Recently, Bhargava (1997), Srikantia & Bhargava (1998) and Bhargava & Bassi (1998) have written useful synopses of the geology of Himachal Pradesh and especially of Spiti.

2.3. Geological setting

The Himalaya is one of the spectacular examples of a continent-continent collision mountain belt on Earth (Fig. 2.1). The collision of India and Asia is the result of the closure of the Tethys after 65 and before 55 Ma ago (Klootwijk *et al.* 1992, Klootwijk *et al.* 1994), the exact timing is still in discussion. Continuos northwards movement of India caused contractional and extensional tectonics, crustal thickening, regional metamorphism, anatexis, thrusting, normal faulting, tectonic exhumation, surface uplift and erosion.

Based on the classic book by Gansser (1964) the Himalaya is divided into several tectonic zones (see also Medlicott & Blanford 1879/1887), which are bounded by major fault zones and on the whole correspond with geo-morphological divisions (Srikantia & Bhargava 1998). These zones from South to North are (i) Sub-Himalaya, (ii) Lesser Himalaya (LH), (iii) Higher Himalaya (HH), (iv) Indus Yarlung Suture Zone (IYSZ), (v) Transhimalaya (Fig. 2.2). The units south of the suture zone are bounded by SW-directed major thrust zones, their activity pattern showing a foreland-directed propagation in-sequence from North to South. The three southernmost zones, with emphasis on the geological situation in the NW-Himalaya of Himachal Pradesh, are described below:

SUB-HIMALAYA

This element represents the outermost zone of the mountain belt that rises up just north of the Indus-Ganges plains constituting of densely vegetated low-altitude foothills with an average altitude of 900-1500 m. Its southernmost part is known as Siwalik Range. This zone is prone with landslides during monsoon.

The Sub-Himalaya tectonic unit comprises Tertiary molasse-type sediments, which are overthrusted by the Lesser Himalaya along the Main Boundary Thrust (MBT) and subsequently they themselves are thrust SW-wards over Holocene sediments of the Indus-Ganges plains by the Main Frontal Thrust (MFT). In the Himachal Pradesh sector, the lower Eocene to lower Miocene Sirmur Group (Subathu, Dagshai and Kasauli Fms.) consisting of foraminiferal limestone, sandstone, and mudstone is succeeded by the mainly terrigenous clastic sediments of the middle Miocene to Pleistocene Siwalik Group (Medlicott & Blanford 1879/1887). Due to the ongoing in-sequence propagation of the thrust activity towards the South, the depositional basin of the Siwalik Group is situated south of the basin of the Sirmur Group, nevertheless it on-laps to some extent on the

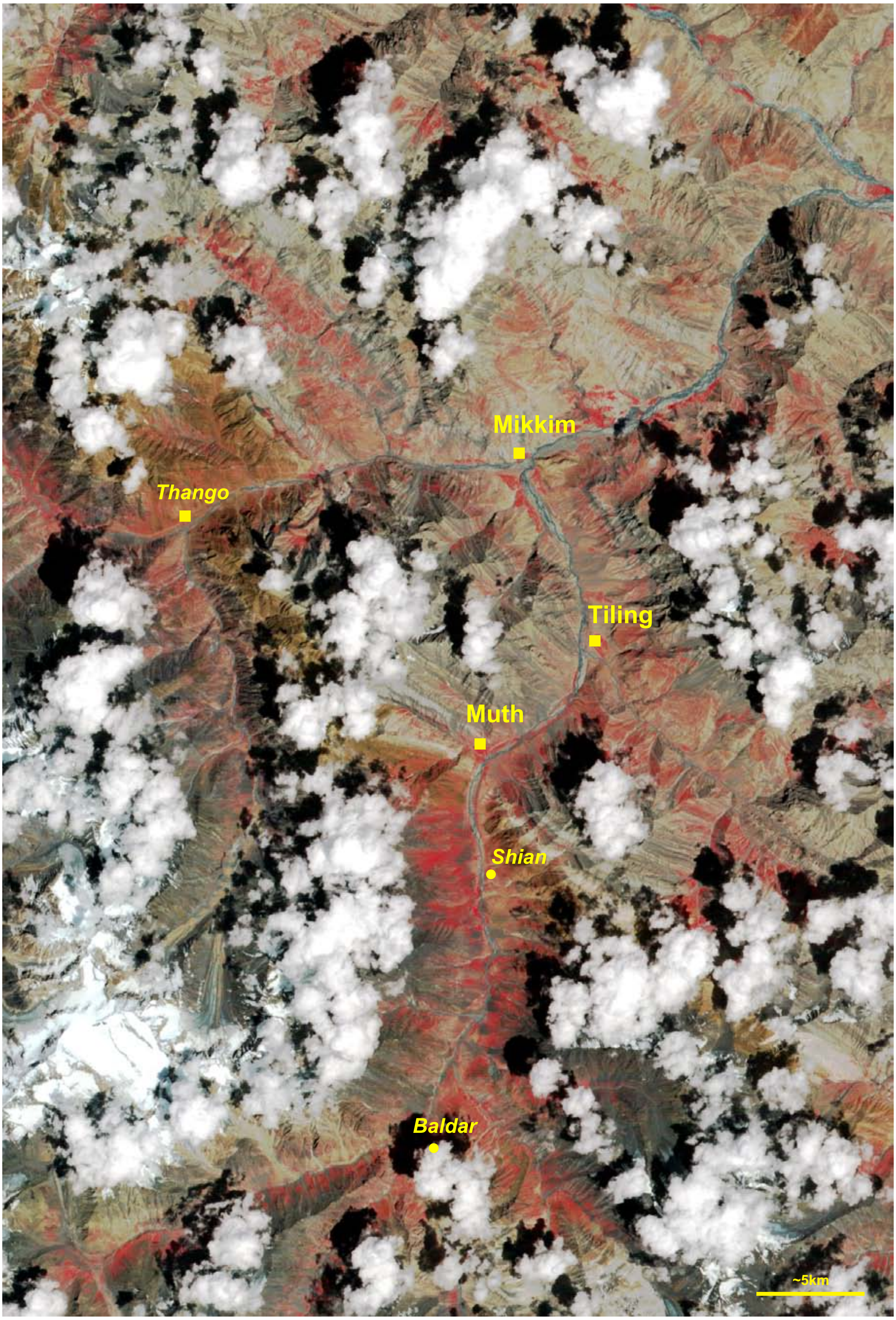


Fig. 2.6: JERS-1 OVN satellite image (acquisition date 24-8-1995) of the Pin and Parahio Valleys. Due to little vegetation and strong lithologic contrasts bedding and geological structures are nicely visible. The white horizon south of Muth represents the Muth Fm.

Sirmur Group. (Srikantia & Bhargava 1998). The sedimentary successions were affected by folding and imbrication.

LESSER HIMALAYA

The Lesser Himalaya shows alpine-type mountain ranges with altitudes ranging between *c*. 1500 to 5000 m. Due to the position directly South of the main range, This zone benefits from a lot of rain during monsoon and thus it is densely covered by vegetation.

The Lesser Himalaya tectonic unit is placed South of the Higher Himalaya, which overthrusted the Lesser Himalaya along the Main Central Thrust (MCT; Heim & Gansser 1939), and North of the MBT, where the Lesser Himalaya tectonic unit is thrust over the Sub-Himalaya. Additionally, Lesser Himalaya lithologies can be found in large tectonic windows below the Higher Himalaya, the Kishtwar Window (Fuchs 1975, Guntli 1993) and the Larji-Kullu-Rampur Window (Auden 1934, Frank *et al.* 1973) indicating a minimum thrusting distance of 100 km on the km-thick MCT-Zone.

The ages of the lithologies range from Precambrian to Eocene with a major break in deposition between middle Cambrian and Eocene, the metamorphic grade is generally low, but can reach lower greenschist conditions in the uppermost nappes (Srikantia & Bhargava 1998). Within the LH several tectonic units can be distinguished, in principal several nappes are thrust above nearly unmetamorphosed, imbricated, para-autochthonous sedimentary series (Frank *et al.* 1995, Srikantia & Bhargava 1998).

The amphibolite grade Jutogh Nappe overlying LH units, is the southernmost tectonic element of the HH even forming a tectonic *Klippe* near Shimla (Pilgrim & West 1928). Its affiliation to the HH tectonic unit is seen for example in the Tons Valley, where it has a position above MCT-mylonites corresponding with the Bharagaon mylonites (pers. comm. Frank & Bhargava 2000). The Jutogh Nappe tectonically overlies lower greenschist grade thrust slices, which themselves are thrust above the imbricated para-autochthonous sedimentary sequences.

Four successive para-autochthonous Proterozoic sedimentary megacycles, bounded by unconformities, have been distinguished (Virdi 1995, Srikantia & Bhargava 1998): (i) Rampur-Berinag cycle (~ 1800 Ma; Miller et al. in press) consist of striking ortho-quartzites and slates associated with basic volcanics; (ii) Shali (= Larji, = Deoban) cycle (*c*. 1400-900 Ma) comprises dolomitic and calcareous stromatolites with very rare siliciclastics; (iii) Shimla cycle (*c*. 900-700) is made of shales and greywackes with minor carbonates and rare volcanics; the cycle ends with redbeds (Nagthat Fm.); (iv) Blaini-Krol-Tal cycle (*c*. 700 Ma to early Cambrian) shows two diamictite horizons (Blaini Group) followed by black shales and carbonates (Infra Krol Fm.) and finally succeeded by dolomites with some siliciclastics.

The Proterozoic sedimentary series of the LH represent thick and far correlatable deposits, for example the Shimla Slates can probably be correlated with the Attock - and Hazara Slates west of the syntaxis in Pakistan (Wadia 1934, Pascoe 1959, Gansser 1964). On the other hand these sediments show little commons with those from the northern part of the Indian Peninsular region, not even equivalents of the Blaini Group have been found yet

Fig. 2.7: Upper Chandra Valley. View towards the South to snow covered peaks of the main range consisting of Ordovician granites intruded into the Haimanta Group.

Fig. 2.8: View from Dankhar Gompa towards the NW to the confluence of the Spiti River (from right to lower left) and the Pin River.

Fig. 2.9: View from Mikkim towards the SE to the main trackway site at the W termination of the anticline above the Pin River, where fresh bedding and foreset surfaces of the Muth Fm. are exposed.

Fig. 2.10: Early in the summer river crossings are a challenge for people and animals.

(pers. comm. Frank 2000). Even more, some lithological, geochemical and geochronological similarities between the Lesser Himalayan Shimla and Krol cycles and the Haimanta Group of the HH suggest a correlation and a deposition in the same basin (Virdi 1995, Frank *et al.* 1995).

Isolated remnants of the Paleocene to lower Eocene Kakara Fm. (Srikantia & Bhargava 1967), which were deposited during a transgression on the Precambrian to middle Cambrian series, can be found in the southern part of the Lesser Himalaya (Srikantia & Bhargava 1998). In the NW Himalayas, sediments of this Formation can be found up to 45 km to the North of the MBT, in Nepal even up to 95 km.

HIGHER HIMALAYA

The HH is a rugged unfriendly region with narrow valleys and ten 8000´s among its spectacular peaks. The Sutlej is the only river that is able to wind its way through these ranges from North to South.

In the general accepted opinion the HH forms the northernmost tectonic unit of Indian continental crust in the Himalayan orogen. The Main Central Thrust marks the southern limit, where the HH is thrust above the LH tectonic unit. The ophiolithic melange of the Indus-Yarlung Suture Zone, which represents the remnants of the Neo-Tethys Ocean, forms the northern limit.

In principal the Higher Himalaya is divided into 2 sub-units: (i) Higher Himalaya Crystalline (HHC), *i.e.* “Central Gneiss” of Stoliczka (1866) and “Vaikrita Group” of Griesbach (1891), and the (ii) Tethyan Himalaya, *i.e*. “Tethys Himalaya” of Auden (1935) and “Tibetan Himalaya” of Gansser (1964).

The HIGHER HIMALAYA CRYSTALLINE is located north of the MCT, where it is thrust above the LH tectonic unit; its northern limit is a diffuse zone of decreasing metamorphic intensity towards the TH; its upper boundary has not been satisfyingly defined yet. The unit comprises amphibolite grade metasediments of the Vaikrita Group in lower levels with gradually decreasing metamorphic grade towards higher levels into hardly metamorphosed sediments of the same succession towards the North, the Haimanta Group (Griesbach 1891, Frank *et al.* 1995).

Abundant high-level intrusions of Early Ordovician, peraluminous granites with minor associated basic intrusions are restricted to the HHC, they are not found in the LH (Frank *et al.* 1995). According to Miller (in prep.) these granites indicate an extensional setting in their geochemistry that fits to the observation of pre-Himalayan deformation in the Pin Valley (Grasemann *et al.* 1997, Wiesmayr *et al.* 1998). Leucogranitic intrusives generated by anatexis during the Tertiary metamorphism are rare and occur near the top of the HHC (Le Fort 1975), but may also intrude basal horizons of the Ordovician Ralam Conglomerate in Kumaon (Griesbach 1891).

The term Higher Himalaya Crystalline is somehow misleading, because this unit experienced its main metamorphism during Tertiary times, thus per definition it can not represent the crystalline basement of the TH sediments. In analogue, according to Parrish & Hodges (1996) there is no real basement found for the late Proterozoic sediments in the

HH of Nepal. The HH rather constitutes the Neoproterozoic (in the sense of Knoll 2000) to Cambrian metasediments (Vaikrita and Haimanta Groups) below the TH with continuous sedimentation into the Palaeozoic, apart from a depositional break in upper Cambrian to lower Ordovician time. In places where the contact between Vaikritas and Haimantas is not complicated by faults, the gradual relationship is clearly evident. *"Between it* [Vaikritas] *and the next following clearly sedimentary rocks, which I have termed the haimanta system, a clearly defined boundary scarcely exists. In nearly all sections which I have hitherto examined between the Kali river and Spiti, the schists seem to pass gradually into the overlying slates, phyllites and quartzites of the haimantas"* (Griesbach 1891).

Wadia (1934) and Gansser (1964) describe the same situation in Kashmir, where the Salkhala Schists and Dogra Slates represent correlatives of the Vaikrita and Haimanta Groups. *"Dogras and Salkhalas may after all represent deeper geosynclinal sediments of the Precambrian which change gradually into shallower deposits with the beginning of the Palaeozoic"* (Gansser 1964).

Recently, Dèzes (1999) and Dèzes *et al*. (1999) pointed out, that *"It is now an established fact that the relation between the HHCS* [HHC Sequence] *and the TH is not one of basement-cover type, but that the metasedimentary series of the HHCS represent the metamorphic equivalent of the lowermost sedimentary series of the TH"*(Dèzes 1999).

This argumentation rises the question of the location of the real basement *senso stricto* for the Higher Himalaya sedimentary sequences, *i.e*. distinctly older metamorphous lithologies below the Neoproterozoic Vaikrita/Haimanta Groups. The most probable candidates for this basement are peculiar lithologies (augengneisses, calcite-marbles and carbonaceous slates) at the base of the HH tectonic unit, which have first been described by Fuchs (1967) from western Nepal, who called this unit the Lower Kathmandu Nappe. According to Frank (pers. comm. 2000), at least parts of these carbonaceous slates represent thrust slices of definitely younger ages. At the Jumla Window (Nepal), broadly 1800 Ma old meta-granites from a considerable part of this unit (pers. comm. Frank 2000). Fuchs & Frank (1970) and Frank & Fuchs (1970) introduced the term LOWER CRYSTALLINE NAPPE for this some 10 to several 100 m thick unit. In the following years comparable lithologies in the same tectonic position at the base of the Higher Crystalline have been found from the Kishtwar to the Anapurna region (Frank *et al.* 1995 and references cited therein).

Apart from the characteristic lithology this unit is typically situated at the base of the HH above the brittle contact of the MCT to the LH, with a more or less gradational increase in metamorphic grades into higher structural levels (Thöni, 1977, Honegger 1983). Whole rock Rb/Sr isochron ages on these augengneisses gave 1960±29 Ma for a sample from the Bharagaon gneiss in the Sutlej Valley and around 1860 Ma for a sample near Bajaura in the Kullu Valley (Frank *et al.* 1995, Miller et al. in press).

The Almora crystalline in Kumaon that is situated above the MCT comprises abundant, moderately deformed granitic intrusions with ages around 1800 Ma (pers. comm. Frank 2000), thus representing a potential candidate for the basement of the Haimanta Group.

In the literature the TETHYAN HIMALAYA comprises nearly continuos sedimentary sequences from Cambrian to Eocene (Hayden 1904, Heim & Gansser 1939, Baud *et al*. 1984) with the main occurrences in NW India in the Kashmir, Ladakh/Zanskar, Spiti and Kumaon synclinoria. During the deposition of the sediments of the TH, this zone formed a vast, continuos, and more or less hardly structured single depositional basin. Today's division into the above synclinoria is just a result of the Himalayan deformation (Bhargava & Bassi 1998), therefore the use of the term "basin" for the different areas of the TH is misleading and should be avoided. The gap in sedimentation from middle Cambrian to lower Ordovician and the associated formation of an angular unconformity (Fig. 4.5) *e.g.* in the Pin Valley is remarkable (Stoliczka 1866, Fuchs 1982, Grasemann *et al.* 1997). Marine sedimentation in Zanskar terminated with Eocene nummulitic limestones (Gaetani *et al.* 1986).

In the NW Himalaya the TH is situated immediately north of the Zanskar Shear Zone (Herren 1987, Dézes *et al*. 1999) and similar normal faults like the Sangla Detachment (Vannay & Grasemann 1998) and the South Tibetan Detachment Zone (STDZ) far to the East (Burg *et al.* 1984, Burchfield *et al.* 1992). There are several minor occurrences of Phanerozoic sediments south of these normal faults, like the Kalhel Syncline (Rattan 1973) and the Tandi Syncline (Stoliczka 1866), but also vast areas like in Kashmir and Puljauki (south of Kathmandu). At least in the NW Himalayas there are indications that these Miocene normal faults represent reactivated thrusts, which had been active during crustal thickening prior to the onset of the MCT (Jain & Manickavasagam 1993, Vannay & Grasemann 1998, Wiesmayr & Grasemann 1999).

Grasemann (pers. comm. 1999) explained the apparent discontinuity and variations in throw of these normal faults in the HH in a simple but fundamental way. Grasemann *et al.* (1999) stated extruding wedge geometry for the observed extrusion of metamorphic rocks at the base of the HH, with the MCT at the base and STDZ-related normal faults at its top. Vorticity analyses of the ductile deformation of the MCT indicate general shear conditions with a considerable amount of pure shear, thus inducing extension in the hanging wall that culminates in previous mentioned normal faults. The amount of extension is related to the position within the normal faults relative to the branching point of this extruding wedge (Grasemann *et al.* 1999, Fig. 11c). Therefore the observed throw on these normal faults depends on the level of erosion, to which depth the normal fault is exposed, or in other words the distance of the exposed part of the normal fault relative to the branching point. This model is supported by the observation that in places with high throw values (*e.g.* Zanskar and Everest area) shallower levels of the hanging wall are exposed than for example in the Sutlej Valley, with lesser throw.

A differentiation between HHC and TH is absolutely necessary, but for the above mentioned reasons the long used division solely based on the metamorphic grade is insufficient because of its gradational nature and it has had fatal effects on the understanding of the basement-cover relationship in the Higher Himalaya. The alternative approach to use the Zanskar Shear Zone and comparable normal faults for a pure tectonic differentiation definitively fails in the existence of the above-mentioned Phanerozoic sediments *e.g.* Kashmir, Puljauki, Kalhel and Tandi Synclines south of these detachment systems.

Therefore a tempting idea is, to draw the boundary between Higher Himalaya Crystalline and the TH at the Cambrian-Ordovician unconformity, *i.e.* in Spiti between the Parahio Fm. (Haimanta Group) and Shian Fm. and their correlatives. The Higher Himalaya tectonic unit could be divided into the Vaikrita Himalaya (Vaikrita Group of Griesbach 1891) consisting of the Lower Crystalline Nappe and the Neoproterozoic to late Cambrian metasediments (*i.e*. Haimanta Group) with granitic intrusions between the MCT and the Ordovician unconformity. The Tethyan Himalaya in this classification would consist of sediments from Early Ordovician to Eocene time.

The unconformity below the Early Ordovician transgressive conglomerates is one of the most widespread (Fig. 4.5) and drastic events in the whole stratigraphy of the HH which can be traced from the Peshawar Basin (Pogue *et al.* 1992), the Karakorum (Le Fort *et al*. 1994), Zanskar (Gaetani *et al*. 1986), Spiti (Hayden 1904, Fuchs 1982) to Kumaon (Heim & Gansser 1939, Shah & Sinha 1974). This event formed a depositional gap from about latest Cambrian to earliest Ordovician, in some places with an angular unconformity *e.g.* in the Pin Valley (Fuchs 1982). The unconformity is possibly related to an orogenic event at the Cambrian-Ordovician boundary (Garzanti *et al.* 1986), post-orogenic extension (Girard & Bussy 1999) or solely extensional tectonics (Miller in prep.). The ages of widespread granitic intrusions in the lower levels of the HH cluster between 470-480 Ma, this means they formed relative during the youngest part of the depositional gap.

In the here suggested definition the throw on normal faults like the Zanskar Shear Zone [*c.* 25 km (Herren 1987); 35 ± 9 km (Dèzes 1999, Dèzes *et al.* 1999)] is regarded as considerable, but the primary relationship of the sedimentary series in the hangingwall and footwall are still recognizable.

The approach based on stratigraphic criterions is very simple and easy to reproduce and avoids many of the above-mentioned problems. A stratigraphic boundary is tempting for the fact, that both the TH and the HHC comprise mainly sediments, although in the lower levels of the latter unit this fact is veiled by metamorphism and granitic intrusions. Additionally, drawing the boundary at the Cambrian-Ordovician unconformity would bring the TH closer to the nowadays-usual understanding of the time span of the Tethys term and even to the original definition by Eduard Suess (Sengör 1998).

2.4. Tectonic development in the Pin Valley

This chapter provides just a cursory overview about the tectonic development in the Pin Valley. For much more information on this topic the reader is referred to Wiesmayr (2000), who carried out detailed structural investigations in this area simultaneously to the present work. The area is clearly dominated by large-scale inclined horizontal folds, with fold axes trending NW-SE

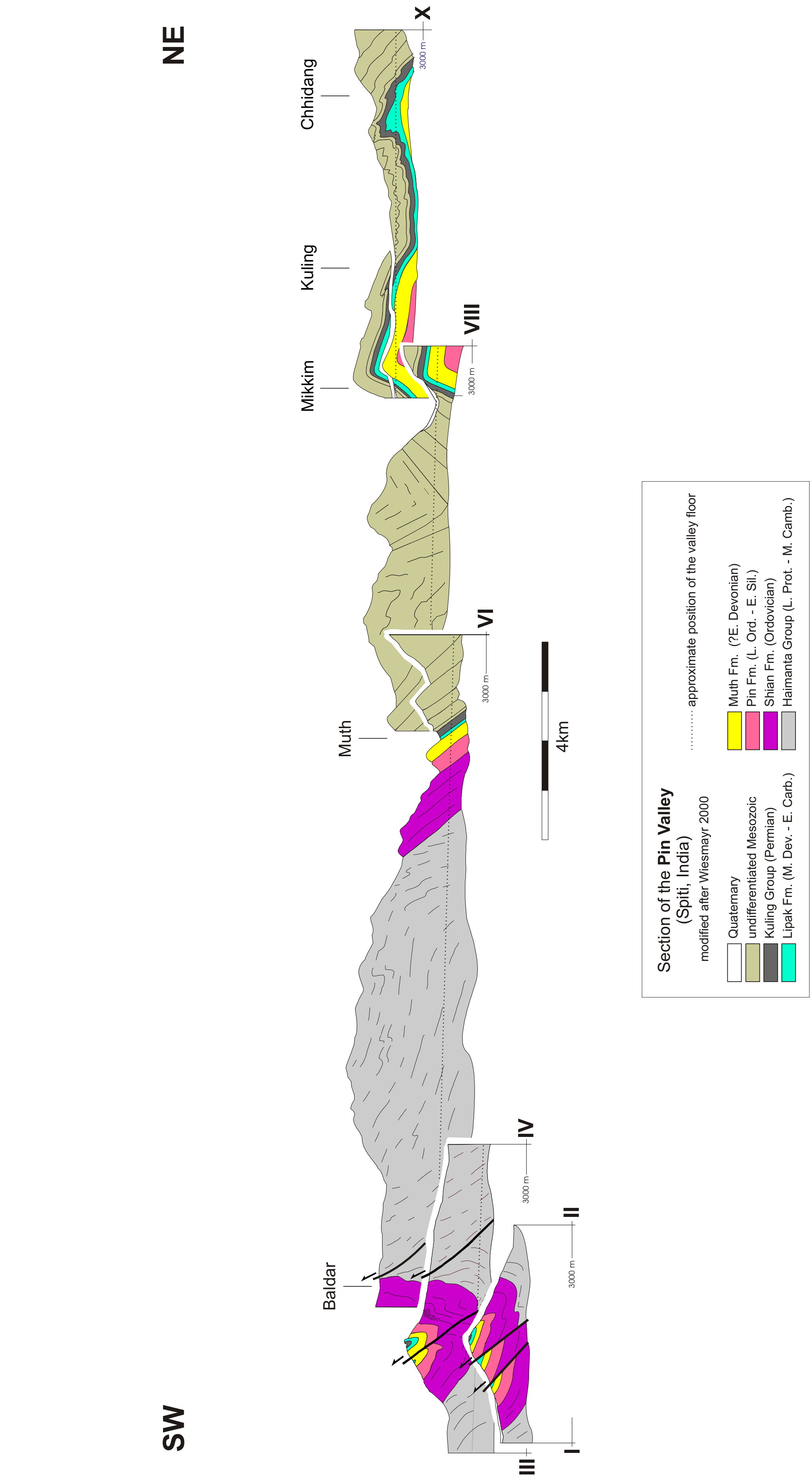


Fig. 2.11: Serial geological cross-sections in the Pin Valley, modified after Wiesmayr (2000).

and wavelengths of approximately 5 km (Figs. 2.6, 2.11), caused by Himalayan orogeny (Fuchs 1982). Nevertheless several older deformation events are recognizable.

The first evident deformation stage is represented by the angular unconformity (Fig. 4.5) below the probably Early Ordovician basal conglomerates of the Shian Fm., which has first been recognized by Griesbach (1891). The angle of the angular unconformity is variable and shows some 15° in the Pin Valley (Wiesmayr 2000), but is nearly at right angle in Kumaon (Griesbach 1891). Additionally to these observations, Grasemann *et al*. (1997) and Wiesmayr (2000) have presented the existence of weak folding in the sediments below the unconformity predating the unconformity. This deformation resulted in a depositional gap of latest Cambrian to **earliest Ordovician** time and broadly coincides with widespread granitic intrusions (470-480 Ma) in the HH, their geotectonic setting being still in discussion (see chapter 2.3.).

Deformation bands (Aydin 1977) can be found in all levels of the Muth Fm., but are most abundant in the uppermost third. These subvertical deformation bands trend consistently E-W and because these structures are thought to develop in still porous arenite, they probably indicate minor tectonic activity during **Devonian** time. For a more detailed description the reader is referred to Chapter 5.8.

The third evident deformation phase is manifested in the angular unconformity below the base of the Kuling Group. This unconformity is related to the opening of the Neotethys Ocean during **Early Permian** time and the development of a pronounced rift shoulder in parts of the TH (Stampfli *et al.* 1991, Garzanti *et al.* 1996a). Dependent to the position relative to the rift shoulder the erosion levels vary considerably from hardly any disconformity visible in the Spiti Valley, Kuling Group overlying the Silurian Pin Fm. in NW Kinnaur and even erosion down to the Neoproterozoic Phe Fm. in the Tandi syncline (Vannay 1993, Bhargava & Bassi 1998).

The fourth event represents the Eo-Himalayan deformation and is responsible for the nowadays so prominent SW vergent large-scale fault-thrust belt in the TH. This deformation has been dated by Ar/Ar stepwise heating method on 2-11 μm small illite from steep N-wards dipping cleavage domains to be **Early Eocene** in age (Wiesmayr *et al.* 1998).

Illite crystallinity and geometric parameters indicate a subsequent passive rotation of this area, correlated with a shallowly NE-dipping crenulation cleavage with increasing intensity towards the South. This rotation is explained by SW-directed thrusting on a detachment horizon, which position coincides with the position of the reactivated normal fault of the Sangla Detachment (Vannay & Grasemann 1998), a **Middle Miocene** age is assumed for this deformation (Wiesmayr 2000).

2.5. Aims of the thesis

The Pin Valley is renown for the excellent outcrop situation, rich fossil content, lithological variations and the almost continuos stratigraphic range from Neoproterozoic to Cretaceous. This specific situation offers great conditions for geological studies. Surprisingly, in spite of the long duration of geological research in this area, there are still many questions open, leading to the following project aims:

(1) Establishment of a clear lower Paleozoic litho-stratigraphy in the Pin Valley to build a base for the correlation with the series in Zanskar and Kumaon. What are the implications of the angular unconformity at the base of the Ordovician Shian Fm? (in collaboration with Gerhard Wiesmayr, Vienna)

(2) Although the Muth Fm. represents one of the largest sand accumulations in the geological history and is an important lithological marker horizon for a vast area of the TH, its sedimentology is still poorly understood. What is the depositional environment of the Muth Fm. and do the incidentally discovered arthropod trackways support these sedimentary models?

(3) What is the systematic ichnology of the trackways? (in collaboration with Derek Briggs and Simon Braddy, Bristol)

(4) The Muth Fm. is practically devoid of body fossils, therefore its age is just constrained by its stratigraphic position. Is it possible to tighten its age by new conodont dating of beds directly below and above the formation? (in collaboration with John Talent, Sydney and Leo Krystyn, Vienna)

3. STRATIGRAPHY

3.1. General remarks

Since the first geological reference about Spiti by Gerard (1827), this region has been a classical study area for stratigraphy attracting the interest of numerous earth scientists from all over the world. However, the case of deliberate mispositioning, misinterpretation and forging of paleontological data, by V.J. Gupta in the 1970´s and 80´s casted a gloom over the stratigraphic database of this area (Talent *et al.* 1988, Shanker *et al.* 1993). The numerous publications by V.J. Gupta and his associates are not considered here.

Where possible, the stratigraphic nomenclature (Figs. 3.1B, 3.10, 3.11) has been established according to the recommendations of the "International Stratigraphic Guide" (Salvador 1994). The modified terminology is mainly based on the stratigraphic manuals by Holland (1926) Pascoe (1959) and the regional publications of Stoliczka (1866), Hayden (1904, 1908), Srikantia (1981), Bagati (1990) and Bhargava & Bassi (1998) about Spiti, of Nanda & Singh (1977), Baud *et al.* (1984), Gaetani *et al.* (1986) about Zanskar and Ladakh and of Heim & Gansser (1939), Shah & Sinha (1974) and Sinha (1989) about Kumaun.

3.2. Haimanta Group (Griesbach 1891)

In his Pin Valley traverse Stoliczka (1866) used "Babeh series" for sediments above higher grade metamorphic rocks near the Sutlej Valley and below purple conglomerates at the base of his "Muth Series". This definition meets the definition of the present work, except that the present definition also includes the higher metamorphosed metasediments of deeper structural levels. He realized the structure of the Baldar Syncline (Stoliczka 1866, Fig. 1), but regarded the lithologies of this syncline as part of his Babeh series and therefore concluded Silurian age from the fossils found there.

Griesbach (1891) established the term "Haimanta System" for sedimentary sequences in the HH above metamorphosed metasediments of his "Vaikrita System" and below lower Silurian beds and concluded Precambrian to Cambrian age from the stratigraphic position. His definition is strongly confusing, because it is sometimes not consistent throughout his memoir, but in today's terms his Haimanta System seems to include the Phe Fm. and the Shian Fm. The purple conglomerate at the base of his Haimanta System seems to represent the basal conglomerates of the Shian Fm. (Hayden 1904).

Hayden (1904) limited his "Haimanta series" to sediments below the Parahio Fm. and he believed in a Cambrian age for the latter together with the Middle and Upper Haimanta series. Srikantia (1981) included late Precambrian sediments and broadly the Cambrian and Silurian systems of Hayden (1904) in the "Haimanta Group", which he divided into Batal Fm., Kunzam La Fm. and Shian Fm. Bagati (1990) followed this definition, but he paid attention to the Ordovician unconformity and excluded his Thango Fm. (= Shian Fm.) from his Haimanta Group.

In the present work all sediments and metasediments in the Higher Himalaya (except *e.g.* the Jutoghs and the quartzite in the Alaknanda Valley, south of Badrinath) above the "Lower Crystalline Nappe" (see chapter 2.3. for details) and below the Ordovician unconformity are assigned to the Haimanta Group with a stratigraphic range from Neoproterozoic to early Late Cambrian (Figs. 3.1B, 3.11).

This definition takes the continuous relationship of amphibolite grade metasediments at structural lower levels and Precambrian sediments at higher levels into account, with a gradational decrease in metamorphic grade towards higher levels (Griesbach 1891, Hayden 1904, Frank *et al.* 1995). Hayden (1904, p. 9) even states: *"There is therefore no reason to suppose that the schists of the Wangar valley* [between Wangtu bridge and the Babeh Pass] *represent the vaikrita system, for they also are probably only altered representatives of the cambrian slates* [part of the Haimanta Group] *…"*. This implies that the Higher Himalayan Crystalline is not the basement of the Tethyan Zone, but it represents the metamorphosed lower portion of it. Deformation, metamorphism and granitic intrusions obscure this very important fact.

In this view, in spite of the variable throw and related jump in metamorphic grade of normal faults like the Zanskar Shear Zone (Herren 1987), Sangla Detachment (Vannay & Grasemann 1998) and the South Tibetan Detachment Zone far to the East (Burg *et al.* 1984, Burchfield *et al.* 1992) these normal faults still leave the primarily correlation recognizable.

In the Pir Panjal Range stratigraphic lower levels of the Haimanta Group are exposed (Figs. 3.1A, 3.1B). On the base of significant lithological and sedimentological differences in this area, Frank *et al.* (1995) divided the metasediments of the Haimanta Group into three formations (Lower, Middle and Upper Haimantas), *i.e.* the flysch-type Chamba Fm., the glaciomarine Manjir Fm. and the Phe Fm. with turbititic sediments in lower parts grading into tidal flat deposits in upper parts (Garzanti *et al.* 1986, Draganits *et al.* 1998b). Due to the conformable contact of the Phe Fm. and the Parahio Fm. in the Parahio Valley, the Parahio Fm. is also included into the Haimanta Group (Hayden 1904, Fuchs 1982).

The correlation of the glaciomarine Manjir Fm. with the Neoproterozoic diamictites of the Blaini Group in the LH is tempting. As a consequence the Chamba Fm. of the HH represents a correlative of the Simla Slate (Medlicott 1864) in the LH by its stratigraphic position. Comparable lithological successions imply the deposition of the Proterozoic sediments of Higher and Lesser Himalaya in the same basin (Frank *et al.* 1995, Virdi 1995).

In detail the Blaini Group comprises two separate diamictite horizons, of which the upper one shows a cap dolomite succession. In Brookfield´s (1994) correlation he compared the upper diamictite of the Blaini Group, capped by pinkish dolomite, with the Australian Sturtian Glaciation (some 700 Ma ago). From stratigraphic considerations this age is quite high, the exact age is still in vivid discussion, but a broad age between 650-600 Ma is generally accepted in Indian literature (pers. comm. Aninda Mazumdar 2000).

According to Preiss *et al.* (1978), there are two diamictite horizons in Australia, of which the lower one (Sturtian) is capped by dark, laminated shaley dolomites, the upper one (Marinoan) by pinkish carbonate, commonly with macroscopic precipitate textures. Kennedy *et al.* (1998) compares the Sturtian and Marinoan diamictites of Australian with the lower and upper Varangian tillites in Europe, respectively. The correlation of the Manjir Fm. (HH) with the upper diamictites

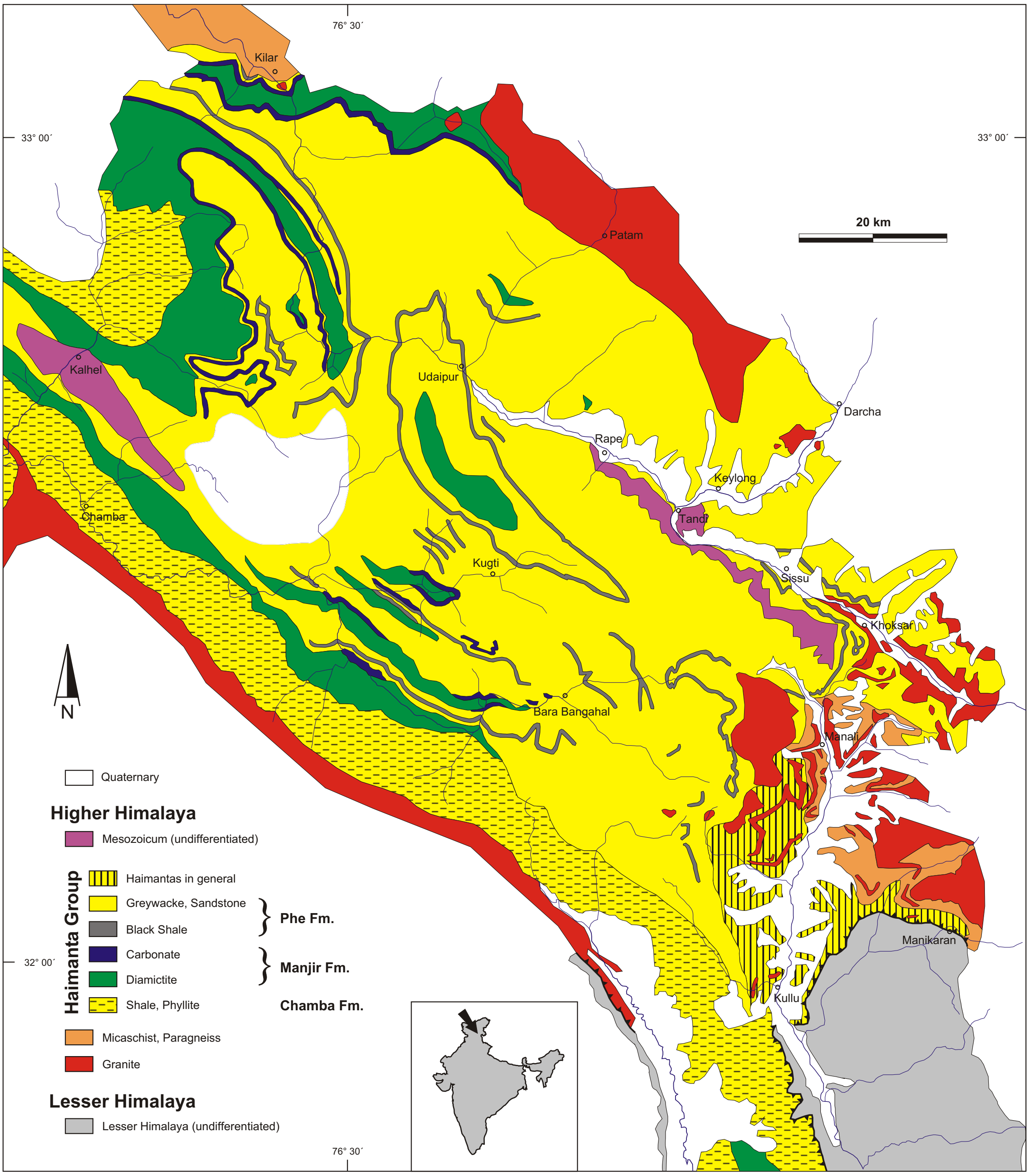


Fig. 3.1A: Geological map of the Chamba-Kullu area, NW Himalaya. The diamictites of the Manjir Fm. and especially the black shale horizon of the Phe Fm. are useful markers for the structural outline of this area. Simplified and modified after Frank *et al.* (1995).

of the Blaini Group (LH) and the Marinoan diamictites in Australia indicates an age around 600 Ma for the Manjir Fm.

The Chamba and the lower Phe Fm. show broadly south-directed sedimentary palaeo-transport (Fig. 3.6 and Grasemann 1993, Fig. 10), in obvious contrast to the more or less N-NE directed paleo-transport in the Simla Slates of the Lesser Himalaya (Valdiya 1970). As a consequence the Chamba Fm. is probably not simply the distal facies equivalent of the Simla Slates, which is also indicated by the general coarser grain-size of the Chamba Fm.

Furthermore, U-Pb and Sm-Nd isotopic studies on samples from equivalents of the Haimanta Group in the HH and samples from the LH revealed different source areas for time-equivalent sediments of the two tectonic units. The age of the majority of detrital zircons in Proterozoic sediments of the HH dated between 0.8-1.0 Ga, in contrast to the 1.87-2.60 Ga old zircons from the LH. In analogue, HH samples gave values of ε_{Nd} between -14.6 and -18.5, but ε_{Nd} -21.4 to -25.9 for LH samples (Parrish & Hodges 1996).

Ar/Ar-dating on detrital white mica on Late Proterozoic sediments in the NW Himalaya indicate differences in the source area of both formations, too. The relatively fine-grained white micas of the Simla Fm. show uniform ages around 860 Ma. In contrast, the coarse-grained (< 2 mm) white micas from the Phe Fm. in the Pin Valley give variable ages scattering between 1280-755 Ma (Grasemann *et al.* 1997). The source area for the clastic sediments in the Simla Slate can probably be identified in the Aravalli-Delhi orogen, while the source area for the Haimanta Group might be a yet unallocated Gondwana fragment in the North of the basin, possibly the Lhasa Terrane (Metcalfe 1999). The pre-thrusting, paleogeographic relationship between the Higher Himalayan and Lesser Himalayan sequences remains poorly understood.

The age of these sediments devoid of any body fossils is not well constraint; due to the stratigraphic position and regional correlation a Late Proterozoic age is probable (Frank *et al.* 1995). The absolutely oldest age limit of the Haimanta Group is constraint by depleted mantle Nd model ages in the range of 1.7-2.2 Ga (Miller *et al.* in prep.). These Nd model ages clearly represent the absolutely oldest age limits. Older age limits that are obviously closer to the real sedimentation age of the Haimanta Group are indicated by the above-mentioned Ar/Ar ages on detrital micas between 1280-755 Ma (Grasemann *et al*. 1997).

The younger age limit is constraint mainly by trilobite faunas, which gave Early to Middle Cambrian age for the uppermost parts of the Phe Fm., just below the dolomitic Parahio Formation, in the Parahio Valley (Hayden 1904) and early Late Cambrian ages for the Kurgiakh Fm. in Zanskar (Hughes & Jell 1999).

The different amount of erosion below the Ordovician unconformity, with decreasing erosion towards NW, results in the situation that no Cambrian carbonates are preserved in the Pin Valley, the dolomites of the Parahio Fm. reach only some 90 m in the Parahio Valley, but more than 1000 m sediments with carbonates are preserved in Zanskar (Hayden 1904, Fuchs 1982, Gaetani *et al.* 1986). Summarizing the observations in the Pir Panjal (Frank *et al.* 1995), the Parahio Valley (Hayden 1904) and Zanskar (Baud *et al.* 1984, Gaetani *et al.* 1986 and Garzanti *et al.* 1986), according to the above-mentioned definition of the Group ("sediments of the Higher Himalaya below the Ordovician unconformity") the Haimanta Group can be divided into 5 formations (from bottom to top): CHAMBA FORMATION (Rattan 1973), MANJIR FORMATION (Sehgal 1966), PHE

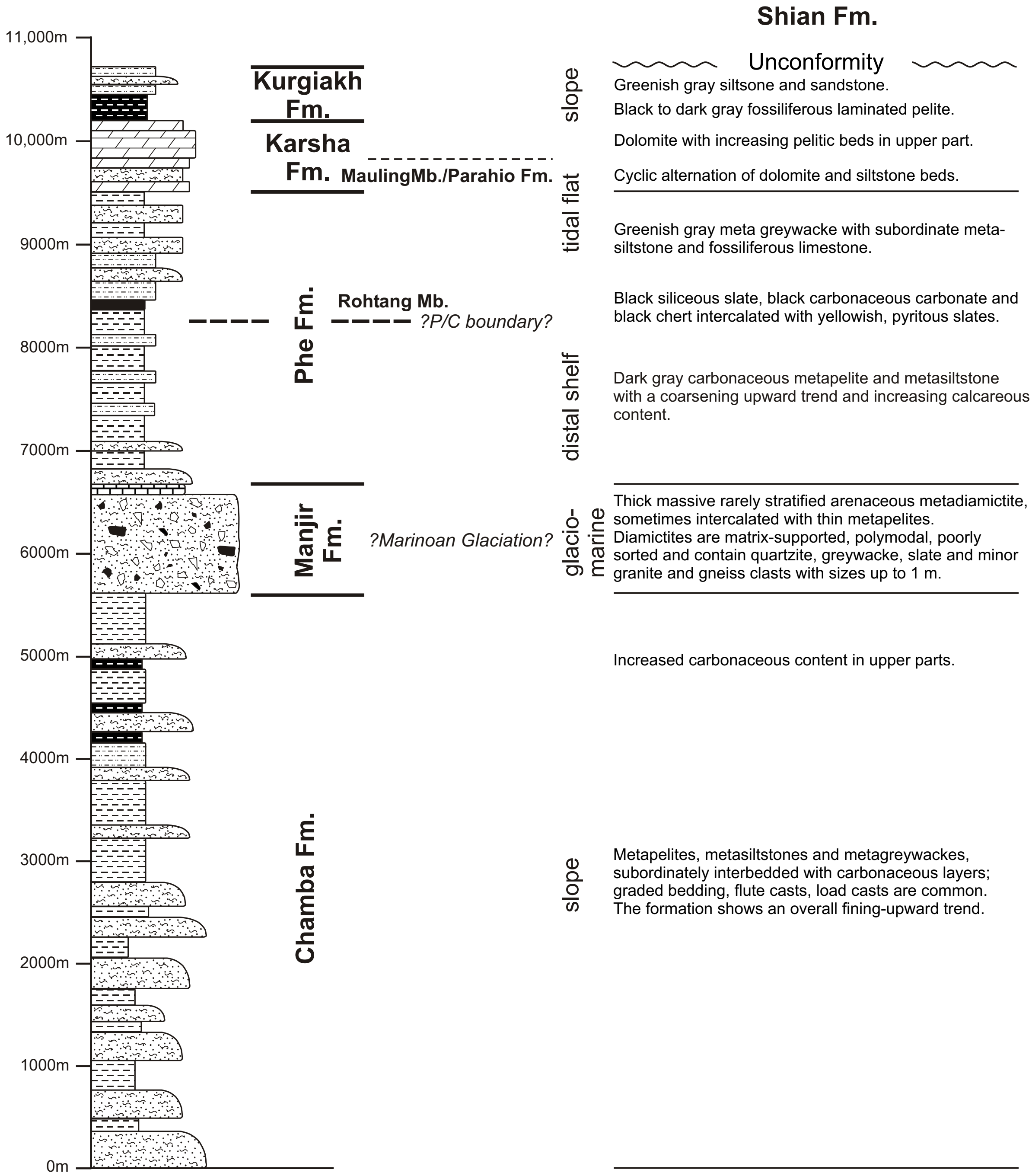


Fig. 3.1B: Composite section of the complete Haimanta Group from outcrops in the Pir Panjal Range, Zanskar and Spiti based on Hayden (1904), Rattan (1973), Garzanti *et al.* (1986), Frank *et al.* (1995) and own observations.

Formation (Nanda & Singh 1977), Karsha Fm. (Nanda & Singh 1977) and Kurgiakh Fm. (Garzanti *et al.* 1986).

The Parahio Fm. represents a correlative of the Karsha Fm./Mauling Mb. (Garzanti *et al.* 1986), but as long as detailed sections have not proofed the exact relation of these two Formations, it is preferred here to describe the Parahio Fm. separately.

	Pir Panjal	Zanskar	Spiti
Haimanta Group	–	Kurgiakh Fm.	–
	–	Karsha Fm.	Parahio Fm.
	Phe Fm.	Phe Fm.	Phe Fm.
	Manjir Fm.	–	–
	Chamba Fm.	–	–

Tab. 3.1: Formations of the Haimanta Group and their regional occurrence.

The Formations of the Haimanta Group are dealed with here in a cursory way, the reader is referred to the detailed descriptions of the Chamba, Manjir and Phe Fms. by Rattan (1973, 1984), Garzanti *et al.* (1986), Grasemann (1993) and Frank *et al.* (1995) and to Garzanti *et al.* (1986) about the Karsha and Kurgiakh Fm.

The **Chamba Formation** comprises the basal part of the Haimanta Group (Figs. 3.1A, 3.1B), above the “Lower Crystalline Nappe” (see chapter 2.3. for details) and below the first occurrence of diamictites of the Manjir Fm. and has been named after the old Maharaja township of Chamba on the Ravi River in the Pir Panjal Range. Towards lower stratigraphic levels the metamorphic grade gradually increases up to lower amphibolite facies. Rattan (1973) has delineated the lower part of the Chamba Formation to form the Bhalai Fm., but as for the “imperceptible gradation between the constituent rocks” (Rattan 1984) they are regarded as one formation by Frank *et al.* (1995). The outcrop of this Formation is evident in the map by Frank *et al.* (1995).

The Chamba Fm. consists mainly of well-bedded monotonous alternations of grayish green to dark gray metapelites, metasiltstones and metagreywackes, the thickness may reach more than 6000 m (Rattan 1984, Frank *et al.* 1995). A general trend of decreasing grain-size associated with an increase of carbonaceous content can be observed throughout the Formation (Rattan 1984). Sedimentary structures like graded bedding, load casts, flute casts and groove casts probably indicate a turbitite type depositional environment (Rattan 1973, Frank *et al.* 1995). The Formation is totally devoid of fossils, but from the stratigraphic position below the Manjir Fm. a correlation with the Simla Slates of the Lesser Himalaya is possible, thus indicating a late Proterozoic age.

Attention has to be paid to the fact, that in the Gorbiachin Valley (Igarka District, Sibiria), the term “Chamba Formation” is also in use for early Silurian shale/limestone series (Johnson *et al.* 1997), but as far as Russian references have been available to the author, the use of this formation name seems to be later there.

The **Manjir Formation** delineates sediments above the Chamba Fm. and below the Phe Fm., its uppermost part is represented by a thin carbonate horizon (Figs. 3.1A, 3.1B). The Formation

Fig. 3.2: Top of the Manjir Fm. in the upper Ravi Valley near Sind. *C.* 10 m thick, white carbonate follows directly above diamictite beds, of which the uppermost part contains large carbonate clasts.

Fig. 3.3: Carbonaceous limestone on top of the Manjir Fm. in the Deda Valley, N of Sach Pass. ?Algal laminations deformed into folds, parallel to the stretching lineation. 50 Paisa coin is 23 mm in diameter.

Fig. 3.4: ?Algal lamination showing strange interference of growth and deformation structures. 50 Paisa coin is 23 mm in diameter.

Fig. 3.5: Concentric ?algal structures with black carbonaceous-rich aggregates (arrow). 50 Paisa coin is 23 mm in diameter.

was named after the village of Manjir in the Siul Valley in the Pir Panjal Range and is easily distinguishable from the formations below and above by its contents of big subangular-rounded clasts. Due to Himalayan deformation the contact to the underlying Chamba Fm. is unclear, Rattan (1984) describes a conformable contact.

The Formation reaches a thickness of some 1000 m and characteristically consists of thick, massive, rarely stratified, matrix-supported diamictites; the matrix is similar to the greywackes of the Chamba Fm. and also contains characteristical violet quartz grains (probably of volcanic origin). Metasilt-, metasandstones, metagreywackes and metapelites are commonly intercalated. Bedding is usually poorly developed. The subangular to rounded clasts with sizes up to 1 m are dispersed in a chaotic manner and mainly comprise white quartzites, vein quartz, greywackes, slates, rarely carbonates and occasionally granites and gneisses. The clast to matrix ratio is generally very low, but may be relatively high in some places (*e.g.* upper Ravi Valley near Sind). Striated clasts have been observed very rarely. No detailed sections across the Manjir Fm. exist and the stratigraphy is complicated by Himalayan deformation, therefore it remains a question for further investigation, whether there exist two separate diamictite horizons in analogue to the LH or not.

Due to its distinct lithology, sandwiched between the relatively similar Chamba - and Phe Fm., the Manjir Fm. represents an important marker horizon within these unfossiliferous, monotonous Proterozoic sequences of the HH. The Formation has been mapped from Kashmir (Raina *et al.* 1990) to the Beas Valley near Mandi (Frank *et al.* 1995), it is thickly developed in the Sutlej Valley (pers. comm. Wolfgang Frank 1999) and even can probably be correlated with the Nanda Kot Cgl. in the Higher Himalaya of Kumaon (Heim & Gansser 1939). The depositional environment of these sediments is probably glacio-marine and might be correlated with the Blaini Fm. in the Lesser Himalaya (Shanker *et al.* 1989, Frank *et al.* 1995).

The gray carbonate horizon with a thickness less than 100 m at the top of the Formation succeeds directly or some tens of meters above the diamictites. The carbonates usually are strongly recristallized, coarse-grained magnesite is common. In some places calcite and/or dolomite is still preserved, *e.g.* in the Deda Valley (north side of Sach Pass); there the carbonate is represented by deformed gray carbonaceous limestone. In less deformed parts, concentric carbonaceous aggregates of unclear origin and stromatolitic structures are preserved (Fig. 3.3-5). Near the village Sind in the upper Ravi Valley the carbonate directly follows on top of coarse-grained diamictite/conglomerate (Fig. 3.2). There the intimate facies relationship of the two lithologies is shown by the existence of carbonate clasts of the succeeding carbonate horizon in the uppermost part of the conglomerate. The carbonate clasts are much larger and less rounded than clasts of other lithologies, thus indicating less transport distance.

Knoll (2000) mentioned the sharp boundary between several Marinoan diamictites and succeeding cap carbonates reported in the literature, implying rapid de-glaciation, possibly representing a level of syncroneity on a continental scale. Thus he recommended separating the cap carbonates from the diamictites in lithostratigraphic sections. Because of the above mentioned intimate relationship of diamictites and succeeding carbonates, with reworked carbonate clasts in-between the carbonates have been included into the Manjir Fm.

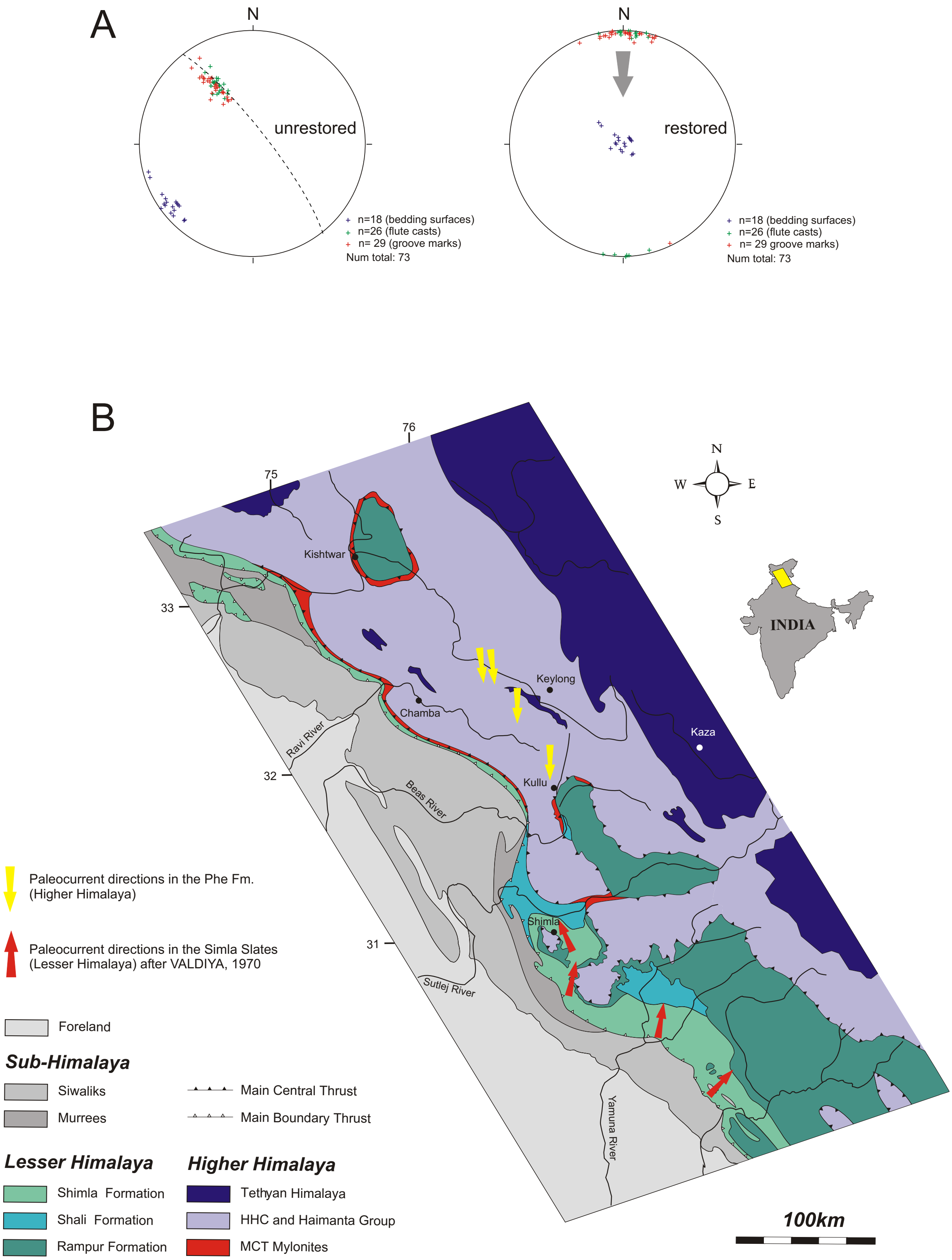


Fig. 3.6: Comparison of paleocurrent directions of late Proterozoic deposits in the Lesser Himalaya and Higher Himalaya. **A)** Stereonet of flute casts and groove marks with south-directed paleocurrent indications in the lower Phe Fm. (HH) in the Chenap Valley near Udaipur. Equal area lower hemisphere projection. **B)** Simplified and modified tectonic map of the NW Himalaya (courtesy Bernhard Grasemann) with yellow arrows indicating paleocurrent directions in the Phe Fm. of the HH after Grasemann (1993) and own data and paleocurrent directions in the Simla Fm. (red arrows) of the LH after Valdiya (1970).

Although the grainsize is slightly coarser than in the Chamba Fm. no diamictites have been found in the Phe Fm. above the carbonate horizon. Similar carbonates directly on top of late Proterozoic glacial-marine diamictites (“cap-carbonates”) have been described from several places in the world and seem to represent characteristic features for these successions (Fairchild 1993, Brookfield 1994). Kirschvink (1992), based on world heat balance models by Budyko (1963) proposed a “Snowball Earth” theory to explain these Proterozoic low-latitude glacial sediments. Hoffman *et al.* (1998) investigated the peculiar trends of the Carbon isotopes in these glaciation related carbonates.

3.2.1. PHE FORMATION (Nanda & Singh 1977)

CORRELATION: **Kashmir:** Dogra slates (Wadia 1934), Marinag Fm. and Lolab Fms. (Shah 1968), Machhal Fm. (Raina & Razdan 1975), Shumahal Fm. (Srikantia & Bhargava 1983), Khaiyar Fm. (Kumar *et al.* 1987); **Pir Panjal:** Lower Salooni Fm. (Rattan 1973), Upper Haimanta (Grasemann 1993, Frank *et al.* 1995) **Zanskar:** Suru Fm./Darcha Mb., Phe Fm. and lowest part of Karsha Fm. (Nanda & Singh 1977); **Spiti:** Babeh series/beds α and β (Stoliczka 1866), Haimanta system/division 2 and 3 (Griesbach 1891), Cambrian slates and quartzites, Cambrian slates and Cambrian trilobite beds (Hayden 1904), Batal Fm. and Kunzam La Fm./Mbs. A-D (Srikantia 1981), Batal and Kunzam La Fm./Debsa Khad Mb. (Kumar *et al.* 1984); **Kinnaur:** Hilap Fm. (Bassi *et al.* 1983), Batal Fm. and lower Kunzam La Fm. (Bassi 1989); **Kumaon:** Azoic slates (Strachey 1851), Martoli phyllite (Heim & Gansser 1939), Martoli Fm. (Shah & Sinha 1974), Martoli Group (Sinha 1989).

TYPE SECTION: Due to the enormous thickness of the Phe Fm., it is impossible to define a single type section, it is even not possible to establish a single type locality. The lower levels of the Formation are well exposed at the road section between Udaipur and Tindi in the Chenab Valley and on the trails from that road to the passes of the Pir Panjal Range to the South (Fig. 3.1A, Frank *et al.* 1995). The middle and upper levels of the Formation are well exposed in the Tsarap Lingti Valley SE of Padam (Fuchs 1987) and further towards SE the Kurgiakh Valley (Gaetani *et al.* 1986) as well as in the Parahio and Debsa Khad Valleys in Spiti (Hayden 1904, Fuchs 1982).

NAME & BOUNDARIES: The Phe Fm. was named after the village Phe in the Doda Valley (Zanskar, NW of Padam) and comprises sediments above the carbonate horizon on top of the Manjir Fm. and below the first appearance of yellowish weathering dolomite of the lower Karsha Fm. (Parahio Fm.).

AGE: As mentioned earlier, the lower part of the Phe Fm. is totally devoid of fossils, therefore the older age limit of the Formation can just be deduced from its stratigraphic position. Frank *et al.* (1995) correlated the Manjir Fm. below with the glaciomarine Blaini Fm. in the Lesser Himalaya. The Blaini Fm. comprises two diamictite horizons separated by sandstone and shale, the upper horizon shows a characteristic cap dolomite on top. Brookfield (1994) correlated the upper diamictite of the Blaini Fm. with other Sinian diamictites in Asia and Australia and concluded an age around 700 Ma (Sturtian Glaciation). According to more recent literature (*e.g.* Kennedy *et al.* 1998, Knoll 2000) a correlation with the Marinoan Glaciation (around 600 Ma) seems more probably. If the correlation between Manjir Fm. and Blaini Fm. is correct, 600 Ma will represent the older age limit for the Phe Fm. although the exact age of the Blaini Fm. is still in discussion with ages ranging between 600-650 Ma (pers. comm. Aninda Mazumdar).

In the Parahio Valley the higher levels of the Phe Fm. yielded rich trilobite faunas, which have been investigated by Reed (1910) and re-investigated by Jell & Hughes (1997). According to the latter the youngest of these trilobites indicate middle Middle Cambrian age. Parcha (1998) even mentions early Late Cambrian trilobites from the Parahio Valley, but their exact position in the section is not clear. At the right riverbank of the Parahio River, oblique opposite of Thango (Fig. 2.6, 4.1), in a thick horizon of light gray, thick-bedded, well-laminated sandstones (Fig. 4.3) with parting lineation, some 600 m below the Ordovician unconformity, a well preserved large *Diplichnites* trackway has been found, again indicating that these sandstones are situated well within the Cambrian already (Fig. 3.7).

At the same riverbank, directly opposite of Thango, some 70 m stratigraphic above this *Diplichnites* trackway, phosphoritic, bioclastic limestone with wavy bedding surfaces, intercalated with dark gray shales (Fig. 4.3), contains abundant trilobites, brachiopods and crinoids. The occurrence of rare protoconodonts additionally indicates middle Cambrian age or younger (Bhatt & Kumar 1980, pers. comm. John Talent 1999).

The situation in Zanskar is more difficult to deduce, because most of the trilobites originate from the upper Karsha Fm. and Kurgiakh Fm. and gave late Middle Cambrian age (Whittington 1986, Jell & Hughes 1997). From the stratigraphic position below the Karsha Fm. and additional information from trace fossils (Bhargava *et al.* 1982, Kumar *et al.* 1984, Bhargava *et al.* 1986, Bhargava & Srikantia 1985, Hughes and Droser 1992), a middle Middle Cambrian age for the top of the Phe Fm. can be inferred.

DESCRIPTION: The Formation encompasses alternating metagreywacke, metasiltstone and metapelite. The contact to the Manjir Fm. below is more or less gradational and also the lithologies are quite similar, but the Phe Fm. lacks large clasts that are characteristic for the Manjir Fm.

As a general tendency the Formation shows a shallowing and coarsening upward trend, associated with a decrease in the carbonaceous content and an increase in the carbonate influence (Garzanti *et al.* 1986, Frank *et al.* 1995, Bhargava & Bassi 1998). In lower stratigraphic levels, argillaceous sediments dominate over arenaceous beds, in some parts of the outcrop (*e.g.* in the upper Debsa Khad), arenaceous deposits seem to exist below the shale unit (Hayden 1904). Monotonous alternations of greywackes and shales associated with common sedimentary structures like graded bedding, load casts, flute casts, chevron marks, groove marks, *etc*. indicate a turbitite type depositional environment. Very well preserved structures of this kind have been found near Udaipur and Tindi in the Chenab Valley. Paleocurrent directions of the lower Phe Fm. defined by flute casts and chevron marks scatter between SW-SE directions, with a dominant direction towards the South (Fig. 3.6).

The high carbon content of the lower Phe Fm., indicating restricted depositional conditions, has been mentioned by several authors from different areas of its outcrop, *e.g.* Pir Panjal (Rattan 1973), Zanskar (Nanda & Singh 1977), Spiti (Bhargava & Bassi 1998), Kinnaur (Bassi 1989) and Kumaon (Kumar *et al.* 1977). One dark carbonaceous horizon turned out to be of regional

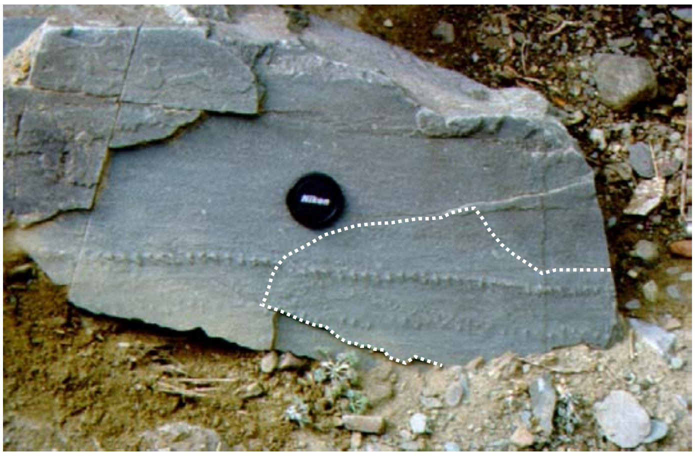


Fig. 3.7A: Large *Diplichnites gouldi* trackway on the bedding surface of a parallel laminated bed, running slightly oblique to the parting lineation. Opposed series of probably 5 tracks are are V-like arranged, the open part of the V pointing in walking direction (from left to right). Dotted white line shows margin of the sketch below. Lens cap diameter is 53 mm.

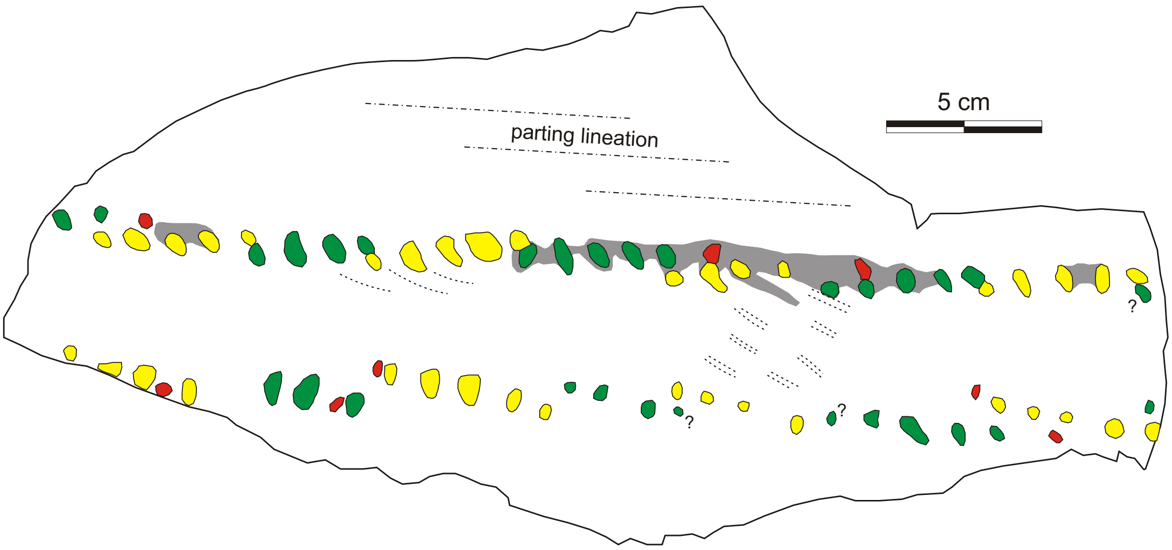


Fig. 3.7B: Sketch from a hand specimen (sample ED98/226) from the right part of the large *Diplichnites gouldi* trackway above. Tracks are arranged in series of 5 tracks. Series show opposite symmetry, sets of series are V-shaped diverging in walking direction (towards the right) and overlap by one track or less. Sets of series show the same color. Tracks with unclear position concerning series are colored red. Slightly raised surface is shaded gray. Dotted lines indicate scratches.

importance and can be traced from the Sach Pass in the West to Kinnaur in the East (Frank *et al.* 1995, Wyss *et al.* 1999, Bhargava & Bassi 1998, Bassi 1989) and is of great help for the recognition of large-scale structures within the otherwise monotonous series (Figs. 3.1A, 3.1B and Grasemann 1993, Frank *et al.* 1995).

The horizon varies in thickness between 10-50 m and consists of black siliceous slates, black carbonaceous carbonates and black cherts intercalated with yellowish, pyritous slates. The horizon can readily be seen from distance on the north side of the Rohtang Pass looking towards the North to the first mountain scenery and near Udaipur in the Chenab Valley. Easy accessible outcrops at road sections, where the lithologies can be studied in detail are for example Dhakog in the Ravi Valley (*c.* 3 km W of Kharamuk) and between Chhota Dara and Batal in the Chandra Valley, the latter being greenschist grade metamorphosed (Wyss *et al.* 1999). Draganits *et al.* (1998b) suggested the name Rohtang Member for this black marker horizon within the Phe Fm.

As a peculiarity the above-mentioned outcrop in the Chandra Valley showed a *c*. 20 cm thick gray/white laminated, fine-grained bed with pyrite containing more than 20 wt.% BaO in XRF whole rock analyses. As the Sulfur content is quite low, it seems that the Barium content is constituted by Ba-silicates and not by Barite. Barium *Lagerstätten* are one of the rare mineralizations in the HH usually found in Neoproterozoic to Lower Paleozoic rocks; most of these occurrences constitute crosscutting Barite veins, conformal beds or Ba-silicates are hardly described (Sinha 1989, Bhargava & Bassi 1998).

Towards higher stratigraphic levels the depositional setting gradually becomes shallower and arenaceous sediments dominate over argilliceous ones. Sedimentary structures like bi-directional cross-bedding, lenticular bedding, symmetrical wave ripples, interference ripples, rill marks and desiccation cracks (Kanwar & Ahluwalia 1978) point a subtidal to intertidal depositional environment on a tidal flat. Paleocurrent directions derived from cross-bedding and current ripples trend ESE-WNW to NNE-SSW (Garzanti *et al.* 1986).

In the Parahio Valley the gradual increase in grain size and decrease in carbon content in the upper part of the Phe Fm. (*i.e.* Kunzam La Fm./Debsa Gad Mb. Kumar *et al.* 1984) is also accompanied by the occurrence of a diverse ichnofauna (Bhargava *et al.* 1982, Kumar *et al.* 1984, Bhargava *et al.* 1986, Bhargava & Srikantia 1985) indicative for Lower Cambrian (Crimes 1989). Due to the occurrence of trilobites, Hayden (1904), who also gives the best lithological description of this part, refers this uppermost part of the Phe Fm. as “Cambrian trilobite beds”. This level of the Phe Fm. shows regular coarsening upward cycles, with shale grading into siltstone and finally sandstone, which are well exposed in the Parahio Valley near Thango. These cycles sometimes end with gray, dark brown weathering limestone (Fig. 4.3), sometimes extremely rich in trilobites, brachiopods and very rare protoconodonts.

DISCUSSION: The enormous thickness, large outcrop, relatively inconspicuous lithologies and the lack of fossils in lower parts resulted in different subdivisions and stratigraphic terminology of the Formation. A detailed discussion is above the scope of this chapter, but the Kunzam La Fm. should be discussed here. Srikantia (1981) separated the arenaceous sediments of the upper part of the Haimanta Group (“Kunzam La Fm.”) from finer-grained, darker sediments below (“Batal Fm.”). Not all, but most of the Indian literature since then has adopted these terms. Kumar *et al.*

Fig. 3.8: Parahio Fm. in the Parahio Valley. View from Thango towards SSE. Black arrows indicate the gradual contact to the Phe Fm. below and the angular unconformity at the base of the Shian Fm. above. Asterisks indicate four distinct cycles starting with dolomite horizons grading into siltstone and pelite.

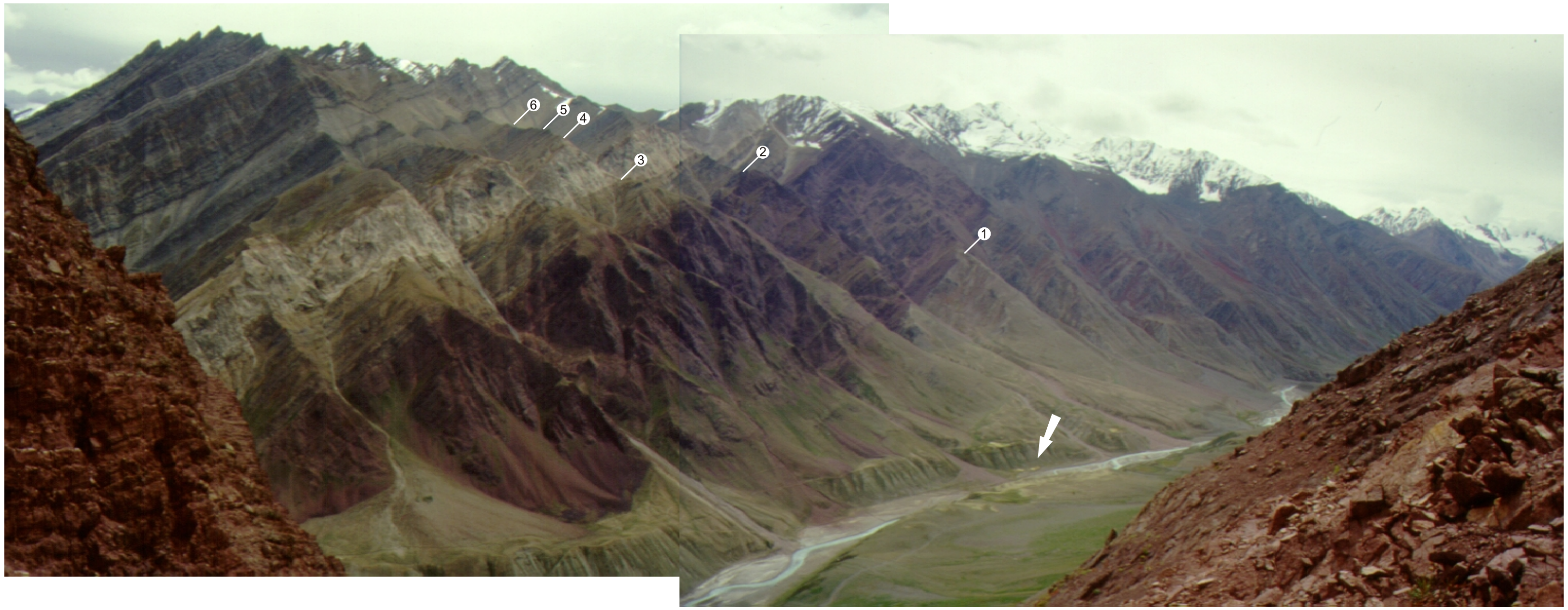


Fig. 3.9: Late Proterozoic to Triassic stratigraphy in the Pin Valley. Gorgeous view from a ridge south of Muth towards SE (Photo courtesy G Wiesmayr). Arrow indicates the position of Shian. Labeled pins indicate boundaries between the different Formations. (1) angular unconformity between Phe Fm. and Shian Fm. (2) gradual but sharp boundary between Shian Fm. and Pin Fm. (3) disconformity between Pin Fm. and Muth Fm. (4) gradual contact of Muth Fm. and Lipak Fm. (5) angular unconformity between Lipak Fm. and Guling Group. (6) Paleozoic-Mesozoic boundary (Kuling Group Tamba Kurkur Fm.).

(1984) divided the Kunzam La Fm. in a lower Debsa Khad Mb. and an upper Parahio Mb. (broadly correlating with Hayden´s (1904) Cambrian trilobite beds and Cambrian dolomite). A stratigraphic division of this part of the thick Haimanta Group is desirable, but the above mentioned units fail to really show clear boundaries that are recognizable in the field and thus rectify the definition of lithostratigraphic units.

The lithological difference between Batal Fm. and the Debsa Khad Mb. of the Kunzam La. Fm. is evident, but very gradual and thus not possible to trace in the field without detailed sections. Additionally the term “Debsa Khad Mb.” is somehow misleading, because this member touches the Debsa Valley just in its westernmost part, in the rest of the valley the Batal Fm. crops out. It is preferred here to use no lithological subdivisions as long as they are not well defined. Therefore the Phe Fm., as used here, includes the Batal Fm. and the Debsa Khad Mb. of the Kunzam La Fm. Further detailed research will show where to set the boundaries. Due to the striking contrast of its yellowish weathering dolomites, the Parahio Fm. is separated from the Phe Fm. below (Fig. 3.8).

Garzanti *et al*. (1986) suggest a shallow tidal flat depositional environment for the upper Phe Fm. in Zanskar. In the Parahio Valley sedimentary structures like parting lineation (Fig. 3.7) also support very shallow water depths, desiccation cracks even indicate temporarily emergent conditions.

3.2.2. PARAHIO FORMATION (Pascoe 1959)

CORRELATION: **Kashmir:** Upper fossiliferous clays, shales and limestones (Wadia 1934), Nutunus and Trahagam Fms. (Shah 1968), Manjhar Fm. (Raina & Razdan 1975), Rangmal Fm. (Srikantia & Bhargava 1983), lower part Karihul Fm. (Kumar *et al.* 1987); **Zanskar:** Karsha Fm./Mauling Mb. (Nanda & Singh 1977, Gaetani *et al.* 1986); **Spiti:** Cambrian dolomite Hayden (1904), Parahio series (Pascoe 1959), Kunzam La Fm./Mb. E (Srikantia 1981), Kunzam La Fm./Parahio Mb. (Kumar *et al.* 1984); **Kinnaur:** Kunzam La Fm./Suti Mb. (Bassi 1989).

TYPE SECTION: Above Maopo campground, Parahio Valley, Spiti (Hayden 1904). Steep mountain slope on the left riverbank directly at the confluence of Khamengar and Parahio Rivers.

NAME & BOUNDARIES: The Formation was named after the type section in the Parahio Valley. Lower boundary is here defined at the first occurrence of gray, yellowish weathering dolomite (Fig. 3.8). The upper boundary is well defined by the sharp lithological break at the unconformity below the basal conglomerate of the Shian Fm. The so defined Formation corresponds exactly to Hayden´s (1904) “Cambrian dolomite”.

AGE: Except some poorly preserved trilobite fragments, hardly any fossils have been found in the Parahio Fm. as defined here (Hayden 1904). A middle Middle Cambrian older age limit of the Formation can be deciphered from trilobites, brachiopods and trace fossils from the uppermost Phe Fm. below (for details and references see the text about the age of the Phe Fm. above).

The younger age limit is even more difficult to define, as the overlying Shian Fm. is principally devoid of fossils in its lower levels. Bagati (1990) mentions trilobite casts without any

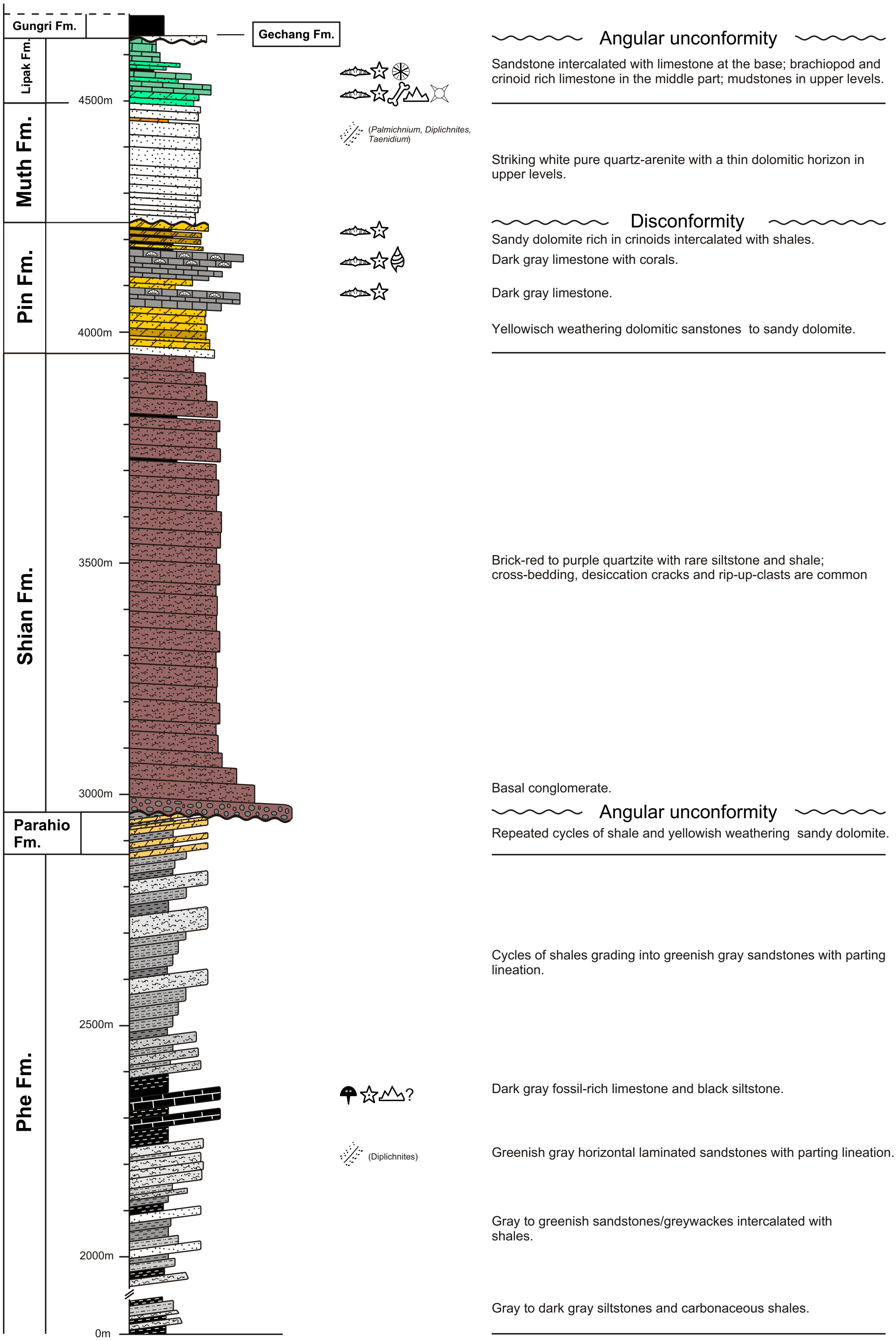


Fig. 3.10: Composite section of the Late Proterozoic and Paleozoic sediments (without the Permian Kuling Group) in the Pin Valley and Parahio Valley. Thickness of the Pin, Muth and Lipak Fms. have been measured, other thickness' have been taken from literature. Note the angular unconformities at the base of the Shian Fm. and Kuling Group.

further details. The Early Ordovician transgression is a common feature at many Gondwana margins (Stampfli *et al.* 1991, Brookfield 1993). The possible correlation of the transgression at the base of the Shian Fm. in Spiti with the similar transgression of the Yarkhun Fm. in the Karakorum, which was dated by acritarchs to Arenigian age (Le Fort *et al.* 1994, Amerise *et al*. 1998), represents the theoretical younger age limit for the Parahio Fm. The probable correlation with the Mauling Mb. of the Karsha Fm. in Zanskar would imply a late Early to early Middle Cambrian age for the Formation (Gaetani *et al.* 1986).

DESCRIPTION: A valuable lithological description of the section can be found in Hayden (1904, p. 14-15). The Formation is quite similar to the uppermost part of the Phe Fm. in showing several relative regular cycles with shale grading into siltstone and finally into sandstone with carbonate on top, but it is differentiated from the latter by the occurrence of yellowish-rusty weathering dolomite (Fig. 3.8). Sandstone are very well laminated and commonly show parting lineation on the bedding surfaces. Towards higher stratigraphic levels the dolomite horizons increase in frequency and thickness till they become the predominant lithology. This observation closely resembles the situation in the Karsha Fm. in Zanskar, where the increase in frequency and thickness of dolomite horizons in the Mauling Mb. towards the Thidsi Mb. is described by Garzanti *et al.* (1986).

DISCUSSION: Pascoe (1959) included the Cambrian trilobite beds and the Cambrian dolomite beds of Hayden (1904) in his Parahio series. The present author has modified this definition, because of the difficulties to draw a clear boundary between the cambrian trilobite beds and the lithologies below (Ranga Rao *et al.* 1982), although there is a gradual decrease in grain size and increase in carbon content noticeable towards stratigraphic lower levels. The definition of the lower boundary of the Parahio Fm. as used here (first occurrence of dolomite) has the advantage of being readily recognizable in the field and it is the same that has been used by Fuchs (1982). A valid alternative could represent the use of the first occurrence of true carbonate that would shift the boundary downwards. However, a clear definition clearly requires further investigations and detailed sections in this area.

A glance on the geological maps of Spiti (Hayden 1904, Fuchs 1982, Bhargava & Bassi 1998) clearly shows the different extent of erosion of the Cambrian below the Ordovician unconformity, with the trend of erosion to deeper levels in the SE part and decreasing erosion towards the NW. For example there are no beds of the Cambrian trilobite beds (Hayden 1904) preserved in the Thanam Valley (NW Kinnaur), in the Pin Valley some beds of this level occur and the Parahio Valley shows even higher levels of the Cambrian. Cambrian beds further to the NW in Zanskar reach a thickness of more than 1200 m (Garzanti *et al.* 1986). The observation that the fold axes of the Pre-Himalayan deformation plunge slightly towards NW might be an explanation for the observation that more of the Cambrian beds are preserved in this part (Wiesmayr 2000).

Garzanti *et al.* 1986 concluded a tidal flat depositional environment for correlatives of the Parahio Fm. in Zanskar.

3.3. Shian Formation (Goel & Nair 1977)

CORRELATION: **Kashmir:** Rishkobal Fm./Mb. A (Srikantia & Bhargava 1983), upper part of Karihul Fm. (Kumar *et al.* 1987); **Zanskar:** Thaple Fm. (Nanda & Singh 1977); **Spiti:** Muth series/Mb. A (Stoliczka 1866), Haimanta system/division 1 (Griesbach 1891), Lower silurian conglomerate and Lower silurian red quartzite Hayden (1904), Thango Fm. (Srikantia 1981); **Kinnaur:** Yamrang La Fm. and Tiwri Fm. (Bassi *et al.* 1983), Thango Fm. (Bassi 1989); **Kumaon:** Conglomerate (Griesbach 1891), Ralam Cgl. and Garbyang Fm. (Heim & Gansser 1939).

TYPE SECTION: Mountain slope east of Shian on the right side of the Pin River (Pin Valley, Spiti). The type section is shown in Fig. 3.9.

NAME & BOUNDARIES: Goel & Nair (1977) named the Formation after the seasonal used shepherds place "Shian" in the Pin Valley (Fig. 3.9). Srikantia (1974) introduced "Thango Fm." after a small village in the Parahio Valley, but in an unpublished survey report, and following the recommendations by Salvador (1994) this reference is not considered here. The first published papers using "Thango Fm." are Srikantia (1978) and Srikantia (1981), except a short reference in Srikantia (1977). In the subsequent years various authors have been using both terms: Thango Fm. by Kumar *et al.* (1983), Bhargava & Bassi (1998); Shian Fm. by Ranga Rao *et al.* (1982), Bagati (1990), Bagati *et al.* (1991), Parcha (1998).

Both terms have been published in the same year, but in the author's opinion, there are several arguments favoring the use of Shian Fm. In the Parahio Valley the Formation is folded and faulted (Fuchs 1982), therefore stratigraphic investigations are limited. The section near Shian in the Pin Valley is easy to access and the section is located at the limb of a large-scale fold, therefore deformation is nearly neclegtable (Figs. 2.11, 3.9). An additional advantage is the concentration of three type sections (*i.e.* Shian Fm., Pin Fm. and Muth Fm.) in one single continuous section within an area of less than 3 km in the Pin Valley, thus facilitating further investigations.

In the Pin Valley at Shian the boundaries of the Formation are relatively clear and well visible from distance (Fig. 3.9). The lower boundary is represented by the sharp lithological break at the angular unconformity comprising purple conglomerate above greenish gray sandstone/siltstone. The upper boundary is delineated at a relatively sharp, but conformable contact of thin-bedded purple siltstone to fine-grained sandstone of the Shian Fm. overlain by thick-bedded, light gray quartzites of the basal Pin Fm. followed by characteristic beige to brown weathering dolomitic sandstone (Fig. 3.9). This definition corresponds exactly with Hayden´s (1904) "Lower Silurian".

AGE: Besides some badly preserved trilobite casts mentioned by Bagati (1990), the Shian Fm. is nearly devoid of any fossils. The correlation of the transgression event at the base of the Shian Fm. in Spiti with the similar transgression of the Yarkhun Fm. in the Karakorum dated by acritarchs (Le Fort *et al.* 1994, Amerise *et al*. 1998) points to a possible Arenigian age of the basal Shian Fm.

The fauna of the lower limestone horizon of the Pin Fm. (*i.e.* Hayden´s (1904) beds 2-3 of the Silurian limestone) probably represents Caradocian age (Hayden 1904 and Reed 1912), thus indicating a Caradocian younger age limit for the Shian Fm.

DESCRIPTION: The thick, purple to brick red Shian Fm. forms a striking horizon within the Tethyan sediments from Zanskar to Kumaon. The basal conglomerate forms a conspicuous horizon that is well developed in the Parahio Valley, in the Pin Valley near Muth and Baldar and at the Baralacha La (Kanwar & Ahluwalia 1978), with variable thickness between some meters to 30 m. The conglomerate is clast-supported with sandy matrix, very similar to the arenites above. Bedding is weakly developed. Clast sizes reach some 25 cm, in rare cases more and show angular to rounded shapes. The internal fabric is more or less chaotic, no imbrication has been observed. In lower levels of the conglomerate, clasts derived from the underlying dolomite of the Parahio Fm. are very abundant, but become rarer into higher levels and are more and more replaced by quartzites and vein quartz. Where dolomitic clasts of the underlying Parahio occur, they are usually larger and more angular than clasts of other lithologies. Ranga Rao *et al.* (1982) mention the occurrence of rare granitic gneiss clasts. Near Thango two conglomerate horizons are separated by a thin sandstone horizon, the lower conglomerate showing many dolomite clasts, the upper one hardly any (Hayden 1904).

The upper part of the Formation, that reach more than 1000 m in some places comprises monotonous, well bedded, fine- to medium-grained quartz arenites, thin dolomite (Ahluwalia & Dasgupta 1982), shale or siltstone layers might occur, but are very rare. Tabular, tangential and trough cross bedding, desiccation cracks, mudchips and various kinds of ripples are common sedimentary structures (see Bagati *et al.* 1991 for details). Bhargava & Bassi (1998) mention casts of gypsum crystals and volcanic glass sherds in the Shian Fm. in Kumaon. In the uppermost part the bed thickness and grain size gradually decrease; Bhargava and Bassi (1988) mention marine trace fossils from this part.

DISCUSSION: Due to the scarcity of fossils, the age of the Shian Fm. is not constrained precisely. What a pity, because the knowledge of the stratigraphic base of the Formation would be of great importance for the interpretation of the relationship of pre-Shian Deformation in the Pin Valley (Grasemann *et al*. 1997, Wiesmayr 2000), depositional gap and intrusion of granites. From the data available at the moment, the hiatus between the Haimanta Group and the Shian Fm. shows regional variations and broadly ranges from early Middle Cambrian to Arenigian. Most of the granite ages cluster between 480-470 Ma, *i.e*. they intruded during the younger part of the depositional gap.

The continental deposits in the lower parts of the Shian Fm. represent an important interruption within thousands of meters of Neoproterozoic-Paleozoic marine sedimentation on the shallow N-Indian shelf. Bagati *et al.* (1991) have worked in detail on this Formation; they recognized three different depositional environments and draw a NNW-SSE trending shoreline for the Spiti area during Ordovician time. In the lower part of the Formation they conclude a braided river type fluvial depositional environment, the scarcity of fine-grained deposits are explained by bed load dominated transport, caused by the lack of terrestrial vegetation, which results in high denudation and run-off rates. The broad range of the paleocurrent directions between N-ENE is explained by channel migration.

In the middle part of the Shian Fm., flaser -, lenticular -, and herringbone bedding are common, additional sedimentary structures are reactivation surfaces, lunate ripples and mudchips.

Bagati *et al.* (1991) suggest a fluvial-marine depositional environment on a tidal flat for these sediments. The paleocurrent directions are very variable. The uppermost part of the Formation with decreased bed thickness and grain size represents true marine sedimentation, an interpretation that is also supported by the occurrence of marine trace fossils.

3.4. Pin Formation (Goel & Nair 1977)

CORRELATION: **Kashmir:** Upper Silurian (Middlemiss 1910), Marhaum Fm. (Shah 1968), Halmatpura Fm. (Raina & Razdan 1975), Rishkobal Fm./Mb. B (Srikantia & Bhargava 1983), Margan Shale (Kumar *et al.* 1987); **Zanskar:** Takche Fm. (Srikantia *et al.* 1978); **Spiti:** Muth series/Mb. B (Stoliczka 1866), Silurian system/coral limestone and flesh coloured and brown limestone, Devonian system/dark coral limestone and Carboniferous system/red crinoid limestone (Griesbach 1891), Silurian limestone (Hayden 1904), Takche Fm. (Srikantia 1981), Pin Dolomite (Bagati 1990), Pin Fm. (Garzanti et al 1996b); **Kinnaur:** Manchap Fm. (Bassi *et al.* 1983), Takche Fm. (Bassi 1989); **Kumaon:** Shiala Fm., Yong Limestone (Shah & Sinha 1974) and Variagated Fm. (Heim & Gansser 1939).

TYPE SECTION: Mountain slope on the right side of the Pin River 1.6 km South of Muth (Pin Valley, Spiti). The exact position of the type section is indicated in the geological map of Fig. 4.1 and the photo of Fig. 4.9.

NAME & BOUNDARIES: Goel & Nair (1977) introduced the term “Pin Limestone” after the nice outcrop of the Formation in the Pin Valley. The name was modified by Khanna *et al.* 1981 to Pin Dolomite, because they regarded all carbonates of this sequence as dolomite. Both formation names are somehow misleading because in fact the Formation comprises a variety of lithologies, like limestone, dolomite, sandy dolomite, sandstone, quartzite and shale, therefore the use of the more general term “Pin Formation” is strongly recommended. Srikantia (1981) introduced “Takche Formation”.

Srikantia (1974) introduced “Takche Fm.” after a valley east of Kunzam La, but in an unpublished survey report, and following the recommendations by Salvador (1994) this reference is not considered here. The first published papers using “Takche Fm.” are Srikantia (1978) and Srikantia (1981), except a short reference in Srikantia (1977).

In the subsequent years various authors have used both terms: Takche Fm. by Bhargava & Bassi (1986), Bhargava & Bassi (1998); Pin Dol./Fm. by Ranga Rao *et al.* 1982), Talent *et al.* (1988), Bagati (1990), Bagati *et al.* (1991), Garzanti *et al.* (1996b), Parcha (1998). Both terms have been published in the same year, but in the author’s opinion, there are several arguments for the use of Pin Fm. “*At the Takche type locality, however, the section is not well developed*” (Bhargava & Bassi 1998), thus the use of this name does not seem advantageous. Additionally, the sections in the Takche Valley are strongly faulted (Hayden 1904).

On the other hand the section near Muth in the Pin Valley is easy to access (the road to Muth is already under construction), the profile is located at the limb of a large-scale fold (Fig. 2.11), therefore hardly affected by deformation and thus offer ideal outcrop conditions. As mentioned above, an additional advantage is the concentration of three type sections (*i.e.* Shian Fm., Pin Fm. and Muth Fm.) in one single continuous section within an area of less than 3 km in the Pin Valley, thus facilitating further investigations.

Due to the distinct colors of the Early Paleozoic Formations (purple to brick red Shian Fm., yellowish brown Pin Fm. and snow white Muth Fm.) the boundaries are easily visible from distance (Fig. 3.9). The lower boundary is delineated at a relatively sharp, but conformable contact of thick-bedded, light gray quartzites of the basal Pin Fm. (appendix B20) above typical reddish siltstone to fine-grained sandstone of the underlying Shian Fm.

The upper boundary is a sharp contact of pinkish gray to orange, yellowish brown weathering dolomite, overlain by white quartz arenite of the Muth Fm. (Figs. 4.10, 4.13, 4.16; appendix B1, B11). Despite of the striking lithological break, the upper boundary of the Pin Fm. seems to be conformal near Muth. In the Mikkim section this boundary was interpreted as a disconformity, because of the reddish oxidation and small relief of the uppermost bed of the Pin Fm. This definition of the boundaries corresponds with Hayden´s (1904) “Silurian limestone”.

AGE: Although rich faunas have been described from the Pin Fm. (*e.g.* Reed 1912, Mehrotra *et al.* 1982 and Bhargava & Bassi 1986), the exact age is not well constrained. Reasons are the difficulty of taking out fossils from these often hard, siliceous lithologies and details of fossils are often obscured by dolomtization. Additionally this section lacks recent reexamination of existing fossil collections and recent re-sampling. Most of the fossils described in the literature originate from the two prominent carbonate horizons (Fig. 4.9; appendix B20).

Hayden (1904) concluded Caradocian age for parts of the lower limestone horizon that he dated by trilobites and brachiopods; Goel *et al.* (1987) mentioned Caradocian algae from similar stratigraphic levels. The calcareous upper levels are tentatively dated into the Llandoverian to Wenlockian by Hayden (1904) and Reed (1912). Berry & Boucot (1972) and Talent *et al.* (1988) regard Hayden´s (1904) faunas of beds 6-8 as broadly Llandoverian with a possible extension of bed 8 into younger ages. Therefore a Wenlockian, possible a Ludlovian age for the uppermost Pin Fm. is likely. For further details and faunal lists the reader is referred to the above mentioned authors.

DESCRIPTION: The Pin Fm. reaches a thickness of some 290 m at the type section and consists of variable lithologies. According to Ranga Rao *et al.* (1982) and Bhargava & Bassi (1998) the Formation shows a terrigenous dominance in the Takche Valley with increasing carbonate content towards the Pin Valley in the SE. Two dark gray weathering limestone horizons within the usually yellowish brown weathering Formation are easily visible from distance (Fig. 4.9). In principal within the first 230 m two large-scale thickening and coarsening upward cycles are recognizable, both starting with dolomitic sandstone gradually leading into coral limestone.

The first cycle is made of some 100 m of fine medium-grained, yellowish brown dolomitic sandstone with a conspicuous light gray relatively pure quartz arenite at the base. Thin shale layers between the beds and mud chips are common, as well as bioturbation on the upper bed surfaces. The contact to the overlying 50 m thick, gray limestone is gradual and marked by an increasing content of bioclasts and grain size. Coral buildups occur mainly in the middle part of the limestone, the upper part comprises nodular limestone.

The boundary to the second cycle is relatively sharp. The cycle starts with dolomitic siltstone at the base grading into dolomitic sandstone; both together being some 25 m thick. The contact to the overlying 55 m limestone is gradual. Coral buildups (for description see Reed 1912 and Bhargava & Bassi 1986) occur mainly in the upper part and are quite abundant there.

The uppermost 60 m of the Pin Fm. consist of dolomitic sandstone to dolomite intercalated with shale/siltstone layers, the boundary to the limestone below is sharp. The siliciclastic influence increases towards the top. High content of crinoidal fragments, increasing silicification (especially strong in the fine-grained layers) towards higher levels and the conspicuous yellowish orange brown weathering color are diagnostic, "Red *Crinoid* limestone" of Griesbach (1891). Bioturbation throughout the beds is very prominent (Fig. 4.6). The upper boundary to the overlying Muth Fm. is a prominent lithological break (Fig. 4.13, 4.16). A more detailed description of this Formation can be found in Chapter 4.1.

DISCUSSION: In principal the Pin Fm. is well exposed in the mountain slopes on both sides of the Pin River south of Muth, but the mountain slope on the right side is preferred because of its better accessibility and because this outcrop is less impaired by faults. The Formation comprises a big variability of lithologies, some of them appear to be quite constant from Spiti to Kumaon (Griesbach 1891, Hayden 1904) therefore a more detailed subdivision on the member level might be possible, but needs further detailed sections. Patching up these variable lithologies into one formation is preferred, because of the relative homogenous appearance of the Formation from distance and the similarity of the dolomitic sandstones at the base and top of the Pin Fm.

3.5. Muth Formation (Stoliczka 1866)

CORRELATION: **Kashmir:** Muth Quartzite (Middlemiss 1910), Muth Fm. (Srikantia & Bhargava 1983), Muth Fm. (Kumar *et al.* 1987); **Zanskar:** Kenlung Fm. (Nanda & Singh 1977), Muth Quartzite (Baud *et al.* 1984); **Spiti:** Muth series/Mb. C (Stoliczka 1866), Carboniferous system/white quartzite (Griesbach 1891), Muth quartzite (Hayden 1904), Muth Fm. (Srikantia 1981); **Kinnaur:** Muth Fm. (Bassi 1989); **Kumaon:** white quartzite (Strachey 1851), Muth Quartzite (Heim & Gansser 1939).

TYPE SECTION: Mountain slope on the right side of the Pin River 1.4 km south of Muth (Pin Valley, Spiti). GPS-datum of the contact between the Muth Fm. and the underlying Pin Fm.: N31°56'43"; E078°02'04"; EPE: 30 m; Altitude: 3965 m. The exact position of the type section is indicated in the geological map of Fig. 4.1 and the photo of Fig. 4.13.

NAME & BOUNDARIES: Stoliczka (1866) has named the Formation after the great outcrops next to the village Muth in the Pin Valley (Spiti). Due to its striking and uniform appearance it is probably one of the at least debated formation in the whole NW Himalaya.

The lower boundary is represented by a sharp contact of pinkish gray, yellowish orange brown weathering dolomite of the Pin Fm. overlain by white quartz arenite of the Muth Fm. Despite the striking lithological break this contact seems to be conformal near Muth. In the Mikkim section this boundary is interpreted as a disconformity, because of the reddish oxidation and small paleo-relief of the uppermost bed of the Pin Fm.

In the Pin Valley a relative gradual decrease in purity of the quartz arenites and an increase in carbonate content represent the upper contact to the overlying Lipak Fm. In sections, which lack the Lipak Fm. due to erosion below Permian sediments (see the valuable geological maps by Bhargava & Bassi 1998), the Permian Kuling Group rests directly on an erosional surface on top of the Muth Fm. In the fully developed type section the actual boundary was drawn at the first appearance of dark carbonaceous, argillaceous siltstone and dolomite of the Lipak Fm. The definition of the boundaries corresponds with Hayden´s (1904) “Muth quartzite”.

AGE: With only a few exceptions (*e.g.* Bassi, 1988), the Muth Fm. in its present definition is practically devoid of fossils. The scarcity of suitable and indubitable (Talent *et al.* 1988) fossils has meant that the Formation has been dated mainly by the fauna of the Lipak Fm. above and the Pin Fm. below (see there). Special attention deserves the disconformity at the lower boundary probably indicating a major time gap below the Muth Fm. (Fuchs 1982; Srikantia and Bhargava 1998).

DESCRIPTION: The Muth Formation in the Pin Valley comprises monotonous striking white, stunning pure, fine- to middle-grained quartz arenite with virtually no clay matrix present. The quartz arenite shows high textural as well as compositional maturity. The only exemption within these quartzites is a thin horizon of sandy/silty dolomites in higher levels of the Formation (Figs. 4.10, 4.11, 4.13). In the Pin Valley the thickness of the Formation reaches 258 m at the type locality and some 305 m in the section SE Mikkim (Pin Valley).

In the Pin Valley the Formation can be divided into 4 facies (FA1-FA4). The lowest 140 m are dominated by relatively thin bedded, horizontally laminated arenites, with an increase of tabular cross-bedded beds into higher levels. This lithologies are followed by some 80 m thick, large-scale tabular and tangential cross-bedded beds. The third part forms a conspicuous horizon comprising about 10 m yellowish to orange weathering very fine-grained dolomite, silty/sandy dolomite and sandstone. This horizon can at least be traced towards Hango in NW Kinnaur (Bhargava & Bassi 1998). The uppermost 35 m of the Muth Fm. consist mainly of horizontally bedded quartz arenite, which shows a gradual increase in impurity towards higher levels. A more detailed description and interpretation of this Formation can be found in Chapters 4.4 and 5.

DISCUSSION: Stoliczka (1866) introduced the term “Muth series” for sediments in Spiti, which in today’s definition included the Shian Fm., Pin Fm. and the Muth Fm. The Shian Fm. was excluded in Hayden´s (1908) fourfold “Muth System” that included (a) Dark coral limestone, (b) Red quartzite, (c) Coral limestone and finally (d) White Muth quartzite. In the present definition series a-c comprise the Pin Fm., while only the uppermost part, the white Muth quartzite, is regarded to represent the Muth Fm.

Recently, Bhargava (1997) regarded only the pure, white quartz-arenite as Muth Fm. and delineated the Lipak Fm. at the first appearance of the carbonates. In the present work the arenaceous dolomite-incursion and the pure, white quartz-arenite above it are included into the Muth Fm. due to its intimate facies relation with the rest of the Formation, a similar concept also suggested by Fuchs (1982), Talent *et al.* (1988), Bhargava & Bassi (1998) and Srikantia & Bhargava (1998).

3.6. Lipak Formation (Hayden 1908)

CORRELATION: **Kashmir:** Syringothyris Limestone (Middlemiss 1910), Wazura Fm. and Aishmuqam Fm. (Srikantia & Bhargava 1983), Aishmuqam Fm. and Syringothyris Limestone (Kumar *et al.* 1987); **Zanskar:** Tanze Fm./Mbs. A and B (Nanda & Singh 1977), Lipak Fm. (Srikantia *et al.* 1978); **Spiti:** Carboniferous system/Dark limestone with Productus sp. (Griesbach 1891), Limestones and quartzites of the Lipak river (Hayden 1904), Lipak series (Hayden 1908), Lipak Fm. (Srikantia 1981).

TYPE SECTION: Due to the lack of fully developed sections of the Lipak Fm. it is impossible to select one single type section, even a single type locality is difficult. Further detailed research is definitely needed. Preliminary, the section in the Lipak Valley (NW-Kinnaur) and the very similar section near Muth in the Pin Valley are regarded to be representative for the lower parts of the Formation. The upper parts of the Formation are well exposed in the lower Spiti Valley near Po and in the upper Spiti Valley near Losar (Fig. 2.3).

NAME & BOUNDARIES: Hayden (1908) introduced “Lipak series” after the nice outcrops near Leo in the Lipak Valley, which is a side-valley of the lower Spiti River in NW-Kinnaur (Fig. 2.3).

A gradual decrease in purity of the quartz arenites and an increase in carbonate content of the underlying Muth Fm. mark the lower boundary of the Lipak Fm. In the present work, the actual boundary has been drawn at the first appearance of dark carbonaceous, argillaceous siltstone and dolomite (Appendix B15).

The upper boundary, where preserved, is a gradual contact with the overlying white sandstones and black pelites of the Po Fm (Garzanti *et al.* 1996b). Bhargava & Bassi (1998) drew the actual upper boundary at the disappearance of carbonate beds in the uppermost part of the Lipak Fm. The definition of the boundaries corresponds with Hayden´s (1904) “Limestones and quartzites of the Lipak river”.

The erosion below the late Permian Kuling Group resulted in a variable preservation of the beds below. In general, there is less erosion in central and NE Spiti, *e.g.* in the Lingti Valley and in the whole Spiti Valley, thus the Carboniferous to early Permian Po Group and Ganmachidam Fm. are best preserved in these places. In the Pin Valley the much deeper erosion cuts off the upper levels of the Lipak Fm., from Muth towards SE the Kuling Group is found to rest on lower and lower levels of the underlying horizons, till in the Thanam Valley they lie on the Pin Fm. (Hayden 1904, Bhargava & Bassi 1998). In Zanskar the Lipak Fm lacks its lower levels and only the upper parts. are developed (Vannay 1993).

AGE: Due to the rich fossil content of the Formation the age is quite well constrained. Hayden (1904) and Fuchs (1982) stated late Devonian to early Carboniferous age for the Lipak Fm. in Spiti, mainly based on the rich brachiopod fauna. Garzanti *et al.* (1996b) found Late Famennian conodonts in lower parts of the Formation in Spiti. The recent re-sampling of the Lipak Fm. for conodonts in the Pin Valley gave well-constrained Givetian age for bioclastic limestone (Fig. 4.19) some 30 m above the base (Fig. 4.20-4.22) and a Famennian, possible early Carboniferous age for the top (pers. comm. John Talent 1999). Radiolarians from the same levels also not exclude a Late Devonian age (pers. comm. Frances Spiller).

Baud *et al.* (1984) found brachiopods and bryozoans in the Lipak Fm. in Zanskar and concluded Early Carboniferous age. Gaetani *et al.* (1986) dated limestone beds directly below the gypsum horizon to middle Tournaisian to Early Visean age. Vannay (1993) carried out detailed conodont sampling and stated early Tournaisian age for the Lipak Fm. in Zanskar. These concurring ages deduced by these authors indicate that in Zanskar only the upper parts of the Lipak Fm. are developed.

Thus the evidence available indicates that the Lipak Fm. covers Late Devonian (Givetian) to Early Carboniferous (early Tournaisian) ages, but shows pronounced regional variations in its extent. For further details and faunal lists the reader is referred to the above mentioned authors.

DESCRIPTION: The basal parts of the Lipak Fm. in the Pin Valley are marked by the gradual transition of the pure quartz arenite of the Muth Fm. to the carbonates of the Lipak Fm. These levels comprise quartz arenite, sandstone and dolomitic sandstone, the appearance of thin carbonaceous shale layers and increasing bioturbation is diagnostic. Interlayering with dolomite is common, some 27 m above the base of the Formation gray, graded bioclastic limestone is found (Fig. 4.21; appendix B17-B19). These limestone beds still show some terrigenous influence and are extremely rich in brachiopods, crinoid stems and bone fragments; acid residues yielded radiolarians, fish scales and conodonts of Givetian age (pers. comm. John Talent 1999).

Towards higher levels the terrigenous influence decreases and gives way to very fine-grained stromatolitic limestone between 37-55 m of the section. The next 30 m are marked by calcareous arenites, rich in crinoids and brachiopods, with some oolite horizons and the occurrence of two sandstone horizons. Calcareous arenites dominate the section between 90-100 m and towards the top they are succeeded by dark gray to black mudstone. In the Pin Valley, 126 m above the base, the Lipak Fm. is truncated by the basal breccia of the overlying Early Permian Gechang Fm. (Fig. 4.21). A more detailed description of these lower parts of the Formation can be found in Chapters 4.5.

In Spiti, no detailed sections of the upper part of the Lipak Fm. exist, therefore the correlatives of the Formation in Zanskar are used for description. There, the basal 125 m comprises dark mudstone and wackestone with some shale layers. This level not only correlates well with the dark mudstones in the upper parts of the Lipak Fm. in the Pin Valley concerning lithology (see Chapter 4.5.), but also the conodont faunas indicate similar ages for both levels (Vanney 1993, pers. comm. John Talent 1999). These mudstones are overlain by a horizon of some 30 m of white pure layered gypsum (Gaetani *et al.* 1986, Vannay 1993), which is characteristic for fully developed sections of the Lipak Fm. The section in Zanskar is finished by some 60 m of dolomite grading into mudstone interlayered with marls, resembling similar lithologies like the mudstone in lower levels of the Formation in Zanskar (Vannay 1993).

As stated earlier, the Formation shows pronounced variations concerning its thickness (Bhargava & Bassi 1998). Due to the lack of fully developed sections, the whole thickness is difficult to state. In a very simplistic approach, just summing up the thickness of the Lipak Fm. in the Pin Valley, which represents the lower part, and the thickness of the Formation in Zanskar (Vannay 1993), which represents the upper part, the Formation reaches some 375 m. Hayden (1904) mentions the value of 530 m for a composite section of the Formation in Spiti. In the author's opinion the true thickness may range within these values.

DISCUSSION: Due to the gradual contact between the Muth and Lipak Fms. the exact boundary is difficult to define and often related to subjective opinions of the authors. Some authors (*e.g.* Bhargava 1997) tend to draw the boundary at the first occurrence of carbonate in the upper levels of the Muth Fm. According to this definition, the dolomite of facies association 3 and the pure quartz arenites of facies association 4 would belong to the Lipak Fm.

In the here presented depositional model (see Chapter 5) the Muth Fm. is interpreted as a barrier island system and there are several arguments to leave F3 and F4 within the Muth Fm., because of their genetic relations. Additionally the dolomites of F3 are devoid of fossils, which is also characteristic for the Muth Fm. The first definitely different sediments appear with thin carbonaceous siltstone and shale interbedded with plant fragments containing sandstone.

The section of the Lipak Fm. in the Lipak Valley (Hayden 1904) is more than three times thicker than in the Pin Valley, but both closely resemble the same lithological succession. The occurrence of gypsum in the upper part of the Lipak Fm. is described from the Lipak Valley in NW Kinnaur, Lower Spiti Valley, Takche Nalla and Zanskar (Hayden 1904, Gaetani *et al*. 1986, Vannay 1993, Bhargava & Bassi 1998). These gypsum beds are generally interpreted as Carboniferous evaporates. Mallet (1866) explains these occurrences by the action of sulphurous thermal springs. The latter interpretation does not seem plausible for the fact that this gypsum is restricted to certain levels of the Lipak Fm. and no other formations are involved. Furthermore the gypsum in the Lipak Fm. has also been described from areas with hardly any thermal spring activity.

3.7. Stratigraphic correlation

It is in the nature of things that litho-stratigraphic correlation with all its complications like lateral facies variation, deformation, metamorphism, incomplete data concerning age and sections, *etc*. is a difficult task to deal with, especially when covering sediments from Kashmir to Kumaon as in this case. The stratigraphic correlation presented in this chapter (Fig. 3.11) is a trial and I apologize in advance my possible overlooking, misunderstanding or misinterpretation of any publications within the existing literature.

Correlation of the Paleozoic formations in the NW Himalaya

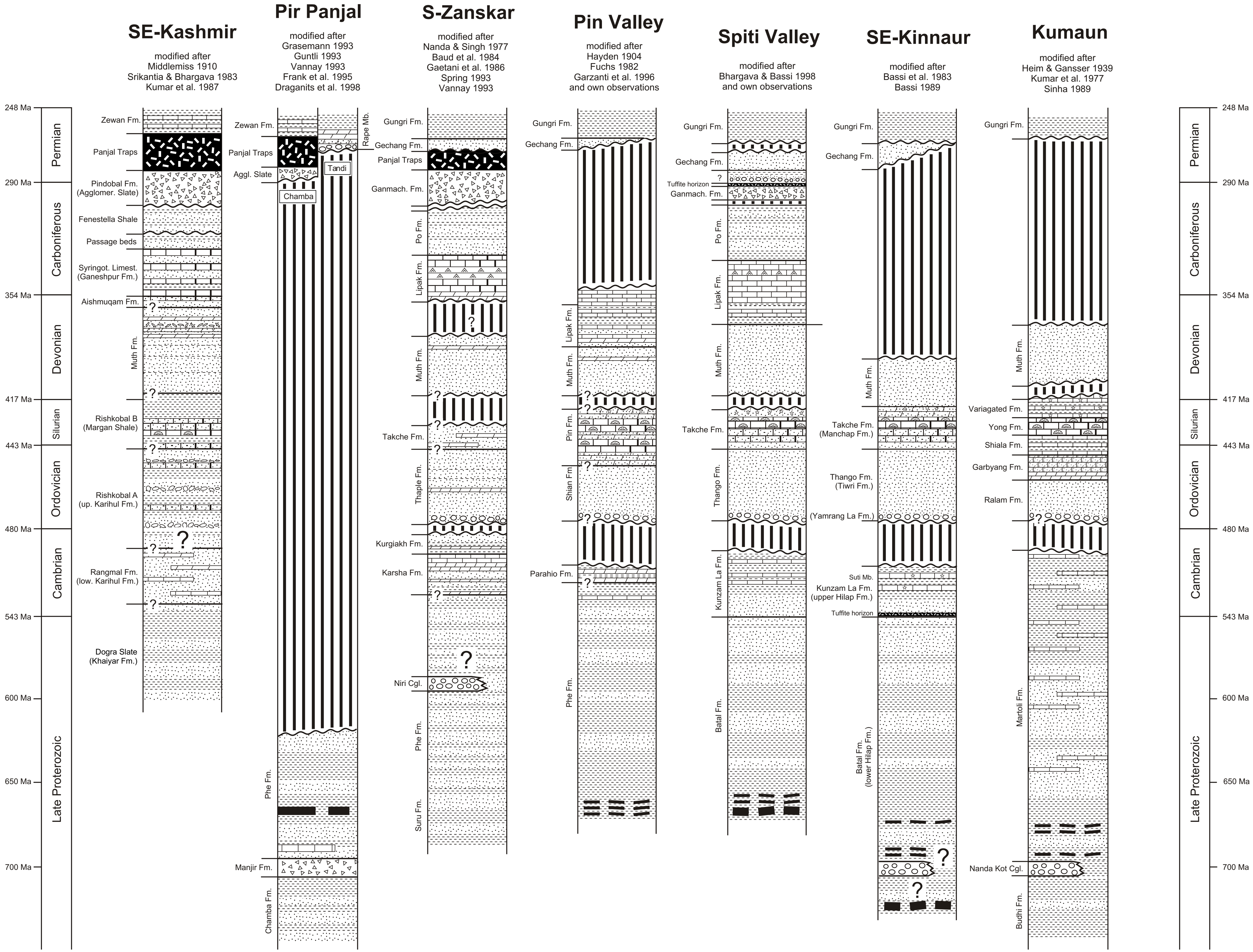


Fig. 3.11: Correlation of Paleozoic Formations in the NW Himalaya, based on literature and own observations in the Pir Panjal Range and Spiti.

In principal, the sediments of the Higher Himalaya are amazingly consistent along strike and they can be traced and correlated from Kashmir to Nepal. In detail, the sequences from Zanskar to Kumaun show the closest relationship concerning lithology, age and depositional environment and represent the best correlatives to the Spiti sections. Higher Himalaya sediments east of Kumaon show increasing differences towards the East.

The correlation of different Paleozoic sections of the Tethyan sediments is mainly based on the striking lithologies of the Shian Fm. and the Muth Fm., the characteristic Permo-Carboniferous diamictites and the easy recognizable Panjal Traps. Unfortunately large parts of the sections lack age-diagnostic fossils, or several fossil descriptions lack modern re-examination. Therefore the correlation still remains preliminary.

SE-Kashmir

Based on the first general account on the geology of Kashmir by Lydekker (1883), Middlemiss (1910) established a Paleozoic stratigraphy for SE Kashmir that was modified by Srikantia & Bhargava (1983) and Kumar *et al.* (1987). Wadia (1934) proposed a Paleozoic stratigraphy for NW Kashmir. Due to the explosive political situation in Kashmir the area lacks recent reinvestigations, as a result the stratigraphy is less well established compared with the sections in Zanskar and Spiti.

Metamorphosed sediments with granitic intrusions form the base of the sedimentary sequences in Kashmir. Wadia (1934) called this unit “Salkhala series” which can be correlated with the Vaikrita System (Griesbach 1891) in Spiti and Kumaon.

With a gradual decrease in metamorphic grade shales with greywacke intercalation have been separated from the Salkhala series. Due to the stratigraphic position a Neoproterozoic to early Cambrian age has been assigned and a correlation with the Phe Fm. in Zanskar and Spiti is reasonable. Various names have been given to these lithologies, Dogra slates (Wadia 1934) and Machhal Fm. (Raina & Razdan 1975) seem to be the most customary ones. Although the contact between Salkhala series and Dogra slates is often veiled by faults, Wadia (1934) stressed out the gradual boundary between the two units.

Increasing arenaceous sediments with common fossiliferous limestone and dolomite beds overlie the shales and sandstones of the Dogra slates. Kumar *et al.* (1987) used Khaiyar Fm. for the lower portion of these lithologies and Karihul Fm. for the upper part. Dating mainly based on trilobite fossils indicate early Cambrian age for the Khaiyar Fm. and an early late Cambrian age for the Karihul Fm. (Jell & Hughes 1997, Kumar *et al.* 1987). For these reasons a correlation of the Khaiyar Fm. with the uppermost Phe Fm., *i.e.* Debsa Khad Member of the Kunzam La Fm. (Kumar *et al.* 1984), and correlation of the lower Karihul Fm. with the Parahio Fm. (Hayden 1904), *i.e.* Parahio Member of the Kunzam La Fm. (Kumar *et al.* 1984) is possible.

Although Brookfield (1993) draws an unconformity with basal conglomerate between Cambrian and Ordovician series in the Kashmir section, the late Cambrian to early Ordovician depositional break seems to be less evident than in Spiti or Kumaon (Middlemiss 1910, Wadia 1934, Kumar *et al.* 1987). The well-developed Ordovician of Spiti apparently is not well recognized in Kashmir (Kumar *et al.* 1987). Middlemiss (1910) describes his “older Silurian” as pale yellowish to purple sandy to argillaceous sediments. According to Srikantia & Bhargava (1983) clastic sediments with common conglomerates dominate their Rishkobal A. In spite of the different appearance of the sediments in Kashmir above the Cambrian, the upper Karihul Fm., *i.e.* Rishkobal A (Srikantia & Bhargava 1983), might represent correlatives of the Ordovician Shian Fm. in Spiti (Bhargava & Bassi 1998).

The subsequent formation comprises the oldest richly fossiliferous beds in Kashmir (Kumar *et al.* 1987) and is characterized by quartzite, impure sandstone, silty shale and brownish calcarenite with sporadic corals (Srikantia & Bhargava 1983). Due to the stratigraphic position directly below the unmistakable Muth Fm. and the broadly determined Silurian age the correlation with the Pin Fm. is evident (Bhargava & Bassi 1998, Kumar *et al.* 1987). The occurrence of corals seems to be characteristic for Silurian sediments of the complete Northern Indian margin (Fig. 3.11).

The easy recognizable Muth Fm. that follows the Silurian with a conformable contact (Srikantia & Bhargava 1983) reaches a thickness of more than 1000 m. Due to its identical lithology, striking white, mature quartz arenites, it has always been correlated with the Muth Fm. of Spiti. In analogue to Spiti the formation is totally devoid of fossils.

The upper boundary of the Muth Fm. is a gradual contact to the sandstone, siltstone and dark limestone of the Syringothyris Limestone (Middlemiss 1910) and characterized by decreasing number of quartzite beds and increase of carbonate and carbonaceous matter (Wazura Fm. of Srikantia & Bhargava 1983; Aishmuqam Fm. of Kumar *et al.* 1987). Most authors assign a Carboniferous age for the Syringothyris Limestone and the formation is broadly correlated with the Lipak Fm. in Spiti. The age would imply, that the Late Devonian part of the Formation is missing in Kashmir, similar to the situation in Zanskar, or this period is confined to the Wazura Fm. of Srikantia & Bhargava (1983), which lacks age diagnostic fossils.

Pir Panjal Range

Phanerozoic sediments in the Pir Panjal range are found in narrow folded synclines NW of Chamba and S of Tandi (Rattan 1973, Vannay 1993, Frank *et al.* 1995). In contrast to the sedimentary sections elsewhere in the HH of the NW Himalaya latest Neoproterozoic to latest Carboniferous are missing. Stampfli *et al.* (1991) and Vannay (1993) explain this huge erosional hiatus with the Lower Permian palaeogeographic position of this area at the rift shoulder during Neotethyan rifting. In this model Zanskar and Spiti were situated ocean-wards of the rift shoulder during that time, while the Kashmir sediments were deposited land-wards on an epicontinental basin. Deposition started with the Agglomeratic Slate in latest Carboniferous to earliest Permian time in the Chamba area, the Phanerozoic sedimentation in the Tandi syncline followed with Permian conglomerates, passing into dolomites and finally shales that that probably be correlatives of the Kuling Group in Spiti (Vannay 1993).

Zanskar

The Tethyan sediments of southern Zanskar show many commons with the Spiti sections and correlation is relatively well established. The Phe Fm. of Nanda & Singh (1977) is easily comparable with the Phe Fm. in the Parahio Valley. The deposits show a similar trend of finer grained silt and shale in lower levels (Tsarap Mb. in Zanskar, Gaetani *et al.* 1986 and Batal Fm. in the Parahio Valley, Srikantia 1981) grading into more arenaceous sediments in higher levels (Doda Mb. in Zanskar, Gaetani *et al.* 1986 and Debsa Khad Mb. of the Kunzam La Fm. in the Parahio

Valley, Kumar *et al.* 1984). In both areas the lower levels are carbonaceous, trace fossils and body fossils appear with the increase in grain size in the higher levels. The Doda Mb. might represent Early Cambrian sediments, the Karsha Fm. higher up in the section, shows increasing dolomite content in its basal Mauling Mb., leading into the dolomites of the Thidsi Mb. and Teta Mb. (Gaetani *et al.* 1986, Garzanti *et al.* 1986). The Mauling Mb. represents a probable correlative of the Parahio Fm. The range of the Cambrian sequences indicates minor erosion below the Ordovician transgression in Zanskar compared with the situation in Spiti.

The Ordovician Thaple Fm. (Nanda & Singh 1977) is nearly identical with the Shian Fm. in Spiti and comprises basal conglomerates above a disconformity and mainly brick red sandstones in higher levels. The Takche Fm. (Srikantia 1981) is not well developed in Zanskar and has just been described as thin pinkish sandstone/siltstone horizon in the Lingti Valley. Corals that are widespread in the Pin Fm. of the Pin Valley have not been mentioned from Zanskar. The contact to the overlying Muth Fm., which is identical, but slightly thinner than in the Pin Valley, seems to be gradual.

The contact to the following Lipak Fm. is generally very sharp (Baud *et al.* 1984, Bhargava & Bassi 1998), but also may be represented by a transition from quartz arenites into carbonates (Gaetani *et al.* 1986, Vannay 1993). Although the contact appears to be gradual in places, conodont ages indicate that the Late Devonian part of the Lipak Fm., which is well developed in the Pin Valley, is missing in the Zanskar sections (Vannay 1993). In strong contrast to the Pin Valley section, the upper boundary of the Lipak Fm. to the overlying Po Fm. seems to be conformal (Baud *et al.* 1984).

Spiti Valley

The stratigraphy of the Pin Valley and Spiti Valley are identical concerning the Neoproterozoic to Upper Devonian sediments. Stoliczka (1866) was the first one, who realized differences in the Upper Palaeozoic of these two areas, introducing “southern facies” for the former and “eastern facies” for the latter, but he failed to give the right interpretation. Stoliczka (1866) correlated the Po Fm. of Lower Spiti with his Babeh and Muth series and described these differences. Hayden (1904) improved the stratigraphic correlation of both areas, stating that the Po Fm. was much younger than Stoliczka (1866) thought and stressed out, that the obvious difference of these two areas, are lying in the different preservation of the Carboniferous below the Permian Kuling Group, a fact that is nicely visible in the maps by Bhargava & Bassi (1998).

In the Pin Valley the upper boundary of the Lipak Fm. is represented by an angular unconformity below the Permian Kuling Group (Griesbach 1891), nearly the whole Carboniferous is missing (Garzanti *et al.* 1996a, pers. comm. John Talent 1999). There is a gradual increase in preservation of the series between the Muth and Gechang Fms. towards the NE, leading to the observation of well-developed sections of this period in the Upper Spiti Valley near Losar, in the Lingti Valley and in the Lower Spiti Valley near Tabo and Po (Stoliczka 1866, Hayden 1904, Garzanti *et al.* 1993). In these areas the upper part of the Lipak Fm. (with evaporites), the whole Po Fm. and Ganmachidam are preserved and the Gechang Fm. of the Kuling Group is thickly developed. There is no unconformity visible below the Kuling Group. Even more a thin tuffite horizon in the Lingti Valley might indicate the presence of Panjal Trap correlatives in Spiti.

Applying the Neo-Tethyan rifting model of Stampfli *et al.* (1991), the different preservation of the Carboniferous in Spiti can be explained by the different positions of these areas relative to the rift shoulder.

SE-Kinnaur

The sedimentary sections of SE-Kinnaur are nearly identical with the sections in the Pin Valley (Bassi 1989). The major difference is represented by the larger erosional hiatus below the Kuling Group; Kinnaur lacks the whole Carboniferous and the Late Devonian sediments. Therefore the Gechang Mb. of the Kuling Group rests directly above an erosional surface on the Muth Fm., in places even the Gungri Mb. follows directly on top of the Muth Fm (Bassi 1989).

Conglomerates, mentioned from lower levels in the Batal Fm. (broadly correlated with the Phe Fm.), might represent the continuation of the Manjir Fm. from the Pir Panjal Range to Kinnaur.

Kumaon

At a short glance the sedimentary sequences of Kumaon seem to be very similar to the sections in Kinnaur and Spiti Valley. Despite this fact there are a lot of discrepancies in the nomenclature and dating of the formations (Griesbach 1891, Heim & Gansser 1939, Sinha 1989). One reason might be that the series in Kumaon show a stronger Himalayan metamorphic overprint than in Spiti. Metasiltstone and metapelite of the Budhi Fm. (Heim & Gansser 1939) represent the base of the sedimentary sequence. Greenish gray conglomerates, Nanda Kot Cgl. of Heim & Gansser (1939), succeed these metasediments, from the stratigraphic position a correlation with the Manjir Fm. is tempting. As a consequence, this correlation implies a correlation of the Budhi Fm. with the Chamba Fm. in the Pir Panjal Range.

Following the argumentation above the following more than 2000 m thick Martoli Fm. (Heim & Gansser 1939), which comprises carbonaceous metapelite and metasiltstone in lower levels, becoming more arenaceous in higher levels, can be correlated with the Phe Fm., which shows a similar trend in lithologies in the Parahio Valley. The upper parts of the Martoli Fm. with increased carbonate content might represent correlatives of the Parahio/Karsha Fms.

The Ralam Fm. (Heim & Gansser 1939) succeeds the Martoli Fm. above an erosional surface and an angular unconformity, in some places the beds show angles up to nearly right angle (Griesbach 1891). This angular unconformity and the “brick-red” quartzite and sandstone above resemble many features of the Ordovician Shian Fm. of Spiti, strongly implying a correlation of both. Surprisingly, because of the uncritical adoption of Griesbach´s (1891) view of a wrong Cambrian age, based just on a wrong stratigraphic position, the Martoli Fm. has only rarely been correlated with the obvious similar Shian Fm. This correlation is also supported by the discovery of early Ordovician fossils directly above the Ralam Conglomerate in Kumaon (Mamgain & Misra 1989).

The Martoli Fm. is conformably succeeded by dolomite and sandy limestone [Garbyang Fm., Heim & Gansser 1939)], limestone and shales [Shiala Fm., Heim & Gansser 1939)], dark limestone with corals [Yong Limestone, Shah & Sinha 1974)] and finally by red crinoidal limestone and shale of the Variagated Fm. [Heim & Gansser 1939; Red crinoidal limestone of Griesbach (1891)] below the Muth Fm. These lithologies closely resemble the various lithologies of the Pin Fm. in Spiti and should be correlated with this Formation.

In some places the boundary between the Variagated Fm. and the overlying Muth Fm. is an angular unconformity (Sinha 1989). The upper boundary of the Muth Fm. is represented by an erosional surface below the Gungri Fm. of the Kuling Group, implying a non-deposition and/or erosion of the whole Late Devonian to Early Permian sediments in Kumaon (Fig. 3.11).

4. STRATIGRAPHIC SECTIONS

4.1. General remarks

Sections of the Pin -, Muth -, and Lipak Fms. have been recorded during the field seasons 1997-1999. All the sections of the Muth and Lipak Fms. represent detailed bed-by-bed examinations and measurements. In contrast, the section of the Pin Fm. has been done in a more cursory way by recording lithological variations and their thicknesses. An emphasis has been laid on accurate thickness measurements and the identification of characteristic lithologies to provide a useful tool for regional comparison. Boundaries of beds are easily identifiable in the variable lithologies

Grainsize analyses have been carried out by a micro lens with a special scale for grainsize analysis from "Fema Salzgitter"; the grain size limits on the scale are 0.063-0.200 mm for fine-grained sand, 0.200-0.630 mm for medium-grained sand and 0.630-2.000 mm for coarse-grained sand. Sample localities are indicated in the sections. Conodont samples with variable success have been taken from the Pin Fm., Muth Fm. (a single sample from the dolomites of FA3) and Lipak Fm.

The exact position of all sections has been marked in the geological map (Fig. 4.1) and on photos; additionally, where possible, GPS data (Garmin GPS III) of the beginning of the sections have been taken. Altitude measurements have been carried out by Avocet Vertech altimeters, in rare cases with Thommen altimeters. For space reasons, sections in this chapter just show summary sections, detailed sections in the scale of about 1:100 are placed at the end in Appendix B.

4.2. Phe Formation

Section F-F´(beds H_b1-H_b15) – Appendix B (page B16)

Section F-F´ represents a short section within the upper part of the Phe Fm. (Kunzam La Fm. of Srikantia (1981). The intention for measuring this section was the discovery of fossiliferous limestones by H.P. Schmid in 1998 and the subsequent sampling for paraconodonts.

The section is located in the Parahio Valley on the right bank of the Parahio River, *c*. 10 m from the waterline directly opposite of the village Thango (Fig. 4.1). GPS data of the first bed: N32°01'56"; E77°56'48"; EPE 39 m; Alt.: 3790 m (Thommen altimeter). The section runs W-E on level ground.

Rock surfaces throughout the section are usually covered by thick desert varnish, which is typical for rocks in this climate close to the water line. The first three beds comprise greenish gray fine middle-grained sandstone with thin dark gray siltstone layers, bioturbation is rare. The uppermost 10 cm of bed H_b4 shows bioturbated gray limestone containing brachiopod and trilobite fragments. The next two beds show greenish gray sandstone, with irregular wavy beds, abundant thin, dark gray shale intercalation with detrital white micas. Thin burrows on top of the beds are common.

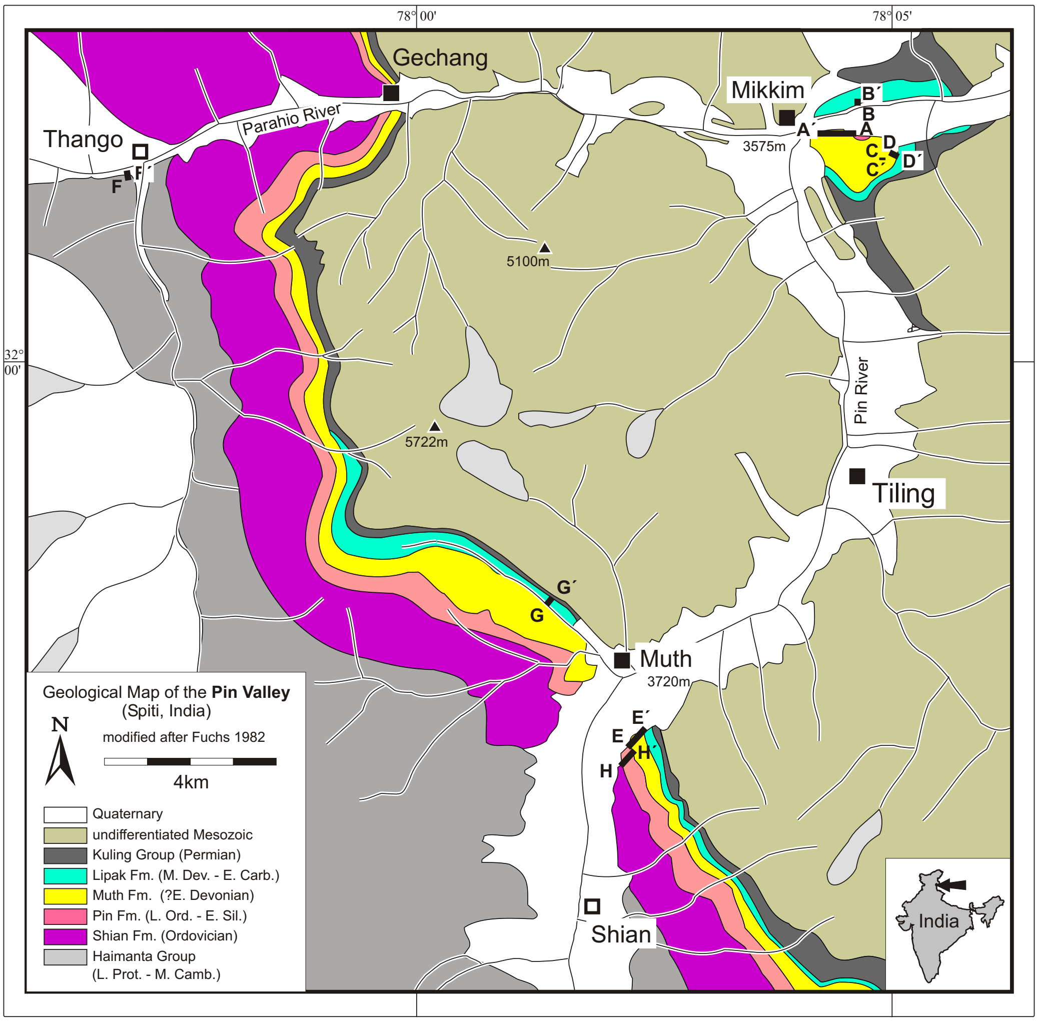


Fig. 4.1A: Geological map of the Pin Valley between Mikkim and Shian with section localities; simplified and modified after Fuchs (1982).

Fig. 4.1B: General view from the Mangling peak (5925 m) towards SW to village Muth, Pin River flowing from upper left to lower right. (1) village Muth. (2) villige Tiling. Brownish weathering Mesozoic rocks in the foreground; grayish white Muth Fm. followed by beige weathering Pin Fm. and purple Shian Fm. Gray Haimanta Group (Phe Fm.) in the background (compare with Figs. 2.6, 2.11, 3.9 and 4.1A). Photo courtesy Gerhard Wiesmayr.

Fig. 4.2: Very well developed horizontal lamination in sandstone in the upper part of the Phe Fm. in the Parahio Valley directly next to the *Diplichnites* trace of Fig. 7. Lens cap is 53 mm in diameter.

Fig. 4.3: Limestone in the upper Phe Fm. in the Parahio Valley. Gray, phophatized, fossiliferous limestone of bed H_b15 in section F-F´ with wavy bedding and thin siltstone intercalations.

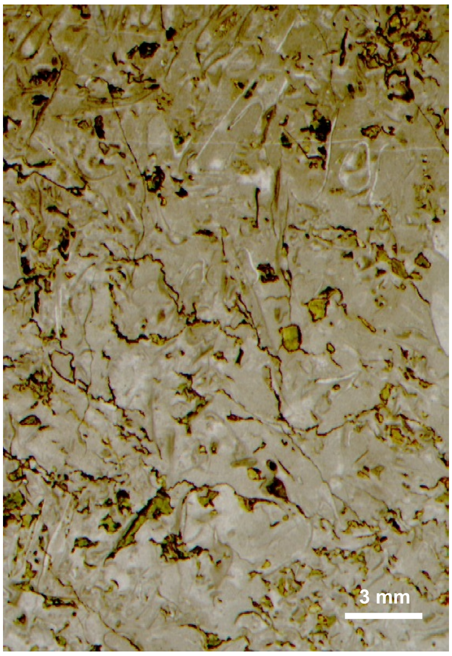


Fig. 4.4: Thinsection of the phosphoritic limestone of bed H_b15 (sample ED98/225) with typical shepard´s crook-shaped cross-sections of trilobite fragments and several brachiopods. Note the abundance of glauconite.

Fig. 4.5: View from the ridge SE of Thango (Parahio Valley) towards the North to the angular unconformity below the Shian Fm. visible in the foreground and in the background.

Beds H_b7-H_b12 mainly consist of greenish gray sandstone with parallel, even bed boundaries, some thin horizons of intraformational breccia occur. Body fossils are rare, but bioturbation is very common. Frequent ripple bedding and lenticular bedding are found.

After a sharp contact the limestone of beds H_b13-H_b15 follows. The first bed comprises an intraformational breccia of up to 5 cm big sandstone clasts in gray bioclastic limestone. Limestone of bed H_b15 shows wavy parallel boundaries and intercalation of dark shale horizons. The strongly phosphoritic limestone contains glauconite and abundant up to 5 mm large brachiopod, crinoid and trilobite fragments (Figs. 4.3, 4.4). Sample ED98/225 possibly yielded rare paraconodonts thus indicating Middle Cambrian age or younger.

These sediments were deposited in a shallow marine setting, probably a tidal flat depositional environment, indicated by rhythmic sand/mud intercalation, lenticular bedding and sandstone with parting lineation. Garzanti *et al.* (1986) concludes similar environments for the Phe Fm. in Zanskar.

4.3. Pin Formation

Section A-A´ (beds P_a1-P_a39) – Appendix B (page B1)

Beds P_a1-P_a39 of section A-A´ comprise the uppermost part of the Pin Fm. at the base of this section (Fig. 4.10). This part of the Pin Fm. is well comparable with litho-units LU-37-LU-40 of section H-H´. Section A-A´ has been made to work out the clear stratigraphic position of the arthropod trackways (Draganits *et al.* 1998a) within the Muth Fm. Due to the lack of fossils in the Muth Fm. the uppermost part of the Pin Fm. has been included into the section and conodont samples have been taken; unfortunately none of them contained conodonts. In general this part of the Pin Fm. comprise gray dolomite to dolomitic sandstone in all variations, with a characteristic yellowish brown weathering color. The whole described section is strongly silicified.

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1 km ESE of the village Mikkim in the core of the prominent anticline of the Muth Fm. (Fig. 4.1, 4.11). GPS data of the first bed: N32°02’15”; E78°04’18”; EPE 44 m; Alt.: 3610 m (Avocet Vertech altimeter). The section starts above the quaternary fluvial gravel and runs uphill *c.* N-S on the steep slope.

Bed P_a1 consist of gray laminated sandy dolomite with few crinoid stems and shows slump folds throughout the bed. Fold axis of five measured slump folds trend *c.* 300-120° (Fig. 5.4C). If the assumption of the position of the open sea somewhere towards the North is correct, these slump folds indicate a slumping direction towards 75° (ENE) and thus a similar direction of the paleoslope, but clearly more data are required to support this preliminary interpretation. Softsediment deformation in similar levels of the Pin Fm. is also described from other parts of its outcrop (Bhargava & Bassi 1986).

The subsequent 4 beds show quite similar lithologies, but they contain slightly more crinoid stems, bed P_a2 shows stromatolitic structures in places and beds P_a4 and P_a6 contain vertebrate bone fragments.

Beds P_a7-P_a28 are characterized by bioturbation throughout the beds, which destroy the primary lamination that might represent former microbial mats and is responsible for the nodular

Fig. 4.6: Outcrop photo of a Pin Fm. dolomite (bed P_a12; section A-A´). Characteristic appearance of relatively small sized bioturbation in the dolomite in the upper part of the Formation.

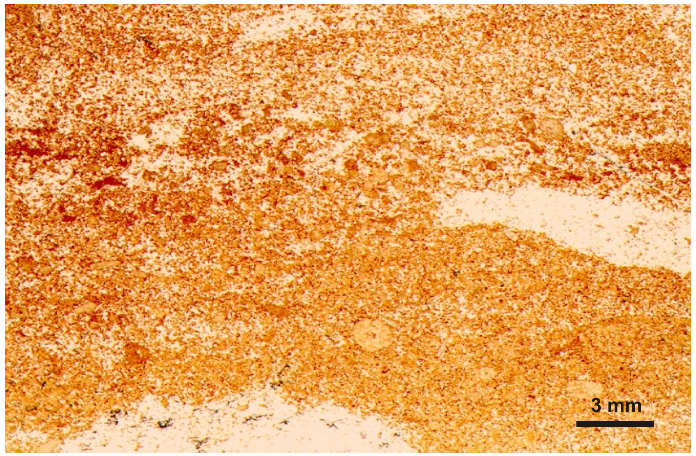


Fig. 4.8: Thinsection of the uppermost bed of the Pin Fm. in section A-A´ (bed P_a39). Note the striking orange color of the sandy dolomite, quartz sand-filled burrows and rare crinoid stems.

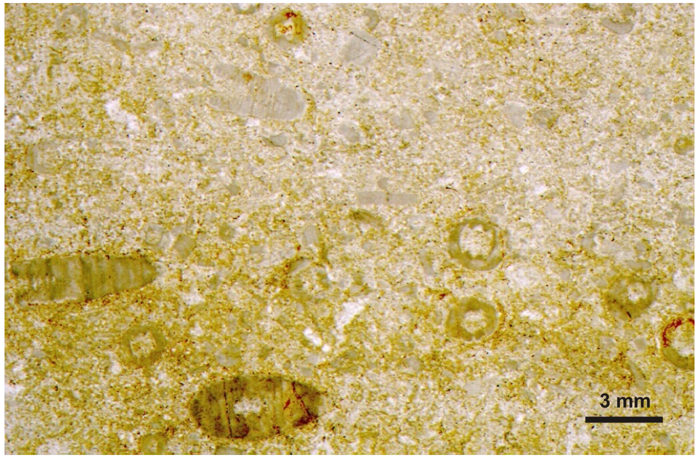


Fig. 4.7: Typical gray dolomite from the upper part of the Pin Fm. (bed P_a30; sample ED97/112).

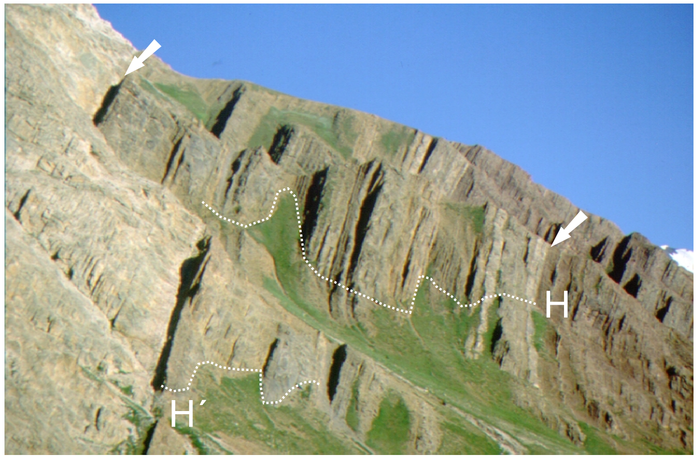


Fig. 4.9: Type section of the Pin Fm. in the Pin Valley. View from Muth towards the Southeast. Dotted lines indicate course of the section. Note the sharp boundaries and the good correspondence of the lithologies from distance with the type section in appendix B20.

appearance of the rock. The bioturbation of beds P_a7-P_a11 consist of burrows with diameters of *c.* 2-3 cm. The size of the burrows diameter increases above P_a11, being in the range of 3-4 cm and the beds are strongly bioturbated up to P_a17 (Fig. 4.6). There the intensity of bioturbation gradually decreases and becomes rare above bed P_a28. The sediment within the burrows is slightly coarser-grained than outside.

Beds P_a29-P_a32 totally lack bioturbation. These beds are coarser grained than the beds before and comprise graded, sandy dolomite with abundant crinoid stems and brachiopods (Fig. 4.7).

The uppermost beds of the Pin Fm. in this section comprise sandy dolomite to dolomitic quartzite. With a relative distinct, but still gradual change to the bioclastic dolomite below, beds P_a33- P_a37 comprise horizontally laminated to cross-bedded dolomitic quartz arenite with hardly any fossils. The youngest two beds are made of sandy dolomite with abundant crinoid stems and burrows on upper bed surfaces. The uppermost bed of the Pin Fm. (P_a39) shows a striking orange color, probably caused by oxidation and an irregular upper surface, thus indicating a disconformity between the Pin Fm. and the overlying Muth Fm. (Figs. 4.8, 4.16). Bhargava & Bassi (1998) describe similar undulatory upper surfaces of the Pin Fm. from the Takche Valley and from the Parahio Valley.

Section E-E´ (beds P_b1-P_b4) – Appendix B (page B10)

Beds P_b1-P_b4 of section E-E´ comprise the uppermost part of the Pin Fm. at the base of the section (Fig. 4.10). Section E-E´ has been carried out to provide a detailed bed-by-bed section of the type section of the Muth Fm. Two conodont samples have been taken; unfortunately none of them contained conodonts.

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1.4 km south of the village Muth, just below the Muth Fm. (Fig. 4.1, 4.13). GPS data of the first bed: N31°56’44”; E78°02’05”; EPE 32 m; Alt.: 3950 m (Avocet Vertech altimeter). The section starts above the quaternary debris at the base of the slope and trends *c.* SW-NE.

Beds P_b1-P_b4 are well-bedded and comprise pinkish gray sandy dolomite, low-angle laminated with abundant crinoid stems and some more than 3 cm large brachiopod shells. All beds are strongly silicified. The contact to the overlying white quartz arenite of the Muth Fm. is a very sharp lithological break, but no indications for a disconformity like in section A-A´ have been found.

Two sets of fractures in the uppermost part of the Pin Fm., which are oriented nearly at right angle to the bedding surfaces trending NNE-SSW and NW-SE seem to have enhanced the paleo fluid flow. The few decimeters of the wall rocks of these fractures have been completely silicified giving the appearance of white quartzite (Fig. 4.34, 4.35B).

Section H-H´ – Appendix B (page B20)

Based on the section by Hayden (1904) at the same locality, section H-H´ has been carried out to provide a cursory section of the complete type section of the Pin Fm. Lithological variations and their thicknesses have been recorded to present an accurate thickness of the Formation. In principal, two large cycles are visible within the variable lithologies of this formation, both starting with dolomitic sandstone grading into coarse bioclastic limestone. These dark gray limestone horizons within the Pin Fm. can be traced from Spiti to Kumaon (Griesbach 1891). Grain sizes usually range in the middle sand fraction except the limestone, which is coarser grained. Two conodont samples have been taken from the upper limestone horizon, unfortunately, they yielded hardly any.

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1.6 km south of the village Muth (Fig. 4.1, 4.3). The section starts above the quaternary debris at the base of the slope and trends *c.* SW-NE.

The Pin Fm. starts with grayish white, horizontally laminated sandstone of litho-unit 1 (LU-1) above a sharp boundary to the underlying purple siltstone/sandstone of the Shian Fm. LU-2 to LU-7 comprise yellowish brown weathering sandy dolomite to dolomitic sandstone, most of the beds are well-bedded and horizontally laminated. Hardly any fossils have been found in this part. These beds closely resemble the orange brown sandy dolomite of the uppermost part of the Formation (LU-40), but lack the characteristic abundant crinoid stems from there.

The consequent LU-8 represents grayish white sandstone similar to LU-1. LU-9 to LU-13 still show similar lithologies as LU-2 to LU-7 below, but the beds are slightly thinner and separated by thin shale/siltstone horizons with strong bioturbation. These argillaceous horizons probably represent the source for common mud-chips. Some thin intraformational breccia occurs. LU-9 consists of sandy dolomite.

LU-14 to LU-19 represent the lower limestone horizon in the Pin Fm. Within LU-14 there is a distinct, but gradual contact to sandy limestone and bio-calcarenites of the beds above. Starting with LU-14 and up to LU-17 there is a gradual change to dark gray, brownish gray weathering limestone. The beds have a more massive appearance and bed boundaries are difficult to find. Fossil content (mainly brachiopods, trilobites, crinoid stems and coral fragments) and grain size increases and in LU-17 corals up to 50 cm in diameter in life position are common. LU-18 and LU-19 are made of dark gray, nodular calcarenite with less fossils and decreased grain size compared with the limestone below.

The following LU-20 and LU-21 are very similar to LU-10 to LU-13, but are slightly finer grained. Relatively thin-bedded sandy dolomite and dolomitic sandstone with shale/siltstone intercalation again show increased bioturbation, but fossils are rare.

The next LU-22 is characterized by a gradual increase in fossil content and limestone content. Fragments of corals rarely occur. LU-23 to LU-32 show gray calcarenite in basal parts they are well-bedded and intercalated with thin marly beds becoming gradually coarser grained with increasing fossil content. LU-28 represents a thin dolomitic sandstone incursion within this limestone interval. Coral fragments are widespread in this upper limestone interval, but become more abundant in upper levels; there, corals in life position are common and bed boundaries are difficult to find.

Above a sharp boundary fine-grained gray dolomites, intercalated with marly beds of LU33 to LU-35 occur. Bioturbation is common in the dolomite beds, hardly any fossils are visible. Bed thickness ranges between 20-30 cm. The following LU-36 comprises yellowish weathering gray dolomitic sandstone with bed thickness' around 40 cm. This interval shows sharp upper and lower boundaries.

Beds from LU 37 upwards towards the top of the Pin Fm. in this section are strongly silicified and correspond with beds P_a1-P_a39 of section A-A´. LU-37 to LU-39 consist of about 70 cm thick gray dolomite beds intercalated with *c.* 20 cm thick marly beds. Towards the top sandy lamination and crinoid content increase, but are still not common in higher levels. Bioturbation is very strong and similar to section A-A´ burrows in LU- 38 have diameters of *c.* 1-2 cm and reach some 3-4 cm in higher levels; towards the top the bioturbation gradually diminishes.

The uppermost part of the Pin Fm. in the type section comprise sandy, well laminated brownish orange weathering gray dolomite, rich in crinoid stems and brachiopods. As a difference to the beds directly below, LU-40 lacks argillaceous intercalation. This level corresponds with Griesbach´s (1891) “Red crinoid limestone”. The boundary to the overlying Muth Fm. in this section is a sharp lithological boundary.

Two sets of fractures in the uppermost part of the Pin Fm., which are oriented at right angle to the bedding surfaces trending NNE-SSW and NW-SE show nearly complete silicification of the wall rocks of these fractures (Fig. 4.34, 4.35B).

4.4. Muth Formation

Section A-A´ (beds M_a1-M_a247 and M_a374-M_a576) – Appendix B (pages B1-B6)

This part of the section represents just the lower parts of the Muth Fm., the upper levels are not exposed in this anticline. Section A-A´ has been made to work out the clear stratigraphic position of the arthropod trackways (Draganits *et al.* 1998a) within the Muth Fm. The section of the Muth Fm. shows monotonous quartz arenite with hardly any lithological variations, the detailed investigation with a micro lens with a special scale for grainsize analysis from “Fema Salzgitter” showed small variations in the grain size that generally varied within the middle sand fraction (in beds M_a74-M_a96 no grain sizes have been measured). Arrows to the right of the stratigraphic columns indicate paleocurrent directions, mainly derived from back-rotated dips of foreset directions.

Beds M_a1-M_a515 are designated to facies association 1 (FA1). Due to a fault (H056/85, L323/32) with *dextral* oblique slip, parts of the section are duplicated and therefore beds M_a248 and M_a373 had to be cut out. FA2 follows above an local angular unconformity (Fig. 4.12) and consists of beds M_a516 M_a-576. The uppermost parts of the Muth Fm. are not exposed in this section.

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1 km ESE of the village Mikkim at the striking anticline of the Muth Fm. (Fig. 4.1). GPS data of the first bed: N32°02’14”; E78°04’18”; EPE 44 m; Alt.: 3640 m (Avocet Vertech altimeter). The section starts in the core of the anticline at the top of the Pin Fm. and runs roughly E-W, the first part (beds M_a1-M_a120) the section line follows the upper edge of the Quaternary debris up and down; from bed M_a121 to the end the section has been taken along the E-W trending base of the cliff (Fig. 4.11).

The base of the Muth Fm. in section A-A´ is represented by a disconformity (Fig. 4.16). The first *c.* 100 m of the Formation are made of white, horizontally laminated quartz arenite, cross-

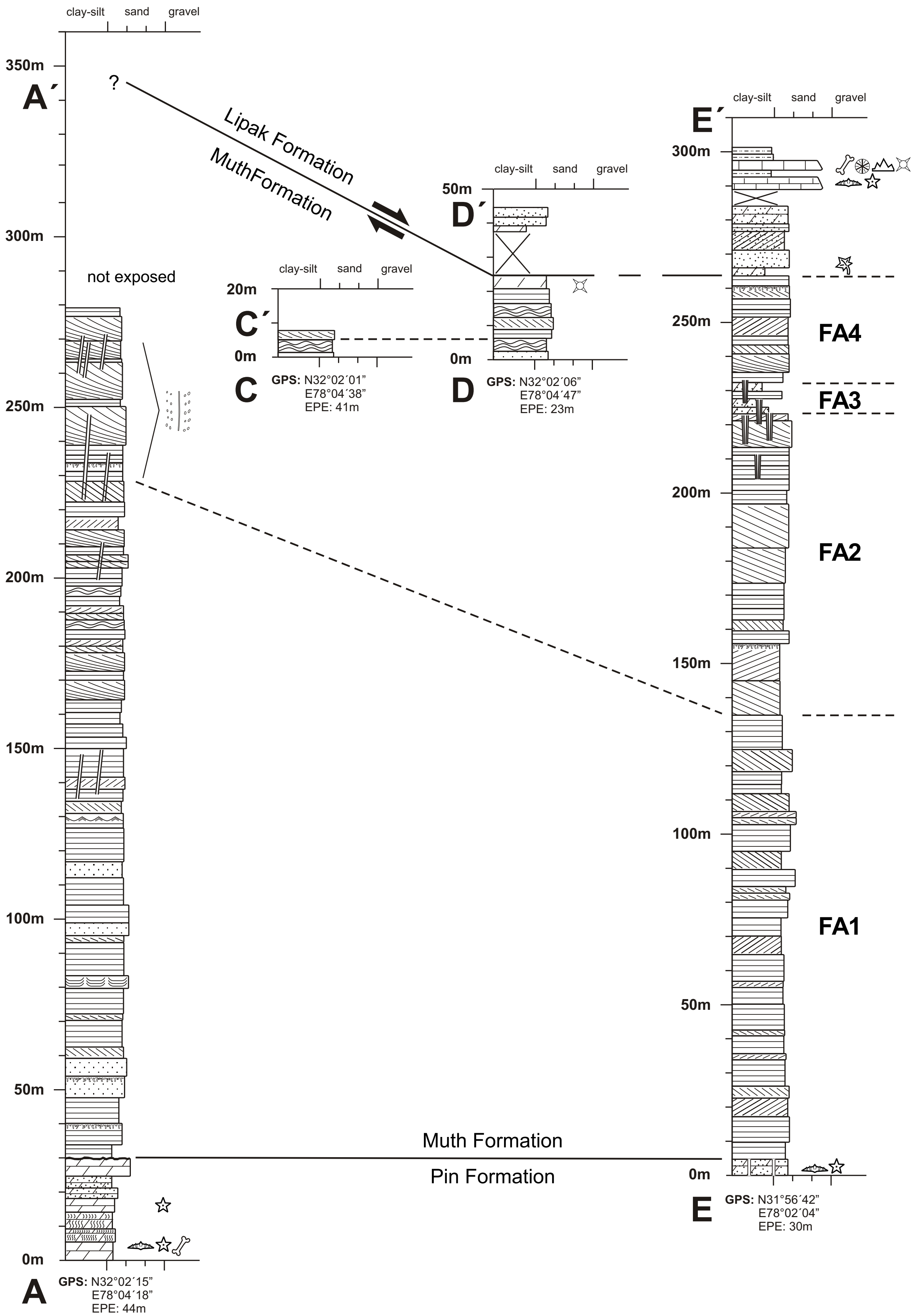


Fig. 4.10: Correlation of Muth Fm. sections in the Pin Valley. Note that in section D-D´ the uppermost part of the Formation has been cut off by a detachment fault. The trackway symbol near section A-A´ indicates the range of occurrence of arthropod trackways.

Fig. 4.11: General view of the Muth Fm. anticline with Pin Fm. in its core SE of Mikkim. View broadly parallel to the fold axis towards SE. Note the relatively shallow dip of the NE limb and the steeper SW limb. Labeled arrows indicate: (1) begin of section A-A´ in the Pin Fm.; dotted line indicates the course of sectionA-A´. (2) Contact between Pin Fm. and Muth Fm. (3) Main trackway locality around the corner. (4) domal stromatolites described in Chapter 5.4.1.

bedding is very rare and confined to some thin beds. Some beds appear to be massive and no internal structures are visible. In the lowest part of FA1, upper bedding surfaces of horizontally laminated beds (beds M_a5-M_a8 and M_a10-M_a13) show surface structures strongly resembling plant root traces (Fig. 4.17). These traces form tunnel shaped *c*. 0.5-1 cm wide, irregular curving, sometimes branching and narrowing structures that penetrate some 5 cm into the bed. Although a formation by faunal burrowing activity cannot be totally excluded, these structures are interpreted as root traces. The subsequent 45 m above bed M_a128, cross-bedding becomes more common and the average bed thickness decreases. Next to beds M_a174, M_a188 and M_a207 deformation bands have been found, bed M_a198 contains two sandstone diatremes.

Beds M_a235 is the first cross-bedded bed with a thickness exceeding 1 m. The following 10 m are characterized by cross-bedded beds. Tangential, concave upward cross-bedding occurs, reactivation surfaces and also herringbone cross-bedding is common. The next 20 m show relatively thick cross-bedded beds with tabular cross-bedding, but horizontally laminated beds are also common.

Beds M_a503-M_a515 comprise thin horizontally laminated beds, bioturbation of the top surfaces of the beds have been found. Between M_a515 and M_a516 there is a local angular unconformity and the difference in the appearance of the beds and the change in dip direction of the foresets indicate a change in the depositional conditions (Figs. 4.12, 5.1). Therefore beds M_a516- M_a576 have been delineated from FA1 below and represent FA2 in this section. All arthropod trackways in this area except one single *Palmichnium* trace in bed M_a515 are confined to strata in FA2 above the unconformity (see Chapter 6).

This part of the section is characterized by large scale tangential cross-bedding with bed thickness sometimes exceeding 4 m. Horizontally laminated beds are common between cross-bedded beds and they often show irregular mounds (see chapter 5.4.1) on the upper surfaces. Deformation bands are abundant, M_a554 shows clear indications for soft sediment deformation.

Section C-C´ (beds M_c1-M_c12) – Appendix B (page B8)

This part of the section covers just a very small part within the upper levels of FA2 of the Muth Fm. Section C-C´ has been prepared to provide a section for the spectacular domal stromatolites found in this part of the Muth Fm.

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1.5 km WSW of the village Guling at the Eastern limb of the easily visible anticline there (Figs. 4.1, 4.11). GPS data of the first bed: N32°02'01"; E78°04'36"; EPE 41 m; Alt.: 3820 m (Avocet Vertech altimeter). The section starts three beds below the stromatolites near the cliff of the Muth Fm. in this area and runs roughly N-S on gently dipping bedding surfaces.

This less than 10 m long section of the Muth Fm. comprises white quartz arenite in fine to medium sand fraction. The section begins with two relative thin, horizontally laminated beds, which are overlain by a massive bed. The following beds show several horizons of domal buildups closely resembling stromatolites, which is surprising in such a terrigenous clastic environment (Figs. 5.10-5.15).

These stromatolites are followed by two horizontally laminated beds and the section is finished by large scale, tangential cross-bedded M_c12. Some 50 m E of the section line, the stromatolites are directly overlain by a cross-bedded bed with a thickness of more than 4 m without any indication of erosion in between (Fig. 5.16), thus implying rapid facies variation in this part of the Muth Fm. More details of these stromatolites can be found in Chapter 5.4.1.

Section D-D´ (beds M_d1-M_d39) – Appendix B (page B9)

This part of the section represents 42 m of the upper levels of the Muth Fm in that area. Section D-D´ has been prepared to present a section for the donut-shaped surface structures (spring pits? Quirke 1930) in lower levels of the section, to provide a more complete section for the spectacular domal stromatolites of section C-C´ and also to describe the uppermost part of FA2 and higher parts of the Muth Fm. in this area of the outcrop. The uppermost part of the Muth Fm. seems to be cut off by a detachment fault (Wiesmayr 2000).

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1.5 km WSW of the village Guling at the Eastern limb of the easy visible anticline there (Fig. 4.1). GPS data of the first bed: N32°02’06”; E78°04’47”; EPE 23 m; Alt.: 3675 m (Avocet Vertech altimeter). The section starts below domal stromatolites near the eastern end of the anticline and runs roughly NNW-SSE on the gently dipping slope.

The basal two beds comprise massive white quartz arenite, followed by two beds with stromatolitic layering. In contrast to the stromatolite of section C-C´ the bio films of beds M_d3 appear as undulatory laminae and do not build single, pronounced domes. M_d3 is followed by a thin cross-bedded bed and then up to M_e15 by horizontally laminated beds. The upper surfaces of M_d5-M_d7 show distinct donut-shaped surface structures with variable diameters between *c.* 10-70 cm (Figs. 5.26-5.29). The beds at this level also show nicely preserved *Taenidium* traces (Fig. 6.22). Except M_d16 with tabular cross-bedding, the next 10 m are horizontally laminated, in some beds, *e.g.* M_d12, the lamination resembles beach lamination.

The thin horizon of M_d19 with stromatolites can probably be correlated with bed M_c6 of section C-C´. The next 7 beds are dominated by horizontal lamination, but herringbone cross-bedding also occurs. The quartz arenite of beds M_d28-M_d38 becomes impure and more or less sandstone to dolomitic sandstone, bed thickness decreases. Two thin, dark gray horizons with fine-grained sandy dolomite occur, the lower one yielded some badly preserved radiolarians in residues from acid-leaching of rocks (ED97/118). Above this level the subsequent 12 m are covered by a green meadow, that covers the location of the detachment fault.

Section E-E´ (beds M_e1-M_e522) – Appendix B (pages B10-B15)

Section E-E´ has been produced to provide a detailed bed-by-bed section of the type section of the complete Muth Fm., because though its considerable importance for regional correlation, the formation lacked detailed investigation so far. Additionally, section E-E´ and the above described section A-A´ have a palinspastic reconstructed distance of 13 km in normal direction to the strike of the Muth Fm. (Wiesmayr 2000) and the continuation of the differentiated facies associations was

Fig. 4.12: H.P. Schmid pointing to the local angular unconformity between FA1 and FA2 in section A-A´.

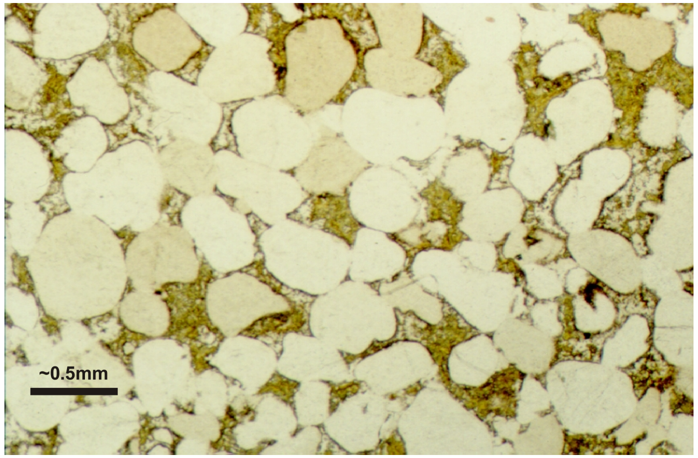


Fig. 4.14: Muth Fm. sample ED97/145, bed M_e437. Rare example of matrix (pore-filling infiltrated former clay) in the super mature arenites. Note the slightly lighter appearance of mono-crystalline quartz compared to poly-crystalline quartz.

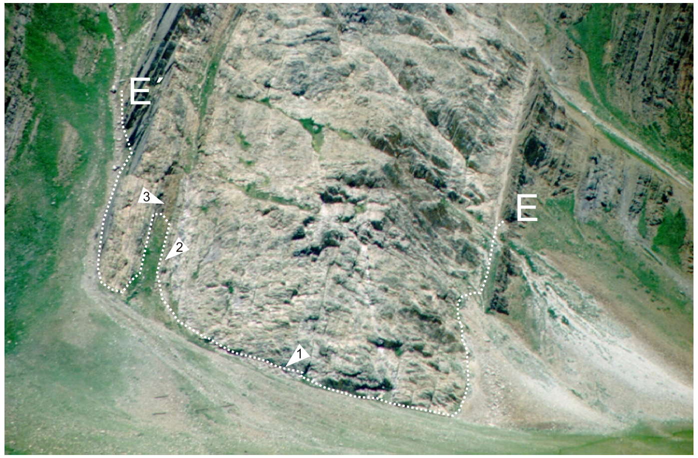


Fig. 4.13: General view towards SE to the type section of the Muth Fm. Arrow (1) indicates boundary between FA1 and FA2; Arrow (2) points to the location of sandstone diatremes. Dotted line indicates course of section. (3) Weathered dolomites of FA3 are easily visibly.

Fig. 4.15: Strange white quartzite boulders in bed M_f1 of unclear origin.

planned to investigate. In general the beds of section E-E´ are slightly thinner than beds in comparable levels of A-A´ and also the grain sizes appear to be faintly finer grained than in the section near Mikkim (Fig. 10). Arrows to the right of the stratigraphic columns indicate paleocurrent directions, mainly derived from back-rotated dips of foreset directions.

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1.3 km south of the village Muth (Fig. 4.1). GPS data from the base: N31°56’44”; E78°02’05”; EPE 32 m; Alt.: 3950 m (Avocet Vertech altimeter). The section starts in the small ravine that marks the contact of the Pin Fm. with the Muth Fm., descends steeply in the ravine and runs roughly in SW-NE direction above the quaternary debris at the base of the slope (Fig. 4.13).

The Muth Fm. in this section starts above a sharp contact with relatively thin beds of quartz arenite in fine medium sand-fraction and in the lowermost 13 m several thin beds show striking pinkish colors. Tangential cross bedding is common, among these, herringbone and trough cross bedding and reactivation surfaces have been noticed. Between 13-133 m of the section the sediment remains relatively similar, but the grain size increases slightly and the arenite is completely white. Horizontally laminated beds alternate with cross-bedded beds.

Between M_e337 and M_e338 the boundary between FA1 and FA2 has been drawn, but the boundary is less evident than in section A-A´ and no local angular unconformity has been found. In general FA2 is characterized by thick beds of cross-bedded arenite, some of them are more than 4 m thick. Cross-bedded sets seem to have slightly coarser grain sizes. Within FA2 there is an alternation of levels with horizontally laminated beds and others dominated by cross-bedding. In the uppermost part of FA2 (beds M_e436-M_e440) several sandstone diatremes have been found, indicating extensive de-watering in these levels (see Chapter 5.4.2. for details).

Above a sharp boundary the uppermost quartzite bed of FA2 is succeeded by fine-grained dolomite of FA3. FA3 represents a striking horizon within the monotonous quartzite of the rest of the Muth Fm. (Fig. 4.13). It comprises mainly of yellowish brown to reddish brown very fine-grained dolomite, alternating with brownish and white quartzite. The sandy beds increase in number towards higher levels; these beds are horizontally laminated or cross-bedded. The dolomite is finely laminated and no bioturbation has been noticed. In the uppermost part of FA3 several sandstone diatremes occur.

Beds M_e472-M_e522 comprise FA4 in the uppermost part of the Muth Fm. with gradual lower and upper boundaries. Beds are white to grayish white, they are horizontally laminated and cross-bedded and show a trend towards stratigraphical higher levels of decreasing purity of the arenite and an increase of bioturbation on upper surfaces. The boundary to the overlying Lipak Fm. has been drawn at the first occurrence of carbonaceous silt and dolomite.

Section G-G´ (beds M_f1-M_f4) – Appendix B (page B17)

Beds M_f1-M_f4 represent the uppermost part of the Muth Fm. at the base of section G-G´. This section has been carried out to provide a detailed bed-by-bed section of the Lipak Fm. for stratigraphic correlation.

The section is located in the Pin Valley on the left bank of the Pin River, in the ravine that separates the Muth Fm. from the Lipak Fm., 0.5 km NW of the village Muth (Figs. 4.1, 4.20). GPS

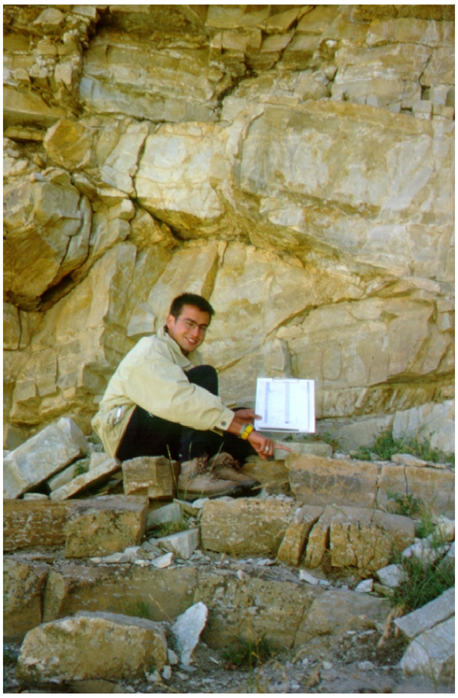


Fig. 4.16: H.P. Schmid pointing to the disconformity between the Pin and Muth Fms. in section A-A´ SE of Mikkim. Note the red color of the uppermost bed of the Pin Fm.

Fig. 4.17: Top surface of horizontally laminated bed (M_a5) showing abundant root traces. Note the variable size, branching and tapering of the structures. Coin diameter is 25 mm.

Fig. 4.18: General view towards the North to the location and course (dotted line) of section B-B´ Lipak Fm.). Arrow pointing to Leo Krystyn for scale.

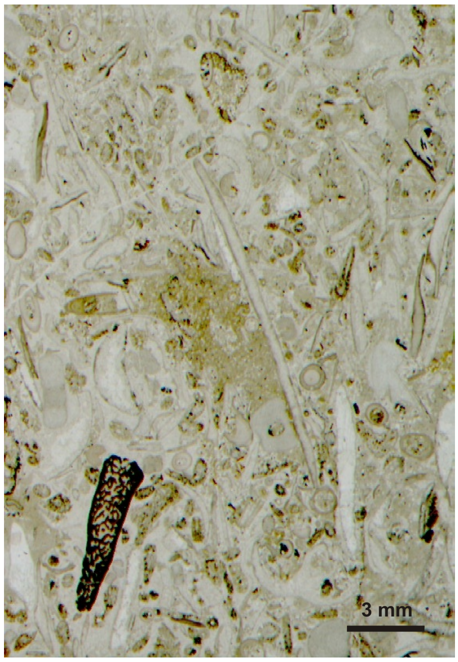


Fig. 4.19: Bioclastic limestone with vertebrate bone fragment (lower left corner) of the Lipak Fm. (section E-E´, bed L_b77, sample ED97/166).

data of the first bed: N31°57’54”; E78°01’17”; EPE 25 m; Alt.: 3980 m (Avocet Vertech altimeter). The section comprises the uppermost beds of the Muth Fm. south of the snowfield that covers the floor of the ravine and trends in a SW-NE direction (Fig. 4.20).

The first bed of the section comprises greenish quartzite with strongly undulatory upper surface. This surface shows abundant holes, in several of which white quartzite boulders sized up to 1 m are found (Fig. 4.15). This distinct horizon correlates with a high probability with the similar upper surface of bed M_e511 in section E-E´. A white, horizontally laminated bed overlies this horizon, followed by a tangential, concave upward cross-bedded bed. The final bed of the Muth Fm. in this section is represented by a thin, massive, white quartz arenite, the actual contact to the Lipak Fm. is covered by snow.

4.5. Lipak Formation

Section B-B´ (beds L_a1-L_a84) – Appendix B (page B7)

Beds L_a1-L_a39 of section B-B´ represent a part of the lowermost levels of the Lipak Fm. near Guling. This part of the Lipak Fm. broadly correlates with the beds of the Lipak Fm. in section E-E´. Section B-B´ has been produced to provide a detailed section of the contact between Muth and Lipak Fms. Later it has been interpreted to comprise solely Lipak Fm. Several conodont samples have been taken, some of them contained conodonts, unfortunately they are badly preserved and less age indicative compared with the conodonts of section E-E´.

The section is located in the Pin Valley on the left bank of the Pin River, *c.* 1 km E of the village Mikkim in a small ravine below the village Gungri (Figs. 4.1, 4.18). GPS data of the first bed, that is situated directly next to the road: N32°02’30”; E78°04’23”; EPE 22 m; Alt.: 3600 m (Avocet Vertech altimeter). The section starts above the road and runs *c.* S-N in the ravine uphill.

Beds L_a1-L_a6 comprise cross-bedded, white quartz arenite resembling typical quartzite of the Muth Fm., but small ferruginous spots and greenish mud chips in basal parts of the beds indicate, that these levels belong to the transition beds of the basal Lipak Fm. already. The actual boundary between these Formations is probably very close below the beginning of section B-B´.

The next 5 m are characterized by a gradual increase in impurity of the arenite beds; they become light gray, show ferruginous spots and alternate with thin carbonaceous marly beds. L_a22 shows nice symmetrical ripples. Beds L_a23-L_a62 show a relative high portion of argillaceous and siltstone beds alternating with more arenaceous beds, graded bedding occurs. Beds are gray to dark gray colored and show an increased carbonate and carbonaceous content. Fine-grained beds often show a lamination of mud- and silt-sized laminae, bioturbation throughout the beds is common. Carbonate beds often comprise very fine-grained dark dolomitic matrix with sand-sized quartz clasts floating in the matrix. Hardly any fossils have been found in these beds so far.

Above a short unexposed part at the 20 m mark of the section up to L_a78, sandy dark dolomite beds alternate with dark shale to siltstone, bioturbation is less than in the beds below. The carbonate content is higher than before and consists of dolomite, therefore HCl still does not fizz on fresh rock surfaces.

Beds L_a79-L_a84 are distinctly different to all the beds below. These beds appear relatively thick and massive, no internal layering is visible and because of their resistance against weathering they form a conspicuous horizon in the field (Fig. 4.18). Bed boundaries are even and parallel. They usually comprise dark limestone with abundant crinoid stems floating in this matrix. Carbonate and quartz filled hollows up to several centimeters are common. Bed L_a84 contains abundant brachiopods at the base and some vertebrate bone fragments and thus might be correlated with the bone-containing beds of L_b77 (section E-E´) and L_c21 (section G-G´). Residues from acid-leaching of rock samples yielded conodonts (ED97/134, (ED97/135 and ED98/218) and radiolarians (ED97/135).

Section E-E´ (beds L_b1-L_b94) – Appendix B (page B15)

Beds L_b1-L_b94 comprise the lower part of the Lipak Fm. in the upper part of section E-E´ (Fig. 4.10). This part of the Lipak Fm. is broadly correlated with section B-B´. Due to the lack of fossils in the Muth Fm. the lower part of the Lipak Fm. has been included into the section and conodont samples have been taken. In general this part of the Lipak Fm. comprise the transition of the arenaceous Muth Fm. into the carbonates of the upper Lipak Fm.

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1 km SE of the village Muth (Fig. 4.1). GPS data are not available. The section starts above the quaternary debris in an altitude of 4245 m (Avocet Vertech altimeter) near a *sinistral* strike slip fault (H177/88; L271/20) that has displaced the SE-part of the section several tens of meters and trends broadly in S-N direction.

The boundary to the Muth Fm. below has been defined at the first occurrence of carbonaceous, argillaceous beds. The first 10 m of the section comprise impure quartz arenite, mostly with a massive appearance intercalated with dark gray siltstone. Some bioturbation occurs on upper bed surfaces. Bed L_b15 contains abundant mud chips.

The following beds L_b31-L_b55 are quite similar to the beds below, but show hardly any siltstone beds, a slightly increasing dolomitic influence is visible. Above some 6 m unexposed area in the section, the terrigenous clastic influence more and more decreases. Beds L_b56-L_b75 mainly comprise gray sandstone, sometimes cemented by calcite. The beds are commonly graded and contain crinoid stems and brachiopods in basal parts. Bioturbation on upper bed surfaces is frequent.

L_b76-L_b83 consist of brownish weathering gray bioclastic limestone (Fig. 4.19), their graded bedding might indicate the tempestite nature of the beds. Some thin, dark siltstone layers occur between the limestone beds. L_b84-L_b88 are made of gray biocalcarenite, sometimes containing intraformational breccia. The final part of the section consists of dark gray siltstone and marl. Beds L_b56-L_b84 are rich in macrofossils (brachiopods, crinoid stems, vertebrate bones), additionally conodont samples yielded nicely preserved conodonts (Fig. 4.22), radiolarians and fish scales.

Section D-D´ (beds L_d1-L_d14) – Appendix B (page B9)

This section was planned to measure the uppermost part of the Muth Fm. and the contact to the overlying Lipak Fm. at the eastern limb of the anticline. During the investigation it appeared that the upper parts of the Muth Fm. have been cut off by a detachment fault (Wiesmayr 2000).

The section is located in the Pin Valley on the right bank of the Pin River, *c.* 1.5 km WSW of the village Guling at the Eastern limb of the easy visible anticline there (Fig. 4.1). The described section represents the upper part of section D-D´ (Fig. 4.10) and starts above the unexposed area at the Eastern end of the anticline and runs roughly NNW-SSE on the gently dipping slope.

The meadow represents the location of the detachment fault. This fault has caused deformation of the uppermost part of the Muth Fm. (*i.e.* FA3 and FA4) and basal parts of the Lipak Fm.; additionally the stratigraphy is unclear because of the bad outcrop conditions. The dolomite and sandy dolomite and the impure quartzite, as well as the fossiliferous limestone bed are indicative for lower parts of the Lipak Fm.

Section G-G´ (beds L_c1-L_c313) – Appendix B (pages B17-B19)

This section has been produced to provide a detailed bed-by-bed section of the Lipak Fm. in the Pin Valley. Several successful conodont samples of previous field seasons have raised hopes for the possibility of an accurate dating of the Formation and therefore 11 more samples have been taken from this continuos section. On this base a reasonable correlation with the Lipak Fm. in Zanskar, which has been dated with conodonts by Vannay (1993), is aimed to.

The section is located in the Pin Valley on the left bank of the Pin River, in the ravine that separates the Muth Fm. from the Lipak Fm., 0.5 km NW of the village Muth (Figs. 4.1, 4.20, 4.21). GPS data at the base of the section: N31°57'55"; E78°01'17"; EPE 25 m; Alt.: 3980 m (Avocet Vertech altimeter). The section starts at the first bed of the Lipak Fm. north of the snowfield that covers the floor of the ravine and also the basal *c*. 21 m of the Lipak Fm. and trends in a SW-NE direction uphill on a steep slope.

The basal *c*. 21 m of the Lipak Fm. in this ravine are covered by snow, but this level is easily to correlate with the nearby Lipak Fm. of section E-E´. Beds L_c1-L_c11 still comprise light gray quartzite, some of them are dolomitic cemented. Above a short unexposed part these beds rapidly become more and more calcareous and consist of gray, graded bioclastic limestone to sandy limestone with increasing bioturbation on upper bedding surfaces. Brachiopods and crinoid stems are very abundant, but remarkable is the occurrence of dark brown vertebrate bone fragments (beds L_c21, L_c25, L_c45) and of corals (beds L_c38, L_c39). Dickins (1993) mentions widespread reefs in the Frasnian and a perhaps universally warm climate during this time. Beds (L_c12-L_c39) are interlayered with thin beds of dark siltstone. This part of the section can probably be correlated with bed L_b77 (section E-E´) by its lithology, the content of vertebrate bone fragments and also by the similar conodont fauna.

There is still a strong detrital, siliciclastic component recognizable up to bed L_c56, but the grainsize and fossil content gradually decrease. Beds L_c54 and L_c56 show intraformational breccia in their basal parts. The following beds up to L_c75 are dominated by fine-grained horizontally laminated stromatolitic limestone, thin marly and sandy beds occur.

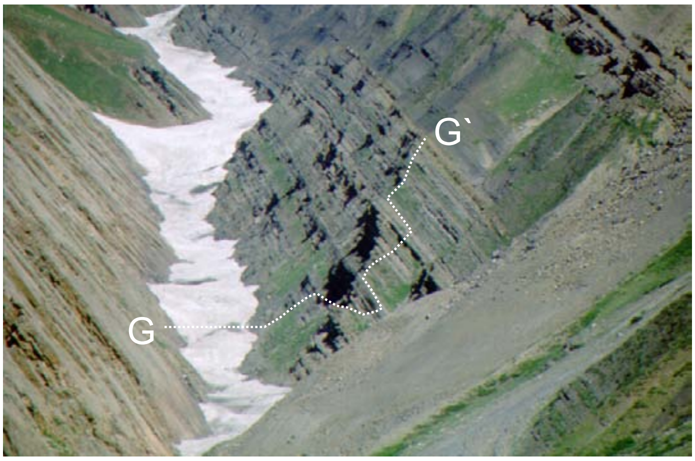


Fig. 4.20: Lipak Fm. south of Muth. View from Pandoshering towards the Northwest to the ravine south of Muth, where the Lipak Fm. has been measured in section G-G´. The dotted line indicates the course of the section, the lowermost part is covered by snow in the ravine.

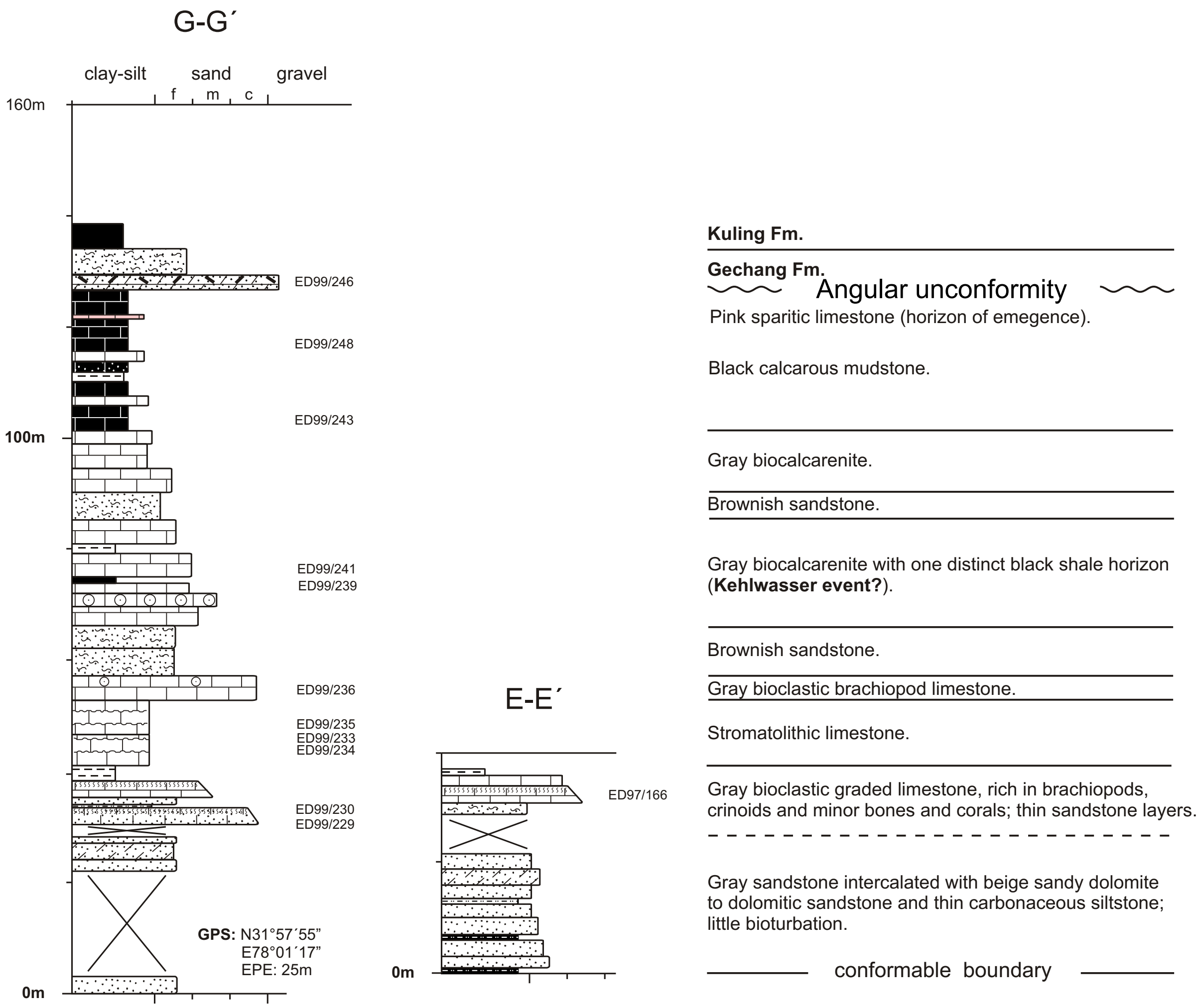


Fig. 4.21: Simplified sections E-E´ and G-G´ of the Lipak Fm. near Muth. Note the two sandstone horizons within the calcareous middle and upper part of the Formation and the black shale horizon, that possibly represents the Kehlwasser event.Only conodont samples are indicated in this figure, for complete sample locations look to detailed sections in the appendix.

Beds L_c76-L_c85 interrupt these stromatolitic limestone beds and represent coarse-grained limestone containing crinoids and brachiopods, L_c84 represents a thick intraformational breccia. The subsequent beds up to L_c113a consist of fine-grained horizontally laminated stromatolitic limestone, similar to the beds below; in the upper part, the limestone becomes coarser grained and ooid limestone occurs.

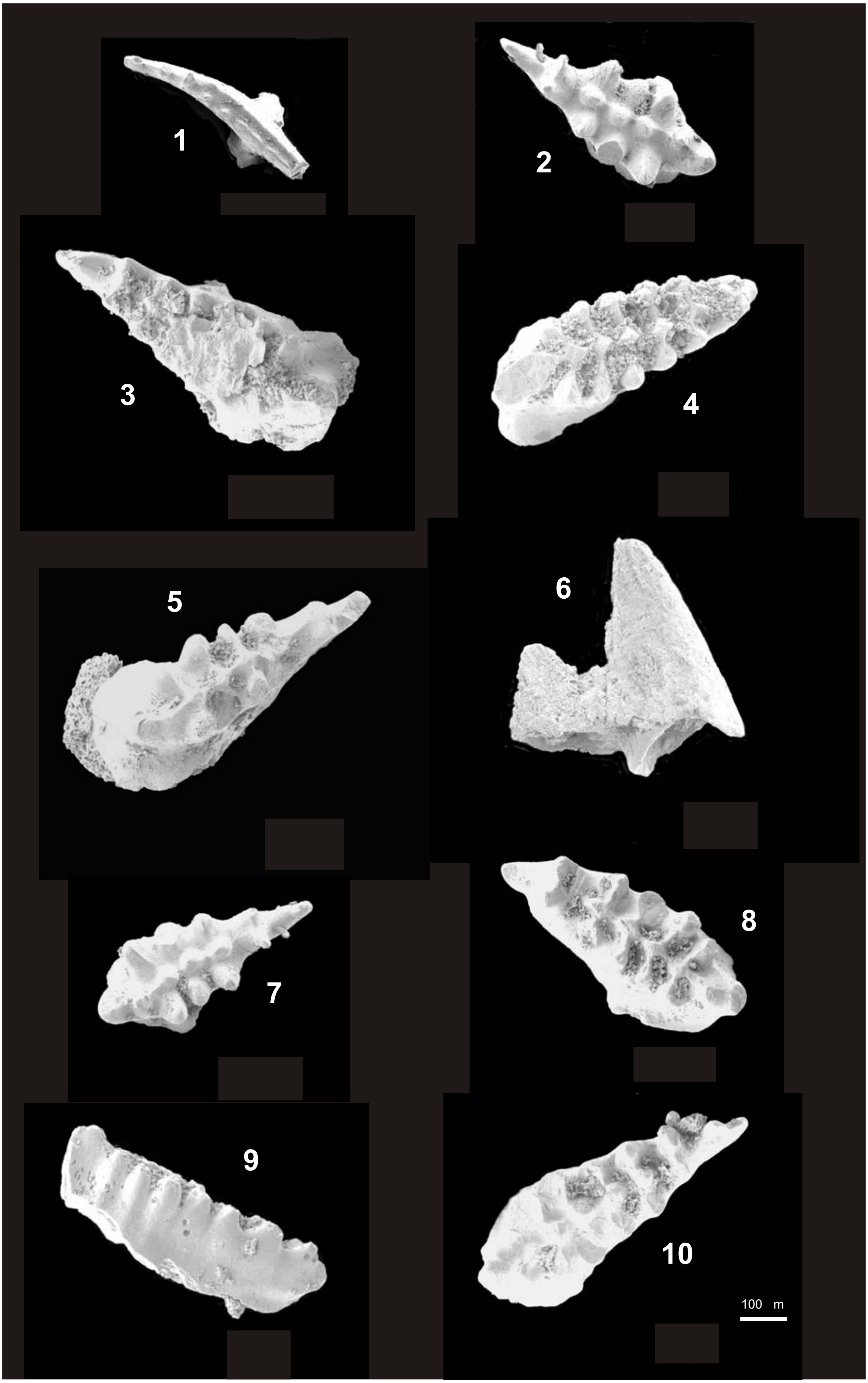
Between the 58-65 m marks of section G-G´ there is a conspicuous horizon of gray, reddish brown weathering sandstone. The sandstone horizon seems to have a gradual lower boundary and a well-defined upper boundary. Hayden (1904) described two similar sandstone horizons in the Lipak Fm. at the type section. The beds above the sandstone incursion up to bed L_c173 comprise calcarenite, with abundant ooid limestone and increased grain size and fossil content in upper parts (Fig. 4.21).

The black carbonaceous shale of bed L_c174 represents a strong and abrupt lithological break, possible marking the Upper Kehlwasser Event. The beds above the black shale up to L_c208 are made of nearly the same calcarenite as below, but contain less ooids; in upper levels some thin sandstone beds occur.

Beds L_c210-L_c219 comprise the upper sandstone horizon in the Lipak formation of this section. It consists of gray, reddish brown weathering sandstone, very similar to the lower sandstone horizon and similarly it also shows a gradual lower boundary and an erosional upper boundary. Above the erosional surface the next 11 m are made of gray calcarenite; fossils are rare and mainly consist of crinoid stems.

The uppermost 25 m of the Lipak Fm. in this section are dominated by black mudstone, except beds L_c279-L_c289 that comprise calcarenite. Hardly any macro-fossils have been found in this part. The upper and lower boundaries of bed L_c305 are very irregular; the bed is made of pink, crystalline calcite and is interpreted as horizon of emergence.

In the uppermost part of the section, the Lipak Fm. is truncated by the basal breccia of the overlying Gechang Fm. The contact is represented by an angular unconformity, representing the lack of nearly the whole Carboniferous deposits in this area.


1
2
3
4
5
6
7
8
9
10
100 m

5. SEDIMENTARY STRUCTURES IN THE MUTH FORMATION

5.1. General remarks

The Muth Fm. with its striking lithological monotony, its thickness of more than 250 m and several 100 ´s of km lateral extend is one of the largest quartz sand accumulations in the geological record and there are hardly any analogue sedimentary units in the world. One of the rare comparable examples is represented by the Ordovician Peninsula Fm. (Table Mountain Group, South Africa) comprising 750 m of mainly quartz arenites; the depositional setting of the Formation has been interpreted as barrier island environment (Hobday & Tankard 1978).

Interpretations of the depositional environment of the Muth Fm. are still controversial because investigation is complicated by the monotonous lithology and Himalayan deformation. Inner- to mid shelf environments (Shanker *et al.* 1993), shallow marine littoral environments (Kumar *et al.* 1977, Bhargava & Bassi 1986, Bagati 1990, Gaetani & Garzanti 1991), shallow marine re-sedimentation of desert sands (Mukherjee & Dasgupta 1972) to eolian terrestrial settings (Dasgupta 1971; Garzanti *et al.* 1996b) have been suggested in the literature.

The Muth Fm. has been investigated in the Pin Valley and detailed bed-by-bed sections near Mikkim and at the Muth type locality have been made (Figs. 4.1, 4.10). Although there are some differences in detail, the sections can readily be correlated, because of the wide lateral extend of the general architecture of the Formation. The two longest sections (A-A´ and E-E´) display a distinctive lower boundary to the shoreface deposits of the uppermost Pin Fm. and a gradual transition to the overlying shelf carbonates of the Lipak Fm. With exception of a thin horizon of sandy/silty dolomites in the uppermost third of the section, the Muth Fm. in the Pin Valley comprises super mature quartz arenites with both, high textural and compositional maturity. Even of the heavy minerals only zircon, well rounded bluish tourmaline and minor xenotime occur, which are considered as very stable during sedimentary processes.

Quartz usually comprises more than 99 % of the detrital clasts, this is definitely higher than the highest quartz contents for example within the Galveston barrier complex (Texas) and a Lower Cretaceous barrier system at Bell Creek Oil Field (Montana) found in the eolian sediments, with 80 % and 92 %, respectively (Davies *et al.* 1971).

According to the Bagati (1990) the Muth Fm. was subjected to sedimentary burial by some 4000 m of Upper Paleozoic to Mesozoic sediments. According to Wiesmayr (2000) temperature did not exceed 200° C. Of course, due to diagenesis the present composition is appreciably different from the composition at the time of deposition; *e.g.* oversize pores (Schmidt & McDonald 1979), now filled mainly with quartz cement (*e.g*. sample ED97/145) indicate complete dissolution of detrital grains, of which the majority probably were feldspars (McBride 1985). In the Muth Fm. oversize pores hardly constitute more than a few percent therefore the rock composition at the time of deposition probably was not very different from the present quartzarenites (classification of McBride 1963), indicating high-energy depositional environment for the Muth Fm. The relative rare occurrence of pressure solution indicates early cementation; possible sources for the quartz cement are discussed in McBride (1989).

In the below presented interpretation the Muth Fm. was deposited in a barrier-island system (for Holocene examples see Davis 1992, Davis 1994). The arguments leading to this conclusion are discussed below. Even at the first general approach, the enormous sand accumulation, the big lateral extend and the stratigraphic position, sandwiched between two marine formations, supports the barrier island model. Based on lithological variations and different sedimentary structures, four facies associations (FA) have been distinguished (beginning at the base):

FA1: HORIZONTALLY LAMINATED AND MINOR CROSS-BEDDED QUARTZ ARENITE
FA2: LARGE-SCALE CROSS-BEDDED QUARTZ ARENITE
FA3: DOLOMITE AND SANDY DOLOMITE
FA4: MIXED CROSS-BEDDED AND HORIZONTALLY LAMINATED QUARTZ ARENITE

The present facies differentiation is a simplified approach based on the sections that have been measured (Fig. 4.10). Although these sections provide a lot of valuable information, they represent just 2-dimensional insight into the sedimentary structures of the Muth Fm. Barrier-island systems typically display a wide variety of depositional structures resulting from the interaction of processes driven by wind, waves and tides, showing a diversified pattern of depositional environments, therefore a more complex 3-dimensional sedimentary architecture has to be expected. Bagati *et al.* (1991) have reconstructed a NNW-SSE trending shoreline for the Spiti area during Ordovician time. Paleocurrent directions in the Muth Fm. during its time of depositional also indicate a similar NW-SE trend.

The understanding of the sedimentary structures of the Lower Devonian barrier island system of the Muth Fm. might provide new insights into this type of depositional environment, that represents one of the potentially most productive sand bodies for hydrocarbons (Davies *et al*. 1971). Sections A-A´, C-C´ and D-D´ of the Muth Fm. have a restored distance to the proximal type section E-E´ of 13 km at right angle to the strike of overall facies trend (Wiesmayr 2000) and therefore provide valuable information about proximal/distal facies variations in this direction.

Grainsize investigations have been carried out, but did not provide the expected success. Due to the syntaxial quartz cement around the quartz clasts, the primary shape of the grains is hardly ever visible, except thin dust layers prior to cementation have coated the grains. Especially the grains of the finer fraction have been difficult to measure. Therefore no detailed grain size analysis is presented. The range of grainsizes in the whole Formation determined with hand lenses in the field is strictly within the sand fraction, only FA3 comprises finer-grained sediments. Some exemplary grainsize analyses have been carried out by a Carl Zeiss KS 300 Imaging System, by measuring the longest grain diameters. The obtained thin section data have been calibrated for conventional sieve data by a Mathematica® notebook developed by Bernhard Grasemann and Peter Faupl.

5.2. Paleocurrent directions in the Muth Formation

Special attention has been directed on paleocurrent indicators in the Muth Fm. Due to the way the Muth Fm. erodes in this area, beds are mainly visible in cross-section, but hardly any set boundaries are visible in plane view (Figs. 4.11, 4.13). In places where ripples have been found, they hardly gave reliable paleocurrent directions. Therefore cross-bedding has been used as easily measurable paleocurrent indicator and the data are presented in plots showing the dip directions of the foresets, which have been restored together with the bedding surfaces (Figs. 5.1-5.3).

The orientation of the outcrop, respectively the orientation of the section, relative to the orientation of the foreset can have a significant influence on the obtained data. For example, on E-W trending outcrops N- ward or S-ward dipping tabular cross-beds would appear as horizontally laminated sets and E- or W-ward dipping foresets would be over-represented. In the case of the Muth Fm. this seems to have a minor effect, because the two longest sections have a different orientation (section A-A´ trends E-W and section E-E´ trends SW-NE), resulting in the good comparability of paleocurrent directions of both sections.

No detailed investigation of the vertical variations of the paleocurrent directions of the Muth Fm. have been carried out so far, nevertheless several authors mention some dominant directions. Among the Lower Paleozoic Formations the Muth Fm., but also the Shian Fm. comprise abundant cross-beds and therefore are well suitable for paleocurrent studies. Bagati *et al*. (1991) divided the Shian Fm. in the Pin Valley into 3 parts, a lower fluvial environment with dominant dip direction of planar foresets towards ENE, a middle fluvial-marine sequence with main dip azimuth towards NNW but also a strong ESE direction and finally a marine upper part with a prevailing E-ward paleocurrent direction. Kumar *et al.* (1977) describe a main NE- to E-ward paleocurrent direction for correlatives of the Shian Fm. (90 readings) and variable directions with a small maximum towards ENE for the Muth Fm. (29 readings) in Kumaon.

Gaetani *et al.* (1986) mention a dominant E-ward directed paleoflow direction in the Muth Fm. of Zanskar, Bhargava & Bassi (1998) measured 30 foresets in the Muth Fm. and state a SE-ward paleocurrent direction in the Pin Valley, but a broad NW-ward direction in the rest of Spiti.

The paleocurrent measurements in the Muth Fm. in the Pin Valley of this study showed distinct variations of mean flow directions related to the vertical position of the measurements in the sections. As a result, the data presented in the literature without indication of their vertical position are regarded of limited value for interpretation.

In the present investigation dip directions of in total 255 foresets have been measured in sections A-A´ (95 readings) and E-E´ (165 readings). The foreset orientations have been measured simultaneously with the work on the sections; therefore changes in the dominant paleocurrent directions can be related to their positions in the sections. Based on variations in sedimentary structures and lithology the Muth Fm. has been divided into four facies associations (Fig. 4.10) and they also differ in their dominant paleocurrent directions. Sections A-A´ and E-E´ show similar trends. Directional data of foresets are presented in rose diagrams showing dip directions of foresets and in stereographic plots with poles of foreset surfaces to provide additional information about dip angles (Figs. 5.1-5.3).

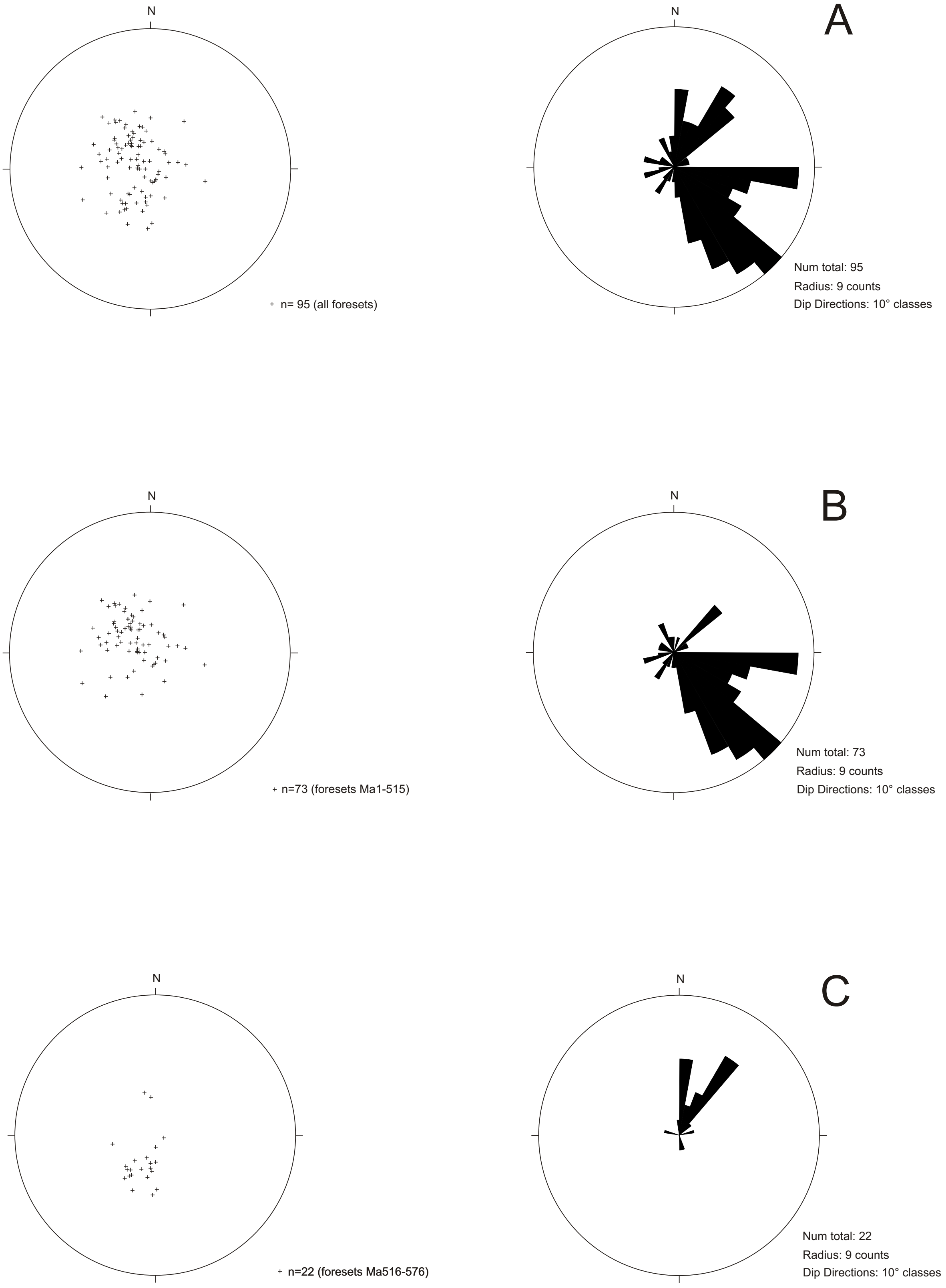


Fig. 5.1: Dip directions of foresets in the Muth Fm. in section A-A´, SE Mikkim. Stereonet plots of forests in the left column give an idea of variations of the dip angles; equal area lower hemisphere projection of foreset poles. All the foresets are restored together with the bedding surfaces. **A)** All measurements of section A-A´. **B)** Foresets of Facies Association 1 (beds M_a5-M_a502) indicate a dominant SE directed paleocurrent. **C)** Foresets of Facies Association 2 (beds M_a517-M_a576) indicate a mean NE directed paleocurrent.

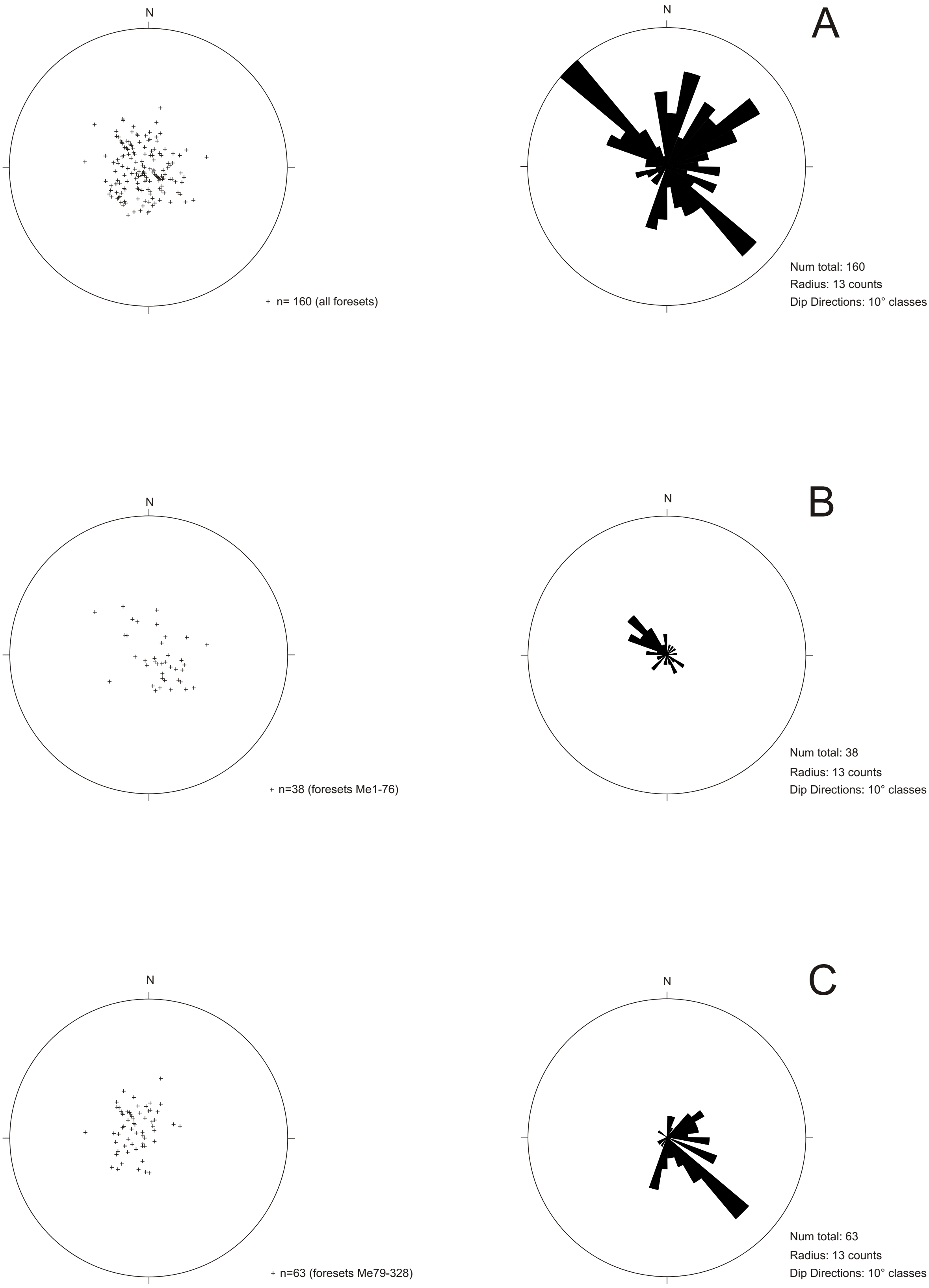


Fig. 5.2: Dip directions of foresets in the Muth Fm. in section E-E´, S Muth. Stereonet plots of forests in the left column give an idea of variations of the dip angles; equal area lower hemisphere projection of foreset poles. All the foresets are restored together with the bedding surfaces. **A)** All measurements of section E-E´. **B)** Foresets of the lower part of Facies Association 1 (beds M_e1-M_e76) indicate a bipolar NW-SE directed paleocurrents. **C)** Foresets of the upper part of Facies Association 1 (beds M_e79-M_e328) indicate broadly a SE-E directed paleocurrent.

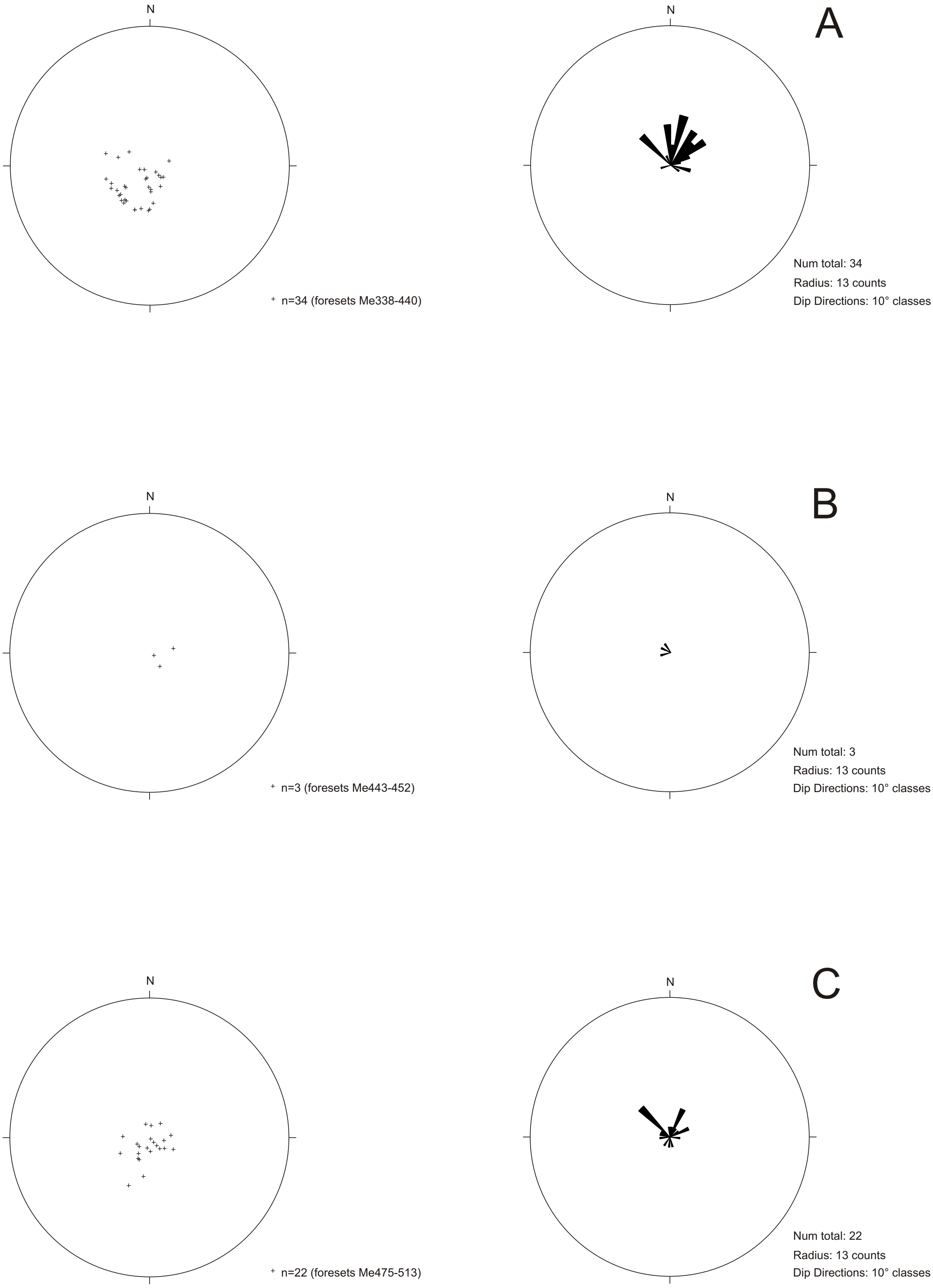


Fig. 5.3: Dip directions of foresets in the Muth Fm. in section E-E´, S Muth. Stereonet plots of forests in the left column give an idea of variations of the dip angles; equal area lower hemisphere projection of foreset poles. All the foresets are restored together with the bedding surfaces. **A)** Foresets of Facies Association 2 (beds M_a338-M_a440) indicate broadly a dominant N directed paleocurrent. **B)** Foresets of Facies Association 3 (beds M_a443-M_a452) are statistically not significant. **C)** Foresets of Facies Association 4 (beds M_a475-M_a513) show very variable directions.

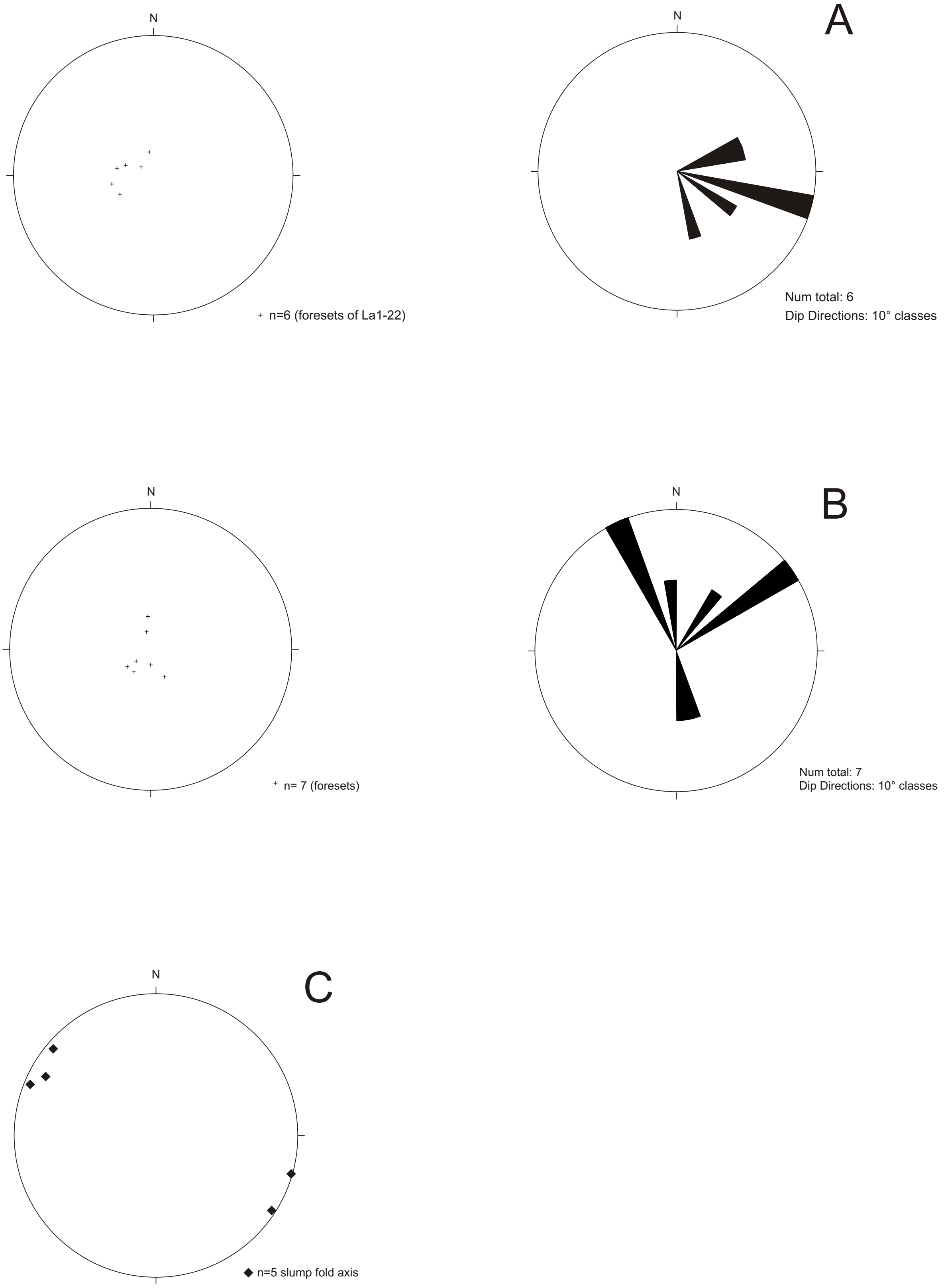


Fig. 5.4: A) Dip directions of foresets in the lower Lipak Fm. in section B-B´, east of Mikkim. All the foresets are restored together with the bedding surfaces. Dip of foresets broadly indicates E to SE directed paleocurrents. **B)** Dip directions of foresets in the basal Lipak Fm. in section E-E´, south of Muth. indicate N to NE paleocurrent directions. **C)** NW-SE trending fold axis of slump folds in the Pin Fm. (bed P_a1) in section A-A´ SE of Mikkim.

FA1 in section A-A´ shows a mean paleocurrent direction towards the SE (Fig. 5.1A). FA1 in section E-E´ gave similar trends, but additionally the facies can be divided in a relative thin lower part with NW-ward paleocurrent direction and a thick upper part with indicators for SE-directed paleoflow (Figs. 5.2B, 5.2C). These SE and NW directed paleocurrents are interpreted as coast-parallel, possibly wave-generated currents.

FA2 that is dominated by large-scale cross-beds shows corresponding paleoflow directions towards NNE (*i.e*. offshore directed) in both sections. In section A-A´ the direction is relatively well defined, but in section E-E´ the directions scatter around N-NE (Figs. 5.1C, 5.3A). Paleogeographic reconstructions of Northern India show the position of this area at about 30° southern latitude, with two possible implications: (1) Nowadays these latitudes receive very little precipitation resulting in vast desert areas (Allen 1997). Assuming similar conditions for the depositional area of the Muth Fm. this could explain the high availability of sand. (2) Considering similar organization of atmospheric circulation cells during this time the Muth Fm. would have been situated in areas dominated by trade winds (Allen 1997) that might be responsible for the consistent paleocurrent directions of the eolian dunes in FA2.

The upper two facies associations of the Muth Fm. are not exposed near Mikkim and foresets have been measured in the Muth section, but only little measurements have been possible, therefore the directions are considered statistically insignificant and an interpretation is hardly possible. Just three foresets have been measured in FA3; they indicate a W-ward paleocurrent direction (Fig. 5.3B). The dip directions of FA4 scatter around in all directions, badly defined maximums indicate NW-ward and NNE-ward directions (Fig. 5.3C), similar trends are also found in the lower part of the overlying Lipak Fm. (Fig. 5.4B).

5.3. Sedimentary structures in facies association 1

There is a sharp erosive contact, with small relief, between the Muth Fm. and the underlying Pin Fm., of which the uppermost bed shows a distinct orange-red color (Figs. 4.8, 4.16), probably indicating a depositional break.

FA1 solely comprises pure quartz arenites, with a thickness of 205 m in Mikkim and 129 m in Muth. Bed thickness varies between 3-360 cm (Fig. 5.5), with an average of 50-39 cm (Mikkim/Muth respectively). Grain size consistently remains in the sand range, mainly about fine to medium-grained sand. In the lowest part of FA1, upper bedding surfaces of horizontally laminated beds (beds M_a5-M_a8 and M_a10-M_a13) show surface structures strongly resembling plant root traces(Fig. 4.17). These traces form tunnel shaped *c*. 0.5-1 cm wide, irregular curving, sometimes branching and tapering structures that penetrate some 5 cm into the bed. Although a formation by faunal burrowing activity cannot be totally excluded, these structures are interpreted as root traces, similar to those described from the Early Devonian of Québec by Elick *et al.* (1998).

Set boundaries are usually even, parallel and well defined, low-angle to horizontal lamination is the dominant sedimentary structure; cross-stratification is rare, but with increasing frequency towards younger beds. Where cross-bedding occurs, it is usually high angle tabular-planar,

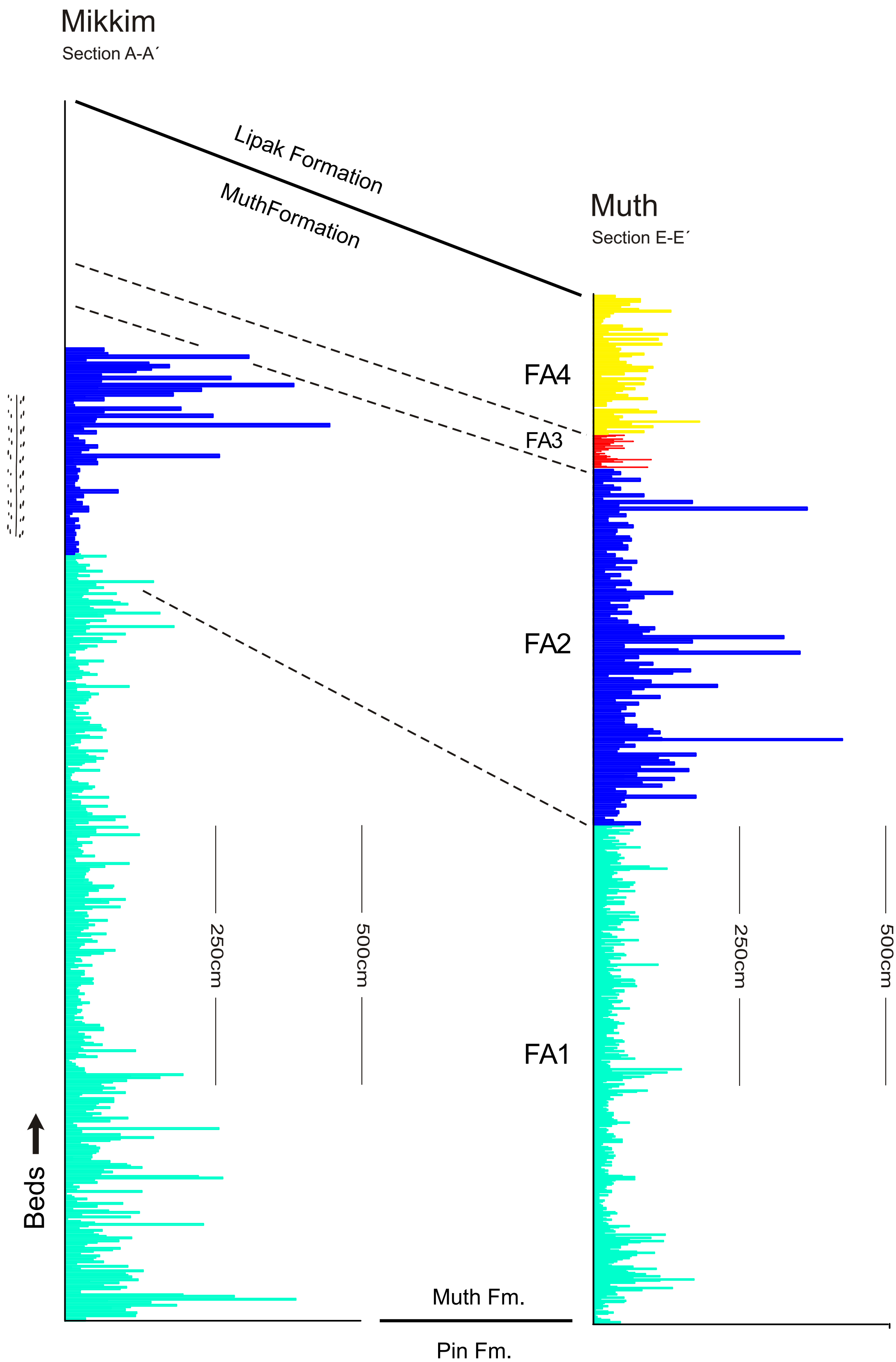


Fig. 5.5: Bed thickness variations in the Muth Fm. SE of Mikkim (section A-A´) and South of Muth (section E-E´). Bed thickness variations broadly correlate in both sections. The thickest beds are found in FA2. Trackway symbols near the left column indicate the occurrence of arthropod trackways in section A-A´.

sometimes tangential, concave-up. Small-scale trough cross-bedding is very rare. Reactivation surfaces are quite common in upper parts of FA1, often in beds with indications for paleocurrent reversals, sometimes forming complex patterns of cross-bedding like in beds M_a465-469. Typical herringbone cross bedding has been found rarely. Upper surfaces of beds are hardly ever exposed, therefore additional sedimentary structures like symmetrical, asymmetrical and flat-topped ripples have been found rarely.

In section A-A´, SE of Mikkim, FA1 is truncated by a small local angular unconformity (Fig. 4.12), which is also related with a change in sedimentary structures and the main paleocurrent direction (Fig. 5.1).

5.3.1. DEPOSITIONAL ENVIRONMENT OF FACIES ASSOCIATION 1

The interpretation of FA1, which lacks good environmental specific sedimentary structures, is difficult. On the base of obtained data it is interpreted as beach, lower foreshore and upper shoreface deposits on a coast with abundant sand supply. Small-scale cross-bedding might have been formed as longshore bars. Sedimentation was strongly influenced by wave-generated longshore currents with minor tidal impact. The amazing thickness of pure quartz arenites throughout the entire FA1 indicates a relatively stable balance of sea level, sediment supply and tectonics during its deposition.

The sharp erosive boundary between the Muth Fm. and the underlying Pin Fm., of which the uppermost bed shows a distinct orange-red color in section A-A´ can be explained in two ways. (1) A sub-aerial exposure of the Pin Fm. due to a relative sea level fall created a disconformity. During the subsequent relative sea level rise, a ravinement surface in terms of sequence stratigraphy was produced. (2) A regression of the siliciclastic nearshore Muth Fm. resulted in the basinward movement of the shoreline. This “forced regression” produced a sharp contact between shoreline sands overlying shelf carbonates, a so-called “regressive marine surface of erosion” (Posamentier *et al.* 1992).

The absence of unequivocal observations makes a decision between the two possibilities difficult. The lack of microfacies evidence, real karst, a basal angular unconformity, or reworked Pin Fm. clasts supports a "forced regression" interpretation, although many of these missing observations could be explained by an arid climate during the sub-aerial exposure. The disconformity model is attractive because it explains the enormous theoretical time span for the deposition of the Muth Fm. between the top of the Pin Fm. and the base of the Lipak Fm. if we assume that the unconformity represents a considerable time-gap. Furthermore this model fits also very well to beach sediments at the base of the Muth Fm., assuming that the interpretation of root traces (Fig. 4.17) in this level is correct.

Fig. 5.6: Upper bedding surface of a crossbed of FA2 in the Muth Fm. in the ravine south of Muth shows foresets of dune cross-bedding in plan view. Photo width *c*. 40 m.

Fig. 5.7: Regular, finely laminated dune foresets in the Muth Fm. (FA2) SE of Mikkim possibly represent daily cycles in coastal dunes. Arrow points to lens cap.

Fig. 5.8: Flat-topped ripples on a loose block in the Muth Fm. South of Muth indicate emergent conditions during formation.

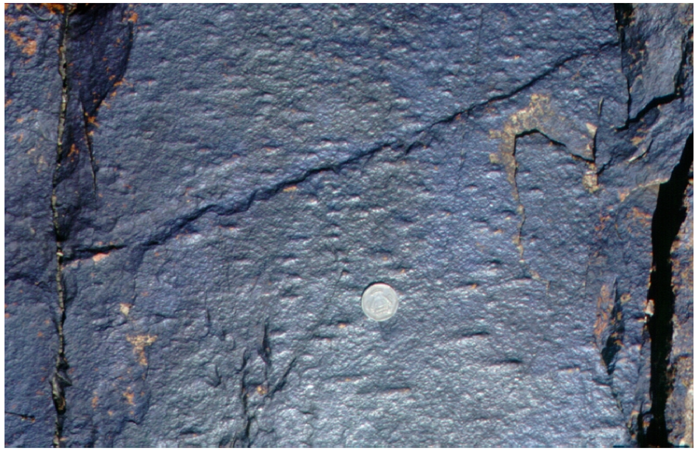


Fig. 5.9: Tear-shaped ridges on an upper bedding plane in the Muth Fm. (FA1) SE of Mikkim are a common feature of wind erosion on modern beaches. Wind from right to left.

5.4. Sedimentary structures in Facies association 2

In analogue to FA1 below, the lithological character of FA2 is very uniform, comprising fine to coarse-grained pure quartz arenite, with the bulk in the medium sand fraction. Beds directly below FA3 show a slightly decreased purity, indicated by low mud content of the arenites; sometimes like in bed M_e437 displaying a texture of pore-filling infiltrated clay (Fig. 4.14), see Wahab (1998, Fig. 4A).

The thickness comes to some 60 m in Mikkim (uppermost part not exposed) and reaches 87 m in Muth. Bed thickness is much higher than in the other facies, varying between 6-455 cm and averaging 70-81 cm (Mikkim/Muth); the average value is strongly influenced by the thin beds in the lowest part of the facies (Fig. 5.5). The occurrence of the arthropod dominated ichnofauna is restricted to FA2 in the Mikkim section (Draganits *et al*. 1998a).

This facies is characterized by large-scale cross bedding, whereas horizontal lamination is rare and nearly exclusively confined to the lowest part. Bed boundaries are well defined and are even parallel, but also wedge-shaped sets occur. SSW of Muth upper bed surfaces are exposed and single, slightly curved foresets can be traced for more than 20 m in plan-view (Fig. 5.6), similar to truncated barchan dunes in the White Sands, New Mexico (McKee *et al.* 1971). Foresets of cross-bedded beds are *c.* 25 cm in average thick (Fig. 5.7), concave-up, tangential and sometimes display bimodal lamination of fine and coarse sand; some laminae display reverse grading, similar to the eolian pin-stripe lamination described by Fryberger & Schenk (1988). The angle between the foresets and bedding surfaces often exceeds 30° (Figs. 5.1C, 5.3A), the dip direction of the foresets is relative uniform towards NE-NNE (*i.e*. offshore directed). Arthropod trackways occur on foresets as well as on bedding surfaces.

Additional sedimentary structures on bedding surfaces include symmetrical ripples, interference ripples, adhesion ripples, donut-shaped structures (see Chapter 5.7.2. for details) and tear-shaped ridges (Fig. 5.6 and McKee 1957), *i.e*. scour-remnant ridges by Allen (1965), as well as rare sand-flows structures on foresets (Fig. 6.24). Domal build-ups occur in some levels within FA2 in the Muth Fm. near Mikkim that have been interpreted as stromatolites (for details see Chapter 5.4.1.). Additional structures, which are related to microbial activity, are desiccation structures (Fig. 5.17) resembling those of desiccated biofilms (pers. comm. Gisela Gerdes 1999). Severe post-depositional de-watering is indicated by sandstone diatremes found in the uppermost beds of this unit (Figs. 5.18-5.24), directly below FA3 (for details see Chapter 5.4.2.).

5.4.1. SEDIMENTARY STRUCTURES RELATED TO MICROBIAL MATS

The lithological monotony of the Muth Fm. has been mentioned several times before; thus the keys for understanding the sedimentology of the Formation are the sedimentological structures. Besides structures like ripples, cross-beds, *etc*. that are manifestations of physical sedimentary processes, other structures have been found that are indications for microbial activity, although no organic matter has been preserved. All of these structures described below have been found in

Fig. 5.10: Irregular rough bedding surface at the base of bed M_c4 (section C-C´). The pustular-wrinkled appearance of this surface might be related to microbial mats. Lens cap is 53 mm in diameter.

Fig. 5.11: The bedding irregularities of the wrinkled surface of Fig. 5.10 gradually become larger and evolve into domal stromatolites. Hammer is 57 cm long.

Fig. 5.12: Simple build-ups in lower levels of the stromatolite interval have an asymmetric shape indicating paleocurrent from right to left (NE to SW). Hammer is 57 cm long.

Fig. 5.13: General view of the domal stromatolites. Note the continuos lamination of these structures. Hammer is 57 cm long.

Fig. 5.14: Geology can be that exciting! Note the large size and regular shape of the domal quartzite stromatolites. Muth Fm., section C-C′ (bed M_c6).

Fig. 5.15: Vertical section of a domal stromatolite showing irregular lamination of white and gray quartzite.

Fig. 5.16: Domal stromatolite directly followed by a large-scale cross-bed with more than 3 m thickness in the eastern part of the outcrop. Hammer is 57 cm long.

Fig. 5.17: Irregular polygonal fracture pattern in bed M_a566 resembles those of desiccated biofilms. The 1 Rupee coin is 25 mm in diameter.

section A-A´ and indicate at least in places former organic content in the recent pure quartz arenites of the Muth Fm.
The lower part of bed M_a566, which also contains abundant arthropod trackways, shows irregular polygonal fracture patterns on the upper bedding surface (Fig. 5.17), at first glance similar to desiccation cracks in fine-grained sediments. Cracks break the surface into *c*. 2-6 cm large irregular, but relative isometric fragments resembling the shrinkage pattern of desiccated microbial mats of Gerdes & Krumbein (1994, Fig. 9b). These structures are thought to have been formed on surfaces that are episodically wetted by tides or by sufficient pore water supply from below. The structures appear in medium sand-sized quartz arenite and therefore simple mud cracks can be excluded.

Another enigmatic structures found on upper surfaces of horizontally laminated beds between large-scale cross-beds of FA2 comprise gently rising mounds on level surfaces; they can reach sizes of *c*. 80 cm diameter with heights of some 30 cm. In cross-sections no internal structures have been found, the arenite has a massive appearance. Crinkled surfaces of these beds might represent indication for microbial activity, the occurrence of pyrite possible point to the former presence of organic matter.

Whereas the above mentioned surface structures represent relatively vague indications for microbial activity, at the Eastern limb of the anticline SE of Mikkim other and even more impressive manifestations of biogenic structures have been found in the upper part of FA2. There, two horizons (appendix B8) show irregular domal build-ups (Figs. 5.10-5.15) growing on level bedding surfaces as spaced lateral-linked-hemispheroids (LLH-S) evolving to large and regular close lateral-linked-hemispheroids (LLH-C; Logan *et al*. 1964).

Beds with these structures are up to 2 m thick, in basal parts the build-ups are small, irregular shaped and spaced (Fig. 5.11), with the trend of becoming larger and more regular towards higher levels. In bed M_c4 the stromatolites are interlayered by a thin horizontal quartzite horizon, that shows no erosion at the base, it seems to cover the domes more or less like a blanket and levels the height differences of the domes. Domes are up to 110 cm in diameter and some 40 cm high and linked by continuos, undulating laminae (Figs. 5.13, 5.14), closely resembling the domal stromatolites presented by Schieber (1998, Fig. 22). Angles between domal surfaces and bedding surfaces average c. 30°, but can also show values up to 45°.

The whole structures consist of quartz arenite; the lamination is made of wavy, irregular alternations of white and grayish laminae (Fig. 5.15). The white parts are made of quartz grains with quartz cement. The grayish layers also consist of quartz grains, but in contrast the matrix comprise micro crystalline aggregates, which are not identifiable by light-microscopy, additionally to quartz.

Although unequivocal evidence for microbial origin is very difficult to find in terrigenous clastic sedimentary environments like the Muth Fm., the shape of the domal build-ups, the wavy-undulose appearance of their laminae and cohesive behavior of some laminae make a microbial origin probable (Schieber 1998, 1999). The present lack of organic matter might be explained by elutriation and oxidation in the originally highly permeable sediment (Gevers & Twomey 1982).

Besides the above described occurrence of domal stromatolites in the upper part of FA2 (sections C-C´ and D-D´) similar structures have also been found in the upper part of FA1 (section A-A´, beds M_a246, M_a410 and M_a430). There, the stromatolites are less well developed than at the Eastern limb of the anticline, but still resemble many features of microbial mats, possibly indicating the close relation of the depositional environments of FA1 and FA2.

5.4.2. Depositional environment of facies association 2

Based on the sedimentary structures facies association 2 is considered to represent shallowest foreshore, backshore and coastal dune sediments. Several distinct sedimentary structures are thought to be indicative for shallow intertidal to supratidal conditions. Desiccated biofilms might be typical for this environment; the domal stromatolites possibly grew in shallow intertidal areas with slightly reduced water energy.

Tear-shaped ridges are up to 4 cm long, 0.2 cm high and 0.7 cm wide and display a streamlined shape (Fig. 5.9), similar to much larger yardangs described by Ward & Greeley (1984), representing a common surface structure found on modern beaches (Allen 1984). They are erosional in origin and are related to eolian deflation, thus indicating at least temporarily emergent conditions (McKee 1957, Berry 1973, Carter 1978, Allen 1984). Adhesion ripples probably form under similar conditions as tear-shaped ridges. Another indicator for coastal environments represents the well-developed beach lamination, *e.g.* in bed M_d12 (ED98/214).

The intimate depositional relationship of shallow marine deposits with the large-scale cross beds that are interpreted as coastal dunes, is indicated by the direct succession of nicely developed domal stromatolites by a more than 3 m thick large-scale cross bed without any sign of erosion (Fig. 5.16). Rare sand flows (Fig. 6.24), high foreset dip angles (Figs. 5.1C, 5.3A), the depth of the arthropod tracks (Figs. 6.14, 6.17, 6.19A) and the relation of the cross-beds with beds with above mentioned indicators for at least temporarily emergent conditions represent major arguments for a marginal marine interpretation of these beds.

Both, *Diplichnites* and *Diplopodichnus* traces are generally considered to have been produced by myriapods, their differences resulting from different surface conditions, favorable differences in surface moisture (Buatois *et al.* 1998b, Johnson *et al.* 1994). This is again an argument more for at least partly sub-aerial depositional environment of FA2.

In the literature the occurrence of frosted quartz grains has been mentioned (*e.g*. Bhargava & Bassi 1998) and related to certain (eolian, beach) sedimentary transport mechanisms. These structures are not regarded conclusive for a special environment, because in the case of the Muth Fm., this might also be explained by chemical leaching of the grain surface (Pettijohn 1957). Bed M_a554 shows indications for local soft sediment deformation. The sandstone diatremes in large-scale cross-beds of section E-E´ are nearly identical to the structures in Late Precambrian eolian dunes of western Mali described by (Deynoux *et al.* 1990).

5.5. Sedimentary structures in facies association 3

Facies association 3 is well exposed near Muth; in section E-E´ the basal contact is represented by a sharp lithological break of fine-grained dolomite above quartz arenite of the underlying FA2 (Fig. 4.13; appendix B14). FA2 is virtually the only horizon in the whole Muth Fm. with grain sizes distinctly smaller than in all other facies and ranges from clay to medium sand size. Beds consist of very fine-grained dolomite, silty/sandy dolomite and minor sandstone/quartzite beds. Towards younger beds, the number of quartzite beds increases and the contact to the quartz arenites of FA4 is gradual.

Near Muth FA3 is only 11 m thick. Bed thickness ranges from 1-98 cm with mean thickness of 31 cm (Fig. 5.5), bed boundaries are usually even parallel and well defined. Fine lamination is a common internal structure; ripple lamination and cross-bedding is rare. Bed M_a442 comprises intraformational breccia in basal parts. Not even one fossil has been found in these beds and bioturbation is absent. In the upper part of FA3 some sandstone diatremes occur.

5.5.1. Depositional environment of facies association 3

Hardly any good indicators for its depositional environment have been found in FA3, therefore the interpretation below is mainly based on its position within the Muth Fm. This horizon is regarded as lagoonal deposits within the barrier system of the Muth Fm.

Dolomitic mudstone beds indicate a low-energy environment. The total lack of fossils in the dolomite and the lack even of trace fossils, which results in the preservation of the fine lamination, possibly indicate hyper-saline water conditions. Quartzite beds within FA3 may be explained as washover fans. If this interpretation is correct, the observation of lagoonal sediments above dune deposits indicates a rise in sea level in combination with progradation due to high sediment supply, resulting in a “depositional regression” (Curray 1964).

5.6. Sedimentary structures in facies association 4

FA4 is well exposed near Muth, where it has a gradual contact to the underlying FA3. Quartz arenites predominate, but they are slightly less pure than in FA1 and 2 and they show a tendency to become less pure in the upper part. The upper contact to the Lipak Fm. is gradual. The thickness of this facies is 33 m in Muth, with bed thickness ranging from 12-180 cm, with a mean of 55 cm (Fig. 5.5).

Both horizontal lamination and cross-stratification occur. Lower parts show some beds with large-scale cross-beds, resembling those of FA2 and in analogue to them they show a NNE-ward paleocurrent direction. The middle part is dominated by cross-beds with variable dip directions, followed by mainly horizontally laminated beds with burrowed upper surfaces in the upper part.

Small-scale sedimentary structures like ripples are scarce. The upper surfaces of beds M_e511 and M_e522 show remarkable hollows of unclear origin.

5.6.1. DEPOSITIONAL ENVIRONMENT OF FACIES ASSOCIATION 4

The lack of univocal environmental indicators leaves the interpretation of FA4 unclear. Due to its position above probable lagoonal sediments below and the proximal offshore basal part of the Lipak Fm. above, it possibly represents transgressive sediments with foreshore to shoreface elements, indicating an effective relative sea level rise in the uppermost part of the Muth Fm. The large-scale cross-beds in the lowermost part might indicate the preservation of sediments similar to FA2.

5.7. De-watering structures

5.7.1. VERTICAL SANDSTONE DIATREMES

In the Muth Fm. de-watering structures have been found in several places, the most spectacular of them, sandstone diatremes (Fig. 5.18-5.24), are described here in detail. Well-developed sandstone diatremes have been found in section E-E´ near Muth (Fig. 4.13; appendix B14). Most of these structures are found in the uppermost part of FA2 (eolian dunes), with the highest abundance directly below FA3 (lagoonal sediments), although some more appear in FA3 and a few more have also been found in section A-A´ about 10 m below near bed M_a423. This restriction to just a few 10´s of meters of the section possibly indicates a relation of these structures to the depositional environment. Several names have been proposed for this structure in the literature (Deynoux *et al.* 1990 and references cited therein), the use of the relative descriptive term “sandstone diatremes” is preferred in the present work.

Sandstone diatremes in the Muth Fm. are relatively uniform, although differences can be found in details. In general they comprise vertical pipe-shaped structures, diameters ranging between 2-80 cm and lengths between 0.05-2 m with crosscutting relationship to the surrounding host rock that consist of large-scale cross-beds. These structures are usually cylindrical, but in rare cases some small conical shapes have also been found with the apices pointing downward. Grainsize in- and out-side of the sandstone diatremes is in principal the same and seems to be controlled mainly by the availability of grainsizes.

The sandstone diatremes directly below FA3 have been investigated in detail. Most of them originate directly above the upper boundary of bed M_a437 (appendix B14), that comprises extremely well-rounded and well-sorted quartz grains with a clay matrix (metamorphosed to chlorite), which shows a texture of pore-filling infiltrated clay (Fig. 4.14). This bed shows very irregular wavy set boundaries possibly indicating deformation related to de-watering. Nevertheless, no clear relationship of this bed with the sandstone diatremes directly above has been found.

Fig. 5.18: On the upper bedding surface of bed M_e440 sandstone diatrems are eroded below surface level forming large circular pits. Hammer is 57 cm long.

Fig. 5.19: Detail of a pit showing slightly domed rims and several thin sand-fissures. Hammer is 57 cm long.

Fig. 5.20: Nicely developed sand diatreme cross-cutting through eolian cross-bed of M_e438 with sharp boundaries to the host rock.

Fig. 5.21: Oblique section of a sandstone diatreme in bed M_e438. Note the sharp boundary to the foresets of the hostrock and the internal fine lamination at the rim. Lens cap is 53 mm in diameter.

Fig. 5.22: Muth Fm. south of Muth. Lower bedding surface of bed M_e438 showing abundant axial sections of sandstone diatremes of various sizes. Hammer is 57 cm long.

Fig. 5.23: Detail of previous photo with axial section of a sandstone diatreme. Note the fine, concentric lamination at the rim and the more massive appearance in the center. Lens cap is 53 mm in diameter.

Fig. 5.24: Axial section of twin sandstone diatremes enclosed by a larger one on the lower bedding surface of bed M_e438. Lens cap is 53 mm in diameter.

Fig. 5.25: Abundant pits on the upper bedding surface of M_e511 possibly related to de-watering. No sandstone diatremes have been found in this bed. Arrow points to H.P. Schmid for scale.

Hardly any of the structures are crosscutting this bed. It is possible, that the clay content of this bed affected the formation of the sandstone diatremes to an unknown extent.
The lower set boundary of M_a438, a large scale cross-bed, shows abundant cross-sections of sandstone diatremes in various sizes (Fig. 5.22). These structures are usually circular to rarer semi-circular in cross-section and usually show fine mm-thick concentric lamination, the central part (usually 1/5 to 1/3 of the diameter) seems to be more massive, or at least thicker laminated (Fig. 5.23). In rare cases the midpoint of the lamination is slightly de-centered, in one case two sandstone diatremes, c. 5 cm in diameter and some10 cm apart, are enveloped by a larger, elliptical sandstone diatreme (Fig. 5.24). In some places at the lower bedding surface of M_a438 the first stages of some smaller sandstone diatremes are half-sphere shaped, with several thin half-sphere shaped laminae sitting one in another, slightly intruding into the bed below (compare with "bullet-shaped" terminations of Gabelmann 1955). From the observations of bed M_a438, it seems, that fluidization has started in several small points at the boundary between beds M_a437 and M_a438 and fluid supported liquefaction cells ascended like ascending bubbles. Gabelmann (1955) described similar cylindrical structures in siltsone that also have a relatively narrow defined horizon, where they begin.

Sandstone diatremes usually are striking well-defined cylinders, with very sharp boundaries to the host-rock (Fig. 5.20-5.21). Deynoux *et al.* (1990) divided his structures into ones that partly preserved primary sedimentary structures and in others that are completely modified by de-watering. Similar observations have been made in the Muth Fm. (see Deynoux *et al.* 1990 for discussion of different formation mechanisms).

In the first case sandstone diatremes show a well-defined cylindrical shape, nevertheless, the lamination of the host-rock (large-scale cross-bed) is still visible, although slightly distorted and indicating a gravitational downward movement and the formation of steep normal faults. The contact to the hostrock is represented by some mm to a few cm thick fluidized cylindrical zones. Primary sedimentary structures commonly show concave-up dish shaped primary lamination in the upper part of the sandstone diatremes. Deynoux *et al.* (1990) explains this structure by a statically liquefaction cell and upward directed flow of water and/or fluidized sediment at the enveloping cylinder surface resulting in a downward movement of the whole structure. In the Muth Fm. in one example, structures of a first stage of such a fluidization cell, which has not evolved to a fully developed sandstone diatreme, seems to have been observed.

In the second case the sandstone diatremes show a well-defined cylindrical shape and additionally the internal structure of the whole cylinder is also strongly affected by de-watering (Fig. 5.21). Deynoux *et al.* (1990) explains this type of structure by an upward moving liquefaction cell, resulting in conical laminae sitting one in another. In the Muth Fm. no real conical structures have been observed, but also cannot be excluded, because conical laminae tend to become cylindrical shaped in higher parts. In the Muth Fm. sandstone diatremes seem to show cylindrical internal lamination.

The upper bedding surface of M_a440, which represents the boundary between FA2 and FA3 is overlain by very fine-grained dolomite, shows noticeable pits eroded into the surface, representing the uppermost part of sandstone diatremes (Fig. 5.18, 5.19). No sandstone diatremes have been found that continue into the dolomite above M_a440. Pits are circular in cross-section and show slightly raised rims (Fig. 5.19); thin sub-vertical fissures are abundant on the upper bedding surface of M_a440 (Fig. 5.19), but do not show any radial orientation to the pits.

The upper bedding surfaces of M_e512 and M_f1 are totally covered with abundant circular pits (Fig. 5.25), that possibly also represent similar de-watering features below this surface, although no clear evidence for the existence of sandstone diatremes have been found in these beds. The striking surface can be traced for several km and represents a marker horizon of the uppermost part of the Muth Fm.

No sand volcano cones have been found directly on the upper bedding surface of M_a440, but some 15 cm above, within the fine-grained dolomite, there is a 5-15 cm thick quartzite horizon with even lower boundary and undulatory upper boundary that possible could represent sand, that has been vented from below, but no conclusive evidence has been found.

Liquefaction is the transformation of a granular material from a solid state into a liquefied state as a consequence of increased pore-water pressure (Youd, 1973). Liquefaction induced upward flow of a water/sediment mixture causes distinct sedimentary structures like sand dykes, sand sills, sand volcanoes, etc. The upward flow of the water is caused by two mechanisms, relief of the high pore-water pressure and reconsolidation of the sediment grains. The vertical settlement caused by sediment densification is commonly only some 3% of the liquified sediment thickness (Obermeier, 1996).

Liquefaction structures originate from seismic as well as nonseismic processes (strong wave action, de-watering due to rapid sediment loading, artesian ground water conditions, slumping, nonseismic landsliding, etc.). Often it is difficult to distinguish between seismic and nonseismic liquefaction structures. Liquefaction structures are most common in sand and silty sand, rare in gravel, but unusual in sediment with more than 15 % clay content (Obermeier, 1996).

Earthquakes with magnitudes of at least 5 are thought to cause liquefaction; magnitudes of about 5.5-6 are given in the literature for seismic events at which liquefaction effects become relatively common. Liquefaction of seismic origin is mainly caused by a cyclic shaking of the ground. Seismological controlling factors include (i) amplitude of the cyclic shear stresses, (ii) duration of seismic shaking, (iii) seismic history, (iv) focal depth; geological factors are (v) degree of compactness, (vi) sand thickness (vii) depth of sediment that liquefies, (viii) grain size, (ix) grain size distribution, (x) grain packing, (xi) depth of the groundwater table, (xii) thickness and properties of a possible overlying cap horizon.

Liquefaction takes place only where the sediment is completely saturated with water; seismic liquefaction strongly affects sediments in a few to about ten meters depth, the susceptibility decreases nearly to nil in depth exceeding 10 m (Obermeier, 1996).

Eolian deposits are regarded good aquifers because of there sorting and grainsize. In the case of the vertical sandstone diatremes in the uppermost part of FA2, the transgression by the lagoonal sediments of FA3 caused a rise of the ground water table; the transgression is thought to has happened rapid, because of the preservation of the coastal dunes of FA2. The exact mechanism that finally triggered the liquefaction is unknown; it can just be speculated about above mentioned possibilities, strong wave action and seismic activity seem to be prime candidates.

Fig. 5.26: Plan view of a donut-shaped mound in the Muth Fm. (section D-D´) on the upper bedding surface of bed M_d6 with a central pit. Note the circular shape and the horizontal erosion on top.

Fig. 5.27: Plan view of a huge Donut-shaped mound on M_d5 with a diameter of 71 cm, raised 7 cm above the bedding surface, the central pit is below the bedding surface. Lens cap is 53 mm in diameter.

Fig. 5.28: Plan view of 3 intersecting donut-shaped mounds on the bedding surface of M_d6. Note also the small circular structure some centimeters below. Lens cap is 53 mm in diameter.

Fig. 5.29: Oblique view of the previous photo. Note the smooth flanks of the structures and the lack of a central vent-like structure. Lens cap is 53 mm in diameter.

5.7.2. DONUT-SHAPED SURFACE STRUCTURE

In the lower part of section D-D´ (beds M_d5-M_d7) and A-A´ (bed M_a557) several enigmatic donut-shaped structures on upper bedding surfaces of characteristically well horizontally laminated beds have been found (Fig. 5.26-5.29). These structures have a striking circular, donut-shaped raised ring with a dish-shaped, central depression. The donut-shaped rings comprise relatively gentle slopes towards the bedding surface, as well as towards the central depression that is slightly deeper relative to the bedding surface (Fig. 5.29). Bed M_d6 shows three of these structures close together, in parts overlapping (Fig. 5.28, 5.29). In one case the structure has slightly been modified by the subsequent formation of ripples. The quartzite of the donut is in principal the same as the arenite of the bed on which it is found.

The outer diameter of the structure ranges between 12-70 cm, the central pit usually amounts to 1/5 to 2/3 of the outer diameter and the bottom of the pit is slightly less deep below the bedding surface, than the donut is raised above it. The largest structure for example comprises 70 cm outer diameter, 31 cm inner diameter, the donut is raised 7 cm above the bedding surface, the bottom of the central pit *c*. 5 cm below (Fig. 5.27).

In principal three modes of formation of these structures appear to be possible. The first possibility is represented by ejection of sand caused by compaction, reduction of pore volume and related expulsion of a water/sand mixture, *i.e*. a sand volcano. The striking regular shape of the structures could argue for a de-watering model. By the way, the term “monroe structure” should not be used in this context and should be restricted to mud volcanoes in tidal flats (Dionne 1973). The lack of a central vent and the usual conical shape of sand volcanoes (Neumann-Mahlkau 1976) that is quite different to the round shape of the donut argue against this interpretation. Bernhard Grasemann (pers. comm. 1997) made similar observations; he watched small structures closely resembling those described by Neumann-Mahlkau (1976) in river sand near the waterline of the Pin River, directly next to foot prints of a yak. There these structures have been formed by expulsion of a water/sand mixture, caused by compaction of the sediment by shock and weight of the yak.

Another possibility is the formation by ascending gas bubbles. Gas might just be entrapped air, or produced by aerobic decomposition of organic matter, it migrates from underneath towards the surface, there the gas/sand/water mixture creates a blister-like mound with a central tiny pit (Häntzschel 1941). Similar structures are also reported from arenaceous sediments, where the surface is covered by biofilms, there the coherent fibrillar mat prevents the escape of the gas into the water or air, causing doming of the mats, sometimes with a central pit, where the gas finally escaped (Gerdes & Krumbein 1994). Both structures usually are much smaller (a few mm to 2 cm in diameter) than the donut-shaped structures and therefore are regarded unlikely mechanisms to explaine the structures in the Muth Fm.

The third possibility is the formation of donut-shaped structures by digging organisms. In this interpretation the raised ring represents the excavated sediment from below, the central pit, the partly filled remnants of the subsurface burrow. The diameter of the central pit of the smaller donut-shaped structures shows similar sizes to some *Metaichnia* Form A traces, this interpretation is not totally unrealistic. Furthermore the abundant *Taenidium* traces on surfaces with donut-shaped structures indicate considerable faunal activity, although the very large size of some the donut-shaped structures would demand very large digging organisms to create these structures, additionally no burrows underneath central pits have been observed (Fig. 5.29). Thus the precise mode of formation of this enigmatic structure remains unsolved.

5.8. Deformation bands

Draganits (1999a,b) and Draganits & Schlaf (1999) have mentioned the occurrence of sand dykes in the Muth Fm. After very instructive discussions with Stephan Matthai (Zürich) these structures are re-interpreted here as deformations bands (Aydin 1977). The additional existence of true sand dykes cannot be ruled out totally, especially some veins near slumped bed M_a544 might represent some, but the overwhelming majority of planar, crosscutting structures in the Muth Fm. are regarded to represent deformation bands.

Deformation bands have mainly been observed in section A-A´ SE of Mikkim, but hardly ever in the other sections. This might be explained by the excellent outcrop conditions, the length of the section (because the other sections are shorter and therefore the probability to find these structure is smaller) or these structures are more common near Mikkim than near Muth. Deformation bands appear throughout the section, but are most common in the lower part of FA 2 (between beds M_a530-M_a555), that is interpreted as eolian deposits. Several more deformation bands occur around bed M_a188, M_a206 and M_a452. The orientation of all observed deformation bands has been measured (Fig. 5.35A).

Deformation bands are mm thin planar, but slightly wavy features, that can be traced for some cm to several m. Although they comprise the same mineralogy as the host rock (*i.e.* monotonous quartzite) they are conspicuous for their lighter color and because of their higher resistance against weathering, they weather slightly above the average rock surface (Fig. 5.30, 5.31). The boundary to the host rock is well defined, but not so distinct like in a brittle fault (Fig. 5.31). In principal, hardly any displacement on these surfaces has been noticed, but in a few cases some mm to several cm of displacement has occurred (Fig. 5.32). Deformation bands hardly ever occur alone; usually they appear in “zones of deformation bands” (Aydin & Johnson 1983) that constitute of many closely spaced deformation bands (Fig. 5.30).

In zones of deformation bands, where the deformation bands are very closely spaced, usually only relicts of the primary sedimentary lamination is preserved. Usually these relicts show thin, lenticular “micro-lithons”, surrounded by slightly undulating deformation bands (Fig. 5.31). In places, possibly with advanced cementation of the arenite prior to the formation of the deformation bands, small angular fragments of the lamination are contorted and slightly rotated (Fig. 5.33). In zones of exceptional abundant and close deformation bands hardly any internal structure is visible and they appear nearly massive (Fig. 5.32).

At the western termination of the Muth Fm., SE of Mikkim, where the Pin River erodes the western limb of this anticline two striking zones of deformation bands occur which are even visible from distance. At the NW corner of this outcrop a zone of deformation bands (F212/76) appears that cross-cutts the bedding at near right angle (Fig. 5.30-5.32). The zone of deformation bands is some 70 cm at its thickest part, pinches out towards East and branches into two thinner zones of

Fig. 5.30: Zone of deformation bands in the Muth Fm. SE Mikkim, near bed M_a544. Deformation bands are lighter than the hostrock and have a wavy appearance. Lens cap is 53 mm in diameter.

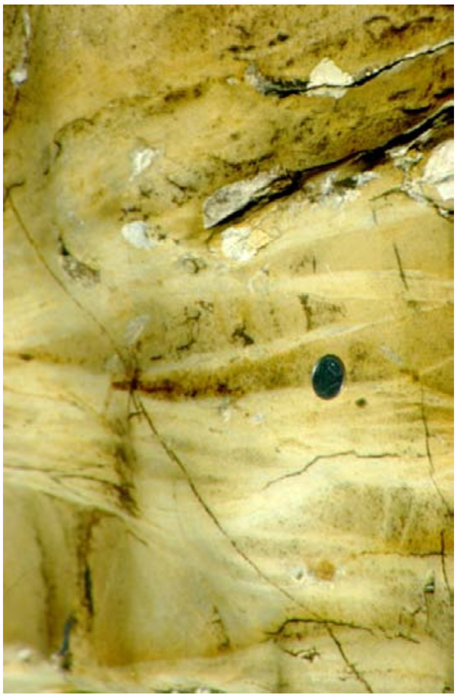


Fig. 5.31: Detail of previous photo (rotated image) showing typical deformation bands, which are lighter and more resistant to weathering than the lenticular preserved hostrock between them. 1 Rupee coin is 25 mm in diameter.

Fig. 5.32: Cross-section of the same zone of deformation bands. Note the massive appearance and the slightly distorted sedimentary lamination. Arrow points to conjugated deformation bands with some displacement.

Fig. 5.33: Deformation bands in the Muth Fm. SE of Mikkim in bed M_a186 showing rare brecciation and distortion of the hostrock. Lens cap is 53 mm in diameter.

deformation bands towards the West. The total displacement may account to several dm to a few m. In the middle part of this outcrop with WSW dipping bedding surfaces, a zone of deformation bands (F195/84) crosscuts the whole outcrop in this place and reaches some 8 m thickness; its southern directed boundary has developed to a slip surface (Aydin & Johnson 1983) with several m displacement. Three additional zones of deformation bands with similar orientation, all less than 30 cm thick, are found between these two prominent ones.

Antonellini *et al.* (1993) divided deformation bands into three groups, (1) with little or no cataclasis, (2) with cataclasis and (3) with fault gauge, according to the degree of grain fragmentation during the formation of this structure. In the Muth Fm. all transitions between the first two groups are observed, but the third group has not been found, but this might be related to the monotonous quartzitic lithology of the Formation. The formation of deformation bands causes grain dilatancy, followed by grain crushing and compaction; the final appearance of the microstructure is controlled by porosity, confining pressure, fluid content and amount of strain (Antonellini *et al.* 1993). The growth of deformation bands has been investigated by Mair *et al.* (2000) by laboratory triaxial compression of sandstone core samples.

Due to the lack of matter in the deformation bands, useful for geochronologic dating methods, the timing of their formation is hardly to constrain. The orientation of 39 deformation bands has been measured in the Muth Fm. SE of Mikkim (Fig. 5.35A). The orientation of the deformation bands does not fit reasonable to any of the orientations of brittle faults related to Himalayan deformation that have been measured in the Pin Valley by Wiesmayr (2000). Additionally, the formation of deformation bands demands porous sandstone, therefore these structures will hardly develop in sandstone that is fully cemented, therefore more indicative for preferable older deformation, than younger.

Both arguments probably indicate that the deformation bands are not related to Himalayan deformation and therefore early Himalayan large-scale folding has been restored by back-rotation of the bedding surfaces to horizontal orientation together with the deformation bands. In the restored orientation the deformation bands consistently trend E-W, with subvertical orientation to the bedding surfaces (Fig. 5.35A). Aydin & Reches (1982) mention the formation of four sets of deformation bands in the Entrada and Navajo Sandstone in Utah, although 39 orientations have been measured in the Muth Fm. a comparable detailed differentiation of the data has not been possible.

Subvertical, silizified fractures forming conjugate sets in the uppermost part of the Pin Fm. near Muth has already been mentioned in Chapter 4.3. These fractures are oriented at right angle to the bedding surfaces trending NE-SW and NW-SE (Fig. 5.34, 5.35B). Though small differences in orientation (Fig. 5.35A, 5.35B) between the deformation bands in the Muth Fm. and the fractures in the Pin Fm. their formation is possibly related.

Fig. 5.34: Plan view of the upper bedding surface of the uppermost bed of the Pin Fm. with fractures at nearly right angle to the bedding surface. Fractures are striking white due to silicification of the hostrock near the fracture. Rasmus Thiede for scale.

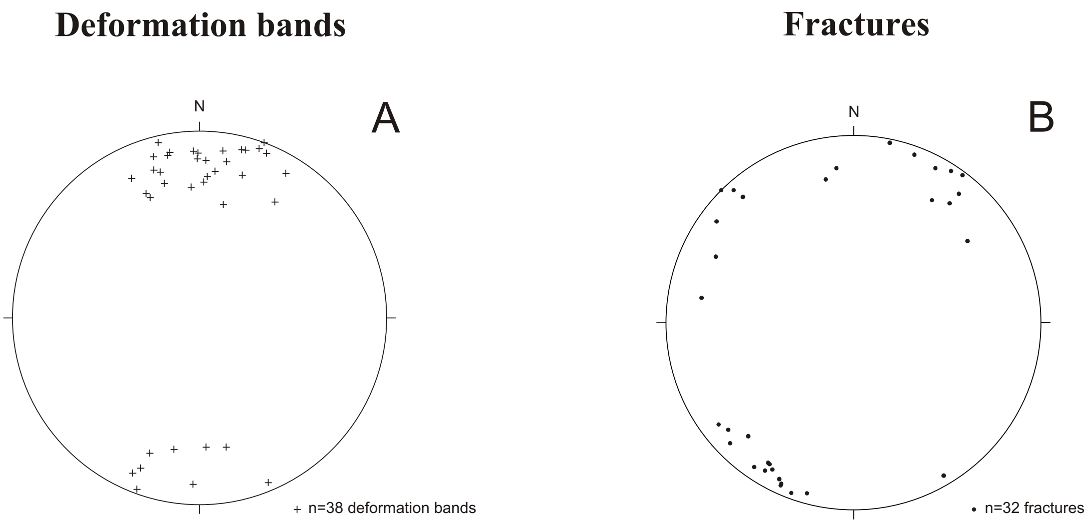


Fig. 5.35: A) Stereo net plot of restored orientations of deformation bands in the Muth Fm. SE of Mikkim (section A-A´). Equal area lower hemisphere projection; crosses represent poles of deformation bands. Note the consistency of data throughout the whole section. **B)** Stereo net plot of restored orientations of fractures in the uppermost part of the Pin Fm. South of Muth. Possibly two conjugate orientations can be differentiated.

6. ICHNOFAUNA OF THE MUTH FORMATION

6.1. Trace fossil association

Most of the trace fossils described herein are found at an altitude of about 3600 m at the western limb of the anticline SW of Mikkim (Fig. 2.9, 4.1) at the river junction of Pin River and Parahio River (GPS: N32°02‘14“; E78°03‘53“; EPE: 22 m). Here the erosion of the Pin River has exposed relatively fresh bedding and foreset surfaces of the quartzite. Rarer and poorly preserved trace fossils also occur in other parts of the outcrop at a similar stratigraphic level. The abundance of trace fossils in this western termination of the Formation near Mikkim, the scarcity in other parts and near total lack in the Muth section can be explained by the outcrop conditions and exposure of foresets and bedding surfaces of FA2 in this place.

Some trace fossils have previously been reported from other parts of the Muth Fm., although never in the abundance recognized at the Mikkim locality. Bhargava & Bassi (1988, table 1) described *Chondrites* and *Phoelus* at Kumaon and *Paleophycus*, *Planolites*, *Skolithos* and “arthropod markings” in the Lipak Valley (NW Kinnaur), *c*. 15 m below the bae of the Lipak Fm. They suggest chelicerates (xiphosurids) as producer for these arthropod trackways.

The ichnofauna in the Pin Valley is discussed here in general, notes on individual traces are mentioned in the discussion in the taxonomic section. Trace fossils have been produced in/on quartz sand substrate, later lithified to quartzites. All trace fossils remain *in situ* at the outcrop, because attempts to split off slabs bearing traces failed; as an alternative Formasil RTV© peels of some very well preserved trackways were produced. The surfaces on which the trace fossils are found are white when fresh, but they are progressively coated with an orange-brown-black shiny rock varnish, which often obscures the trace fossils and sedimentary structures (Fig. 6.17, 6.23). Rock surfaces exposed to the wind have experienced abrasion (Fig. 6.15).

Most of the trackways are interpreted as being preserved on the actual surfaces on which they were produced, indicated by undisturbed sand laminae directly above. Undertracks are rare. It seems that the sand grain-size of the Muth Fm. did not facilitate the development of undertracks, which tend to be more widespread within laminations of finer grained sediments (Goldring & Seilacher 1971; Johnson *et al.* 1994).

The distribution of the ichnofauna in the Muth Fm., dominated by *Palmichnium* and *Diplichnites* trackways, is clearly facies controlled as indicated by its nearly exclusive restriction to FA2 (Fig. 4.10). Sedimentological evidence, indicates a very shallow intertidal to barrier-island dune environment for FA2 (Chapter 5.4.2).

Palmichnium trackways occur in the whole range of FA2 in the Mikkim section, where just the lower part of the facies association is exposed (Fig. 4.10). Their first appearance coincides with the beginning of FA2. They can be found on bedding surfaces but are more widespread on foresets, often producing deep tracks with pronounced pushback mounds. The *Palmichnium* trackways show a large size range (Table 3). Additionally the number of tracks per series, shape of the tracks, symmetry and gait display variations, sometimes even within one trackway.

Diplichnites are rarer than *Palmichnium* and are more common on bedding surfaces than foresets. There are some examples of *Diplichnites* and *Palmichnium* on single surfaces, but with no

Trace	Name	Bed	Surface	Orientation	Ext. width	Int. width	Stride	Ser. ang.	Shape of Tracks	Tr. Orient.	Tr./Ser.	Sym	Med.	Remarks
3	Diplichnites Form A	Ma543	SS263/40	L207/26 >?	31	16	11*	-	ellipsoidal	oblique	?	O-A	-	well preserved
4	Diplichnites Form A	Ma543	SS263/40	-	33	15	12*	-	?	?	?	?	-	curves, poorly preserved
5	Diplichnites Form A	Ma543	SS263/40	L220/12 >?	48	28	15*	-	elongate	oblique	?	?	-	curves, poorly preserved V-series
11	Diplichnites Form A	Ma548	SS250/50	L002/03 >?	90	57	11*	-	crescentic	?	?	?	-	ridges parallel to trackway
27	Diplichnites Form A	Ma566	Sfor282/25	L260/25 >080	140	75	27*	-	D-shaped	?	?	?	-	poorly preserved
28	Diplichnites Form A	Ma566	Sfor282/25	L240/19 >?	31	21	10*	-	ellipsoidal (6x4)	?	?	A	-	
38	Diplichnites Form A	Ma544	Sfor299/27	L009/14 >?	38	15	15*	-	ellipsoidal	oblique	?	?	-	poorly preserved
48	Diplichnites Form A	Ma543	SS263/40	L263/40 >?	18	15	8*	-	ellipsoidal	?	?	O-A	-	poorly preserved
86	Diplichnites Form A	Ma565	Sfor290/44	L230/23 >230	51	22	15*	-	ellipsoidal (14x11)	oblique	?	O-A	-	very well preserved
69	Diplichnites Form B	-	SS224/12	L228/12 >228	51	44-49	12*	-	circular	?	?	?	-	
70	Diplichnites Form B	-	SS224/12	L235/13 >235	68	51-66	13*	-	circular	oblique	?	?	-	poorly preserved-V
71	Diplichnites Form B	-	SS224/12	L235/14 >235	80	50-77	16*	-	circular	?	?	?	-	
72	Diplichnites Form B	-	SS224/12	L174/08 >?	77	36-70	17*	-	circular	?	?	?	-	
73	Diplichnites Form B	-	SS224/12	L191/12 >191	107	51	32*	-	circular	?	?	?	-	poorly preserved-V
77	Diplichnites Form B	-	SS246/09	L204/08 >204	90	50	102/12*							
								-	circular	oblique	?	?	-	
19	Diplopodichnus	Ma556b	Sfor281/21	L218/09 >?	20	15	3.6*	-	-	-	-			well preserved
1	Palmichnium Form A	Ma543	SS263/40	L234/25 >234	177	100	70	53	tear-drop - ellipsoidal	parallel	2(3)	A	yes	well preserved, back-push mounds
2	Palmichnium Form A	Ma543	SS263/40	L214/11 >214	140	83	54	79	tear-drop	parallel	2(3)	O	-	well preserved, back-push mounds
8	Palmichnium Form B	Ma515	-	-	74	28	52	74	circular	-	4	A	-	loose block, asymmetric gait, similar to trace 29
9	Palmichnium Form B	Ma506	-	-	90	43	49	49	circular	-	3(4)	O	-	loose block
12	Palmichnium Form A	Ma553	-	-	100	60	38	64	elongate - Ellipsoidal	parallel	2	S	yes	loose block
13	Palmichnium Form A	Ma555	SS268/31	L310/22 >310	?	?	90	36	circ/ellipsoidal	?	2	?	-	poorly preserved
14	Palmichnium Form A	Ma556b	Sfor286/35	L192/01 >192	84	48	19	34	comma	parallel - obl	2	O	?	
15	Palmichnium Form A	Ma556b	Sfor286/35	L354/14 >174	107	62	27	68	circular - ellipsoidal	?	2	O	-	
16	Palmichnium Form A	Ma556b	Sfor286/35	L174/01 >?	?	25	46	86	?	?	2	?	-	poorly preserved
17	Palmichnium Form A	Ma555	SS283/35	L196/01 >16	188	120	30	81	rear-drop	oblique	2	O	-	
18	Palmichnium Form A	Ma555	SS283/35	L310/32 >?	250	110	91	?	D-shaped (63x38)	perpendicular	?	-	-	fairly preserved
20	Palmichnium Form A	Ma558	Sfor270/26	L218/16 >218	211	131	44	88	linear	?	2	?	-	poorly preserved
21	Palmichnium Form A	Ma558	Sfor270/26	L301/20 >?	198	138	42	41	oval	?	2	?	-	poorly preserved
22	Palmichnium Form A	Ma558	Sfor270/26	L220/19 >?	245	115	58	?	?	?	2	?	-	poorly preserved
24	Palmichnium Form A	Ma560	Sfor264/36	L234/28 >234	215	175	66	71	ellipsoidal - elongate	oblique	2	O	?	
25	Palmichnium Form A	Ma566	Sfor274/30	L355/02 >355	270	120	70	68	ellipsoidal - elongate	oblique	2	?	-	poorly preserved
26	Palmichnium Form A	Ma566	Sfor274/24	L004/02 >?	63	37	15	?	?	?	?	-	-	curves, poorly preserved
29	Palmichnium Form B	Ma566	Sfor290/26	L322/20 >322	170	46	82	59	circular	-	4(1)	S	-	backpush mounds, asymmetric gait, similar to trace 8
30	Palmichnium Form A	Ma565	SS262/47	L242/41 >062?	125	75	51	46	ellipsoidal (20x14)	oblique	2	O	-	

Tab. 2: Trackway attributes using the nomenclature of Trewin (1994). Abbreviations: bedding surface (SS), foresets (Sfor); Orientation: trackway alignment with probable walking direction; external width (Ext. width); internal width (Int. width); Stride: values with "*" are paces of Diplichnites, series angle to the midline; mainly mid and outer tracks measured (Ser. ang.); orientation of tracks to the midline (Tr. orient.); number of tracks per series (Tr./Ser.); symmetry of opposing series (Sym.); opposite (O); staggered (S), Alternate (A); medial impression (Med.).

Trace	Name	Bed	Surface	Orientation	Ext. width	Int. width	Stride	Ser. ang.	Shape of Tracks	Tr. Orient.	Series	Sym		Remarks
31	Palmichnium Form A	Ma565	SS262/47	L226/40 >046	498	245	255	62	ellipsoidal (35x20)	perpendicular	3(2)	O	-	largest trackway
32	Palmichnium Form A	Ma566	Sfor248/16	L226/15 >226	212	116	53	72	elongate	oblique	2(3)	A	yes	very well preserved, backpush mounds
33	Palmichnium Form A	Ma566	Sfor247/16	L237/16 >237	215	78	45	71	tear-drop/circular	oblique	3	A	yes	very well preserved, backpush mounds
34	Palmichnium Form A	Ma566	Sfor245/26	L335/26 >155	215	105	112	24	crescentic	perpendicular	2	S	-	big stride
35	Palmichnium Form A	Ma566	Sfor274/29	L271/26 >?	118	53	40	?	elongate	oblique	2	?	-	poorly preserved
36	Palmichnium Form B	Ma535	SS257/36	L268/35 >088	90	45	60	29	circular	-	4	S-A	-	
37	Palmichnium Form A	Ma543	SS263/40	L004/05 >184	190	140	67	47	ellipsoidal (20x10)	parallel	2(3)	A	-	well preserved, backpush mounds
39	Palmichnium Form A	Ma544	Sfor299/27	L295/25 >295	320	135	214	29	crescentic	perpendicular	3(4)	?	-	backpush mounds
40	Palmichnium Form A	Ma510	-	-	104	67	48	44	circular	-	3	?	-	loose block
41	Palmichnium Form A	Ma554	Sfor282/20	L227/14 >227	145	88	54	74	?	?	2	?	-	poorly preserved
42	Palmichnium Form A	Ma553	SS268/26	L268/26 >088	172	118	150	26	elongate	oblique	2	S	-	poorly preserved
43	Palmichnium Form A	Ma559	SS284/30	L235/21 >?	?	?	54	30	?	?	2	?	-	poorly preserved
44	Palmichnium Form A	Ma561	SS270/25	L213/10 >213	?	?	31	68	?	?	2	?	-	
45	Palmichnium Form A	Ma561	SS270/25	L230/23 >230	385	240	125	90	?	?	2(3)	O	-	poorly preserved
46	Palmichnium Form A	Ma556b	Sfor280/35	L198/03 >198	198	140	47	79	?	?	2(3)	S	-	
47	Palmichnium Form A	Ma543	SS263/40	L288/32 >288	208	113	113	42	crescentic	perpendicular	2	A	-	backpush mounds
49	Palmichnium Form A	Ma555	SS283/35	L268/31 >088	165	77	118	40	D-shaped (25x12)	perpendicular	2	S	-	
50	Palmichnium Form A	Ma555	SS283/35	L196/04 >196	131	82	30	23	?	?	2	?	-	poorly preserved
51	Palmichnium Form A	Ma555	SS283/35	L203/04 >203	132	88	36	63	tear-drop	parallel	2	A	-	
52	Palmichnium Form A	Ma555	SS283/35	L013/02 >193	183	128	46	64	elongate?	parallel	2	?	-	poorly preserved
53	Palmichnium Form A	Ma556b	Sfor286/35	L204/14 >?	90	40	36	90	elongate	parallel	2	O	-	
54	Palmichnium Form A	Ma556b	Sfor285/41	L308/39 >128	190	130	40	68	tear-drop	oblique	2	S	-	backpush mounds
55	Palmichnium Form A	Ma556b	Sfor285/38	L014/06 >194	145	85	50	70	elongate	parallel	2	?	-	
56	Palmichnium Form A	Ma556b	Sfor285/38	L012/06 >192	180	110	53	71	elongate	parallel	2	A	-	backpush mounds
57	Palmichnium Form A	Ma556b	Sfor285/38	L194/01 >194	74	58	44	67	circular	-	3	?	yes	
58	Palmichnium Form A	Ma566	Sfor176/25	L330/12 >150	-	-	35	67	?	?	?	?	-	poorly preserved
59	Palmichnium Form A	Ma566	Sfor176/25	L333/13 >153	-	-	49	?	?	?	2	?	-	poorly preserved
60	Palmichnium Form A	Ma566	Sfor268/29	L357/02 >?	400	-	-	-	D-shaped/elongated	perpendicular	-	-	-	abundant superimposed Palm.
61	Palmichnium Form A	Ma566	Sfor278/30	L228/21 >?	265	145	153	?	circular?	-	2	S	-	
62	Palmichnium Form A	Ma566	Sfor280/29	L006/04 >186	375	230	50	55	?	?	3	?	-	
63	Palmichnium Form A	Ma566	Sfor240/60	L255/12 >255	175	108	45	74	elongate	parallel	2	?	yes	
64	Palmichnium Form A	Ma566	Sfor250/21	L241/18 >061	260	115	60	84	elongate	?	2	?	-	
65	Palmichnium Form A	Ma566	Sfor250/21	L248/17 >248	-	-	56	64	elongate	?	2	?	-	
66	Palmichnium Form A	Ma566	Sfor250/21	L251/21 >251	196	120	43	?	crescentic	?	?	?	-	backpush mounds
67	Palmichnium Form A	Ma566	Sfor232/16	L265/14 >265	240	158	49	85	elongate	?	2	?	-	
68	Palmichnium Form A	Ma568	Sfor250/18	L171/07 >171	210	115	?	?	?	?	?	?	-	
87	Palmichnium Form A	Ma565	Sfor290/44	L224/21 >224	160	85	39	67	ellipsoidal	oblique	2	A	-	poorly preserved

Tab. 2 (continuation): Trackway attributes using the nomenclature of Trewin (1994). Abbreviations: bedding surface (SS), foresets (Sfor); Orientation: trackway alignment with probable walking direction; external width (Ext. width); internal width (Int. width); Stride: values with "*" are paces of Diplichnites, series angle to the midline; mainly mid and outer tracks measured (Ser. ang.); orientation of tracks to the midline (Tr. orient.); number of tracks per series (Tr./Ser.); symmetry of opposing series (Sym.); opposite (O); staggered (S), Alternate (A); medial impression (Med.).

evidence of any interaction. *Diplichnites* tracks show hardly any pushback mounds (Fig. 6.5-6.10), thus their walking direction is difficult to decide, but even the strong scatter of the trackway orientation indicates that no comparable general walking direction like in the *Palmichnium* trackways is present (Fig. 6.2B). The only *Diplopodichnus* trace is found on a foreset of a large-scale cross-bed (Fig. 6.7).

Taenidium are not so abundant as *Palmichnium* and *Diplichnites* traces, but they still are relative common and occur mainly on bedding surfaces of horizontally laminated beds, sometimes on the same surfaces with spring pits. In the rare occurrences on foresets they appear in the lowest part of the tangential foreset. The best-preserved ones are not found at the western termination of the Muth Fm., but occur near the stone cairn with a prayers flag at the crest line of the Muth Fm. anticline, SE of Mikkim and at the eastern termination of the anticline. In the occurrence near the stone cairn with a prayers flag, on the upper surfaces of a horizontally laminated bed, *Taenidium* (Fig. 6.22) is noted in cross-cutting near axial orientations. At the western termination of the anticline, transverse sections through *Metaichnia* Form B, forming elongated shallow hollows (Fig. 6.11, 6.12), occur on the bedding surface of the horizontally laminated bed M_a567, directly above the large-scale cross-bed of M_a566, with abundant *Palmichnium* and *Diplichnite*s and very rare *Didymaulichnus*.

Vertical burrows are quite common both in horizontally laminated units and in large-scale cross-bedded units (Fig. 6.23). No expressive vertical sections of these traces were found, thus their ichnotaxonomy remains unclear. Vertical burrows are sometimes concentrated on the asymptotic bases of eolian dunes and gradually become rarer towards the top of the slip face (Fig. 6.24). A single meandering grazing trail of unclear affinity was recorded on the bedding surfaces of a large-scale cross-bedded unit (Fig. 6.13). Sub-parallel orientated *Selenichnites* are found in the Muth section in the uppermost part of FA4, close to the base of the Lipak Fm. on bedding surfaces (Fig. 6.21).

6.2. Comparison with other ichnofaunas

Devonian terrestrial trace fossils are rare; most are described from marginal marine, fluvial and lacustrine settings. Besides the ichnofauna presented here, only two, with a comparable ichnoassemblage, facies and age, are mentioned in the literature: (1) An ichnofauna in the Beacon Supergroup of Antarctica (Gevers *et al.* 1971), and (2) an ichnofauna from the Tumblagooda Sandstone of western Australia (Trewin & McNamara 1995). All three ichnofaunas are broadly assigned to Late Silurian to Middle Devonianm age and a marine-terrestrial or marine-fluvial transition depositional environment, dominated by sand. Paleogeographical reconstructions for the Early Devonian show the Australian and Indian ichnofaunas at around 30° southern latitude, and the Antarctic ichnofaua at 45° southern latitude, all three of them at the margin of Gondwana (Fig. 6.1A).

Diverse ichnofaunas have been described from several outcrops of the Taylor Group (lower part of the Beacon Supergroup), Antarctica (Gevers *et al.* 1971, Bradshaw 1981, Gevers &

Twomey 1982, Bradshaw *et al.* 1990, Woolfe 1990), which are discussed in detail by Trewin & McNamara (1995). Based on dating of fish fossils, molluscs and plants, an Early to early Late Devonian age of the Taylor Group is suggested (Trewin & McNamara 1995 and references cited therein).

The probable Early Devonian ichnofauna of the Junction Fm. and the Upper Hatherton Fm., at the base of the Beacon Supergroup in the Darwin Glacier area, seems to represent a good equivalent to the ichnofauna of the Muth Fm. Trace fossils are found in fine- to medium-grained mature sandstones and orthoquartzites, which comprise thick cross-bedding with interbedded thin horizontally bedded beds (Gevers *et al.* 1971), similar to FA2 in the Muth Fm. In analogue to the Muth Fm. Gevers & Twomey (1982) state a 99% purity of quartz arenite of the Upper Hatherton Sandstone. Gevers *et al.* (1971) assigned a marine environment mainly based on the hypothesis that the large *Beaconites* (= *Taenidium* Keighley & Pickerill 1994) were made by polychaete annelids, *i.e.* marine organisms. In contrast, Gunn & Warren (1962), Plume (1982) and Woolfe (1990, 1993) suggested a non-marine environment including fluvial, estuarine, littoral and eolian facies due to investigations of sedimentary structures and a reinterpretation of some trace fossils. This ichnofauna includes large *Palmichnium* trackways, up to 37 cm wide (Bradshaw *et al.* 1990), smaller *Diplichnites* and *Diplopodichnus* trackways, *Taenidium*, *Skolithos* as well as some *Aulichnites* and *Heimdallia*. In contrast to the abundant trackways on foreset surfaces in the Muth Fm., the Darwin Mountain *Palmichnium*, *Diplichnites* and *Diplopodichnus* trackways exclusively have been found on the upper surface of horizontally bedded beds (Gevers *et al.* 1971).

The most impressive traces, both in the Muth Fm. and in the Upper Hatherton Fm. are the large *Palmichnium* trackways. One slab, described by Gevers *et al.* (1971, Plate 20, fig. 3) as a "veritable stamping ground" shows several *Palmichnium* trackways on the same surface. Similar surfaces with several *Palmichnium* have also been found in the Muth Fm. (Fig. 6.14).

Gevers *et al.* (1971) described *Taenidium* of different size, narrow sinuous forms and bigger ones, the latter are relatively straight with widths up to 130 mm. *Taenidium* in the Muth Fm. is mainly narrow, with widths rarely exceeding 30 mm. The ratio of *Diplopodichnus* to *Diplichnites* is much higher in the Upper Hatherton Fm. than in the Muth Fm., which probably reflects moisture differences of the sediment surface during trackways formation (Brand 1979, Johnson *et al.* 1994).

Trewin & McNamara (1995) noted the similarity of the Darwin Glacier ichnofauna to the one of the Tumblagooda Sandstone of Western Australia. The Tumblagooda Sandstone represents the lowest unit of the Kalbarri Group. As an analogue to the Muth Fm. it comprises mainly arenaceous sediments with virtually no mud. The unit was tentatively dated as late Silurian on the basis of regional correlation, although an early Devonian was not completely dismissed (Trewin & McNamara, 1995 and references cited therein). Trewin & McNamara (1995) divided the Tumblagooda Sandstone into four facies associations, the uppermost three containing trace fossils.

This highly diverse ichnoassemblage can be divided into two groups, an arthropod dominated *Heimdallia-Diplichnites* ichnofauna, with many similarities to the Indian ichnofauna, and a filter feeder dominated *Skolithos-Diplocraterion* ichnofauna. The *Heimdallia-Diplichnites* ichnofauna is concentrated in facies association 2 and especially in its lithological association B (eolian sandsheet/pond sub-facies). Facies association 2 is interpreted as mixed waterlain and eolian sedimentation in a coastal fluvial outwash area (Trewin & McNamara 1995).

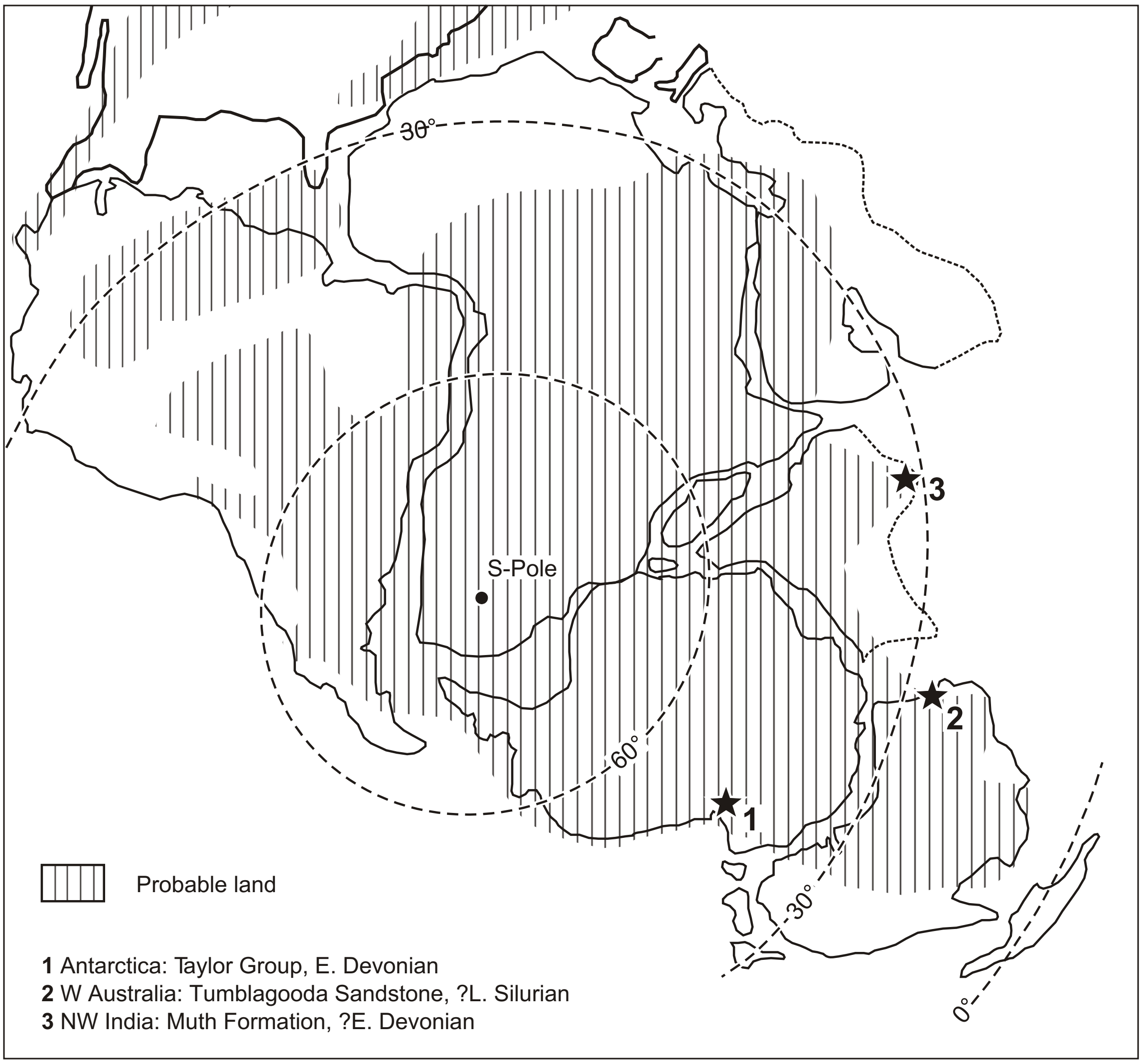


Fig. 6.1A: Distribution of well comparable Early Devonian ichnofaunas in similar sedimentary environments. Paleogeographic map modified after Young (1990).

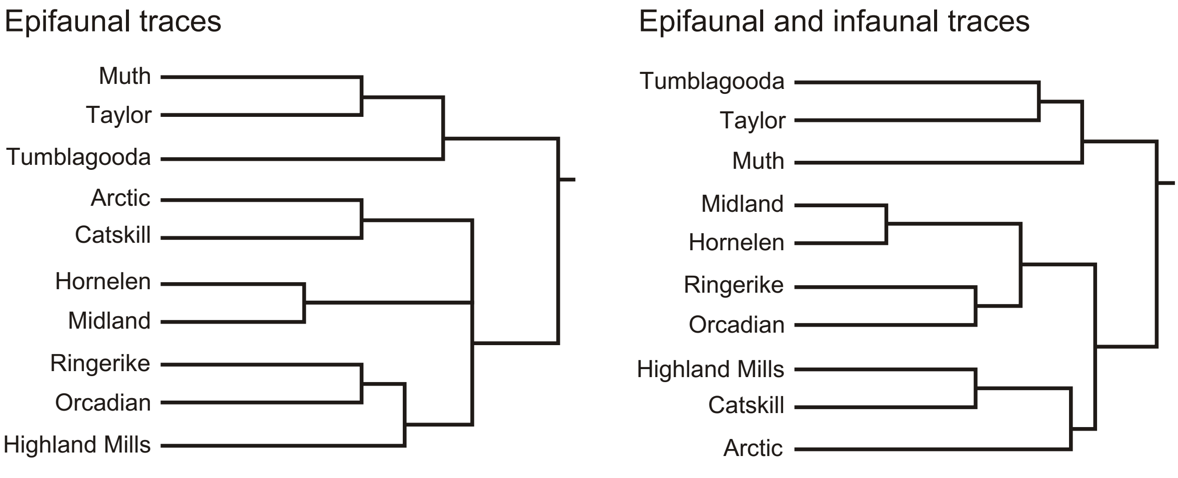


Fig. 6.1B: Cluster analyses of ichnoassemblage composition (Table 2) for 10 Upper Silurian and Devonian ichnoassemblages (after pers. comm. Simon Braddy 2000). Distance metric is Euclidean distance (0-1). Epifaunal traces only (left). Epifaunal and infaunal traces (right). See text for details.

In analogue to the Indian ichnofauna *Diplichnites* traces are common. They vary in their morphology and can be found on horizontally as well as low angle cross-bedded units. The main contrast to the ichnoassemblage in the Muth Fm. is the scarcity and relative small size of *Palmichnium* trackways and the abundant occurrence of *Heimdallia*. A possible explanation for these differences might be that the depositional environment of the Muth Fm. lacked quiet, shallow pools, the dominant environment in which *Heimdallia* in the Tumblagooda Sandstone occurs (Trewin & McNamara 1995). Additionally the barrier island setting of the Muth Fm. provides an environment possibly much closer to the shoreline, than the coastal fluvial outwash area of facies association 2 in the Tumblagooda Sandstone, which perhaps suited more to the habit of eurypterids (see below).

Compared with the Antarctic and Indian ichnofaunas, the Tumblagooda ichnofauna displays a much greater trace fossil diversity, especially in the infaunal component. This can probably be explained by differences in depositional environment. Alternatively, the number of ichnotaxa described from these ichnofaunas could be related to the amount of fieldwork done in these areas, the outcrop situation and the accessibility of the localities.

Several other broadly contemporaneous (Upper Silurian to Devonian) marginal to terrestrial ichnoassemblages are known from: (1) Ringerike Group, Norway, Late Silurian, fluvial to littoral settings (Hanken & Størmer, 1975, Pollard & Walker 1984). (2) Cape Storm Fm. and Leopold Fm., Somerset, Cornwall and Devon Islands in the Canadian Arctic, Late Silurian, only intertidal ichnofauna considered (Narbonne 1984). (3) Old Red Sandstone, Midland Valley of Scotland, Early Devonian, continental ephemeral to shallow lacustrine settings (Pollard & Walker 1984, Walker 1985). (4) Esopus Formation, Highland Mills, New York State, Early Devonian, only *Skolithos-Planolites* ichnofacies very shallow subtidal considered (Marintsch & Finks 1982). (5) Hornelen Basin, Norway, Middle Devonian, fluvial to shallow lacustrine settings (Pollard *et al.* 1982). (6) Middle Old Red Sandstone, Orcadian Basin, Scotland, Middle Devonian, fluvial to shallow lacustrine settings (Trewin 1976, Rogers 1990, Pollard & Walker 1984). (7) Catskill Group, New York State, Late Devonian, only fluvial and tidal ichnofaunas considered (Miller, 1979).

In order to test the ichnofaunal similarity of these ten ichnofaunas (*i.e.* Muth Fm., Taylor Group, Tumblagooda Sandstone and seven others above) data on the composition (trace fossils present) of each ichnofauna were compiled by Simon Braddy (pers. comm. 2000). In each case, the presence or absence of ichnotaxa (or trace fossil types, *i.e*. morphologically similar ichnotaxa) was recorded (Table 2). These data were then analysed via a cluster analysis (SYSTAT 5.2.1, complete linkage method, distance metric equals Euclidean distance between 0-1).

The Muth Fm. ichnofauna consistently clusters with the Antarctic and Australian ichnofaunas (Fig. 6.1B). According to an analysis based on only epifaunal traces, the Muth ichnofauna clusters closest to the Taylor Group ichnofauna of Antarctica. If the entire (epifaunal and infaunal) ichnofauna is considered, however, the Australian ichnofauna clusters closest to that of Antarctica. This is because the Muth Fm. traces tend to be preserved on bedding and foreset surfaces only; the relatively homogenous lithology of the Muth Fm. biases the data (pers. comm. Simon Braddy 2000).

The similarity between these three ichnofaunas (Antarctica, Australia, India), and their occurrence on the northern margin of Gondwana, identifies a recurrent Gondwanan ichnocoenosis. Although the precise age of these Formations, however, due to the scarcity of body fossils in these sediments, remains a question for future research.

Ichnoassemblage	**Ichnotaxa/trace type present** 123456789012345678901
Taylor Group, Antarctica	111101001111111101000
Tumblagooda Sdst., Australia	111011010111111011100
Muth Formation, India	110001001011001000000
Orcadian Basin, Scotland	101000100000101000000
Midland Valley, Scotland	011010110000000000000
Hornelen Basin, Norway	001010110000000000000
Ringerike, Norway	100010100000100000010
Arctic Canada	001111000001101010011
Catskill, New York State	001010000001000111001
Highland Mills, New York State	000000000001000001001

Table 2: Ichnoassemblage composition [ichnotaxa in common: presence (1), absence (0)] for Upper Silurian - Devonian ichnoassemblages (after pers. comm. Simon Braddy 2000)]. Ichnotaxa 1-9 epifaunal traces. Ichnotaxa 10-21 infaunal traces. Ichnotaxa Key: 1. *Palmichnium* (*cf. Paleohelcura*), 2. *Diplichnites* and *Diplopodichnus*, 3. Bilobed trails (*i.e*. *Cruziana*, *Isopodichnus* and *Tumblagoodichnus*), 4. *Petalichnus*, 5. Small horizontal meandering (looping) trails (*e.g*. *Mermia*, *Gordia*), 6. Arthropod resting traces (i.e. *Selenichnites* and *Rusophycus*), 7. *Merostomichnites*, 8. *Siskemia*, 9. Grazing trails *(i.e*. "epichnial ridge" of Bradshaw (1981), *Helminthoidia*, *Helminthopsis*), 10. *Heimdallia*, 11. *Didymaulichnus*, *Didymauloponomous* and *Fucusopsis*, 12. Simple vertical burrows (e.g. *Skolithos*), 13. *Diplocraterion*, 14. *Tigillites*, 15. *Taenidium* (*cf. Beaconites*), 16. *Thallasinoides*, 17. Paired burrows (*e.g*. *Arenicolites*), 18. *Planolites* and *Agrichnium*, 19. *Lunatubichnus*, 20. *Polarichnus*, 21.*Chondrites*.

There is also considerable paleobiogeographic similarity between the other ichnofaunas (*i.e*. Scandinavian-Scottish and North American clusters), particularly if overall trace composition is considered (Fig. 6.1B). Although provincialism is supported by this analysis, the diversity of ichnotaxa described from these settings also reflects the amount of fieldwork done in each area, the available outcrop, and the accessibility of the localities.

6.3. Trackway preservation

The lithology in which these trackways are preserved is quite atypical for trackway formation. Most fossil trackways reported in the literature occur in fine-grained, finely laminated sediments. Sand, especially dry sand, is regarded unsuitable for the preservation of trackways unless a suitable binding agent is present, such as water, from dew, a light rain fall or saline surf-spray, infiltrating clay minerals (McKeever 1991) or algal slime (Gevers & Twomey 1982).

Brady (1939, 1947) and Sadler (1993) conducted various subaerial neoichnological experiments with arthropods. Neoichnological investigations with arthropods walking subaqueosly have not been conducted, although Brand (1979, 1996) carried out extensive studies on the production of trackways produced by modern amphibians and reptiles with different substrates, slope angles and moisture contents, even under subaqueous conditions. The different substrate conditions caused a big variation in the appearance of trackways produced by one single species. Comparison of the trackways considered here with those of Brand (1979) reveals a convincing similarity concerning track depth and shape of pushback mounds between some uphill directed *Palmichnium* trackways found on foresets in the Muth Fm. (Fig. 6.19A) and those produced by salamanders walking uphill on sloping dry sand in the laboratory studies (Brand 1979, Fig. 6b).

We suggest saline surf-spray and/or microbial mats as probable candidates for the stabilization of trackways in the Muth Fm. During periods with slight wind these would have stabilized the surface, allowing trackway preservation. The existence of microbial mats in the pure quartzite is indicated by large domal stromatolites at the eastern limb of the anticline SE of Mikkim, or some smaller structures such as desiccated biofilms (pers. comm. G. Gerdes, 1999) on sedimentary surfaces directly at the outcrop where the trackways have been found.

6.4. Environmental implications

Only about twenty non-marine Devonian invertebrate ichnogenera are currently known (Buatois *et al.* 1998a). Most of these trace fossils occur in marginal marine and lacustrine settings, although traces in subaerial environments are also known (Rolfe 1980, Pollard 1985). Of these ichnogenera, only a small number has been assigned to arthropods. The recent discovery of the coastal ichnofauna of the Muth Fm. extends our knowledge of Early Devonian marginal marine arthropod communities.

The Himalayan traces indicate that a low diversity community was present in the coastal paleoenvironment of the Muth Fm. The most widespread trackways described herein are *Palmichnium antarcticum* (Fig. 6.3, 6.14-6.17, 6.19), probably made by stylonurid eurypterids. *Diplichnites gouldi* (Fig. 6.6.5, 6.6, 6.8-6.10), *Diplopodichnus biformis* (Fig. 6.7) and possibly also *Taenidium* (Fig. 6.22) can be assigned to the activity of myriapod-like animals (?eoarthropleurids). Vertical burrows (Fig. 6.23, 6.24) of different sizes represent the infaunal component, in plan view they resemble *Diplocraterion*, *Skolithos* and/or *Tigillites*, but due to the lack of expressive vertical sections the ichnotaxonomy is unclear and possible producers are not discussed here.

The diversity of the Muth Fm. ichnofauna seems low compared with the highly diverse fauna of the Lower Devonian coastal Nellenköpfchen Schichten of Alken (Germany), interpreted as an elongate lagoon bordered by a sandy bar, similar to the environment of the Muth Fm., but belonging to the Euramerican province (Størmer 1976). This poverty reflects the fact that the traces of the Muth Fm. probably record the activities of only a small fraction of the original faunal community.

Early Devonian arthropod trackways are also known from continental settings, for example the Midland Valley of Scotland ichnofauna (Walker 1985). This ichnofauna occurs in red-brown, thinly bedded siltstones and sandstones, with intervening mud horizons, on which the delicate trackways are preserved. The paleoenvironment is interpreted as shallow ephemeral lacustrine pools prone to periodic desiccation and exposure, as evidenced by rain prints and desiccation cracks, within a volcanic complex. This ichnofauna is considerably different from the Himalayan ichnofauna, yielding ichnotaxa such as *Stiaria*, *Siskemia*, *Mitchellichnus* and *Isopodichnus*, probably recording the activities of early arachnids, myriapods, crustaceans and possibly primitive insects. All these ichnotaxa are small, with the exception of *Mitchellichnus* (several cm wide), and infaunal componets are absent.

Eurypterids are considered to have had an amphibious habit, short excursions out of water possible by the possession of protected accessory aerial respiratory tissues (Manning & Dunlop 1995, Braddy *et al.* 1999). Ichnological evidence such as trackways formed under at least temporarily dry conditions support their amphibous habit (*e.g.* Sharpe 1932).

The Himalayan *Palmichnium* trackways on the foresets show very deeply impressed tracks with pronounced pushback mounds (Fig. 6.17, 6.19), very similar to the subaerial, uphill traces described by Brand (1997, Fig. 6b). We regard the underwater formation of the very deep tracks unlikely, as the buoyancy of the water would have reduced the weight of the eurypterids considerably, which would have resulted in much shallower tracks. This supports the sedimentological evidence (see above) for subaerial production of most of the Muth Fm. *Palmichnium* trackways.

Why were the arthropods "invading" the land? Størmer (1976) suggested three possible reasons: (1) obtaining new areas for food, (2) escaping enemies, and (3) protecting the next generation from marine predators. Gevers & Twomey (1982) stressed the importance of an abundant terrestrial food supply. However, eurypterids would not have been able to feed subaerially as they did not possess a pre-oral cavity (as in arachnids, myriapods and insects), which is regarded as an important adaptation to terrestrial life (Størmer 1976). Trewin & McNamara (1995) argued that the driving force behind short subaerial excursions was due to the drying up of marginal marine pools that the arthropods inhabited, causing high population densities and predatory pressure, forcing the animals to seek safer places.

Briggs & Rolfe (1983) proposed that eurypterids might have undertaken seasonal nuptial walks. This may not necesserily has involved only adult individuals. Evidence exists that eurypterids acquired sexual maturity before their final moult stage (Braddy & Dunlop 1997). A "mass-moult-mate" hypothesis was proposed by Braddy (1996), analogous with the behavior of extant xiphosurans and some semi-terrestrial crabs, whereby these animals would may migrated *en-masse* into nearer shore waters or quiet lagoons, moulted and mated before returning to their usual (marginal marine) habitats. This hypothesis might explain the abundant accumulations of exuviae (*e.g.* the Bertie waterlime in New York and Saaremaa in Estonia), which were previously considered mass death assemblages (Andrews *et al.* 1974).

The size range of the trackways varies considerably (Table 3). This indicates that various instars of one eurypterid species and/or several eurypterid species were involved in these subaerial excursions. Keeping in mind the considerable variations of the *Palmichnium* trackways (Figs. 6.15-

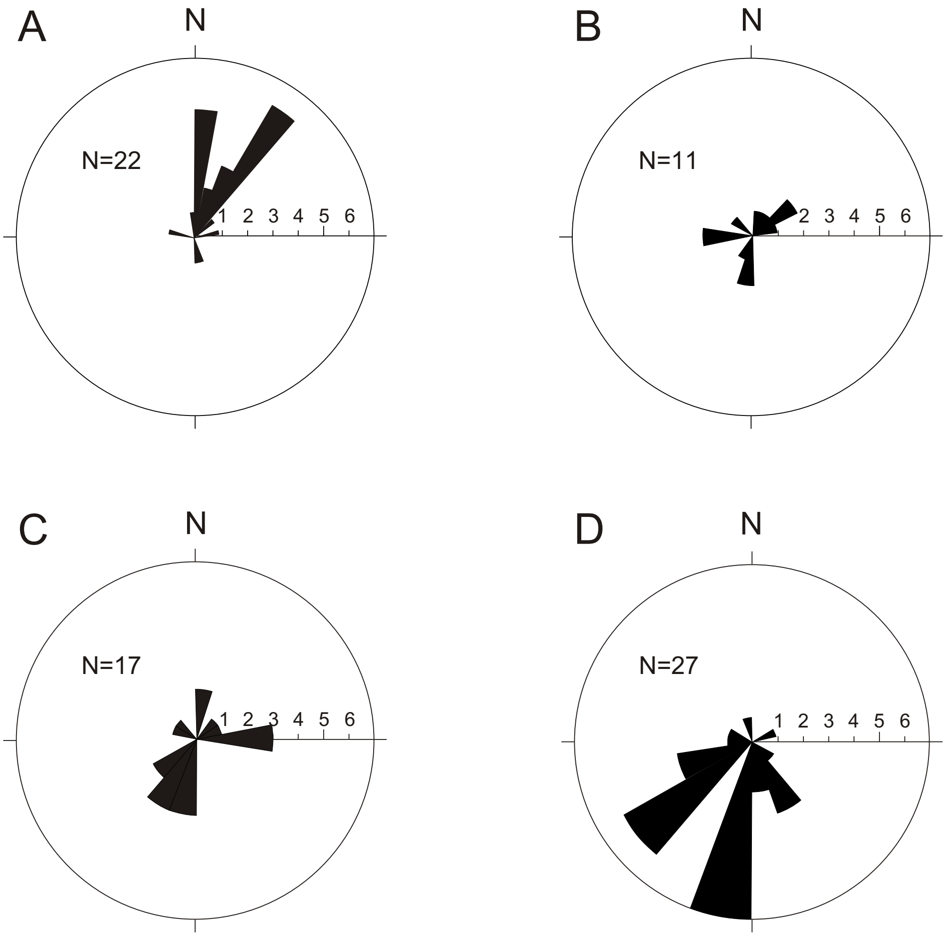


Fig. 6.2: Rose diagrams (linear scale, 20° classes) of foreset dip directions and trackway walking directions in the Muth Fm. near Mikkim. All measurements have been corrected by rotation of bedding surfaces back to horizontal. (A) Dip directions of foresets in facies association 2 (see also Fig. 5.1). (B) Orientation of *Diplichnites* traces without walking direction. (C) *Palmichnium* walking directions on bedding surfaces. (D) *Palmichnium* walking directions on foresets. Note that most of the *Palmichnium* traces are directed opposite to the dip of the foresets indicating uphill walking directions.

Fig. 6.3: *Palmichnium antarcticum* Form B trackway (trace 29) oriented along strike of a foreset with unusual gait, due to adjustment to the sloping surface. Walking direction from right to left (Coin diameter 25 mm).

6.17), as well as the multi-species eurypterid faunas at this time (*e.g.* Alken assemblage in Germany; Størmer 1976), the latter possibility seems more likely.

The abundance, straightness and sub-parallel orientation of most of the *Palmichnium* trackways suggests that the eurypterids responsible were not wandering about, nor spent a long time in this environment, but instead were migrating across it for a specific purpose. This might indicate that the beach/dune environment of FA2 did not represent the normal habitat of the eurypterids responsible for these trackways. In contrast, the orientation of *Diplichnites* traces does not show any preferred alignment (Fig. 6.2B). They are commonly curved, possibly indicating that the producer of these trackways ambled about more in this environment than the animals responsible for *Palmichnium*.

The main walking direction of the eurypterids, indicated predominantly by pushback mounds behind the tracks, are consistently towards the Southwest. This trend is clearly visible on bedding surfaces (Fig. 6.2C) and even stronger on dune foresets (Fig. 6.2D). The slip faces of the dunes are directed in the opposite direction towards the Northeast (Fig. 6.2A), as a result most of the *Palmichnium* traces run uphill, close to the dip line of the foresets, possibly indicating that eurypterids may have been migrating across the shoreline, likely to the lagoon behind the dunes. The observation that most *Palmichnium* trackways are oriented close to the dip line (Appendix 1) might be explained by comparisons with modern arthropods attempting to walk uphill on the slip faces of eolian dunes; they lost balance and rolled down the slope if they walked uphill oblique to the dip line (pers. comm. B. Grasemann 1997).

The *Palmichnium* trackways are generally relative uniform, with similar strides and series angles (Table 3). However, several trackways show modification of the gait, apparently influenced by the slope of the surface. Trackways walking uphill on dune slip faces generally show smaller stride values and higher series angles (Table 3). *Palmichnium* trace 29 crossing a foreset surface along strike clearly shows the influence of the slope on the producer; the series of the downhill side show four tracks with sediment mounds on the downhill-directed side, whereas on the uphill side the tracks were all overprinted (Fig. 6.3). A part of trace 36 shows a similar gait variation (Fig. 6.16). *Palmichnium* trace 2 consists of series of three tracks. The outer and middle tracks are teardrop-shaped, with pronounced pushback mounds. The pointed ends might be explained by scratching the surface in front of the tracks during the recovery stroke.

Hunter & Richmond (1988) investigated daily cycles of summer winds, which are caused by the pressure gradients set up by differential heating and cooling of water and land. These daily variations of the wind intensities and directions produce characteristic cyclic foreset deposits in coastal dunes. These cyclic variations involve foresets produced by climbing-wind-ripples, grainfall, sandflow, or a mixture of these types of deposition. In a typical daily cycle, a period of morning calm is followed by increasing onshore winds that reaches a maximum in mid-afternoon and decreases to an evening calm. The offshore winds blow during night (Hunter & Richmond 1988). Generally, the onshore sea breezes are stronger than the offshore land breezes, but the superposition of a synoptic wind system can modify this relation, or even suppress one direction. In their model one foreset represents one day.

In the case of the eolian dunes (FA2) in the Muth Fm. the foresets are relative uniform with thickness around 30 cm, possibly representing daily cycles, the cyclicity mainly defined by grainsize variations (Fig. 5.7). In our environmental interpretation slip faces were directed offshore, towards the (recent) Northeast (Fig. 5.1C), implying that they were produced by offshore land breezes during night. Arthropod trackways are nearly exclusively visible on foreset boundaries. This may be explained by the fact that foresets hardly ever split internally, therefore trackways on these surfaces cannot be found. Alternatively, trackways produced during periods of little or no sedimentation, forming the foreset boundaries (*i.e.* periods with less wind), are preferentially preserved. Consequently following these hypotheses and assuming suppressed sea breezes, the trackways on the foreset boundaries would have been produced anytime between the end of the land breeze early in the morning and the onset of the next land breeze in the evening, which covered them with the subsequent foreset.

Palmichnium traces 32 (Fig. 6.17) and 33 (Fig. 6.19) have nearly identical proportions and the walking directions differ by just 11 degrees (Table 3). The former occurs directly above the latter, separated by only two foresets (together 54 cm). If the hypothesis of daily cycles in the eolian dunes of FA2 were correct, this would imply that trace 32 has been formed two days after trace 33. It can be argued that two different individuals of the same species with by chance exactly the same size are responsible for the two traces, or exactly the same individual walked the same direction twice. The surprising constant walking directions of *Palmichnium* traces has been noted already, but if the latter assumption were right, it can be speculated that this could represent an exciting record of traces of exactly the same individual using the same route twice within two days.

The so-called "beaten track" (Draganits *et al.* 1998a, Fig. 6) apparently consists of several superimposed sub-parallel *Palmichnium* trackways occurring on a single bedding surface, *c.* 30 cm wide and traceable for more than 1.5 m long. Although the superimposition of the trackways and some erosion of this surface complicate the distribution of tracks, we suggest that this is more evidence that eurypterids used the same routes when they made these trackways.

6.5. Systematic ichnology

Ichnogenus *Didymaulichnus* Young, 1972

Didymaulichnus cf. *lyelli* (Rouault 1850)

(Fig. 6.4)

Material and localities: Low abundance. One example described herein (trace number 85). Lower part of a foreset at the western termination of the anticline of the Muth Fm., 0.4 km southeast of Mikkim, Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (GPS: N32°02'14"; E78°03'53"; EPE 22 m). All material remains *in situ* at the locality.

Description: Narrow bilobed epichnial, straight to slightly curved, parallel grooves, commonly 9 mm wide and 45 mm long, fading out at both ends. They appear as dark grooves in bright quartzite with a median ridge about 2 to 3 mm wide. Details of the traces have been abraded by the wind.

Fig. 6.4: *Didymaulichnus* cf. *lyelli* (trace 85) on a foreset surface at the western termination of the anticline (Coin diameter 25 mm).

Fig. 6.5: *Diplichnites gouldi* Form A (trace 86) on a foreset surface. Note the relative straight trackway segments separated by abrupt curves (Coin diameter 26 mm).

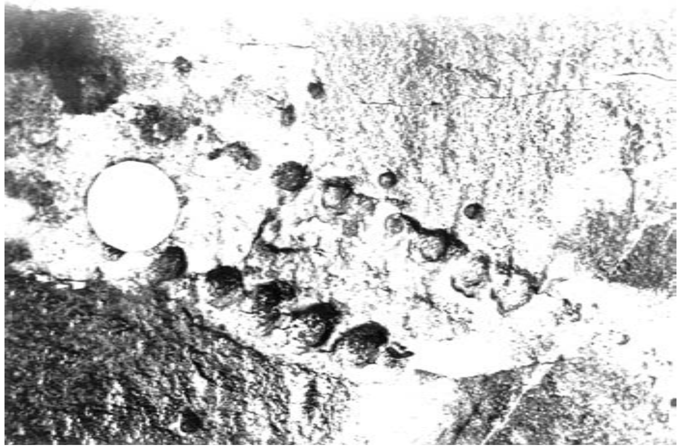


Fig. 6.6: Detail of Fig. 6.5. Walking direction probably towards the right, uphill; slightly raised ridge flanks on one side of the track rows (Coin diameter 26 mm).

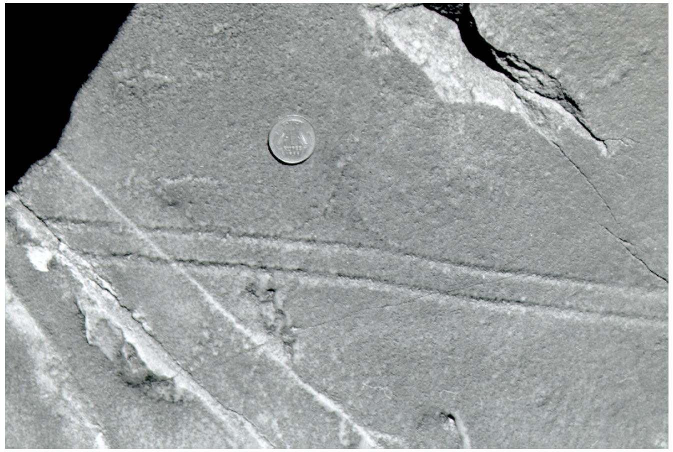


Fig. 6.7: *Diplopodichnus biformis* (trace 19) on a foreset surface (Coin diameter 25 mm).

Discussion: These traces closely resemble *Didymaulichnus* described by Trewin & McNamara (1995, Fig. 15), with the difference that the grooves there are weathered out. Trewin & McNamara (1995) suggested a possible formation by arthropods and noted that *Didymaulichnus* possibly represent the bottom half of *Didymaulyponomos* burrows. The occurrence of very similar *Didymaulyponomos* traces in the Muth Fm. support this interpretation.

Ichnogenus *Didymaulyponomos* Bradshaw 1981

Didymaulyponomos cf. *rowei* Bradshaw 1981

Material and localities: Low abundance. One example described herein (trace number 76). Bedding surface at the eastern termination of the anticline, 1.7 km ESE of Mikkim, Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (no GPS data). All material remains *in situ* at the locality.

Description: Straight to slightly curved, parallel grooves, 12 mm wide and sometimes more than 100 mm long. Burrows are preserved as open tunnels with a 3 mm wide median ridge between the two grooves. Burrows commonly show crosscutting relationships. No internal details are preserved.

Discussion: See the discussion of *Didymaulichnus*.

Ichnogenus *Diplichnites* Dawson 1873 (emended by Briggs *et al.* 1979)

Diplichnites gouldi (Gevers *et al.* 1971)

(Figs. 6.5, 6.6, 6.8-6.10)

Material and locality: Nine trackways (trace numbers 3, 4, 5, 11, 27, 28, 38, 48 and 86) on bedding surfaces and foresets at the western termination of the anticline of the Muth Fm, 0.4 km southeast of Mikkim (GPS: N32°02'14”; E78°03'53”; EPE 22 m). Six trackways (traces 69, 70, 71, 72, 73, 77) on a bedding surface near the crest line of the anticline directly at the edge of the prominent cliff, Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (no GPS data). All material remains *in situ* at the locality. One Formasil RTV© peel of trace 3 is housed in the collections of the Institute of Geology, University of Vienna, Austria (not yet catalogued).

Description: Terminology follows Trewin (1994). Trackways vary considerably in their external width, series arrangement and the shape of single tracks. Following Trewin & McNamara (1995) and Buatois *et al.* (1998b) these trackways are not divided into different ichnotaxa, to avoid a profusion of names. Within the variation of the trackways two end-members have been recognized, which are described separately as Forms A and B, with no taxonomic significance.The size distribution of the traces is presented in Appendix 1.

Fig. 6.8: *Diplichnites gouldi* Form A (trace 3) on the same bedding surface as several *Palmichnium antarcticum* trackways (not visible hear. Coin diameter is 25 mm.

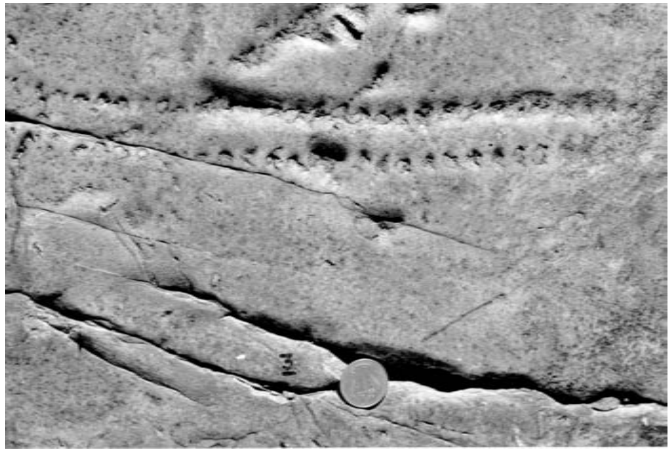


Fig. 6.9: Detail of figure 6.8.

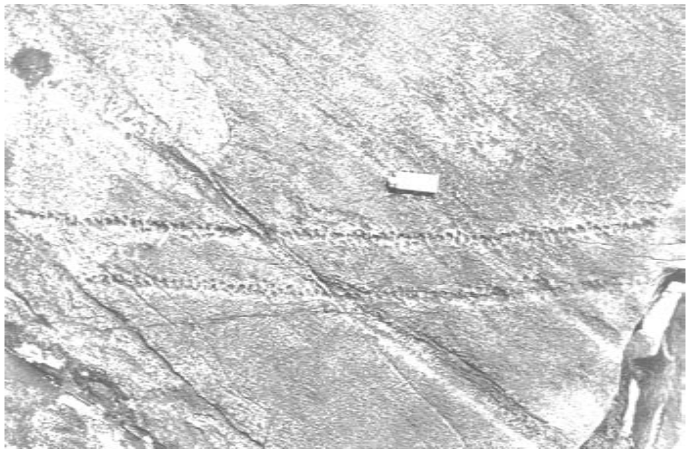


Fig. 6.10A: *Diplichnites gouldi* Form B (trace 77) on a bedding surface with parting lineation. Lighter is 60 mm long. Note the difference to Form A traces.

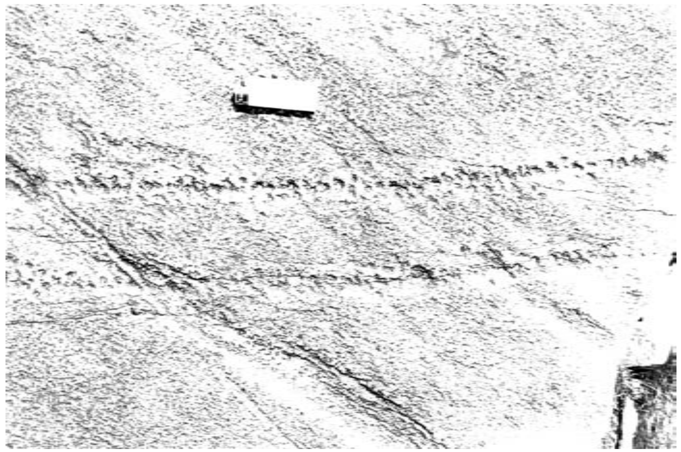


Fig. 6.10B: Detail of figure 6.10A.

Diplichnites gouldi Form A (trace numbers 3, 4, 5, 11, 27, 28, 38, 48 and 86; Figs. 6.5, 6.6, 6.8, 6.9) consists of two parallel rows of closely spaced tracks. External width ranges from 18 to 140 mm, the internal width is 15 to 75 mm. Some trackways can be traced for nearly 1 m. Trackways are straight to gently curved, it is common that straight parts are separated by relatively abrupt curves. In some trackways a slightly raised ridge flanks one or both outer margins of the track rows (Fig. 6.6). Well-preserved tracks vary from ellipsoidal to circular pits. Tracks are relatively large compared with the external width, they are arranged obliquely, typically inclined around 60° to the mid-line, pace ranging between 8 to 27 mm. The symmetry of opposing tracks may change from opposite to staggered within a single trackway. No clearly discernible series were detected.

Diplichnites gouldi Form B (trace numbers 69, 70, 71, 72, 73, 77; Fig. 6.10) consists of two parallel rows of small conical to elongate tracks, the external width ranging from 44 to 77 mm, the internal width is 51 to 107 mm. Trackways are straight and can be traced for about 60 cm across the surface. Individual tracks are commonly superimposed, elongate tracks are arranged obliquely with angles around 60° to the mid-line, and pace ranging between 12 to 32 mm. The symmetry of the opposing tracks is opposite to slightly staggered. Series are difficult to determine because of the straightness of the trackways, but in some cases there is evidence for at least 8 tracks per series (trace 77), arranged at a low angle to the mid-line.

Discussion: *Diplichnites* Form A and B were found in different parts of the Muth Fm., but at a similar stratigraphic level. Form A occurs in combination with *Palmichnium* and the majority of all other traces at the western end of the outcrop, whereas Form B was found in only one place, on bedding surfaces near the crest of the anticline directly at the edge of the prominent cliff. Traces 69-73 appear on just one single surface, trace 77 is just 5 m away, their appearance resembles closely *Diplichnites* Form B described by Trewin & McNamara (1995).

These trackways were most probably produced by myriapod-like animals (see Johnson *et al.* (1994) and Buatois *et al.* (1998) for a discussion of the possible trailmakers). Differences between the two different forms of *Diplichnites gouldi* described herein may indicate that they probably were produced by different types of myriapods, although behavioural and preservational differences of one producer might also account for these variations.

Ichnogenus *Diplopodichnus* Brady 1947

Diplopodichnus biformis Brady 1947

(Fig. 6.7)

Material and localities: Very low abundance. One example described herein (trace number 19). Foreset at the western termination of the anticline of the Muth Fm, near the water line of the Pin River, 0.4 km southeast of Mikkim, Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (GPS: N32°02'14"; E78°03'53"; EPE 22 m). All material remains *in situ* at the locality.

Description: This trail consists of two epichnial parallel grooves, each 2.5 mm wide, with an external width of 20 mm and an internal width of 15 mm. The trail runs straight on a foreset surface, hardly any details can be seen because of the small size relative to the sand-sized sediment. In some parts faint tracks, 3.6 mm apart, are detectable within the track rows.

Discussion: *Diplichnites gouldi* and *Diplopodichnus biformis* were considered by Buatois *et al.* (1998b) to represent compound ichnotaxa capable of being produced by the same animal, probably a small myriapod. Buatois *et al.* (1998b, Figs. 2A-B, 3A) documented examples of specimens that laterally grade from one ichnogenus into the other. Their differing morphology is explained by variations related to substrate consistency (Johnson *et al.* 1994). In their interpretation *Diplopodichnus* has been produced under moistened to wet surface conditions, *Diplichnites* under dryer conditions.

Ichnogenus *Metaichna* Anderson 1975
Metaichna isp.
(Figs. 6.11, 6.12)

Material and localities: Moderate abundance. Trace number 75 on bedding surface at the crest line of the anticline southeast of Mikkim. Four traces (trace numbers 80, 81, 83 and 84) on bedding surfaces at the western termination of the anticline of the Muth Fm, 0.4 km southeast of Mikkim, Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (GPS: N32°02'14”; E78°03'53”; EPE 22 m). All material remains *in situ* at the locality.

Description: Traces included in this ichnospecies appear in two different forms, (A) vertical tube-like and (B) hemispherical with elliptical shape in plan view. These two endmembers are described separately as Forms A and B. These forms are not indented to have taxonomic significance.

Metaichnia Form A (trace number 75). This trace consists of vertical tube-like structures weathered into the surface. They are irregular circular in plan view with a diameter of about 70 mm. Margins are irregular, no special wall construction is visible. Pits are up to 40 mm deep, most of them shallower, with a relatively sharp contact between vertical tube and bottom, implying that the bottom is not the end of the trace, but the level of downward erosion. *Metaichnia* Form A is found on a bedding surface, single holes show a distance of at least 20 cm from neighboring holes. There are abundant sub-horizontal, small *Taenidium* traces on the same surface, some of them are crosscutting *Metaichnia* Form A, but the exact relation between them is not clear.

Metaichnia Form B (trace numbers 80, 81, 83 and 84). This trace shows elliptical to elongated shallow hollows in bedding surfaces, with a length of 101-203 mm and 39-142 wide. They have a convex downward shape with smooth, unstructured downward tapering margins (Fig. 6.11, 6.12).

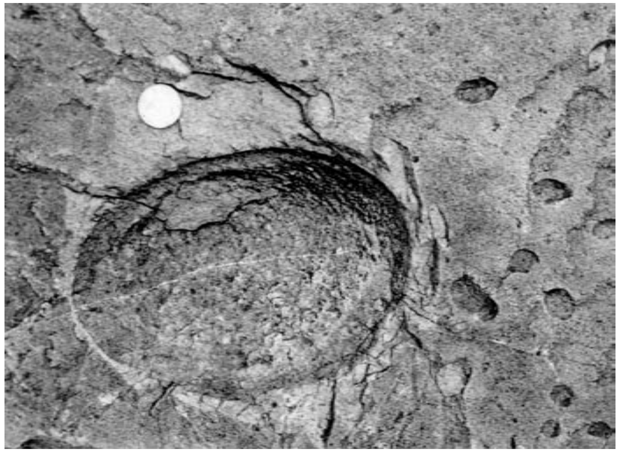


Fig. 6.11: *Metaichnia* isp. Form B with typical sharp boundaries (trace 83) on the upper bedding surface of bed M_a566. Coin diameter is 25 mm.

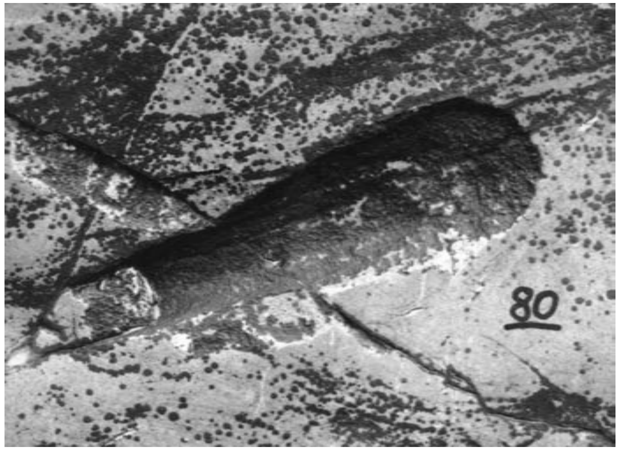


Fig. 6.12: *Metaichnia* isp. Form B on the upper bedding surface of bed M_a566.

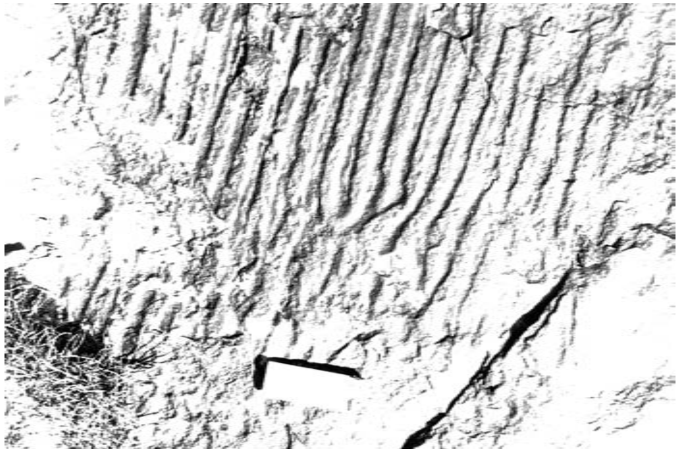


Fig. 6.13: Trace with unclear taxonomy. Regular meandering trace (trace 74) on a bedding surface. Lighter is 60 mm long. Trace resembles Palaeohelminthoidea, but is much larger and occurs usually in a different environment.

Fig. 6.14: Several *Palmichnium antarcticum Form A* trackways (traces 2 and 37) on a single bedding surface (Lens cap diameter 53 mm).

Except some vertical burrows no other traces have been found on the same surfaces. Sandfill is massive and comprises the same grain size as the surrounding sediment.

Discussion: The different forms of *Metaichna* occur in the Muth Fm. southeast of Mikkim in different places, but both within FA2. Form A is found associated with *Taenidium* near the crest line of the anticline, Form B at the western termination together with all other traces. As a contrast to the description of Anderson (1975) the sandfill in the Indian traces is the same material, as the surrounding sediment, but this feature seems to be controlled by the availability of different sediment, thus not possessing taxonomic value. Rust (1967) interpreted the South African examples as burrows excavated by trilobites. Anderson (1975) suggested a resting or sheltering burrow (cubichnial) interpretation, perhaps involving an arthropod.

Metaichnia in the Muth Fm. can tentatively be related with *Taenidium* traces, because both comprise burrows often appearing on the same surfaces. In *Taenidium* usually the fill of the burrow is preserved, in contrast *Metaichnia* are weathered into the surfaces and no back-fill structures are visible. Additionally, in the Mikkim section they show considerable difference concerning size. On the other hand, *Taenidium* (trace 88) on a loose block near Muth shows similar sizes to *Metaichnia* traces and from the curved appearance, it might be possible to compare the curved part of this trace with *Metaichnia* Form B and the vertical part with *Metaichnia* Form A.

Ichnogenus *Palmichnium* Richter 1954

Palmichnium antarcticum (Gevers *et al.* 1971)

(Figs. 6.14-6.19)

Material and localities: Very high abundance (trace numbers 1, 2, 8, 9, 12, 13, 14, 15, 16, 17, 18, 20, 21, 22, 24, 25, 26, 29, 30, 31, 32, 33, 34, 35, 36, 37, 39, 40, 41, 42, 43, 44, 45, 46, 47, 49, 50, 51, 52, 53, 54, 55, 56, 57, 58, 59, 60, 61, 62, 63, 64, 65, 66, 67, 68 and 87). Bedding surfaces and foresets at the western termination of the anticline of the Muth Fm., 0.4 km southeast of Mikkim, Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (GPS: N32°02´14"; E78°03´53"; EPE 22 m). All material remains *in situ* at the locality. Formasil RTV© peels of traces 2, 33, 36 and 37 are housed in the collections of the Institute of Geology, University of Vienna, Austria (not yet catalogued).

Description: The trackway attributes are presented in Table 3 according to the terminology of Trewin (1994). Although these trackways vary considerably in their width, series arrangement, number of tracks per series and the shape of single tracks, they have not been divided into different ichnotaxa, to avoid a profusion of names. However, the variations within the numerous trackways allowed the recognition of two endmembers, described separately as Forms A and B, which are not indented to have taxonomic significance.

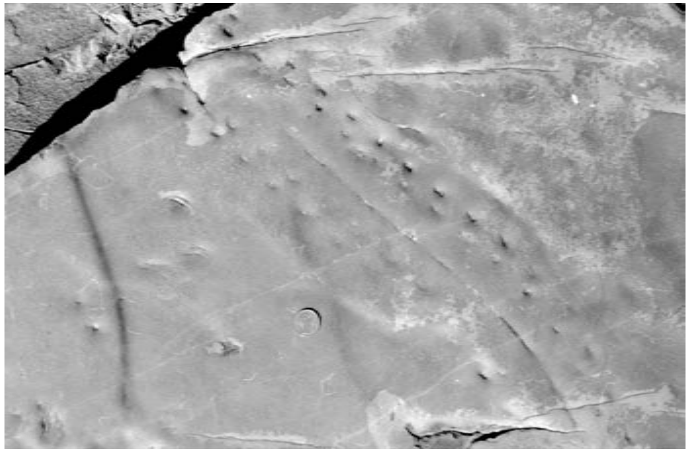


Fig. 6.15: Nice *Palmichnium antarcticum* Form A trackway (trace 12); note regular variation of the telson imprint depth related to the gait and 2 elongate tracks per series. First trace found in this outcrop.

Fig. 6.16: *Palmichnium antarcticum* Form B (trace 36) shows 4 track per series, series are V-like arranged. Note the overlap of tracks in the upper right. Coin is 25 mm in diameter.

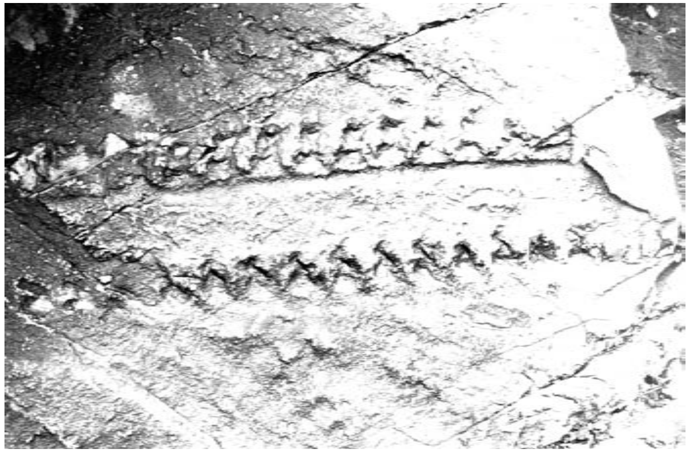


Fig. 6.17: *Palmichnium antarcticum* Form A (trace 32) on a foreset walking uphill (to the right). Note the telson drag far left from the mid-line and the stunning similarity to Fig. 6.18.

Fig. 6.18: H.P. Schmid carefully removing a Formasil RTV peel of trace 2 and 37.

Palmichnium antarcticum Form A (trace numbers 1, 2, 12, 13, 14, 15, 16, 17, 18, 20, 21, 22, 24, 25, 26, 30, 31, 32, 33, 34, 35, 37, 39, 40, 41, 42, 43, 44, 45, 46, 47, 49, 50, 51, 52, 53, 54, 55, 56, 57, 58, 59, 60, 61, 62, 63, 64, 65, 66, 67, 68 and 87) consist of straight or rarer slightly curved trackways consisting of two parallel rows of track series, found on bedding surfaces and foresets. External width ranges from 63 to 498 mm with internal width between 25 to 245 mm. Series generally show opposite or staggered symmetry (rarely alternate) and consist of three tracks in well preserved examples and of two tracks in badly preserved examples (*i.e.* lacking the shallower inner track). Series are mainly straight but also slightly curved series are present, with the concave side directed in the walking direction. Angle between series and the mid-line ranges from 23 to 90°, with most examples between 40 to 75°.

The outer and inner tracks are very similar, but differ from the inner track. The former are relatively large, deep and mainly elongate to teardrop-shaped, arranged sub-parallel to oblique to the mid-line (Fig. 6.19). In places where the trackway is eroded to a deeper level, it is visible that these have been made by a limb with bifid distal spines. Well preserved trackways display pushback mounds behind the outer and middle track, some often more than 6 mm high (Fig. 6.17, 6.19).

The inner tracks, circular-ellipsoidal imprints, where preserved, are smaller and shallower than the outer tracks and lack pushback mounds (Fig. 6.19). Inner tracks are up to 1 cm in diameter and where ellipsoidal shaped, arranged at right angle to the mid line.

Very few trackways display a medial impression. Those that do rarely occur along the mid-line, more commonly displaced to one side. Where they occur, the medial impressions vary in width from 0.2 to 1.5 cm, with a depth of 0.1 to 0.4 cm, and are semi-circular in cross-section. Some medial impressions show weak regular variations in depth (*e.g.* trackway 12), reflecting variations in the force that the telson was applied to the substrate during walking (Fig. 6.15).

Palmichnium antarcticum Form B (trace numbers 8, 9, 29 and 36) consists of straight to slightly curved trackways composed of two parallel rows of track series, found on bedding surfaces and foresets. External width ranges between 74 to 170 mm, internal width is between 28 to 46 mm. Series comprise four circular conical tracks of equal size with an opposite to staggered (occasionally alternate) symmetry, and a stride of 49 to 82 mm. Series are generally straight but may grade into slightly curved series within the same trackway, the concave sides directed in the walking direction. The angle between series and the mid-line ranges from 29 to 74°. No medial impressions were found associated with this form.

Discussion: Braddy & Milner (1998) recently referred various trackways of this type to this ichnospecies. Both forms of *Palmichnium* described herein fall within their emended diagnosis, although they show a range of morphologies. *Palmichnium* Form A was produced by large heteropodous, hexapodous animals, whereas *Palmichnium* Form B was produced by smaller, octopodous animals. Trackways with series of four tracks (Form B) are less than 200 mm in external width and 100 mm in internal width (Fig 6.20), whereas those of Form A occur across the

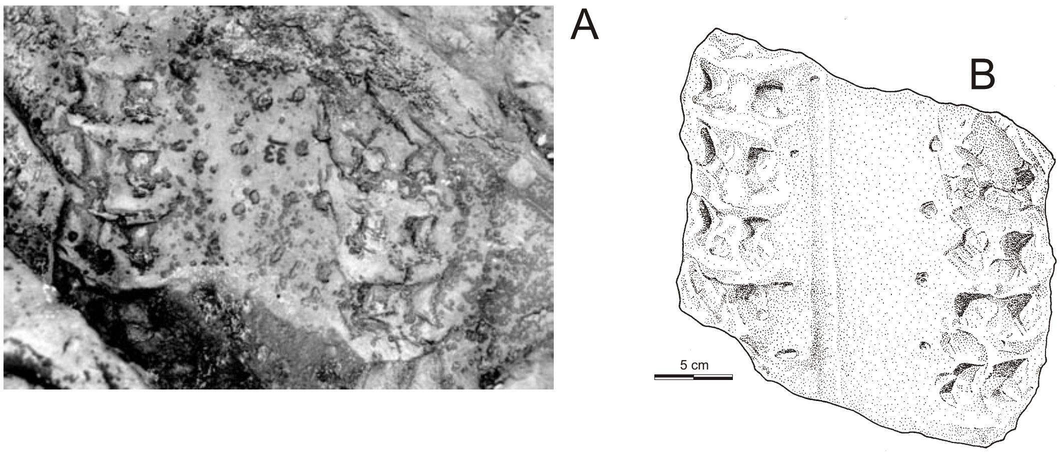


Fig. 6.19: Best preserved *Palmichnium antarcticum* Form A trackway (trace 33). **A)** Photo of the trackway on a foreset of bed M_a566 running uphill (walking direction from bottom to top) with large back-push mounds. Note the similar elongated middle and outer track oriented obliquely to the middle line, the much smaller, circular to elliptical inner track and the telson drag left of the middle line. **B)** Drawing from a replica by Leo Leitner.

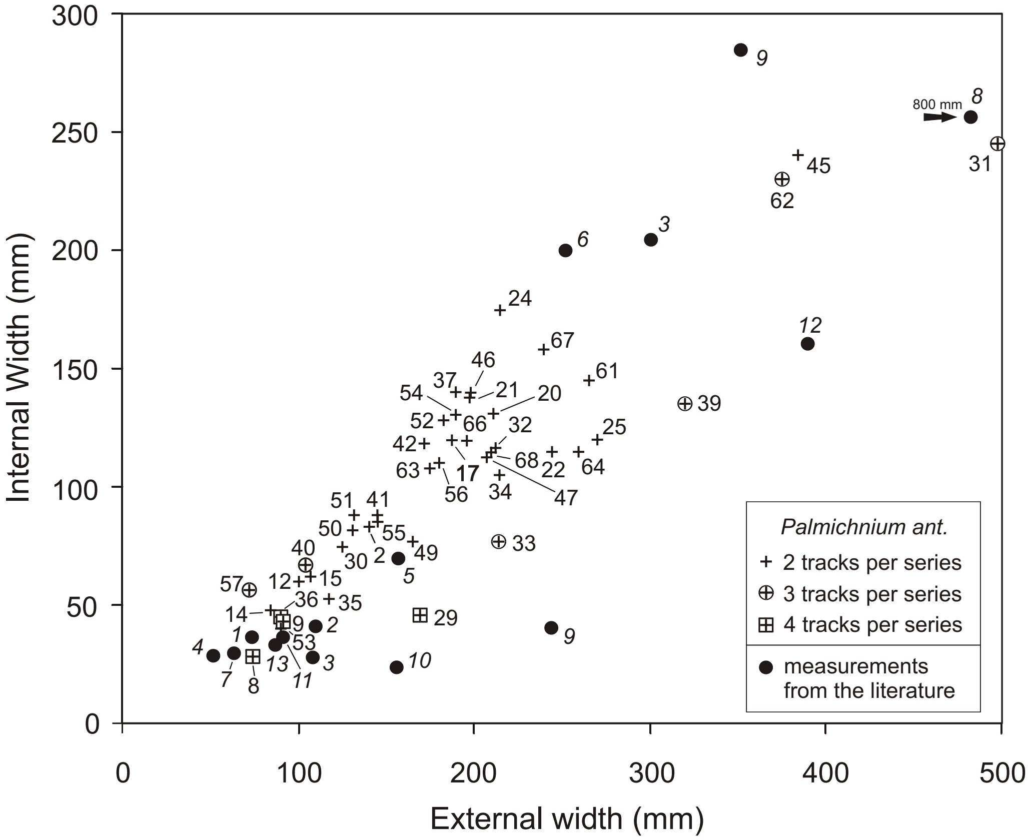


Fig. 6.20: Diagramm of external width *versus* internal width of *Palmichnium antarcticum* trackways. Trackway attributes from the Muth Fm. seem to cluster in 3 size ranges. Note that trackways with four tracks per series are confined to the smallest size group. *Palmichnium* size measurements from the literature with italic numbers. 1: Sharpe (1932), 2: Richter (1954), 3: Gevers (1971), 4: Greiner (1972), 5: Hanken & Størmer (1975), 6: Friend (1976), 7: Bradshaw (1981), 8: Briggs & Rolfe (1983), 9: Rogers 1987, (1990), 10: Trewin & McNamara (1995), 11: Braddy & Anderson (1996), 12: Braddy & Milner (1998), 13: J. Waddington (pers. comm., 1998).

whole size range (Fig. 6.20). Fig. 6.20 compares the Indian *Palmichnium* trackways with other eurypterid trackways described in the literature, which show a similar range in size.

Only a few Devonian arthropods were big enough to have produced these trackways. Eurypterids or scorpions are prime candidates. Besides their size, there are several arguments in favour of eurypterids. Gevers *et al.* (1971) preferred a eurypterid to a trilobite as the producer of the Antarctic trackways, however cited the fact that “the regular pattern of footprints here indicates an animal capable of an extremely well balanced gait, in no way encumbered with an unduly long abdominal section such as characterised by most eurypterids” (Gevers *et al.* 1971, p. 90). Briggs *et al.* (1979), Rolfe (1980) and Trewin & McNamara (1995) considered that large amphibious scorpions produced these trackways. Bradshaw (1981), however, suggested that these trackways were produced subaqueously by stylonurid eurypterids.

The large size of the tracks, the morphology of the trackways, and absence of any paddle-shaped outer tracks indicate the Indian *Palmichnium* traces were produced by stylonurid eurypterids, many Devonian forms attaining sufficient size to have produced these trackways (cf. Waterston 1979). The bifid tracks indicate that the distal podomere of the leg had two movable spines or claws. The differences between *Palmichnium* Forms A and B may indicate that they were produced by different species of eurypterids, although the variations in trackway morphology (*e.g.* number of tracks per series) may also have been in response to the size, and locomotory capabilities of different instars of one species. The first possibility seems to be more probable calling back to mind for example the multi-species Devonian eurypterid fauna of Alken (Størmer 1976).

The Indian *Palmichnium* trackways generally show opposing series diverging in the direction of movement, the walking direction determined by pushback mounds behind the tracks. This is in clear contrast to the general assumption that eurypterid trackways display opposing series converging in the direction of movement, as in the xiphosuran trackway *Kouphichnium*. However, the morphological diversity of eurypterids would have ensured that their walking techniques, and hence their trackways, were equally diverse (Braddy & Milner 1998). Different forms could have produced trackways with series that converge or diverge in the direction of travel.

Most examples of *Palmichnium antarcticum* show an opposing or staggered symmetry. Several examples from this ichnofauna, however, show opposite series gradually grading into staggered or even alternating series within a single trackway, apparently reflecting differences in slope angle, surface conditions and walking speed. Symmetry alone, therefore is not a significant ichnotaxobase to discriminate ichnotaxa such as *Paleohelcura* and *Palmichnium*, although if used in conjuction with external width and track size, it may be useful (Braddy & Milner 1998).

Ichnogenus *Selenichnites* Romano & White 1987

Selenichnites isp.

(Fig. 6.21)

Material and localities: Low abundance (trace number 89). Several traces on single bedding surface in the upper most part of the type section of the Muth Fm., 1.2 km south-southeast of Muth, Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (no GPS data). All material remains *in situ* at the locality.

Description: Poorly preserved (weathered) horseshoe-shaped epichnial hollows on bedding surfaces, with lateral margins of trace parallel or diverging (at about 30°). Width ranges between 25 to 90 mm, mostly from 40 to 60 mm. Trace deepest at anterior, but in some cases also deepest laterally, up to some 5 mm below the bedding surface. Traces consistently arranged with the apices pointing towards the Southwest.

Discussion: In contrast to all other traces described herein, *Selenichnites* occurs within FA4, interpreted as a shallow marine environment (see above), and is not part of the trackway-dominated ichnoassemblage of FA2. Although paleocurrent directions vary in FA4, the preferred southwestern alignment of *Selenichnites* might indicate an orientation of the apices against the dominant depositional currents.

Selenichnites is known from five ichnospecies: *S. rossendalensis* (Carboniferous, U.K.) *S. cordoformis* (Ordovician, Colorado) *S. bradfordensis* (Carboniferous, U.K), *S. hundalensis* (Jurassic, U.K; see Romano & Whyte 1987 for review), and *S. langridgei* (Siluro-Devonian, Western Australia; Trewin & McNamara 1995). Additionally, *Selenichnites* isp. has been reported from the ?Lower Ordovician of western Antarctica (pers. comm. B. Weber 1999).

Comparing the size and overall shape, the Indian material is most similar to *S. langridgei*, although the lack of the medial trefoil-shaped feature (a characteristic of this ichnospecies), and their poor preservation, prevents assignment to this (or any other) ichnospecies. *Selenichnites* is generally regarded as representing digging traces of a xiphosurid, although other arthropods (?eurypterids, crustaceans) could conceivably leave similar traces by employing similar behaviour.

Ichnogenus *Taenidium* Heer 1877 (emended by Keighley & Pickerill 1994)

Taenidium barretti Bradshaw 1981 (emended by Keighley & Pickerill 1994)
(Figs. 6.22)

Material and localities: Moderate abundance in various orientations. Five well-preserved near axial sections (trace numbers 6, 7, 23,79 and 88), on bedding and foresets surfaces. Two near axial sections (traces 79 and 88), on bedding surfaces at the eastern termination of the anticline, 1.7 km ESE of Mikkim (GPS: N32°02'06"; E78°04'47"; EPE 23 m), Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India. All material remains *in situ* at the locality.

Description: Gently curved endichnial burrows with horizontal to slightly oblique orientation to the bedding surfaces, several tens of centimeters in lateral extent, commonly fading-out across the bedding plane. Traces usually are crowded and crosscutting relationships are common. Burrows

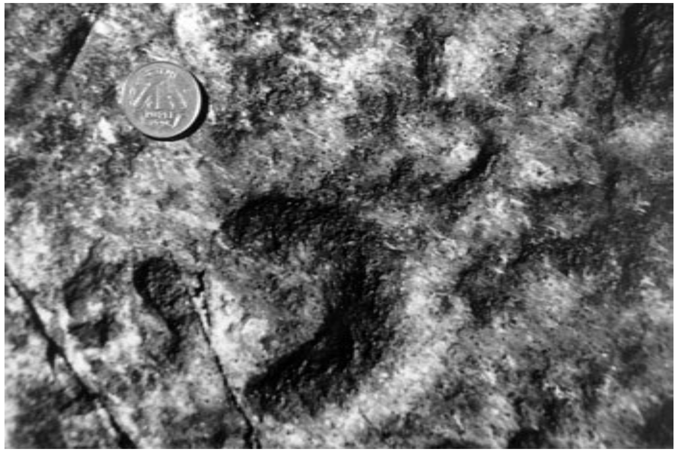


Fig. 6.21: *Selenichnites* isp. (trace 89) on a bedding surface in the uppermost part of the Muth Fm. (bed M_e522) near Muth. Coin diameter is 25 mm.

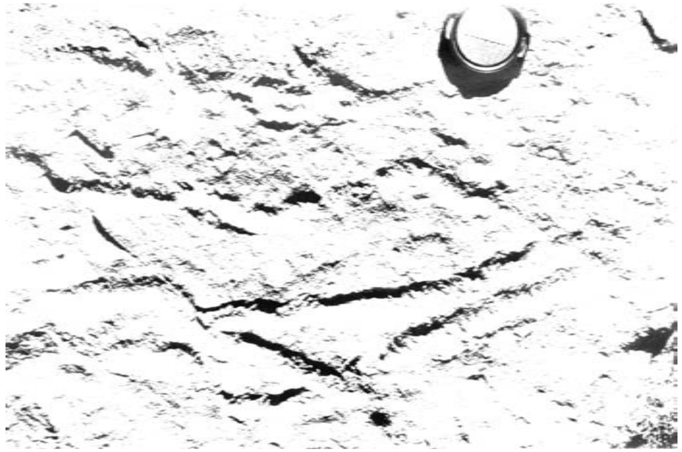


Fig. 6.22: *Taenidium* (trace 79) on a bedding surface. Note the meniscate back-fill packets and the lack of any special wall structures. Coin diameter is 25 mm.

Fig. 6.23: Vertical burrows with unclear taxonomic position in plan vie w on a bedding surface. Note the raised rims that indicate slightly cohesive behavior of the arenite during formation.

Fig. 6.24: Plan view of a tangential foreset. Openings of vertical burrows are abundant at the base of the foreset (right), but gradually become less towards the top (left). Note the sandflow in the center.

circular in cross-section, usually are some 4 to 30 mm wide, commonly weathering in negative relief, although some examples show weak positive relief. The sand fill of the burrows is massive or shows poorly defined meniscate back-fill packets. Neither special wall structure nor differences of the sediment within the burrow, compared to the host-rock, is observed (Fig. 6.22).

A loose block from the Muth section shows a part of a 65 mm wide burrow starting oblique to the bedding surface bending to a vertical direction; due to different colored, well laminated quartzite layers, the internal structures are clearly visible, showing very nice meniscate back-fill packets.

Discussion: The ichnotaxonomic problems associated with meniscate filled burrows were discussed by Keighley & Pickerill (1994). Following their definition of *Beaconites* (possessing a wall structure) and *Taenidium* (lacking a wall structure) the burrows in the Muth Fm. are attributed to *Taenidium*.

Trewin & McNamara (1995) noted two types of *Beaconites* (=*Taenidium*; Keighley & Pickerill 1994) from the Tumblagooda ichnofauna. Smaller ones (*B.* cf. *antarcticus*) were attributed to the *Heimdallia* animal, whereas larger ones were attributed to the animal responsible for the larger *Diplichnites* trackways. Indeed, Gevers *et al.* (1971) recorded *B. antarcticus* on the same bedding planes (and predating) small *Diplichnites* trackways, which occasionally ended in small rounded vertical pits. Rounded terminations of myriapod trails were also reported from the Lower Permian of southern New Mexico, and interpreted as representing the point where the animal began burrowing beneath the substrate (Braddy 1995).

The association of sub-horizontal meniscate backfilled burrows (*Taenidium*) and vertical burrows (*Metaichnia* Form A) on a single bedding surface might indicate some relationship of their formation. Indeed reports of similar burrows (*e.g.* O'Sullivan *et al.* 1986; Allen & Williams 1981) show these traces in various orientations to the bedding surface. On the other hand all observed *Taenidium* traces in the Muth Fm. are considerable smaller than *Metaichnia* that additionally lack meniscate backfill structures, therefore both traces are described separately.

Taenidium has been variously interpreted as having been produced by polycheate worms (Gevers *et al.* 1971), a worm-like animal (*e.g.* a amphisbaenid (worm lizard) or a blind snake) which ingested the substrate, or burrowed through it (Ridgeway 1974), limbed reptiles or amphibians (Pollard 1976), ostracoderms (Allan & Williams 1981), lungfish (O'Sullivan *et al.* 1986), “desert dwelling arthropods” (Rolfe 1980), “myriapod-like arthropods” (Bradshaw 1981) and arthropleurids (Pearson 1992).

In the Muth ichnofauna, the narrow width of the near-axial *Taenidium* sections is similar to that of the narrower *Diplichnites* and *Diplopodichnus* trackways (here attributed to myriapods). Larger examples (*e.g.* trace 88) also have a similar width to larger *Diplichnites*. It is possible, therefore, that these *Taenidium* burrows may be assigned to myriapods or arthropleurids. It is possible that different types of animals may have formed the different ichnospecies of *Taenidium*. The precise producer(s) and mode of formation of this enigmatic ichnotaxon remains unsolved.

Meandering trail
(Fig. 6.13)

Material and locality: One example (trace 74), on bedding surface near the western termination of the anticline of the Muth Fm., 0.6 km southeast of Mikkim, *c.* 50 m SSE of the prominent chorten (Tibetan stupa), Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (GPS: N32°02'01"; E78°04'38"; EPE 41 m). All material remains *in situ* at the locality.

Description: Very regular, parallel, closely spaced, meandering grooves extending for more than 20 cm. Grooves are some 10 mm wide and show a narrow median cord-like ridge, which is lower than the bedding surface. The preserved total relief is less than 3 mm.

Discussion: The regularity of the trail implies a characteristic grazing behaviorial pattern of the animal responsible. Similar traces such as *Helminthoida* or *Palaeohelminthoida* are interpreted as surface feeding traces, but are of smaller size and are found mainly in flysch-type deposits. The affinities of this trace remain unclear.

Vertical burrows
(Fig. 6.23, 6.24)

Material and localities: Very high abundance, but due to the uniform appearance only two traces (traces 10 and 82) are mentioned here. Bedding surfaces and foresets at the western termination of the anticline of the Muth Fm., 0.4 km southeast of Mikkim, Pin Valley (Lahaul-Spiti district, Himachal Pradesh), India (GPS: N32°02'14"; E78°03'53"; EPE 22 m). All material remains *in situ* at the locality.

Description: Vertical or near vertical burrows, on bedding and foreset surfaces, circular to oval in cross-section, diameter 6 to 22 mm. They commonly occur in dense concentrations, sometimes paired or in rows. In some cases they seem to be crowded on the toe of dune foresets, becoming gradually less abundant upwards. No vertical section was found. Sandfill slightly weathered below surface level, but no differences between the fill and surrounding sediment were recognized. Some examples show steeply raised rims around entrances of burrows (Fig. 6.23), implying that there was some binding agent for stabilization of the excavation mound.

Discussion: These traces are very common in the Muth Fm. as small depressions in bedding and foreset surfaces. Various vertical burrows have been described from marine to non-marine environments. As long as the ichnotaxonomy remains unclear, given the lack of vertical sections, their environmental implications remain speculative. In analogy to the Australian (Trewin & McNamara 1995) and Antarctic ichnofaunas (Bradshaw 1981), *Skolithos* sp., *Diplocraterion* or *Tigillites* are probable candidates for these burrows.

7. CONCLUSIONS

In the HH, there is no clear boundary between HHC and TH; these units show a conformal contact in several places, with a gradual increase in metamorphic grade towards lower structural levels. Therefore metasediments of the HHC represent equivalents of lower stratigraphic levels of the TH and the use of metamorphic grade to define the boundary between these two units is rejected due to its gradual nature. A stratigraphic definition of this boundary is suggested and defined at the angular unconformity between Late Cambrian and Early Ordovician sediments that indicates widespread deformation, erosion and granitic intrusions in this time interval. Furthermore it is recommended to replace the misleading term “Higher Himalaya Crystalline” by the term “Vaikrita Himalaya”, representing metasediments and intrusions of the HH below the Early Ordovician unconformity, but keep the well-known and useful term “Tethyan Himalaya”, representing sediments and volcanics above this unconformity.

The prominent diamictites of the Manjir Fm. (Haimanta Group) enable the reliable correlation of HH metasediments with less deformed and less metamorphosed late Proterozoic series in the LH. As a consequence the Manjir Fm. is tentatively correlated with the upper diamictite of the Blaini Group of the LH, possibly representing remnants of the Marinoan Glaciation (some 600 Ma ago), the carbonates at the base of the Phe Fm. are characteristic features of the de-glaciation of this period.

Paleocurrent directions in the Chamba Fm. and Phe Fm. (Haimanta Group) defined by flute casts, chevron marks, foreset-dip show a dominant sediment transport direction broadly to the South, whereas the Simla Slates of the Lesser Himalaya have a dominant sedimentary transport direction to the NW-NE. As a consequence the model of a Proterozoic sedimentation of the Higher and Lesser Himalaya both on the northern shelf of India, with the Haimanta Group representing just distal equivalents of Lesser Himalayan series, turns out to be oversimplified.

The Cambrian to Triassic sequences in the Pin Valley show several major depositional breaks. Besides the above mentioned angular unconformity at the base of the Shian Fm. (Late Cambrian to Early Ordovician), there is a disconformity at the base of the Muth Fm. (Late Silurian?) and again an angular unconformity between the Lipak and Gechang Fms. (Early Carboniferous to Early Permian).

Except two calcareous horizons in the Pin Fm. the whole sedimentary sequences up to the Lipak Fm. are strongly dominated by siliciclastic sediments, deposited on a shallow shelf, with a wide continuity of the lithologic units. Two thickening and coarsening upwards mega-cycles exist, starting with thin-bedded sandy dolomite grading into coarse-grained bioclastic limestone with corals.

The Muth Fm. represents one of the largest sand accumulations in the world with an amazing purity of the arenites and lateral extent. In the Pin Valley the Formation is interpreted as Early Devonian barrier island system. This thick succession of monotonous highly mature quartz arenites indicates a relatively stable balance of sea level, sediment supply and tectonics during its deposition. Besides a thin dolomitic horizon in the upper part, the Formation consists of pure quartz arenites. Based on variations of dominant sedimentary structures four facies associations have been distinguished that correlate within two sections measured at the type locality and in a restored distance of 13 km near Mikkim.

Arthropod trackways in facies association 2 of the Muth Fm. represent one of the best preserved trackways of this age in the world. Many of the trackways are interpreted to have been produced subaerially and therefore represent an exciting record of early terrestrialization. Due to the big number of trackways, variations of size, gait modifications and track orientation within single ichnospecies have been recorded. This ichnofauna is very similar to other trackway associations in Australia and Antarctica and probably indicate the existence of a recurrent Lower Devonian ichnocoenosis around the margins of eastern Gondwana.

Paleocurrent analyses indicate a NW-SE trend of the coastline during the deposition of the Muth Fm. Dominant SE directed paleocurrents with minor NW directed paleoflow are interpreted as coast parallel currents related to wave action. The NE directed dip of large scale crossbeds of facies association 2 represent offshore directed dune migration.

Conodont ages of the Lipak Fm. in the Pin Valley indicate an age range between Givetian to late Famennian/earliest Tournaisian for the Formation. This Givetan age of the basal Lipak Fm. near Muth additionally represents the younger age limit for the unfossiliferous Muth Fm. in this area. Tournaisian conodont ages for the whole Lipak Fm. in Zanskar thus indicate, that different levels of this Formation are preserved in both areas, a conclusion which is also supported by the different lithological successions of these regions.

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9. APPENDIX

Appendix A: Curriculum vitae

Appendix B: Stratigraphic sections

B1-B6:	**Section A-A´**	(parts of the Pin Fm. and Muth Fm.) Anticline SE of Mikkim, Pin Valley.
B7:	**Section B-B´**	(part of the Lipak Fm.) 1 km east of Mikkim, Pin Valley.
B8:	**Section C-C´**	(part of the Muth Fm.) E-limb of the anticline SE of Mikkim, Pin Valley.
B9:	**Section D-D´**	(upper part Muth Fm. and lower part Lipak Fm.) E-limb of the anticline SE of Mikkim, Pin Valley.
B10-B15:	**Section E-E´**	(top Pin Fm., complete Muth Fm., base of Lipak Fm.) E of Pandoshering, south of Muth, Pin Valley.
B16:	**Section F-F´**	(part of the Phe Fm.) South of Thango, Parahio Valley.
B17-B19:	**Section G-G´**	(top Muth Fm., complete Lipak Fm., base Gechang Fm.) Ravine NW of Muth, Pin Valley.
B20:	**Section H-H´**	(complete Pin Fm.) SE of Muth, Pin Valley.

CURRICULUM VITAE

Personal data:

Place of birth: Austria; Nationality: Austria

Education:

Primary school in Lackendorf.

High school in Oberpullendorf (leaving exams broadly equivalent to A-levels).

Study of Geology and Prehistoric Archaeology at the University of Vienna.

Diploma thesis (advisor Prof. Martin Thöni): „Kristallingeologische Neubearbeitung des südlichen Ödenburger Gebirges, Burgenland (Österreich).“

Ph.D. thesis (advisors Prof. Wolfgang Frank and Prof. Richard Lein): „The Muth Formation in the Pin Valley (Spiti, N-India): Depositional Environment and Ichnofauna of a Lower Devonian Barrier Island System.“

Additional courses and professional experiences:

22.-26. April 1992: „Mikrostrukturen und Rheologie“, University of Vienna, Prof. Neil Mancktelow (ETH Zürich).

8.-12. Februar 1993: „Pyroklastische Gesteine“, GEOMAR in Kiel, Prof. Hans-Ulrich Schmincke.

Since 1986 abundant archaeological excavations in Austria and abroad, together with Burgenländisches Landesmuseum, Bundesdenkmalamt, University of Vienna and Deutsche Forschungsgemeinschaft.

September 1989 and 1990: holiday-job at Ilbau AG in Gdynia, Poland.

1992-1999: Demonstrator in courses of the Institute for Geologie: „Geologische Methodik“, „Geologische Kartenkunde“ and „Geologische Kartierung im Gelände“.

22.-26. September 1993: Hydrogeological fieldwork at DKW: hydroelectric power plant Vienna.

August 1993: Geological mapping for the Austrian Geological Survey.

June 1995/June 1996: Excavation-manager in Project P 10 814-HIS (FWF): „Das spätantike Gräberfeld in Halbturn.“ Principal investigator Prof. Falko Daim.

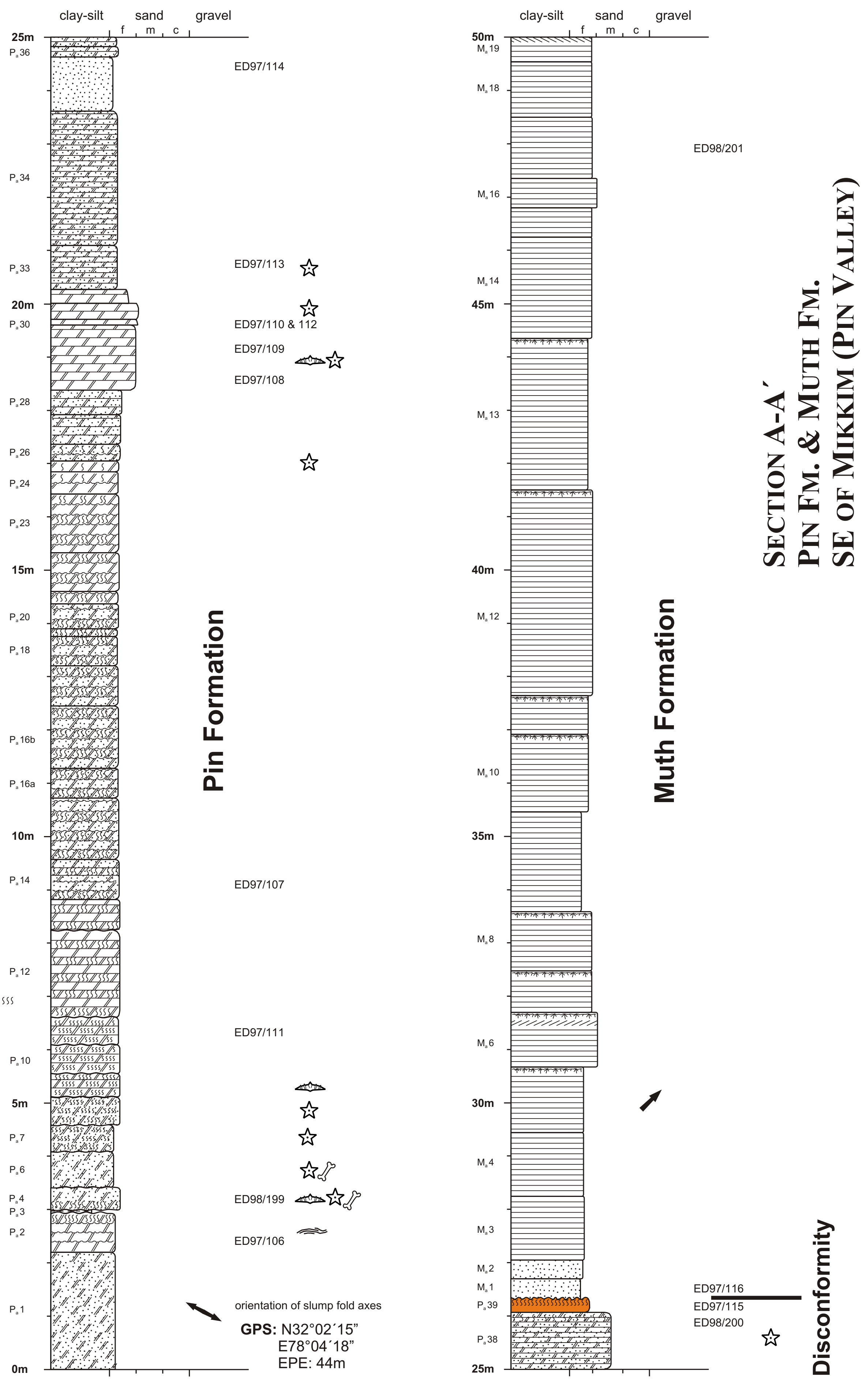
1996-1999: Research position in Project P11765-GEO (FWF: „Investigation of the Alpine deformation and cooling history of the NW-Himalayas and its bearing on accommodation mechanisms of India-Asia collision - a study from the Sutlej cross section.“ Principal investigator Dr. Bernhard Grasemann.

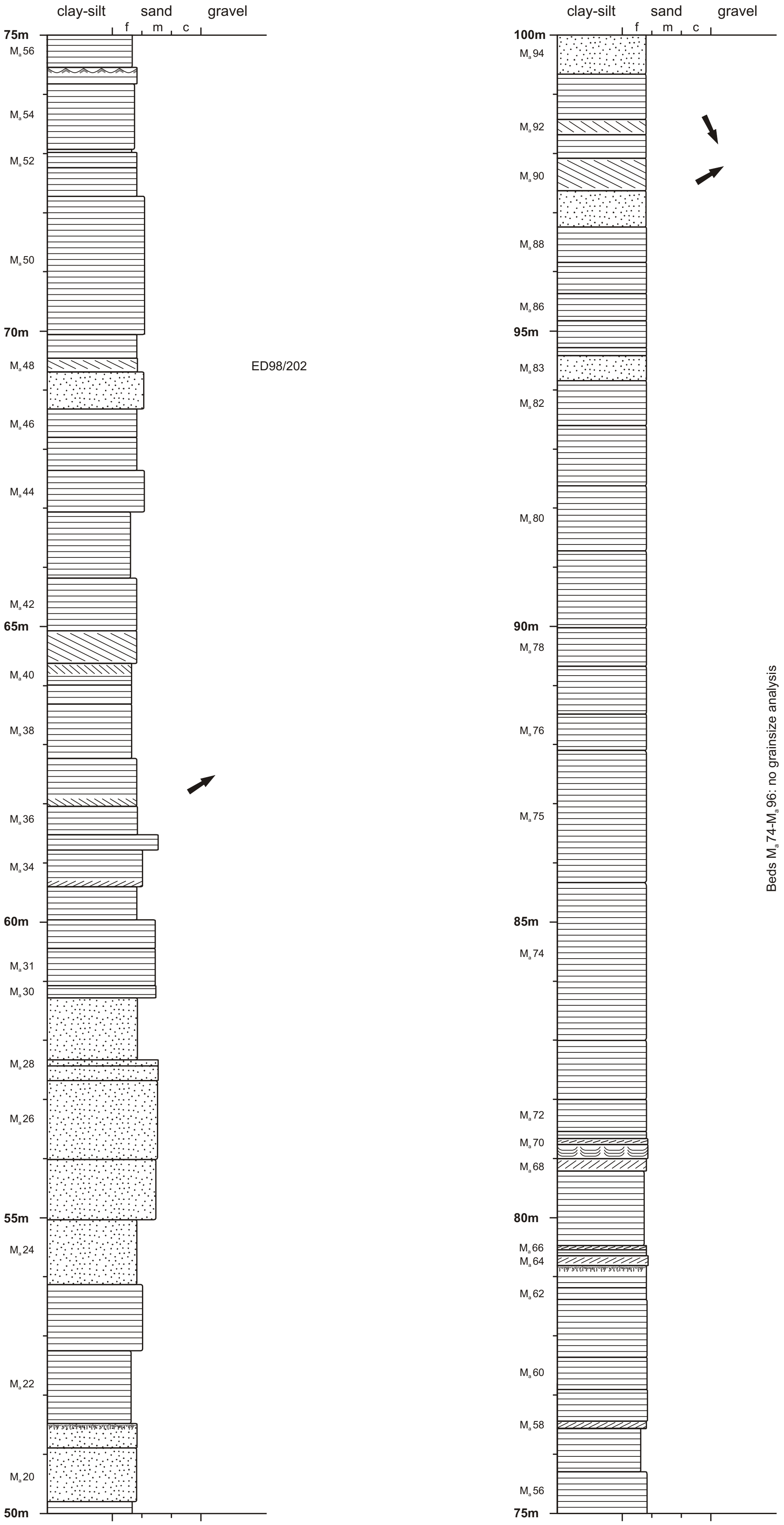
Attended field trips:

Several field trips to the Northern and Central Alps (Austria); S-Turkey (stratigraphy, sedimentology); Canada (mining geology); Cyprus (ophiolites); Crete (regional geology); S-Italy (volcanology); N-India (tectonics, stratigraphy, metamorphism); Andalucia, Spain (tertiary carbonates); N-Norway (structural geology); Mallorca (sequence stratigraphy); Oman (stratigraphy, sedimentology, tectonics); N-Pakistan (regional geology); Cyclades, Greece (metamorphism, exhumation).

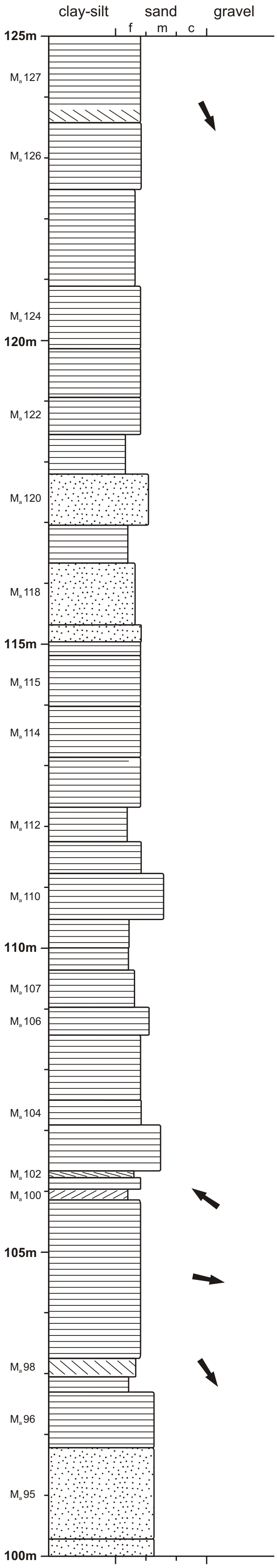
National Service:

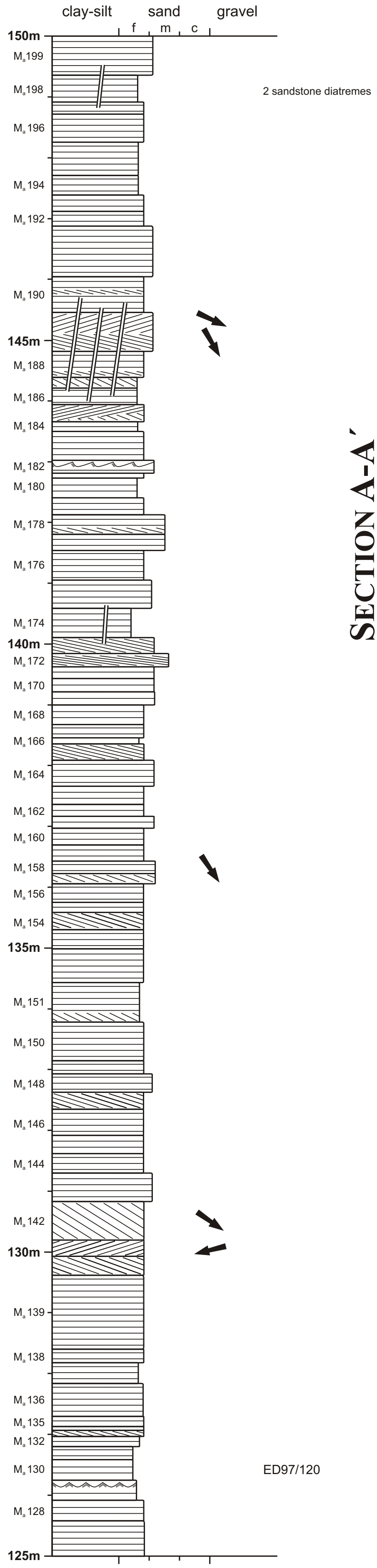
October 1994 - May 1995 (Guards of Honor, Military Geographical Institute).


Section A-A´
Pin Fm. & Muth Fm.
SE of Mikkim (Pin Valley)
clay-silt
sand
gravel
f
m
c
25m
20m
15m
10m
5m
0m
Pa36
Pa34
Pa33
Pa30
Pa28
Pa26
Pa24
Pa23
Pa20
Pa18
Pa16b
Pa16a
Pa14
Pa12
Pa10
Pa7
Pa6
Pa4
Pa3
Pa2
Pa1
ED97/114
ED97/113
ED97/110 & 112
ED97/109
ED97/108
ED97/107
ED97/111
ED98/199
ED97/106
Pin Formation
orientation of slump fold axes
GPS: N32°02´15”
E78°04´18”
EPE: 44m
50m
45m
40m
35m
30m
25m
Ma19
Ma18
Ma16
Ma14
Ma13
Ma12
Ma10
Ma8
Ma6
Ma4
Ma3
Ma2
Ma1
Pa39
Pa38
ED98/201
Muth Formation
ED97/116
ED97/115
ED98/200
Disconformity

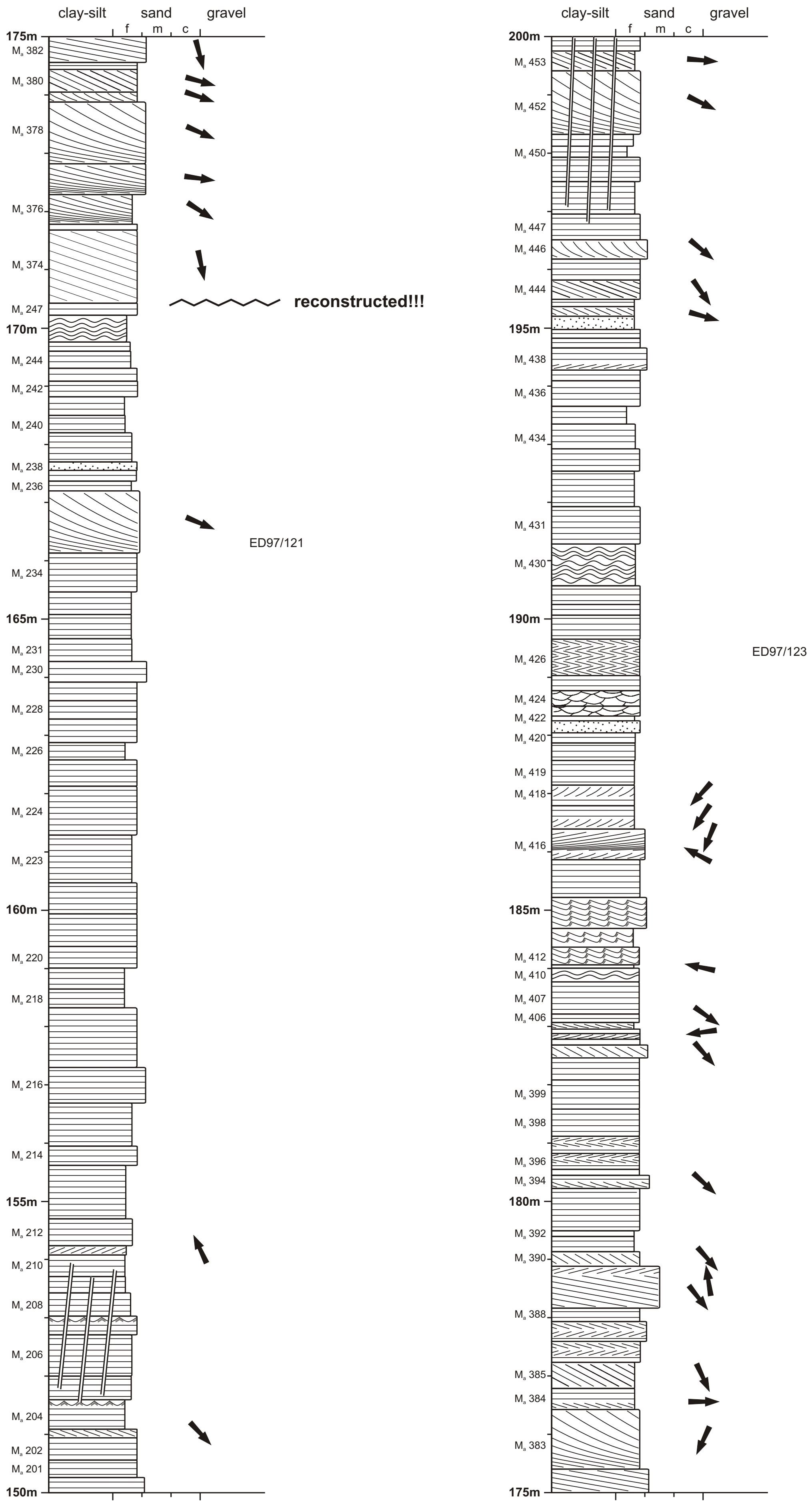


SECTION A-A´
MUTH FM.
SE OF MIKKIM (PIN VALLEY)

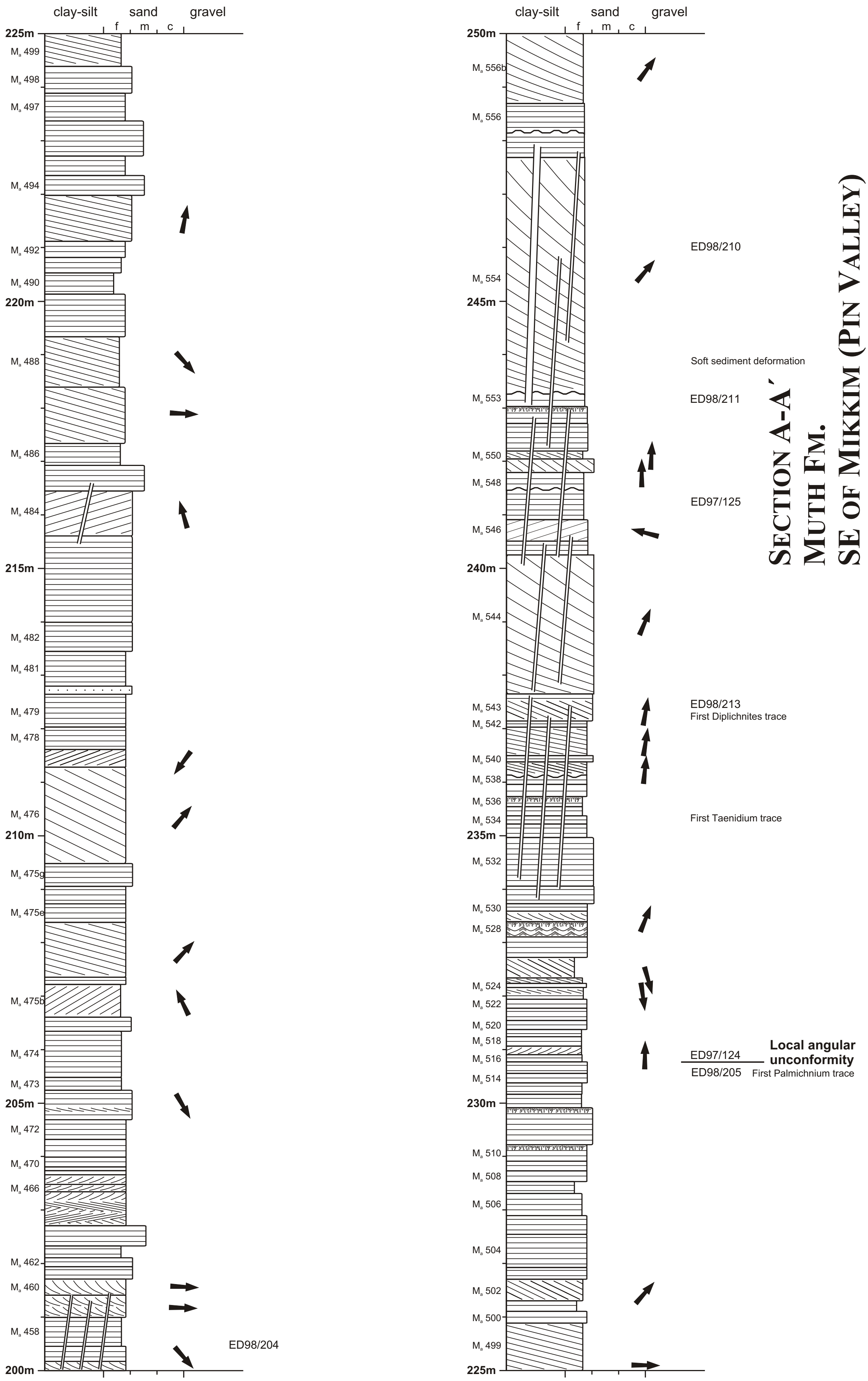




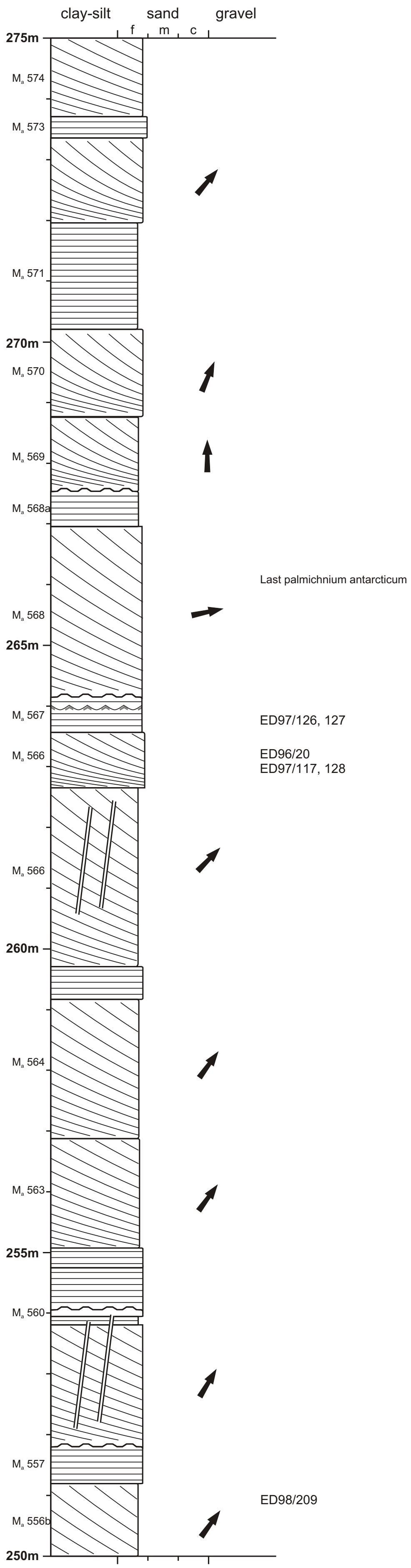
SECTION A-A´
MUTH FM.
SE OF MIKKIM (PIN VALLEY)

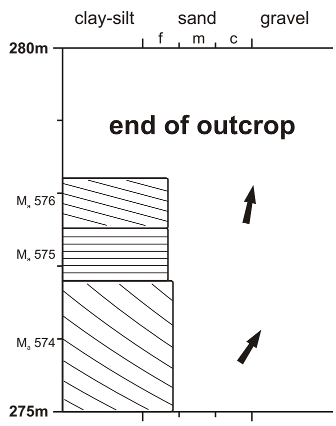


SECTION A-A´
MUTH FM.
SE OF MIKKIM (PIN VALLEY)

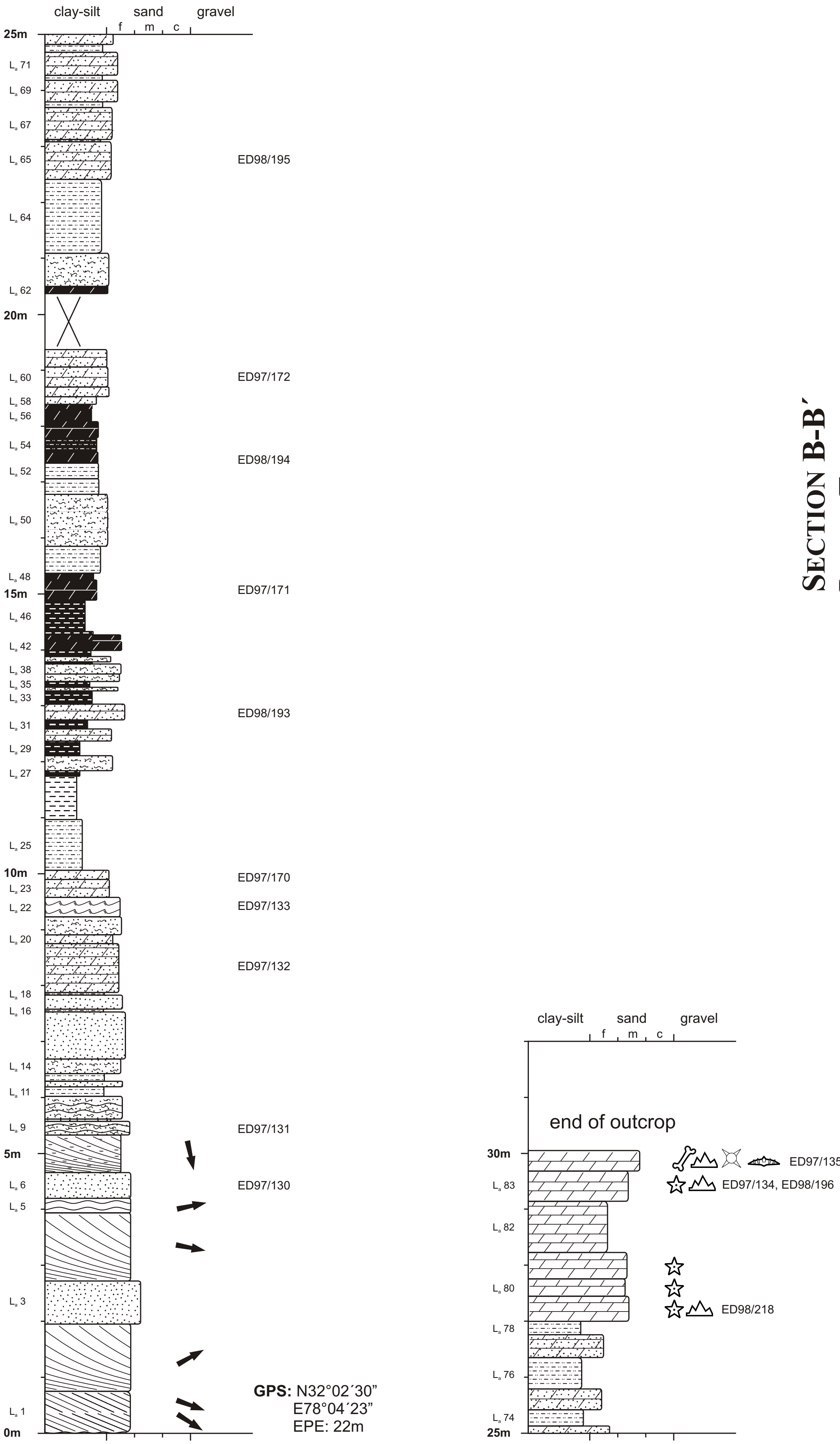

Section A-A´
Muth Fm.
SE of Mikkim (Pin Valley)
clay-silt
sand
gravel
f
m
c
225m
220m
215m
210m
205m
200m
250m
245m
240m
235m
230m
225m
Ma 499
Ma 498
Ma 497
Ma 494
Ma 492
Ma 490
Ma 488
Ma 486
Ma 484
Ma 482
Ma 481
Ma 479
Ma 478
Ma 476
Ma 475g
Ma 475e
Ma 475b
Ma 474
Ma 473
Ma 472
Ma 470
Ma 466
Ma 462
Ma 460
Ma 458
ED98/204
Ma 556b
Ma 556
Ma 554
Ma 553
Ma 550
Ma 548
Ma 546
Ma 544
Ma 543
Ma 542
Ma 540
Ma 538
Ma 536
Ma 534
Ma 532
Ma 530
Ma 528
Ma 524
Ma 522
Ma 520
Ma 518
Ma 516
Ma 514
Ma 510
Ma 508
Ma 506
Ma 504
Ma 502
Ma 500
Ma 499
ED98/210
Soft sediment deformation
ED98/211
ED97/125
ED98/213
First Diplichnites trace
First Taenidium trace
ED97/124
ED98/205
Local angular unconformity
First Palmichnium trace

Section A-A´
Muth Fm.
SE of Mikkim (Pin Valley)

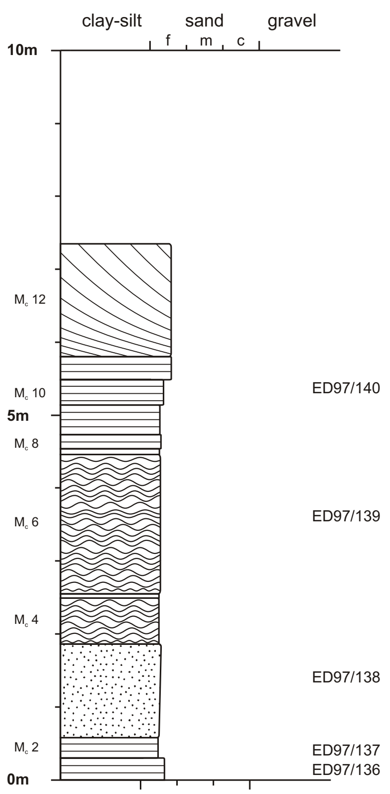




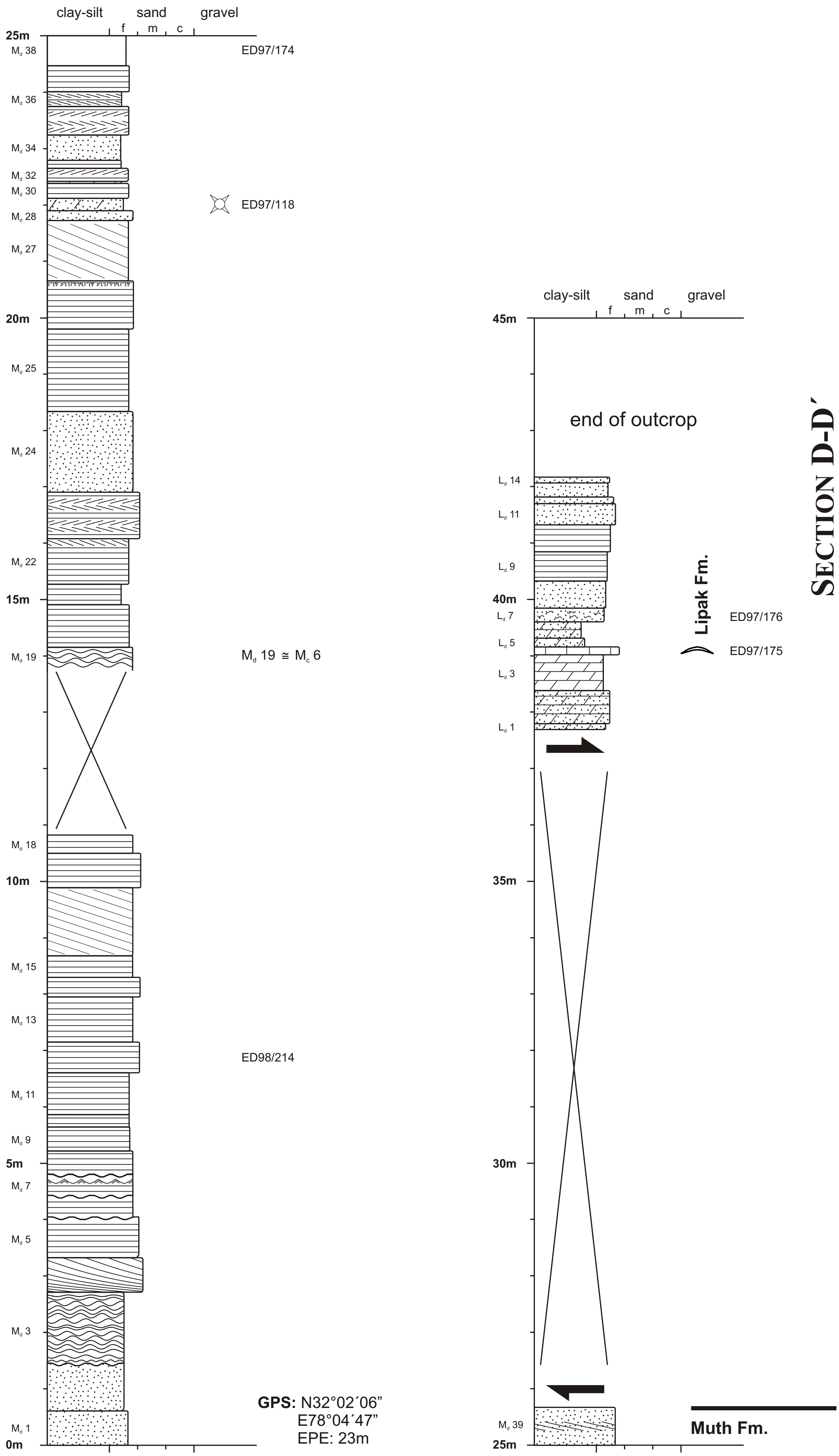
Section B-B´
Lipak Fm.
1 km E of Mikkim (Pin Valley)



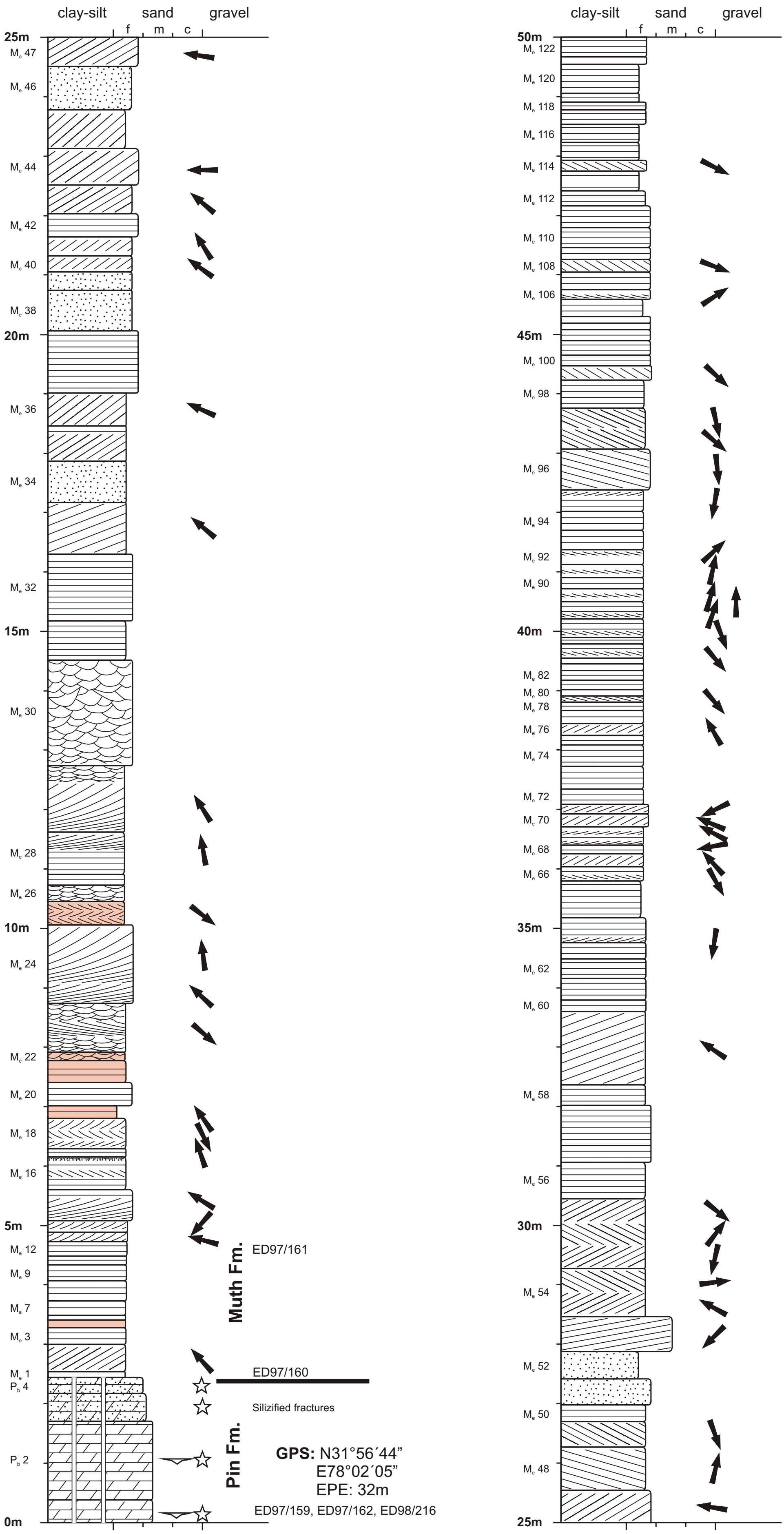
SECTION C-C´
MUTH FM.
SE OF MIKKIM (PIN VALLEY)



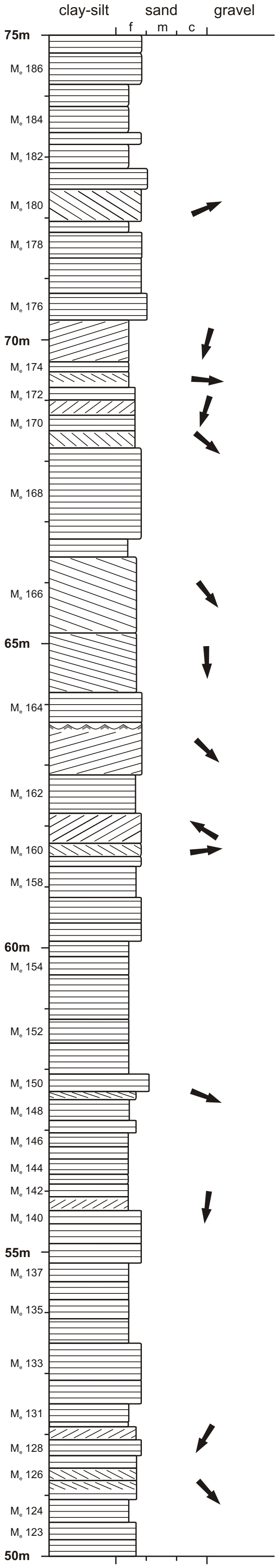
GPS: N32°02´01”
E78°04´38”
EPE: 41m

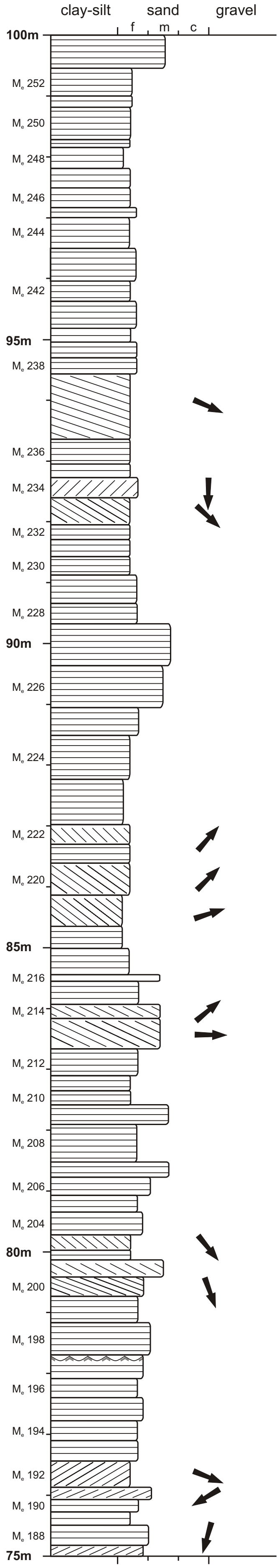


Section D-D´
Muth Fm. and Lipak Fm.
SE of Mikkim (Pin Valley)



SECTION E-E´
PIN FM. AND MUTH FM.
1.3 KM S OF MUTH (PIN VALLEY)





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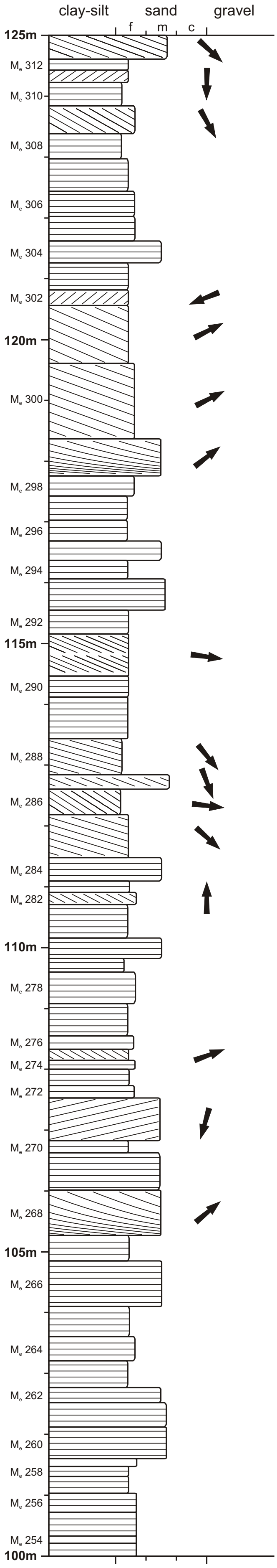
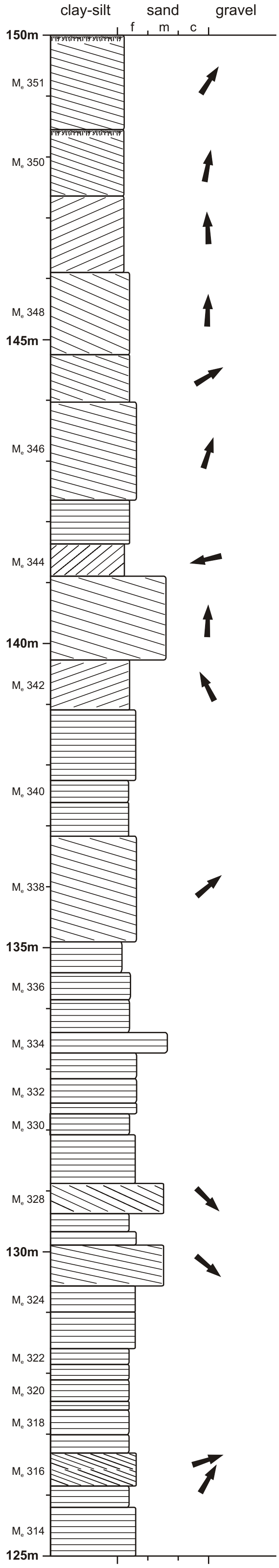
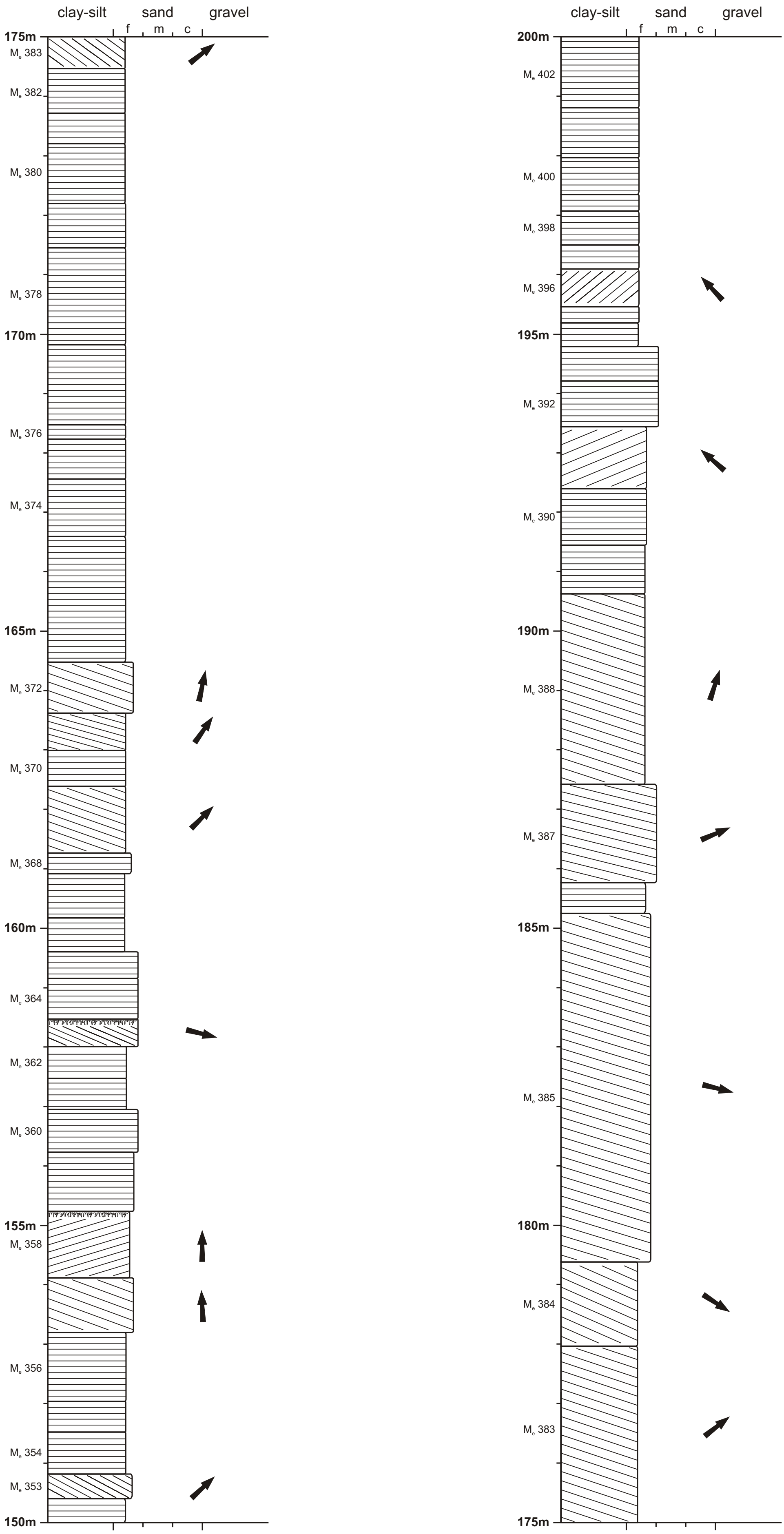
MUTH FM.
1.3 KM S OF MUTH (PIN VALLEY)

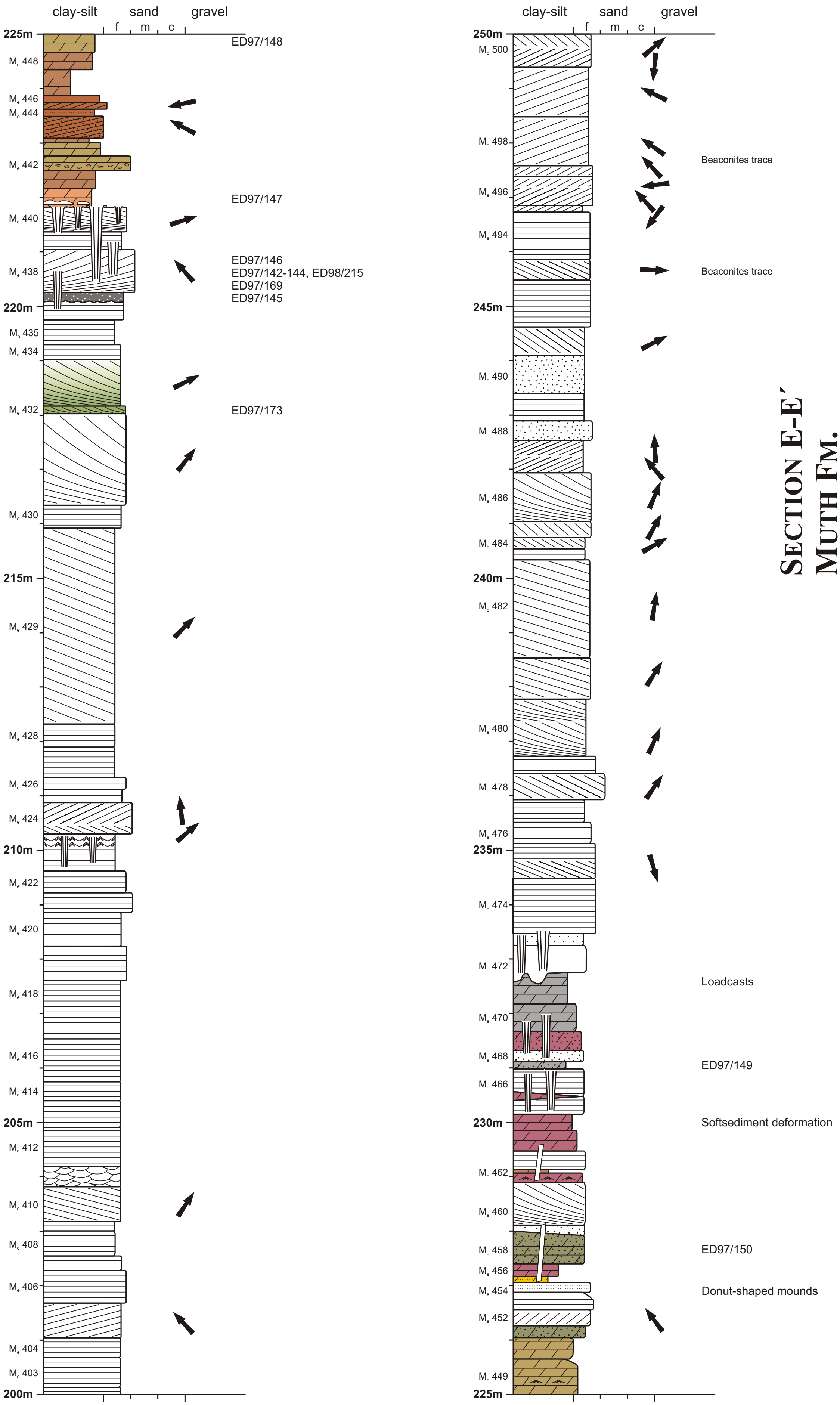




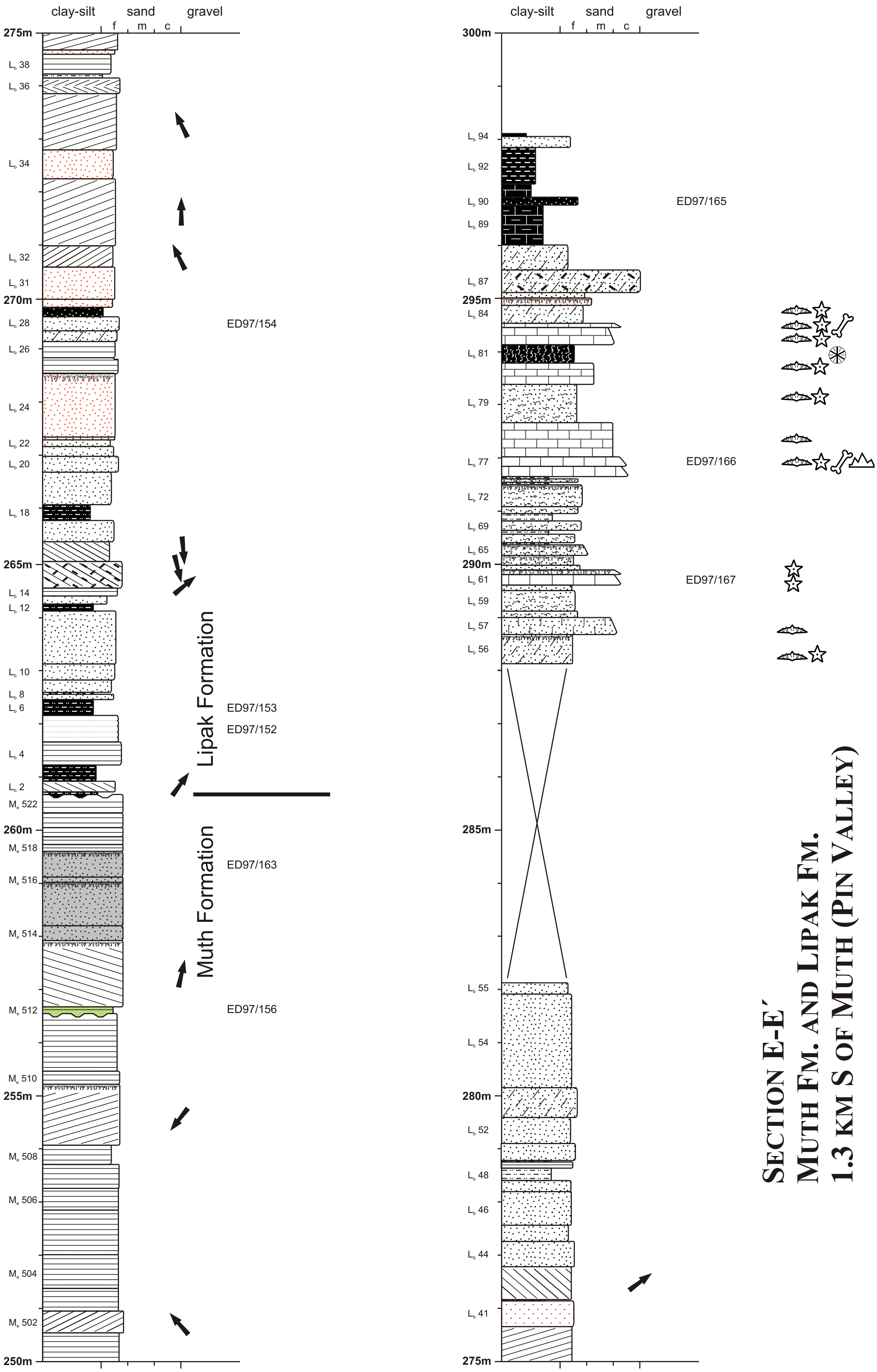
SECTION E-E´
MUTH FM.
1.3 KM S OF MUTH (PIN VALLEY)



SECTION E-E´
MUTH FM.
1.3 KM S OF MUTH (PIN VALLEY)

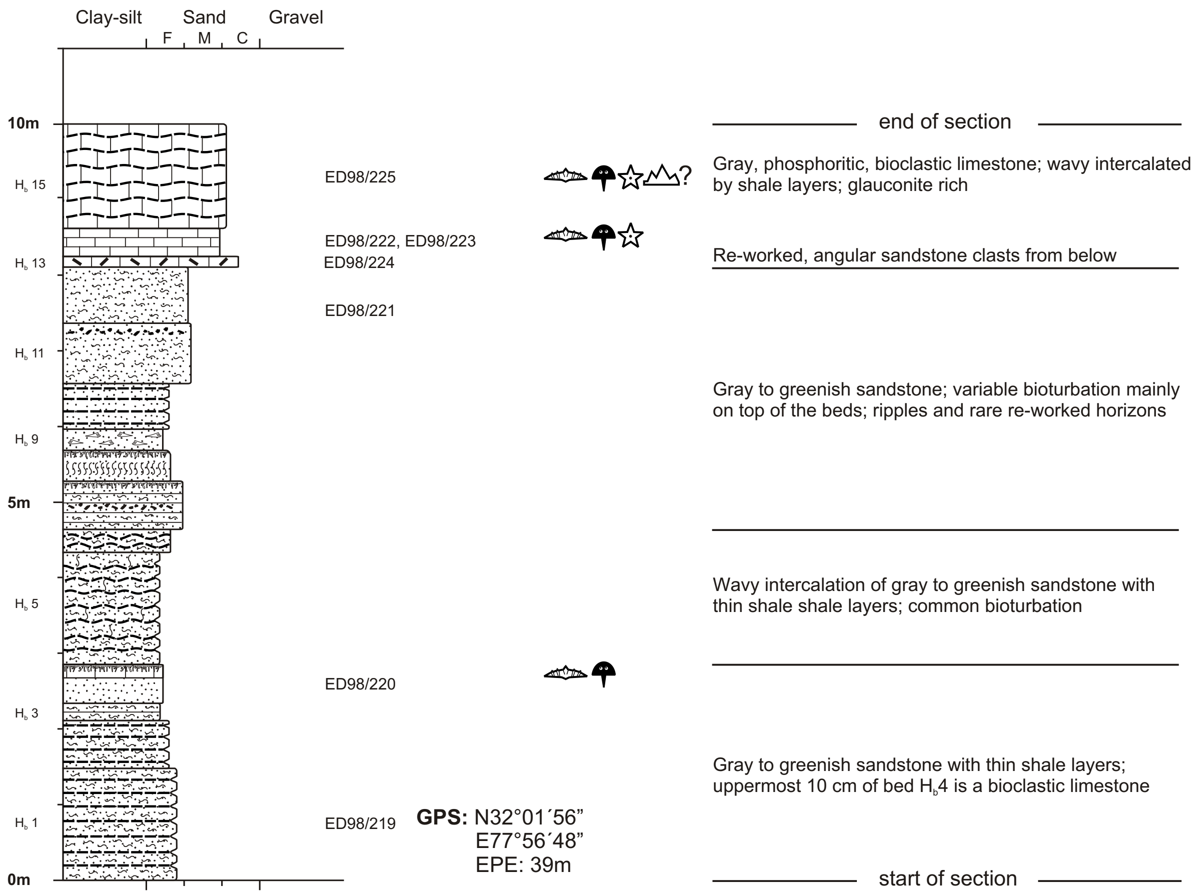


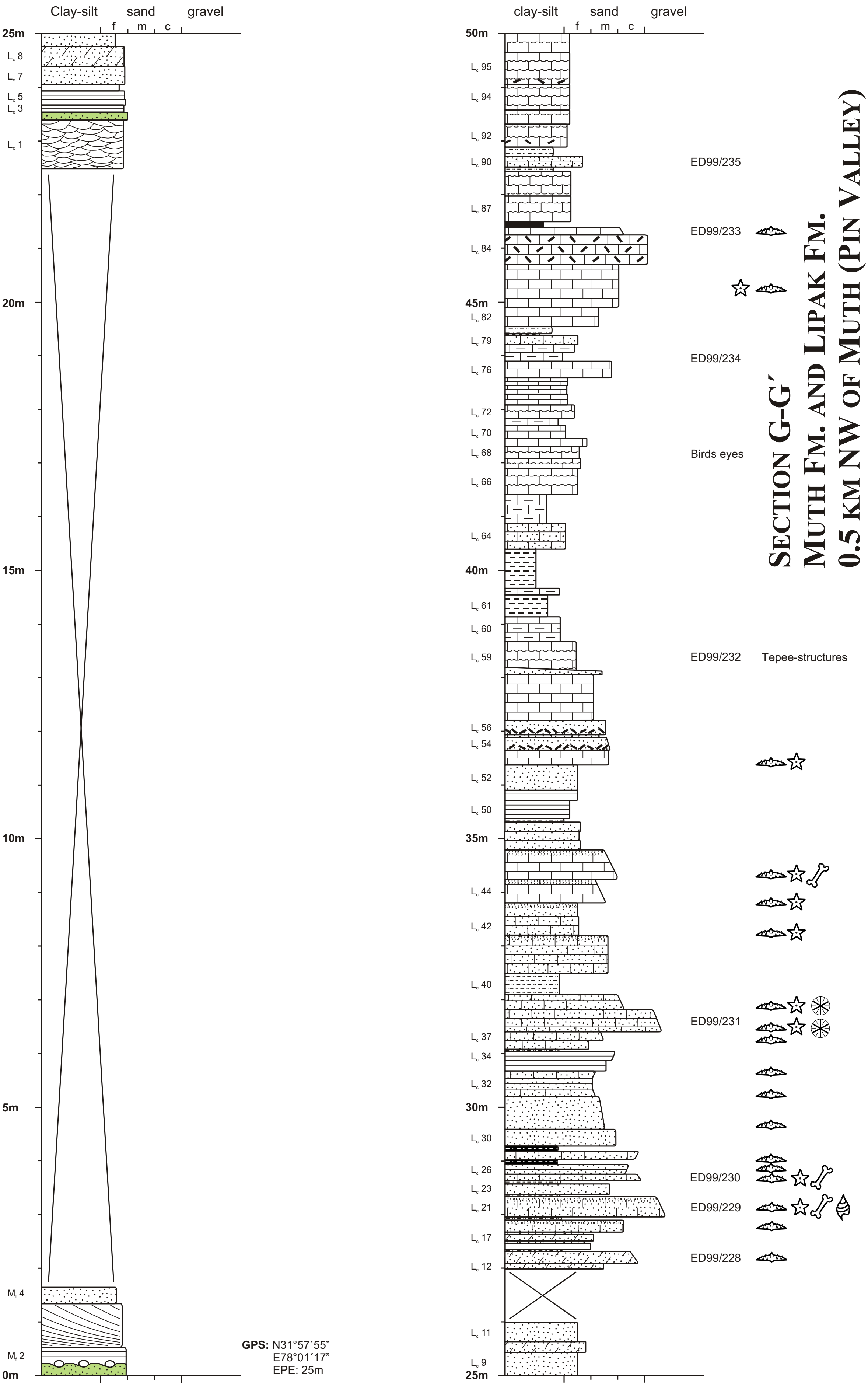
Section E-E´
Muth Fm.
1.3 km S of Muth (Pin Valley)



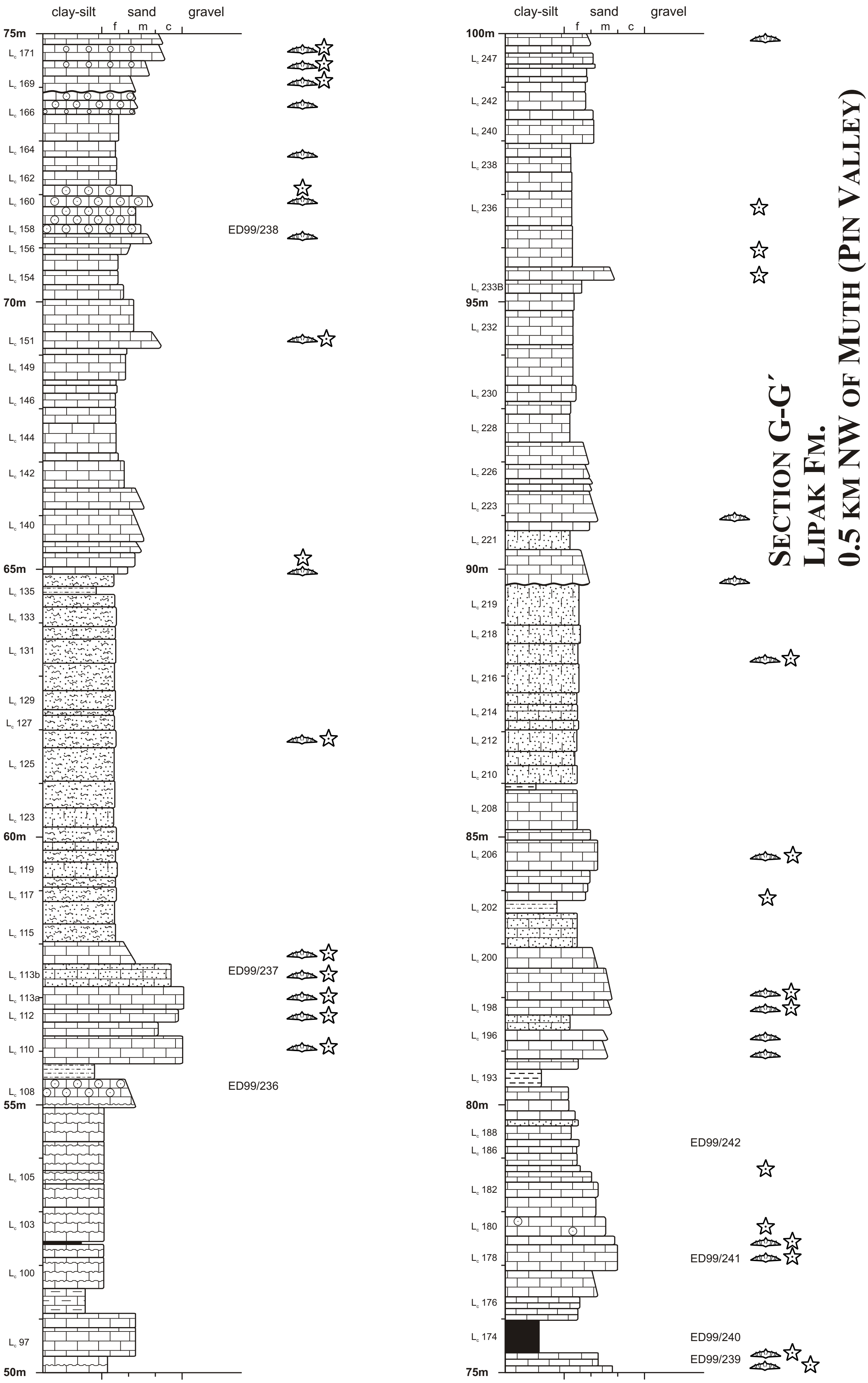
SECTION E-E′
MUTH FM. AND LIPAK FM.
1.3 KM S OF MUTH (PIN VALLEY)

SECTION F-F´
PHE FM. SOUTH OF THANGO (PARAHIO VALLEY)





Clay-silt sand gravel
f m c
25m
L_c 8
L_c 7
L_c 5
L_c 3
L_c 1
20m
15m
10m
5m
M_f 4
M_f 2
0m
GPS: N31°57´55”
E78°01´17”
EPE: 25m
clay-silt sand gravel
f m c
50m
L_c 95
L_c 94
L_c 92
L_c 90
ED99/235
L_c 87
ED99/233
L_c 84
45m
L_c 82
L_c 79
ED99/234
L_c 76
L_c 72
L_c 70
L_c 68
Birds eyes
L_c 66
L_c 64
40m
L_c 61
L_c 60
L_c 59
ED99/232
Tepee-structures
L_c 56
L_c 54
L_c 52
L_c 50
35m
L_c 44
L_c 42
L_c 40
ED99/231
L_c 37
L_c 34
L_c 32
30m
L_c 30
L_c 26
ED99/230
L_c 23
L_c 21
ED99/229
L_c 17
ED99/228
L_c 12
L_c 11
L_c 9
25m
SECTION G-G´
MUTH FM. AND LIPAK FM.
0.5 KM NW OF MUTH (PIN VALLEY)


Section G-G´
Lipak Fm.
0.5 km NW of Muth (Pin Valley)
clay-silt
sand
gravel
f
m
c
50m
55m
60m
65m
70m
75m
80m
85m
90m
95m
100m
Lc 97
Lc 100
Lc 103
Lc 105
Lc 108
Lc 110
Lc 112
Lc 113a
Lc 113b
Lc 115
Lc 117
Lc 119
Lc 123
Lc 125
Lc 127
Lc 129
Lc 131
Lc 133
Lc 135
Lc 140
Lc 142
Lc 144
Lc 146
Lc 149
Lc 151
Lc 154
Lc 156
Lc 158
Lc 160
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Lc 164
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Lc 196
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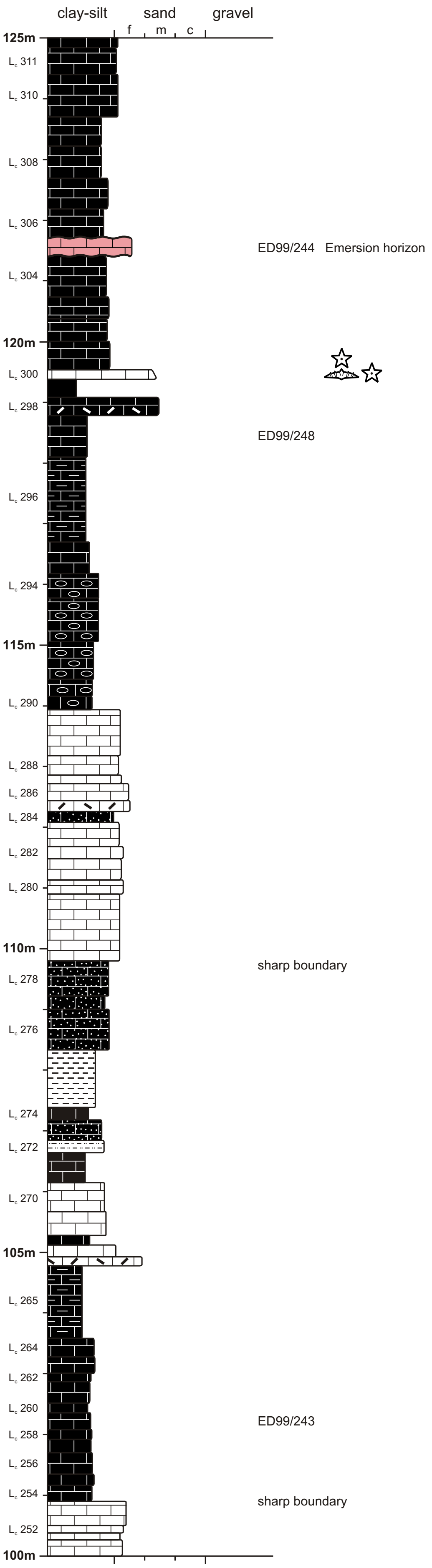
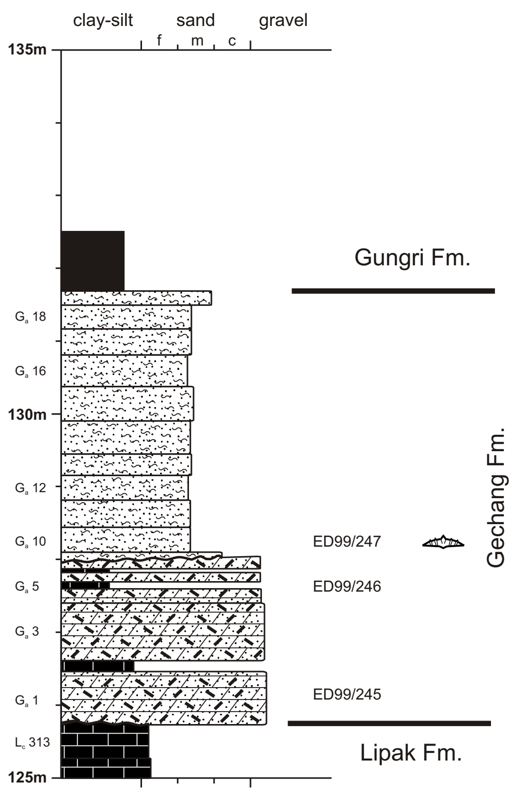
Lc 208
Lc 210
Lc 212
Lc 214
Lc 216
Lc 218
Lc 219
Lc 221
Lc 223
Lc 226
Lc 228
Lc 230
Lc 232
Lc 233B
Lc 236
Lc 238
Lc 240
Lc 242
Lc 247
ED99/236
ED99/237
ED99/238
ED99/239
ED99/240
ED99/241
ED99/242

Section G-G′
Lipak Fm. and Gechang Fm.
0.5 km NW of Muth (Pin Valley)





Stratigraphic section H-H´ of the Pin Formation 1.6 km S Muth, Pin Valley (Spiti)