OVERVIEW

Tectonics of the Himalaya and southern Tibet from two perspectives

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ABSTRACT

The Himalaya and Tibet provide an unparalleled opportunity to examine the complex ways in which continents respond to collisional orogenesis. This paper is an attempt to synthesize the known geology of this orogenic system, with special attention paid to the tectonic evolution of the Himalaya and southernmost Tibet since India-Eurasia collision at ca. 50 Ma. Two alternative perspectives are developed. The first is largely historical. It includes brief (and necessarily subjective) reviews of the tectonic stratigraphy, the structural geology, and metamorphic geology of the Himalaya. The second focuses on the processes that dictate the behavior of the orogenic system today. It is argued that these processes have not changed substantially over the Miocene–Holocene interval, which suggests that the orogen has achieved a quasi–steady state. This condition implies a rough balance between plate-tectonic processes that lead to the accumulation of energy in the orogen and many other processes (e.g., erosion of the Himalayan front and the lateral flow of the middle and lower crust of Tibet) that lead to the dissipation of energy. The tectonics of the Himalaya and Tibet are thus intimately related; the Himalaya might have evolved very differently had the Tibetan Plateau never have formed.

Keywords: Himalaya, tectonics, India, Asia, mountain building, orogenic belts, Tibet.

INTRODUCTION

Ask an undergraduate student of geology to name a mountain belt produced by continent-continent collision, and the likely answer will be the Himalaya. In our science, the Himalayan-Tibetan orogenic system has become an icon of sorts, and the models proposed for its evolution strongly influence our interpretation of the tectonics of older belts. For this reason, it is important from time to time to review the state of our understanding of the Himalaya and Tibet and also to ask what more we can do here to better inform the next generation of models of collisional orogenesis.

In this attempt at a synthesis, the problem will be approached from two very different perspectives that I like to think of as analogous to the approaches of two major schools of painting in late nineteenth century France. The first, and most familiar of the two, is that of Impressionism. The great Impressionist masters dedicated their efforts to crystallizing the flow of time on canvas as they perceived it. Whether painted in a bustling Paris café or on the banks of a quiet pond, Impressionist art preserves an instantaneous sensation. Claude Monet, perhaps the greatest of the Impressionists, tried to go further by expressing the passage of time in his series paintings, like those of the façade of the Rouen Cathedral. If Impressionism is a form of historical documentation, we might think of one of the great traditions of tectonics research—the description of orogeny as a temporal progression of deformational episodes—as an essentially Impressionist enterprise. Our ability to use the developmental sequence of major structures in one setting to predict the sequence in others, as is the case for foreland fold-and-thrust belts worldwide (Dahlstrom, 1970), is ample testimony to the value of the Impressionist perspective.

A second approach to the study of orogeny bears conceptual similarities to Neo-Impressionism, the avant garde artistic movement in France near the end of the nineteenth century. Exploding onto the scene at the last of the great Impressionist exhibitions in Paris in 1886, Neo-Impressionism was an organized response to the subjectivity of Impressionist art. The leader of this revolution, which many historians regard as paving the way for what we now call Modern Art (Ward, 1995), was an unlikely 27-year-old, classically trained painter named Georges Seurat. Rather than presenting the "impression" of a scene through the eyes of the artist, Seurat reasoned, shouldn't it be possible to translate the substance of the scene to a painting and then depend on the perception of the viewer to reconstruct the essence of the original? Seurat became a serious student of the physics and chemistry of light, color, and visual perception, and his careful, experimental approach toward painting helped establish his reputation as the preeminent "scientific" artist of his time (Homer, 1964). Despite his tragic death at the age of 31, Seurat's approach had a profound influence on the development of a remarkably diverse group of artists—Dali, Gauguin, Matisse, Kandinsky, Picasso, and Van Gogh, among others—and some of his theories helped lay the foundation for modern digital imagery.

A close look at one of Seurat's later paintings, like *Les Poseues,*[1] reveals one of his most famous innovations: the use of small, closely spaced specks of primary colors that blend optically to create a dramatic, almost luminous image in the eyes of an observer positioned some distance from the painting. Introduced just when reductionism was developing into the prevailing philosophy of modern science, Seurat's "pontillist" technique has been interpreted as a commentary on the atomistic nature of the world around us: to know the whole, you must first understand its constituent parts. I see Seurat's works as something more: a celebration of how those parts inte-

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[1]Seurat's work does not reproduce well, but those who do not have the opportunity to experience the original at the Barnes Foundation near Philadelphia can find images of this and other Neo-Impressionist masterpieces on the World Wide Web at http://www2.iinet.com/art/index.html.

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grate to define the whole. In the field of tectonics, a Neo-Impressionist perspective provides special insights into the behavior of mountain ranges by emphasizing the synergy of thermal, deformational, and erosional processes during orogenesis.

The first part of this paper is a traditional, Impressionistic view of the evolution of the Himalayan-Tibetan orogenic system, with emphasis on the past fifty million years of history. Although unapologetically subjective, this first part is designed to reflect the state of our knowledge at the end of the twentieth century. The second part is a Neo-Impressionistic attempt to answer some questions: What are the essential processes that define the behavior of this particular system today? How far backward in time can they be traced? What can they tell us more generally about orogeny?

OROGRAPHIC IMPRESSIONS

The Himalayan-Tibetan orogenic system is the most distinctive landform on our planet (Fig. 1). Covering an area comparable in scale to that of the Iberian peninsula, the Tibetan Plateau is dominated by an internally drained central region with very low relief and a mean elevation of over 5 km (Fielding et al., 1994). Neither the eastern nor the western boundaries of the plateau are particularly well defined. From its central high, Tibet slopes gently eastward, over a distance of nearly 1000 km, to a mean elevation of ~3.5 km before ending at the Longmen Shan escarpment at about long 105°E. To the west, the subdued topography of the plateau passes gradually into the rugged terrain of the Karakoram and Pamir Mountains (Searle, 1991; Burtman and Molnar, 1993). The northern edge of Tibet, in contrast, is an extremely sharp topographic boundary defined by the narrow (~100 km wide) eastern and western Kunlun Mountains.

This paper is focused on the southern topographic front of the Tibetan Plateau, the Himalaya, which are traditionally defined as the 2500-km-long arc of mountain ranges stretching between two structural syntaxes named for major peaks: Namche Barwa (7782 m) on the east and Nanga Parbat (8125 m) on the west (Fig. 1). The Himalayan arc can be further subdivided into western, central, and eastern sectors on the basis of regional variations in geomorphology. The sharpest transition between the Tibetan Plateau and the Indo-Gangetic foreland occurs in the central Himalaya, between long 76°E and 91°E, where the width of the mountain belt is at its narrowest, a relatively uniform 100–150 km. The morphology of the central Himalaya is dictated by longitudinal river systems that flow directly off the southern plateau through the central Himalaya, carving some of the world's deepest canyons and segmenting the region into distinct mountain ranges that include eight of the ten highest mountains on Earth (Fig. 2). The average relief is significantly lower in the eastern and western sectors of the Himalaya, 7000 and 8000 m peaks are less numerous, and the river drainage systems are typically more complex. At Namche Barwa (Fig. 1), the eastern Himalaya sector passes through a tight orographic bend before diffusing into a series of north- and northeast-trending ranges and high plateaus in the Assam State of India, the Kachin and Shan States of Myanmar, and the Yunnan Province of China (Tapponnier et al., 1986; Mitchell, 1993; Bertrand et al., 1999). Northwest of Nanga Parbat, the main Himalayan ranges merge with elements of the Karakoram, Hindu Kush, and Pamir Mountains to form the westernmost Tibetan Plateau without a clearly defined "Himalayan topographic front."

The Himalaya and Tibet have a strong effect on regional climate. Sweeping northward from the Bay of Bengal in the spring and summer months, the Indian monsoon strikes the eastern Himalaya with full force before losing energy as it is diverted westward along the mountain front. As a consequence, the climate of the Himalayan foreland varies progressively from wet tropical in the east, to temperate in the central Himalaya, and to semiarid in the far west. Heavy monsoonal rainfall and snowfall in the eastern Himalaya almost certainly contribute to the low elevations of this sector. Although the central Himalayan ranges create one of the most impressive rain shadows on Earth, the eastern Himalayan peaks are insufficiently high to form a completely effective barrier to precipitation. Thus, eastern Tibet experiences heavy precipitation, while central and western Tibet are high-altitude desert. The latitudinal variations in climate, combined with the microclimatic effects of increasing elevation that are common to all mountainous regions, have important implications regarding our ability to reconstruct the geologic history of the Himalaya. Exposures of foreland rock sequences and structures are best in the western Himalayan foothills and are progressively poorer toward the east. As a consequence, much of our perception of the geology of the Himalayan foreland is shaped by studies in the west, and we know far less about the geology of the eastern Himalayan foreland (Burbank et al., 1997). Exposure improves dramatically in the Higher Himalaya, especially above

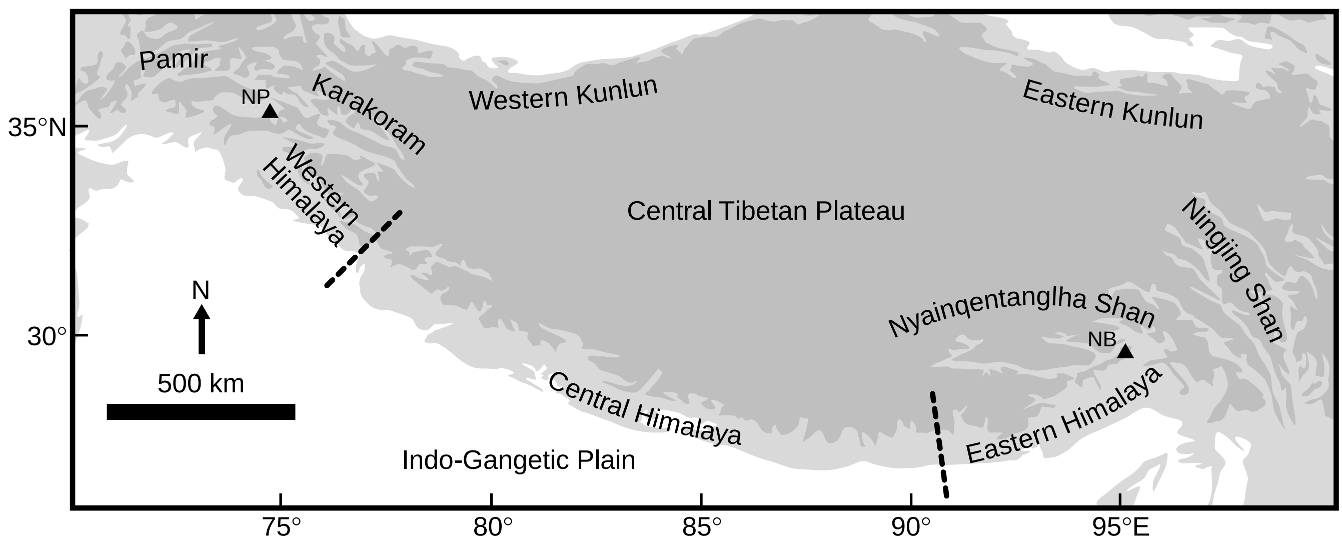

Figure 1. Generalized topographic map of the Himalaya and Tibetan Plateau. Dark shading indicates elevations above 4000 m; light shading indicates elevations between 1000 and 3000 m. Dashed lines represent boundaries between the western, central, and eastern Himalaya. NP—Nanga Parbat; NB—Namche Barwa.

3000 m, and reaches nearly 100% on the southern Tibetan Plateau north of the central Himalaya. Thus, the geologic record of orogenesis is most complete in northwest India (Searle, 1986), almost as good in the Transhimalayan region of south-central Xizang (political Tibet; Burg and Chen, 1984), adequate in the structurally and topographically highest parts of the metamorphic core of the Himalaya (Le Fort, 1986; Pêcher, 1991), and least well preserved (or well exposed) in the piedmont regions east of approximately long 78°E (Stöcklin, 1980; Valdiya, 1980; Gansser, 1983).

Our level of understanding of the orogen is further hampered by politics. Skirmishes involving India, Pakistan, and China have been common throughout the past half-century, and many international borders in the Himalayan region are disputed. China's occupation of Xizang, India's persistent difficulties with ethnic insurgencies, and the reluctance of Tibet, Bhutan, and Nepal to welcome foreigners at various times in this century have discouraged systematic research. In the Himalaya and Tibet, geologists often work in remote areas of extraordinarily high relief, sometimes at very high elevations, where access roads are few. Such deterrents have contributed to one of the great ironies of modern tectonics research: the orogen that inspires most models of collisional orogenesis is also one of the most incompletely mapped.

AREAL GEOLOGY

Such sobering thoughts notwithstanding, Figure 3 represents my current impression of the areal geology of the Himalaya and adjacent areas of Tibet between long 73° and 89°E, where the present state of mapping seems to justify an attempt at synthesis. Influenced by numerous maps published over the past 35 years—starting with that of Augusto Gansser (Gansser, 1964), and guided in part by remote-sensing data (e.g., Landsat and SPOT imagery)—Figure 3 portrays the geology of the region by dividing it into a series of longitudinal tectonostratigraphic domains that are bounded by major fault systems. In doing so, it follows a tradition established by Himalayan geologists in the first quarter of this century and effectively codified through the seminal works of Gansser (1964) and Le Fort (1975). Readers should be aware, however, that these tectonostratigraphic divisions were developed largely as a consequence of research in the well-exposed Punjab and Kumaun regions of the Indian Himalaya between long 73° and 80°E (Auden, 1937; Heim and Gansser, 1939; Wadia, 1939); the application of these tectonostratigraphic divisions to other parts of the Himalaya is not entirely without controversy (Yeats and Lawrence, 1984; Pogue et al., 1999).

Transhimalayan Zone

Prior to its collision with India, the southern margin of Eurasia was marked by a continental arc that developed as a consequence of the northward subduction of Neo-Tethys oceanic crust (Dewey and Bird, 1970; Tapponnier et al., 1981). The Transhimalayan zone consists of volcanic and plutonic elements of this arc, their variably metamorphosed Precambrian–Mesozoic country rocks, and less commonly preserved Cretaceous–Tertiary forearc basin sequences (Burg et al., 1983; Searle, 1991). Reconstructions of the precollisional configuration of this margin are uncertain, largely because research on pre-Himalayan geologic problems in the Transhimalayan realm has focused on only two relatively accessible areas: the region around Xizang's capital of Lhasa (Burg et al., 1983) and the Kohistan-Ladakh sector of Pakistan and India (Searle, 1991; Honegger et al., 1982).

Largely as a consequence of geologic explorations undertaken by Chinese scientists (Chang et al., 1982) in collaboration with American (Bally et al., 1980), French (Allègre et al., 1984), and British (Shackelton, 1981; Dewey et al., 1988) delegations, it is now widely recognized that Tibet was assembled by the accretion of a series of exotic terranes to Eurasia in Mesozoic time. The last pre-Himalayan accretion event was the collision of an island-arc complex, now represented by the Kohistan-Ladakh terrane, along the Shyok suture zone in Late Cretaceous time (Treloar et al., 1989b; Rolfo et al., 1997). The Kohistan region (Fig. 2), west of the Nanga Parbat syntaxis, provides an especially good cross section through the arc complex (Coward et al., 1982). The structurally highest rocks, exposed along the northern margin of the terrane, include Lower Cretaceous island-arc volcanic and sedimentary units that are intruded by gabbroic to granitic plutons of the Kohistan batholith (Searle, 1991). Farther south are progressively deeper elements of the Kohistan arc, including the spectacular gabbroic and ultramafic stratiform plutons of the Lower Cretaceous Chilas Complex (Khan et al., 1989; Mikoshiba et al., 1999) and highly deformed, mafic metavolcanic and metaplutonic rocks of the Kamila Amphibolite, which may include remnants of the Neo-Tethyan oceanic basement of the arc (Treloar et al., 1996). East of the Nanga Parbat syntaxis, the Ladakh part of the terrane (Figs. 2, 3) preserves plutonic elements of the arc in the form of the Ladakh batholith (Honegger et al., 1982), as well as representatives of the volcanic and sedimentary arc carapace (Dietrich et al., 1983).

Geochronologic data suggest that calc-alkalic magmatism in Kohistan and Ladakh began before, but significantly outlasted, Late Cretaceous docking of the terrane with Eurasia. The oldest reliably dated elements of the Kohistan and Ladakh batholiths are ca. 100 Ma, and the youngest are of late Paleocene age (Honegger et al., 1982; Schärer et al., 1984; Petterson and Windley, 1985). This age range suggests that, subsequent to collision, the then-newly-accreted Kohistan-Ladakh island-arc complex evolved into a continental arc marking the southern border of Eurasia. Volumetrically, most of the Kohistan and Ladakh batholiths developed in this continental-arc setting in latest Cretaceous time.

East of long 80°E, the continental arc is represented principally by the Gangdese batholith of southern Tibet. The basic geology of the western edge of this batholith, in the Kailas region (Fig. 2), was outlined by Heim and Gansser (1939). Their documentation of the existence of Gangdese granitoid rocks cropping out north of Mount Kailas and of evidence for a volcanic carapace in the form of clasts within the Kailas conglomerate has been expanded upon greatly by subsequent researchers (Ryerson et al., 1995; Murphy et al., 1997b; Miller et al., 1999; Yin et al., 1999). However, most of these studies have focused on postcollisional geologic problems, and relatively little is known still about the pre-Himalayan geology of the Kailas sector.

The best-studied sector of the Gangdese batholith and its country rocks is in southern Xizang between about long 84° and 93°E. This part of the continental arc developed on stabilized crust of the Lhasa terrane that had accreted with Eurasia during the Late Jurassic (Dewey et al., 1988). The northern Lhasa block includes poorly characterized Precambrian–Cambrian metamorphic rocks unconformably overlain by Devonian–Upper Cretaceous shallow-water continental and marine strata with some intercalated volcanic rocks of Carboniferous, Triassic, and Jurassic age (Burg et al., 1983; Pearce and Mei, 1988; Yin et al., 1988). Farther south, the basement for the Paleozoic–Mesozoic stratigraphic succession is unexposed, but Burg et al. (1983) have suggested that parts of it may have been deposited on an oceanic crust. Igneous rocks along the southern margin include the calc-alkalic Linzizong volcanic rocks and their plutonic substrate, the Gangdese batholith (Harris et al., 1988; Pearce and Mei, 1988). Geochronologic data for these units document magmatic activity from at least 94 Ma to as recently as 42 Ma, with most dated units having Paleocene or Eocene ages (Schärer and Allègre, 1984; Xu et al., 1985; Coulon et al., 1986; Copeland et al., 1995). Compared with data from regions farther west, such findings suggest that the most intense period of magmatic activity in the Transhimalayan continental arc was older (latest Cretaceous) in the west and younger (early Tertiary) in the east and that the cessation of arc magmatism occurred earlier in the west (late Paleocene) than in the east (middle Eocene). This last observation is consistent with the notion that the end of Transhimalayan arc magmatism corresponds closely in time with the collision of India, which occurred earlier in the west than in the east (Rowley, 1996).

Exactly how the Lhasa and Kohistan-Ladakh terranes were related to one another prior to India-Eurasia collision is one of the great unanswered questions of Himalayan-Tibetan tectonics. At present, the two terranes are juxtaposed by the right-lateral Karakoram fault system, one of the most spectacular structures in the orogen (Figs. 2, 3), and how one interprets their relationship depends largely on how much slip has occurred on the Karakoram system since Late Cretaceous time. West of the Karakoram system and north of the Shyok suture lies the Karakoram terrane (Fig. 3), which includes Carboniferous–Upper Cretaceous marine to continental strata, but is dominated by the calc-alkalic Jurassic–Cretaceous Karakoram batholith (Searle, 1991). Some researchers have regarded the Karakoram batholith as a westward extension of the Gangdese batholith that has been offset hundreds of kilometers by the Karakoram fault system (Peltzer and Tapponnier, 1988). In such a model, the Lhasa and Karakoram terranes would be correlative, and the Banggong-Nujiang suture (marking the northern boundary of the Lhasa terrane in central Xizang [Fig. 2; Girardeau et al., 1984b]) corresponds to the Rushan-Pshart suture between the central and southern Pamir Mountains (Shvolman, 1981; Şengör et al., 1988; Gaetani et al., 1990; Sinha et al., 1999). This interpretation has been criticized by Searle (1996) because all available geologic evidence suggests that the Karakoram fault system developed in Neogene time and because several offset markers of Miocene–Holocene age suggest no more than 120–150 km of displacement on the fault system. Correlating the Gangdese and Karakoram batholiths also is made difficult by contrasting magmatic histories; for example, the oldest phases of the Karakoram batholith are much older than any dated intrusive rocks in the Gangdese batholith (Searle et al., 1989). Moreover, this interpretation precludes a natural, nearly along-strike correlation between continental-arc rocks of similar age and composition in the Ladakh and Gangdese batholiths.

An alternative model, promoted by Searle (1996) and Burtman and Molnar (1993), makes the Banggong-Nujiang and Shyok sutures correlative, as well as the Gangdese and Ladakh batholiths, and it implies that the Rushan-Pshart suture correlates with a different suture along the southern margin of the western Kunlun Mountains. Unfortunately, this interpretation also has its drawbacks. The most important is that current estimates of the ages of suturing along the Shyok and Banggong-Nujiang zones are substantially different—Late Cretaceous (Treloar et al., 1989b) vs. Late Jurassic (Allègre et al., 1984; Dewey et al., 1988)—as are estimates of the probable ages of suturing along the Rushan-Pshart and southern Kunlun zones—Late Jurassic–Early Cretaceous (Burtman and Molnar, 1993) vs. Late Triassic–Early Jurassic (Dewey et al., 1988). Evaluation of competing models of the paleogeography of the Transhimalaya will require better constraints on the ages of major sutures in western Tibet and systematic mapping in the region north of the Banggong-Nujiang suture in western Xizang, where the model of Searle (1996) predicts the locations of offset equivalents of well-characterized units in the Karakoram terrane.

The youngest bedrock in the Transhimalaya includes Neogene volcanic rocks that have been characterized best in western Xizang (Fig. 2; Coulon et al., 1986; Pearce and Mei, 1988; Turner et al., 1996). In this region, they include ultrapotassic, potassic, and high-potassium calc-alkalic lavas with chemical characteristics indicating a combination of crustal- and mantle-lithosphere source regions (Miller et al., 1999). K-Ar, $^{40}Ar/^{39}Ar$, and Rb-Sr geochronologic data suggest a broad range of eruption ages (10–25 Ma), but most dates seem to cluster in the 16–23 Ma interval, with some indication that calc-alkalic lavas are younger than potassic and ultrapotassic lavas (Coulon et al., 1986; Miller et al., 1999).

Indus-Tsangpo Suture Zone

The suture that marks the zone of collision between the Indian and Eurasian plates can be traced discontinuously for a distance of at least 3000 km from Myanmar to Afghanistan. The first detailed observations of the complex geology of this zone were made by Augusto Gansser during two surreptitious expeditions into the Kingdom of Tibet in the late 1930s (Heim and Gansser, 1939). Gansser recognized massive allochthonous sheets containing "exotic blocks" of radiolarian chert, flysch, limestone, and mafic intrusive and extrusive rocks just south of the Kailas region. By 1964, he had interpreted these blocks as part of a major overthrust sheet of ophiolitic material that rooted into a narrow, near-vertical shear zone exposed along the upper Indus River drainage (Gansser, 1964). Gansser regarded this "Indus suture zone" as marking a fundamental discontinuity between the Transhimalayan realm and India, but the plate-tectonic significance of the suture zone was not recognized until some years later (Dewey and Bird, 1970; Molnar and Tapponnier, 1977; Gansser, 1980).

Today, most of our understanding of the Indus-Tsangpo suture zone derives from studies done in south-central Xizang, near Lhasa and Xigaze (Bally et al., 1980; Shackelton, 1981; Tapponnier et al., 1981), and in Ladakh (Frank et al., 1977a; Searle, 1983; Thakur, 1981). In both of these areas, the suture zone comprises three major rock sequences. Separated by fault systems of both Mesozoic and Cenozoic age, they represent the Neo-Tethyan Ocean basin and its northern and southern continental margins.

Transhimalayan Components. In south-central Xizang, Cretaceous turbidites of the Xigaze Group have been interpreted as a forearc sequence deposited along the southern margin of the Gangdese continental arc and subsequently incorporated into the Indus-Tsangpo suture zone during collision (Bally et al., 1980; Shackelton, 1981; Burg and Chen, 1984; Wan et al., 1998). In Ladakh, the forearc basin is represented by middle Cretaceous–earliest Eocene turbidites of the Indus Group (Garzanti and Van Haver, 1988). In the same region, a second sequence of Upper Cretaceous(?) volcaniclastic strata (the Nindam Formation) grades into stratigraphically high units within the Jurassic–Cretaceous Dras Volcanics (Dietrich et al., 1983; Searle, 1983). The Nindam–Dras package may represent an island arc and its accretionary prism that were positioned just south of the Indus forearc basin by Late Cretaceous time and structurally juxtaposed with Indus units at the time of collision (Garzanti and Van Haver, 1988; Robertson and Degnan, 1994).

Neo-Tethyan Ocean-Floor Components. Ophiolites, ophiolitic melange, and deep-ocean sedimentary rocks in the Indus-Tsangpo suture zone are of Jurassic–Cretaceous age (Gansser, 1980; Le Fort, 1997; Corfield et al., 1999). Well-preserved ophiolites are rare in the Himalaya, but those that do occur (Fig. 3) are spectacular: the Xigaze ophiolite of south-central Xizang (~2000 km^2 in outcrop—Nicolas et al. [1981] and Girardeau et al. [1985]), the Kiogar ophiolite of southwestern Xizang (~3500 km^2 in outcrop—Gansser [1964, 1980]), and the Spontang ophiolite of Ladakh (~200 km^2 in outcrop—Reuber [1986]). Of these, only the Xigaze ophiolite occurs exclusively within the Indus-Tsangpo suture zone sensu stricto; the others are found as klippen or half-klippen in the Indus-Tsangpo suture allochthons (discussed in more detail in a subsequent section). Oceanic rocks in the Indus-Tsangpo suture were metamorphosed to varying degrees at low temperatures. Greenschist-facies metamorphic assemblages are widespread, and blueschist-facies assemblages have been reported from several localities (Honegger et al., 1989; Jan, 1990).

Indian Plate Components. The southern margin of Neo-Tethys is represented in the suture zone by Triassic–Cretaceous turbidites deposited on the north Indian shelf and slope (Frank et al., 1977a; Burg and Chen, 1984; Robertson and Degnan, 1993). Some of the most spectacular exposures occur in Ladakh as the Lamayuru flysch sequence, which contains decameter-

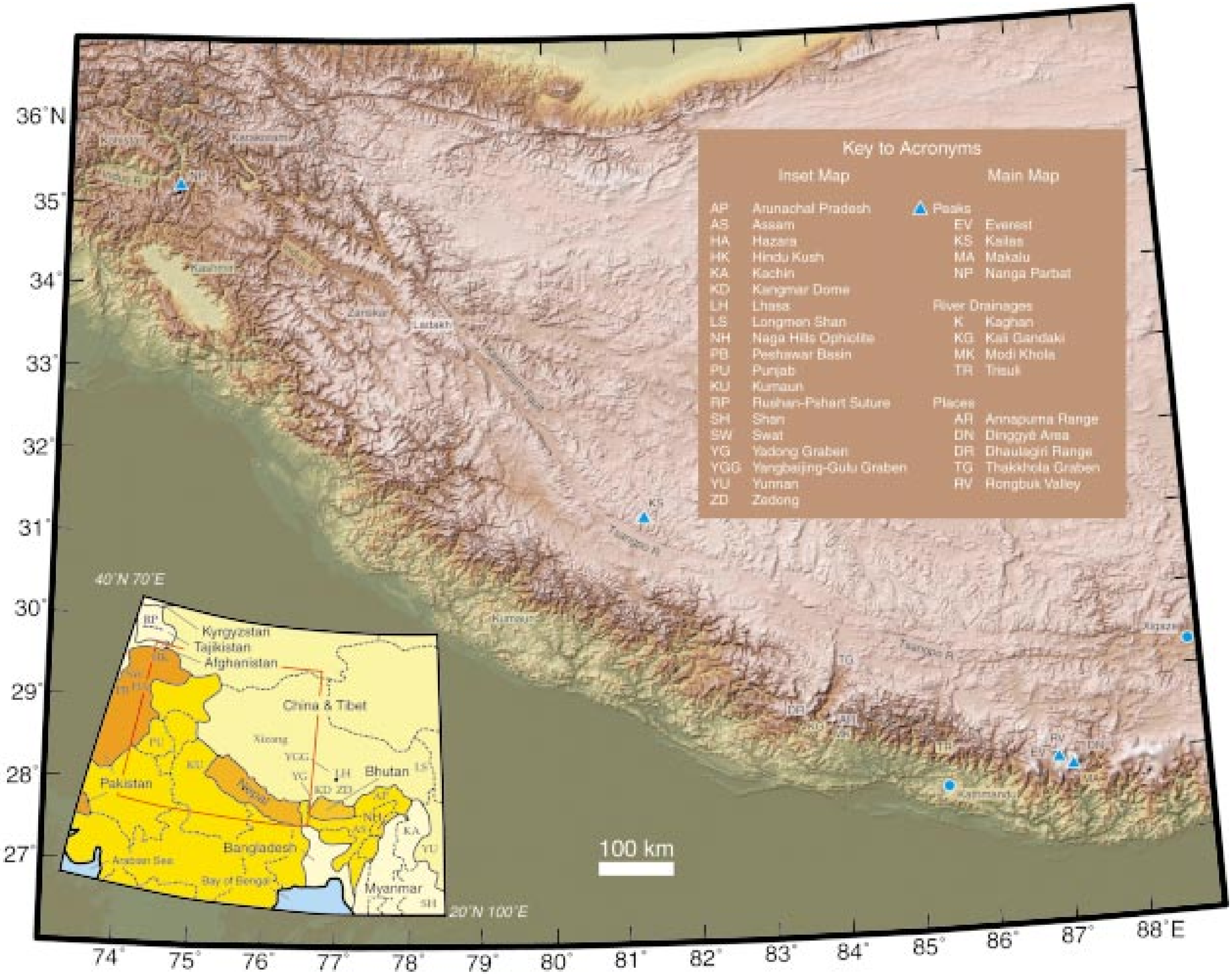

Figure 2. Raised relief map of the Himalaya and a part of Tibet, constructed from GTOPO 30 digital elevation data with an Albers conical projection. Inset is a smaller-scale political map of the Himalayan-Tibetan region showing features not lying within the area of the main relief map or the tectonic map in Figure 3.

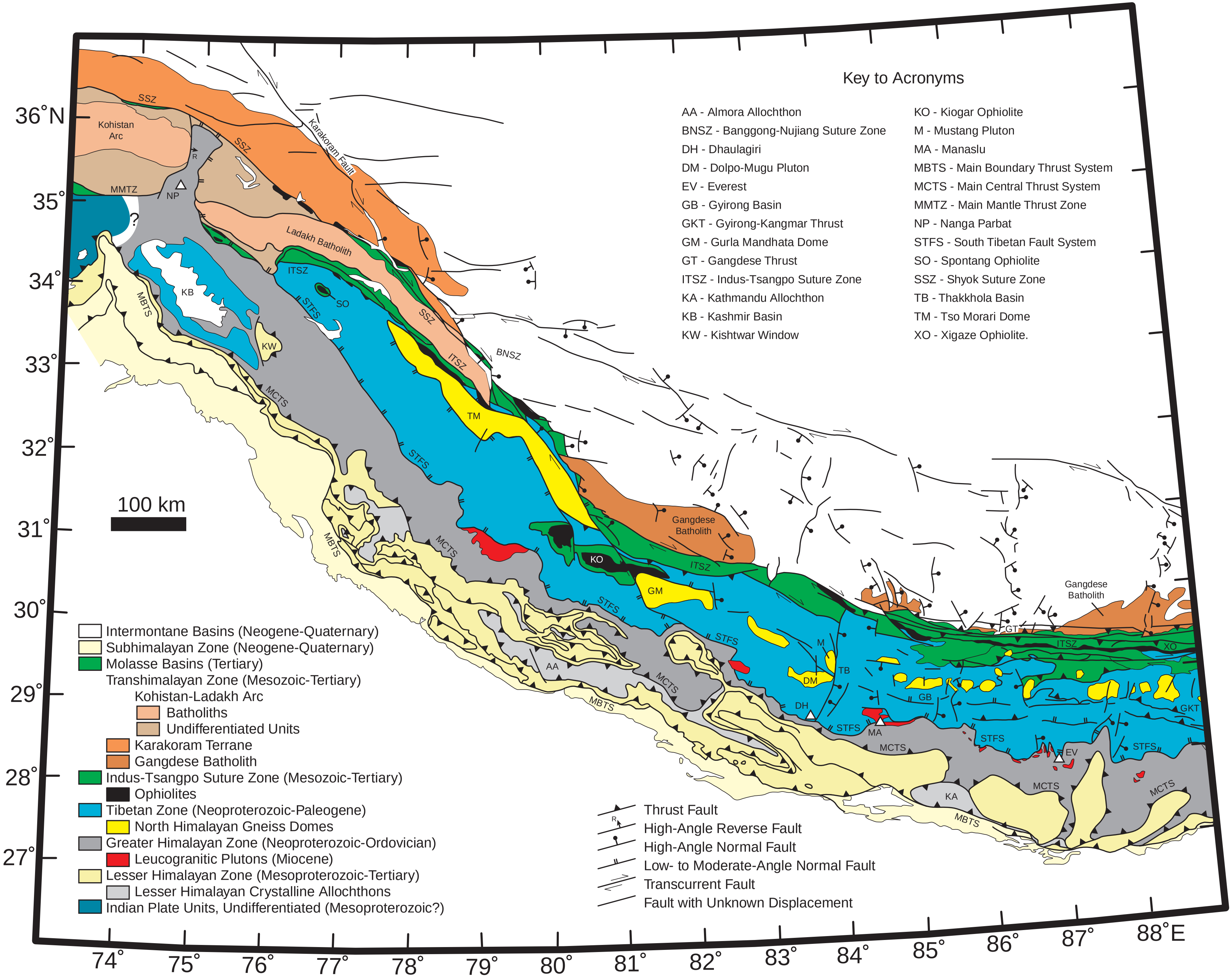

Figure 3. Tectonic map of the Himalaya and southern Tibet compiled (with extensive reinterpretations) from a variety of sources. Scale and projection are the same as for Figure 2.

to kilometer-scale exotic blocks of Permian–Triassic limestone (Bassoullet et al., 1981; Robertson, 1998). Similar rocks occur within the suture zone in the Kailas and Lhasa-Xigaze regions of Xizang (Gansser, 1964; Shackelton, 1981; Burg and Chen, 1984).

Indus-Tsangpo Suture Zone Allochthons. In the western Himalaya, a system of steep backfolds and backthrusts of Miocene and younger age has modified the original geometry of the suture between India and Eurasia such that suture-zone rocks occur in two distinctive structural settings: the steep structural belt of the Indus-Tsangpo suture zone sensu stricto and a series of erosional remnants of shallowly dipping, composite thrust sheets lying structurally above rocks of the Tibetan zone (Heim and Gansser, 1939; Frank et al., 1977a; Searle et al., 1988). These klippen and half-klippen preserve two of the most extensive tracts of Neo-Tethyan ocean floor in the Himalayan-Tibetan orogen (the Kiogar and Spontang ophiolites), as well as a remarkable record of the paleogeography of the Indian margin of Neo-Tethys (Searle et al., 1997a). Although most of the units in the allochthons have correlatives within the suture zone itself, some do not. These include remnants of a Cretaceous island arc thought to have developed near the southern margin of Neo-Tethys (the "Spong arc" volcanic rocks), as well as a sedimentary and tectonic melange of carbonate, volcaniclastic, and alkalic volcanic rocks that has been interpreted as an accretionary complex that developed above the north-directed, intraoceanic subduction zone responsible for the Spong arc (Corfield et al., 1999).

West of Ladakh, the Indus-Tsangpo suture zone narrows substantially as it wraps around the Nanga Parbat syntaxis and continues westward into the Hazara and Swat regions of Pakistan (long 72°–74°E). Here the suture is referred to as the "Main Mantle thrust zone," or sometimes as the "Southern suture," in order to differentiate it from the "Shyok" or "Northern" suture that separates the Kohistan-Ladakh and Karakoram arc terranes (Gansser, 1980). Most researchers have mapped sedimentary and volcanic tectonites defining the Main Mantle thrust collectively as the "Indus melange" (Tahirkheli et al., 1979; Coward et al., 1986; DiPietro et al., 1999;), but there also have been some successful attempts to define mappable, regionally extensive tectonostratigraphic units within the suture zone (for example, Anczkiewicz et al., 1998a).

The geology of the Indus-Tsangpo suture zone is poorly known between south-central Tibet and the Namche Barwa syntaxis of eastern Tibet. Burg et al. (1998) mapped the suture at Namche Barwa as a mylonite zone containing lenses of metamorphosed mafic and ultramafic rocks. It apparently continues southward into the Indo-Burman ranges, near the border between Myanmar and the Indian state of Assam, where it is marked by the Naga Hills ophiolite belt and the synorogenic "flyschoid" sedimentary rocks of middle Eocene–Oligocene age (Acharyya, 1997).

Post-Collisional Molasse Basins

Spatially associated with the Indus-Tsangpo suture zone is a discontinuous belt of continental molasse basins that place important minimum age constraints on India-Eurasia collision (Rowley, 1996). Deposits in such basins include the Kailas conglomerates (which form the bedrock of the mountain after which they are named [Gansser, 1964; Honegger et al., 1982]), and the Liuqu conglomerates and associated continental clastic rocks in south-central Xizang (Shackelton, 1981). The most thoroughly investigated sections, however, have been those of the Ladakh region, where the Indus Group displays evidence for an early Eocene transition from forearc-marine to continental-nonmarine depositional environments (Brookfield and Andrews-Speed, 1984). Postcollisional continental strata in Ladakh (the "Indus molasse") include red beds, conglomerates, and lacustrine deposits that suggest deposition in intermontane basins (Garzanti and Van Haver, 1988). Fossil control on the age of the Indus molasse and similar deposits in southern Tibet is limited, and many of the preserved sections may include components as young as late Miocene or even Pliocene (Searle et al., 1997a). From three areas in southern Xizang, $^{40}Ar/^{39}Ar$ cooling histories of clasts in the molasse deposits and the minimum ages of crosscutting dikes have been used to bracket the depositional age of part of the molasse sequence between about 24 and 17 Ma (Harrison et al., 1993; Yin et al., 1999).

Tibetan Zone

The broad region of the southern Tibetan Plateau lying between the Indus-Tsangpo suture zone and the crest of the Himalaya—or Tibetan zone—contains a nearly complete stratigraphic record of the northern continental margin of India over the Paleozoic–Eocene interval (Gaetani and Garzanti, 1991). Exposures of this sedimentary succession have been studied extensively in the Zanskar Range of India, south of Ladakh (approximate long 76°–78°E; Searle, 1983; Gaetani et al., 1985), in north-central Nepal (approximate long 83°–84°E; Bordet et al., 1975; Fuchs, 1977; Fuchs et al., 1988), and in south-central Xizang (approximate long 86°–88°E; Gradstein et al., 1992; Liu and Einsele, 1994; Willems et al., 1996; Jadoul et al., 1998).

For most of this century, Himalayan researchers regarded the base of this "Tibetan sedimentary sequence" as a profound unconformity developed above high-grade metamorphic rocks of the Greater Himalayan sequence (Gansser, 1964; Stöcklin, 1980). Recognition of a major fault system along the southern boundary of the Tibetan zone in the 1980s (Burg et al., 1984a; Burchfiel and Royden, 1985; Searle, 1986; Herren, 1987) has required reassessment of this view, and, as of today, no unambiguous exposure of the base of the Tibetan sedimentary sequence is known to exist. The oldest exposed units are Lower Cambrian, shallow-marine, terrigenous rocks in the Zanskar Range, which pass upward, with minor dolomite intercalations, into Middle Cambrian–Upper Cambrian deep-marine strata (Gaetani and Garzanti, 1991). An unconformity separating these rocks from Cambrian–Ordovician continental deposits has been interpreted as the product of an important phase of orogenesis in north India (Garzanti et al., 1986). Shallow-marine to coastal depositional conditions continued from Ordovician time until the Late Carboniferous–Permian disruption of Gondwana, and development of Neo-Tethys (Şengör et al., 1988).

Rift-related basalts (the Permian "Panjal Traps") are widespread in the Tibetan sedimentary sequence of Zanskar and Kashmir, India (Fig. 2), and rare Permian alkali granites have been found to intrude the section (Spring et al., 1993). Upper Permian–Lower Jurassic strata record first the development of the passive Neo-Tethyan margin and then its deepening to accommodate extensive carbonate platforms (Gaetani and Garzanti, 1991). Deposition of alternating transgressive and regressive sequences marked the Middle Jurassic–Early Cretaceous time period. Increasing continental-margin instability in Aptian–Albian time is indicated by the formation of regionally important unconformities, the influx of continental clastic sediments, and the inception of alkalic volcanism (Garzanti, 1987). These events, probably marking the separation of the Indian plate from Gondwana and the beginning of its northward drift toward Eurasia, were followed in Late Cretaceous time by a major marine transgression (Searle et al., 1988). Marine conditions persisted on the part of the Indian margin exposed in the Zanskar Range until early Eocene (Ypresian) time, when red beds containing ophiolitic debris first appeared in the stratigraphic succession (Gaetani and Garzanti, 1991).

Much of this record is repeated in other preserved sections of the Tibetan sedimentary sequence, but there were some variations in depositional environment along strike. The most significant appears to have occurred in Cambrian–Middle Ordovician time, when that part of the margin now exposed in north-central Nepal and south-central Xizang hosted near-continuous platform-carbonate deposition with no obvious interruption by a major Cambrian–Ordovician orogenic event (Stöcklin, 1980).

Stratified rocks of the Tibetan zone are generally unmetamorphosed, but limited metamorphism has been documented in a few areas. In some places near the crest of the Himalaya, lower Paleozoic rocks contain regional metamorphic assemblages consistent with middle- to lower-amphibolite-facies conditions (Coleman, 1996; Hodges et al., 1996; Carosi et al., 1998; Godin et al., 1999a; Searle, 1999; Searle et al., 1999b). Usually these units are in structural horses within the South Tibetan fault system (see the Structural History section). In some areas, thrust imbrication and large-scale backfolding have produced sufficient structural thickening to promote the development of low- to medium-grade metamorphic assemblages (Schneider and Masch, 1993; Godin et al., 1999b). However, the highest grades of metamorphism in Tibetan sequence rocks are restricted to the carapaces of the so-called North Himalayan gneiss domes.

The North Himalayan Gneiss Domes

A discontinuous belt of metamorphic culminations, referred to as the North Himalayan gneiss domes, can be traced across southern Tibet from at least as far east as long 89°E to at least as far west as long 78°E (Fig. 3). Most occur within the northern half of the Tibetan zone, but at least three have been mapped within the Indus-Tsangpo suture zone in south-central Xizang (Jiao et al., 1988). The earliest directed studies of these features were conducted in south-central Xizang, where about 15 of them were mapped by Chinese and French research teams in the early 1980s (Burg et al., 1984b). One of the most accessible—and thus most extensively studied—of these is the Kangmar dome (lat 28°40′N, long 89°40′E). Burg et al. (1984b) showed that the core of the dome consisted of deformed augen orthogneiss (with a U-Pb zircon age of 562 ± 4 Ma; Schärer et al., 1986) and that it was mantled by progressively less metamorphosed, Carboniferous–Triassic rocks of the Tibetan sedimentary sequence.

A few other domes in the belt display basement complexes of variably deformed orthogneisses and paragneisses with protolith ages that are Early Ordovician or older (Baldwin et al., 1998; Debon et al., 1986). Many also contain low- to medium-grade metamorphic equivalents of upper Paleozoic–Mesozoic Tibetan sedimentary sequence rocks. In the Tso Morari dome of Ladakh and adjacent Xizang (Fig. 3)—the largest of the North Himalayan gneiss domes at well over 12 000 km^2 and perhaps as much as 20 000 km^2—early eclogite-facies assemblages have been overprinted by amphibolite-facies assemblages at conditions similar to those at Kangmar (De Sigoyer et al., 1997; Guillot et al., 1995).

Many of the North Himalayan gneiss domes mapped by the Chinese and French teams in south-central Xizang are dominated by muscovite-biotite granites and leucogranites of Cenozoic age (Burg et al., 1984b; Debon et al., 1986). U-Th-Pb monazite dates are available for three of these, and they are highly discrepant, ranging from ca. 9.5 to 17.6 Ma (Schärer et al., 1986; Harrison et al., 1997a). The contact relationships of these plutons to the surrounding rocks are not well known because none of the granite-cored domes has been mapped in detail. Burg et al. (1984b) described them as intruding into the Tibetan sedimentary sequence country rocks and causing limited contact metamorphism. Similar relationships have been proposed for two of the largest plutons, the Dolpo-Mugu and Mustang leucogranites of north-central Nepal (Fig. 3; Le Fort and France-Lanord, 1994). On the other hand, both Chen et al. (1990) and Burchfiel et al. (1992) speculated that there may be a structural discontinuity between the igneous and metamorphic infrastructures of the domes and their relatively low-grade superstructures.

Greater Himalayan Zone

The metamorphic core of the Himalaya goes by many names—“Central crystallines,” “Higher Himalayan gneisses,” and “Tibetan slab” are a few—but this continuous belt of high-grade metasedimentary and meta-igneous rocks and associated leucogranites is referred to here as the Greater Himalayan zone. Despite having a complex deformational history, the succession displays a remarkably uniform tectonic stratigraphy along strike. The best-characterized sections are found in the deep river gorges that drain the southern flank of the central Nepalese Himalaya. Several decades of French research (Le Fort, 1994) allow these sections to be described in terms of three principal units.

Formation I. The base of the Greater Himalayan sequence consists of predominantly clastic metasedimentary rocks of Formation I. Although mica schists and phyllites, calc-schists, quartzites, para-amphibolites, and subordinate impure marbles are also present, the major rock type in Formation I is biotite-muscovite gneiss. Compositional layering in the unit dips moderately northward in most outcrops. Facing indicators are rare and the quality of exposures is not always conducive to detailed mapping, but Formation I has been regarded traditionally as an intact crustal section with a cumulative thickness ranging from about 1 km to greater than 20 km along strike (Le Fort, 1975). Over the past decade, however, researchers have become increasingly cognizant of low-angle structural discontinuities within the section, with most being interpreted as thrust-sense shear zones (e.g., Burg et al., 1984a; Reddy et al., 1993; Grujic et al., 1996; Searle, 1999). At least some of the extreme variation in along-strike thickness of Formation I may be attributable to these structures, and the rest is probably due to lateral ramping of the Main Central thrust system that defines the base of the Greater Himalayan sequence (Fig. 3).

Formation I rocks typically contain mineral assemblages consistent with middle- to upper-amphibolite-facies metamorphism. The upper part of the unit typically consists of migmatitic gneisses containing between 20% and 75% concordant leucosomes or discrete leucogranitic dikes and sills. Field, petrologic, and geochemical studies of these rocks strongly support the interpretation that they are anatexites (Le Fort, 1975; Le Fort et al., 1987a). That part of Formation I in which melt products first appear varies from place to place. In a few areas, such as the Modi Khola transect of Nepal (Fig. 2), leucogranitic leucosomes occur throughout the section (Hodges et al., 1996). In most others, anatexites are abundant only in the upper part of the section (e.g., Pognante and Benna, 1993).

Formation II. In many areas of the central Nepalese Himalaya, Formation I gneisses are overlain by a 2–4-km-thick sequence of middle- to upper-amphibolite-facies calcareous rocks referred to as Formation II. The predominant rock type is banded calc-silicate gneiss; other lithologies include marble, calc-schist, quartz-rich psammitic schist, para-amphibolite, and orthoquartzite. The contact between Formation I and Formation II is sharp and parallel to compositional layering in both packages. The lack of a metamorphic discontinuity at the transition and the absence of localized tectonite fabrics lead most researchers to join Colchen et al. (1986) in regarding Formation I and Formation II as a conformable package. Formation II is missing entirely from several central Himalayan sections, particularly those in eastern Nepal and adjacent parts of southern Xizang (Burg et al., 1984a; Burchfiel et al., 1992; Lombardo et al., 1993), probably as a consequence of displacements on the overlying South Tibetan fault system, a family of principally extensional structures that marks the contact between the Tibetan and Greater Himalayan zones (Fig. 3).

Formation III. One of the most enigmatic tectonostratigraphic units in the Himalaya, Formation III is a nearly homogeneous augen orthogneiss horizon (with a few metasedimentary intercalations) that usually occurs within the uppermost part of the Formation II sequence or, where Formation II is absent, above the uppermost part of Formation I. It can be traced almost continuously from eastern to central Nepal over a distance of several hundred kilometers (Le Fort et al., 1986), and similar rocks occur at the same structural level as far east as Bhutan (Gansser, 1983) and as far west as Zanskar (Pognante et al., 1990). Many outcrops have the appearance of a deformed granite sill, but the persistence of Formation III as a mappable unit over great distances seems to support the interpretation of Colchen et al. (1986) that it is a volcano-sedimentary horizon within the Greater Himalayan sequence. Available geochronologic data do not shed additional light on this problem. Several Rb-Sr studies are consistent with a Cambrian–Ordovician age for Formation III samples (e.g., Frank et al., 1977b; Ferrara et al., 1983; Pognante et al., 1990). However, attempts to date these rocks by using the U-Pb method (e.g., Hodges et al., 1996) have yielded results equally consistent with either early Paleozoic or Neogene crystallization ages.

Other Regions. Outside the central Himalaya, most researchers have found it difficult to map distinctive equivalents of the Formation I–III trinity, although similar lithologies have been identified. Paragneisses become increasingly predominant in the Arunachal Pradesh State of India (long 92°–95°E; Singh, 1993), and the core of the Namche Barwa syntaxis exposes quartzofeldspathic gneisses, paragneisses, amphibolites, and rare metacarbonate and meta-ultramafic horizons (Burg et al., 1998). In western Ladakh, amphibolite-facies paragneisses like those of Formation I, invaded by numerous leucogranitic rocks, are the predominant rock types (Searle and Fryer, 1986). Much of the Greater Himalayan sequence in this region is covered by erosional outliers of the Tibetan zone and the Neogene–Quaternary Kashmir Basin (Fig. 3). Farther west, upper-amphibolite-facies orthogneisses and paragneisses are abundant in the core of the Nanga Parbat culmination (Misch, 1949; Madin et al., 1989; Wheeler et al., 1995).

It has proved to be impossible, with any degree of confidence, to correlate the metamorphic core of the orogen west of the Nanga Parbat syntaxis to the Greater Himalayan zone of the central Himalaya. Variable proportions of pelitic and psammitic schists and gneisses, orthogneisses, amphibolites, marbles, and quartzites characterize most of this terrain (Treloar et al., 1989a; DiPietro and Lawrence, 1991). Attempts have been made to match these rocks with the Greater Himalayan succession (Coward et al., 1988; Greco et al., 1989), but they also may be metamorphosed equivalents of rocks within the Tibetan zone or Lesser Himalayan zone exposed farther east in the orogen (Pogue et al., 1999). One problematic correlation bears special mention. Metamorphosed mafic rocks containing eclogite-facies assemblages from the upper Kaghan Valley (Fig. 2) are regarded as the metamorphosed equivalents of feeder dikes or flows of the Permian Panjal Traps (Pognante and Spencer, 1991), rocks that occur in the Tibetan zone or North Himalayan gneiss domes but not in the Greater Himalayan sequence in other parts of the western Himalaya. However, Pognante and Spencer (1991) preferred to interpret the metamorphosed mafic rocks as part of the Greater Himalayan sequence. Whether this interpretation is viable has important implications regarding the geodynamics of Himalayan-Tibetan orogenesis because recent investigations have led to the discovery of the ultrahigh-pressure mineral coesite in the upper Kaghan eclogites (O'Brien et al., 1999). If these rocks are indeed part of the Greater Himalayan sequence, they constitute prima facie evidence of the subduction of Indian plate continental crust in the Himalaya to depths of >100 km (Schreyer, 1995).

Age of the Greater Himalayan Sequence. No unambiguous fossils have been found in the Greater Himalayan zone, and its age remains poorly constrained. Parrish and Hodges (1996) showed that Formation I rocks from the central Himalaya contain abundant 0.8–1.0 Ga detrital zircons and must have a Neoproterozoic or younger depositional age. If the Formation I–Formation III succession is more or less structurally intact, as many researchers suggest, and if Formation III is indeed of Cambrian–Ordovician age, it seems probable that most Formation I and II protolith sediments were deposited in Neoproterozoic-Ordovician time.

Greater Himalayan Leucogranites

Besides the migmatitic leucosomes in many exposures of Formation I, discrete leucogranite bodies can be found within all units of the Greater Himalayan sequence and, in a few cases, within basal strata of the Tibetan sedimentary sequence (Dietrich and Gansser, 1981; Le Fort et al., 1987a; Burchfiel et al., 1992; Guillot et al., 1993; Hodges et al., 1996). They occur at all scales, ranging from sills and dikes a few centimeters across to plutons with dimensions of several hundreds of kilometers. Because these granites were produced by the anatectic melting of Greater Himalayan sequence rocks (especially the Formation I pelitic gneisses) during orogenesis (Le Fort et al., 1987a), their age range, relationships to major deformational structures, and spatial distributions have strongly influenced models of the evolution of the Himalaya (Molnar et al., 1983; England et al., 1992; Harris and Massey, 1994; Huerta et al., 1996; Harrison et al., 1997a; Hodges, 1998). As a consequence, they have been subject to extensive field and laboratory research.

Petrology and Geochemistry. Concordant migmatitic leucosomes in Formation I rocks typically contain the assemblage[2] Qtz + Kfs + Pg + Ms + Bt ± Tur ± Grt ± Sil, but kyanite occurs instead of sillimanite in some leucosomes at deep structural levels. Some of the discrete leucogranites (Qtz + Kfs + Pl + Ms ± Bt ± Tur ± Grt ± Sil ± Crd) can be traced into anatexites with high proportions of melt, but most display field characteristics suggestive of the mobilization and transport of leucogranitic magma over distances of meters to kilometers (Le Fort et al., 1987a; Scaillet et al., 1990a). Crosscutting relationships show that multiple generations of leucogranites occur in any single area. Several research groups have divided the discrete bodies into three groups: Ms + Bt granites with little or no tourmaline; Tur + Ms granites; and Ms + Bt + Tur granites (Scaillet et al., 1990b; Hodges et al., 1993; Inger and Harris, 1993; Guillot and Le Fort, 1995). These mineral-assemblage distinctions are not reflected by great differences in major element chemistry; in general, samples with tourmaline have slightly higher SiO_2, Na_2O, and P_2O_5 and slightly lower TiO_2, MgO, CaO, and K_2O compared to those without (Scaillet et al., 1990b; Inger and Harris, 1993; Guillot and Le Fort, 1995; Searle et al., 1997b). All studied samples contain 70–75 wt% SiO_2 and >13 wt% Al_2O_3. Trace element analyses for multiple samples from the same area show wide variations, although tourmaline-bearing samples are generally depleted in Sr and Ba relative to tourmaline-free samples (e.g., Guillot and Le Fort, 1995).

There has been considerable controversy over the role of fluids in the generation of Himalayan leucogranites, and resolution of this problem is important to better understand the thermal evolution of the Himalaya. Early researchers suggested that the leucogranites were produced by fluid-saturated melting of Formation I rocks at temperatures of between 600 and 700 °C (Le Fort et al., 1987a). This model is consistent with the mineral assemblages, major element chemistry, and some of the trace element and isotopic geochemistry of most studied examples (Deniel et al., 1987; Le Fort et al., 1987a; Vidal et al., 1982), as well as phase equilibria and thermobarometric data obtained from most Formation I outcrops, which show evidence for in situ melting (Hodges et al., 1988b, 1988c, 1993; Inger and Harris, 1992; Searle et al., 1992; Metcalfe, 1993; Pognante and Benna, 1993; Macfarlane, 1995; Rai et al., 1998; Vannay and Grasemann, 1998; Manickavasagam et al., 1999). On the other hand, some trace elements (es-

[2] Mineral abbreviations throughout the paper follow the conventions of Spear (1993) and Kretz (1983).

pecially Rb, Sr, and Ba) exhibit behaviors in the Himalayan leucogranites that strongly support a model involving fluid-undersaturated (dehydration) melting at very high temperatures (≥750 °C; Harris and Inger, 1992; Harris et al., 1993; Harris and Massey, 1994). Such a model is also consistent with dehydration-melting experiments conducted on Formation I protoliths (Patiño Douce and Harris, 1998) and with crystallization experiments performed on remelted Greater Himalayan leucogranites (Scaillet et al., 1995). Proponents of fluid-undersaturated melting suspect that most of the available thermobarometric data for Formation I assemblages pertain to final equilibrium temperatures or are artifacts of disequilibrium during cooling (Hodges, 1991; Spear and Florence, 1991) and thus significantly underestimate peak temperature conditions.

Age Constraints. Leucogranite geochronology in the Himalaya stretches back to the 1970s, when the method of choice among most geochronologists for dating granites was whole-rock Rb-Sr (Hamet and Allègre, 1976). This approach had been largely abandoned by the mid-1980s because the Himalayan leucogranites, having complex metasedimentary protoliths, rarely achieved isotopic equilibrium at the whole-rock scale during the melting process (Vidal et al., 1982; Deniel et al., 1987). U-Th-Pb geochronology of accessory minerals such as zircon, monazite, and xenotime has proven to be substantially more robust (Schärer, 1984; Parrish, 1990; Harrison et al., 1995b), but even this approach is not without complications. Most zircons, many monazites, and at least some xenotimes are inherited from the magmatic source regions of the leucogranites or incorporated during emplacement (Parrish, 1990; Copeland et al., 1988), and dating such minerals can therefore overestimate the magmatic ages of their host leucogranites. These same minerals may lose radiogenic Pb by high-temperature diffusion (Parrish and Carr, 1994), and could thus underestimate magmatic ages. With these caveats in mind, apparently reliable U--Th-Pb ages for the Greater Himalayan leucogranites in the central Himalaya range from 22–23 Ma (Harrison et al., 1995b; Hodges et al., 1996; Coleman, 1998; Searle et al., 1999b) to 12–13 Ma (Edwards and Harrison, 1997; Wu et al., 1998). The youngest Greater Himalayan leucogranites (<4 Ma) crop out at the eastern and western ends of the orogen at Nanga Parbat and Namche Barwa (Zeitler et al., 1993; Burg et al., 1998).

Lesser Himalayan Zone

The Lesser Himalayan zone constitutes the foothills of the Himalaya, a physiographic province that is heavily forested or intensely cultivated in most places and thus usually poorly exposed. It mainly consists of lower-greenschist to lower-amphibolite-facies clastic metasedimentary units that define a structurally complex system of fold-and-thrust nappes. Palinspastic reconstructions suggest a cumulative stratigraphic thickness of more than 8–10 km (e.g., Schelling, 1992). The predominant rock types are impure quartzites and psammitic phyllites and schists, with subordinate impure marbles, metamorphosed mafic rocks, and augen orthogneisses (Gansser, 1964; Stöcklin, 1980; Valdiya, 1980; Colchen et al., 1986). Although the basement of this succession is unexposed, it is traditional to assume that it, like the Tibetan zone sequence, was deposited on the north Indian passive margin (Gansser, 1964). However, the dramatic change in sedimentary facies between the Lesser Himalayan and Tibetan sequences with no obvious transition preserved in the Greater Himalayan realm has hampered paleogeographic reconstructions of the north Indian margin throughout the twentieth century (Brookfield, 1993).

Much of the controversy about the stratigraphic relationship between the Lesser Himalayan and Tibetan sequences can be attributed to poor age control for the former. Fossils are extremely rare throughout most of the Lesser Himalayan succession, although some rich and paleoecologically important assemblages represent the Neoproterozoic–Cambrian transition (Tewari, 1993, 1996; Mathur et al., 1997; Gautam and Rai, 1998). Many purported finds of Paleozoic fossils have been proven fraudulent (Jayaraman, 1994), however, and it now appears that much of the sequence consists of rocks of Mesoproterozoic to Early Cambrian age (Brasier and Singh, 1987; Brookfield, 1993; Frank et al., 1995; Parrish and Hodges, 1996; Singh et al., 1999). Thus, the Lesser Himalayan rocks represent a part of the north Indian marginal sequence that is both older and more proximal than that represented by Tibetan zone strata.

In the eastern Lesser Himalaya, the Proterozoic–Cambrian succession is overlain unconformably by a relatively thin (2–3 km) carapace of fossiliferous, Carboniferous–Permian detrital strata that are related, in part, to the opening of the Neo-Tethys (Acharyya and Sastry, 1979; Gansser, 1983).This upper Paleozoic–lower Mesozoic section thins to the west and disappears altogether beneath a pre–Early Cretaceous unconformity near the eastern edge of the western Himalaya (Brookfield, 1993). Considerable controversy surrounds the ages of the limestones and calcareous sandstones above this unconformity. Although some of these strata are fossiliferous, inconsistencies in age assignments and uncertainties in the correlation of sections have led to estimates ranging from late Paleozoic to Paleocene (Stöcklin, 1980; Valdiya, 1980). Above a second unconformity, Eocene–middle(?) Miocene shallow-water turbidites and overlying continental strata represent the earliest stages of Himalayan foreland-basin development (Critelli and Garzanti, 1994; DeCelles et al., 1998a; Najman et al., 1993, 1997).

Lesser Himalayan Crystalline Allochthons

Early Himalayan geologists distinguished three “subzones” within the Lesser Himalayan zone: northern and southern outcrop belts of low-grade Lesser Himalayan strata separated by large, discontinuous tracts of medium-grade metasedimentary rocks, granitic gneisses, and granites (Auden, 1937; Heim and Gansser, 1939; Gansser, 1964). Most of the crystalline terrains are structurally complex synformal klippen that have been thrust southward over less metamorphosed Lesser Himalayan sequence rocks, and they are widely regarded as erosional outliers of the Greater Himalayan sequence (Gansser, 1964; Stöcklin, 1980; Schelling, 1992). However, many of the largest “Lesser Himalayan crystalline allochthons” actually have an internal tectonic stratigraphy that is difficult to relate in any simple way to tectonostratigraphic elements in the Greater Himalayan zone to the north. For example, the Almora allochthon of Kumaun, India (Fig. 3), can be subdivided into basal psammitic and pelitic phyllites and metagraywackes with intercalated augen orthogneisses, followed by porphyritic, cordierite-bearing monzogranite with dikes of tourmaline leucogranite, and an upper sequence of carbonaceous phyllites and quartzites (Valdiya, 1980). Although the metasedimentary rocks of the basal unit may be low-grade equivalents of Formation I rocks, and it could be argued that the augen gneisses are somehow correlative with Formation III, there is no obvious counterpart in the Greater Himalayan sequence to the carbonaceous metasedimentary rocks of the upper unit. In the Kathmandu allochthon of central Nepal (Fig. 3), the succession from bottom to top is (1) pelitic to psammitic schists and phyllites, with subordinate marbles, that are intruded by several large (20–300 km^2) cordierite monzogranites and monzogranitic augen gneisses; (2) very low grade, fine-grained, clastic metasedimentary rocks; (3) argillaceous limestones with rare Middle Ordovician–Late Ordovician fossils; and (4) shales and impure limestones with abundant Silurian fossils (Stöcklin, 1980). The first of these sequences bears superficial resemblance to the basal part of the Greater Himalayan sequence at this longitude (Macfarlane et al., 1992), but the other units are impossible to match with equivalents in the Greater Himalayan realm and instead seem similar to some age-correlative rocks within the Tibetan sedimentary sequence.

Perhaps the two best arguments against correlating the Lesser Himalayan crystalline allochthons with the Greater Himalayan sequence are the difference in metamorphic grade between the two (typically greenschist or lower amphibolite facies in the former, middle to upper amphibolite facies in the latter) and the absence in the Greater Himalayan zone of cordierite monzogranites. These intrusions crop out not only in the Kathmandu and Almora allochthons, but also in nearly every one of the Lesser Himalayan crystalline allochthons (Le Fort et al., 1986). Reliable U-Pb dates for these granites range from 470 to 492 Ma (Schärer and Allègre, 1983; DeCelles et al., 1998b). Although this interval is broadly equivalent with some of the estimated ages for Formation III augen gneisses in the Greater Himalaya, the two granitic suites have distinctive mineral assemblages (Le Fort et al., 1986) and inherited zircon systematics (Hodges et al., 1996; DeCelles et al., 1998b), and it is highly improbable that they are strictly correlative. It seems likely that the Lesser Himalayan crystalline allochthons represent stratigraphically high levels of the north Indian margin that occupied a paleogeographic position north of the source region of the Lesser Himalayan lower-grade nappes and south of the source regions of the Greater Himalayan and Tibetan zones (Upreti and Le Fort, 1999).

Subhimalayan Zone

For the purposes of this paper, the Subhimalayan zone is defined as that part of the Neogene and Quaternary foreland basin of the Himalaya lying between the Lesser Himalayan zone and the "active" thrust front of the orogen. The best-studied sections of the Subhimalayan zone are those in the western Himalaya, which are typically described in terms of two stratigraphic packages: (1) uppermost Paleocene or lower Eocene to lower Miocene siltstones and sandstones of the Rawalpindi Group and (2) lower Miocene to Pleistocene sandstones, conglomerates, siltstones, and mudstones of the Siwalik Group (Burbank et al., 1997). Both packages thicken from south to north, such that the entire sequence ranges in thickness from considerably less than 2 km near the frontal thrust region to more than 10 km near the Lesser Himalayan zone contact. Farther east, the Subhimalayan zone is dominated by the Siwalik Group molasse (DeCelles et al., 1998b).

Intermontane Basins

Neogene–Quaternary intermontane basins occur throughout the Himalaya and southern Tibet (Fig. 2). They can be divided into three broad categories: (1) extensional basins just north of the Himalayan crest that are related to the approximately east-striking South Tibetan fault system (Burchfiel et al., 1992); (2) basins associated with kinematically linked displacement on northwest- and northeast-striking strike-slip faults and north-trending rift systems in southern Tibet (Molnar and Tapponnier, 1978; Fort et al., 1982; Armijo et al., 1986); and (3) "thrust-top" or "piggy-back" basins lying north of the Himalayan thrust front and south of the range crest (Burbank et al., 1997). Basins of the first category are known to contain Pliocene and younger sedimentary fill (Chen, 1981), but are generally not well understood. Most basins of the second type are restricted to topographic lows in southern Tibet and are thus poorly exposed. A notable exception is the Thakkhola basin of north-central Nepal (Fig. 3). Excavated by the Kali Gandaki River (Fig. 2), the Thakkhola basin contains alluvial, colluvial, and lacustrine fill ranging in age from more than 10 Ma to Holocene (Fort et al., 1982; Garzione et al., 1999; J. M. Hurtado, K. V. Hodges, and K. X. Whipple, unpublished data). Of the numerous examples of the third category, only the Kashmir Basin is shown in Figure 3. Both it and the larger Peshawar Basin of Pakistan contain strata equivalent in age and similar in lithology to the upper Siwalik Group. Alluvial-fan, braided-stream, and lacustrine deposits younger than 3 Ma predominate in the Peshawar Basin (Burbank and Tahirkheli, 1985; Pivnik and Johnson, 1995). The basal part of the Kashmir Basin sequence is slightly older, perhaps 5 Ma, but much of that succession is also of late Pliocene–Holocene age (Burbank and Johnson, 1983).

STRUCTURAL HISTORY OF THE HIMALAYA AND SOUTHERN TIBET

Just as Monet's serial paintings provide an incomplete record of time's passage, any attempt to divide the Tertiary structural history of the Himalaya and southern Tibet into discrete deformational episodes may misrepresent the continuous nature of the process. Nevertheless, it seems natural to distinguish three broad phases of deformation in the orogen that are separated by major transitions in deformational style: Protohimalayan, Eohimalayan, and Neohimalayan.

Protohimalayan Phase (Cretaceous–Early Eocene)

The Protohimalayan phase is defined here to include deformation just prior to India-Eurasia collision in the Transhimalaya, Indus-Tsangpo suture zone, and the Tibetan zone. South- to southwest-directed fold-and-thrust structures of Cretaceous age are found throughout the Transhimalayan region, and the available data suggest that most of the documented shortening in the Transhimalayan zone may be of Protohimalayan age (England and Searle, 1986; Searle, 1991; Murphy et al., 1997a). Collision of the Karakoram and Kohistan terranes along the Shyok suture zone has been dated at ca. 75 Ma (Petterson and Windley, 1985; Coward et al., 1987). Thrusting of the Kohistan terrane southward over the north Indian margin along the Main Mantle thrust probably took place in latest Cretaceous or Paleocene time and was certainly completed by ca. 55 Ma (Beck et al., 1995; Searle et al., 1999a).

Protohimalayan structures found in the Zanskar region of the western Himalaya are related to the obduction of the Spontang ophiolite over north Indian margin rocks of the Tibetan zone (Searle, 1986). The Protohimalayan allochthons of Zanskar include not only the relatively intact ophiolite, but a structurally complex sedimentary-tectonic melange of Triassic–Upper Cretaceous rocks as well (Searle et al., 1997a). The melange includes slope-facies rocks of the Indian passive margin, alkalic mafic rocks that may represent ocean-island volcanism, and the remnants of a Neo-Tethys intraoceanic arc (Robertson and Degnan, 1993; Corfield et al., 1999). Several important, nearly flat-lying thrust faults separate distinctive rock packages within the allochthonous stack, and all units have been internally deformed by less significant thrust faults and south-vergent folds (Corfield et al., 1999). The entire stack is unconformably overlain by lower Eocene marine limestones, providing a minimum age for the obduction event, whereas the Late Cretaceous age of the oldest allochthonous strata provides a maximum age (Searle et al., 1997a). A similar age for obduction of the Xigaze ophiolite and related Indus-Tsangpo suture zone melange in south-central Xizang was proposed by Burg and Chen (1984).

Eohimalayan Phase (Middle Eocene–Late Oligocene)

The Eohimalayan phase represents the main India-Eurasia collision and subsequent imbrication of the Indian plate prior to the initiation of north-south extension in the physiographic Higher Himalaya. The precise timing of India-Eurasia collision along the Indus-Tsangpo suture zone has been controversial, largely owing to disagreements regarding the definition of collision. One definition, which will be adopted here, is the transition from marine to nonmarine sedimentation in the suture zone between 54 and 50 Ma (Rowley, 1996; Searle et al., 1997a). However, direct field evidence for suture-zone deformation in this age range is sparse. On the basis of a K-Ar age for synkinematic muscovite, Ratschbacher et al. (1994) assigned a ca. 50 Ma age to southward obduction of the Indus-Tsangpo suture zone rocks and to related south-directed thrusting and tight-to-isoclinal folding of the northernmost sections of Tibetan zone rocks in southern Xizang.

Indirect evidence for the subduction of Indian continental-margin rocks northward beneath the Transhimalayan zone exists in the form of Eohimalayan (55–44 Ma) high- and ultrahigh-pressure eclogite-facies metamorphism of Tibetan zone (and possibly Greater Himalayan zone) rocks in the western Himalaya (Pognante and Spencer, 1991; Guillot et al., 1995). Exactly how these eclogite-facies rocks were juxtaposed with the mid-crustal rocks that currently surround them is unclear, although their exhumation is thought to be an Eohimalayan phenomenon because $^{40}Ar/^{39}Ar$ dates for phengites and biotites from the high-pressure rocks suggest cooling below ~300 °C by ca. 30 Ma (De Sigoyer et al., 1997). Steck et al. (1998) have suggested that the uplift of eclogite-facies units of the Tso Morari dome was accomplished by their buoyant rise between an upper, extensional shear zone and a lower, thrust-sense shear zone, citing as a model the experimental results of Chemenda et al. (1995).

A discrete extensional shear zone of appropriate age for Eohimalayan eclogite exhumation has not yet been mapped in the Tso Morari dome, but candidates for the deep-level thrust structure are abundant and well characterized. In fact, at least three major Eohimalayan fold-and-thrust nappes, all south vergent, have been identified in eastern Ladakh between the Indus-Tsangpo suture zone and the Greater Himalayan zone at the longitude of the Tso Morari dome (Steck et al., 1998). These enormous structures, with fold amplitudes in excess of 10 km and displacements on individual thrusts of as much as several tens of kilometers, can be traced westward as far as western Ladakh and eastward to at least long 79°E (Searle et al., 1988; McElroy et al., 1990; Steck et al., 1993b). In general, the ages of these structures are progressively younger from north to south, and their geometries change as well. To the north, near the Indus-Tsangpo suture zone, the Tibetan zone rocks are deformed by generally upright folds and steeply north-dipping reverse faults. Southward, the axial planes of the folds and the thrust faults dip more shallowly to the north. In the southern one-third of the outcrop belt of the Tibetan zone, the Eohimalayan structures exhibit the classical ramp-flat geometries characteristic of foreland fold-and-thrust belts (Searle et al., 1988; McElroy et al., 1990).

At the southern margin of the Tibetan zone of northwest India, geologists from the University of Lausanne have mapped a stack of northeast-vergent folds and thrust faults that they referred to collectively as the Shikar Beh nappe (Steck et al., 1993a). According to Steck et al. (1999), these structures include shallow-level thrust faults at high structural levels in southeastern Ladakh that cut to progressively deeper structural levels to the northwest and there involve deep-seated rocks of the Greater Himalayan zone. Because the shallow structures are overridden by the frontal thrusts of the southeast-directed allochthons, Steck et al. (1998) and Wyss et al. (1999) regarded the Shikar Beh nappe as an early Eohimalayan feature. However, other research groups working in the region have not recognized the Shikar Beh nappe, and at least one has questioned its existence altogether (Fuchs and Linner, 1995). This controversy notwithstanding, palinspastic reconstructions of the Tibetan zone in the Ladakh region suggest significantly more than 100 km and perhaps more than 200 km of Eohimalayan shortening (Searle, 1986; Searle et al., 1997a; Steck et al., 1998).

South-directed thrust faults and subordinate south-vergent folds of demonstrable or probable Eohimalayan age have been identified throughout the Tibetan zone of the western and central Himalaya (Bally et al., 1980; Shackelton, 1981; Burg and Chen, 1984; Coward and Butler, 1985; Colchen et al., 1986; Ratschbacher et al., 1994; Vannay and Hodges, 1996; Godin et al., 1999b; Yin et al., 1999). Two of these structures deserve special mention. Burg (1983) and Burg and Chen (1984) postulated the existence of a major, east-striking, south-vergent thrust lying south of the band of North Himalayan gneiss domes and separating the Tibetan zone into distinctive northern and southern domains. Ratschbacher et al. (1994) adopted this structure as a major intracontinental thrust system—the Gyirong-Kangmar thrust—in their palinspastic reconstructions of the north Indian margin as exposed in southern Xizang. On the basis of their cross sections and those of Burg and Chen (1984), this structure dips beneath the North Himalayan gneiss domes and is thus a basement-involved thrust, with a minimum displacement of at least 20 km. However, the surface expression of this structure has not been described in detail by geologists who have worked in southern Xizang, and the principal evidence for its existence seems to be a difference in the apparent stratigraphic thickness of Mesozoic strata of the Tibetan zone in that region and the increase in metamorphic grade around the Kangmar dome (Burg and Chen, 1984; Ratschbacher et al., 1994). Given the active debate concerning the origin of the North Himalayan gneiss domes (as discussed subsequently), the relatively limited understanding of the cause of variations in Mesozoic stratigraphic thicknesses in the Xizang sector of the Tibetan zone, and the fact that no unambiguous reflector that might represent the downdip projection of the Gyirong-Kangmar thrust is apparent in the INDEPTH seismic reflection profile through the region (Hauck et al., 1998), more work is needed to demonstrate the regional significance of the thrust system.

Major structural significance also has been attributed to the Gangdese thrust, a south-directed feature along which intrusive rocks of the Gangdese batholith, its Transhimalayan country rocks, and Cretaceous forearc deposits of the Indus-Tsangpo suture zone moved southward over Tibetan zone strata (Yin et al., 1994). With a well-exposed, shallowly north-dipping outcrop trace marked by mylonites and cataclasites containing fabrics indicative of southward displacement (between S20°E and S20°W), the Gangdese thrust has the appearance of a major rooted thrust system in the Zedong region of south-central Xizang (approximately lat 29°15′N, long 92°E; Yin et al., 1999). On the basis of the inference that an abrupt termination of the Xigaze Group forearc strata in this region can be attributed to overthrusting along the Gangdese fault, Yin et al. (1994) have inferred a minimum displacement of ~46 km. The age of the Gangdese thrust in the Zedong region is constrained to be younger than a 31 Ma hanging-wall granodiorite, and, by attributing a rapid phase of cooling inferred from multidomain diffusion modeling of K-feldspars from hanging-wall rocks to erosional unroofing related to thrust emplacement, Yin et al. (1994, 1999) suggested that most of the thrust displacement occurred during the 28–24 Ma interval. Although Yin et al. (1994) regarded the Gangdese thrust as an important crustal-scale feature, this interpretation has been difficult to confirm through geologic mapping in other parts of southern Tibet. In the Kailas region of western Xizang, no comparable structure can be found, although Yin et al. (1999) inferred that it exists at depth but has been overridden by Neohimalayan, north-directed thrusts. Still farther west, mapped features at the appropriate structural level include Protohimalayan thrusts within the suture zone and Neohimalayan backthrusts, transcurrent faults, and extensional detachments, but no obvious correlative to the Gangdese thrust (Gansser, 1964; Frank et al., 1977a; Thakur and Virdi, 1979; Searle et al., 1988; Searle, 1991; Steck et al., 1993b, 1998).

Neohimalayan Phase (Early Miocene–Present)

Neohimalayan structures are found throughout all tectonostratigraphic zones of the Himalaya, and activity on these structures has largely dictated the structural architecture of the orogen.

South-Vergent Shortening Structures. The best known and most significant north-south shortening structures of Neohimalayan age are the east-striking thrust systems that separate the Greater Himalayan, Lesser Himalayan, and Subhimalayan zones from one another. Less significant Neohimalayan thrusts have been identified in all three of these zones, and both mesoscopic and macroscopic folds provide additional evidence for significant internal strain (Burg et al., 1984a; Brun et al., 1985; Macfarlane, 1993; Reddy et al., 1993; Vannay and Steck, 1995; Coleman, 1996; Hodges et al., 1996; DeCelles et al., 1998a; Searle, 1999; Wyss et al., 1999; among many others).

The structurally highest and oldest of the major faults is the Main Central thrust system (MCTS in Fig. 3), which marks the Greater Himalayan–Lesser Himalayan contact from Bhutan to the Kashmir region of the western Himalaya. Farther east, the quality of mapping is not yet sufficient to demonstrate how the system extends into the eastern syntaxial region. Farther west, the existence and significance of the Main Central thrust system are unclear (Pogue et al., 1999). Exposures of the Main Central thrust system are generally not good in the eastern Himalaya because the structure crops out along the heavily forested or heavily cultivated transition between the Higher Himalayan ranges and their foothills. Some of the best outcrops can be found along lateral ramps in the system, such as that in the Trisuli River drainage of the central Nepalese Himalaya (Fig. 2; Macfarlane et al., 1992). The level of exposure is dramatically better in the western Himalaya, where the Main Central thrust system crops out both along its main east-trending trace and around the margins of large fensters through the Greater Himalayan thrust sheet, such as the Kishtwar window (Kündig, 1989; Searle and Rex, 1989; Staübli, 1989; Stephenson et al., 2000; Wyss et al., 1999). In all well-studied examples, the Main Central thrust system consists of a broad shear zone, ranging from several hundreds of meters to several kilometers in thickness, that is developed in a tectonic melange of units derived from both the Greater and Lesser Himalayan sequences (Arita, 1983; Brunel, 1986; Grujic et al., 1996; Hodges et al., 1996; Hubbard, 1989; Jain and Manickavasagam, 1993; Macfarlane et al., 1992; Pêcher, 1978; Schelling and Arita, 1991; Stephenson et al., 2000; Valdiya, 1980; Vannay and Grasemann, 1998; Vannay and Hodges, 1996; Wyss et al., 1999). The roof and sole faults of the shear zone dip moderately northward, subparallel to intense shear fabrics internal to the zone. Kinematic indicators in these tectonites typically indicate southwestward or southeastward displacement. Detailed structural analyses show that the Main Central thrust system has had a complex, polyphase deformational history (e.g., Brunel, 1986; Brunel and Kienast, 1986; Grasemann et al., 1999; Hodges et al., 1996; Macfarlane et al., 1992; Wyss et al., 1999). The oldest dated structures are discrete, amphibolite-facies shear zones that developed between 23 and 20 Ma, synchronous with regional metamorphism and the early stages of Neohimalayan anatexis in the Greater Himalayan sequence (Hodges et al., 1996; Hubbard and Harrison, 1989). There is ample evidence for additional southward displacement along less well-defined shear zones that developed at garnet-grade (or lower) metamorphic conditions, and cataclastic faults of uncertain vergence are found throughout the Main Central thrust system shear zone as well (Brunel and Kienast, 1986; Hodges et al., 1996; Macfarlane et al., 1992; Wyss et al., 1999). It has been known for many years that $^{40}Ar/^{39}Ar$ cooling ages for minerals in rocks within the Main Central thrust system shear zone are substantially younger (late Miocene or Pliocene) than those for minerals from structurally higher rocks (Hubbard and Harrison, 1989; Maluski et al., 1988; Vannay and Hodges, 1996). Although some workers have attributed these young ages to late-stage deformation along the Main Central thrust system (e.g., Macfarlane et al., 1992), others have suggested that they may be related to late hydrothermal activity in the zone (Copeland et al., 1991). Th-Pb ion-microprobe ages for synkinematic monazites now confirm the significance of late Miocene–Pliocene slip on the Main Central thrust system in many sectors of the Himalaya (Catlos et al., 1999; Harrison et al., 1997b). The age of the youngest deformation in the Main Central thrust system shear zone is unknown. Seeber and Gornitz (1983) pointed out that a distinctive knickpoint in the gradients of major rivers draining the southern flank of the Himalaya generally corresponds to the trace of the Main Central thrust system, and they suggested that at least some segments of the system may still be active. Their hypothesis is supported by a high concentration of landslide and hydrothermal activity along the system, by a discontinuity in the slope of Himalayan topography across the Main Central thrust system, and by the fact that geodetic studies imply a sharp transition in the modern kinematics of the Himalaya at the approximate position of the Main Central thrust system (Bilham et al., 1997). Many relatively straight strands of the Miocene Main Central thrust system show evidence of recent displacement, but others with more complex geometries have been abandoned in favor of newer, less contorted slip planes.

Evidence for large displacements on the Main Central thrust system is provided by a series of half-klippen exposed by longitudinally inconsistent erosion through the Greater Himalayan thrust sheets (Fig. 3). Their geometries imply a minimum of several tens of kilometers to a maximum of 150–250 km of cumulative slip on the Main Central thrust system (Brunel and Kienast, 1986; Molnar, 1984; Schelling, 1992), although how this displacement was partitioned among different deformational phases over middle Miocene to Holocene time remains unknown.

The contact between the Lesser Himalayan zone and the Subhimalayan zone is marked by north-dipping thrust faults of the Main Boundary thrust system (MBTS in Fig. 3). This system can be traced for even longer distances along strike than the Main Central thrust system (Gansser, 1983; Meigs et al., 1995; Valdiya, 1992), but good, continuous outcrops are found only in the western Himalaya. Where exposed, the Main Boundary thrust system is generally marked by a narrow (~100 m or less) zone of cataclasis that typically dips moderately to steeply northward and is, in some cases, overturned to dip steeply southward (Schelling, 1992; Valdiya, 1992). A shallowly (<35°) northward regional dip of the Main Boundary thrust system is inferred from palinspastic reconstructions of the frontal thrust system of the Himalaya (Schelling, 1992; DeCelles et al., 1998a; Srivastava and Mitra, 1994). Faults in the system typically place low-grade Lesser Himalayan rocks on different members of the Siwalik Group with a sharp discontinuity in stratal dips. The most recent movement on the Main Boundary thrust system is constrained to be younger than the Pliocene molasse strata that it cuts (e.g., DeCelles et al., 1998a), but there is little hard evidence regarding its Pliocene–Holocene movement history. On the basis of sedimentation patterns in the Subhimalayan zone, Meigs (1995) suggested that the Main Boundary thrust system may have developed as early as 11–9 Ma. Total amounts of thrusting on the Main Boundary thrust system are unknown, because no rocks in the hanging wall can be matched to rocks in the footwall and because no large, dip-parallel exposures that might provide geometric constraints on structural overlap have been identified. However, reconstructions across the orogenic front interpret the throw on the Main Boundary thrust system to have been at least several tens of kilometers and perhaps much more (DeCelles et al., 1998b; Molnar, 1984; Srivastava and Mitra, 1994).

The Main Frontal thrust system separates the Subhimalayan zone from the Indo-Gangetic Plain and represents the toe of the Himalayan orogenic wedge. Actual exposures of the Main Frontal thrust system are extremely rare—so much so that the Main Frontal thrust system is not drawn as a continuous feature in Figure 3—but those exposures that do exist have well-defined scarps cutting river terraces and alluvial fans (Nakata, 1989). More commonly, the geometry of the system is inferred from the geomorphology and structural geology of its hanging wall (Yeats et al., 1992). There is no direct geologic evidence pertaining to the initiation age of slip on the Main Frontal thrust system, although it is usually assumed to be a Pliocene–Holocene structure (Molnar, 1984). Published cross sections of the Himalayan front typically show the Main Frontal thrust system as a decollement thrust with no basement involvement at least as far north as the downdip projection of the surface trace of the Main Boundary thrust system (e.g., Yeats and Lillie, 1991). Cross sections drawn across the entire Himalayan orogen typically depict the Main Frontal thrust system as the surface expression of a low-angle, basal thrust along which the Indian plate is subducted beneath the Himalaya and southern Tibet and into which the Main Boundary thrust system and Main Central thrust system root (Coward et al., 1988; DeCelles et al., 1998a; Molnar, 1984; Schelling, 1992; Schelling and Arita, 1991; Srivastava and Mitra, 1994). In this model, the basal thrust—referred to hereafter as the Himalayan Sole thrust—must become basement-involved north of the downdip projection of the Main Central thrust system, or approximately at the latitude of the Himalayan range crest. It has become a tradition among Himalayan geologists to invoke a ramp in the Himalayan Sole thrust just south of this position, both to explain the basement involvement and to provide a mechanism for the generally steeper dip of rock units north of the transition (e.g., Lyon-Caen and Molnar, 1983; Molnar, 1984), but other geometries are possible. Geodetic measurements imply that much of the modern convergence between India and Eurasia is concentrated just south of the Himalayan range crest and that substantially faster rates of uplift prevail north of the surface trace of the Main Central thrust system (Bilham et al., 1997). These phenomena have been attributed to locking and strain accumulation on the ramp in the Himalayan Sole thrust (Jackson and Bilham, 1994), but at least some of the geodetic data could be explained just as reasonably by recent activity on the Main Central thrust system. Although the reflection seismic data gathered during the first phase of the INDEPTH project (Zhao et al., 1993) did not extend far enough south to help constrain the subsurface geometry of the Himalayan Sole thrust where the ramp is thought to occur, they do reveal a set of reflectors, extending northward beneath the High Himalaya, that are interpreted as confirming the northward projection of the Himalayan Sole thrust to depths of at least 45 km before it disappears at approximately lat 28.6°N (Hauck et al., 1998).

The relatively well-defined initiation age for the Main Central thrust system (early Miocene), the less well-constrained initiation age of the Main Boundary thrust system (late Miocene–Pliocene), and the inferred initiation age of the Main Frontal thrust system (Pliocene–Holocene) are consistent with traditional models of fold-and-thrust belts in which the thrust front propagates toward the foreland with time (Dahlstrom, 1970). However, temporal variations in the principal location of shortening appear more complex when studied in detail. As outlined above, there is much evidence that the Main Central thrust system accommodated significant shortening in late Miocene, Pliocene, and perhaps even Pleistocene–Holocene time. Out-of-sequence thrusts of Miocene age, younger than the structurally lower Main Central thrust system, have been mapped at near the top of the Greater Himalayan sequence in Nepal and Bhutan (Brun et al., 1985; Grujic et al., 1996; Hodges et al., 1996; Searle, 1999; Vannay and Hodges, 1996).

One of the most important unresolved questions of Himalayan tectonics is how the thrust structures within and at the base of the Lesser Himalayan crystalline allochthons relate to the Main Central thrust system. In the traditional view, rock units in the allochthons are klippen of Greater Himalayan zone rocks, and the basal thrusts of the allochthons are part of the Main Central thrust system (Gansser, 1964). This interpretation increases substantially the amount of structural overlap of the Greater Himalayan sequence relative to the Lesser Himalayan sequence and would require a minimum of 125 km of slip on the Main Central thrust system (Lyon-Caen and Molnar, 1983). Alternatively, as reviewed in a previous section, several lines of evidence may be used to argue instead that the Lesser Himalayan crystalline allochthons have a provenance different from that of the exposed Greater Himalayan sequence. Upreti and Le Fort (1999) suggested that the basal thrusts of the Lesser Himalayan allochthons represent a separate thrust system (their "Mahabarat thrust"), with an initiation age intermediate between that of the Main Central thrust system and that of the Main Boundary thrust system. However, their contention that the Mahabarat thrust has no exposed root zone, as well as their developmental cross sections, require that the latest movement on the Main Central thrust system must actually postdate the latest movement on the Mahabarat thrust. This hypothesis should be testable through detailed geochronologic investigations of fault-related fabrics.

The extent of Neohimalayan, south-directed thrusting within the Tibetan zone, the Indus-Tsangpo suture zone, and the Transhimalaya is poorly known. Many north-dipping thrusts in these zones affect Eocene and older strata and may be either Eohimalayan or Neohimalayan features, or a combination of the two (e.g., Ratschbacher et al., 1994; Searle, 1986). Indirect evidence for Neohimalayan shortening and concomitant erosional denudation in the Transhimalayan zone comes from $^{40}Ar/^{39}Ar$ cooling ages for intrusive rocks of the Gangdese batholith (Copeland et al., 1995).

North-Vergent Shortening Structures. South-dipping reverse faults and kilometer-scale upright folds, overturned to the north, are common in the northern Tibetan zone and Indus-Tsangpo suture zone of southern Tibet (Heim and Gansser, 1939; Gansser, 1964; Bally et al., 1980; Searle, 1983; Burg and Chen, 1984; Girardeau et al., 1984a). The highest concentration of these structures occurs along the southern boundary of the Indus-Tsangpo suture zone in the Ladakh-Zanskar region of India (Searle, 1986; Searle et al., 1988, 1997a), in western Xizang (Yin et al., 1999), and in south-central Xizang (Ratschbacher et al., 1992; Yin et al., 1994; Quideleur et al., 1997). Yin et al. (1999) regarded these structures as marking an orogen-scale fault system that they referred to as the Great Counter thrust system. Although some of the principal faults of this system dip as shallowly as 28° southward at the surface (Yin et al., 1999), most are actually moderate- to high-angle reverse faults (Girardeau et al., 1984a; Searle et al., 1997a; Yin et al., 1999), and their significance with respect to overall Neohimalayan shortening in the Himalaya appears limited. An inability to match units unambiguously across the larger structures in the system precludes quantitative estimates of displacement. Most published cross sections require no more than a few kilometers (Ratschbacher et al., 1994; Yin et al., 1999), but a recent geometric interpretation by Makovsky et al. (1999) would require net displacements that are as much as an order of magnitude higher. Detailed studies in the Ladakh-Zanskar area suggest that north-vergent and south-vergent Neohimalayan shortening structures in the northern Tibetan zone and Indus-Tsangpo suture zone are related and together define a large-scale "pop-up" structure responsible for about one-third of the total Tertiary shortening across the Tibetan zone in this area (Searle et al., 1990, 1997a). About half to two-thirds of this amount (perhaps 20–30 km) might be attributable to the north-vergent structures.

The age of the backthrusts and related folds is debated. Noting that these structures deform the entire sequence of Indus Group molasse in the Zanskar-Ladakh region, Searle et al. (1997a) assigned them a Pliocene–Pleistocene age. Quideleur et al. (1997), on the other hand, documented a younging of $^{40}Ar/^{39}Ar$ biotite and K-feldspar dates near one strand of the backthrust system in south-central Xizang—the Renbu-Zedong thrust—that they attributed to 19–10 Ma faulting. Yin et al. (1999) suggested that the principal backthrust in the Kailas region of western Xizang—the Kailas thrust—was active during and after deposition of the upper part of the lower Miocene Kailas conglomerates. Multidomain diffusion modeling (Lovera et al., 1989) of K-feldspar $^{40}Ar/^{39}Ar$ data for a clast from the Kailas conglomerates in the footwall of the Kailas thrust is consistent with post–20 Ma burial and heating, which Yin et al. (1999) attributed to thrust displacement. A minimum age for the Kailas thrust was inferred to be ca. 4 Ma by Yin et al. (1999) because the thrust is truncated by the Karakoram fault, which may have an inception age of no more than 4 Ma (Searle, 1996). Searle et al. (1998) refined the estimated inception age of the Karakoram fault upward to ca. 11 Ma, which might suggest an even older age for backthrusting. However, there are few hard constraints (at present) on how slip on the Karakoram fault has been partitioned over the Miocene–Holocene interval, and thus the relative age relationship between that fault and the Kailas thrust does not preclude substantial post-Pliocene slip on the backthrust system.

Structures Related to North-South Extension. The most thought-provoking deformational features in the Himalayan orogen are north-dipping normal faults and related folds of Neohimalayan age. Although most widely distributed in the Tibetan zone (Burchfiel et al., 1992), they also have been found in the Indus-Tsangpo suture zone, the Greater Himalayan zone, and the Lesser Himalayan zone (Nakata, 1989; Guillot et al., 1997; Steck et al., 1998). In addition to the poorly understood Eohimalayan extensional shear zones that may have played a role in the exhumation of high- and ultrahigh-pressure eclogites in the western Himalaya (Steck et al., 1998), there are five classes of north-south extensional structures in the Himalaya.

Class I: The South Tibetan Fault System and Related Structures. The existence of normal faults separating the Tibetan and Greater Himalayan zones was first recognized in north-central Nepal (Caby et al., 1983) and later documented in southern Xizang (Burg et al., 1984a; Burchfiel et al., 1992) and northwest India (Searle, 1986; Herren, 1987; Valdiya, 1989). They are referred to here collectively as the South Tibetan fault system (STFS in Fig. 3). In most well-studied examples, the basal detachment of the South Tibetan fault system—which typically separates unmetamorphosed or weakly metamorphosed Tibetan zone strata of the hanging wall from upper-amphibolite-facies gneisses and leucogranites of the Greater Himalayan sequence footwall—is exposed very near the crest of the Himalaya (Hodges et al., 1992; Pognante and Benna, 1993; Searle et al., 1997b; Searle, 1999). It is represented by a shallowly north-dipping brittle fault underlain by a subparallel mylonitic carapace in the uppermost 500–1000 m of the footwall; in most cases, well-developed shear-sense indicators are consistent with northeastward or northwestward displacement of the hanging wall in a normal sense (e.g., Burchfiel et al., 1992). Because of the geographic coincidence of many of the basal detachments with the range crest and the relatively subdued relief north of the crest, most of these structures cannot be traced far downdip, and their net displacements are thus poorly known. Important exceptions occur in the Mount Everest region, where components of the South Tibetan fault system can be traced parallel to their slip vectors from the summit region of Mount Everest to the northern end of the Rongbuk Valley of southern Xizang (Carosi et al., 1998; Hodges et al., 1998; Searle, 1999). The fact that footwall and hanging-wall rocks cannot be reconstructed along this traverse requires minimum displacements of ~35–40 km.

Many segments of the South Tibetan fault system are marked by complex arrays of synthetic and antithetic splay faults that divide the immediate hanging wall of the basal detachment into extensional riders. Several examples of this phenomenon were documented by Burchfiel et al. (1992), and imbricate South Tibetan structures have been mapped subsequently by Hodges et al. (1996), Carosi et al. (1998), Steck et al. (1998), and Searle (1999). The cumulative extension represented by such hanging-wall features may be of comparable magnitude to the slip on the basal detachments (Hodges et al., 1998; Girard et al., 1999; Searle, 1999).

Kinematic analyses of some segments of the South Tibetan fault system indicate a more complicated history than simple downdip extension. In north-central Nepal, there is clear evidence for multiple displacement episodes with either (1) alternating top-to-the-north, predominantly normal displacement and dextral or sinistral transcurrent displacement or (2) oblique displacement with greater or lesser dip-slip components (Stutz and Steck, 1986; Pêcher, 1991; Coleman, 1996). In the Annapurna Range, an episode of south-directed, break-back thrusting occurred along the Tibetan zone–Greater Himalayan zone contact both before and after extensional faulting at the same structural level (Hodges et al., 1996). Such complex deformational histories suggest that the South Tibetan fault system is best interpreted as the surface trace of a long-lived, crustal-scale decoupling horizon between the upper crust and the middle-lower crust of the Tibetan Plateau, such that the fault system's kinematics may vary in time and space to accommodate the differential response of the two layers to an evolving stress field (Hodges, K. V., Hurtado, J. M., and Whipple, K. X., unpublished data).

Although most data suggest that the South Tibetan fault system was active by Miocene time (Guillot et al., 1994; Harrison et al., 1995c; Hodges et al., 1996, 1998; Edwards and Harrison, 1997; Searle et al., 1997b; Coleman, 1998; Wu et al., 1998; Murphy and Harrison, 1999; Walker et al., 1999), the duration of activity on the system remains poorly understood. In the Annapurna and Dhaulagiri Ranges of central Nepal (Fig. 2), Pleistocene displacement can be demonstrated on one segment of the system (J. M. Hurtado, K. V. Hodges, and K. X. Whipple, unpublished data). It seems likely that much of the system has been active episodically over the Miocene–Holocene interval.

Class II: Marginal Faults of the North Himalayan Gneiss Domes. As part of their study of the Kangmar dome of southern Xizang, Burg et al. (1984b) documented a change in fabric orientations across the contact between the orthogneiss in the core of the dome and its metasedimentary carapace. Unable to decide whether the contact was structural or depositional, they suggested that the fabric discontinuity may be caused by cleavage refraction. Some subsequent workers have regarded the contact as an important extensional detachment (Chen et al., 1990; Wang et al., 1997; Guillot et al., 1998), whereas others have documented brittle and ductile tectonite fabrics at the contact but have interpreted the relationship as a modified unconformity with limited displacement (Lee et al., 1998, 1999). At Tso Morari in Ladakh, the only other North Himalayan gneiss dome studied in detail as of this writing, the comparable infrastructure-superstructure contact is marked by well-defined extensional structures, including brittle faults similar to the basal detachments of the South Tibetan fault system that are underlain by subparallel mylonitic shear zones (Guillot et al., 1997; Steck et al., 1998).

Such structural relationships, as well as the geomorphology of the North Himalayan gneiss domes, are strikingly similar to those of the metamorphic core complexes of western North America (Coney, 1980), so much so that Chen et al. (1990) were prompted to propose a similar origin for the Kangmar dome. Burchfiel et al. (1992) suggested that all of the North Himalayan gneiss domes may be metamorphic core complexes. Some workers have adopted this interpretation (Wang et al., 1997; Guillot et al., 1998), but others have not. Burg et al. (1984b) suggested that the Kangmar dome was a fault-bend fold developed above either a simple ramp or a thrust duplex system on the Gyirong-Kangmar thrust. On the basis of an interpretation of INDEPTH deep seismic reflection profiles, Hauck et al. (1998) related the doming to a ramp or duplex developed on the structurally lower Himalayan Sole thrust. They went on to suggest that the detachment described at Kangmar by Chen et al. (1990) is an exposure of the basal detachment of the South Tibetan fault system that was exhumed by the doming process. In contrast, Makovsky et al. (1999) attributed the doming to duplex development along a north-vergent backthrust system. Lee et al. (1999) explicitly rejected both the core-complex model and the duplex model, instead interpreting "the formation of the extensional fabrics [at Kangmar] as a consequence of maintaining a stable wedge geometry or dynamic equilibrium between vertical thinning and horizontal stretching at midcrustal depths and underplating and thickening at deep crustal levels." Edwards et al. (1999) have suggested that the detachment exposed at Kangmar is a regionally important structure—the Karo-La decollement—that crops out in the cores of several different North Himalayan gneiss domes because it has been domed by the emplacement of discrete granite plutons.

Class III: Longitudinal Normal Faults North of the South Tibetan Fault System. Among the most poorly characterized extensional features in southernmost Tibet are roughly east-striking, north-dipping normal faults that occur sporadically throughout the region north of the Himalayan crest and south of the Indus-Tsangpo suture. Although their displacement histories are unknown, several of these structures have surface traces that are several tens of kilometers in length, and one fault that appears on several maps of south-central Xizang and crops out a few kilometers north of the Kangmar dome has a surface trace in excess of 300 km (Burg and Chen, 1984; Jiao et al., 1988; Burchfiel et al., 1992). Cross sections through a large imbricate fan of class III faults, mapped as the Dutung-Thaktote extensional fault zone in the Ladakh region of the northwest Himalaya (Steck et al., 1998), suggest that this structural suite was responsible for as much as 16 km of extension (Girard et al., 1999).

Cenozoic intermontane basins are common in southernmost Tibet, and although many are rift basins associated with the generally north-striking faults described in the next section, at least some are supradetachment basins related to class III faults. The best-documented example thus far is the Gyirong basin of south-central Xizang (Fig. 3; Burchfiel et al., 1992). The Gyirong basin, with a total stratigraphic thickness of about 1 km, includes basal megabreccia deposits overlain by fluvial, colluvial, and lacustrine strata of Miocene(?)–Pleistocene age (Chen, 1981; Wang et al., 1981; Mercier et al., 1987). Its principal growth fault is well exposed along the southern margin of the basin as an east-striking, 42°N-dipping normal fault placing footwall-derived megabreccia sheets and other basinal strata on Jurassic limestones of the Tibetan sedimentary sequence (Burchfiel et al., 1992). Although the field relationships require a Miocene–Pliocene initiation age for this particular fault, most examples of class III extensional structures remain undated.

Class IV: Rift Systems of Southern Tibet. A series of prominent, north-trending rifts were first recognized on satellite images of the Tibetan Plateau a quarter-century ago, and seismicity in southern Tibet suggests that east-west extension is the dominant mode of modern deformation in the plateau region (Molnar and Tapponnier, 1975, 1978; Ni and York, 1978). The structural characteristics and kinematics of these fault systems have been the subjects of several regional studies in southern Xizang (Armijo et al., 1986; Mercier et al., 1987; Ratschbacher et al., 1994), and local studies have focused on the Thakkhola graben of north-central Nepal (Fort et al., 1982; Garzione et al., 1999; J. M. Hurtado, K. V. Hodges, and K. X. Whipple, unpublished data), the Yadong graben of southern Xizang (Burchfiel et al., 1991; Wu et al., 1998), and the Yangbaijing-Gulu graben of central Xizang (Pan and Kidd, 1992; Harrison et al., 1995a).

Because models of Tibetan Plateau evolution commonly attribute east-west extension in Tibet to gravitational spreading after the plateau had reached its maximum elevation (Harrison et al., 1992; Molnar et al., 1993), arguments regarding the timing of plateau uplift often revolve around the initiation age of east-west extension. Estimates based on the cooling histories of rocks from one rift flank in central Xizang have figured prominently in papers promoting a ca. 8 Ma date for the maximum elevation of the plateau (Harrison et al., 1992, 1995a; Molnar et al., 1993). However, $^{40}Ar/^{39}Ar$ cooling ages for micas that provide a minimum age for east-west extension in the Tibetan Plateau region of north-central Nepal have been used to suggest a pre–14 Ma age for plateau uplift (Coleman and Hodges, 1995; Searle, 1995). Harrison et al. (1995a) dismissed the evidence from Nepal and preferred instead to relate pre–late Miocene, east-west extension to "incipient collapse of a narrow mountain belt" prior to development of the Tibetan Plateau (Yin et al., 1994) and even argued that the Thakkhola graben was not produced by the same mechanism as more northern grabens because it seemed to have been active prior to 8 Ma. The fact of the matter is that there are simply too few pertinent geochronologic data at present to justify assumptions regarding the initiation age of rifting in Tibet. We have no reason to believe, for example, that any of the handful of east-west extensional features that have been dated in southern Tibet are the oldest. Even if we eventually achieve a generally comprehensive knowledge of the timing of east-west rifting on the plateau, understanding its significance with regard to plateau uplift is far from straightforward. Although a direct connection between plateau uplift and east-west extension is geodynamically sensible (England and Houseman, 1989; Houseman and England, 1993b; Royden et al., 1997), other interpretations of the cause of east-west extension are plausible (McCaffrey and Nabelek, 1998). Moreover, evidence is growing in support of pre–middle Miocene, perhaps even pre-Neohimalayan, uplift of some parts of the plateau (e.g., Chung et al., 1998)

Class V: Neotectonic Features of the Himalaya South of the South Tibetan Fault System. Tectonic geomorphology along the southern flank of the Himalaya has revealed a remarkable array of neotectonic features north of the Main Frontal thrust system. They include some northwest-striking faults with relatively straight topographic expressions and evidence of both right-lateral and normal-sense movement, but the preponderance of neotectonic features mapped in the Lesser Himalaya are generally west- to northwest-striking, steeply north- or south-dipping normal faults (Nakata, 1989; Yeats et al., 1992). Although concentrated along the surface traces of the Main Boundary thrust system, such features also have been mapped within the Lesser Himalayan and Greater Himalayan zones. The age range, amount of displacement, and overall tectonic significance of these faults remain poorly understood.

Major Transcurrent Faults. Although transcurrent faults have played a fundamental role in the development of Tibet (Molnar and Tapponnier, 1975; Peltzer and Tapponnier, 1988; Armijo et al., 1989; Avouac and Tapponnier, 1993), relatively few have been mapped in southernmost Tibet and the Himalaya. The best known of these is the Karakoram fault, a northwest-striking structure that extends over a distance of 1000 km from the Pamir in the northwest to Gurla Mandhata, one of the larger of the North Himalayan gneiss domes, in southern Xizang (Fig. 3). Numerous offset geomorphic and geologic features demonstrate that the dominant displacement along the fault is dextral, although both transtensional and transpressional segments have been identified (Searle et al., 1998). Peltzer and Tapponnier (1988) deduced a right-lateral displacement of roughly 1000 km on the basis of proposed correlations of offset granitic rocks, but Searle (1996) questioned the correlations and, thus, the estimate of slip. Better known offset markers imply much less displacement (<150 km; Searle et al., 1998). Avouac and Tapponnier (1993) used offset geomorphic features to estimate a modern slip rate of ~3.2 cm/yr for the Karakoram fault. If this rate is extrapolated backward in time, the fault need not be older than ca. 4 Ma (Searle, 1996). On the other hand, Searle et al. (1998) postulated an inception age of ca. 11 Ma, which would imply either that the slip rate of Avouac and Tapponnier (1993) is an overestimate or that the rate of slip has accelerated substantially with time.

Some researchers have argued for large dextral displacements on the South Tibetan fault system, in some cases emphasizing the greater importance of transcurrent slip compared to normal slip (Steck et al., 1993a). Most evidence comes from the trajectory of tectonite fabrics in the Greater Himalayan zone footwall; for example, Pêcher (1991) documented a rotation of the dominant lineation in the Greater Himalayan sequence of central Nepal from approximately north trending in the middle of the sequence to nearly east trending at the top of the sequence near the basal detachment of the South Tibetan fault system. In the absence of offset markers, it is difficult to evaluate the hypothesis that such rotation implies large-scale dextral slip. Working in the Annapurna Range to the west of Pêcher's study area, Coleman (1996) found evidence for phases of both normal and sinistral—not dextral—displacement on the South Tibetan fault system.

Although the Indus-Tsangpo suture zone was established in Eohimalayan time, there is substantial evidence for reactivation of faults within it over Neohimalayan time and for the development of new fault systems, like the backthrusts described above (Searle, 1986; Ratschbacher et al., 1994; Yin et al., 1999). Transcurrent faulting along the suture, particularly in the central and eastern Himalaya, is generally underemphasized in papers on south Tibetan geology (e.g., Yin et al., 1999), but may have been extremely important during the Pliocene–Holocene interval (Tapponnier et al., 1986).

Perhaps the strongest evidence for such kinematics is the change in the pattern of Neohimalayan rift systems across the suture zone (Fig. 2): no major rift system of Tibet extends across the zone without disruption, many rifts are truncated at the suture, and the trends of the rifts are generally perpendicular to the Himalayan arc south of the suture zone, but are less so to the north. In the long 84°–89°E sector of the orogen, the pattern of rifts is consistent with dextral offsets on the order of several tens of kilometers, but, to the best of my knowledge, no recent attempts have been made to investigate the neotectonic evolution of the suture.

ESTIMATES OF POSTCOLLISIONAL SHORTENING IN THE HIMALAYA

The paleomagnetic record suggests that India has continued to move northward with respect to Tibet, with a significant counterclockwise rotation, since the early stages of collision (Patriat and Achache, 1984; Klootwijk et al., 1992). During that time, roughly 1800 km of shortening has occurred between the Indian subcontinent and stable Eurasia in the western part of the Himalayan-Tibetan orogenic system and as much as 2750 km of shortening has occurred in the east (Dewey et al., 1989). This contraction must have been accommodated by shortening in the Himalaya, shortening in Tibet, and the removal of lithosphere from the system by erosion, by continental subduction, by eastward extrusion of Tibetan lithosphere (Tapponnier et al., 1982), and/or by the foundering of Tibetan lower lithosphere (England and Houseman, 1988). The relative importance of these processes is hotly debated among students of the tectonics of the Himalaya and Tibet, largely because each hypothesis is difficult to test in a quantitatively meaningful way.

Estimating the amount of crustal shortening in the Himalaya provides an illustration of the problem. Early estimates of the total shortening across the range varied from a few hundred to as much as 1000 km (Seeber et al., 1981; Lyon-Caen and Molnar, 1983; Molnar, 1984). Balanced cross sections for the thin-skinned fold-and-thrust belt of the Lesser Himalayan and Subhimalayan zones of Pakistan were used by Coward et al. (1988) to suggest a minimum of ~470 km of shortening. and they argued that an additional 150 km of shortening may have been accommodated by structures between the Lesser Himalayan zone and the Main Mantle thrust zone. Their estimate of total shortening (~620 km) is highly dependent on the amount of slip that is presumed to have occurred on the basal thrust of the Main Central, Main Boundary, and Main Frontal thrust systems. Because no tectonostratigraphic units can be matched across any of these structural systems, such presumptions are highly speculative. For the Kumaun region of India, Srivastava and Mitra (1994) estimated the amount of shortening between the Main Central thrust system and the Himalayan front as 414–550 km. Although this range compares well with the Coward et al. (1988) value, it includes a poorly defined estimate on the amount of slip along the basal thrust of the Almora allochthon; as was the case for the Main Central thrust system, Main Boundary thrust system, and Main Frontal thrust system, a robust estimate of the displacement on this structure is impossible. In western Nepal, DeCelles et al. (1998b) calculated ~228 km of shortening across the Lesser Himalayan and Subhimalayan zones, but were forced to rely on the interpretations of Srivastava and Mitra (1994) for Kumaun in order to estimate 193–260 km of shortening on the basal thrust(s) of the Lesser Himalayan crystalline allochthons and the Main Central thrust system. For eastern Nepal, Schelling (1992) estimated only 70 km of shortening for the Lesser Himalayan and Subhimalayan zones, but inferred 245–280 km of shortening on the Main Central thrust system.

No attempts have been made to restore the internal strain of the Greater Himalayan zone, but the well-defined stratigraphy of the Tibetan zone in Ladakh-Zanskar and southern Xizang invites attempts to calculate the total shortening between the South Tibetan fault system and the Indus-Tsangpo suture zone. In southern Xizang, features like the Gyirong-Kangmar thrust are problematic because, again, the footwall and hanging-wall stratigraphies cannot be matched across the fault. Nevertheless, Ratschbacher et al. (1994) estimated shortening of ~258 km across the Tibetan zone. An inability to adequately account for internal strain of the Tibetan zone in the Ladakh-Zanskar region jeopardizes the reliability of quantitative estimates of shortening based on thrust reconstructions, but Searle et al. (1997a) suggested a minimum of 150–170 km. Combining all estimates for shortening on the northern and southern flanks of the Himalaya, it is possible to calculate a 465–808 km range of shortening amounts for the region between the foreland and the Indus-Tsangpo suture zone. However, these estimates largely ignore contractional deformation within the Greater Himalayan and Indus-Tsangpo suture zones, and extensional deformation throughout the orogen. These two shortcomings have opposing implications for total shortening calculations, such that 465–808 km may be either a gross underestimate or a significant overestimate. Given geophysical evidence that the Himalayan Sole thrust extends at least as far north as the surface trace of the Indus-Tsangpo suture zone (ITSZ, Fig. 3) (Hauck et al., 1998) and perhaps substantially farther (Makovsky et al., 1999), as well as petrologic evidence for the subduction of the north Indian margin to mantle depths (O'Brien et al., 1999), it seems reasonable to speculate that as much as one-third to one-half of the total convergence between India and Eurasia over the past 50 m.y. was accommodated by shortening in the Himalaya. However, the data necessary to support that speculation are not now—and may never be—available.

MESOZOIC–TERTIARY METAMORPHIC HISTORY

Metamorphism in the Himalaya also can be partitioned into Protohimalayan, Eohimalayan, and Neohimalayan phases. Protohimalayan metamorphism produced scattered examples of blueschist-facies metamorphism in the Indus-Tsangpo suture zone of the western Himalaya (Shams, 1980; Honegger et al., 1982). Only samples from Ladakh have been studied in some detail; estimates of their metamorphic conditions range from 9 to 11 kbar at 350–420 °C (Honegger et al., 1989). Few isotopic age determinations are available for blueschist-facies metamorphism in the Himalaya. Limited K-Ar and $^{40}Ar/^{39}Ar$ data provide cooling dates ranging from 67 to 100 Ma (Desio and Shams, 1980; Maluski and Schaeffer, 1982; Maluski and Matte, 1984; Honegger et al., 1989), and Rb-Sr amphibole-phengite mineral isochrons yield cooling dates of 77–79 Ma for high-pressure rocks along the Main Mantle thrust of Pakistan (Anczkiewicz et al., 1998b).

Protohimalayan high-pressure (>10 kbar) metamorphism also is manifested in the structurally deepest rocks of the Kohistan-Ladakh arc, immediately above the Main Mantle thrust, as garnet granulites and retrogressed eclogites (Jan and Howie, 1981; Le Fort et al., 1997; Rolfo et al., 1997). Although this metamorphism is generally regarded as having taken place in an arc environment prior to India-Eurasia collision, there is no geochronologic confirmation. North of the Shyok suture zone, in the Karakoram terrane, amphibolite-facies metamorphism of the country rocks of the Karakoram batholith has been dated by U-Pb monazite geochronology at 64 Ma and may represent the impact of Kohistan-Karakoram terrane collision (Fraser et al., 1999).

The oldest, well-documented Eohimalayan metamorphic assemblages are in the high-pressure and ultrahigh-pressure eclogites of the western Himalaya. In the upper Kaghan Valley of Pakistan, the eclogite-facies assemblages are developed in dismembered mafic dikes and sills that intrude orthogneisses and paragneisses of presumed Greater Himalayan zone affinity in the immediate footwall of the Main Mantle thrust (Pognante and Spencer, 1991). The recent identification of coesite inclusions in omphacite from one sample of the upper Kaghan Valley eclogites by O'Brien et al. (1999) was the first documentation of ultrahigh-pressure metamorphism in the Himalaya (~680 °C; 27 kbar). Sm-Nd and U-Pb geochronology suggests a 44–49 Ma age for eclogite-facies metamorphism in Pakistan (Tonarini et al., 1993; Spencer and Gebauer, 1996). In the Tso Morari dome, lithologically similar dikes and sills contain eclogite-facies assemblages indicative of pressure-temperature (*P-T*) conditions of ~20 kbar and ~580 °C and show evidence of retrogression in the glaucophane stability field at ~11 kbar and ~570 °C (De Sigoyer et al., 1997). A similar Eohimalayan *P-T* evolution was deduced for associated metasedimentary rocks (Guillot et al., 1997). Lu-Hf and Sm-Nd mineral–whole-rock isochrons for the Tso Morari eclogites suggest that peak metamorphism occurred at ca. 55 Ma (De Sigoyer et al., 1998).

In addition to evidence for an amphibolite-facies metamorphic overprint in the core gneisses of the Tso Morari dome (De Sigoyer et al., 1997; Guillot et al., 1997), there is abundant evidence for an early, high-pressure, amphibolite- to granulite-facies metamorphic event in the gneisses of the Greater Himalayan zone from at least as far east as the Arun Valley of Nepal (Fig. 2) to at least as far west as the Nanga Parbat syntaxis (Hodges et al., 1988a; Pêcher, 1989; Guillot et al., 1999). Because a later Neohimalayan overprint has obliterated all but vestiges of the Eohimalayan assemblages in central and eastern Nepal, the most thorough studies of the Eohimalayan event have been done in northwest India and Pakistan. Granulite-facies assemblages (~9–13 kbar, ~650–700 °C) are best developed in northern Pakistan (Treloar et al., 1989a; DiPietro and Lawrence, 1991; Pognante et al., 1993). In the Himalaya of northwest India, Eohimalayan metamorphic temperatures and pressures were generally lower (~500–650 °C, ~6–11 kbar—Hodges and Silverberg, 1988; Pognante et al., 1990; Searle et al., 1992; Metcalfe, 1993; Walker, 1999; Walker et al., 1999; Wyss et al., 1999). However, samples from the Greater Himalayan sequence around the Kishtwar window of Zanskar suggest temperatures as high as 740 °C (Stephenson et al., 2000). U-Pb and Sm-Nd data imply an early Oligocene age for Eohimalayan amphibolite- to granulite-facies metamorphism in the Zanskar Greater Himalayan zone (Vance and Harris, 1999; Walker et al., 1999).

Estimates of the *P-T* conditions for Eohimalayan metamorphism are less reliable in the central and eastern Himalaya. In general, pressures for amphibolite-facies assemblages in Greater Himalayan zone rocks may have ranged from 4 to 10 kbar, and temperatures may have ranged from 475 to 700 °C, depending on structural level (Brunel and Kienast, 1986; Hodges et al., 1988b, 1993, 1994; Pêcher, 1989; Pognante and Benna, 1993; Vannay and Hodges, 1996). In the Kharta region of Xizang, just east of Makalu (Fig. 2), Lombardo et al. (1999) have found evidence for the existence of Eohimalayan eclogite-facies assemblages, now completely reequilibrated as Neohimalayan granulite-facies assemblages, just above the upper bounding fault of the Main Central thrust system. In central Nepal, gneisses of the Greater Himalayan sequence and Neohimalayan leucogranites frequently contain two populations of Tertiary monazites, one of early Oligocene age and one of Miocene age (e.g., Hodges et al., 1996; Coleman, 1998), and it is probable that the older dates represent Eohimalayan metamorphism. From the Dinggyê area of southern Xizang (Fig. 2), Hodges et al. (1994) obtained late Oligocene $^{40}Ar/^{39}Ar$ hornblende cooling dates for an unusually pristine suite of Eohimalayan metamorphic rocks collected within the uppermost Greater Himalayan sequence.

The Greater Himalayan zone, structurally higher parts of the Lesser Himalayan zone, structurally lower parts of the Transhimalayan zone, and the metamorphic cores of the North Himalayan gneiss domes all contain a record of Neohimalayan metamorphism. The Neohimalayan thermal histories of the gneiss domes are not well understood in general, although the Kangmar complex in Xizang is a notable exception. Building on petrographic evidence reported by Chen et al. (1990) for polymetamorphism at Kangmar, Guillot et al. (1998) deduced a three-phase metamorphic evolution for highest-grade metasedimentary rocks in the dome: (1) an early, low-temperature, amphibolite-facies event (8.3–8.8 kbar, 500–550 °C); (2) a higher-temperature, amphibolite-facies event (7.2–7.4 kbar, 600–650 °C); and (3) a late retrograde event (3.5–4.5 kbar, ~500 °C). A thermobarometric study across the entire range of metamorphic grades preserved at Kangmar (kyanite to garnet) yielded slightly different *P-T* estimates for high-temperature metamorphism of the kyanite-grade rocks (~8.5 kbar; ~625 °C) and estimates of ~3.75 kbar and ~450 °C for the garnet-grade rocks (Lee et al., 1999). On the basis of $^{40}Ar/^{39}Ar$ data reported by Maluski et al. (1988), Chen et al. (1990) and Guillot et al. (1998) inferred an early Miocene age for peak metamorphism at Kangmar. However, Gans et al. (1998) have shown the $^{40}Ar/^{39}Ar$ systematics at Kangmar to be very complex, and more geochronologic data obtained through the use of more robust systems (e.g., U-Pb) are required to understand the thermal evolution of the Kangmar complex and other North Himalayan gneiss domes.

Both the Greater Himalayan and Lesser Himalayan sequences display inverted Neohimalayan metamorphic gradients (Heim and Gansser, 1939; Le Fort, 1975; Pêcher and Le Fort, 1986), but the two are not of the same age. In the Greater Himalayan zone, near the roof thrust of the Main Central thrust zone, pelitic rocks typically contain kyanite-grade assemblages. At progressively higher structural levels, characteristic subassemblages change first to Sil + Ms, then to Sil + Kfs, and finally to Sil + Kfs ± Crd (Pêcher, 1989). In most sections, the sillimanite isograd roughly corresponds to the first appearance of anatectic leucosomes in rocks of pelitic composition, and the proportion of leucogranitic melt increases upsection. However, kyanite-bearing leucogranites have been identified within the kyanite zone, near the base of Formation I, in some parts of the Annapurna Range of north-central Nepal (Hodges et al., 1996). In general, peak metamorphic temperatures range from 500–550 °C near the Main Central thrust system to >650–700 °C in the upper half of the Greater Himalayan sequence (Brunel and Kienast, 1986; Le Fort et al., 1987b; Hodges and Silverberg, 1988; Hodges et al., 1988c; Hubbard, 1989; Kündig, 1989; Mohan et al., 1989; Pognante and Lombardo, 1989; Staübli, 1989; Pognante et al., 1990; Swapp and Hollister, 1991; Inger and Harris, 1992; Searle et al., 1992; Spring and Crespo-Blanc, 1992; Hodges et al., 1993; Lombardo et al., 1993; Metcalfe, 1993; Pognante and Benna, 1993; Macfarlane, 1995; Treloar, 1995; Winslow et al., 1995; Vannay and Hodges, 1996; Davidson et al., 1997; Lombardo et al., 1999; Vannay and Grasemann, 1998; Manickavasagam et al., 1999; Walker, 1999; Walker et al., 1999; Wyss et al., 1999). The temporal association of amphibolite- to granulite-facies metamorphism with anatexis provides a straightforward way to date the inverted metamorphic gradient in the Greater Himalayan sequence. Reliable ages for the leucogranitic melts range from 23 to 12 Ma in the main outcrop belt of the Greater Himalayan zone, and it seems likely that the Neohimalayan metamorphic event was similarly long lived. In the syntaxial regions on either end of the Himalayan orogen, the Neohimalayan event may extend to early Miocene or even Pleistocene (Smith et al., 1992; Zeitler et al., 1993; Wheeler et al., 1995; Winslow et al., 1995, 1996; Burg et al., 1998; Schneider et al., 1999a, 1999b).

Metamorphic studies of the Main Central thrust zone are complicated because late Neohimalayan slip on Main Central thrust system structures have shuffled lower and middle Miocene metamorphic rocks of the Greater Himalayan zone with Lesser Himalayan rocks having a different metamorphic history (e.g., Brunel and Kienast, 1986; Hubbard and Harrison, 1989; Macfarlane et al., 1992). Lesser Himalayan sequence metapelites contain the characteristic subassemblages Grt + Bt + Ms ± St near the roof fault of the Main Central thrust system schuppen zone (Pêcher and Le Fort, 1986; Macfarlane, 1995; Vannay and Hodges, 1996); kyanite- and sillimanite-bearing assemblages reported from this structural level (Hubbard, 1989; Vannay and Grasemann, 1998) probably represent structurally disrupted Greater Himalayan rocks rather than Lesser Himalayan units. At deeper structural levels, Lesser Himalayan metamorphic assemblages range progressively through garnet, biotite, and chlorite grades (Pêcher and Le Fort, 1986). Recent Th-Pb ion-microprobe dates for metamorphic monazites in unambiguous Lesser Himalayan sequence rocks demonstrate that the Lesser Himalayan inverted metamorphism is a late Miocene–Pliocene phenomenon (Harrison et al., 1997b; Catlos et al., 1999).

Numerous models have been proposed to explain the apparent inversion of metamorphic isograds and metamorphic field gradients in the Greater Himalaya. Le Fort (1975) developed a very influential hypothesis that emplacement of the Greater Himalayan allochthon along the Main Central thrust system led to inverted metamorphism of the Lesser Himalayan footwall and "refrigeration" of the hanging wall. Subsequent thermal modeling of time-dependent thrust-sheet emplacement cast doubts on Le Fort's model (Shi and Wang, 1987; Ruppel and Hodges, 1994). Jaupart and Provost (1985) suggested that high temperatures near the top of the Greater Himalayan sequence might be related to a difference in thermal conductivity of Greater Himalayan and Tibetan units, leading to the focusing of heat near the contact. However, postmetamorphic displacements have been large on brittle detachments of the South Tibetan fault system, leaving some question as to the thermal properties of the overlying rocks at the time of high-grade metamorphism of the footwall. Models invoking shear heating along the Main Central thrust system also have been popular from time to time (Bird et al., 1975; England et al., 1992), but they provide no explanation for the presence of the highest-temperature hanging-wall assemblages at structural levels substantially above the Main Central thrust system.

Another class of models appeals to recumbent folding, thrust imbrication, or distributed shearing of a preexisting, right-way-up metamorphic sequence in the Greater Himalayan zone (Frank et al., 1973; Searle and Rex, 1989; Jain and Manickavasagam, 1993; Grujic et al., 1996; Hubbard, 1996; Grasemann and Vannay, 1999). Although the recumbent folding model may be appropriate for western Zanskar (Searle et al., 1992), the closures of postulated folds in isograd patterns have yet to be mapped in Zanskar, and it has proven similarly difficult to defined folded isograds elsewhere in the Himalaya. Most isograds in the Greater Himalayan sequence do not correspond to mapped structural discontinuities, as would be expected if discrete thrust imbrication of preexisting isograds was responsible for inverted metamorphism. At present, the most plausible hypotheses seem to be those that involve distributed shearing of the Greater Himalayan sequence. In a series of papers, Huerta and coworkers (Huerta et al., 1996; Huerta et al., 1998; Huerta et al., 1999) showed that the accretion of material from the footwall of a major intracrustal shear zone like the Main Central thrust system, coeval with the erosion of hanging-wall material from high structural levels, could produce inverted thermal structures in the hanging wall with a temperature maximum spatially removed from the most recently active plane of shearing. Such structures are most pronounced when the level of radiogenic heat production is high in the accreted materials, which is clearly the case for the Greater Himalayan sequence (Jaupart and Provost, 1985; Pinet and Jaupart, 1987). If such a model is applicable to the Greater Himalayan inversion, it would require that the Greater Himalayan zone—at least the pelitic gneisses of Formation I—is the intracontinental equivalent of an accretionary complex, deforming continuously over millions of years. The documentation of kinematically complex, general shear flow of the Greater Himalayan zone in areas as far removed as Bhutan (Grujic et al., 1996) and northwestern India (Grasemann et al., 1999) support such a tectonothermal history, but the remarkable lateral persistence of the Greater Himalayan zone tectonic stratigraphy would be unexpected in a rock package that is presumably so highly deformed.

Without modification, the Huerta et al. models do not provide a convenient explanation for the late Miocene–Pliocene inverted metamorphism in the uppermost Lesser Himalaya. It seems probable that this phenomenon is, in some way, related to late, out-of-sequence faulting on the Main Central thrust system (Harrison et al., 1997b). Recently, Harrison et al. (1997a) proposed a thermal model for inverted metamorphism in the Main Central thrust system footwall, leucogranitic plutonism in the Greater Himalayan zone, and leucogranitic plutonism in the North Himalayan gneiss domes. Requiring that the currently exposed Greater Himalayan zone resided at mid-crustal levels in the hanging wall of a low-angle thrust decollement (analogous to the Himalayan Sole thrust) throughout early and middle Miocene time, the model attributes the partial melting that led to the emplacement of leucogranites in both the Greater Himalayan zone and the cores of the North Himalayan gneiss domes to shear heating along the decollement. Final emplacement of the exposed parts of the Greater Himalayan sequence, as well as burial metamorphism of the Lesser Himalayan footwall, is related to break-back thrusting on the Main Central thrust system. Although this model admittedly incorporates many ad hoc assumptions regarding the pre-Neohimalayan thermal structure of the orogen, the geometry of the Main Central and Main Boundary thrust systems, and evolution of this geometry with time (Harrison et al., 1997b), it nevertheless reproduces many of the geologic characteristics of the Himalayan metamorphic hinterland. Unfortunately, it is inconsistent with others. In particular, it does not predict the observed inverted metamorphism of the Greater Himalaya (structurally higher rocks always remain at lower temperatures than structurally lower rocks in the model), it does not explain widespread evidence for anatexis in the middle exposed parts of the Greater Himalayan sequence (all anatexis is along the basal decollement in the model), and it does not provide a ready explanation for the long duration of melting and Neohimalayan metamorphism in the Greater Himalayan sequence, both of which overlap in time with metamorphism and leucogranitic plutonism in the North Himalayan gneiss domes.

One conundrum that arises as a consequence of the documentation of late Neohimalayan metamorphism of the Lesser Himalayan sequence is generally underappreciated and deserves special mention. Abundant $^{40}Ar/^{39}Ar$ data for the Greater Himalayan sequence suggest that these rocks cooled from early Miocene peak conditions to temperatures of less than 350 °C several million years prior to late Miocene–Pliocene amphibolite-facies metamorphism of the subjacent Lesser Himalayan rocks (Hubbard and Harrison, 1989; Maluski et al., 1988; Vannay and Hodges, 1996). This timing implies that the roof fault of the Main Central thrust system juxtaposes hanging-wall rocks that were at a high structural level in the late Miocene and Pliocene with footwall rocks that were at deeper structural levels and higher temperatures at the same time. The observed structural relationship is thus more consistent with late-stage normal faulting than with out-of-sequence thrust faulting. One possible scenario that deserves further scrutiny is that the metamorphosed Lesser Himalayan rocks within the Main Central thrust system shear zone originated in a more northerly position in late Miocene–Pliocene time, and that they were exhumed during normal-sense reactivation of the Main Central thrust system roof fault in late Pliocene–Holocene time.

A NEO-IMPRESSIONISTIC PERSPECTIVE

Previous Impressionistic sections of this paper represent an approach to the study of mountain ranges that emphasizes lithologic and structural "taxonomy" presented from a historical perspective. A complimentary Neo-Impressionistic approach to understanding orogens involves identifying the processes that are responsible for orogenesis over a specific time interval in its evolution and asking how they work together to define the behavior of the orogen. I emphasize the Neohimalayan interval here, but similar exercises could be carried out for the Eohimalayan or Protohimalayan phases.

Any successful attempt to understand the behavior of the Himalayan orogen in Neohimalayan time must consider the broad synchroneity of seven processes. These are (1) north-south shortening on the Main Central thrust system, Main Boundary thrust system, Main Frontal thrust system, and lesser thrust and fold nappes; (2) north-south extension at various structural levels in the orogenic system, but especially along the South Tibetan fault system and within the Tibetan zone; (3) east-west extension restricted to the Tibetan, Indus-Tsangpo suture, and Transhimalayan zones; (4) high-grade metamorphism and anatexis in the Greater Himalayan zone that began in earliest Miocene time (e.g., Harrison et al., 1996; Hodges et al., 1996) and lasted to at least late-middle Miocene time (ca. 12 Ma; Edwards and Harrison, 1997); (5) late Miocene–Pliocene amphibolite-facies metamorphism within the Main Central thrust system schuppen zone and in the immediately subjacent rocks of the Lesser Himalayan zone (Harrison et al., 1997b; Catlos et al., 1999); (6) rapid erosion of the south flank of the Himalaya and rapid transport of the detritus to distant depocenters like the Bengal Fan (Copeland and Harrison, 1990); and (7) large-scale melting of the middle crust of Tibet, as indicated by INDEPTH seismic reflection data (Nelson et al., 1996). Models of accretionary-wedge development and their extrapolated equivalents for orogenic belts (Platt, 1986; Dahlen, 1988, 1990; Harrison et al., 1998) provide a valuable mechanical framework for some processes, but they fail to predict some aspects of Neohimalayan orogenesis. Without amplification or modification, they do not provide a satisfactory explanation for the development and persistence of the South Tibetan fault system as a major decoupling horizon between the middle and upper crust, they do not explain the longevity of high-grade metamorphism and anatexis at intransient positions in the wedge, and—perhaps most significantly—they do not predict that major fault systems, once established, should be active episodically over many millions of years. For example, the Main Central thrust system was established at least as early as 23–20 Ma (Hubbard and Harrison, 1989; Hodges et al., 1996), but may have been active in late Miocene–Pliocene time (Harrison et al., 1997b), and shows geomorphic evidence for Holocene activity (Hodges, K. V., Hurtado, J. M., and Whipple, K. X., unpublished data). Because an absence of evidence for faulting does not preclude its having occurred, we do not know if displacements on the Main Central thrust system have been quasi-continuous or episodic with relatively long periods of quiescence. We can say with confidence, however, that this one structure has been an important feature of Neohimalayan orogenesis over a time period corresponding to nearly 40% of the entire evolutionary history of the Himalaya. The same appears to be true of the South Tibetan fault system: although it began moving prior to 22 Ma (Hodges et al., 1996), middle Miocene displacement has been documented in several areas (Edwards and Harrison, 1997; Hodges et al., 1998), and Quaternary displacement can be documented in at least one area (J. M. Hurtado, K. V. Hodges, and K. X. Whipple, unpublished data).

Why the Main Central thrust system and South Tibetan fault system should remain active for so long, rather than be abandoned permanently in favor of slip on structures farther to the south in the accretionary wedge, as conventional fold-and-thrust-belt theory would predict, is a fascinating question; its answer might lead to a deeper understanding of Himalayan orogenesis. As outlined next, it is possible to construct a working model for the Neohimalayan behavior of the orogen, consistent with all documented processes, by modifying classical accretionary-wedge models to account for the geodynamic effect of the Tibetan Plateau.

It has been argued for many years that overthickened, isostatically compensated continental crust has a tendency to flow laterally under its own weight (Artyushkov, 1973; England, 1982; Fleitout and Froidevaux, 1982; Molnar and Lyon-Caen, 1988; Bird, 1991). Given the great crustal thickness of Tibet and geologic evidence for Neohimalayan extension of the plateau, it has become popular to relate the extension to gravitational collapse (England and Houseman, 1988, 1989; Harrison et al., 1992; Molnar et al., 1993). Most of these models carry with them the assumption of convective delamination of the lower lithosphere beneath Tibet in late Miocene time, a physically attractive but geologically untestable hypothesis. However, numerical experiments also suggest that channelized flow of the middle and lower continental crust is a viable mechanism for the dissipation of potential energy stored in overthickened crust, even in the absence of lithospheric delamination (Bird, 1991; Royden et al., 1997). Seismic evidence for partially molten lower crust in southern Tibet (Nelson et al., 1996) lends credence to the notion that the lower crust of Tibet is capable of lateral flow.

Most contributions about the spreading of Tibet have focused on the dynamics of the northern and eastern margins of the plateau and on investigating the significance of “lateral extrusion” of the Tibetan lithosphere (Tapponnier et al., 1982; Peltzer and Tapponnier, 1988; Houseman and England, 1993a; Royden et al., 1997); however, the potential for southward extrusion is generally underappreciated. After all, the southern flank of the Himalaya is an exposure of the edge of the Tibetan Plateau, and its examination is one of the most direct ways to deduce the behavior of the Tibetan middle and lower crust. When we consider the tectonic and erosional processes active along this margin in the recent past, it is not difficult to imagine that the currently exposed Greater Himalayan zone is the leading edge of a channel of Tibetan middle and lower crust, bounded above and below by the South Tibetan and Main Central fault systems, that is being expelled southward by a pressure gradient between the plateau and India (Hodges and Hurtado, 1998; Wu et al., 1998).

What is most remarkable about such a hypothesis, if it is correct, is that southward extrusion appears to have persisted over a period of at least 20 m.y., since the early stages of movement on the Main Central thrust system and the South Tibetan fault system (Burchfiel and Royden, 1985; Burchfiel et al., 1992). Such stability almost certainly implies the existence of a dynamical steady state defined by a rough balance among processes responsible for energy accumulation (e.g., crustal thickening related to India-Eurasia convergence) and energy dissipation (e.g., southward extrusion of the middle crust and rapid erosion along the Himalayan front). One of these classes of processes may have been more important than the other from time to time. For example, periods of rapid energy accumulation may have corresponded to periods of quiescence on the South Tibetan fault system, to renewed slip on the Main Central thrust system (leading to "anomalously" young metamorphism of the footwall; Harrison et al., 1997b), and to widespread north-south shortening between the Indus-Tsangpo suture zone and the Main Frontal thrust system. Periods of rapid energy dissipation may have been marked by accelerated extension on the South Tibetan fault system, by extensional deformation over a broader region, or simply by more rapid erosion along the Himalayan front. Transitions between such periods provide the best available explanation for the observation that extension and contraction alternate at similar structural levels in the Himalaya over timescales of no more than a few million years (Hodges et al., 1996).

Whether such behavior is characteristic of evolving continent-continent collisional orogens remains unclear. The Himalaya and the Tibetan Plateau form an unusual system. Their evolutionary pathways have been intertwined since the development of sufficiently thick and weak crust beneath the plateau to accommodate gravitational spreading. It may be that traditional evolutionary models of collisional orogenesis are appropriate for the Himalaya prior to development of the plateau, but they are less than adequate to explain the evolution of the system after plateau development. As a consequence, we would be well advised to exercise some restraint when pointing to the Himalaya as the “definitive” example of a continent-continent collisional orogen; although narrow orogenic belts are common in the geologic record, analogues for the Tibetan Plateau are not. In the end, the Himalayan-Tibetan orogenic system may be a special case rather than an archetype, and its value may be limited as a guide to interpreting the structural evolution of other collisional orogens that are less well developed, less well exposed, or more deeply eroded.

On the other hand, the Himalaya and Tibet may be the best available laboratory for exploring how feedback relationships among structural, thermal, and erosional processes dictate the behavior of a collisional system. For example, the orogen is young enough that modern erosional patterns can be extrapolated backward in time over a significant portion of the orogenic interval, yet old enough to display surface exposures of molasse basins with a rich sedimentary record of its earlier history. Moreover, the modern orography of the Himalaya is probably similar to that which has characterized the range throughout much of Neogene–Quaternary time, yet the erosional level is deep enough in some places to expose the broad tracts of the middle crust in the Greater Himalayan zone, as well as scattered remnants of the lower crust and upper mantle in Kohistan, the upper Kaghan Valley, and the Tso Morari dome.

Fully exploiting the opportunity provided by the Himalaya and Tibet for a deeper understanding of collisional orogenesis will require carefully designed research programs. Like most interesting dynamical systems, orogens are not characterized by regular behavior. We cannot expect to be able to study one small segment of the Himalaya in great detail and then develop tectonothermal models that can be extrapolated to the scale of the entire orogenic system. At the same time, reconnaissance studies of large tracts of the system provide such a coarse data set that it is practically valueless for testing and refining modern, sophisticated models of Himalayan-Tibetan orogenesis. Given what we know at present about the behavior of the system—and given the sensitivity of our models to parameters such as erosion rates, bedrock cooling rates, and the distribution and displacement histories of major fault systems—it seems likely that major advances in our understanding of Himalayan-Tibetan orogenic processes will require the development of robust data sets pertinent to timescales of less than 1 m.y. and length scales of no more than a few tens of kilometers. One gauge of the magnitude of this challenge is that less than 1% of the area shown in Figures 2 and 3 has been mapped geologically at a scale of 1:50 000 or larger. Besides sufficient detail, the best-designed future studies will integrate structural geology, geochronology, petrology, geomorphology, and geodesy. For example, a detailed study of the *P-T* evolution of a tract of gneisses within the Greater Himalayan zone is wasted effort without the coordination of detailed mapping and geochronology to provide both structural and temporal contexts.

As the twenty-first century begins, we are approaching the end of an era of geological exploration in the Himalaya and southern Tibet. It is no longer satisfying to paint the history of this remarkable orogen in broad, ill-defined strokes on the basis of severely limited data sets from a handful of study areas. We must have the patience to collect detailed, comprehensive data sets for specific regions and the persistence to repeat this process until enough regions have been characterized to define the evolution of the orogenic system as a whole. If there is to be a unified theory of mountain building, I believe it will emerge from careful analysis of the spatial and temporal patterns of deformational, thermal, and erosional processes at a variety of scales. Like the great Neo-Impressionist masterpieces, orogens must be viewed from multiple perspectives to be fully appreciated.

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