A Review of the Seismicity and the Rates of Active Underthrusting and Deformation at the Himalaya

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Abstract: The Himalaya have been built by underthrusting of slices of India’s crust onto the intact Indian shield, and this process continues. Fault plane solutions of most moderate earthquakes (5.5 ≤ M ≤ 7) show such thrust faulting. Assuming the great earthquakes in the Himalaya rupture by the same mechanism, crude estimates of seismic moments are consistent with rates of underthrusting determined from geologic data. The Ganga Basin owes its existence to the flexure of the Indian shield as it is underthrust beneath the Himalaya, and the progradation of the Siwalik sequence of sedimentary rock onto the shield implies a rate of convergence of 10-20 mm a between the shield and the Himalaya. Geologic cross sections and precise dating of the Upper Siwalik sequence in Pakistan using magnetostratigraphy suggest that the Salt Range is thrust southward onto the Indian shield at about 10 mm a. The likely possibility of further shortening within the Potwar Plateau and the Himalaya farther north makes this rate a lower bound. Rates of uplift of rock or of the Earth’s surface in the Himalaya with respect to sea level have only been qualitatively estimated. Nevertheless, geomorphic observations suggest that the rock of the Greater Himalaya may rise more rapidly than that of the Lesser Himalaya, except at the southern front of the Lesser Himalaya. Both geodetic measurements and folding and tilting of Late Cenozoic sedimentary rock imply that the southern edge of the Himalaya is ramping rapidly up onto the Siwalik sequence in the Indo-Gangetic Plains. The variation in vertical speeds across the Himalaya require internal deformation of the crystalline nappes comprising the mountain belt, and if the compressive strain rate is as rapid as the apparent tilting, then as much as 5 mm a of shortening between the southern edge of the Himalaya and southern Tibet may occur within the Himalaya. Thus, the rate of convergence between the Indian shield and southern Tibet is between 10 and 25 mm a, between 20% and 50% of the convergence rate between the Indian and Eurasian plates. The rate of underthrusting of the Indian shield beneath the Himalaya and the seismic moments of the great Himalayan earthquakes in this century imply average displacements of few meters for these earthquakes and recurrence intervals of 200-1000 years at any locality along the chain.

The Himalaya not only constitute the world’s highest mountain range, but they also are the locus of the world’s greatest intracontinental earthquakes. The seismic moments of three great earthquakes in 1905, 1934, and 1950 seem to be the largest among intracontinental earthquakes in this century, and the Himalaya appear to be the only intracontinental setting where great earthquakes occur on gently dipping thrust faults, like those at island arcs. Moreover, the rapid rate of underthrusting of 10 to 20 mm a makes the tectonics of the Himalayan region among the most rapid in intracontinental environments. It follows that this region is one of the world’s best laboratories for understanding how mountain belts form following continental collisions.

The data summarized below indicate that the processes that built the Himalaya continue, and they suggest that these processes do not vary much along most of the chain. Although lateral differences will become clearer as study of the Himalaya continues, there is at present no obvious reason for supposing that the gross aspects of the tectonic processes, such as the rates of underthrusting or the average seismicity along some segments of the Himalaya, are very different from those of other segments of the chain.

The following review summarizes evidence that bears on rates and styles of active deformation in the Himalaya. Following a brief review of the geologic structure and history of the Himalaya, I discuss the seismicity and fault plane solutions of moderate earthquakes and then summarize what is known about the great earthquakes. Geologic studies of the distribution and age of the Siwalik sedimentary sequence allow crude limits to be placed on rates of convergence between the Indian shield and the Himalaya, and a combination of various geologic and geophysical observations permit the inference that rates of uplift vary across the Himalaya. These latter data, therefore, suggest that the Himalaya is undergoing modest internal deformation, in addition to the underthrusting of the shield beneath the range.
Brief summary of the geologic structure and history of the Himalaya

Prior to the collision of the Indian subcontinent with southern Eurasia, a large ocean basin lay north of India and was subducted northward beneath southern Eurasia. The rocks comprising the Himalaya were part of India’s northern continental margin, and slices of the sedimentary cover and its underlying basement were scraped off the leading edge of the Indian subcontinent as it collided with and plunged beneath the southern margin of Eurasia (e.g., Gansser, 1964, 1966; LeFort, 1975; Mattauer, 1975). This convergence has continued, apparently unabated, since the collision (e.g., Molnar & Tapponnier, 1975; Patriat & Achaëché, 1984; Powell, 1979), with different thrust faults in the Himalaya being active at different times. At present, the Indian plate is flexed down to form the Ganga basin south of the Himalaya and underthrusting beneath the Lesser Himalaya occurs at a very gentle angle ($\delta = 3^\circ-5^\circ$) (Fig. 1).

Fig. 1. Simple, typical geologic cross section across the Himalaya showing the basic tectonic units. In the south, the Ganga Basin is filled with Cenozoic sedimentary rock, material largely derived by erosion of the Himalaya. The crystalline basement of the basin, the Indian shield, extends beneath the front of the Himalaya and the Lesser Himalaya. About it are two main nappes, separated by two main faults. The Main Boundary fault separates the pre-Tertiary rock from the Tertiary sedimentary rock of the Indo-Gangetic plains, and the Main Central Thrust separates mostly low grade metamorphic rock of the Lesser Himalaya from higher grade rock and its sedimentary cover in the Greater Himalaya. Above this sedimentary sequence are thrust sheets containing fragments of ophiolites. Farther north, the ancient southern edge of southern Asia is defined by the Kangdese granites and overlying sedimentary rock. (This cross section is derived largely from Gansser’s [1964] and is modified from Molnar’s [1984] simplified version of it.)

In simple terms, the Himalaya consist of three units separated from one another and from the rest of the Indian subcontinent by three major fault zones (Fig. 1) (e.g., Gansser, 1964; LeFort, 1975). The northern thrust zone separates the ophiolitic mélange of the Indus-Tsangpo Suture Zone from the underlying Mesozoic and Palaeozoic sedimentary rock of India’s ancient northern margin. Slip on the now inactive Main Central Thrust placed India’s ancient northern margin onto a more internal part of the Indian subcontinent, part of which is now exposed in the Lesser Himalaya. The Main Boundary Fault separates the pre-Tertiary rock comprising most of the Lesser Himalaya from the Tertiary sedimentary rock, mostly deposited on the Indian continent after its collision with southern Eurasia. Although the Main Boundary Fault presently crops out as a steep fault, it seems to flatten at depth and to dip gently beneath the Lesser Himalaya (e.g., Molnar, 1984, 1988; Molnar & Chen, 1982; Seeber et al., 1981). This gently dipping part continues to be active, but in most localities, the surface trace of the Main Boundary Fault seems no longer to be active. Instead, slip now occurs on listric thrust faults that cut upward from the flat segment of the Main Boundary Fault into the Siwalik sequence in the Ganga Basin (e.g., Baker et al., 1988; Delcaillau, 1986a, 1986b; Lilie et al., 1987; Pennock et al., 1989). This southward migration of the locus of major active deformation at the surface has characterized the history of the Himalaya since the collision. At the same time, the folding and minor thrust faulting of essentially all units within the Himalaya attest to additional deformation that is not part of this southward migration.

Data of several types constrain the styles and rates of active deformation. Studies of earthquakes provide the strongest constraints on the position and shape of the principal thrust fault separating the underthrusting Indian shield from the overriding crystalline nappes of the Himalaya. What little is known about great earthquakes in the Himalaya yields a weak constraint on the rate of underthrusting. Tighter constraint are provided by the ages of onlap of sediment in the Ganga Basin (Lyon-Caen & Molnar, 1985) and by a detailed study of the timing and amount of thrust faulting in one locality (Baker et al., 1988). These data indicate an average rate of underthrusting of at least 10 mm a. At the same time, deformation also actively occurs within the Himalaya, a phenomenon implied by evidence...
showing variations across the chain in the speed with which rock moves upward, away from the centre of the earth (Molnar, 1987c).

**Seismicity and Fault Plane Solutions**

Most well located earthquakes in the Himalaya have occurred roughly 60 to 100 km north or northeast of the southern margin of the range, and roughly 10 to 50 km south or southwest of the crest of the Greater Himalaya. Thus, along most of the chain, the earthquakes occur south of the surface trace of the Main Central Thrust (e.g., Baranowski et al., 1984; Molnar, 1984; Molnar et al., 1977; Ni & Barazangi, 1984). This fact and the lack of clear evidence for active faulting along the surface trace of the Main Central Thrust implies that this fault is inactive.

Fault plane solutions of most moderate earthquakes (5.5 ≤ M ≤ 7) in the Himalaya (Table 1) indicate thrust faulting, with one nodal plane dipping gently northward or northeastward and with the other plane dipping steeply southward or southwestward (Fig. 2) (Baranowski et al., 1984; Chandra, 1978; Ekström, 1987; Fitch, 1970; Molnar & Lyon-Caen, 1989; Molnar et al., 1973, 1977; Ni & Barazangi, 1984; Rastogi, 1974; Rastogi et al., 1973). Obviously, the gently dipping plane, the one more nearly parallel to the faults mapped in the Himalaya, is more likely to be the fault plane than is the steeper, south or southwesterly dipping plane. Thus, these earthquakes indicate continued underthrusting of material beneath the Himalaya. Note that small earthquakes occur throughout much of the crust of the Himalaya (e.g., Armbruster et al., 1979; Gaur et al., 1986). Because they provide information only about orientations of strain within the crust, and neither about rates of deformation nor about slip on major faults, I ignore them in this discussion.

**Dip of the main underthrust zone**

The focal depths and the fault plane solutions of moderate sized earthquakes are consistent with most of them having occurred on the top surface of the Indian plate as it slides beneath the overriding crystalline

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**Table 1. Source parameters of earthquakes in the Himalaya**

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Seismic moments (M0) are given in units of 10⁸ Newton-meters. The nodal plane listed first is assumed to be the fault plane, except where marked by an asterisk (*). Strikes (str), dips, rakes, azimuths (az), and plunges (pl) are given in degrees.

*Baranowski et al. [1984]; Chen and Molnar [1989]; Dziewonski et al. [1989]; Ekström [1987]; Jackson and Yielding [1983]; Molnar et al. [1977]; Molnar and Lyon-Caen [1989]; Ni and Barazangi [1984].
nappes comprising the Lesser Himalaya (Baranowski et al., 1984; Molnar, 1984; Molnar & Chen, 1982; Ni & Barazangi, 1984). Accurate focal depths of moderate-sized earthquakes have been determined, in most cases, using a comparison of synthetic and recorded P-waveforms (Baranowski et al., 1984; Ekström, 1987; Jackson & Yielding; 1983; Molnar & Lyon-Caen, 1989; Sipkin & Needham, 1991). For smaller events, however, intervals between P and pP phases are clearly recorded by short period seismographs (Chen & Molnar, 1990; Ni & Barazangi, 1984). Nearly all measured focal depths of earthquakes in the Himalaya lie between 12 and 18 (±4) km below the Earth's surface. For more than half of the moderate earthquakes east of 76°E, the dips of the gently dipping nodal planes are less than 15°, in general with uncertainties of at least 5°. The top surface of the basement of the Indian shield lies at a depth of about 5 km beneath the Ganga Basin (Karunakaran & Ranga Rao, 1979; Sastri et al., 1971). The average dip of the continuation of a surface at that depth to the hypocenters of the earthquakes would be about 6° ± 3° (Fig. 3). This average dip is similar to the dips of the northward or northeastward dipping nodal planes of many earthquakes (Fig. 4) and to the very gentle dip of the Moho beneath the Lesser Himalaya deduced from profiles of gravity anomalies (Lyon-Caen & Molnar, 1983, 1985; Warsi & Molnar, 1977). Thus, it seems sensible to conclude that these earthquakes occurred on the top surface of the underthrusting Indian plate. The variations in the dips of the gently dipping nodal planes for different earthquakes (Fig. 4) and the uncertainties in the hypocenters allow this inference to be false, but because it is sensible, I accept it.

The dips of the gently, northerly or northeasterly dipping nodal planes for some events appear to be too steep (25°-35°) for these events to have occurred on the top surface of an intact Indian plate (Fig. 4) (Baranowski et al., 1984; Molnar & Caen, 1982), but see Ni and Barazangi (1984) for a contrary opinion. These steeper planes may reflect an abrupt steepening of the Main Boundary fault beneath the Greater...
Fig. 3. Simple profile across the Ganga Basin and the Himalaya showing the relationship of the depth of the basin to the depths of moderate earthquakes in the Himalaya, and the projected top surface of the Indian Shield. Note that the depths of the earthquakes are consistent with the boundary between the Indian plate and the overriding Himalaya being the top surface of the intact Indian shield as it is underthrust beneath the Himalaya.

\[ \theta \approx \arctan \left( \frac{\text{Depth of Earthquakes} - \text{Height} - \text{Depth of Basement}}{\text{Distance}} \right) \]

\[ \approx \arctan \left( \frac{(15-2.5 \text{ km})}{90 \text{ km}} \right) \approx 6^\circ \pm 3^\circ \]

Fig. 4. Projections of hypocenters and nodal planes onto a profile across the Himalaya. Note that most nodal planes are roughly parallel to the surface defined by the top of the Indian Shield beneath the Indo-Gangetic Plains and the earthquake hypocenters (Figure 3), but that some nodal planes are steeper than it. Either those events with steeper nodal planes did not occur on the top surface of the Indian plate, or the top surface dips much more steeply where those earthquakes occurred. Because of the uncertainties in both the hypocentral locations and their projected positions onto one profile, we cannot rule out the possibility that the earthquakes with steeper planes occurred north or northeast of those with gentler planes (Modified from Ni and Barazangi [1984]).

Himalaya (e.g., Molnar & Chen, 1982). The steeper gradient of Bouguer gravity anomalies across the Greater Himalaya (≈ 2 mgal/km) than across the Lesser Himalaya (≈ 1 mgal/km) and the corresponding inference that the Moho dips more steeply beneath the Greater than the Lesser Himalaya are consistent with a steepening of the fault (Kono, 1974; Lyon-Caen & Molnar, 1983; 1985; Warsi & Molnar, 1977). Such a
steepening would imply that this, northern, segment of the fault does not mark the top surface of the intact Indian Shield. Instead, it seems to mark the location where the Indian plate is flexed much more than farther south. Because this corresponds to the location where the Main Boundary Fault seems to steepen and to have cut into the Indian plate, the southern edge of the Greater Himalaya would, in the jargon of structural geology, appear to overlie the "footwall cut-off" (Lyon-Caen & Molnar, 1983; Molnar, 1984).

If the apparently abrupt increase in the dip of the Main Boundary Fault beneath the southern edge of the Greater Himalaya defines the footwall cut-off, the throw on that fault is 80-100 km. This estimate of the throw ignores deformation of the overriding material and therefore is a lower bound. If the Lesser Himalaya had undergone shortening to half its dimension before the Main Boundary Fault became active (Johnson, 1986), the total amount of underthrusting on this fault might be as much as 200 km. In any case, if the apparent steepening does mark the footwall cut-off, then large amounts of slip (= 500-1000 km) cannot have occurred.

Unfortunately, the uncertainties in epicentres, of 10 to 20 km, and the likelihood of a long-strike variations in the shape of the thrust plane do not allow the cross sectional shape of the thrust fault to be resolved, and in particular this inferred steepening of it cannot be defined with confidence.

The occurrence of earthquakes with relatively steep planes might simply indicate internal deformation of either the overriding nappes or the underthrusting Indian Shield (Baranowski et al., 1984; Molnar & Chen, 1982). There is a suggestion from wide-angle seismic reflections that the interface between the Indian Shield and the Himalayan nappes does not dip as steeply as 25°-35°, but only 10°-15° beneath the Greater Himalaya. Lépine et al. (1984) showed a series of seismograms with a pair of strong, reflect signals recorded in Nepal at distances of 120 to 150 km from a shotpoint in southern Tibet. The apparent velocity of 7 km/s of the reflected phases, intermediate between the 6 and 8 km/s expected for wide-angle reflections from shallow or deep, horizontal surfaces, renders a horizontal reflector unlikely. Hirn and Sapin (1984) showed that both the apparent velocity and the amplitude of the earlier phase could be matched using a structure with a strong reflector dipping north at 12°, such that its depth roughly 50 km south of the crest of the Himalaya would be 10 km, and its depth some 30 km north of the crest would be 30 km. Hirn and Sapin (1984) inferred a southward dipping reflector for the other reflected phase, but the two reflected phases, consistently 1.3 s apart, could be reflected from parallel northward dipping surface (Molnar, 1988). If these reflected phases did mark an interface dipping 12° northward, and if that interface were the northward continuation of the Main Boundary Fault, then the earthquakes showing steep nodal planes of 25° to 35° must reflect internal deformation of the Indian Shield or the overriding crystalline nappes of the Himalaya.

Azimuths of slip vectors and orientations of convergence

For most earthquakes, slip vectors (directions perpendicular to the auxiliary planes and parallel to the orientations of slip on the faults) are oriented perpendicular to the local trend of the range (Fig. 2 & 5) (e.g., Baranowski et al., 1984; Molnar & Chen, 1982; Molnar & Lyon-Caen, 1989). For earthquakes in the eastern Himalaya, the northerly plunging slip vectors imply a southward overthrusting of the Himalaya onto the Indian plate in this area. For earthquakes in western Nepal and India, the northeasterly plunging slip vectors indicate a southwestward direction of such overthrusting. Thus, if the Indian plate behaves rigidly, the diverging directions of overthrusting require east-west extension of the area north of the Himalaya (Armijo et al., 1986; Baranowski et al., 1984; Molnar & Chen, 1982; Molnar & Lyon-Caen, 1989). This east-southeast-west-northwest extension is amply demonstrated by the normal faulting in southern Tibet (Armijo et al., 1986; Molnar & Chen, 1983; Molnar & Lyon-Caen, 1989; Molnar & Tapponnier, 1975; 1978; Ni & York, 1978; Tapponnier et al., 1981, 1986). Moreover, the rate of extension of 18 ± 9 mm a. calculated from the variations in the direction of overthrusting (Molnar & Lyon-Caen, 1989) and from estimates of rates of underthrusting at the Himalaya (Lyon-Caen & Molnar, 1985) accords with the rate of extension obtained from offsets on faults in southern Tibet (Armijo et al., 1986).

It is difficult to know whether this radially oriented overthrusting in the Himalaya is a Late Cenozoic phenomenon that began when the north-south trending normal faulting in Tibet developed (Armijo et al., 1986; Mercier et al., 1987), or is older and occurred, for instance, during slip on the Main Central Thrust. Brunel (1983, 1986) reported that stretching lineations and linear fabrics in the shear zone, some 5-10 km thick, of the Main Central Thrust also show a radial orientation, but Pécher and Scaillet (1989) inferred that such lineations trend nearly north-south all along the Himalaya. Thus, Brunel’s (1983, 1986) data suggest that east-west extension occurred during the late stages of slip on the Main Central Thrust zone, which is difficult to date, but surely occurred before 10 Ma and before the north-south trending normal faults in southern Tibet became active. Alternatively, Pécher and Scaillet’s (1989) observations imply that the east-west extension is a young phenomenon. Unfortunately, a date of the initiation of normal faulting in Tibet cannot resolve this controversy. A divergence of stretching lineations and
the corresponding east-west extension in southern Tibet does not require that extension occurred by normal faulting; conjugate strike-slip faulting in Tibet could have occurred.

**Exceptional fault plane solutions**

The fault plane solutions of a few earthquakes in the Himalaya do not show the simple underthrusting of India beneath the Himalaya. The solutions for one event (15 August, 66) indicates normal faulting with a T-axis approximately perpendicular to the chain, and that for a less well constrained solution (19 June, 79) appears to be similar (Fig. 2, Table 1). These events occurred within the Indian plate south of the Himalaya, and their solutions apparently reflect a stretching of the top surface of this plate in response to the flexing of it (Chandra, 1978; Molnar et al., 1973, 1977; Ni & Barazangi, 1984). The earthquakes of 19 November, 1980 and 20 August, 1988 also appear to have occurred within the Indian plate, but unlike those south of the range, these occurred deep below the Himalaya, at depths of 40 km or more and possibly within the upper mantle of the Indian plate. A large component of strike-slip faulting with P-axes trending roughly north-south (Diewonski et al., 1989; Ekström, 1987; Sipkin & Needham, 1991) indicates north-south compression of Indian lithosphere. Thus, although the majority of the fault plane solutions indicate an underthrusting of India beneath the Himalaya, a few of them reflect internal deformation of the Indian plate.

**Great earthquakes in the Himalaya**

Three earthquakes dominate the historical seismicity of the Himalaya in the twentieth century, the 1905 Kangra earthquake, the 1934 Bihar-Nepal earthquake, and the 1950 Assam earthquake, and another in 1897 is associated with the rupture an extensive area just south of the Himalaya, beneath the Shillong Plateau (Fig. 2). In addition, there have been three other earthquakes with M ≥ 7.0: in 1916, 1936 and 1947. Among these earthquakes, the three in 1897, 1905 and 1934 received thorough studies of the macroseismic effects associated with them. The 1950 Assam earthquake occurred in the most remote part of the Himalaya, its eastern end, but for it the limited macroseismic investigations can be supplemented by deductions from seismically recorded seismograms. Unfortunately, the types of faulting and amounts of slip associated with all four great earthquakes are poorly constrained, and I am
aware of little information that allows tight constraints on the seismic moments of any of these earthquakes. I give here only a brief discussion of them and of modifications to the parameters listed in Table I of Molnar and Deng (1984).

15 June, 1897 Assam earthquake (and the formation of the Shillong Plateau)

Molnar and Deng (1984) did not discuss this event because it occurred in the last century, but a review here seems appropriate. This earthquake clearly occurred near the Shillong Plateau, for ground motion was strongest there. The only evidence for surface faulting, probably only secondary faulting, also was from that plateau, and aftershocks were most frequently reported as felt at towns on the plateau (Molnar, 1987b; Oldham, 1899).

The style of faulting that occurred in 1897 is not well constrained, and no useful constraint can be put on the seismic moment of the 1897 earthquake. Oldham (1899), Seeber and Armbruster, (1981), and Molnar (1987b) all assumed that thrust faulting occurred on a plane dipping gently northward beneath the Shillong Plateau, but there is little evidence to substantiate this. Fault plane solutions of recent moderate earthquakes in this area (Fig. 2) indicate thrust faulting with planes dipping steeply north or south, or strike-slip faulting on planes striking northeast or northwest, and focal depths of 25-55 km suggest that these events occurred well below any gently dipping thrust fault in this region (Chen & Molnar, 1990). Thus these recent moderate earthquakes probably cannot be associated with rupture of the same fault as in 1897. Gravity anomalies from the Shillong Plateau indicate large deviations from isostatic equilibrium (e.g., Verma, 1985; Verma et al., 1976), which require support of a large excess mass beneath the plateau either by dynamic processes in the underlying mantle or by a strong lithosphere. An underthrusting of strong Indian lithosphere could support the excess mass of the Shillong Plateau, but such an explanation is not unique. Finally, thick Late Miocene, Pliocene, and Quaternary deposits in the Sylhet trough south of the Shillong Plateau imply a Late Cenozoic accelerated subsidence of this trough, and S. Y. Johnson and Alam (1990) inferred that underthrusting of the Shillong Plateau onto the Sylhet trough provided the force that caused the subsidence. They deduced that several tens of kilometres of underthrusting had occurred.

Oldham's (1899) thorough investigation of damage associated with the 1897 earthquake yielded a very complete map of isoseismals except in the area east of the Shillong Plateau. From the dimensions of the area severely shaken, the east-west dimension of the rupture area is at least 200 km across the plateau, and perhaps 300 km, if it extends farther east into the area not examined by Oldham and his deputies (Molnar, 1987b). Although the greatest destruction was centered on the Shillong Plateau, there was also much destruction north, west, and south of it. Seeber and Armbruster (1981) concluded that the entire area encompassed by the isoseismal VII on the Modified Mercalli scale, 560 km in its east-west dimension and 300 km in its north-south dimension, defines the extent of the rupture zone. Oldham (1899), however, showed convincingly that the vast majority of this destruction was associated with slumping and fissuring of unconsolidated sediment, and not with deep seated faulting. Consequently, I think that Seeber and Armbruster vastly overestimated the east-west dimension of the rupture zone (Molnar, 1987b).

Seeber and Armbruster (1981) also inferred that the rupture extended northward beneath the Himalaya, north of where intensities were measured. Oldham's lack of data in that area permits an extension of the rupture beneath the Himalaya, but by no means requires it. In any case, as Seeber and Armbruster (1981) suggested, I would not expect the next great earthquake in the Himalaya to occur north of the Shillong Plateau. At the same time, it probably would be unwise to design a program for seismic risk that neglected the possibility of major earthquakes in that segment of the Himalaya (Molnar & Pandey, 1989). The evidence for a rupture on a gently dipping plane beneath the Shillong Plateau in 1897 is too incomplete to neglect such a possibility.

5 April, 1905 Kangra earthquake

Fault plane solutions of moderate earthquakes near the epicenter of this earthquake (Fig. 2) indicate underthrusting on planes dipping gently northeast (Ekstrom, 1987; Molnar & Lyon-Caen, 1989). Let us assume such a fault plane. From a compilation of the evidence given by Middelmis (1910) for the destruction associated with this earthquake, Molnar (1987a) concluded that this event was not as great an earthquake as the 1934 Bihar-Nepal earthquake. The distribution of damage certainly permits a long rupture (300 km) (Seeber & Armbruster, 1981), but by no means does it require a rupture longer than about 120 km. The destruction was much greater in the western third of the region enclosed by the isoseismal for an Intensity VIII on the Rossi-Forel scale than in the rest of that region, and only in that western third was the intensity (I = IX-X) comparable with that of the 1934 Bihar-Nepal earthquake, discussed below. Because Richter (1958) assigned comparable magnitudes to the earthquakes of 1905 and 1934, I assign comparable seismic moments to them, but because of the shorter zone of high intensity in 1905, I use a smaller seismic moment for that earthquake: 2 × 10^21 Nm, which is half that assigned to the 1934 Bihar-Nepal earthquake.
Each, however, is uncertain by at least a factor of two. If the fault length were only 120 km and the width of the rupture were 100 km, the area of the ruptured plane would be $1.2 \times 10^{10}$ m$^2$. From the definition of the seismic moment, $M_o = \mu A \Delta u$, where $\mu = 3.3 \times 10^{10}$ N/m$^2$ is the shear modulus, $A$ is the fault area, and $\Delta u$ is the average slip on the fault plane (Aki, 1966), the average displacement in 1905 would have been 5 m.

15 January, 1934 Bihar-Nepal earthquake

From the distribution of destruction reported by Rana (1935) for Nepal and by Dunn et al. (1939) for India, Pandey and Molnar (1988) showed that damage due to shaking, as opposed to slumping and fissuring of unconsolidated sediment, was greatest in the Lesser Himalaya of eastern Nepal. Dunn et al. (1939) had reported the highest intensity in the plains of northern India, where slumping was extensive, and seeber and armbruster (1981) inferred that slip occurred on a nearly flat plane lying beneath the Indo-Gangetic Plains. Among Dunn’s colleagues, only J. B. Auden visited Nepal, and although he made three traverses across parts of Nepal, he was not able to examine the areas of Nepal where Rana (1935) reported the greatest damage and the most numerous casualties. Hence, it seems much more likely that the epicenter lay beneath the Himalaya (Pandey & Molnar, 1988), than south of the range, as is commonly believed (Dunn et al., 1939; Richter, 1958; Seeber & Armbruster, 1981; Singh & Gupta, 1980). In fact, Chen and Molnar’s (1977) relocated epicenter lies within this area of very high intensity in eastern Nepal. Moreover, the association of extensive damage with extensive slumping in the plains of India (Dunn et al., 1939) is also consistent with the rupture zone of this event not extending beneath the plains. Consequently, I assume that this event can be associated with thrust faulting on a plane dipping gently beneath the Himalaya (Pandey & Molnar, 1988).

Molnar and Deng (1984) assumed such underthrusting on a gently northward dipping plane ($\delta = 5^\circ$). Using Chen and Molnar’s (1977) measured amplitude spectra of long-period surface waves, they estimated a seismic moment of $4.1 \times 10^{12}$ Nm, but that value is uncertain by at least a factor of two. The amplitude spectra are not well constrained, and the inferred scalar moment is inversely proportional to the assumed dip of the fault. Thus, the error in the moment could be larger than a factor of two. From the reports of macroseismic effects by Rana (1935) and Dunn et al. (1939), Pandey and Molnar (1988) inferred that the length of the rupture zone was at least 100 km, but no more than 300 km. They suggested 200 (± 100) km. For dimensions of 200 km by 100 km, the calculated average displacement is 6.2 m.

29 July, 1947 earthquake (28.63°N, 93.73°E, M = 7.9) Chen and Molnar (1977) determined a seismic moment of $1 \times 10^{13}$ Nm for this earthquake assuming underthrusting on a plane dipping 10° north. Molnar and Deng (1984) suggested that a dip of 5° might be more likely, for which the moment would be $2 \times 10^{12}$ Nm.

15 August, 1950 Assam earthquake

Relocated epicenters of earthquakes that occurred in the eastern Himalaya and surroundings in the year following this earthquake are dispersed over a broad area extending both westward and south-southeastward from the epicenter of the mainshock (e.g. Ben-Menahem et al., 1974; Chen & Molnar, 1977). Because fault plane solutions of moderate earthquakes in this region (Fig. 2) indicate underthrusting on gently northward dipping planes, it is logical to assume such a style of faulting for at least part of the rupture of the 1950 earthquake (e.g. Chen & Molnar, 1977). The group of earthquakes located south-southeast of the mainshock, however, led to the deduction that the strike of at least one plane that ruptured in 1950 trends south-southeast from the epicenter of the mainshock (Ben-Menahem et al., 1974; Molnar & Deng, 1984; Molnar et al., 1977).

A re-assessment of the locations of these earthquakes, however, indicates that virtually all of the well recorded aftershocks (by more than 50 stations), and therefore presumably the better located ones, occurred beneath the Himalaya, and not southeast of the mainshock (Fig. 6). Those to the south-southeast apparently were smaller, and presumably more poorly located, events. Thus, probably the principal faulting was underthrusting of the Indian shield beneath the Himalaya, and the component of right-lateral strike-slip on the north-northwest trending plane deduced by Ben-Menahem et al. (1974), and assumed also by Molnar and Deng (1984), might be overestimated. This inference, however, is not consistent with the spectral ratios of Love to Rayleigh waves reported by Ben-Menahem et al. (1974). Thus, in the absence of new information constraining the seismic moment, I, nevertheless, assume the same parameters given by Molnar and Deng (1984): $M_o = 6 \times 10^{12}$ Nm on a gently northward dipping thrust fault, some 200 km long, and $M_o = 2 \times 10^{12}$ Nm on a north-northwesterly trending right-lateral strike-slip fault some 400 km long. The average slip on a thrust fault 100 km wide would have been 9 m.

On the basis of the magnitudes assigned to earthquakes on 28 August, 1916 (29.82°N, 80.71°E, M = 7.5) and 27 May, 1936 (28.27°N, 83.37°N, M = 7.0), I assume seismic moments of $8 \times 10^{12}$ Nm and $2 \times 10^{12}$ Nm, each with an uncertainty of a factor
of two (Molnar & Deng, 1984). I assume that thrust faulting occurred on gently northward dipping planes, but the small moments make their contributions to the calculated rate of convergence negligible.

Seismic slip rate
The sum of the scalar moments of earthquakes along a single fault can be used to infer the amount of slip associated with these earthquakes (Brune, 1968):

\[
\text{slip} = \frac{\text{sum of seismic moments}}{\text{shear modulus} \times \text{area of fault}}
\]

Dividing this estimate of the slip by the interval of time in which the earthquakes occurred gives an estimate of the rate of slip.

The sum of the seismic moments of major earthquakes in this century (1905, 1916, 1934, 1936, 1947, and 1950) is $1.24 \times 10^{22}$ Nm, ignoring the possible strike-slip component associated with the 1950 Assam earthquake. Nearly half of the sum, nevertheless, is contributed by the poorly constrained estimate of the moment for the part of the 1950 earthquake associated with thrust faulting. Dividing this sum by the shear modulus ($3.3 \times 10^{10}$ N/m$^2$), by the length of the chain (2500 km), by the approximate width of the zone (100 km), and by the duration of the period (90 years) yields an average rate of convergence of 17 mm/a, essentially the value reported by Molnar and Deng (1984). Its uncertainty of at least a factor of two, however, gives it little quantitative value, except that it is less than the convergence rate between India and Eurasia of about 50 mm/a (DeMets et al., 1990).
Geologic constraints on rates of convergence between the Indian shield and the Himalaya

At present, the only estimates of rates of convergence of India toward the Himalaya, or of underthrusting of the shield beneath the range, exploit the ages of the Siwalik sequence of sedimentary rock. Quantitative studies of Holocene or late Quaternary faulting in the Himalaya are sparse, and as noted above, the seismic history of the Himalaya does not place a tight constraint on the rate of underthrusting. Thus, the estimates of average rates of convergence at the Himalaya must be associated with longer durations than those described below for other parts of Asia and must also be associated with large uncertainties.

Migration of the flexure of the Indian plate beneath the Ganga Basin

Both the cross sectional shape of the basement of the Ganga Basin and the gravity anomalies across it imply that the basin formed by a flexure of the Indian plate, as it has been underthrust beneath the Himalaya (e.g., Duray et al., 1989; Karner & Watts, 1983; Lyon-Caen & Molnar, 1983; 1985). The depth of the basin should depend on the load that flexes the plate down. Insofar as the Indian plate can be treated as an elastic plate overlying an inviscid fluid, however, the width of the basin should depend only on the flexural rigidity of the Indian plate. Thus, regardless of changes in the height of the Himalaya or in the forces that have flexed the plate since 10 to 20 Ma, if the flexural rigidity of the Indian plate were essentially constant, the width of the Ganga Basin also should not have varied much in that interval.

As India moves north, is flexed down, and underthrusts the Himalaya, sediment of progressively younger age is deposited directly on the Indian Shield at the southern edge of the basin. Once deposited, this sediment is carried north, and still younger and coarser material is deposited atop it. Insofar as the width of the basin has remained constant, the age of the oldest sediment deposited on the Indian plate should increase northward or northeastward beneath the Indo-Gangetic Plains. If the rate of convergence of India and the southern edge of the Himalaya were constant, then a plot of the age of the oldest Cenozoic sediment in the Ganga Basin vs the distance to the southern or southwestern edge of the basin should yield a straight line with the slope giving the convergence rate (e.g. Lyon-Caen & Molnar, 1985). Using Karunakaran and Ranga Rao's (1979) estimates of ages for the sedimentary rocks at the bottoms of deep drill holes in the Ganga Basin and an average smooth curve for the southern edge of the basin (Fig. 7), Lyon-Caen and Molnar (1985) found a systematic northward or northeastward increase in ages consistent with a rate of convergence of 15 ± 5 mm/a (Fig. 8).

Fig. 7. Map of the Indo-Gangetic Plains showing locations of drill holes. The dot-dashed line approximates the southern edge of the Ganga Basin, from which distances to the drill holes were measured. (From Lyon-Caen and Molnar [1985].)

Fig. 8. Plot of inferred ages of sediment at the base of the drill holes in Fig. 7 vs. the distance to the southern edge of the basin. Also shown are lines corresponding to constant rates of convergence between the Himalaya and the Indian shield, assuming that the flexural rigidity of the lithosphere has not changed with time. Despite the scatter and the large uncertainties in ages and distances, these data indicate a convergence rate of 10-20 mm/a. (From Lyon-Caen and Molnar [1985].)

The major uncertainty in this rate derives from poor constraints on the age of material at the bottoms of the drill holes. The Siwalik Sequence in Pakistan has been very accurately dated using fission-tracks and magnetostratigraphy (e.g., G. D. Johnson et al., 1979.
Rates of progradation onto the Indo-Gangetic Plain

Burbank and Raynolds (1984; Burbank et al., 1986, 1988) reported an abrupt change in sedimentation southwest of the Pir Panjal between 3.0 and 1.5 Ma. Conglomerate with distinctive crystalline clasts progressively prograded 50 km southwestward from the front of the range onto the plains at about 30 mm/a. If the topography and sedimentation were in steady state, then this rate of progradation would imply a convergence rate of 30 mm/a, perpendicular to the chain (Fig. 9). Given both that the Pir Panjal may have emerged in front of the rest of the Himalaya only at about 5 Ma (Burbank & Johnson, 1983), and that the rate of erosion could well have changed dramatically.

Fig. 9. Simplified profile of stratigraphic sections showing the dates of initiation of deposition of conglomerate as a function of distance from the Main Boundary fault in the foothills of the Pir Panjal (northeastern Potwar Plateau). Notice how gravel prograded southwestward onto the plains at about 30 mm/a. Paleomagnetic stratigraphy using the magnetic polarity time scale (MPTS) defines the time scale. (From Burbank et al. [1988].)
because of climatic changes in the last few million years (e.g., Molnar & England, 1990), neither the topography nor sedimentation are likely to have been in steady state. Thus, while this progradation of sedimentary facies is consistent with convergence rates of tens of millimetres per year, it probably should not be used to place a quantitatively strict bound on the convergence rate.

**Rate of convergence between the Salt Range and the Indian Shield**

The sediment cover of the Potwar Plateau in northwestern Pakistan (Fig. 10), south of the Main Boundary Fault, is clearly detached from the underlying basement along a layer of Cambrian salt, and the western part of the Potwar Plateau constitutes a thin thrust sheet, some 120 km wide, overthrust nearly rigidly over the Indian Shield (e.g., Baker et al., 1988; Jaumé & Lillie, 1987; Lillie et al., 1987; Pennock et al., 1989; Yeats et al., 1984). The thrust fault ramps up into the sedimentary cover at the Salt Range.

![Fig. 10. Map of the Potwar Plateau in Pakistan, showing the locations of the Soan syncline, the profile A-A' used by Baker et al. (1988) to measure a balanced cross section, and the Kalabagh fault (KF) studied by McDougall and Khan (1990). Other abbreviations include BH-basement fault, HR-Hill Ranges, IA-Lilla antiline, NPDC-north Potwar deformed zone, PH-Pabbi Hills, RI-Rawat fault, SR-Sugaran Range, SR1-Salt Range thrust, Islamabad, R-Rawalpindi, and exploration wells at Dhariala (D) and Lilla (L). (From Baker et al. [1988]).](image)

Dating of unconformities in the Siwalik Sequence near the frontal thrust of the Salt Range in Pakistan (G. D. Johnson et al., 1986; Raynolds & Johnson, 1985) and a minimum estimate of displacement on that fault allow a bound to be placed on the rate of underthrusting there (Baker et al., 1988; Lillie et al., 1987, 1989; Pennock et al., 1989). G. D. Johnson et al. (1986) interpreted the abrupt increase in sedimentation rates at sites near the Salt Range between 2.1 and 1.6 Ma as evidence that anticlines were emerging and that ramping on the frontal thrust at the Salt Range began in this interval. Using seismic reflection profiles to constrain a balanced cross section, Baker et al. (1988) estimated a displacement of 19 to 23 km on that ramp-overthrust (Fig. 11). Thus, they deduced a convergence rate between the southern part of the Potwar Plateau and the plains farther south of about 10 mm a.

The simplicity of the interpretation given by Baker et al. (1988) is not unanimously accepted and does not apply throughout the Salt Range. Burbank and Beck (1989a, 1989b) questioned the inference of a latest Pliocene emergence of the Salt Range. Using magnetostratigraphy to date unconformities near the Salt Range, they inferred that deformation in the Salt Range began at about 5 Ma. In such a case, the average rate of overthrusting would be slower than 10 mm a. Nevertheless, they too suggest that the overall rate of convergence across the Potwar Plateau may be about 10 mm a.

East of the profile studied by Baker et al. (1988), a smaller amount of overthrusting at the Salt Range of 18 km since about 2.5 Ma yields a lower rate of 7 mm a (Pennock et al., 1989). In this area, however, the sedimentary rocks of the Potwar Plateau have not been rigidly overthrust onto the underlying basement, and folding and thrust faulting account for an additional 6 km of shortening. Some of this shortening clearly occurred since 2.5 Ma, such as that along the Soan syncline (G. D. Johnson et al., 1986; Raynolds & Johnson, 1985), but some appears to be older.

A north-northwesterly trending right-lateral strike-slip fault, the Kalabagh fault (Fig. 10), truncates the west end of the Salt Ranges, and 12-14 km of middle to late Quaternary strike slip on this fault is also consistent with a convergence rate of 7-10 mm a (McDougall & Khan, 1990).

Thus, the overall rate of convergence between the northern margin of the Potwar Plateau and the Indian shield seems to be roughly 10 mm a, but the distribution of that shortening varies from place to place. Moreover, recall that this value is a minimum for the Himalaya alone, because it does not include possible convergence farther north. In fact, folding of the sedimentary rock on the north limb of the Soan syncline sequence between 2.1 and 1.9 Ma (Burbank & Raynolds, 1984; Johnson et al., 1986) suggests that shortening may have occurred simultaneously with overthrusting of the Salt Range onto the Indian shield. Moreover, Yeats and Hussain (1987, 1989) inferred that Pleistocene, but not Holocene, faulting and folding occurred yet farther north in the Peshawar basin (Burbank & Tahirkheil, 1985). Thus, the convergence rate between the shield and the crystalline rock of the Himalaya might be as high as 15 or 20 mm a.
Summary of constraints on the rate of underthrusting

There seems little doubt that the rate of convergence between the Indian shield and the southern edge of the Himalaya is at least 10 mm a. The southward decrease in ages of sediment in the Ganga Basin implies faster convergence, but probably no faster than 20 mm a. The well defined rate of about 10 mm a at the Salt Range does not seem to permit a slower rate, but given the possibility of convergence between the Potwar Plateau and the rest of the Himalaya, a faster rate is possible. The high rates of progradation onto the Indo-Gangetic Plain are qualitatively consistent with convergence at tens of mm a, but they probably are not accurate enough to be definitive. Thus the rate seems to be between 10 and 20 mm a, but it need not be constant along the range.

Rates of vertical movement within the Himalaya

The word “uplift” has been used in a number of different ways, in reference to the Himalaya and other regions, but let us concern ourselves here either with changes in mean elevations (uplift of the Earth’s surface) or with the speed with which rock moves away from the reference ellipsoid (such as sea level) of the Earth (uplift of rock) (England & Molnar, 1990). These differ from one another by the rate of denudation, which clearly is rapid in parts of the Himalaya. The dramatic morphology of the Himalaya attests to relatively recent, rapid denudation. Cooling ages determined from fission tracks and $^{40}$Ar/$^{39}$Ar retention spectra have allowed that rapid denudation, of several millimeters per year, to be quantified in some segments of the Himalaya (e.g. Mehta, 1980; Saina et al., 1979; Sharma, 1984; Sharma et al., 1978; Zeitler, 1985; Zeitler et al., 1982a, 1982b).

These observations of rapid denudation are often taken as evidence that much of the Himalaya, if not the entire range, has recently been uplifted (e.g. Gansser, 1982). Moreover, palaeobotanical data indicating recent, drastic local climate change on the north slope of the Himalaya sometimes are extrapolated to apply to the entire Tibetan Plateau (e.g. Xu Ren, 1978, 1981). This review begins with two convictions. First, evidence of denudation provides no useful information about elevation changes and only weak, generally qualitative, constraints on the average vertical speeds of the rock now exposed at the surface (England & Molnar, 1990). Second, vertical speeds of rock, if not rates of elevation change, probably vary systematically across the Himalaya (Lyon-Caen & Molnar, 1983; Molnar, 1984, 1987c; Molnar & Lyon-Caen, 1988; Whitehouse, 1990).
Thus, to facilitate this discussion, let us consider different portions of the Himalaya separately, beginning with the north slope of the Greater Himalaya, where mean elevations may have changed in the last few million years. Then I discuss the southern front of the Lesser Himalaya, where vertical movements of rock might also be relatively rapid. Finally I consider the Lesser Himalaya, for which constraints on vertical movements are weak, but where vertical displacements seem to have been less than in the Greater Himalaya or at the southern edge of the Lesser Himalaya.

Uplift of the north slope of the Greater Himalaya

Most of the data bearing on elevation changes of Greater Himalaya are remnants or fossils of plants whose present species grow in warmer climates. In general, no consideration for global climatic change has been made, and present-day elevations of nearest living relatives have been used to assign paleo-elevations to fossil plants. Systematic cooling of the Earth is likely to make such estimates of paleo-elevations underestimate (Molnar & England, 1990). Moreover, the use of nearest living relatives, which ignores evolution, makes most published paleobotanical estimates of paleo-elevations qualitative at best (e.g., MacGinitie, 1962; Wolfe, 1971).

Fossils of plants with nearest living relatives in warm tropical climates have been used to infer low elevations in the Early Tertiary Period. Late Cretaceous and Eocene coal-bearing shales contain tropical flora, from which Kong and Du (1981) inferred paleo-elevations of 1000 m. In contrast, Guo Shuang-xing (1981) reported Eocene “angiosperms, containing broad-leaved evergreen trees with a couple of deciduous species” from the same area, and he inferred elevations of 100-300 m. Song and Liu’s (1981) and Xu Ren’s (1981) descriptions of flora are similar to Guo’s, and Xu Ren (1981) inferred Eocene tropical to subtropical climates with elevations less than 500 m. Finally, Lakhanpal et al. (1983) reported fossils of palms in the Late Oligocene basal part of the Indus molasse on the north slope of the Ladakh Himalaya, and from nearest living relatives growing in tropical latitudes, they inferred elevations of 500 m or less. Recall, however, that the global climate was much warmer in the Eocene Epoch than it is now (e.g., Haq et al., 1987; Shackleton, 1984; Shackleton & Kennett, 1975; Wolfe, 1978), and that these rocks were 2000 km or more south of their present latitudes. Thus, these assignments of paleo-elevations could be underestimated by 2000 m.

By Late Miocene time, when rocks in the Greater Himalaya probably were only a few degrees south of their present latitudes, “subtropical” forests apparently had been replaced by temperate broad-leaved deciduous forests. Guo Shuang-xing (1981) inferred paleo-elevations of 500 m for the north slope of the Himalaya in southern Tibet, and Lakhanpal et al. (1983) inferred paleo-elevations of 600-2100 m for the Ladakh Himalaya. Again because the global climate was warmer than now, and mean annual temperatures might have been 6-10°C higher, these paleo-elevations might also be underestimated by more than 1000 m.

In the Middle and Late Pliocene Epoch, conifers, including fir and spruce, and later Alpine trees and shrubs became dominant; accordingly Chen (1981), Kong and Du (1981), Song and Liu (1981), and Xu Ren (1978, 1981, 1984) inferred elevations of 3000 m or more. Once again, because global temperatures were higher, possibly by 6-7°C, than now, these paleo-elevations could be 1000 m too low.

Even considering the climatic change and global cooling in the Late Cenozoic Era, the rapid change from Late Miocene “temperate” forests to Pliocene alpine conifer forests, to no forests at all at present elevations of about 5000 m probably requires a change in mean elevation. For instance, changes of mean elevations of 1000 m to 3000 m since 3 to 10 Ma imply average rates of 0.1 to 1 mm a, and faster rates are possible. Unfortunately, the methods used to infer paleo-elevations and the uncertainties in corrections to them do not allow a truly quantitative analysis, and these data are inadequate to disprove a steady uplift of, for instance, 0.2 mm a since 25 Ma. Thus, we may conclude that the Himalaya rose in Cenozoic time and that the rate of uplift might have been somewhat faster in Neogene than Palaeogene time, but more precise quantitative constraints are not available.

Southern front of the Lesser Himalaya

Several facts suggest very high vertical speeds of rock at the southern front of the range.

The only reported geodetic measurement of uplift of the Himalaya is from a short leveling profile (Fig. 12) from the Indo-Gangetic Plain across the front of the range and into the Lesser Himalaya (Chugh, 1974). Releving of the profile after 12 years revealed an increase of about 50 ± 10 mm in elevation across the southwestern 25 km of the Lesser Himalaya, corresponding to a rate of relative uplift of benchmarks of about 4 ± 1 mm a. During this interval, no large earthquakes occurred in this area; thus the vertical speeds could be due to steady deformation. The measured vertical speeds strictly refer to the material on which the benchmarks were fixed. Because erosion seems to be rapid, the rate determined by releveling probably overestimates the rate of change of the mean elevation. Moreover, Chugh’s (1974) unpublished report does not provide enough information to evaluate the uncertainties in the releveling. Nevertheless, if we accept his data, they imply that the material comprising the front of the range is rising rapidly with respect to the Indo-Gangetic Plain to the south.
Relatively rapid uplift of rocks along the southern margin of the Lesser Himalaya is also implied by the altitudes and elevations of river terraces along major rivers traversing the Himalaya. The clearest examples are terraces adjacent to rivers that have cut through the small anticlines in front of the Himalaya (Deccaillau, 1986a, 1986b; Iwata & Nakata, 1985). Several terraces, which surely formed adjacent to southward flowing streams, now dip northward (Fig. 13) (Deccaillau, 1986a; Iwata & Nakata, 1985), indicating a tilt of the front of the range, with the largest vertical displacement at the front. This pattern of tilting suggests that slip occurs on listric thrust faults that flatten northward beneath the Lesser Himalaya (Fig. 14), and Nakata (1989) inferred that the rock at the front of the Lesser Himalaya might be rising above the Indo-Gangetic Plains at 3-4 mm/a.

Some of the most rapid average vertical speeds of rock in the Himalayan region apply to small folds within the fold-and-thrust belt of the Siwalik sequence at the foot of the Range. Accurate dating of when anticlines grew, from local changes in rates of sedimentation, imply relative vertical speeds of about 1 mm/a or more for limbs of some folds (G. D. Johnson et al., 1979). The northern limb of the Soan syncline in the Potwar Plateau, Pakistan (Fig. 10), is exposed in a sequence of vertically dipping beds with ages from older than 3.5 Ma to about 2.1 Ma and is capped by flat-lying sequence younger than 1.9 Ma (Fig. 15) (Burbank & Raynolds, 1984; G. D. Johnson et al., 1986). In this 0.2-Ma interval, the Upper and Middle Siwalik sequence must have been folded and tilted, such that the base of the sequence rose 3000 m or more (at roughly 15 mm/a), and the tilted sequence was eroded flat (Burbank & Raynolds, 1984; G. D. Johnson et al., 1986). This type of uplift, where anticlines grow quickly in detached sedimentary layers appears to be common in the Indo-Gangetic plains (Deccaillau, 1986a, 1986b; Iwata & Nakata, 1985; G. D. Johnson et al., 1979; Nakata, 1972, 1989).

From fission-track ages and the magnetostratigraphy of Quaternary sediment in the Kashmir Basin, Burbank et al. (1986; Burbank & Johnson, 1983; Burbank & Raynolds, 1984) showed that a recent emersion of the Pir Panjal range, which separates the Kashmir Basin to its north from the Indo-Gangetic plain. Paleocurrents show that lower two-thirds of the Karewa sequence in the Kashmir Basin, largely fluvial and deltaic material deposited between about 4 and 1.8 Ma, were derived from the Himalaya to the northeast, but paleocurrents in sediment deposited between about 1.7 and 0.4 Ma indicate filling of the basin from its present margins, including a large part from the Pir Panjal to the southwest. Hence, they require that the Pir Panjal existed as a source of sediment by that time. Thus, an uplift of about 3500 m of the Pir Panjal above the plains to the southwest must have occurred since 1.7
Ma, corresponding to an average rate of about 2 mm/a. Burbank and Johnson (1983) inferred a minimum average rate of uplift of 4 mm/a of the Pir Panjal with respect to the Kashmir Basin from the difference in relative elevations of the Karewa series on the north slope of the Pir Panjal range and in the Kashmir Basin and from the age of 0.4 Ma of the youngest material of the Karewa series. This rate presupposes that all tilting occurred after the youngest material was deposited, but the clear evidence that the Pir Panjal already existed well before sedimentation in the Kashmir Basin stopped makes me doubt the rate of 4 mm/a. Nevertheless, it cannot be ruled out.

All of the observations cited in this section suggest that material comprising the southern front of the Lesser Himalaya is rising rapidly with respect to the Indo-Gangetic Plain (e.g., Nakata, 1989), but that much of the southern margin of the Lesser Himalaya is being
tilted down to the north, as if slip occurred on a northward dipping listric thrust fault (Figs. 14 and 15) (e.g., Delcaillau, 1986a, 1986b).

The Lesser Himalaya

The evidence reported above suggests relatively rapid vertical speeds of material both in the Greater Himalaya and at the southern margin of the Lesser Himalaya, but similar evidence from most of the Lesser Himalaya is absent. Vertical movement of Lesser Himalayan rock seems to be relatively slow, but the evidence that is most suggestive of such stability bears more on erosion than on uplift. A typical cross section of topography across the Himalaya suggests marked variations in both relief and denudation. Mean elevations rise rapidly from the plains at the southern edge of the range into the Lesser Himalaya and then again at the southern edge of the Greater Himalaya, but between them mean elevations over much of the Lesser Himalaya are remarkably constant. Moreover, the relief at the southern edges of both the Greater and the Lesser Himalaya indicates youthful stages of river incision, with steep, V-shaped canyons or even convex walls of gorges. The landscape of the Lesser Himalaya, however, seems to be more mature. Although canyons can be steep and relief large, in many parts the topography is not rugged, and in some areas it is very gentle. In addition, two quantitative observations suggest that erosion rates in the Lesser Himalaya are lower than those in the Greater Himalaya.

First, river terraces along rivers crossing from the Greater to the Lesser Himalaya suggest that the vertical velocity of material varies across the range. The heights of river terraces above the present stream bed increase northward across the Lesser Himalaya and seem to reach a maximum where the rivers cross the crest of the Greater Himalaya (Yamanaka & Iwata, 1982). Farther north, across the Greater Himalaya, differences in altitudes of terraces and streams decrease. If these terraces can be correlated reliably, then the variation in heights implies a maximum rate of incision where the rivers cross the crest of the Greater Himalaya, at least since the terraces formed. If the river profiles were in steady state, in the sense that the elevations of the river beds did not change with time, then the vertical velocity of the incised rock would reach a maximum where downcutting is most rapid. Clearly, this latter deduction requires a number of assumptions that are difficult to test, but the simplest interpretation of the rapid increase in stream gradient where the terraces are high above the stream is that the rock is rising most rapidly there (Iwata, 1987; Molnar, 1987c).

Second, fission track and $^{40}$Ar/$^{39}$Ar ages of minerals with different closure temperatures also imply variations in rates of denudation across the Himalaya. The youngest ages, and therefore the most rapid measured rates of denudation, are for samples from the Nanga Parbat area, where a suite of ages for different closure temperatures implies an accelerating denudation rate reaching about 4-5 mm/a since about 0.7 Ma (Zeitler, 1985, Zeitler et al., 1982a). Significantly older ages from samples in the Lesser Himalaya (Mehta, 1980; Saini et al., 1979; Sharma, 1984; Sharma et al., 1978) and north of Nanga Parbat (Zeitler et al., 1982b) require slower average rates of denudation ($\leq$ 1 mm/a) since 5 to 40 Ma. Thus, the rate of erosion does not seem to be constant across the range.

Whitehouse (1990) interpreted the variation in erosion rates across the Himalaya as indicating that in the Lesser, but not the Greater, Himalaya, erosion keeps pace with uplift of the rock. I suspect instead that both erosion rates and vertical speeds are greater in the Greater than Lesser Himalaya, but in both the Greater and Lesser Himalaya, the vertical speeds and the denudation rates are similar to one another.

Summary of uplift rates

There seem to be good reasons to believe that the vertical speed of rock (relative to the geoid) varies across the Himalaya, with maxima over the front of the Lesser Himalaya and over the crest of the Greater Himalaya, and with only low values over most of the Lesser Himalaya between them. Consequently, questions like “when did the Himalaya go up?” are too imprecise to allow a simple answer. Moreover, we cannot yet measure, or infer, uplift of the Earth’s surface in the Himalaya accurately, and therefore we cannot isolate quantitatively the part of the uplift of rocks that is due to deep-seated tectonic processes from that due to erosion and isostatic rebound. Nevertheless, the obvious variations in denudation rates, if treated as manifestations of spatially varying vertical material velocities, are easily interpreted in terms of ramp overthrusting of the Himalaya on a fault of variable dip (Molnar, 1987c). If the variations in uplift of rock across the Himalaya were due to the underthrusting of a relatively rigid Indian plate beneath the Himalaya, then the pattern of uplift would suggest that the underlying thrust fault consists of several listric splay beneath the southern margin of the Lesser Himalaya with the Indo-Gangetic Plain, a gently dipping segment beneath the lesser Himalaya, and a steeper dipping ramp beneath the Greater Himalaya (Fig. 16).

If this configuration does exist, then the curved profile of the fault should yield a simple relationship between the rate of slip on the thrust fault and the local dip of the fault. Let us, for the moment, ignore possible shortening within the Himalaya, and assume that the Indian plate plunges beneath the Himalaya at about 5°. The vertical speed ($V$) of the overriding material relative to the Indian plate should be given by

$$V = H \sin (\delta - 5^\circ)$$

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where $^{14}$H ($=10-20$ mm/a) is the rate of underthrusting beneath the Himalaya and $\delta$ is the dip of steeper segments of the fault (Fig. 16). For dips of the basal thrust fault of 250, $^{14}$V $= 4$-8 mm/a. This rate ignores isostatic compensation and therefore gives the rate of uplift relative to the Indian plate. The vertical speed relative to sea level should be much smaller, only 1-2 mm/a. It also ignores erosion, which could make the vertical speed of the rock relative to sea level much greater than the 1-2 mm/a given above. Finally, this calculated rate ignores internal deformation within the Himalaya except on vertical planes, and if deformation occurred on inclined planes, the vertical speed relative to the lower plate could be smaller, but such that the region in which uplift occurs is distributed over a broader area than otherwise. See Molnar (1987c) for a discussion of the interrelationships among $^{14}$V, $^{14}$H, $\delta$, and the assumed dip of the pervasive shearing in the hanging wall.

**Shortening within the Himalaya**

Varying rates of vertical velocities across the Himalaya require that the region be undergoing active deformation. For differences in vertical speeds of rock of 1 to 5 mm/a over distances of 50 to 100 km, average rates of tilting and of vertical shear strain would be at least $10^{-8}$/a and might reach $10^{-7}$/a. Where rates of compressional strain comparable to the rates of tilting or vertical shear strain, they would imply rates of shortening within, and across, the Himalaya of 1 to 5 mm/a (Molnar, 1987c). The extensive folding within the Lesser Himalaya implies that shortening has occurred, possibly totaling 50% (M. R. W. Johnson, 1987), but there are no useful constraints on when that shortening occurred. In any case, it seems logical to allow for the possibility that the rate of convergence of India with southern Tibet could be 5 mm/a more rapid than the rate of underthrusting of India beneath the southern edge of the Himalaya. Nakata (1989) inferred a comparable rate from Late Quaternary faulting. The rate of convergence between India and southern Tibet may best be described at $18 \pm 7$ mm/a (Molnar, 1987c), instead of $15 \pm 5$ mm/a, deduced for the underthrusting of the Indian Shield beneath the front of Himalaya (Lyon-Caen & Molnar, 1985).

**Summary**

Underthrusting of the Indian Shield beneath the Himalaya continues. Fault plane solutions of earthquakes, focal depths, and gravity anomalies imply that the Indian Shield is flexed down in front of the Himalaya and extends intact at least 80-100 km beneath the Lesser Himalaya. Thus, the Lesser Himalaya comprise a crystalline nappe, whose thickness reaches roughly 15 km, and which is overthrust onto the Indian Shield. The interface between the Indian Shield and the
overriding Himalaya may dip more steeply farther north beneath the Greater Himalaya than it does beneath the Lesser Himalaya. The abrupt increase in height from the Lesser to the Greater Himalaya appears to be a consequence of the more rapid vertical component of motion of rock above the steeper segments of the fault, than above the gentler segment farther south. Slip along the gently dipping segment beneath most of the Lesser Himalaya leads to only small vertical movement, and therefore to the relatively mature topography there. Vertical movement seems to be rapid also at the southern edge of the Lesser Himalaya; some river terraces are tilted up and dip north, implying that slip occurs on northward dipping listric thrust faults. Thus, the interface between the crystalline nappé of the Himalaya and the underlying Indian Shield appears to be curved, with steep segments beneath areas of marked increase in elevation and flatter beneath areas of more mature landscape.

The rate of convergence between the Indian Shield and the Himalaya is at least 10 mm/a. The average rate of overthrusting since 2 Ma is constrained to be about 10 mm/a at one major thrust fault, at the Salt Range (Baker et al., 1988), but additional shortening farther north within the Himalaya is likely. The age of sedimentary rock at the bottom of the Ganga Basin increases northward, and if the cross sectional shape of that basin, due to flexure of the Indian lithosphere, has not changed as the Himalaya has overridden it, then the sequence of ages imply a rate of convergence of 15 ± 5 mm/a between the Indian Shield and the southern edge of the Himalaya. Clearly, only a fraction of India’s convergence with Eurasia, at about 50 mm/a (DeMets et al., 1990), is absorbed by this underthrusting.

Great earthquakes apparently can be associated with slip on the gently dipping (δ < 10°) interface between the intact Indian Shield and the overriding crystalline nappé (Seeber & Armbruster, 1981). The historic record of seismicity in the Himalaya is too short to allow the average recurrence interval of great earthquakes to be constrained, except that it must exceed 100 years, and probably exceeds 200 years. I am aware of no evidence suggesting that the gross aspects of the tectonics vary along the Himalaya in such a way that underthrusting occurs in some segments without great earthquakes. Hence, I think that it would be foolish to assume that great earthquakes will not occur in segments of the chain where such earthquakes have not occurred in the last 200 years.

The seismic moments of great earthquakes in this century also are not well enough known that they can be used together with an estimate of the convergence rate to constrain recurrence intervals. For underthrusting at 10 to 25 mm/a, average displacements of 2.5 to 10 m associated with great earthquakes yield average recurrence intervals of 100 to 1000 years for them. Thus, the likely interval between earthquakes with M ≥ 8 is a few hundred years but cannot yet be made more precise (e.g., Molnar & Pandey, 1989).

The curved cross sectional shape of the main active thrust zone beneath the Himalaya is suggested (1) by the differences in the dips of the northward dipping nodal planes of moderate earthquakes in the Himalaya (Baranowski et al., 1984; Ni & Barazangi, 1984), (2) by the steepening of the gravity gradient and therefore of the dip of the Moho beneath the Greater Himalaya (Lyon-Caen & Molnar, 1983), and, most importantly, (3) by the apparent variations in vertical speeds of rock across the Himalaya. The apparently rapid uplift of the southern margins of the Greater and the Lesser Himalaya with respect to the Indo-Gangetic Plains, but the apparently negligible uplift of the rest of the Lesser Himalaya, suggests that the thrust fault beneath this region dips steeply beneath the rapidly rising belts and gently beneath the region between them (Fig. 16) (Lyon-Caen & Molnar, 1983; Molnar, 1984, 1987c).

A small fraction of India’s penetration into Eurasia seems to be absorbed by deformation within the crystalline nappé that forms the lesser Himalaya. The intense deformation of the rock there attests to shortening, but little evidence allows it to be dated. Nevertheless, vertical movements seem to vary markedly across the range, with high rates at the southern margin of the Lesser Himalaya and where the Greater Himalaya abruptly rises and lower rates across the Lesser Himalaya and north of the Greater Himalaya. This variation across the Himalaya implies tilts or strains of 10^-6/a to 10^-7/a on vertical planes within the Himalaya, and if compressive strains within the Himalaya occurred at comparable rates, as much as 5 mm/a of shortening across the range could be accomodated there (Molnar, 1987c). Thus, let us assume a rate of 18 ± 7 mm/a of convergence between India and southern Tibet.

A curved cross section of any major thrust fault is required by the finite displacement on it, and therefore by its formation as a ramp-overthrust. The steeper segment remains active while slip occurs, but as rock is thrust onto the Earth’s surface, which necessarily dips only very gently, this gentle segment must grow in width as the displacement increases. The existence of a curved fault requires that at least one of the upper or lower blocks deform during slip. Because the lower block of the Himalaya consists largely of intact lithosphere, it is likely to be strong. Hence, the curved cross-sectional shape of the main thrust zone beneath the Himalaya implies that the upper block, in this case the Himalaya itself, deforms. In some cases, poor exposure makes recognizing the deformation difficult, but it is especially clear in the Potwar Plateau of Pakistan (e.g., Baker et al., 1988; Lillie et al., 1987; Fennock et al., 1989) and in the seismicity within the
Himalaya (Armbruster et al., 1979; Gaur et al., 1986).

Such curved thrust faults (ramp-overthrusts) are common in orogenic belts, as is deformation of their hanging walls. Some of this deformation, however, is often referred to as "out-of-sequence-thrusting", which seems to be a misnomer, given that such deformation is a predictable consequence of deformation involving thrust faulting of large amplitude. Moreover, such deformation is required to maintain the cross sectional shape of an eroding and sliding nappe (Davis et al., 1983).

The direction of overthrusting of the Himalaya onto the Indian Shield, determined from fault plane solutions, is radially outward all along the chain. Because the Indian Shield behaves as an essentially rigid plate, this variation in direction requires east-west extension of southern Tibet at a rate of about 18 ± 9 mm/a (Molnar & Lyon-Caen, 1989), an inference corroborated well by studies of the Tibetan Plateau (Armijo et al., 1986).

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Reference


— 1987c. Inversion of profiles of uplift rates for the geometry of dip-slip faults at depth, with examples from the Alps and the Himalaya, Ann. Geophysicae, 5, 663-70.


Rastogi, B. K., 1974. Earthquake mechanisms and plate tectonics in the Himalayan region, Tectonophysics, 21, 47-56.


