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Thrust tectonics, thin skinned or thick skinned, and the continuation of thrusts to deep in the crust

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Abstract—The paper analyses the geometry of thin-skinned thrust zones, where the thrusts shallow out at depth and of thicker-skinned fault zones where much of the crust is involved and where the thrusts are frequently observed to become steeper downwards. In the interiors of many orogenic belts the steep dip of faults is not original but due to the folding above lower decoupling zones. The energy involved in the internal deformation of hanging-wall rocks may prohibit many faults becoming more shallow upwards. Such shallowing-upwards faults may occur in more ductile rocks to maintain compatibility between zones which have experienced different deformation intensities, but displacements on the faults are unlikely to be large.

Another mechanism for producing faults which steepen downwards is proposed for the major thrusts which form the southern margin to the Himalayas. These carry large thicknesses (30 to 100 km) of crustal and upper mantle rocks to the south, causing flexuring and isostatic depression of the Indian plate. The steeply dipping thrusts are not footwall ramps; these may be some distance behind the steepened zone. This thrust-induced isostatic bending of the crust has important implications when considering regional seismic interpretations as well as thrust mechanics and kinematics.

INTRODUCTION

THERE are two opposed schools of thought related to thrust tectonics: one is that major thrusts flatten at depth to join with some decoupling horizon which gradually works its way back by some form of staircase trajectory to the original source of thrust movement. The other is that thrusts steepen at depth, presumably to die out in ductile strains in the lower crust or mantle. The first thrust model is generally termed a 'thin-skinned tectonic model' or 'allochthonous tectonic model'. The second model which involves nearly vertical movements at depth without much horizontal displacement, may be termed a 'vertical model' or 'autochthonous model'.

Examples of the two models are given in Fig. 1 which shows cross-sections through the Munchberg Massif of central Germany. This is a synclinal structure of high-grade metamorphic rocks, with some eclogite rocks in the core. The basic structure was recognized as long ago as 1817 by Goldfuss & Bischoff, but has since been interpreted using an autochthonous model, as a diapir-like metamorphic uplift, thrusting out high-grade gneisses over surrounding Palaeozoic sediments (Cloos 1927, Scholtz 1929, von Gaertner 1951, Kraus 1951).The other view uses the allochthonous concept and considers the Munchberg Massif to be a far-travelled, fold and thrust sheet of lower crustal rocks (Wurm 1926, Scheuman 1935, Thiele 1966, Behr *et al.* 1980).

Similar alternative views have been put forward for large masses of lower crustal rocks in the Caledonides of Norway (Smithson *et al.* 1974) and the Hercynides of north Spain (Ries & Shackleton 1971). The Moine thrust zone has been considered to steepen at depth (Watson & Dunning 1979) or follow a gently dipping trajectory (Coward 1980). Much of the evidence for the character of thrusts at depth comes from the inner parts of orogenic belts, where field observations show thrusts more steeply dipping than in the foreland and marginal fold and thrust belts. Shackleton (1969) notes that the boundaries between cratons and orogens are commonly sharp and are often marked by thrusts or shears which, when deeply eroded, appear steeply inclined. Similarly, Beloussov (1960) stresses the importance of vertical tectonics in orogenic belts and Ramberg's (1967) centrifuge experiments show that deformation similar to that seen in orogenic belts may occur without crustal shortening but by diapiric movement of material. This view contrasts with the thin-skinned models for the Rocky Mountains (Bally *et al.* 1966, Price 1981), Appalachians (Hatcher 1981, Brewer *et al.* 1981) and Norwegian Caledonides (Hossack 1978), where inferred horizontal displacements are large. Are we dealing with different types of mountain belt?

The aim of this paper is to show that the observations of steeply dipping thrusts in orogenic belts are not

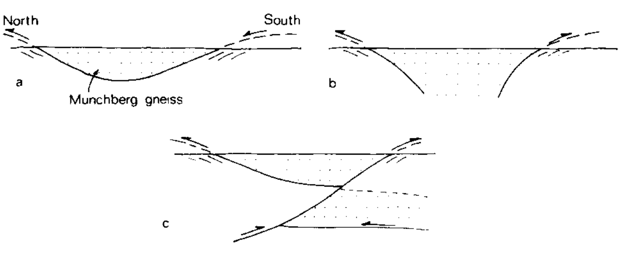

Fig. 1. (a) Allochthonous or thin-skinned model to explain the origin and emplacement of the Munchberg Massif, a block of eclogites and gneisses about 20 km across. (b) Autochthonous or diapiric model to explain the emplacement of the Munchberg. Note that it would appear easy to distinguish the two models by identifying the sense of movement on the thrusts at the southern edge of the mass. However, even if these thrust outwards, it does not rule out a thin-skinned model as shown in (c) (modified after Scheumann 1935).

incompatible with a thin-skinned or allochthonous tectonic model. Some aspects of the geometry of the different thrust models will be examined and then applied to the 'thicker-skinned' allochthonous tectonics related to continental collision, taking the western Himalayas as an example, modifying some of the geometrical concepts of thinner-skinned allochthonous models where necessary.

THE THIN-SKINNED THRUST MODEL

The basic features of a thin-skinned thrust model have been described recently by Boyer & Elliott (1982) and some aspects of the geometry have been summarized by Butler (1982) and Hossack (1983). The main concept is that thrusts follow a series of flats and ramps, the flats generally occurring along easy slip horizons such as bedding planes but possibly also along metamorphic boundaries in high-grade rocks (cf. Boyer & Elliott 1982). This flat-ramp trajectory causes the development of an anticlinal bulge over the ramp, due to the displacement of the hanging-wall over the footwall. With continued movement, this antiform grows as a large flat-topped structure (cf. Rich 1934), or alternatively the footwall to the ramp may fail leading to the development of a new flat-ramp structure. The new thrust will thus develop beneath the older thrusts carrying them forward in piggy-back style (Dahlstrom 1970). Therefore, the thrusts generally propagate into the footwall, gradually accreting slabs of footwall rock to the thrust belt (Fig. 2).

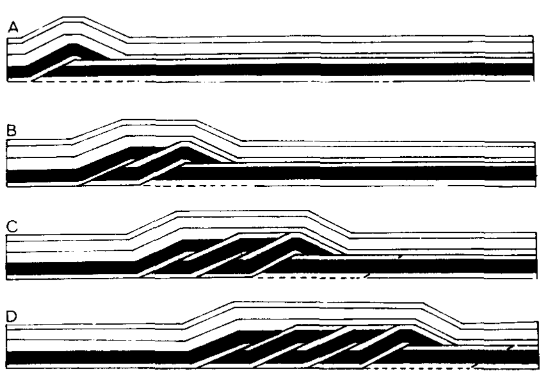

Fig. 2. Development of a duplex zone (after Boyer & Elliott 1982).

The thrust sequence, as shown in Fig. 2, can lead to the production of a duplex with roof and floor faults joined by imbricate faults, together enclosing 'horses' or 'fault-bounded packages' of rock (Boyer & Elliott 1982). A classic example of this type of structure is the Mount Crandell–Lewis thrust system of the Canadian Rockies (see Fig. 3), as described by Douglas (1952). The section shown in Fig. 3 demonstrates that the dip of early imbricate faults may vary due to the accretion of later horses or due to the movement of the duplex over an irregular thrust-ramp trajectory in the floor thrust. Boyer & Elliott (1982) show that duplex zones may be tilted and sometimes completely refolded by the development of underlying thrust systems. Boyer & Elliott (1982) also note that in the foreland parts of orogenic belts, thrusts rarely cut up the stratigraphy at angles greater than 35° except close to where the faults reach the surface. Any steeper dip to the thrust is generally a result of later folding. This gives rise to one explanation for the presence of some steeply dipping thrusts in the inner part of an orogenic belt; they have been steepened by the subsequent development of thrusts in their footwall.

STRAINS DEVELOPED IN THIN-SKINNED THRUST SHEETS

As the rocks of the thrust sheet climb from flat to ramp, they experience some strain due to bending. This strain may be comparable to flexural shear or tangential longitudinal strain in buckle folds (cf. Ramsay 1967, pp. 391–402). Assuming that most of the deformation is taken up by flexural-shear processes, and assuming a kink-band type geometry, the shear strain on the ramp is given by 2 tan $(\alpha/2)$ where α is the dip of the ramp (cf. Fischer & Coward 1982, Sanderson 1982).

Assuming deformation associated with tangential longitudinal strain, with compression on the inner arc of a fold, the thrust plane would represent a neutral surface. The strain would increase with increasing curvature and distance from the thrust plane (cf. Ramsay 1967, p. 400). If this type of deformation were to affect a thrust sheet passing into a ramp, the rocks of the sheet would first be strained and then unstrained, the strain intensity depending on the sharpness of the ramp climb and the

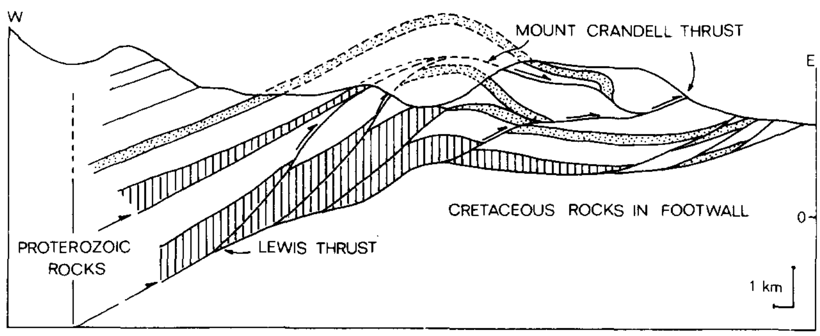

Fig. 3. Mount Crandell–Lewis thrust system, showing tilted and folded thrusts in the duplex zone (after Douglas 1952).

distance from the thrust surface. This type of deformation is unlikely to affect thick thrust sheets where this distance may be large.

Thus, the strains in thrust sheets are kept low by having only moderate curvatures to the ramp, avoiding intense longitudinal strains and/or having gentle dips avoiding intense shear strains. It is possible that in some rocks the resistance to strain may well limit the dip of thrust ramps and their curvature.

So far, only the strain in the thrust ramp has been discussed. As the thrust sheet climbs out of the ramp on to the upper flat the rocks may be strained and then unstrained again, assuming a mechanism of longitudinal strain. If flexural shear is the dominant mechanism during ramp climb, then these flexural shear strains must be removed on the upper flat. This removal of strain may cause additional structures to develop; it will be difficult to remove a strain associated with cleavage production, metamorphic phase changes or pressure solution and these structures may be modified, possibly crenulated, during the unstraining. Sanderson (1982) suggests that a crenulation cleavage may develop on the flat above a ramp, as the flexural shear strain, associated with the unstraining process, has the same magnitude though opposite shear sense, to the shear strain due to ramp climb.

As the rocks move from the flat on to the ramp other strains are developed on and behind the ramp. A simple model for their development is given by Fischer & Coward (1982), based on the shape of the Pine Mountain thrust block in Tennessee (Rich 1934, Harris & Milici 1977) and a discussion of this type of strain has been given by Elliott (1976). There must be stretching of the rocks over the ramp or extra displacement and therefore also shear strain behind the ramp. The extra displacement is given by length $2t \tan (\alpha/2)$ where t is the thickness of the layer carried over the ramp (see Fischer & Coward 1982). These strains therefore increase with increase in dip of the ramp, and in some rocks the resistance to this strain may be enough to stop movement along a particular thrust trajectory, causing rocks to fail elsewhere.

THE GEOMETRY OF SHALLOWING-UPWARDS FAULTS AND VERTICAL TECTONICS

The general concept of the shallowing-upwards faults, as described by Thom (1923), Chamberlain (1945), Osterwald (1961), Prucha *et al.* (1965) and Stearns (1975) is one of block uplift where rigid basement blocks have been uplifted along high-angle faults, sometimes reactivating earlier features in the basement. They have been described from fault zones near the margins of orogenic belts, such as the southern Alps and Pyrenees (de Sitter 1964) and probably one of the most quoted areas is that showing the Laramide uplifts in Wyoming and Montana. Here reverse faults flank areas of basement uplift (Prucha *et al.* 1965, Stearns 1978). Many of the associated smaller faults have not been considered as imbricate faults, branching from a low angle detachment fault, but as secondary structures, related to the folding and arching developed as the fault became more shallow upwards (Wise 1963). Sanford (1959) conducted experiments of block uplift with sand and other materials to show how this shallowing upwards could be related to the branching of the faults in the upper layers.

Even in this Laramide fault area of Wyoming there has been some disagreement over the concept of block uplift. Tourtelot & Thompson (1948) proposed that there was a major low-angle thrust beneath the Laramide ranges and Berg (1961, 1962) considered the uplift to be due to folding of the basement and cover, producing large, overturned fold limbs and local thrusting. In this concept of fold-thrust uplift, Berg (1962) considered the faults as secondary phenomena and still drew them steepening downwards on cross-sections. More recently, deep seismic reflection profiling by the COCORP group (Smithson *et al.* 1978, Brewer *et al.* 1981) has shown that the bounding fault to the west flank of the Wind River Mountains, neither shallows nor steepens at depth but dips at a constant angle of about 30–35° to a depth of 25 to 30 km. This certainly resolves the controversy in favour of the overthrust hypothesis, though not with such a low angle-fault as seems to occur beneath the Appalachian or Rocky Mountain belts (cf. Brewer *et al.* 1981).

A simplified model for a shallowing-upwards fault is shown in Fig. 4. Where the fault is steep, as at A–B in Fig. 4, the cut-off angle with the original horizontal layering should be large. There may be no strain in this zone, in contrast to the thin-skinned model, where with a steep dip to faults, high flexural-slip strains may be expected. If there was uplift on the fault without change in dip of fault, bed A would be displaced to say point D. With change in dip and the same displacement, A moves to A′. If there is no change in length of the beds along the hanging-wall of the fault and, for simplicity, we assume a kink band geometry, then bed length D-C will have to expand to length A′–C. Using the dips of the fault shown in Fig. 4 and applying simple trigonometry

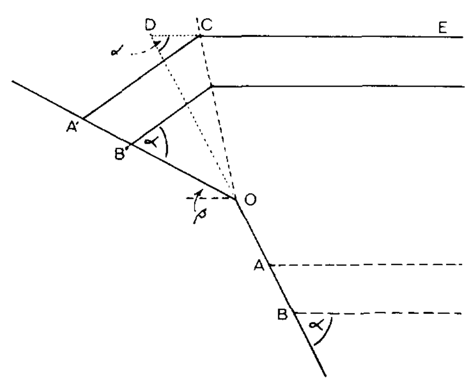

Fig. 4. Sketch to show how extensional strains are developed above a shallowing-upwards fault, assuming a simple kink band geometry. See text for details.

$$\frac{CA'}{CD} = \frac{\sin\left[90 - \left(\frac{\alpha-\beta}{2}\right) - \beta\right]}{\sin\left[90 - \left(\frac{\alpha-\beta}{2}\right) - \alpha\right]}.$$

Thus, the stretching over the top of such a shallowing-upwards fault increases with increase in α and decrease in β. This extension may occur over the whole zone, producing ductile thinning of the upper layers or numerous extension faults, the keystone faults as described by Wise (1963). Thus, for moderate to steeply dipping faults in the basement to shallow out upwards requires considerable extra strain. The extra work involved in producing this deformation may prohibit some faults from shallowing out or from having large displacements.

Another problem concerning sections through faults which shallow upwards is that generally they do not balance. In Fig. 4, as A–B moves to A′–B′ and all the beds beneath are also uplifted, what fills the void at the base of the fault? Possibly this void may be infilled by some intrusive material and possibly the intrusive material may cause the fault. If there is no infill, then ductile strains are required in rocks adjacent to the deeper levels of the fault (Fig. 5). Indeed variations in ductile shortening and thickening in the lower crust, as shown in Fig. 5, may be the best way to explain the development of steep faults. A decrease in the amount of shortening upwards would cause the fault to curve to become more shallow. Examples of this type of deformation are rare, though Shackleton (1953) describes fault-bounded blocks of basement and cover in the Caledonian rocks of Anglesey and North Wales, where the deformation varies from fault block to fault block; many of the faults are necessary discontinuities to allow for the inhomogeneous strain. Shackleton (1953) notes that many of these faults appear to shallow upwards.

Unless the strain variations are immense, such thrusts developed from strain inhomogeneities at depth are unlikely to have large displacements, though they could explain the presence of steep faults which bound the margins of some orogenic belts (Shackleton 1969). Where displacements are large, other methods of developing steep dips to faults must be found. One explanation invokes folding above or by the development of later lower-level thrusts. Another explanation using examples from the Himalayas is given below.

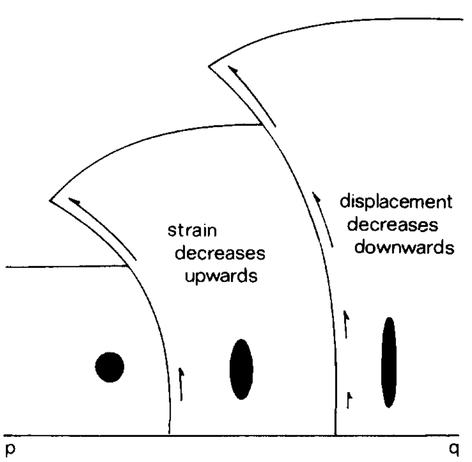

Fig. 5. Shallowing-upwards faults produced to accommodate variable amounts of crustal shortening in more ductile rocks deep in the section. If shortening in the crust decreases upwards then the faults must shallow upwards. Presumably this deformation must die out somewhere at depth and line p–q must represent some kind of decoupling plane or wide decoupling zone.

FAULT PATTERNS IN THE HIMALAYAS

General description

Having argued that thin-skinned tectonics are more likely than vertical tectonics in high-level thrust belts, it is proposed to examine the geometry of these structures when traced into the major thrusts, folds and ductile shears in the lower crust, and their development during crustal collision processes. The western Himalayas will be taken as a typical example of an orogenic belt formed by collision. The model may or may not apply to cordilleran-type orogenic belts, such as the Andes or Rockies where gravity spreading from a magmatic arc, as well as collision, has been suggested as a driving mechanism (see Bally 1981 for discussion).

The main outline of the west Himalayan geology is shown in Figs. 6 and 7. The thrusts are arranged symmetrically about the Hindu Kush–Karakorum zone. To the north, in the Pamirs, there are northward-directed thrusts. To the south, in Kohistan, Kashmir and the Northwest Frontier Province, the thrusts are directed to the south. Some of the major thrusts mark sutures between small plates, possibly small island arcs such as the Kohistan complex (Tahirkheli *et al.* 1979, Bard *et al.* 1980) trapped between the Asian and Indian continents (Ganser 1980a, b, Bard *et al.* 1980, Windley *et al.* in press). Thus, there are probably suture lines between the Karakorum and Kohistan plates and between the Kohistan and Indian plates. Within the Indian plate there are several major thrust sheets, some of the more important thrusts, the Oghi, Tarbela and Hazara thrusts in Pakistan, and the Main Central thrust and Main

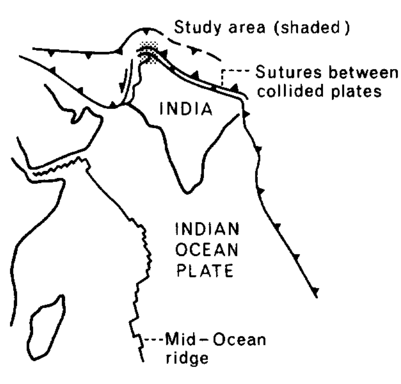

Fig. 6. Map of the Indian plate and the main plate suture zone. Note, at least two sutures are proposed for the Himalayas (see Windley *et al.* in press), a southern one which closed in late Cretaceous–Eocene times to produce the Indus ophiolite and Kohistan arc and a later northern one, active till at least Miocene times, generating masses of calc–alkaline magma in the Kohistan–Karakorum region. The area depicted in Fig. 7 is also shown.

Boundary thrust, their counterparts in India, are shown in Fig. 7. These thrusts are highly sinuous on a map tracing out large re-entrants or 'syntaxes'. The large Nanga Parbat syntaxis folds the Indian Plate gneisses and the Main Mantle thrust, which forms the southern boundary of the Kohistan complex, in a large NNE-trending antiform. The Hazara syntaxis is a large, nearly isoclinal fold of Indian Plate gneisses, originally thrust over Mesozoic sediments and Eocene–Miocene molasse.

These syntaxes allow one to make minimum estimates for displacements on the thrusts. The thrust direction in north Pakistan, as shown by stretched objects and sheath folds near the fault zone, is generally towards 160° in Kohistan, towards 200° in Kashmir, with localized variations (Coward *et al.* 1982). Thus, the Main Mantle thrust has a minimum horizontal displacement of about 100 km, as the Kohistan Complex overlies the Indian gneisses for at least this distance in the transport direction. Similarly the Tarbela and Hazara thrusts have minimum displacements of 100 km each as on the Tarbela thrusts, gneisses overlie Mesozoic sediments throughout the section and on the Hazara thrust, Mesozoic sediments overlie Miocene molasse. Thus the total shortening on the major thrusts alone is well over 300 km.

The main period for continental and small plate collision was from the late Cretaceous to the present (Windley *et al.* in press). The convergence rate as defined by magnetic anomaly patterns in the Indian Ocean, slowed down from 100 mm a^{-1} to approximately 50 mm a^{-1} about 50 Ma ago (Molnar & Tapponnier 1975, Powell 1979). This presumably marked the time of the main collision, trapping the Indus–Tsangpo ophiolite belt in southern Tibet (Gansser 1974) and the Kohistan arc in Pakistan (Coward *et al.* 1982). However, there was additional motion of India northwards, for at least 200 km. Some of this may be taken up by the major thrusts mentioned above, but other shortening may have been accommodated by the sideways displacement of Asia as India squeezed into it (Tapponnier & Molnar 1977, Molnar & Tapponnier 1978). Molnar & Tapponnier explain some of the large strike-slip faults (Fig. 7) as due to this indentation of India into Asia. It is possible however, that some movement of India relative to Asia may have been taken up by the closing of small oceans north of Kohistan and the Indus suture. Much of the calc–alkaline magmatism in Kohistan post-dates the major folds and thrusts and may have been generated from the subduction of more oceanic material (see Windley *et al.* in press for discussion). This eases the problem of having to find vast amounts of crustal shortening in the Himalayas and Tibet. Final ocean closure must have occurred by late Miocene–Pliocene times (Windley *et al.* in press) though the site of the proposed suture is uncertain.

An approximate displacement rate may be estimated for the thrusts in north Pakistan. The Main Mantle thrust must have been initiated before late Eocene times, as undeformed Eocene rocks occur in central Kohistan (Windley *et al.* in press). Likewise, movement

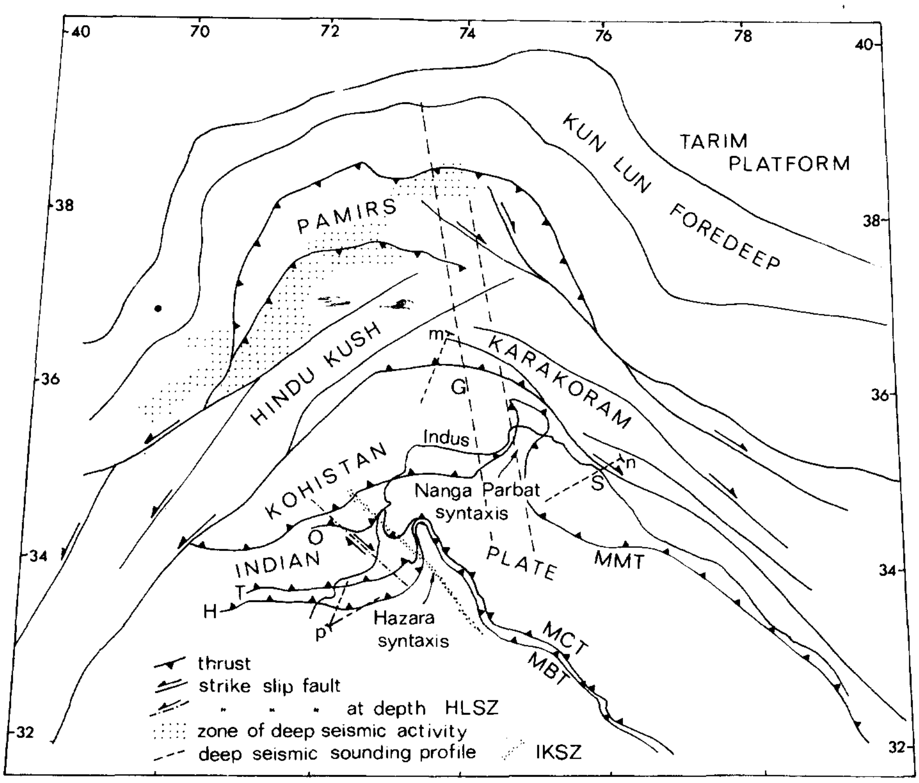

Fig. 7. Outline map of the western Himalayas (see Fig. 6 for location) showing the main zones of moderate to deep seismic activity in the Hindu Kush and Pamirs and the shallower seismic zones in Pakistan (IKSZ, Indus–Kohistan seismic zone; HLSZ, Hazara lower seismic zone). The main thrusts in Pakistan are the MMT (Main Mantle thrust), the MCT (Main Central thrust) probably equal to the O (Oghi thrust) and T (Tarbela thrust) and the MBT (Main Boundary thrust) probably equal to the H (Hazara thrust). G, Gilgit; S, Skardu. Section line n–p is given in Fig. 8, line m–p in Fig. 11.

on the Tarbela and Hazara thrusts occurred after Miocene times; they over-ride Miocene molasse. These thrusts are now inactive, displacement having been transferred to a seismic zone at about 10–30 km depth. Assuming these major faults ceased activity during the Pliocene, this suggests a time-averaged displacement rate of over 300 km in under 40 Ma, that is over 7 mm a^{-1}. This is similar to the time-averaged displacement suggested for the Norwegian Caledonides (Hossack *et al.* 1981) but obviously far less than the 50 mm a^{-1} suggested by Indian ocean spreading rates. Presumably ocean consumption north of Kohistan and the indentation of India into Asia took up the remainder of the displacement.

The continental crust, south of the Himalayas, is on average about 35–40 km thick (Chaudhury 1965), but under the main Himalayan belt it is estimated to be 60–70 km thick (Kono 1974). Beneath Nanga Parbat, and northern Kohistan, Beloussov *et al.* (1980) and Kaila (1982) estimate a crustal thickness of over 70 km. These depths, based on a deep seismic sounding profile, help constrain the large-scale geological cross-sections.

Recent seismic activity may also be used to estimate the positions and orientations of the major movement zones at depth. The majority of recent earthquakes had epicentres north of Kohistan; Yeilding *et al.* (in press) note that of 371 recent recorded earthquakes, two-thirds lie in the Hindu Kush area (Fig. 7) at intermediate depths (70–300 km). They appear to trace a sub-vertical slab which may represent a recently subducted plate (Chatelain *et al.* 1980, Tapponnier *et al.* 1981). Earthquakes occur at depths of approximately 65 km beneath Nanga Parbat, near the base of the crust, but active thrusting continues along the Himalayan frontal thrusts (Seeber & Armbruster 1979, Seeber *et al.* 1981). In this active seismic zone, earthquakes occur at 10–30 km depth, though sometimes down to 70 km.

From telemetric networks of seismic stations in the Hazara region of Pakistan, numerous and, according to Seeber *et al.* (1981), well-located hypocentres have been obtained. Seeber & Armbruster (1979) and Seeber *et al.* (1981) have traced the active faults in the basement along an arc in front of the Himalayas in India to beneath the western limb of the syntaxis and into southern Kohistan. They call this zone of active faults the Indus–Kohistan seismic zone (IKSZ) and consider it to be primarily a thrust fault dipping at about 30° to the NE. They define a second seismic zone, the Hazara lower seismic zone (HLSZ) to the southwest of the syntaxis. This HLSZ is steeply dipping and has dominantly dextral displacement. Some of the seismic data from both these zones as well as scattered shallow seismic activity are projected on to the cross-sections in Fig. 8. The shallow seismic data seem to define an active detachment zone at about 10–30 km depth, which probably connects with the shallow-dipping thrusts recognized to the south in the Salt Ranges (Wadia 1961) and to the north with the IKSZ. In more detail, Seeber *et al.* (1981) recognize steep offsets, or lateral ramps, in the detachment fault, where the motion is strike slip and north of where the detachment joins the IKSZ, they consider there to be a low-angle back thrust extending beneath the Hazara syntaxis.

The eastern section

Figures 8 and 11 show two cross-sections through the Himalayas of northern Pakistan, using the above seismic data, published maps and sections by Calkins *et al.* (1975), Gansser (1964, 1980a) and detailed and reconnaissance work by Brian Windley and myself. Much of the original data are given in Coward *et al.* (1982) and Windley *et al.* (in press). The section in Fig. 8 crosses the eastern part of the Kohistan complex and the Hazara syntaxis. The Kohistan complex here shows at least two phases of major recumbent folds, deforming sediments and volcanics of the island arc (cf. Coward *et al.* 1982). This deformation however predates many of the gabbroic, dioritic and granitic rocks, related to the later subduction event (Windley *et al.* in press). The suture between the Kohistan and Karakorum plates is here obscured by the Karakorum strike-slip shear zone, but is better observed to the east in Tibet where it consists of a mélange incorporating ophiolitic rocks. In Tibet and Indian Kashmir, the recumbent structures are modified and steepened by back-folds and back-thrusts which carry the ophiolites and Kohistan complex over the northern plate.

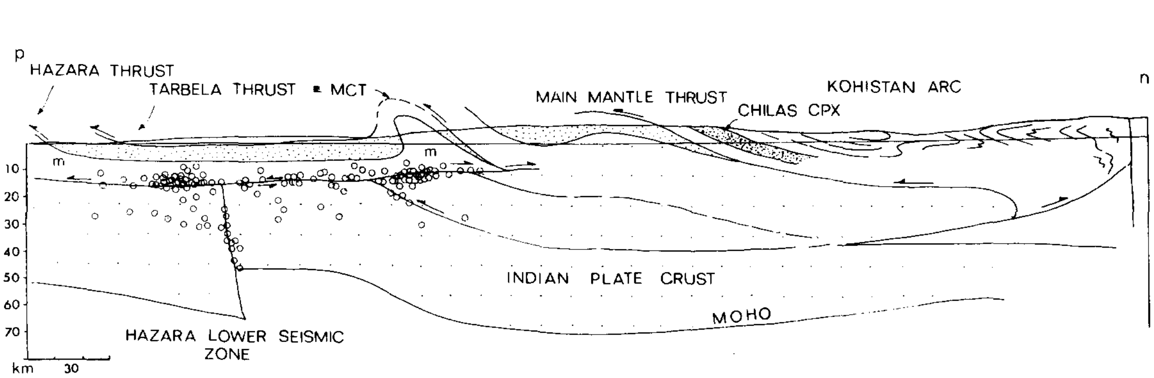

Fig. 8. Cross-section east of the Nanga Parbat syntaxis. Earthquake hypocentres shown by circles (after Seeber *et al.* 1981). m, Miocene rocks overlying an unknown thickness of deformed Phanerozoic sediments; Indian plate crust is shown with fine stipple; Mesozoic sediments of the Hazara (= MBT) sheet and Tarbela (= MCT) sheet are shown with coarser stipple.

To the south, the Kohistan arc sequence is thrust over the Indian plate. The structures are gently dipping and re-folded by large open antiforms and synforms, so that klippes of ophiolitic material lie more than 50 km south of the main suture (Gansser 1979, Fuchs 1981, Andrew-Speed & Brookfield 1982). The Main Central thrust carries the Indian gneisses some 100 km to the SSW over Mesozoic sediments which are themselves thrust by the Main Boundary thrust over Eocene–Miocene molasse. As these thrusts carry only a thin sheet of basement along with a shortened sedimentary cover, there must be shallow thin-skinned structures to the north of the Hazara syntaxis. The present seismic activity occurs on a deeper-level thrust (the IKSZ), dipping at about 30° to the NE. The Hazara syntaxis, a tight to isoclinal fold in molassic shales and sandstones, and the backthrust proposed by Seeber *et al.* (1981), probably formed above this deeper thrust. From the amount of folding and the numerous thrusts recorded in the Salt Ranges and the Siwalik molasse, there must be considerable displacement along this detachment zone, although no exact figure can yet be given. A suggested model for these structures is shown in Fig. 9.

The IKSZ and its continuation eastwards into northern India, is clearly associated with the prominent step in the topography, forming the main Himalayan mountains (Seeber *et al.* 1981). Beneath the IKSZ, the *Moho* is also depressed and this leads to the basic model for crustal-scale thrusting as shown in Fig. 10. The Indian crust is obviously too thick and too light to be carried down into the lithosphere and so must somehow underplate the over-riding Himalayan sheet (Bird 1978, Powell 1979, Seeber *et al.* 1981). During a thin-skinned thrust process, the overriding plate would be elevated, but if this were to occur in Pakistan we would need mountains some 20–30 km high. Thrust-induced isostatic depression has been recorded on restored cross-sections through thin-skinned thrust zones in the Rockies (Elliott 1976). By

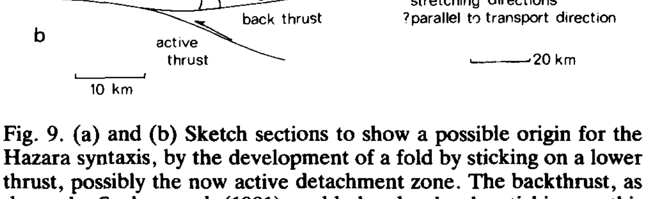

Fig. 9. (a) and (b) Sketch sections to show a possible origin for the Hazara syntaxis, by the development of a fold by sticking on a lower thrust, possibly the now active detachment zone. The backthrust, as drawn by Seeber *et al.* (1981) could also develop by sticking on this thrust. (c) Sketch map of the syntaxis, showing that the hinge is oblique to the main thrust movement as shown by stretched pebbles, etc. in the Hazara thrust zone. This interpretation suggests that the lower thrust stuck in the northwest but continued to move more in the southeast, crumpling the higher thrusts into oblique folds. MBT, Main Boundary thrust; MCT, Main Central thrust; T, Tarbela thrust; T, Tarbela thrust.

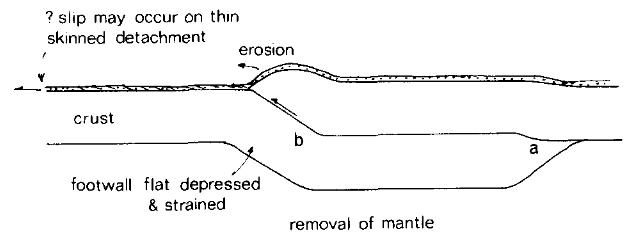

Fig. 10. Model for the development of overthrusting on a crustal scale. Thrust movement would be aided by isostatic depression of the footwall slabs as well as by erosion of the hanging-wall rock. a, footwall ramp: b, footwall beneath hanging-wall ramp.

creating the depression, thrusts make it much easier to move forwards. In the Himalayan crustal-scale thrust zone depicted in Figs. 8 and 10, the isostatic effects need to depress the lower crust and lower lithosphere by some 20–30 km. Movement of the Himalayan thrust sheet over India must be partly controlled by the deformation rate in the footwall to the thrust sheet, that is by the rate at which the crust can be depressed and flexured. Much of the seismic activity in the Indian crust, immediately south of the IKSZ, may be related to this bending strain. The footwall thrust ramp across the lower crust is not where the fault steepens, but is some distance to the rear; the exact distance depends on the amount of shortening in the upper part of the Indian plate, above the detachment horizon. Thus, on this scale and according to this model, thrust faults steepen not because of ramping but by the depression of the footwall, due to the mass of the overriding material.

The western section

The section from the western part of the Hazara syntaxis, through the main part of the Kohistan complex, west of the Nanga Parbat syntaxis, is shown in Fig. 11. In this area, the folding of rocks of the Kohistan arc is more extreme. There are two important fold phases (Coward *et al.* 1982). The first isoclinally folded lower-crustal rocks; evidence for the fold is given by way-up criteria in layered basic–ultrabasic rocks of the Chilas complex. This was then refolded by a major syncline–anticline pair, the Jijal syncline–Gilgit anticline; probably one of the largest fold structures in the world. It is some 100 km across, though this figure has been slightly exaggerated by the intrusion of late to post-tectonic granites, tonalites and diorites. A suggested model for the development of this structure is shown in Fig. 12, it involves buckling and thrusting above a decoupling horizon in lower crust–upper mantle. The Main Mantle thrust carried these Kohistan rocks partly over the Indian plate, tightly interfolding cover and basement and producing the major Hazara and Tarbela thrusts as described above (cf. Coward *et al.* 1982).

As shown in Fig. 11, the bedding, igneous layering and tectonic and metamorphic fabrics in the Kohistan arc are no longer flat but have been folded to become vertical. This steepening-up may be due to two processes.

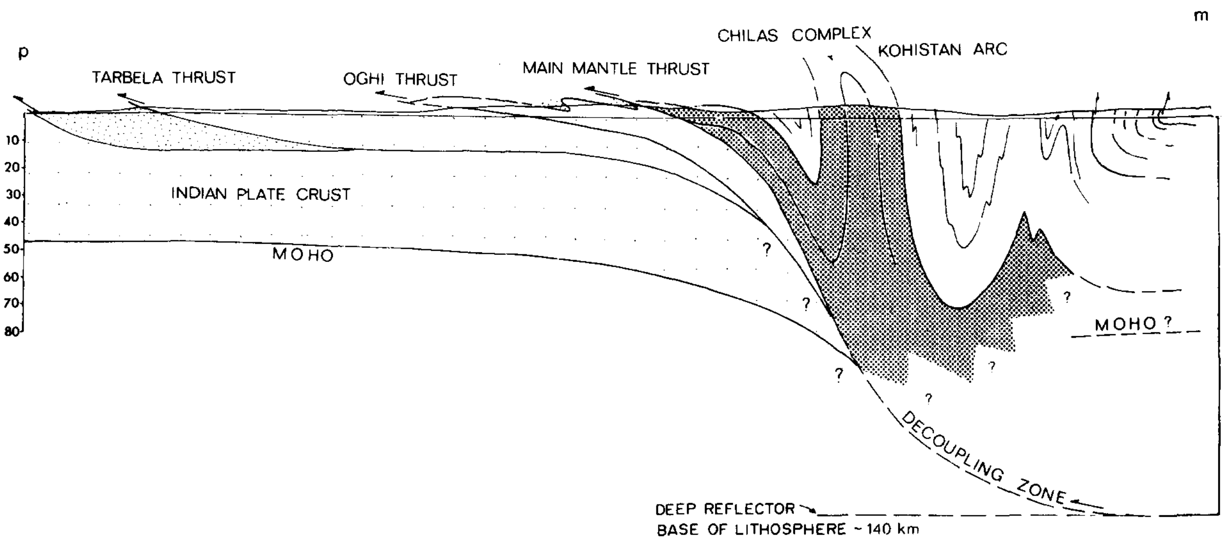

Fig. 11. Section west of the Nanga Parbat syntaxis partly after Coward *et al.* (1982). The seismic data have not been projected into the section. The dense stipple shows basic and ultrabasic rocks of the Chilas and associated bodies in the Kohistan Arc. The form of the MMT is drawn using a simple Busk construction. The form of the Indian plate below the MMT is unknown.

(i) *An isostatic bending process.* Because of the excess thickening of the Kohistan plate due to large-scale folding, the Main Mantle thrust would have carried well over 40 km of rock above its hanging-wall, thus causing major bending of the Indian plate.

(ii) *A phase of back folding.* There are numerous crenulations and minor thrusts directed northwards (cf. Coward *et al.* 1982). Some of the steepening could be associated with a major back fold related to these.

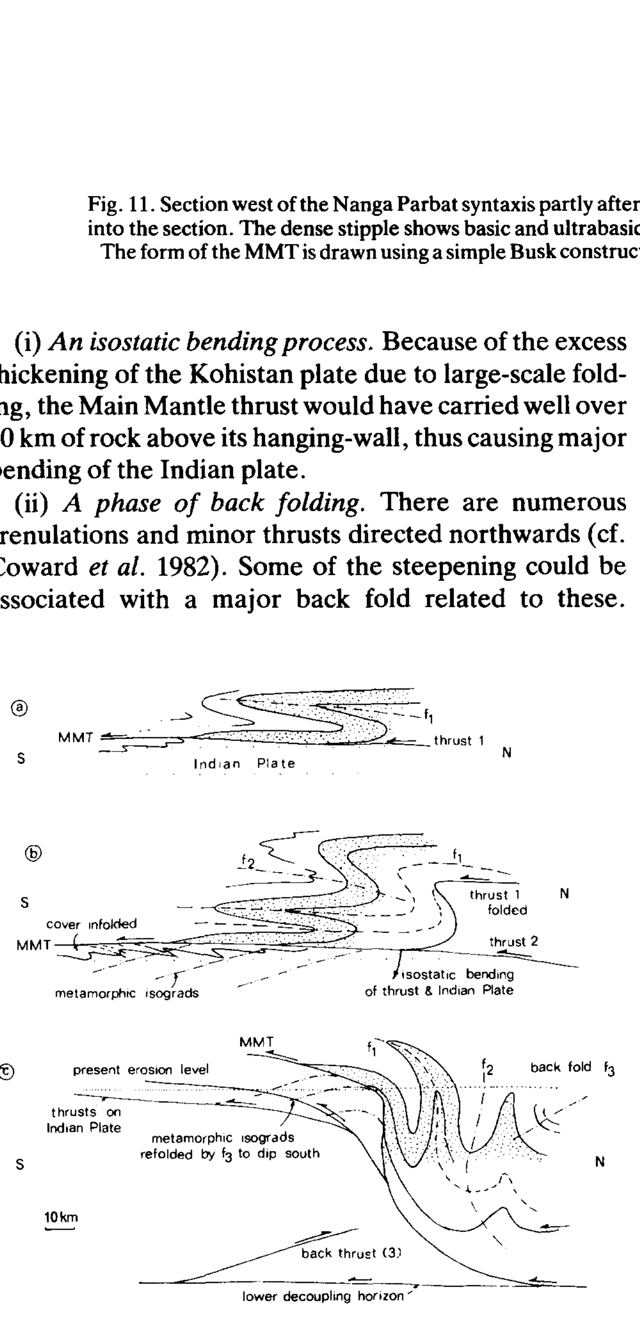

Fig. 12. Sketches to show the development sequence of structures shown in Fig. 11. In (b) for thrust 2 to move over the Indian plate, there needs to be considerable isostatic depression. This may account for the local pattern of metamorphic isograds which show an increase in metamorphic grade to the north (see Calkins *et al.* 1975, Coward *et al.* 1982). Due to the tilting of the Indian plate, these isograds may become markedly oblique (at a yet unknown angle) to the thrusts and fold axial planes but then become uplifted and folded by the late phase of back folding (f_3) shown in (c).

Similar back folds and back thrusts have been recorded east of the Nanga Parbat syntaxis.

Using a simple Busk construction, the depth of the decoupling zone, beneath the central part of the Kohistan complex, must be about 140 km. This is not the depth of the *Moho*, as mantle-type rocks may be involved in the folded Kohistan complex. Indeed the basic–ultrabasic masses of the Chilas and related complexes may represent the upper part of the mantle. Kaila (1982) suggests that there may be a deep seismic reflector at about 140 km depth and this coincides with that predicted on structural grounds. Note the steep Main Mantle thrust bounding the Kohistan arc is not a thrust ramp. As drawn in the sections it lies almost parallel to the original horizontal datum and therefore is flat which has been bent due to the mass of the overlying material. If it were to be drawn as a ramp, climbing though the structures in the Kohistan arc, the decoupling zone would have to be even deeper.

The main difference between the Kohistan complex, west and east of the Nanga Parbat syntaxis, lies in the degree of folding and thickening during early thrust

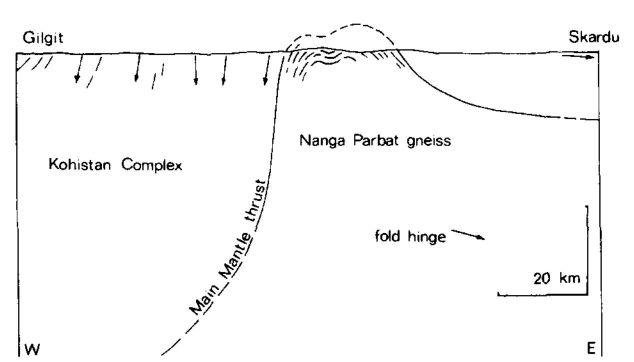

Fig. 13. West-east section through the Nanga Parbat syntaxis from Gilgit to Skardu. The arrows give the plunges of the folds in the Kohistan arc and presumably their variation reflects the change in dip of the Main Mantle thrust.

movements. This is associated with a varying degree of crustal depression, east of the syntaxis down to probably only a few tens of kilometres, west of the syntaxis down to over 100 km. The reason for this is not yet known; it is possibly due to differences in Indian plate crust east and west of the syntaxis, a difference in structure of the Kohistan arc, or different rates of plate convergence. A longitudinal section (Fig. 13), drawn normal to the thrust transport direction, shows how this difference may partly account for the Nanga Parbat syntaxis, the syntaxis having formed as a bulge next to the depressed crust.

CONCLUSIONS AND DISCUSSION

(1) Much of the energy involved in thrust movements is taken up by gravity, during the uplift of the thrust mass, by resistance to slip on the fault and by internal strain within the rock. Thus, the movement of gently dipping thrust sheets involves less energy that the uplift on steeply dipping thrusts. In high-level rocks, such as in foreland thrust and fold belts, the strain necessarily developed in hanging-wall rocks due to the flattening out of a thrust may stop movement on that particular thrust. These strains increase with increase in dip of the thrusts. It is concluded therefore, that for large thrusts developed in upper crustal rocks, vertical tectonics or autochthonous tectonics are unlikely. From mechanical points of view, the thin-skinned tectonic models are preferable. From a consideration of fault geometry, the thin-skinned models are also better; they are easier to balance and restore without involving something mysterious at depth! They have also been proven in many areas where exploration drilling and seismic reflection profiling have taken place.

Faults which steepen downwards may occur in more ductile rocks, to maintain compatibility between zones which have experienced different amounts of deformation. These faults are unlikely to have large displacements.

(2) In the inner parts of orogenic belts, faults may be steep but commonly their steep dip is not original but due to later folding. The faults may be steepened by the development of lower and later thrusts and duplex zones or by folding on lower decoupling horizons. Sometimes the early thrusts may be steepened and folded by thrusts with a movement direction opposite to that of main thrust movement.

(3) Where large thicknesses of crust are carried as part of the hanging-wall, there may be considerable thrust-induced isostatic bending of the footwall. This model is proposed for the downward-steeping thrusts at the southern edge of the Himalayas. The steeply dipping thrusts are not footwall ramps, which actually may be some distance behind the steepened zone.

This model for crustal-scale thrusting involving isostatic bending of the crust has important implications.

(a) If the active fault shown in Figs. 8 and 10 were to lock up, where would a new fault grow? Would it grow from the original ramp zone (point a of Fig. 10), at the base of the depressed crust, or would it cut through the bent and strained crust (point b of Fig. 10)? In this latter model, the late thrust would cut down through the stratigraphy of the footwall and with later uplift of the thrust zone, it may appear to be an extension fault.

(b) In thin-skinned tectonics, where an anticline is formed above a ramp, material is uplifted and there is no space problem. In the thicker-skinned tectonics, shown in Fig. 10, where a synform is produced, then there is a space problem and some mantle needs to be removed. What would be the petrogenetic and geophysical implications of this movement of mantle? Would this be pushed back to some old subduction zone, or, in the case of the indentation of India into Asia, would it be squeezed out sideways beneath the Pamirs and Afghanistan? Could this removal of mantle material be responsible for some of the deep earthquakes to the north and west of the Himalayas (Fig. 7)? Could it be the driving mechanism for some of the deep faults, such as the strike motion on the Hazara lower seismic zone (Fig. 8) which appears unrelated to detachment tectonics in the upper 20 to 30 km of the crust?

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