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Conjugate, cataclastic deformation bands in the Lower Devonian Muth Formation (Tethyan Zone, NW India): evidence for pre-Himalayan deformation structures

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Abstract – The purpose of this study is to use the mechanisms of deformation band formation to help with interpreting the timing of phases of deformation in an area with a complex geological history. Deformation bands and zones of deformation bands are described from the quartzites of the Lower Devonian Muth Formation in the Pin Valley, NW Himalayas. Thin-section analyses show that the deformation bands in the Muth Formation formed early in the diagenetic history before porosity was lost. Deformation mechanisms involved cataclasis, translation, rotation of quartz grains and effective porosity reduction. The orientations of the deformation bands cannot be reasonably grouped with the orientations of faults related to Himalayan deformation in the Pin Valley. Additionally, the deformation bands are deformed by Eo-Himalayan (Eocene) folds, which in turn are cut by later faults. The later faults that cross-cut the Eo-Himalayan folds developed in the already-cemented Muth Formation at much higher temperature and pressure conditions by crystal plastic deformation mechanisms, indicated by quartz crystals with undulatory extinction, abundant kink bands, dislocation glide, elongated subgrains, slightly curved deformation lamellae and pronounced shape-preferred orientation. These two completely contrasting deformation mechanisms on the microstructural scale characterize two distinct fault sets that formed at different depths in the crust. Based on these differences, a pre-Himalayan origin of the deformation bands is concluded, thus representing a set of rare pre-Himalayan deformation structures. After unfolding to remove Eo-Himalayan crustal shortening, the orientation of the deformation bands and restored relative offsets of sedimentary bedding are most compatible with ∼E–W-oriented shortening associated with N–S extension. The age of the deformation bands in the Muth Formation is bracketed by an early Devonian sedimentation age of the Muth Formation and a middle Cretaceous age of considerable cementation as deduced from compiled burial histories. Accepting a pre-middle Cretaceous age of the deformation bands, maximum conditions of about 80 °C and 60 MPa lithostatic pressure during their formation are estimated from the amount of overburden during the middle Cretaceous. We suggest the deformation bands are a result of either the Neo-Tethys rifting event beginning in the early Carboniferous or the extension related to late Carnian/early Norian rapid subsidence, although a hitherto unknown deformation event cannot be excluded.

Keywords: deformation bands, pre-Himalayan deformation, India, quartz arenite, Muth Formation.

1. Introduction

Tectonic evolution studies of the Himalayan mainly focus on the Tertiary deformation and kinematics resulting from the Indo-Asian collision (for reviews see Gansser, 1964; Fuchs, 1981; Hodges, 2000; Yin & Harrison, 2000 and references cited therein). Due to the magnitude and intensity of Himalayan tectonics in the rocks that comprise the Himalayan orogen, any pre-existing structures deformed by pre-Himalayan events are obscured or only partly preserved. As a result, pre-Himalayan deformation episodes are poorly documented and mainly inferred from lithostratigraphic anomalies such as tectonic unconformities (Griesbach, 1891; Stampfli, Marcoux & Baud, 1991; Garzanti, Angiolini & Sciunnach, 1996*b*) and variations in tectonic subsidence rates (Garzanti *et al*. 1995). Reports on pre-Himalayan structural field data are extremely rare (Vannay, 1993; Steck, Epard & Robyr, 1999; Wyss, Hermann & Steck, 1999; Wiesmayr & Grasemann, 2002). In recent years, increasing interest has been focused on possible pre-Himalayan deformation structures and their attendant influence on the Tertiary kinematic evolution of the Himalayas (e.g. Gehrels *et al*. 2003; Myrow *et al*. 2002). Such structures are the target of our study. We investigate deformation bands in quartzites of the Lower Devonian Muth Formation and sheared joints with the same orientation in the underlying dolomites of the Upper Ordovician–Lower Silurian Pin Formation (Pin Valley, NW Himalayas).

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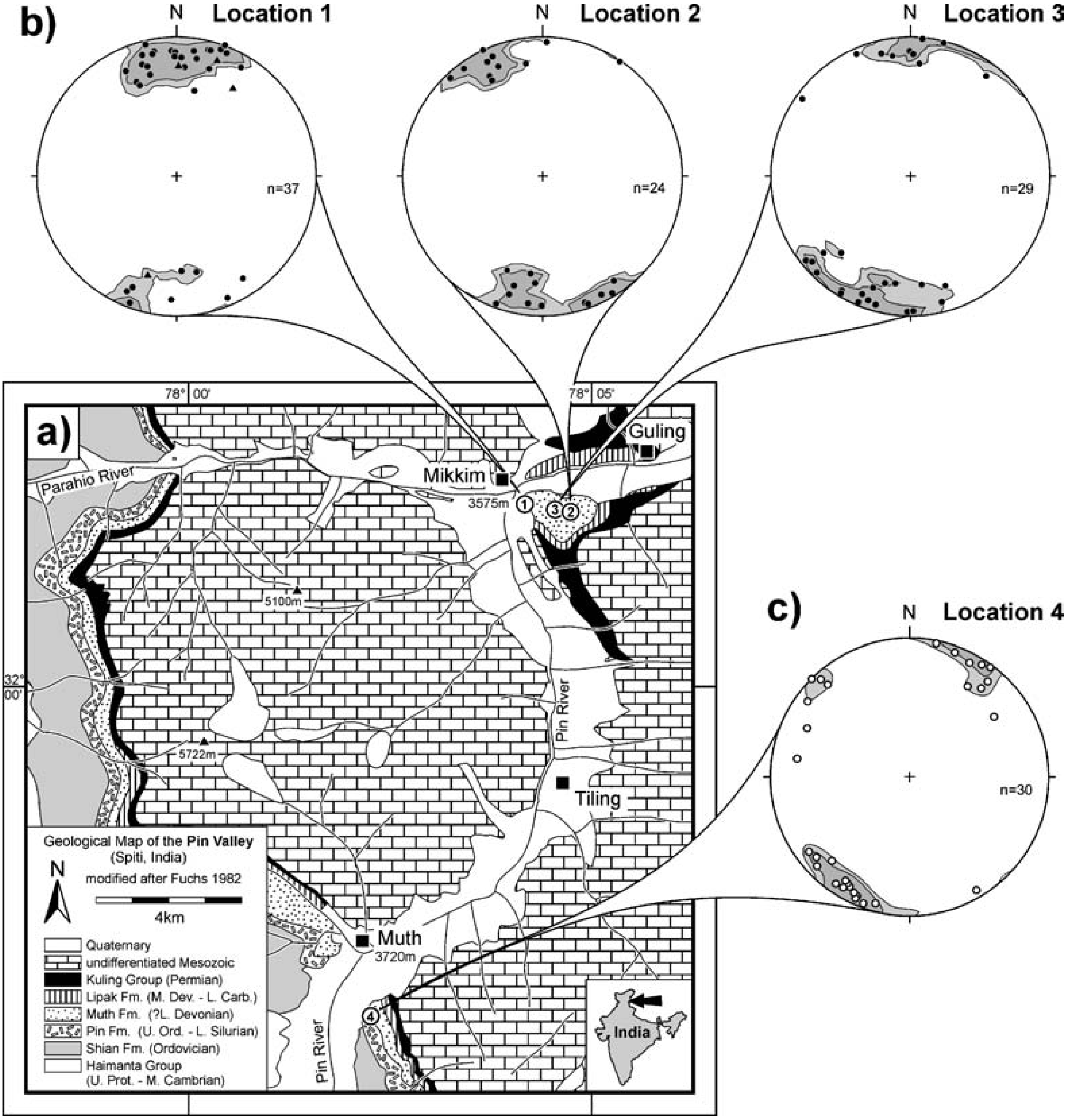

Figure 1. (a) Geological map of the Pin Valley modified after Fuchs (1982). (b) Stereo plots (equal area projections; lower hemisphere; contours at 3× and 5× the random distribution) of deformation bands (filled circles) and zones of deformation bands (filled triangles) in the Muth Formation, at locations 1 to 3, southeast of Mikkim. Deformation bands have been rotated to account for Eocene folding. (c) Stereo plots of sheared joints (open circles) in the uppermost part of the Pin Formation at location 4, south of Muth. Faults have been rotated to account for Eocene folding.

Deformation bands (cf. Aydin, 1978) and zones of deformation bands (cf. Aydin & Johnson, 1978) described in this study occur in the Lower Devonian Muth Formation, NW Himalayas (Fig. 1). The purpose of this paper is to constrain the orientations, kinematics and microstructural characteristics of these deformation bands. Based upon a clear separation of these structures from later faults in the same formation that have clearly different orientations as well as deformation mechanism, a pre-Himalayan age for the deformation bands is concluded. To understand the differentiation between these two types of faults, the development, appearance and characteristics of deformation bands are briefly summarized below.

2. Characteristics of deformation bands

Generally, brittle faults are loosely described as 'discontinuous deformation', whereas continuous deformation is frequently termed 'ductile'. Because this definition is clearly scale-dependent, we follow the suggestion of Rutter (1986) and Schmid & Handy (1991),

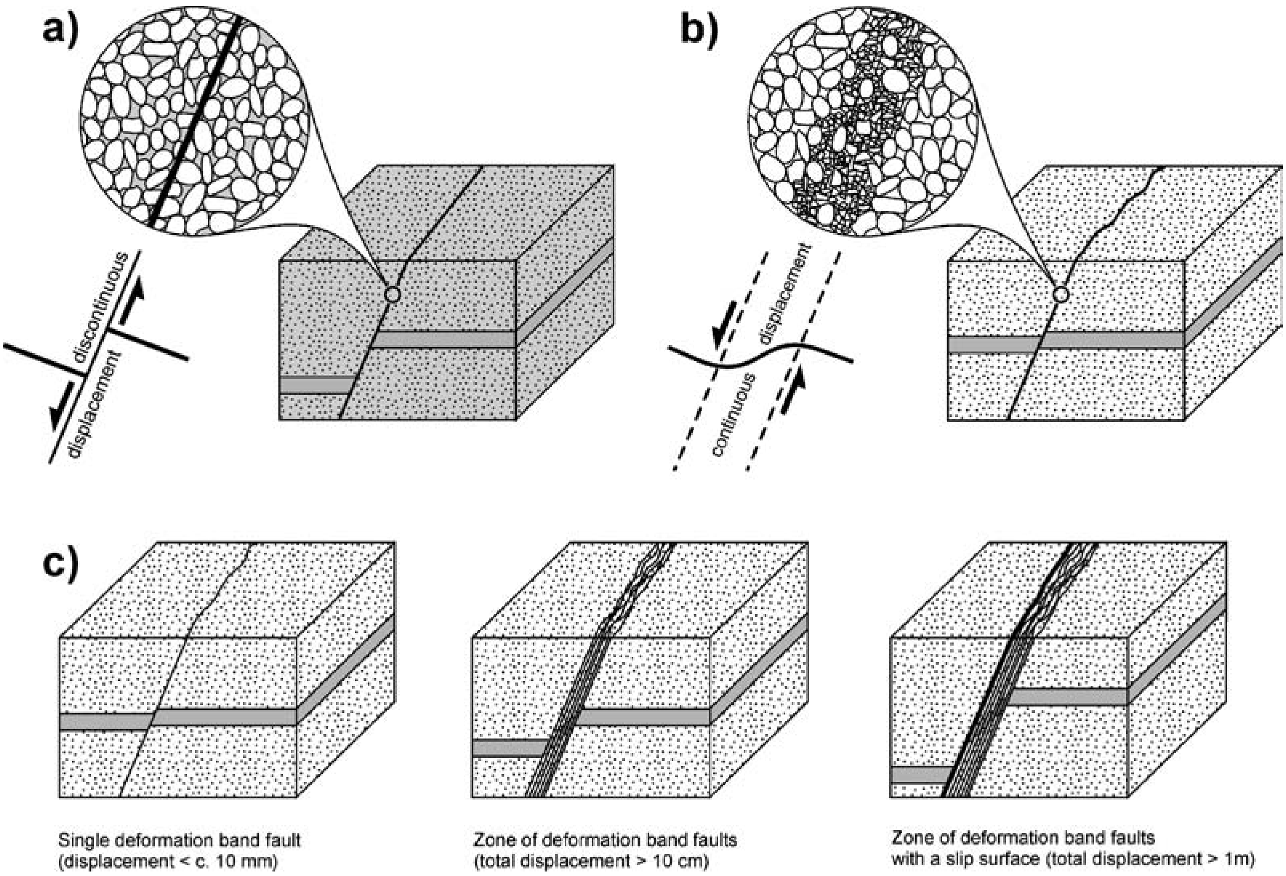

Figure 2. Theoretical diagram demonstrating the difference between 'ordinary' faults and deformation bands. (a) Faults usually form discrete slip-surfaces in fully cemented rocks. Once a slip-surface is formed, subsequent strain is focused on this surface because of the strain localization mechanisms of this kind of deformation. (b) Deformation bands are typical for deformation in porous or even completely uncemented granular material, where strain is accommodated by the formation of slightly undulating deformation surfaces. Porosity reduction, grain rotation and grain fracturing result in an overall strain hardening type of deformation. (c) In the model of Aydin & Johnson (1978) the strain hardening mechanisms result in a sequential growth of deformation bands with deformation widths of few millimetres into zones of deformation bands with deformation widths of up to several tens of centimetres and finally into zones of deformation bands with a slip-surface on either side with offset of up to several tens of metres (modified after Aydin & Johnson, 1978, fig. 7). See Shipton & Cowie (2001, fig. 16) for a modified model.

using the term 'frictional faults' for deformation mechanisms involving fracturing and volume change, which are strongly dependent on effective pressure. In contrast, 'viscous flow' encompasses non-frictional, thermally activated deformation mechanisms (see also Mandl, 2000).

Deformation bands are one kind of frictional deformation structure in the Earth's uppermost crust. Deformation bands, which were first described by Aydin (1978), are commonly found in sand and porous sandstone and are defined as tabular structures of finite width resulting from strain localization. Three end-member cases of these structures are distinguished: (1) deformation bands (Aydin, 1978) with clear shear offset, which have been termed 'deformation band faults' by Mollema & Antonellini (1996), (2) compaction bands that refer to tabular bands of localized porosity reduction that lack shear offset, termed 'compaction bands' by Mollema & Antonellini (1996) and (3) tabular bands of localized increase in porosity that lack a macroscopic shear offset, termed 'dilatation bands' by Du Bernard, Eichhübl & Aydin (2002). According to the degree of grain fragmentation and clay content, Antonellini, Aydin & Pollard (1994) further divided deformation bands into three groups: (a) with little or no cataclasis, (b) with cataclasis and (c) with clay smearing. In the following we use 'deformation bands' to describe deformation bands with shear offset.

Deformation bands (for recent reviews, see Mair, Main & Elphick, 2000; Main *et al*. 2001) are typically about 1 mm wide, roughly planar deformation structures that show shear deformation in the range of millimetres to a few centimetres. The slip- to fault-length ratios are low compared to ordinary faults with slip-surfaces in dense lithologies (Fossen & Hesthammer, 1997; Wibberley, Petit & Rives, 2000). Single cataclastic deformation bands accommodate deformation across the entire band width by collapse or increase of porosity, grain fracturing, grain size reduction and cataclastic flow that lacks a discrete discontinuity surface (Fig. 2) (Aydin, 1978; Aydin & Johnson, 1978; Antonellini, Aydin & Pollard, 1994).

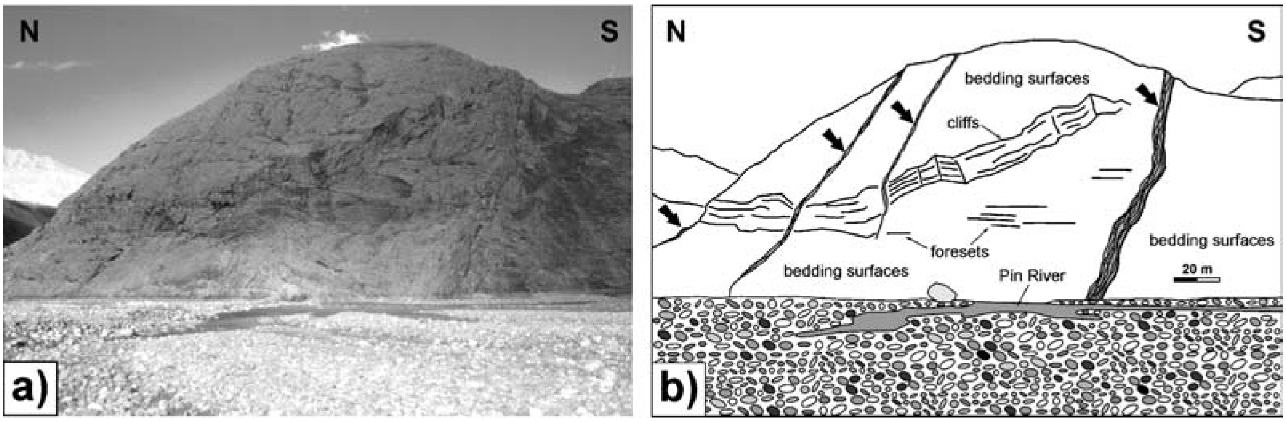

Figure 3. Photo (a) and line drawing (b) of the outcrop at the western termination of the Muth Formation anticline near Mikkim (location 1 in Fig. 1); view to the east across the fluvial gravel of the Parahio River. The outcrop is some 140 m high. Bedding surfaces are dipping at about 45° towards the southwest and are cut by four zones of nearly perpendicular deformation bands (indicated by arrows). Prominent zone of deformation bands in the right part of the picture is up to 8 m thick.

Under low stresses, grain fracturing may be absent and deformation is accommodated also by grain rotation, grain sliding and porosity reduction. These deformation mechanisms are typical for frictional faults in high-porosity rocks (e.g. Wong, David & Zhu, 1997). Deformation bands are thus structures that represent continuous modes of deformation on a macroscopic scale, as opposed to joints or slip-surfaces that are discontinuous modes of deformation. In other words, deformation bands are solely the result of a displacement gradient (over a narrow tabular zone), in contrast to faults with slip-surface which must include a displacement discontinuity (Fig. 2a, b). 'Ordinary' frictional faults (e.g. slickensides) with slip-surface and displacement discontinuities that usually develop in low-porosity rock will be termed 'faults' in the text below.

Generally, deformation bands may occur alone, but usually they group in multiple, sub-parallel, closely spaced zones (e.g. Figs 2c, 3, 4d, 5). The formation of zones of deformation bands, each of which have limited slip, is explained by Aydin & Johnson (1978). They suggest that strain hardening due to cataclasis and porosity collapse requires the formation of new bands in order to accommodate bulk strain (Fig. 2c). The zones of deformation bands are thought to grow by addition of new deformation bands, side by side, thus the thickness of the zone depends on the number and spacing of individual deformation bands (Mair, Main & Elphick, 2000) (Fig. 2c). Usually, single deformation bands accommodate just a few millimetres of deformation and the finite deformation of a zone of deformation bands is the sum of the offsets on each individual band and possibly also related slip-surfaces (Aydin, 1978; Shipton & Cowie, 2001). The strain hardening is due to increased grain friction in the bands during grain breaking and porosity reduction processes (e.g. Aydin, 1978; Aydin & Johnson, 1983; Wibberley, Petit & Rives, 2000). In the model of Aydin & Johnson (1978), the strain hardening mechanisms result in a sequential growth of deformation bands with deformation widths of a few millimetres (Fig. 4e) into zones of deformation bands with deformation widths of up to several tens of centimetres (Fig. 4a–d) and finally into zones of deformation bands with a slip-surface on either side with offsets of up to several tens of metres (Fig. 2c). Shipton & Cowie (2001) expand this classical model and mention that small slip-surfaces may already nucleate at a relatively early stage in the evolution of a zone of deformation bands. With increasing strain, the slip-surfaces propagate and link within a growing zone of deformation bands (Shipton & Cowie, 2001, fig. 16). In the Shipton & Cowie (2003) model, both types of faults grow contemporaneously and interdependently with each other, controlled by the transition from strain hardening to strain softening and strain localization.

In recent years, the investigation of deformation bands has changed from mainly academic curiosity about their formation, to their tectonic implications, as well as economic interests regarding their spatial orientations and petrographic properties influencing hydrocarbon migration and reservoir compartmentalization (Aydin, 2000; Hesthammer, Johansen & Watts, 2000; Fisher & Knipe, 2001; Ogilvie & Glover, 2001). In this context, the recognition of deformation bands (usually with low permeability) and their separation from faults (commonly pathways for fluids) is a basic requirement in oil and groundwater exploration.

3. Geological setting

Our area of study is NW India, investigating the Palaeozoic Pin and Muth formations. The Muth Formation is part of the Tethyan Zone of the Higher Himalaya tectonic unit, which can be traced all along the strike of the Himalaya from Pakistan to Bhutan.

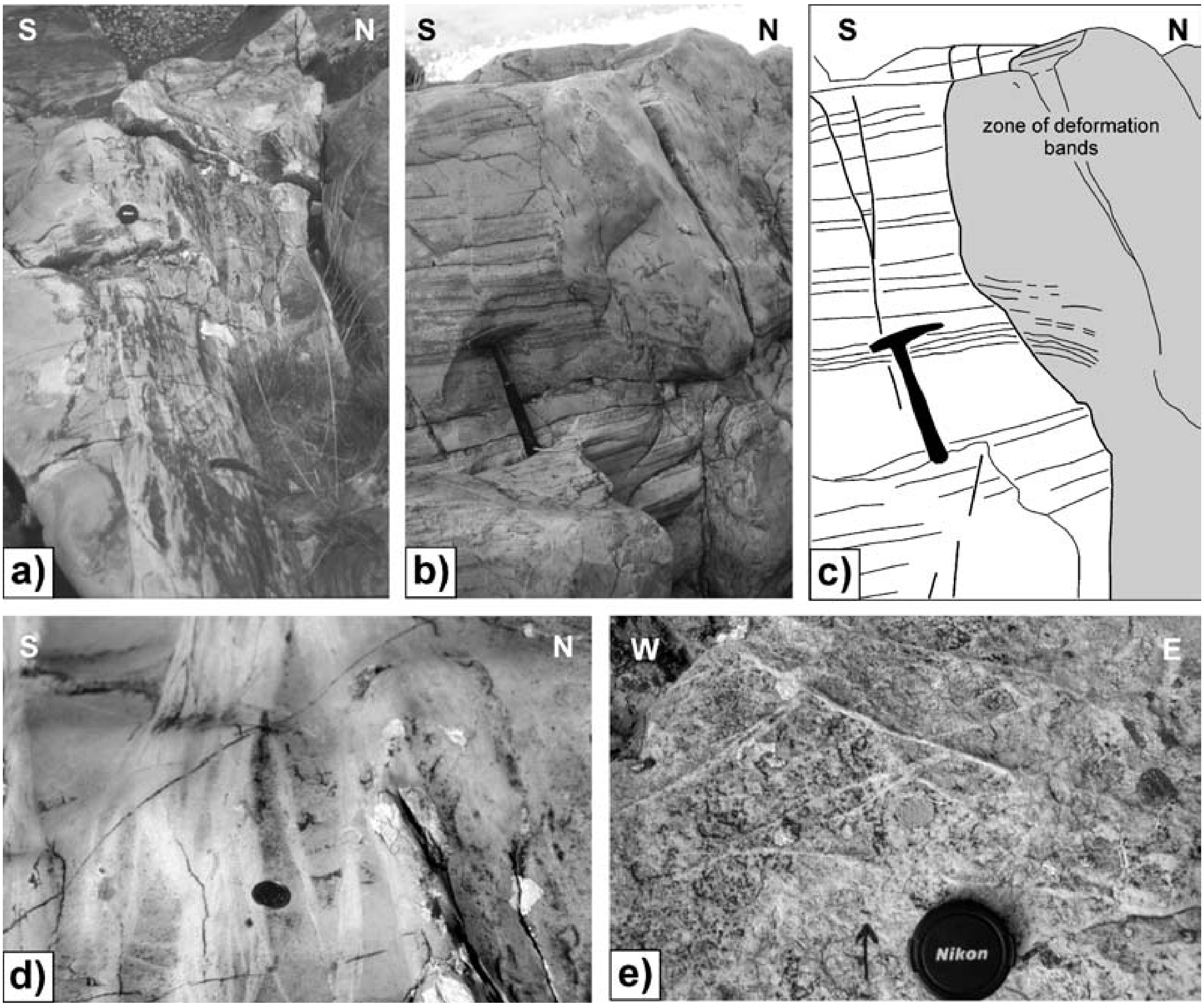

Figure 4. (a) Oblique plan view photo of a 70 cm thick zone of deformation bands in the Muth Formation comprising closely spaced parallel deformation bands. Between individual bands, elongated lenses of the primary sediment are preserved. Note the lighter colour of the bands compared with the surrounding quartzite. 55 mm lens cap for scale. (b) Photo and (c) linedrawing of a vertical section of the same zone of deformation bands. Bands are so closely spaced that they have effectively amalgamated together. Note the small normal offset of the primary bedding on individual bands to the right (north). Note single deformation bands slightly displacing sedimentary bedding left of the zone. Grey part of line drawing indicates the zone of deformation bands with very high deformation band fault density. 33 cm hammer for scale. (d) Photo shows detail of outcrop in (a). Note the lighter colour of the bands compared with the undeformed surrounding quartzite and lenses of primary sediment between individual bands. Coin diameter 25 mm. (e) Plan view photo of conjugate deformation bands at location 2; north to the top. Deformation bands form lighter coloured ridges with positive relief on the rock surface. 55 mm lens cap for scale.

The Tethyan Zone comprises more than 10 km of Neoproterozoic to Eocene sediments, deposited upon the former northern Indian passive margin (Gansser, 1964; Bhargava & Bassi, 1998). To the north, the zone is bordered by the mélange of the Indus-Yarlung suture zone, which represents the surface expression of the boundary between Indian and Asian continental crust (Gansser, 1964). In the generally accepted view, the southern limit of the Tethyan Zone is formed by the normal faults of the Southern Tibetan Detachment System and equivalents (Burg *et al*. 1984; Burchfield *et al*. 1992).

The stratigraphic context of the Muth Formation has been discussed recently by Bhargava & Bassi (1998), Draganits, Braddy & Briggs (2001) and Draganits *et al*. (2002). In the Pin Valley, the thickness of the Muth Formation reaches about 300 m near the village of Mikkim (Draganits *et al*. 2002) and with the exception of a thin dolomitic interval in its upper levels, the formation consists of exceptionally uniform, completely cemented, pure quartzites (Fig. 6a). Based on sedimentary structures from several bed-by-bed sections in the Pin Valley, the Muth Formation has been divided into four facies associations, FA 1 to FA 4, from

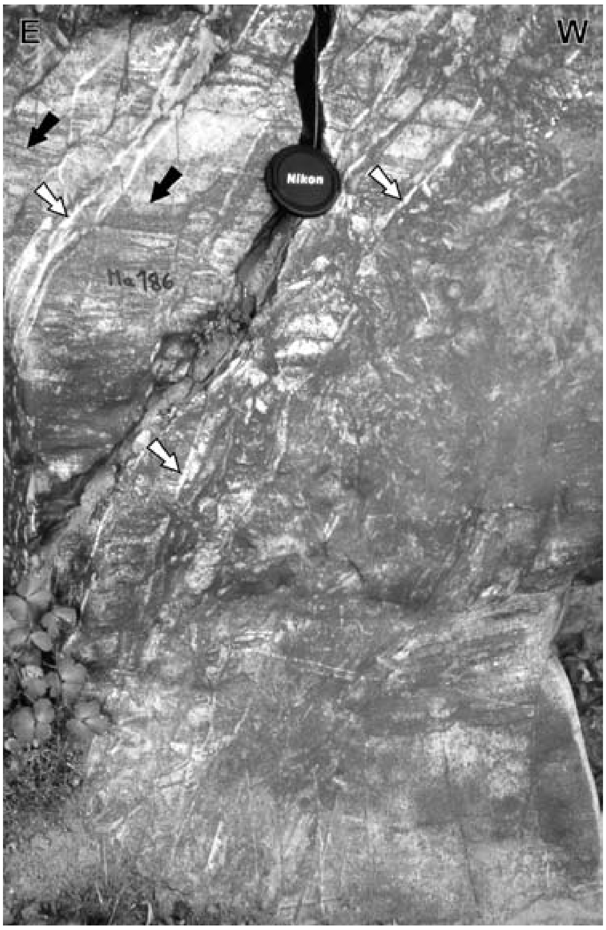

Figure 5. Photo of a vertical section of a zone of deformation bands (white arrows) cutting horizontal bedding (black arrows) at high angles. The visibility of the structures is amplified by selective rock varnish cover. Note that bedding is only slightly displaced. The zone of deformation bands shows incipient breccia formation with centimetre-sized angular fragments of sediment having been preserved between the bands. The fragments show evidence of fracturing, along with small amounts of rotation and translation; the difference from most other deformation bands in the Muth Formation (compare with Fig. 4a, b, d) might be explained by a local slightly higher degree of cementation when the deformation bands formed. 55 mm lens cap for scale.

bottom to top (E. Draganits, unpub. Ph.D. thesis, Univ. Vienna, 2000). In general, the formation is interpreted as peritidal sediments of a barrier island system. The stratigraphic position and all available evidence for the age of the Muth Formation in Spiti indicate an Early Devonian age, although its lowermost part may be of late Silurian age (Goel *et al*. 1987; Talent *et al*. 1988; R. Mawson & J. A. Talent, pers. comm. 2001; Draganits, Braddy & Briggs, 2001; Draganits *et al*. 2002).

The deformation structures and the sedimentary record in the Pin Valley (Spiti, NW Himalaya) indicate the existence of three pre-Himalayan and three Himalayan deformation events: (1) a weak pre-Ordovician deformation event, indicated by slight buckling and tilting of strata, and by the prominent angular unconformity between the Upper Proterozoic Haimanta Group and the Ordovician Shian Formation (Griesbach, 1891; Fuchs, 1982; Garzanti, Casnedi & Jadoul, 1986; Wiesmayr & Grasemann, 2002); (2) early Carboniferous to middle Permian rifting and break up of the Neo-Tethys Ocean (Stampfli, Marcoux & Baud, 1991; Garzanti, Angiolini & Sciunnach, 1996*b*; Metcalfe, 1996; Gaetani *et al*. 2004); (3) drastic subsidence related to extensional tectonics in the late Carnian/early Norian (Garzanti *et al*. 1995); (4) dominant deformation structures of the Eocene crustal thickening event ('Eohimalayan') with large scale, SW-vergent folds with wavelengths of approximately 5 km that are associated with shallowly NE-dipping, SW-directed thrusts and subvertical, NE–SW-trending tear faults (Wiesmayr & Grasemann, 2002); (5) Early Miocene ('Neohimalayan') deformation related to thrusting at the Main Central Thrust forming shallowly NE-dipping crenulation cleavage in rocks of the Upper Proterozoic to Middle Cambrian Haimanta Group in the southeastern part of the Pin Valley (Wiesmayr & Grasemann, 2002); (6) Late Miocene, broadly E–W-directed extension that affected the Pin Valley only in its northwesternmost parts (Neumayer *et al*. 2004).

4. Deformation structures

4.a. Deformation bands in the Muth Formation

Deformation bands have been studied in the Muth Formation in the Pin Valley southeast of Mikkim (Fig. 1a). They are relatively abundant at the western termination of the outcrop (location 1 in Fig. 1), where many fresh rock surfaces are exposed by the Pin and Parahio rivers and at the upper areas of the outcrop (locations 2 and 3 in Fig. 1).

The deformation bands are found in the exceptionally pure quartzites of the Muth Formation (Fig. 6a). The present day petrography of this formation shows a medium-grained and well-sorted quartzite with syntaxial quartz cements and virtually zero porosity. Single bands are about 1 mm wide (Fig. 6b), can be traced up to several metres length and cut bedding surfaces at high angles (Fig. 4). On the scale of several metres, the bands are planar, although they undulate slightly on smaller scales, forming characteristic eye and ramp structures (Antonellini, Aydin & Pollard, 1994).

Two conjugate sets of deformation bands are documented (Figs 1b, 4e). The acute angles of individual conjugate deformation band pairs show a large range; their mean values at locations 1 to 3 are 40°, 46° and 33°, respectively. After unfolding of Eocene deformation (see Wiesmayr & Grasemann, 2002), the reoriented sets of deformation bands show WNW–ESE and WSW–ENE trends in all outcrop areas (Fig. 1b).

Both deformation bands and the surrounding rock are completely cemented by quartz; during sampling the rocks never split along the bands. The deformation

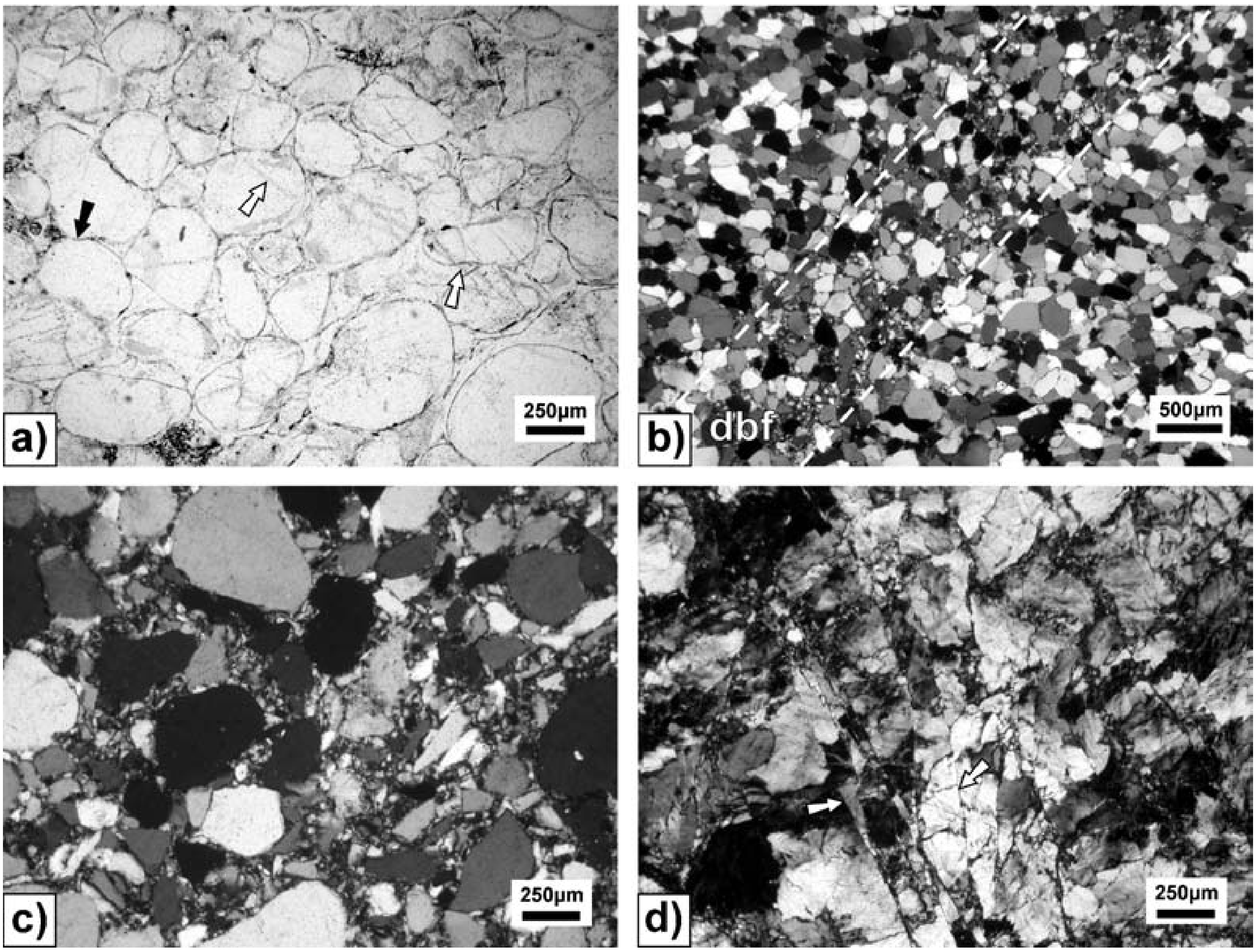

Figure 6. (a) Thin-section photomicrograph of quartzite typical of the Muth Formation without deformation bands. Note compositional and textural maturity of sediment (plain polarized light); sedimentary grain shapes are identifiable by dust and fluid inclusion rims. Truncating and impinging grain boundaries (filled arrow) are indicative of diffusive mass transfer by solution. Note the common intra-granular and trans-granular fluid inclusion trails (open arrows). (b) Thin-section photomicrograph of a deformation band fault (area highlighted by dashed lines) in the quartzites of the Muth Formation running from the upper right to the lower left (crossed polarized light). (c) Detail of a deformation band fault with well-developed cataclasis but absence of any internal foliation. Note the increase of grain size distribution due to refinement of some grains and porosity decrease compared to the undeformed quartzite in (a). (d) Photomicrograph of thin-section of a protocataclasite in the Muth Formation that formed clearly after cementation of the quartzite. Microfaults cross-cut the quartzite displacing aggregates of grains (open arrows) (crossed polarized light).

bands comprise the same pure quartz mineralogy as the parent rock outside the bands, but are conspicuous due to their lighter colour (Fig. 4d). Because of their higher resistance to erosion, which is a commonly observed feature of deformation bands (e.g. Jamison & Stearns, 1982), they show elevated relief on exposed surfaces (Fig. 4e). Each band represents a few millimetres to a few centimetres lateral offset indicated by displacements of sedimentary laminae (Fig. 4b) or of other conjugate bands (Fig. 4e). The bands have neither well-defined slip-surfaces nor slickensides on a mesoscopic scale. Deformation bands are distinct in thin-section, where evidence of effective porosity reduction and compaction by cataclasis, translation and rotation of quartz grains can be observed (Fig. 6c). The grain size is markedly smaller inside the deformation band fault compared to the host rock (Fig. 6b). Sorting is also much poorer than in the host rock and results from incomplete and unequal grain fracture, leaving large unbroken grains in a matrix of finer-grained fragments (Fig. 6c), thereby further reducing porosity by allowing denser packing of grains. Based on these microstructural characteristics and the lack of clay minerals, the bands classify as cataclastic deformation bands (Antonellini, Aydin & Pollard, 1994).

Deformation bands may occur alone, or appear in zones that are made up of many closely spaced parallel deformation bands (Fig. 4a). Southeast of Mikkim, at the western termination of the Muth Formation outcrop (location 1 in Fig. 1), four major zones of deformation bands are present (Fig. 3). They transect bedding surfaces at a high angle and can be traced for hundreds of metres. The zones are spaced several tens of metres apart with rare single deformation bands present in between. After folding of Eocene deformation is removed, the re-oriented zones of deformation bands show consistent WNW–ESE trends, and dip steeply towards SSW and NNE (Fig. 1b).

Whether the prominent, 8 m thick, elevated ridge in the southern part of location 1 (Fig. 3) represents a zone of deformation bands or a damage zone around deformation band-dominated faults is obscured by weathering and thick rock varnish. The stratigraphic offset of different facies associations of the Muth Formation on either side of this ridge indicates an offset of 20–30 m, with downthrow on the south side of the zone of deformation bands. The amount of offset implies that there is also a slip-surface at the southern side of this zone of deformation bands, although details of the slip-surface as well as details of the zone have also been obscured by weathering.

Three additional zones of deformation bands present in location 1 are less than 1 m wide. One example is about 70 cm at its thickest part (Figs 3, 4a); this zone pinches out to the east and branches into two thinner zones of deformation bands towards the west. The cumulative offset of this zone is difficult to measure in this monotonous lithology, but is of the order of about a metre. In these zones of deformation bands, the bands are closely spaced and usually only lenses of the primary sedimentary lamination are preserved between individual bands. Usually these relicts are elongated, lenticular bodies, surrounded by anastomosing deformation bands (Fig. 4a, d). In zones with very abundant and closely spaced deformation bands, hardly any primary sedimentary structures are preserved, and the deformation bands are so closely amalgamated that they appear almost as a massive single unit (Fig. 4b).

Two zones of deformation bands show incipient breccia formation with centimetre-sized angular fragments of sediment having been preserved between deformation bands; these show evidence of fracturing, along with small amounts of rotation and translation of the fragments (Fig. 5). The bands are spaced several centimetres apart, and are thus more widely spaced compared to all other deformation band zones in the Muth Formation (Fig. 4a, d). We are intrigued by the exciting possibility that these represent rare examples of the transition from deformation bands to fracture-dominated faults, the deformation style being mainly controlled by the amount of cementation and porosity (Flodin, Prasad & Aydin, 2003).

4.b. Sheared joints in the Pin Formation

Abundant sheared joints have been observed within the upper 15–20 m of the Pin Formation at the type locality, at the right bank of the Pin River, 1.3 km south of the village of Muth (Figs 1a, 7). This part of the Pin Formation comprises well-laminated, sandy, grey dolomite containing many crinoid ossicles and brachiopods (E. Draganits, unpub. Ph.D. thesis, Univ. Vienna, 2000).

Figure 7. Photograph of outcrop of steeply dipping bedding surface of the Pin Formation perpendicularly cut by sheared joints (arrows) (compare with stereo plot in Fig. 1c). View towards the south to the uppermost bedding surface of the Pin Formation at the type section, south of village Muth. Note the lighter colour of the alteration zone on both sides of the sheared joints, relative to the darker dolomite of the host rock. White rocks in the lower left are quartzites of the overlying Muth Formation. H. P. Schmid for scale.

The rocks show prominent brownish-orange weathering colours and are well cemented by dolomite and silica.

The joints in the Pin Formation are planar structures that can be traced for several metres to a few tens of metres. They represent joints or sheared joints with very little offset, because any offsets are barely visible and slickenlines have not been observed. Nevertheless, the joints represent true discontinuity surfaces, contrasting with the deformation bands in the Muth Formation. No continuation of these sheared joints into the directly overlying basal beds of the Muth Formation has been observed. The joints are noticeable in the field due to their light grey colour resulting from strong wall rock alteration by silicification on both sides of the joints (Fig. 7). The thickness of the alteration zone varies with different bedding lithologies and can reach some 60 cm thickness in siltstone beds and about 30 cm in coarser-grained bioclastic dolomite, at both sides of the joints surfaces. Additionally, the boundaries of silica alteration in siltstone beds are irregular and diffuse, but well defined in dolomite beds.

All sheared joints (30 measurements) are oriented nearly perpendicularly to the bedding surfaces and can be separated into two well-defined joint sets. After the folding of Eocene deformation is removed, they show NE–SW and WNW–ESE trends (Fig. 1c). The WNW–ESE-trending sheared joint suites are more abundant and better developed than the NE–SW orientation. Acute angles of conjugate joints show mean values of 74°.

4.c. Faults in the Muth Formation

In addition to the deformation bands, there are also several frictional faults in the Muth Formation. Their appearance is completely different to the deformation bands. These faults represent distinct, planar surfaces of discontinuity; they have polished surfaces with slickenlines and offsets in the range of metres to several tens of metres. They show many features of common frictional faults reviewed by Hancock (1985). Most of the faults belong to shallowly NE- and SW-dipping thrust planes and NE–SW-striking, sub-vertical tear folds, both related to Himalayan fault-propagation fold deformation (Wiesmayr & Grasemann, 2002) or to steeply dipping normal faults cross-cutting Eo-Himalayan folds (Neumayer *et al*. 2004).

In broader fault zones, these faults develop cohesive cataclasites with abundant, irregular, centimetres- to tens of centimetres-spaced sub-parallel faults. Hand specimens of fault material show quartz with grey to bluish colour and opalescent appearance. The weathering properties of these faults contrast with those of deformation bands in the same rock: the faults have lower weathering resistance resulting in negative relief on exposed rock surfaces (Fig. 8).

The faults in the Muth Formation are spatially separated from the deformation bands and show different orientations at high angles to deformation bands. Thus these faults cannot be the slip-surfaces that develop within zones of deformation bands, as occur in the model of Shipton & Cowie (2001).

4.d. Microstructural characteristics of deformation bands and faults with slip-surfaces in the Muth Formation

Today, the Muth Formation is completely cemented by quartz, resulting in the present-day porosity of virtually zero. At the microscopic scale the quartzites show complete cementation by crystallographically continuous quartz overgrowths, where the original shapes of the quartz grains are still indicated by dust rims and fluid inclusions (Fig. 6a).

Several observations of deformation structures at the grain-scale post-date the deformation bands. For example, structurally late fluid inclusion planes associated with faults cross-cut grain boundaries and also deformation bands, therefore identifiably post-dating the deformation band formation. Truncated and indented grain boundaries are particularly obvious on the grains where the original grain boundary and the crystallographic continuous overgrowth are both removed (Fig. 6a) (see also Blenkinsop, 2000). These processes are also visible within deformation bands (Fig. 6c), indicating that the complete cementation post-dates the formation of the deformation bands.

Figure 8. Photograph of outcrop southeast of Mikkim hosting example of a Himalayan (Tertiary) fault (thick dashed line) displacing bedding (thin dashed lines) in the Muth Formation quartzite. The large fault (F053/84, L330/38) shows several tens of metres of dextral oblique slip indicated by offset of sedimentary beds. Offset resulted in the repetition of parts of the Muth Formation section (E. Draganits, unpub. Ph.D. thesis, Univ. Vienna, 2000). Note that this fault shows much lower weathering resistance compared to the undeformed quartzite, which is in strong contrast to the higher weathering resistance of deformation bands (Fig. 4e).

In clear contrast to the cataclasis of uncemented sand grains observed in the deformation bands described above, the later faults in the Muth Formation that transect Eo-Himalayan (Tertiary) folds show microscale fracturing of the already completely cemented quartzites (Fig. 6d). These faults cross-cut the quartzite and displace already cemented aggregates of grains (Fig. 6d). These faults that belong to the Himalayan deformation even record evidence for intra-crystalline plasticity in quartz (e.g. quartz crystals with undulatory extinction, slightly curved deformation lamellae, abundant kink bands and elongated subgrains due to dislocation glide) preceding frictional faulting suggesting corresponding elevated temperatures for at least parts of the Tertiary deformation. This intra-crystalline deformation of the quartz grains results in a pronounced shape preferred orientation, thus no relics of the original shape of the detrital quartz grains and no overgrowth around the grains is preserved (Fig. 6d).

5. Discussion

5.a. Comparison with other deformation bands

The deformation bands in the Lower Devonian Muth Formation closely resemble deformation bands described by Aydin (1978) from the Entrada and Navajo sandstones in Utah (see also Davis, 1999). The bands in the Muth Formation are organized in a conjugate system, a characteristic that has been described for deformation bands in several other regions (Davis *et al*. 2000; Olsson, Lorenz & Cooper, 2004 and references therein). Unfortunately, due to the limited accessibility of the Muth Formation outcrop southeast of Mikkim and the strong Himalayan deformation, the observations are fragmentary (Fig. 1), and many details of the three-dimensional orientation of the conjugate deformation bands (cf. Davis *et al*. 2000) are not seen.

As mentioned in the introduction, deformation bands commonly group into multiple, sub-parallel, closely spaced zones, that is, zones of deformation bands (Aydin & Johnson, 1978). There are several zones of deformation bands in the Muth Formation; most of them are less than 0.5 m thick. Only one zone reaches 8 m thickness (Fig. 3) and it is also the only one accompanied by a slip-surface. This thick zone is probably comparable with complex damage zones around deformation band-dominated faults with thicknesses even greater than 10 m that are reported from other regions by Fossen & Hesthammer (2000) and Shipton & Cowie (2001).

The common observation of large, unfractured grains surrounded by smaller, fractured particles in deformation bands of the Muth Formation (Fig. 6c) agrees with the 'nearest neighbour theory' of Sammis, King & Biegel (1987), stating that the probability of grain fracture decreases as the grain size of neighbours decreases. Therefore, well-sorted sediments like the Muth Formation (Fig. 6a) are thought to be well suited for the formation of cataclastic deformation bands (see also Antonellini, Aydin & Orr, 1999).

5.b. Pre-Himalayan deformation events

The Tethyan Zone of the Higher Himalaya tectonic unit is clearly dominated by structures formed by Himalayan crustal thickening during Tertiary time (Searle, Cooper & Rex, 1988; Wiesmayr & Grasemann, 2002; Murphy & Yin, 2003); therefore prominent, preserved pre-Himalayan deformation structures are very rare. There are, however, a few distinct pre-Himalayan deformation features:

(1) Early Palaeozoic angular unconformity: Griesbach (1891) was the first to recognize the prominent angular unconformity and related depositional gap separating the upper Proterozoic to upper Cambrian Haimanta Group from the Ordovician Shian Formation. Widespread early Palaeozoic magmatic intrusions in the Himalayas (Miller *et al*. 2001 and references therein), as well as surface uplift and erosion (Garzanti, Casnedi & Jadoul, 1986) are generally related to this unconformity. Recently, Wiesmayr & Grasemann (2002) described folds and extensional faults from the Pin Valley that unequivocally pre-date the Ordovician unconformity and were interpreted by these authors as deformation structures related to the Cambrian/Ordovician hiatus.

Jain, Goel & Nair (1980, fig. 5) described another angular unconformity between the Muth and the Lipak formations on the right bank of the Pin River, opposite Muth village. However, subsequent detailed mapping, lithological sections and measurements of the orientation of the bedding surfaces reveal a gradual, conformable contact (Fuchs, 1982; Bhargava & Bassi, 1998; Draganits *et al*. 2002).

(2) Neo-Tethyan rifting: The rifting event between northern Gondwana and the Peri-Gondwana blocks (Cimmerian microcontinents), beginning in the early Carboniferous, was the most tectonically significant event along the former northern Indian passive margin before the Himalayan orogeny in Tertiary time. This event resulted in the effusion of Early Permian Panjal Traps flood basalts and the formation of the Neo-Tethys ocean (e.g. Lydekker, 1878; Şengör, 1984; Stampfli, Marcoux & Baud, 1991; Pogue *et al*. 1992; Garzanti, Angiolini & Sciunnach, 1996*b*; Metcalfe, 1996; Garzanti, Le Fort & Sciunnach, 1999; Gaetani *et al*. 2004). The rifting event had pronounced effects on the sediments of this region (Fig. 9) and fragmented the late Proterozoic to early Palaeozoic depositional area of the former northern Indian passive margin. Surface uplift of the rift shoulders resulted in widespread non-deposition, erosion and the formation of angular unconformities and depositional gaps in the stratigraphic record (Stampfli, Marcoux & Baud, 1991; Gaetani, Garzanti & Tintori, 1990; Garzanti, Angiolini & Sciunnach, 1996*b*).

(3) Late Triassic subsidence event: Finally, during the late Carnian/early Norian, large areas of the northern Indian continental margin were affected by a pronounced increase in tectonic subsidence, indicated by very high sedimentation rates (Garzanti *et al*. 1995; see also Fig. 9); knowledge of the origin and kinematics of this event is still very poor. Garzanti *et al*. (1995) interpret a disconformity and sandstone dykes on top of the early Norian Nimaloska Formation as result of an extensional tectonic event.

5.c. When did the deformation bands form?

Usually the interpretation of frictional faults suffers from the fact that these structures are difficult or often impossible to date by geochronological methods (e.g. van der Pluijm *et al*. 2001) and therefore faults are mainly dated by cross-cutting relationships. The separation of different generations of frictional faults is important for the reconstruction of the tectonic history

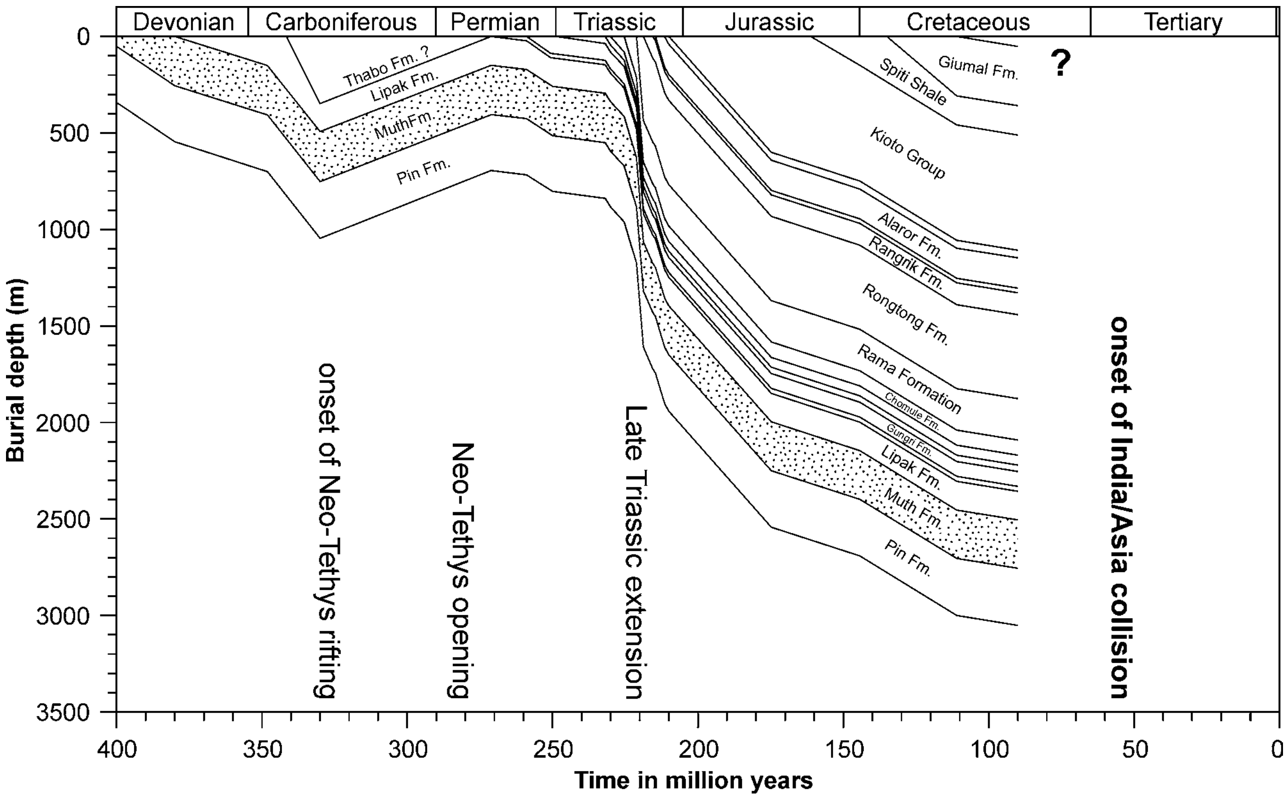

Figure 9. Burial curve for the Middle Palaeozoic to Late Cretaceous succession in the Pin Valley. Compaction and water depths not considered. Note uplift and erosion in the late Palaeozoic resulting from rifting and break up of the Neo-Tethys Ocean (Stampfli, Marcoux & Baud, 1991; Garzanti, Le Fort & Sciunnach, 1999) and the sharp increase in deposition rate in the late Carnian/early Norian (Garzanti *et al*. 1995). Curve constructed based on lithostratigraphic and age data from Garzanti *et al*. (1993, 1995), Garzanti, Angiolini & Sciunnach (1996*a*,*b*), Bhargava & Bassi (1998) and Draganits *et al*. (2002). Lithostratigraphic terminology of the Triassic sequences according to Bhargava *et al*. (2004).

of an area. Different microstructural characteristics of deformation bands have in the past been used to distinguish between different generations of deformation bands (Antonellini, Aydin & Pollard, 1994; Davatzes, Aydin & Eichhübl, 2003), but studies of tectonic histories based on separating deformation bands from other frictional faults are rare.

As the deformation bands in the Muth Formation cannot be absolutely dated by any presently practised conventional geochronological method, the age of these structures has to be constrained by relative criteria like the age of the Muth Formation, estimates of pressure and temperature conditions during deformation, cross-cutting relationships and overprinting structures of the Himalayan orogeny. The main arguments (see also Table 1) to separate the deformation bands from those Tertiary faults related to the Himalayan orogeny from the Muth Formation are (1) cross-cutting relationships: deformation bands are folded by Eo-Himalayan folds, while the faults commonly cross-cut these folds (e.g. Fig. 8) and (2) different microstructural characteristics and conditions of deformation: deformation bands deform by cataclasis, translation, rotation of quartz grains and effective porosity reduction. In contrast, the faults involve crystal plastic deformation mechanisms (Fig. 6d), which are mechanisms that can only be explained by pressure and temperature conditions reached during or after Eo-Himalayan crustal thickening at about 180 °C and 190 MPa (calculated from illite crystallinity data from Wiesmayr & Grasemann (2002) converted into depth and temperature). If we accept a pre-Himalayan age of the deformation bands, the maximum conditions during their formation might be estimated at 62–88 °C and 58 MPa lithostatic pressure, calculated from the maximum amount of overburden in the late Cretaceous, before the onset of Himalayan crustal thickening (see burial curve in Fig. 9), using average geothermal gradients (25–35 °C/km) of passive margins and densities of 2.4 g cm^{-3} for the sediments (Turcotte & Schubert, 2002). The different deformation mechanisms of deformation bands compared to 'ordinary' faults in the Muth Formation therefore indicate the existence of two different fault generations that formed in the same rock in different crustal levels.

In recent studies it has been shown that deformation bands typically form in either unconsolidated sand (Cashman & Cashman, 2000; Rawling & Goodwin, 2003) or porous sandstones within the upper 2 km of the crust (Aydin & Johnson, 1983; Antonellini, Aydin & Pollard, 1994; Mair, Main & Elphick, 2000;

Table 1. Comparison chart of the properties of deformation bands and faults in the Muth Formation as well as sheared joints in the Pin Formation

	Faults	Deformation bands	Conjugate sheared joints in the Pin Formation
Outcrop appearance	Distinct, planar, cohesive cataclastic faults; up to several hundreds of metres long, up to few metres thick; abundant, irregular parting surfaces sub-parallel to orientation of fault; slickenlines; grey to blueish colour with opalescent appearance; relatively low weathering resistance results in negative relief to the rock surface (Fig. 8)	Planar to slightly undulating, about 1 mm thick faults white colour; well visible on weathered surfaces (Figs 4d, 4e) but hardly on fresh rock; weathering resistant with positive relief on rock surfaces (Fig. 4e); never part along fault surface	Distinct planar, faults with strong quartz wall rock alteration up to 60 cm thick (Fig. 7)
Orientation and cross-cutting relationship	Shallowly NE- and SW-dipping thrustplanes related to fault-propagation folds; NE–SW-striking, subvertical tear folds related to fault-propagation folds; steep normal and transtensional faults cross-cutting Eo-Himalayan folds (Wiesmayr & Grasemann, 2002; Neumayer *et al*. 2004)	Deformation bands are folded by Eo-Himalayan folds; orientations corrected for Eo-Himalayan folding show two sets of subvertical dipping deformation bands striking WNW–ESE and WSW–ENE, respectively	Sheared joints in the Pin Fm. are folded by Eo-Himalayan folds; orientations corrected for Eo-Himalayan folding show two sets of subvertical dipping joints striking NW–SE and SW–NE, respectively
Offset	Metres to several tens of metres	< 10 mm	< 10 cm
Overall deformation orientations	NE–SW directed shortening accomodated by fault-propagation folds, post-dated by E–W extension (Wiesmayr & Grasemann, 2002; Neumayer *et al*. 2004)	Broadly E–W-oriented shortening (orientation of acute angles, Fig. 1b) associated with N–S extension (normal offset of sedimentary bedding, Fig. 4b)	Broadly E–W-oriented shortening (orientation of acute angles, Fig. 1c)
Microstructural deformation characteristics	Crystal plastic deformation indicated by undulatory extinction, abundant kink bands, dislocation glide, elongated subgrains; slightly curved deformation lamellae and pronounced shape preferred orientation in quartz crystals (Fig. 6d); crystal plastic microfaults preceding frictional faulting cross-cut the already completely cemented quartzites (Fig. 6d)	Deformation is accomodated by cataclasis, translation, rotation of quartz grains and effective porosity reduction (Fig. 6a, c)	Microstructures are obscured by strong quartz wall-rock silicification
Deformation conditions	About 180 °C and 190 MPa during Tertiary crustal thickening (illite crystallinity (Wiesmayr & Grasemann, 2002) and conodont alteration observations (Draganits *et al*. 2002))	Maximum conditions of 62–88 °C and 58 MPa estimated from the amount of 2.5 km sedimentary overburden during the Middle Cretaceous (Fig. 9)	Maximum conditions of 62–88 °C and 60 MPa estimated from the amount of 2.5 km sedimentary overburden during the Middle Cretaceous (Fig. 9); differences to deformation bands probably result from the early diagenetic cementation of the dolomitic Pin Fm.
Age	Eo-Himalayan: Middle Eocene–Late Oligocene (Hodges, 2000)	Pre-Himalayan: within the interval of Early Devonian (sedimentation age) to Cretaceous? probable age of complete cementation)	Probably contemporenaeous with deformation bands

Main *et al*. 2001). Generally, deformation bands with cataclasis are thought to develop at higher confining pressures than deformation bands without cataclasis (Zhang, Wong & Davis, 1990; Antonellini, Aydin & Pollard, 1994). However, Mair, Elphick & Main (2002) carried out triaxial deformation experiments on large sandstone samples with 7.8 % axial strain and at variable confining pressures from 13.5 MPa to 54.8 MPa. Their results show that individual deformation bands produced at low and at high confining pressures have similar microscopic characteristics, but differ in their macroscopic spatial organization. Microscopic observations further support that deformation bands can form at very low confining pressures in poorly lithified and even completely unconsolidated sediments (Rawling & Goodwin, 2003). Cashman & Cashman (2000) describe a modern example of deformation bands with little cataclasis that formed at very low confining pressure, at less than 50 m burial depth, equivalent to some 1 MPa. Consequently, because deformation bands can form at near-surface conditions, the age of deformation band formation might be close to the age of deposition of sediment.

Based on our arguments in the discussion about the microstructural characteristics of the deformation bands, and from the comparison with several field observations and experiments (see references above), we conclude that the deformation bands in the quartzites of the Muth Formation must have formed before its complete cementation. Therefore the age of the deformation bands in the Muth Formation is bracketed by the early Devonian sedimentation age of the Muth Formation (the older age limit) and the timing of complete cementation of the Muth Formation (the younger limit), whose age we now consider.

Investigation of fluid inclusions in the quartz cement is likely to be unrevealing, as these rocks have been overprinted by the temperature and pressure conditions reached during Eo-Himalayan crustal thickening (Wiesmayr & Grasemann, 2002). Therefore the age of the cementation may be estimated from indirect observations. Generally, pressure dissolution of quartz grains, followed by feldspar alteration and dissolution of amorphous silica may represent the main internal SiO_2 source for the cementation of the Muth Formation (Worden & Morad, 2000). Burial curves (Fig. 9) indicate that the Muth Formation reached depths of more than 2500 m (with maximum conditions estimated at 62–88 °C and 58 MPa lithostatic pressure; see above) not later than middle Cretaceous times. Although massive quartz cementation typically begins at some 80 °C, the pure quartz composition, the scarcity of grain coatings, together with the high maturity and permeability of the Muth Formation, may have supported quartz cementation at lower temperatures (Worden & Morad, 2000; Trewin & Fallik, 2000; Giles *et al*. 2000).

If we accept that the Muth Formation has been cemented no later than the middle Cretaceous and that deformation bands only develop in porous sandstone, we have to look for possible tectonic events in this area between early Devonian and middle Cretaceous times. There are reports of two possible candidates:

(1) The Neotethys rifting event between northern Gondwana and the Peri-Gondwana blocks (Cimmerian microcontinents) began in the early Carboniferous (e.g. Şengör, 1984; Stampfli, Marcoux & Baud, 1991; Pogue *et al*. 1992; Garzanti, Angiolini & Sciunnach, 1996*a*; Gaetani *et al*. 2004), soon after the deposition of the Muth Formation (see also Fig. 9). The deformation bands in the Muth Formation and the sheared joints in the Pin Formation may belong to this event that affected large parts of the northern Indian continental margin. The report of mafic dykes in the Muth Formation of the Pin Valley by Fuchs (1982), from the Lipak Valley in southeastern Spiti (Hayden, 1904) and from Kinnaur (Bhargava & Bassi, 1998), which are possibly equivalent to Panjal Trap, may support this interpretation. Major Carboniferous tectonic activity in the Spiti area is also indicated by stratigraphic observations, because 'the distribution of the Po Formation in the Spiti Valley suggests that an area between Po and Losar was uplifted during the late to post-Lipak period to form a NW–SE-trending subaerial high' (Bhargava & Bassi, 1998).

How much of this Middle Carboniferous to Early Permian gap in the Pin Valley, contrasting with the deposition of more than 900 m of sediments in the northwest and northeast of the Pin Valley during this interval (Bhargava & Bassi, 1998; Garzanti *et al*., 1993), resulted from primary non-deposition or was formed by sedimentation and subsequent erosion is difficult to determine. The burial curve in Figure 9 assumes the sedimentation just of the Thabo Formation (lowermost Po Group) and non-deposition during the rest of this interval that seems more realistic (Eduardo Garzanti, pers. comm; Om Bhargava, pers. comm.) than the sedimentation of the complete Middle Carboniferous to Early Permian and following erosion as suggested by Draganits *et al*. (2004, fig. 4).

(2) Large areas of the northern Indian continental margin have been affected by a pronounced increase in tectonic subsidence during the late Carnian, indicated by very high sedimentation rates (Garzanti *et al*. 1995; see also Fig. 9). The structures of this event are still poorly investigated, but a disconformity as well as sandstone dykes on top of the early Norian Nimaloska Formation are interpreted to result from an extensional tectonic event by Garzanti *et al*. (1995).

Summarizing the available evidence (see also Table 1), the deformation bands must be pre-Himalayan (pre-Tertiary age). Both the early Carboniferous rifting event and the deformation related to a late Carnian/ early Norian extensional tectonic event are plausible candidates for their formation. However, the possibility cannot be ruled out that the deformation bands in the Muth Formation represent a hitherto unknown deformation event.

5.d. Kinematic interpretation of the deformation bands

Based on several arguments summarized in Table 1, the deformation bands are interpreted as pre-Himalayan structures; we can therefore restore them to their pre-Himalayan orientation by rotation of the bedding surfaces into a horizontal orientation. After the folding of Eocene deformation (see Wiesmayr & Grasemann, 2002) is removed, the deformation bands in the Muth Formation show two WNW–ESE- and WSW–ENE-trending, steeply dipping sets (Fig. 1b). Cross-cutting relationships of the two existing sets of deformation bands indicate that bands of both orientations offset each other (Fig. 4e), thus implying the contemporaneous conjugate nature of the deformation band sets.

However, several studies of natural examples (e.g. Reches, 1978; Aydin & Reches, 1982) and rock samples deformed in the laboratory by three-dimensional strain fields (e.g. Oertel, 1965; Aydin & Reches, 1982) show more than two (typically four) distinct fault sets. Fault theories like those of Anderson (1951), based on the Coulomb criterion, or Brace (1960), based on Griffith criterion, which both predict two sets of faults, fail to accommodate three-dimensional strain (Reches, 1978; Aydin & Reches, 1982; Krantz, 1988). Reches' (1978), Aydin & Reches' (1982) and Krantz's (1988) data indicate that conjugate faults (Anderson, 1951) are only a special case of more general fault deformation and that four faults with orthorhombic geometry are required in a three-dimensional strain field.

Kinematic analysis (e.g. Krantz, 1988; Olsson, Lorenz & Cooper, 2004) of the deformation bands in the Muth Formation is complicated by the complete lack of slickenlines and therefore any information about exact displacement directions. Additionally, it is clear that the deformation bands formed when the Muth Formation still was porous, but the actual porosity during the deformation is unknown and therefore important rock mechanical properties are not available (see Dunn, LaFountain & Jackson, 1973). The reconstruction of the incremental palaeostrain is thus based on (1) the orientation of the deformation bands, (2) the orientation of their acute angles, (3) the orientation of the longest axis of lenses of undeformed sediment between individual bands in zones of deformation bands and (4) small-scale offsets of sedimentary laminae.

The acute angles of the conjugate deformation bands in the Muth Formation show broadly E–W orientations, indicating horizontal shortening in the E–W direction. This orientation of offset is clearly evident at location 2 (Fig. 4e). Generally, traces of deformation bands are regular and straight in sections normal to the band and parallel to the direction of offset, but they are wavy in sections normal to the band and normal to the direction of offset (Aydin & Johnson, 1978). Therefore, the longest axes of lenses of undeformed sediment between individual deformation bands usually are parallel to the displacement direction (Aydin & Johnson, 1978). In the Muth Formation at location 1, the relatively steep orientations of the longest axes of lenses of undeformed sediment (Fig. 4a, d) and the straight appearance of zones of deformation bands in sections perpendicular to bedding surfaces (Fig. 4a, b) indicate a considerable normal displacement component of the deformation bands. These rare vertical sections of zones of deformation bands also show considerable amounts of normal offset of individual deformation bands, indicated by offset of sedimentary bedding (Fig. 4b, c). Summarizing all available evidence, the spatial orientation of the deformation bands and field observations of offsets of sedimentary bedding are most compatible with broadly E–W-oriented shortening associated with N–S extension (Krantz, 1988; Olsson, Lorenz & Cooper, 2004).

5.e. Kinematic interpretation of the sheared joints in the Pin Formation

In the Pin Formation, all measured sheared joints are oriented nearly perpendicularly to the bedding surfaces and can be separated into two well-defined conjugate joint sets. After the folding of Eocene deformation is removed, these sets show NE–SW and WNW–ESE trends (Fig. 1c). The WNW–ESE orientation is more clearly defined; these joints are more abundant and better developed compared with the NE–SW-orientated set. Acute angles of conjugate sheared joints show mean values of 74°, which is much higher than the mean value of acute angles of the deformation bands in the Muth Formation. Probably due to the strong wall rock alteration associated with these joints, no slickenlines or useful offset criteria have been found.

The spatial orientation of the sheared joints in the Pin Formation is most compatible with broadly E–W-oriented shortening (Krantz, 1988; Olsson, Lorenz & Cooper, 2004), similar to the palaeostrain orientation of the deformation bands in the Muth Formation. Accepting that both structures belong to the same deformation event, the difference in the values of the acute angles may be explained by differences in rheology (quartz arenite versus dolomite, different porosity and/or differences in hydrostatic pressure) (Krantz, 1988; Mandl, 2000).

6. Conclusions

Conjugate sets of deformation bands are described from the Lower Devonian Muth Formation, Tethyan Zone, NW Himalayas. As the deformation bands are folded by Eo-Himalayan folds, a pre-Himalayan (pre-Tertiary) origin of the deformation bands is evident. The deformation bands in the Muth Formation therefore represent clear field data of rare pre-Himalayan deformation structures. After the folding of Eocene deformation is removed, the deformation bands show WNW–ESE- and ENE–WSW-trending, steeply dipping orientations. The orientations of the deformation bands and field observations of offsets of sedimentary bedding are most compatible with broadly E–W-oriented oblique shortening, associated with N–S extension.

Differences in microstructural scale deformation mechanisms between the deformation bands and the faults related to Himalayan orogeny (Tertiary) provide clear evidence for the existence of two separate fault sets that formed at different depths in the crust. Deformation bands in the Muth Formation formed in porous sediment and were deformed by cataclasis, translation, rotation of quartz grains and effective porosity reduction. In contrast, Tertiary (Himalayan) faults in the Muth Formation were deformed by crystal plastic mechanisms, indicated by quartz crystals with undulatory extinction, abundant kink bands, elongated subgrains, slightly curved deformation lamellae and pronounced shape-preferred orientation. Pressure and temperature conditions during Eocene crustal thickening are estimated at about 180 °C and 190 MPa (calculated from illite crystallinity data from Wiesmayr & Grasemann (2002) converted into depth and temperature).

Deformation bands in the Muth Formation are characteristic deformation structures in unconsolidated sand and porous sandstone. Thus the age of the deformation bands in the Muth Formation is bracketed by the Early Devonian sedimentation age of the Muth Formation and the timing of complete cementation of the Muth Formation probably not later than the middle Cretaceous, as indicated by compiled burial curves. Among the known pre-Himalayan (pre-Tertiary) deformation events, the early Carboniferous rifting event and the late Carnian/early Norian extensional tectonic event are plausible candidates for their formation, although a hitherto unknown deformation event cannot be excluded.

Our example from the NW Himalayas shows that deformation bands can be separated from other, frictional deformation structures by their characteristic microstructural properties, spatial architecture and stratigraphic position. Their correct interpretation, in combination with studies on the stratigraphy and sedimentology, essentially contributes to the reconstruction of the tectonic history of multi-deformed areas, even in severely folded and faulted orogens like the Himalayas.

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