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MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

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The tectonic evolution of the Tibetan Plateau

By John F. Dewey[1], F.R.S., Robert M. Shackleton[2], F.R.S., Chang Chengfa[3] and Sun Yiyin[3]

[1] *Department of Earth Sciences, Parks Road, Oxford OX1 3PR, U.K.*
[2] *The Croft Barn, Church Street, East Hendred, Oxfordshire OX12 8LA, U.K.*
[3] *Academia Sinica, Beijing, People's Republic of China*

The Tibetan Plateau, between the Kunlun Shan and the Himalayas, consists of terranes accreted successively to Eurasia. The northernmost, the Songban–Ganzi Terrane, was accreted to the Kunlun (Tarim–North China Terrane) along the Kunlun–Qinling Suture during the late Permian. The Qiangtang Terrane accreted to the Songban–Ganzi along the Jinsha Suture during the late Triassic or earliest Jurassic, the Lhasa Terrane to the Qiangtang along the Banggong Suture during the late Jurassic and, finally, Peninsular India to the Lhasa Terrane along the Zangbo Suture during the Middle Eocene. The Kunlun Shan, Qiangtang and Lhasa Terranes are all underlain by Precambrian continental crust at least a billion years old. The Qiangtang and Lhasa Terranes came from Gondwanaland. Substantial southward ophiolite obduction occurred across the Lhasa Terrane from the Banggong Suture in the late Jurassic and from the Zangbo Suture in the latest Cretaceous–earliest Palaeocene. Palaeomagnetic data suggest successive wide Palaeotethyan oceans during the late Palaeozoic and early Mesozoic and a Neotethys which was at least 6000 km wide during the mid-Cretaceous.

Thickening of the Tibetan crust to almost double the normal thickness occurred by northward-migrating north–south shortening and vertical stretching during the mid-Eocene to earliest Miocene indentation of Asia by India; Neogene strata are almost flat-lying and rest unconformably upon Palaeogene or older strata. Since the early Miocene, the northward motion of India has been accommodated principally by north–south shortening both north and south of Tibet. From early Pliocene to the Present, the Tibetan Plateau has risen by about two kilometres and has suffered east–west extension. Little, if any, of the India–Eurasia convergence has been accommodated by eastward lateral extrusion.

1. Introduction

This tectonic synthesis of the work done during and after the 1985 Royal Society–Academia Sinica Tibet Geotraverse has been written by Robert Shackleton (pre-Cretaceous tectonics) and John Dewey (Cretaceous–present). The views expressed are, unavoidably, to some extent theirs but the work on which the synthesis is based was done by all of the team. When discussing regions away from the Geotraverse, we have relied especially on a major synthesis of the tectonics of the whole Tibetan Plateau and beyond by Chang Chengfa and Sun Yiyin, which, it is hoped, will be published in full elsewhere.

It is now generally accepted that the Tibetan Plateau north of the Indus–Zangbo Suture consists of a series of microplates that were accreted to Asia before the India–Asia collision. The basis for this interpretation was the recognition by Chinese geologists (Chang & Zheng 1973; Chang & Pan 1981, 1984) of many ophiolites, some concentrated along zones which were

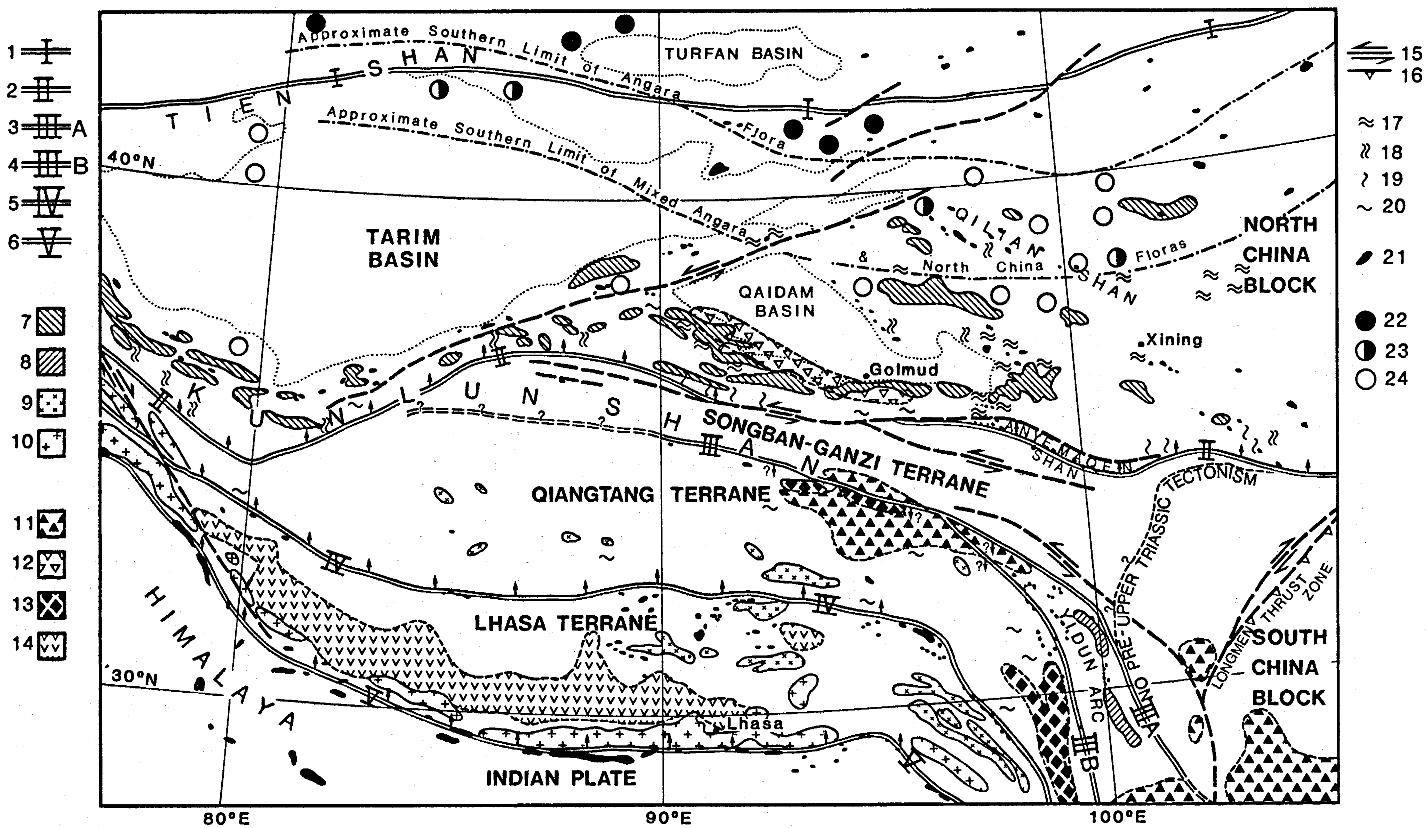

FIGURE 1. Tectonic map of Tibet, based on 1:1.5 M Geological Map, Qinghai–Xizang (Tibet) Plateau (1980); 1:1 M Geological Map, Qinghai Province (1981); 1:8 M Tectonic Map of Asia (1982); Watson *et al.* 1987; Zhang & He 1985 and Geotraverse data. 1: Tien Shan–Hegen Suture, 2: Kunlun–Qilian Suture, 3 & 4: (III) Jinsha Suture (3: Litang Suture, 4: Jinsha Suture, continued), 5: Banggong Suture, 6: Indus–Zangbo Suture, 7: Upper Palaeozoic plutons, 8: Triassic plutons, 9: Cretaceous plutons, 10: Late Cretaceous–Tertiary plutons, 11: Permian rift-related volcanic rocks, 12: Permian subduction-related volcanic rocks, 13: Triassic subduction-related volcanic rocks, 14: Tertiary subduction-related volcanic rocks, 15: strike-slip faults, 16: thrusts, 17: Lower Palaeozoic tectonism, 18: Carboniferous tectonism, 19: Permian tectonism, 20: Late Triassic (or early Jurassic) tectonism, 21: ophiolite, 22: Angaran flora, 23: Mixed Angaran and North Cathaysian flora, 24: North Cathaysian flora. Small arrows on sutures indicate inferred direction of subduction.

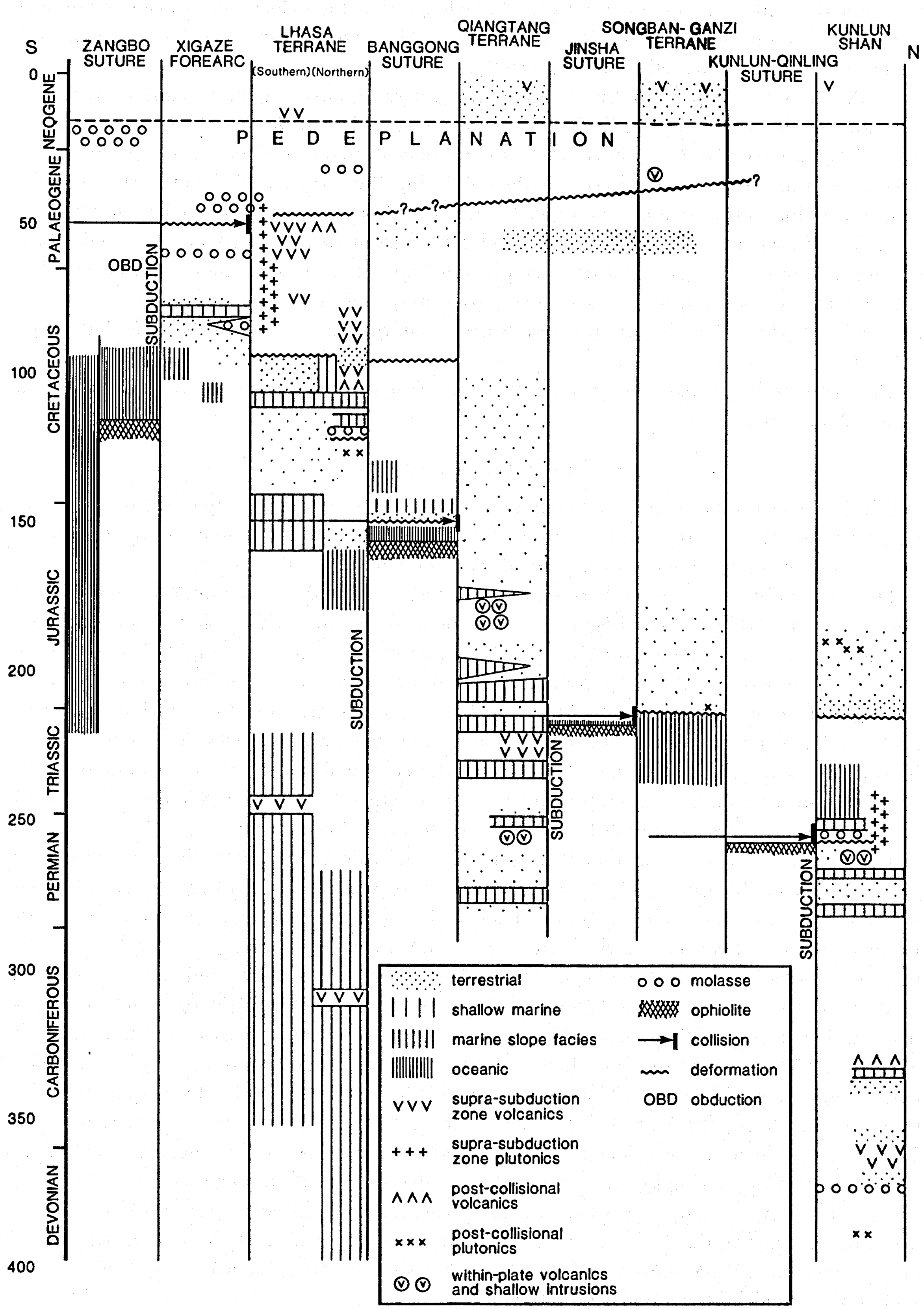

FIGURE 2. Summary of the tectonic chronology of Tibet.

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interpreted as sutures, combined with the knowledge that the deformations become younger southwards across the Plateau. One major aim of the 1985 Geotraverse was to define the history of microplate accretion more precisely.

In the previous papers in this volume, the Lhasa, Qiangtang and Kunlun Terranes, separated by the Banggong and Jinsha Sutures, are accepted as established. However, ophiolites are widely scattered across the Plateau, rather than concentrated in distinct linear zones (figure 1); few of them are dated; the criteria for distinguishing the terranes are inconclusive and the supposed ophiolites may not all represent fragments of oceanic crust. Therefore, because we are uncertain whether the ophiolites of the Lhasa Terrane are the klippen of a single sheet obducted from the Banggong Suture, our previous separation into terranes may not be valid. There may be more sutures; those recognized may not have been correctly located or extrapolated. Here, therefore, we do not assume *a priori* that the sutures and terranes have been defined.

The account is arranged chronologically. The supposed sequence of events is shown in figures 2 and 10.

2. Precambrian tectonics

Within the Geotraverse area, only one undoubtedly Precambrian group of rocks, the Amdo Gneisses, was seen. Gneisses in the southwest Nyainqentanglha Shan and a sedimentary series in the Kunlun Shan have been interpreted by some workers as Precambrian.

The gneisses in the Nyainqentanglha Shan include granitoid and migmatitic gneisses. The isotopic data (Harris, Xu, Lewis, Hawkesworth & Zhang, this volume) indicate the emplacement, approximately 50 Ma ago, of granitic melt, of anatectic origin, from a middle crustal source with an age of approximately 1000 Ma. The gneiss contains inherited zircons dating from about 1000 to 2000 Ma ago. These gneisses are not Precambrian. No clear evidence has been published that indicates that any Precambrian rocks are exposed in the Nyainqentanglha Shan. The age of the metasediments, which near to the granitoid gneiss contain staurolite, garnet, andalusite and sillimanite (Harris, Holland & Tindle, this volume) is uncertain but there is nothing to suggest that they are Precambrian.

The Amdo Gneisses comprise pelitic, psammitic and calcareous metasediments, amphibole gneisses and tonalitic gneiss. The latter has given a zircon age of 531 ± 14 (Xu *et al.* 1985) and Nd model ages of 1242 and 1646 Ma. These Nd model ages are taken to indicate a mid-Proterozoic crustal source (Harris, Xu, Lewis, Hawkesworth & Zhang, this volume). The intense D_1 foliation was isoclinally folded on N-striking axial surfaces and then refolded on E–W axes (Coward *et al.*, this volume). These structures are likely to represent continued deformation during cooling. Later biotite grade shear zones and thrusts are thought to be Mesozoic, perhaps at the time of sphene growth, 171 ± 6 Ma ago (Xu *et al.* 1985) or ophiolite obduction (*ca.* end-Jurassic). The whole assemblage was intruded by the Amdo bimodal granite suite dated 130 ± 10 Ma (Xu *et al.* 1985). The anatexis of tonalitic magma from Proterozoic crust and the associated sequence of intense deformations suggest a collisional environment. Considering the time involved, from the start of collision to anatexis in thickened crust and rise of the melt, it seems clear that at least the metasediments in the Amdo Gneisses are late Precambrian. The assemblage invites comparison with the Pan-African of Gondwanaland whence the Lhasa Terrane came, as well as with the array of granites in the Himalayas dated between 550 and 450 Ma.

The sequence in the Kunlun Shan, attributed to the late Precambrian and Cambro-Sinian, comprises marbles, pelites, greywackes and possibly basic volcanics. Stromatolites thought to indicate a Precambrian age have been found in the marbles at several localities (Yin *et al.*, this volume). The extent and precise age of these rocks is uncertain: they include some that are almost certainly Ordovician because, near the top, they include highly characteristic mass flow deposits with limestone clasts (Leeder *et al.*, this volume), and some that are thought to be Permian (Smith & Xu, this volume). Whether or not some of these rocks are Precambrian, there is no evidence that they were affected by any Precambrian tectonism. We conclude that the only exposed rocks in the Geotraverse area that show Precambrian tectonism or metamorphism are the Amdo Gneisses.

As well as the evidence from exposures, there is important isotopic evidence concerning the age of the deeper crust under the Plateau. Inherited zircons in the Nyainqentanglha granite gneiss give upper intercept ages of *ca.* 1000 and 2000 Ma (Xu *et al.* 1985) and Nd model ages also indicate the presence of Precambrian (1000 Ma or older) crust beneath the Lhasa and Kunlun Terranes. Evidence from the Qiangtang Terrane is not available but there can be little doubt that it too is underlain by Proterozoic or older crust. The Nd model ages show significant variations across the Plateau. Some of these differences are attributed to fractionation processes but those from the Amdo Gneisses in the north of the Lhasa Terrane (1242 and 1646 Ma) and from the Golmud Granite in the north of the Kunlun Shan (source model age 1100–1250 Ma) are significantly higher than the others, which are about 1000 Ma (Harris, Xu, Lewis, Hawkesworth & Zhang, this volume). From this evidence, it can be concluded that the whole of the Plateau is underlain at depth by Precambrian crust at least 1000 Ma old. The Amdo Gneisses in the Lhasa Terrane show exposed evidence of an end-Proterozoic ('Pan-African') collision. Neither the extent nor trend of this end-Proterozoic domain is known.

There is no evidence to show the existence of Archaean rocks under the Plateau: an Archaean component in the Carboniferous glaciomarine diamictite in the Lhasa Terrane is presumed to have come from the Indian shield. From the palaeolontological data, it seems that the Lhasa and Qiangtang Terranes were formerly attached, with India, to Gondwanaland, but the Precambrian exposures in Tibet are too limited to suggest any pattern of connections with the Precambrian of India.

3. Lower Palaeozoic tectonism

About 20 km south of Golmud, the basal conglomerate of the Upper Devonian rests unconformably on late Precambrian or Cambro–Sinian beds which include sericite schists and marbles (Zhang, appendix to Coward *et al.* this volume). The Devonian and Carboniferous sediments and volcanics of that area are folded and commonly dip steeply but they are not cleaved and are only metamorphosed near contacts with plutons. Farther south between Naij Tal and the Xidatan, the supposed Cambro-Sinian sequence (Yin *et al.*, this volume) appears to pass up continuously into rocks identified as Lower Ordovician by the presence of distinctive mass-flow deposits with limestone clasts (Leeder *et al.*, this volume). These relationships indicate a post-Ordovician, pre-Upper Devonian, cleavage-producing deformation and low-grade metamorphism. No Silurian is known in the part of the Kunlun Shan reached during the Geotraverse so the timing is imprecise. Another indication of such a deformation event is that the Ordovician rocks of the central Kunlun consistently show high finite strains. In the mass-flow deposits, the limestone clasts usually show $X:Z = \sim 3:1, X$ (the extension direction) plunging steeply down dip on the cleavage. Fold axes often plunge 70° or more either eastwards or westwards, indicating sheath folds such as are not seen in the Upper Palaeozoic or Triassic rocks of the area. In the Triassic rocks of the middle Kunlun, finite strains are low except in narrow shear zones, and in the northern Kunlun, Triassic dykes are undeformed. Fragments of limestone, probably Lower Palaeozoic (as the largest clasts, up to 5 m long, certainly are) show cleavages in different directions in different clasts in only weakly deformed Triassic boulder conglomerates, where rotation during deformation is unlikely (M. Leeder pers. comm. 1985; RMS, loc. S551). Two cleavages at an acute angle can be seen in some of the Ordovician rocks of the middle Kunlun Shan. These various observations, taken together, show that in the parts of the central and northern Kunlun studied during the Geotraverse, there was a phase of strong deformation in the interval between the Ordovician and the Upper Devonian. Because the attitude of the main cleavage in the Palaeozoic and Triassic in those areas is essentially the same, the pre-Devonian and post-Triassic cleavage and folds must have been virtually coaxial. Southwards through the Kunlun Shan, the later, post-Triassic deformation becomes increasingly intense and the early Palaeozoic deformation, if present, has not been recognised.

Some of the magmatic activity in the northern part of the Kunlun Shan may be related to pre-Upper Devonian collisional tectonism; the Qinbazhan granite has yielded an early Devonian age of 389±10 Ma (C. L. Lewis, pers. comm. 1988) and the supposedly Upper Devonian (?) volcanics are thought, from their chemistry, possibly to indicate an active continental margin while those of supposedly early Carboniferous age are post-collisional (Pearce & Mei, this volume). Such a relationship of this magmatism to the post-Ordovician tectonism suggests that the tectonism occurred not long before the unconformable Upper Devonian was deposited. The volcanism occurred along a belt extending far to the ESE (figure 1). Evidence of mid-Palaeozoic tectonism extends over a large area (figure 1) from the West Kunlun to the Qilian Shan and beyond. Its southern limit is not well defined but is either at the Kunlun–Qilian Suture or slightly further north.

4. Late Carboniferous or early Permian tectonism

No indication of late Carboniferous or early Permian tectonism was seen on the Geotraverse and the Tienshan Suture, of late Carboniferous or early Permian age (Watson *et al.* 1987) lies far to the north of the area covered (figure 1). Nevertheless a collision between a terrane which included the northern Kunlun and a plate to the north seems to be necessitated by the palaeomagnetic data and by the evidence on the geological maps (1:1.5 M Qinghai and 1:1 M Tibet) of widespread deformation at this time (figure 1). The palaeomagnetic data came from the red beds of the Dagangou Formation, which is either late Devonian (Yin *et al.*, this volume) or early Carboniferous (Smith & Xu, this volume). The data are interpreted as showing a palaeolatitude of $-23° \pm 16$ (Lin & Watts, this volume).

The palaeomagnetic evidence from the Tibetan Plateau is presented in a diagrammatic form in figure 3. This is intended to show that the data can be interpreted in terms of the successive accretion to Eurasia of a series of terranes, some certainly derived from the south. The positions suggested for the terranes are drawn schematically, with the control of the palaeomagnetic

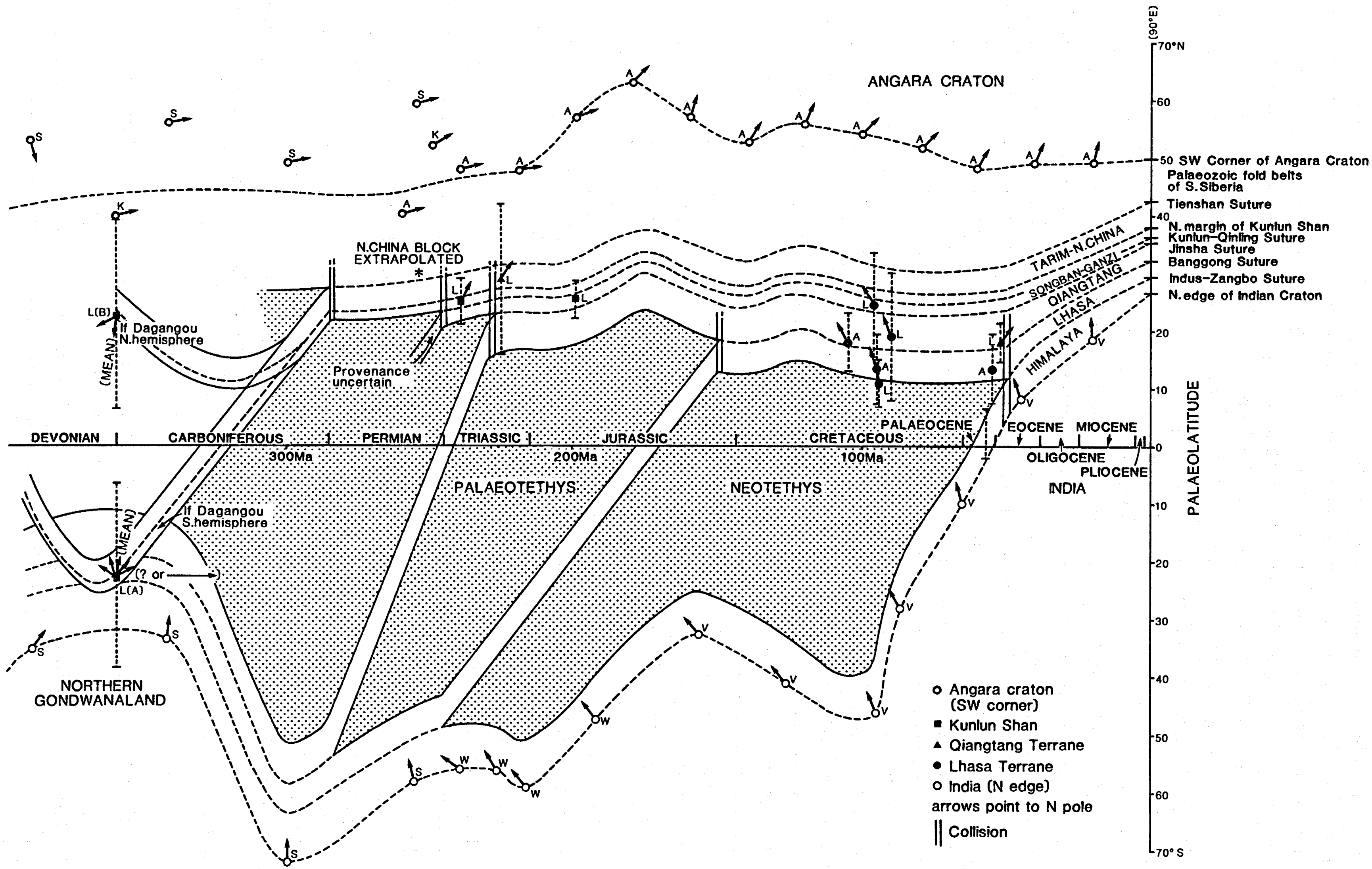

FIGURE 3. Diagrammatic interpretation of changes of latitude of Tibet terranes (now at 90° E) relative to India and the Angara craton. Based on Lin & Watts, this volume (L); Acache *et al.* 1984 (A); Khramov *et al.* 1981 (K); Scotese *et al.* 1979 (S); F. J. Vine (pers. comm. 1983) (V); and Westphal *et al.* 1983 (W). For explanation see text.

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data, so as to follow the assumed path of the Eurasian plate when they were supposedly attached to it. Note, however, that only latitudinal differences can be shown. Because of possible rotations, these would not necessarily remain constant as indicated; nor need apparent convergences mean real convergences; two terranes may converge to the same latitude but be far apart longitudinally. Note too that the diagram shows the supposed palaeolatitudes of points that are now on 90°E meridian. There is not reason to suppose that these points preserved this longitudinal relation through time.

Although the inclinations from two sites on the Dagangou Formation are in close agreement, the declinations differ widely. Values from opposite limbs of a major fold, 245° and 125°, are reduced to 215° and 168° (all means) by correcting for a plunge of 70° to the west (Lin & Watts, this volume). However, with a Northern Hemisphere position (B on figure 3) any of these declinations would imply a very large rotation of this part of the Kunlun Shan before the Permian (see figure 3). This is the main reason for interpreting the results as indicating a southern latitude (see Lin & Watts, this volume). If the southern hemisphere position (A on figure 3) for the Dagangou Formation is accepted, that block must have moved 5000 km northwards during the next 100 Ma, with no great rotation. A northern latitude interpretation (B on figure 3) would imply little change in latitude in the same interval but about 180° rotation. Either interpretation would imply motion quite different from the Asian Plate to the north and therefore a suture between the two. Two problems arise, firstly the position of the suture which must bound the Terrane containing the Dagangou Formation to the north and secondly the previous attachment of this part of the Kunlun Shan, whether to Asia, to Gondwanaland or to neither.

There are two possible positions for the suture. It may be the late Carboniferous or early Permian Tianshan Suture, in which case this part of the Kunlun Shan formed part of the Tarim and North China Terrane, or it may be a suture which is suggested by a line of small basic masses, mapped as late Palaeozoic (1:1 M Geological Map, Qinghai; Geotraverse geological map, this volume), through the Qimantage Shan and extending at least 130 km ESE of Golmud. This line is about 70 km north of the Anyemaqen Shan ophiolite zone (see below) from which it is clearly distinct (cf. Zhang *et al.* 1984). The evidence obtained on the Geotraverse does not help to decide between these alternatives but, on balance, it seems more probable that the whole region between the Kunlun–Qinling Suture to the south (see below) and the Tienshan Suture to the north formed a single terrane, the Tarim–North China Terrane (itself possibly composite) during the late Palaeozoic.

The attachment, earlier in the Palaeozoic, of the Kunlun area and the Tarim Terrane, seems likely to have been to the Asian Plate, primarily because there are Lower Palaeozoic ophiolite belts, tectonism and magmatism (figure 1) that seem to be continuous, allowing for some displacement on the Altyn Tagh fault, with similar ophiolite belts, tectonism and magmatism within the Asian Plate, where they encircle the Angara nucleus. It therefore seems that, if the Kunlun block was in the Southern Hemisphere in the early Carboniferous, it had previously been much farther north and part of the Asian plate.

5. End-Permian–early Triassic collision: the Southern Kunlun ophiolite belt

From late Devonian to late Permian times, shallow-marine and terrestrial sedimentation proceeded, though how continuously is uncertain, across the northern part (north of the

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Middle Kunlun Fault; figure 1) of the area covered during the Geotraverse. Some of the Lower Carboniferous beds are regarded as molassic, derived from mountains to the north (Leeder *et al.*, this volume).

Two different cycles of magmatic activity can be recognized in the northern Kunlun Shan: the earlier, late Devonian and early Carboniferous, perhaps 370–320 Ma ago, is regarded here as the last phase of the pre-mid-Devonian tectonism (see above). The later cycle is represented by the North Kunlun batholithic intrusions, dated at 260–240 Ma (Harris, Xu, Lewis, Hawkesworth & Zhang, this volume), and the closely related NNW-trending basic dyke swarm which transects the plutons. The chemistry of the plutonic rocks implies an active continental margin; the only slightly younger dykes are regarded as post-collisional (Harris, Xu, Lewis & Jin, this volume). If this is correct, this major phase of magmatism in the northern Kunlun Shan cannot be related to the end-Triassic or earliest Jurassic (Indo–Sinian) collisional tectonism which affected a vast region to the south (see below), but rather to an end-Permian or early Triassic tectonism.

Late Palaeozoic deformation is clearly demonstrated by the folding, strong enough to produce local northward overturning (loc. S549) of the lower Carboniferous sediments and volcanics, though not strong enough to produce a cleavage, before the injection of the undeformed NNW-trending basic dyke swarm, which cuts through the folded rocks, and before the intrusion of the batholith which hornfelsed already vertical beds. Debris from the batholith and from now eroded volcanics is seen in the massive conglomerates at or near the base of the Triassic sequence to the south (loc. S551 etc).

These magmatic and tectonic events appear to be related to the most impressive ophiolite belt on the Tibetan Plateau apart from that on the Indus–Zangbo Suture, namely the Anyemaqen Shan belt. This strikingly linear belt, along which more than a hundred ophiolite bodies have been mapped, extends for about 400 km through the Anyemaqen Shan (figure 1). It reaches to within about 300 km east of the Geotraverse line, where it intersects, and is truncated by, the eastward continuation of the Xidatan fault. Associated with the ophiolite belt is glaucophane schist as well as melanges, one of which has Lower Permian exotics in a Permian flysch matrix; another has Permian limestone exotics in a Triassic matrix. Clearly, this zone represents an important suture, which is known as the Kunlun–Qinling Suture (Watson *et al.* 1987). These ophiolites have not been dated but from published maps (1:1 M Geological Map, Qinghai) they appear generally to occur within a continuous band of Permian rocks, although locally in older or younger ones. It seems likely that they are late Permian or early Triassic.

The band of Permian rocks continues about 300 km farther west than the ophiolites, almost to the Geotraverse line, before it too is truncated by the Xidatan fault. On the north side of the fault the ophiolite belt is not clearly identifiable on the published maps. The total displacement on this fault is not known. Kidd & Molnar (this volume) demonstrate a Quaternary left-lateral slip of 30 km and suggest a total offset of *ca.* 75 km, based on the tentative recognition on satellite imagery of the phyllonite belt that has been mapped just south of the Xidatan fault. There is no indication of any extension of the ophiolite zone north of the supposed phyllonite.

In spite of its apparent failure to continue, the suture cannot have stopped at the Xidatan fault. Even if the fault dates back to the Permian, it is at too acute an angle to the suture to be taken for a transform fault, considering that all the indications of transport direction, from stretching lineations, are almost north–south. It is possible that the western continuation of the suture is represented by the ophiolite (age unknown) immediately north of Ulugh Muztagh (Molnar *et al.* 1987*b*). Alternatively (Watson *et al.* 1987), the continuation of the suture may be towards the WNW through the Qimantagh Shan, along the southwest side of the Qaidam Basin, where several ophiolites of late Palaeozoic age are indicated on the maps (1:1 M Geological Map, Qinghai, 1:1.5 M Geological Map, Tibet). This seems improbable because it would take the suture obliquely across an apparently continuous belt of plutons and across the belt of Devonian volcanics, and also because of evidence both of Lower Palaeozoic and Carboniferous tectonism well to the south of the Qimantagh ophiolites (figure 1).

Indications of Lower Palaeozoic tectonism continue SW along the NW side of the Altyn Tagh and between the Karakoram Fault and the Tarim Basin, where unconformities indicate significant tectonism during the Carboniferous. Because these pre-Permian tectonisms, which affected extensive regions to the north, cannot have affected the terrane south of the suture before it arrived there, the suture must continue south of the Kunlun Shan, approximately as shown in figure 1. If the position of the Kunlun–Qinling Suture has been correctly identified, it is likely that the subduction zone dipped northwards because Upper Palaeozoic I-type plutons lie to the north of the suture.

The late Permian Kunlun–Qinling Suture separates areas with the Northern Cathaysian flora from those with the Southern Cathaysian flora (Watson *et al.* 1987; figure 1). A line can be tentatively drawn to represent the southern limit of mixed Angaran and Northern Cathaysian floras, and about 200 km still farther north, a line at the southern limit of the Angaran floras. The fact that in the late Permian (*ca.* 225 Ma ago) these three zones appear to have been gradational, but separate from the Southern Cathaysian flora of the South China and Qiangtang Terranes (Smith & Xu, this volume) supports the view that the Kunlun–Qinling Suture marks the southern limit of the composite Eurasian plate, before the accretion of the Terrane to the south.

Recognition of the Kunlun–Qinling Suture makes it necessary to separate by name the terranes to the north and south of it, as has previously been done by Chinese geologists (Li *et al.* 1979) but not previously by the Geotraverse team (Chang *et al.* 1986; previous papers in this volume). The Terranes will be referred to as the Tarim–North China (probably composite) and Songban–Ganzi Terranes, north and south of the Kunlun–Qinling Suture (figure 1) respectively.

In the part of the Songban–Ganzi Terrane seen on the Geotraverse, only Triassic and younger rocks are exposed. The Trias is represented by an extensive flysch wedge (Bayan Har Group) presumed to have been transported from the north (Leeder *et al.*, this volume). However, farther east, older rocks appear where maps (1:1 M Geological Map, Qinghai, 1:1.5 M Geological Map, Tibet) show that there is an extensive area west of the Longmen Shan thrust belt which is taken to mark the edge of the South China Block, where there is no sign of any significant deformation in the Palaeozoic, nor below the Upper Trias. It must be suspected that the same is true father west, where there is no exposed evidence. On this view, the Songban–Ganzi Terrane is an extension of the South China block.

From the middle Kunlun Fault to just south of Wudaoliang, a distance of a little over 100 km across strike, the Triassic rocks are strongly deformed. This deformation took place at the end of the Triassic: in the Kunlun Shan, Middle Triassic rocks, and father south, Upper Triassic (Bayan Har Group) are affected by the main deformation, while a post-tectonic tonalite intrusion is dated at 213±6 Ma (Harris, Xu, Lewis, Hawkesworth & Zhang, this volume), which corresponds to the Triassic–Jurassic boundary. Farther east, and southeast, the region affected by the end-Triassic tectonism is much wider (figure 1).

The southern limit of this strong end-Triassic deformation cannot be seen in the Geotraverse area because of the cover of Neogene and Palaeogene deposits in the Fenghuo Shan. An inlier, which we were not able to reach, northwest of Erdaogou exposes an ophiolite (associated with Triassic sediments), which is taken to mark the approximate position of the Jinsha Suture to which the end-Triassic deformation is related. The area was subsequently mapped by a group led by Colin Stark (pers. comm. 1988). Their work showed that the ophiolite occurs as numerous separate angular masses, in a matrix of Triassic sandstones. The ophiolites appear to be in a melange. South of the Fenghuo Shan, the Triassic rocks which reappear from under the Tertiary cover are quite different from those to the north of it. The end-Triassic deformation is much weaker, the sedimentary facies is different and there are volcanic rocks which are not seen north of the suture. The suture clearly separates two different domains. It is taken to represent a Palaeotethyan ocean, which before collision separated the Eurasian Plate from the Qiangtang Terrane. Indications of weak end-Triassic folding can be recognized from published maps right across the Qiangtang Terrane.

The recognition of the Jinsha Suture zone is based primarily on the occurrence of ophiolites along it. In the region of the Geotraverse these are few and far apart – one 60 km NW of Erdaogou, another 200 km farther WNW, others 200 and 400 km ESE. However, from Yushu southeastwards, there are very many ophiolites. They occur (figure 1), along either side of, and to some extent within, a wedge shaped area known as the Yidun Island Arc Belt. The concentration of ophiolites along the northeast side of the wedge marks the Yushu–Litang Suture zone. Here, a nearly uninterrupted ophiolitic melange with greenschist facies metamorphism has been traced for about a thousand kilometres through Yushu, Garze and Litang to Wuli. At least one fairly complete, though dismembered, ophiolite, is recognized, at Zimenda, with harzburgite, gabbro, pillow lavas and cherts. Exotic blocks include Permian, Carboniferous and even Silurian limestones and radiolarites and basalts. The matrix of this melange consists of Triassic clastics and flysch; a rich microfauna in siliceous rocks overlying the basic lavas is early Triassic. The ophiolites are regarded as late Permian or early Triassic: if so their position is difficult to explain. It would seem to imply that the Yidun Island Arc was accreted to Eurasia in the late Permian or early Triassic, but this cannot be so because the arc, which is thought to be composite, consists of late Triassic arc volcanics and I-type plutons and flysch. The underlying sequence extends down at least to the Ordovician (Yin *et al.*, this volume) suggesting that it was built on continental crust.

The ophiolites along the southwest side of the Yidun Arc define the Jinsha River Suture Zone. This is best seen in a belt about 40 km wide extending more than 700 km SSE. Along the western side of the belt is an ophiolitic melange with Devonian, Carboniferous and Permian limestone exotics in an Upper Triassic matrix. A convincing ophiolite, as well as many ultramafic diapirs, is recognized. Because island arc rocks occur between the two supposed sutures, it has been suggested that Benioff zones, above which the volcanics were erupted, dipped inwards under the arc from both sides. An apparent continuation of the Jinsha River Suture Zone is seen, beyond a gap, in the Ailao Mountains-Tongtian River fracture zone where phyllonites and mylonites are associated with many ultramafic bodies in an ophiolite melange, which is covered unconformably by late Triassic red beds. Blueschists to the southwest and high-T metamorphism to the northeast suggest a paired metamorphic belt and subduction northeastwards under the Yangzi Block.

This summary of the relationship along the Jinsha River Suture zone far to the southeast of the Geotraverse area is included here because the evidence obtained along the Geotraverse does not show clearly whether Palaeotethys 1 was subducted northwards (Coward *et al.*, this volume), southwards (Pearce & Mei, this volume) or both (Leeder *et al.*, this volume). The argument from structures is based on the recognition, in the southern Kunlun Shan and south of it, of steep north-side up faults and a phyllonite zone with the same displacement sense. It is proposed (Coward *et al.*, this volume) that these structures were initiated as gently dipping thrusts and have been rotated up northwards to their present steep attitude in a forearc prism.

Towards the southern margin of the Triassic flysch wedge, north of Wudaoliang, the early folds are refolded but appear to have been recumbent, and the structures face southwards. The associated cleavage is folded but often dips south. Farther east in Qinghai, the foliation dips about 40 °N or NW in the Bayan Har Group (observations by Chinese members of the team, after the Geotraverse).

The magmatic evidence is that arc-type volcanics occur south of the Jinsha Suture (Pearce & Mei, this volume). They are thought to represent the edge of the Upper Triassic Baitang Group, although not so mapped on the 1:1 M Geological map of Qinghai. They continue along the SW side of the suture for over 1000 km to the southeast (figure 1).

It seems likely that the exotics in the melanges along the sutures southeast of the traverse came from the southwest, where such rocks are known, rather than from the huge area of Bayan Har Group muds and turbidites to the northeast. This suggests that ophiolites were being obducted along the southwest side of the suture zone.

On balance it seems more likely, despite the structural evidence, that subduction on the Jinsha Suture Zone was southwards. If so, the strong deformation was in the downgoing plate and the weaker in the overriding one, as is commonly the case.

6. End Jurassic (Yenshanian) tectonism

The southern limit of the Qiangtang Terrane is marked by the sudden appearance of ophiolites (figure 1). They are not, however, restricted to a narrow linear zone, but are distributed across strike for about 150 km. Those in the north, in the area studied, dip northwards and show a shear-sense towards the SSE (Kidd *et al.* this volume). Especially between Dongqiao and Gyanco, the ophiolite forms a flat thrust sheet. It is thought that the ophiolite was obducted southwards and later re-imbricated (Coward *et al.*, this volume). For these reasons, the suture zone is thought to be along the northern limit of the array of ophiolites. The projection of the suture westwards from just north of Siling Co, to the ophiolites south of Banggong Lake, is based on the occurrence of one ophiolite mapped north of Dong Co., about midway in a stretch of 800 km, and on map indications of late- or post-Triassic deformation nearly as far south as the supposed line (figure 1). The age of the ophiolites and their obduction, in the area studied, is well-established as late Jurassic (Girardeau *et al.* 1984; Smith & Xu, this volume). From the structural evidence, it seems clear that subduction was northwards.

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During the Jurassic, much of the Qiangtang Terrane in the areas studied to the north of the suture was a coastal plain over which molassic continental clastics were spread from the north and northeast; while farther south, nearer the oceanic area represented in the Banggong Suture zone, shallow-marine carbonates were deposited (Leeder *et al.*, this volume). In sharp contrast, over much of the Lhasa Terrane south of the Banggong Suture, muds and turbidites, the 'Lake Area Flysch' (Yin *et al.*, this volume) were deposited. The oceanic sediments, seen overlying the ophiolite, include radiolarites. Island arc magmatism was not recognised on either side of the Banggong Suture.

The Banggong Suture is not marked by a zone of strong deformation. To the north, in the Qiangtang Terrane, mapping (1:1 M Geological Map, Qinghai) indicates that widespread moderate folding of the Jurassic beds took place before the deposition of the unconformably overlying Palaeogene. This folding is presumably associated with the collision along the Banggong Suture. South of the suture in the Lhasa Terrane, there seems to be no clear evidence of significant deformation at the time of the collision. It can only be concluded that pre-collisional subduction was so slow, and post-collisional convergence so slight, that no magmatism and very little deformation was produced.

7. Cenozoic tectonism: the India–Asia collision

Since Argand's (1924) recognition that the Cenozoic tectonics of Asia are principally the result of the convergence of Gondwanan continental fragments with Laurasia, there has been general acceptance of the view that the approximate doubling of crustal thickness, uplift, and roughly north–south shortening of the Himalayas and Tibet have been caused by the collision of the Indian subcontinent with the bulk of Asia along the Indus–Zangbo Suture following the subduction of a substantial ocean, the Tethys (Allègre *et al.* 1984; Carey 1955; Dewey & Bird 1970; Dewey & Burke 1973; Holmes 1965; Molnar & Tapponnier 1975), or rather its Mesozoic Neotethyan tract (Sengör 1979).

The Tibetan Plateau, part of a broader zone of deformation from the Himalayan thrust front to Lake Baikal (figure 4), is a remarkably level plateau, with a dissected average elevation just below 5 km, of approximately 7×10^5 km^2, an area about equal to the US Cordillera and Rockies. Relief is somewhat subdued, the plateau consisting mainly of rolling hill terrain with about 1.0 km of relief, local plateaux between 5 and 6 km and local mountain ranges such as the Tanggula Shan rising to above 7 km with permanent snow and ice. Localized consequent drainage systems feed mostly into larger meandering rivers that drain into lakes whose long term evaporation rates exceed fluvial input in the dry Tibetan climate in the precipitation shadow of the Himalayas. Only a few larger rivers meander across the plateau either to plunge off the eastern edge of the plateau through spectacular gorges or to cut through the Himalayas as antecedent rivers. Therefore, the Tibetan landscape is one of low ablative relief as a consequence of which there has been little Tertiary denudation except in the marginal thrust belts. Most of Tibet, except the Gangdese and Kunlun belts, consists of high-level sequences. The plateau, containing 70% of the Chinese permafrost, is about one fifth karstic probably formed at a time when Tibet enjoyed a wetter climate at a lower elevation.

There is general agreement that the Tibetan Plateau is underlain by a continental crust between 60 and 85 km thick with a small free-air gravity anomaly that implies near isostatic equilibrium. Love and Rayleigh wave velocities are much lower beneath Tibet than beneath

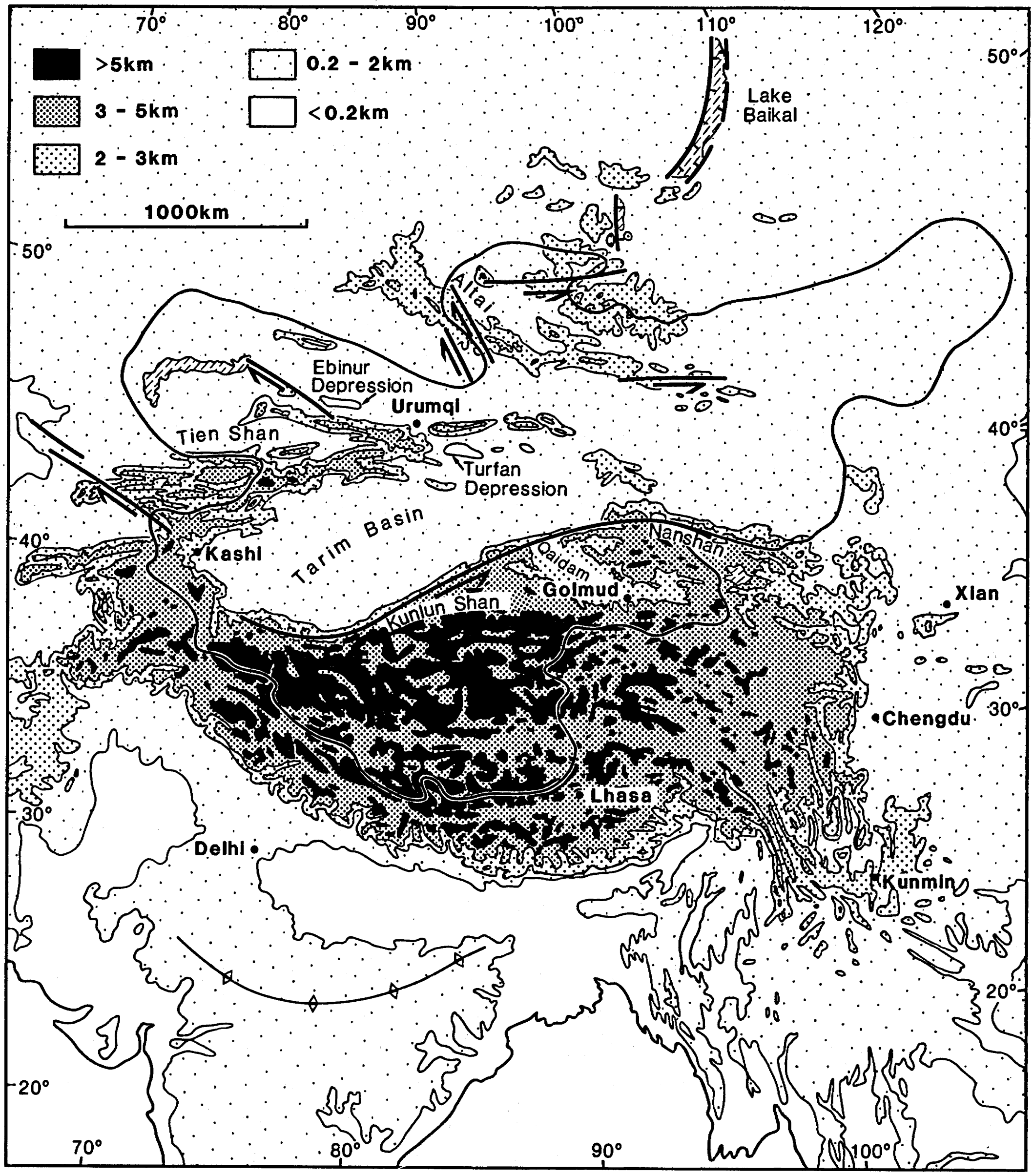

FIGURE 4. Simplified topographic map of the India–Eurasia collision zone. Lakes indicated by oblique ornament; area of internal drainage enclosed by continuous line; line with anticline symbol indicates crest of flexural bulge.

peninsular India and somewhat lower than north of Tibet implying a thinner lithospheric boundary conduction layer and a higher geothermal gradient beneath Tibet (Romanowicz 1982). Furthermore, there is a Magsat crustal negative anomaly field over Tibet (Achache *et al.* 1987), which, in a region where the crust is thick but magnetically thin, suggests a high geothermal gradient. A generalized crust/mantle profile from peninsular India across the Himalayas and the Tibetan Plateau to the Qaidam Basin and the Nan Shan (figure 5) is derived from the data and arguments of Barazangi & Ni (1982), Bird & Toksoz (1975, 1977), Brandon & Romanowicz (1986), Chen & Molnar (1975, 1981), Choudhury (1975), Chun &

McEvilly (1986), Chun & Yoshii (1977), Gupta & Narain (1967), Hirn *et al.* (1984*a*, *b*, *c*), Jobert *et al.* (1985), Lyon-Caen (1986), Lyon-Caen & Molnar (1983, 1984, 1985), Matthews & Hirn (1984), Min & Wu (1987), Molnar *et al.* (1987*b*), Ni & Barazangi (1983), Pines *et al.* (1980), Romanowicz (1982), Ruzaikeri *et al.* (1977), Shaw & Orcutt (1984), Teng (1980), Teng *et al.* (1981, 1983) and Zhou *et al.* (1981). Data on the seismicity of the plateau is from Chen & Molnar (1983), Chen *et al.* (1981), Molnar & Chen (1978), Molnar & Deng (1984) and Ni & Barazangi (1984). Data and views on the heatflow, geothermal structure and recent magmatism of the plateau are from Chen & Molnar (1981), Coulon *et al.* (1986), Deng (1978), Dewey & Burke (1973), Francheteau *et al.* (1984), Hennig (1915), Jaupart *et al.* (1985), Kidd (1975) and Van *et al.* (1986).

The crust thickens from about 33 km beneath the Indian Shield to a maximum of about 75 km beneath the Himalayas and a regional average of 65 km beneath Tibet, with about 20 km denuded from the Himalayas, about 10 km from the Gangdise Belt in Southern Tibet and an average of about 2 km from the Tibetan Plateau between the Gangdise Belt and the Kunlun. The crust and lithospheric mantle are multilayered in six zones (figure 5). A low velocity veneer about 5 km thick, probably consisting mainly of supracrustals (1) is underlain by a higher velocity layer (2) to about 17 km. Crustal earthquakes are concentrated in this layer mostly between 5 and 10 km with normal and strike-slip first motion solutions. A crustal low velocity zone (3) between about 17 km and 30 km with a conductivity anomaly and low Poisson's Ratio may be a partial melt zone or, less likely, a zone of high pore pressure. A higher velocity lower crust (4) between 30 km and a Moho at about 65 km show steep shear wave velocity gradients. The upper mantle has a high velocity lid (5) to about 100 km with two earthquake hypocentres at about 85 km giving normal fault solutions beneath which the lower lithosphere (6) is completed by a 4.4 shear wave velocity (Vs) layer to 150 km. Therefore, the crust of Tibet is about twice the thickness of the Indian crust, whereas the Tibetan lithosphere is about three fifths the thickness of the Indian shield lithosphere. A Tibetan lithosphere of about 150 km is consistent with a convex-up geothermal gradient of about 27 °C km^{-1} in the uppermost crust and 18 °C km^{-1} in the lower crust with a Moho temperature of about 750 °C. The widespread recent volcanism across the plateau supports the idea of a thinner boundary conduction layer beneath Tibet.

There has been less agreement about the timing, mechanism and rates of crustal thickening and uplift during the collisional process although the timing is generally believed to have been during the latest Cretaceous to late Eocene interval with a recent rapid uplift. Three broad classes of hypotheses have been suggested to account for the Tibetan Plateau in its regional Asian tectonic setting, namely crustal underthrusting, crustal shortening and thickening, and lateral eastward crustal extrusion.

Argand (1924) proposed that the Tibetan Plateau is underlain by a double normal thickness crust produced by the underthrusting of the northern margin of peninsular India or 'Greater India' beneath the Asian crust (figure 6A), a view supported by Barazangi & Ni (1982), Holmes (1965), Powell & Conaghan (1973, 1975) and Ni & Barazangi (1984). Bird (1978) introduced the variant of lithospheric delamination and underthrusting. Powell (1986) suggested wholescale subduction of the Indian continental crust followed by its buoyant uprise to underplate the Asian Tibetan continental crust. Cohen & Morgan (1987) and Zhao & Morgan (1987) suggested the further variant of the injection of the Indian crust into a softer Asian Tibetan lower crust.

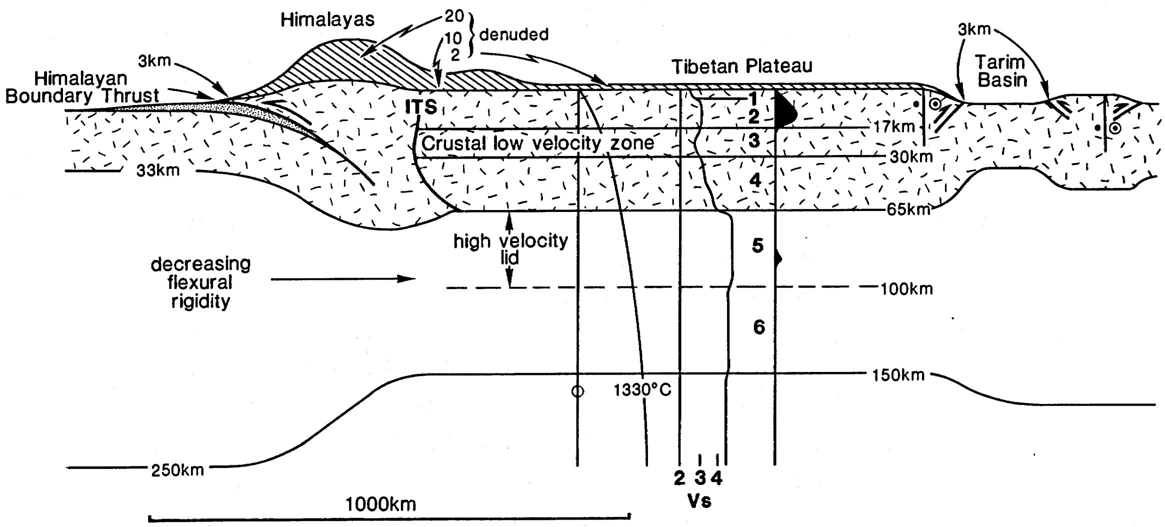

FIGURE 5. Simplified and idealised cross section of the Himalayas, Tibetan Plateau and Tarim Basin showing crustal and lithospheric thickness variations. Vs, shear wave velocity.

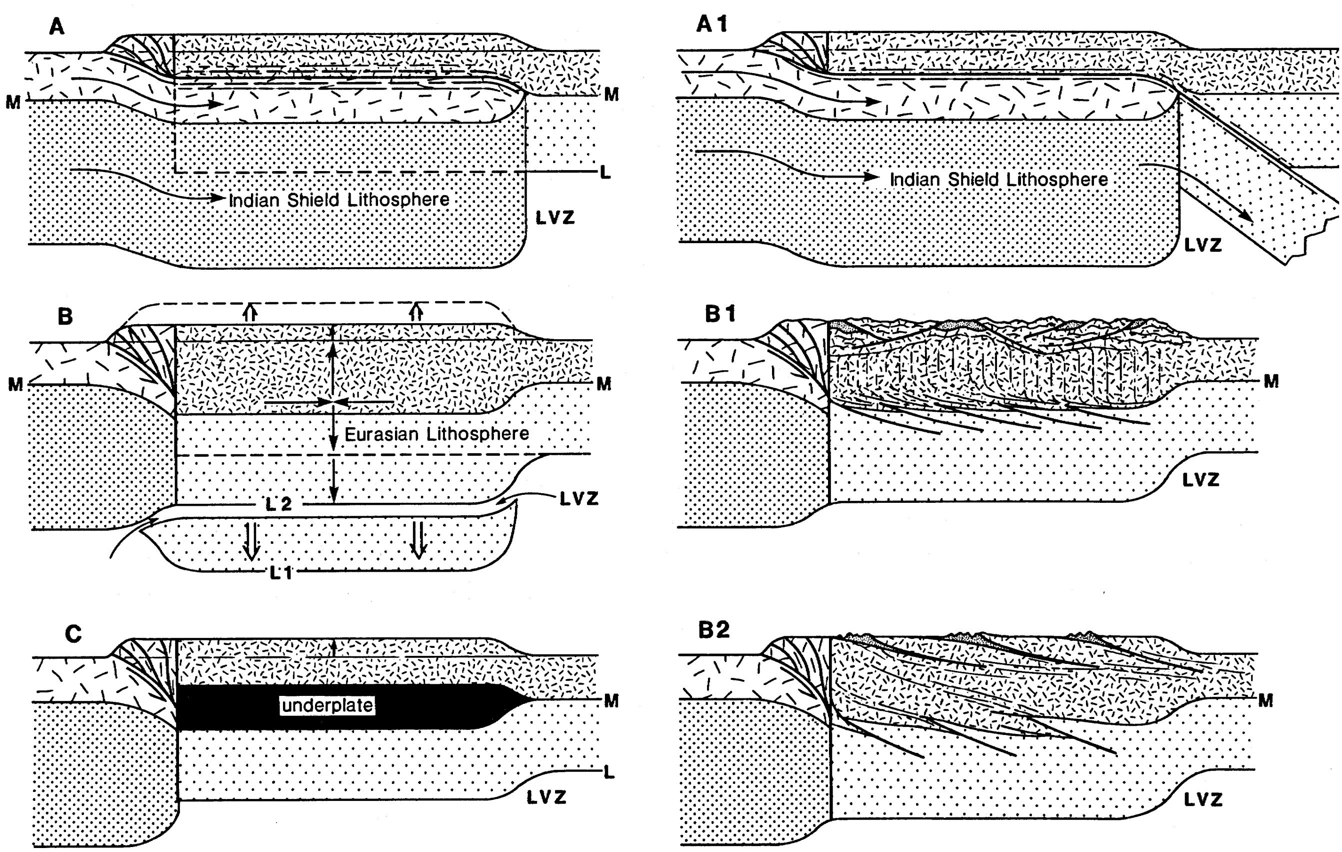

FIGURE 6. Mechanics for crustal and lithospheric thickening and thinning, and uplift. A: Indian lithosphere underthrusts Eurasia following delamination of the Eurasian lithospheric mantle along the Moho. B: Eurasian lithosphere thickens by vertical plane strain north-south shortening. B1 & B2: Structural expressions of B. C: Tibetan crust thickens and rises by magmatic underplating. M – Moho, L – base of thermal lithosphere, L1 – base of thermal lithosphere at end of thickening phase, L2 – base of thermal lithosphere after delamination, LVZ – low velocity zone.

Dewey & Bird (1970) and Dewey & Burke (1973) argued that the thick Tibetan crust was formed by horizontal shortening and vertical stretching of the Asian crust (figure 6B) in advance of the 'bull-dozing' Indian subcontinent. The shortening/thickening hypothesis was further supported by Bird & Toksoz (1975, 1977), Bird, Toksoz & Sleep (1975), Chen & Molnar (1981), Vilotte, Daignières & Madariaga (1982) and Vilotte *et al.* (1984, 1986). The quantitative modelling work of England & McKenzie (1982, 1983), England & Houseman (1985, 1986), Houseman & England (1986), Houseman, McKenzie & Molnar (1981) and England & Houseman (1988) has led to an integrated model of viscous vertical plane strain shortening and thickening of the Asian lithosphere progressively spreading northwards into Asia, with strong inhomogeneities like the Tarim Basin that resist thickening. In this model, the Tibetan Plateau is uplifted by two mechanisms, slow uplift caused by crustal/lithospheric thickening followed by rapid uplift caused by catastrophic lithospheric advective thinning (England & Houseman 1988).

The alternative mode of horizontal plane strain of the Tibetan lithosphere to accommodate the northward motion of India was suggested and developed by Molnar & Tapponnier (1975, 1977), Tapponnier & Molnar (1976, 1977) and Tapponnier *et al.* (1982). This involves the eastward lateral extrusion of Tibet and parts of Asia to the north along an array of giant transforms such as the Kunlun and Altyn Tagh Faults.

There are sufficient consistent palaeomagnetic data to help to constrain our choice of Tibetan models (Achache, Courtillot & Besse 1983; Achache, Courtillot & Zhou 1984; Allègre *et al.* 1984; Besse *et al.* 1984; Bingham & Klootwijk 1980; Klootwijk 1979; Klootwijk, Conaghan & Powell 1985; Klootwijk & Pierce 1979; Klootwijk *et al.* 1979; Pozzi *et al.* 1982; Westphal *et al.* 1983, Lin & Watts, this volume). The palaeomagnetic data are supported by relative motion solutions based upon magnetic anomaly and fracture zone studies in the Atlantic and Indian Oceans (McKenzie & Sclater 1971; Patriat & Achache 1984; S. Cande and W. C. Pitman, pers. comm). Since the late Cretaceous (84 Ma), India has moved northwards with respect to stable Eurasia by some 52° (5720 km) and rotated counterclockwise by about 35°. Since the time of the India–Asia collision (about 45 Ma, argued below) India has moved northwards with respect to 'stable' Eurasia by about 22° (2420 km) and rotated counterclockwise by 21°. Between 84 Ma and 45 Ma, the northward motion of about 3300 km was absorbed principally by northward subduction beneath the Gangdise arc of Neotethyan ocean between India and the Lhasa Block. Since 45 Ma, the northward motion has been absorbed along an intracontinental convergent boundary north of the Himalayan Boundary Fault. The southern edge of the Lhasa Terrane has moved northwards by 18° and rotated counterclockwise by up to 30° during the last 45 Ma and, therefore, little of the northward motion of India can have been absorbed by Greater India underthrusting Tibet. Lin & Watts (this volume) can detect no Tertiary latitudinal separational change between Lhasa and Erdaogou from palaeomagnetic data; however, the errors are large and the present separation is 500 km and the maximum Eocene separation would have been only about 1000 km according to the shortening/thickening model. The 4° (440 km) difference between the northward motion of India and the southern Lhasa Terrane since collision is probably taken up by shortening in the Himalayas. The involvement of basement in the thrust sheets and the preservation of Mesozoic continental margin facies in the Himalayas, south of the Indus–Zangbo Suture suggest, also, that little of the Indian crust has vanished beneath Tibet. Arguments based on balanced sections for the cover stripped from the basement in the outer

Himalayan zones of Pakistan (Coward & Butler 1985) indicate a northward subduction of the Indian crust by about 470 km, very close to the 440 km, indicated from palaeomagnetic and oceanic data. This is likely to have been taken up by crustal thickening in the Himalayas.

Therefore, about 1980 km of intracontinental convergence has been absorbed by the Asian lithosphere north of the Indus–Zangbo Suture. The principal mechanism cannot have been eastward lateral extrusion from Tibet by horizontal plane strain for two main reasons. First, extrusion cannot account for the crustal thickening in Tibet and in areas such as the Tien Shan (Nelson, McCaffrey & Molnar 1987) and the Nan Shan to the north. Second, the high plateau of Tibet is largely immediately north of and opposite the Indian 'indenter'. Only a maximum of about 15% of the plateau can be considered to lie east of the indented margin and, therefore, had extrusion been a major factor, it would have predated crustal thickening and uplift.

Therefore, because most of the northward motion of India relative to Eurasia cannot be accounted for by underthrusting of Greater India or by eastward lateral extrusion from Tibet, vertical plane strain during north–south shortening seems the likely dominant mechanism in that it accommodates the displacement and accounts for crustal thickening. Furthermore, the Tibetan Plateau is only a part, albeit an important and spectacular part of a zone of north–south compressional deformation (figure 4) that extends northwards to Lake Baikal (Molnar & Deng 1984; Molnar & Chen 1978). Clearly the Indian crust cannot underthrust as far as Lake Baikal and it seems unlikely, therefore, that wholly different mechanics thickened the crust of Tibet and the areas north of Tibet. Within this zone of deformation, rigid regions such as the Tarim and possibly the Qaidam Basins, the former probably underlain by Precambrian lithosphere, resisted deformation and it is possible that major strike-slip zones, such as the Altyn Tagh and Kunlun Faults, result not from extrusion *per se*, but from jostling between hard lithospheric zones and from compatibility problems resulting from adjacent deforming and non-deforming zones. If this were the case, the Altyn Tagh and Kunlun Faults would show an increasing eastwards displacement; existing data are insufficient to test this hypothesis.

Lastly, the shortening and thickening of the Eurasian lithosphere by an Indian indenter is supported by the regional form and deformation of Tethyan sutures between India and Eurasia (figure 7). India moved northwards relative to Eurasia between two giant transforms that linked the Indus–Zangbo subduction zone with contemporaneous subduction zones in the Gulf of Oman–Zagros to the west and a zone to the east through the East Indies. If India had advanced mainly by underthrusting Tibet, the Tethyan sutures would have today approximately their pre-collisional form. It is difficult to accept that the suture offsets that coincide with the western and eastern transform boundaries of the Indian protrusion are pre-collisional and that the Indian protrusion slipped precisely into a pre-determined embayment in Eurasia. Furthermore, the underthrust model would imply that displacements on the two bounding transform zones should drop abruptly at each end of the Indus–Zangbo suture; according to the shortening/thickening model the displacements should diminish northwards gradually and terminate on line with the northward limit of Eurasian shortening. On the western edge of India, the Owen Fracture Zone is continued on land as the Chamaun and associated faults, which form a 1500 km sinistral transpressional zone linking the Indus–Zangbo suture with the Gulf of Oman. Allowing for post-45 Ma subduction–accretion in the Makhran and some crustal shortening west of the Chamaun Zone, the offset of the Indus–Zsangbo Suture is about 1300 km, that of the Kunlun–Qinling Suture about 700, and that of the

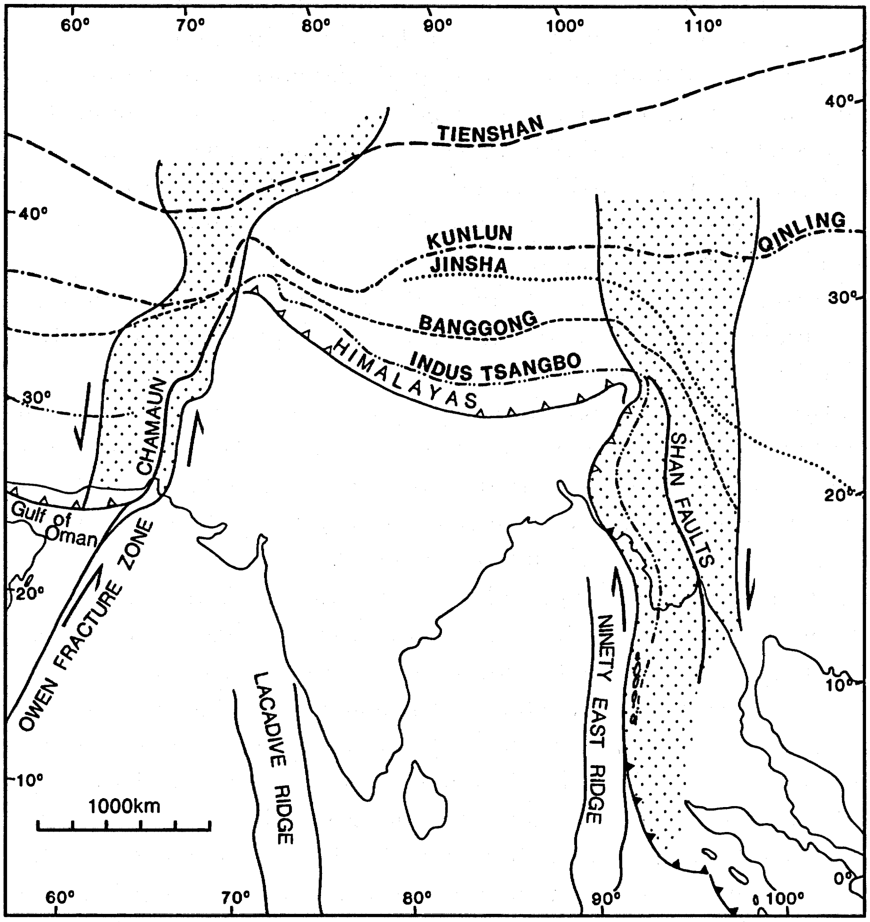

FIGURE 7. Map showing the distribution of the principal sutures from the Indus–Zangbo to the Tien Shan and their deformation by the indentation of Eurasia by India. The dotted ornament indicates the zones of northward-decreasing shear strain that bound the Indian indenter.

Tienshan a maximum of about 500 km. Therefore, we view the western dotted zone of figure 7 as representing a left-lateral transpressional zone of distributed shear bounding the Indian indenter. Similarly, the eastern margin of the Indian indenter is a 400 km wide right-lateral transform complex consisting of the Andaman and Shan Faults approximately on line with the Ninety East Ridge, originally a right-lateral transform between the Indian and Australian Plates. Across this dextral zone of transpressional shear, the Indus–Zangbo Suture is displaced at least 2000 km, the Jinsha Suture by about 1000 km and the Kunlun–Qinling Suture by perhaps 400 km, whereas the Tienshan Suture appears not to be displaced. Thus the eastern dotted zone in figure 7 is believed to represent a zone of distributed right-lateral shear that forms the eastern boundary of the Indian indenter.

The underthrusting and vertical plane strain mechanisms of crustal thickening should produce quite different structural geometries in the Tibetan crust. Underthrusting, in its simplest form (figure 6A), would involve little or no north–south shortening of the upper Tibetan crust and supracrustals, but would generate substantial simple shear deformation of the thrusted interface region of the Indian and Tibetan crust to develop widespread horizontal or sub-horizontal fabrics at the base of the Asian and top of the Indian crust. In contrast, vertical plane strain by north–south horizontal shortening and bulk vertical stretching must involve the Tibetan crust and supracrustals in substantial north–south shortening by thrusting and folding; doubling of crustal thickness would involve 50% north–south shortening. This model cannot predict, simply, the way in which the shortening is likely to occur, whether fairly homogeneous (figure 6, B1) or partitioned in upper crustal thrust sheets (figure 6, B2), but it is unlikely that simple homogeneous vertical plane strain would generate the widespread horizontal shear fabrics predicted by the underthrust model.

Many deeply eroded Precambrian terranes of high-grade metamorphic rocks, probably developed in collisional tectonic environments, show widespread flat or gently dipping axial surfaces and fabrics which would seemingly support the underthrust model for the origin of these terranes. However, Moho imbrication and thrust segmentation of the brittle upper crust either at a fine (figure 6, B1) or a coarse (figure 6, B2) scale is likely to generate gently dipping fabrics in adjacent crustal rocks. This raises, however, the problem of the great contrast in structural style between the Himalayas and Tibet, at least in the style observed at present erosion levels. The Himalayas have suffered some 80% shortening and are dominated by south-verging, gently northward dipping thrusts. By contrast, Tertiary structures in Tibet are thrust ramp basins containing folded red beds in a crust that has suffered a maximum 50% shortening. We suggest that the style contrast results from the presence or absence of pre-collisional lithospheric/crustal asymmetry. The Himalayas are, basically, a pile of basement-cored thrust sheets stacked on the northward-thinning crust of India above a northward thinning lithosphere; hence the thrust asymmetry was predetermined by the asymmetrical structure of the northern edge of India. In contrast, Tibet has resulted from the bulk homogeneous shortening of the Asian lithosphere, in which the thrust polarity within the brittle upper crust varies, although it is commonly to the south (Coward *et al.*, this volume).

Finally, the role of the lithospheric mantle cannot be ignored in discriminating among models. In general, convergence must thicken the lithosphere as it thickens the crust, whereas most underthrust solutions to crustal thickening avoid the problem of the Asian lithospheric mantle as India underthrusts. To slide the Indian crust neatly beneath the Tibetan–Asian crust involves the prior disposal or delamination of the Tibetan lithospheric mantle cleanly along the Moho (figure 6A) or a thrust ramp/flat model with a mega-flat along the Moho (figure 6A1) if a simple crustal doubling is to be achieved by this mechanism. Moreover, as argued above, it is clear that a thick underthrust Indian shield lithosphere does not underlie Tibet. If the Tibetan crust has been thickened by vertical plane strain, the underlying mantle lithosphere must have been similarly thickened. If, for example, the Tibetan lithosphere, prior to north–south shortening, was 125 km thick, it should now be 250 km following a 50% shortening. Simple thermal recovery to its present 150 km would take far too long and, hence, some form of rapid delamination or advective thinning seems to have been likely (England & Houseman 1988; Houseman, McKenzie & Molnar 1981). Rapid lithospheric thinning by delamination or stoping could also account for the rapid recent uplift of the Tibetan Plateau, its widespread recent volcanism and hot springs and its east–west extension (England & Houseman 1988). The deep earthquakes seen beneath the Tethyan collisional system in Rumania, southern Iberia, Iran and the Hindu Kush may be in sinking delaminated mantle lithosphere fragments. Many orogenic belts show a post-shortening morphotectonic (Hills 1956) phase of late uplift, synchronous with bimodal basaltic/silicic magmatism and extensional tectonics, which could result from catastrophic thinning of a lithosphere thickened by shortening. Magmatism, particularly during advective lithospheric thinning, could contribute to crustal thickening and uplift by gabbroic underplating (Cox 1980; McKenzie 1984; Hildreth & Moorbath, in press) in the lower crust (figure 6C). It is remarkable that post-collisional intracontinental shortening is taken up principally by the Asian lithosphere north of the Indus–Zangbo Suture, excluding the strong Tarim and Qaidam enclaves; the Indian shield shows little post-Archaean deformation, except foreland basin flexure, south of the Himalayan Boundary Thrust. We suggest that this is the result of the collisional impingement of a strong Indian shield lithosphere, which has undergone secular thickening since the Archaean, with a complex collage of accreted Asian fragments with a thinner lithosphere north of the Indus–Zangbo Suture. Thus, the Indian lithosphere may be thought of as an indenting buttress with a thinner northern edge generated by Neotethyan Triassic/Jurassic rifting, which collapsed to form the Himalayan Zone of shortening. A thinner lithosphere along the northern edge of India is suggested by the gravity anomaly/flexural studies of Lyon-Caen & Molnar (1983, 1985) and Molnar & Lyon-Caen (in press) that show a northward-decreasing flexural rigidity of the Indian Shield beneath the Himalayas.

Present deformation of the Tibetan crust (figure 8) is principally by strike-slip and extension, seen in both seismic first motion solutions (Chen *et al.* 1981; Molnar & Chen 1983; Molnar & Deng 1984; Tapponnier & Molnar 1977) and in surface displacements (Armijo *et al.* 1982, 1986; Chang *et al.* 1986; Molnar *et al.* 1987*a*, *b*; Molnar & Tapponnier 1978; Ni & York 1978; Norin 1946; Shackleton 1981; Tapponnier *et al.* 1981*a*, *b*; Tapponnier & Molnar 1977). A widespread Miocene erosion surface is very gently warped and faulted across the width of the plateau (Shackleton & Chang, this volume). Roughly north–south grabens effect an approximately east–west extension, across steeply dipping normal faults. Along the eastern flank of the Nyainqentanglha, an inactive gently south-east dipping extensional detachment draws down early Tertiary volcanics across an older metamorphic and granitic basement. The uplift of the Nyainqentanglha may have resulted from footwall uplift during this phase of extensional detachment. Young north–south normal faulting continues south into the Himalayas but appears to be confined to elevations above 3 km. The grabens commonly control the positions and shape of lakes and are the sites of extensive hydrothermal activity.

Strike-slip faulting occurs on dextral NW-striking faults and on sinistral ENE-striking faults. The two sets appear to comprise a conjugate wrench regime (Rothery & Drury 1984) but with the bisector of the obtuse angle being the shortening direction. To the south in the Himalayas, a conjugate wrench regime consists of dextral NNW-striking faults and sinistral NE-striking faults (Dasgupta, Mukhopadhyay & Nandy 1987), a geometry in which the acute angle is bisected by the shortening direction as predicted if the faults developed on planes of maximum shear stress (Anderson 1948). The transition from the Tibetan to the Himalayan wrench geometry occurs roughly along the Yarlung–Zangbo suture, but it is not clear whether the transition is effected by a zone of intersection or by a gradual change of strike of individual faults. The strike-slip faults have, commonly, pull apart and, less commonly, transpressive segments. Several terminate at the ends of the north–south grabens, indicating that these at least have evolved synchronously with the graben.

There is little sign of active or recent thrusting in the Tibetan Plateau. Southwest of Amdo, Cretaceous or Palaeogene red sandstones and conglomerates are thrust southeastwards over Pleistocene–Holocene brown and yellow sandstones and conglomerates with frost-heave

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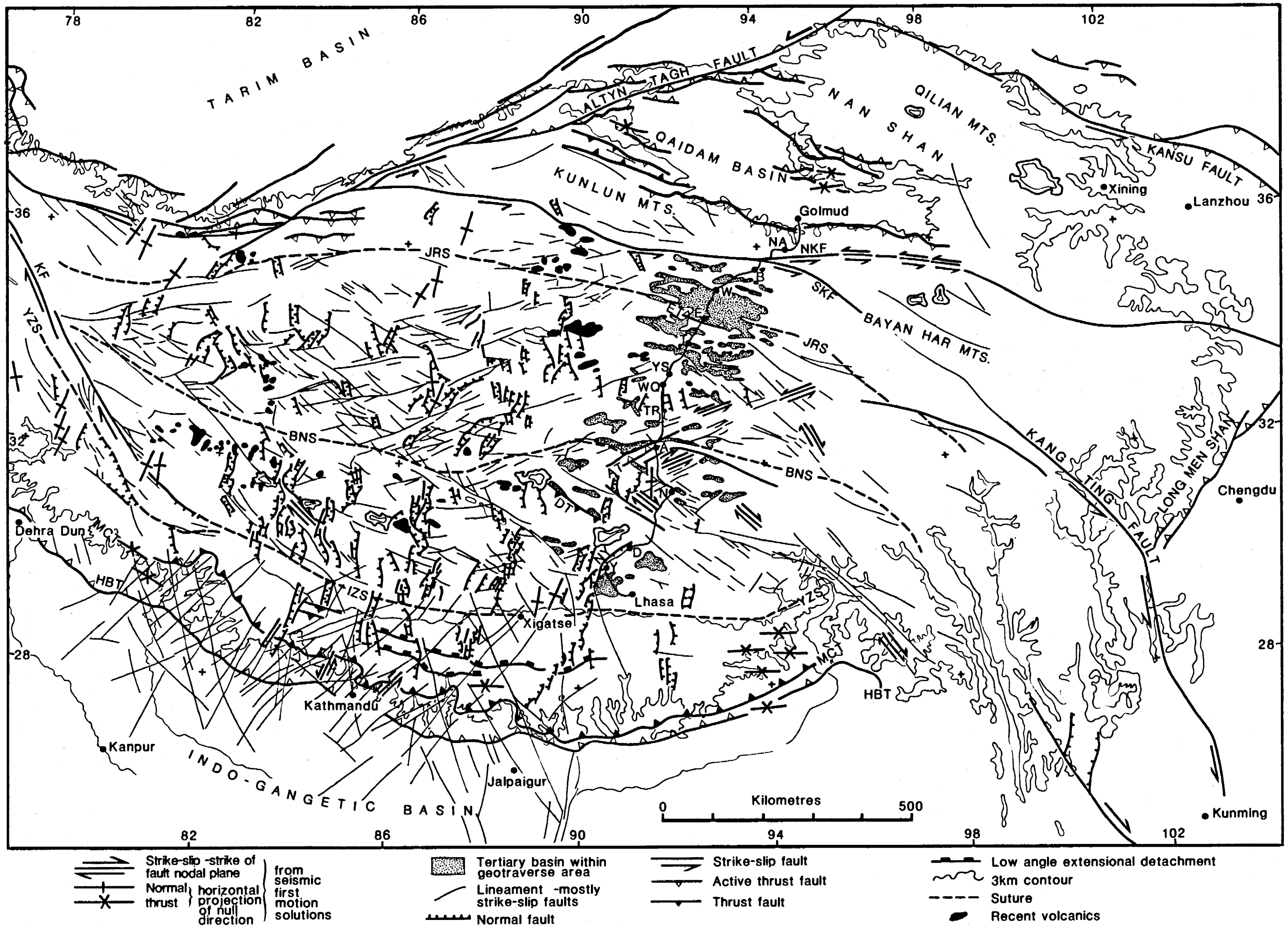

FIGURE 8. Map showing the principal neotectonic features of the Tibetan Plateau and adjacent regions. Himalayan low-angle extensional detachment zones from Burchfiel & Royden (1985). A – Amdo, AT – Amdo Thrust, B – Budongquan, BNS – Banggong-Nujiang Suture, D – Damxung, DT – Duba Thrust, E – Erdaogou, ET – Erdaogou Thrust, HBT – Himalayan Boundary Thrust, JRS – Jinsha River Suture, KF – Karakoram Fault, MCT – Main Central Thrust, N – Nagqu, NKF – North Kunlun Fault, NY – Nyainqentanglha, SKF – South Kunlun Fault, T – Tuo Tuo River, TP – Tanggula Pass, W – Wudaoliang, WQ – Wenquan, Y – Yangbajain, YS – Yanshiping, IZS – Indus–Zangbo Suture.

structures; the frost-heave structures are deformed below the thrust (Kidd & Molnar, this volume). South of Erdaogou, Palaeogene red sandstones are thrust to the south across Neogene lake beds. In both cases, young local relative uplift is indicated by antecedent drainage and incised meanders. These examples of thrusting do not appear to be characteristic of regional relationships and both instances of thrusting may be on transpressive segments of strike-slip faults. On the thrust plane, south of Erdaogou, striae pitch east at about 20° in the thrust surface indicating a dominant left-lateral strike-slip component on the thrust fault. Therefore, if north–south shortening is occurring today within the Tibetan crust, it is accommodated by strike-slip faulting, not by thrusting; this is supported by the dominant strike-slip first motion solutions. East-trending, thrust-bounded, basins involve Palaeogene red beds and appear to be of pre-Miocene age. Elevation, therefore, seems to be a critical factor in controlling tectonic style. Thrusting occurs at or just below 3 km, strike-slip faulting occurs at all elevations, whereas east–west extension occurs above 3 km. The way in which the present/Recent motion of India relative to Asia is partitioned into displacement and strain on, and to the north of, the Himalayan Boundary Thrust, has been examined by Molnar *et al.* (1987*a*). A modified version of their principal conclusions is shown in figure 9. A recent finite difference study of the Atlantic and Indian Oceans by S. Cande and W. C. Pitman III, pers. comm. has shown that a point on the Indian Plate just south of Kathmandu has moved northwards by about 2° since the time

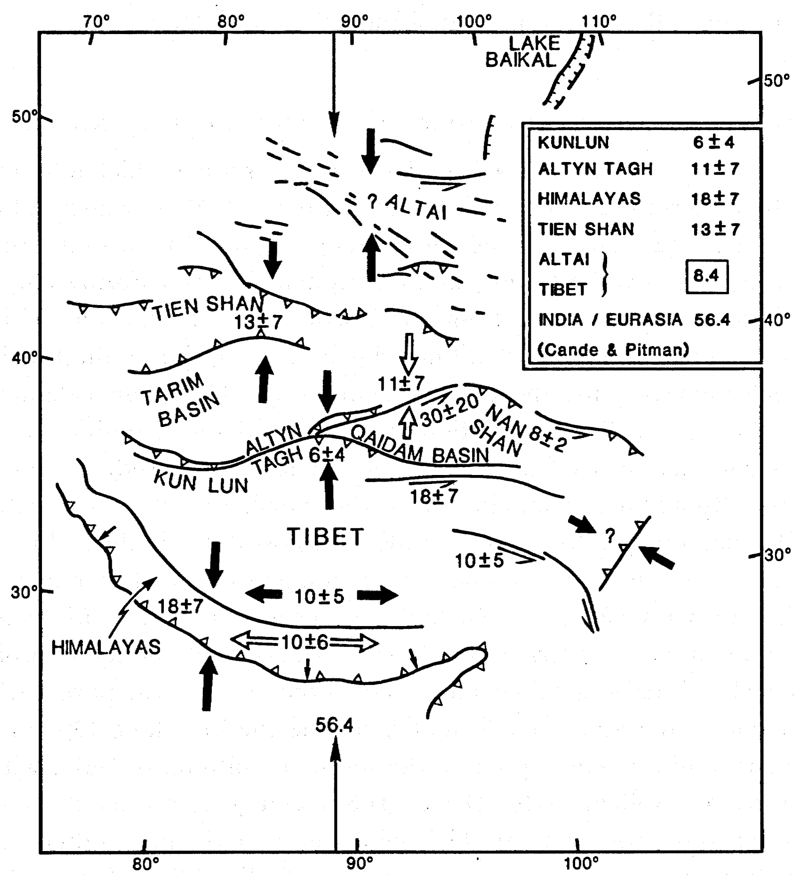

FIGURE 9. Neotectonic displacement in millimetres per year across and within the India–Eurasia convergent zone. Black arrows, shortening and extension values from thrust zones and graben zones respectively; open arrows, from strike-slip faults.

of Anomaly 3 (3.9 Ma), relative to 'stable' Eurasia, an average rate of 56.4 mm a^{-1}. 18 ± 7 mm a^{-1} is absorbed by thrusting in the Himalayas, 13 ± 7 mm a^{-1} by thrusting in the Tien Shan, 6 ± 4 mm a^{-1} by thrusting in the Kunlun and Altyn Tagh, and 11 ± 7 mm a^{-1} as the north–south convergent component of strike–slip faulting on the Altyn Tagh Fault (Molnar *et al.* 1987*a*). Excluding any possible shortening in Tibet, this gives a north–south shortening rate of 48 ± 25 mm a^{-1}, not identical to but consistent with the 56.4 mm a^{-1} deduced by Cande and Pitman, particularly if a small amount of shortening is absorbed by strike-slip faulting in Tibet. East–west extension of about 10 ± 6 mm a^{-1} is occurring within the Himalayas resulting from the curvature of the thrust belt (Armijo *et al.* 1986) in which fault plane solutions indicate a fanning of the slip direction everywhere normal to the thrust front as it changes strike. The continuation of the north–south grabens from the Himalayas across Tibet and the absence of a strike-slip boundary than would take up any differential motion between Tibet and the Himalayas suggest that Tibet is extending east–west at the same 10 mm a^{-1} rate with a small amount of north–south shortening contributed by the strike-slip faults.

The rate of eastward extrusion resulting from east–west extension in Tibet and left-lateral slip on the Kunlun Faults (*ca.* 18 mm a^{-1}; Kidd & Molnar, this volume) and Altyn Tagh Fault, (30 ± 20 mm a^{-1}; Molnar *et al.* 1987*a*) is uncertain. It is not known whether slip on the Kunlun Faults results from the compatibility termination of east–west extension in Tibet, or is in addition to it. It seems unlikely that the rate exceeds 10 mm a^{-1} because there is no obvious right-lateral slip zone at the southern edge of the Tibetan Plateau.

8. The history and timing of shortening, uplift and extrusion

That southern Tibet was at, or near, sea-level with a normal thickness continental crust during the mid-Cretaceous is indicated by widespread shallow water Albian limestones (Hennig 1915; Norin 1946), such as the Lingbuzong (figure 10). In southern Tibet, Albian limestones are succeeded by the Cenomanian–Turonian Takena Formation, a mostly fine-grained molasse derived principally from the north. The Takena was folded during the Campanian to early Palaeocene interval, but prior to the eruption of the Linzizong volcanics (figure 10), synchronously with the development of the Gangdese calc-alkaline arc, the development of the Xigaze flysch accretionary prism and the deposition of shallow water carbonates on a continental shelf on the northern edge of the Indian sub-continent. The obduction of the Spongtang ophiolite onto the northern edges of India during the Maastrichtian/Danian interval (Searle *et al.* 1987) was post-dated in the Indus–Zangbo Suture by the Palaeocene accumulation of ophiolitic conglomerates and late melanges. Therefore, until at least the late Cretaceous, the Neotethyan oceanic tract was subducted northwards beneath the Gangdise arc in southern Tibet. Northward oceanic subduction probably persisted until at least 50 Ma (Linzizong Volcanics). The Zongpu and Zhepure carbonates of the Zanskai shelf of northern India unconformably overlie the Spontang Ophiolite. We believe that all these events and sequences predate the terminal collision of India with Eurasia. We visualize a compressive Andean-style (Dewey 1980) Gangdise arc on the southern edge of Tibet in which intra-arc compression thickened the crust of the southern Lhasa Terrane (England & Searle 1986) and led to the pre-Linzizong folding of the Takena Formation. The obduction of the Spontang ophiolite is seen, therefore, as unrelated to the India–Eurasia collision and like other giant ophiolite nappes such as the Ordovician Bay of Islands Complex

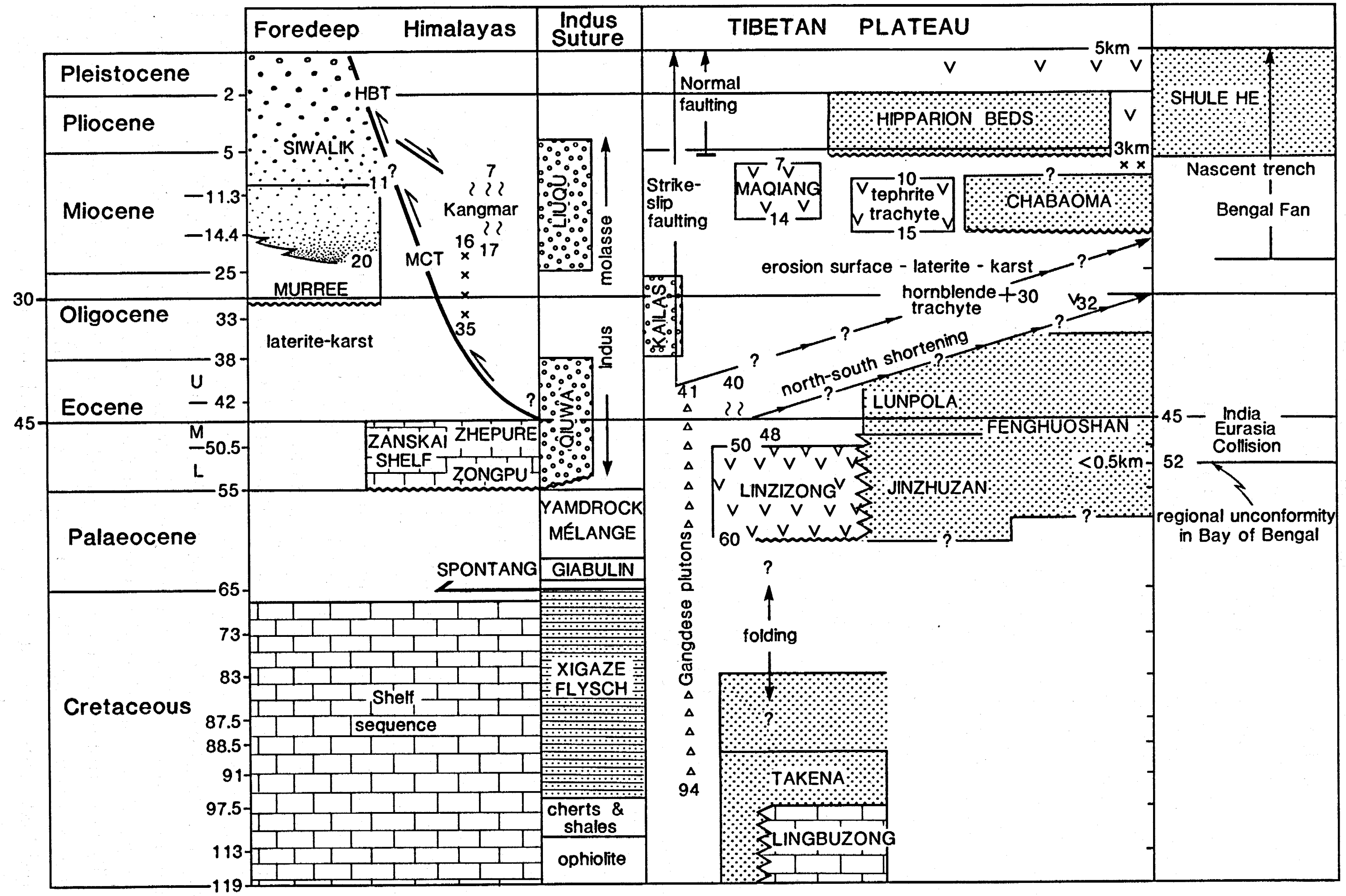

FIGURE 10. Simplified chronological chart for late Cretaceous to Present events in the Himalayas and Tibet. Triangles – calc-alkaline plutons, crosses – other plutons, V – volcanics, lazy S – metamorphism, bricks – marine carbonates, circles – coarse non-marine molasse, dots – non-marine clastics. Details from all papers in this volume; Allègre *et al.* 1984; Copeland *et al.* 1987; Le Fort 1975; Maluski *et al.* 1982; Searle *et al.* 1987; Xu, Schärer & Allègre 1985.

(Dewey & Bird 1971) was emplaced by the subduction of a rifted continental margin beneath the ophiolitic fore-arc of a primitive ensimatic arc. The obduction of the Spontang Ophiolite was part of a late Cretaceous obduction event that, although not precisely synchronous, occurred along the southern margin of Neotethys from Cyprus to India.

There is a widespread agreement that the collision of India with Eurasia occurred during the Eocene (e.g. Achache *et al.* 1983, 1984; Allègre *et al.* 1984; Burg & Chen 1984; Burg *et al.* 1983; Burke, Dewey & Kidd 1974; Klootwijk 1979; Molnar & Tapponnier 1975; Patriat & Achache 1984; Powell & Conaghan 1975; Sengör 1979; Tapponnier *et al.* 1981*a*). That the timing of the collision may be narrowed to the mid-Eocene at about 45 Ma is suggested by several lines of evidence.

First, relative plate motions deduced from finite difference studies using fracture zones and magnetic anomalies in the Atlantic and Indian Oceans indicate changes in the rate and direction of motion of India relative to Eurasia and changes in spreading rate in the Indian Ocean between Anomaly 22 (52 Ma) and Anomaly 20 (44.7 Ma) (McKenzie & Sclater 1971; Patriat & Achache 1984). The new detailed study by S. Cande and W. C. Pitman III (pers. comm.) of the motion of India relative to Eurasia, working the circuit from Eurasia to India through the spreading histories of the Atlantic and Indian Oceans, also shows a change in direction and rate of motion with a slowdown at about 52 Ma from 140 to 95 mm a^{-1}, a further slowdown to 71 mm a^{-1} and a sharp change in direction at about 45 Ma, and generally lower but roughly northward velocities since about 35 Ma combined with an anticlockwise rotation of 21°. We take the change of direction and slowdown at about 45 Ma as the approximate time during collision at which collisional buoyancy forces become sufficiently large to cause the beginning of 'indenter' tectonics (Molnar & Tapponnier 1975) and the beginning of collisional lithospheric thickening. This timing is supported by the cessation of the Gangdise magmatic arc between 50 and 41 Ma and the termination of the Zanskai carbonate shelf during the mid-Eocene (figure 10).

The kinematics, timing and amount of post-collisional north–south crustal shortening in Tibet is ill-defined and the subject of some disagreement among the authors of this paper. This is partly because there is little palaeontological control on the age of Palaeogene red-bed sequences, only the strongly deformed Fenghuoshan red-beds have yielded Eocene charophytes; we were unfortunately unable to examine much of the Palaeogene across the plateau. Also it is difficult or impossible to date, unequivocally, much of the deformation of the Mesozoic strata. There is, however, a clear contrast between virtually undeformed Neogene sequences and generally strongly deformed Palaeogene and older sequences. Flat-lying or gently dipping Lower Miocene and younger volcanics and sediments (Chabaoma, Majang, Hipparion beds) rest unconformably upon deformed older rocks, cut by a 30 Ma trachyte plug north of Erdaogou (Harris, Xu, Lewis, Hawkesworth & Zhang, this volume). The Neogene sequences mainly occupy faulted depressions or erosional hollows and form wide, flat, poorly exposed plains between gently undulating hill ranges that consist of older, more deformed, rocks. The widespread Tibetan erosion surface truncates pre-Neogene, but never Neogene, rocks and is believed to be Miocene in age (Shackleton & Chang, this volume; figure 10). Where Palaeogene red-bed sequences outcrop, they are folded and thrust. At Duba, Amdo and Erdaogou, red-beds occur in thrust-bounded ramp basins with both northward- and southward-directed thrusts. North–south shortening in the Fenghuo Shan ranges is a minimum of 50% and may be considerably more (Coward *et al.*, this volume). It is not considered likely that pre-

Miocene shortening is confined to areas of exposed Palaeogene and older rocks; shortened Palaeogene and older rocks probably lie beneath the Neogene basins because the basins close eastwards and westwards in places onto deformed Palaeogene rocks. Therefore we believe that an average figure of 50% shortening is appropriate for the whole plateau, although it is clearly not homogeneously developed. Modelling suggests that the shortening and thickening of the Tibetan crust probably proceeded northwards from a pre-collisionally-thickened Gangdise crust (England & Searle 1986; England & Houseman 1988) rather than synchronously across the plateau. Evidence in support of this idea comes from Ulugh Muztagh (Molnar *et al.* 1987*b*) near the northern edge of the plateau. Here, an undeformed late Miocene two-mica granite intrudes a deformed basement overlain unconformably by undeformed Pliocene quartz sanidine tuffs, whereas just to the north, thrusting is extant suggesting a northward thrust migration.

In the authors' view, there is, as yet, no definitive evidence to determine the rate and timing of uplift of the Tibetan Plateau. However, several lines of evidence cumulatively suggest that the uplift was accomplished by at least two quite different mechanisms at different rates; a pre-Miocene uplift to about 3 km caused by northward-migrating crustal shortening and thickening and a very rapid Plio-Pleistocene uplift of some 2 km at an average rate of 0.4 mm a^{-1}. The second phase was accompanied by volcanism and east–west extension. Pedeplanation post-dated the pre-Miocene uplift and predated the Plio-Pleistocene uplift (Shackleton & Chang, this volume). First, floral evidence indicates evergreen broad-leaved forests growing in a hot and moist climate at less than 0.5 km during the Eocene (Li *et al.* 1981) succeeded by a later Eocene period of subdued low mean relief landscapes with sub-tropical grassland basins. By early Miocene times, broad-leaved evergreen forests had given way to coniferous alpine forests (Xu Ren 1981; Xu Shuying 1981). Miocene broad-leaved deciduous forests persisted into the early Pliocene (Axelrod 1981; Guo 1981; Song & Liu 1981). During the Pliocene, the Tibetan climate became rapidly drier, partly because of the increasing elevation of the Himalayas, with increasingly sparse deciduous vegetation (Xu Ren 1981; Xu Shuying 1981) culminating today in a plateau with an exceedingly sparse flora of diminutive plants. Second, coarse Pliocene to recent clastics (e.g. Shule He Formation, see figure 10) in basins around the plateau suggest fast late Tertiary–Quaternary uplift. Thirdly, terraces in the Kunlun and knickpoints in rivers at the edge of the plateau indicate fast recent uplift. Fourthly, fission track data in the Himalayas (Zeitler 1985) indicate a rapid acceleration of uplift rates during the last 5 Ma from about 0.2 mm a^{-1} to 0.9 mm a^{-1}; although uplift coupling between the Himalayas and Tibet cannot be proved, these high rates are too great to have resulted from thickening caused by crustal shortening and indicate a mechanism such as catastrophic lithospheric mantle thinning (England & Houseman 1988).

The time relationship between post-mid Eocene northward-migrating shortening in Tibet and southward-migrating thrusting in the Himalayas is not clear. The late Oligocene Murree sediments of the Indo–Gangetic foredeep are derived from the Indian Shield to the south, whereas the upwards-coarsening Siwalik sediments derived from the north began in the early Miocene (figure 10). Within the Himalayas, radiometric ages of structural, metamorphic and igneous events related to Tertiary–Quaternary shortening are mainly younger than 35 Ma (Le Fort 1975; Searle *et al.* 1987). Himalayan large rivers are antecedent from Tibet through the Himalayas (Seeber & Gornitz 1983; Wang, Shi & Zhou 1982) indicating that Tibet was a high area earlier than the Himalayas. Also, during the early Miocene, the Bengal Fan spread as a

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coarsening upwards sequence of increasingly rapidly deposited sediments (Curray *et al.* 1980; Curray & Moore 1974; Cochran *et al.* 1987). Possibly the mid-Miocene development of a nascent trench in the Indian Ocean (Weissel, Anderson & Geller 1980) resulted from an increase in forces resisting the India–Asia convergence by collisional tightening in the Himalayas. We suggest that thrusting spread southwards in the Himalayas from early Miocene times as a consequence of the temporary 'blocking' of north-spreading deformation by the stronger lithosphere of the Tarim and Qaidam Basins. Thrusting and crustal thickening probably began in the northern Himalayas during late Eocene–Oligocene times (Maluski & Matte 1984), principally by the restacking of the thinned crust of the north Indian continental margin, the principal uplift of the Himalayas beginning in early Miocene times. However, the Gangdise belt underwent a phase of accelerating uplift during the early Miocene (Copeland *et al.* 1987, 1988), probably related to early Himalayan events, possibly to underthrusting of northern Himalayan basement beneath the Gangdise belt.

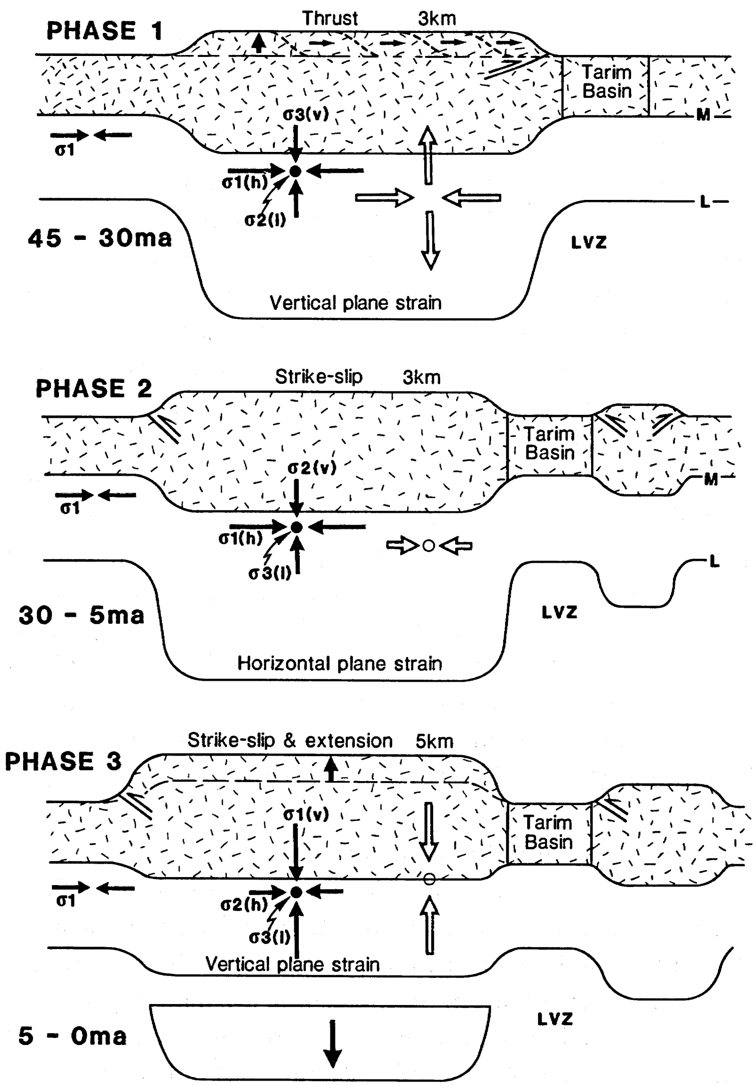

FIGURE 11. Three suggested phases of the Tertiary tectonic evolution of the Tibetan Plateau.

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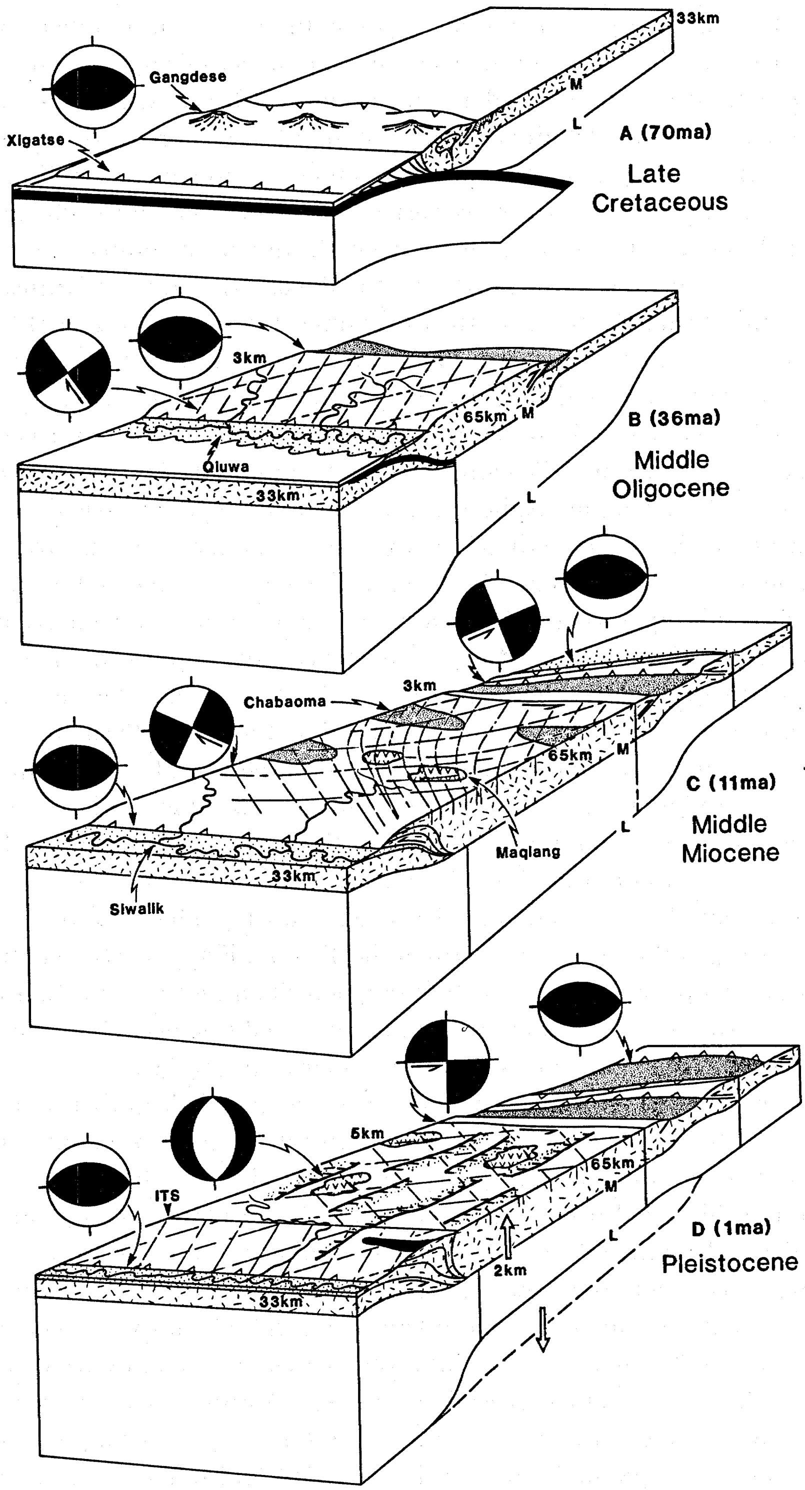

FIGURE 12. Schematic block diagrams illustrating suggested tectonic evolution of the Tibetan Plateau, the Himalayas, the Altyn Tagh and the Tarim Basin.

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Lastly, we summarize our perception of the overall tectonic evolution of the Tibetan Plateau in the context of how finite displacements and strain in the India–Eurasia deforming zone since collision were partitioned in space and time, expressed as a tectonic facies sequence in figure 11 and as a series of four schematic block diagrams in figure 12. We wish to emphasize that this evolutionary model cannot be definitively substantiated but is more consistent than are other suggested models with geological and geophysical data currently available. From 45 to 30 Ma (mid-Eocene to late Oligocene) an India–Eurasia convergence of about 1000 km at an average rate of 66 mm a^{-1} was taken up in Tibet by a northward-propagating north–south crustal shortening and thickening and, in the northern Himalayas, by southward thrusting (Phase 1). Phase 1 was a plane strain, almost pure thrust, regime (figure 11, figure 12B) by the end of which the Tibetan crust between the Qaidam/Tarim Basins to the north and the Indus–Zangbo Suture to the south had doubled its thickness to about 65 km. As the crust/lithosphere progressively reached double thickness, thrusting ceased and a very slow north-south shortening was effected by strike-slip faulting. By about 30 Ma, shortening had reached, and was blocked by, the strong lithosphere of the Tarim and Qaidam Basins and, in Phase 2, spread southward to further develop the Himalayan thrust belt, and northwards into the Altyn Tagh and Tien Shan (figure 11, figure 12C). Phase 2 in Tibet was one of slow horizontal plane strain by strike-slip faulting, during which a very small amount of east–west extrusion occurred and the conjugate strike-slip fault system was shortened north–south to change the dihedral angle from acute to obtuse. In Phase 3, during the last 5 Ma (figure 11, figure 12D) north–south shortening has propagated into Asia north of the Tarim Basin and Tien Shan and southwards to involve more of the Indian subcontinent in the Himalayas. During Phases 2 and 3, some 1420 km of India–Eurasia convergence had been partitioned into about 440 km of shortening in the Himalayas and about 980 km north of Tibet. During Phase 3, a minimum of 50 km of eastward extrusion from Tibet is indicated by the 10 mm a^{-1} east–west extension; we do not believe that the total amount of extrusion during Phases 2 and 3 has exceeded 200 km, or about 15% of the east–west length of the Tibetan Plateau. Much of this 'extrusion' may have been taken up by thrusting in the Longmen Shan and we have not discovered, in Tibet, strains and displacements of the magnitude that would be needed to generate the huge amounts of eastward extrusion claimed by Tapponnier *et al.* (1982).

Phase 3 was a period of east–west extension across north–south grabens and of conjugate strike-slip faulting, when the Tibetan Plateau underwent an uplift of about 2 km (figure 11, figure 12D). Uplift was accompanied by volcanism and has been explained by the catastrophic delamination of the thickened lithospheric root beneath the Plateau (England & Houseman 1988). Late and rapid uplift, extension, bimodal magmatism, and post-orogenic minimum-melting granites superposed upon earlier phases of shortening, crustal thickening and prograde metamorphism is a feature common to many Palaeozoic and Cainozoic orogenic belts. It may be an important mechanism, generated by lithospheric delamination, by which orogenic zones weaken and extend, so as to return crustal thicknesses to normal without large amounts of denudation. In an extreme form, it may lead to total orogenic collapse and continental separation, a phenomenon pointed out by Wilson (1966) and inherent in the concept of the Wilson cycle of ocean opening and closing, where zones of crustal thickening in orogenic belts became weak zones susceptible to lithospheric extension (Dewey, in press).

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Maps

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PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES